Mechanical Weathering in Cold Regions with Special Emphasis on the Antarctic Environment and the Freeze-Thaw Mechanism in Particular.

by

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DECLARATION

The work presented in this D.Sc. thesis constitutes original work that, outside of where specified as such, has not been submitted elsewhere for any degree or diploma at any University or College in any country.

The introductions to each chapter comprise original work by the author except where use is made of references (which are duly cited).

The body of the work comprises papers and reports that have been published in a wide range of journals, some in conjunction with co-authors. In all cases the references to those papers identify clearly the authors and place of publication.

Where use has been made of the work of others it is duly acknowledged in the text.

Kevin Hall

Kevin J. Hall
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Abstract

"Science must begin with myths, and with the criticism of myths"
Sir Karl Popper

Consideration of almost any geomorphology textbook will show the fundamental argument that in cold environments mechanical weathering processes, usually freeze-thaw, will predominate and that chemical weathering will be temperature-limited, often to the point of non-occurrence. These basic concepts have underpinned geomorphology for over a century and are the basis for the development of many landforms in periglacial regions. With the introduction of data loggers so field data became more readily available but, sadly, those data were not of a quality to other than justify the existent assumptions and thus did little more than reinforce, rather than test, the nature of our understanding of cold region weathering. Factors such as rock properties were dealt with to a limited extent but rock moisture was all but ignored, despite its centrality to most weathering processes. Here the results of field studies into weathering in cold regions, coupled with laboratory experiments based on the field data, are presented. An attempt is made to overcome the shortcomings of earlier studies. Temperature, moisture and rock properties have all been considered. Processes were not assumed but rather the data were used to evaluate what processes were operative. The results, both in terms of weathering process understanding *per se* and of its application to landform development, significantly challenge our long-held perceptions.

Information is presented that shows that it is not temperature, but rather water, that is the limiting factor in cold region weathering. Indeed, in the absence of water, many cold environments have attributes akin to a hot desert. The relevance of this is that weathering processes other than freeze-thaw may play a significant role and that in the presence of water chemical weathering can play a far greater role than hitherto thought. Overall, the whole concept of zonality with respect to weathering is questioned. Finally, the attributes of weathering are put within the context of landform development and questions raised regarding the origin of some forms and of their palaeoenvironmental significance. Attributes of periglacial, glacial and zoogeomorphic processes and landforms in present and past cold environments are also presented.

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"The real voyage of discovery consists not in seeking new landscapes, but in having new eyes."

Marcel Proust

"What we see depends on mainly what we look for."

John Lubbock

"Its not what you look at that matters, its what you see."

Henry David Thoreau

Chapter 1

Introduction

"First get your facts, then you can distort them at your leisure."

Mark Twain

The work presented in this thesis is primarily aimed at a consideration of mechanical weathering in cold regions, with a special emphasis on the role of freeze-thaw, and on the Antarctic region in particular. The questions raised are not new, others such as Grawe (1936) brought attention to the unquestioned assumption of the role and efficacy of freeze-thaw weathering in cold regions, whilst such as Warren (1914a) long ago discussed the meaning and validity of laboratory undertakings. Yet, as will be shown by multiple quotations presented below, the bulk of texts today still rely on freeze-thaw weathering as an explanation for landforms and sediments in cold regions. In an attempt to 'set the scene' a brief discussion will be presented regarding what is meant by the term "weathering" and this is followed by a more focussed consideration of freeze-thaw weathering in particular. As the whole of the thesis is in the context of 'cold regions' a discussion is presented as to exactly what this may, or may not, mean. In truth, once considered, it became apparent that it was very difficult indeed to find any universally acceptable way of defining a 'cold region'. Then, to create a foundation against which the presented papers can be viewed, extensive quotations from a wide range of literature are given. As noted above, it will be seen that the overwhelming consensus is the assumed role of freeze-thaw weathering in cold regions as the main cause of non-glacial landforms and sediments.

1.1 Weathering

Weathering includes all those processes which cause disintegration and alteration of rock in the upper part of the Earth's crust (Ollier, 1984). It is possible, however, to distinguish those that require movement - abrasion and erosion - from those that involve *in situ* chemical or physical changes. It should be noted that Chorley *et al.* (1984) state that the break up of rock *in situ* is impossible in so far as some movement must always take place. They assert that as weathering occurs within a gravity field there has to be a component of mass movement involved. Patently weathering must involve *some* movement; the springing apart of rock during dilatation, the movement of molecules during solution, the propagation of a crack due to salt or ice crystal growth all have movement implicit within them. However, weathering does occur *in place* without the requirement of a moving medium (e.g. ice, water, wind) to facilitate its operation. Yatsu (1988, p.2) concludes with a working definition very similar to that employed here: "Weathering is the alteration of rock or minerals in situ, at or near the surface of the earth under the conditions that prevail there."

Weathering produces changes at a local level, altering or breaking rock. The action of erosion and transport upon this modified material then produces actual landforms. The physical (or 'mechanical') weathering of rocks results in progressive, but probably non-linear (Colman, 1981), fragmentation without chemical alteration of the mineral components, whilst chemical weathering, on the other hand, causes decomposition of the rock, culminating in new mineral forms which have less free energy (Curtis, 1976). Biological (or 'biotic') weathering does not really constitute a separate entity, but is rather a subset of the other two groups and comprises *biologically - induced* mechanical or chemical weathering processes. These three principal processes (chemical, mechanical, and biologically-induced weathering) are inextricably linked together in a variety of ways.

Weathering is a major factor in both the natural and the anthropogenic landscape, and it exerts an influence in a number of ways. Whilst weathering does not itself produce landforms, it has an intimate association with a number of specific features. For instance, small depressions on horizontal or vertical surfaces of boulders and cliffs in many cold environments are variously referred to as "weathering hollows" (Juckes, 1969) or "weathering pits" (Watts, 1983a & b; Samuelsson and Werner, 1978; Fahey, 1986), with the term "cavernous weathering" being used as both a verb and a noun (Calkin and Cailleux, 1962; Mercer, 1963). For the formation of features such as these two stages are necessary - firstly weathering to breakdown and/or transform the material *in situ*, and secondly the removal of the debris to give the hollow. Recognition of this is important, for the mechanism of debris removal, even by as straightforward a process as gravity fall, is an intimate a part of the formation of that feature as is the weathering. However, in the case of weathering pits or hollows much emphasis has been placed upon the weathering processes (e.g. Mustoe, 1982) but far less on the mechanism of debris removal.

A further result of transporting weathering products is that they must be deposited elsewhere, which may result in the formation of landforms *composed* of weathered material. For instance, Gordon and Birnie (1986), from a study on South Georgia, show how talus and talus-related forms, gelifluction features, rock glaciers, and supraglacial debris supply, together with the ensuing morainic landforms, are all associated with the provision of material due to weathering of the available bedrock. It is shown that the rate of debris supply, which helps control the degree of landform development, is related to the type and amount of weathering that has taken place. Transport is required for, and is an intimate part of, landform creation *but* the origin of the formative material is weathering, without which none of the listed features could have been produced.

1.2 Freeze-thaw Weathering

Here is considered weathering that takes place due to the freezing and thawing of water. Water is central to the process and in its absence, although temperatures may be conducive, this form of weathering cannot take place. Some confusion often arises as a result of, so-called, freeze-thaw cycles being monitored with respect to *air* temperatures and these being adjudged representative of what is taking place in the rock. In fact, thawing of saline solutions inside of rocks will occur at sub-zero temperatures, the exact value of which is a function of the salt and the solution molarity. Equally, the freezing point will also be depressed due to the presence of salts such that temperatures will need to be below 0°C before freezing will occur. Pore size also exerts an influence on freezing temperature, with lower temperatures required for freezing to occur in smaller pores. In addition, as has already been mentioned, air temperatures may bear little or no relationship to rock temperatures due to such factors as the influence of incoming solar radiation. Thus, overall, in order to fully justify the occurrence of freeze-thaw weathering, knowledge of the presence of water within the rock is required as too is some manner of deducing whether freezing and thawing of that water actually took place.

The terms "microgelivation" (small-scale frost weathering) and "macrogelivation" (large-scale frost wedging) are also used by some authors following their initial usage by Tricart (1956). The former operates independently of geological structure and results in the breakdown of sound rock into particles that range in size from silt to fine gravel. The latter exploits pre-existing structures such as joints and bedding planes and results in the production of clastic debris of variable size. The actual manner whereby "freeze-thaw" causes rock breakdown is still far from clear, but there are a number of major theories each of which may operate at some time or in some rock with the same overall effect. The main ideas have been summarised by McGreevy (1981) and Ugolini (1986). The actual application of any one of the available mechanisms is largely constrained by our knowledge of the controlling factors, e.g. rock temperature, rate of fall of temperature, moisture content, solute chemistry, moisture distribution within the rock, pore size, tensile strength of the rock, and the synergistic operation of other processes. Later it will be seen that it is our lack of these fundamental data which is a stumbling block to our understanding of the operation of freeze-thaw in Nature; the discussions regarding moisture content and rock temperatures by McGreevy and Whalley (1985 and 1982 respectively) outline some of these basic problems.

The recent hypotheses of Hallet (1983), Walder and Hallet (1986) and Tharp (1987) regarding the mechanism of frost wedging, as with the earlier idea of Powers (1945), rely on the effects of *unfrozen water.* However, as the process is 'driven' by the formation of ice during the freezing phase and the destructive forces are relieved during ice melt in the thaw phase, these processes are still considered as a function of "freeze-thaw". Various recent studies point out the inter-connectivity of freeze-thaw with other mechanical (and chemical) processes, namely salt weathering (e.g. Williams and Robinson, 1981; McGreevy, 1982) and wetting and drying (e.g. Mugridge and Young, 1983). Thus it may be appreciated that freeze-thaw is a complex process with many questions still unanswered, but that its central theme is that of the freezing and thawing of water within rock, the effects of which are, in some manner, to cause weakening and possible ultimate failure of that rock.

A special type of frost weathering is that of "frost bursting" (Michaud, *et al.*, 1989) whereby rocks are said to fail in an explosive manner due to strains imposed by freezing. Frost bursting (termed 'éclatement' by Lautridou, 1985) was mathematically justified by Bertouille (1972) but has been little noted outside of the study of Michaud *et al.* (1989). This mechanism requires that the rock be sufficiently strong such that the hydrostatic pressure developed in pores and cracks under freezing conditions allows strain energy to be stored until the release of this energy takes place in an explosive manner, thereby shattering the rock. A rapid rate of freezing is thought to seal the rock to produce a closed system which, as water cannot be extruded, can ultimately lead to a dramatic failure of the rock. This explosive release of energy is said to disperse the gelifracts so producing a 'frost burst feature'.

1.3 What is a 'Cold Environment'?

Intuitively it would appear easy to decide if a particular environment is ‘cold’. In the context of low temperature weathering, attention cannot be limited to polar and high alpine regions. Many temperate areas have winter temperatures low enough to have a marked effect on weathering processes and rates. Climatic classifications (see Table 6.2 of Oliver, 1973) such as those of Köppen (1923), Thornthwaite (1948), Miller (1951) or Strahler (1969) are of limited use in determining the distribution of cold climates for present purposes. For example, in Köppen's classification there are "frost climates" with the warmest month between 0°C and 10°C and "cold boreal forest climates" which experience a coldest month below -3°C and the warmest below 10°C. Whilst a complex scheme when fully applied (see Petterssen, 1958) it does, nevertheless, have somewhat arbitrary temperature limits based upon a variety of criteria which are far from precise (Barry and Chorley, 1971) and which are *not* related, in any direct manner, to weathering. Later classifications are more rigorous but, based as they are on environmental features relevant to plant growth, have, despite the attempts by such as Wilson (1969), limited use for the determination of geomorphic processes. Climatic classifications such as those cited above do not readily account for situations where cold-based processes can be expected for part of the year. In such environments the climate is not *characterised* by cold but, in terms of weathering, it constitutes a major element for *part* of the year.

Burdick, *et al.* (1978), in a consideration of cold region engineering in the Northern Hemisphere, suggest that the southern limit of cold should be considered as 40°N. They note that climatologists utilise the isotherm for the average temperature of the warmest month being above 0°C but not above 10°C to identify the southern boundary. Alternatively, some engineers in the U.S.A. use the 150-300 mm depth of frost penetration or soil freezing to derive a southern boundary. However, neither of these methods satisfies the problem of characterising areas where cryogenic weathering operates. According to Burdick, *et al.* (1978, p.1) a more 'practical definition' of a cold environment could be one based upon the design and operation requirement essential to the maintenance of the industrial and social economy. In other words, if a city or state needs to spend large sums of money to facilitate snow removal then that place is situated in a 'cold region'. This is clearly an approach suited to urban environment fiscal practice, but it could be modified to shift the emphasis from snow removal to that of building codes requiring frost protection as the distinguishing criteria. Even then, this is an approach more applicable to buildings than landforms, and many marginal areas may still not be recognised.

Engineers offer other methods, some elements of which are useful for the evaluation of weathering. For instance, such factors as 'degree-days', 'air freezing index', 'surface freezing index', 'mean freezing index', 'thawing index' (Oliver, 1973), 'frost index' (Nelson and Outcalt, 1983), 'frost days', 'ice days', 'freeze-thaw days' (Wexler, 1982), and the 'frost number' (Nelson and Outcalt, 1987) can all be calculated for any given area or locality. The freezing index, a measure of the combined duration and magnitude of below freezing temperatures, can be used to calculate the depth of ground freezing and may be useful for estimating the potential for weathering. Johnson and Hartman (1971) produced maps of the Northern Hemisphere to show freezing indices and thawing indices, as too did Corté (1969). A further refinement, the 'design-freezing-index' (cumulative degree days of air temperature below 0°C for the coldest year in a 10 yr cycle) is another manner by which the possible frost hazard can be expressed. Conversely, the length of the 'freeze-free season' (Schmidlin and Dethier, 1986) could be calculated, and the shorter this season the greater is the potential for damage from frost related processes. These sorts of indices are particularly useful for highway design and can give a good estimate of probable damage to such as roads (see Johnson, 1952 for a review of techniques and applications). However, the above techniques give no information regarding the frequency, duration or amplitude of freezing and thawing events.

Although not an expression of a cold climate, freeze-thaw is a component that is of particular significance to weathering studies. Some areas, such as continental Antarctica, may suffer prolonged low temperatures but few freeze-thaw cycles (but see McKay and Friedmann, 1985). Conversely, other regions may not attain a particularly low temperature but will experience substantial across-freezing oscillations. Brochu (1986) even suggests that the "periglacial zone" should be defined upon the basis of a minimum of 10 freeze-thaw cycles per year (averaged over 10 years). However, as Washburn (1979, p.71) points out, a measure of how many times air temperature crosses the freezing point is not an adequate measure of its effectiveness. What is required for weathering studies is a measure of the amplitude, duration, rate of change of temperature and frequency of sub-zero events actually *on* and *in* the material concerned (Russell, 1943), plus some evidence as to whether freezing *actually* took place. Surface temperatures are often radically different from those of the air due to the effects of insolation (Taylor, 1922; Souchez, 1967) and so normal meteorological screen data are of but limited value. However, with only a few exceptions, it is this standard meteorological data that is normally used to evaluate freezing and thawing.

Arndt (1943) monitored the number of freezing and thawing cycles in the air and at the top and bottom of a concrete pavement over a period of five years. He defined the cycle limits as -0.56°C (31°F) and +0.56°C (33°F) but did not assess if freezing actually took place. In a like manner Swanberg (1945) also monitored temperatures of the air, top and bottom of a slab and at various depths down to 1.52m (60 inches) and calculated the number of freeze-thaw cycles roads and pavements might be subjected to.

The above studies were not however directly concerned with weathering processes. More recently a number of geomorphological studies have attempted to relate the number of freeze-thaw cycles to weathering or other geomorphic processes (e.g. Herschfield, 1974). Amongst the more notable are those of Cook and Raiche (1962) for Resolute (N.W.T., Canada) and Fraser (1959) for the whole of Canada, and Russell (1943) and Visher (1945) for the U.S.A. In a similar fashion, Hewitt (1968) investigated freeze-thaw frequencies for the Karakoram Himalaya, Mathys (1974) for the Jungfrau in Switzerland, and Barsch (1977) for the Swiss Alps. More recently, Herschfield (1974, 1979) analysed the number of freeze-thaw days for 1300 weather stations in the U.S.A. and related them to the amount of road damage that took place. A freeze-thaw day was defined as one within which air temperature crossed 0°C and then returned to the original side. However, all of these studies relied upon *air temperature* records for the determination of freeze-thaw frequency. In addition, a variety of criteria were used for defining a freeze-thaw event. For example, Hewitt used three crossings of 0°C, Fraser a rise to 1.1°C (34°F) following a drop to -2.2°C (28°F), Russell a freeze at -2.2°C (28°F) following a thaw at 0°C (32°F), Visher the range -3.9°C (25°F) to 1.7°C (35°F), whilst Cook and Raiche analysed their data according to each of the criteria of Russell, Fraser and Visher. The number of freeze-thaw cycles varies enormously dependent upon the criteria adopted, with (for the same period) 15 according to the definition of Russell, nine for that of Fraser and only three when following Visher. This problem of the band width of the cycle markedly influencing the perceived number of freeze-thaw cycles was also noted by Walton (1982) where an increase in band width from -0.5/+0.5°C to -1.0/+1.0°C for soil temperatures roughly halved the number of cycles recorded. Based upon available American studies, Schmidlin *et al.* (1987) undertook an analysis of freeze-thaw days in the U.S.A. and recognised such complicating factors as altitude, longevity of snow lay and the height of the shelter used for the recording instruments. Despite the multiplicity of hazards they conclude that freeze-thaw days should be determined by a minimum temperature of -2.2°C (or less) during any given day and a maximum of 0°C (or greater) in an instrument shelter at 1.5m above the ground. However, as important as the defining of cycle parameters is, in the absence of *actual* monitoring of freeze and thaw in soil, rock or building material, analyses of air temperatures provide at best only a poor guide to temperatures on or in the medium under study.

Ground, rock or building material temperatures are influenced by their albedo, incoming and outgoing radiation, and insulation by vegetation or snow. The inadequacy of air temperature data has been shown by Hall (1980a) where insulation provided by a snow cover inhibited any freeze-thaw cycles on underlying rock surfaces during three months when the air experienced 48 cycles. These results, associated with an investigation of nivation processes (Hall, 1974, 1980a, 1985; Thorn and Hall, 1980), suggested that earlier studies (e.g. Lewis, 1939) which had relied upon air temperature data may have misinterpreted the timing, nature and effectiveness of freeze-thaw weathering. Examples of temperature variability between the air, soil, stone and various types of vegetation are given by Mølgaard (1982) and clearly illustrate the enormous range that can occur (e.g. air @ 2m = 6.5°C, exposed stone = 10°C and the apex of *Dryas integrifolia* = 16.5°C). Equally, Mølgaard shows that there is also a distinction between sloping and level ground, and between slopes of different aspects. Thus, not only is air temperature an ineffective indicator of ground or rock conditions, but care must be taken in extrapolating from one microclimatic site to another as great variability can occur over very short distances (Hall, 1980a). In another study, Friedmann *et al.* (1987) found short-term peaks of +5°C on a northeast-facing surface in the Ross Desert area of Antarctica during a time when air temperatures fluctuated between -45°C and -10°C, clearly demonstrating the dangers of presuming that temperature cycling does not occur in the cold Antarctic. It has also been shown (e.g. McGreevy, 1985) that there can be considerable temperature differences between different rock types as a function of their albedo, whilst thermal conductivity differs not only between rock types but within rocks of the same type due to differences of mineralogical structure.

Thus it can be seen that the definition, for weathering studies, of a 'cold environment' is far from simple. Certainly it goes beyond the description of Tricart (1970) that states:

"Great frozen expanses of the ice sheets, the chaos of séracs on mountain glaciers, snow and névé fields which last through the summer in sheltered hollows on slopes, ground which is by turns soft, marshy, and hardened by the frost of the tundra, flows of sodden earth, great unvegetated debris slopes, peculiar geometrical patterns of stones - these are the pictures called to the geomorphologist's mind by the term 'cold environments'."

Rather, a modified version of a later statement by Tricart (p.xii)

"...cold environments may be defined as those in which the conversion of water to the solid state plays a predominant geomorphological role" would seem to better express the framework within which this study is undertaken. However, even within this framework of a 'cold climate' further information is required to characterise it in terms of weathering. Whilst recognising the importance of actually knowing that freezing of water takes place, it is still required to know how often and for how long freezing occurs. This temperature-based information must come from the rock itself for not only does the air temperature give no indication of rock conditions but the data has implications for other than water-controlled weathering (see below). A component of the freezing of water must be the origin, frequency, duration, chemistry and amount of moisture that is made available. Whilst not a direct 'control' upon a cold climate it is, nevertheless, an attribute of that climate that has an effect upon the resultant weathering. *However, the absence of moisture need not imply an absence of weathering.* Here again, detailed information regarding the range of temperatures, the rate of change of temperature and frequency of change, all factors that can influence freeze-thaw, will have an effect equally upon salt weathering and thermal fatigue. Thus it is not enough, from the point of view of weathering, to simply state a climate is 'cold' for information on its attributes with respect to the rock or building material is also required. However, even without a uniform definition, as long as all workers state their working criteria comparison will be possible.

1.4 Cold Region Weathering Information from General Textbooks

In most textbooks and publications, weathering in cold regions is perceived as being dominated by mechanical processes and by the freeze-thaw mechanism (or any of its synonyms - see below) in particular. To exemplify the situation, the following are typical examples extracted from a number of representative textbooks. It should be noted that many of the citations below are but part of an extensive discussion and some cover the subject extremely well. The aim is to show a number of "typical" themes that become apparent from consideration of the sort of basic text that any reader might first consider. The references provide a selection from readily accessible texts. It is likely that a reader anywhere within the English speaking world would use one or more of the cited texts if perusing a library for general information on weathering. More process specific and region specific texts are dealt with after this section.

Ahnert (1998, p. 65)

"Temperature fluctuations are particularly effective if they cross the freezing point and humidity is also present. Water increases its volume by 10 per cent when it becomes ice. The pressure created in pores and fissures near the surface that contain water causes frost shattering."

Bland and Rolls (1998, p.87)

"This is the process of rock disintegration that takes place when water freezes and so expands within the rock."

Bloom (1998, p.127)

"...two popular concepts - (1) that expansion of confined water by freezing breaks rocks, and (2) that the process is enhanced by frequent cycles through the freezing temperature - are both probably wrong."

p.311 "Despite continued uncertainty about the exact process by which rocks are fractured by freezing and thawing, there is no doubt that regions now in the periglacial environment are characterized by great quantities of angular, fractured rock detritus."

Clark and Small (1982, p. 17)

"Frost weathering: this is the most widespread type of pure physical weathering."

p.17 "A traditional view of frost weathering is that it results from 'frost wedging' or 'splitting', which occurs when water penetrates joints and bedding planes, undergoes a phase change from liquid to ice, and thus has the potential to expand by approximately 10% if unconfined by the rigidity of the surrounding rock."

p.17 "However, experiments indicate that frost processes do not produce very fine particles.."

p.17 "The potential power of frost weathering is shown by the fact that theoretically freezing water in a totally confined cavity can exert a maximum pressure of 2,100 kg/cm^2 at -22°C (below -22°C the ice contracts, reducing stress)."

Clowes and Comfort (1982, p. 21)

"If temperatures fall below freezing point, the ice formed at 0°C expands about 9 per cent. Potentially this can create great pressures against the confining walls and in theory a maximum of 2100 kg/cm^2 is reached at -22°C"

p.21"There are very few rocks that could withstand such pressures."

De Blij and Muller (1996, p.489)

"Frost wedging, the repeated freezing and thawing of water in rock cracks and joints, loosens pieces of bedrock."

p.493 "The presence of water in soil and rock, and its freezing and thawing, is the key disintegrative combination in periglacial environments."

p.493 "frost wedging is capable of dislodging boulders from cliffs, of splinterring boulders into angular pebbles, of cracking pebbles into gravel-sized fragments, and of reducing gravel to sand and even finer particles."

Easterbrook (1993, p. 16)

"When water freezes, its volume is increased by about 9 per cent. The expansion of the ice being frozen *(sic)* in a confined space thus exerts great pressure against the sides of the material enclosing the ice."

p18 "Optimum conditions for the wedging effect of freezing require a supply of water, many alternations of freezing and thawing, and yet enough sustained freezing at temperatures well below 0°C that masses of ice will grow."

p.18 "Repeated freezing and thawing, with resulting shattering of rock, is especially common in high mountains....a layer of angular rubble can be produced..."

Getis, *et al.* (1996, p. 76)

"If water that soaks into a rock...freezes, ice crystals grow and exert pressure on the rock. When the process is repeated - freezing, thawing, freezing, thawing and so on - the rock begins to disintegrate."

Monroe and Wicander(1995, p.100)

"Frost action involves the repeated freezing and thawing of water in cracks and crevices in rocks. When water seeps into a crack and freezes, it expands by about 9% and exerts great force on the walls of the crack, thereby widening and extending it by frost wedging."

p.100 "Frost action is most effective in areas where temperatures commonly fluctuate above and below freezing."

p.100 "The debris produced by frost wedging in mountains...are simply angular pieces of rock from a larger body.."

Ollier (1984, p. 13)

"Water expands about 9% on freezing at 0°C. The great change in volume has a potentially disruptive effect, and frost shattering is one of the greatest mechanical agents in weathering."

p.14 The formation of ice can itself prize rock fragments apart. This works along planes of fissility in rocks, and produces angular rock debris."

Press and Siever (1986, p. 114)

"One of the most efficient physical weathering mechanisms is freezing and thawing of ice. Water expands as it freezes, and the expansive force exerted during freezing is enough to crack...rocks..."

Renton (1994, p. 156)

"During the spring, throughout the temperate, more humid parts of the world, roadways below road cuts and the bases of cliffs are commonly littered with boulders and rock fragments, the end products of one of the most common mechanical weathering processes, frost action."

p. 156 "A volume of water will expand about 9% as it freezes. This may not sound like much, but water freezing within a completely filled, enclosed container could theoretically generate a pressure of about 1,500 pounds per square inch (680 kg/cm^2)." "Expanding ice will also break the strongest rock" "Most of the rock litter that frequently covers the ground in high mountainous areas is produced by frost action."

p. 243 "In both arid and humid temperate regions, frost action is a major mechanism of physical weathering."

Rice (1988, p. 96)

"This phase change from water to ice thus involves a volumetric expansion of just over 9 per cent, but the temperature at which it occurs varies according to the confining pressure; for every increase of 10 MN m^{-2}, the temperature at which ice forms declines by about 1°C."

p.96 "Theoretically with a temperature of -22°C, a pressure of 216 MN m^{-2} could be exerted on the confining walls of a rock joint..."

Ritter, *et al.* (1995, p. 95)

"The most significant processes of physical weathering involve forces generated by crystallization of ice (frost action).." "In a perfectly closed system, water experiences a 9 percent increase in volume upon freezing and almost certainly produces hydrostatic pressures that exceed the tensile strength of common rocks."

Scott (1996, p. 392)

"Frost wedging results from the growth of ice crystals within rock fractures or hollows. This process can generate pressures of more than 100 kilograms per square centimeter (1,400 lb/in^2) of rock surface."

p.392 "Frost wedging can be highly effective in regions where daily freeze-thaw cycles occur during a large part of the year..."

p.392 "Active frost wedging in mountainous regions can litter the surface with angular rock fragments of all sizes..."

Selby (1982, p. 16)

"Hydrofracturing and frost action are two of the most important and widely recognised processes of physical weathering."

Selby (1985, p. 396)

"Frost-wedging is the prying apart of materials, commonly rock, by the expansion of water upon freezing."

p.396 "The result is that large accumulations of angular rock debris are characteristic of alpine and polar environments..."

Skinner and Porter (1995, p. 200)

"Wherever temperatures fluctuate about the freezing point for part of the year, water in the ground periodically freezes and thaws. When water freezes to form ice, its volume increases by about 9 percent. The high pressures resulting from this volume increase lead to disruption of rocks."

Small (1972, p. 25)

"Thus, frost weathering can occur only where there are atmospheric freeze-thaw cycles.."

p.27 "Arctic climates. The dominant weathering process here is frost action..."

Sparks (1986, p.27)

"In cold regions the most effective action is caused by the crystallisation of water into ice, and the expansion accompanying this."

Summerfield (1991, p. 146)

"In arctic and alpine environments the surface is often seen to be composed of a layer of angular rock fragments commonly described by the term felsenmeer and attributed to the operation of frost weathering."

p.147 "...experimental work has failed both to clarify fully the exact mechanisms involved in frost weathering and to define precisely the climatic conditions under which the process is likely to be most effective."

p.147 "Uncertainties about the efficacy of volume expansion on freezing in rock shattering has encouraged the examination of other possibilities."

Thompson and Turk (1995, p. 247)

"Water collects in natural cracks and crevices in rocks. If the outside temperature drops below 0°C, the water may freeze. Water expands when it freezes. Thus, water freezing in a crack pushes the rock apart in a process called frost wedging."

p. 248 "Anyone who has spent time in the mountains has noticed large piles of broken, angular rock at the bases of the cliffs....broken from the cliffs, mainly by frost wedging." (Repeated (p. 195) in **Thomson, *et al.,*** 1995)

Thompson, *et al.,* (1986, p. 147)

"The freezing of water in a confined space generates an outward force of about 1500 t m^{-2} and as this mainly acts near the surface of the rocks, it sets up pressures between outer and inner layers which can lead to exfoliation...The results of such breakdown can be seen in the rubble and talus slopes formed in mountain areas subject to frequent freezing."

Trenhaile (1997, p. 44)

"It is generally assumed that rocks in cold regions are split or shattered by the alternate freezing and thawing of water contained in crevices or in small voids and capillaries."

p.45 "Effective frost action can only occur in environments with a plentiful supply of water and suitable fluctuations in temperature. Many sites appear to lack at least one of the essential requirements."

Weyman and Weyman (1981, p.21)
"If water is lodged in a crack in a rock and then freezes, it will expand and press against the walls of the crack, causing the rock to break."

White, *et al.*, (1984, p. 231)
"Water freezing to form ice crystals undergoes a volume expansion of 9% and can develop pressures in a confined space theoretically in excess of 200 MN m^{-2}.

p.231 "Rock breakdown by this process...will tend to be most effective in conditions with frequent alternations of temperature about 0°C."

Yatsu (1988, p. 74)
"Leaving aside the question of the conditions under which the process of frost shattering takes place, the most fundamental factor is the 9 per cent volume expansion upon freezing...and the occurrence of pressure associated with this volume expansion."

From the above it is clear that there are a number of dominant themes that, in one way or another, are deemed part of, or associated with, the freeze-thaw weathering process. Listed, those main themes are:

- the *c.*9% increase in volume as water changes to ice
- very large stresses can be exerted (although values seem to vary)
- rock breakdown can occur (Figs 1 & 2)
- the products of that breakdown are usually angular
- freeze-thaw weathering is a very common process in cold regions
- it is particularly common in cold, mountain regions
- temperature fluctuations need to cross either 0°C or some undefined freezing point
- frequently it is atmospheric freeze-thaw cycles that are considered as defining what is occurring in the rock

- some authors identify that the maximum (theoretical) pressures are only attained at temperatures close to -22°C, others do not identify this
- water needs to be present, usually in some amount approaching saturation
- that, in fairness, many of the authors *do* provide great detail regarding the processes and controls, but
- the above outlines the basic foundation of the freeze-thaw concept.

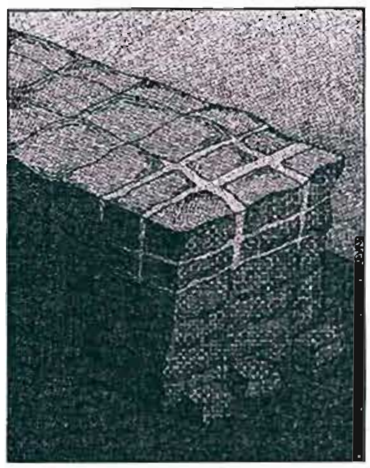

FIGURE 5-5 Frost wedging occurs when water seeps into cracks and expands as it freezes. Angular pieces of rock are pried loose by repeated freezing and thawing.

Fig. 1

Fig. 2.3 Frost action

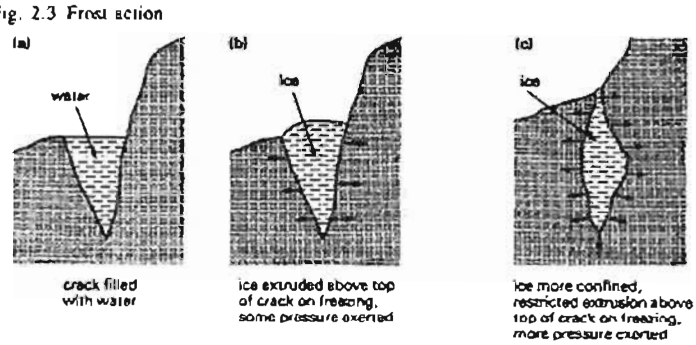

Fig. 2

Figs. 1 & 2

Examples of the way freeze-thaw weathering is depicted in texts.

Fig1 is from Thompson and Turk (1995) and Fig 2 from Clowes and Comfort (1982)

1.5 Weathering Information from Cold Region Texts

More specialised cold regions texts also cover the action of freeze-thaw. Even here, the fundamentals remain much the same as outlined from more general texts. Following are a some examples, for the period 1970 to 1996, from what might be recognized as the main "cold regions/periglacial" texts:

Ballantyne and Harris (1994, p.163)
"Traditionally, the breakdown of rock in periglacial environments has been attributed to frost weathering, the mechanical disintegration of well-lithified rock as a result of repeated freezing and thawing."

Davies (1972, p.24)
"Frost achieves maximum significance as an agent of rock weathering in periglacial conditions."

Embleton and King (1975, p.4)
"Freeze-thaw action is undoubtably the most important process of rock weathering in the periglacial zone; it is the primary agent responsible for such features as talus accumulations and blockfields, and the breakdown of debris into particles fine enough to be handled by running water and wind."

French (1976, p.37)
"The disintegration and mechanical breakdown of rock by the freezing of water present within pore spaces, joints and bedding planes has long been thought of as a particularly potent geomorphic agent in the periglacial environment. The presence of extensive upland surfaces of angular frost shattered rocks and boulders in both present-day and Pleistocene periglacial environments is the most dramatic morphological features formed by intense frost wedging."

French (1996, p.31)
"Frost action is a collective term used to describe a number of distinct processes which result mainly from alternate freezing and thawing in soil, rock and other materials."

p.41 "The disintegration and mechanical breakdown of rock by the freezing of water present within pore spaces, joints and bedding planes is widely regarded as a particularly potent geomorphic agent in periglacial environments. Extensive surfaces of angular rock fragments....are the most dramatic features...".

Ryder (1998, p.4)

"Frost action.....contributes to shattering of bedrock..."

p.23 "Frost shattering of bedrock is ubiquitous in alpine areas. This weathering process widens cracks in bedrock and liberates angular rock fragments."

Tricart (1970, p.113)

"Rock Shattering: This is almost entirely the work of freeze-thaw; the role of other processes is negligible."

p.112 "This morphogenetic system is marked by the dominance of mechanical weathering (frost shattering)..."

p.112 " The morphogenetic system acts on the slopes through the shattering of the bed rock, mainly by freeze-thaw..."

Washburn (1973, p.62)

"Frost wedging characteristically produces angular fragments that can be of varying size..."

A number of points emerge from the above:

- freeze-thaw is perceived as the major weathering process in cold regions
- it is considered the prime cause of many landforms
- the presence of water is assumed (hence the ability to presume freeze-thaw)
- spatial and temporal accessibility to water (by the rock) is not intrinsic to the arguments
- angular rock fragments are the product of freeze-thaw weathering
- the finding of certain landforms (e.g. angular screes or felsenmeer) identify the operation of freeze-thaw weathering.

1.6 Information from Antarctic Books

Books dealing with the periglacial geomorphology of the Antarctic are few; most rather deal with specific areas or landforms. One book does deal with weathering, in general, within the Antarctic (although, this is mostly restricted to the continent) and another considers weathering within the framework of biological studies. Comments from these are presented:

Campbell and Claridge (1987, p.97)

"Due to low temperatures and arid conditions, water-based processes are not very effective in Antarctica. One of the main agencies of physical disintegration, freeze and thaw, which is the cause of features such as talus or scree slopes, blockfields or felsenmeer, and of the breakdown of debris to fine particles, is comparatively restricted in Antarctica, because temperatures are continuously below freezing in winter and, in most places, are seldom above freezing in summer."

p.102 "Even on old surfaces, at high elevations, there is often no evidence of significant bedrock shattering by frost action. Locally, however, bedrock shattering may be extensive, forming areas of rubbly debris or felsenmeer."

p.104 "Spalling is assumed to occur through freezing of moisture in thin fine cracks or fracture planes within the rock..."

Fogg (1998, p. 65)

"Water, entering fine cracks, fracture planes, or pores, expands on freezing, breaking the rock."

p.65 "In some situations freeze-thaw cycles can exceed 100 per year."

p.70 "Under more moist but still cold conditions, as in the maritime Antarctic and northern Arctic coasts, chemical weathering and frost action are more evident...."

Taylor[1] (1916, p. 136)

"I slid down the steep eastern face of the Riegel, where King Frost had gnawed away the cliff and built up a steep ramp of talus...".

p. 380 "It was obvious that frost action was now leading to a great deal of erosion..."

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1 Numerous anecdotal comments in the diaries of the early Antarctic explorers cite the role of frost action. I give examples from just one such diary.

P. 388 "...I think it is merely the result of frost cleavage..."

From the above a number of points are clear:

- that the above considers mainly the continent
- the role of freeze-thaw is spatially and temporally limited
- that landforms due to the action of freeze-thaw (e.g. scree, felsenmeer, etc) are present
- that the action of freeze-thaw is the cause of observed spalling
- the use of the process on the continent is much more limited than it is for other cold regions.

1.7 Landforms Associated with Mechanical Weathering in Cold Regions

As is becoming obvious from the above, a number of landforms are intimately associated with the freeze-thaw process. Not only is freeze-thaw frequently argued as the process operative within the development of that landform but often the landform is used to identify the former (or present) operation of freeze-thaw. Comments and discussion regarding this are extensive and so a few typical examples only are cited to exemplify the issue:

Bird (1969, p.347)

"In dry arctic environments special landforms in limestone and other sedimentary rocks are primarily a consequence of mechanical weathering believed to be associated with frost riving."

Boch and Krasnov (1994 (translation of 1943 article), p.179)

"Altiplanation terraces...appear only as a result of combined action of solifluction and frost weathering..."

p. 180 "Under the conditions of frequent temperature fluctuations around the point of freezing, the work of frost weathering is particularly active here."

Boyé (1994 (translation of 1952 article) p.214) re nivation: "As far as the attack on rocky relief is concerned, the majority of authors agree with us concerning the power of rock shattering, or more exactly its cryoclastic activity."

p.215 "The production of this debris is attributed to the reduction (comminution) by gelivation (freeze-thaw processes) of material in contact with the snow."

DeWolf (1988, p. 103)

re stratified slope deposits:- "There is now consensus of opinion that bedding and grading can only be acquired by the action of selective processes working on material originally produced by frost action."

p.106 "Since the grain sizes of the gelifracts thus identified are less than 2.5 cm, these are genuine grèzes and imply the existence of sufficient freeze-thaw cycles to effect the necessary rock breakdown."

Priesnitz (1988, p. 56)

re cryoplanation:- "Rock disintegration seems to result mainly from frost weathering."

Taylor (1916, p. 175,)

"Pronounced erosion by "thaw and freeze" (= nivation)...", and continued on p. 176 with "...further erosion by nivation will produce basins with level bottoms...".

Waters (1978, p. 157)

"a typical periglacial landscape possessed not only of an abundance of features indicative of Pleistocene frost action but also a morphology which was moulded by that frost action..."

p. 158 "Cryoplanation terraces and their relations with frost-riven cliffs and tors..."

1.8 Synonyms

With respect to 'freeze-thaw weathering', numerous synonyms are used in the literature on cold region weathering processes. Van Everdingen (1998, p.27), in the recent "Multi-Language Glossary of Permafrost and Related Ground-Ice

Terms", notes the following terms as synonymous with freeze-thaw weathering: frost shattering, frost wedging, congelifraction, frost bursting, frost prying, frost riving, frost splitting, and gelifraction. Others terms found in the literature include gelivation, microgelivation, macrogelivation, cryoclastis, and frost weathering whilst Dylikowa and Olchowiak-Kolasińska (1954) provide a list of comparable terms in English, Polish, French, German and Russian. Although all of these terms certainly "overlap" in meaning with respect to "freeze-thaw weathering" and are frequently used solely in that context, some of the terms have additional connotations. For example, 'frost bursting' is the explosive breakdown of rock as a result of the pressures developed during freezing. Macrogelivation is where 'freeze-thaw' exploits the texture of the rock and fragments (e.g, along stratification) whilst microgelivation is where breakdown occurs without any visible link to the texture (Tricart, 1956 in Evans, 1994). In one way or another, all of these terms do, however, identify the breakdown of rock (weathering) as being the result of freezing and thawing of water within the rock system.

1.9 Discussion

This acceptance of freeze-thaw as the dominant weathering mechanism has led to its use in almost any discussion regarding the origin of cold region landforms or sediments. Thus, in the absence of any empirical testing, a number of criteria have evolved that are now considered indicative of the past or present action of freeze-thaw weathering. Such criteria include, for a cold region present or past, the finding of angular clasts or attributes to a landform, the origin of any non-glacial feature that requires weathering, and the causative mechanism for the breakdown of rock (bedrock or transported).

Thorn (1988, 1992) provides extensive, reasoned accounts regarding the definition and historical context of the freeze-thaw concept and its application with regard to a number of landforms. As much of what he writes underpins the rationale for this thesis, an extensive presentation will be made of his thoughts. Thorn (1992, p.10) synthesizes the perceived problem where he states, "From its inception, periglacial geomorphology has been dominated by the concept of frost wedging, a synonym for weathering by freezing and thawing. The story is one of casual empiricism gathering respectability by repetition until it attained the stature of an article of faith." Historically, the concept that (Thorn, 1992, p.11) "...freeze-thaw weathering dominates cold regions gained respectability long before there was the ability to test it in the field." As Thorn (1992, p.11) then explains, "..the most common argument to substantiate the importance of freeze-thaw weathering is both circumstantial and circular." Thus, the angular rock fragments found in the field "...were assumed to be the product of the dominant process, namely freeze-thaw weathering. Today, it is common to assume that angular rock fragments are definitive evidence of frost weathering." As a brief aside, texts concerning processes in *hot* deserts (e.g. Abrahams and Parsons, 1993) also show highly angular rock and debate the relationship of process to landform. The discussions in these texts are almost identical to that found in cold region texts. Thus, if a geomorphologist working in a hot desert is plagued by almost identical questions, how can a periglacial geomorphologist simply assume that *his/her* angular clasts must be the product of frost action when the desert geomorphologist may well be considering clasts with an identical form the product of thermal stress fatigue or salt weathering? The whole problem justifies Thorn's (1992, p.11) assertion that, "...what periglacial geomorphologists need more than any other single item is a way to determine in the field whether or not bedrock fragments have been frost weathered." To date, no such test exists. Everything is based on assumption.

The application of laboratory studies to the concept of freeze-thaw weathering is not without its problems. "Foremost among these problems has been uncertainty concerning the thermal and moisture regimes which actually prevail within natural bedrock and regolith fragments. Consequently, freeze-thaw cycles used in laboratory samples may or may not reflect natural temperature ranges. A similar problem has overshadowed the moisture issue, and most laboratory experiments have embraced very crude approaches to moisture conditions and supply" (Thorn, 1992, p.10). This issue was outlined in papers by McGreevy and Whalley (1982) regarding temperature, and McGreevy and Whalley (1985) dealing with moisture. These authors observed that thermal conditions used in experiments rarely ever reflect *rock* conditions but are usually related to temperature variability monitored in the air (McGreevy and Whalley, 1982). As Thorn (1988, p. 13) states, "In this context the entire range of published papers that use meteorological screen temperatures as a surrogate for bedrock and/or regolith temperatures is irrelevant, since the variables that intervene between air and surface temperatures are too numerous and too complex in their interaction to permit reliable extrapolation. This deficiency places all laboratory research in jeopardy, as the supply of field data from bedrock sites is entirely inadequate to reliably validate laboratory results". With regard to moisture conditions McGreevy and Whalley (1985) make the point that, at the time of writing, almost no data regarding rock moisture content or rock moisture chemistry (which will affect the freezing point) were available. Thus, given "..almost total ignorance of both field conditions and applicable theory, the experiments have not even precluded the possibility of other mechanisms.." (Thorn, 1992, p.11). The key to all of the problems surrounding laboratory experiments regarding freeze-thaw was, in a similar context, stated by Warren (1914a, p.413) who warned that, it would be unsound "..to assume that the results of a certain experiment must also be produced by natural agencies, without evidence that similar conditions exist in Nature to those employed in the experiments". Sadly, hardly any laboratory experiments have taken heed of this; they have assumed conditions that may really have no relationship to either the thermal or moisture conditions the rocks under experimentation ever experience(d) in Nature. Key issues in this are such factors as the amount of water, the chemistry of the water, the rock temperatures and, particularly, the rate of change of temperature ($\Delta T/t$), and the thermal gradient. Without such data, it cannot be known whether the laboratory experiments replicate the field situation or not.

Thus, "Field corroboration is something of a misplaced concept with respect to frost weathering. At present there is no adequate criterion to establish that bedrock weathering or further comminution of rock fragments has been dominated by freeze-thaw weathering. Nevertheless, it is clear that the majority of periglacial researchers believe that freeze-thaw weathering of bedrock is an established fact, and that it is an acceptable premise upon which to base many secondary concepts (e.g. cryoplanation)" (Thorn, 1992, p. 11). The problems cited above with regard to laboratory experiments (i.e. the need for data on rock temperatures, moisture content, etc.) apply equally to the use of the freeze-thaw concept in the generation of landforms. Rather than deduction based on empirical data, it is usually the observation of 'angular clasts' that provides the assumption of 'freeze-thaw' weathering as factor in landform origin. Thus, the invocation of freeze-thaw as a central tenet of nivation, cryoplanation, blockfields, tors, etc. is without empirical foundation (Thorn, 1988, 1992).

1.10 The Present Research Undertaking

The present undertaking involved the collection of the data necessary to investigate weathering in cold regions, and the freeze-thaw process in particular. Especial consideration was given to studies in Antarctica due to logistical opportunities and also because this region should be ideal for such an undertaking as a result of its wide spectrum of "cold climates". It was necessary to consider *all* aspects of weathering (although mechanical processes were the major consideration) otherwise the predominance of processes other than freeze-thaw might not be recognised nor the potential synergistic relationships or the temporal/spatial variability in process identified. While this study deals almost exclusively with the mechanical processes, it is recognised that chemical processes must also play a role. Logistics and expertise did not allow equal consideration of the chemical weathering component.

Those factors that were monitored, as a foundation for understanding the nature and timing of the weathering, included rock temperature, rock moisture content and chemistry, rock properties and the study of various landforms and sediments associated with cold environments (past and present). Rock temperature was monitored both at the rock surface and at various depths, and the rate of change of temperature ($\Delta T/t$) and the thermal gradient were measured. Air temperature, as well as potentially pertinent climatic factors that might influence rock temperatures, such as radiation and wind, were monitored where possible. Spatial and temporal variability of the above attributes were also investigated as key factors whenever possible (i.e. a variety of sites were monitored at different frequencies and for varying lengths of time in order to determine spatial and temporal variability of process). Moisture data included estimation of rock moisture levels and gradients (and their variability through time, as well as spatially), rock moisture chemistry, and ancillary studies of the geochemistry of moisture sources. Rock properties included, where possible, tensile strength (determined from point load compressive strength tests via existent correlation equations), micro-indenter strength tests, permeability, porosity saturation coefficient, absorption coefficient, ρ-wave ultrasonic velocity values, and porosimetry. In some instances it was also possible to actually measure, or obtain an indirect measurement of, weathering rates in the field.

Together, the above data provide the field and laboratory background against which determinations regarding weathering processes and rates could be attempted. These data could then be used to facilitate meaningful laboratory experimentation to investigate the processes and rates of weathering (Fig. 3). Laboratory results could then be, in some instances, compared to field results to further refine both the laboratory experimentation and the overall assessment of weathering process(es) and rate(s). This was seen as an extremely important component of the study for, prior to this, few (if any) laboratory experiments were based on measured field parameters. This meant that the results could be applied back to the field with some degree of certainty that they were meaningful for that situation. Other laboratory experiments were undertaken to try and filter out the specific components of weathering - i.e. where one process impacts on another (e.g. the role of wet/dry weathering within freeze-thaw weathering).

The above approach enabled the monitoring and measurement of field conditions considered to be those fundamental to the understanding of weathering. Complementary laboratory experimentation, based on the field data, provided additional information. Field studies also included investigation of landforms and sediments where weathering, particularly freeze-thaw weathering, were argued to be the major, if not the sole, factor in their development. With respect to landforms, and to a lesser extent weathering, consideration was also given to the role of animals/organisms. Although somewhat peripheral to the main theme of this research, weathering due to algae was found to be a major factor at one locality whilst at a number of sites animals (penguins, elephant seals, albatrosses, musk ox, pika, marmots, yak, goats, and grizzly bears) were seen to play a significant role in the development of some cold region landforms. As animals had been rarely considered in this regard and may play a greater role than hitherto thought, the findings are included as part of the overall picture of cold region landform development. The work presented here provides an integrated look at the complexity of cold region weathering processes and the development of landforms and sediments in such regions.

1.11 Conclusions

The thesis will first examine various field attributes associated with weathering, then consider the laboratory findings. All of this will then be integrated to re-evaluate cold region landform and sediment production. Within this, details of the laboratory and field verification undertakings will be presented. Where appropriate, the role and impact of animals will be considered. The nature of the presented papers is such that it is inevitable that material in one paper may have attributes pertinent to more than one section. Where this occurs, reference will be made to the already presented paper. Thus, the thesis covers (in the broadest sense) the following attributes:

- field data regarding rock temperatures
- field data regarding rock moisture
- field and laboratory data regarding rock properties
- field data regarding p-wave ultrasonic velocities
- field data regarding weathering rates
- field data regarding degree of weathering
- monitoring, in the field, of:
    - freeze-thaw weathering
    - weathering by wetting and drying
    - salt weathering
    - thermal stress fatigue
    - thermal shock

- biological weathering
- chemical weathering

- laboratory simulations of :
    - freeze-thaw weathering
    - weathering by wetting and drying
    - salt weathering
    - thermal stress fatigue
    - thermal shock

- laboratory simulation data regarding
    - weathering rates
    - weathering mechanisms

- the proposal of a new weathering mechanism
- the proposal of new techniques for monitoring rock moisture chemistry
- the application of a new technique for deducing palaeoenvironments based on weathering rinds
- the application of the above to general theory regarding weathering in cold regions
- the application of the above to the interpretation of Quaternary sediments
- the application of the above to the origin and palaeoenvironmental meaning of certain cold region landforms
- field observations regarding a number of cold region landforms
- the role of animals in the formation of certain cold region landforms

The material presented in this chapter provides the background against which the aims of the weathering studies in this thesis can be viewed. Substantial quotes and references have been provided to show that freeze-thaw weathering is frequently perceived as "proven" to be the dominant weathering mechanism in cold regions. Experience has shown that questions regarding the veracity of freeze-thaw as the causative mechanism for landform genesis and sediment formation frequently meets with scepticism. To try and change an 'article of faith' is not easy but it is hoped that the material presented in the following chapters may provide some alternative answers as well as beg more realistic questions.

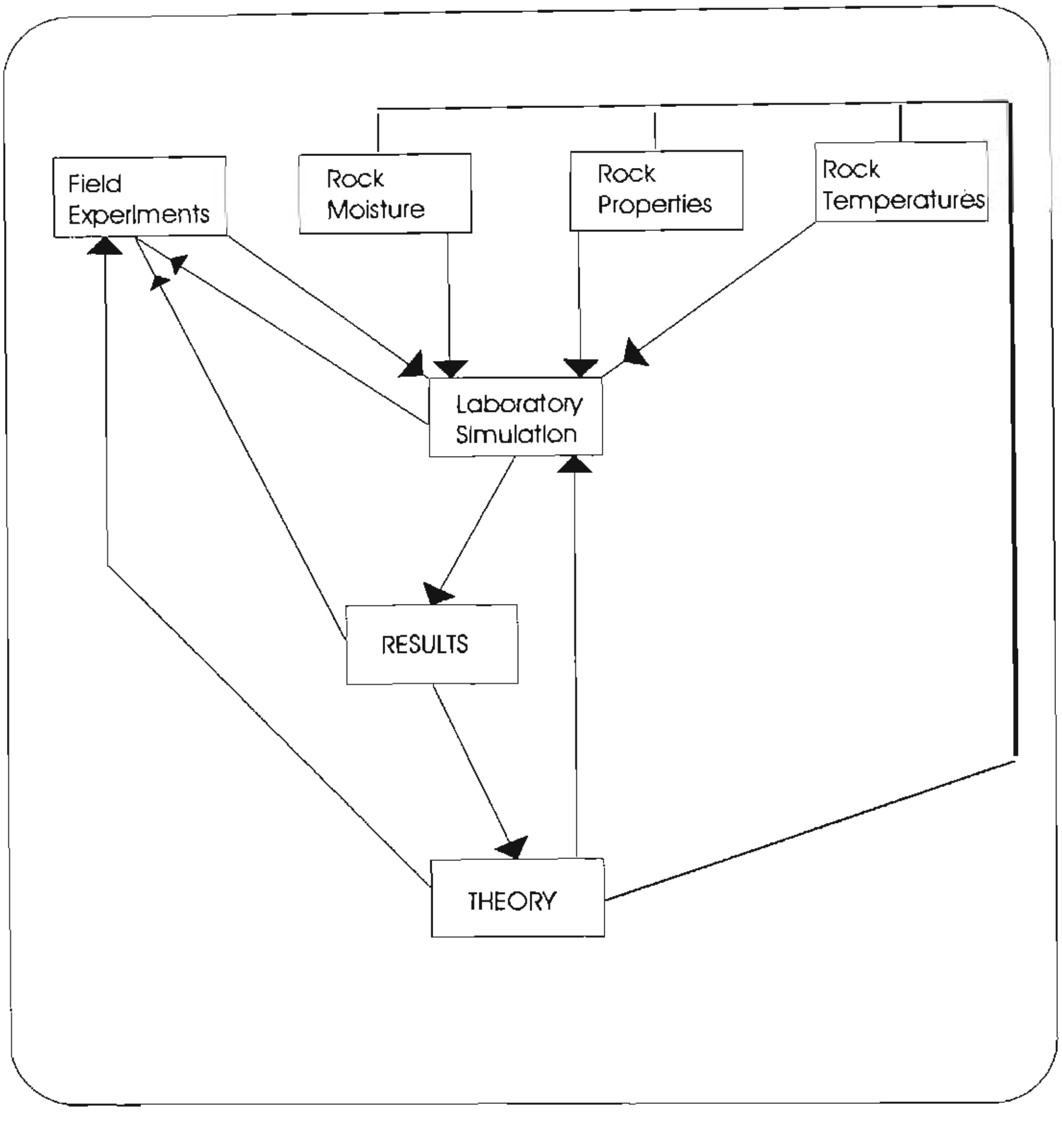

Fig. 3

A simple flow chart to show the relationship of the collected field data to both laboratory simulation and field experimentation and how this was integrated with available theory.

Chapter 2

Field Data

"We must make sure our theories accord with the facts as they are, not with imaginary facts which might conceivably be but which are not."

S.H. Warren (1914b)

With regard to the field investigation of weathering in cold regions, Bland and Rolls (1998, p. 86) state that "Many of the earlier field studies were based on the acceptance of the simple idea that the expansion by freezing of existing water caused rock destruction. Investigators collected information about the number and temperature range of freeze-thaw cycles, and the freezing rate, at and near rock surfaces". All in all this was frequently a very subjective undertaking wherein it was *assumed* that the weathering process was frost action and this was substantiated by temperature shifts across 0°C as derived from *air* temperatures. This use of air temperatures then led many studies to assume that the period of greatest activity was in the spring as that was when the largest number of such cycles were monitored. That the ground was frequently covered by snow at this time, and that air temperatures are not a surrogate for rock temperatures, were not an issue. Rather, the finding of these cycles substantiated the argument that it was indeed frost action that was the cause of weathering and hence many landforms. Rock temperature data with respect to the rock surface and at depth within the rock, plus the rate of change of temperature with time, are all required for any meaningful understanding of rock weathering. Rarely, however, were any such data presented.

Concomitant with the assumption of frost action, and its justification by means of temperatures measured in the air, was the presumption that there was water present in the rock to actually freeze. White (1976, p. 5) was one of the first to really question the assumed presence of rock moisture: "In how many mountain ranges or on how many arctic plains will bedrock be fortuitously ever become >50% water-saturated from melting snow or rain and then undergo rapid freezing to crack the rock?" As fundamental as temperature is to the understanding of rock weathering in cold regions so too is information pertaining to rock moisture - rock moisture content, distribution and chemistry. Water chemistry plays a role at a number of levels. Not only will impurities depress the freezing point such that, despite sub-zero temperatures, any water present may not freeze, but impurities also impact chemical weathering and influence processes such as salt weathering. Salt content depresses the freezing point and salts can precipitate out during drying of the rock or during the freezing process (to create a cryohydrate) and so facilitate salt weathering of that rock. Although not considered in any detail within this study, solutes within the interstitial rock water will also play a role in chemical weathering. Apart from the study by Kinniburgh and Miles (1983) data regarding interstitial rock water chemistry are not available. Equally, the distribution of water within the rock is of great significance. A block of rock that has, for example, only 25% saturation with respect to the whole block is likely to have 100% saturation in the outer shell due to the moisture gradient within the rock. This has important ramifications for the freeze-thaw mechanism, where degree of saturation influences process, as well as for *all* other water-based weathering processes. Data regarding moisture distribution are not available.

Rock properties clearly must influence weathering. Attributes such as porosity, permeability, albedo, strength, saturation coefficient, etc. all play a role. A number of factors influence what processes can or cannot take place, or the degree to which they can operate (e.g. permeability influences the ability of water to penetrate the rock and so facilitates or constrains water-based weathering processes). Attributes such as the tensile strength of the rock define the stress levels that must be exerted for failure to occur (i.e. if the stress exerted by frost action is below the tensile strength of the rock then failure will not occur). Changes in these strength/stress levels help understand the impact of fatigue (i.e. that multiple replications of a stress below the failure strength of a material will ultimately cause failure through fatigue). Many of these data are available from detailed tables of rock properties (e.g. Bell, 1983) but, where possible, it is still better to obtain data for the rocks actually under investigation; this can be both in the field (this section) and in the laboratory (Chapter 3). Information regarding ultrasonic p-wave velocity can also be obtained in the field. Ultrasonic velocity is useful for determining the presence of water and/or ice as well as for showing changes in sample dimensions as a result of heating by solar radiation.

As clearly shown in the Introduction, and cited above, the presumption of many cold region weathering studies has been that the dominant weathering process is freeze-thaw. The obtaining of data on the controlling factors (temperature, moisture, and rock properties) allows for a meaningful evaluation of the freeze-thaw process and for some investigation of the freeze-thaw mechanism itself. Equally important, though, is that it also allows for the evaluation of *other* weathering processes that might be occurring at that site. By obtaining data of sufficient detail (e.g. high-frequency rock temperatures, daily or hourly rock moisture variability) it is possible to also evaluate the role of weathering processes other than freeze-thaw (e.g. thermal stress fatigue) and thus the synergistic interaction of processes through time.

The papers presented in this section provide field data which were collected with the aim of answering questions pertaining to rock thermal conditions, moisture content and properties. Many of the data obtained here were then used as a foundation for (meaningful) laboratory simulations (Fig. 3) and to evaluate general theory with respect to the nature and timing of weathering in cold regions. In some instances (e.g. the data regarding moisture content fluctuations) these data allowed evaluation of whether certain processes could indeed occur (e.g. in the absence of water so freeze-thaw weathering could *not* occur) and in other instances gave direct information on the activity of a weathering process (e.g. in the example of moisture fluctuations, that of wetting and drying). As thermal and moisture conditions are fundamental to almost all weathering processes a significant amount of effort was spent on deriving as much information as possible on these attributes.

In this section the following papers that deal primarily with collection of field data are presented:

★ **Thermal Data**

- ♦ Hall, K.J. 1980a. Freeze-thaw activity at a nivation site in northern Norway. *Arctic and Alpine Research*, 12, 183-194.
- ♦ Hall, K.J. 1985. Some observations on ground temperatures and transport processes at a nivation site in northern Norway. *Norsk Geografisk Tidsskrift*, 39, 27-37
- ♦ Hall, K. 1993a. Rock temperatures from Livingston Island (Maritime Antarctic): Implications for cryogenic weathering. *Proceedings of the 6th International Permafrost Conference, Beijing*, 1, 220-225.
- ♦ Hall, K. 1997a. Rock temperatures and implications for cold region weathering: I. New data from Viking Valley, Alexander Island (Antarctica). *Permafrost and Periglacial Processes*, 8, 69-90.
- ♦ Hall, K. 1997b. The impact of temperature record interval and sensor location on weathering inference in periglacial environments. *Supplementi di Geografia Fisica e Dinamica Quaternaria*, III, 196.

- Hall, K.1998a. Rock temperatures and implications for cold region weathering: II. New data from Rothera, Adelaide Island (Antarctica). *Permafrost and Periglacial Processes*, 9, 47-55.
- Hall, K. 2001a.The necessity for high-frequency rock temperature data for rock weathering studies: Antarctic and northern examples. *First European Permafrost Conference, Abstracts,*102.
- Hall, K. 2001b. The conceptual fallacy of "weathering in cold climates" - the error in the assumption of zonality. *Western Division, Canadian Association of Geographers, Annual Meeting, Abstracts*, 23-24.
- Hall, K. and Hall, A. 1991. Thermal gradients and rates of change of temperature in rock at low temperature: New data and significance for weathering. *Permafrost and Periglacial Processes*, 2, 103-112.
- Hall, K. and André, M-F. In Press. New insights into rock weathering as deduced from high-frequency rock temperature data: An Antarctic study. *Geomorphology,* 1059, . (Paper presented at the *"Weathering 2000"* Conference in Belfast)

★ **Moisture Data**

- Hall, K.J. 1986a. Rock moisture content in the field and the laboratory and its relationship to mechanical weathering studies. *Earth Surface Processes and Landforms*, 11, 131-142.
- Hall, K. 1988a. Daily monitoring of a block of indigenous rock at a Maritime Antarctic site: moisture and weathering

results. *British Antarctic Survey Bulletin*, 79, 17-25.

- Hall, K. 1991a. Rock moisture data from the Juneau Icefield (Alaska), and its significance for mechanical weathering studies. *Permafrost and Periglacial Processes*, 2, 321-330.
- Hall, K. 1993b. Rock moisture data from Livingston Island, (Maritime Antarctic) and implications for weathering studies. *Permafrost and Periglacial Processes*, 4, 245-253.
- Hall, K. 1995a. Rock moisture: The missing data in rock weathering studies, *In* B.Hallet and P. Black (eds): *Frozen Ground Workshop: Our Current Understanding of Processes and Ability to Detect Change. (Abstracts).* U.S.A Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire, 18.
- Hall, K., Verbeek, A., & Meiklejohn, K. 1986. A method for the extraction and analysis of solutes from rock samples and their implication for weathering studies : an example from the maritime Antarctic. *British Antarctic Survey Bulletin*, 70, 79-84.
- Meiklejohn, I. and Hall, K.1997. Aqueous geochemistry as an indicator of chemical weathering on southeastern Alexander Island, Antarctica. *Polar Geography*, 2, 101-112.

★ **Rock Properties**

- Hall, K. 1986b. The utilisation of the stress intensity factor (K_{IC}) in a model for rock fracture during freezing: an example from the maritime Antarctic. *British Antarctic*

Survey Bulletin, 72, 53-60.

- ♦ Hall, K.J. 1987a. The physical properties of quartz-micaschist and their relationship to mechanical weathering studies. *Earth Surface Processes and Landforms*, 12, 137-149.

★ **Weathering Rates**

- ♦ Hall, K. 1990'a. Mechanical weathering rates on Signy Island, maritime Antarctic. *Permafrost and Periglacial Processes*, 1, 61-67.

The first two papers are extensions of the work undertaken for an M.Phil. thesis at the University of Reading. Although the papers are based upon that research undertaking, the detail and discussion extend beyond that which was presented in the final thesis. The work is incorporated here, as information pertaining to thermal conditions at a nivation site, to add to the general background regarding thermal data and also with respect to the later discussions regarding nivation and its relationship to cryoplanation (Chapter 5). Thermal data were obtained from a variety of locations (Fig. 4); information from Livingston Island, Alexander Island, Adelaide Island (Antarctica) and the Canadian Rockies are cited here but other references (e.g. Walton and Hall, unpubl., Hall, 1999a (discussed in Chapter 4), and Hall, 1991a (given above under rock moisture data)) provide thermal data for Signy Island (Antarctica), Canada, and the Juneau Icefield (Alaska) respectively. These data offer direct information on rock temperatures from field situations and, in some instances (e.g. Hall, 1997a; Hall and André, In Press), include data at sufficient frequency (two minutes, one minute or 30 second intervals) as to allow evaluation of thermal stress fatigue/thermal shock. One importance of the data is that it provides a clear manifestation that air temperatures are no surrogate for rock temperatures. Long data records are important for indicating winter conditions (e.g. Hall, 1997a), while data from various aspects are crucial for understanding microclimatic variability at a site (e.g. Hall, 1998a). As fundamental as these data are, few cold region studies have obtained actual rock temperature data from the site(s) under investigation and even when these have been obtained they are frequently of short duration, of widely-spaced record intervals and do not show the spatial variability about the site. The data presented here are, so far, unique to cold region weathering studies and particularly for Antarctica.

As stated in the opening arguments, the monitoring of rock thermal conditions without a knowledge of the moisture status would not provide an adequate foundation for the evaluation of weathering, and of the role of freeze-thaw in particular. From the perspective of weathering in cold regions, no data whatsoever were available pertaining to interstitial rock water chemistry and almost none regarding rock moisture content. Thus an attempt was made to undertake monitoring of rock moisture content in the field, including its temporal and spatial variability, and to derive a method for the determination of rock moisture chemistry. Monitoring was undertaken primarily during the summer season (for logistical reasons) but, in one instance (Hall, 1988a), data were collected over a whole year. In addition, on several occasions, at different work sites, short-term changes in rock moisture content were monitored via frequent observations through a 24 hr period (e.g. Hall, 1991a). Some field data regarding rock moisture content were also obtained by means of ultrasonics (Hall, 1997a, see Fig. 5) but this was mainly restricted to laboratory undertakings (see Chapter 3).

A new technique was established for determination of interstitial rock moisture chemistry (Hall, *et al.,* 1986), while suggestions have also been proposed for a number of other new methods including the use of high pressure vacuum pumps plus industrial microwave units (Hall, 1995a). As a measure of chemical weathering, and to complement the work on the actual interstitial rock water chemistry, work was also undertaken regarding the solutes being removed from the weathering system (Meiklejohn and Hall, 1997). As spatially and temporally limited as these data are, when considering the complexity of cold regions, they, nevertheless, provide the first look at rock moisture chemistry and moisture content variability within the study area. In this capacity they provided the foundation for laboratory simulations and a basis for the evaluation of weathering processes operative in the study areas.

Two papers are identified as dealing specifically with rock properties, although much information on these attributes is disseminated in many of the other publications, including those dealing with laboratory simulations. The two papers cited cover the application of the stress intensity factor (K_{Ic}), from linear elastic fracture mechanics, to determine the sort of stress required to cause rock failure, as well as the rock temperatures and crack sizes that would be necessary. This approach provides a background against which the temperature data can be evaluated, to see if temperatures sufficiently cold to cause failure actually occur. The basis for much of the data utilised in the determination, together with further detail on porosity, permeability, frost susceptibility (S-value) and spatial variability of rock strength, was provided in Hall (1987a). Almost all the field studies cited in other chapters include some direct or indirect measures of rock properties from the field; these are complemented by the details provided in associated laboratory investigations.

The only paper dealing specifically with the determination of weathering rates is Hall (1990a). Logistical constraints prohibited the on-going monitoring (except in the case of Signy Island cited in Hall, 1990a) that would be required. Some papers (e.g. Hall, 1993a. or Hall, Subm. a) provide details regarding the spatial variability of weathering as deduced from such data as micro-indenter, Schmidt hammer measurements, or based on taffoni sizes and frequency of occurrence, but not of weathering rates *per se*. Discussion regarding weathering rates on Livingston

Island (South Shetland Islands, Antarctica) can also found in Hall, 1992a and in Hall, 1992b that are discussed in Chapter 4. The latter of the above two papers specifically deals with the difference in amount of weathering between north- and south- facing aspects.

This material thus provides a foundation of field information necessary for any investigation of weathering in cold regions. Not only does it provide detailed data regarding rock thermal and moisture conditions as well as information pertaining to rock properties but it also introduces new techniques for the determination of rock moisture chemistry. Data are provided that deal with Antarctic weathering rates and discussion regarding the meaning of these data is given. Without such data it would have been impossible to undertake meaningful laboratory simulations or to provide an objective basis against which to evaluate the processes associated with landform development in the areas studied.

Fig. 4
Collecting rock temperature data on Alexander Island (Antarctica)

Fig. 5

Collecting ultrasonic pulse velocity data in the field, Alexander Island (Antarctica

Arctic and Alpine Research, Vol. 12, No. 2, 1980, pp. 183-194

FREEZE-THAW ACTIVITY AT A NIVATION SITE IN NORTHERN NORWAY

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ABSTRACT

Monitoring of rock-face temperatures at various points within an arctic nivation site indicate that there is great variation in number, duration, and amplitude of freeze-thaw cycles experienced. The air-temperature record of freeze-thaw cycles provided no indication of temperature conditions at the rock face, because much of the nivation site was under a protective snow cover for a large period of time during which such oscillations were taking place. Available data suggest that the rock faces are subjected to freeze-thaw activity principally in the autumn to early winter period before they become covered by snow. Rock-face temperature records indicate wide fluctuations in subzero temperature conditions and it is suggested that these oscillations, across threshold values, may equate to crossing of the 0°C isotherm.

INTRODUCTION

In periglacial regions, physical weathering of rock is usually considered to be a major process in landscape development (Tricart, 1970: 113; French, 1976: 12). In nivation specifically, the role of rock disintegration is generally attributed to accentuated weathering by freeze-thaw action (Thorn, 1975, 1976, 1979; Thorn and Hall, in press, Table 1). However, despite the considered role of freeze-thaw activity in the nivation process there has been remarkably little quantitative investigation; Ives (1973: 1-2) has defined this as one of the topics currently in most need of study.

The actual mechanism of freeze-thaw destruction is still largely open to question (White, 1976), despite the number of landforms believed to result from this process (French, 1976). In particular, as noted by White (1976), there is the clear distinction in the efficacy of the mechanism between breaking apart previously split rock and initiating cracks in undamaged rock. However, a multitude of destructive mechanisms and controlling conditions have been suggested by laboratory investigations (Collins, 1944; Tricart, 1956; Wiman, 1963; Dunn and Hudec, 1966; Martini, 1967; Potts, 1970; Connell and Tombs, 1971; Fukada, 1971, 1972; Latridou, 1971; Brockie, 1972; Martin, 1972; Ritchie, 1972; Blaga and Yamasaki, 1973; Hudec, 1973; Mellor, 1973). Field measurements of freeze-thaw activity are extremely limited (Battle, 1960; Cook and Raiche, 1962; Hewitt, 1968; Gardner, 1969; Thorn, 1974; Hall, 1975) and in many instances offer information which conflicts with theoretical laboratory studies and subjective field evaluations (Thorn, 1976; Thorn and Hall, in press).

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0004-0851/80/020183-12$01.80

As part of a study of nivation processes in northern Norway (Hall, 1975), a detailed investigation of freeze-thaw activity and the mechanism involved was undertaken. During a 2-yr period, an attempt was made to monitor rock-face temperatures at various points within a nivation site. These data were then used to estimate the freeze-thaw conditions operative within the nivation process and their variation through the year.

STUDY AREA

An investigation was undertaken at a transverse snowpatch site on the north-facing slope of Austre Okstindbre valley, Okstindan, northern Norway (Figure 1). Located approximately 80 km from the sea, the area experiences a relatively maritime climate. Extrapolation from the nearest meteorological station, Hattfjelldal (380 m a.s.l.) some 55 km to the south, suggests a mean annual temperature at Okstindan of between −2 and −4°C, annual precipitation approaching 1500 mm, and snow cover on approximately 180 d yr^{-1} (Worsley and Harris, 1974; Worsley and Ward, 1974). At the Hattfjelldal station, mean January and July temperatures (1963 to 1970) were −9.6°C and 10.6°C, respectively (Harris, 1974), but these may be lower at Okstindan due to its higher altitude.

The transverse snowpatch (Figure 2) is located at approximately 800 m above sea level

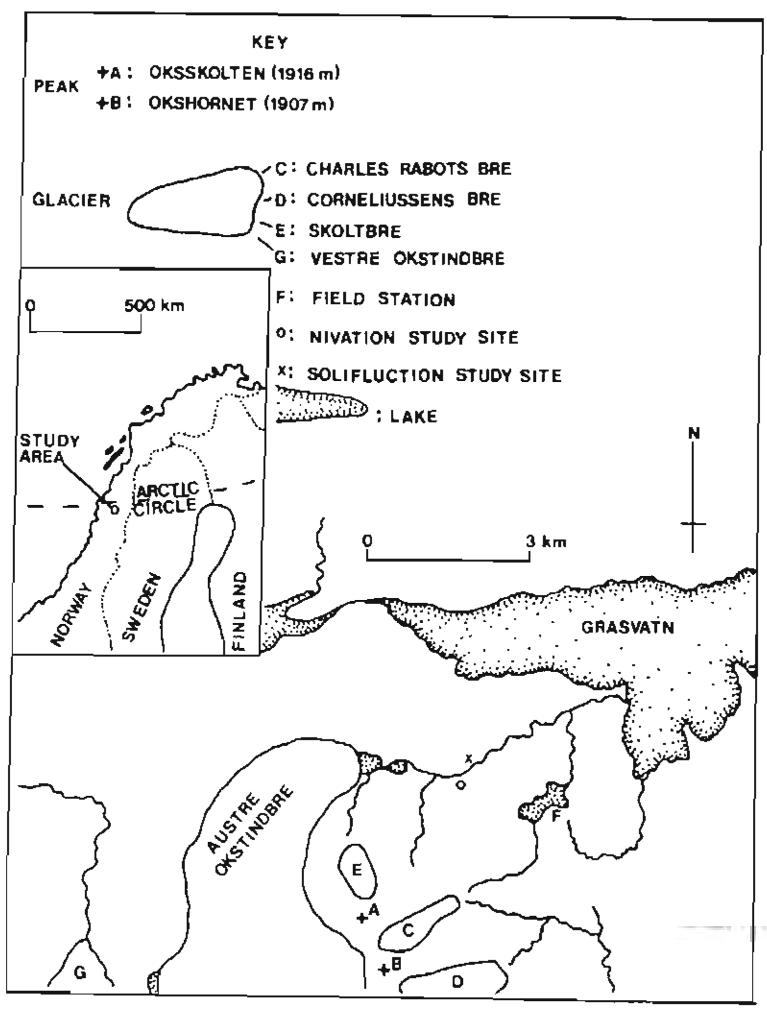

FIGURE 1. Location of study area.

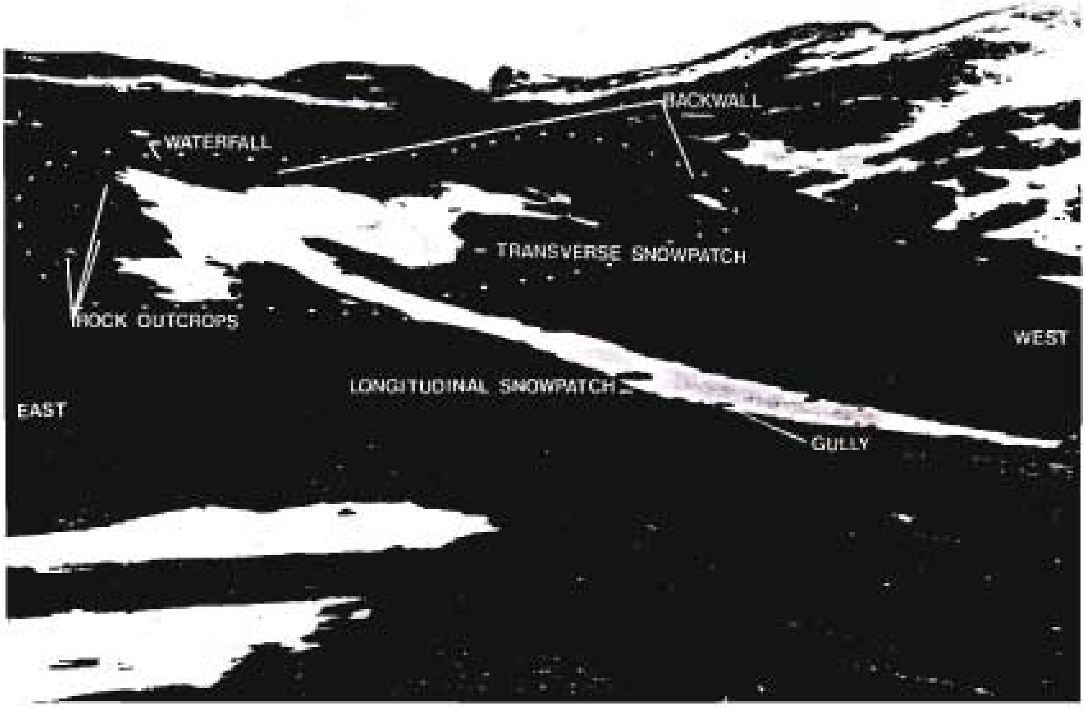

FIGURE 2. View of the study site (July 1973) indicating the major features described in the text. The dotted outline denotes the approximate position of the transverse snowpatch at maximum extent.

FIGURE 3. The study site during 1972 (5 August) when the snow had almost completely ablated, showing the location of thermistors A to G.

and is 150 to 180 m along its maximum cross-slope extent and 50 to 100 m downslope. It is backed by a north-facing cliff up to 18 m high, composed of gneiss and schist, cut through at its eastern end by a waterfall which continues down a gully to the valley floor. A longitudinal snowpatch occurs along this gully (Figure 2). A number of small, garnetiferous mica-schist and gneiss outcrops are found in the stone pavement area on the eastern side of the gully which demarcate the lateral extent of the snowpatch in that direction (Figure 2). The vegetation-free stone pavement describing the main snowpatch position (Figure 2) gradually gives way downslope to an increasing vegetation cover (Figure 3). The vegetation is dominated by billberry (*Vaccinium myrtillus*), dwarf birch (*Betula nana*), and *Carex* spp. No evidence of permafrost was found at any point within the nivation site.

It is thought that the nivation site was initiated as a structural depression in the north-facing valley wall. The increased accumulation area plus drifting of snow from the rock bench above and the north-facing aspect result in a location where snow cover persists after that over most of the surrounding area. The backwall and rock outcrops which margin the snowpatch show varying degrees of jointing and fissility. Along the backwall there also appears to be some evidence for dilatation jointing, parallel to the valley wall. This is probably a result of the effects of erosion and loading by the outlet glacier which occupied the valley during the last glaciation.

INSTRUMENTATION

Three Grant model D temperature recorders with a combined total of 60 thermistors were used to monitor temperatures at various points above, around and beneath the transverse snowpatch. The instrument accuracy, according to the manufacturers, is considered to be within ±0.6°C. During the spring to autumn period, when field workers were available to change recording charts, temperatures were monitored every 2 h. During the winter period recordings were obtained at 6-h intervals in an effort to conserve chart and battery life. Due to a combination of instrument failure and loss of battery or chart life, records were obtained only for the periods 29 August to 25 October 1972, 30 May to 8 December 1973, and 6 July to 10 September 1974, together with occasional short periods during early 1973 and 1974.

Thermistors were located in a variety of positions in an attempt to monitor temperature changes quantitatively at the rock surface; the location of those described in this study is shown in Figure 3. Along the backwall, disc-type thermistors, with a surface coloring similar to that of the rock, were cemented such that their unobstructed sensor-faces, directed outwards, were flush with the rock surface. Thermistors were also cemented into cracks and basal niches, and onto the rock faces of the outcrops at the eastern extremity of the study site. Thermistors were left unshielded so that they would monitor temperature variations resulting from all climatic elements incident at each location. Thus it was possible to compare upward facing surfaces in the lower part of the study area, which were warmed by direct insolation, with locations on the vertical rock faces and at the back of basal niches, which were not so warmed.

Thermistor probes, shielded against insolation, were used to obtain air temperatures close to the ground surface, and at 1 and 2 m above the ground, near the lower edge of the eastern stone pavement area (Figure 3). An unshielded probe was used, close to the ground surface, to monitor the effect of insolation at the same location. In addition, at both the study site and the nearby Field Station (Figure 1), a thermograph, in a Stevenson screen, was used to obtain a continuous record of air temperatures during the period when personnel were in the field.

FREEZE-THAW ACTIVITY

At the end of the winter, the backwall and rock outcrops are completely beneath an insulating snow cover which gradually thins during the spring-to-summer ablation period. During spring, monitoring of air temperatures indicates many freeze-thaw cycles

(Table 1). It is, however, suggested that the geomorphic effectiveness of these cycles is limited at the nivation site because the thick snow cover supresses air-temperature fluctuations prior to their reaching the rock surface. Limited data obtained from the study area during March, April, and May failed to indicate either positive temperatures or freeze-thaw activity at the snow-covered rock faces, but the information is sparse and possibly cycles may have been missed. Rock face temperatures recorded varied between −2.3 and −0.7°C. In a study of solifluction processes a few hundred meters to the northeast (Figure 1) both ground and air temperatures were monitored (Harris, 1972: Figure 5; Harris, 1974: Figure 5). The data and interpretation presented by Harris (1972, 1974) indicate that during the period the area was snow covered, freeze-thaw cycles monitored in the air were not mirrored by cycles at the ground surface; the snow cover had damped out the oscillations. This is the same situation as is suggested, with limited data, for the nivation site rock faces.

As summer progresses, air temperatures rise, and the frequency of freeze-thaw cycles decreases until mid-June when temperatures are at their highest (≤28°C) and only rarely, if ever, are freeze-thaw cycles monitored (Hall, 1975: Figure 4-18). Consequently, as the rock faces emerge from the ablating snow cover during late June and early July they do not experience a period of intense freeze-thaw activity but rather a transition from near-zero to high positive temperatures (Figure 4). A similar situation was observed by Harris (1972, 1974) with respect to the emergence of the ground from beneath the melting snow cover.

During the latter stages of snow-cover thinning, the nivation site rock-surface temperatures oscillate about 0°C (Figure 4). However, the freeze and thaw amplitudes are only in the order of 0.2°C, and so are probably of limited geomorphological significance. As the rock becomes snow free, surface temperatures rise rapidly and during the summer to early autumn period remain positive; those surfaces affected by direct insolation may attain temperatures in excess of 20°C. Unfortunately, no method of sensing temperatures at depth within the rock was utilized, and so information on heat penetration is unavailable.

As autumn progresses, temperatures gradually begin to fall and the rock surfaces experience freeze-thaw activity (Figures 4 and 5). In general terms, the period during which a point on the rock surface experiences freeze-thaw activity is dependent upon how long it remains snow free. Freedom from snowpack insulation is a function of how soon and to what extent snow falls and the location of the point on the rock face relative to the buildup of the snowpack. That is to say, a horizontal surface will, if not kept clear by wind action, obtain a protective cover earlier in the autumn than a point on a vertical face where the controlling factor for snowpack buildup is height above the ground. In addition, those areas subject to direct insolation will experience

Table 1

Summary of the air temperature freeze amplitude and wavelengths recorded during the period May to July 1973

Freeze Amp. (°C)	Freeze wave (h)	Freeze Amp. (°C)	Freeze wave (h)
−8	12	−5	10.75
−1	1	−4.5	10
−1.5	1	−1	5
−8	14.5	−8	39
−2	6.5	−9	11
−4	19	−4	11
−2	3	−3	9
−6	13	−1.5	10
−2	1	−0.5	3
−0.5	1	−4	8
−3.5	21	−3	8
−2.5	1	−2	13
−6	26.5	−2	10
−7	15	−3	3
−5	?	−2	5
−2	3	−1	7
−8	14	−2.5	8
−0.5	1	−1	3
−4	11	−1.5	8
−1.5	3	−1.5	31
−0.5	1	−2.5	25
−0.5	1	−4.5	131
−1	1	−1.5	16
−1	1	−2	10

n = 48.
$\bar{x}$ freeze temperature = −3.1°C (s = 2.37).
Total hours of freeze = 566.25.
Definitions:
Freeze Amp.—Maximum freeze amplitude
Freeze Wave—Maximum freeze wavelength

more rapid snowmelt than shaded areas. Within the nivation site these factors, together with the microtopography, result in a great variation between exposures of number, amplitude, and wavelength of freeze-thaw cycles at the rock surface.

Figure 5 shows the temperature curves obtained on four rock outcrops at the eastern end of the stone pavement area (Figure 3). Curves A and B, obtained from the vertical faces of the outcrops, appear similar while C, from a basal niche, is much subdued, and D, from a horizontal surface, indicates a radical difference in temperature conditions during the latter part of the record. Specific details of the number, wavelength, and amplitude of the freeze phases illustrated in Figure 5 are given in Table 2. It can therefore be seen that there are very big differences in freeze-thaw temperature conditions among these four points.

Locations A and B both experienced nearly the same duration of freeze time, but B experienced one more freeze-thaw cycle than A and had 40% overall lower freeze amplitude (Table 2). Location C experienced approximately 25% fewer freeze-thaw cycles than A or B, nearly 100 h less total freeze time and had the smallest mean freeze amplitude of all four locations. The thaw phases appear fairly similar between C and A or B (Figure 5) but the freeze phases do not, those of C being greatly subdued with respect to either A or B. This difference is believed to result from the thermistor at location C being situated at the very back of a semienclosed basal niche where there would be a greater stored body of heat to be overcome than on the exposed vertical face.

Location D, a horizontal surface (Figure 3), indicates the effect of snow cover. From 20 October (Figure 5) the snow on top of this site appears to have remained at a sufficient thickness to effectively damp out the freeze-thaw cycles experienced at the other points such that, for the duration of the record shown, location D experienced 57% fewer freeze occurrences and 60% less total freeze time than location A. At the end of the available record, curve D is approaching 0°C (±0.2°C), and data obtained in the spring of the following year indicate a rock surface temperature at this point of −1.1°C. While the dampening effect of the snow cover at point D is important with respect to geomorphological implications, it is important to note that during the period D was snow free it experienced the

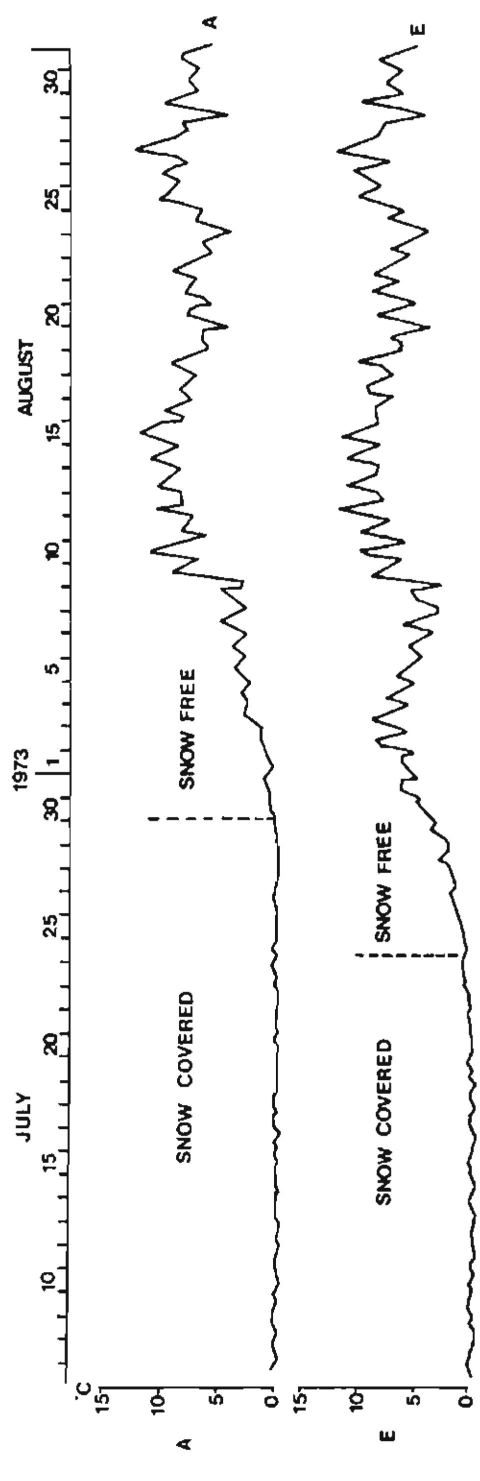

FIGURE 4. Six-hourly rock surface temperature records for thermistors A and E during the spring to summer ablation period.

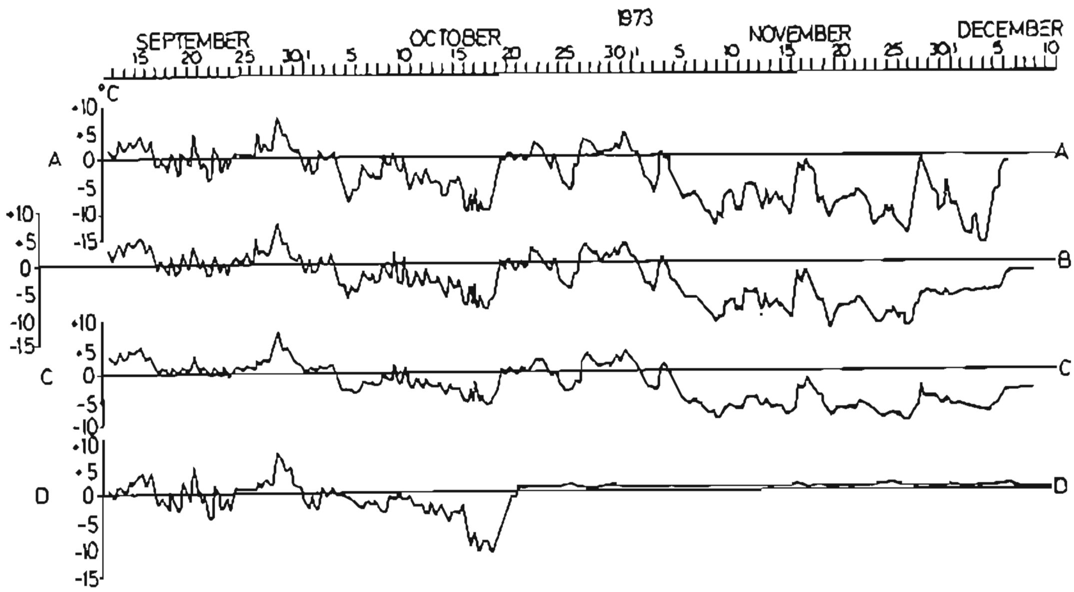

FIGURE 5. Six-hourly rock surface temperature records for thermistors A to D, located on the rock outcrops, during the autumn to winter period.

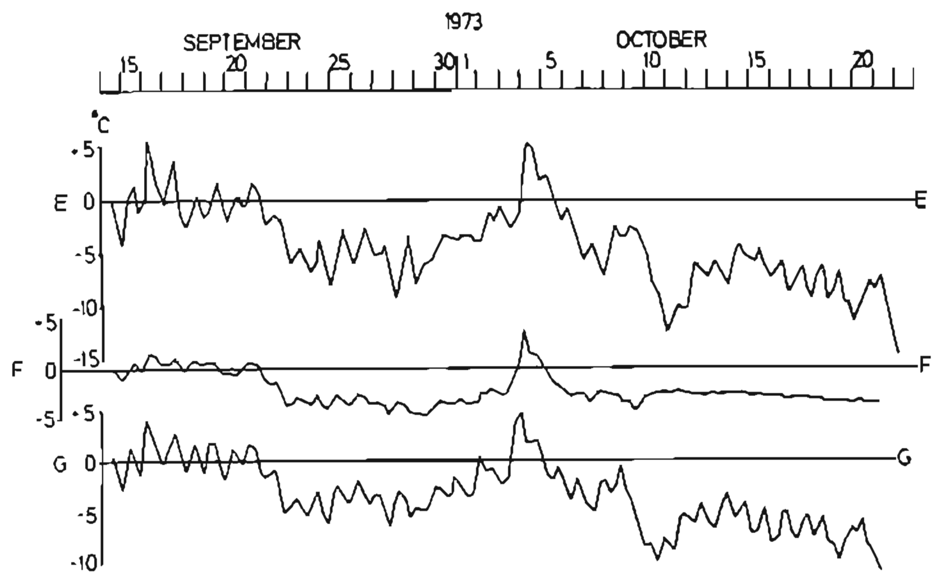

FIGURE 6. Six-hourly rock surface temperature records for thermistors E to F, located on the backwall, during the autumn to winter period.

largest mean freeze amplitude and the longest total freeze time of the four points, although not the largest number of individual freeze-thaw cycles (Table 3).

Figure 6 shows the temperature curves obtained from three locations on the backwall (Figure 3). Locations E and G are on the vertical face while F is at the base of a vertical crack 0.2 m wide, 1.8 m long, and 0.45 m deep. As with the rock outcrops, the three positions on the backwall show variation in number, duration, and amplitude of freezes

Table 2

Freeze occurrences monitored by thermistors A to D on rock surface outcrops during the period 12 September to 10 December 1973

A		B		C		D	
Amp (°C)	Wave (h)	Amp (°C)	Wave (h)	Amp (°C)	Wave (h)	Amp (°C)	Wave (h)
- 2.0	36	- 1.0	12	- 0.2	6	- 0.8	18
- 3.2	18	- 1.3	18	- 0.3	12	- 3.5	36
- 1.3	12	- 1.5	18	- 0.3	12	- 4.2	24
- 4.5	36	- 0.2	12	- 0.9	12	- 1.7	12
- 2.9	12	- 0.6	6	- 0.2	6	- 5.0	36
- 3.2	18	- 2.0	18	- 4.2	132	- 3.2	30
- 2.7	18	- 1.3	12	- 2.8	24	- 3.5	36
- 3.6	24	- 1.9	6	- 6.8	216	- 1.1	18
- 0.8	18	- 1.8	12	- 1.1	12	- 11.4	402
- 9.0	108	- 1.4	18	- 0.9	12	n = 9	
- 2.0	18	- 6.3	108	- 0.2	6	$\bar{x}$ Amp = -3.8°C	
- 11.0	240	- 3.2	18	- 1.5	18	(s = 3.17)	
- 1.0	8	- 3.9	18	- 4.2	60	Total freeze = 612 h	
- 0.9	12	- 8.7	216	- 4.0	48		
- 0.6	12	- 1.2	12	- 9.3	840 +		
- 3.0	24	- 1.1	18	n = 15			
- 6.9	60	- 0.5	12	$\bar{x}$ Amp = -2.5°C			
- 7.4	48	- 2.2	18	(s = 2.76)			
- 14.6	576	- 5.0	60	Total freeze = 1416 h			
- 16.4	204 +	- 5.5	48				
n = 20		- 12.0	840 +				
$\bar{x}$ Amp = -4.9°C		n = 21					
(s = 4.63)		$\bar{x}$ Amp = -3.0°C					
Total freeze = 1502 h		(s = 3.01)					
		Total freeze = 1490 h					

Definitions:
Amp (°C) = Maximum freeze amplitude.
Wave (h) = Maximum freeze wavelength.

(Table 4) during the record available. Locations E and G are fairly similar in response to temperature changes, but location F experiences 33% fewer freeze-thaw cycles and a smaller mean freeze amplitude. Detailed consideration of curves E and G show that there is 25% difference in mean freeze amplitude and that G experienced one more freeze-thaw cycle while E had 42 h more total freeze time.

The quintessence of the data is that the freeze-thaw environment within the nivation site is complex and shows great variability. The situation becomes even more complex when temperature curves for the downslope outcrops (Figure 5) and the backwall (Figure 6) are compared for their overlap period (14 September to 21 October). Clearly, the backwall and outcrops do not experience the same regime, even within an area as small as that under study, due to the variability of microrelief and exposure to the climatic elements. The rock outcrops did not register the two cold phases of 21 September to 3 October and 5 October onwards which were so evident on the backwall (Figure 6). Conversely, the earlier fluctuations appear to have been more severe on the outcrops than the backwall. During the two main cold periods experienced on the backwall, the outcrop surfaces in fact indicate peaks of warm temperatures; at the time of these peaks there was a complementary rise of backwall temperatures, although they remained negative (≤ -3°C). Other data show that during the same period ground-surface temperatures in the lower area did not remain negative but experienced warm peaks similar to the rock outcrops. Thus, in this instance, it is suggested, the

TABLE 3
Summary of freeze occurrences monitored by thermistors A to D for the period (14 September to 22 October) during which location D was snow free

	A	B	C	D
Number of freezes	14	16	9	9
$\bar{x}$ freeze temperature	−3.4°C	−2.3°C	−1.9°C	−3.8°C
s freeze temperature	3.02	2.25	2.31	3.17
Number hours of freeze	578	522	432	612

TABLE 4
Freeze occurrences monitored on the backwall by thermistors E to G during the period 14 September to 22 October 1973

E		F		G	
Amp (°C)	Wave (h)	Amp (°C)	Wave (h)	Amp (°C)	Wave (h)
−4.5	24	−1.2	18	−3.0	24
−1.4	18	−0.2	6	−1.8	6
−0.5	6	−0.3	6	−0.3	6
−2.6	18	−0.9	30	−1.3	6
−1.8	24	−4.6	306	−1.1	12
−2.2	18	−4.2	402	−2.2	12
−0.7	12			−0.4	6
−9.2	294			−6.7	258
−14.0	396			−2.4	36
				−11.0	402

Mean freeze amplitude at E = −4.1°C (s = 4.57); n = 9.
Mean freeze amplitude at F = −1.9°C (s = 1.98); n = 6.
Mean freeze amplitude at G = −3.0°C (s = 3.34); n = 10
Total hours of freeze at E = 810.
Total hours of freeze at F = 768
Total hours of freeze at G = 768
Definitions:
Amp (°C) = freeze amplitude (°C).
Wave (h) = freeze wavelength in hours.

higher outcrop temperatures are a result of heating by direct insolation (as with the ground), while the lower backwall temperatures derive from the continual shadowing effect there.

DISCUSSION

Two principal aspects of the freeze-thaw regime emerge: first, monitoring of air temperatures is no indication of temperature fluctuations occurring at the rock face and, second, that within a nivation site very large differences in freeze-thaw regime occur from point to point. Unfortunately, many investigations of freeze-thaw activity are based upon meteorological station air-temperature records which often imply a rigorous regime with a spring peak (Fraser, 1959). As has been shown, however, a nivation site is very likely to be protected by a thick snow cover during the spring period and so the geomorphological repercussions of the air-temperature fluctuations appear to be minimal. At the nivation site, rock outcrops lost their snow cover later than those in the surrounding area, and available subsnow temperature data do not appear to indicate freeze-thaw activity during spring. Thus it is suggested that, despite the large number of freeze-thaw cycles monitored in the air, the spring period is not an active time for freeze-thaw oscillations at the rock surface. Although data pertaining to temperature conditions and variations at depth within the rock are absent, the information on rock-face temperatures does offer a starting point for the consideration of freeze-thaw rock destruction comparable with most other field (e.g., Gardner, 1969) or laboratory (e.g., Potts, 1970) investigations.

Within-site variation of freeze occurrences during the autumn and winter, prior to the establishment of the seasonal snow cover, produces a complex problem. The observed character of the freeze-thaw regime at the nivation site is dependent upon where temperatures are monitored. Maximum freeze amplitude has been considered by some as one of the main factors in the destructive mechanism (Grawe, 1936) and yet this varies between −14°C for location E and −4.2°C for location F on the same date (22 October)—a 70% variation. Similarly, consideration of the last freeze recorded for locations A and C gives a 7.1°C difference in amplitude. If number of freeze occurrences is considered a major factor in freeze-thaw activity (Schaffer, 1932; Thomas, 1938) then monitoring of location D instead of B would have suggested 9 subzero oscillations instead of 21. Comparable variations for duration of freeze also occur (Tables 2 and 4). Thus, whether amplitude, duration, or number of freezes is considered to be the most important factor in freeze-thaw rock destruction, the estimation of activity for a particular nivation site could be greatly in error if only limited observations were obtained.

It is apparent that freeze-thaw activity at a nivation site is extremely complex and that the widely held generalizations of its place within nivation are in need of review. It is a difficult problem and variables other than temperature fluctuations must be considered. It is a paradox that the uppermost sections of rock faces (those first to be exposed from beneath a melting snow cover and last to be snow covered) which experience the greatest number of freeze-thaw cycles should show less rock destruction than occurs at the rock base (the first covered and last exposed). Nevertheless the extensive basal niches with their apron of rock debris show that destruction is more potent at the rock base. This phenomenon is perhaps related to moisture availability, but it still begs the question, as expounded by White (1976), whether the destructive mechanism is freeze-thaw or, perhaps more likely, hydration shattering.

While the available data are inadequate for detailed consideration of the rock-breakdown mechanism, a number of comments and observations pertaining to freeze-thaw activity can be made. The term "freeze-thaw" implies temperature fluctuations across the 0°C isotherm, but various minimum necessary freeze amplitudes, for the freeze to be effective in terms of rock destruction, have been suggested by laboratory investigations: −5°C (Collins, 1944), −4°C (Fukada, 1971), −5°C (Fukada, 1972), −5°C (Latridou, 1971), and −6 to −10°C (Dunn and Hudec, 1966). Consideration of Figure 5 shows that towards the ends of curves A, B, and C temperatures fluctuate from strongly negative (approx. −14°C) to only just negative (−0.8°C). Thus there are fluctuations across the various threshold temperatures cited above although no positive temperature is attained. This implies the possibility that a freeze-thaw effect is taking place with crossings of the considered boundary value during the period that temperatures remain below 0°C. The implications of this are twofold: first, that a negative crossing of 0°C must attain the critical value before it can be considered an effective freeze and, second, that the temperature need not go positive to effect the freeze-thaw action.

CONCLUSIONS

Qualitative impressions of the freeze-thaw regime at a nivation site are in need of reconsideration. Spring has usually been postulated as the most active time in terms of rock breakdown but evidence available from this study suggests the autumn-winter period to be more important. Air temperature records are limited indicators of temperature conditions prevailing at the rock face and so are of doubtful use in judgment of geomorphological activity. Within the nivation site there is a great range of localized freeze-thaw environments such that great care must be exercised in interpretation when information is available from a limited number of points. The possibility that a freeze-thaw action is taking place without positive crossings of the 0°C isotherm needs more detailed investigation and, if proven, adds a further complicating factor to the interpretation of temperature records.

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Some observations on ground temperatures and transport processes at a nivation site in northern Norway

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Ground temperature data, at various depths, beneath and around a transverse snowpatch at a nivation site in northern Norway are presented. This information is integrated with observations of debris movement, of which four main types are identified. A temporal and spatial continuum of transport processes from spring to early winter is suggested. Overall, it is hypothesised that enhanced transport may be characteristic of nivation sites.

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Introduction

Nivation is a collective term which encompasses all aspects of weathering and transport which take place within the presence of late-lying snow (Matthes 1900) and is a process which has been widely postulated as a major agent in landscape development (Tricart 1970, p. 113; French 1976, p. 12). Despite its apparent pervasion throughout past and present periglacial areas, the limited number of studies undertaken on this topic (see Thorn & Hall 1980, Table I) have been largely qualitative. The recent quantitative work on nivation (Thorn 1974, 1975, 1976, 1979a, 1979b, Thorn & Hall 1980, Hall 1975, 1980) has questioned some of the basic tenets relating to the role, timing, interrelationship and results of the processes assumed to be involved.

In the context of transport within nivation, available information (Thorn & Hall, 1980, Table I) indicates that, with few exceptions, solifluction and sheetwash are deemed to be the major agents. However, the interrelationship and specific timing of these two processes, certainly within the Alpine context, is considered to be largely in doubt and mainly conjectural (Thorn 1979a). Whilst detailed work on mass wastage and meltwater activity within periglacial regions is extensive (see Embleton & King (1975) and Washburn (1979) for review of literature), little research on these topics has been undertaken at nivation sites despite the apparent close association. Integral with the consideration of the transport processes is the investigation of ground temperatures which, again, are limited with respect to nivation sites, Thorn (1979b) being the only major study freely available.

In this study, ground temperatures, at various depths, beneath and around the snowpatch during periods of accumulation and ablation, together with observations and measurements of transport activity, are integrated to give some assessment of the overall process combination. The study, although lacking in detail in certain sections, was an attempt to follow the temporal and spatial variations of processes and so to develop a comprehensive picture of nivation transport activity.

The study area

Investigation of the conditions and processes in association with a semi-permanent snowpatch were undertaken on the north-facing slope of Austre Okstindbre valley, Okstindan, Northern Norway (Fig. 1). The study site is located approximately 80 km from the coast and thus experiences a realtively maritime climate. Extrapolation from the nearest meteorological station Hattfjelldal (380 m a.s.l.), which is situated some 55 km to the south (Worsley & Harris 1974; Worsley & Ward 1974) suggests a mean annual temperature at Okstindan in the range of $-2°$ to $-4°C$, annual precipitation approaching 150[illegible] mm and a snow cover of approximately 18[illegible] d yr^{-1}. The mean January and July temperature at Hattfjelldal were $-9.6°C$ and $10.6°C$ respectively (Harris 1974), but these may be lower at Okstindan due to the greater altitude.

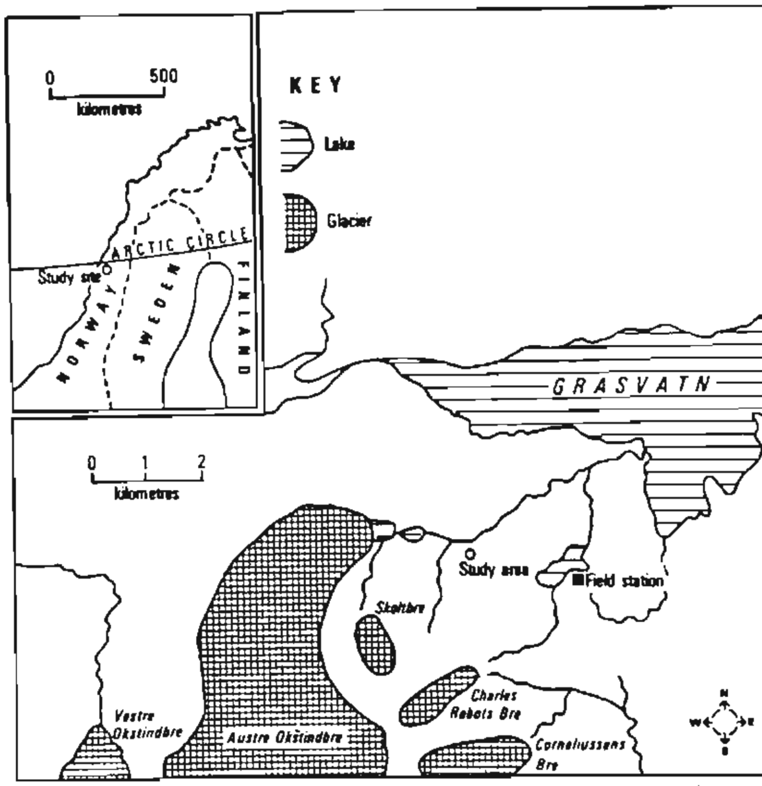

Fig. 1. Location of study area.

The study site comprises two snowpatches (Fig. 2), an upper transverse patch at about 800 m a.s.l. and a longitudinal patch running down slope from the transverse to the valley floor. Within the context of this study only the transverse patch is pertinent, as the longitudinal snowpatch resides in a rock gully occupied by a stream in summer. The transverse patch is 150–180 m in cross-slope extent, and 50–100 m downslope.

The snowpatch is backed by a north-facing gneiss and schist cliff up to 18 m in height which is cut at the eastern end by a waterfall, the water from which flows down the rock gully to the valley floor (Fig. 2). At the eastern end of the snowpatch, and at the western end to a lesser degree, are found a number of small outcrops composed of gneiss or garnetiferous mica-schist, which demarcate the maximum extent of the snow (Fig. 2). These outcrops vary in height from approximately 0.70 m to 2.5 m. The area in which the snowpatch resides comprises a stone pavement, which shows an increase in vegetation cover with distance downslope. The vegetation is dominated by billberry *(Vaccinium myrtillus)*, dwarf birch *(Betula nana)* and *Carex spp.*

Instrumentation

Temperature conditions at the study site were monitored via 3 Grant model 'D' temperature recorders and a varying number of thermistors. According to the manufacturers, the instrument accuracy was in the order of ± 0.6°C. Temperature observations were taken at 2-hourly intervals when personnel were in the field (usually spring to autumn), and at 6-hourly intervals during the winter in order to conserve chart and battery life. Thermistors were located at 0.05, 0.1, 0.25, 0.5, 0.75 and 1.0 m depth in the ground just below the lower edge of the snowpatch (Fig. 2), and at 0.05, 0.1, 0.25 and 0.5 m depth at the top of the backwall, i.e. above the snowpatch site. A further 6 thermistors were situated at 5 mm depth across the area usually covered by the snowpatch with additional thermistors, also at 5 mm depth, located next to ground movement devices (see below). Shielded thermistors were used to obtain air temperatures at 1 m and 2 m above the ground and grass minimum temperature in the area just below the snowpatch (Fig. 2). At the same location an unshielded thermistor was utilized to monitor the effect of insola

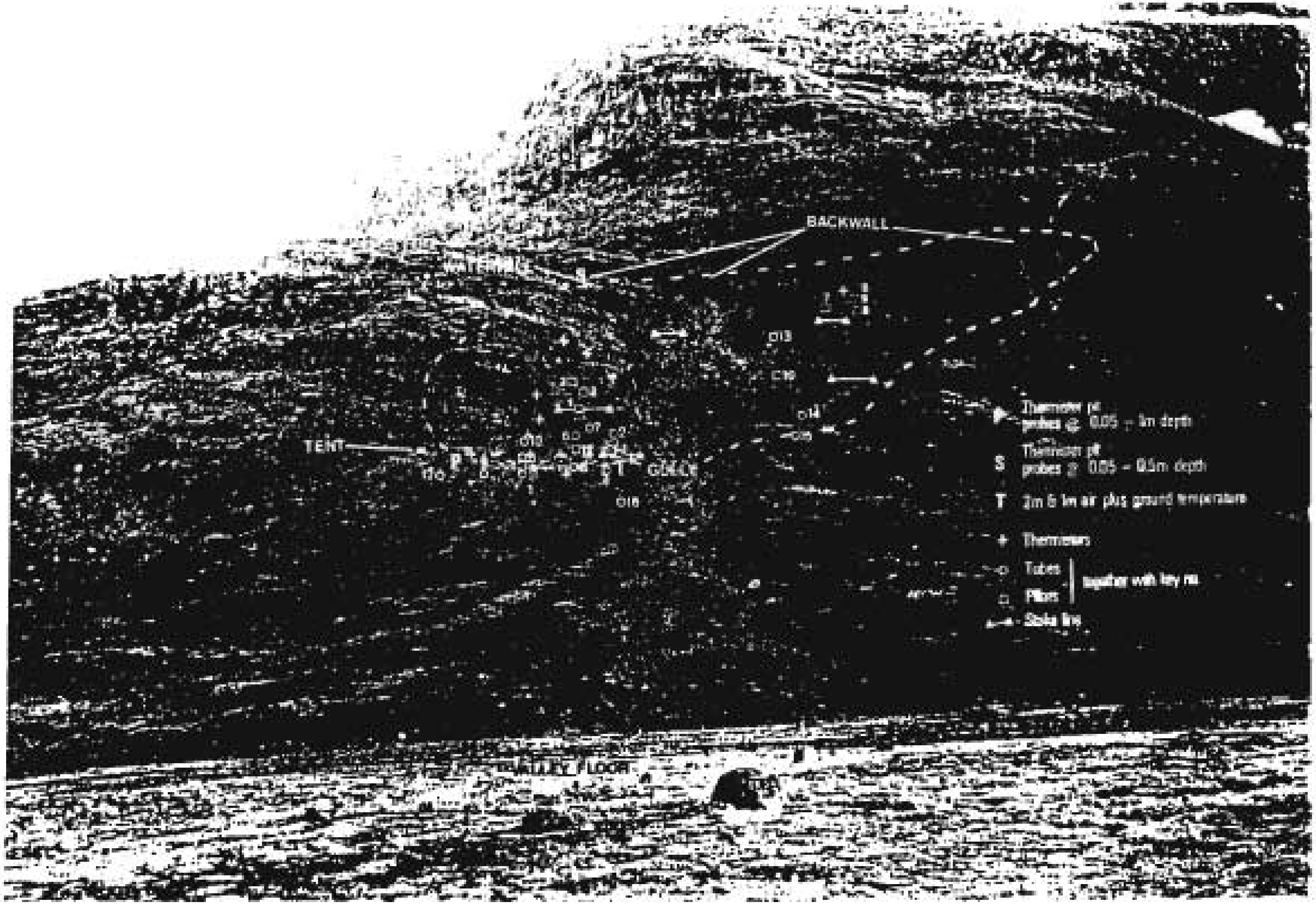

Fig. 2. Study area during a snow free period, showing location of monitoring instruments.

Table I. Temperatures at various depths during ground thaw (without snow cover).

Date	Depth 0,05 m	0,10 m	0,25 m	0,50 m	0,75 m	1,00 m
June 10	+0,2	-0,2	-0,3	-7,7	-7,3	-8,0
12	+0,7	-0,1	-0,2	-7,3	-7,3	-7,9
14	+1,3	-0,3	-0,3	-6,8	-7,3	-7,5
16	+1,0	-0,1	-0,2	-5,5	-6,0	-6,5
18	+2,6	+1,6	-0,3	-4,0	-4,0	-4,4
20	+4,3	+3,9	-0,3	-3,7	-3,8	-3,9
22	+9,0	+7,0	+0,7	-4,0	-3,2	-4,4
24	+9,1	+6,5	+3,3	-5,0	-5,0	-5,3
26	+8,0	+5,0	+2,3	+4,0	-4,6	-4,7

Table II. Cubic regression analysis of temperatures at various depths.

Depth (m) x	y	Result r	r^2
0,05	0,10	0,98	0,97
0,10	0,25	0,82	0,67
0,25	0,50	0,87	0,76
0,50	0,75	0,89	0,79
0,75	1,00	0,88	0,77
0,50	0,25	0,54	0,29
0,75	0,50	0,98	0,96
1,00	0,75	0,98	0,96

Equation of the form:
$y = a + bx + cx^2 + dx^3$

All results significant at 0,05 level (t-test of r)

tion. Rock-face temperatures were also obtained (Hall 1980), but these are not directly relevant to this study. Finally, at both the study site and the nearby Field Station (Fig. 1), a thermograph was used to obtain a continuous record of air temperatures during the periods personnel were in the field.

Downslope movements of the ground were monitored by three main methods, namely lines of wooden stakes, Rudberg Pillars, and deformable tubes. Sixteen deformable tubes, each 0.7 m in length, were enplaced in the lower area. These tubes allowed continual monitoring (Thorn & Hall 1980, p. 115) without recourse to their removal, as was the case with the 12 sets of Rudberg Pillars. The pillar sets were of varying length, dependent upon possible depth of penetration into the ground, and consisted of 2 cm high blocks of polythene tube. These were situated primarily within the stone pavement area. Six lines of wooden stakes were set up at various locations near the foot of the snowpatch to monitor both downslope movement and ground heave. Figure 2 shows the approximate location of all the temperature and ground movement monitoring devices.

Soil samples were analysed for their grain-size distribution. Samples of snow, rain and run-off were investigated by means of an atomic absorption spectrophotometer for their solute content.

Observations

As an aid to clarity, the data will be considered in five main sections and then resolved into a 'unit' within the ensuing discussion. This division into individual process groups is not meant to suggest mutual exclusivity, for, as shall be seen later, they are all part of a spatial and temporal continuum.

Temperature conditions

Record of ground thaw with depth is limited as a result of either instrument or battery failure. That record which does exist indicates a warming from both the surface and beneath. Information of temperature variation with depth, in two-day intervals (Table I), was subject to regression analysis with, initially, the point below considered to be dependent (the y value) upon temperature changes at the point above (the x value). A cubic regression gave the best fit (Table II) for the data set. However, whilst the uppermost 0.25 m clearly showed the effect of warming from above, the highest regression coefficients obtained for the 0.5 to 1.0 m levels were based upon warming from below (Table II). Thus, for the limited period available, it would appear that warming of the ground takes place from both above and below, but that the ground actually thaws from the surface downwards.

Temperatures obtained from a depth of 0.05 m in the ground beneath the ablating snowpatch (Fig. 3) show, as might be expected, thawing of the ground at the snowpatch edge as the snow recedes. Temperatures are seen to oscillate about 0°C (Fig. 4), but the fluctuations are only

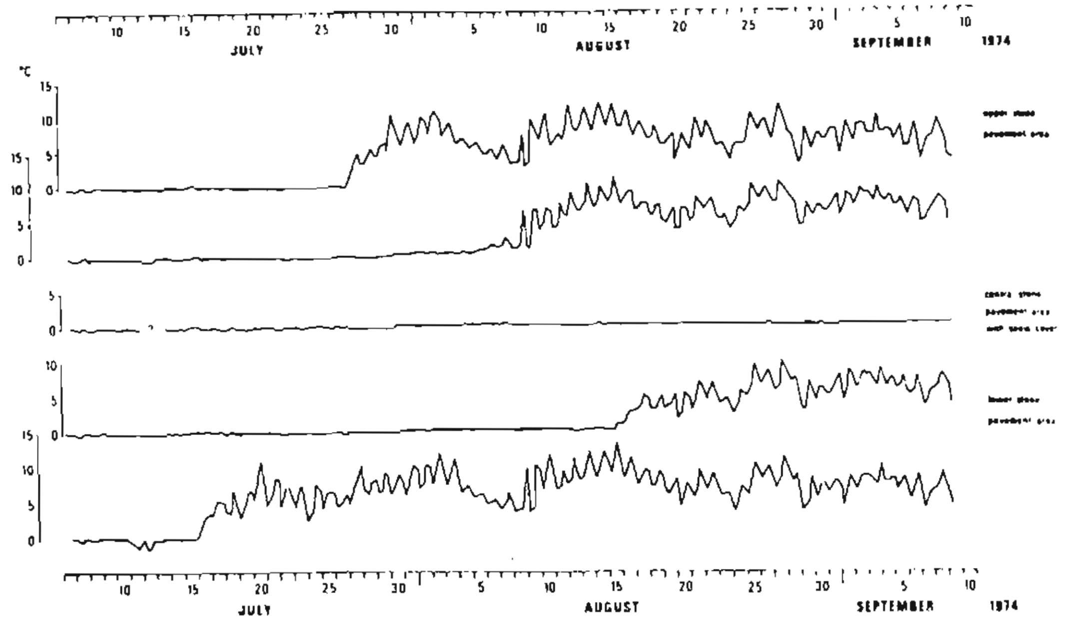

Fig. 3. Temperature curves for ablation phase in stone pavement area.

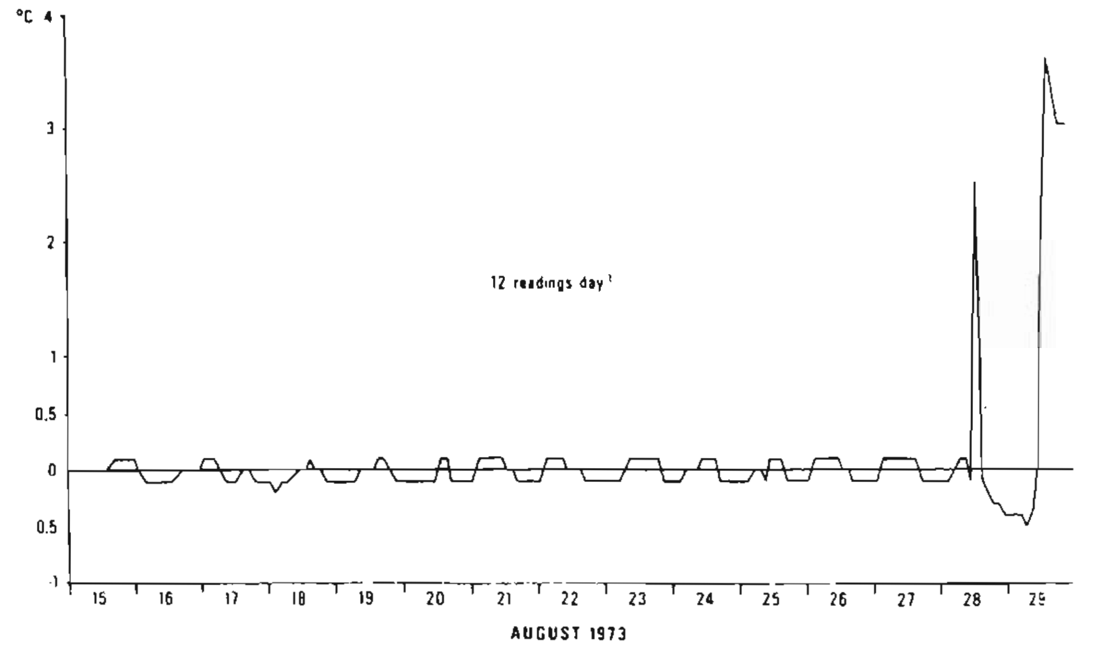

Fig. 4. Thermistor readings at snow-edge showing minor oscillations about 0°C.

in the order of ± 0.1°C and so are not thought to be geomorphologically significant. What is considered to be important is that the observed temperatures, even beneath the central part of the snowpatch which did not ablate completely, are conducive to allowing the movement of water without hindrance due to freezing. However, thawing of the ground at depth appears only to take place once that location has become snow-free.

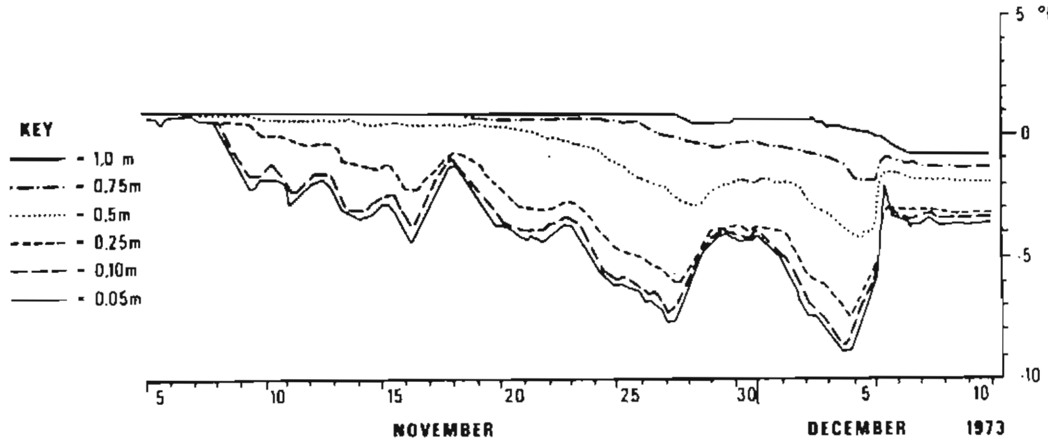

Fig. 5. Temperature profiles showing winter freezing of ground with depth.

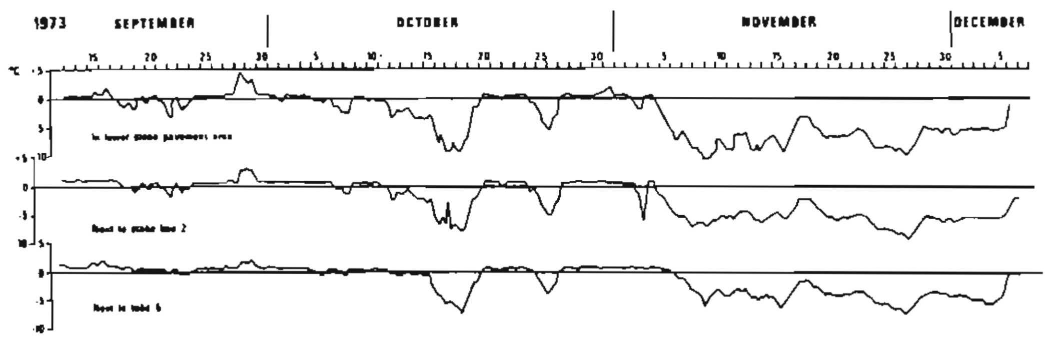

Fig. 6. Three temperature profiles showing autumn to winter freezing of the ground.

Freezing of the ground during the autumn to winter period clearly takes place from the surface downwards (Fig. 5). Whilst the 0.05 m and 0.10 m levels both went subzero on the same date (November 8th) there was a 2-day delay for the 0.25 m level, 14 days for the 0.05 m, 19 days for 0.75 m and 28 days for the 1.0 m level. Once negative temperatures were achieved they persisted till the end of the record, but with frequent, often rapid, subzero oscillations. This subzero cycling is thought to indicate snow-free conditions, with the levelling of the record (at 0.05, 0.10 and 0.25 m depths) on December 8th indicating covering by snow.

Temperatures at 0.03 m depth next to ground movement instruments (Fig. 6) showed essentially the same responses with minor variations due to local differences. The ground was subject to freezing and thawing until November 4th, when it became continually frozen but still experienced subzero oscillations.

Movement of material across the snow surface

Although this form of transport might not b(considered a nivation process per se (Thorn & Hall 1980, p. 121), nevertheless it was active a the nivation site and so must be taken into ac count within the totality of debris movement Weathered-free material was seen to slide, rol or bounce across the snow surface. Althougl debris in motion across the snow was primaril limited to cobble size and larger, slumps of fin material were observed (Fig. 7) which had a ne downslope movement when let down by the ab lating snow surface. For rockfall movement ove the snow, four factors had to be satisfied: ther must be active weathering of the backwall (t generate material), snow must abutt or be ver close to the backwall (to offer a surface fo movement), the snow slope must be relativel steep (to aid movement), and the snow itself be i

Fig. 7. Movement of a boulder and fines down the steep snow slope.

a hard condition (so as not to retard momentum).

The snow slope at the study area varied between 40 and 60 degrees, and the ground at its foot from 15 to 45 degrees. Thus, many clasts had sufficient momentum to leave the snow foot and travel 20 or 30 m out beyond the snowpatch. During movement beyond the snow foot, moving blocks often impacted against stationary ones and caused them to travel a short distance downslope. Qualitatively, rolling appeared to be the most common means of movement.

It was not possible to sample the moving material and thus estimate the mass moved. The largest block seen to move across the snow surface was of the size 1.0 × 0.6 × 0.45 m. The period during which this form of transport occurred varied from year to year as a function of when a free rock face was exposed behind the ablating snowpatch, the release of weathered material, and the rate of ablation of the snowpatch.

Transport by wind

Transport by wind was only seen to be effective on particles up to coarse sand size and was spatially and temporally extremely limited. Activity was restricted to dry, snow- and vegetation-free sites on windy, precipitation-free days. This limited the zone of activity to the vegetation-free stone pavement area after the snow cover had melted. Despite the many limitations there were days following a warm, dry period during which the air was seen to be full of small mica flakes. These clay to silt-size particles were carried tens of metres above the ground and were only visible due to their reflecting the sun. No quantitative estimate of the amount of material involved is available but, qualitatively, it is thought to be very small.

Ground-surface movement of fine to coarse sand was observed and an elementary form of sorting took place. Sorting resulted from the transport away of the finer particles, thus leaving material of granule size or larger, and its redeposition at irregularities in the ground. At deposition sites there was a gradation from coarser to finer material in a down-wind direction. Build-up of deposits took place until the irregularity was filled and then material would continue to be moved beyond that point by the wind. Unlike the other transport processes, movement by wind was not directly downslope but was a vector between the cross-slope wind and the downslope force of gravity.

Transport by water

Water was seen to issue from beneath the snowpatch either from a tunnel as a concentrated flow along an incised channel, or as a sheet from below sections of the feather edge. All the observed tunnels were greater than 0.12 m in width. One tunnel observed in 1972 was 0.13 m wide, 0.085 m high and had water to a depth of 0.01 m; another in 1973 was 0.5 m wide, 0.15 m high with water 0.03 m deep. During June of 1973, when ablation was well advanced, five small tunnels were observed on the eastern side of the longitudinal gully and eight larger ones on the western side. Those rivulets emerging from tunnels on the eastern side were in shallow channels but then usually spread out into sheets. On the western side the water exited from the tunnels in well-defined channels up to 0.4 m deep and 0.3 m wide. These larger channels appeared to be in use in successive years but the existence of some which were dry suggests that some grow at the expense of others, as in a normal drainage network.

One tunnel, excavated back into the snowpatch, showed water continually in contact with the ground which was thawed to a depth of 0.05 m. Excavation also indicated that rivulets joined together, beneath the snow, to produce an anastomosing network, but that as any individual channel was traced back upslope it became shallower until, ultimately, somewhere within the stone pavement, it was no longer evident.

The sheetflow from under the snow edge was, due to its extreme shallowness and low velocity, observed to rarely move debris and then only of fine sand or silt size. The rivulets, however, were often seen to carry material in the fine to coarse sand size range, but the colloidal nature of the finer material inhibited movement except when subject to disturbance by such as falling blocks of snow from the tunnel edge. In addition, inferential evidence of transport was found in both the stone pavement and vegetated areas. Within the former were observed, after the snow cover had disappeared, sinuous ridges of sand (<0.05 m high) which slowly increased in width and/or height downslope or where two ridges joined together. These ridges of fine material (termed 'aleurite' by Kachurin (1959)) ran downslope but, in so doing, often ran across blocks or mounds which involved a short upward section. This, then, implies transport and deposition in a subsnow channel under hydraulic pressure with the ridges forming in much the same manner as an esker. That this material was in motion down the slope is shown at the tunnel exits where water flowed out over a vegetation cover and spreads of debris, trapped in the grasses, were observed. Of interest was that in these spreads of debris large amounts of gravel size material were found, indicating that earlier in the ablation cycle conditions were such that material of this size could be transported.

Analysis of the solute content of the water (Thorn & Hall 1980, Table VI) showed that, although the amounts were not high, there was removal of material in solution. It is argued that as this material is being removed by water released from the ablating snowpatch then a greater amount of material can be removed from a nivation site than from surrounding snow-free areas.

Mass wasting

Analysis of ten soil samples from the top 0.5 m of ground in the lower area indicated a range from silty-sand to gravelly-sand, but with a varying percentage of silt evident in each sample (Hall 1975, Fig. 6–17). Detailed investigation of the more common finer soil indicated a liquid limit of 27.8 % and a plastic limit of 22.1 %. These values gave a plasticity index of 3.55 % which, on the Casagrande Plasticity Chart (Casagrande 1932), approximates to a soil of low cohesion. During 1973, tensiometer records of pore water tension in the top 0.3 m of the ground in the lower area showed that from early to late June there was a saturated condition. There was loss of tensiometer record in July, but a saturated state was again indicated from late August to late September, when readings were terminated. This saturated soil of low cohesion gave morphological expression of mass wasting in the form of terracettes, solifluction lobes and a mudflow.

Terracettes were observed in the lower, vegetated sector. Solifluction lobes were found to the west of the gully and a section through one indicated and overriden organic horizon and a downslope clast a-axis orientation fabric. The mudflow, also to the west of the gully, was a large (c. 25 m long and 8 m wide) multiple feature resulting from a number of successive slope failures. It comprised several lateral ridges and terminal banks of debris, up to 0.5 m high, composed of material from clay to small boulders in size.

Monitoring of ground surface movement by means of deformable tubes showed a movement range of 0 to 55 mm in one year (Fig. 8), but no relationship between the amount of movement and slope angle (Table III). For the tubes as a whole, no movement took place below a depth of 0.35 m. The mean rate of movement for all the tubes monitored during the two-year record period was 22.5 mm yr^{-1}. The recorded tube movements relate to duration of ground saturation as a function of interstitial ice melt and, more importantly, snowmelt water supply from upslope. The

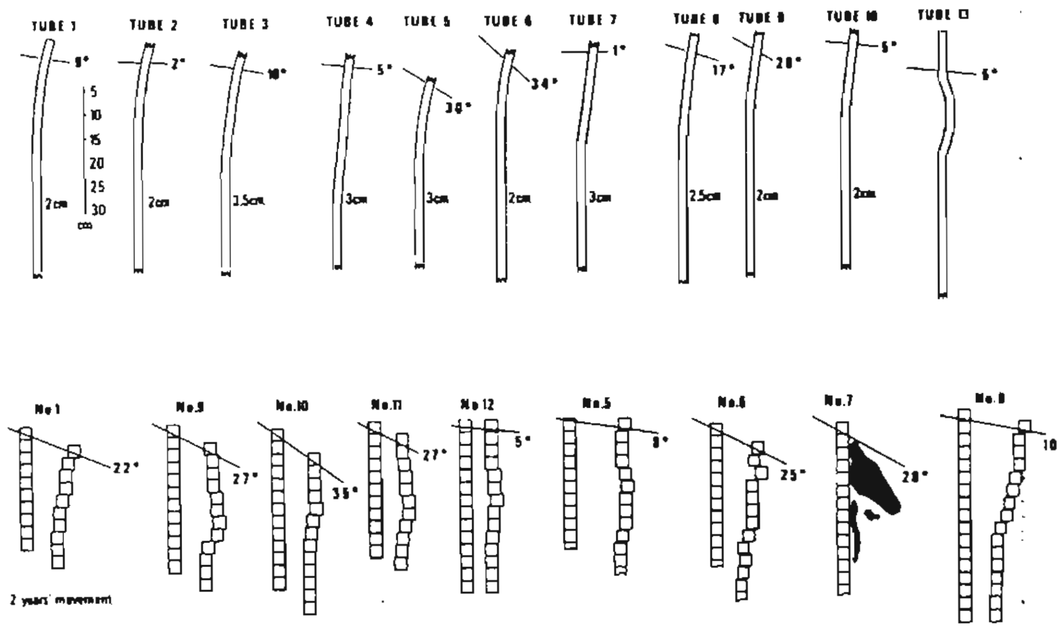

Fig. 8. Tube profiles and detail of ground movement as shown by Rudberg Pillars after one year.

relationship between saturation and movement was best exemplified by Tube 13 (Fig. 8) which showed no surface displacement but distinct downslope deformation between 10 and 180 mm depth. The tube was initially straight as the inserted metal rod at time of emplacement would not have allowed any bending. Excavation of the tube showed that there was a clearly observable zone of throughflow in a band corresponding to the curved section of the tube. Water flowed into the excavation pit at a fast rate and slumping-in of the saturated ground took place, so that after about five minutes the throughflow band had cut back 70 mm. At the base of the flow was a 40 mm clay layer which was acting as an impervious boundary. Surface movement is thought to have been inhibited by the vegetation mat and so there must have been a zone of shearing between the moving and non-moving sections which could not be expressed by the tube.

That shear took place within the ground was clearly shown by sets of Rudberg Pillars, whose discrete nature allowed differential movement to be recorded (Fig. 8). Surface rates of movement varied between 10 and 50 mm but in many instances maximum downslope displacement took place below the surface. This may, in part, be explained by surficial retardation by vegetation, as Pillar set 8 (Fig. 8), in an unvegetated locality,

Table III. Ground surface movement for one year as shown by tube profiles.

	Movement (mm)		
Tube No.	1972-1973	1973-1974	Slope (°)
1	20	15	9
2	20	20	2
3	35	20	10
4	30	30	5
5	30	20	30
6	20	25	34
7	30	20	1
8	25	20	17
9	20	25	29
10	20	10	5
13	0[1]	-[2]	5
14	15	-	20
15	20	-	5
16	55	-	30

[1]: movement at 0,11 depth = 20 mm

[2]: tubes removed or damaged

Correlation between amount of movement and slope angle (1972-3) indicates r = +0,33 which is not significant at 0,05 showed maximum movement at the surface. However, as with the tubes, it is the pressure of water creating zero pore water tension which is the prime control, because, again, there is no relationship between amount of movement and slope angle.

The lines of wooden stakes were not such good indicators of downslope movement as either the tubes or pillars, because their rigid nature inhibited clear expression. However, they were able to give a measure of residual ground heave, i.e. winter heave less summer settling. Surface movement rates recorded varied between 0 and 50 mm ($\bar{x} = 12.6$ mm, $s = 15.8$, $n = 39$) whilst residual heave was between 0 and 30 mm ($\bar{x} = 9.5$ mm, $s = 6.5$, $n = 30$). Within any one line of stakes there was great variety in both downslope movement and heave.

Conclusion

From the available evidence it would appear that material is moved out of the nivation site by a combination of over-snow transport, wind action, solifluction and confined or unconfined water action. With the onset of spring and snowpatch ablation so water transport is initiated. Water entering the back of the snowpatch from melt along the backwall is able to move beneath the snow cover. At this stage the ground beneath the snow, apart from the top few millimetres where the water runs, is still frozen and so solifluction is not yet active. During this period, the proximity of the snowpatch to the backwall allows weathered-free material to move down the snow slope. Thus, initial activity is in the form of water and over-snow transport.

With the loss of snow cover the ground thaws, from the surface down, and solifluction is initiated. The inherent low cohesion of the ground, plus its saturation by interstial ice melt and addition of water from snow melt, leads to downslope movement in various forms: terracettes, solifluction lobes, slumping and mudflow. During the period of ground movement, water transport is still active and so may be over-snow movement. However, as time progresses and the snowpatch ablates, the over-snow form of movement decreases and then finally ceases when weathered or slumped material falls into the randkluft, at the backwall foot, rather than onto the snow surface.

As summer progresses and the ground thaws, so the movement of saturated ground over a sti frozen layer (solifluction) ceases and is replace by creep. During this period warm, dry days ma allow some movement of the fine material l wind action, particularly if it is a year in whic the whole snowpatch ablates (as in 1972), then by exposing the vegetation-free stone paveme area. Towards the end of summer, transport limited to creep, rock fall, some water activit and occasional but rare, wind action. All transpo processes, except perhaps for rock fall, come to halt as the ground begins to freeze once mo and then becomes snow covered. No over-sno movement was observed prior to the onset of tl succeeding ablation period, presumably becau the mechanically weathered material was eith held by an ice bond or was beneath the ba edge of the snowpatch.

Thus there is a continuum of transport mech nisms operative throughout from spring to ear winter. The processes vary both temporally ar spatially as the snowpatch ablates. Essential fluvial and over-snow transport occurs at the b ginning to be superseded by mass wasting. Co trary to most observers (see Thorn & Hall 198 Table I), erosion by meltwater was observed the lower vegetated area. All of the transpo mechanisms are the same as those active in tl surrounding areas with the exception that at tl nivation site the wasting of the snowpatch exe a distinct influence over the location and timi of their operation. In addition, the snowpat enhances the effect in that it acts as a surfa over which material may move and that it is reservoir of moisture which continues to be r leased when surrounding sites may be dry. Thu it is suggested that a nivation hollow is a locatic of enhanced transport activity compared to tl surrounding snow-free regions. It is further su gested that it is largely because of this enhance transport, rather than the usually cited enhance weathering, that the nivation hollow exists ar grows. In fact, in the light of recent quantitati studies (Thorn 1979a) which indicate that m chanical weathering may not be as effective originally postulated, it may well be that it enhanced transport which characterizes nivatic rather than the oft-cited freeze-thaw activity.

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ROCK TEMPERATURES FROM LIVINGSTON ISLAND (MARITIME ANTARCTIC): IMPLICATIONS FOR CRYOGENIC WEATHERING

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The present-day understanding of cryogenic weathering of rock is constrained not so much by theory as by actual field data regarding the controlling factors (rock properties, rock temperatures and rock moisture content, chemistry and distribution). Attempting to overcome this limitation, the British Antarctic Survey has initiated a series of undertakings, at various locations in the Antarctic, in which these parameters are monitored. The rock temperature component of the information data base is presented for three sites on the ice-free Byers Peninsula of Livingston Island in the South Shetland Islands.

At these study sites, on different aspects, rock temperatures were monitored several times each day. In addition, climatic data regarding air temperature, radiation, humidity, wind speed and direction and moisture availability were also monitored. This information made it possible to undertake within and between site comparisons plus to consider the relationship of the various rock temperatures to climatic conditions.

INTRODUCTION

In a discussion regarding frost shattering, McGreevy (1981, p. 71) concluded that "Priority ought to be given...to measurements of rates, amplitudes and magnitudes of temperature oscillations around freezing point on and below rock surfaces...". Seven years later, Thorn (1988, p.13), following a detailed discussion by McGreevy and Whalley (1982), was still led to state that "...there is a pressing need for field records of bedrock temperature..." and four years further on to state "Foremost among...problems has been uncertainty concerning the thermal and moisture regimes which actually prevail within natural bedrock.." (Thorn, 1992, p.10). Many studies rely on air temperature records and yet these are of no use with respect to rock weathering (McGreevy, 1985) and so it is imperative that actual rock temperatures are obtained. Whether the older notions of rock damage due to a 9% volumetric increase resulting from the phase change of water during freezing or the newer hypothetical models based on segregation ice growth within rock as the disruptive influence (e.g. Hallet, 1983) are considered, then without such data it is impossible to determine the mode of frost weathering in the field or to undertake meaningful laboratory simulations. In fact, without data on both temperature and moisture the presumption of frost action as the main cause of rock breakdown in cold regions may itself be erroneous (Thorn, 1992). Rock temperature data are, in general, relatively rare although McGreevy (1985) has examined the role of the thermal properties of rocks on weathering processes. For cold regions there are some data, of varying degrees of longevity and complexity, relating to rock temperatures but considering the extent and diversity of such regions it is still an extremely small data base.

As part of a recent study of weathering processes in a maritime Antarctic environment, rock temperature data were collected for part of the summer period. Data were collected for various aspects, allowing comparison with the controlling climatic factors of radiation, humidity, air temperature, wind speed and direction. In much the same vein as papers by Jenkins and Smith (1990) for the Canary Islands and Balke et al. (1991) for the continental Antarctic, the aim here is to show the variability of rock surface temperatures that can occur as a function of aspect and to discuss the findings in the context of the generally perceived nature of weathering for this region. Data on ancillary factors pertinent to the consideration of weathering, such as rock moisture content and rock properties have been collected but are presented in detail elsewhere (Hall, In press).

STUDY AREA

The study was undertaken on the Byers Peninsula, located at the western extremity of Livingston Island. The Byers Peninsula is the largest (c. 50 km²) ice-free area in the South Shetland Islands (Lat.62°40'S, Long.61°00'W). According to John and Sugden (1971) this area has a mean annual temperature of -3°C and annual precipitation is in the order of 100 to 150 cm water equivalent. Thom (1978) states that for most years the mean daily temperature from December to March is above freezing and that permafrost is present below an active layer of 0.3 to 0.7m thickness. However, whether there is truly permafrost, as opposed to seasonally frozen ground, has not been verified.

The bulk of upstanding rock outcrops in the landscape comprise volcanic dykes, sills or plugs. By providing a major obstruction to the snow-bearing winds, these outcrops collect a substantial lee-side accumulation of snow, predominantly on the south-facing side. In addition to the presence of snow the southerly aspect also has low radiation receipts, is influenced by the cooling effect of occasional cold, southerly winds and is protected from rain-bearing northerly winds during summer. Thus different aspects have markedly different microclimatic environments and these are reflected by the observed differences in weathering (Hall, 1992, In press).

With respect to weathering, several authors have argued strongly in favour of freeze-thaw action dominating in the South Shetlands. Simonov (1977) indicated that despite a mean annual air temperature of -2.9°C the rock surfaces experience frequent crossings of 0°C. Blümel (1986) showed that in 1979 there were 286 frost days consisting of 164 when no thaw occurred and 122 with both freezing and thawing. These were with respect to the air, however, and thus take no cognisance of rock heating due to insolation. However, the extensive cloud cover in this region limits the effect of direct radiation although Simonov (1977) did find that with solar heating the rock surface temperature could rise to 20°C whilst the shadow-side remained below 0°C. These temperatures do not necessarily mean effective frost action occurred as, in addition to subzero temperatures, there must be water available in the rock to freeze. Nevertheless, with respect to the South Shetlands, Araya and Hervé (1966), Corté and Somoza (1957), Olsacher (1956), Vtyurin and Moskalevskiy (1985), Simonov (1977), Stäblein (1983), Blümel (1986), Dutkiewicz (1982) and Blümel and Eitel (1989) all cite freeze-thaw as the major agent in modifying the landscape.

METHODOLOGY

Rock temperature data were collected during January and February of 1991 from three sites: an E-W trending dyke, a volcanic boss and on a raised beach. On the dyke, temperatures were recorded, by means of thermocouples (with an accuracy of ±0.2°C) for the north- and south-facing aspects at the rock surface and at a depth of 2cm at c. 0900, 1300 and 1800 hrs each day plus at others times as the opportunity arose. On the volcanic boss, rock surface temperatures were monitored simultaneously on the four cardinal aspects. At these two sites rock moisture content was ascertained for each rock face at the same time the temperature measurements were taken. At the raised beach site two logging systems were used. A Campbell logger was utilised to record air (at 1m above the ground surface), ground surface and 5 cm depth temperatures, together with wind speed and direction, humidity, and radiation. Being a site exposed though the full 360° the raised beach served as a control site to give the general meteorological conditions against which the other sites could be compared. Data were collected at 30 minute intervals for the first 25 days (Julian days 7 to 31) and then at 5 minute intervals for the remaining 20 days. Cobbles of local rock were used at this site to monitor non-aspect controlled variability in rock moisture content. In addition to the Campbell, sporadic use was made of a Grant "Squirrel" system to monitor the differences between the top surface and north-facing surface of a cobble. Although the two sensors were only c.2 cm apart they showed distinctly different thermal regimes on the two faces. Both the Campbell and the Squirrel loggers used thermistors with an accuracy of better than 0.1°C.

Ancillary data on factors such as rock properties (Hall, in press) and the nature of the sorted pattered ground (size and orientations) were also obtained as these provide information that is useful in determining the weathering regime (Hall, 1992).

RESULTS AND DISCUSSION

At the dyke, the mean surface temperature for the northern aspect was 3.1°C whilst that for the southern aspect was 1.6°C. The minimum for the south (0°C) was lower, but not substantially so, than for the north (1°C) whilst the maximum for the north (19°C) was much greater than that for the south (9°C). Despite the disparity in ranges between the two sites (18° vs. 9°C) a comparison of sample means indicates the two samples come from the same population (with $\rho = 0.05$). Although the north has a larger spread of values, the concentration of temperatures is mainly in the range 2.5° to 4.5°C whilst that for the south is in the range 3.5° to 5.5°C.

For the rock boss, the mean temperature of the northern aspect (6°C) was higher than that of the southern (3.9°C) but not so different from the values for the east and west aspects (which were both the same at 5.3°C). A comparison of sample means with meaningfully paired observations for the north and south aspects showed that (with $\rho = 0.05$) they could not be considered to be from the same population. The temperature range for the north was the greatest (16°C) whilst that for the south was the least (7°C) with that for the west and east both being 13°C. Statistically, the observations for the west and east showed no difference (with $\rho = 0.05$).

Rock surface temperatures for the dyke are different from those for the boss (Fig. 1). The mean temperature for the northerly aspect of the boss was only 1°C higher than for the dyke but a comparison of sample means with meaningfully paired observations indicated that the two could not be considered to come from the same population ($\rho = 0.05$). Similarly, the southern samples from the dyke and the boss were statistically different but here the mean for the dyke (4.6°C) was greater than that for the boss (3.9°C). The main reason for the southern aspect of the boss having lower temperatures than that of the dyke is that the former is overshadowed by a large rock outcrop that protects it from much of the late-day, low-angled sun. The slight difference between the northern aspects at the two sites is a result of the angles of their respective faces. Whilst the dyke had a vertical to slightly overhanging face, that of the boss was at ±50° and so received radiation at a more perpendicular angle. However, at both sites rock surface temperature maxima correspond to peaks in received radiation with, at the dyke (Fig. 1), the southern face experiencing a higher temperature than the northern face in the afternoon.

At the open site, the ground surface temperatures broadly replicate that of the air (Fig. 2) but with highs closely mirroring peaks in received radiation. The effect of the incoming radiation in heating the ground is clearly expressed by the difference in temperatures between the air and the ground surface with the maximum ground surface temperature (14.1°C) substantially greater than that for the air (4.9°C). Conversely, the minimum values show radiative cooling from the ground surface with the air temperature (1.2°C) being greater than that for the ground surface (0.45°C). The mean wind speed (22 km hr^{-1}, s=4) reflects a continuous wind throughout the day, with only 2% of observations below 14.4 km hr^{-1} but with 56% above 21.6 km hr^{-1} and 6% above 29 km hr^{-1}. The strongest winds were between 0030hrs and 1000hrs, the lowest between 1230hrs and 1530hrs.

At the open site, measurements on the top and north-facing sides of a stone show differences of >1°C despite their having been taken only 2cm apart from each other. On January 8 (Fig. 3) the north-facing side of the stone was marginally (±0.6°C) warmer than the top surface until c.1445hrs when the top became warmer, culminating in a peak at 1510hrs (when it was 1°C hotter). The two sides then attained similar temperatures during the cooling phase until c.1900hrs, when the north face became marginally (0.1°C to 0.4°C) warmer than the top surface. Another example of the marked difference between these two aspects is given by the data for February 5 (Fig. 4). On this day the northern aspect became warmer than the stone

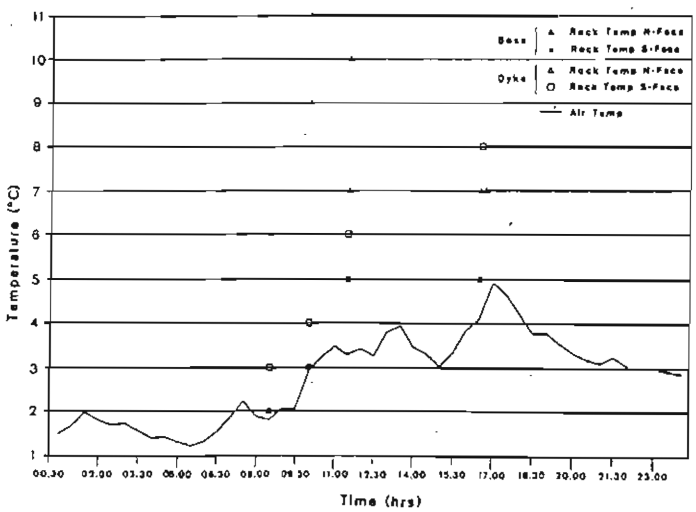

Fig.1: North and south face rock temperatures as measured at the dyke and the rock boss together with the air temperature record for January 8th, 1991.

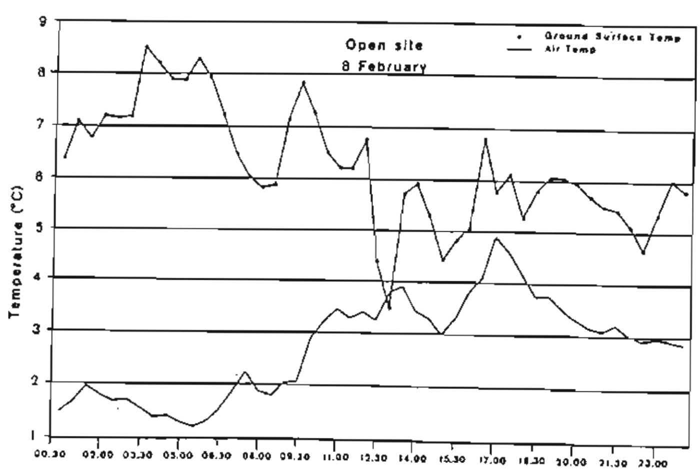

Fig.2: Temperatures recorded for the air and soil surface at the open site on January 8th, 1991.

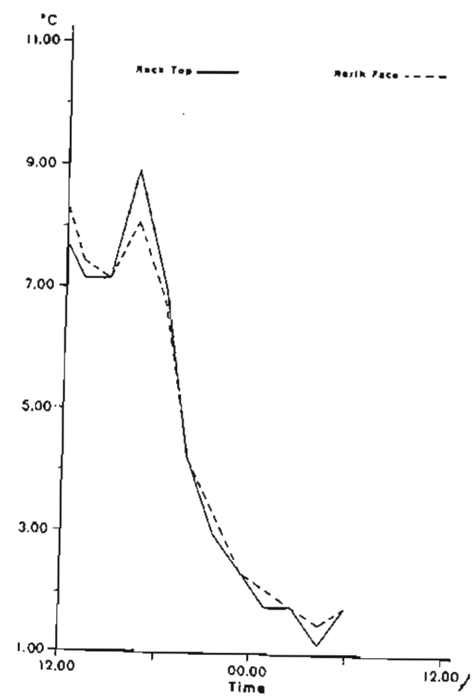

Fig.3 : Temperature measurements for the top and north-face of the stone at the open site for January 8th, 1991.

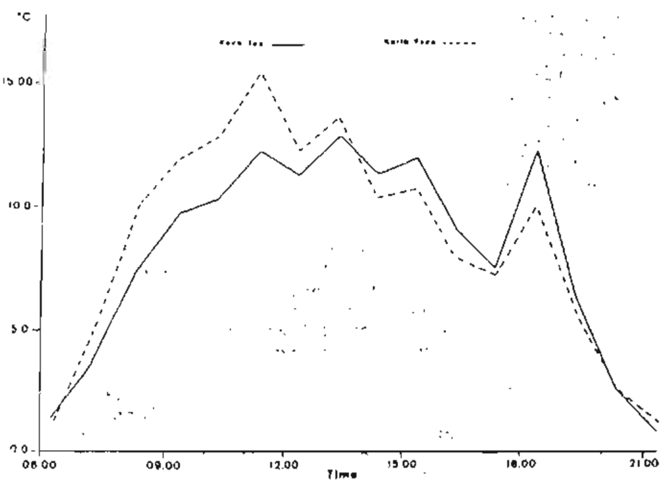

Fig.4 : Temperature measurements for the top and north-face of the stone at the open site for February 5th, 1991.

top at 0630hrs, after which the difference between these two faces continued to increase until a peak at 1120hrs when there was a 3°C difference (14.6°C vs. 11.6°C) between the two faces. After this peak the two records begin to coincide until at 1340hrs they cross over and the rock top attains the higher temperature with the maximum differences at 1510hrs (1.4°C) and 1815hrs (2.1°C). Subsequently the two curves coincide once more with the northern face becoming marginally warmer (±0.1°C) after 2015hrs.

The importance of these data are twofold. First, marked differences in temperature can occur over very short distances and this, in turn, could have an influence on the character and rate of the operative weathering processes. Second, high recorded temperatures could be conducive to chemical, rather than physical, weathering. Clearly the above effects are the result of heating of the rock by incoming radiation and night-time heat loss due to radiative cooling. The northern face has the greatest receipts in the first part of the morning but is superseded by the top in the afternoon when the north face is in shadow. The night-time crossover is a result of radiative cooling affecting the upper surface of the rock more than the vertical northern face. In addition, as with the data from the dyke and the rock boss, it shows how much higher rock temperatures can be as compared to the air. On February 5 when the north face attained a temperature of 14.6°C, the air was only 4.9°C. Later, at 1815hrs when the top surface of the rock was 11.6°C the air was 4°C. Conversely, as night approached and the air temperature dropped to 2.3°C (@2100hrs) the stone was ±0.9°C as a result of radiative cooling.

Another attribute shown by the detailed data from the logger at the open site is the difference in the rate of change of temperature. During the cooling phase from 1810hrs to 2010hrs on February 5 the rate of change of temperature of the rock surface was 4.8° hr^{-1} for the first hour and 4.5°C hr^{-1} for the second whilst that for the air showed no temperature decline during the first period and only 1.2°C hr^{-1} during the second. Thus the rock is clearly cooling more rapidly than the air, mainly as a result of the loss of incoming radiation. Although this was not a freezing event it is interesting to note that the recorded rate of change of temperature for the rock (0.08°C min^{-1}) was very close to the 0.1°C min^{-1} suggested by Battle (1960) to be necessary for effective frost action., i.e. it could be that rates of cooling are more conducive to the hypothesis of Battle (1960) rather than the slow rates required by the model of Walder and Hallet (1985). However, as argued later, slow rates of fall of temperature in line with the suggestion of Walder and Hallet (1985) are more likely to occur when autumn snow provides some insulation, which slows the penetration of the winter freeze.

Thus it can be seen that the air temperature record does not indicate what is happening to the rock. Consequently, the air temperature record of frost events as cited by Blümel (1986) for the South Shetlands may grossly overestimate the role of freeze-thaw weathering and underestimate other processes. Another noticeable fact is the differences in rock temperature that can occur as a function of aspect, including over very short distances. These temperature differences may be important with respect to weathering processes such that, temporally and spatially, different processes may be operating. As moisture variability was often less than temperature variability during the period of record it would seem that the former exerts a stronger influence on weathering than the latter. One exception was the presence of snow on the lee side of obstacles that acted as a moisture source during warm, dry periods. This availability of moisture combined with high rock temperatures resulted in weathering not usually considered with respect to the Antarctic environment, namely chemical weathering.

The presence of rock moisture combined with the relatively high (5-15°C) rock temperatures indicates that chemical weathering may be active during the summer, a suggestion also made by Balke et al. (1991) for the ice-free areas of the continental Antarctic. Evidence for the action of chemical weathering was provided by the presence of weathering rinds on both the dyke and rock boss. It was found that the rinds were consistently thicker on the snow accumulation side of obstacles (Hall, In press). Thus, as recently suggested by Balke et al. (1991) with respect to the Antarctic continent, it may be that chemical weathering is far more active, although limited to the summer period, than has been considered previously. Certainly the wet summer conditions combined with the relatively high rock temperatures experienced on Livingston Island explain the presence of weathering rinds and chemically altered bedrock outcrops.

It is noticeable that despite the number of authors who cite freeze-thaw as the major process in the South Shetlands, no frost events with negative temperatures lower than -1°C took place during the 45 days of record. Rather than freeze-thaw, it was wetting and drying that was the most commonly recorded event, largely as a result of the frequent rain and strong, desiccating winds. Wetting and drying is a mechanical weathering process about which very little is known (Hall, 1988), but experiments currently in progress on samples of bedrock from this area have shown that a mass loss of 3.2g resulted from 82 wetting and drying cycles indicate a change in the perception of weathering for this area, nevertheless it is still insufficient to make any sound detailed judgements. However, this is a universal problem with respect to cold regions and epitomises both the need for more data and the care that should be accorded in simply assuming the action of freeze-thaw.

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Rock Temperatures and Implications for Cold Region Weathering. I: New Data from Viking Valley, Alexander Island, Antarctica

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ABSTRACT

A principal tenet of cold region weathering studies is that of temperature. Unfortunately, despite the appearance to the contrary, actual data are still sadly lacking in many instances. Here data are provided for the best part of two Antarctic winters plus, at two minute resolution, for one summer, from a variety of positions within a dry valley. The data clearly show the dangers of using air temperature as a surrogate for thermal conditions either at the rock surface or at depth in the rock. Although detailed rock moisture data are absent in this study, indirect evidence from both observation and non-destructive ultrasonic testing shows that water is extremely limited during the period of freeze–thaw cycles. Thus, despite the occurrence of the thermal events no damage can result from frost action. Detailed data at two minute intervals show the importance of such high resolution observations. It is argued that without data acquisition at two minute or preferably one minute intervals, it is not possible to discern the weathering regime, including interpreting the freeze–thaw process. These data show that processes such as thermal stress fatigue/shock are possible and that rates of change of temperature $\geqslant 2\,°C\ min^{-1}$, as required for thermal shock, do occur. Measurements of taffoni size and occurrence, coupled with Schmidt hammer rebound values, show that the eastern aspect experiences the least weathering whilst the northern and western exposures have the greatest amount. This observation is in accord with separate findings of aspect-controlled orientation of "cryoplanation" terraces at higher elevations. Some speculative suggestions for the cause of this aspect-controlled weathering are given, the most important of which is that it is unlikely that freeze–thaw plays any significant role – a factor that underscores its qualitative presumption in cold regions and questions the origin and development of cryoplanation forms. Beyond anything else, this paper indicates the complexity of rock temperature regimes and suggests that it is the synergistic relationships between different weathering processes that are important. © 1997 by John Wiley & Sons, Ltd.

RÉSUMÉ

On croit généralement que l'altération des roches dans les régions froides est contrôlée par les fluctuations de températures. Malheureusement malgré les apparences, des observations précises manquent cruellement dans bien des cas. Dans le présent article, des données sont fournies pour différentes situations dans une vallée sèche de l'Antarctique pendant la plus grande partie de deux hivers, ainsi que, avec une résolution de deux minutes, pendant un été. Les données montrent clairement les dangers d'utiliser les températures de l'air et non les températures observées à la surface des roches ou en profondeur. Bien que des données détaillées d'humidité soient absentes dans la présente étude, des évidences indirectes résultant à la fois d'observations et de tests par ultrasons indiquent que l'eau est extrêmement limitée pendant la période où se produisent les cycles de gel-dégel. De ce fait malgré l'occurrence d'événements thermiques, aucun dommage ne peut résulter de l'action du gel. Des données détaillées avec un intervalle de mesures de deux

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minutes montrent l'importance d'observations à haute résolution. Il est démontré que sans observation avec cet intervalle, ou ce qui serait encore mieux avec des intervalles d'une minute, il n'est pas possible de discerner le régime d'altération, et même d'interpréter les processus de gel-dégel. Ces données montrent que des processus comme la fatigue (due aux chocs thermiques) sont possibles et que des vitesses de changements de température supérieures ou égales á 2 °C par minute, valeur nécessaire pour avoir un choc thermique, existent dans l'environnement étudié. Des mesures de la taille et de l'occurrence des taffonis, mises en rapport avec les valeur obtenues avec le marteau de Schmidt montrent que les expositions à l'est présentent le moins d'altération tandis que les expositions au nord et à l'ouest en subissent le plus. Cette observation est en accord avec des observations indépendantes sur l'orientation des terrasses de cryoplanation à des altitudes plus élevées. Quelques suppositions quant au contrôle de l'altération sont données, la plus importante étant qu'il est improbable que le gel-dégel joue un rôle significatif. Tout cela minimise l'importance de ce facteur dans les régions froides et pose des questions sur l'origine et le développement des formes de cryoplanation. Avant tout, le présent article montre la complexité des régimes thermiques des roches et suggère qu'il y a des relations importantes de synergie entre différents processus d'altération.

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KEY WORDS: Antarctic; weathering; thermal conditions; aspect; cryoplanation; taffoni; Schmidt hammer

INTRODUCTION

In studies of cold region weathering it is frequently "freeze–thaw" (or one of its many synonyms) that is cited as the dominant process and, in most cases, it is air temperature that is used as the foundation for this premise. Both of these assumptions err. First, it may well be that freeze–thaw is not the dominant process, indeed it may not even be operative. Second, air temperatures are not a surrogate for rock temperatures (Debenham, 1921; Thorn, 1988) and thus any deductions therefrom are doubly misleading. Frequently it seems that frost action is universally considered the dominant process in polar or mountain areas. Högbom (1914) suggested that the more rigorous the climate the more effective the frost weathering but van Autenboer (1964, p. 101), after visiting Spitsbergen, states he "was struck by the fact that in spite of a much milder climate the frost activity was considerably greater" than the more "rigorous" (i.e. colder) climate of the Antarctic. Van Autenboer goes on to say that, with respect to the Antarctic, "the extreme severity or rather the consequent aridity ... is even less favourable to the rapid evolution of landscape forms by frost action."

In any consideration of freeze–thaw weathering, moisture and rock properties are as important as the thermal conditions (McGreevy and Whalley, 1982; Hall and Walton, 1992; Hall, 1993). However, in this present discussion it is solely the thermal component that is discussed in detail. With respect to the thermal conditions, Thorn (1980, p. 85) states "The presence of thresholds in the temperature–weathering relationship would produce substantial changes in the likely zones of maximum weathering." This holds true not only for freeze–thaw weathering but for other processes as well, and it is the recognition of these thermal thresholds for processes other than freeze–thaw that is significant. Another key issue is the complexity of rock temperatures and a variety of factors must be considered in any evaluation of weathering where thresholds regarding the magnitude of warming or cooling events, the rates of change of temperature and the thermal gradients are all important (e.g. Matsuoka, 1994).

In this study it was possible to obtain temperature data from a variety of positions and depths at relatively short time intervals (two minutes) as well as longer term data at hourly intervals. Although the latter lack the detail required for a satisfactory analysis of freeze–thaw or thermal stress fatigue (or shock) they do, nevertheless, provide extremely good insight into the character of the weathering regime in an Antarctic dry valley for the greater part of the winter. The combination, presented here, of winter rock temperature data and short time interval summer data allow for a better insight into the nature of Antarctic weathering. In addition, the variety of temperature data obtained

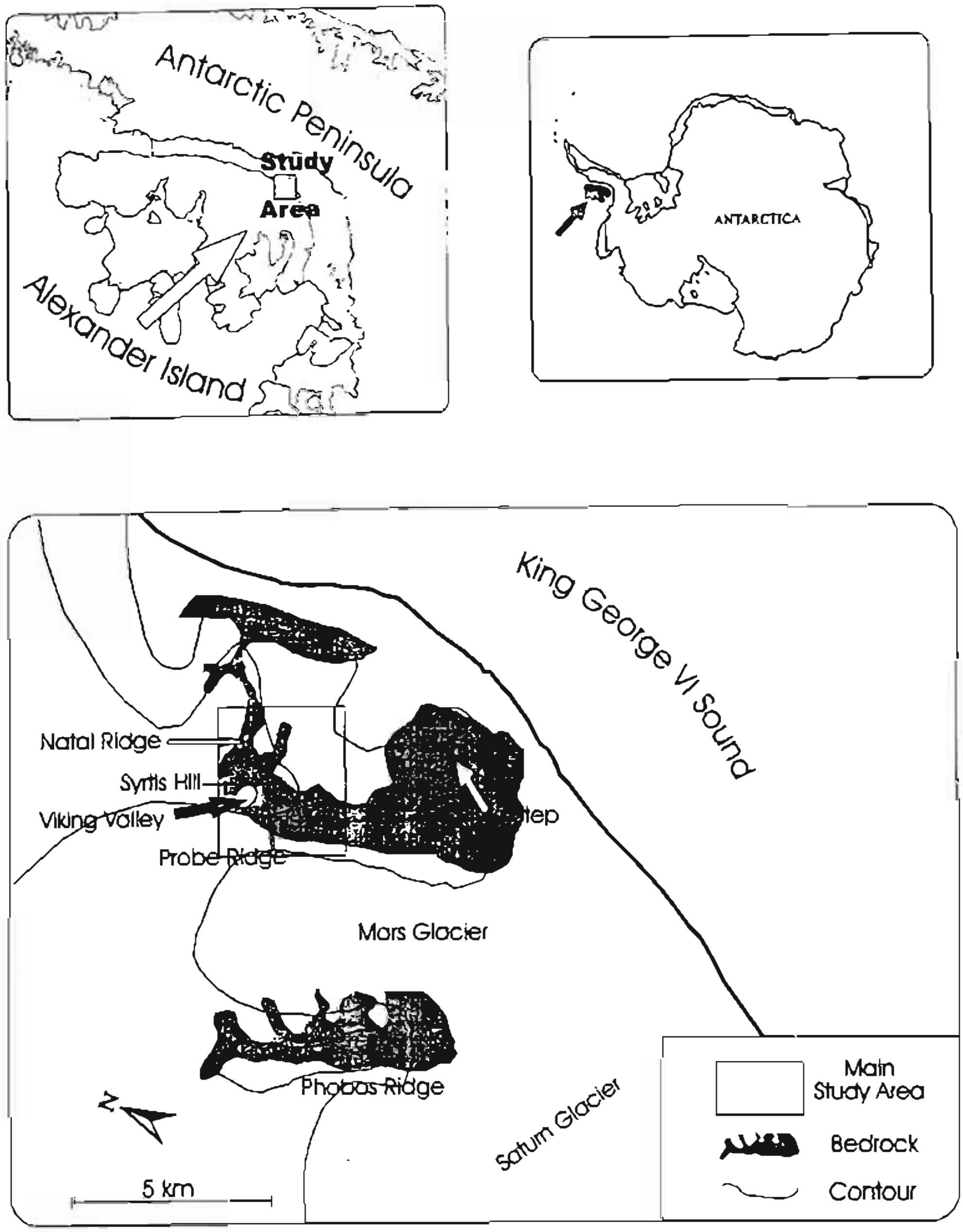


Figure 1 Location of the study area.

allows for a discussion of the interpretation of such data in weathering studies.

STUDY AREA

The study was undertaken in Viking Valley which is located on a north–south aligned nunatak (Figure 1) along the north-eastern side of the Mars Glacier at the southern end of Alexander Island (71°50′S, 68°21′W). This east–west orientated "dry valley" is situated approximately 70 m above the Mars Glacier but has a very small remnant glacier (unnamed) at its northern end between Syrtis Hill and Probe Ridge and a lake (Secret Lake) at the western end (Figure 2). The valley is very small, being less than 200 m between the lake and glacier boundaries, and roughly 40 m

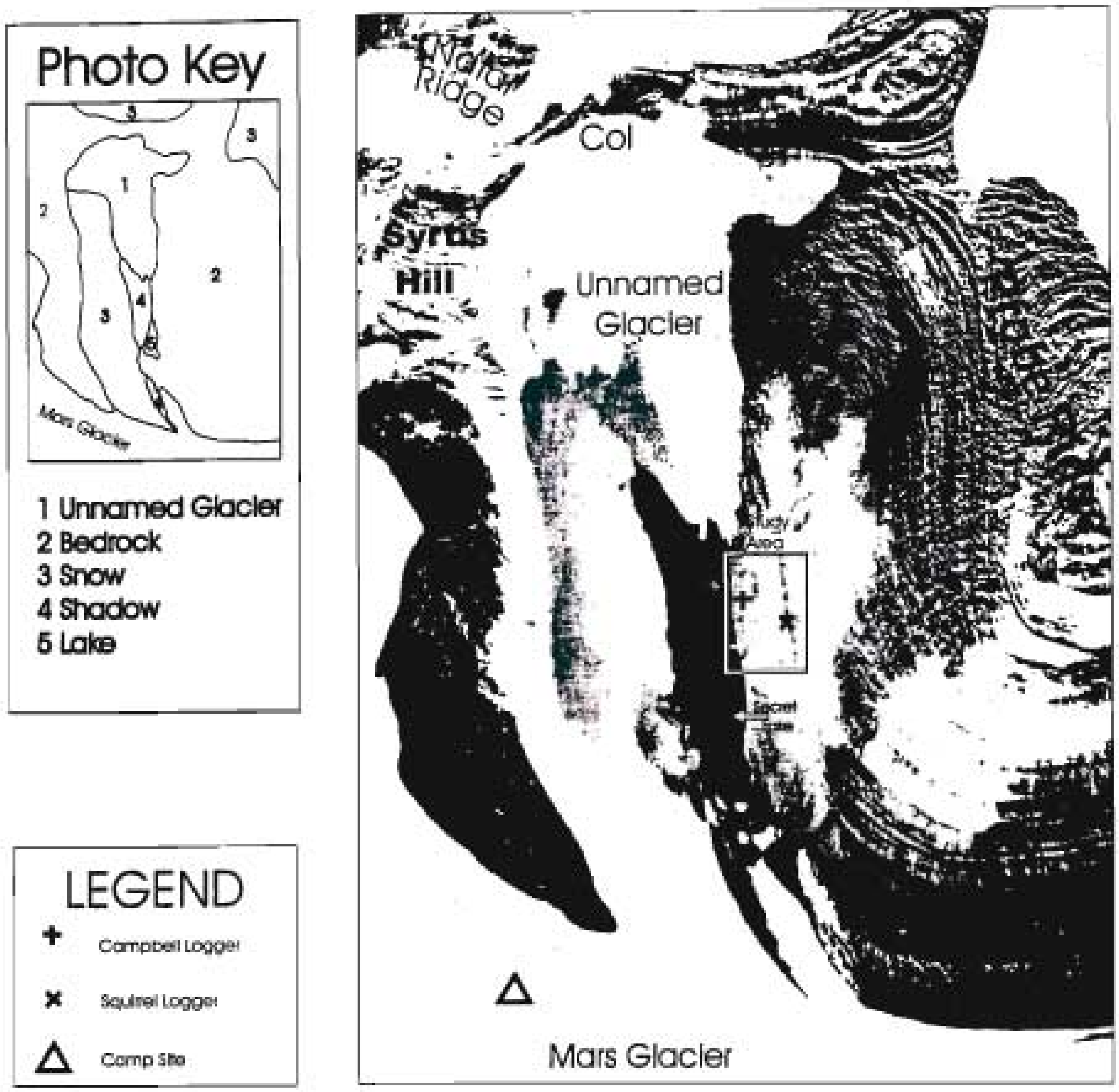


Figure 2 Detailed view of Viking Valley and surrounding area from a vertical air photograph.

wide with a slope from the remnant glacier down to the lake. The lake is ponded at the Mars Glacier end by a rock ridge at an elevation of 60 m above the glacier. The valley is approximately 300 m ASL and although it is only *c.* 8 km from King George VI Sound, the sea here is ice-covered all year round. The area is one of continuous permafrost, there being very little snowfall and mean temperatures of *c.* −2.5 °C in summer and *c.* −11.5°C in winter. Frozen ground and a 0 °C temperature were encountered at 0.27 m depth on 13 December 1992 on Syrtis Hill (Figure 1); the active layer in this area attained a maximum thickness in the order of 0.3 to 0.4 m but aspect exerts a large influence (Meiklejohn, 1995). Geologically the area comprises sandstones, conglomerates and argillaceous sedimentary rocks (Taylor *et al.*, 1979). The dominant lithology is an arkose sandstone that has sub-spherical, postcompaction concretions, comprising siliceous nodules with a ferruginous cement to which the name "cannonball sandstone" is sometimes applied (Thomson, 1964; Horne, 1965; Moncrieff, 1989). Mudstones, shaly mudstones and both light-coloured and dark-coloured orthoquartzitic sandstones are also found. These sedimentary rocks are all horizontally, or near horizontally, bedded. The retreat of the surrounding ice has resulted in extensive vertical dilatation joints that cut the bedding at *c.* 90°.

FIELD APPROACH

The study was undertaken as part of a multidisciplinary investigation of the evolution of soils and the colonization by life of an Antarctic dry valley; Viking Valley was particularly apt as the area was recently deglaciated. As part of the overall study a Campbell 21X micrologger with a full assemblage of meteorological transducers was positioned at an elevation of *c.* 300 m (70 m above the Mars Glacier and *c.* 300 m below the top of Probe Hill). The Campbell was situated 17.5 m from the south-facing valley wall, 13.6 m from the north-facing wall, 85.6 m from the lake edge and *c.* 31 m from the edge of the unnamed glacier. Set to record every hour, the Campbell recorded air temperature, relative humidity, radiation, wind speed and direction as well as temperatures on a dark-coloured sandstone (surface), a light-coloured sandstone (surface), and at the surface and at depths of 5, 10, 15, 30 and 40 mm of a block of local cannonball sandstone. This latter piece of sandstone had the holes for the thermistors drilled through from the "bottom" and then, after the thermistors were emplaced, it was buried in the valley floor such that the top surface was flush and thermal exchange occurred unidirectionally from the top surface. This logger was left to run throughout the summer of 1992–93 and for the winters of 1993 and 1994.

Next to the Campbell logger there was placed another piece of cannonball sandstone to which was attached a pair of 1 MHz ultrasonic transducers linked to a PUNDIT ultrasonic test apparatus. Unfortunately this could not be controlled by the Campbell but had to be run by the operator. As the transducers were fixed to a block of known dimensions it is possible to convert the pulse time readings (in μs) to a velocity (m s^{-1}). Readings were taken at various times on a number of days and the comparable real time values from the Campbell were noted (i.e. for the actual times rather than the hourly readings) for temperatures, radiation, etc. The ultrasonic data are potentially extremely important as they, non-destructively, indicate changes to rock properties (e.g. changes in length due to heating and cooling, cracking and the changes in moisture content). Thus, used in conjunction with the climatic data they can provide valuable information on the behaviour of the rock and the conditions it is experiencing.

On the north-facing valley wall, at a distance of approximately 120 m and *c.* 30 m higher elevation, an 11 channel Squirrel logger was set up to monitor rock temperatures on a rock outcrop pitted with taffoni. On a north-facing exposure, a block of predrilled local sandstone (as with the blocks on the valley floor) with thermistors at the surface and at 5, 10, 15, 30 and 40 mm was positioned flush with the rock exposure. In addition, one thermistor was used to measure the rock surface temperature on the top, east- and west-facing exposures of the same rock outcrop. Two other thermistors were used to monitor temperatures on the bottom and top of the inside of a small taffoni. Temperatures were recorded every two minutes throughout the period in the field (7 December 1992 until 8 January 1993).

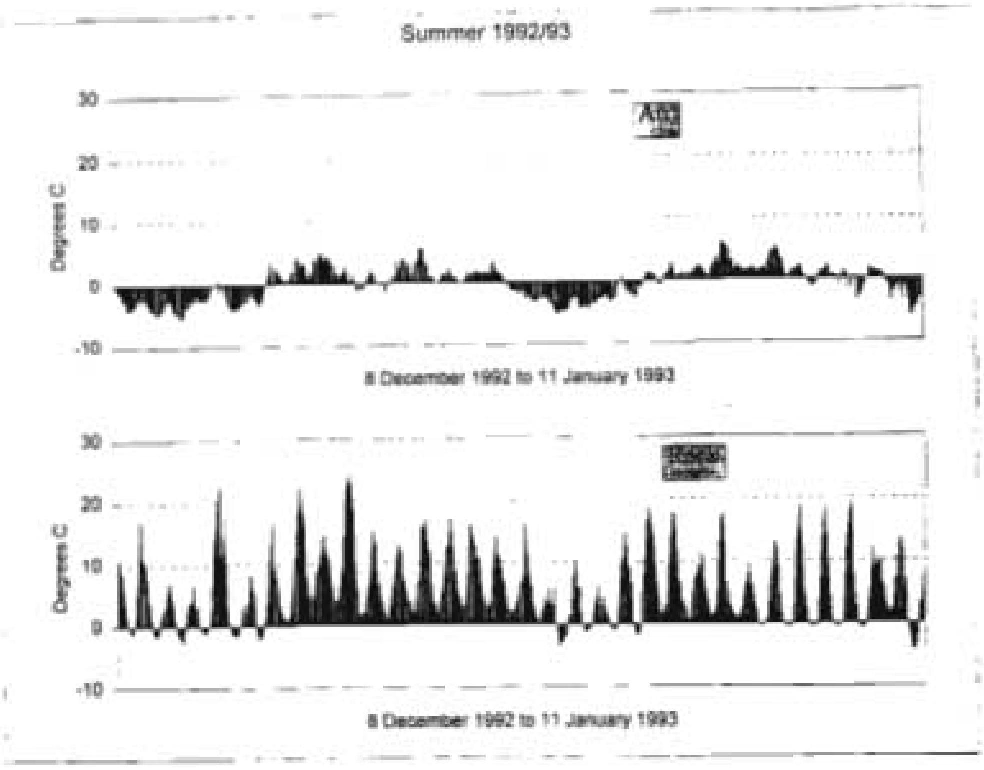
As a further adjunct to the overall study, Schmidt hammer readings were taken at a number of locations throughout the valley. In addition, measurements of size and orientation of the taffoni were collected as well as general observations regarding the distribution of features and the nature of the weathering. Details of Schmidt hammer and taffoni data will be presented more extensively elsewhere but information appropriate to this discussion will be given in outline here. Samples of precipitates (sometimes up to 23.9 mm thick) found on the rocks were taken for analysis by means of X-ray diffraction and all were found to be gypsum (calcium sulphate).

RESULTS AND DISCUSSION

The information is such that it is more convenient to discuss it under a series of headings rather than as one issue. For clarity, the material can be considered in the context of: (i) the weathering regime found in Viking Valley, (ii) the perception of Antarctic weathering, and (iii) implications for weathering studies in general.

Weathering in Viking Valley

Valley floor (one hour interval) data. The Campbell data (i.e. at hourly intervals from the valley bottom) for the summer of 1992–93 and the winters of both 1993 and 1994 for the air temperature and the rock surface temperature are shown in Figure 3. The first and most obvious difference between the rock and the air is the


Summer 1992/93
30
20
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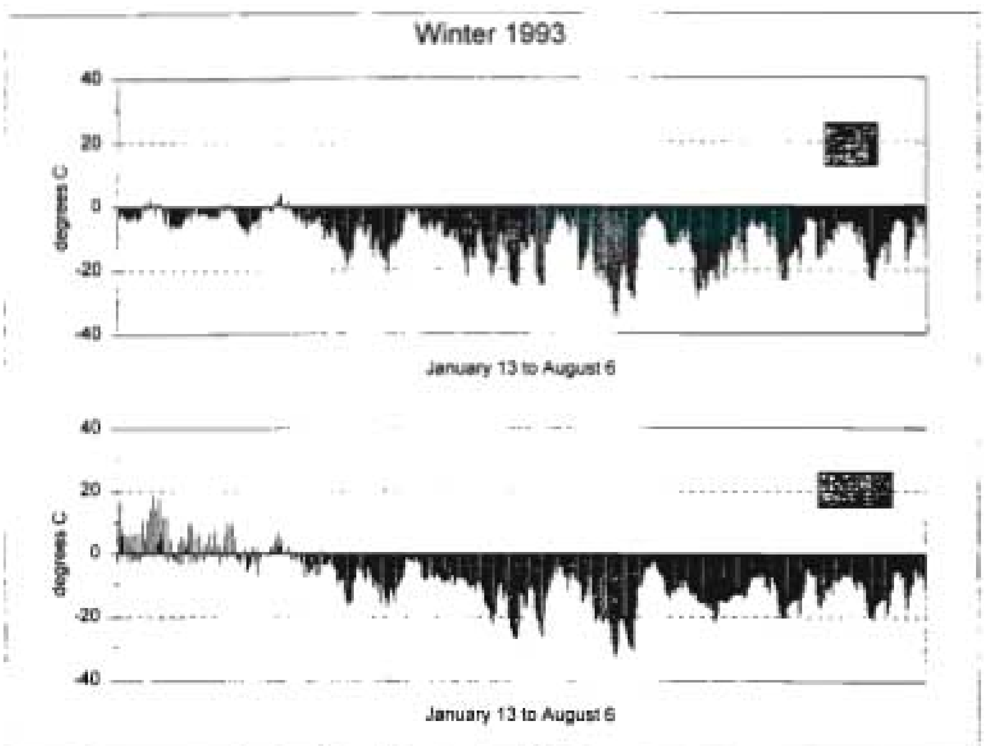
Winter 1993
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20
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40
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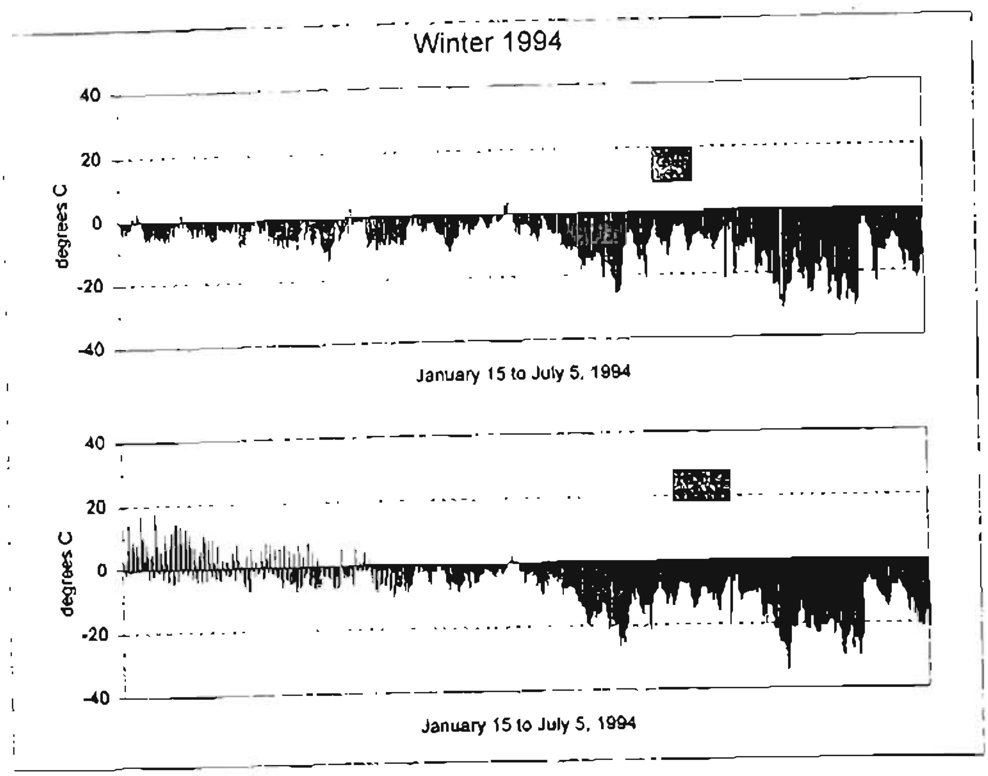

Figure 3 Campbell temperature records (one hour intervals) for the summer of 1992–93 and for the winter records of 1993 and 1994.

amplitude of the rock temperatures and the difference in crossings of the 0 °C isotherm. Considering the summer first, details of the rock surface and air temperature values are given in Table 1. From this it can be seen that rock temperatures are significantly higher than those of the air (5.86 versus −0.24 °C) and that the maximum rock temperature is noticeably higher than that of the air (24.38 versus 6.03 °C); the range of the rock temperatures is more than twice that of the air. Using a threshold of 0 °C (but not implying any geomorphic response), it is interesting to see that both the air and the rock experienced 16 freeze–thaw cycles – but *not* always at the same time. The rock temperature data clearly show diurnal variability but do not always drop below 0 °C; the air temperatures, however, do not so clearly show diurnal variation and for two periods remain below 0 °C whilst the rock temperatures fluctuate across this boundary. Further, the amplitude of the sub-zero rock temperatures is, with only one or two exceptions, small (*c.* 2 °C) and so their geomorphic significance is very limited. Conversely, in considering the air temperatures, if the small (and frequently very short) sub-zero depressions are ignored the air still indicates 12 freeze–thaw cycles of amplitudes $\geqslant -5$ °C – a value clearly more likely to be geomorphologically significant, the rock experiencing only one cycle during this period. Thus, the air temperatures are clearly *not* a surrogate for what is happening to the rock (this will be discussed in more detail later). In short, the available data indicate that, as a purely thermal event, the summer is *not* a particularly active period for freeze–thaw weathering in this area. Lastly, even if and when freeze–thaw cycles of adequate amplitude and duration are found to occur in the rock (presuming water is also present) the one hour record intervals are not adequate to provide the

Table 1 Analysis of temperature data for the summer of 1992–93 (°C).

	Mean	*s*	Range	Min.	Max.	Cycles[1]
Air	−0.24	2.60	12.17	−6.14	6.03	16
Rock	+5.86	5.85	28.79	−4.41	24.38	16

[1] Number of freeze–thaw cycles is here based on crossings of 0 °C boundary. See discussion for significance or otherwise of these values.

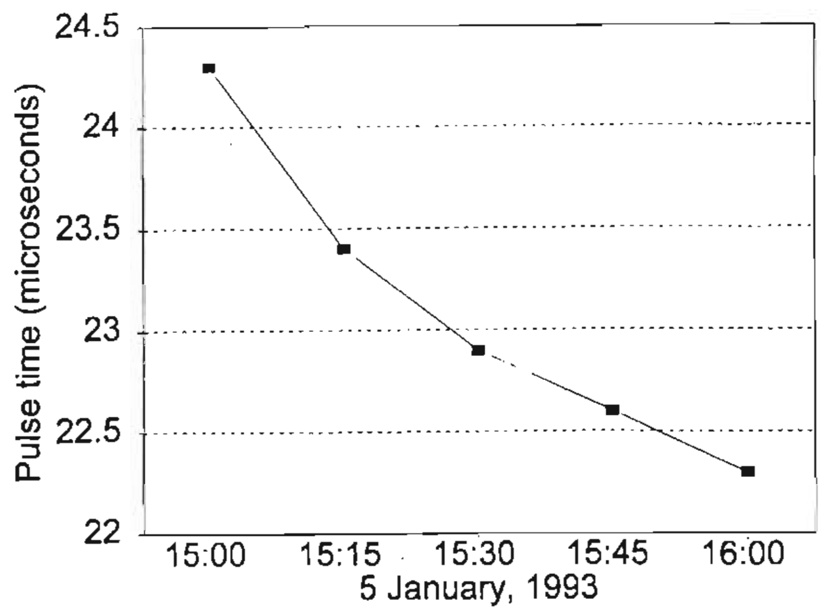

Figure 4 Ultrasonic pulse time to show the impact of water on the rock (a decrease in pulse time equates to an increase in pulse velocity owing to the presence of the better propagating medium, water).

information necessary to determine the actual freeze–thaw mechanism (see below).

In terms of weathering processes, the frequency of freeze–thaw weathering is limited by thermal constraints alone; however, moisture availability constrains the process even further. This being a "dry valley" snowfall is low (see below) and moisture is extremely limited. Some summer snow was seen to melt on contact with rock on the valley bottom but the rock temperatures remained above 0 °C and the moisture was subsequently lost due to the radiative heating of the rock. The *only* source of water on the valley bottom was from glacier melt and this only occurred on the warmer days with high radiation inputs. The impact of the meltwater on the rocks is clearly shown by the ultrasonic data (Figure 4). As water enters a rock so the pulse time decreases as water is a better transmitting medium than air. On 5 January as the glacier meltwater ran over the rock block with ultrasonic transducers so the pulse time can be seen to drop from 24.3 μs to 22.3 μs over one hour as the water penetrated the rock. Both air and rock temperatures remained above 0 °C and so the only forms of weathering likely to occur are wetting and drying and, possibly, salt weathering. During sub-zero periods the rocks were dry and ultrasonic data indicated an absence of water. Unfortunately the one hour record intervals are not adequate for any consideration of weathering by thermal stresses (see below).

The valley floor data for the winter periods are extremely interesting and very valuable as rock temperature data, particularly from the Antarctic, are rare. In broad terms the 1993 data are very similar to those of 1994 (Table 2) – there being a period of frequent crossings of 0 °C for the latter part of the summer in to autumn (roughly mid January through to early March) followed by the winter freeze. However, it is very clear that both air and rock temperatures fluctuate significantly through the winter freeze period and can even become positive, albeit for only a very short period (as in 1994).

Considering the first part of the records (i.e. the summer to autumn section) both years clearly show frequent crossings of 0 °C for the rock data but significantly fewer for the air. Detailed graphs for the early period in both 1993 and 1994 (Figure 5) indicate that the rock surface experiences nearly a

Table 2 Analysis of the temperature data for the total record for the winters of 1993 and 1994 (°C).

	Air 1993	Rock 1993	Air 1994	Rock 1994
Mean	−9.65	−8.72	−9.23	−7.89
s	7.19	7.97	7.22	8.82
Range	39.85	54.9	35.55	56.48
Min.	−35.18	−33.28	−31.71	−35.14
Max.	4.67	21.62	3.84	21.34

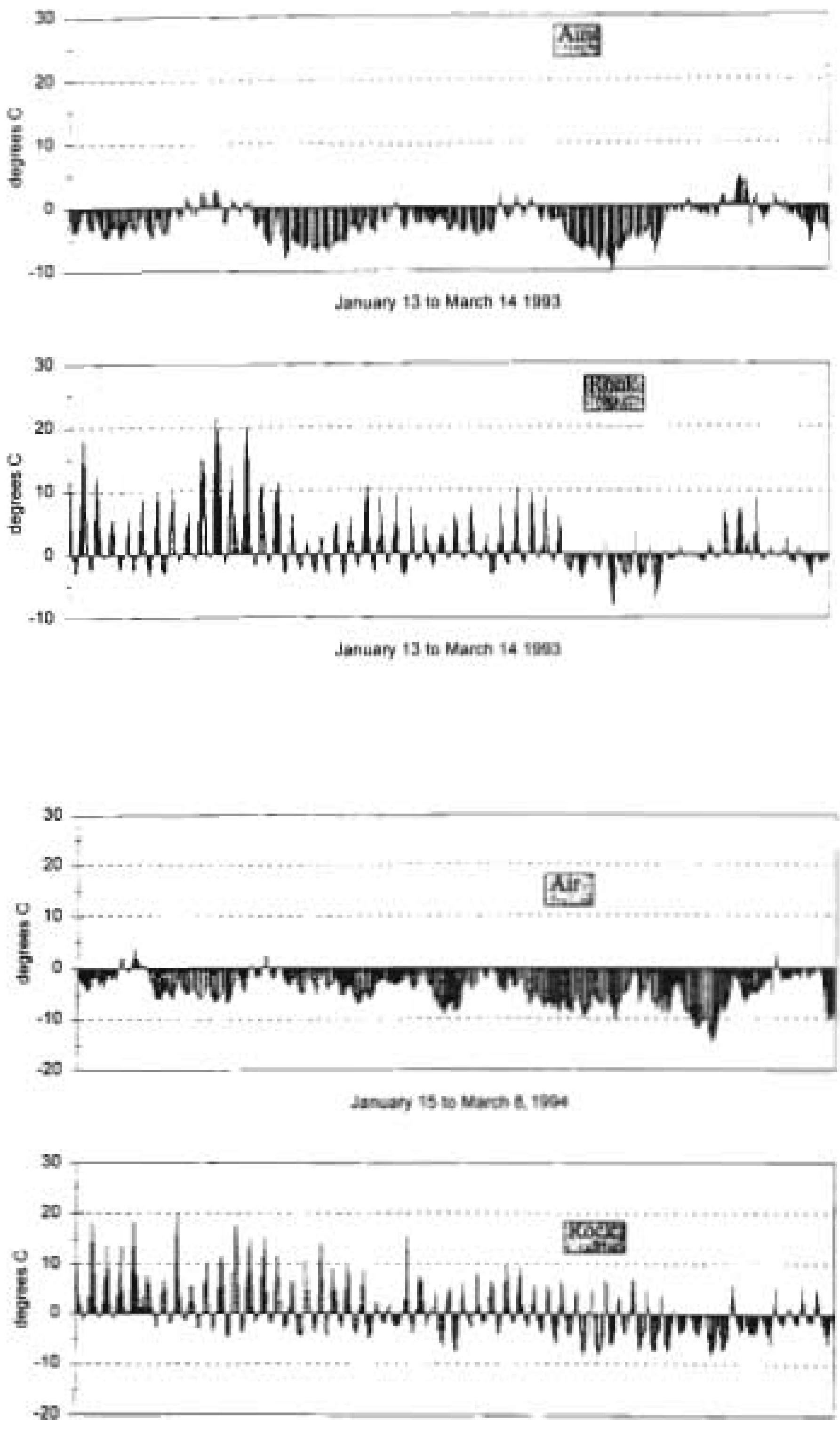

Figure 5 Detail of the late summer through to early winter period of the air and rock temperatures (at one hour intervals) for the winters of 1993 and 1994 (the first section of the winter graphs in Figure 3).

Table 3 Analysis of the temperature data for the periods of across 0 °C oscillations for 1993 and 1994.

	Air 1993	Rock 1993	Air 1994	Rock 1994
Mean	−2.65	1.51	−4.31	0.54
s	2.64	4.61	2.93	5.19
Range	14.54	29.91	17.94	30.06
Min.	−9.87	−8.29	−14.28	−8.72
Max.	4.67	21.62	3.66	21.34
Cycles 0°C[1]	14[2]	46	7[3]	51
Cycles −2.5°C	22	20	23	39
Cycles −5°C	8	2	20	15

[1] Freeze–thaw cycles derived from the record in Figure 5 using thresholds of 0 °C, −2.5 °C and −5 °C.
[2] From 51 days of record (13 January to 4 March 1993).
[3] From 53 days of record (15 January to 8 March 1994).

Table 4 Details of temperature records for (a) a period with no apparent snow cover and (b) a period with snow cover (°C).

(a) Data for 5 May to 20 May 1995 to show strong relationship between air and rock: correlation $r = +0.937$.

	Air 1994	Rock 1994
Mean	−8.78	−10.95
s	3.58	3.11
Range	19.10	16.69
Min.	−21.33	−21.95
Max.	−2.23	−5.56

(b) Data for 11 June to 6 July 1993 to show "buffering" of the rock by snow: correlation $r = +0.787$.

	Air 1993	Rock 1993
Mean	−13.77	−12.95
s	6.40	3.47
Range	26.61	17.98
Min.	−29.56	−21.75
Max.	−2.95	−3.87

diurnal cycle whilst the air only rises above 0 °C on a few occasions. Details of the thermal conditions for these periods are given in Table 3. From the table it can be seen that the data for 1994 suggest more severe conditions, with lower temperatures and far more (thermal) freeze–thaw cycles on the rock. It is interesting that 1994 differs from 1993 in that it indicates the danger of using data for only one year (and clearly data from more than the two used here would be far better). Using the arbitrary and non-geomorphic boundary of 0 °C it can be seen that the rock surface experiences many more cycles than does the air (32 more in 1993 and 44 in 1994). Thus the use of air temperature would, with this threshold, give a very false picture. Using a threshold of −2.5 °C, which is potentially geomorphologically meaningful, 1994 still shows more (16 more) on the rock than in the air but 1993 has the rock with marginally fewer (only 2). Then, using a threshold of −5 °C, at which some water will likely freeze (if water is present), the air indicates more cycles than the rock experiences in both years (6 more and 5 more respectively). Thus, the air is not reflecting what is happening on the rock. In all instances here the discussion is purely with respect to the *thermal event* and no presumption regarding the presence of water and thus the geomorphic effectiveness is made.

Following the early winter period of intense thermal oscillations both the air and rock temperatures drop, to remain, with the one short exception in 1994, below 0 °C. However, the records indicate some very interesting activity that has geomorphic implications. First, they indicate that for much of the time there cannot be any significant snow on the ground (as suggested by the limited snow observed in the field in December). The air and the rock records both show a "peakedness" with a strong correlation (values as high as +0.937 °C being obtained with only a marginal difference between actual temperature values: Table 4a). Conversely, for short periods the rock temperature data are "smoother" and less spiky than the air and this is thought to indicate the presence of snow (Table 4b). However, although snow is considered to be present it cannot be very thick as it does not obliterate completely the fluctuations shown in the air temperatures ($r = +0.787$) and its influence can rapidly disappear which, as it cannot be due to melting (owing to the sub-zero temperatures and the limited radiation input in winter), suggests it is removed by wind. Wind speed data for the time of the transition from "buffered" to "spiky" tend to support this as wind speeds changed from values close to 4 m s^{-1} to values close to 18 m s^{-1} at the time of transition. Thus, during the winter the rock can experience substantial temperature changes, particularly as it is not protected by a thick snow cover. This also means, as observations during 1992–93 suggested, that there is very little snow available in the valley or on the nunataks to melt during the summer and provide water for the rocks.

If the winter data as a whole are considered, it can be seen that there are significant oscillations of temperature within the rock, albeit all below 0 °C.

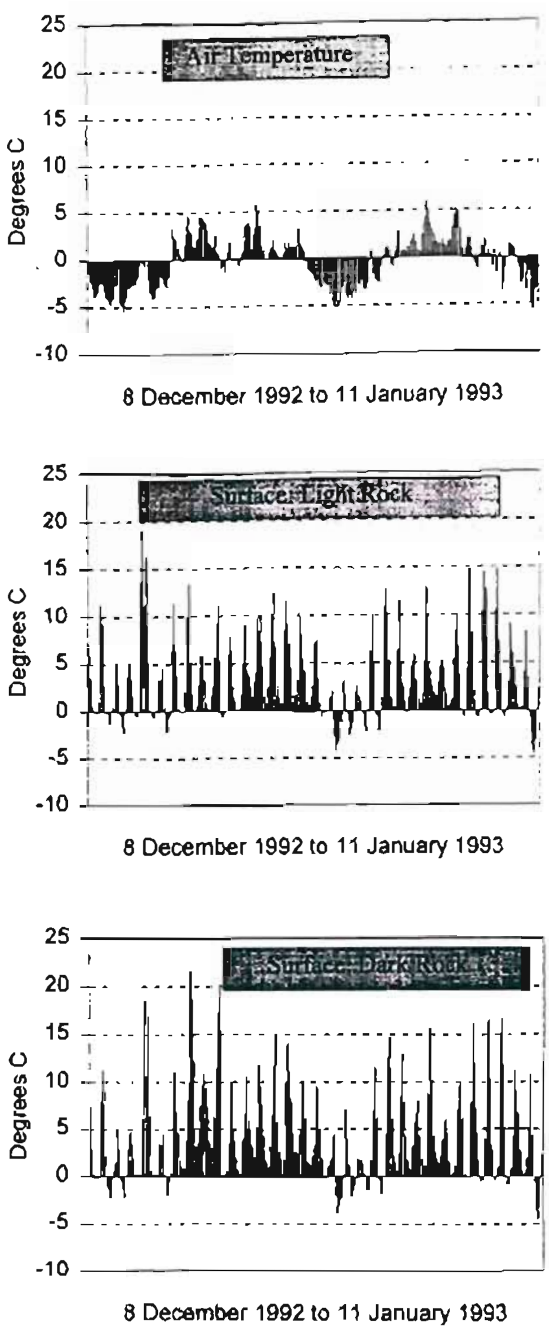

Figure 6 Graph of the temperatures recorded on a light-coloured and a dark-coloured sandstone, located on the valley floor, for the 1992–93 summer (at one hour intervals).

If there is any water available in the rock then it is not pure and so it will freeze and thaw at a temperature below 0 °C. Without data on the type and amount of salts in solution (assuming that there is any water in the rock) it is not possible to calculate the freezing point depression. However, as the oscillations can, particularly in 1994, come close to −1 °C it is possible that thawing of any water could occur. Further, as the stress that any frozen water can exert is a function of the temperature (with maximum stresses being exerted in the region of −22 °C), so the fluctuations between *c.* −5 °C and *c.* −25 °C indicate that there is a change in the stress regime that could cause damage to rock. All of this, it is emphasized, is dependent upon water being present in the rock and here it seems unlikely that any is actually available in this area.

Lastly, from the valley floor data there is the information regarding the difference between the light-coloured and the dark-coloured sandstones (Figure 6). So far these data are only available for the summer of 1992–93. The graphs clearly show the diurnal fluctuations, which are less pronounced in the air, as well as the larger amplitude of the events on the dark rock compared with the light-coloured rock. Details of the temperature data (Table 5) show that the dark rock was generally warmer, had a marginally larger range and did not become quite so cold, although many of these differences are actually very small. The light-coloured rock, based on a 0 °C threshold, experienced marginally more freeze–thaw cycles than did the dark rock. The differences were not as large as had been expected but may have been more significant had readings been taken at more frequent intervals (see below) in so far as there may be implications for thermal stress fatigue/shock if the dark rock warms up faster.

Table 5 Details of rock temperatures for the dark- and light-coloured sandstones (together with air temperatures) for the summer of 1992–93 (°C).

	Air	Light rock	Dark rock
Mean	−0.24	+3.5	+4.61
s	2.60	4.37	5.12
Range	12.17	26.96	27.75
Min.	−6.14	−4.62	−4.72
Max.	+6.03	+22.34	+23.03
Cycles (0 °C)	16	16	14

High frequency data from the valley side. For the period in the field it was possible to collect temperature data on the north-facing valley side at two minute intervals. Data of this frequency (or, preferably, at a higher frequency) are required for any real understanding of weathering processes in

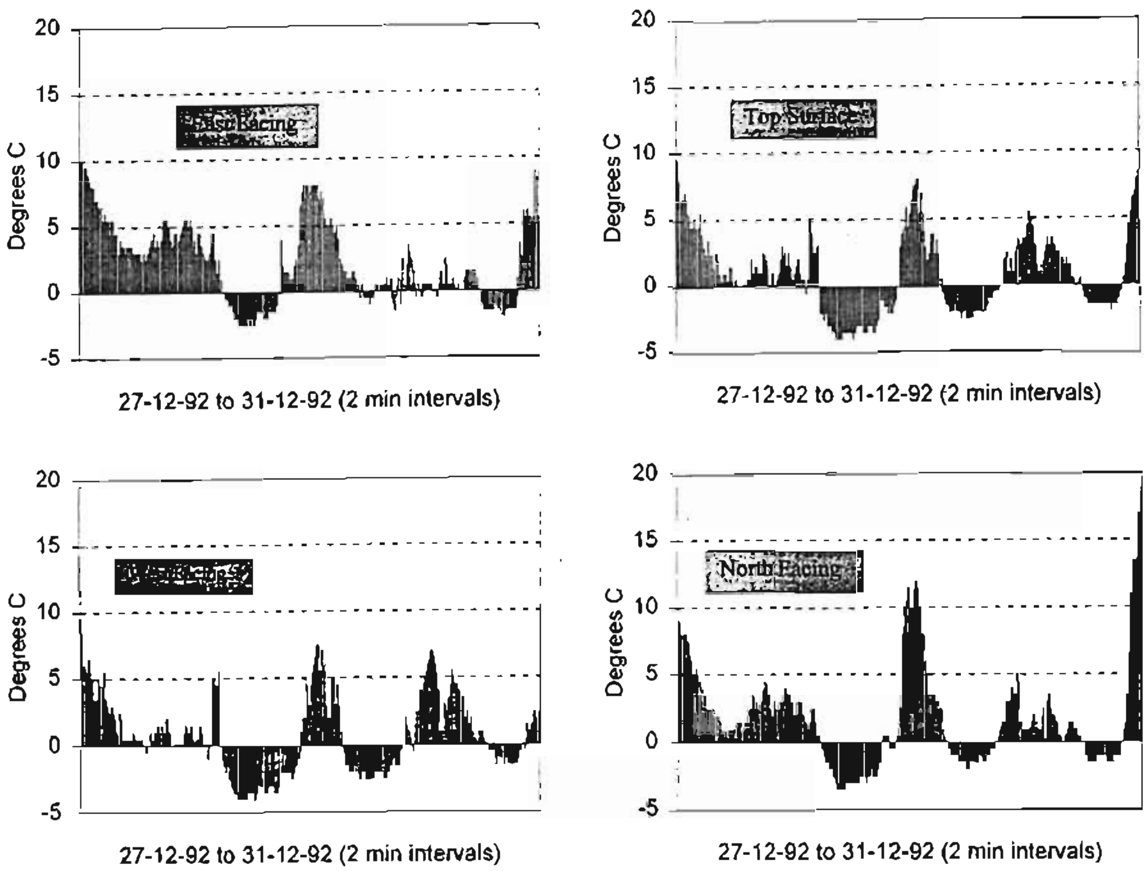

Figure 7 An example of a five day record of temperatures for east, west and north aspects, together with a horizontal surface, collected at two minute intervals.

the field. This is for three main reasons. First, they provide information on the short term changes that are missed by observations at even five minute intervals and are certainly lost at measurement intervals of 15 minutes or more. Second, they provide the sort of information required for evaluation of weathering due to thermal stress/shock. Lastly, data of this sort are required if any understanding of the freeze–thaw mechanism is to be attempted as all the available processes are constrained by different rates of fall of temperature, the information for which cannot be derived with any certainty from temperature records collected at intervals of 15 minutes or more.

Figure 7 shows a typical example of five days records with observations every two minutes for the east-, west- and north-facing sides of a rock outcrop together with the top (horizontal) surface. Using the 0 °C boundary it can be seen that the east-facing side experienced eight freeze–thaw cycles, the west-facing seven, the top surface five and the north-facing side only four. Thus, the different aspects would (given the presence of water and if freezing occurred) experience different weathering regimes. If a more realistic threshold of −3 °C is used then the four locations show very similar results with between one and two events each. It is very clear that the north-facing exposure experiences the highest temperatures as well as the largest range (Table 6). The ranges for both east- and west-facing exposures are comparable but the west experiences lower minimum and maximum temperatures. The horizontal surface, which

Table 6 Analysis of temperature data for north-, east- and west-facing exposures together with a horizontal surface for the period 27 to 31 December 1992 (°C).

	Mean	s	Range	Min.	Max.
East-facing	2.07	2.953	12.5	−2.5	10.0
West-facing	0.65	2.759	12.5	−4.0	8.5
North-facing	1.74	3.759	23.5	−3.5	20.0
Horizontal	0.77	2.869	14.5	−4.0	10.5

receives low angle radiation for most of the day, has values, except for the mean, higher than either east or west. The east-facing exposure has the highest mean value, possibly as a result of receiving radiation input at an angle very close to normal to the face such that, although it does not achieve the maxima experienced by the north-facing exposure, it does maintain a relatively warm, stable condition during the period of direct sunshine. Although the northern exposure achieves the highest temperatures they are of shorter duration than those of the eastern exposure, owing to the short duration of optimum radiation, and so the mean is diminished.

Using the same data presented in Figure 7 for the north-facing exposure, analysis was undertaken of the influence of recording interval on the resulting record. The original data were collected at two minute intervals but by extracting the data value at the 10, 20, 30 and 60 minute record points so it was possible to simulate data records of that time interval (Figure 8). These derived data could then be used to show the difference in thermal regime that the record interval produces. Analysis of the extracted data from each of the data sets is given in Table 7. Although the example presented here is not as clear as when data are recorded at one minute intervals (analysis of such data from another site is in progress), nevertheless it can be seen that the information degenerates with increasing record interval in a number of ways. First, and most obvious, the shorter term variations are lost with each increasing step (and two minute records lose much over one minute records). This is particularly important in determining the role of thermal stress and in trying to assess what freeze–thaw mechanism might be operative. For thermal shock, a rate of change of temperature in the order of 2 °C min^{-1} is required (Richter and Simmons, 1974; Yatsu, 1988; Hall and Hall, 1991) and this can *only* be ascertained if record intervals are two minutes or less (one minute or less is preferable). From this current record no such occurrences were found (but see below regarding taffoni temperatures), although the two minute record does show an example of 5 °C in four minutes which, had one minute intervals been used, might have shown a value sufficient for thermal shock (as *has* been found with the ongoing analysis of one minute data). The resolution of one minute data is far superior to that of the two minute data presented here such that even this level of data recording is seen as inadequate. With respect to freeze–thaw weathering, it is imperative that the rate of fall of temperature during freezing be known. It is possible to derive a good indication of the rates from the two minute data (Figure 9) which shows that, in all three freeze events, a rate of fall of temperature of 0.5 °C per two minutes or 0.25 °C min^{-1} occurred. That all graphs show this 0.5 °C stepwise decrease in temperature suggests that this is the optimal sensor resolution with this observation interval (although other data from the same logging system, to be presented elsewhere, indicate otherwise). Nevertheless, a rate of fall of temperature of 0.25 °C min^{-1}, assuming any water were present, would suggest that the hypotheses of such as Walder and Hallet (1985), Connell and Tombs (1971), and Powers (1945) would not be operative as the rates recorded are in excess of those that these mechanisms require, whilst mechanisms such as that of Battle (1960) might be possible (requiring rates of fall $\geqslant 0.1$ °C min^{-1}). Recognizing that the presence of water is here considered unlikely, this conjecture regarding process is based solely on the thermal event and, if water were present, it would be in the very outer shell of the rock near to where such surface temperature values occur and so the judgement would be realistic. Identification of the freeze–thaw process without such data to confirm its operation is highly questionable.

Table 7 Analysis of temperature data at time intervals of 2, 10, 20, 30 and 60 minutes (°C).

	Mean	s	Range	Min.	Max.
2 minutes	1.7	3.74	23.5	−3.5	20.0
10 minutes	1.8	3.77	22.0	−3.5	18.5
20 minutes	1.8	3.75	22.0	−3.5	18.5
30 minutes	1.8	3.73	21.0	−3.5	17.5
60 minutes	1.7	3.63	19.5	−3.5	16.0

The significance of this two minute record interval in elucidating process can be shown by the data recorded in the taffoni (Figure 10) where one thermistor was located (always in the shade) at the top of the hollow and the other (sometimes directly in the sun) at the bottom of the hollow. The two sets of data show a very close similarity ($r = +0.942$) but it can be seen that the bottom location has generally higher temperatures and lower freeze amplitudes, and is also more "spiky" at times (Table 8). Both locations experienced two freeze cycles and, although the range of temperatures for both positions was the same, the bottom sensor had higher mean and maximum values together with a lower freeze amplitude. The difference between the two locations within the

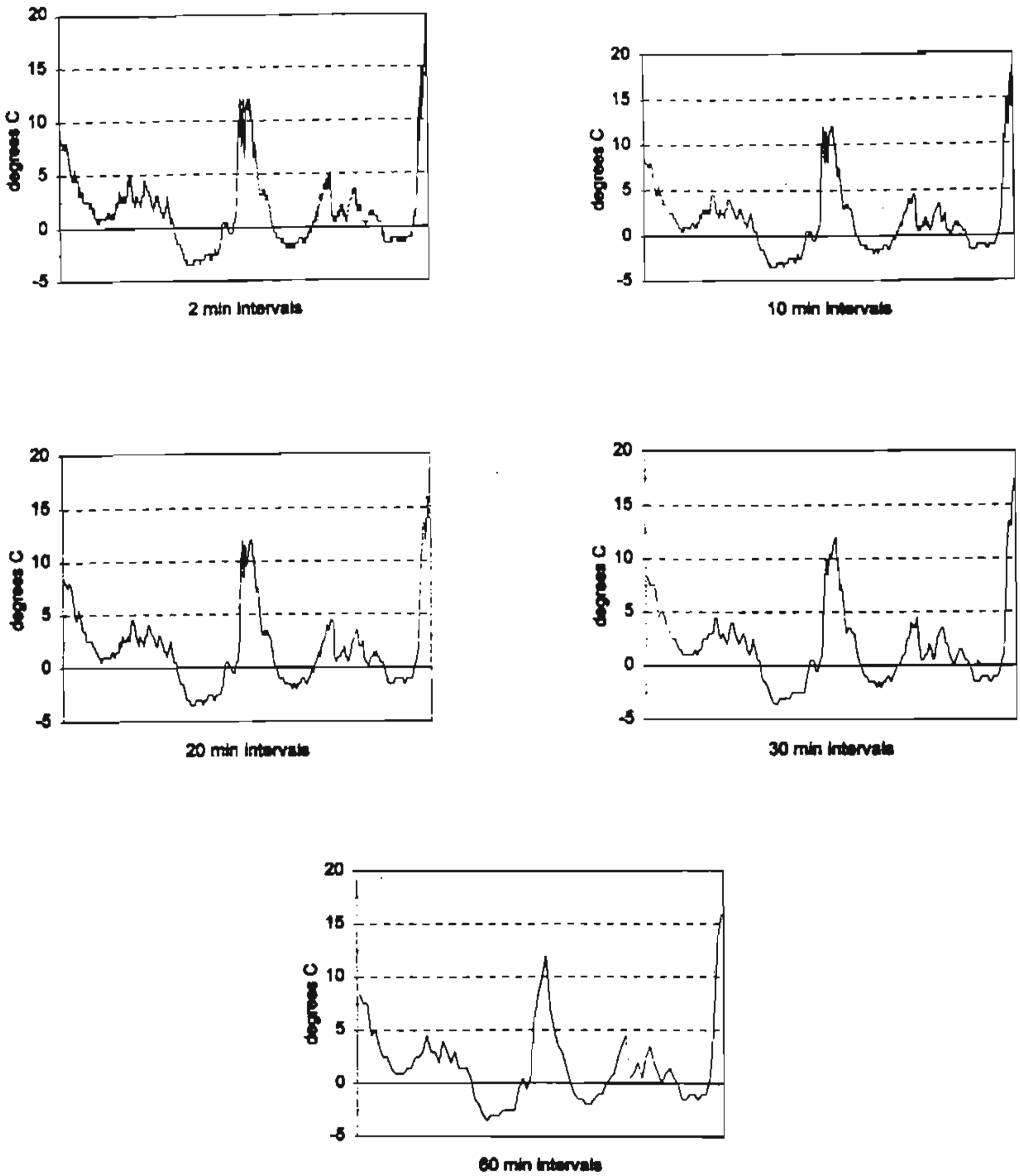

Figure 8 Graphs to show the sort of records, extracted from the original two minute data, that would have resulted from recordings at 10, 20, 30 and 60 minutes.

taffoni is shown clearly by the record detail given in Figure 11. Here the difference in temperature at midday, when receiving direct radiation, can be seen. Such differences may have implications for the processes operative in the development of taffoni but this will be dealt with in detail elsewhere. What is of major significance is the large cooling and warming event at minutes 8 to 12

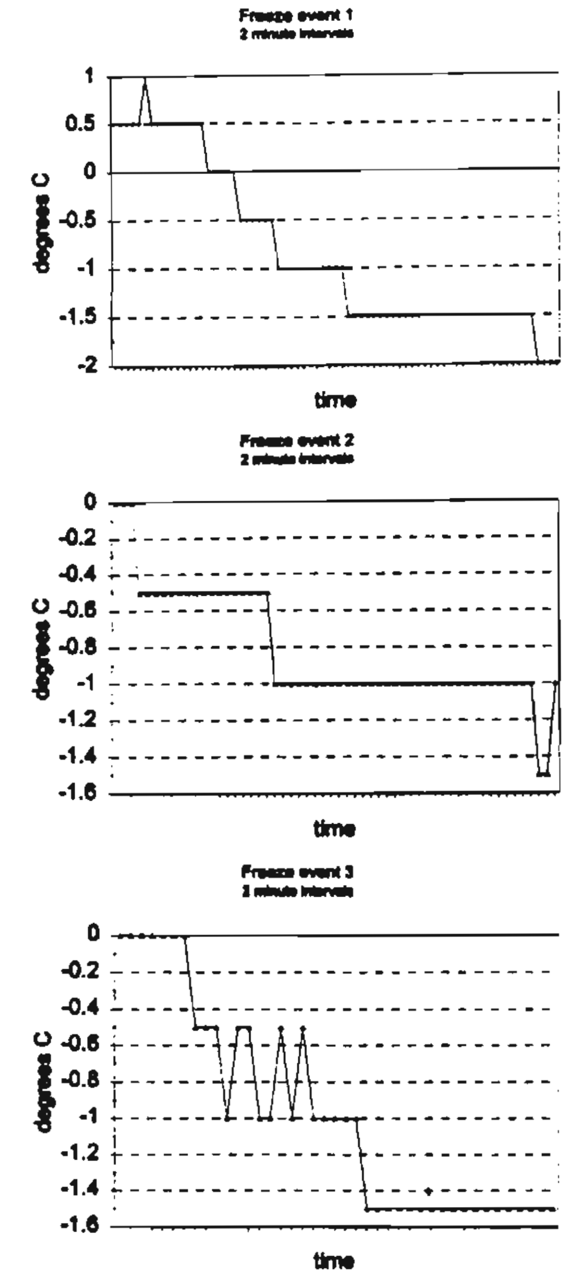

Figure 9 Examples of the rate of fall of temperature as can be extracted from freeze events shown in Figure 7.

(records being at two minute intervals and the first record being considered as record 0), where there is a 2 °C min^{-1} fall in temperature followed by a >2 °C min^{-1} rise in temperature. Thus, it can be shown that, at the rock surface, temperatures alternated at rates suggested to be required for thermal shock (see above). The significance (discussed in detail below) is that, by using such a high frequency record interval it is possible to show: (1) that thermal shock can occur, and (2) that it is occurring in an area where freeze–thaw is usually (qualitatively) thought to occur. Such findings would *not* have been possible with record intervals greater than two minutes and thus any study that uses longer record intervals is not able to judge the full scope of the operative processes.

Table 8 Analysis of the temperature data from the taffoni for the period 27 to 31 December 1992 (°C).

	Mean	s	Range	Min.	Max.
Top	0.77	2.54	12.0	−3.5	8.5
Bottom	1.60	2.38	12.0	−2.5	9.5

The two minute data also provide information regarding the influence of depth as recordings were taken at 5, 10, 15, 30 and 40 mm (Figure 12). The depth of 40 mm was used partly for practical reasons and partly because it was believed that this encompassed the zone in which the majority of weathering activity would occur in this area. The main feature, as might be expected, is that of diminishing of freeze and thaw amplitudes with depth as well as a decrease in "peakedness". Detailed analysis of the data shows that there is never more than a 1 °C temperature differential between the surface and 40 mm depth and that usually it is only 0.5 °C. The time lag for the 40 mm depth to reflect the surface temperature varied, with it taking 10 minutes for a surface temperature of −0.5 °C to reach 40 mm, 18 minutes for −1 °C, 56 minutes for −1.5 °C, 28 minutes for −2 °C and 28 minutes for −2.5 °C. The important factor here is that there was never, within the outer shell (during the available record periods and for this position), a large temperature differential within the outer 40 mm of the rock. Thus, at least during this time period, thermal stresses resulting from large temperature gradients seem to be absent. However, it is considered too small a data sample for this determination to have any meaning other than for the record period under consideration.

The Schmidt hammer rebound values and the measurements of taffoni are to be dealt with in detail elsewhere but some of the data are, in synopsis, presented here as they reflect the impact of weathering (Table 9). The most noticeable component of the Schmidt hammer data is that the north-facing exposure has the lowest rebound value, indicating the greatest amount of weathering. This is reflected in the taffoni measurements,

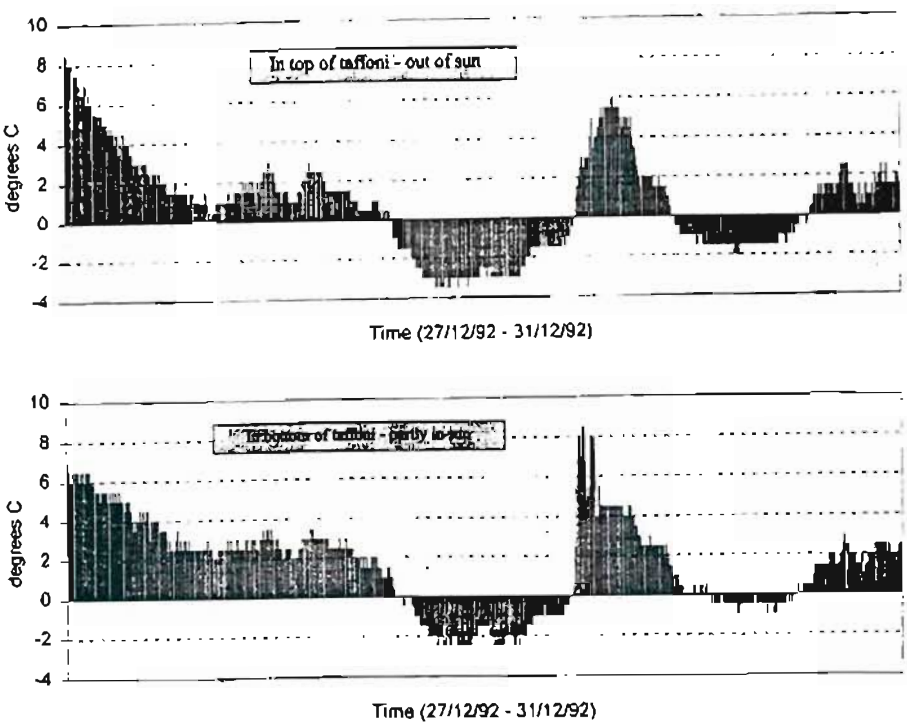

Figure 10 Temperature data from the top and bottom of a taffoni (data at two minute intervals).

where the largest taffoni are seen on the north-facing exposure. The large standard deviation associated with taffoni width for this exposure is due to the taffoni being notably "elongate" along the exposure. The west-facing taffoni were also elongate but did not have such a spectrum of sizes (Table 9). The taffoni on the north-facing exposure were also deeper than the other exposures although the values for the western aspect come close and actually exceed the north for heights. It was also noticeable that the eastern exposures had very few taffoni on them and that the northern and western had the most, with, frequently, the northern face having marginally more. These data tend to indicate the greatest amount of weathering, in terms of both taffoni size (and numbers) and low rebound values, on the north-facing exposure, closely followed by the western aspect and with the eastern exposure significantly less weathered as shown by the higher rebound values and the smaller (and fewer) taffoni. Thus, whatever the processes that are causing weathering in this valley, they clearly have the largest impact on the northern aspect, a very similar impact on the western aspect but minimal impact on the eastern aspect.

These data regarding taffoni and Schmidt hammer rebound values, indicating significant weathering on the north- and west-facing aspects are most interesting. A study on the nunatak tops regarding "cryoplanation terraces" (Hall, In Press)

Table 9 Synopsis of Schmidt hammer data and taffoni measurements.

	Schmidt		Taffoni (cm)					
	$\bar{x}$[1]	s	Depth	Depth s	Width	Width s	Height	Height s
North-facing	15	2.6	11.0	7.6	27.9	29.8	10.9	7.9
East-facing	22	2.7	3.7	2.1	5.5	2.9	4.4	2.9
West-facing	20	3.1	12.3	6.2	22.8	9.6	11.1	3.9

[1] Schmidt hammer rebound value (type N hammer).

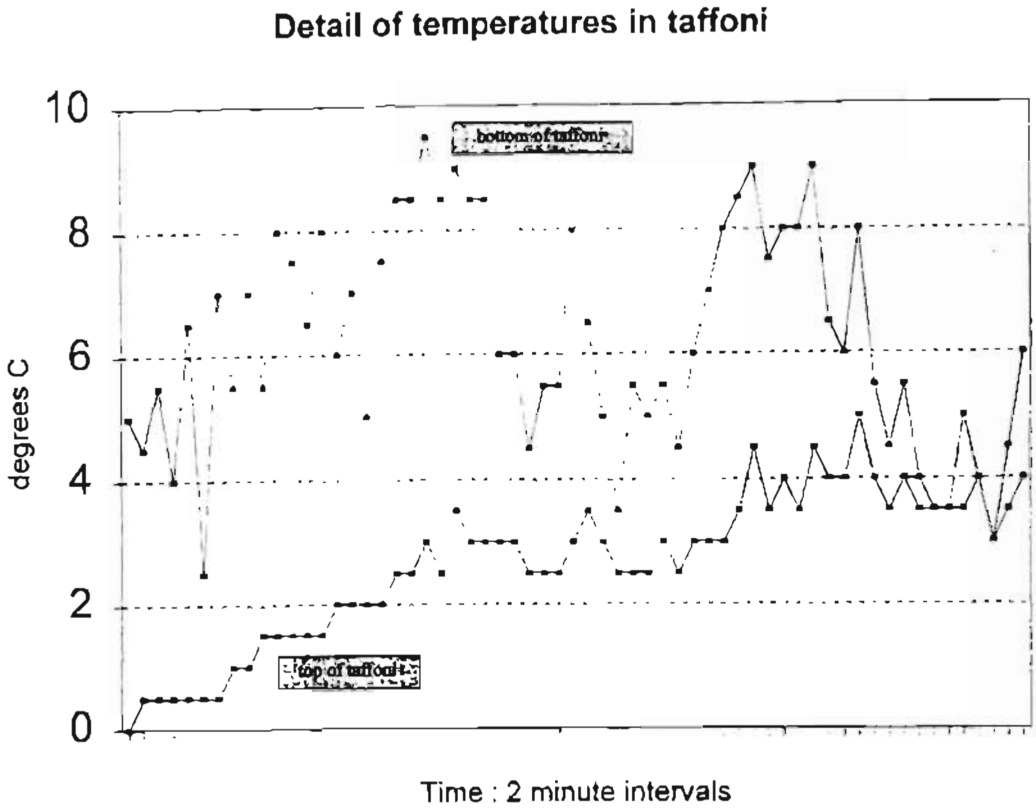

Figure 11 Detail of the temperature differences between the top and base of the taffoni together with an example of a rate of change of temperature $\geq 2\,°C\ min^{-1}$.

clearly showed that these terraces only occurred in an arc from north through to west/south-west (010° through 360° to 250°) and with *none* orientated to the east (045° through 90° to 200°). As the terraces occurred along lithologic junctions on horizontally bedded sediments that outcropped on all aspects of the nunatak, their occurrence must be related to process. Further, where the data outlined above were collected (in the valley bottom) there are no terraces, probably as a result of not having been exposed for sufficient time (by glacier retreat). With elevation up the side of the nunatak so there is a specific size distribution of these features, with the largest being found at the highest elevations (i.e. furthest from the retreating glacier). The implication is that the process association, or at least the weathering component of that process association, that is responsible for the terraces is operative, as a function of aspect, in the valley bottom. As snow is almost non-existent (see discussion above) and was certainly not seen on the valley wall at the start of summer, it is unlikely to be freeze–thaw weathering within nivation (a process normally cited as operative in cryoplanation terrace formation) that is a major weathering component. As an aside, it could be suggested that the main transport component(s) of the process association responsible for the terrace formation (solifluction and wash) results from the small body of snow that begins to accumulate (as seen in the field) along the riser of the terraces. Significantly though, the riser is only present once cutting back by weathering (of some form) and removal of the products has taken place. This is very important for, on the eastern and southern aspects weathering has not yet developed a riser; consequently terraces are not found on those aspects despite uniformity of rock outcrop. Whatever the process synergy and sequence, it seems clear that the indirect weathering evidence supplied by Schmidt hammer rebound values and the size and distribution of taffoni indicates a preferential aspect-controlled weathering regime that is highly unlikely to be dominated by freeze–thaw. This is contrary to nearly all discussions regarding the perceived formation of cryoplanation terraces Hall.

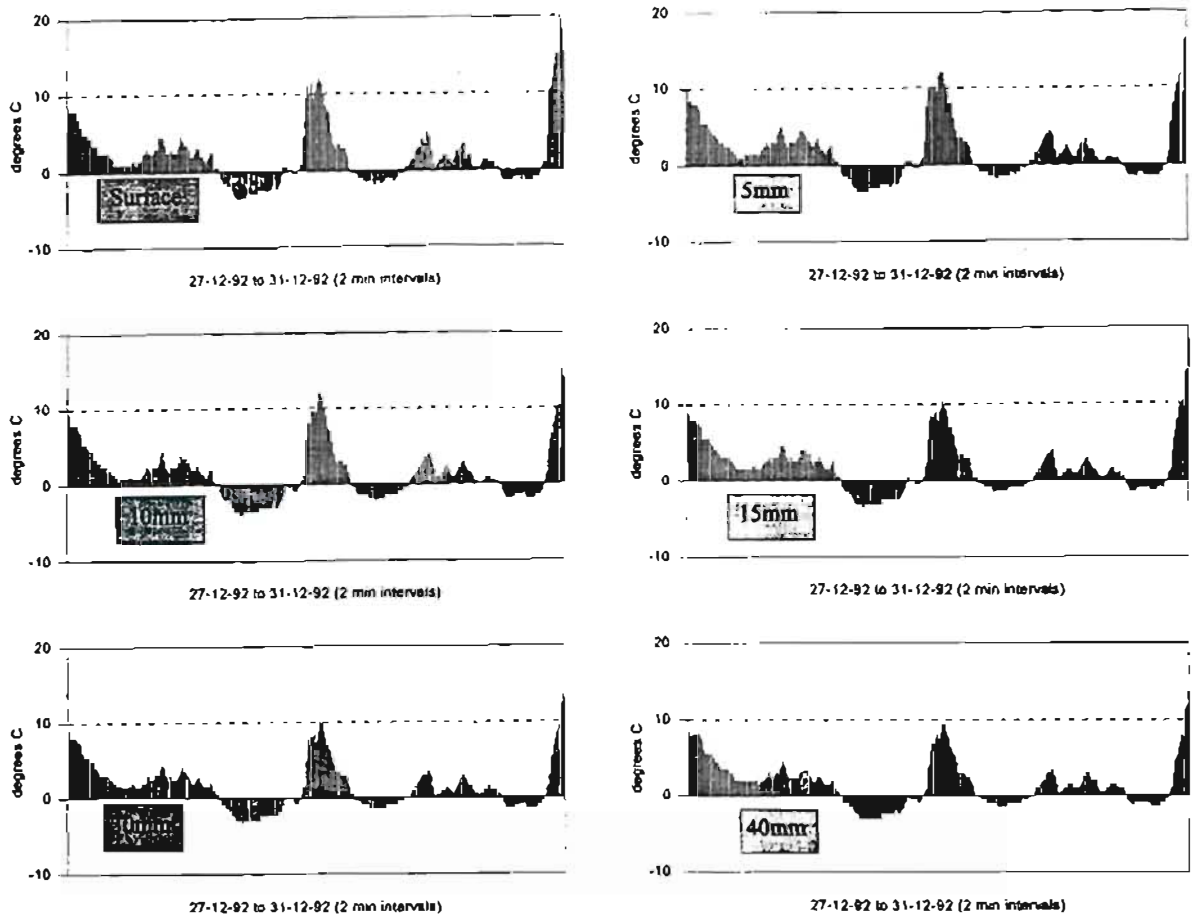

Figure 12 To show the variation of temperature with depth (from the surface down to 40 mm) based on two minute recordings.

One important aspect of weathering in this area that has not been considered is that of salt weathering. Being a "dry valley", with all the associated limitations on water and the high evaporative conditions, there is every expectation of salt weathering being operative. This is all the more so as thick layers of gypsum were found on some exposures, especially at the undercut bases of rock outcrops where snow was found to reside (owing to the protection from the sun). The large diurnal temperature ranges monitored for the rock surfaces indicate great potential for the thermal expansion and contraction of the salts in cracks and pores near the surface of the rock. Further, the limited but occasional wetting of the surface zone of the rock would also facilitate hydration expansion and contraction of those salts. Thus, there is every expectation that salt weathering is operative in the valley and, potentially, on the rest of the nunatak as salt efflorescences were observed on the treads of some terraces. However, beyond the identification of the dominant deposit (gypsum) no other data or information are available and so it is not possible to quantify or more rigorously interpret the role of salt weathering.

Antarctic Weathering

In the Antarctic, as with most cold regions (see Hall, 1995), it is frequently argued that the cause of rock breakdown is a result of freeze–thaw weathering although, as van Autenboer (1964) states in the opening paragraph, many Antarctic workers are not so sure of its efficacy owing to the widespread aridity of the continent. It is worth noting three examples of early Antarctic studies that cite other than freeze–thaw as the operative mechanism for rock breakdown in this environment. First, Bernacchi (1901, p. 107) writes, "The daily change in temperature caused the porous volcanic rocks of the cliffs to alternately expand and contract, and the rapid nocturnal contraction produced such a superficial strain as to cause the surface to crack, peel off in irregular pieces, and fall." Second, Shackleton (1909, p. 294) states, "Such a diurnal range of temperatures, combined with the effects of summer thaw followed by the severe frosts of winter, exerts a powerful disrupting force upon the rocks." Finally, Gunn and Warren (1962, p. 60), in a discussion of the breakdown of fine-grained rocks, note that, "Although experimental evidence suggests that the forces exerted by rapid temperature change are too small to fracture rocks ... it is difficult to find an alternative explanation for some ... features." Others who reference the role of thermal fatigue/shock include Avsyuk *et al.* (1956), Markov and Bodina (1961), Stephenson (1966), McCraw (1967), Black (1973), Miotke (1980; 1982; 1984), Aleksandrov and Simonov (1981), Robinson (1982), Miotke and von Hodenberg (1983), Blümel (1986), Brunk (1989) and Picciotto *et al.* (no date). In fairness, the list of authors who refer to the (suspected) role of freeze–thaw is substantially greater! Of this latter group, many *do* provide good evidence in favour of freeze–thaw where the prime constraint is not temperature but rather the presence of water. Many of the early explorers *did* note the occurrence of water on the rock; for example, Armitage (1905, p. 172) states, "Water could be heard trickling down amongst the rocks; it had formed a pond at the base of the cliffs about 4 inches deep" and observes that the rocks in this area were "exceedingly weathered". The key here is not that water is *totally absent* but rather, as Scott (1905, p. 141) observed, that thawing only lasts for a very short time (about three weeks in his observations) and is highly spatially constrained; these same points are more recently reiterated by Balke *et al.* (1991) in respect to chemical weathering in the Antarctic where they too note the only limitation on this process is the temporal and spatial availability of moisture.

The data here presented tend to support the notion, outlined above, that the main limitation on freeze–thaw weathering in Viking Valley is not thermal but rather moisture. The winter data indicate that snow was limited and transient, as suggested by visual observation on entering the valley at the start of summer. Thus, even though the rock showed a significant number of freeze–thaw cycles they were of no (cryoclastic) effect as there was little or no water available in the rock to freeze. Any water which may have been present could have been, owing to its highly limited nature, only in a superficial surface layer and thus could not explain the larger-scale breakdown of the rock observed in the area. Thus, the qualitative judgement of the role of freeze–thaw in this area (and comparable locations) is spurious and not supported by the data. Nonetheless, the rock *does* break down and shows a marked aspect-controlled weathering effect. The winter data, at one hour intervals, are probably not adequate for an accurate determination of weathering.

Consideration of the two minute data record suggests processes such as thermal stress and/or thermal shock are highly possible. Certainly the available record shows the occurrence of changes of temperature ($\geq 2\,°C\ min^{-1}$) of a rate thought to cause thermal shock. Other data under analysis from another site nearby, where measurements were taken at one minute intervals, certainly supports the occurrence of such temperature changes. In addition, where salts are present these fluctuations in temperature are likely to be causing the expansion and contraction of salt crystals/accumulations. Such changes, occurring as they do with such frequency during the latter part of the summer, might be very important for salt weathering. Even the winter fluctuations at sub-zero temperatures could possibly produce a varying stress field, owing to the expansion and contraction of the salts relative to that of the rock, that could cause weathering. This salt weathering is, though, spatially constrained to the occurrence of the salts and this was not ubiquitous but rather seemed to be concentrated at certain locations within the valley. For instance, no visual indication of the presence of salts could be found in the taffoni but gypsum in layered accumulations up to 29 cm thick was found at the undercut base of some outcrops.

The main question that these data suggest is, why is there evidence of such extensive weathering on the northern and western aspects but not on the eastern? This question is fuelled by the separate finding of the preferred orientation of the terraces higher up on the nunatak that complements this aspect-controlled weathering in the valley bottom. From the preceding discussion it was suggested that the east side had the greatest number of (thermal) freeze–thaw cycles *but* that, as no water was present, they were not effective. The other aspects received fewer of these thermal events but as they too were devoid of moisture this was of no consequence. The role of freeze–thaw could thus be discounted for other than a highly superficial effect affecting only the extreme outer shell of the rock where, as a result of surface ice extrusion during freezing, it is likely of limited geomorphic impact. Conversely, the eastern aspect was found to have "warm, stable" conditions whilst the north and west had more varied temperatures, with larger and more frequent oscillations. Thus, it is possible that processes such as thermal stress fatigue, thermal shock and salt weathering (where salts are present) could be more operative and efficacious on the northern and western aspects.

It should be pointed out that no obvious signs of chemical weathering (e.g. weathering rinds) could be found anywhere in the area, nor were there any indications of chasmoendolithic organisms that might affect weathering. As another observation (which will be dealt with in detail elsewhere), it was extremely noticeable in the field that the light-coloured sandstones were weathering in a very "rounded" form whilst the dark-coloured sandstone was weathering to a very angular form. The light-coloured sandstones exhibited this roundness not only on loose clasts but also on bedrock, along the bedding and jointing intersections, on both horizontal and vertical surfaces. Further, at a number of sites the two lithologies could be seen only centimetres apart, along a lithologic junction, and yet the two weathering manifestations were quite distinct; the conditions to which they were subjected having, at these places, to be identical. The taffoni developed in the light-coloured sandstone exhibited this "smooth, rounded" character. The implications of these weathering responses are also now under investigation and this too may help in understanding the overall nature of weathering in the region. The palaeoenvironmental implications of such differences are also marked, for whilst the angular debris would have readily been identified as "frost-produced", in most studies the same cannot be said for the rounded clasts (see Hall, 1995 for discussion) and yet here they are contemporaneous and in close juxtaposition.

It is clear that there is some form of aspect-controlled weathering, and some indication of what processes may or may not be active is available, but it is far from clear exactly what the actual causes of breakdown are and exactly when and how they operate. Based upon these findings new data acquisition will soon be initiated to gather year round temperatures and another field season is expected in which an attempt will be made to obtain other data which will help solve the many problems outlined above and to integrate the findings into the explanation for the terraces and the other resultant weathering forms.

Weathering in General

From the above study a number of generalizations can be derived that may apply to cold region weathering studies and the interpretation of cold region (past and present) sediments. First, it is clear (as other authors have suggested) that the air temperature is not a surrogate for rock temperatures and that, in the context of freeze–thaw, the threshold value for the freeze cycle can have a large impact on the number of effective cycles with the arbitrary use of 0 °C over-emphasizing the number of events. There is also a marked aspect effect on the nature of the thermal environment experienced by the rock. Second, it is highly subjective to presume freeze–thaw weathering just because it is a "cold region": the evidence here indicates that this is an unlikely process and yet the rock is still weathering. Third, as an adjunct to the preceding statement, other processes, particularly thermal stress fatigue, need more recognition. Fourth, for any meaningful judgement of process (of any kind) there is a need for temperature recording at very frequent intervals (one minute seeming to be optimal). Fifth, the monitoring of rock moisture content is a necessity and in its absence some indirect observation (e.g. via ultrasonics) is necessary. Without the adjunct of moisture with temperature the ability to determine and examine weathering processes is impossible. Sixth, clast shape is no (apparent) indicator of weathering process and hence no indicator of present or past climatic conditions. It certainly is *not* an indicator of process (e.g. angular clasts can be produced by a whole range of processes other than freeze–thaw). Lastly, aspect-controlled weathering can have an effect on larger landforms, such as terraces, but the terraces themselves do not indicate process. Again, this has implications for palaeoenvironments where frost action is presumed to be associated with cryoplanation terrace formation.

Taken as a whole, it would seem that no one process is responsible for the weathering. Rather, as in most situations, there is a synergistic combination of different processes that cause the breakdown and, although not discussed here, interact with processes that remove the weathering products (e.g. especially in the case of the taffoni and in terrace formation). As Thorn has intimated on a number of occasions (e.g. 1988, p. 13) and as was recently discussed by Hall (1995), we need to re-evaluate cold region weathering processes and not simply presume freeze–thaw, particularly without any evidence to support the contention. Freezing conditions and angular clasts are *not* adequate evidence of frost weathering and we cannot presume processes such as that modelled by Walder and Hallet (1985) without being able to clearly show that the rock, thermal and moisture conditions *were* met. With increased data acquisition, it is likely we will find that processes other than freeze–thaw are often responsible and that even where that process does dominate it is working synergistically with other processes such that the breakdown and landforms are the product of process combinations, not a singular entity.

ACKNOWLEDGEMENTS

The fieldwork on Alexander Island was undertaken in conjunction with a larger study run by the British Antarctic Survey. Dr David Walton of BAS is sincerely thanked for allowing my participation in this project and for facilitating the collection of the winter temperature data. I would also like to thank all the BAS personnel who helped in the field and who maintained and collected the climatic data. Dr Ian Meiklejohn is thanked for the time he spent with me in the field and for all the help and companionship on our daily trek to the study site. Nancy Alexander is thanked for all her advice in the production of the maps and graphs: without her help I could not have managed. Dr Joe Ackerman kindly wrote a program for the extraction of temperature data at varying time intervals. Two anonymous referees greatly helped in correcting errors and English in the original text and I am extremely grateful for their help and advice.

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GEOGRAFIA FISICA E DINAMICA QUATERNARIA

SUPPLEMENTO III - 1997

TOMO I

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L'Association Internationale des Géomorphologues

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QUATRIÈME CONFÉRENCE INTERNATIONALE DE GÉOMORPHOLOGIE

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GRUPPO NAZIONALE GEOGRAFIA FISICA E GEOMORFOLOGIA
del Consiglio Nazionale delle Ricerche

Bologna (Italia) - 28-VIII / 3-IX, 1997

ABSTRACTS / RÉSUMÉS

COMITATO GLACIOLOGICO ITALIANO - TORINO
1997

SHOICHI HACHINOHE[1] & TAKASUKE SUZUKI[2]

Weathering rates of Tertiary sedimentary bedrock in Japan

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The rates of weathering of bedrock have been considered to be one of the most important factors that influence the modes and rates of denudational processes such as erosion and mass movements. The weathering rates were investigated in the following way.
(1) The weathering rates were examined for two kinds of the definition: first the rate (dZ/dt) at which thickness of weathered zone of bedrock (Z) increases with time (t) and second the rate (dR/dt) at which strength of weathered materials at a given depth (R) decreases with time (t). (2) The emergence age of marine-erosional terrace is assumed to be equal to the weathering time (t) for the bedrock under the terrace surface. (3) The weathered materials except the soluble of bedrock under the terrace veneer are assumed to have scarcely been eroded away. (4) The weathering profiles were observed for the drilling cores obtained by the present authors. (5) The drilling cores were examined in the laboratory for the changes in the needle penetration hardness with depth. (6) The weathering profiles were divided into four weathered zones according to the change in the mechanical property: i.e. highly weathered zone (H), moderately (M), slightly (S) and faintly (F). (7) The thickness of weathered zones is defined as the depth from the bedrock surface to the weathering front of each of the four weathered zones.
The data were obtained for the bedrock of the marine erosional-terraces in the Boso Peninsula, Japan. The terraces are divided into five levels. The bedrock in this area is the Pliocene marine sedimentary rocks of the interbedded mudstone, sandstone and conglomerate. The drilling cores were obtained at the 11 sites for sandstone and at the 13 sites for mudstone on the terrace surfaces with three different ages.
Mode of deceleration in the weathering rates (dZ/dt) with weathering time (t) differs between mudstone and sandstone. In the faintly weathered zone, mudstone is weathered faster than sandstone at the beginning of weathering. After about 400 years in the weathering time, sandstone exceeds mudstone in the weathering rates.
The rates of decrease in mechanical properties (dR_S/dt) with weathering time (t) also differs between sandstone and mudstone. At the shallow zone, e.g. 3 cm and 10 cm deep, mudstone begins to be weathered earlier than sandstone, but after the certain elapsed time from the beginning of weathering, i.e. about 70 years for 3 cm deep and about 200 years for 10 cm deep, mudstone will be exceeded by sandstone in the rates of decrease in R_S. At the zone deeper than about 30 cm, however, sandstone starts to be weathered earlier and faster than mudstone.

KEVIN J. HALL

The impact of temperature record interval and sensor location on weathering inference in periglacial environments

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In many weathering studies, particularly those in periglacial regions, much emphasis is placed on the thermal conditions. The reality is that it is moisture, not temperature, that is the limiting factor. Nevertheless, with respect to the thermal conditions, many deductions are based on air temperatures which are, in reality, meaningless as a surrogate for rock conditions. This use of air temperatures has resulted in subjective interpretations of weathering environments/processes that have served to reinforce, rather than question, the presumption of freeze-thaw weathering in periglacial environments. Further, almost all available rock temperature data are inadequate for any meaningful deduction of process, particularly that of freeze-thaw. Without information regarding the presence of water within the rock (including its amount, distribution and chemistry) together with that on the rate of fall of temperature within the rock as well as the amplitudes of the freeze and thaw cycles, so it is impossible to assume the operation of freeze-thaw or to be able to deduce which freeze-thaw mechanism was active.
Detailed temperature data, obtained at 30 second or one minute intervals, from recent studies in Antarctica and the Canadian Rockies, show the importance of such high frequency data acquisition for the evaluation of weathering processes. Not only are such data necessary for the establishing of which freeze-thaw process is operative (if any) but they also show that mechanical processes other than that of freeze-thaw may be operative and, possibly, more important. Data analysis of freeze events allowed for the determination of the rate of fall of temperature and thus deduction of possible freeze-thaw mechanism(s). More importantly, the detailed data provide evidence for the operation of thermal shock ($\Delta t \geq 2$ C° min^{-1}) as well as thermal stress fatigue. The significant control of aspect on temperature, a factor often ignored in weathering process interpretation, can also have an impact on thermal stress fatigue. There are major thermal differences between aspects (≥ 18 C°) and these not only have implications for process differentiation but also for the implementation of buttressing that can enhance the role of thermal stress fatigue. It is suggested that in many periglacial environments processes other than freeze-thaw (e.g. thermal stress and/or wetting and drying) are more active and important in sediment and landform development. Data to indicate the impact of record interval and aspect will be presented together with examples of its importance for process understanding. The need for these type of data to replace the qualitative presumptions of cold (and other) region weathering will be emphasized.

The Impact of Temperature Record Interval and Sensor Location on Weathering Inference In Periglacial Environments.

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Determination of weathering in periglacial environments is dependent (primarily) on rock properties, rock moisture content and the thermal conditions to which the rock is subjected.

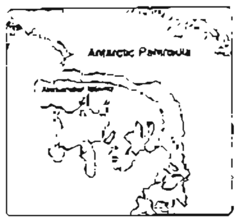
Information presented here relates solely to the thermal conditions.

For any realistic understanding of rock weathering it is necessary to monitor the actual rock temperatures (i.e. air temperatures are *not* a surrogate for rock temperatures)

Data shown here indicate the influence of aspect and record interval on the interpretation of the weathering environment.

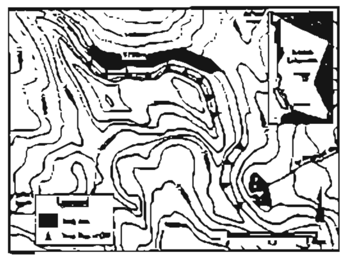
Data were collected from:

Alexander Island (Antarctica)



The Canadian Rockies

and



Panel 2

Panel 1 = Background Information *Panel 2 = Influence of record interval*

Panel 3 = Influence of aspect *Panel 4 = Influence of rate of change of temperature*

Deduction of weathering regime is often based on temperature information. For example, frequent crossings of zero degrees is perceived as conducive to freeze-thaw weathering - often, though, without information on rock moisture content! The perception of the thermal oscillations are frequently based on temperature measurements at intervals of one hour or, at best, 15 minutes. Examples here show the difference in what the rock actually experiences as a function of the record interval ( ).

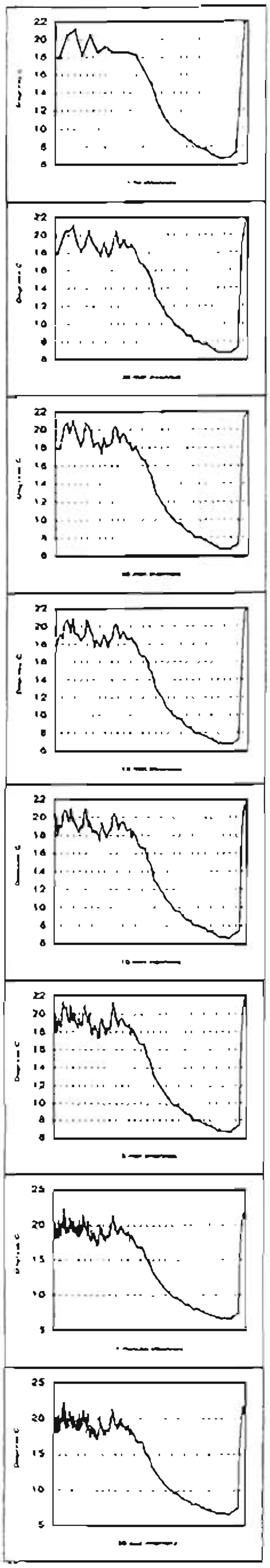
Aspect also plays a significant role. Not only are there marked aspect differences in weathering regime that can influence the formation of landforms and/or sediments but lack of recognition of aspect in placing of monitoring sensors can give a biased view of the weathering regime (Panel 3).

The rate of change of temperature is extremely important. Any deduction of the freeze-thaw process is dependent upon the rate of change of temperature. Equally, however, the rate of change of temperature can, in its own right, cause weathering - thermal stress fatigue or thermal shock. The ability to determine thermal effects is greatly dependant on record interval. Lack of high frequency data has produced a bias view of rock thermal conditions in cold regions (Panel 4).

Based on data collected at 30 second intervals, it was possible to generate a series of graphs to depict what would have been found had the record intervals been 1,5,10, 15, 20 30 and 60 minutes respectively ( ). These graphs were generated for the north-facing and south-facing exposures of a rock outcrop in the Canadian Rockies for the same day and time period (thus aspect differences are also evident). Although these are summer data (and hence not applicable to freeze-thaw) the impact of record interval in showing an ever-increasing number of thermal oscillations is obvious. A marked difference can even be seen between the 5 minute record and the one minute record. Consider the different perceptions of weathering environmentgenerated by each graph. A one minute record interval is considered optimal. This record interval is also deemed important (see Panel 4) for determination of actual weathering processes. It is argued that record intervals greater than 1 minute are of limited value.

Aspect also plays an important role (Panel 3). Whilst it is obvious that there are large thermal differences as a function of aspect, data are rarely provided to show this and sometimes a single orientation is used to depict a weathering regime. Data from Antarctica are presented to indicate the role of aspect including, based on this summer information, how the 24 hours of daylight in high latitudes can impact on what would normally be the shadow side of outcrops

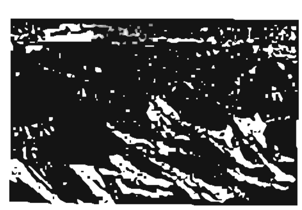
The rate of change of temperature is extremely important for deducing the freeze-thaw process and for consideration of thermal stress fatigue or shock. Unless data are collected at one minute intervals it is not possible to evaluate this (Panel 4). As water was not present during these studies and these are summer data no consideration was given to freeze-thaw. Rather, it is the role of thermal stresses in cold environments that were considered the significant factor. Various studies have indicated that a rate of change of temperature greater or equal to 2 degrees C per minute is the threshold for thermal shock. Rates of change lower than this threshold can still induce fatigue, particularly if changes are frequent and of large amplitude. The graphs, from both Canada and Antarctica, show examples of thermal events that meet or exceed the threshold as well as various examples of large amplitude variability. Without data at one minute intervals this role of thermal stress/shock would not have been evident.

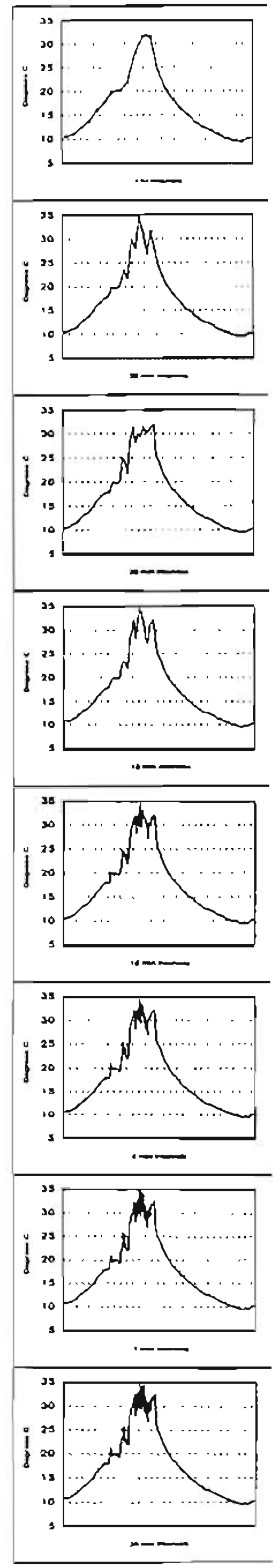


DATA COLLECTED AT 30 SECOND INTERVALS

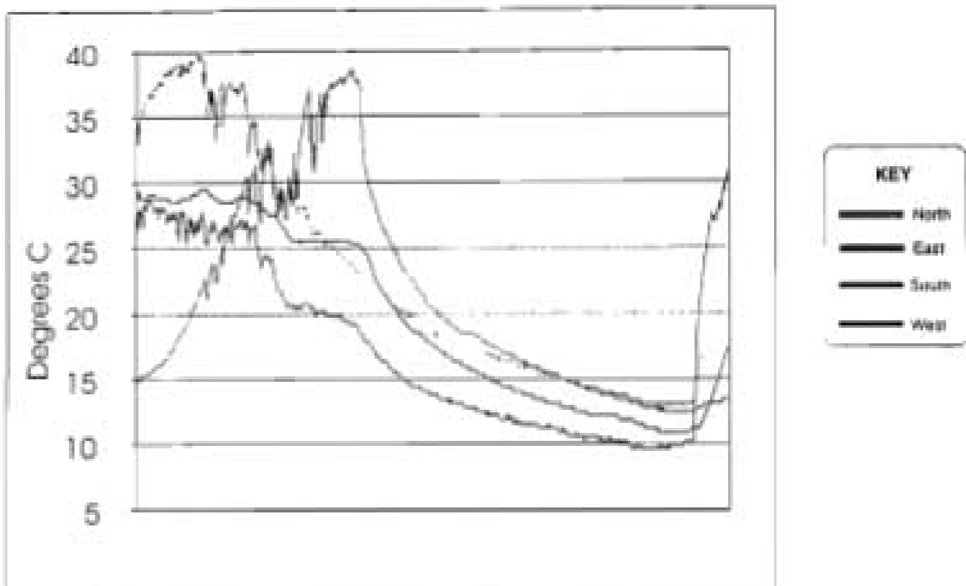
GRAPHS TO SHOW WHAT WOULD HAVE BEEN RECORDED IF DATA HAD BEEN COLLECTED AT INTERVALS OF 1, 5, 10, 15 20, 30 AND 60 MINUTES TOGETHER WITH THE ORIGINAL 30 SECOND DATA.

DATA FROM CANADIAN ROCKIES COLLECTED:
0722 HRS 25 JULY 1996 TO
0726 HRS 26 JULY 1996



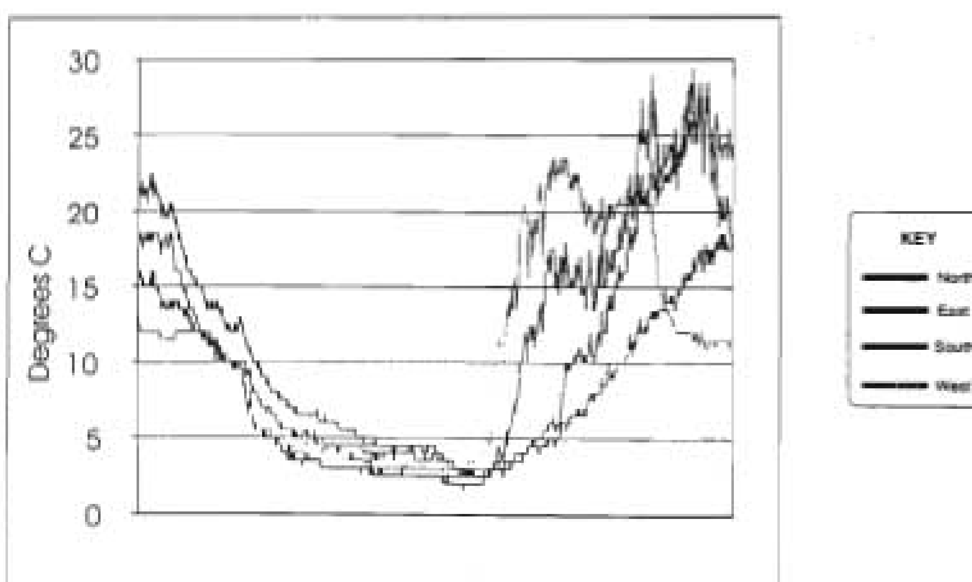


Data from
Canadian Rockies



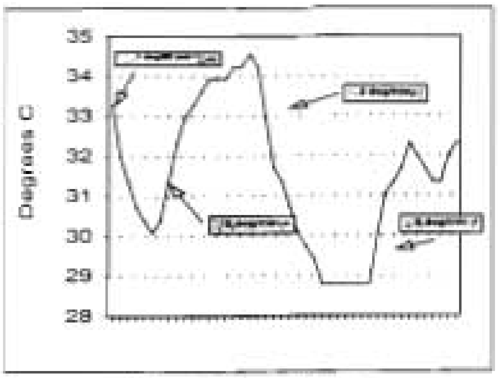
1041 hrs 27 July to 0749 hrs 28 July 1996
Data at 30 second intervals

Data from
Alexander Island
(Antarctica)

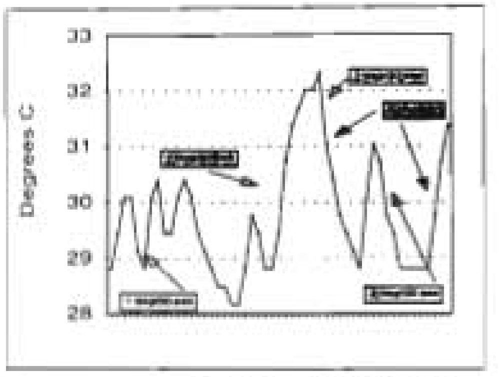


1640 hrs 26 January to 1632 hrs 27 January 1993
Data at 1 minute intervals

DETAILS FROM CANADIAN STUDIES



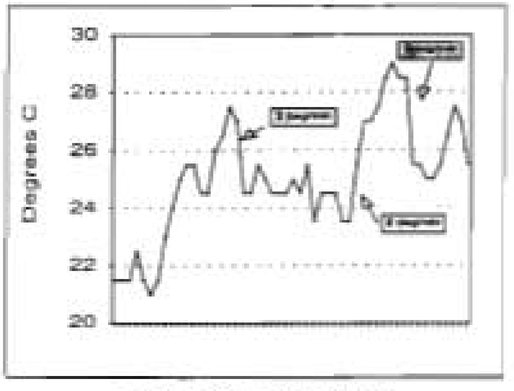
1712 hrs to 1733 hrs 25 July, 1996
South-facing



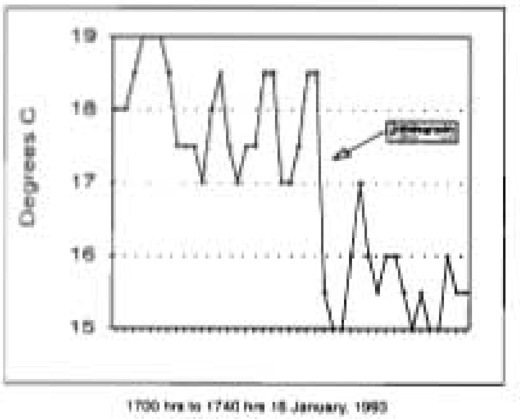
1506 hrs to 1630 hrs 27 July, 1997
South-facing

Record at 30 second intervals

DETAILS FROM ANTARCTIC STUDIES

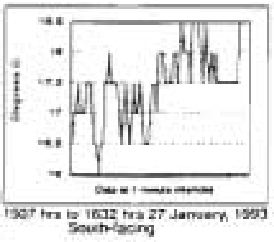


1237 hrs to 1327 hrs 27 January, 1993
North-facing



1700 hrs to 1740 hrs 18 January, 1993
North-facing

Record at 1 minute intervals



1507 hrs to 1632 hrs 27 January, 1993
South-facing

To show the large fluctuations in temperature over an 85 minute period

1239 hrs to 1429 hrs 29 December, 1992
South-facing

To show the fluctuations in temperature recorded in the top and bottom of a south-facing tafoni

PERMAFROST AND PERIGLACIAL PROCESSES
Permafrost Periglac. Process. 9: 47–55 (1998)

Rock Temperatures and Implications for Cold Region Weathering. II: New Data from Rothera, Adelaide Island, Antarctica

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ABSTRACT

Rock temperature data collected at one-minute intervals from both the horizontal surface and the four cardinal directions of a rock outcrop show the influence of record interval and aspect on the thermal regime of bedrock as it applies to cryogenic weathering. High frequency data are necessary to identify components of thermal stress fatigue and thermal shock events that play a significant role in rock breakdown. The northern aspect exhibits the lowest temperatures despite its apparent preferential orientation. At the 2 cm depth, temperatures on the northern and horizontal surfaces sometimes stayed above those for the rock surface despite the daytime energy input from solar radiation. Short-term wind fluctuations are considered as a possible explanation. Because the rock temperatures are quite different from those of the air the latter can, in no way, be used as a surrogate for rock thermal conditions. The argument is made that one-minute record intervals are required for thermal data if use is to be made of this information to help explain and understand the weathering regime. © 1998 John Wiley & Sons, Ltd.

RÉSUMÉ

Les enregistrements de la température des roches obtenus avec des intervalles de 1 minute sur des surfaces horizontales et des affleurements exposés aux quatre directions cardinales, démontrent l'influence de l'intervalle d'enregistrement pour la compréhension du régime thermique du bedrock et, de ce fait, de l'altération cryogénique. Une grande fréquence de mesures est indispensable pour identifier les éléments responsables de la fatigue thermique, ainsi que les chocs thermiques qui sont très importants dans la rupture des roches. C'est pour des expositions au nord que les températures enregistrées ont été les plus basses, bien que cette orientation semble plus favorable. A 2 cm de profondeur, les températures de surfaces horizontales et de roches exposées au nord ont été parfois supérieures à celles de la surface de la roche et cela malgré l'énergie solaire. Des fluctuations du vent à courte période pourraient expliquer cette observation. Parce que les températures de la roche sont tout à fait différentes de celles de l'air, celles-ci ne peuvent en aucune manière remplacer les températures de la roche. Des enregistrements des températures avec un intervalle d'une minute sont nécessaires si l'on veut utiliser ces données pour comprendre les phénomènes d'altération. © 1998 John Wiley & Sons, Ltd.

KEY WORDS: weathering; thermal conditions; thermal stress fatigue; record interval; Antarctica

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INTRODUCTION

In an earlier paper (Hall, 1997a) details were given of rock temperature data collected from sites in a dry valley on Alexander Island, Antarctica. In that paper winter observations were outlined, which showed the thermal regime of the rocks for the first part of two successive winters, while detailed summer temperature data indicate the probability of thermal shock and thermal stress fatigue as important weathering processes. These observations were then put in the context of Antarctic weathering. It was suggested that the frequency of data recording impacts upon the ability to realistically deduce weathering process(es). In that study temperature measurements were collected at two-minute intervals and it was shown that weathering processes such as thermal shock (which requires $\Delta T/t \geqslant 2\,°C/min$) could not be recognized with measurement intervals of five minutes or longer. It was argued that 30-second or one-minute record intervals would be better, not only for the identification of thermal shock but also for the assessment of freeze–thaw weathering for which the rate of change of temperature is critical (McGreevy and Whalley, 1982; Hall, 1995).

In this paper data are presented which were collected at one-minute intervals for the rock surface and at 2 cm depth on the four cardinal aspects, plus on a horizontal surface, for nearly three weeks from a knoll overlooking the airstrip at the British Antarctic Survey Rothera base (Figure 1), Adelaide Island (67°34′S, 68°07′W). In this region of Antarctica the sea remains

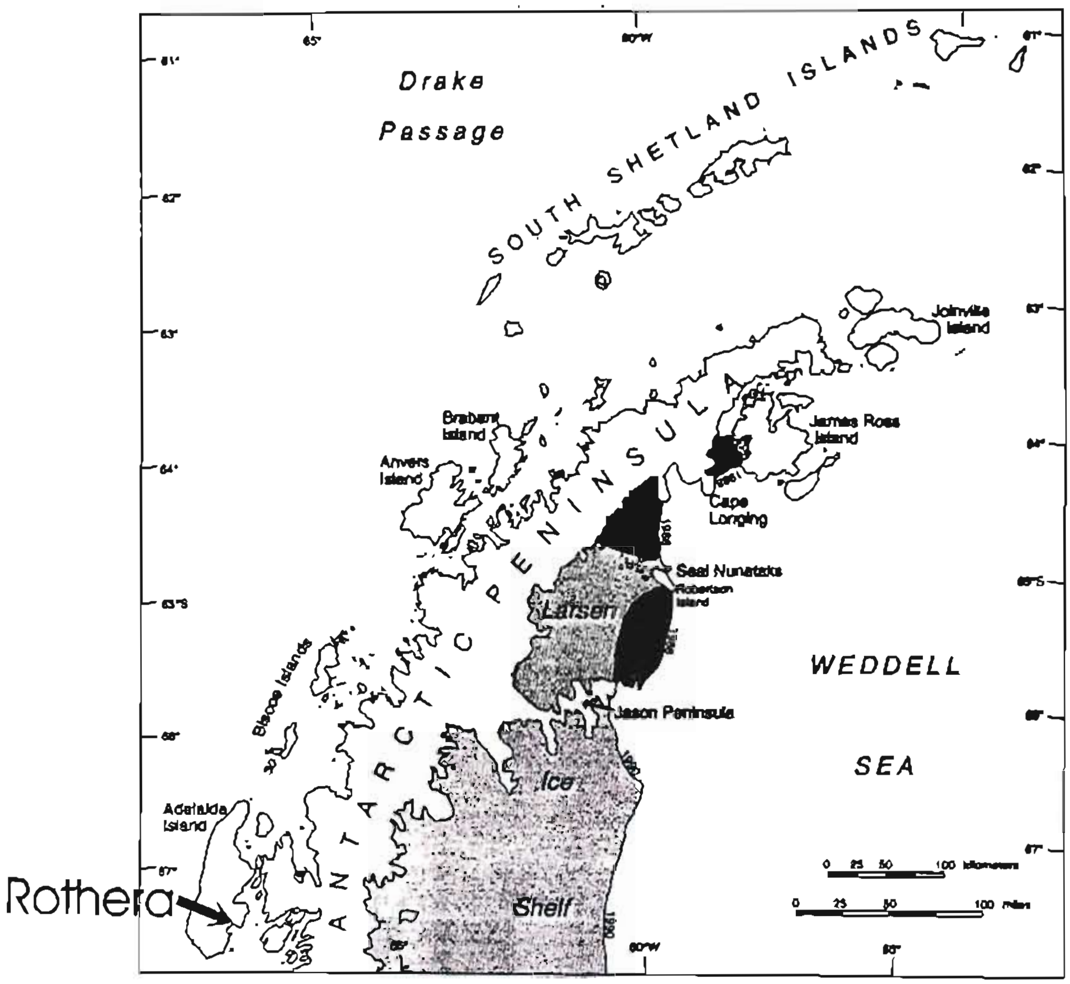

Figure 1 Map to show the location of Rothera. Map reproduced from *Antarctic Science Handy Atlas*, Map 12 (1995, p. 180).

ice-covered for the greater part of the year and the site was *c.* 50 m above sea level. Bedrock consisted of gabbro and granodiorite – with the temperature data presented here being recorded on a medium-grained granodiorite.

FIELD METHODS

Data were collected using a 'Squirrel' data logger and 10 bead-type thermistors. On a single rock outcrop, a thermistor was attached to the rock surface and another was located at 2 cm depth for each of the cardinal directions: thermistors were also located at the surface and 2 cm depth on the upper, horizontal surface of the rock outcrop. Data were then recorded at one-minute intervals, with downloading to a laptop computer once per day. It was not possible to use pre-drilled blocks, nor were measurements made of moisture content, ultrasonic *p*-wave velocity or Schmidt hammer rebound values as in the initial study (Hall, 1997a). The aim was to focus purely on high frequency measurement of the thermal events in the surface layer of the rock as an *indicator* of weathering process. For comparison with the rock temperature data, daily climatological data are available from the Rothera meteorological station.

RESULTS AND DISCUSSION

The impact of aspect on rock thermal conditions was not, at least for the record period shown here, as straightforward as might be expected. For a Southern hemisphere site, it should be expected that the north-facing aspect should have the highest temperatures and the south-facing the lowest. However, this is not the case (Figure 2). For the example shown (0000 h to 2359 h on 14 January 1993), the lowest temperature (−1.5 °C) was recorded for the northern aspect and the highest temperature (27 °C) on the western aspect.

A temporal shift of influence can be seen as the sun rises and then sets, from east through north and then west, to finish with south. Although this is to be expected, the actual rock temperatures offer a surprise. The influence of direct radiation can be seen from *c.* 0600 h on the eastern aspect as the temperature rapidly rises from a low of 0 °C to achieve a maximum from *c.* 0730 h through to 0845 h, after which temperatures decline and return towards zero close to midnight. The temperature for the northern aspect then begins to rise around 0700 h and attains its maximum near 1300 h and subsequently cools from *c.* 1800 h. Despite its favourable orientation, the northern aspect does *not* achieve the highest temperatures but experiences the lowest. It was the only aspect to have subzero temperatures. By contrast, temperatures on the western aspect were some 4 °C higher. These began to rise around 0800 h to attain a maximum close to 1800 h. The western aspect experiences an almost asymmetric temperature profile, with a (relatively) slow rise through to the maximum and then a relatively rapid decline. The southern aspect starts to warm as air temperatures rise in late morning, and peak a short time after that of the western aspect. The horizontal surface does not have the lowest overall temperature but it does have the lowest maximum values. This is the result of never receiving radiation at other than an acute angle. Noticeably, the maximum warming of this surface is *not* at the time when the sun is at its highest but when the sun is in the west, the temperature profile being one of radiation 'accumulation' throughout the day. As the sun drops after 1900 h so temperatures (relatively) rapidly decline.

In addition to the role of aspect, the present data exemplify the influence of record interval. The 'smoothing' effect of increasing time between records is clearly shown in Figure 3. Based on the original one-minute record interval data, it is shown what would have been recorded had the logging interval been 5, 10, 15, 30 or 60 minutes. Figure 3 indicates a significant decrease in detail between one-minute data and that of five minutes. There is a greater variability at the record start and end for the one-minute data. This information is lost with the five-minute record. For instance, the two drops in temperature at the record end look unidirectional and 'dramatic' on the five-minute record but the one-minute data show that there were, in fact, rises and falls within the overall decline. Information loss with increased record interval dramatically changes the perception of the thermal environment. Thus, the 30-minute data are little more than a rough guide to what actually took place. Certainly any attempt to generate values of $\Delta T/t$ from 15- or 30-minute data would be fruitless.

Examples of the important detail loss in records other than the one-minute record interval are shown in Figure 4. This 41-minute record shows one event with a $\Delta T/t$ value of 3 °C/min and one of 2 °C/min. Both values are greater than, or equal to, the theoretical threshold (2 °C/min) for thermal shock (Yatsu, 1988). Other events, while not achieving the 2 °C/min boundary for thermal

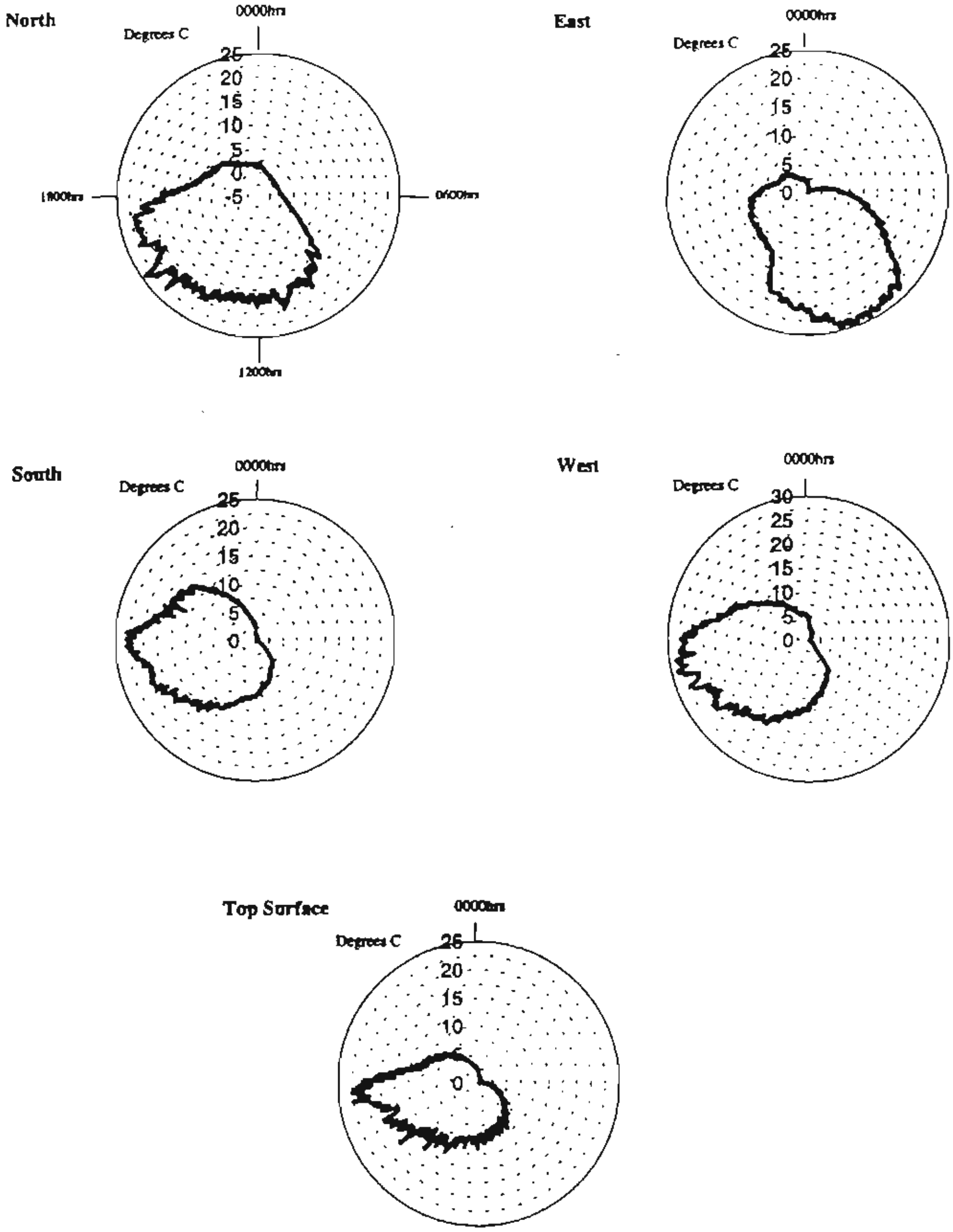

Figure 2 Temperature data from 0000 h to 2359 h for 14 January 1993 for each of the aspects plus the top surface.

shock, nevertheless indicate values $\geqslant 1$ °C/min that must exacerbate thermal stress fatigue. In the 41-minute record shown here there are eight such events with a further 16 of 0.5 °C/min. Several of these involve consecutive positive and negative temperature changes each of 0.5 °C/min magnitude.

At depth (Figure 5), the largest temperature differential (between the surface value and that recorded at *c*. 2 cm depth) is found on the eastern aspect (± 12 °C) and the smallest (± 4 °C) on the southern. Although it is clear from Figure 5 that the surface fluctuations are largely damped out by 2 cm depth there are still variations of the order of 1 to 1.5 °C/min at times. By obtaining high frequency data one can demonstrate that values of $\Delta T/t$ are such that thermal shock could occur.

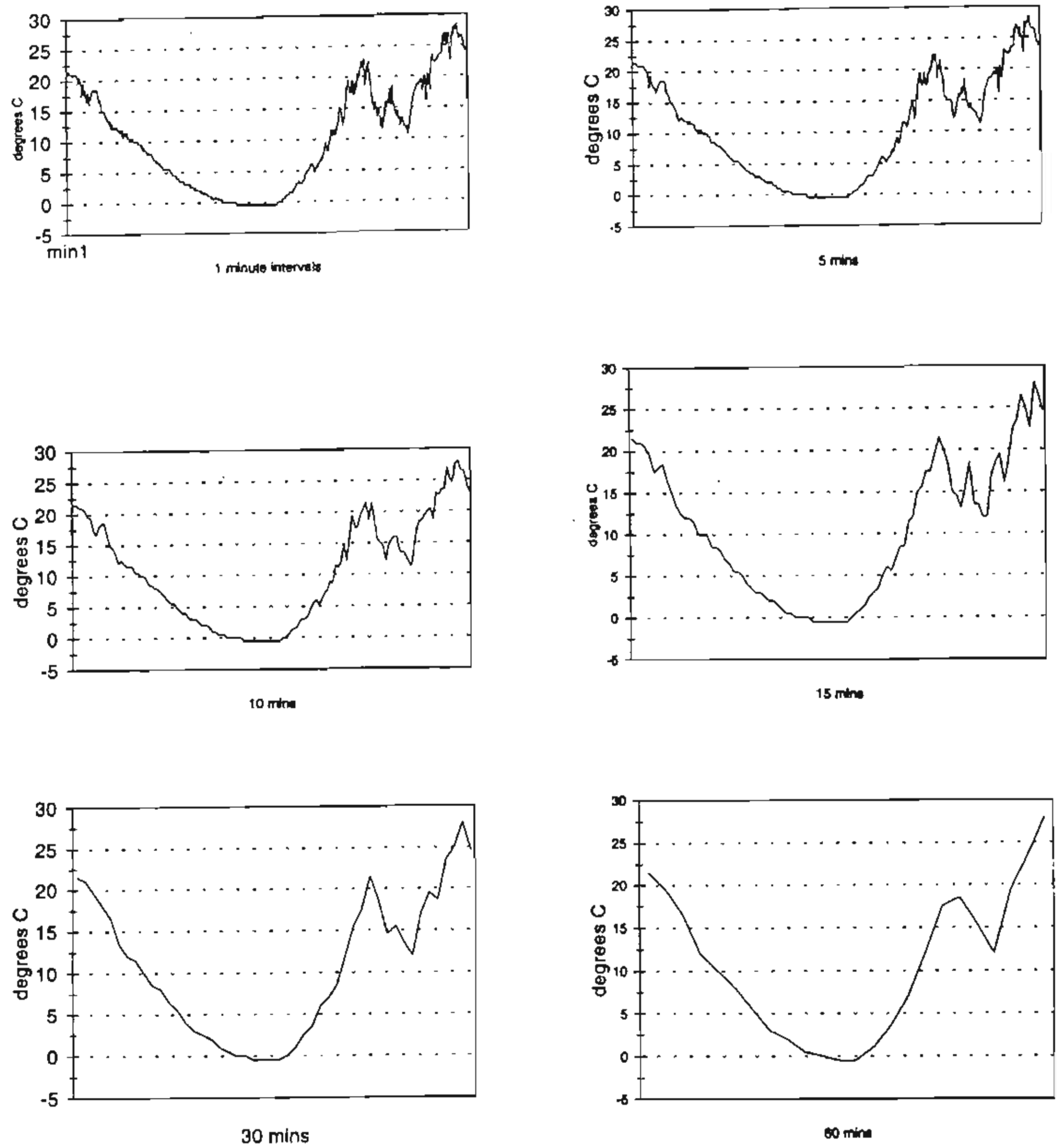

Figure 3 Temperature data for the period 1656 h on 15 January to 1639 h on 16 January 1993 to show, based on the original one-minute data, what would have been recorded with collection times of 5, 10, 15, 30 and 60 minutes.

Rapid thermal fluctuations produce transient high thermal gradients in the outer layer that facilitates thermal spalling (Thirumalai, 1970).

The surface and 2 cm depth rock temperature data show that complex stress fields must occur in the outer shell of the rock. For example, the southern aspect shows the 2 cm level is warmer than the rock surface at the record start. The temperature differential of *c.* 4 °C indicates that the rock at depth is experiencing greater expansivity than the surface. Later, there is a crossover and the surface is warmer than the 2 cm level: at this point, the exterior is expanding while the interior is contracting (its temperature having declined from the higher, earlier value). The resultant differential stresses could play an important role in aspect-constrained weathering. Without this one-minute data it would not be possible to attribute any

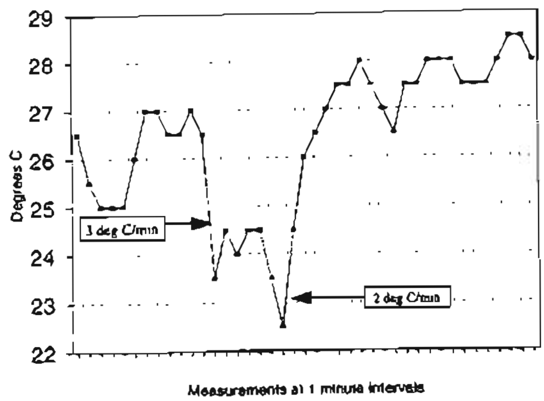

Figure 4 Small section of the data presented in Figure 3 (one-minute data) to show the rapid thermal changes that can occur at the rock surface.

observations of aspect influence on weathering to these thermal fatigue effects.

The north-facing record (Figure 5) may indicate other factors. Here, in the example shown, the rock at 2 cm depth stays *warmer* than the surface for the greater part of the day. The effect did not occur all the time and surface temperatures show the typical rapid changes associated with the rock surface while those at 2 cm depth show the expected dampening. Consideration of the rest of the records for the northern aspect shows that, while the same effect *did* occur on other days, it did not occur every day. On some days (e.g. 20 to 22 January), the surface temperatures were warmer than the interior, and on other days (e.g. 25 to 26 January) the thermal conditions were mixed, with the interior warm but with significant periods when the surface was warmer than the interior.

These apparently random thermal fluctuations are possibly related to the effect of the wind. A number of Antarctic biological studies (e.g. McKay and Friedmann, 1985; Kappen *et al.*, 1981; Friedmann *et al.*, 1987) have collected temperature data at intervals of one minute or less on rock surfaces and at depths to *c.* 4 cm for different aspects. Modelling of thermal changes as a function of temperature, wind speed and internal light gradients was subsequently undertaken by Nienow (1987). The models suggest that the observed temperature oscillations at the rock surface can be caused by wind fluctuations of short duration and can be effective to a depth of *c.* 4 mm. Other factors such as heating at depth by light penetration (due to translucent minerals) may also play a role although Nienow (1987) did not consider this significant for the sandstone that he studied. Although the effect of wind offers some degree of explanation for the observed effects at Rothera, more data are needed to elucidate why the effects were only observed on the horizontal and northern surfaces.

The Rothera data also show distinct differences between the air and the rock temperatures (Figure 6). The air temperature graph is based on five-minute record intervals. The diurnal cycle is very clear. It is also very clear that maximum temperatures for the air ($\leqslant 7$ °C) are substantially lower than those for the rocks (e.g. see Figure 2, $\leqslant 27$ °C; or Figure 4, $\leqslant 28.5$ °C). Equally, the minima for the rock can be lower than that of the air although the differences here are much less than for the maxima: Figure 2 shows the northern aspect going to -1 °C whilst the air does not go lower than $+0.3$ °C. For the other aspects shown in Figure 2, although the air went to -0.3 °C, none of them went below 0 °C. The most important factor here is that rock temperatures are substantially higher than those of the air and they show many and varied fluctuations. Thus, in terms of weathering, the air temperature data are all but useless for any interpretation of the weathering regime despite the recent arguments by such as Arnold *et al.* (1996).

IMPLICATIONS FOR WEATHERING

In this paper, the western and northern aspects exhibited extremes, with the north having the lowest temperatures and the west the highest. When the 2 cm data are considered, the eastern aspect has the largest surface to depth temperature differential and thus the greatest potential for stress-induced (as a result of the thermal gradient) weathering. The higher 2 cm depth temperature than the surface for the northern aspect and the horizontal surface may have implications for weathering. Interestingly, further analysis of the thermal data referred to in Hall (1997a) failed to show the same effect on the northern exposure although the values were very close. It is possible that there is a temporal shift for this effect and that it is negligible during the summer but increases through the autumn towards the winter. Equally, it may reflect a wind effect. Accepting that thermal maxima can occur *below* the surface (as also argued from the biological studies) then this does offer a probable weathering effect. That maximum expansion could occur below the surface would help

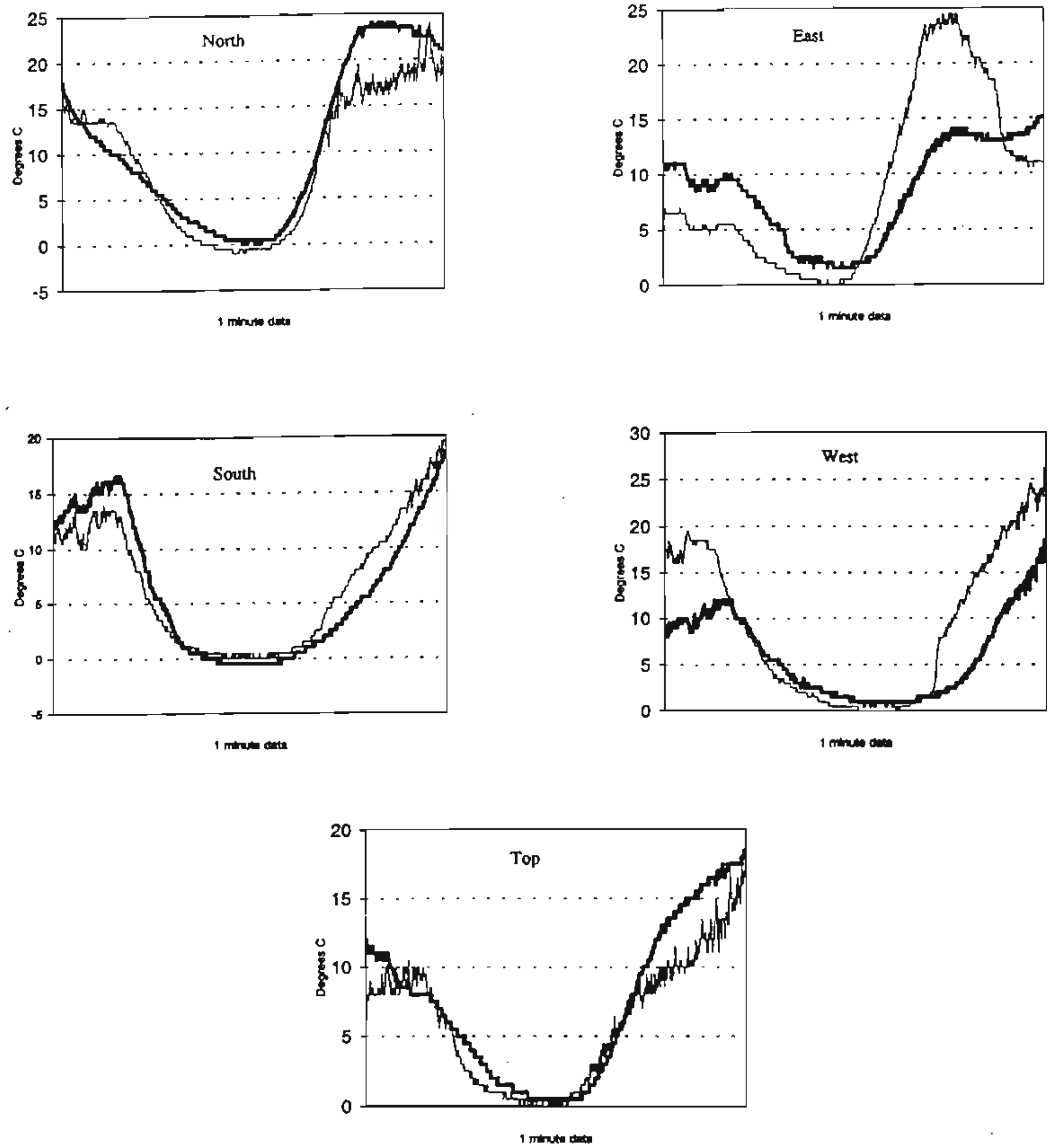

Figure 5 Data for the period 1639 h on 13 January to 1631 h on 14 January 1993 for the different aspects plus top surface to show the surface (thin line) and 2 cm depth (thick line) temperature curves.

explain spalling of the rock. This form of weathering could be particularly effective in dry areas where water-based processes are minimized.

The data presented here justify the questions posed by Thorn (1988; 1992) and White (1976) who have both questioned the ubiquitousness of freeze–thaw. The data presented here offer solutions/causes *other* than freeze–thaw. In the case outlined both here and in Hall (1997a; 1997b), thermal stress fatigue or thermal shock are identified as causes worthy of detailed examination; frost weathering is unlikely owing to the aridity of the study areas. Other studies suggest other mechanisms, e.g. wetting and drying (Hall,

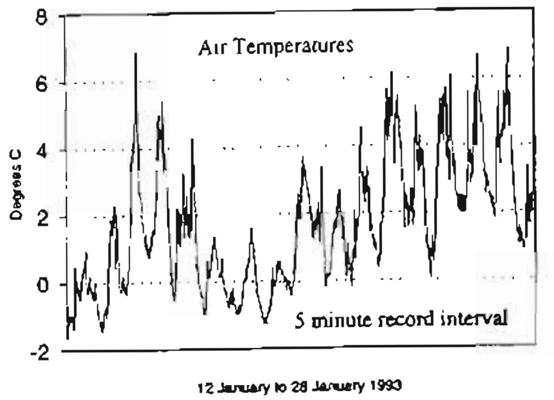

Figure 6 Air temperature record (five-minute intervals) for Rothera station

1991; 1993) or biological agencies (e.g. Hall and Otte, 1990). None of this negates the occurrence of freeze–thaw but rather questions its common acceptance and explanation for landforms.

CONCLUSIONS

The information presented in this adjunct to the original study further indicates the need for more detailed studies of the rock thermal regime. These data will provide the baseline required for determination of the freeze–thaw mechanism and may also indicate the occurrence of other, possibly more effective, mechanical weathering processes. Certainly, the combination of high values of $\Delta T/t$ with high thermal gradients suggests the potential for thermal stress fatigue and for thermal shock. Ultimately, these thermal data are only one of the two significant data sets which are required, the other being moisture and its attributes and how they vary in time and space. With these sorts of data a more meaningful understanding of the role of weathering in landform and sediment origin can be achieved.

ACKNOWLEDGEMENTS

This work comprised part of a larger British Antarctic Survey undertaking and I would like to sincerely thank Dr David Walton and the then Director, Dr David Drewry, for allowing my participation. Dr Ian Meiklejohn, my colleague and friend in the field, helped with the daily readings and I thank him for his support. Dr John King of BAS kindly provided the Rothera climate data. I would also like to thank Dr Hugh French who suggested *many* changes to the original manuscript that greatly improved its content and presentation and an anonymous referee who offered some very valuable suggestions regarding some of the problems that the data presented.

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Present-Day Soil Frost Activity on Marion Island, maritime sub-Antarctic

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University of the Western Cape

Marion Island (46°54'S, 32°45'E, 290km^2) is one of a few small islands in the Southern Ocean and a rich location for geomorphological research on climate change. Marion consists of a shield volcano rising to 1230m asl. and has an extremely oceanic climate. Almost entirely glaciated during the last Glacial it still has a few small permanent ice bodies near its summit. Periglacial morphology include extensive solifluction lobes and terraces, blockfields and –streams, mostly developed on glacial till during the early Holocene and are now relict (Boelhouwers and Holness, 1998). Currently active morphology includes small solifluction forms and patterned ground and are widespread on the island.

Objective of this study is to establish quantitative relationships between morphology/sedimentology of currently active soil frost forms, the processes involved and their associated ground climate and materials controls. This is supported by two years of field observations by one of the authors (S.H.).

Morphology/sedimentology data on patterned ground show a steady increase in lateral and vertical sorting, but no change in particle size, with altitude. Process activity is here shown in the form of dowel heave indicating a close correlation between effective depth of frost heave and depth of vertical sorting.

Clast movement data, ground thermal profiles and frost susceptibility data quantify the effectiveness of needle-ice and ice-lens induced frost creep. Needle-ice activity dominates up to 300m asl, with ice lens growth under diurnal frost becoming increasingly important above these altitudes. This pattern correlates strongly with a change in ratio of coarse versus fine stripe width in sorted stripes. Above 800m asl mild seasonal frost dominates with snow limiting frost action and creating isothermal conditions in winter around –1 to –2oC, down to at least 40cm depth. Moisture and particle size distribution forms no limit to frost activity on the island.

Overall results allow for a first zonation of the Marion Island frost environment. Further work on establishing statistical correlation between morphological and climatic parameters is now underway, while future work will attempt physical modeling and application towards palaeoenvironmental reconstructions.

The necessity for high-frequency rock temperature data for rock weathering studies: Antarctic and northern examples.

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To a very large extent our perceptions of rock weathering in cold regions are the product of untested assumptions coupled with the artefacts of our methods of temperature monitoring (Hall and Andre, In Press; Hall, *et al*, In Press). Two attributes have worked to produce a poor scientific approach to rock weathering among geomorphologists. First, the assumption of process(es) and hence a frequency of temperature observation suited to that assumption coupled with, secondly, the logistical and technical constraints of data acquisition in cold regions. Most available rock temperature data are at a frequency that (a) is not suitable for the deduction of some processes, (b) produces a false "image" of the weathering regime, and (c) becomes a self-fulfilling prophecy of the initial assumption due to inadequate data. Rock temperature data collected at one-minute intervals for extended periods (3 months to over one year) at various depths within rocks, including on different aspects, has opened a whole new perception of bedrock weathering. Thermal stress fatigue and thermal shock can, based on one-minute data, be shown to occur and to, particularly in the Antarctic context, be major contributors to rock breakdown. This is all the more so as water, rather than temperature, is the limiting factor with respect to weathering in the Antarctic (Balke, *et al.*, 1991). It will also be shown that at high latitudes weathering regimes vary greatly both spatially and temporally, with, for example, the poleward exposure being warmer and experiencing less freezing events than the equator-facing exposure for part of the year. Such findings may have major ramifications in terms of understanding landform development through time. Lastly, while contemporary discourse has moved from diurnal freeze-thaw events to the annual freeze-thaw cycle as being of the greatest geomorphic effect, recent data have shown that this may be, in part, an artefact of available data. Using one-minute data recording it has been found that frequent exotherms, indicative of water freezing, can be identified in the field. The question remains, as it does for the annual cycle, of their effectiveness in any given situation, but, nevertheless, there is unequivocal evidence that diurnal freeze-thaw *does* take place. Thus, the argument is made that, for any meaningful investigation of weathering, it is necessary to obtain thermal data at a high frequency.

The necessity for high frequency rock temperature data for rock weathering studies

Kevin Hall, UNBC, Canada

Our perceptions of cold region weathering are largely a product of:

- Our untested assumptions
  - Freeze-thaw dominates
  - Chemical weathering is limited
- Our methods of temperature monitoring

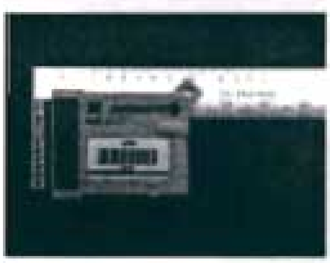
Our assumption of process has constrained the methods of temperature acquisition and this has further been affected by logistical problems

- Process assumption has led to low frequency observation commensurate with deducing freeze-thaw wavelengths and amplitudes
- Battery operation and memory capacity have constrained our ability to monitor at high frequency for extended periods

The result is that data are at a frequency:

- Not suitable for the deduction of (m)any processes
- Produce a "false" image of the weathering regime
- Become a self-fulfilling prophecy of the initial assumption due to the inadequate data

New technologies in both batteries and loggers have facilitated our ability to measure at high frequency for extended periods



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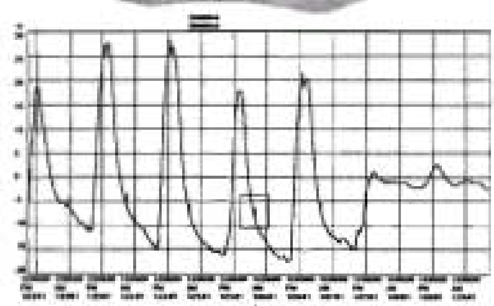

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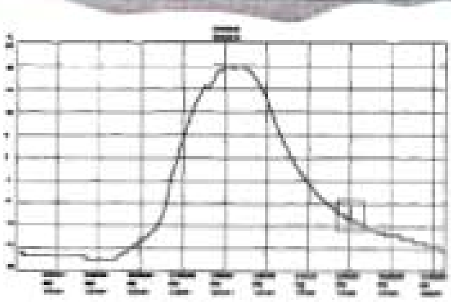

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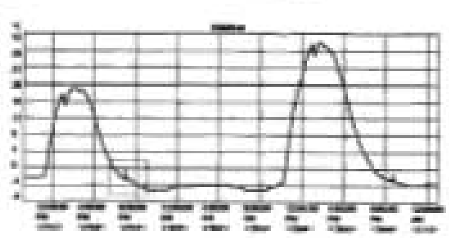

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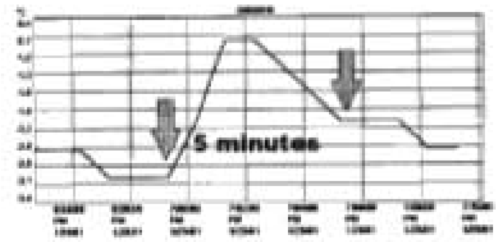
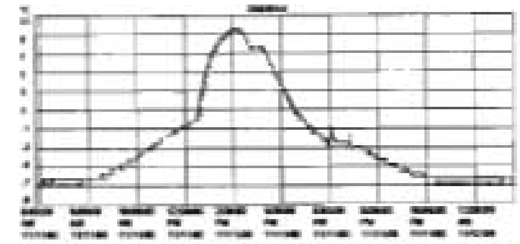
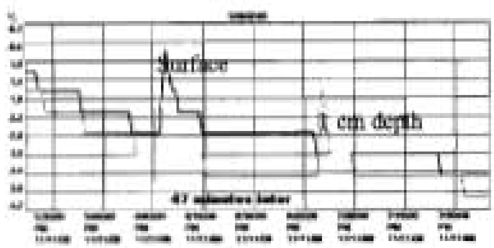


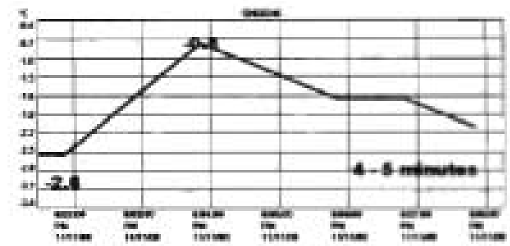
Exothermsatboththerocksurface andat1cmdepth



Detailoftheexothermsattherock surfaceandat1cmdepth



Detailofthesurfaceexotherm

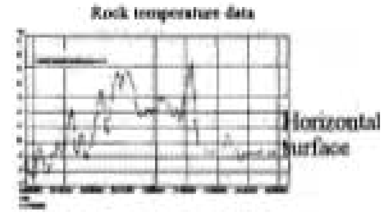


Inadditiontobeingabletoactually *show*thatafreeze-thaweventtruly tookplace,theoneminutedataalso

- Facilitate discernment of thermal shock
- Show just how unreliable data measurement at longer time intervals is for discernment of weathering processes

OneminutedatafromtheAntarctic

Rock temperature data



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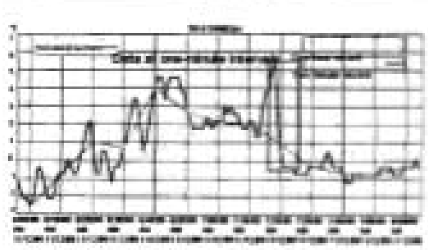
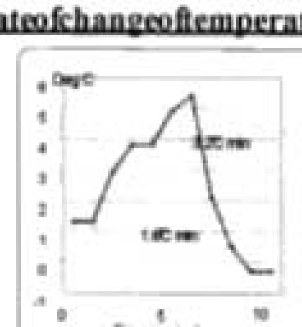


Rateofchangeoftemperature



“Itisacapitalmistaketotheorize beforeonehasdata.Insensiblyone beginstotwistfactstosuittheories, insteadoftheoriestosuitfacts”

Sir Arthur Conan Doyle

- Or
- *“First get your facts, then you can distort them at your leisure”*

  Mark Twain

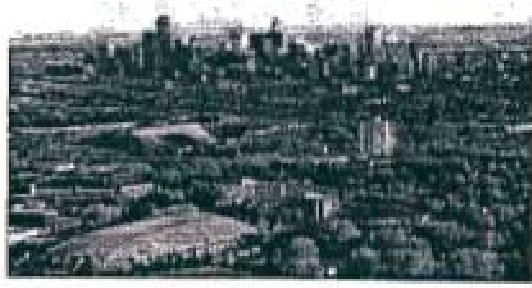
WESTERN DIVISION
OF THE
CANADIAN ASSOCIATION OF GEOGRAPHERS

ANNUAL MEETING
PROGRAM AND ABSTRACTS

UNIVERSITY OF
CALGARY

University of Calgary
Department of Geography
March 8-10, 2001

The Department o
GEOGRAPHY



European explorers and settlers arrived in what became southwest British Columbia, they utilized this existing trail network as a means of establishing and initially supporting permanent colonial settlements. Records of the location of the original web of trails utilized by the Tsleil-Waututh should thus be identifiable in the maps and journals of the early colonists. More specifically, the routes of ancient trails should be able to be found in materials prepared by explorers, land and marine surveyors, mine inspectors, and government cartographers active in the period between the mid 1800s and the early 1900s.

Perceptions of Risk When Living Near a Hazardous Waste Treatment Facility

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One of the most urgent environmental problems facing modern society today is the disposal of toxic, hazardous waste. However, the technology developed to treat such substances may represent a great concern to those living near a hazardous waste plant. The present study investigates how three rural communities construct views of risk with regards to the Swan Hills Special Waste Treatment Facility, near Swan Hills Alberta. Semi-structured interviews of 55 residents from the towns of Kinuso, Fort Assiniboine, and Barrhead were conducted to examine the judgements people make about risk within the context of their communities. A comparative analysis methodology is used to determine how risk is defined by residents in each community, and by what means "risk" is constructed. Preliminary results indicate that the smaller, more rural communities of Kinuso and Fort Assiniobine perceive the facility with great concern, whereas the respondents in Barrhead feel little concern, if any at all. These contrasting perceptions of risk are argued to be contextual and influenced by both individual and collective interpretations of belief systems. This research aims to provide a better understanding of how people think and respond to the risk of a technological hazard. Such information will facilitate communication of information between lay people, technical experts, and policy makers to resolve the issue of how to manage hazardous technologies.

Geographic Change in the Gulf Islands British Columbia

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Department of Geography, University of Victoria

The Gulf Islands, which are situated between Vancouver Island and the Mainland have become subject to increasing pressures of development. The Islands Trust is the government body responsible for managing land use planning arrangements and responding to these pressures. This paper reviews the findings of recent research into specific aspects of the general process of change namely: water shortages, forest harvesting, and planning for an aging population.

The conceptual fallacy of "weathering in cold climates" - the error in the assumption of zonality.

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Consideration of almost any geomorphology text, including specialist volumes, suggests some degree of zonality with respect to weathering such that there is perceived to be an entity under the rubric

"weathering in cold regions". By implication and, I would argue, general perception, this is considered "unique" and is considered to have attributes that make it so. Classic characteristics would be a lack of chemical weathering, a dominance of mechanical weathering, with a generally accepted dominance of freeze-thaw weathering. These arguments then extend into evaluation of landforms, the "recognition" of cold-origin sediments, and palaeo-climatic reconstructions (e.g angular clasts therefore frost weathering therefore cold climate). The circularity and self-fulfilling prophecy of such arguments seems to escape us! The naïvety of the assumption really is that in, such as, Quaternary studies those angular clasts are used to justify a cold climate, via frost action, *only* when other criteria (e.g. geographical location) have already shown that the area was cold. Essentially, there are no known criteria that can associate a mechanically weathered, or in many instances, a chemically weathered, product with a specific climate. The argument will be made for, and evidence provided in support of, the azonality of weathering in cold regions. Chemical weathering occurs and is not temperature limited but rather moisture limited. Rock, rather than air, temperatures paint a very different "climate" with respect to weathering such that many weathering forms are identical to those in hot deserts. Finally, the perceived angularity of resultant weathering debris may not be as ubiquitous as we frequently think and thus we "filter" our observations with the in-built assumptions of what we "should see". Serious reconsideration of weathering regimes is required, perhaps all the more so for those of us working in cold environments.

A Crucial Lesson Not Learned: loss of memory in resource-town planning.

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Single industry communities are a common part of the Canadian rural landscape. In many cases, these places were developed by a single firm or industry to provide a focal point for local extraction and processing operations, as well as to house the needed workforce. But why do so many post world war II resource towns share such similar townsite plans? And, why have they almost always faced similar forms of economic devastation? This paper examines the planning foundations of Kitimat, BC and compares them to the later resource towns of Mackenzie and Tumbler Ridge. The purpose is to identify which of the early planning tenets have been forgotten and whether this may have some explanatory value in why "crisis" remains the watchword of so many such towns.

Measuring Spatial Accessibility to Urban Amenities: Does Aggregation Error Matter?

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Neighbourhood spatial accessibility (NSA) refers to the ease with which residents of a given neighbourhood can reach amenities. The widespread availability of Geographic Information Systems (GIS) has increased the usefulness of NSA indicators to inform urban policy issues, such as amenity provision and spatial equity. Spatial accessibility measurements at the neighbourhood level (or other similarly aggregated units) are susceptible to numerous methodological problems. Failing to consider such problems may result in inaccurate spatial accessibility measures, and hence produce erroneous

"It is not what you look at that matters, its what you see."

Henry David Thoreau

"Cold regions are characterized by cold"

- PROCESSES ARE THUS DRIVEN BY COLD

- Hence: Mechanical processes predominate
- Freeze-thaw predominates
- Chemical weathering is limited.

Is this premiss correct?

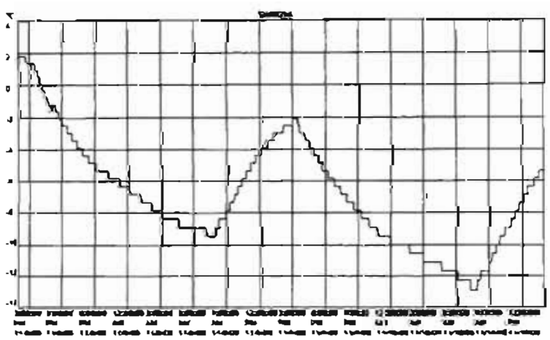
- Surely cold region processes are driven by heat not cold?
- Without heat, cold "rules" and nothing happens.

If "cold" regions are really, in terms of processes, driven by heat

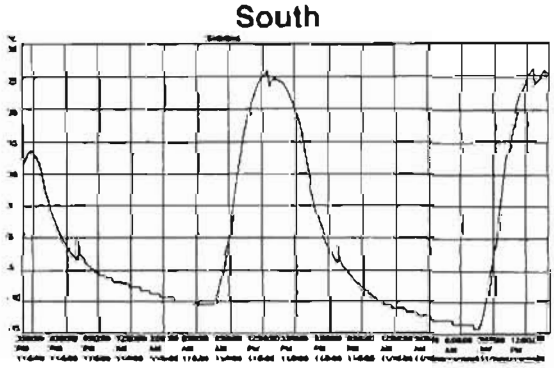
Then

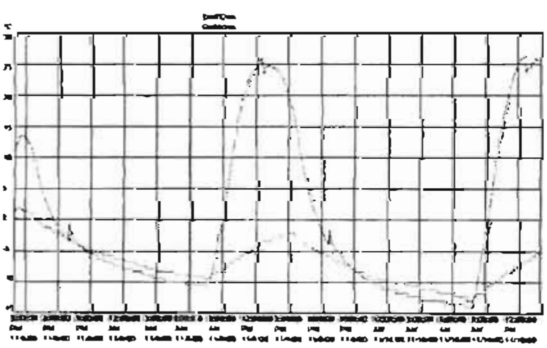
- Our frame of reference (and hence concepts and paradigms) change dramatically
- Now we must monitor warmth not cold
- We must now consider the impact of heat, not cold, on landform development.
- We must look at processes very differently and measure accordingly.

North



South





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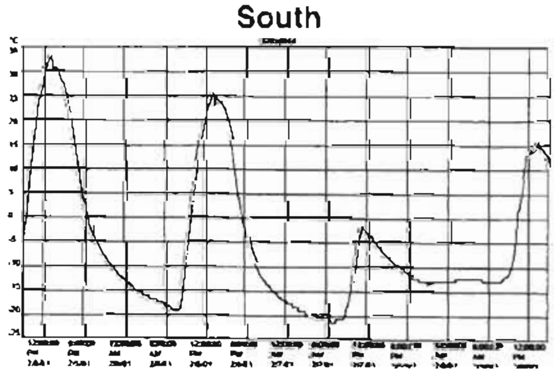

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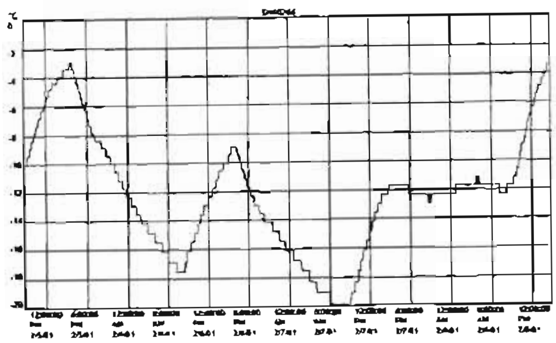
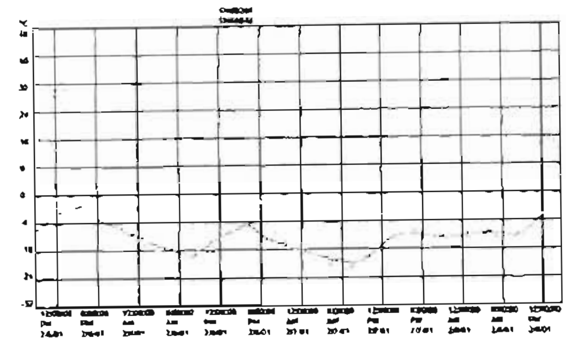




Thus, using rock temperatures rather than air temperatures, a very different weathering environment is seen.

So

- "Summers" are significantly longer
- Temperatures are high
- Temperature is *not* the limiting factor

If temperature is not the limiting factor then:

WATER is that which limits weathering

- Water limits freeze-thaw
- Water limits chemical weathering
- Water limits biological weathering
- Water limits wetting and drying
- Water limits salt weathering

Water does *not* limit thermal effects

Thus

- Thermal stresses are possible
- Thermal fatigue is possible
- Thermal shock is possible

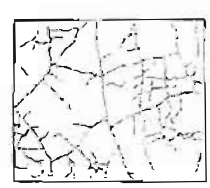
If water is the limiting factor rather than temperature

Then

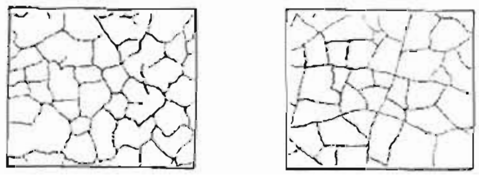
- A "cold" environment is no different to a "hot" one - 'cold' is *not* the factor as we have seen the rocks are 'hot'
- The same suite of weathering processes are cited for hot deserts as for cold - water limits in both, not temperature
- Cold deserts are no different to hot deserts
- Processes are azonal rather than zonal

There is physical evidence to support the azonality of weathering...

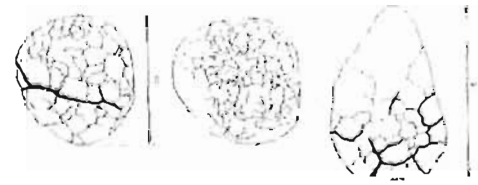
Rock fracture patterns



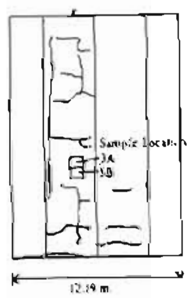
Rock surface fracture patterns



Rock fracture patterns



Fractures in an Arctic road (note scale)



Fracture patterns produced by thermal stress in a laboratory

Fracture patterns on rock in the high Andes

Fracture patterns found on the rock in the preceding figure

Fracture patterns on rock in the high Andes

Thus, if water, rather than temperature, is the limiting factor for weathering, and fracture patterns from both cold and hot environments are comparable, then:

- Processes are azonal
- Concepts of 'cold region' weathering are misleading
- In 'cold regions' we need to give more recognition to chemical weathering and processes other than freeze-thaw.

"It is a capital mistake to theorize before one has data. Insensibly one begins to twist facts to suit theories, instead of theories to suit facts"

Sir Arthur Conan Doyle

- Or
- *"First get your facts, then you can distort them at your leisure"*

  Mark Twain

Permafrost and Periglacial Processes, Vol 2: 103-112 (1991)

Thermal Gradients and Rock Weathering at Low Temperatures: Some Simulation Data.

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ABSTRACT

The heating of rock by insolation during subzero air temperatures may cause thermal stresses within that rock. The values of Δt may be such that fracturing due to thermal shock may occur. The uneven heating of a rock body may cause buttressing of the heated faces such that thermal stresses are accentuated. Upon removal of the heat source, rapid cooling may occur and values of Δt may be sufficient to cause thermal shock.

Replications of these thermal stresses may lead to fatigue and failure. The zone within which these stresses may occur is also one within which freeze-thaw can take place if water is present. It is suggested that thermally induced fracturing of rock in cold environments may be a significant but underrated process. Thermal results of laboratory simulation experiments are presented during which values of $\Delta t = > 500\,°C/h$ occurred for short periods.

RÉSUMÉ

L'échauffement de roches par insolation pendant des périodes où la température de l'air est inférieure à 0 °C peut causer des tensions dans les roches. Les valeurs de variations de température peuvent être telles que des fracturations des roches par choc thermique peuvent se produire. L'echauffement inégal d'une masse rocheuse peut provoquer des tensions des faces échauffées qui accentuent les stress thermiques. Lorsque la source de chaleur disparaît, un refroidissement rapide peut survenir et les variations de températures peuvent être suffisantes pour causer des chocs thermiques.

Des répétitions de ces efforts peuvent conduire à une fatigue des roches et à leur rupture. La zone dans laquelle ces efforts surviennent est aussi celle dans laquelle les alternances gel-dégel peuvent prendre place si de l'eau est présente. Il est suggéré que la fracturation des roches par fluctuations thermiques dans les régions froides peut être un processus significatif quoique sous-estimé. Des résultats d'expériences simulant ces fluctuations thermiques sont présentées. Pendant ces expériences, les valeurs des fluctuations de la température ont été égales ou supérieures à 500 °C par heure pour de courtes périodes.

KEY WORDS: Thermal shock Thermal fatigue Freeze-thaw Weathering Cold regions

INTRODUCTION

In recent years there has been a significant increase in detailed, quantitative studies of mechanical weathering processes (Yatsu, 1988). However, one aspect that has received little attention, particularly in cold regions, is that of thermal gradients and rate of change of temperature within rock. These factors greatly influence a number of mechanical weathering processes and, in some cases, certain theories rely upon specific rates. In the case of freeze-thaw weathering, the rate of change of temperature and

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the temperature gradient within the rock are both of central importance.

Thomas (1938), from an early study of the effects of freezing on building materials, noted that it was the cooling rate in the early part of the freeze cycle that exerted greatest influence on rock damage. This rate of cooling had far more effect than the actual temperature ultimately reached. Thomas argued that with a low rate of freezing, ice extrusion would occur and thus large pressures would not be achieved. Conversely, with more rapid temperature changes, less extrusion takes place and more unfrozen water would be available within the rock to generate high interstitial pressures. Newton's Law of Cooling states that the rate of heat loss from a warm body is proportional to the difference in temperature between that body and the surrounding medium. Thus, with large temperature differences between two mediums (i.e. the rock and the air), the temperature gradient within the rock can also be expected to be great and the rate of temperature change, particularly in the outer part of the rock, to be very large. Thomas (1938, p. 72) also argued that the ratio between the exposed surface and the total volume of the specimen affects the temperature gradient. If the temperature difference between the specimen and its surroundings is large, then large temperature gradients result, owing to the outer faces cooling more rapidly than the rock interior. One consequence of this is the increased likelihood of unfrozen water being trapped within the rock and high pressures being developed.

More recently, Michaud *et al.* (1989) suggest that 'frost bursting', the explosive failure of intact, massive rock, can occur when saturated rock is subject to intense, rapid freezing. The hydrostatic pressure developed in the pores and cracks of the rock generates strain energy which, if able to be stored by the rock, may ultimately be released in an explosive manner. According to Michaud *et al.*, a fast rate of freezing helps increase the strength of the rock by sealing pores and fractures with ice, thereby transferring the mass into a continuous rock medium which is thus able to store strain energy. Bout (1982) also refers to 'thermal shock' associated with a +20 °C to −30 °C temperature change taking place within a 24 h period while Le Ber and Oter-Duthoit (1987) note the operation of thermal shock on saturated rocks at temperatures below −3 °C.

In other studies the rate of temperature change is often cited as significant. For instance, Battle (1960) suggested that this rate needed to be of the order of ≥0.1 °C/min for breakdown to occur, while Walder and Hallet (1985) hypothesize that the rate should be very slow, preferably in the region of 0.025-0.1 °C/h. However, a rock does not experience a singular rate of temperature change, but rather, at any given depth, the rate changes linearly with time (Walder and Hallet, 1985; Hall, 1988)—the so-called 'omega component' lag effect described by Jerwood *et al.* (1987). It is worth noting that Jerwood *et al.* (1987, p. 142) also state that: 'Rock freezing and thawing rates cannot be predicted directly from air cooling and warming rates. Rates of freezing and thawing are related to temperature differentials, and these in turn are dependent on the point at which freezing and thawing begin and end ...'. However, although many workers note the importance of the rate of temperature change upon the rock body, actual data are extremely rare (Table 1). Data which are missing relate to the change of temperature with depth (the temperature gradient) and its change through time together with the rate of temperature change at given depths and how this changes with time.

Yatsu (1988) shows that sudden increases of heating or cooling of a rock body create steep temperature gradients. If a high gradient is set up within a thin layer, then spalling can occur. Thus, with surface heating, the outer shell of the rock expands and tensile forces are created between it and the cooler, inner part of the rock (Bahr *et al.*, 1986). With the removal of the heat source, the outer layer starts to contract but if the inner part is still warming owing to lag effects, then a zone of compressive stress will develop. When the temperature change is ≥2 °C/min, then the rock cracks, usually along grain boundaries (Richter and Simmons, 1974; Yatsu, 1988). In the case of anisotropic rock '.... if there is a mismatch in the thermoelastic behaviour of minerals across their grain boundaries, internal thermal stresses may be generated when the rock is subjected to different temperatures, and the stresses thus induced may be large enough to promote the formation of new cracks' (Yatsu, 1988, p. 132).

Thus, within what is generally considered to be the freeze-thaw process *per se*, it is possible that there is a synergistic component of thermal fatigue. It follows that considerations regarding the rate of temperature change and the thermal gradient within the rock need to be related not only to their effect on the freeze-thaw mechanism but also to their role in thermal stress fatigue. While this is not the case in many environments, two situations merit attention. The first is high-altitude and/or high-latitude

Table 1 Rates of change of temperature as found by different authors.

Author	Location	Rate (°C/h)*	Depth (cm)	Other details
Michaud *et al.* (1989)	Canada	0.5	5	Cooling rate past 0°C
Hare (1985)	Canada	0.7–2.01	5	
Thorn (1975)	USA	0.26 °C/d	1	
Whalley *et al.* (1984)	Karakoram	1.9	SE-facing rock surface	
"	"	2.5	W-facing rock surface	Under clear conditions
"	"	1.99	0.3 cm wide crack	
"	"	1.5	0.5 cm wide crack	
"	"	1.3	10 cm cavity	
"	"	1.0	SE-facing rock surface	
"	"	0.7	W-facing rock surface	
"	"	0.6	0.3 cm wide crack	Cloudy conditions
"	"	0.5	0.5 cm wide crack	
"	"	0.5	10 cm cavity	
"	"	4.6	Basalt surface	
"	"	1.4	Basalt 5 cm depth	
"	"	2.7	Sandstone surface	
"	"	2.4	Sandstone 5 cm depth	
Myagkov (1973)	Antarctica	0.8 °C/min	Surface	15–20 $°C/h^{-1}$
Van Autenboer (1964)	Antarctica	16.3	Surface	49 °C in 3 h
Jonsson (1985)	Antarctica	*c.*15	2–3	In crack

* Unless stated otherwise.

locations where large radiation inputs can occur during times of subzero temperatures, thereby creating steep temperature gradients, and which, when that heat source is removed, cause very rapid changes in temperature within the outer layer of the rock. The second is with respect to freeze–thaw simulations where high rates of heating and cooling are employed. In both cases the possibility arises that rock breakdown is directly related to thermal change, rather than to the freezing and thawing of interstitial water.

As part of the British Antarctic Survey 'Fellfield Programme' a range of simulations were undertaken based upon temperature and moisture conditions monitored in the field (Hall, 1986a, 1988). Large blocks of rock were heated by infrared lamps during subzero temperatures to simulate Antarctic conditions. The results of these simulations, in terms of rates of temperature change and thermal gradients, are presented here.

METHODOLOGY

By use of a computer-controlled climatic simulation cabinet (Hall *et al.*, 1989) temperature conditions similar to those experienced on Signy Island (Maritime Antarctic) were replicated. Typical rock moisture content and chemistry are already known (Hall, 1986a; Hall *et al.*, 1986). In the early experiments these conditions were replicated and air-based freeze–thaw cycles were used (Hall, 1988). Subsequently, two rock types (details given in Hall, 1988) were subject to heating by variable control infrared lamps once the samples had attained a temperature of *c.* −19 °C. The cabinet air temperature was maintained within the range −19 °C to −10 °C (some warming occurred during use of the lamps). Some warming cycles were short and intense; others of longer duration and less intense. Heating by the lamps was removed by turning them off. In this way an attempt was made to simulate various forms of heating by the sun during periods of subzero air temperatures. The heat source, being suddenly removed, simulated the case of a cloud covering the sun or the rock going into shadow. During these simulations temperatures of the air, the rock surface, and at depths of 2.8 cm and 3.3 cm (sample 1) and of 2.8 cm and 3.1 cm (sample 2) were measured every minute. In these experiments the rocks were dry so as to exclude any complications that the presence of water could induce.

RESULTS

Freeze–thaw simulations utilized cycles of −3 °C (at 4.2 °C/h), −6 °C (at 3 °C/h) and −20 °C (at 1 °C/h and 3 °C/h). Data obtained during multiple replication of these cycles on quartz-micaschist show rates (Table 2) that cause no thermal stresses because they are much too small (within the range 0.2–3.0 °C/h); details regarding these experiments and their results are given in Hall (1986a, 1988). It has been shown (Hall, 1988) that rock temperature decreases linearly with time (e.g. $r = -0.98$ for the −6 °C cycle) and that the rate of temperature change also varies in a linear fashion ($r = 0.92$). The implications of these rates with respect to the freeze–thaw mechanism are also discussed in Hall (1988). Suffice it here to state that within the simulated *air temperature* freeze–thaw cycling used in these experiments, the monitored thermal gradients and rate of change of temperature were not sufficient to cause thermal stress fatigue. In additions, during none of the freeze cycles were conditions conducive to those required by Thomas (1938) or Michaud *et al.* (1989) to produce any form of 'frost bursting'. If anything, during some freeze cycles the attributes were more akin to those suggested by the hypothetical model of Walder and Hallet (1985).

Table 2 Examples of rates of change of temperature measured during the −20 °C freeze cycle.

Location	Temperature (°C)	Time (decimal h)	Rate (°C/h)
Surface			
	0 to −1	0.417	2.4
	0 to −3	1.084	2.8
	−1 to −3	0.667	3.0
	−3 to −6	1.268	2.4
	−6 to −10	1.680	2.4
	−10 to −15	1.935	2.6
	−15 to −19.6	15.726	0.29
2.8 cm depth			
	0 to −1	0.416	2.4
	0 to −3	1.182	2.5
	−1 to −3	0.766	2.6
	−3 to −6	1.232	2.4
	−6 to −10	1.619	2.5
	−10 to −15	1.927	2.6
	−15 to −19.2	15.065	0.28
3.1 cm depth			
	0 to −1	0.416	2.4
	0 to −3	1.165	2.6
	−1 to −3	0.749	2.7
	−3 to −6	1.222	2.5
	−6 to −10	1.646	2.4
	−10 to −15	1.927	2.6
	−15 to −19.4	15.065	0.29

Examples of the sort of rates of temperature change measured when infrared heating was turned on and off are given in Tables 3 and 4. As expected, the rate of change was greatest at the rock surface and decreased with depth. The rate of change also decreases linearly with time in both cases ($r = 0.99$ and $r = 1.0$, respectively). However, rates are extremely high *for short periods of time*, with values far in excess of the 2 °C/min (120 °C/h) deemed sufficient for thermal fatigue by Yatsu (1988) and others. Using linear regression, it appears that rates of 2 °C/min operate to a depth of *c.* 2.2 cm. Thus, the outer 2 cm shell of the rock experiences alternating tensile and compressive stresses associated with the heating and cooling phase, respectively. This situation is highly likely to produce thermal stress fatigue.

In the outer shell of the rock there is a very steep temperature gradient ($\bar{x} = 7.5$ °C/cm for the outer 3.3 cm). Within this shell, freezing and thawing can also take place if water is present. From the available data it would seem that it is approximately the outer 1 cm of the rock within which interstitial water can thaw and then be subject to very sudden freezing. Because the zone within which thermal stress fatigue can operate encompasses that within which freeze–thaw can also occur, it may be extremely difficult to discern the role played by either in rock breakdown.

Two further factors may operate to aid weathering via thermally-induced stresses. First, the rock is not heated equally, and, second, parts of the rock may be actively warming while others are cooling. In the first instance, as replicated in the simulations, certain faces receive heat while others are still in shadow (Figure 1). Thus, the temperatures presented in Figure 1 are with respect to the heated surface. However, those surfaces not receiving direct incoming radiation will not expand and will act to constrain the lateral expansion of those which are heated. In other words, a form of 'buttressing' (Folk *et al.*, 1982) takes place. With respect to the second factor, the temperature data show (Figure 2) that the interior of the rock continues to experience warming when the outer part

Table 3 Rates of change of temperature experienced when infrared lamps turned on, with initial air/rock temperature at − 19 °C.

Location	Temperature (°C)	Time taken (decimal h)	Δt (°C)	Rate (°C/h)
Surface				
from	−19.1			
to	−18.1	0.017	1.0	58.8
	−9.6	0.017	8.5	500.0
	−2.5	0.013	7.1	546.2
	+2.3	0.019	4.8	252.6
	+4.0	0.015	1.5	100.0
2.8 cm depth				
from	−18.6			
to	−18.4	0.017	0.2	11.8
	−17.3	0.017	0.7	41.2
	−16.3	0.013	1.0	76.9
	−15.3	0.019	1.0	52.6
	−14.5	0.016	0.8	50.0
	−13.9	0.015	0.6	40.0
	−13.3	0.019	0.6	31.6
	−13.2	0.017	0.1	58.8
	−12.9	0.016	0.3	18.7
3.1 cm depth				
from	−18.8			
to	−18.7	0.017	0.1	5.9
	−17.6	0.017	0.9	52.2
	−16.7	0.013	0.9	69.2
	−15.6	0.019	1.1	57.9
	−15.0	0.016	0.6	37.5
	−14.3	0.015	0.7	46.7
	−13.7	0.019	0.6	31.6
	−13.3	0.017	0.4	23.5
	−13.3	0.016	0	0

cools following the heat source removal. This creates the situation whereby the inner body is attempting to expand while the outer part is contracting. The result is a zone of compressive stress which, because it is not uniform, may cause shearing.

Michaud *et al.* (1989) describe frost bursting as a result of strain energy developed within a saturated rock, subject to rapid freezing. The hydrostatic pressure of the unfrozen water develops in the pores and cracks of the rock body. Both Le Ber and Oter-Dethoit (1987) and Michaud *et al.* (1989) specify that the rock must be saturated for frost bursting to occur. The presence of saturated rock during winter in Antarctica, on other than perhaps a wave-cut platform or beach, must be questioned (Hall, 1986b; Trenhaile and Mercan, 1984). Intrinsically, there does not appear to be a problem with the theory; rather, it is the practicality of saturating the rock which is in question. However, if the rate of temperature change is rapid and the rock is able to constrain the energy, there is no reason why strain energy should not develop, even in the total absence of water during either the heating or cooling events. During either phase, the rock mass is neither expanding nor contracting in a uniform fashion; rather, the outer part experiences the greatest change, while with increasing depth the amount of attempted change decreases (Hockman and Kessler, 1950). This can be modelled as a series of zones each experiencing different amounts of change (Figure 3). If the rock can withstand this without shear, then strain energy builds up only to be released catastrophically, in the manner described by Michaud *et al.* (1989), along a pre-existing line (or lines) of relative weakness.

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Table 4 Rates of change of temperature experienced when infrared lamps

Location	Temperature (°C)	Time taken (decimal h)	Δt (°C)	
Surface				
from	+9 2			
to	+2 4	0.017	6.8	
	−5.2	0.018	7.6	
	−8.9	0.016	3.7	
	−10 7	0.015	1.8	
	−12.1	0.020	1.4	
	−13 3	0.031	1.2	
	−14.1	0.034	0 8	
	−15.1	0.049	1.0	
	−16.0	0.069	0.9	
	−17.1	0.148	1.1	
	−18.0	0.500	1.0	
	−19.0	2.599	1 0	
2.8 cm depth				
from	−13.3			
to	−14.0	0.017	0.6	35.
	−15.0	0.217	1.0	4.(
	−16.0	0.215	1.0	4 7
	−17.0	0.317	1.0	3.2
	−18 0	0.667	1.0	1 5
	−19 0	2.698	1.0	0 4
3.3 cm depth				
from	−15.5			
to	−15.0	0.151	0.5	3.3
	−16.0	0.627	1.0	1.6
	−17.0	0.303	1 0	3.3
	−18.0	0.565	1.0	1.8
	−19.0	1.335	1.0	0.8

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Michaud *et al.* (1989) cite rates of temperature change of the order of 0.5 °C/h. These are far below that required for thermal stress fatigue (2 °C/min: Yatsu, 1988). However, these authors were attempting to develop a hypothesis with respect to *frost* action in a *saturated* rock. In this study, the rocks were not saturated but rates of temperaure change were, for short periods of time, in excess (by as much as a factor of 4) of that considered to be the threshold for normal strain fatigue. That the rocks used in the simulation experiments did not shatter does not negate the hypothesis of strain energy; multiple replications are needed to weaken the rock sufficiently until it ultimately fails catastrophic… when no longer able to constrain th…
In the Dry Valleys of … Antarctica and … frac…

cluded) and experience very low air temperat… but with strong radiation inputs. Although weathering is probably operative in these locatic… no signs could be found. It was noticeable, partic… larly in the Andes, that the cracking pattern w… often polygonal, somewhat akin to frost cracking i… permafrost regions. Qualitatively, these observa… tions suggest that many of these rocks shatter by thermal stress fatigue associated with rapid tempe… ature changes. Certainly, the crack … served are similar to those … (1986, Figure 2) … duced …

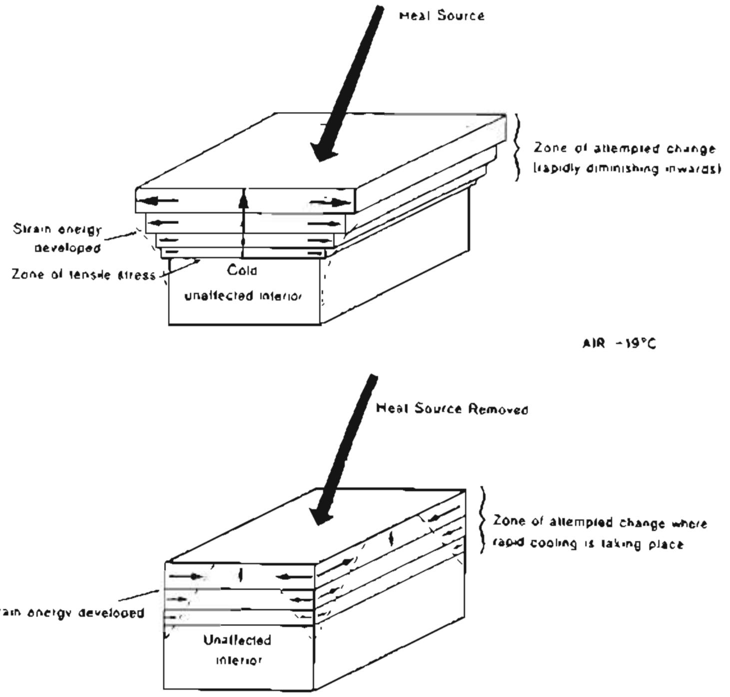

Figure 3 A simple block model, showing the development of tensile stresses during the heating phase and the compressive stresses during cooling

continued heating while the outer shell is cooling generate a zone of compressive stress. Because these forces are not uniform, they might cause shear stress.

In summary, the main results of these simulation experiments indicate:

(i) Freeze-thaw cycles induced by changes in air temperature are insufficient to cause thermal stress fatigue.
(ii) The concept of frost bursting as proposed by Michaud *et al.* (1989) appears viable *if the rock is close to saturated.*

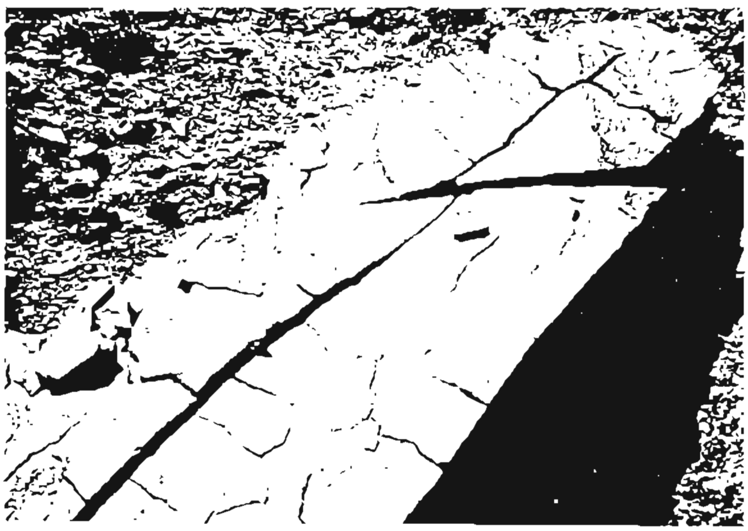

Figure 4 A large rock block at 4300 m a.s.l. in the Andes, showing fracture patterns considered to result from thermal stress fatigue.

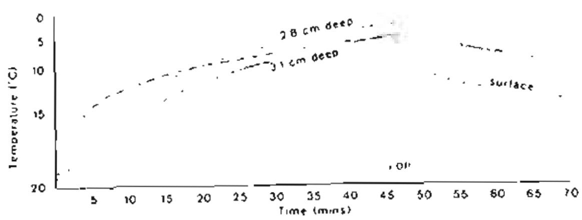

Figure 5 Temperature data from a rock block subject to angled heating, showing that internal temperatures can be greater than at the top surface of the rock.

(iii) Sudden removal of radiative heating to rock at subzero air temperatures generates high values of Δt.
(iv) The zone within which thermal stresses are generated encompasses that within which freeze–thaw can also take place *if water is present*.
(v) Because of uneven heating (due to an angled heat source), it is possible that (a) a form of buttressing takes place which enhances surface shear stresses, and (b) subsurface temperatures can be greater than the upper surface of the rock.

DISCUSSION

These results have implications with respect to the perception of weathering in cold climates. Eichler (1981), from work on Ellesmere Island, showed that rock surface temperatures could be very high (39.7 °C) and that, owing to shadow effects, large thermic diurnal rhythms take place. Those rocks on slopes which were *not* radiated equally all round were said to show signs of thermal strain fatigue as expressed by very strong weathering features. Eichler (1981, p. 442) concluded that: 'Not the frost but the insolation appears to be the main agent in the high arctic temperature weathering.' Bout (1982) also discusses the role of thermal shock in causing the breakdown of basalt in Iceland, and Le Ber and Oter-Duthoit (1987) refer to the role of thermal shock on rock wall weathering in France. Gunn and Warren (1962, p. 60) consider that the breakdown of fine-grained rocks in Victoria Land (Antarctica) is due to the 'forces exerted by rapid temperature change' with a temperature range of 60 °C being cited.

In more general terms, greater cognisance must be taken of thermal changes beyond their direct role within freeze–thaw. For instance, fractured rock present within cold environments need not, of necessity, imply freeze–thaw weathering alone, or in cold arid environments the action of salt weathering. The possibility arises of thermal stresses, via fatigue or catastrophic failure, either operating alone or synergistically with other weathering processes. Ultimately, the problem rests upon acquiring field data on moisture content and chemistry, rock temperatures and a rate of change of temperature (Hall, 1986c). Until such data are available, the ability to apply theoretical models of freeze–thaw action such as that of Walder and Hallet (1985) or mechanisms such as that of 'frost burst' (Michaud *et al.*, 1989) are fraught with difficulties, as too is the understanding of the role of thermal stress effects on rock breakdown. In summary, simplistic and qualitative judgements regarding the nature of weathering in cold regions can no longer be accepted.

CONCLUSIONS

Rapid freezing of a saturated rock may cause severe stresses that are released catastrophically, the so-called 'frost bursting' process. This thermally-induced stress can either cause catastrophic cracking or operate through time via fatigue. Frozen rock can also be subject to thermal stresses in the outer shell when warmed by direct insolation and then suddenly cooled when that heat source is removed. Freeze–thaw can also take place in this outer shell if water is present. Buttressing of rock that is attempting to expand by unheated zones may add to the tensile forces generated. Simulation data indicate that the potential for thermal stress fatigue can be present and that cracking of rock might take place as a result of thermal stresses even in the absence of water and thus the freeze–thaw component. It is tentatively suggested that many cracked rocks seen in cold environments owe their origin to thermal stresses but more data are required on rock temperature and moisture conditions to verify this.

ACKNOWLEDGEMENTS

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New insights into rock weathering from high-frequency rock temperature data: an Antarctic study of weathering by thermal stress

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Abstract

A major limitation of many weathering studies has been the acquisition of rock temperature data at insufficiently frequent intervals for the meaningful determination of the rate of change of temperature ($\Delta T/t$). Equipment and/or logistical constraints frequently facilitate temperature measurement at only hourly intervals or, at best, every 10 min. Such data are not adequate for the determination of $\Delta T/t$ required for the evaluation of the freeze–thaw mechanism or thermal stress fatigue. Recent undertakings at different sites in Antarctica (and at other cold-region locations) provide rock temperature measurements at 1-min intervals, which indicate that the perception of the weathering regime would be very different from that generally assumed from longer-interval geomorphological data. These data clearly show that thermal stress fatigue and thermal shock may be more active components of the Antarctic weathering regime than have generally been recognised: the aridity of the study area limits the role of freeze–thaw weathering. Values of $\Delta T/t$ of ≥ 2 °C min^{-1} that suggest thermal stress fatigue/shock is operative were recorded; observations of rock flaking are thought to reflect the impact of thermal stress. Further, the data show that contrary to general perceptions, the southern aspect can, in summer, experience higher rock surface temperatures than the north-facing exposure. An examination of rock fracture patterns found in the field shows great similarity to fracture patterns developed in the laboratory as a direct result of thermal shock. The argument is made that greater cognizance should be given to thermal effects. © 2001 Published by Elsevier Science B.V.

Keywords: Weathering; Rock temperatures; High-frequency measurements; Thermal stress; Thermal shock; Antarctica

1. Introduction

In any consideration of rock weathering, there are three key factors, namely rock temperature, rock moisture and rock properties. These attributes, independently and in synergy, exert a major influence on the type, degree and rate of weathering that takes place at any given site. For any meaningful investigation of rock weathering, detailed data pertaining to each of these attributes are required. Unfortunately, consideration of much of the available literature indicates that one or more of these key components will usually be assumed (e.g. that water is both present and in sufficient quantities for freeze–thaw weathering to occur under freezing conditions) and that as a result, the nature of the weathering process is inferred rather than proved. The same problem is

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present with many laboratory simulations: temperature and moisture conditions rarely mimic natural conditions simply because those attributes have never been measured. Studies of weathering require that the various attributes of the three key parameters be measured in the field. Without such data, simulations cannot be related back to any specific site and knowledge of the weathering taking place at that site is unfounded. Here, attributes of the rock temperature data that change the general perception of rock weathering in cold regions are presented.

In any consideration of rock temperatures, there are five main factors that must be recognised:

- air temperatures are not a surrogate for rock temperatures;
- the minimum requirement is that rock surface temperatures be measured;
- rock temperature at various depths should be measured;
- the thermal gradient within the rock should be calculated; and
- values of the rate of change of temperature ($\Delta T/t$) are required.

All too frequently, air temperatures are used to exemplify the weathering regime. Unfortunately, such undertakings are meaningless (Thorn et al., 1999) for they take no account of rock warming by the sun (even at times when air temperatures may be severely sub-zero), rock to rock variations resulting from albedo affects (Kelly and Zumberge, 1961; André, 1993), or situations where the rock may in fact not experience any thermal fluctuations because it is insulated by a thick snow cover. Even when rock temperatures are recorded, this rarely includes sufficient information to evaluate thermal gradients. Equally, for a variety of reasons, collection of rock temperature data is rarely, if ever, at a frequency sufficient for any meaningful evaluation of the rate of change of temperature ($\Delta T/t$). Both of these attributes, thermal gradient and $\Delta T/t$, are important for evaluating the weathering process(es). Indeed, measurements of $\Delta T/t$ are imperative for any understanding of the freeze–thaw mechanism as the various mechanisms have a variety of controlling rates which constrain whether they can operate, e.g. 0.1 °C min^{-1} as suggested by Battle (1960). Beyond the simplistic assumption of freeze–thaw (Hall, 1995), other weathering processes, notably thermal stress fatigue and thermal shock, are constrained by (among others) the rate of change of temperature. It is the measurement and importance of $\Delta T/t$ that will be discussed here.

Paradoxically, in so-called "cold climates," it may well be heat which is a major factor with respect to geomorphic processes. The very name "cold climates" implies, almost exclusively, the role of 'cold.' As a result, most discussions of cold region weathering tend towards consideration of only mechanical processes, and usually, within those arguments, the dominance of freeze–thaw weathering (see Hall, 1995, 1999 for discussions). However, such an approach negates the impact of the summer, short as it may be at high latitudes, plus the influence of rock warming within the otherwise cold environment. Recent studies (e.g. Balke et al., 1991) have shown that in cold regions, it is not the temperature that is the limiting factor for chemical weathering, even in the Antarctic, but rather it is the availability of water. Thus, the measurement of rock temperature, and its temporal and spatial variability, is critical for the understanding of rock weathering. Central to any measurement of rock temperatures must be the evaluation of $\Delta T/t$. Not only is this critical to understanding any freeze–thaw activity, but can also itself be a cause of weathering—by thermal stress and thermal shock. Thermal stress fatigue and/or thermal shock are usually considered under the synonym "insolation weathering." Unfortunately, this term generates wrong perceptions. First, insolation does not "weather"—it is only one of the driving forces for the thermal changes that actually cause the weathering. This is very important indeed, for a number of studies have, for example, shown that sub-surface rock temperature fluctuations are driven by variations in wind speed even where air temperature and radiation input are held constant (e.g. Nienow, 1987). Second, the role of 'insolation' seems more appropriate to "hot environments" than to "cold" ones—as shown by any comparison of "Hot Desert" geomorphology texts (e.g. Abrahams and Parsons, 1994) with "Cold Environment" geomorphology texts (e.g. French, 1996). Thirdly, the general perception of the role and significance of "insolation weathering" has been severely reduced, at least in geomorphology, by on-going reference to the studies of Blackwelder (1933) and Griggs (1936). The unquestioning acceptance of these studies (see Ollier, 1984 for arguments about this) has led to many geomorphologists discounting the possible role of thermal stresses in the breakdown of rock (see Hall, 1999 for a wider discussion, including the impact of differential variations in $\Delta L/°C$ as a function of crystal axes). Once recourse is made to engineering and ceramics studies, where thermal stress and thermal shock are central to many investigations, the potential of the role of thermal variations in causing rock disintegration becomes evident. Thus, if the negative attributes of the three discussion points above can be overcome, the impact of $\Delta T/t$ upon weathering in cold environments can be considered.

The role of $\Delta T/t$ is even more significant when it is realised that it is 'temperature independent,' i.e. that the rate of change of temperature that might effect damage can be anywhere on the temperature scale. Studies have shown (e.g. Richter and Simmons, 1974; Yatsu, 1988) that the threshold value for thermal shock approximates to a rate of temperature change of 2 °C min^{-1}. Values equal to or greater than this cause the rock to try and adjust at a rate that is greater than its ability to deform plastically and so the rock fails. That value can, however, be anywhere on the temperature scale: from +32 to +34, from −15 to −17 °C; it is not constrained to freezing temperatures or to positive temperatures, and does not require the presence of water. Once this is accepted, then it can be seen that 'cold environments' may well be ideal locations for such events. A typical scenario, particularly for the Antarctic, would envisage rock exposed to sub-zero air temperatures (perhaps as low as −30 °C) but being heated by incoming radiation on a clear day. That source of incoming energy is then "switched off" by cloud covering the sun or the sun moving behind a peak. At that point, the temperature differential between the outer layers of the rock and the air is very large (e.g. +10 °C rock to −30 °C air) and so, following Newton's Law of Cooling, the rate of change of temperature ($\Delta T/t$) will be very high, potentially ≥ 2 °C min^{-1}. The same happens, but in the opposite temperature direction, when the sun then hits the cold rock once the cloud has passed or the sun emerges from behind a peak. A whole range of differential stress fields result from these temperature changes (see Hall, 1999), and it is these which offer an explanation for the flaking observed on the rocks at the study site. Indeed, Dragovich (1967, p. 801), in a discussion regarding flaking, cites the suggestions by Kvelberg and Popoff (1937) and Cailleux (1953) that "... the surface–rock layer ... is affected by cool air which descends rapidly over it. This abrupt lowering of temperature forces the rock surface to contract and buckle outward from the underlying rock, thereby causing flakes to develop."

Recognizing from above, the frequency of temperature measurement then becomes the key issue. The obtainment of high frequency temperature data has hitherto been limited by a number of factors. At the conceptual level, the need to record at intervals of 1 min has not arisen. Assuming freeze–thaw weathering, most studies have required some measurement of the amplitude and length of freeze and/or thaw; in the field, $\Delta T/t$ has not been seen as significant in this regard. This conceptual component has been compounded by practical and logistical constraints. The availability of sensors that can resolve temperatures, and loggers that can record at such intervals, are reasonably recent innovations. Even then, logger memory size and battery maintenance in cold environments, exert extreme limitations on the logistics required for such undertakings; battery changing and memory downloads become frequent, perhaps daily. Under such practical and philosophical constraints, it has hitherto been deemed expedient to worry more about freeze and thaw durations and amplitudes to the detriment of any consideration of $\Delta T/t$. Interestingly, it is from biological studies that the best high-frequency data come (e.g. Kappen et al., 1981; McKay and Friedmann, 1985; Friedmann et al., 1987; Nienow, 1987), although more recently, a number of geomorphological studies have collected 1-min records for short periods of time (e.g. Warke et al., 1996). In Antarctic and Canadian studies, Hall (1997, 1998, 1999) has collected temperature records at 2-min, 1-min and 30-s intervals for periods ranging from 1 week to 2 months and, more recently (Hall, unpublished) for a period of 1 year. In this paper, recent data (Hall, 1999) collected from the Antarctic will be used to show the value of 1-min temperature data, and to help justify the argument regarding the role of thermal stresses in cold environments.

2. Study area and procedures

Data were collected from the top horizontal surface of a rock outcrop as well as from the north, south, east and west aspects at (Fig. 1) the British Antarctic Survey Rothera base (67°34'S, 68°07'W). The site (Fig. 2) comprises a medium-grained granodiorite at an elevation of approximately 30 m a.s.l. Thermistors were located on the rock surface and data were collected every minute; a number of thermistors were also located in cracks, lichen cover, fine debris, etc. but are not discussed here. Data were collected by an "ACR Systems" 'SmartReader' logger that has a resolution better than 0.1 °C and can store 32,768 readings; accuracy of the thermistors was ±0.2 °C. Additional data on humidity at the rock/air interface were also collected at 1-min intervals but are not discussed here.

3. Results and discussion

The data set for part of 11th December 1999 are ideal for showing the value of 1-min data (Fig. 3). Part 1 of Fig. 3 shows the complete data set for December 11, 1999 while part 2 shows a 2-h record for the five sample locations. Part 3 is the detail for the horizontal surface for that 2-h period. It is this

Fig. 1. Location map to show the position of the study site at Rothera Scientific Station.

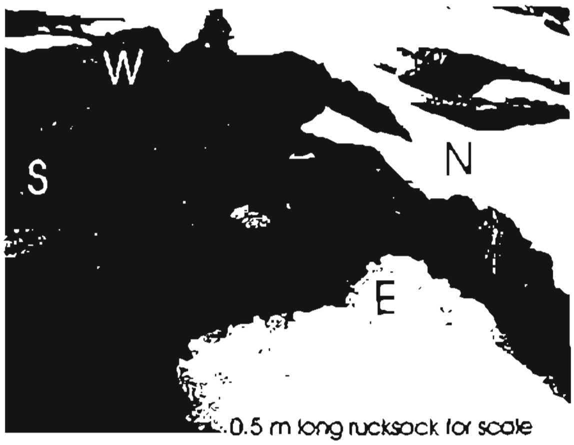

Fig. 2. Photograph of the rock outcrop where the sensors were located. The four cardinal directions are marked and a rucksack is shown for scale.

that will be used here to indicate the value of the 1-min data. Fig. 3, part 3, shows that over the 2-h period, there were a number of fluctuations. Specific among the many fluctuations is that from 0718 to 0721 h, during which there was a marked drop in temperature (from 5.4 to -0.3 °C), which will be discussed shortly. Fig. 4, part 1, shows the detail of temperatures that would be shown for that same 2-h period if temperatures had been recorded every 10 min; such a rate being reasonably high in most published studies. Evaluation of that record (Fig. 4, part 1) would indicate the highest rate of change of temperature to be 6.3 °C/10 min for the rising limb from 0630 to 0640 h. If, however, these data are then placed on top of the actual information (Fig. 4, part 2), it can be seen what an abstract of reality the 10-min data represent. Many fluctuations are missed by the 10-min record. For example, during the first hour (0600–0700 h), the 10-min data show a rise to 4 °C and then a decline to 2.2 °C while, for that same period, the 1-min data show nine rises and falls. Equally, the 0700–0800-h record for the 10-min data shows a fall in temperature while the 1-min information indicates eight rises and falls. During 0700–0800 h of falling limb for the 10-min data, there was, in reality, a significant peak to the highest temperature in this record period followed by a rapid fall. Clearly, the 10-min data are inadequate for any meaningful determination of thermal variation at the rock surface.

The details of the temperatures during the period 0712–0722 h on December 11 are shown in Fig. 5. The importance of these data are that during the drop in temperature between 0718 and 0719 h, $\Delta T/t$ was 3.2 °C min^{-1} and the following minute showed a $\Delta T/t$ value of 1.6° min^{-1}; 4.8 °C over 2 min. The value of 3.2 °C min^{-1} exceeds the theoretical threshold for thermal shock as does the composite for 2 min, 0718–0720 h Thus, not only would the 10-min record have completely missed all the fluctuations that actually took place, but it would have shown no indication whatsoever of the high $\Delta T/t$ value that may be significant for rock failure. Another example of just such an event is that for the eastern aspect on 11th of December between 0852 and 0854 h when successive values of $\Delta T/t$ were 2.48 and 2.87 °C min^{-1}, respectively. Again, longer interval data measurement would have failed to resolve these events. For comparison, consider the recent data presented by French and Guglielmin (1999, Table 2b, p. 334). Here, hourly data are used to determine the number of crossings of various thermal thresholds (0, -2 and -4 °C) which are then applied to an evaluation of weathering, particularly in relation to freeze–thaw weathering. If the data presented here for a 10-min record frequency can generalise, and thereby eliminate major variations to the extent shown in Fig. 5, then the question must arise as to the meaningfulness of an hourly evaluation. In fairness, French and Guglielmin (1999, p. 335) note that the aridity of the region would limit any freeze–thaw weathering. However, the evaluation of any cycling based on hourly data must be fraught with over generalization.

Significantly, the high magnitude thermal events were not found (within this data set) on all aspects. The eastern and horizontal surfaces cited above were the only ones that experienced events with a $\Delta T/t$ of ≥ 2 °C min^{-1}. On the western aspect, the greatest values recorded were 1.6 °C min^{-1} during a 3-min period (1348–1351 h on December 11) when temperatures changed by 4 °C; values of 1 °C min^{-1} were fairly common. On the southern aspect, the most common $\Delta T/t$ value was of the order 0.4 °C min^{-1}, with 0.6 °C min^{-1} being the highest recorded. The high southern value was at the same time (0852 h) as that recorded for the northern aspect when a $\Delta T/t$ rate of 1.82 °C min^{-1} was recorded, the

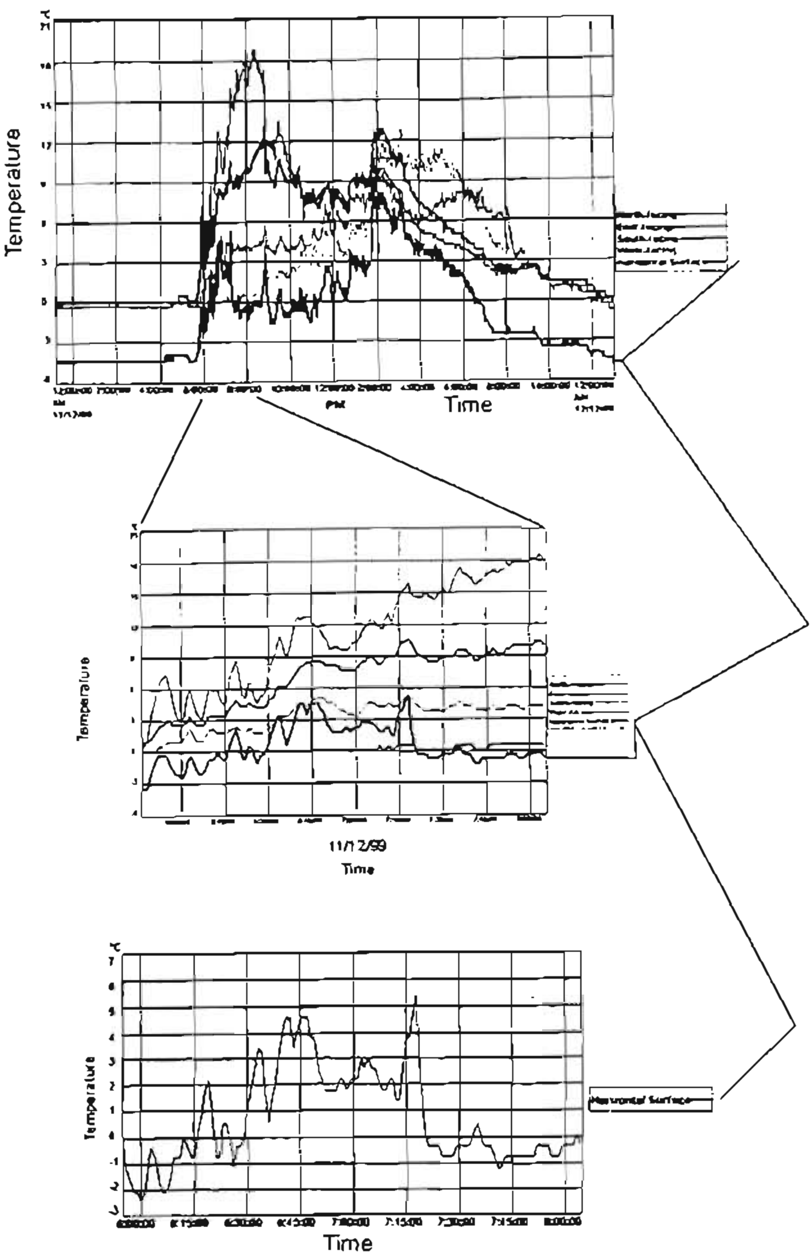

Fig. 3. Detail of the temperature data for the four cardinal directions plus the horizontal surface for the 11th of December, 1999. The temperature data for the 2-h period, 0600–0800 h, are then shown together with those for the horizontal surface during that 2-h period.

highest for this aspect. The more common $\Delta T/t$ rate for the northern aspect was in the order of 0.7–0.8 °C min^{-1} Thus, it would appear that the high $\Delta T/t$ values may not be found simultaneously on all aspects and that aspect preferences may occur depending on the time of year and the local climatic conditions. Indeed, Hall (1998, Fig. 2) showed distinct aspect differences at the Rothera site, with the northern aspect having the lowest recorded temperature and the western aspect the highest for that record period. Data for the 11 December 1999, from this present study, show a different aspect distribution:

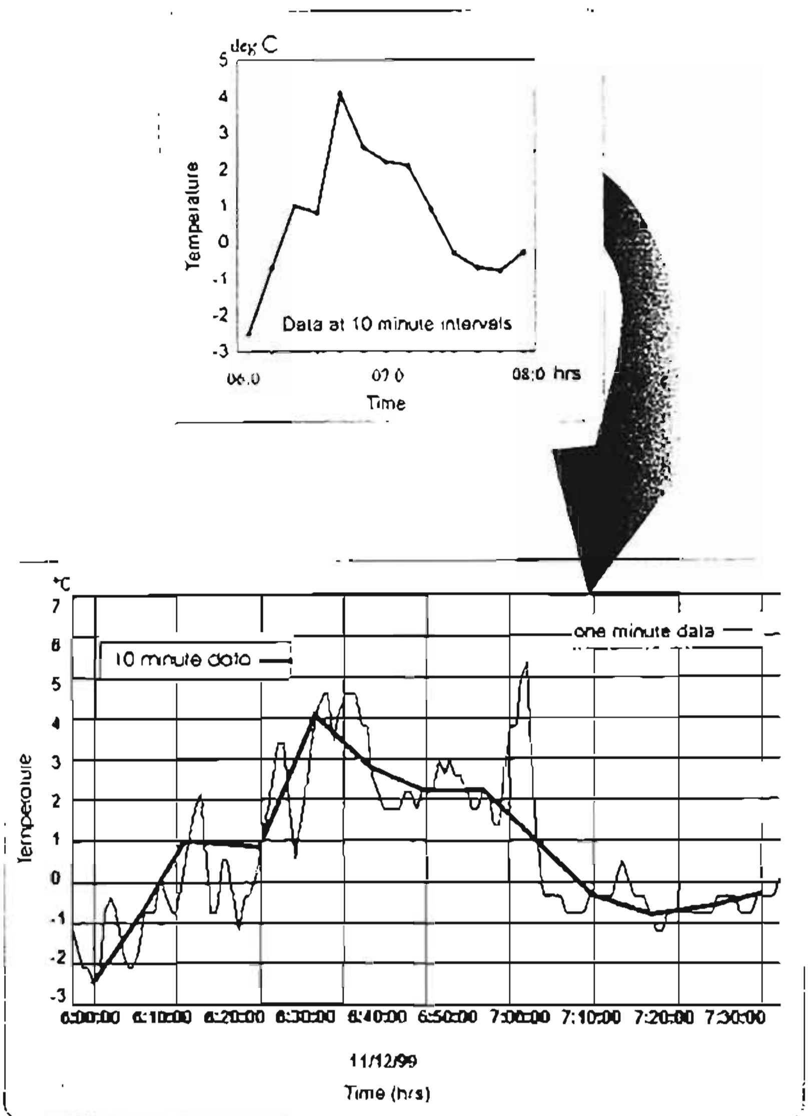

Fig. 4. The temperature curve for the 0600–0800-h period, shown in Fig. 3, based upon 10-min temperature data together with that same curve superimposed on the 1-min data recorded over the same period.

the southern aspect had the lowest temperature as well as the highest. The aspect influence is clearly more complex than has simplistically been presented here; not the least being the need for daily comparisons over a longer time period to help filter out day-to-day variations from those that might be seasonal (Hall, in preparation). However, it would appear that with respect to $\Delta T/t$ values, there is an aspect influence, and that this may vary through the spring to autumn period in terms of which aspect experiences the greatest thermal variations.

Elsewhere in the Antarctic, a number of studies provide information in support of weathering by thermal stress fatigue and/or shock. Gunn and Warren (1962, p. 60) state, with respect to the weathering of fine-grained rocks in Victoria Land, "Although experimental evidence suggests that the forces exerted by rapid temperature change are too small to fracture rocks ... it is difficult to find an alternative explanation for some ... features." In support of this argument, they note that temperature ranges as large as 60 °C may occur. Myagkov (1973), also in Victo-

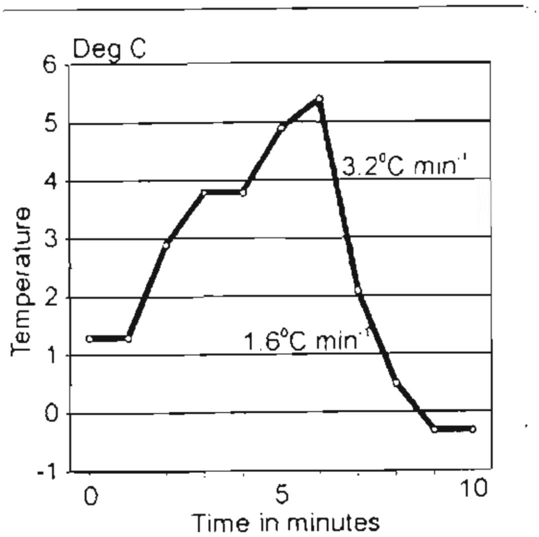

Fig. 5. Detail of a 10-min (0712–0722 h) from Fig. 4 showing the rate of change of temperature that occurred during the main thermal peak at that time.

ria Land, monitored rates of heating and cooling of 0.8 °C min^{-1} and 15–20 °C h^{-1}. Bardin et al. (1965) monitored a daily temperature range of 42 °C (+30– – 12 °C) in the mountains of Queen Maud Land, and ascribe breakdown, in part, to the thermal effects. On the Black Coast of Palmer Land, Singleton (1979) refers to the action of thermal fatigue in causing rock breakdown; the large thermal gradient causes differential expansion that results in failure. Indeed, a number of authors all refer to the role of thermal stresses in causing rock breakdown within the Antarctic: Avsyuk et al. (1956), Markov and Bodina (1961), Stephenson (1966), McCraw (1967), Bardin (1968), Black (1973), Miotke (1980, 1982), Aleksandrov and Simonov (1981), Robinson (1982), Miotke and von Hodenberg (1983), Blümel (1986), Moriwaki et al. (1986), Brunk (1989), Picciotto et al. (no date). Black (1973), Aleksandrov and Simonov (1981) and Miotke and von Hodenberg (1983) all refer to the formation of a network of fissures on the rock surface resulting from thermal stresses that may, in some instances, show a polygonal pattern. Miotke and von Hodenberg (1983) also identified a "reverse" situation whereby the thermally induced cracking appeared to be greater *inside* the rock and diminished towards the outer margins. Suggate and Watters (1991) also identified a similar "inside-out weathering," but this being on an outwash plain in New Zealand, ascribed it to hydration effects. Lastly, Ford and Andersen (1967, p. 722) refer to the cracking of ice "...consequent on tensional stresses produced by rapid temperature drops that accompany the passage of shadows over sun-warmed, debris-mantled ice slopes." This situation must be analogous to that taking place in sun-warmed rock subject to comparable rapid cooling. It is worth noting that the water-limited Antarctic environment has led to a greater consideration of thermal weathering than the case for the Arctic where, due to the moister conditions, the preoccupation has been with freeze–thaw weathering. Unfortunately, the Antarctic literature has had less impact on northern or high altitude studies, such that inadequate consideration has been given to weathering other than freeze–thaw in these areas.

A field observation that appears to support arguments in favour of a thermal stress origin for cracking is that of the crack patterns found on boulders (as noted above by Aleksandrov and Simonov, 1981 and Miotke and von Hodenberg, 1983) in a number of arid cold regions. Fig. 6 shows the crack pattern observed on a single boulder from an arid, cold region (the high Andes) and compares it with the crack patterns developed as part of a thermal shock experiment by Bahr et al. (1986). A number of key parameters are evident in Fig. 6. First, on the photograph of the boulder, the cracking cuts across pre-existing structures in the rock; evidence of folding can be seen in addition to that demarcated by white lines. It would be hard to justify any water-based weathering process (e.g. frost weathering, wetting and drying, or salt weathering) that apparently ignores existing lines of ingress/egress for water. Second, the crack patterns show a remarkable angularity of junctions; many are almost at 90° to each other (the "...angular heat-cracked..." form referred to by Holman, 1927). This, too, is hard to explain by any water-based weathering process, especially when the cracks cut across pre-existing structures. Third, there is a relative 'hierarchy' of cracks, comprising major cracks with small ones at right angles to them. In the light of these two points, this too would be hard to explain by water-based weathering processes. However, if the crack pattern produced by thermal shock

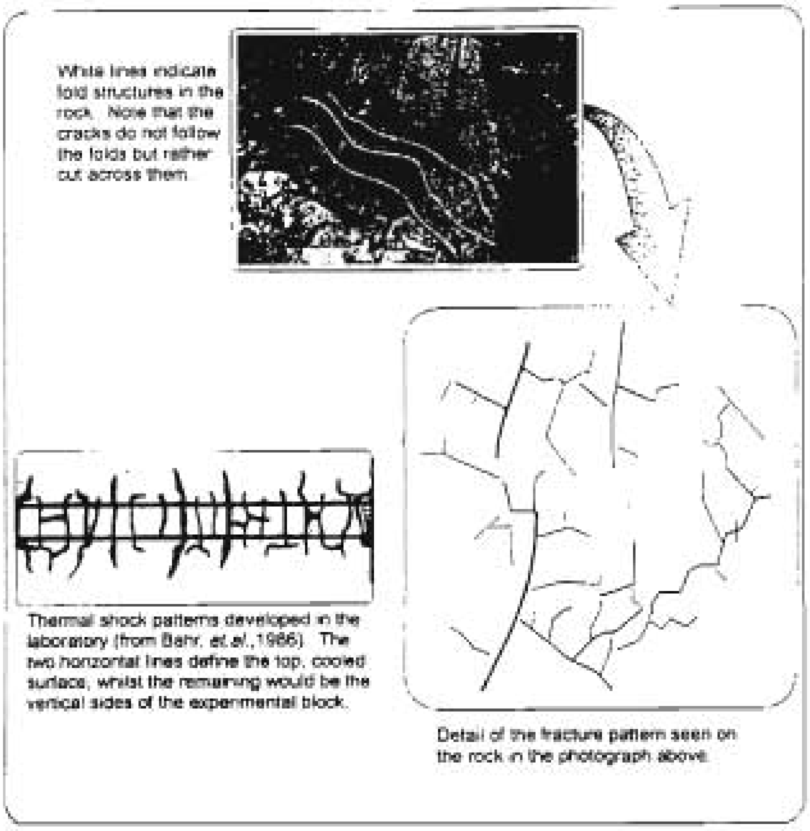

Fig. 6. The fracture patterns developed on a rock block together with those created in the laboratory during a thermal shock experiment. It is seen that the fracture patterns on the rock do not follow any of the obvious pre-existing lines of weakness in the rock and that the patterns closely resemble those generated in the laboratory.

in the laboratory by Bahr et al. (1986, Fig. 2) is compared with that for the boulder, then two immediate similarities are apparent: cracks intersect at right angles, and there is an apparent hierarchy of crack size. Consideration of Fig. 2 in Bahr et al. (1986), which shows a variety of crack patterns as a result of different $\Delta T/t$ values, will provide a better "feel" for the character of thermal shock fracture patterns, and it is suggested to be a good comparison with that shown for the boulder in Fig. 6. Marovelli et al. (1966, Fig. 16) also show fracture patterns resulting from thermal shock that are, although on circular disks, very similar to those of Bahr et al. (1986) and those shown here in Fig. 6. Hall (1999, Figs. 1 and 3) shows other comparable, fracture patterns in boulders of other lithologies from a cold, arid environment. It is difficult to envisage how water-based processes, especially freeze–thaw weathering, could produce such fracture patterns even if water was not a limiting factor at the study sites. Ollier (1963, Plate 1B) shows fracture patterns in a quartzite boulder from a hot desert that has similarities to those produced in the laboratory by Bahr et al. (1986). More significantly, Ollier (1963, p. 378) observes, as in the discussion above, that "... the quartzite has traces of cross lamination, and the direction of this, or other features such as joints, *has no bearing on the position of the cracks*" (our italics). Ollier (1963) argues that the fractures he observed are a result of thermal stresses as the central Australian desert is a water-limited environment.

"Thermal shock resistance is that property of a body which enables it to withstand sudden and severe temperature change without fracturing" (Marovelli et al., 1966, p. 8). Certainly, different rock types have different degrees of thermal shock resistance and this must play a role in any field situation. As an example, Marovelli et al. (1966) undertook experiments on basalt, quartzite and taconite and found that quartzite and taconite were less susceptible to thermal shock failure than basalt was. Consideration of the thermal fracture literature suggests that cold environments are, in fact, ideally suited for thermal fracture and may be even more so than hot deserts. Pertinent to this consideration of thermal shock effectiveness in cold regions was the finding that "... about twice as much temperature difference will be required to initiate cracks in heating shock than is required in a cooling shock" (Marovelli et al., 1966, p. 12). During cooling shock, failure will be initiated on the surface as a result of the maximum tensile stresses occurring here (Marovelli et al., 1966, p. 3). Brittle materials fail more readily (by a factor of at least 10 times) under tension than compression, and so the cold region cooling environment is, theoretically, highly conducive to thermal failure. Thus, so long as the thermal change causes the tensile stress to exceed the tensile strength of the rock, failure must occur. Theoretically, complete failure can occur as a result of a single event, but usually, multiple events are required. In that regard, differences in lithology will exert a control. In the experiments of Marovelli et al. (1966), the greater susceptibility to thermal shock exhibited by the basalt was considered to have resulted from its higher thermal conductivity. Thus, thermal failure will be as a result of the combination of rock properties (e.g. thermal conductivity), tensile strength, coupled with the magnitude and duration of the stress generated by the thermal change.

A significant factor in thermal weathering processes must be the duration of the thermal stress event. For example, the large $\Delta T/t$ events presented in this study are of very short duration. The laboratory studies of Richter and Simmons (1974) suggest that the experiments, from which the thermal threshold of ≥ 2 °C min^{-1} was derived, had the thermal change occurring for 30 min. Other studies (e.g. Marovelli et al., 1966) have almost instantaneous cooling, with very high values of $\Delta T/t$ (23.89 °C min^{-1}) across large temperature ranges (141.6–100 °C) that may be unrealistic in the sort of field situations under consideration here. The principles do, though, remain and do appear to have some field expression as exemplified by the crack patterns presented and discussed above. What needs greater clarification is the significance of the shorter duration occurrences of high $\Delta T/t$ as found in this study. A better understanding will also be aided by more extensive measurement of rock temperatures at these high frequency intervals for this will provide better insights into the occurrence (or not) of longer duration periods of thermal change and/or the frequency of shorter term events that may, collectively, play a role.

4. Conclusions

The availability of low cost data acquisition systems that allow high frequency data recording coupled with a memory capacity that facilitates storage over extended periods now offers the opportunity for the collection of temperature (and other, e.g. humidity) data at intervals that may offer new insights into weathering processes. Using a record interval of 1 min, it is possible to derive data that provide meaningful information regarding $\Delta T/t$ and facilitate a better understanding of freeze–thaw processes. This opens new avenues of study with respect to weathering by thermal processes. The information presented here clearly demonstrates the inadequacy of thermal data collected at even 10-min intervals, and shows that potentially damaging rates of $\Delta T/t$ (≥ 2 °C min^{-1}) are revealed when temperatures are recorded at 1-min intervals. While more information is still required (e.g. the impact of short-term high values of $\Delta T/t$, as shown here, versus such rates for longer time periods), nevertheless, it does offer a new perspective on weathering conditions and one which might be particularly relevant to cold environments. At the same time, as the temperature data indicate the potential for thermal stress fatigue and/or thermal shock, consideration of fracture patterns observed on rocks in cold, dry environments appears to show very similar forms to those produced in thermal shock experiments in the laboratory. Not only do these fracture patterns replicate the laboratory-produced ones, but the character of the fractures would be very difficult to explain by water-based weathering processes such as freeze–thaw or salt weathering; the water-limited nature of the environment would also limit the effectiveness of freeze–thaw as a potential fracturing mechanism. The coupling of the observed fracture patterns with the thermal data appears to suggest that greater credence should be given to weathering by thermal processes in cold, arid environments. Even the rock flaking observed in these environments can be explained by differential stresses along a temperature gradient, and although salts may also offer an explanation in some situations, their absence or limited occurrence at some localities may relegate their effectiveness to less than that of the thermal effects. Thus, the need is for more detailed measurement of rock thermal conditions over long time periods (at least 1 year) coupled with laboratory experimentation to ascertain the effectiveness of observed thermal events on the rock type under consideration. It is suggested that the acquisition of more frequent thermal data will lead to new perspectives on cold region weathering processes.

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ROCK MOISTURE CONTENT IN THE FIELD AND THE LABORATORY AND ITS RELATIONSHIP TO MECHANICAL WEATHERING STUDIES

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ABSTRACT

Rock moisture content is a major control of mechanical weathering, particularly freeze–thaw, and yet almost no data exist from field situations. This study presents moisture content values for rocks, taken from a variety of positions and conditions, in the maritime Antarctic. Additional information regarding the amount of water the rock could take up, as observed from laboratory experiments, is also presented. The results show that the approaches used in simulation experiments, particularly that of soaking a rock for 24 hours, may produce exaggerated results. It was found that the saturation coefficient (S-value) was a good indicator of frost susceptibility (based on water content) but that the derivation of that value may underestimate the potential of some rocks. The distribution of moisture within rocks is seen as an important, but unkown, factor. The results of these field moisture contents suggest that for simulations of freeze–thaw or hydration to be meaningful then they should have rock water contents based on field observations.

KEY WORDS Rock moisture content Rock properties Schist Freeze–thaw Hydration Signy Island Maritime Antarctic

INTRODUCTION

Central to studies of freeze–thaw weathering is the presence of water within the rock which, upon freezing, causes damage to that rock. As fundamental as this may be there is, nevertheless, an enormous lack of data pertaining to rock water content under natural conditions; the work of Ritchie and Davison (1968) and Trenhaile and Mercan (1984) being the only ones currently available. It is not just the 'presence' of water but the actual amount available, and its distribution within the rock, that is important. This lack of field data from geomorphological studies, and even from such as engineering investigations of building materials (as noted by Litvan, 1980), has put the interpretation of field situations and the validity of laboratory simulation studies in some doubt (Trenhaile and Mercan, 1984). The present status of studies, both in the field and the laboratory, and their implications has been dealt with in depth by McGreevy and Whalley (submitted) and so will not be repeated here. Suffice to say, the present position with respect to freeze–thaw and the role of rock moisture is one of 'fumbling in the dark' for, without actual values from rocks under field conditions, the applicability of simulation experiments and the consideration of weathering mechanisms is somewhat subjective.

As part of a long term, multidisciplinary approach to rock breakdown and the production of soil in the maritime Antarctic (Walton and Hall, submitted) a study was undertaken of mechanical weathering processes both in the field and in the laboratory. As part of that study, data on rock water content for the different local rock types under a variety of conditions were obtained as a control for laboratory experiments and as an aid to process investigation. In addition, physical properties of the same rocks were tested in the laboratory to find 'absolute' values for comparison with that found in the field.

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The main rock involved in this study on Signy Island is quartz–micaschist. This is a rock, however, for which there appears to be a dearth of information pertaining to its physical properties, even with respect to engineering aspects (Hall, in preparation). Regarding information on moisture content and its controls, Goodman (1980) cites the permeability of schist as measured in the laboratory to be 10^{-8} cm/s and in the field as 2×10^{-7} cm/s, whilst 'fissured schist' was in the range of 1×10^{-4} to 3×10^{-4} cm/s. Selby (1982) refers to the porosity of schist as being between 0·001 and 1·00 per cent with a permeability of 10^{-9} to 10^{-5} m/day. Fahey (1983) presents the greatest amount of available data (see Discussion) with respect to moisture conditions and freeze–thaw but sums up the whole situation (1983, p. 543) by saying 'Since this was a laboratory study, the experimental conditions are not particularly representative of actual periglacial environments'. The aim here is to present *both* laboratory and field data, to analyse these data and relate them to mechanical weathering processes and, particularly, to simulation studies of those processes.

STUDY AREA

The study was undertaken on Signy Island (Lat. 60°43'S, Long. 45°38'W), one of the smaller islands in the South Orkney group (Figure 1), as part of the Fellfield Programme (Walton and Hall, submitted). Signy is approximately 6·4 km north to south, 4·8 km east to west, with an area of 19·94 km^2 and rising, at Tioga Hill, to a height of 279 m. Roughly one-third of the island is covered by an ice cap and many small areas are subject to long-term snow lay. A few streams, usually along the margins of the ice cover, flow for a short period during the summer months and a number of small lakes may be ice-free during this time. The island is an area of continuous permafrost with numerous periglacial features (Chambers, 1967).

The vegetation is typical of the maritime Antarctic, comprising cushion mosses, fructicose lichens, and occasional outcrops of the grass *Deschampsia antarctica* (Smith, 1972). Geologically, the island comprises metamorphosed sediments, mainly quartz–micaschist with smaller areas of amphibolites, marbles, and quartzites (Mathews and Maling, 1967). There is a typical cold, oceanic climate with a mean monthly temperature of *c.* −4°C but the summer three months have means slightly above freezing (Watson, 1975). Rain predominates in January and February but the rest of the average 0·4 m y^{-1} precipitation is in the form of snow. Sunshine levels are low ($\bar{x} = c.\ 1\frac{1}{2}$ h day^{-1}), whilst average wind speeds are in the region of 26 km h^{-1} (Watson, 1975). For specific details of climate, particularly microclimate, see Walton (1982) and Collins *et al.* (1975). Sea ice surrounds the island during winter.

On the island there are three major reference sites: Factory Bluffs, Moraine Valley, and Jane Peak Col (Figure 1). Details of these sites are given in Walton and Hall (submitted) but in this study sample collection was not limited to the reference sites but encompassed the whole island.

APPROACH

Rock samples were collected from outcrops at the three study sites (Figure 1), from a variety of locations all over the island. The samples were also obtained under a variety of conditions, i.e. from a rivulet after a prolonged period of melt, under a new snow cover, under the same cover at a later stage and then during melt, from non-snow covered locations, and even from 19 m depth on the seabed. In short, an attempt was made to sample the range of rock types, in a variety of physiographic positions, under a mixture of climatic conditions.

Small rock samples collected in the field were immediately weighed (using a portable electronic balance) to obtain their wet weight. Larger samples were sealed in plastic bags and returned to the island's laboratory where they were subject to the point load test to assess their compressive and shear strengths at field water content (Hall, in preparation). The larger fractured rock pieces were then weighed to obtain their wet weight and, together with the small, ready-weighed field samples, then placed in an oven set at 105°C. The samples were then weighed again after 24 hours and then after 48 hours. As some rocks showed a continued, albeit very small, decrease in weight between the two weighings the 48 hour value was then taken. The difference between the wet and dry weight was then expressed as a percentage. A large number of these rocks were then retained for more detailed analysis of their physical properties.

For the retained samples the following physical attributes were ascertained, following the procedures

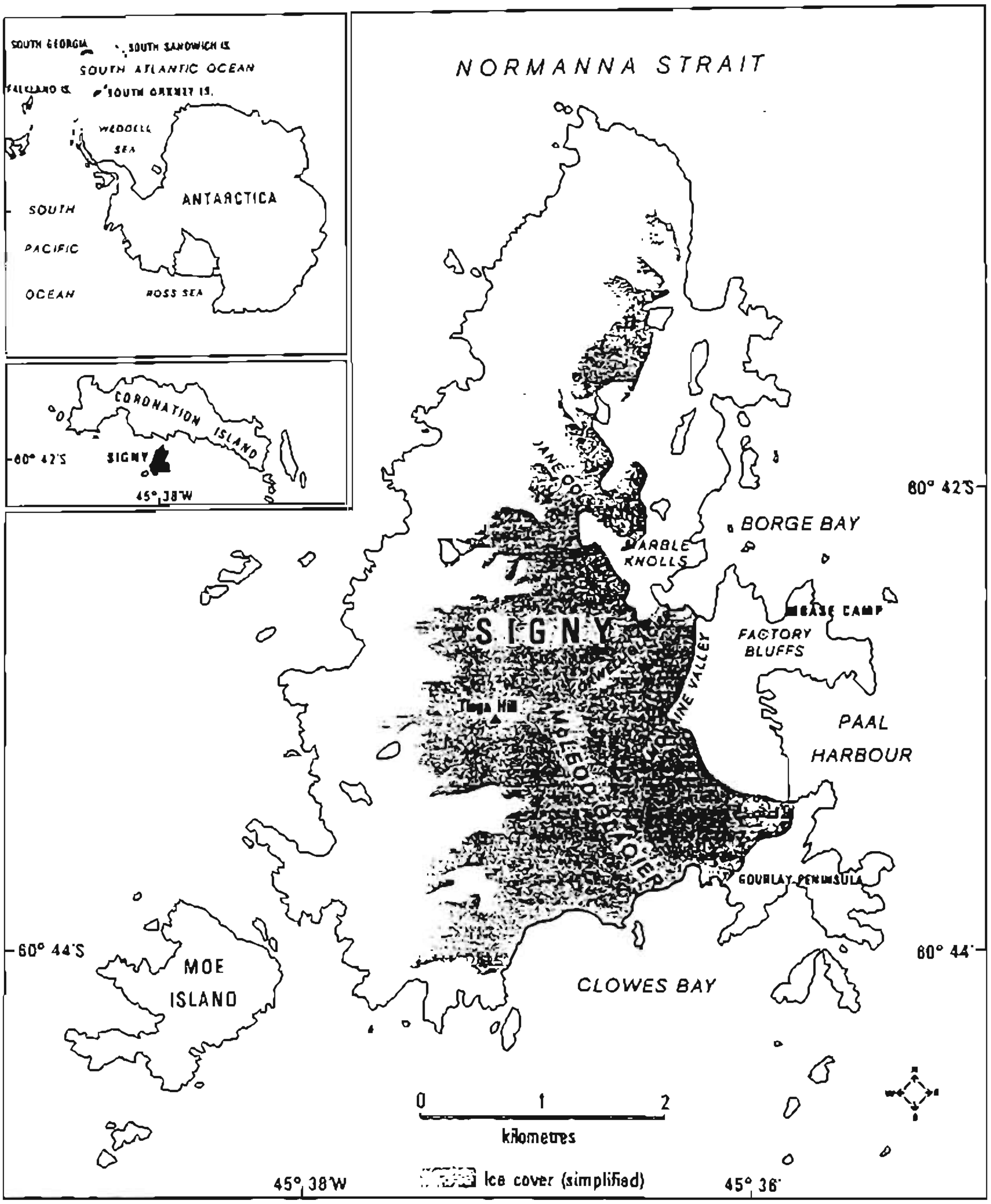

Figure 1. Signy Island and its location

suggested by Cooke (1979): porosity, water absorption capacity, saturation coefficient, microporosity, bulk density, and microporosity expressed as a per cent of total bulk volume. Where possible a note was made of minor lithologic differences within a rock type, i.e. in the quartz–micaschist whether there were pods of quartz, thick or thin bands of quartz, almost total absence of quartz, etc.

In addition, these same rock properties were obtained for 85 rock tablets cut from blocks of indigenous rock obtained from the field the preceding year. These rock tablets were returned to the field and a number will be withdrawn each year and their properties recalculated in an attempt to get some idea of rates of weathering. A comparison of the cut rock and field rock data also proved informative.

Thus it was possible to obtain information pertaining to the amount of water found in a rock under a variety of field conditions and the properties, related to water content, of that same rock. It was not possible to state the distribution of the water within the rock; however, a subjective estimation was attempted.

RESULTS

A total of 155 rocks was collected from the field and their water content ascertained. The mean water content for the total sample was 0·53 per cent, by weight ($s = 0{\cdot}40$) but this could be further divided for the quartz–micaschists (abbreviate to 'qms') as a product of small lithologic variations (Table I). The data in Table I suggest that, with respect to water content: quartzite < quartz < qms with thin quartz bands < qms with thick quartz bands < qms with pods of quartz < qms with very little quartz.

Table I. Average water content of rocks, subdivided by lithologic variations, collected in the field

Lithology	n*	$\bar{x}$ (%)	s
Quartzite	3	0·14	0·16
Quartz	11	0·30	0·20
Qms with thin bands quartz	22	0·32	0·30
Qms with thick bands quartz	7	0·46	0·34
Qms with pods of quartz	11	0·52	0·45
Qms† with very little quartz	12	1·17	0·34

* The residue of the 155 rocks sampled were either uncertain as to category they fell in or were not adequately described at time of collection to allow division.
† Qms = quartz–micaschist.

For the rocks collected in the field, data relating to the quartz–micaschist are presented in Table II, where the 13 environmental/physiographic situations from which they were obtained is also shown. The average moisture content for each of the 13 groups is shown in Table III which also indicates that the samples can be subdivided into three groups. The first group (*1 in Table III) reflects those in a wet situation for a 'prolonged' (> 24 hours) period of time, the second (*2) those in a wet but well drained environment, and the third (*3) a sunny or water deficient situation. Thus the rock moisture content can be seen to vary both within and between samples. For comparison Table VII shows the moisture content for six sets of samples obtained during spring as the snow cover begins to ablate.

In Table IV the physical properties, pertinent to moisture content, of some of the rocks noted in Table III are presented. It is noticeable that in the section regarding clasts 'in rain for *c.* 12 hours and under snow 24 hours' there is a preponderance of clasts whose field water content is *greater* than that found in the laboratory experiments. Of the 11 samples for which there are data, nine exhibit a larger moisture percentage than was obtained by immersing the same rocks in water for 24 hours. The nine have a range of moisture contents between 14·14 per cent and 51·43 per cent greater than those found in the laboratory ($\bar{x} = 29{\cdot}91$ per cent, $s = 13{\cdot}29$). Of the five samples from 'under > 24 hours snow', two exhibit a similar status. Conversely, the bulk

Table II. Moisture content obtained in the field at a variety of environmental locations

Conditions	Rock type	Moisture content (%) $\bar{x}$	s	$\bar{x}$ dry rock weight (g)	Range of rock wts (g)	n
After 24 hours snow: blown-free patterned ground site	Quartz	0·28	—	179·9	—	1
After 24 hours snow: blown-free patterned ground site	Qms	0·43	0·16	137·0	37·7 to 309·0	16
After > 24 hours snow: from under snow	Quartz	0·44	—	67·3	—	1
After > 24 hours snow: from under snow	Qms	1.0	0.38	145·3	23·5 to 345·0	10
After 24 hours snow: from faces, partly under snow	Qms	0·48	0·27	425·7	51·7 to 758·0	5
From a rock face during snowmelt	Qms	0·56	0·25	368·0	227·7 to 585·0	6
From ground surface melting out from snow	Quartz	0·12	—	164·3	—	1
From ground surface melting out from snow	Qms	0·53	0·25	167·4	28·8 to 398·0	13
From an area of wet, sorted stripes	Quartz	0·30	0·03	208·8	140·3 to 277·3	2
From an area of wet, sorted stripes	Qms	0·52	0·43	148·9	37·9 to 284·0	7
From a rock outcrop, after snowmelt	Qms	0·41	0·05	385·3	45·5 to 848·0	5
From base of cliff below an overhang	Qms	0·32	0·11	458·1	265·6 to 687·0	6
From 19 m depth on sea bed	Qms	0·97	0·54	2245·7	1534 to 2903	6
From a rivulet running for > 36 hours	Qms	1·19	0·45	236·0	155·7 to 271·2	4
From ground in rain c. 12 hours then under snow > 24 hours	Qms	1·27	0·42	156·3	82·5 to 267·1	4
From ground in rain c. 12 hours then under snow > 24 hours	Quartz	0·61	0·0	73·3	64·8 to 81·7	2
From face of weathered boulder in sun	Qms	0·29	0·25	255·6	192·1 to 298·6	3
Blocks of scree in sun	Quartz	0·15	—	66·1	—	1
Blocks of scree in sun	Qms	0·18	0·25	20·5	13·1 to 27·9	2

of the samples show a laboratory-derived moisture content greater than that found in the field; in fact 61·1 per cent show laboratory values larger than those that were found in the field. Both situations, where laboratory results are either greater or lesser than field values, have serious implications for simulation studies.

In addition to the actual moisture values, Table IV also shows the 'Saturation Coefficient' or 'S-value'. Hirschwald (1912) and Thomas (1938) have shown that a rock with an S-value of 0·8 or greater indicates that rock to be frost susceptible. Others, notably Kreuger (1923) and Tourenq (1970) have suggested the S-value should be set at 0·85 whilst MacInnes and Beaudoin (1968) suggest 0·9 (for a discussion see McGreevy and

Table III. Average water contents of 13 different environments with statistical comparisons

Environment	$\bar{x}$ water content (%)	s	
Snow-free patterned ground site	0·42	0·16	*2
After 24 hours snow from under snow	0·95	0·40	*1
After 24 hours snow, partly snow covered	0·48	0·27	*2
From rock face during snowmelt	0·56	0·25	*2
From sloping ground surface during snowmelt	0·50	0·26	*2
From area of wet, sorted stripes	0·47	0·38	*2
From rock outcrop after snowmelt	0·41	0·05	*2
From overhang at cliff base	0·32	0·11	*3
From 19 m depth on sea bed	0·97	0·54	*1
From rivulet running for > 36 hours	1·19	0·45	*1
From ground in *c.* 12 hours rain, > 24 hours snow	1·05	0·47	*1
From face of boulder in sun	0·29	0·25	*3
From blocks of scree in sun	0·17	0·18	*3

*1 Kruskall–Wallis test indicates these four to be of the same population.
*2 Kruskall–Wallis test indicates thes six to be of the same population.
*3 Kruskall–Wallis test indicates these three to be of the same population.

Whalley, submitted). For a value of 0·8 then 30·2 per cent of the rocks tested here can be considered frost susceptible, at 0.85 then only 18.6 per cent and at 0.9 a bare 7 per cent. However, this *S*-value does not take cognisance of the distribution of the water within the rock (see below), a factor of great importance. For those rocks with *S*-values $\leqslant$ 0·8 there is no apparent lithologic subdivision within the samples tested.

For the 54 rock tablets cut from quartz–micaschist that were placed in the field, there was a noticeable difference in some of the measured physical properties (Table V). Whilst porosity values remained relatively similar those of the Saturation Coefficient and the Water Absorption Coefficient were much lower than was found for the uncut, field samples. This too may have serious implications for laboratory simulations and their applicability to field situations. An attempt is being made to investigate this by means of scanning electron microscope analysis of cut sections.

Assuming the amount of water taken up by the rock under vacuum must approximate to filling all the readily available space in the rock, then it is possible to compare this with the *actual* amount of water found under field conditions to find the degree of available space unused (Table VI). This still does not take cognisance of *where* the water is in the rock, and it may well be (see below) that some has concentrated in a peripheral zone whilst in others it is disseminated. However, Table VI shows that, in response to White's (1976) question re how many rocks have > 50 per cent water content, in this instance 40·4 per cent do, with 17·0 per cent being > 90 per cent full. Conversely 59·6 per cent have less than 50 per cent moisture content with 27·7 per cent exhibiting > 70 per cent of available space unfilled.

Table VI also shows the calculated S-value for 46 of the presented samples. Of these, 13 (28·3 per cent) appear to have, when compared with actual field moisture content, an *S*-value which is too low whilst four (8·7 per cent) have an *S*-value apparently too high. The bulk of the samples (63·0 per cent) have an *S*-value correctly reflecting the percentage of unused space and there is a reasonable correlation ($r = -0{\cdot}60$) indicating this (where x = percentage unused space and y = *S*-value). Those samples for which the *S*-value is too high may show a better correlation if sampled at another time, for the data reflect solely the conditions prevailing at the time of collection. Those where the *S*-value is too low are thought to be reflecting the use, in the derivative equation for the *S*-value, of the value of rock saturated in water for 24 hours which, as has been noted, may well be less than that found in the field if water is available for a longer period.

Table IV. Properties relating to moisture content for a number of rocks from a variety of environmental locations

Location	Rock	Porosity (%)	Water absorption coefficient (%)	*S*-value	% H_2O after 24 hours	% Field H_2O
Scree	Quartz	1·36	0·81	0·60	0·31	0·15
Scree	Fine-grained qms	1·99	1·42	0·71	0·54	0·15
Under snow > 24 hours	Qms	2·95	2·90	0·90	0·98	0·91
Under snow > 24 hours	Qms	3·34	2·76	0·83	1·04	1·13
Under snow > 24 hours	Qms	1·56	1·19	0·76	0·43	0·35
Under snow > 24 hours	Qms	1·10	0·97	0·88	0·36	0·45
Under snow > 24 hours	Qms	0·60	0·41	0·69	0·16	0·10
In rain 12 hours and under snow 24 hours	Quartz	1·14	0·76	0·67	0·29	0·15
In rain 12 hours and under snow 24 hours	Fractured quartz	1·59	1·27	0·80	0·49	0·61
In rain 12 hours and under snow 24 hours	Fine banded qms	1·70	1·43	0·84	0·53	0·63
In rain 12 hours and under snow 24 hours	Small pods qms	1·90	1·42	0·75	0·54	0·72
In rain 12 hours and under snow 24 hours	Quartz	1·38	0·79	0·57	0·31	0·46
In rain 12 hours and under snow 24 hours	Qms large % SiO_2	1·13	0·68	0·60	0·25	0·37
In rain 12 hours and under snow 24 hours	Small pods, qms	2·59	1·83	0·71	0·68	1·40
In rain 12 hours and under snow 24 hours	Qms, high % SiO_2	1·87	1·58	0·85	0·58	0·21
In rain 12 hours and under snow 24 hours	Qms, low % SiO_2	2·72	2·04	0·75	0·78	1·1
In rain 12 hours and under snow 24 hours	Qms, low % SiO_2	2·56	2·20	0·86	0·85	0·99
In rain 12 hours and under snow 24 hours	Qms, low % SiO_2	2·78	2·55	0·92	0·95	1·86
From ground in sunny weather	Quartzite	2·19	1·50	0·69	0.57	0·31
Vertical cliff	Qms	2·59	1·18	0·45	0·41	0·32
Vertical cliff	Qms	1·69	1·18	0·70	0·45	0·32
From ground	Qms, small pods	2·46	1·48	0·60	0·54	0·13
From ground	Fine banded qms	1·22	0·96	0·79	0·36	0·27
From ground	Qms, large pods	1·27	0·85	0·67	0·32	0·11
From ground	Qms, small % SiO_2	0·55	0·41	0·75	0·15	0·15
From ground	Qms, big pods	1·51	1·06	0·70	0·39	0·19

Table IV. (*Contd.*)

Location	Rock	Porosity (%)	Water absorption coefficient (%)	S-value	% H_2O after 24 hours	% Field H_2O
From ground	Qms, small pods	2·28	1·99	0·88	0·72	0·39
From ground	Qms, high % SiO_2	1·27	1·09	0·86	0·4	0·31
From ground	Qms, thick bands	1·22	0·81	0·67	0·29	0·07
From ground	Qms, thin bands	1·02	0·76	0·75	0·29	0·19
From ground	Qms, large pods	1·37	1·14	0·83	0·42	0·15
From ground	Qms, small bands	0·74	0·44	0·60	0·17	0·23
From ground	Qms, folded	2·43	1·97	0·81	0·71	0·13
From ground	Qms	1·98	1·32	0·67	0·48	0·68
Rock face	Qms	4·65	3·45	0·74	0·39	0·17
Rock face	Qms	1·34	1·06	0·79	0·39	0·17

Table V. Properties of the cut rock tablets

Rock	Porosity (%)	Water absorption coefficient (%)	Saturation coefficient
Qms	1·84	0·53	0·27
Qms	1·07	0·58	0·31
Qms	2·29	0·92	0·41

Qms = quartz–micaschist.

DISCUSSION

The most significant finding is that of the existence of rocks, in a subaerial environment, with > 50 per cent moisture content (Table VI) within a climate where they are subject to freezing. Thus, in response to the question of White (1976, p. 5) as to '. . . will bedrock ever become water-saturated from melting snow or rain and then undergo rapid freezing to crack the rock', it can be said that in the maritime Antarctic this may be the case. Of the samples obtained, 40 per cent indicated moisture contents in excess of 50 per cent (with 17 per cent of them above 90 per cent saturation) near to an area shown by Walton (1982, Table II) to be subject to freeze–thaw during the sampling period. Recent detailed measurements at one of the reference sites have shown that short-term freeze–thaw cycles are a typical feature of summer conditions (Walton, personal communication); thus frost shattering must be considered a potentially active mechanism. It may even be that, in the light of the test procedures and the rocks themselves, an even greater percentage of the rocks were prone to frost damage.

If the data indicate < 50 per cent saturation then this is with respect to the whole rock and does not consider the actual distribution of that moisture within the rock. If the moisture is evenly distributed within the sample then that value is meaningful but what if there is a concentration in the peripheral zone? Laboratory testing of samples indicated that those from the field had higher S-values and water absorption capacities than did cut blocks of the same material. Considering the schistose nature of the rock it could be envisaged that weathering prises the laminae apart at the margins and that this accounts for the higher values found for the field samples. This in turn would tend to suggest that moisture would be more readily available at the periphery and that breakdown would take place here and gradually work inwards. Unfortunately, no means of accurately assessing water distribution within the rock, in the field, is known to the author and so the specifics of water location within the rock remain unknown.

Table VI. The degree of saturation, percentage of unused space and the S-value for rocks collected in the field

Mx % H_2O	Field % H_2O	% of space unused	S-value	Saturation 50%	90%	*If field > lab with 24 hours sat.
0·44	0·15	65·91	0·71			N
0·61	0·61	0·33	0·80		✓	Y
0·63	0·63	0·0	0·84		✓	Y
0·72	0·72	0·42	0·75		✓	Y
0·54	0·46	15·37	0·57	✓		Y
0·41	0·37	10·00	0·60		✓	Y
0·69	0·21	69·28	0·85			N
0·04	0·03	1·25	0·75		✓	Y
1·0	0·99	0·80	0·86		✓	Y
0·52	0·15	71·15	0·67			N
0·76	0·15	80·66	0·60			N
1·0	0·91	9·0	0·90		✓	N
1·27	1·13	11·02	0·83	✓		Y
0·57	0·35	38·60	0·76	✓		N
0·23	0·10	56·52	0·69			N
0·46	0·26	43·48	—	✓		N
0·89	0·32	64·04	0·45			N
0·65	0·32	50·77	0·70			N
0·52	0·17	67·31	0·67			N
0·26	0·12	53·85	0·43			N
0·91	0·13	85·71	0·60			N
0·45	0·27	40·00	0·79	✓		N
0·47	0·11	76·6	0·67			N
0·20	0·15	25·0	0·75	✓		N
0·57	0·19	66·67	0·70			N
0·83	0·39	53·01	0·88			N
0·47	0·31	34·04	0·86	✓		N
0·44	0·07	84·09	0·67			N
0·38	0·19	50·0	0·75			N
0·50	0·15	70·0	0·83			N
0·28	0·23	17·86	0·60	✓		Y
0·57	0·16	71·93	0·67			N
0·88	0·13	85·23	0·81			N
0·67	0·10	85·07	0·78			N
0·73	0·68	6·85	0·67		✓	Y
1·58	0·54	65·82	0·74			N
0·50	0·17	66·0	0·79			N
0·86	0·52	39·53	0·91	✓		N
1·37	0·34	75·18	0·67			N
0·97	0·28	71·13	0·82			N
0·43	0·12	72·09	0·43			N
1·00	0·50	50·0	0·70			N
1·01	0·60	40·59	0·72	✓		N
2·45	1·36	44·49	0·67	✓		N
0·58	0·32	52·84	0·73			N
1·36	0·79	57·53	0·57			N
0·43	0·13	69·77	0·50			N

* If field water content is greater than the laboratory derived value after immersion in water for 24 hours.

The results and interpretations noted above are pertinent to weathering simulations, particularly where cut blocks are used. Cutting may remove that part of the rock within which breakdown was particularly active and so the study would show an initial slow rate of destruction which would then accelerate as the laminae are prised apart. Fahey (1983, p. 541) found that in his simulations the pre-cut schist bars showed great resistance to both frost action and hydration; a further indicator that cut blocks may not be good reflectors of process rates in nature. However, perhaps more important is the moisture made available to the specimens during the study, the size of the samples and the temperatures they are subject to. Ideally, the temperatures should mirror those found at the study area and accelerated cycling may be unrealistic and introduce exaggerated effects (McGreevy and Whalley, submitted). Rock samples should, preferably, be large so as to include such properties as bedding planes, fissures, etc.; the 'massive rock' rather than 'intact rock' (McGreevy and Whalley, 1982). The size of the sample is not only pertinent for the inclusion of geologic considerations but also for thermal effects and moisture conditions.

This unrealistic approach to rock moisture conditions in simulations has been extensively discussed by McGreevy and Whalley (submitted) and the data noted here from the field reflect these inadequacies. For an experiment to be meaningful, then the rock moisture conditions should reflect those found in the field, for otherwise extrapolation to the field may be unfounded. To date, no experiments have clearly reflected field conditions and, in most instances, moisture status has probably been greatly exaggerated (McGreevy and Whalley, submitted; Trenhaile and Mercan, 1984). This study has shown there to be variations in moisture content within a rock type (here, quartz–micaschist), as a product of small lithologic differences, and between sites, as a result of environmental conditions (Table III). Certainly the variations in environment within any study area require consideration for some localities (as in Tables II and III) will be wetter whilst others receive more radiation or wind, which will remove moisture; Smith (1977) indicated such variations to be important with respect to insolation weathering in the Sahara.

When comparing field moisture content with that obtained from immersing the sample in water for 24 hours, then 52 per cent of the rocks were found to have an excessive moisture content and so simulation would likely be unnatural; yet the freezing of samples saturated for 24 hours is a common approach (e.g. Wiman, 1963; Fahey, 1983). It is very possible that rocks in many positions (e.g. high on a vertical cliff face) may, despite the most conducive of rock properties, never attain high moisture contents. Conversely, other samples indicated water contents between 14 and 53 per cent *greater* than that found after immersion for 24 hours and in these instances simulations might show diminished results. Moisture content is clearly a major factor in mechanical weathering processes and so values cannot just be assumed. For field interpretation of processes and meaningful simulations then actual field moisture contents must be measured in a variety of locations.

The moisture content found for samples during the spring (Table VII) are, in a number of cases, higher than was found during the summer. It may well be that many of the rocks are at their wettest during this ablation period. If that is so, then, with the temperature cycling that is known to occur during this period (Walton, 1982),

Table VII. Moisture content for quartz–micaschist during spring meltout

Location	Date	n	$\bar{x}$ water content (%)	s
Factory Bluffs: rock outcrop	18/9/84	10	0·6	0·15
Factory Bluffs: soil surface	18/9/84	10	1·0	0·36
Moraine Valley: rock outcrop	20/9/84	10	0·7	0·48
Moraine Valley: soil surface	20/9/84	10	0·95	0·27
Moraine Valley: rocks uncovered most of winter but frequently covered in rime	25/9/84	10	0·7	0·16
Moraine Valley: rocks uncovered most of winter but subject to meltout	20/9/84	10	0·75	0·29

Rock samples varied in dry weight between 282 and 789 g.

this may be the time of active freeze–thaw weathering. However, at the least, it does indicate the need to consider moisture content on a temporal as well as spatial basis.

The calculated *S*-value (Tables IV and VI) appears to offer a reasonable correlation with the degree of rock saturation found in the field and hence is a potentially good indicator of frost susceptibility. *S*-values obtained for weathered schists in an earlier study (Hall, 1974) in northern Norway indicated a range (0·66–0·98) reasonably comparable with that found for this investigation. The marginally high values from Norway possibly reflect more extensive weathering, due to greater precipitation and higher temperatures, than is found in the maritime Antarctic.

Fahey (1983, Table I) in his consideration of hydration and freeze–thaw weathering of schist obtained laboratory-derived rock property data similar to those found for the cut blocks in this study. Data for schist porosity presented by Goodman (1980) (0·5 to 1·9 per cent) and that by Wolff (1981) (0·62 to 3·12 per cent) are comparable with both the data of Fahey (1983) and this study. However, all of these are 'laboratory' values and do not reflect what might actually be the moisture content in the field. For instance, Fahey (1983, p. 538) assessed '. . . whether the schist was capable of becoming critically saturated . . .' but did not determine whether it actually ever did. Yet he concluded (p. 544) that 'frost action associated with the volumetric expansion of pore water upon freezing was three to four times more effective than hydration . . .' despite the simulations possibly making available far more water than the rocks would normally be likely to receive and '. . . then undergo rapid freezing . . .' (White, 1976, p. 5). However, Fahey (1983, p. 543) does point out that 'Since this was a laboratory study, the experimental conditions are not particularly representative of actual periglacial environments.'; but surely this then begs the question as to the applicability of the findings back to the real world? It is this applicability of experiments to the real world which is the main problem. Wright and Gregory (1955) showed that accelerated freezing and thawing produced more rapid results and that when specimens were frozen immersed in water then damage occurred more rapidly than when the specimens were frozen in air. Unless the study environment is one of rapid freeze–thaw cycles with rocks immersed, or partially immersed, then the experiments show little or no relevance to the study site. This has been the problem with many freeze–thaw experiments (e.g. Wiman, 1963). Trenhaile and Mercan (1984, p. 329), for consideration of frost action in coastal regions, show that '. . . experiments on saturated rock samples provide a poor representation of field conditions . . .'.

CONCLUSIONS

Moisture content is clearly a major factor in rock weathering studies and, as such, there is a great need for detailed field studies to make laboratory simulations and interpretation of landforms more meaningful. To date, almost all geomorphological studies have not been based on field data and therefore are of unknown value. In many instances, it would appear that moisture values may have been exaggerated and consequently weathering rates and mechanisms, particularly if accelerated freezing cycles were also used, are not a reflection of reality. In other cases, moisture conditions are possibly underrated and so, again, simulations do not mirror that which are taking place in the field. The monitoring of temperature fluctuations for simulation and process interpretation without knowledge of moisture content, its variation through space and time, and, preferably, its actual distribution within the rock makes any resulting hypothesis of doubtful value.

From this study it would appear that, of the laboratory derived parameters, the *S*-value is the best indicator of potential frost action. If it is not possible to obtain measurement of rock moisture content in the field, then use should be made of the *S*-value when considering how much water should be made available to the rock for simulation studies. However, this is by no means a substitute for actual measurement of field moisture content. The most important aspect of rock moisture still to be resolved is that of the distribution within the rock. It is very likely that as more field evidence of moisture content and its variations becomes available then our consideration of processes and their rates, and landform development may have to be re-evaluated.

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DAILY MONITORING OF A ROCK TABLET AT A MARITIME ANTARCTIC SITE: MOISTURE AND WEATHERING RESULTS

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Abstract. The mass of a tablet of indigenous rock was monitored daily for one year in order to study changes in moisture content and timing of weathering losses. The broad climatic conditions to which the tablet was subjected were noted. It was found that, within the maritime Antarctic, freezing temperatures occur when rocks have high moisture contents. Although rock tablets may be considered 'unnatural', they can provide valuable information on daily variability of moisture status which is essential for the planning and interpretation of realistic weathering simulations.

Introduction

Rock tablets, despite being somewhat 'unnatural', have been used in a variety of geomorphological studies such as the investigation of erosion rates (e.g. Trudgill, 1975; Crowther, 1983), the study of rock weathering in soils (e.g. Day and others, 1980), for the measurement of rock temperatures by means of internally located thermistors (Whalley and others, 1984) and for the monitoring of rock moisture variation on a wave-cut platform (Trenhaile and Mercan, 1984). Weathering studies, particularly of building materials, have frequently utilized exposure trials of tablets of rock or man-made products (see McGreevy and Whalley, 1984, for a review). The classic study of Ritchie and Davison (1968) utilized bricks of various materials to monitor moisture changes as a function of aspect and, with thermocouples fitted within them, to record the incidence of freezing and thawing. Other cold-environment studies of this type include those of McMahon and Amberg (1947) and Ritchie (1972) who studied the effects of frost action on bricks, and Cook (1952) who investigated the effects of freeze–thaw upon concrete in a marine environment. In a similar manner, Trenhaile and Mercan (1984) considered the moisture content of rock on a wave-cut platform as a control upon frost damage, whilst Chatterji and Christensen (1979) evalulated the frost damage to limestone nodules as a function of different moisture contents. In the Antarctic, Miotke (1982) has obtained temperatures around a small piece of rock on the floor of one of the dry valleys (Taylor Valley). However, probably due to logistical constraints, no long-term study of daily changes to tablets of indigenous rock has been undertaken in the Antarctic.

As part of the British Antarctic Survey Fellfield Ecology Research Programme a study is being undertaken of the mechanical weathering processes operative in a maritime Antarctic environment (Hall, 1986*a*, *b*, *c*, 1987*a*, *b*; Hall and others, 1986), and involves concurrent field and laboratory investigations. As an extension of this programme, a project was initiated to investigate, albeit in a very simple manner, the daily changes in the mass of a tablet of local rock and to record the general climatic conditions to which it was subjected. It was hoped that such data, for the period of a whole year, would provide preliminary information on daily moisture variability and on weathering rates.

Methodology

Rock samples collected on Signy Island (60° 43′ S, 45° 38′ W; see Hall, 1986*a*, for details of study area) were cut into small tablets and, following the procedure of

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Cooke (1979), the porosity, microporosity, water absorption capacity and saturation coefficient of each ascertained. The tablets were then returned to the field in the austral summer 1983–4 with the aim of retrieving some of their number each year and re-testing them to see, as a measure of weathering, if any of the properties had changed. In January 1985, one block was moved to outside the scientific station to allow its daily mass to be recorded. Unfortunately it was not possible to use micrometeorological logging equipment to detail the specific conditions to which the tablet was subject. Thus, only broad, generalized statements concerning temperature, snow or rain, wind, dryness, sun, or whether the tablet was covered (by snow or ice) were possible. Nevertheless, this information is comparable with that which is usually available for the bulk of field sites and is certainly equal to what is used for the general description of large areas.

Only one tablet was used, but the results are considered to be typical for quartz-micashist, which comprises the bulk of the island. The tablet was kept in an open environment with its schistosity parallel to the ground surface, replicating the attitude of most exposures.

Results and Discussion

Despite the simplicity of approach, the data offer, for the first time from this environment, daily information for a single piece of rock over the span of a whole year. Details of the properties of the tablet left in the field at experiment start (1983–4) are given in Table I, whilst graphs of daily change of mass together with the broad

Table I. Properties of the rock tablet

Property	*Value*
a-axis (cm)	11.02
b-axis (cm)	3.30
c-axis (cm)	2.38
Cailleux flatness index	300.80
Oblate–prolate index	18.20
Maximum projection sphericity	0.54
Porosity (%)	1.46
Water absorption capacity (%)	0.31
Saturation coefficient	0.21
Microporosity (%)	16.82
Water absorbed in 24 hrs (g)	0.23
Water absorbed under vacuum (g)	1.07

climatic conditions to which the tablet was subjected are given in Fig. 1. Occurrences of 'major' (> 0.20 g) changes in mass, either on a daily or a cumulative (i.e. several successive days) basis, are presented in Table II. The value of 0.2 g was chosen upon the recognition that a rise of this magnitude constitutes a significant (> 87%) saturation of the rock in terms of potential for damage if that rock is then subject to freezing (White, 1976). A decrease in mass by this amount (0.2 g) would imply a relatively 'dry' state such that it unlikely to be damaged by freezing conditions.

The significance of the 0.2-g increments of moisture in abetting frost action may seem contentious in the light of the recent comments by Walder and Hallet (1986). However, it is argued that whilst the model of Walder and Hallet (1985) is a significant step forward in our understanding of rock damage due to freezing water,

Table II. Details of major weight gain or loss experienced by the rock tablet

Gain/loss	*Weight (g)*	*Dates(s)*
Maximum weight gain in 24 h	0.35	9 April 1985
Gain of ⩾ 0.20 g in 24 h	0.22	7 February 1985
Gain of ⩾ 0.20 g in 24 h	0.21	10 March 1985
Gain of ⩾ 0.20 g in 24 h	0.21	12 March 1985
Gain of ⩾ 0.20 g in 24 h	0.21	25 March 1985
Cumulative gains of > 0.20 g	0.20	25–28 February 1986
Maximum wt loss in 24 h	0.23	24 February 1985
Loss of ⩾ 0.20 g in 24 h	0.22	8 February 1985
Loss of ⩾ 0.20 g in 24 h	0.20	11 March 1985
Cumulative loss of > 0.20 g	0.22	6–8 April 1985
Cumulative loss of > 0.20 g	0.22	27–30 April 1985
Cumulative loss of > 0.20 g	0.20	16–19 October 1985
Cumulative loss of > 0.20 g	0.29	2–5 January 1986
Cumulative loss of > 0.20 g	0.25*	14–21 January 1986
Cumulative loss of > 0.20 g	0.20*	3–4 July 1985

* Weight loss includes material weathered free.

it is by no means universally applicable. Within this experimental programme, ultrasonic evidence has been obtained at freezing (Hall, 1987*b*) which clearly indicates a singular, massive water-to-ice phase change of > 80% of the water available to freeze within the rock. Under these conditions, it is those rocks with the greater moisture content which, despite some extrusion of ice, will suffer the greatest damage due to the 9% volume change which occurs as the water turns to ice. In addition, there is the possibility that rocks subjected to omnidirectional freezing may be damaged by hydrofracture (Walder and Hallet, 1985; Hall, 1986*c*) and, once again, it would be those rocks with the higher moisture contents that should sustain greatest damage. This is not to negate the detrimental effects of other weathering processes (e.g. wetting and drying) that are also operative, but rather to point out the significance of moisture content with respect to freeze–thaw. However, at the same time, the data presented here must be viewed in the context that it is simply a daily change of mass that is recorded. No account of any changes of pore volume, due to frost damage, during this period are quantified although this is something which could be expected to increase with time (Walder and Hallet, 1986).

Consideration of Fig. 1 indicates a number of broad trends with respect to changes in tablet mass. First, it can be seen that there are relatively frequent, large, short-term changes from spring through to autumn, whilst the winter period (late May to early September) has much more subdued daily variations. Secondly, it is evident that there is a gradual overall diminishing of tablet mass through the year, after an initial rise in April. Thirdly, it is apparent that the loss of tablet mass is progressive, rather than a stepwise diminution resulting from incidents of large particle loss.

That the greatest daily changes in mass took place during the spring-to-autumn period is not particularly surprising as it is during this time that some precipitation falls as rain and positive temperatures predominate (Fig. 1). Much of the winter period (*c*. 80%) has below zero temperatures, with values as low as −27 °C being recorded (21 June), and limited sunshine with very limited potential for local melting. There is thus very little unfrozen water available to penetrate the rock. During much of the winter period the tablet was covered by snow or ice, thereby further precluding water availability. During the spring-to-autumn phase there is a significant occurrence of rain (15% of the days) at times when the rock is not covered by snow or ice and

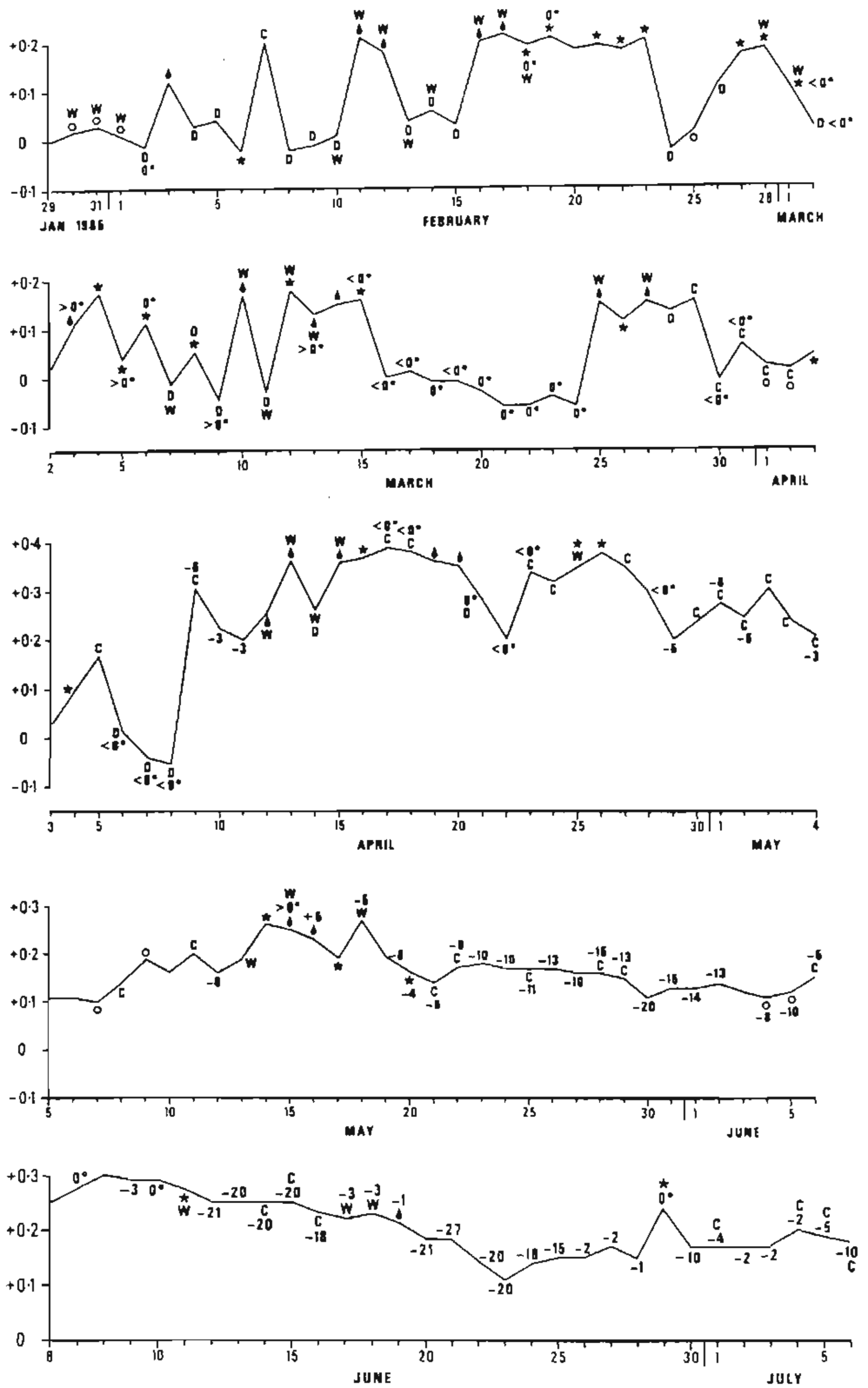

Fig. 1. (Caption on p. 22.)

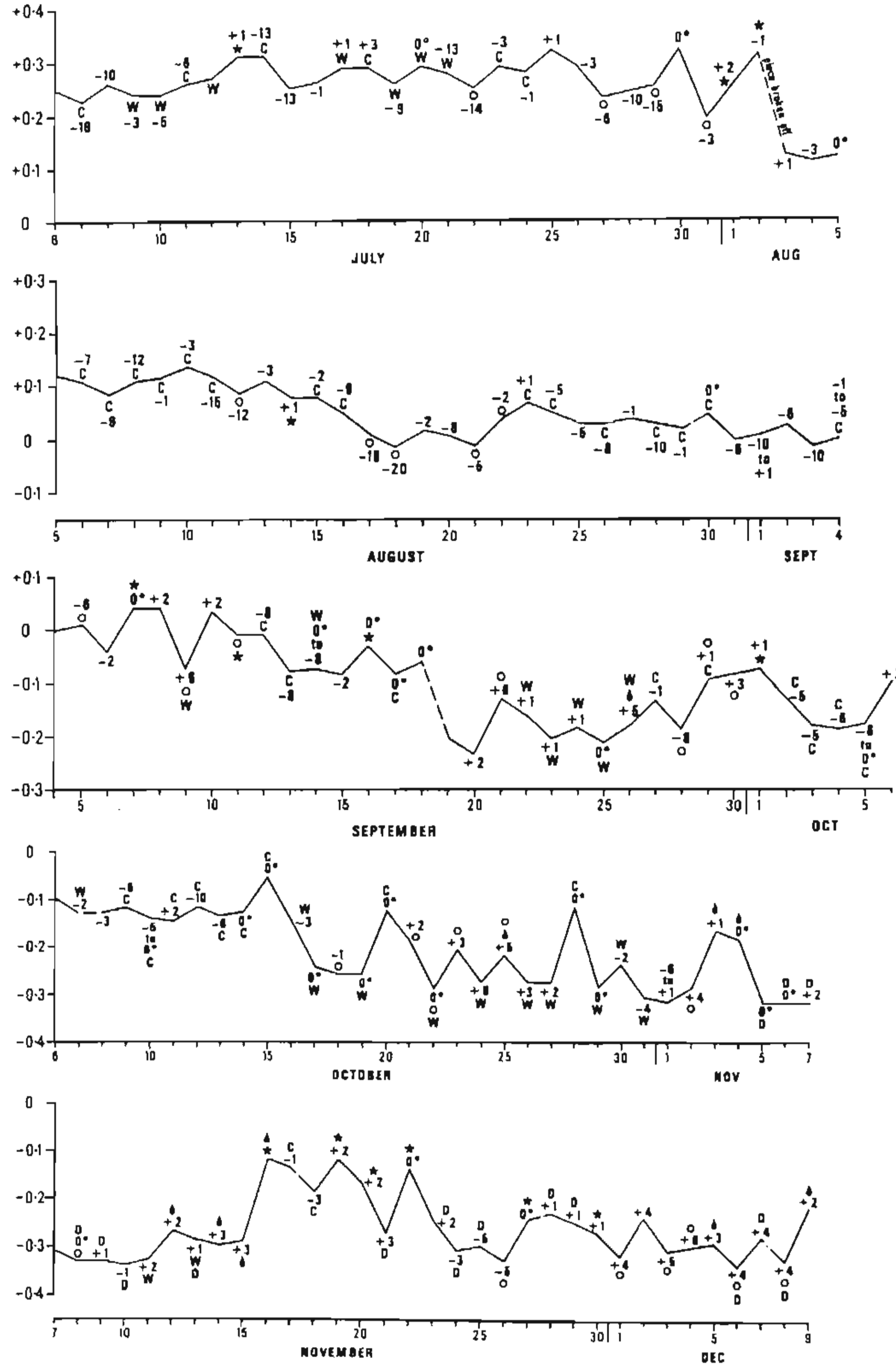
JULY
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Fig. 1. (cont.)

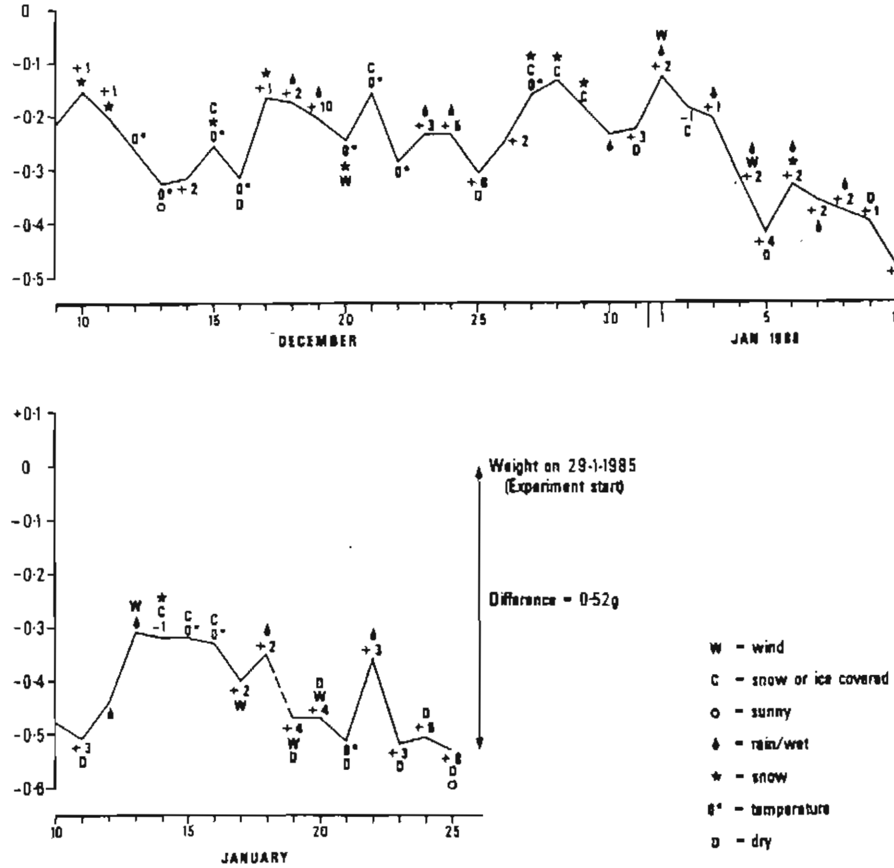

Fig. 1. Graphs of the daily change of mass (g) of the rock tablet together with the broad climatic conditions to which it was subjected.

thus able to take up moisture and so show a rapid gain of mass (e.g. 2–3 November). When the tablet is exposed, dry and windy days cause rapid moisture loss, particularly if there are positive air temperatures (e.g. 21–23 January 1986).

Although the tablet shows frequent significant mass gains during the spring-to-autumn period (Table II), the number and magnitude of freeze events are lower than in winter. Of the four recorded losses of 'large' particles of material, two took place in winter (2 August and 18 September) and two in the following summer (7 October and 18 January). Of the two summer events, one (7 October) took place during continued negative temperatures, but was of a very small effect (0.02 g), whilst the other occurred during a thaw phase.

The progressive diminishing of tablet mass is interesting insofar as, if there were an increase in internal voids due to frost action, as suggested by Walder and Hallet (1986), then some gain in mass might have been anticipated due to the greater water-holding capacity of the sample. This increase of mass can only be attained though if the following three constraints can be met: (1) internal frost damage *does* occur, (2) water is made available to the rock, and (3) the rock is able to take up the water. However, in practical terms, monitoring of mass gain due to the above effects may be masked by loss of rock material resulting from weathering, and so some increase of water-holding capacity may have taken place. Consideration of Fig. 1 suggests that the only time there was an unequivocal mass gain was on 9 April. From 8 to 9 April there was a gain of 0.35 g, an increase of 0.12 g above what the rock was initially found to absorb under non-vacuum conditions in 24 hrs (Table I). After that sudden jump, there was a gradual decrease in mass, with various gains and losses dependent upon climatic conditions, until the first recognized loss of rock material on 2 August (0.19 g). A downward trend in mass then continued, through 4 September, when the original start mass was attained once more, to the end of the study period.

Patently frost and/or salt action are not the only possible weathering mechanisms to which the tablet was subject. Such processes as thermal fatigue are possible during the summer period when relatively high daytime temperatures alternate with cooler night-time conditions, particularly when shading may take place to cause rapid cooling. Wetting and drying is an inherent part of freeze–thaw insofar as the moisture status of the rock varies. There is some evidence (Hall, in press) to show that this absorption and desorption causes changes in the elasticity of the rock and may, via hysteresis, cause fatigue to the rock. However, neither of these processes are well understood and, apart from some laboratory information on the latter (Hall, in press), no data are available. In addition to these mechanical weathering processes it is recognized that some degree of chemical weathering is possible, particularly during the summer months, and that biotic processes may also occur. Although the time span of only one year is relatively short such that, in an evironment of this kind, mechanical processes might be thought to predominate, ancillary studies of chemical and biotic weathering currently in progress may later throw some light on their relative contributions.

With respect to the process of freeze–thaw, it is very likely that under omnidirectional freezing the water is not able to move to the freezing front as required by Hallet (1983). Conversely, the schistose nature of the rock may, in some instances, facilitate hydrofracture (Powers, 1945; Walder and Hallet, 1985). If crack expansion is insufficient to accommodate the volume increase as water changes to ice then there will be forcible expulsion of water ahead of the freezing front which may cause damage to the rock – hydrofracture. If ice can only grow along the laminate mineral interfaces, then the tensile forces will oppose each other and so crack expansion due to ice growth may not be possible (see Hall, 1986*c*, fig. 3). This situation would be conducive to hydrofracture. At the same time, it may well be that due to the relatively small size of the tablet, freeze penetration may be so rapid that a massive phase change occurs thereby precluding hydrofracture. If this were the case then it is probable that due to a moisture gradient the bulk of the water is in the outer part of the tablet and so ice extrusion may well occur (Davidson and Nye, 1985) thereby limiting the amount of damage to the rock.

McGreevy and Whalley (1985, p. 344), in their highly pertinent discussion on rock moisture content, noted that in the study of Ritchie and Davison (1968) high degrees of sample saturation rarely coincided with freezing conditions. In this respect the maritime Antarctic may differ, for Fig. 1 shows that the mass of the tablet, and hence its moisture content, was relatively high on a number of occasions when freeze–thaw took place. For example, from the end of April through to early August the mass is high and negative temperatures, down to −27 °C, are seen to occur. All in all the tablet appears to have been subjected to 39 freeze–thaw cycles during the course of the year, but it must be kept in mind that it had been in the field one year prior to experiment start and therefore subject to weathering that cannot be evaluated here.

A possible complicating factor is that the frequent oscillations in moisture content that took place during parts of the year (e.g. early March) would appear to be conducive to weathering by wetting and drying. Further, the limited data available on interstitial water solute chemistry (Hall and others, 1986) suggests an NaCl content of between *c.* 0.3 and 0.6 M. This, then, adds the further possibility, as has been suggested by Williams and Robinson (1981), McGreevy (1982) and Fahey (1985), of the interoperation of frost and salt weathering. Thus, although the data from the tablet do provide new information, much more needs to be done before the relative contributions and interactions of the various weathering processes are known.

Conclusions

Overall, this experiment attempted to provide data in answer to the questions and criticisms posed by McGreevy and Whalley (1985) with respect to the moisture status of rocks in the field. Although a very simplistic approach and lacking in climatic detail, the information presented here does, nevertheless, give a record of daily variations in the mass of a tablet of rock resulting from gain and loss of moisture together with material loss due to weathering. As a cut block, open on all sides, the results cannot be construed as relating to cliff-face situations but they do approximate to the loose, fallen blocks found over much of the island. In fact, the tablet probably provides a base-line value insofar as the irregular surface of the naturally occurring blocks probably hold, and lose, more moisture.

The data indicate the possibility, in the maritime Antarctic environment, of rocks having relatively high moisture contents at a time when they may experience freezing. There are seen to be daily changes in moisture content, particularly during the spring-to-autumn period, and this wetting and drying may promote rock fatigue. Salts are present in the rock and weathering due to salt activity may also take place. The interaction of at least these three weathering processes (freeze–thaw, wetting and drying, salt) appear to produce progressive weathering of the tablet. However, it may well be that, due to the 'cut' rather than 'natural' nature of the tablet, the experimental period was not long enough to discern true weathering rates.

It is suggested, based upon the evidence obtained in this pilot study, that long-term field measurement of representative samples could be of great use to the understanding of weathering processes and rates. More particularly, it can help to provide a firm data base from which to construct laboratory simulations. The results presented here relate only to the weathering of *small* blocks in the maritime Antarctic. A complementary study of large blocks, with more detailed monitoring of environmental factors, would prove an ideal extension of this current experiment.

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Rock Moisture Data from the Juneau Icefield (Alaska) and Its Significance for Mechanical Weathering Studies

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ABSTRACT

Information was obtained regarding rock moisture content from nunataks along a west to east transect across the Juneau Icefield in Alaska. The data, together with rock temperatures and general climatic information, were collected for east-, west-, north- and south-facing samples. For one day data were collected every hour. The resulting information indicates the spatial and temporal variability that can exist over both short distances and short time-spans. It is suggested that this variability can have important repercussions with respect to weathering processes. The number of wetting and drying cycles monitored greatly depends upon the level of sample saturation that is considered significant. Available information suggests that wetting and drying can be operating at more than one level within the rock and that this will result in different weathering products. Wetting and drying may be more important than has been previously thought, both in terms of a weathering agent in its own right and as one that interacts with other processes, thereby speeding up their effect upon the rock.

RESUME

Des informations ont été recueillies concernant le contenu en eau des nunataks selon un transect ouest-est au travers de l'Icefield Juneau en Alaska. Les données concernant l'humidité ont été notées en même temps que des observations concernant les températures des roches et les composantes climatiques, et cela pour des échantillons recoltés en des lieux exposés à l'est, à l'ouest, au nord et au sud. L'information rassemblée indique une grande variabilité spatiale et temporelle des observations aussi bien sur des distances courtes que sur des périods de temps réduites. Il est suggéré que cette variabilité peut avoir des répercussions importantes en ce qui concerne les processus d'altération. Le nombre de cycles de séchage et d'humidification observé dépend pour beaucoup du degré de saturation des echantillons qui est, de ce fait, un facteur considéré comme significatif. L'information disponible suggère que l'humidification et le séchage peuvent se produire à plus d'un niveau dans la roche et que cela détermine des productions différentes de débris. Ces processus d'humidification et de séchage peuvent être plus importants que ce qui a été supposé précédemment, aussi bien comme agent d'altération simple que comme processus se combinant avec d'autres.

KEY WORDS: Rock moisture Weathering Wetting and drying Alaska

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INTRODUCTION

Although chemical and biological weathering have been recognized as operative on the Juneau Icefield (Dixon *et al.*, 1984; Hall and Otte, 1990), the bulk of studies have emphasized the role of mechanical processes, with gelifraction, in particular, being stressed (Hamelin, 1964; Shenker, 1979; Klipfel, 1981; Linder, 1981). Central to any consideration of these weathering processes are data on rock moisture content and temperature (McGreevy and Whalley, 1982, 1985). Although in recent years there has been an increasing attempt to monitor actual rock temperatures (e.g. Francou, 1988), field-based data regarding rock moisture content, and its variability in both time and space, have received little attention.

While freeze-thaw is relatively rare on the Juneau Icefield during the summer months and no data are available regarding the presence and character of salts, it would appear that climatic conditions, particularly in the west, are conducive to weathering by wetting and drying. Periods of hot, bright weather alternate with times of low cloud, high humidity and heavy precipitation that produce a number of wet-dry cycles. The process of wetting and drying is, however, poorly understood (Ollier, 1984) and little is known regarding its mode of operation (Hall, 1988a). In some rocks water uptake causes rock expansion, while drying results in contraction (Nepper-Christensen, 1965), but as the rocks may not return to their original length, this may ultimately cause breakdown (Venter, 1981; Hames *et al.*, 1987). In addition, the bonding strength of the component minerals can also be diminished such that with many wetting and drying cycles there can be a decrease in rock strength that may ultimately lead to failure (Pissart and Lautridou, 1984; Hall, 1988a). Wetting and drying also operates synergistically with other weathering processes (e.g. within freeze-thaw: Hall, 1991), but its role and contribution are, as yet, unknown, although Hall and Otte (1990) showed that it played a major role in enhancing biological weathering.

Fundamental to any consideration of weathering due to wetting and drying, as with other mechanical weathering processes, is the acquisition of data pertaining to actual rock moisture content and its temporal and spatial variability. However, essential as such data may be, very little are actually available. Subsequent to the pioneering work of Ritchie and Davison (1968), who monitored moisture changes in masonry materials exposed to each of the cardinal points, the only known studies are those of Trenhaile and Mercan (1984) and Hall (1986). In an attempt to add to this meagre supply of information, rock moisture data are presented for a summer period on the Juneau Icefield in Alaska and an assessment is made of its significance with respect to weathering.

STUDY AREA

The Juneau Icefield (Figure 1) is a relict of the great Cordilleran ice sheet and covers an area of approximately 4000 km^2 along the Alaska-Canada Boundary Coast Range (Marston, 1983). The Icefield is situated within a maritime environment along the southern and western edges of the Coast Range but becomes more continental with distance inland towards the east. While no specifics regarding the climate have been published, details regarding the general climatic conditions as well as the mountains and glaciers of this region can be found in Miller (1964). The present studies were undertaken at four sites along a west to east transect across the Icefield (Figure 1). C17 (Camp 17) is situated at the western margin of the icefield on a ridge above a small cirque glacier. C10 is on a nunatak that rises to *c.* 426 m above the surrounding ice, while C18 is located on a nunatak just to the south of the Alaska-Canada border, in the region of the Gilkey Glacier at an altitutde of *c.*1700 m a.s.l. C26 is close to the western extremity of the icefield and is located on a nunatak approximately 30 m above the ice.

METHODOLOGY

Twelve visually comparable specimens of granodiorite were collected at C17, and, of these, three pieces were set out facing each of the cardinal points. The rocks were marked so that at each study site they could be placed facing the same aspect as at the experiment start. They were positioned in such a way that they were open through a 180° arc centred on their cardinal orientation (i.e. an east-facing stone would be open in an arc from north through east to south). These stones were then each weighed three (at 0800, 1300 and 1800 hours) or four times each day (as above and 2200 hours) by means of a portable electronic balance accurate to 0.1 g. The rock samples having been dried and weighed and then saturated and again weighed, it was possible to convert the daily rock

Figure 1 Simplified map of the Juneau Icefield (shaded area) to show the location of the four study sites.

mass to a measure of percentage saturation. The rock samples (Table 1) varied in thickness between c.2 cm and 5 cm, and so, being set on the ground surface, equated to the outer shell of the bedrock.

Air temperature in a Stevenson Screen was recorded together with the unscreened ambient temperature at the rock surface, taken by means of a meteorological thermometer at an open (360° exposure) site. For a short period rock surface temperatures were also measured at each of the four aspects. Wind direction and speed, cloud cover and whether it was sunny, overcast or raining were also recorded. For one day, measurements were taken hourly of temperature and rock mass for each of the rock samples.

RESULTS AND DISCUSSION

The daily record for the period 11 July through to 18 August is shown in Figure 2. From 11 July until 18 July the record is for Camp 17 at the western (maritime) extremity of the Icefield. From 18 July data acquisition is at Camp 10, where it remains until 28 July, when it moves eastward to Camp 18 until 6 August. From 10 August until 18 August the record is from Camp 26 at the eastern (continental) extremity of the Icefield. Data shown include wind speed and direction, rain or sun, cloud cover, air and rock temperature (for a 360° exposed site), and the average percentage saturation for the three blocks exposed at each of the four aspects.

From Figure 2 it can be seen that daytime rock temperatures are higher than air temperatures, reflecting the effect of incoming solar radiation in warming the rock. Conversely, night-time radiative cooling sometimes produces rock temperatures lower than those of the air. On days with clear skies, low wind speeds and large radiation receipts the rock temperatures were substantially higher than those of the air (cf. 12 July, when a 117% difference was recorded), indicating the inadequacy of using air temperatures as a surrogate for rock conditions. Unfortunately, rock temperature data from this study are still not adequate for any detailed discussion regarding their spatial and temporal variability and their effect upon moisture changes and weathering in general, as the time between readings is simply so large that it can hide many variations that could have taken place.

Consideration of rock moisture conditions (Figure 2) indicates that levels on the continental side of the Icefield (C26) were very low and correspond to the sunny conditions and high rock temperatures

Table 1 Sizes and shapes of the samples used in the experiments.

	Axes (cm)						Shapes			
Sample	a	b	c	d^1	D^2	$\bar{x}$size	F^3	R^4	OP	S^5
N1	10.6	5.6	4.9	0.1	2.7	7.03	165.3	0.04	8.16	0.74
N2	15.2	6.6	2.4	0.5	3.1	8.07	454.2	0.16	10.9	0.39
N3	10.2	7.0	2.6	0.1	3.0	6.6	330.8	0.03	−3.1	0.46
S1	10.9	6.3	4.6	0.3	3.2	7.27	187.0	0.09	5.45	0.68
S2	12.0	5.5	3.3	0.5	2.4	6.93	265.2	0.21	8.99	0.55
S3	10.0	6.4	4.1	0.3	2.9	6.83	200.0	0.1	2.69	0.64
E1	9.5	5.4	2.5	0.1	2.6	5.8	298.0	0.04	3.26	0.5
E2	9.3	7.4	3.2	0.2	3.2	6.63	260.9	0.06	−5.5	0.53
E3	12.1	5.0	4.0	0.2	2.4	7.03	213.8	0.08	11.4	0.64
W1	10.4	4.0	2.5	0.3	1.9	5.63	288.0	0.16	12.9	0.53
W2	8.1	6.0	2.2	0.6	3.0	5.43	320.5	0.2	−5.3	0.46
W3	9.2	8.4	2.3	0.1	4.0	6.63	382.6	0.03	−15.4	0.41

Axis Measurements:
[1]Diameter of the sharpest corner of the a/b plane.
[2]Diameter of the largest inscribed circle of the a/b plane.
Shape Indices:
[3]Cailleux's Flatness Index.
[4]Modified Wentworth Roundness.
[5]Maximum Projection Sphericity.
OP = oblate/prolate index.

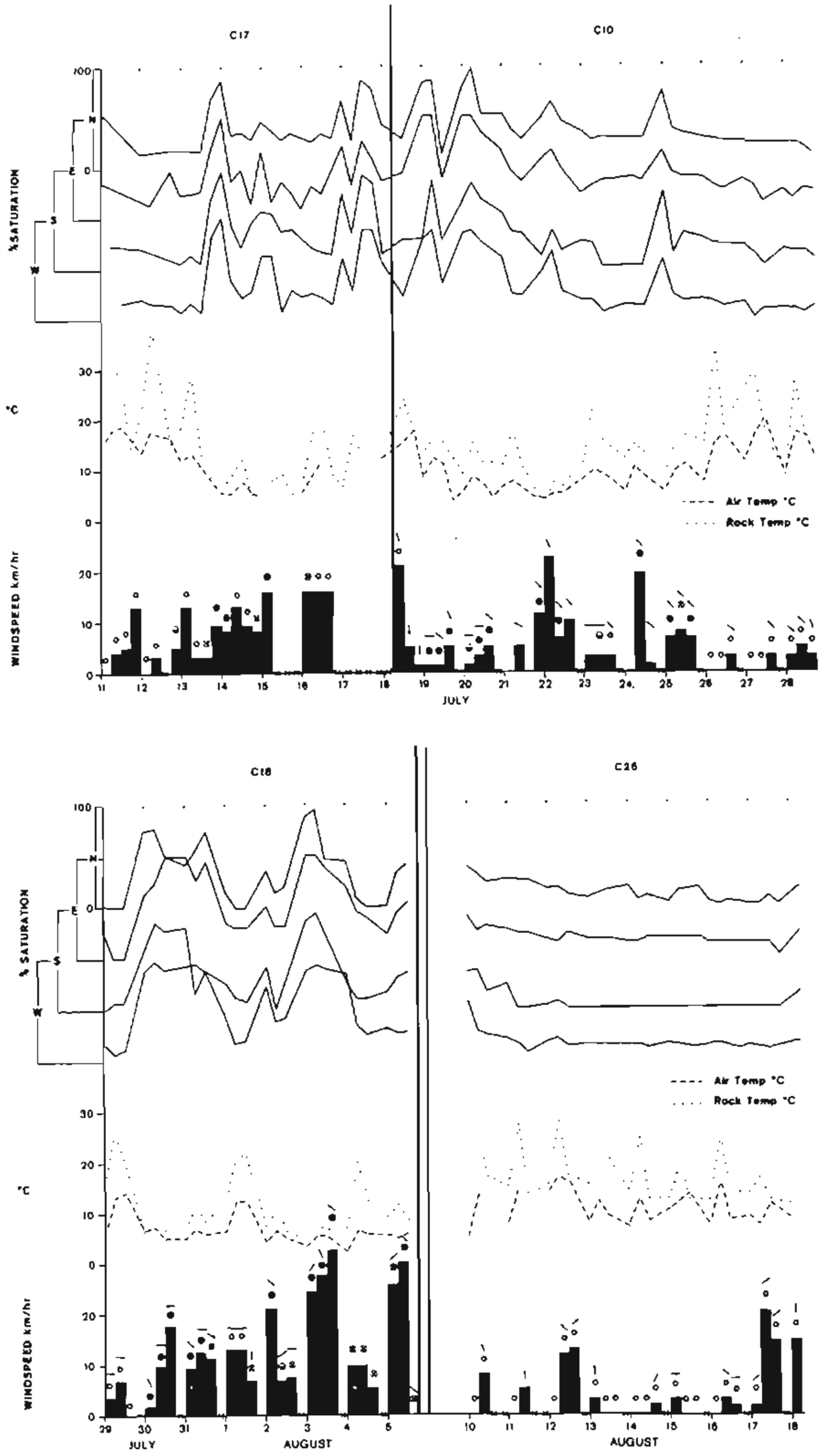

Figure 2 Graphs depicting the variation in moisture content for the four aspects together with rock surface and air temperatures plus a generalized picture of climatic conditions (vertical bars indicate wind speed; arrows indicate wind direction; × is no data available; ● = precipitation; sun symbol = sunshine; and circled × = overcast).

experienced there. At the three other sites there were periods of wet weather (e.g. 13 and 14 July) which produced high degrees of saturation. However, while the basic trends in moisture content for the four aspects are often similar, important variations do occur. For example, on 31 July there was rain and a westerly wind that resulted in those samples on the western side having a high ($\leq 95\%$) moisture content. The samples to the south, however, showed an initial *drying* phase followed by a small peak indicative of wetting, but the moisture content was still less than 50%. The north-facing samples showed a wetting response greater than those of the east or south, owing to the frequency of prevailing northwesterly winds. On the other hand, a northwesterly wind on a day with sun (27 July) resulted in a low, flat response from the north and west aspect samples, while those for the south and east showed small peaks indicative of slight moisture increases.

The responses are not always so obvious or simple. For instance, on 3 August there was a strong southeasterly wind accompanied by intermittent precipitation, and yet the north, south and eastern aspects show a *decrease* in moisture content due to the drying effect of the wind predominating over the wetting by intermittent rain. At the same time the leeward, western aspect maintained a high level of rock moisture content due to the absence of wind to promote drying. For times of high radiation inputs, with consequent high rock temperatures (e.g. 12 August), it is the southerly aspect that exhibits the lowest rock moisture content. Overall, the rock response with respect to its moisture content will be a funtion of moisture type (i.e. rain, mist, snow, etc.), wind direction and speed, the time of day and the radiation input (here shown indirectly by rock surface temperature). Short-term changes in any of these factors (e.g. fluctuations in wind direction) can rapidly alter the moisture status of the rock. In spite of the greater insight into rock moisture content that these data offer, a clearer understanding of temporal and spatial variability requires much more detailed rock and climatic data than are available here. However, for one day (12 August) measurements were taken every hour (Figure 3) and these start to show some of the complexity and variability that can occur.

Figure 3 clearly shows that the rock surface temperatures follow the progression of the sun through the day. The east-facing rocks heat up first, followed sequentially by those of southerly, westerly and northerly aspects. The eastern and southern aspects experience the highest temperatures, while the north has the lowest. The loss of heating due to direct radiation is shown by the sharp drop in temperature as exhibited by the south-facing rocks after 1700 hours. After initially calm conditions, a southeasterly wind probably caused the slight drop in the temperatures of the eastern aspect, while a shift to the southwest a little later is reflected by the slight rise prior to afternoon cooling. As the temperatures show a marked variability, so too does the rock moisture content despite this being a day *without* any form of precipitation. For example, the eastern samples show an *increase* in moisture content from 16% to 24% at 0900 hours as the sun warms the rock, followed by a decrease to 14% by 1200 hours, a plateau and then an increase from 1500 hours to 26% at 1600 hours, after which it remained constant. The south-facing samples start in an almost dry state but then exhibit a sudden rise to 17% by 0900 hours, a return to almost dry by 1000 hours and then back to 7% by 1100 hours. This is followed by a gradual rise to 10% at 1300 hours and then back to an almost dry state at 1600 hours, where it remains until 2000 hours, after which it suddenly rises to 12%. The south-facing samples appear to slightly mirror the changes in the eastern samples but with a different magnitude of change and the timing generally several hours later. The west-facing samples show a gradual *decrease* through to 1100 hours followed by a slight rise to 1300 hours and a subsequent decrease. Finally, the north-facing samples show an initial drop from 25% to 17%, followed by a rise and a fall that parallel the changes in the eastern samples (and are 1 h later than those of the south).

There are three main attributes apparent from Figure 3.

First, the north-, east- and west-facing samples all have moisture contents substantially higher than that of the south ($N > W > E \gg S$).

Second, moisture contents can *rise* subsequent to initial heating of the rock. This phenomenon, not yet clearly understood, has also been found by Meiklejohn (personal communication, 1991) from work on sandstones heated by early morning sun during winter in the Drakensberg Mountains of southern Africa. It may be that the heating of the cool rock sets up a water vapour transfer due to the temperature gradient in the rock. Moist air moves into the rock, where it cools and the moisture condenses. Later, as the rock continues to heat up, this temperature gradient disappears and the moisture is driven off.

Third, there are significant temporal and spatial variations in moisture content that could have

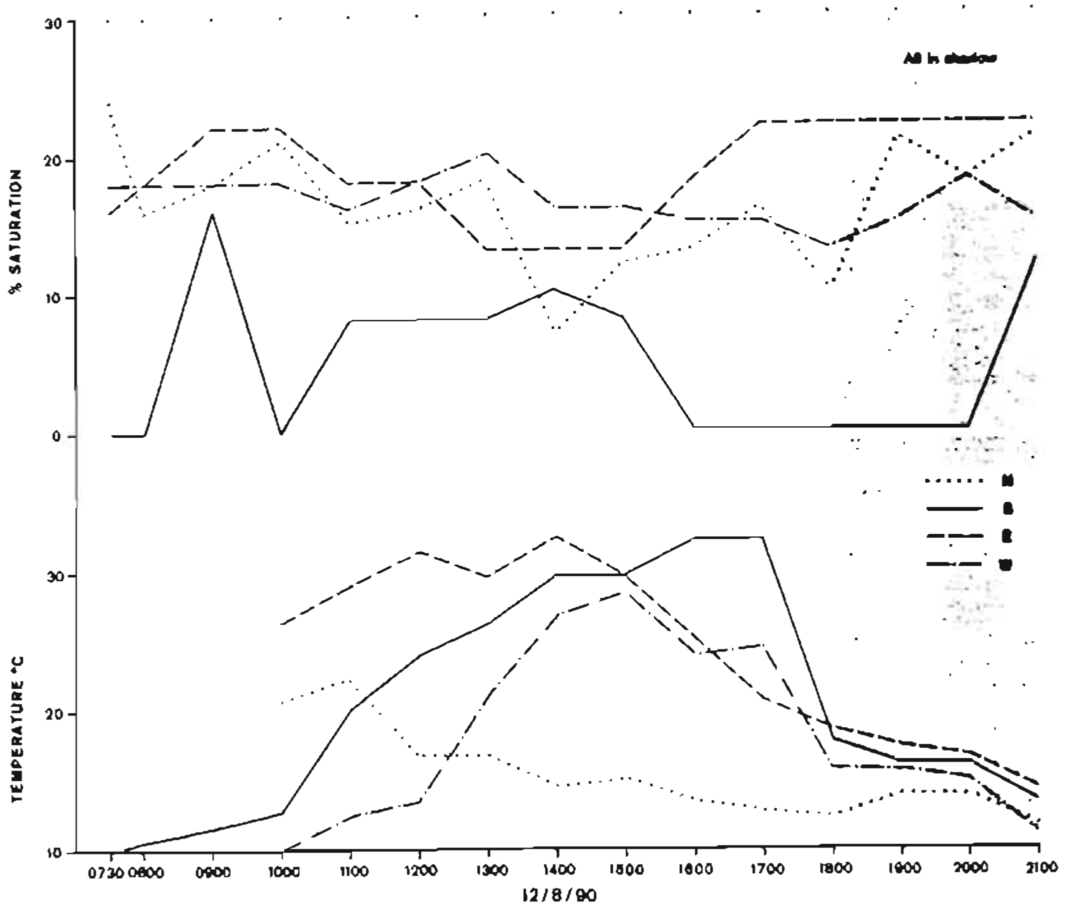

Figure 3 Readings of rock moisture content and rock temperature for each of the four aspects taken at hourly intervals on 12 August.

significant repercussions with respect to weathering.

In terms of weathering it was noticeable that during the record period, even recognizing that it was during a summer, no freeze-thaw events were recorded at any sites. Had any taken place then, as would be the case later in the year when they do occur, rocks of different aspects would have been affected in different ways as a result of the combination of varying moisture regimes and temperature conditions. Thus, it is suggested that great care should be taken in appropriating the freeze-thaw process to any location in the absence of *detailed* rock temperature and moisture information. The small-scale variability in both these factors can be so great (see Hall, 1992, for more details) that it may be found that rocks of differing aspects may or may not freeze in the presence of freezing air temperatures. Equally, the recording of subzero rock surface temperatures is no indicator that the freezing conditions were of any significance, as there may be little or no water available to freeze. Small-scale, aspect-controlled, variability in rock moisture could result in one aspect suffering freeze-thaw weathering, while another a short distance away (< 2 m) does not. Conversely, it could be that the wetter rocks do *not* suffer temperatures conducive to the freezing of available moisture, while the drier rocks of another aspect do. Finally, the amount of water present, and the magnitude, duration and rate of freeze combine to determine the nature of the freeze-thaw mechanism (see Hall, 1991). Without such data, it is impossible to understand the manner in which the rock is being broken and how this varies temporally for any one given location.

Although freeze–thaw did not take place during the study period, wetting and drying events certainly occurred at the more maritime locations. Again, the information from this study, despite being more detailed than in other weathering studies, is still not sufficient to give an accurate picture of the number of wetting and drying cycles, as it was possible for changes to take place within the time between readings. However, from the data that are available it would appear that the actual number of events experienced by any aspect is constrained by the threshold of moisture content that is adopted. Consideration of Table 2 indicates that, on the basis of information presented in Figure 2, the number of events for any given aspect can vary significantly, dependent upon what level of saturation is chosen. However, a lowering of the threshold does not always imply a greater number of cycles (cf. E for C17), as it may be that the moisture level will remain *above* the lower limit but cycle back and forth across a higher level. The problem is really one of the meaning of the threshold adopted. In other words, is the more readily accepted higher level actually of any greater significance? It must be recognized that the moisture levels cited here *refer to the whole block*, although, in reality, water is concentrated within the outer shell of the rock rather than being disseminated throughout the block. Thus, the only real distinction between a 10% saturated rock and a 75% saturated rock is the *depth* to which the rock is wetted. This, then, implies that cycling through, say, a 10% or 25% threshold relates to the outermost margin of the rock and may result in granular disintegration or very thin flaking of that rock. On the other hand, cycling through a 50% or 75% (or higher) level may actually be causing the wetting and drying effect to be taking place *below the surface layer* and result in the production of a thicker weathered skin. In reality, both will be operating at most sites during any one year but it may help to explain, as was found on the nunataks of the Juneau Icefield, how flakes produced from the bedrock (i.e. due to higher moisture levels) subsequently broke down to material of sand grain size (due to the lower moisture levels).

Table 2 Wetting and drying cycles for different degrees of saturation monitored at the four study sites.

Site	10%	25%	50%	75%
C17				
N	0	2	3	2
E	1	7	4	2
S	2	3	4	3
W	3	5	4	2
$\bar{x}$	1.5	4.25	3.75	2.25
C10				
N	1	2	4	2
E	0	4	4	2
S	2	5	3	2
W	2	3	4	0
$\bar{x}$	1.25	3.5	3.75	1.5
C18				
N	2	5	3	1
E	1	2	3	2
S	1	4	2	2
W	0	2	3	3
$\bar{x}$	1.0	3.25	2.75	2.0
C26				
N	4	0	0	0
E	1	0	0	0
S	2	0	0	0
W	3	0	0	0
$\bar{x}$	2.5	0.0	0.0	0.0

Thus, consideration of moisture level variations within rock are much more complex than has been suggested previously. Not only are the variations spatially and temporally controlled such that significant variability can occur over short distances and small time-scales, but also the effects upon weathering processes might be more important than has been hitherto thought. In addition to being a controlling factor upon the nature and effect of freeze–thaw weathering, the degree of saturation and *its* variability may control the nature and extent of weathering due to wetting and drying. Wetting and drying can also work synergistically with freeze–thaw, the former being operative during wetter, milder conditions and accentuating the effects of freeze–thaw, which operates during colder periods, owing to the resulting fatigue. Wetting and drying also exerts a control upon, and must interoperate with, biological weathering. Hall and Otte (1990) found that chasmoendolithic algae were a major cause of flaking in granitic rocks on some of the nunataks of the Juneau Icefield. Expansion and contraction of the mucilage of algae living parallel to the rock surface but at a depth of several millimetres was as a direct result of wetting and drying. The resulting flakes were found to break down to sand-sized grains and it may be that low-amplitude wetting and drying cycles aided in this process.

From available studies it would appear that weathering as a direct result of wetting and drying is likely to be slow (Pissart and Lautridou, 1984; Hames *et al.*, 1987; Hall, 1988b). However, the number of cycles that might occur in any one year may well be very large indeed at some locations and so, through fatigue, could still be effective in causing breakdown, both by their own direct action and also by weakening the rock, thereby lessening the extent of activity required for other processes. What is required, apart from more detailed field data on rock moisture content itself, is information on how wetting and drying actually operates, what the effect of different moisture levels are, and how wetting and drying interoperates with other weathering mechanisms.

CONCLUSIONS

The data presented here, albeit from only a short summer period, give some insight into the complexity and variability of rock moisture content that can occur in cold regions. Significant differences are seen to occur between rocks of different aspects, even though the samples were physically only a short distance apart, and this may have important ramifications with respect to the resulting weathering. The more detailed data for a single day show that this variability can be very large and thus indicate the need for a shorter sampling period if a true picture of wetting and drying is to be obtained. With respect to wetting and drying, it is apparent that the saturation level adopted will greatly affect the perceived number of wetting and drying cycles thought to affect the rock. However, it may well be that rather than a single threshold a range of different levels should be examined in order to evaluate the zones within which the respective cycles are effective, as this may have some bearing on the nature of the resulting breakdown.

Ultimately it will be necessary to obtain information on rock temperatures, radiation input, wind speed and direction, together with precipitation type and amount, in order to explain the temporal and spatial variations in rock moisture content that are measured. However, this level of information is needed to fully investigate not just wetting and drying but also the role of *all* mechanical weathering processes, as well as those of chemical and biological weathering. With respect to wetting and drying itself, there is clearly still a need for extensive laboratory investigation into the manner in which this process operates such that the field data can then be properly evaluated.

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Rock Moisture Data from Livingston Island (Maritime Antarctic) and Implications for Weathering Processes

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ABSTRACT

Rock moisture content was determined for rock samples on different aspects of rock outcrops on Livingston Island during a summer season. As a result of the dominant rain-bearing northerly winds the southern aspect usually has rock moisture levels lower than the northern. The southern aspect, however, experiences high rock moisture levels during periods of snowmelt; snow accumulates on the southern, lee-side of the rock outcrops. Wetting and drying events are more frequent on the northern exposure, although not as common as at a site open through the full 360°, while the southern aspect tends to experience continuous, low moisture levels with infrequent dry events. Contrary to earlier suggestions, freeze–thaw weathering does not appear to be a major factor during the summer. Although rock moisture levels are conducive to freeze–thaw, rock temperatures rarely go below 0°C. Rather, it appears that weathering due to wetting and drying may be more common on the northern aspects than was previously thought while chemical weathering is active on southerly aspects. Rock moisture levels may support rock damage due to segregation ice during the winter freeze when the rate of freezing is slowed by the overlying snow cover.

RÉSUMÉ

Le contenu en eau de roches a été mesuré sur des échantillons divers prélevés pendant l'été sur des affleurements de l'Ile Livingston. Comme les vents pluvieux dominants viennent du nord, les roches exposées au sud ont habituellement des teneurs en eau plus basses que celles exposées au nord. Sur les versants exposés au sud, des teneurs en humidité sont cependant élevées pendant les périodes de fonte de neige car la neige s'accumule sur les versants sud qui se trouvent sous le vent. Les phénomènes de séchage et d'humidification sont plus fréquents sur les affleurements exposés au nord, tout en n'étant pas aussi nombreux que sur les sites exposés à tous les vents (360°). Sur les affleurements exposés au sud, les niveaux d'humidité restent peu élevés et les assèchements sont rares.

Contrairement à ce qui a été suggéré précédemment, l'altération par gel-dégel ne semble pas être, l'été, un facteur de désagrégation principal. Quoique les niveaux d'humidité soient favorables aux actions de gel-dégel, les températures des roches y descendent rarement sous 0°C. Au contraire, il apparaît que l'altération due aux alternances sèchage / humidification doit être plus fréquente sur les versants exposés au nord, tandis que l'altération chimique serait surtout active sur les versants exposés au sud. Les niveaux d'humidité des roches peuvent engendrer une désagrégation due à la formation de glace de ségrégation pendant l'hiver parce que la vitesse de gel est ralentie par la couverture de neige.

KEY WORDS: Maritime Antarctic Rock moisture Cryogenic weathering

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INTRODUCTION

Quantitative data regarding weathering processes in the Antarctic are relatively rare. Where they do exist, they focus primarily on the continent and the dry valleys of the McMurdo region in particular (e.g. Campbell and Claridge. 1988). The dry valleys are certainly very unusual from the point of view of weathering owing to their extreme aridity and cold. For this reason, they are doubly important in that they are perceived as an analogue for the hyper-arid inner planets of our solar system (Vishniac and Mainzer, 1973; Hall, 1989). However, the dry valley environment does not constitute the 'norm' for the Antarctic. Even within the various continental oases significant differences in environmental conditions occur: note the comments of Pickard (1986) with respect to the summer wetness of the Vestfold Hills, a condition rarely experienced in the McMurdo region. However, above and beyond these differences on the continent there is the distinction between the continental and the maritime Antarctic. Although recognized as a biologically active zone, and of geological significance, little attention has been paid to the distinctly different weathering regime experienced on the islands as compared with the continent proper.

Rock moisture is difficult to measure in the field (Thorn, 1988). As a consequence, field determinations of weathering processes and laboratory simulations have both been put in question (McGreevy and Whalley, 1985; Thorn, 1992). However, following the pioneering work of Ritchie and Davison (1968) some attempts have been made to monitor rock moisture content (e.g. Trenhaile and Mercan, 1984; Hall, 1986, 1991a; Humlum, 1992). Nevertheless, empirical data are still extremely limited and most deductions as to the nature and rates of weathering in cold regions remain subjective and unverified (Hall, 1986, 1991b; Thorn, 1992). Following a pilot study undertaken in Alaska (Hall, 1991a) an attempt was made to obtain information on the spatial and temporal variability of rock moisture from a maritime Antarctic site. Although significant in itself, the rock moisture data are of even greater value when combined with information pertaining to rock properties (Hall, 1993a) and rock temperatures (Hall, 1993b), as obtained here.

STUDY AREA

The Byers Peninsula, located at the western extremity of Livingston Island (62°40'S, 61°00'W), constitutes the largest ice-free area (50 km^2) in

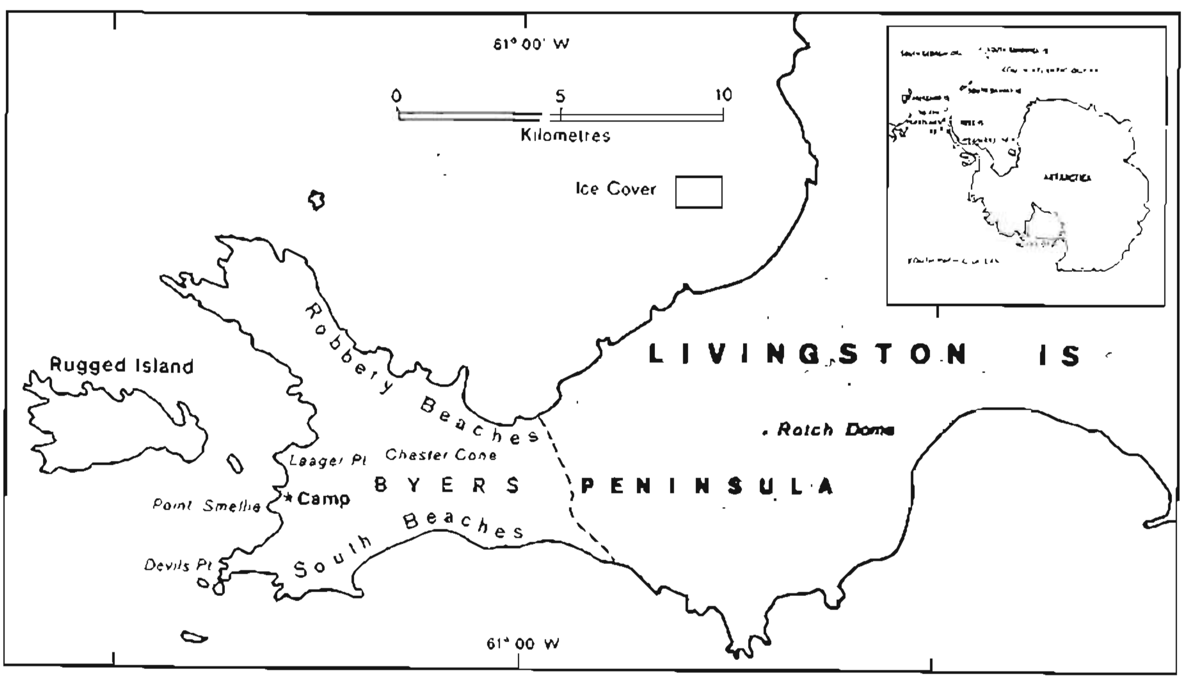

Figure 1 Location of study area.

the South Shetland Islands and is one of the largest in lesser Antarctica (Figure 1). According to John and Sugden (1971) the mean annual temperature is −3°C and the annual precipitation is of the order of 100 to 150 cm water equivalent. The study area consists of an extensive assemblage of raised beach platforms at a number of levels, with the highest (54 m a.s.l.) dated at 9700 years BP, interspersed with upstanding volcanic plugs that frequently exhibit columnar jointing and a number of dykes and sills. The geology, mainly volcanics interbedded with conglomerates and sandstones, is well described by Hobbs (1968) and Smellie *et al.* (1980). During most years the mean daily temperature is above freezing from December to March and permafrost is believed to be present below an active layer of 0.3 to 0.7 m thickness (Thom, 1978). Snow accumulation is particularly pronounced on the lee side of obstacles, this normally being the south side as a result of dominant northerly winds. Extensive cloud cover in this region limits daily radiation receipt and this is particularly pronounced on the southern aspects of outcrops. Thus, snow accumulations in the lee of obstacles survive well into January and the next season's snowfall begins soon after the snow of the preceding winter has finally ablated.

METHODOLOGY

Fundamental to the study of weathering processes, be they mechanical, chemical or biological, is knowledge regarding rock moisture content and its variability both spatially and temporally. An initial study of rock moisture content on Signy Island (Hall, 1986) was unstructured and was thus unable to show variation of moisture with respect to both time and space, although the subsequent daily monitoring of a rock tablet for one year did provide some useful data for an open site (Hall, 1988). In the present study three sites were utilized to monitor rock moisture content. The first, at the site of the field camp, was on a raised beach at *c.* 12 m a.s.l. and approximately 400 m from the sea. The second site was a dolerite dyke, with an east to west trend, a further 200 m inland at *c.* 25 m a.s.l. The third site was to the south of the first two and was on a small volcanic rock boss at *c.* 30 m a.s.l. and 500 m from the coast. At site 1 three blocks of rock from each of site 2 and site 3 were placed on the horizontal ground surface, open through the whole 360°, next to an automatic weather station. At site 2 three pieces of dyke rock were placed on each of the north and south faces of the dyke at a height of 1 m above the ground surface. At site 3, three pieces of local rock were positioned 1 m above the ground surface on each of the north, east, south and west faces of the rock boss. All of these rocks were weighed several times each day. Each rock sample had been previously oven-dried for 24 hours at 105°C and then weighed to get its dry weight. Each had also been saturated by immersion in water for 24 hours and weighed again to get its saturated weight. Thus it was possible to transform each daily weighing to a percentage saturation for that rock. By this method variations in rock moisture content for sites of differing exposure were obtained for a period of 45 days during January and February.

At site 1 an automatic weather station was used to log incoming radiation, wind direction and speed, together with air, ground surface and 5 cm depth temperatures throughout the field season. In addition, a Squirrel logger was also used at site 1 to monitor the upper surface and north-facing surface of a piece of dolerite positioned on the ground surface. At site 2 thermocouples were affixed to the rock surface and, via drill holes, at 1 cm depth on both the north and south faces; temperatures were read at the same time as the rocks were weighed. At site 3 thermocouples were affixed to the rock surface on each of the cardinal faces and also read when the rocks there were weighed. This instrumentation therefore provided standard climatological data (from the automatic weather station at site 1) together with actual temperatures experienced by the rock at each of the rock moisture monitoring sites (Hall, 1993b).

A measure of weathering was obtained from both sites 2 and 3 as well as from a variety of locations on the peninsula by means of a Schmidt hammer, an indentor, by measuring weathering rind thicknesses, and from the collection of weathered-free material from the rock faces (Hall, 1993a). In addition, analysis of interstitial rock water chemistry is being undertaken to gain some insight into the nature and concentration of salts present in the rock and their variability as a function of aspect together with their possible influence on both mechanical and chemical weathering processes. A number of general observations regarding periglacial processes were obtained as these give information both in their own right and indirectly with respect to the weathering regime of the island (Hall, 1992).

RESULTS AND DISCUSSION

For the monitoring of rock moisture content three samples were used at each site. This was partly as a safeguard against sample loss or damage (i.e. in such an instance there would not be total loss of data as would occur had only one sample been used) and partly as a means of establishing sample variability. However, it also provided an important insight into the effect of sample mass upon considered percentage saturation and, via sample size, indirect information pertaining to the depth of bedrock wetting. As an example, the dry mass of samples S1 to S3, from the south side of the dyke, were 211.7 g, 256.1 g and 366.6 g respectively. Consideration of Figure 2 indicates that if only a sample of mass S1 had been used then this would have indicated a generally high level of saturation (>80%) while if S3 had been used then a lower level would have been obtained (<70%). While the values for S2 were notably lower than S1 and marginally higher than S3, all three samples followed the same trends (Figure 3) with the *degree* of saturation varying as a function of rock mass. In other words, all three samples reflect the same changes in rock moisture content but differ in the degree of saturation.

With similar surface areas the thicknesses of the three samples S1 to S3 (50 mm, 55 mm and 65 mm respectively) could be considered to crudely equate to a comparable depth of bedrock. Thus saturation of S1 but *not* S2 implies saturation to a depth of approximately 50 mm whilst saturation of sample S3 would imply saturation of bedrock to a depth of at least 65 mm. For instance, on Julian day 35, S1 was saturated and so was S2 but *not* S3, thereby indicating sufficient moisture was available to saturate the rock to a depth of about 55 mm. Unfortunately, for the other test sites in this undertaking either the rock sample thicknesses were very similar or samples became damaged and/or weathered such that these types of comparisons and deductions were not possible.

White (1976, p. 5) begged the question 'In how many mountain ranges or on how many arctic plains will bedrock fortuitously ever become >50% water-saturated . . . and then undergo rapid freezing to crack the rock?' In essence, White argued that it was adsorbed water (hydration) that was a major cause of rock breakdown in cold regions as the hydration mechanism was able to generate stresses sufficient to fracture the rock but did not require the high levels of rock moisture necessary for 'classic' freeze–thaw (i.e. the

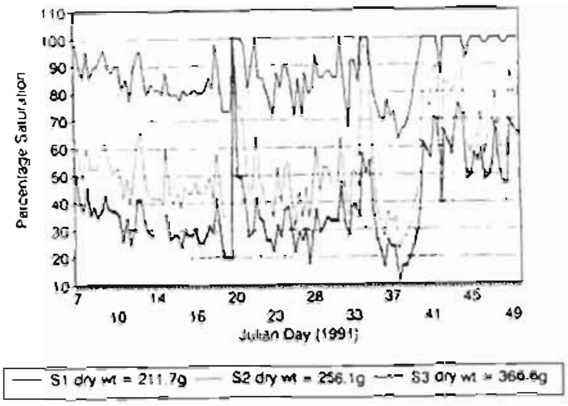

Figure 2 Percentage saturation for the three samples (S1–S3) for the south side of the rock dyke.

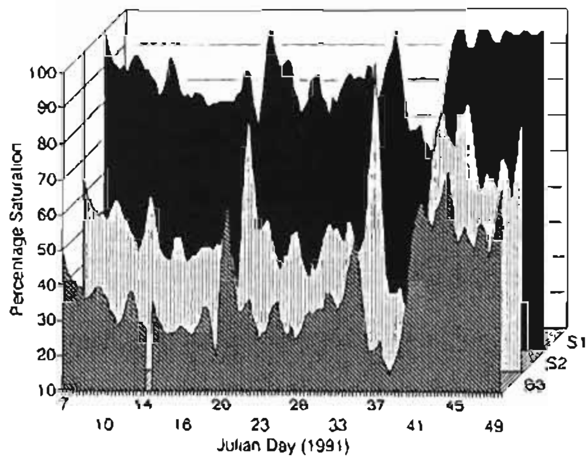

Figure 3 Three-term running mean for samples S1–S3 showing that the trends of the percentage saturation are the same.

9% volumetric expansion as water turns to ice). While not negating the basic premise of White (i.e. the necessity of knowledge concerning rock moisture content) the above data indicate that the question is not as simple as it may first appear. Clearly there is a rock moisture gradient, with the greatest amount of water near to the rock surface. Thus, there can be high moisture levels in the outer part of a rock and a zone of saturation, or near-saturation, that would vary in thickness as a function of the amount of water made available. For example, if sample S1 (thickness = 50 mm) is saturated but not sample S3 (thickness = 65 mm) then this implies that for S3 the outer 50 mm has a moisture content at or close to 100% and a lesser, unknown amount in the next 15 mm. As argued above, the 65 mm of sample S3 would crudely equate to the outer 65 mm of bedrock and so reflects the moisture gradient found there. Thus, while a large block of rock is *not*, as a whole, >50% saturated, the outer shell of that rock may well be. High rock moisture levels are also required by the ice segregation model of Hallet (1983) and so, again, knowledge of the moisture gradient is important. Although, following the question of White, freezing did *not* occur during the record period when rock moisture levels were high, it must be recognized that it *is* a possibility within this region.

Using clasts of similar mass and surface dimensions for the north and south sides of the dolerite dyke (Figure 4) it can be seen that the south side has a markedly lower degree of saturation when compared with the north. The lowest levels attained on the north side were of the order of 36% saturation, but most were above 40%, while on the south side levels below 20% saturation were attained and, except for the latter part of the record, most values were below 40%. As meteorological data indicate the dominant, rain-bearing winds to be from the north, it is not surprising that the northern aspect of the dyke had the higher degree of saturation. The south face exhibited high levels of rock saturation when the snow that accumulated on this lee side started to melt. Such a situation is shown by sample S1 for Julian days 43 to 49 where rock samples positioned prior to snowfall and buried by the snow indicate high moisture contents as the snow melted. The high rock moisture levels are *only* found in the outer part of the rock, there not being sufficient water available to penetrate to any substantial depth owing to the snow losing contact with the rock as thaw progresses. Thus, the northern aspect experienced frequent high levels of wetting as a consequence of the northerly, rain-bearing winds while the southern aspect received substantial wetting mainly as a result of snowmelt.

The data from the rock boss (Figure 5) still indicate the northern aspect to be wetter than the southern but the highest saturation levels occur on the eastern aspect. That the east experienced the highest moisture levels was not due to it receiving more precipitation but was rather a function of fewer drying events. The eastern aspect was little affected by dry northerly or southerly winds and received little in the way of direct radiation to facilitate evaporation. Although on clear days the eastern side *was* warmed by the sun

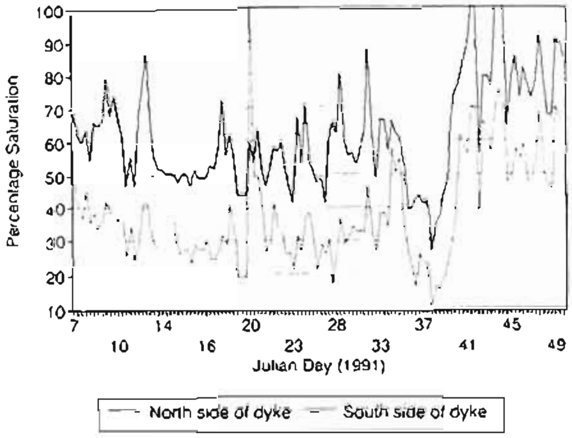

Figure 4 Actual values of percentage saturation for samples of similar size and mass for the north and south sides of the rock dyke.

and had temperatures markedly higher than the southern aspect, and only marginally below that of the northern aspect, the number of instances when this occurred were very few. The poor weather, with overcast skies and rain, that limited radiation to the eastern aspect often cleared by the late afternoon such that the western aspect received direct radiation on a number of occasions and was affected by the dry, northerly winds blowing past. In turn, this resulted in the western aspect having the (marginally) lowest saturation levels.

For the rock boss the sizes of the samples used equate to approximately the outer 60 mm of the bedrock. This then implies that during the record period the eastern aspect was always above 40% saturated, the northern was mostly above 30% and the south and west were, except for a few short periods prior to day 42, below 30% saturated. The north and south sides of the boss show reasonably similar patterns to that of the dyke, with comparable levels of saturation.

The data from the open site reflect a combination of frequent moisture inputs and rapid drying that result in a highly fluctuating graph (Figure 6). Being exposed through the full 360° the rocks freely received any and all precipitation, resulting in the high moisture contents. Equally, drying occurred due to the non-precipitation-bearing winds of any direction together with heating by incoming radiation at any time during daylight hours. Thus the rocks show both rapid wetting and rapid drying. Figure 6 also indicates a general decrease in moisture content from Julian day 7

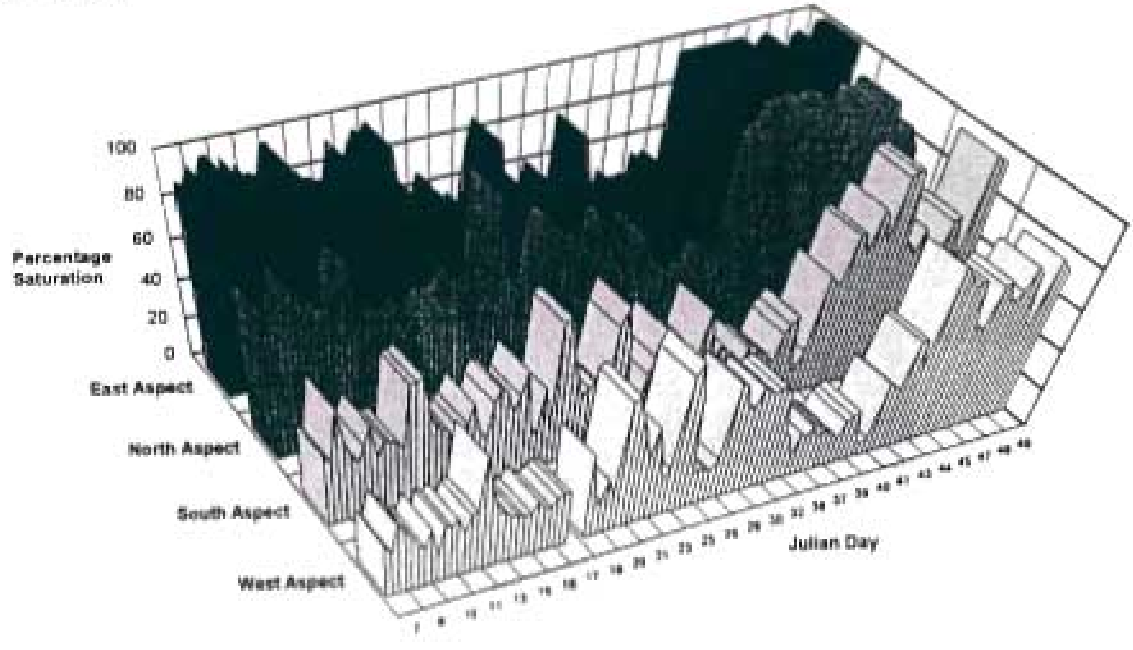

Figure 5 Percentage saturation for samples of similar size and mass for the four aspects of the rock boss.

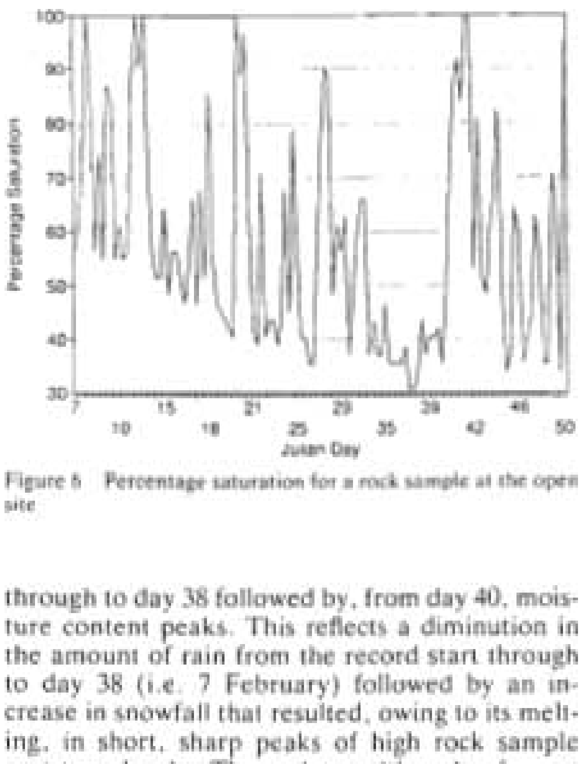

Figure 6 Percentage saturation for a rock sample at the open site

through to day 38 followed by, from day 40, moisture content peaks. This reflects a diminution in the amount of rain from the record start through to day 38 (i.e. 7 February) followed by an increase in snowfall that resulted, owing to its melting, in short, sharp peaks of high rock sample moisture levels. These data, although of great value, suggest that care should be taken in using data from samples in such an exposed situation (as was obtained in the earlier study of Hall, 1988) and then extrapolating the results to any particular site where aspect exerts control.

With respect to weathering, the rock moisture data help explain the cause of rock breakdown for this area and also show the importance of aspect upon weathering in general. Throughout the record period temperatures rarely fell below 0°C, either in the air or at the rock surface. In fact, there were only five events when rock surface temperatures went below 0°C and the lowest value recorded was −1.3°C. Only near the end of the study period were low temperatures starting to occur but they were in conjunction with snowfalls that covered much of the rock surfaces. At the start of the study period a large percentage (*c.* 60%) of the area was still snow covered. Thus, many of the upstanding rock outcrops, like the dyke and the rock boss used in this study, are insulated from air temperature fluctuations for a substantial part of the year by the protective snow cover. Only those rocks that become exposed in early spring or remain uncovered into late autumn are subject to freezing temperatures other than that of the annual freeze. On the Byers Peninsula it would appear that it is not the absence of water that is the constraint upon effective freeze–thaw action but rather that, owing to the insulating snow cover, the rock is not exposed to conducive thermal conditions (i.e. freeze–thaw cycles). Conversely, during the short summer period the rock surface temperatures were frequently greater than 5°C during the day, and values in excess of 15°C were measured on several occasions (Hall, 1993b). Thus, the rocks were frequently in a wet state and subject to warm, rather than freezing, temperatures.

The apparent limited role of freeze–thaw on the Byers Peninsula is at odds with the suggestions of other workers in the South Shetland Islands (e.g. John and Sugden, 1971; Araya and Hervé, 1972; Simonov, 1977; Hansom, 1983; Stäblein, 1983) who emphasize the importance of this process. Although researchers have recorded frequent freeze–thaw events based on air temperatures, the thick snow cover effectively insulates the majority of the rock outcrops from its effect. Thus, as Thorn (1988) has noted, air temperatures are *not* a surrogate for rock temperatures. Rather, it would seem that the high moisture contents coupled with the high rock temperatures that occur in summer are conducive to chemical weathering. The available data regarding weathering rinds (Hall, 1993a) certainly substantiate the proposal of active chemical weathering in the Antarctic. This was also recently proposed for the Antarctic continent by Balke *et al.* (1991). Although the absence of freezing during the short summer and the insulating effect of the snow during spring and autumn inhibit frequent freeze–thaw events (which effect destruction by the 9% volumetric increase as water changes to ice) the presence of moisture coupled with slow rates of fall of temperature below the autumn snow cover may be conducive to frost action resulting from the segregation ice model of Hallet (1983). For the model of Hallet, slow rates of fall of temperature are required and, if sufficient moisture is available, the growth of ice will continue so long as water can be drawn to the freezing zone. As ice growth is in excess of the simple 9% volume change, it will ultimately exceed the available space and thus cause rock fracture. Practical limitations on the applicability of this model are that rock moisture contents are often low at the onset of the freeze period and that rates of fall of temperature exceed those required by the model. On the Byers Peninsula, however, conditions may well be conducive to this form of frost weathering and, recognizing the limited frequency of other forms of freeze–thaw action, could help explain the clearly fractured rock that is observed on the peninsula.

The frequent wetting to various depths of the outer shell of the rock, coupled with the drying by either wind or sun, makes weathering by the process of wetting and drying highly likely. During the short summer period over 100 wet–dry cycles can occur (Hall, 1992) and laboratory experiments currently in progress indicate a mass loss of 3.2 g from 82 wet–dry cycles applied to a 917.9 g block of local rock. While wetting and drying can clearly operate to loosen surface material it is also possible, as postulated by Hall (1991a), that it can cause subsurface weakening and even failure. Where the rocks are experiencing frequent wetting and drying cycles, but the rock never actually attains a totally dry state, then the mechanical weathering effect will be experienced at some depth below the rock surface. Such a situation could clearly operate synergistically with other mechanical weathering processes, such as freeze–thaw (when it occurs), to effect a faster rate of breakdown. Weathering rinds, found to depths of *c.* 4 mm are indicative of the presence of moisture, and suggest that the wetting and drying mechanism could also be operative at or near this level.

CONCLUSIONS

The lack of data regarding rock moisture constitutes one of the major stumbling blocks with respect to our ability to determine the nature of cryogenic bedrock weathering. Like rock temperature, rock moisture varies both temporally and spatially and so data are required before any judgement can be made regarding what processes are operative. Data obtained from this study indicate that rock moisture varies significantly as a function of aspect and can influence the nature of the weathering that will take place. It is found that the outer shell of the rock often attains high moisture levels such that, had freezing taken place, then 'classic' frost action could have occurred. Equally, the high moisture levels would be conducive to rock damage due to segregation ice should the rate of fall of temperature be slow enough, as may occur during autumn. However, although many authors have argued for the role of frost weathering in this region, it would appear that wetting and drying may be a major contributor to mechanical weathering. More importantly, chemical weathering also appears to be a major factor on the southern aspects where rock moisture is maintained in the outer part of the rock, owing to it being protected from the sun and the drying northerly winds, and rock temperatures remain above zero for long periods during the summer despite the southerly aspect. The data provided here, although more detailed than those obtained from any other Antarctic location, are still highly limited temporally. Thus, a call is made for more extensive data acquisition with particular emphasis on long-term recordings. Until this is obtained, judgements regarding the nature of weathering remain highly speculative and it is not possible to test the applicability of theoretical models.

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A major element, rarely ever quantified, in rock weathering studies is that of interstitial rock moisture. Essentially there are four factors that need to be monitored: 1) Rock moisture content, 2) Rock moisture distribution, 3) Rock moisture chemistry, and 4) Status of the moisture (i.e. frozen, liquid or gas). As fundamental as these factors are most are never monitored in the field and for (2) there is no known field technique available to facilitate data acquisition. An outline of methods that allow for moisture content, chemistry and status to be monitored are given. Two techniques, one crude and one more sophisticated (but more problematic), that provide on-going data regarding moisture distribution will be suggested. Finally, all of the above are presented in the context of the fundamental need for such data.

A METHOD FOR THE EXTRACTION AND ANALYSIS OF SOLUTES FROM ROCK SAMPLES WITH SOME COMMENTS ON THE IMPLICATIONS FOR WEATHERING STUDIES: AN EXAMPLE FROM SIGNY ISLAND, ANTARCTICA

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Abstract. A new method is presented that allows determination of the solutes available inside a rock sample. Such results have not previously been available and yet they are of great importance for investigation of both chemical and mechanical weathering. The data also provide a useful input into studies of mineral cycling. The findings from this pilot study suggest that chemical weathering on Signy Island is in a very early stage. The indicated range of molarity for NaCl (0.34–0.57) is exactly that suggested by some workers to be most potent in aiding freeze–thaw action.

Introduction

A major problem in weathering studies is the accurate assessment of rock water content (McGreevy and Whalley, 1985; Hall, in press *a*) for this is an important control of both physical and chemical weathering. In addition, knowledge of the chemical composition of the rock water is extremely important as it helps determine what chemical reactions have taken place within the rock (Selby, 1982). Further, the types and amount of salts present in the aqueous content have a direct control on the rate, and the actual mechanism, of both freeze–thaw and salt weathering (Hallet, 1983; Williams and Robinson, 1981). Yet, despite its obvious significance, the analysis of interstitial water from rocks is limited, as far as the authors know, to one sample from chalk undertaken by Kinniburgh and Miles (1983).

Whilst there are data available on the chemistry of precipitation and soil moisture content (Walton, 1984; O'Brien and others, 1979; Mumford and Peel, 1982; Dixon and others, 1984) and of run-off in cold environments (e.g. Reynolds and Johnson, 1972) these do not take cognisance of changes that may take place within the rock. For instance, the interaction of water availability, rock properties and chemical weathering could give solute accumulation within the rock or may result in the flushing-out of the weathering products in solution. Such information provides essential data for the investigation of mineral cycling and pedogenesis as it aids quantification of the chemical input to the soil directly from the rock rather than as a product of soil processes *per se*. In addition, the role of chemical weathering in cold climates is gaining significance for localities where adequate water and hydrogen ion supply are available (Dixon and others, 1984; Thorn, 1976; Dixon, 1983) and so there is a pressing need for information on rock water chemistry. This same information is also urgently required for the undertaking of laboratory simulations. Fahey (1985, p. 103), in trying to determine what solutions to use for laboratory simulation of frost and salt weathering, sums up the present situation: 'Unfortunately there are virtually no published data available on the concentration of salt solutions in low temperature environments.' Consequently discrepancies in simulated weathering experiments (e.g. Williams and Robinson, 1981; McGreevy, 1982) are largely a result of the lack of knowledge on what salt concentrations to use in the experiments.

The aim here is to present a new, relatively simple, method whereby the solutes from inside a rock may be analysed. The results of four pilot tests are given and then considered in terms of their implications for weathering in the maritime Antarctic.

Study Area

The samples were collected from Signy Island (60° 43′ S, 45° 38′ W), one of the smaller islands in the South Orkney Islands group (Fig. 1). Geologically the island comprises metamorphosed sediments, primarily quartz-micaschist but with smaller areas of amphibolites, marbles and quartzites (Mathews and Maling, 1967; Storey and Meneilly, 1985). The island is small in size, about 8 km north–south and 4.8 km east–west, with an area of 19.94 km^2 and rising to a height of 279 m. About one-third of the island is currently ice covered but much of the remainder is subject to long-term snow cover. Precipitation is approximately 0.4 m yr^{-1} but with running water in summer mainly as a result of snow and ice melt.

Methodology

Samples of quartz-micaschist were collected from the cliff-face and scree at the base of Factory Bluffs in Factory Cove (Fig. 1). The cliffs are north-facing and only a short distance (*c.* 20 m) from the sea. Upon collection the rocks were sealed in several plastic bags and returned to the island laboratory. A sample of snow, that had fallen during the preceding 24 h, was also collected from the surface of the scree for analysis.

Laboratory Procedure

Upon return to the island laboratory the rocks were removed from their bags and weighed. They were then dried for 72 h at 105 °C to drive-off any moisture and leave all solutes as precipitates. The rocks were then sealed in bags and returned to South Africa, where they were then weighed, crushed and milled to a coarse powder. A subsample of 50 g of the powder was added to 50 ml of deionized water; the whole was then shaken for 15 min to dissolve all soluble material. The aqueous solution was then filtered under vacuum and forced through a 0.4 μm filter to remove any fine material in suspension. One millilitre of concentrated HNO_3 was added to the filtrate to dissociate the cations from the salts. The resulting sample was then analysed for the elements Ca, Na, Mg, K, Si, Fe and Cu using an Instrument Laboratory Plasma-100 sequential inductively coupled Argon plasma emission spectrometer.

Sub-samples of the rocks were analysed, following the procedure of Cooke (1979), for their porosity, water absorption, capacity, saturation coefficient and microporosity, as all of these properties are related to water availability and movement within the rock. Analysis of the chemical composition of two quartz-micaschist samples was undertaken by means of electron excitation X-ray fluorescence spectrometry (Walton, pers. comm.) whilst X-ray diffraction, using peak height percentages, was undertaken on the four samples to obtain the percentage composition of plagioclase, silica, mica and chlorite.

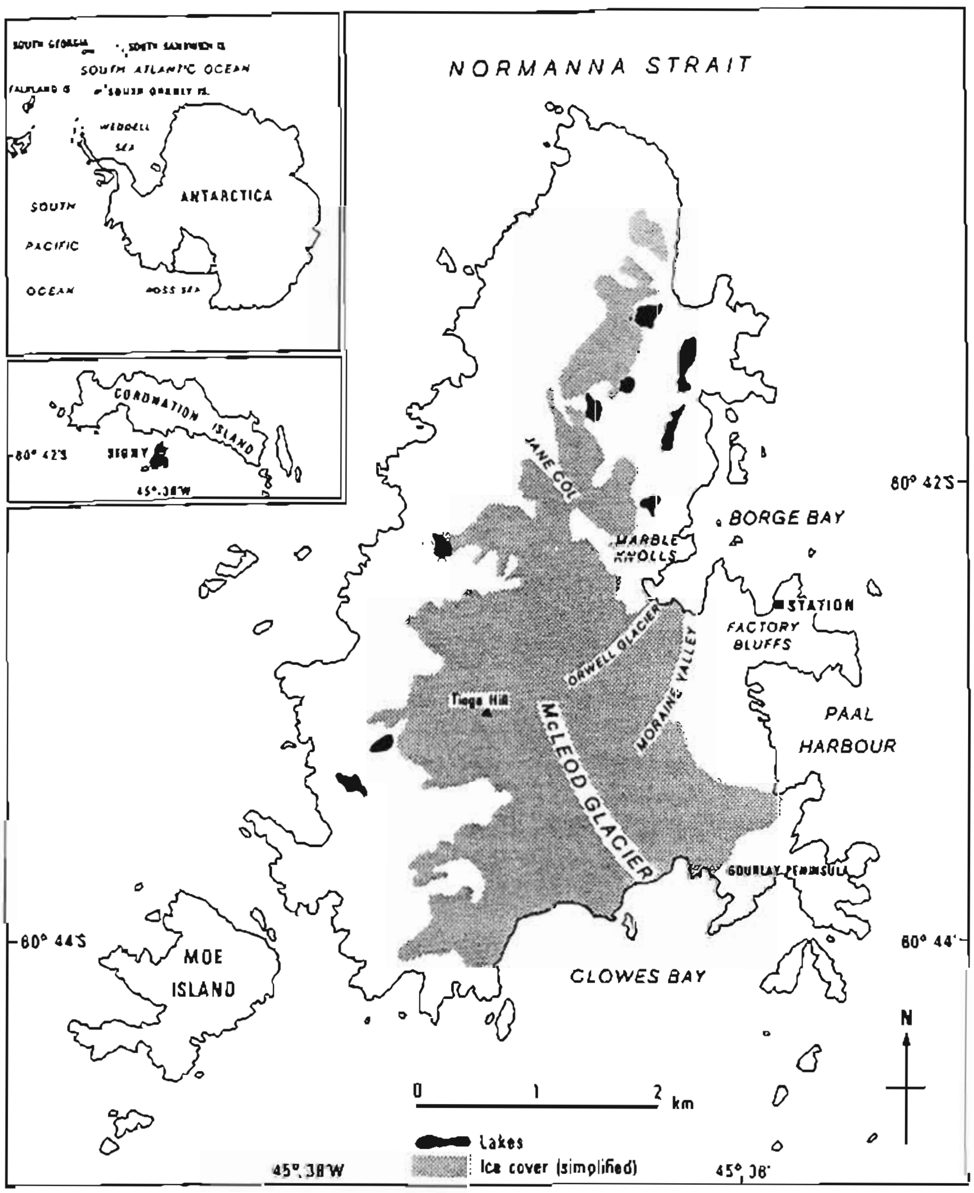

Fig. 1. Location map for Signy Island.

Results and Discussion

Results of the spectrometer analyses of the rock solutes and the snow sample, together with the properties of the rocks, are presented in Table I. X-ray diffraction analyses of the four rock samples are given in Table II and the Betaprobe results for two schist samples are shown in Table III. Whilst it must be considered that water from sea spray, if it entered the rock, could introduce salts, particularly those of Na and Mg, there is, nevertheless, a noticeable difference between the solute content of the snow and the rock. Whilst the origin of the salts (i.e. externally introduced or as a product of weathering) is important with respect to comprehension of chemical

Table I. Rock properties, moisture content and spectrometer analysis of solutes for four Signy Island schist samples.

Property	Sample 1	2	3	4	Snow
Porosity (%)	2.63	1.48	2.01	—[1]	—
Water absorption capacity (%)	2.16	0.99	1.85	—	—
Saturation coefficient	0.82	0.67	0.92	—	—
Microporosity (%)	75.0	58.3	44.0	—	—
Microporosity M[12]	1.97	0.86	0.89	—	—
Rock dry weight (g)	144.7	227.7	330.0	484.0	—
Moisture content (g)	0.80	1.30	2.0	1.98	—
Moisture content (%)[3]	0.55	0.57	0.61	0.41	—
Element (p.p.m.)					
Ca	9.1	10.8	6.8	6.6	0.56
Na	65.1	74.8	62.9	32.0	2.55
Mg	2.47	2.52	1.23	2.16	0.12
K	41.7	67.9	46.8	41.6	0
Fe	0	0.12	0	0.37	0
Si	1.0	1.4	1.8	2.0	0
Cu	0.07	0.10	0.14	0.20	0.016

[1] Rock crushed before these properties found.
[2] Microporosity M[1] is microporosity as per cent of total bulk volume.
[3] Moisture content as a percentage of rock dry weight.

Table II. X-ray diffraction analyses to show the percentage mineral composition of the four rock samples detailed in Table I.

Rock component	1	2	5	6
Plagioclase	37[1]	48	50	31
Silica	22	26	33	38
Mica	27	20	15	29
Chlorite	14	6	2	2

[1] Peak height percentages.

Table III. Betaprobe analysis of two characteristic schist samples from Signy Island.[1]

Compound	Rock A (%)	B (%)
S_1O_2	60.0	64.3
Al_2O_3	15.4	15.2
Total Fe as Fe_2O_3	7.52	7.29
MgO	2.69	2.90
CaO	5.04	1.94
Na_2O	3.48	2.28
K_2O	2.00	2.64
MnO	0.12	0.10
TiO_2	1.28	0.73
P_2O_5	0.24	0.17
F	0.06	0.06
S	0.04	—

[1] Data supplied by Walton (pers. comm.) and do not represent analyses of samples 1–4 shown in Tables I and II.

weathering processes and rates, it is rather what salts are available, and in what amounts, *within* the rock that are pertinent to mechanical weathering considerations (see below). One factor that introduces an error into the spectrometer analyses is that some Na and K may be liberated from the crushed plagioclase and/or K-feldspar during the dissolution treatment. This, though, is considered to be minimal as comparison of results of different shaking times up to one hour showed no significant differences in solute content. If highly accurate results are required then a similar amount of feldspar to that found in the rock could be subject to the same treatment, the solution analysed and the resulting p.p.m. of Na and K subtracted from that found for the crushed rock. In this pilot study this was not possible.

Utilizing the known weight of the original rock and its moisture content, it was possible to change the data units from p.p.m. to moles. Since 1 mole Na derives from 1 mole NaCl the molarity of NaCl for the four samples was calculated (0.57, 0.51, 0.45 and 0.34 M for samples 1–4 respectively). This allowed a determination of the salt most active in mechanical weathering in units which allow comparison with simulation studies of other workers. Some authors (e.g. Fahey, 1985) utilized $MgSO_4$ for polar weathering simulation but the low molarity found here (0.012, 0.018, 0.027 and 0.04 M for samples 1–4) suggests it is not an active salt in this particular maritime Antarctic environment.

It is thought that Ca and Na are derived from the plagioclase, which is found in high concentrations (Table II) and, compared to the other minerals, is relatively soluble. The Mg comes from the chlorite and the K is derived from the micas. Si could come from any of the minerals listed in Table II and may, in fact, be derived from them all. The Fe could come from either the chlorite or the micas whilst the origin of the Cu is uncertain. Overall, the impression is one of a very early stage in chemical weathering, particularly as extensive leaching would have given higher Si values. This conclusion is borne out by the results of tests on the mechanical strength of the rock (Hall, in press *b*) which, on the engineering grade classification of Day (1980), classify them as only 'slightly weathered'.

It is worth noting that McGreevy (1982), in experiments on rock breakdown due to the effects of freezing salt solutions, found that the greatest destruction was associated with 0.25–0.5 M solutions of NaCl. Exactly how and why such a concentration may abet freeze-thaw weathering is uncertain, and it could even be that the damage occurs due to hydration ($NaCl \rightarrow NaCl \cdot 2H_2O$) at +0.2°C (McGreevy, 1982). However, Hallet (1983) suggests that at low solute concentrations the freezing temperature is lowered and so the unfrozen water content is increased, as too is water mobility at sub-freezing temperatures. By increasing water transport to freezing interfaces, crack growth speed would be accelerated and thus more rapid destruction would occur. Whatever the actual mechanism, the importance of this technique is that it has, for the first time, allowed some estimate of the actual salt content of the rock to be determined. Thus, laboratory simulations can now be undertaken with a closer approximation to reality in an endeavour to elucidate what processes are operative.

Conclusions

A new method for determining the solute content of rocks is presented. In its current form it is not a high precision technique but refinements could be added to improve accuracy. Nevertheless, these data are unique and, as such, give a valuable guide to both the understanding of weathering processes (particularly freeze–thaw and salt weathering) on Signy Island and to what salt concentrations to use in laboratory simulations of mechanical weathering. The solute concentrations that have been found suggest limited chemical weathering of the rock, which is perhaps to be expected from a cool Antarctic environment and rocks with low water contents, and this is consistent with the findings of little diminished rock strength. With respect to mechanical weathering, the analyses show NaCl to be the dominant salt and its concentration to be within the range suggested to be the most effective in aiding freeze-thaw. Upon the encouraging basis of these preliminary findings further experiments are planned to enlarge the data base to a statistically valid size.

Acknowledgements

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AQUEOUS GEOCHEMISTRY AS AN INDICATOR OF CHEMICAL WEATHERING ON SOUTHEASTERN ALEXANDER ISLAND, ANTARCTICA[1]

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Abstract: Chemical alteration of rock in the Antarctic is considered to be less dominant than physical weathering processes. The research presented in this discussion represents investigations to determine the extent of contemporary chemical weathering in an area that hitherto has not been investigated. A glacial outwash stream on southeastern Alexander Island was investigated to determine the extent of active chemical weathering. Data indicate that solution of minerals does take place during the Antarctic summer when water is present. Contrary to studies elsewhere in the Antarctic, there is little evidence of maritime or biological influences on the observed weathering regime. Although chemical weathering is active, its relative importance compared to that of physical weathering could not be determined.

INTRODUCTION

The weathering environment in cold regions is usually assumed to be dominated by physical weathering processes, and this is particularly the case in the Antarctic (Campbell and Claridge, 1987; Thorn, 1992; Hall, 1995). Low temperatures have resulted in the assumption that frost shattering is the dominant Antarctic weathering process (Simonov, 1977). However, a number of studies have also indicated the role of salt weathering and thermal stress fatigue (e.g., Gunn and Warren, 1962) and some have questioned the possibility of frost action in an arid environment (van Autonboer, 1964). The problem with the assumption of freeze-thaw is twofold: (1) the absence of data to indicate the presence of water; and

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(2) the fact that rock temperatures in the summer are often extremely high despite the low air temperatures, exhibiting diurnal fluctuations across the 0°C boundary. The key is that the presence of water (as required for frost action) combined with high (10°C) rock temperatures provide an environment that is conducive to chemical weathering. Balke et al. (1991) recently demonstrated that water, not temperature, is the limiting factor in facilitating chemical weathering in the Antarctic, even though the period in which such weathering can take place (summer) is relatively short.

As noted above, chemical weathering is acknowledged as taking place in the Antarctic (Green and Canfield, 1984; Weed and Ackert, 1986; Campbell and Claridge, 1987; Green et al., 1988; Balke et al., 1991; de Mora et al., 1991, 1994) and recent research in aqueous geochemistry has shown that bedrock weathering contributes to the chemical composition of stream and standing water in the dry valleys of eastern Antarctica (e.g., Green and Canfield, 1984; Green et al., 1988; Balke et al., 1991; de Mora, et al., 1991, 1994). However, studies of streams in the dry valleys (Green and Canfield; 1984; Green et al., 1988; de Mora et al., 1991), meltwater on the Ross Ice Shelf (de Mora et al., 1991), snow near the Antarctic coast (Piccardi et al., 1994), and pools in the maritime Antarctic (Jones et al., 1993; Caulkett and Ellis-Evans, 1997) all have found that aerosols and salts from a maritime source are a significant source of the ions.

Geological controls on weathering processes are immediately apparent in the study area. Both vertical and lateral dilatation, resulting from deglaciation, is observed on ridge tops and along the sides of valleys, respectively. The results of chemical weathering also are particularly evident along joints and bedding planes. Iron oxide stains are found in and adjacent to some horizontal beds. Furthermore, gypsum deposits, sometimes substantial (≥24 mm thick), are found in joints and close to ground level around bedrock or large boulders. Evidence suggests that these precipitates originate from the solution of calcite cements and subsequent precipitation as gypsum (Horne, 1967, 1968; Taylor et al., 1979). It has been hypothesized that the precipitates are paleo-features indicative of accumulation under stagnant conditions (Taylor et al., 1979). However, it is equally likely that they are the result of precipitation of solutes from mobile solutions. This is suggested by their occurrence at the edges of water channels and at places where moisture from melting snow and ice was (or is) available. The origin of the water to provide the precipitates, particularly considering their relationship to joints and bedding, plus their occurrence around the base of boulders in or near the valley bottom as well as the observed thicknesses, is likely to be meltwater during deglaciation. The loss of the extensive ice cover would have produced substantial melt, and its movement along joints and bedding planes, plus concentration on the valley bottom, could be expected.

The observed landscape is the result of former as well as contemporary weathering and other geomorphic processes. The questions thus arise as to what are the nature and role of present-day weathering and transport processes and their effect on the landscape. Hall (1997a, 1997b) provides some answers regarding mechanical weathering and its relationship to cryoplanation forms found in this area. A major unresolved problem is that of the nature and impact of chemical weathering and the role it plays in landscape development. In an attempt to obtain answers to this question, chemical analyses of meltwater streams in Viking Valley and of standing snow-melt pools in the surrounding areas were undertaken.

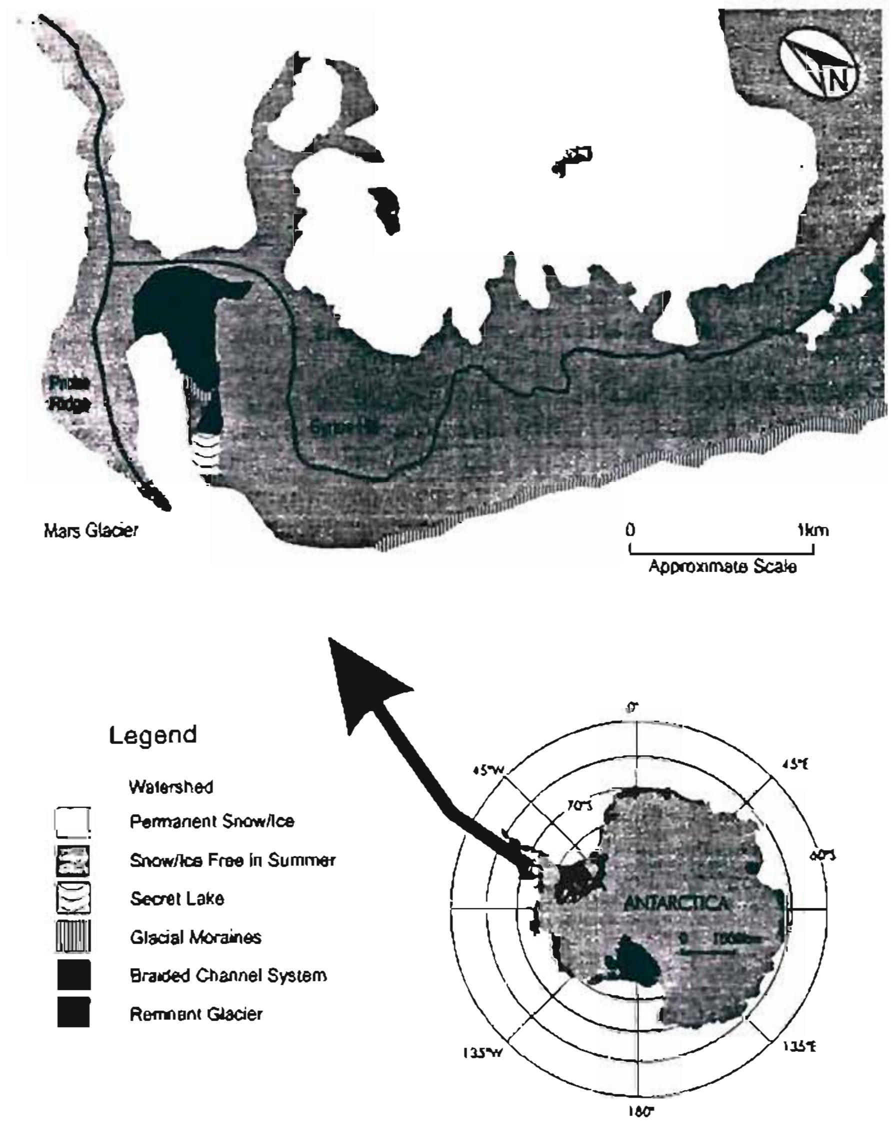

Fig. 1. The study area in southeastern Alexander Island, Antarctica.

SETTING

The study site, located at 71°50' S, 68°21' W, is one of several east-west-oriented tributaries of the Mars Glacier on Alexander Island (Fig. 1). These east-west-trending valleys are all markedly asymmetrical, with steep south-facing slopes and more gentle north-facing slopes (Meiklejohn, 1994b). Viking Valley contains a small (700 m long) remnant glacier at its head (eastern end), which feeds summer meltwater into Secret Lake via a braided channel system. The lake developed as a result of water pooling against a rock bar within the glacially overdeepened basin

(Meiklejohn, 1994b). The flow of water exhibits diurnal cycles, with the strongest flow occurring in the afternoon after the sun has reached its highest point above the northern horizon, while the weakest flow is in the early hours of the morning after the sun has reached the lowest point above the southern horizon. Water flow in the 1992-1993 Austral summer in Viking Valley was observed to increase from early summer (early December) until reaching a peak in early January, after which the onset of cooler weather resulted in no further melting. This has been observed to be the general seasonal pattern in areas to the north of this particular site (Taylor et al., 1979).

The geology of the exposed ground surface from Syrtis Hill in the north to Two-Step Cliffs, some 5 km^2 in extent, appears almost identical to that described to the north (Horne, 1967, 1968; Taylor et al., 1979). The study area is part of a belt of Jurassic and Cretaceous marine sedimentary rocks bounded on the east by King George VI Sound and to the west by the Le May Range (Horne, 1968; Taylor et al., 1979). Deglaciation of the well-jointed sedimentary rocks has, with weathering, produced a generally rounded style of topography (Taylor et al., 1979). Quartz, alkali-feldspar, and plagioclase are the main mineral constituents of sandstones of this study area, and over 50% were found to have calcite cement. Of significance to the present discussion regarding chemical weathering are the following observations. The area is one of continuous permafrost, with a shallow active layer (ca. 5 to 20 cm). An unusual form of non-sorted patterned ground, characterized by pseudo-sorting around the non-sorted margins, is abundant in this area, and observations appear to indicate that both thermal contraction and frost action (weathering and/or sorting) have contributed to the observed polygonal patterns (Hall, 1997a). Details of the weathering regime and the role of mechanical weathering in Viking Valley can be found in Hall (1997b). A distinctive landform on the nunataks in this area is the so-called "cryoplanation terrace" and details of observations and a discussion regarding their formation can be found in Hall (1997a).

METHODOLOGY

Water samples (of ca. 250 ml) were collected at weekly intervals from various specific points in Viking Valley between 11 December 1992 and 10 January 1993 (Fig. 2). In addition, pH, electrical conductivity, and water-temperature measurements were taken at hourly intervals at 17 locations for much of the study season. Solute analysis followed the procedure adopted by Hall et al. (1986) and Meiklejohn (1994a). The specific locations from which water was sampled and measurements taken were selected so that it would be possible to identify any changes in chemistry that occurred in glacial and snow melt over a distance of ca. 150 m. The water samples were analyzed for soluble cations (K^+, Na^+, Mg^{2+}, Ca^{2+}, and Si^{4+}) and anions (Cl^-, NO_3^-, C_3^{2-}, and SO_4^{2-}) using atomic absorption and ion chromatography, respectively. Each sample was stored in refrigerated conditions and filtered through a 0.45 μm micropore filter before analysis. Cation analysis was conducted in a Plasma 100 atomic absorption spectrophotometer. Prior to analysis for cations, each filtered sample was ionized using concentrated nitric acid (0.2 ml conc. nitric acid was added to 20 ml of the sample). Anion analysis was conducted in a Waters liquid ion chromatograph.

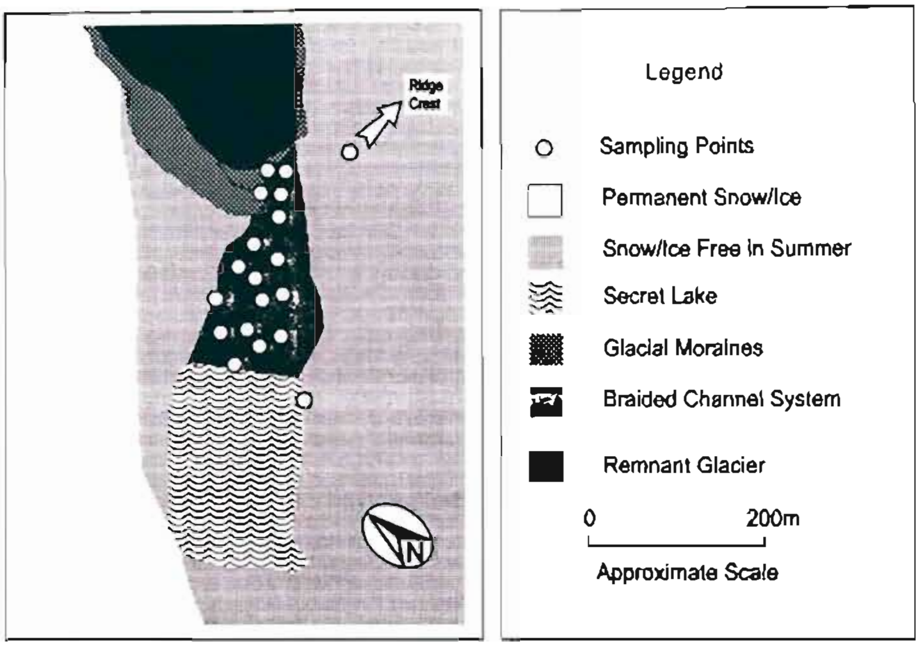

Fig. 2. The location of sampling points in Viking Valley, Alexander Island.

RESULTS AND DISCUSSION

Studies in the dry valleys of East Antarctica and in the Maritime Antarctic indicate a maritime influence on the aqueous geochemistry (Green and Canfield, 1984; Green et al., 1988; Balke et al., 1991; de Mora et al., 1991). However, the low Na^+ and K^+ cation concentrations from the Viking Valley indicate that there is negligible maritime influence (Table 1, Fig. 3). The major cation concentrations were calcium and magnesium, whereas the most common anions were chloride and sulfate. Nitrates also were found, but because of a lack of any specific observed tendencies and the instability of nitrates in light, it was decided not to consider them in this discussion. Given the low concentrations of ions in ice and snow samples, the results indicate that bedrock is the likely source for the observed chemical composition of the meltwater.

The source of the calcium is likely the result of calcite solution or the resolution of gypsum precipitates along the watercourse (comment from an anonymous referee). Data show that chemical concentrations increase over a seemingly short distance (ca. 150 m) (Table 1; Fig. 3). Snow had the lowest concentration of ions, whereas that of the glacier ice was slightly higher (Table 1). Water samples from the stream source had a similar chemical composition to that of the fresh snow, but with increased concentrations where melting glacier ice contributed to the flow. The observed increase in chemical composition down the length of the braided channel system was reflected in increased electrical conductivity readings (Fig. 4); pH values exhibited a similar increase, averaging 7.3 at the start of the outwash stream and 7.8 at the lake entrance, probably as a result of the elevated alkalinity resulting from augmented calcium and magnesium-ion concentrations.

TABLE 1

Average Concentrations of Certain Ions (in mg/l) in Water Samples from Viking Valley, Alexander Island[a]

Sample	n	Mg^{2+}	Ca^{2+}	Cl^-	SO_4^{2-}
Snow	7	0.11	0.70	0.36	0.00
Glacier ice	7	0.14	1.00	0.82	0.12
Source	7	0.13	0.92	0.28	0.12
Braided channel	21	0.39	2.95	3.41	0.41
Lake entrance	7	0.69	5.13	4.24	0.78
Melt pool (ridge crest)	5	17.40	37.50	9.66	12.52

[a]Individual samples were determined $\pm 10^{-6}$ mg/l; averages were calculated and the values rounded.

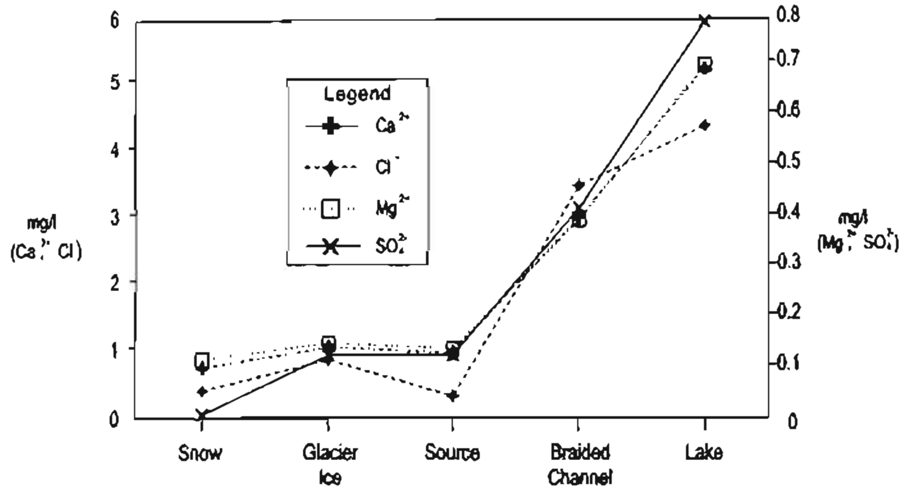

Fig. 3. Ion concentrations of water samples from Viking Valley, Alexander Island.

Alternatively, the increase in pH could be due to the hydrolysis of CO_3^{2-} (comment from an anonymous referee). It is more difficult to identify the sources of the other ions identified in the analysis. There has been no documented volcanic activity, which may account for them. Magnesium ions may originate from the solution of alkali feldspars, although the possibility exists that sulfate ions may originate from re-solution of gypsum or the solution of gypsum originating from calcite. Furthermore, the presence of the sulfur ion may be due to the equilibrium of gypsum and water (Hem, 1992; comment from an anonymous referee).

The significance of these data is that they show that the concentration of dissolved solids increases along the length of the stream; in other words, solution is taking place as the water flows along the braided channel system. It would thus

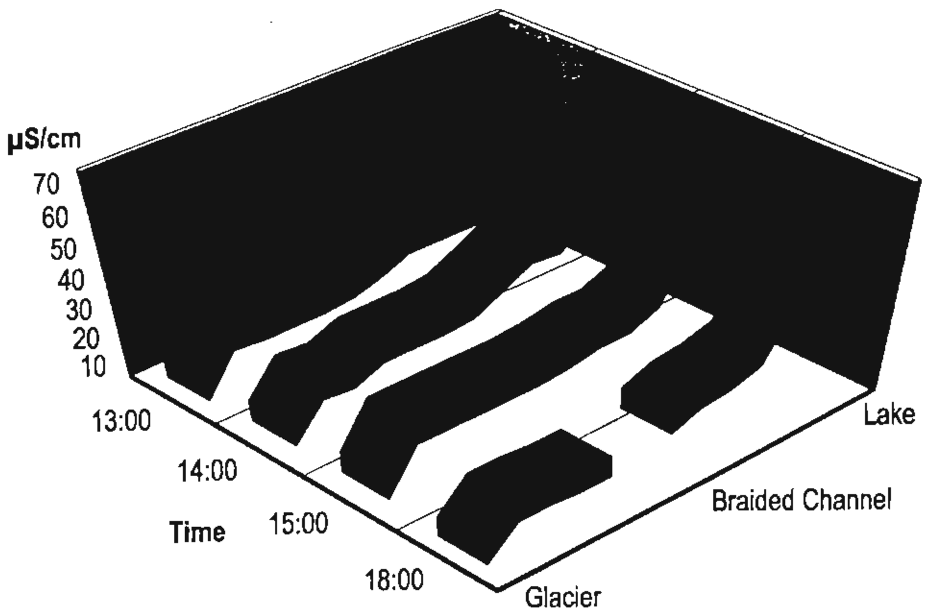

Fig. 4. Electrical conductivity of meltwater in Viking Valley, Alexander Island.

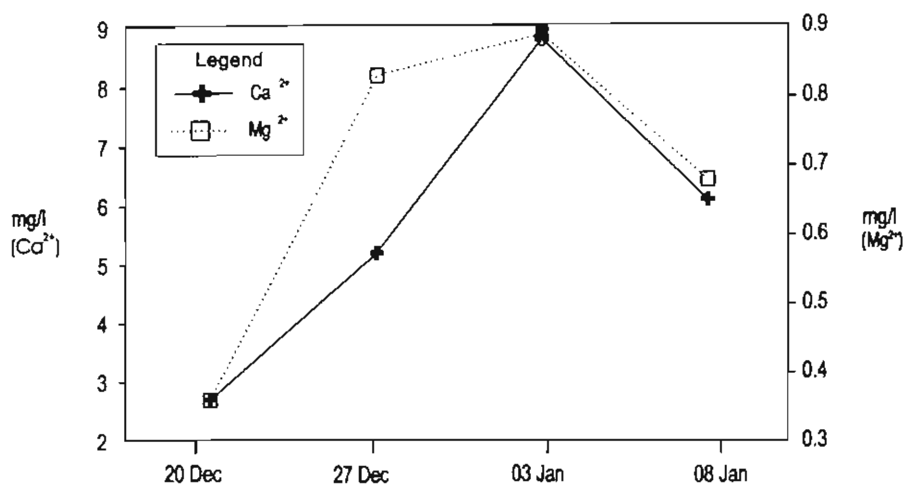

Fig. 5. Ion concentrations of water samples from Secret Lake, Viking Valley, Alexander Island.

seem that the ground over which the water flows contains soluble material that is, despite the relatively short distances involved, taken up by the water. If these ions were derived from re-solution of precipitates, the first meltwater samples would be expected to show the highest concentrations and these concentrations would then decrease through the summer as the source was depleted. However, the opposite was found; ion concentrations increased through the summer (Fig. 5). The contrary decrease in chemical concentrations of the lake water (Fig. 5) is likely to be the result of dilution following the melting of ice that covered its surface at the beginning of summer and an influx of water from the glacial outwash stream. This observation was supported by a noticeable rise in the level of the lake (of ca. 50 cm) during the monitoring period. The ice, which was anchored on the lake floor near its banks, had the effect of damming the inflow of water; following melting of the ice, mixing with the rest of the lake was thus possible and so concentrations became diluted. The implication of the above data is that solution of parent material takes place as the water flows over the valley floor—chemical weathering is taking place. Observed temporal variations of stream water chemical composition show that although ion concentrations increased down the length of the braided channel system during the day, these concentrations reached a peak in the early afternoon and then decreased (Fig. 4). A probable explanation is that as the flow of water increased, so there was no increase of water contact with soluble material, but rather a deepening of the water column, and hence less solution occurred relative to the volume of water (i.e., the amount of solution did not decrease but the volume of water increased such that there was, as a function of water volume, a decrease in soluble material).

Stream temperatures varied little during water flow, averaging close to 1°C, while that of the lake gradually increased from 0.2°C at 12:00 on 7 December 1992 to 3.8°C at 13:00 on 7 January 1993. It must be noted that the lake temperature was measured only near the edge, where it was shallow, and thus the temperature is only representative of the surface layer (particularly as the colder water would sink). Temperature data indicated no significant relationships with other variables (Table 2).

Possibly, the most significant data were obtained from analysis of standing water located between a melting snowpatch and the bedrock wall. The ion concentrations of these samples were more than 10 times that of the stream water (Table 1; Fig. 6). This water is in contact with bedrock longer than is the stream water in the valley bottom, and so a greater amount of solution is possible. The recorded high concentrations of calcium (Table 1) are likely derived from solution of calcite cement and alkali feldspars. Although it is accepted that the low temperatures may enhance the solubility of calcite, the relatively high concentrations of calcite compared to those of the glacial outwash stream are important. The key observation is that chemical weathering is enhanced where water is in contact with bedrock behind a late-lying snowpatch. In addition to chemical processes, the moisture is potentially responsible for a whole number of other weathering processes, such as wetting and drying, hydrolysis, and (given the thermal conditions) frost shattering. However, it is worth noting that during the study period the recorded temperature regime would not have been conducive to cryogenic weathering (Hall, 1997a). Nevertheless, the finding of chemical weathering at such a location indicates that weathering processes other than solely mechanical ones do take place in cold environments and consequently their synergistic role in landform development must be taken into account. Recognizing the direct association of both nivation and cryoplanation landforms with snow (Hall, 1997b), and the occurrence of such features in this area, a more holistic approach to weathering is required than the simplistic recourse to only freeze-thaw as the landform originator.

The evidence regarding the action of chemical weathering helps explain enhanced weathering at the base of cliffs and around large boulders as well as some

TABLE 2

Water Temperature (in °C) at 13:00 for the Stream in the Viking Valley Braided Channel System, Alexander Island, during the 1992-1993 Austral Summer[a]

Date	Source	15 m	26 m	35 m	50 m	70 m	90 m (seepage)	Lake entrance	Lake
21 Dec. 1992	0.2	0.4	stream frozen		2.7	— stream frozen —			0.6
24 Dec. 1992	0.5	0.8	1.0	1.1	1.3	2.8	2.9	3.8	1.0
26 Dec. 1992	0.4	0.5	0.8	0.9	1.0	2.3	3.2	3.5	1.3
27 Dec. 1992	0.3	0.3	1.2	1.2	1.3	2.6	3.7	3.9	2.0
29 Dec. 1992	0.3	0.3	1.1	1.6	1.8	3.5	5.6	2.5	1.7
31 Dec. 1992	— stream frozen —						6.2	3.5	2.4
1 Jan. 1993	0.2	1.1	2.0	3.1	stream frozen		6.4	6.1	2.8
2 Jan. 1993	0.3	1.0	2.1	2.7	4.0	4.5	5.4	4.7	2.9
3 Jan. 1993	0.3	0.2	1.1	1.2	1.4	3.3	2.6	3.2	2.5
4 Jan. 1993	0.5	0.2	1.7	1.8	2.1	4.7	4.2	5.6	2.6
5 Jan. 1993	0.5	0.2	1.2	1.3	1.6	3.6	3.9	4.1	2.8
6 Jan. 1993	0.3	0.6	1.4	1.6	2.1	4.8	2.7	2.4	2.6
7 Jan. 1993	0.6	0.4	1.5	1.7	2.1	5.5	3.4	4.4	2.7
8 Jan. 1993	0.4	0.4	1.3	1.6	2.0	4.8	3.5	4.6	3.8

[a]Note that the stream was frozen and not flowing on the days not indicated.

of the gypsum precipitates. Considering the observed occurrence of snow along the risers of terraces in this area, the moisture from snowmelt is highly likely to facilitate both mechanical and chemical weathering that play a combined role in terrace development. The observation of gypsum precipitates along the drainage runnels of the terrace treads indicates both chemical weathering and transport in solution of the weathering products. Their re-deposition by evaporation (from the transporting water) also may enhance mechanical weathering on the treads by crystallization pressures and ensuing thermal and/or hydration effects on the precipitates. Clearly, there is a need for greater consideration of chemical weathering within the developmental model of such terraces.

In the Allan Hills area of East Antarctica, pits observed in dolerites are ascribed to solution etching (Conca and Malin, 1986) thereby recognizing the effects of chemical weathering. Earlier research from this general region (e.g., Taylor et al., 1979), however, has suggested that the extensive honeycomb weathering seen here is likely the result of frost shattering, the moisture being supplied by wind-blown snow. Therefore, we would argue, given the effects of moisture as indicated by this discussion, the actual process(es) causing honeycomb weathering may be one of any number of weathering processes. Taffoni could thus be formed by a synergistic suite of weathering processes combined with wind- and/or water-based transport of the weathered products. Recent temperature data from a taffoni have indicated the role of thermal processes rather than frost shattering (Hall,

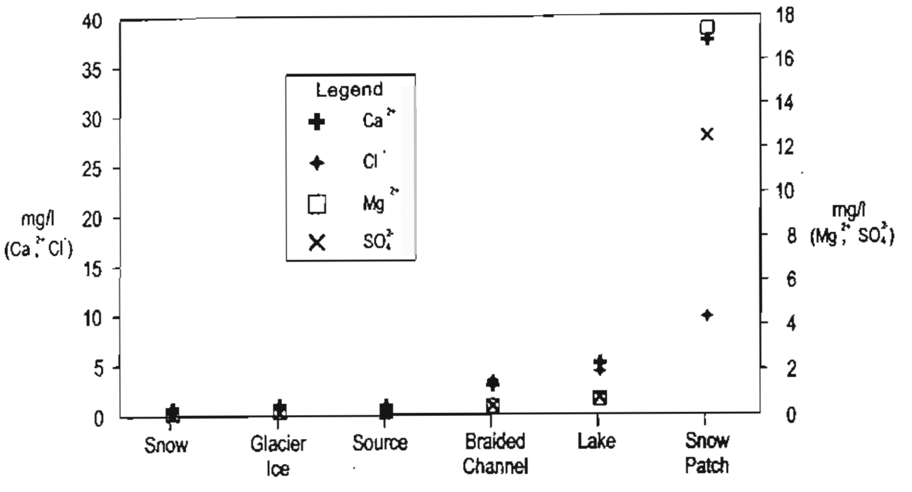

Fig. 6. Ion concentrations from sampling points in Viking Valley and a site on a ridge crest on southeastern Alexander Island.

1997a). Thus, as is the case with the terraces (noted above), there is a need for a more comprehensive approach to the initiation and development of taffoni. Indeed, even the very occurrence of gypsum precipitates along jointing and bedding, which is attributed to meltwater during deglaciation, may have played a role in initiation, whereas present-day weathering may be quite different—possibly based on thermal stresses in the absence of any water supply and the transport away of the precipitates during taffoni growth.

Whereas it can be seen that chemical weathering is indeed active during the Antarctic summer, its relative importance when compared to physical processes could not be determined. Although there are some data pertaining to mechanical weathering processes (Hall, 1997a), neither the chemical data presented here nor the mechanical weathering information are sufficient grounds for any meaningful judgment. Indeed, it will be extremely difficult to differentiate between the various weathering processes in the field situation, owing to their synergistic interaction and to the changes in that relationship over time (seasonal, annual, and on the longer, origin/developmental scale). It can, nevertheless, be stated that chemical weathering is a component of contemporary processes that are shaping the Antarctic landscape in this area.

CONCLUSION

The chemical analysis conducted from water samples in Viking Valley, Alexander Island indicate that contemporary chemical weathering processes are active in Antarctic environments where moisture is present. These findings support recent weathering studies (e.g., Green and Canfield, 1984; Weed and Ackert, 1986; Campbell and Claridge, 1987; Green et al., 1988; Balke et al., 1991; de Mora et al., 1994), which have shown that chemical weathering processes may indeed be significant in terms of bedrock alteration in cold environments. However, as has been shown with physical weathering processes (e.g. Hall, 1995), moisture is a constraining factor when considering the effectiveness of chemical weathering. It is, therefore, only possible for contemporary weathering processes to occur during the warmest part of summer when there is sufficient meltwater to provide moisture. It should be noted, however, that this also is a time of high rock temperatures and thus an environment highly conducive to chemical weathering is created for this short summer period. The importance of chemical weathering in creating the Antarctic landscape should, therefore, not be understated: it does occur and is a crucial component.

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THE UTILIZATION OF THE STRESS INTENSITY FACTOR (K_{IC}) IN A MODEL FOR ROCK FRACTURE DURING FREEZING: AN EXAMPLE FROM SIGNY ISLAND, THE MARITIME ANTARCTIC

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ABSTRACT. A recent theoretical model to explain rock fracture during freezing (Hallet, 1983) offers a potentially valuable means for explaining weathering in the field. The model utilizes the stress intensity factor (K_{IC}), and a value for this is derived for quartz-micaschist, the main rock on Signy Island. It is shown that the main problem of the model for field applications is the attainment of a saturated state. However, it does offer a theoretical framework against which field data can be evaluated, and it provides a mechanism to explain why the cliff faces are mainly unweathered whilst loose talus blocks show extensive breakdown.

INTRODUCTION

Freeze–thaw weathering is a process usually cited (e.g. Washburn, 1980) as a major weathering agent in polar and high-altitude locations. Despite the apparent ease with which it is quoted it is, nevertheless, a mechanism which is as yet little understood and for which there are few quantitative field observations (McGreevy, 1981). In broad terms there are three major controls: temperature, moisture and rock properties. Although especially germane to the mechanism, data on temperatures are few (see McGreevy, 1981 for a review) and those on moisture even rarer (Trenhaile and Mercan, 1984; Hall, in press *a*; McGreevy and Whalley, in press). When considering moisture it is apparent that its chemical composition is important (Williams and Robinson, 1981; McGreevy, 1982; Fahey, 1985) and yet information on solutes of rock moisture *in situ* is limited to one sample from chalk (Kinniburgh and Miles, 1983) and the recent analysis of four samples from this current study (Hall and others, 1986). The properties of the rock have been considered to varying extents in different studies with porosity, water absorption capacity, saturation coefficient and microporosity being the most common ones cited (e.g. Fahey, 1985; Hall, in press *a*). There are, however, other parameters such as compressive strength, specific energy index (Szlavin, 1974) and fracture toughness which play an important role. Here it is the use of the fracture toughness index (K_{IC}) which is to be considered.

Gunsallus and Kulhawy (1984) have stated clearly that the development of fracture mechanics is based on the precept that a single parameter, K, the stress intensity factor, can be used to represent the stress field ahead of a crack. The value of K is mainly a function of the level of applied stress and crack size. The fracture toughness of a rock is thus its critical stress intensity factor value. With knowledge of the value of K_{IC} (critical intensity factor mode I: the opening mode (Schmidt and Rossmanith, 1983)) for the particular rock under study and measurements of the crack length it is possible, by means of a model suggested by Hallet (1983), to calculate the pressure inside the crack for propagation to take place, and the temperature required to generate that pressure.

The moisture content of the rock, and its chemical composition, affect both the strength of that rock (Broch, 1979) and the operation of the model (Hallet, 1983), and therefore data on interstitial rock water is of vital importance. In addition, if the rock is saturated or nearly so, then temperature is the major control on the tensile

Br. Antarct. Surv. Bull. No. 72, 1986, pp. 53–60

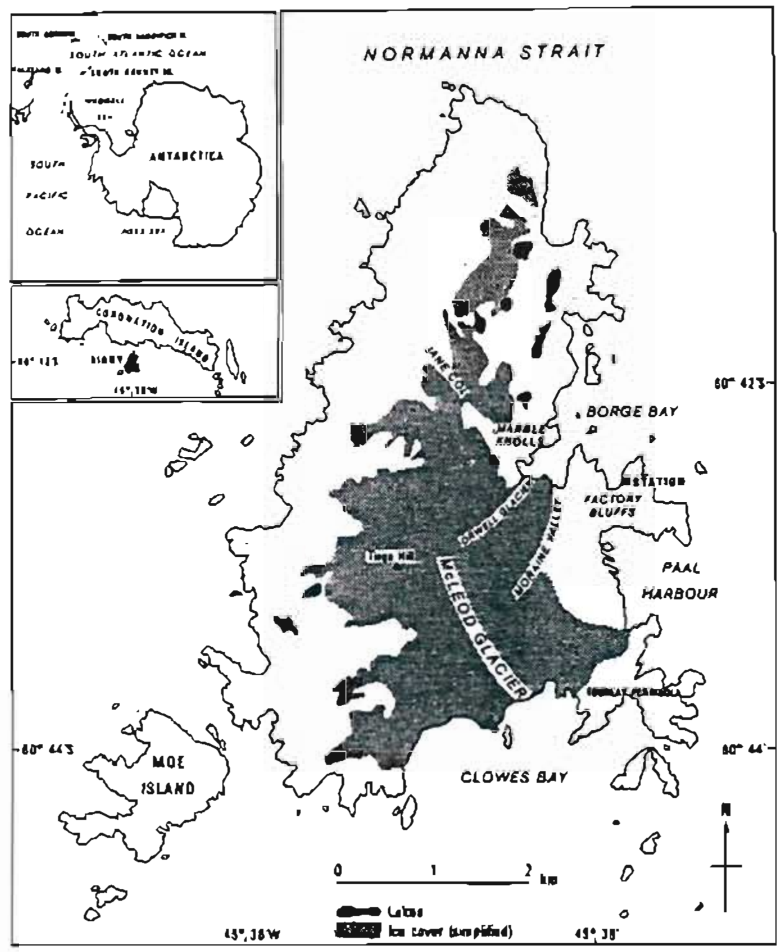

Fig. 1. Location map of Signy Island.

stress exerted by the growth of ice within the rock. So, for a valid utilization of K_{IC}, moisture and temperature data are required. In this study information has been obtained, for the first time in a polar environment, on both of these parameters and so the application of K_{IC} can be attempted. However, the details on temperature (Walton, 1982), moisture (Hall, in press *a*) and rock properties (Hall, in press *a* and *b*) are given elsewhere and it is only the derivation and application of K_{IC} to freeze–thaw studies within a maritime Antarctic environment which are presented here.

Study Area

Data and rock samples were collected from Signy Island (Lat. 60° 43′ S, Long. 45° 38′ W), one of the smaller islands in the South Orkneys group (Fig. 1). Signy is *c*. 8 km north to south, *c*. 4.8 km east to west, has an area of 19.94 km^2 and rises to a height of 279 m. Approximately one-third of the island is covered by an ice cap, but much of the rest of the island is subject to long-term snow cover and the whole is an area of discontinuous permafrost. Running water is limited and mainly confined to a few small streams along the margins of the ice cover.

Geologically the island comprises metamorphosed sediments, primarily quartz-micaschist with smaller outcrops of amphibolites, quartzites and marbles (Mathews and Maling, 1967; Storey and Meneilly, 1985). Climatically it is a typical cold, oceanic island with a mean monthly temperature of $-4°C$ but with the summer three months slightly above freezing (Collins and others, 1975). The precipitation is 0.4 m yr^{-1} and is mainly in the form of snow, except for January and February when rain can predominate. The mean amount of sunshine is only around 1.5 hr day^{-1} with average wind speeds of about 26 km hr^{-1}. The vegetation is typical of the maritime Antarctic and is comprised of fructicose lichens, cushion mosses and occasional areas of the grass *Deschampsia antarctica* (Smith, 1972). There are extensive areas of patterned ground (Chambers, 1967).

Techniques and Approach

Samples of quartz-micaschist were collected from various positions on rock outcrops and the ground surface for a variety of altitudes and aspects. In addition, sampling was undertaken in different environmental conditions during spring and summer, i.e. from under snow cover, during snowmelt, in a meltwater rivulet, during thawing of frozen ground, etc. The field moisture content of all samples was obtained (Hall, in press *a*) and the rocks subjected to the irregular lump point-load test in order to find their compressive strength (Broch and Franklin, 1972). As quartz-micaschist is anisotropic tests were carried out both parallel and transverse to schistosity (Hall, in press *b*). Data on temperature conditions on Signy at two vegetated sites are already available (Walton, 1977 and 1982) but year-round micro-meteorological data logging, by means of 'Datacapture' 16-channel recorders, is currently in progress at the three reference Fellfield sites (Factory Bluffs, Moraine Valley and Jane Col: Fig. 1). Finally, information on the solutes available *inside* the rock has been obtained (Hall and others, in press) as these may play a role in inhibiting (Mcgreevy, 1982) or abetting (Williams and Robinson, 1981) frost action. Thus, measures of rock moisture content *in situ*, and its chemical composition, together with point-load strength and a range of temperatures to which the rock may be subject, have been obtained.

To obtain the fracture toughness (K_{IC}) of the rock, the regression equation of Gunsallus and Kulhawy (1984) was used, where

$$K_{IC} = 0.0995 I_{S_{50}} + 1.11 \tag{1}$$

All point-load readings were index referenced to a standard 50 mm size (designated $I_{S_{50}}$) following Broch and Franklin (1972, fig. 25). Although the regression equation of Gunsallus and Kulhawy (1984) was not derived from quartz-micaschist it was utilized here because no data are available for this rock type (see Atkinson, 1984). Thus it was possible to obtain a value of K_{IC} which could be incorporated into the model for the breakdown of rock due to freezing produced by Hallet (1983):

$$K_{IC} = \left(\frac{\pi l}{2}\right)^{\frac{1}{2}} (P + \sigma) \tag{2}$$

where Hallet (1983) defines l as the length of a 2-dimensional crack, P as the pressure inside the crack, and σ as the 'applied' normal stress perpendicular to the crack plane (tensile stresses being positive). Hallet's equation could then be rewritten as

$$P = K_{IC}\left(\frac{2}{\pi l}\right)^{\frac{1}{2}} - \sigma \tag{3}$$

so as to obtain the value of P required in the crack for propagation to take place.

Table I. Values of P (MPa m^{-2}) found for cracks of different lengths with a variety of overburdens.

Overburden thickness (m)	*Crack lengths (m)*					
	0.0001	0.0005	0.001	0.005	0.01	0.1
1	105.53	47.13	33.34	14.92	10.56	3.36
2	105.37	47.15	33.36	14.94	10.58	3.38
5	105.46	47.24	33.45	15.02	10.67	3.46
10	105.59	47.37	33.52	15.10	10.80	3.60
50	106.67	48.45	34.66	16.24	11.88	4.68
100	108.03	49.81	36.02	17.60	13.24	6.04
150	109.38	51.16	37.37	18.95	14.59	7.39

As overburden pressure plays a role in determining the value of P, the mean density of the rock was determined and so the compressive pressure was then calculated for various thicknesses of overburden and entered into equation 3. Finally, accepting that internal ice pressure increases with negative temperatures at a rate of 1.14 MPa deg^{-1} (Hallet, 1983) for any saturated rock, it was possible to calculate what temperatures would theoretically be required to generate these pressures and then to compare them with the island temperature records.

Results and Discussion

The mean of 50 K_I values derived from point-load tests via equation 1 was found to be 1.32 MN $m^{-\frac{1}{2}}$. The range of K_I values was low, varying between 1.4891 (for $I_{S_{50}}$ normal to schistosity 3.81 MN/m^2) and 1.2000 (for $I_{S_{50}}$ parallel to schistosity 0.1 MN/m^2) despite taking values both transverse and parallel to schistosity. Therefore, utilizing this value of K_I together with crack lengths (l) of 0.0001, 0.0005, 0.001, 0.005, 0.01 and 0.1 m, together with σ set for overburdens of 1, 2, 5, 10, 50, 100 and 150 m, it was possible to generate a variety of results for P (Table I). The value of σ was derived from the density of the quartz-micaschist ($\bar{x}$ of 20 samples = 2.76 g/cm^3) multiplied by h (the height of a column of uniform cross-section) times acceleration due to gravity.

The values of P shown in Table I are based upon a saturated state. If this presumption is accepted then, knowing the rate of pressure increase with decrease in temperature (1.14 MPa deg^{-1}) it is possible to calculate the theoretical temperatures required to achieve these pressures (Table II). If the rock is not saturated then this model (equation 2) is said not to give adequate results (Hallet, 1983) but, nevertheless, it does provide a hypothesis for comparison against observed temperatures.

The theoretical values shown in Table I are constrained by the degree of saturation of the rock and temperatures from actually achieving the required levels. Results of moisture determinations (Hall, in press *a*) show that most samples were less than 50% saturated. However, of a sample of 47, eight showed > 90% saturation, with four fully saturated, and a further 10 had > 50% moisture content. Those samples with the higher moisture contents all comprised small blocks residing on the ground either in melt rivulets or from under melting snow. Samples taken from the wetter, outer faces of vertical cliffs and from small rock outcrops all showed low (15–30%) moisture values. Consequently, on Signy Island it would appear that the model of Hallet (1983) can only be practically applied to blocks that have been released from the cliffs.

Qualitative observations and the results of engineering geology tests (Hall, in press *b*) all indicate that the cliffs have been subject to very little weathering. Further, the

Table II. The calculated temperatures required to generate P (as given in Table I) for different crack lengths and overburden pressures.

Overburden thickness (m)	*Crack lengths (m)* 0.0001	0.0005	0.001	0.005	0.01	0.1
1	−92.41	−41.34	−29.25	−13.09	−9.26	−2.95
2	−92.43	−41.36	−29.27	−13.11	−9.28	−2.96
5	−92.51	−41.44	−29.34	−13.18	−9.36	−3.04
10	−92.62	−41.55	−29.40	−13.25	−9.47	−3.16
50	−93.57	−42.50	−30.40	−14.25	−10.42	−4.11
100	−94.76	−43.69	−31.60	−15.44	−11.61	−5.30
150	−95.95	−44.88	−16.62	−12.80	−12.80	−6.48

analysis of solutes from inside rocks taken from a cliff face (Hall and others, 1986) also shows that minimal chemical weathering has occurred. What weathering has taken place on the cliffs appears to be associated with large blocks bounded or associated with discontinuities (lithologic, structural, etc.). Certainly these discontinuities may be preferential sites for moisture and, with the larger crack size, require less severe temperatures for crack propagation. However, the model of Hallet may not be applicable to this situation.

What is apparent is that it is the small released blocks and the undersides of overhangs that show the greatest amount of weathering. The overhangs are certainly subject to gravity on the unconstrained lower edge where schistosity is parallel to the ground, and this must abet any weathering agent. The small blocks on the ground, which are most vulnerable to weathering along the planes of schistosity, appear to show the greatest amount of damage. This may be for a number of reasons: greater moisture content, damage during fall, a greater number of effective temperature events, freezing from several faces and the possible additional effects of hydrofracturing (Powers, 1945; Hallet, 1983).

If, as was found to be the case, some of the blocks had a high moisture status then the main control on rock destruction by crack propagation is that of temperature. Temperatures measured at the ground surface show values as low as −25°C, with events between 0°C and −15°C fairly common at sites with little snow cover. However, these are surface temperatures, whereas the crack-wall temperature and the rate of cooling (slower rates being preferential) are the actual controlling factors. Walder and Hallet (1985, p. 342) state 'Clearly, crack growth for a fixed *final temperature* is greater, the lower the cooling rate' but experiments currently in progress on the quartz-micaschist show (Fig. 2) that the rate of fall towards a final temperature is faster for a large-amplitude freeze than for a small-amplitude freeze. This is the case even where the small-amplitude freeze has a faster environmental rate of fall of temperature (4.2 deg h^{-1} as against 3.0 deg h^{-1}). Inside the rock the rate of fall is some three times faster, over the same range, when the final temperature is −6°C as opposed to −3°C and both rates (*c.* 1.6 deg h^{-1} and 0.45 deg h^{-1}) are in ranges suggested to be too fast to be efficient by Walder and Hallet (1985, p. 342).

However, an unconstrained block sitting on the ground is subject to cooling from the surrounding air on five faces, and not one as on a cliff face. In this case, water may be forced ahead of the freezing fronts towards the centre of the block and produce a closed system. The effect of this water pressure could be to cause hydrofracture during freezing and such a situation would be most likely '... during rapid freezing in rocks that are saturated, or nearly so, when slow crack growth due to ice segregation

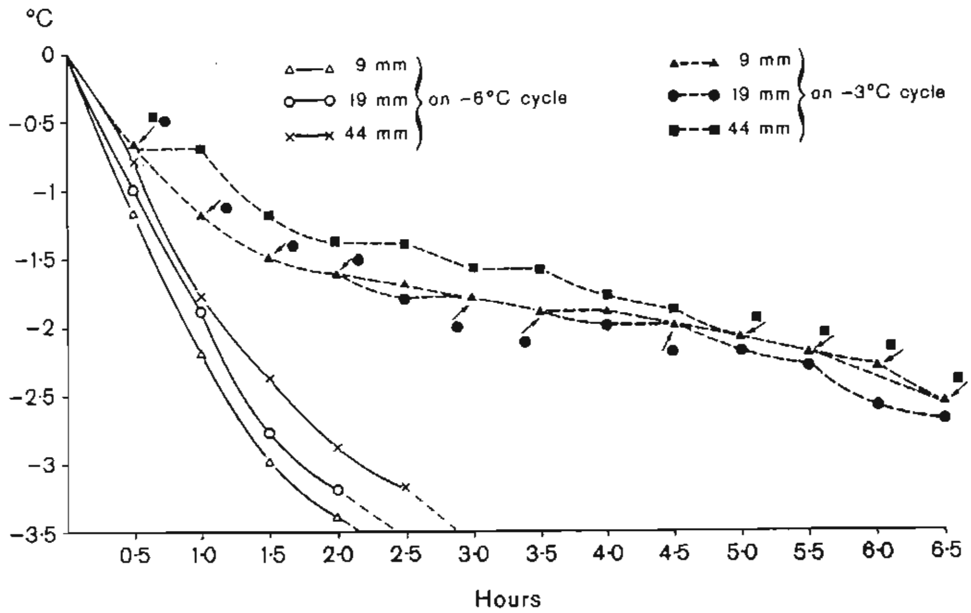

Fig. 2. Plots of temperatures at various depths within a quartz-micaschist block. The environmental rate of fall of temperature for the $-6°C$ cycle was 3 deg h^{-1} and for the $-3°C$ cycle 4.2 h^{-1}.

is unable to accommodate the volumetric increase associated with the H_2O-phase transition' (Walder and Hallet, p. 343). Thus, whilst slow rates of fall of temperature are required for the cliffs, but are not apparently found above a depth of *c*.0.6 m where the rock is almost dry (Hall, submitted), it is the faster rate which would be most destructive for the loose blocks. It is suggested that hydrofracturing is the cause of rock destruction for the loose block, when saturated (or nearly so) and experiencing a relatively rapid rate of fall of temperature inside the rock from several faces, thereby producing a closed system. This hypothesis, based on the utilization of the stress intensity factor and its application to cliff-faces and talus blocks, appears justified both upon the basis of field observation and on empirically derived data.

The above discussion appears to fit the field evidence and, as such, offers an explanation for the pattern of rock breakdown found. However, the role of other mechanisms such as salt or insolation weathering in aiding breakdown cannot be ignored. Further field and laboratory studies are currently in progress and determine the role of these other agents and how they interact with freeze–thaw. In addition, the role of events at levels below apparent threshold levels must be considered, for as Mura (1981, p. 268) states 'It is well known that materials fail under repeated (cyclic) loading and unloading at stresses smaller than the static fracture stresses. The magnitude of the stress required to produce failure decreases as the number of cycles increases.' Thus, whilst the stress intensity factor and its application in models is a great step forward in our theoretical understanding of 'freeze–thaw', great care must be taken in its application to a field situation. Indeed, Walder and Hallet (1985) have pointed out that thaw is not a prerequisite for continued crack growth.

Conclusion

The use of the stress intensity factor (K_{IC}) via such models as that suggested by Hallet (1983) appears to offer a good theoretical framework for more sophisticated investigation of freeze–thaw weathering. The greatest difficulty in the application of the model to the real world is its requirement for a saturated state. To date the utilization of the stress intensity factor has been towards explaining open system, bedrock breakdown (Walder and Hallet, 1985) where abundant water supply is said to be 'quite plausible'. However evidence from the maritime Antarctic would suggest that this saturated condition is definitely not the case in bedrock and so its applicability for bedrock may be limited in this environment. Nevertheless, it does offer a means of explaining why small, unconstrained blocks do show breakdown when cliffs do not and, importantly, it offers a means of deriving important theoretical data against which to test field data. At present, the testing and application of both K_{IC} and the models utilizing it is limited by insufficient field data on temperature, moisture and moisture solutes. Once we obtain extensive information on these controls it will be possible to better determine the applicability of the models to the real world.

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THE PHYSICAL PROPERTIES OF QUARTZ–MICASCHIST AND THEIR APPLICATION TO FREEZE–THAW WEATHERING STUDIES IN THE MARITIME ANTARCTIC

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ABSTRACT

As part of a study on freeze–thaw weathering in the maritime Antarctic an investigation was made of the physical properties of the local rock. Tests were made of point-load compressive strength, Schmidt hammer *in situ* rock strength, moisture content, indentor resistance and the size range of weathering products. The resulting data were used to consider the form of freeze–thaw weathering operative in the field and its relationship to laboratory simulations. A distinct difference between 'massive rock' and 'intact rock' is observed in the field. It is suggested that future studies should generate a greater database pertaining to rock properties as it is an invaluable aid in the study of mechanical weathering.

KEY WORDS Maritime Antarctic Freeze–thaw Hydrofracture Engineering techniques

INTRODUCTION

Part of British Antarctic Survey's Fellfield Programme in the maritime Antarctic involves a field and laboratory investigation of mechanical weathering of quartz–micaschist (Walton and Hall, submitted). As part of this study it was felt that a better understanding of how and why the rocks were weathering could be obtained if some knowledge of their physical properties was established. To date, many weathering studies have neglected the properties of the rock(s) under investigation and this gap in data hinders detailed understanding and analysis of results. The aim of this present project was to establish a database of information relating to both the rocks themselves and the conditions which they were experiencing; this need became even greater when it was found that there was a dearth of published information pertaining to quartz–micaschist. Having obtained the data on rock properties these could then be applied to the hypotheses and findings relating to freeze–thaw.

Data on the engineering properties of most rocks are available, but geomorphologists involved in weathering studies have, as shown by McGreevy and Whalley (1984), been reticent to use such information. In the context of this present study, there is information available on the properties of schist and the effects of schistosity (e.g. Bell, 1983; Brown *et al.*, 1980; Deere and Miller, 1966; and Goodman, 1980 being recent examples) but a number of early simulation experiments which used schists (e.g. Brockie, 1972; Wiman, 1963; Martini, 1967) neglected to consider the properties of the rocks under test. More recent studies (Lautridou, 1982; Mottershead, 1982; Fahey, 1983) have detailed the grain size analysis of the weathering products whilst Fahey (1983) has also given data on water absorption and retention for schist. Finally, Trenhaile and Mercan (1984) obtained, for a coastal environment, actual field saturation percentages of schist which, dependent upon sample position with respect to the tide, were seen to vary between 25·6 and 93·0 per cent.

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However, as informative as the results of the above studies are, they refer to schist and not quartz–micaschist for which there appears to be a distinct lack of information. Wolff (1981, Tables 4.3 and 4.6) gives data on the porosity (46·9 per cent) and permeability ($3{\cdot}8 \times 10^{-6}$ m s^{-1}) of weathered quartz–micaschist, but no data on unweathered rock could be found. Hall (in press) showed that variations in quartz content and the form of the quartz (i.e. pods, thick or thin bands, etc) had an effect on the properties of that rock with respect to moisture content. Some of those data will be briefly repeated here together with new information on compressive strength, indirectly derived estimation of tensile strength, rock mass strength, indentor tests and weathering product sizes. This data will then be used to evaluate the weathering of quartz–micaschist, but the resulting findings will then be considered in the context of maritime Antarctic weathering in general.

STUDY AREA

All of the rock samples were collected from Signy Island (60°43'S. 45°38'W), one of the smaller islands in the South Orkney group (Figure 1). Geologically the island comprises metamorphosed sediments, mainly

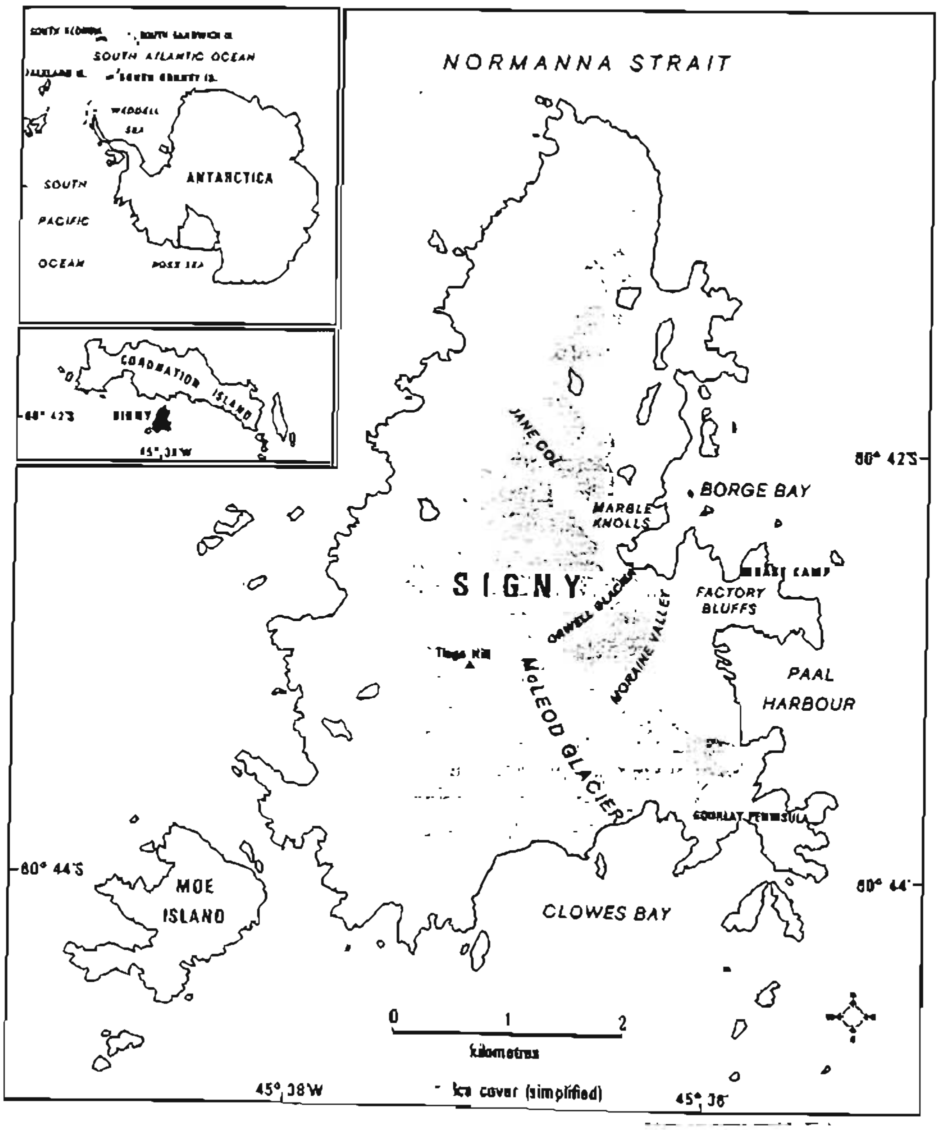


Figure 1. Location map of Signy Island

quartz–micaschist with smaller areas of amphibolites, marbles and quartzites (Mathews and Maling, 1967). The island is small, being only 8·0 km north to south and 4·8 km east to west, with an area of 19·94 km^2 and rising to a height of 279 m. Approximately one-third of the island is ice covered and much of the rest is subject to long-term snow lay, and it is an area of discontinuous permafrost.

TECHNIQUES

Samples of rock were collected from all parts of the island and included both loose blocks obtained from screes and moraines as well as material derived directly from rock outcrops. An attempt was made to collect samples representative of the observed variations in quartz content and form. Upon collection, samples were weighed so that, after drying, an evaluation of their field moisture content could be obtained. Samples were then subject to the irregular lump point-load strength test following the procedures set out by Brock and Franklin (1972). As quartz–micaschist is anisotropic, tests were undertaken both parallel to, and normal to, the direction of schistosity. Results are subject, for a number of reasons (see Broch and Franklin, 1972) to a percentage error but an attempt was made to minimize this by conducting a number of tests and taking their mean value. According to Deere and Miller (1966) there is a strong, positive correlation between compressive strength and point-load tensile strength with, according to Szlavin (1974, Figure 4), tensile strength being approximately one-fifth of the compressive strength. Using a National Coal Board indentor, tests were undertaken on rock flakes, both collected in the field and derived from point-load tests, to gain an evaluation of their hardness and degree of weathering at a microscale.

In the field, estimation of the compressive strength of the *in situ* rock was obtained by means of a Schmidt hammer. Although the instrument actually measures hammer rebound due to rock hardness, this value is directly correlated with compressive strength (Day, 1980). However, the use of the Schmidt hammer was limited by the nature of the quartz–micaschist, for its uneven surface texture has an effect on the readings (Williams and Robinson, 1983). In addition, use of the hammer parallel to schistosity was frequently inhibited by the destructive collapse of the mica laminae under hammer impact, thereby lessening the reading considerably. Despite these problems, a limited number of readings were obtained which were able to be used and integrated into a Rock Mass Strength (Selby, 1980) study.

Samples of weathered material, collected from a variety of locations in the field, were analysed for their grain size distribution. As the size range of the weathering products is a function of the rock concerned and the processes that it has been subject to, the data allowed for comparison with the results of other workers and with the material derived from simulation experiments. Finally, in the laboratory, tests were undertaken on the rock samples, following the procedure of Cooke (1979), to determine their porosity, microporosity, water absorption capacity and saturation coefficient.

RESULTS

Blocks of quartz–micaschist were subject to the point-load strength test with, as the rocks are anisotropic, tests carried out both parallel and transverse to schistosity (Table I). Using results from the same rock sample it was found that, on average, the rock was 70·2 per cent ($s = 26{\cdot}17$) weaker when the load was applied parallel, rather than normal, to schistosity. Following the strength designations of Broch and Franklin (1972) the results of the tests are presented in Table II. It can be seen that when tested normal to schistosity the bulk of the samples (74·3 per cent) are in the compressive strength range 'High' to 'Very high' whilst only 17·4 per cent of those tested parallel to schistosity managed to achieve the rating 'High'. In all tests moisture contents were determined and found not to be sufficient to diminish point-load strength according to the findings of Broch (1979).

Deere and Miller (1966) show that there is a very strong positive correlation (+0·917 to +0·947) between uniaxial compressive strength and tensile strength with, according to Szlavin (1974) tensile strength being about 20 per cent of the compressive strength. Thus, the tensile strength of the rock is dominantly 'Low' to 'Medium' when tested normal to schistosity and 'Very Low' to 'Low' when stressed parallel to schistosity (Table II).

Data on the *in situ* compressive strength of the rock was obtained via the Schmidt hammer. Despite its limited practical use in this present situation a number of readings were obtained, the analysis of which showed

Table I. Some results of point-load tests

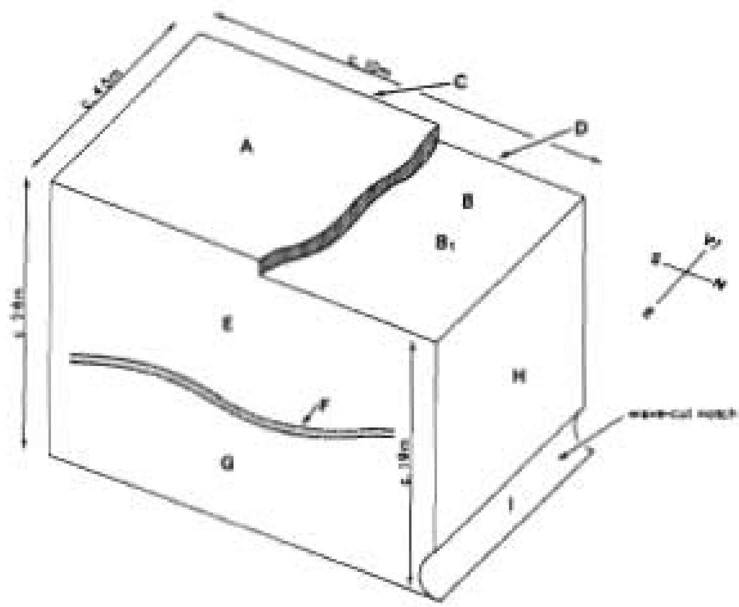
Rock	Direction	MN/m²	Moisture content (%)
Quartz–micaschist		1·31	
" "		0·15	
" "		3·17	
" "		0·101	= 0·32
" "		3·84	(0·5 g)
" "		5·06	
" "		1·71	
Quartz–micaschist		1·74	= 0·32
" "		0·05	(0·4 g)
Garnet rich Qms		0·79	
" " "		1·10	
" " "		1·15	0·17
Quartz–micaschist		1·55	
" "		0·48	0·09
Quartz–micaschist		0·25	0·22
" "		1·14	
Rich in quartz Qms		1·2	0·23
" " "		0·98	
Quartz–micaschist		0·94	
" "		4·01	0·23
Qms—rich in quartz		0·39	
" " "		0·94	
" " "		0·043	
Qms		4·98	0·15
Qms—2·5 mm laminae + garnets		1·08	
" " " "		3·61	
" " " "		2·73	
Gently folded, many small bands		0·25	
" " "		2·25	0·13
" " "		0·55	
" " "		0·19	
Folded thin bands		2·05	
" " "		0·41	0·10

Table II. Compressive and tensile rock strength for quartz-micaschist

Strength designation	Very low	Low	Medium	High	Very high
	Percentage occurrence				
Strength range (MN/m^2)	0·03–0·1*	0·1–0·3	0·3–1·0	1·0–3·0	3·0–10
Compressive strength					
Parallel to schistosity	13·4	30·4	39·1	17·4	0
Normal to schistosity	0	2·9	22·9	54·3	20·0
Tensile strength					
Parallel to schistosity	52·0	40·0	8·0	0	0
Normal to schistosity	9·3	48·8	39·5	2·3	0

* These designations from Broch and Franklin (1972) and in this instance no values fell on boundaries.

that large resultant variations, due to the quartz inclusions, could be found within small areas. An example of hammer rebound variability is shown in Figure 2 where the readings were obtained from a large block on the beach at Factory Cove. It can be seen that the average hammer readings for micaschist are lower than those of quartz-micaschist which are, in turn, less than those of quartz. Thus, care must be taken in the use of the hammer rebound values as readings taken directly on quartz would indicate strength ratings greater than that of the rock as a whole. Whilst the role of quartz inclusions in increasing strength values could be seen, no significant correlation between hammer value and the form of the quartz inclusion was found. Hammer



A { micaschist = 38·3 (n=11); quartz inclusions = 63·6 (n=13); average = 44·4 (n=24)
B micaschist = 39·4 (n=20)
B₁ quartz inclusions = 51·6 (n=13)
C quartz-micaschist = 42·9 (n=20)
D quartz-micaschist = 54·9 (n=20)
E quartz-micaschist = 50·6 (n=22)
F quartz band = 61·6 (n=20)
G { micaschist = 53·6 (n=16); quartz inclusions = 66·9 (n=13); average = 58·4 (n=29)
H quartz-micaschist = 54·6 (n=22)
I quartz-micaschist = 53·1 (n=25)

Figure 2. Details of Schmidt hammer readings from a large block at Factory Cove

readings for quartz–micaschist at other localities around the island indicated mean values between 52·5 and 57·7 taken normal to schistosity and 17·4 to 44·4 parallel to schistosity; details of results are given in Table III.

Although the hardness of the rock is itself of little relevance to weathering studies, the use of an indentor allowed a semiquantitative indication of the degree and depth of weathering at a microscale. Simplistically, those areas which had a brown, rusty colouration were considered to show 'weathering' whilst those of a silver–grey colour were not. Although this test is more applicable to the assessment of the chemical deterioration, it was nevertheless felt to be a useful addition insofar as it was at the microscale and, in a sense, the resistance of the flake to the pressure of the indentor needle gave some idea of strength at this level. Graphs of the movement of the needle ('Turn') versus amount of penetration are given in Figure 3 and they show that, generally speaking, greater penetration was achieved for brown areas. However, some surficially grey spots indicated weaker, more weathered zones beneath the surface and some brown zones showed weathering to only a shallow depth (*c.* 0·2 mm) and then they had a hardness more similar to the grey sites. For a needle movement up to 0·6 mm, all the brown-coloured positions showed greater penetration than did the grey. One brown (7.5 YR 5/8) spot on a flake from a block of highly-weathered (i.e. soft and friable) quartz–micaschist, which had a low quartz content, showed a 0·885 mm penetration for a turn of 1·4 mm and an 85 per cent greater penetration when compared with a grey site for the same amount of turn.

Details of the findings on rock moisture content, rock porosity, water absorption capacity and saturation coefficient have already been presented (Hall, in press) but as much of this information is pertinent to this discussion a brief summary will be given. It was found that moisture content varied with the form and amount

Table III. Details of Schmidt hammer tests

Orientation	Rock	Hammer reading $\bar{x}$	s	n	Details	Code in Figure 2
↓	Micaschist	38·3	7·4	11	Finely schistose	A
↓	Quartz	62·6	3·5	13	Nodules of quartz	A
↓	Micaschist	39·4	8·0	20	'Slabs' of schist	B
↓	Quartz	51·5	8·9	13	Slightly fractured bands of quartz	B_1
→	Qms*	42·9	9·4	20	Highly foliated, weathered	C
↓	Qms	54·9	5·9	20	Schistosity dipping at 38°	D
→	Qms	50·6	9·3	22	Shows signs of weathering	E
→	Quartz	61·5	3·1	20	Quartz bands	F
→	Qms	58·4	4·5	29	Thin quartz bands stand proud	G
→	Micaschist	52·6	6·4	16	Finely laminated section	G
→	Quartz	66·9	2·6	13	Quartz bands	G
→	Qms	54·6	8·1	22	Quartz bands stand proud; signs of weathering	H
→	Qms	53·1	5·4	25	Very fine schistosity + fine quartz bands	I
→	Qms	17·4	4·2	14	Very soft and highly weathered	–
↓	Qms	57·7	7·3	22	Very fine foliation + quartz bands	–
→	Qms	44·4	11·9	16	Very fine foliation + quartz bands	–
↓	Qms	52·5	6·7	17	Coarse foliation + quartz bands	–
→	Qms	47·8	8·0	19	Coarse foliation + quartz bands	–

* Qms = Quartz–micaschist.

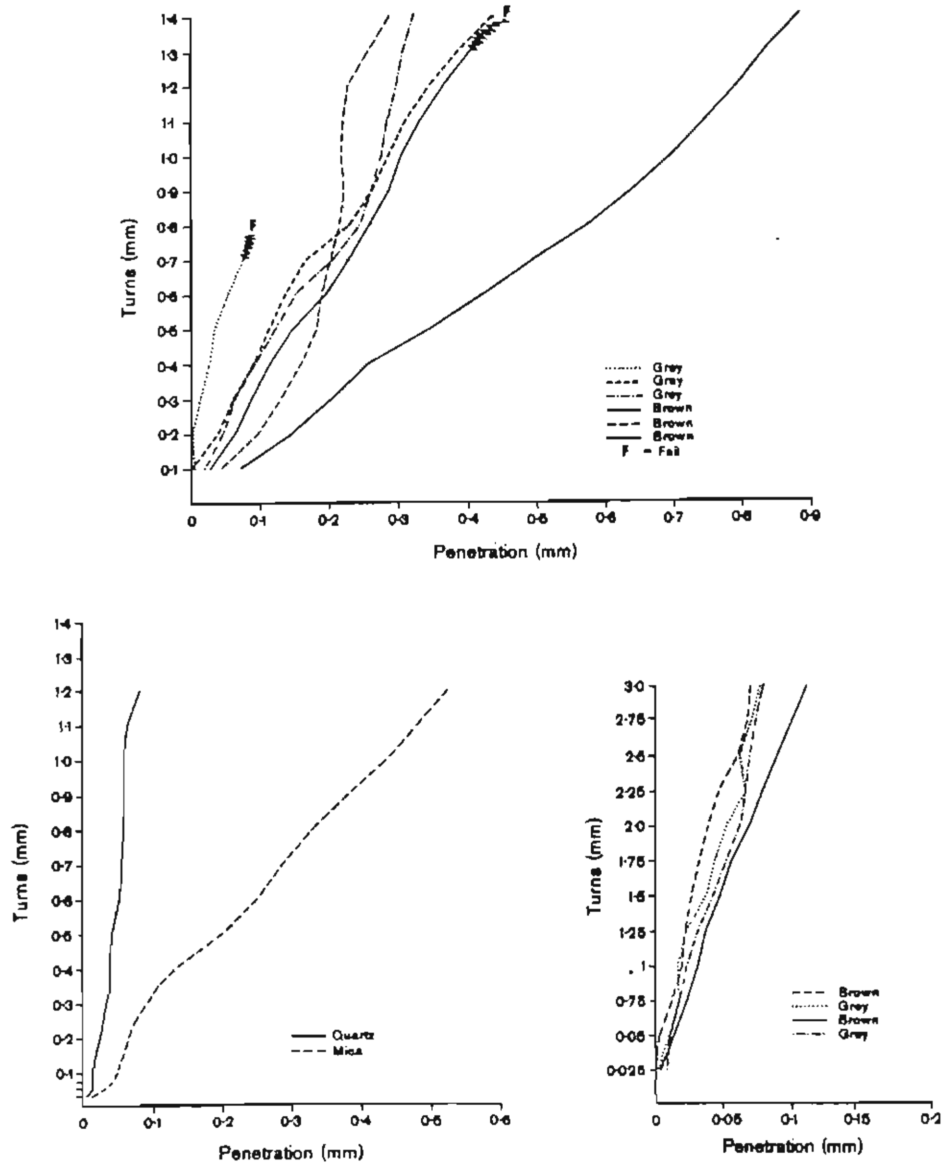


Figure 3. Results of indentor tests

of quartz inclusions, with the quartz-rich end of the spectrum having lower water contents ($\bar{x}$ = 0·30 to 0·32 per cent) than the micaceous end ($\bar{x}$ = 1·17 per cent). Although actual water amounts appeared low (range 0·12 to 1·27 per cent), the degree of saturation varied from fully saturated through to only 14·3 per cent saturation. Rock porosity ranged between 0·55 and 4·65 per cent with water absorption capacities (percentage) from 0·41 and 3·45. Saturation coefficients appeared to reflect the observed amount of field moisture for the bulk (63 per cent) of samples and to be a good indicator of potential frost susceptibility. The biggest problem was the lack of a method to determine the distribution of the water within the rock; to tell whether it was disseminated or within a specific zone.

Finally, analysis of the size range of weathered material collected in the field was undertaken as this is the end product of the weathering processes. Although there may have been some additional comminution of material resulting from abrasion during sieving, it is thought that the results still closely indicate the true size range. The results of the size analyses are shown in Figure 4 in which it can be seen that the coarse fraction (*c.* 70 per cent > granule) predominates. Whilst the coarse component may well reflect mechanical weathering, the result of samples (Nos. 111a and 111b) from a site with a strong chemical weathering component showed a greater fine fraction. There appeared to be a small peak of material at the $+3\phi$ size grade which may reflect the lower size range of quartz particles.

DISCUSSION

Bell (1983, p. 88) states "The type of weathering which predominates in a region is largely dependent upon climate, which also affects the rate at which weathering proceeds. The latter is also influenced by the rock mass concerned, which in turn depends on its mineral composition, texture and porosity, and the incidence of discontinuities within it'. In geomorphological studies the role of climate has been recognized and attempts to monitor significant parameters, although to a lesser extent than one would think (McGreevy, 1981), have been undertaken but the properties of the rock mass have, by and large, been ignored (McGreevy and Whalley, 1984). In this study an endeavour has been made to take cognisance of some of the physical properties of the rock concerned, and an attempt will now be made to consider them in the context of mechanical weathering of quartz–micaschist in the maritime Antarctic. The findings will then be considered in the wider connotation of Antarctic mechanical weathering in general.

The Schmidt hammer results (Table III) give a good reflection of the strength of the massive rock with, on the engineering-grade classification of Day (1980, Table 4), the bulk of the quartz–micaschist falling into the range of 'slightly weathered' to 'moderately weathered'. The qualitative descriptions for these groupings, as given in Fookes *et al.* (1971, Table 2), perfectly fit the field observations with an extensively weathered site showing a low average hammer reading ($\bar{x} = 17{\cdot}4$) which equates to 'highly weathered' on the engineering-grade classification. That hammer readings parallel to schistosity did not show significantly lower readings than those normal to schistosity is attributed to, in most cases, the limited amount of weathering and the quartz inclusions which greatly increased hammer rebound. Whilst the hammer values for micaschist in Table III are fairly similar to those of Bell (1983) and Deere and Miller (1966) those of the quartz–micaschist are significantly higher, reflecting the presence of the hard quartz inclusions. Thus, although there is variety in the

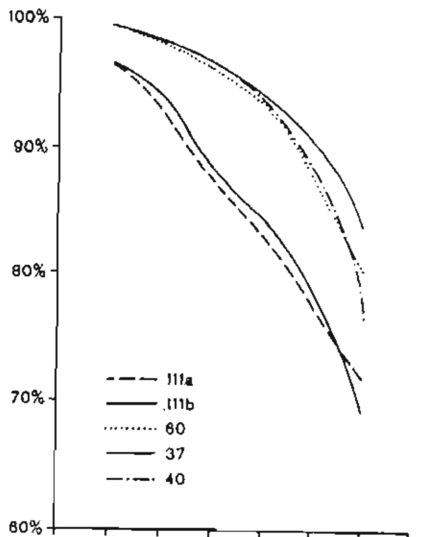


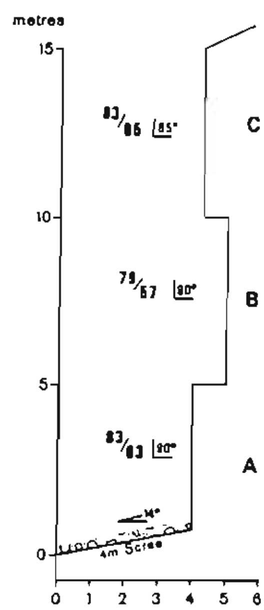
Figure 4. Results of grain size analyses for five samples

readings, and care must be taken in reading the quartz only (too high an average) or collapsing schist edges (too low an average), the hammer readings indicate that the cliffs have not been subjected, in most cases, to extensive weathering. The hammer readings substantiate the point load test results (see below) and so may be used as a rapid guide to the generalized state of rock; a point made by Day (1980, p. 79; 1981, p. 170).

An attempt was made at investigating the Rock Mass Strength (RMS), following the procedure of Selby (1980), incorporating Schmidt hammer values as indicators of intact rock strength. Results of this (Figure 5) gave an RMS score of 79–83 for the cliffs at Factory Bluffs when values for joints (spacing, orientation and width) were given for the major discontinuities and values of 57–63 when values for the schistosity itself were used. These values then rate the cliffs as 'strong' and 'moderately strong' respectively (Selby, 1980, Table 5). When the RMS rating is plotted against the angle of slope they sit almost exactly on the regression line of Selby (1980, Fig. 4), thereby fitting with the test values found for other locations in Antarctica and New Zealand and so indicating the (p. 48) '. . . appropriateness of the classification parameter and the values assigned to them'.

At the microlevel, the indentor tests gave a quantitative representation of hardness and an indication of rock strength. Unfortunately, the National Coal Board indentor that was used, which has a needle-point indentor and does not constrain the rock flakes, does not lend itself to direct comparison with the results of Cook *et al.* (1984) who used a large indentor on confined samples. However, it can be seen (Figure 3), as would be expected, that the quartz is much stronger than the mica with the needle of the indentor penetrating/flexing the mica up to 85 per cent more than the quartz. Direct comparison between the 'brown' (weathering rind?) and 'grey' sites does show the former to be initially softer, but then there is variety with some grey locations being softer beneath the surface whilst some that were surficially brown were harder than the grey at depth. However, this information pertains largely to chemical weathering at the microscale and does not indicate any significant distinctions with respect to the overall mechanical strength of the massive rock. What the indentor record may indicate, are the positions where small-scale surficial weathering might first take place as not only is the rock weaker but also the chemical weathering may have produced a greater water absorption coefficient thereby improving the potential for frost action. It is hoped to test these ideas in order to evaluate the potential of the indentor as a tool in weathering studies.

Particle size analysis of weathered material collected from on or below quartz–micaschist outcrops (Figure



DETAILS

	A	B	C
Strength	14	10	14
Weathering	9	9	9
Joint spacing	25 (10)	25 (8)	25 (12)
Joint orientation	20 (14)	20 (14)	20 (14)
Joint width	5 (6)	5 (6)	5 (6)
Joint continuity	5	5	5
Groundwater flow	5	5	5
TOTAL	83 (63)	79 (57)	83 (65)

( ) – Schistosity value
RMS values according to Selby (1980)

Figure 5. Details of the RMS investigation at Factory Bluffs

4) clearly showed a preponderance of coarse material (> granule), as might be expected from an environment where mechanical weathering is dominant. The samples (111a and b) collected from a ledge on a cliff face at Factory Bluffs which, qualitatively, appeared to be subject to a greater amount of chemical weathering (the rock is 'soft', friable and severely discoloured) showed a marginally greater (c. 4 to 14 per cent) amount of finer material than the other samples. The percentage of products finer than $+4\phi$ in 111a and b was also greater (c. 2 per cent), indicating a further breakdown as a probable result of chemical action. The large amount of coarse material in all samples must be, in part, attributed to the quartz inclusions which are not readily or rapidly diminished in size by the mechanical weathering. The small peak at the $+3\phi$ level found in some samples may represent the lower level (very fine sand) of quartz communication.

In simulation studies Lautridou and Ozouf (1982) found that schists usually produce very little fine material except in the case of the freezing of chemically weathered schists; a situation somewhat analogous to the Signy samples 111a and b except for the incorporation of the extra, coarse-sized quartz particles. Overall, the results of this current study were fairly similar to those of Lautridou (1982, Fig. 2) for a freeze–thaw simulation of schist, with freeze temperatures of $-8°C$, for 500 cycles. The application, though, of Lautridou's 500 cycles to the field situation of a cliff-face could translate to an extensive period of time when considering local moisture contents (Hall, in press) and temperatures (Walton, 1982). For, on Signy, even assuming a preferential location with adequate moisture, although temperatures do go well below $-8°C$ they rarely do this on other than an annual freeze–thaw cycle. Therefore, ignoring the potential effects of negative temperature oscillations about a subzero freeze-point and stress variations of negative temperature cycling (Hall, 1980) which are currently being investigated, comparable breakdown would probably be on a scale of greater than one cycle equating to one year. In other words, weathering would be relatively slow; an attribute indicated above from other lines of evidence.

Moisture is, particularly in the context of freeze–thaw weathering, a prerequisite because, even if temperatures are conducive, '. . . in the absence of water no frost damage occurs' (Litvan, 1974, p 21). However, data on rock field moisture contents are extremely rare (Trenhaile and Mercan, 1984; Hall, in press) and, as a consequence, many simulation studies have little resemblance to real world situations (Trenhaile and Mercan, 1984; McGreevy and Whalley, in press). Thus, in a study of this kind knowledge of moisture levels and their variability is of prime importance, for without this information interpretation of processes and rates is almost impossible.

It was found (Hall, in press) that the moisture content of cliff-face rocks was low whilst that of loose blocks residing on the ground was much higher, with some even being fully saturated. The low degree of saturation for the cliffs is thought to be due to four main factors: (1) the water absorption coefficient for the rock is low, (2) the schistosity is normal to the direction of water movement which thus does not increase the ability of the rock to take-up water, (3) water moving down the steep cliff-face has only a short reside time, and (4) available moisture is limited (precipitation is c. 0·4 m y^{-1}). Despite their low absorption coefficients, many of the loose blocks that reside in the soil, or in areas along which water moves and ponds, are able to take up more moisture as they are in contact with the water for a longer time, they have a larger contact area, and water may be able to enter directly along the lines of schistosity.

Finally, the compressive strength values (Table II), as shown by the point-load test, indicated a high strength range when tested normal to schistosity. As all the cliffs examined had schistosity normal to the cliff-face, this explains the ability of the rock to maintain high, steep faces. However, it is the much weaker tensile strength parallel to schistosity which is exploited by crack tip propogation in forcing the laminae apart during mechanical weathering.

With respect to destruction by mechanical weathering processes it is important to know both the forces that the mechanisms can exert and the strength of the rock which is available to resist them, and the factors influencing these interactions. Hallet (1983) has shown that the theoretical stress intensity factor (K_t) for fissile metamorphic rocks is low and, as a consequence, if sufficient moisture is available then only small pressures are required to facilitate crack propogation under freezing conditions. As the tensile strength of the rock is very low, when stressed normal to the schistosity, then neither the forces available to propogate the cracks nor the forces available to resist them are high and so failure along lines of schistosity is, subject to the controls of temperature and moisture, theoretically relatively 'easy'. According to the equation of Hallet (1983) the pressures required for crack extension would be low and that means temperatures need not go severely negative to achieve sufficient stress to cause failure; the relationship of pressure to temperature being 1·14 MPa $°C^{-1}$. However, compressive overburden pressure increases the force required to promote crack tip extension (Hallet, 1983; Hall, submitted) where, as in the case here, schistosity is normal to the cliff-face. The effect of overburden may, in fact, not appear unduly great with, for example, temperatures required for crack extension being depressed by 1·2°C and 3·5°C for overburdens of 50 m and 150 m respectively (Hall, submitted). However, these are crack wall temperatures and, subject to the availability of moisture, may be sufficient to exercise the difference between an effective or non-effective event, particularly with respect to the smaller, more rapid subzero oscillations. In addition, subjective observation and available data tend to indicate that it is the base of the cliff that may be wetter, but there the overburden is greatest and so cooler temperatures are required for crack growth compared to the higher, but drier, locations. It is the loose blocks, which have no overburden effects and appear to have more frequent higher moisture contents, that will experience the greatest number of effective events.

However, the loose blocks may be subject to a number of deleterious effects that the cliff does not experience. The blocks may be fractured or damaged by the fall itself, estimation of which is difficult to assess. The damage may comprise extension of crack or schistosity length, and this means according to the equation of Hallet (1983), that temperatures need not be so severe to produce crack propogation (Hall, submitted, Table II). Perhaps more significant though, is that the loose blocks, if they are able to obtain moisture, may be subject to hydrofracturing as they have simultaneous freezing-plane penetration from several surfaces (Walder and Hallet, 1985; Hall, submitted).

It is suggested that the combination of greater moisture availability, the potential for hydrofracturing, and the lower temperatures required for crack tip extension explain why many of the loose blocks show greater breakdown than do the cliff faces. Almost certainly it is moisture which is the major control, for small blocks in positions of poor moisture acquisition (i.e. on the top of a moraine, surface of a large scree slope, etc.) also rarely show any sign of breakdown. Conversely, some blocks which are found in areas of good water supply, even sometimes standing in pools, may show no apparent sign of breakdown. When tested, these rocks are found to have very poor absorption coefficients and so, despite its availability, the rocks had low interstitial water contents. Thus, weathering is a result of the complex interrelationship of temperature, moisture and rock properties; all three need to be favourable for damage to occur.

An implication of all the foregoing is that, with respect to Signy Island, the bulk of the breakdown to finer material takes place after large blocks have been detached. These large blocks are broken-free along major boundaries (e.g. joints, lithologic, faults, etc) where there is adequate moisture available. The released block and the pieces broken off during its fall are then more readily broken down. The loss of the block from the cliff now means that there are overhangs/notches within which, on the floors and the roofs, the rock is unconstrained. At these positions the temperatures required for effective frost action are not so low and the splitting away of material is aided, particularly on the roof, by gravity and dilatational forces. It was noticeable in the field that at such locations slabs of material from *c.* 1 mm to 0·2 m in thickness were readily pulled away and that numerous loose pieces were scattered around.

In general, weathering studies require detailed information on temperatures, moisture content and rock properties in both a temporal and spatial context. Limited sampling, in either time or space, may give erroneous results and hence incorrect process interpretation. In fact it is likely that, for any given location, processes may change through time, both seasonally as the controls of temperature and moisture alter, and on a longer timescale as rock properties change due to weathering effects. Information on these three main controls are extremely rare in geomorphological studies to date with even the most ubiquitous, that of temperature, being limited to mainly surface data thereby negating knowledge of crack wall temperatures and rate of fall of temperature with depth. Thus, the utilization and application of theoretical models such as that of Hallet (1983) are limited by our lack of field data; a point overlooked by Walder and Hallet (1985, p. 342) when they stated '. . . the important conditions of . . . abundant water supply are quite plausible . . .'. At the moment it would appear that theory has outstripped field observation and until adequate data are obtained from the real world, simulations and theoretical modelling are going to be of limited value. Certainly the data obtained in this study on all of the considered controls indicate a complex and varying situation.

CONCLUSIONS

The consideration of the engineering properties of rocks offers an important input into mechanical weathering studies and the utilization of rock fracture mechanics. The relatively simple point-load test allows for information on compressive strength, tensile strength and the fracture toughness index to be obtained, thereby producing data describing the rock properties exerting an influence on crack propagation. With point-load data as a background it may be possible to utilize the more convenient Schmidt hammer to more readily obtain comparable information within the study area. The indentor test has potential as a useful tool in weathering studies far greater than was exploited here and more research on this is necessary. The monitoring of rock water content, together with its chemistry, and the knowledge of the rock properties controlling moisture acquisition are of central importance to mechanical weathering studies and no study is really viable without such data. This database of information on the rock and its properties allows for a far greater insight into the operative weathering processes.

Results from this study indicate that there is a significant difference between the responses of 'massive rock' and 'intact rock' to mechanical weathering. The cliffs (massive rock) appear to be affected primarily along major discontinuities whilst loose blocks are being exploited along the lines of schistosity. Explanation for these differences is given by a combination of moisture contents, required temperatures and the effects of overburden as described by the model of Hallet (1983). The relatively high compressive strength of the rock (with load normal to schistosity) explains the ability of the cliff to maintain steep faces. At the same time, the overburden pressure involved would increase the tensile stress needed, and hence require low temperatures, for the effects of freezing to promote crack extension along the planes of schistosity. Conversely, in unconfined blocks the tensile stress for failure is lower and, assuming adequate moisture, very low temperatures are not required for crack propagation. A major controlling factor is water content and this, in turn, is controlled by both its availability and the absorption coefficient of the rock.

The distinction between massive and intact rock found in the field has serious implications for laboratory simulations. Weathering rates found for blocks, even relatively large ones, may not be directly applicable to the same material in the confined situation of a cliff. The cliff is an open system with unidirectional freezing whilst the loose block is, due to omnidirectional freezing plane penetration, a closed system and hence may be experiencing an entirely different form of 'freeze–thaw' weathering. Thus great care must be taken with respect to experimentation and its relationship to the field situation.

It would appear that the use of engineering geology techniques to the investigation of mechanical weathering processes is a profitable one. The extended database helps not only the understanding of the field situation but also offers a valuable background for the development of realistic laboratory simulations. The obtaining of such data pertaining to the properties of the rocks also allows for the application of rock fracture mechanics techniques, and hence a new insight into the mechanisms and rates of rock breakdown. Without such a database pertaining to the properties of the rock and its environmental situation, then simulations have extremely limited application to the field, hypotheses are based on a great deal of conjecture, and the field of study cannot really advance as theory cannot be justifiably applied. There is currently a great wealth of techniques at our disposal and these need to be utilized in mechanical weathering studies, in all environments, so that our explanations and ideas can rest on a more sound background of empirical information.

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Mechanical Weathering Rates on Signy Island, Maritime Antarctic

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ABSTRACT

By means of re-evaluating a number of properties of rock tablets left in the field for varying time periods, an estimation of rock breakdown rates is attained. From data obtained during the last five years, it would appear that rates are very slow, only of the order of 2% mass loss per 100 years. These rates refer to omnidirectionally frozen, relatively wet samples and, on the basis of laboratory simulation results, are over 50 times greater than for unidirectionally frozen bedrock. It is suggested that mechanical weathering rates in the maritime Antarctic are very slow.

RESUME

En réévaluant différentes propriétés de tablettes de roche laissées sur le terrain pour des périodes de temps variées, une estimation des vitesses de fragmentation de ces roches a été obtenue. A partir des données recueillies pendant les cinq dernières années, il apparait que les vitesses de désagrégation sont très lentes et atteignent seulement par siècle une perte de l'ordre de 2% de la masse. Ces vitesses se rapportent à des gels dans toutes les directions d'échantillons relativement humides. Elles sont cependant 50 fois supérieures à ce qui se produit lorsque la roche gèle dans une seule direction. Ces données suggèrent que la désagrégation mécanique des roches dans l'Antarctique maritime est très lente.

KEY WORDS: Maritime Antarctic Mechanical weathering Freeze-thaw Rock tablets Weathering rates

INTRODUCTION

One of the many questions concerning the nature and efficacy of freeze-thaw weathering regards the rate of operation. In recent years there have been a number of attempts to quantify the rate of frost shattering in the field (e.g. Fahey and Lefebure, 1988). Other studies have investigated temporal and spatial variations in bedrock shattering but have not monitored rates (e.g. Gardner, 1983). The most detailed undertaking to date (Matsuoka, 1990) monitored the rate of frost weathering over a five-year period and related the findings to the controlling factors of rock temperature, rock moisture content and the physical and strength properties of the rock. A combination of these field-derived data together with empirical relationships obtained from laboratory simulations (Matsuoka, 1988, 1989) allowed Matsuoka (1990) to suggest a predictive model of the frost shattering rate. However, these studies, like those of other researchers (see Matsuoka, 1990, for a review) have all been with respect to relatively dynamic environments characterized by low-magnitude, high-frequency freeze-thaw events and abundant moisture availability.

As part of the British Antarctic Survey 'Fellfield Ecology Research Programme', an attempt is in progress to investigate mechanical weathering rates in the maritime Antarctic, with particular emphasis on the role of freeze-thaw. Studies within this programme have so far included the analysis of

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sites (Figure 1) reflect different environmental conditions on the island, and each has a microclimatic data logger that provides details of the conditions to which the rocks are subject. The Factory Bluffs site is about 80 m a.s.l. and 200 m inland from the coast. It is blown snow-free for much of the time and has no ice or mountains overshadowing it. The Moraine Valley site is about 50 m a.s.l. and 650 m inland from the coast. The valley has ice at its inland (southern) end and along the whole of the western side. There is a glacial debris-veneered bedrock wall along the east side and the valley is open to the sea at its northern extremity. The valley walls to the east and west rise to *c.* 150 m above the valley bottom. Jane Col is at an altitude of *c.* 150 m a.s.l. and is *c.* 700 m from the coast. This is an ice-surrounded location subject to a long period of snow cover.

Although the altitudes and distances are small, they generate significant environmental differences. The island itself is small (*c.* 19.9 km^2) and rises to only 279 m a.s.l. at its highest point (Tioga Hill). Degree of exposure to winds and sun, sea spray, shadow effects, cold air drainage from the ice, aspect and longevity of snow cover, all generate marked differences in the conditions to which the rocks are subject.

METHODOLOGY

Three large blocks of indigenous quartz-micaschist and two of marble were cut into 86 small tablets with dimensions in the region of 5 × 5 × 2 cm. Following the procedures of Cooke (1979), the dry weight, porosity, microporosity, water absorption capacity and saturation coefficient were determined for each tablet. In the austral summer of 1983–4 between 26 and 30 of these cut tablets were located at each of the three study sites detailed above. The tablets were placed on the ground in a small area (*c.* 3 × 2 m) close to where the micrometeorological transducers are situated. Tablets were not positioned with any particular orientation but were all equally exposed to the climatic conditions of each site, and so no significant microvariations that could cause differential weathering are thought to occur. Some of these tablets were then retrieved in 1985 (15 tablets), 1987 (13), 1988 (11) and 1989 (5), and their properties were retested. It is hoped that, as sufficient tablets remain (42), collection and reanalysis will continue for at least the next five years.

RESULTS AND DISCUSSION

The reweighing and evaluation of the tablets after collection is considered to provide information regarding weathering rates. Changes in dry mass define the material lost during the intervening period, while changes in properties such as porosity are indicative of internal modification to the rock. Although it is recognized that a number of other mechanical weathering processes, as well as chemical and biological processes, are involved, it is assumed that freeze–thaw in some guise or other is the dominant agent. Abrasion by blowing snow or sand could also cause mass loss but is thought unlikely, as long-exposed outcrops show no wind-moulding effects, it is a relatively wet environment such that blown particles are rare, and winter temperatures are not cold enough to give very hard snow that could cause abrasion.

Graphical presentation of the mean percentage change in dry weight (based on three tablets per sample per sample year) shows a general increase in weight loss with time (Figure 2). Sample A, a marble, shows the greatest percentage loss, while the two quartz-micaschists (samples D and E) show slower rates, which still increase over time. Why the marble responds at a faster rate is not certain, but the reason is probably related to the extensive microfracturing found in the marble (this appears comparable to the extensive fracturing and rapid failure due to frost action experienced by marble in Svalbard: Seppälä, personal communication, 1989). However, this generalized information, although giving a picture of the rate of mass lost from the samples, hides the variations that occur between the three sites as a result of different local environmental conditions.

The percentage change (based on one sample per site per sample year) in dry weight of the marble (which has the same form of change as the quartz-micaschist but expresses it more clearly) for the three study sites is shown in Figure 3. It can clearly be seen that the greatest amount of weight loss was found for the samples located on the exposed cliff-top, which is blown-free of snow for much of the year and thus experiences frequent temperature oscillations. In addition, the melt of snow that is trapped in the lee of obstacles, together with the lack of steep gradients and the presence of fines in the soil, all combine to produce relatively high rock moisture contents at times when freezing can occur (Hall, 1986a, 1988b). The valley situation, where snow lay is longer than on the cliff-top, indicates a slower rate of breakdown. It is still a moisture-

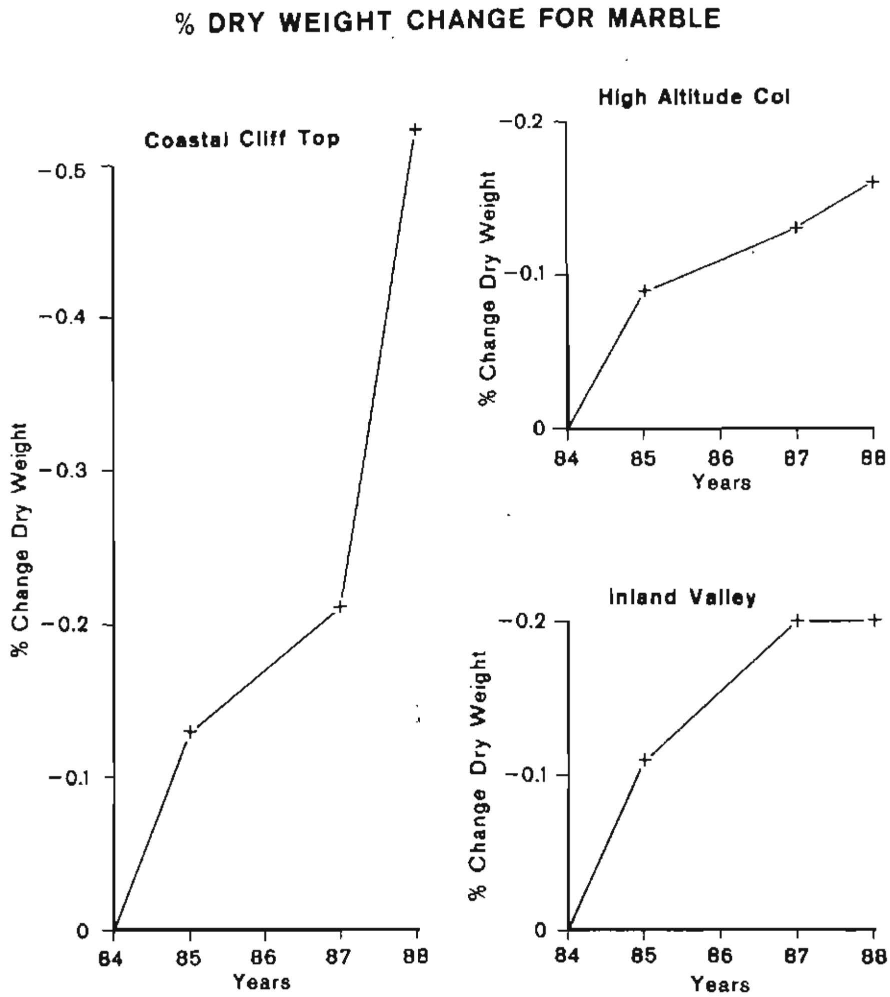

Figure 3 Weathering rates for the three study sites.

retaining environment in so far as much snowmelt moves through and across the study site, plus the fact that there are fines that aid water retention. The study site is snow-covered for much longer than the cliff-top and so experiences far fewer freeze–thaw oscillations (also fewer wetting–drying cycles). Why the 1987 and 1988 values should have been so similar is unknown and can only be ascribed to a chance effect. The samples from the col show the slowest rate of breakdown, which is in accordance with this site being snow-covered for the greater part of the year. The combination of low temperatures and low radiation inputs inhibit moisture availability during the short period that the site is snow-free.

Using multiple linear regression of the form

$$t = a + bx + cy$$

where t is weight loss (g), x is time (yr) and y is surface area (cm^2), an r value of $+0.69$ was obtained for the marble and of $+0.75$ for the quartz-micaschist. The 'total surface area' gave better correlations than mean size or the combined area of the two largest faces. It is accepted that rates are highly unlikely to be linear, some form of exponential relationship being more probable, but the database is not yet sufficient to deduce the nature of the curve. Thus, for the time being a linear relationship is presumed. Using this, it is possible to calculate mass loss for longer time periods. Mass loss from the test samples to date is of the order of 0.1–0.5 g, and thus a best estimate from the regression equation gives a loss of only 2.2 g from a 94.2 g tablet after 100 years.

This suggests that the rate of rock breakdown, at least initially, is very slow. Although rates for the

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Chapter 3

Laboratory Simulations

"It is also a good rule not to put too much confidence in experimental results until they have been verified by theory."
Sir Arthur Eddington

Laboratory simulations have long been a major consideration in cold region weathering studies (e.g. Thomas, 1938; Wiman, 1963; Martini, 1967; Potts, 1970; Lautridou, 1971; Brockie, 1972; Steijn, 1979; Lautridou and Ozouf, 1982; Fahey, 1983; Swantesson, 1985; Matsuoka, 1988) and much has been deduced from these experiments regarding the nature and rate of freeze-thaw weathering. The major problem is that, as Fahey (1983, p.103) states, "...experimental conditions are not particularly representative of actual periglacial environments" and so "laboratory experiments are ultimately only worthwhile if it can be shown that they mimic natural conditions successfully" (Thorn, 1988, p.13). A significant problem in the majority of laboratory experiments regarding freeze-thaw weathering has been the assumption that freeze-thaw *was* the critical process and so experimentation was modeled to only investigate this. Further, in most freeze-thaw simulations the broad, idealized climatic themes of 'Icelandic' or 'Siberian' cycles are used, based upon the recommendations of Tricart (1956): the former being high frequency, low amplitude freezes and the latter low frequency, high amplitude freezes. Whilst this certainly facilitates, at least in principle, comparison between undertakings, the reality is that the cycles themselves may have little relationship to the field situation. The variability in freeze rates, and even freeze amplitudes, for each of the cycles as used by different researchers introduces complexities that make comparisons hazardous. Thus, even the application of these "theoretical" cycles to create a standard for comparison did not work. Warren (1914a, p. 413) warned that, it would be unsound "...to assume that the results of a certain experiment must also be produced by natural agencies, without evidence that similar conditions exist in Nature to those employed by the experiments." With respect to the simulation of freeze-thaw in the laboratory, McGreevy and Whalley (1982, 1985) point out that there are inadequate data regarding rock temperatures and rock moisture for any meaningful undertaking. Fahey (1985) also noted that there is a lack of information pertaining to the chemistry of interstitial rock water.

The argument is sometimes made that simulations, even if not based on field data, are meaningful insofar as they still allow an evaluation of the mechanism(s) associated with the process. Although clearly true, the problem is that the researcher is still left with the dilemma as to whether this mechanism can ever take place for it is not known whether the required conditions occur at the study site(s). Further, the issue may be compounded by others taking those results and, by repetition, adding an "authority" to the findings that is, in fact, potentially false. The classic examples of just such a situation are the laboratory studies of Griggs (1936) and Blackwelder (1933) on thermal stress fatigue which did *not* replicate a 'natural' situation but the results of which, through endless repetition, became "fact" (see the discussion on this in Chapter 4). In truth, as argued in Hall, (1999a), these very experiments probably stifled the idea of thermal stress fatigue as an active agent in cold regions such that the concept of freeze-thaw gained even greater credence as an explanation for the broken rock. In much the same way, the findings from freeze-thaw experiments wherein rocks certainly broke down helped reinforce the freeze-thaw concept for they appeared to 'prove' what was found in the field. The "frost susceptibility" of rocks, based on the laboratory studies of Icelandic and Siberian cycles, became accepted such that the findings of, for instance, a broken, angular sandstone in a periglacial area thus must, *de facto*, be the result of frost action for it had been 'proved' that sandstone was indeed frost susceptible. The two fallacies of the argument are: first, that no other process(es) was/were operative, and, second, that the conditions utlized in the simulation were meaningful to *that* field situation.

The main problem with freeze-thaw simulations was, and continues to be, the allocation of moisture. As McGreevy and Whalley (1985) point out, rock moisture data are all but non-existent. Laboratory experiments tend to have samples submerged in water, in trays half-full of water and/or, in some instances, sprayed with water; often a dry sample is also used as a 'reference'. That the dry sample showed little or no breakdown compared to the wetted samples 'proved' the efficacy of freeze-thaw weathering. The reality of samples being frozen in trays of water and the applicability of this to a field situation seems to have rarely been questioned (McGreevy, 1982 being one such exception). Further, as McGreevy (1982) clearly identifies, freeze-thaw is *not* the only mechanism causing rock breakdown and even within the freeze-thaw simulation there exists the likelihood of weathering due to wetting and drying (resulting from the gain and loss of moisture), salt (from the precipitation out of salts during drying sequences and/or as a cryohydrate during freezing of saline water), and thermal stresses (due to the temperature cycling driving the 'freeze-thaw' cycle). Thus, many experiments may well have been recording rock breakdown due to more than just freeze-thaw but this was neither recognised nor evaluated.

A number of papers deal with some of the issues raised in the above discussion regarding laboratory weathering simulations:

- Hall, K. 1986c. Freeze-thaw simulations on quartz-micaschist and their implications for weathering studies on Signy Island, Antarctica. *British Antarctic Survey Bulletin*, 73, 19 - 30.
- Hall, K. 1988b. The interconnection of wetting and drying with

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- Hall, K. 1988e. A laboratory simulation of rock breakdown due to freeze-thaw in a maritime Antarctic environment. *Earth Surface Processes and Landforms*, 13, 369-382.
- Hall, K. and Hall, A. 1996. Weathering by wetting and drying: Some experimental results. *Earth Surface Processes and Landforms*, 21, 365-376.
- Hall, K., Cullis, A. and Morewood, C. 1989. Antarctic rock weathering simulations: simulator design, application and use. *Antarctic Science*, 1, 1-9.

The last paper cited above deals with the design and possible applications of a simulator for weathering studies. Rather than attempt to utilize a commercially available cabinet, which might not have been able to do all that was required, a cabinet was designed that would have the range of options considered necessary for the undertaking of weathering simulations and the collection of data during experiments (Fig. 6). The cabinet was designed to facilitate cold temperatures (to at least -20°C) with the ability to warm or cool at any predetermined rate. The cabinet was also made such that the set cold temperature could be held whilst radiative heating of the rocks, via infra-red lamps, was undertaken (to simulate thermal stresses due to radiative heating in cold environments). Cabinet dimensions enabled the use of (relatively) large rock samples (± 0.75 x 0.75 x 0.5 m). A range of transducers to monitor cabinet and rock parameters (e.g. temperatures, ultrasonic p-wave velocities, rock cracking, acoustic emissions, etc.) were incorporated within the system. Although unsophisticated by comparison to present-day computing capabilities and power, the cabinet and transducers were controlled by a computer and the software to run the complete system (cooling, heating and transducers) was especially written. In addition to the computer-controlled programmed temperature conditions, the computer also monitored all of the other transducers and recorded their outputs. Thus, for instance, it was possible to know the temperature at which rock cracking (as shown by a fiber optic crack detection system) took place. Later, a second cabinet was built that allowed more complex temperature cycling and these two cabinets provided the foundation for the weathering simulations cited here.

The data provided by the cabinet simulations offered many new insights into freeze-thaw weathering mechanisms, rates, and interactions. Further, via the data output it was possible to test new hypothetical models (e.g. Hallet, 1983) regarding the mechanism and the controlling factors regarding freeze-thaw weathering. By means of unidirectional freezing experiments it was possible to evaluate weathering of rock walls based on data from Signy Island (Antarctica). Testing of the influence of anisotropy on unidirectional weathering was also undertaken. The ultrasonics proved invaluable in determining, via pulse velocity changes, if and when freezing of interstitial water took place and also how this changed when NaCl solutions were used to replicate interstitial rock water chemistry measured in the field. The ultrasonic data, combined with monitored exotherms from latent heat release upon freezing of the water, were also able to indicate the extent and rate of interstitial water freezing as a function of the rate of change of temperature. These findings were significant in evaluating the hypothetical model of Hallet (1983). All in all, a substantial amount of new information pertaining to all aspects of freeze-thaw were generated by the laboratory simulations - as provided in the papers cited above.

The moisture data were so significant to weathering studies and to the undertaking of meaningful laboratory simulations, that an attempt was made to investigate the role of wetting and drying both as an integral part of freeze-thaw and as a mechanism in its own right. The finding of multiple wet-dry cycles in some field situations (Chapter 2) suggested that this should also be investigated in the laboratory, particularly as little is known regarding the role of this weathering mechanism (Prick, 1999). Several of the papers cited above deal with wetting and drying, as a process on its own and also as an integral part of freeze-thaw weathering. Hall (1988b) used ultrasonics in an attempt, initially, to determine interstitial rock water distribution. Although the concept worked in principle when tested on a perspex model it failed when rock was used. As a result of the 'failure' of the test when applied to the rock a number of new insights into the wet-dry weathering mechanism were forthcoming. Further to this work, the experimental sequence reported in Hall and Hall (1996) provides information regarding the impact of water allocation as used in freeze-thaw weathering experiments. It is shown that the method of water allocation can, through time, impact upon porosity, pore-size distribution, water absorption capacity and the actual weathering of the rock. It is argued that wetting and drying effects can enhance and exacerbate the freeze-thaw weathering such that future studies of freeze-thaw need to take this into account.

Thus, the simulation information presented here builds on, and adds to, the data collected in the field. The simulations take on some meaning with respect to the study sites as they are based on rock samples collected from, and temperature-moisture conditions prevailing at, those sites. A number of new findings with respect to freeze-thaw, the role of salts, and of wetting and drying were obtained. Significantly, these findings helped set the foundation for a better understanding of the synergistic relationship between the various weathering processes. Within the initial study framework, the laboratory simulations also give new insights regarding weathering processes and rates that could then be evaluated in the field (see Chapter 7). It is suggested that the material found here adds to our body of theory and to our better understanding regarding weathering in cold regions.

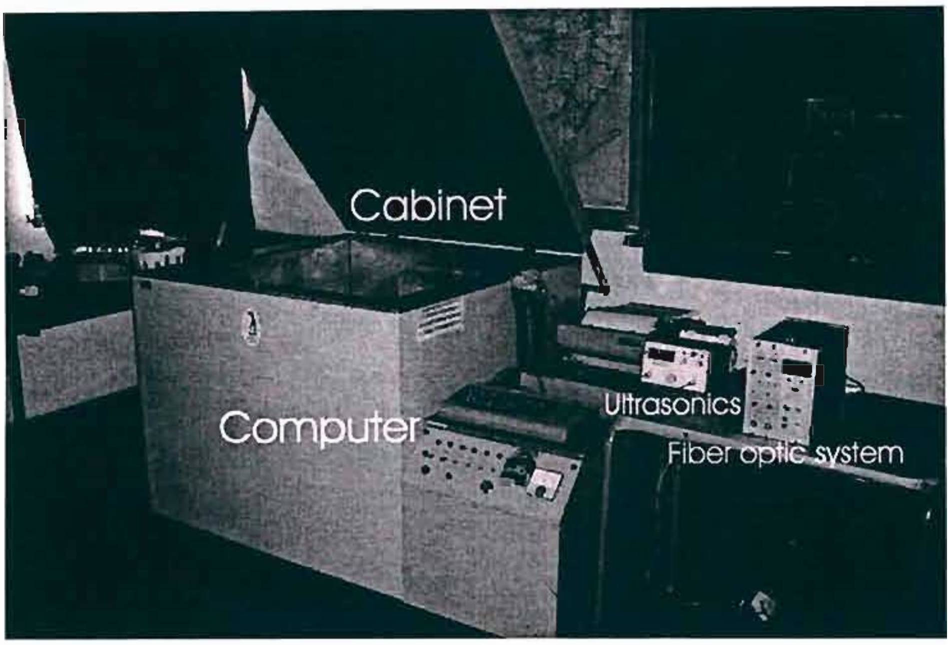

Fig. 6

View of the first purpose-built climatic simulation cabinet. Details regarding the cabinet and the transducers are provided in Hall, *et al.,* 1989.

FREEZE-THAW SIMULATIONS ON QUARTZ-MICASCHIST AND THEIR IMPLICATIONS FOR WEATHERING STUDIES ON SIGNY ISLAND, ANTARCTICA

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ABSTRACT. Results of two series of freeze-thaw simulations on quartz-micaschist indicate that there is a significant difference in the rate of freeze penetrating depending upon whether the plane of schistosity is normal or parallel to the advancing freezing front. Rate of fall of temperature is up to five times faster when schistosity is parallel to the freeze advance. In these stimulations it was found that the rate of fall of temperature within the rock was controlled primarily by the amplitude of the freeze event rather than the environmental rate of fall of temperature. A distinction is made between open systems (e.g. cliffs) and closed systems (e.g. loose blocks) with respect to processes and rate of breakdown. It is suggested that, with the very low porosity of this rock, there is a difference in the freeze mechanism based upon schistosity orientation but that, overall, moisture content plays a crucial role in determining whether any frost weathering will occur.

INTRODUCTION

A series of laboratory simulations of freeze-thaw cycles were undertaken, as part of the investigation of the mechanisms and rate of weathering of quartz-micashist in the maritime Antarctic environment of Signy Island. To date, the results of many simulations have been more reflections of experimental design and procedures rather than the environmental conditions which rocks might experience in nature (McGreevy, 1982). However, in the present instance the study of freeze-thaw weathering comprises part of a larger study, the Fellfield programme (Walton and Hall, submitted), and consequently, for the first time, a large data base was available to relate the simulations to the field situation.

A number of early freeze-thaw studies utilized schists (e.g. Wiman, 1963; Martini, 1967; Brockie, 1972) but no data were presented on either the properties of the rocks themselves or the environments that were being simulated. Thus, the results of these early experiments could not be related to any particular environment. Recent studies have shown the importance of information on such factors as rock water content (e.g. McGreevy and Whalley, 1985), rock thermal properties (McGreevy, 1985), interstitial rock water chemistry (Williams and Robinson, 1981; McGreevy, 1982; Fahey, 1983), and the engineering properties of rock (McGreevy and Whalley, 1984). In this study most of the necessary background data were either already available or produced as part of this study, e.g. field micrometeorological conditions (Walton, 1982), rock moisture content and properties controlling this (Hall, 1986*a*), the chemistry of the interstitial rock water (Hall and others, 1986), the physical properties of the rock (Hall, in press) and the application of rock fracture mechanics (Hall, 1986*b*). Thus the planned simulations could be closely related to the physical and environmental constraints characteristic of quartz-micaschist on Signy Island. The data presented here relate to two sets of simulations, and are concerned with the rate of fall of temperature within the rocks and the effects of schistosity orientation upon this and subsequent processes.

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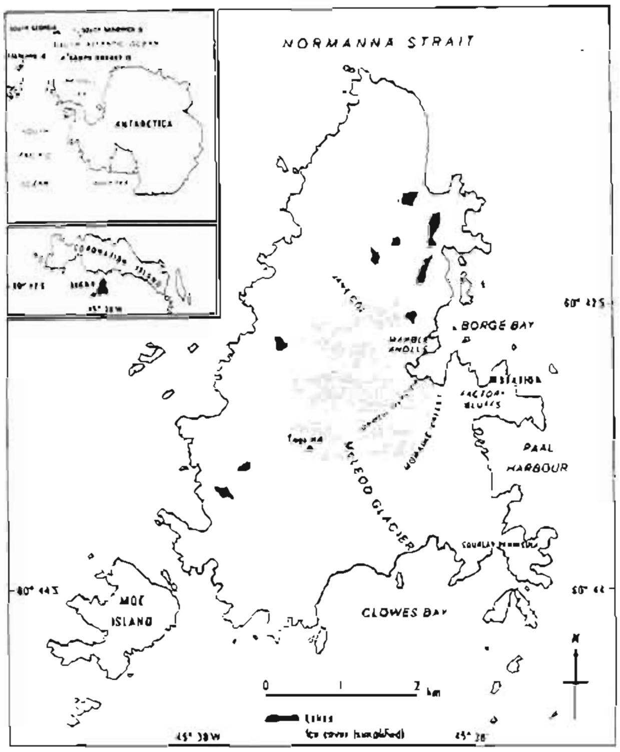

Fig. 1. Location map for Signy Island.

Field Situation

Rock samples and field data were collected from Signy Island (60° 43′ S. 45° 58′ W), one of the smaller islands in the South Orkneys (Fig. 1). As the bulk of the island's metamorphosed sediments comprise quartz-micaschist (Mathews and Maling, 1967; Storey and Meneilly, 1985), this rock was chosen for the initial simulations. In the field, all cliffs and rock outcrops observed had the plane of schistosity normal to the cliff face and parallel to the ground (Fig. 2), with minor variations due to localized folding. Thus, over the face of the cliff schistosity was normal to freezing plane penetration. On blocks that were weathered – free from the cliffs, and at the very cliff top, schistosity was sometimes parallel to the direction of freeze penetration. On the cliffs, once some material has been removed, gravity aids weathering (Hall, in press), and the orientation of the schistosity abets this. This effect of gravity is recognized but is not considered in any detail in the following text.

Climatically, there is a typical cold, oceanic regime with a mean monthly temperature of *c.* −4°C, but the summer three months have means slightly above freezing (Watson, 1975; Collins and others, 1975). Rain predominates in the summer but during the rest of the year the precipitation is in the form of snow. Wind speeds average 26 km h^{-1} and mean sunshine levels are less than 1.5 h per day.

Fig. 2. Example of a typical quartz-micaschist cliff. Note the exploitation along the lines of schistosity that are parallel to the ground and normal to the face. In addition, a number of vertical cracks can be seen that define the edges of large blocks in the process of being weathered-free

METHODS

Samples of quartz-micaschist were collected from different environmental positions at a number of locations of Signy Island (Hall, 1986*a*). For the bulk of the samples the collected (i.e. 'wet') weight was found and then subsamples were subjected to the irregular lump point load test (Broch and Franklin, 1972) in order to gain a measure of the strength of the rock at field moisture content (details of rock strength tests are presented in Hall, in press). The rocks were then dried at the island laboratory so that it was possible to calculate the actual field moisture content (Hall, 1986*a*). Later, the rocks were tested for porosity, microscopy, saturation coefficient and water absorption capacity following the procedures described by Cooke (1979). In addition, a new technique, was used to obtain information about the interstitial rock water chemistry (Hall and others, 1986). Thus, for the rocks that were to be used in the simulations there was a data base pertaining to the field properties of the rock concerned.

Data on the climate and microclimate at various selected reference sites were available (Walton, 1977, 1982). However, in 1983–84 'Datacapture' micro-meteorological data loggers (Walton and Hall, submitted) were installed at the main study areas (Factory Bluffs, Moraine Valley and Jane Col; Fig. 1). These data were used in planning the temperature cycles to be used in the simulations. The temperature of the environment chamber is controlled via a microcomputer that continually monitors the chamber, compares the measured and programmed temperatures and initiates corrective action when required. The same microcomputer, via additional hardware, monitors, logs (on disk) and prints out the sensor data (cabinet and 6 rock temperature sensors, and cabinet humidity) at pre-programmed intervals; details of the equipment are presented in Walton and Hall (submitted).

Simulation I was carried out on a 12.0145 kg block of quartz-micaschist with sensors on the rock surface and at depths of 9, 19, 44 and 92 mm. The rock was encased in polystyrene such that only one a/b plane, with schistosity parallel to the cooling front, was exposed, and wetted to a typical field moisture content (0.11% by weight). In Simulations II a 9.5836 kg block, at field moisture (0.14% by weight), was encased in polyurethane foam such that only one face (the b/c plane) was exposed. In this instance schistosity was normal to the freezing plane, a situation found on most of the cliff exposures observed in the field. Thin film platinum resistance temperature sensors were placed at the rock surface and at 5, 10, 20 and 60 mm depths behind the exposed face. In addition, a small sample of this rock (dry wt = 91.9 g) was 'saturated' by immersion in water for 48 h (i.e. the rock absorbed the maximum amount of water possible under non-vacuum conditions) and was then placed in a tray of water to half its depth as a means of comparison with earlier experiments (e.g. Wiman, 1963). The two samples were very similar with respect to their composition, both having thin silica laminae with the occasional small silica pod.

Both simulations, comprised the same temperature sequence, as shown diagrammatically in Fig. 3, which was considered to approximate to conditions occurring in the spring to autumn poeriod. Simulation I ran for 12 full cycles (600 h) and generated 349368 data readings whilst Simulation II ran for 10.5 cycles (528 h) with 196376 data recordings. The frequency of data collection during Simulation I is presented in Fig. 2; data read rates remained the same in Simulation II except during the long, warm phases when fewer readings were obtained in order to conserve disk space.

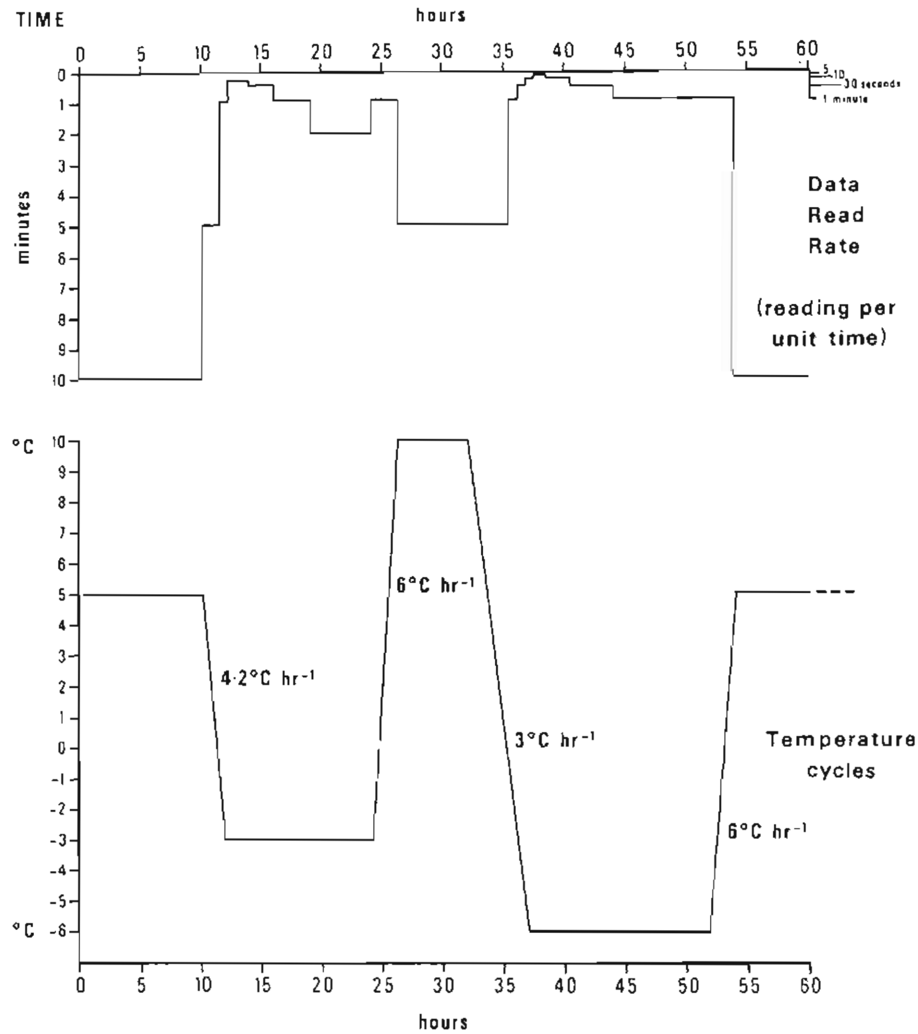

Fig. 3. Graphs to show the temperature sequence, including rate of change of temperature, to which the samples were subjected and the rates at which the computer monitored data during those cycles.

Results and Discussion

In Simulation I sudden temperature rises during cooling phases occurred at 44 mm depth in rock when the temperature was between $-1.2°$ and $-1.4°C$. The rises varied between 0.7° and 0.9°C, and are thought to reflect the release of latent heat during the water-to-ice phase change (Fig. 4). It was noticeable that no signs of this latent heat release were detected in the early cycles, only after four complete cycle sequences were the first peaks recorded, but they were consistent thereafter. During the $-3°C$ cycle the chamber temperature at the time when the exotherm was produced averaged $-2.6°C$, a level it had held during the preceding two hours. The high during the $-6°C$ cycle occurred at a cabinet temperature of approximately $-5.3°C$ during the continual fall towards $-6°C$. Some small, apparent peaks were recorded for other depths.

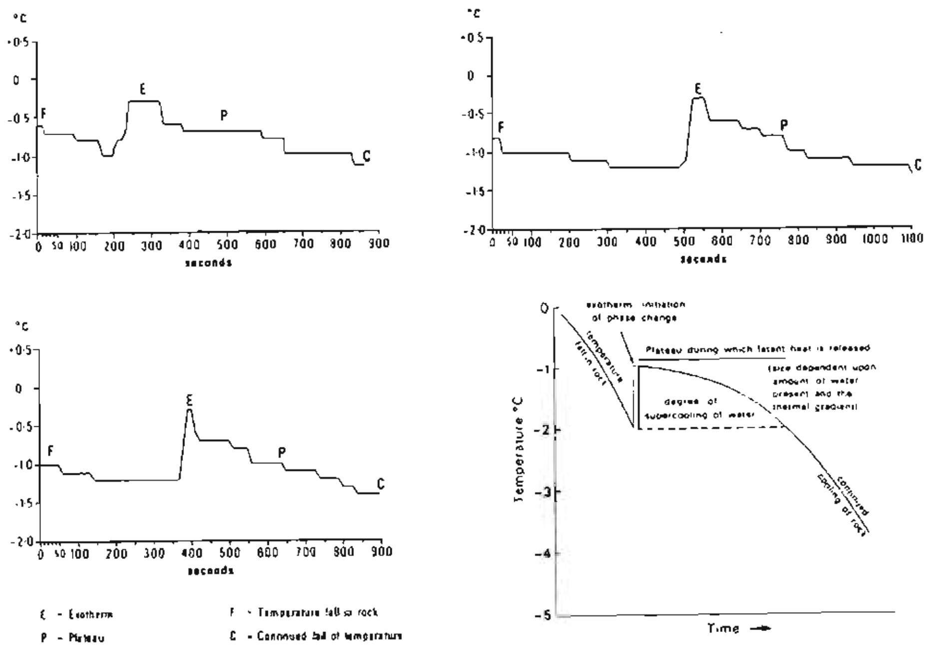

Fig. 4. Examples of some exotherms monitored at 44 mm depth during Simulation I together with a simplified graph detailing the various parts of the temperature curve.

notably a 0.3°C rise at −1.1°C for the 19 mm depth sensor, but none like those observed at 44 mm depth.

Whilst it might be thought that the temperature peaks were due to water freezing in the drill holes, rather than in the rock itself, this is not considered to be the case for four reasons: Firstly, the presence of a tight-fitting sensor covered with a coating of thermal grease would allow very little moisture to enter into the drill hole. Secondly, Douglas and others (1983) show that freezing was initiated in a 5 mm hole (the width of the present drill holes) at −0.6°C, a temperature somewhat higher than the −1.1° to −1.4°C found in this experiment; thirdly, if the freezing points were related to water in the drill holes then it would be expected to be noticeable in all holes and most likely at the first freeze after wetting, when the holes would still contain some moisture. However, four full cycles were required before thermal peaks were observed. Finally, whilst it is possible that some water from within the rock was, after a number of cycles, forced into the drill holes under hydraulic pressure in front of the advancing freezing plane (Powers, 1945), this is thought to be minimal due to the presence of the tighly fitting sensor tube and its covering of thermal grease.

Consideration of the resultant temperature data allowed the rate of fall of temperature to be calculated for different depths during both freezing cycles (Table I). What is apparent is that the rate of fall of temperature *within* the rock is partly controlled by the final temperature to which the freeze is going. For instance, for all sensor depths, the rate of fall of temperature over the range 0–3°C is faster, up to 5.5 times faster, during the −6°C cycle than during the −3°C cycle. This more rapid fall of internal rock temperature found for the −6°C sequence occurs despite the environmental rate of fall of temperature being slower (3°C h^{-1}) than during the −3°C cycle (4.2°C h^{-1}). The marginally faster cooling rate observed at the 19 mm depth may be due to the presence of quartz which allows for a faster passage of heat,

Table I. Typical cooling rates observed at various depths in the rock for Simulation I

Range	Chamber rate	Cooling rate (°C h^{-1})			
		At surface	9 mm	19 mm	44 mm
0 to −6°C cycle					
0 to −3	3°C	1.1	1.7	1.7	1.4
−3 to −6	3°C	0.3	0.3	0.4	0.4
0 to −3°C cycle					
0 to −3	4.2°C	0.4	0.3	0.4	0.4

Properties: Rock weight, 12.045 kg; Porosity, 0.83%; Saturation coefficient, 0.6; Moisture content, 0.11%; Water absorption capacity, 0.38%; Compressive strength, 1.98 MN/m² normal to schistosity, 0.4 MN/m² parallel to schistosity

Table II. Some thermal peaks observed during freeze phases

Experiment no.	Sensor depth (mm)	Temperature exotherm (°C)	
		From	To
I	44	−1.2	−0.3
I	44	−1.0	−0.3
I	44	−1.0	−0.3
I	44	−1.2	−0.3
I	92	−1.0	−0.7
I	92	−1.1	−0.8
II	60	−3.1	−2.9
II	5	−4.0	−3.7
II	10	−4.0	−3.5
II	10	−3.9	−3.3
II	20	−3.7	−3.2
II	60	−3.6	−3.4
II	10	−3.0	−2.7
II	20	−2.6	−1.6
II	60	−2.4	−2.1
II	60	−2.2	−1.9

Note: for all peaks shown the time taken to return to original pre-exotherm temperature varied from 8 to 34 min.

due to its thermal conductivity being higher than that of the mica. Although this is by no means certain, quartz is distributed throughout the rock with numerous localized concentrations, and so it could well be that a drill hole coincided with one of these.

In Simulation II thermal peaks during freezing were found at various times, for all depths (5, 10, 20 and 60 mm). However, unlike Simulation I, the temperatures at which peaks occurred were much lower and the degree of water supercooling was not as great (Table II). Although the exotherms were much smaller they are not 'instrument errors' insofar as the time taken to return *gradually* to the pre-exotherm value varied between 8 and 34 min. Rates of fall of temperature (Table III) again show that, despite the faster environmental decline, the rate of fall of temperature within the rock was faster (*c.* 3.5 times) for the lower end temperature (−6°C).

Comparison of Tables I and III clearly show that the rate of fall of temperature within the rock is slower in Simulation II. For the 0° to −3°C range (of the −6°C freeze), rates for the smaller block were about half those found in the larger block.

Table III. Typical cooling rates observed at various depths in the rock for Simulation II

Range	Chamber rate	Cooling rate (°C h^{-1}) 5 mm	10 mm	20 mm	60 mm
0 to −6°C cycle					
0 to −3	3°C	0.84	0.90	0.83	0.85
−3 to −6	3°C	0.23	0.24	0.22	0.24
0 to −3°C cycle					
0 to −3	4.2°C	0.24	0.25	0.24	0.26

Properties: Rock weight, 9.5836 kg; Porosity, 0.54%; Saturation coefficient, 0.71; Moisture content, 0.14% Water absorption capacity, 0.39%; Compressive strength (normal to schistosity), 1.89 MN/m^2.

For the −3° to −6°C range they were only reduced by about one-third. The main distinction between the two simulations, both rocks being similar in comparison, is that of the relationship of freezing plane penetration to schistosity orientation. In Simulation I schistosity was parallel to the freezing plane whilst in II it was normal to it. As the rate of fall of temperature in the rock can be crucial to both the type and rate of destruction (Walder and Hallet, 1985), it is apparent that the orientation of schistosity in relation to freezing plane penetration could exert a significant influence. On Signy Island all cliffs and rock outcrops observed were comparable to that of Simulation II. However, for loose blocks on the ground and for rock near the top of cliffs, penetration would be both parallel and normal to schistosity.

The reason for the effect of schistosity orientation upon the rate of fall of temperature is not clear. However, it may be related to the presence of air/water along the mineral interfaces of the laminae. When the plane of schistosity is parallel to freeze penetration the air/water 'layers' are very thin and are possibly 'bridged' at many places by the quartz particles. This, then, would allow for a relatively rapid transfer of heat. When penetration is normal to schistosity then the gain/loss of heat in the air/water *along* the laminae, as the freezing plane progresses inward, would be slower since the air/water is a poorer conductor than the rock and the 'bridging' by quartz particles would not help in this instance. However, the importance of schistosity orientation with respect to freeze penetration extends beyond its effects upon the rate of fall of temperature.

Where the laminae are parallel to the direction of freeze penetration then there is a closed (i.e. water migration is confined) rather than an open (where water is free to migrate) (Fig. 5) system, even in the situation of a cliff which would otherwise be recognised as an open state (Walder and Hallet, 1985). Due to its low permeability and the presence of silicate laminae, migration of moisture between layers is highly improbable. Thus, the combination, particularly in cliffs, of low rock moisture content (Hall, 1986*a*), negligible between laminae water movement, and the relatively simultaneous freeze along each plane of schistosity, means that little destruction will occur as the ice will fill the dry sections rather than exert a tensile force (Fig. 5). Water migration along laminae to freezing points is almost non-existent due to simultaneous freezing. Only in situations of high moisture content could tensile forces be developed (i.e. saturation of $\geqslant 91\%$). Conversely, where schistosity is normal to freeze there is a greater potential for frost weathering, despite the frequent low moisture contents. In this instance freeze penetration is progressive rather than simultaneous and so it may be possible to have water migration to the point of freezing or the forcing away of moisture in front of the advancing ice front (Powers, 1945), both of which could cause damage to the rock (Fig. 5). Essentially it will be the amount of moisture which

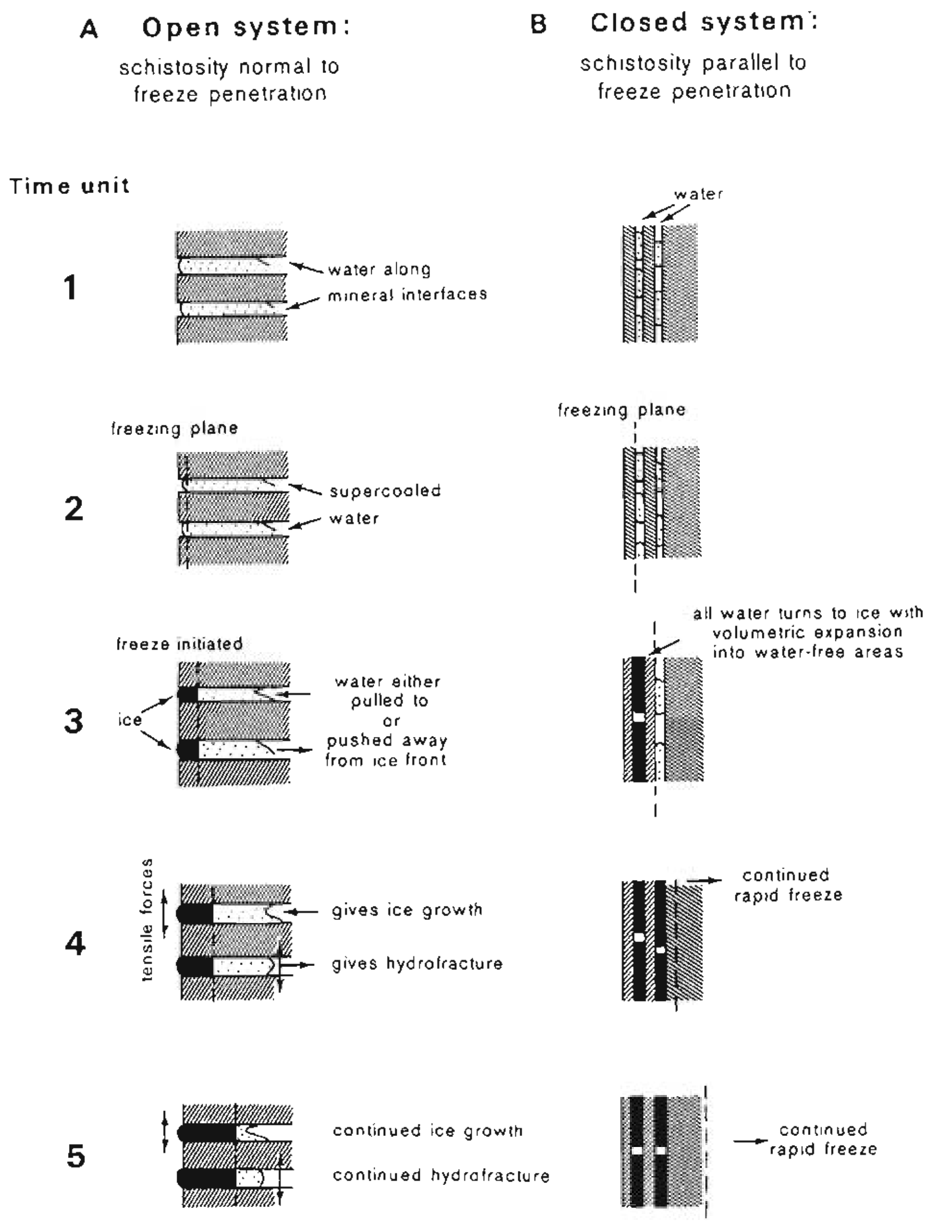

Fig. 5. Diagrammatic representation of the types of freezing envisaged for schistosity normal to, and parallel to, the freezing plane (not to any scale).

is the primary control of the degree of damage produced: the greater the moisture content the greater the potential for tensile forces. In a sample with adequate moisture the slow rates of fall of temperature associated with this schistosity orientation would be conducive to the mechanism of frost shattering suggested by Walder and Hallet (1985).

A further factor mitigating against extensive damage to the rock, particularly in the case of freeze penetration parallel to schistosity, is that of 'opposing forces'. Hallet (1983), in his model of rock breakdown due to freezing states that there is expulsion of water ahead of the freezing front, in a saturated rock, as long as crack expansion is insufficient to accommodate the volume increase due to the ice-water transition. In a laminated rock, like quartz-micaschist, where moisture is concentrated along the lines of mineral interfaces, this would mean that upon freezing the ice in one laminate would be exerting pressure against ice in another (Fig. *5b*). Thus, with schistosity parallel to freeze penetration it is only the outer layer and the edges of the laminae that might be subject to damage, even in a saturated rock.

During Simulation II a small sub-sample of the large block was first saturated and then placed in a tray of water to half its depth. At the end of the experiment, despite only experiencing seven full cycles, as opposed to the 10.5 cycles of the large block, the sample had lost 0.76% (by weight) material whilst the block lost only 0.013%. Two factors help to explain the greater loss from the small sample: a higher moisture content and omnidirectional freezing. The simultaneous freezing from all sides would produce a closed system and hence the potential for hydrofacturing (Walder and Hallet, 1985), and the greater moisture content would abet this process. This would help explain why loose blocks in the field with relatively high moisture contents exhibited a greater degree of breakdown than did the cliffs. Theoretically, with the asymmetric freeze penetration of small blocks there should be the greatest potential for frost damage towards the base of the block. There water would freeze last and so lateral water migration would take place over a longer period.

The anisotropic response of quartz-micaschist and schist to freezing is something not previously recognized and, as such therefore, has not been taken into account in earlier studies (e.g. Fahey, 1983). The evidence available here suggests that, with unsaturated small blocks subject to rapid freezing little damage will occur, as was found by Fahey (1983, p. 541). This is because the volumetric increase at the water to ice phase change is taken up by the available free space along the laminae. With small samples that experience rapid falls of temperature, giving freeze penetration parallel to schistosity, there is no mechanism available to cause localized tensile stress. Thus, in any future study which encompasses laminated rocks it would be necessary to consider the asymmetric nature of freeze penetration.

CONCLUSIONS

The orientation of the plane of schistosity to the freezing front appears to affect both the rate of fall of rock temperature and the actual process operating. In addition, the rate of fall of temperature within the rock is seen to be more a function of freeze amplitude than the rate of environmental temperature decline. A clear distinction, with respect to freeze mechanisms and potential for rock damage, can then be made between the 'open' system of a cliff and the 'closed' system of a loose block. It is suggested from the simulation data, that loose blocks should undergo greater and faster weathering than cliff faces and this is borne out by the author's field observations. Due to anistropy of the rock, loose blocks, which are subject to simultaneous freezing from all sides, are frozen 'asymmetrically', since freeze penetration is faster when entering parallel to schistosity, and as a consequence it is possible that the greatest damage may be towards the base of a block. Generally it is concluded that weathering rates, particularly for non-saturated cliff situations, are likely to be slow. It is suggested that a major control on the degree of weathering is moisture content. The simulation results appear to agree with field observations. This underlines the importance in weathering studies of using field data to plan laboratory simulations.

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The interconnection of wetting and drying with freeze-thaw: some new data

by

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with 3 figures

Summary. Ultrasonic monitoring of moisture adsorption and desorption of a quartz-micaschist produced results contrary to expectations. Upon water uptake rather than an ultrasonic velocity increase due to the introduction of a better propagating medium there was a velocity decrease. This velocity decrease is interpreted as being a consequence of changes in rock elasticity brought about by a diminution in the bonding strength of the rock. This means that wetting, prior to freezing, may weaken a rock. In addition, because of hysteresis, a rock will experience strength reduction and possibly ultimately failure, resulting from wetting and drying alone.

Key words: Weathering, freeze-thaw, Wetting and drying, Ultrasonic monitoring.

Introduction

The presence of water within a rock is a prerequisite for the action of weathering by freeze-thaw (HUDEC 1979, p. 148). Most, if not all, rocks in cold regions are subject to fluctuations in their moisture content (HALL 1988a), and in some locations such as tidal areas may experience marked periodic variability (TRENHAILE & MERCAN 1984). This change in moisture status, namely cyclic wetting and drying, itself comprises a weathering process, albeit one about which little is known (OLLIER 1984). Several authors have investigated the relative contributions to rock breakdown of wetting and drying and temperature fluctuations (e.g. FAHEY & DAGESSE 1984; PYE & SPERLING 1983; UUSINOKA & ERONEN 1979). However, despite the obvious intimacy of water with frost action, few authors other than Moss et. al. (1981) and MUGRIDGE & YOUNG (1983) acknowledge the inter-operation of these two processes.

Water has been recognised as having a number of effects upon the rock with which it is associated. BROCH (1974, 1979) clearly demonstrated that there is a

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decrease in the strength of rock with an increase in moisture content. DUNN & HUDEC (1966) showed that the cyclic expansion and contraction of adsorbed water associated with wetting and drying could cause deterioration of carbonate rocks. PISSART & LAUTRIDOU (1984) and HAMES et. al. (1987) found that some rocks could expand as a result of water uptake, whilst DAVISON & SEREDA (1978) showed that rock expansion could also take place upon the freezing of the interstitial water. The amount and distribution of water within a rock is also known to exert an influence on the form and rate of freeze-thaw weathering (e.g. RITCHIE & DAVISON 1968; AIRES-BARROS 1978; FAGERLUND 1975; MATSUOKA 1984; HALL 1986a and b). Thus, interstitial rock water has been recognised as having both direct and indirect effects with respect to both wetting and drying and freeze-thaw.

Water has also been shown to have an effect upon sound wave propagation within rocks. YANG & KING (1986) show that there is a marked increase in compressional wave (Vp) velocity upon rock saturation but MASSON (1979) found that the type of rock influenced whether there was a velocity increase or decrease. NUR & SIMMONS (1969) demonstrated that Vp was lower in rocks at or near atmospheric pressure compared to rocks subjected to pressures of a few kilobars. However, BERRYMAN (1986) shows that in the low frequency range of 1–100 Hz there is an attenuation of wave velocity associated with fluids in rocks. The monitoring of ultrasonic wave propagation has been utilised, with very good results, as a means of discerning the nature, timing and degree of water to ice and ice to water phase changes (FUKADA 1971; TOURENQ et. al. 1971; MASSON 1979; LETAVERNIER 1984; MATSUOKA 1984; HALL 1988b).

As part of a long-term, multi-disciplinary study of rock breakdown and soil development in the Maritime Antarctic (HALL 1986a, b, c, 1987, 1988a and b; HALL et. al. 1986) an attempt was made to find a method of monitoring moisture gradients within rocks using non-destructive ultrasonic techniques. During these experiments evidence was obtained regarding changes within the rock due to absorption and desorption of water. That data will now be presented and its significance considered with respect to freeze-thaw weathering.

Experimental Procedure

Ultrasonic wave velocity in air is $\pm$ 330 m sec^{-1} whilst in water it is $\pm$ 1300–1400 m sec^{-1} (PRESS 1966). Thus, it was theorised that if a block of rock was dried and its ultrasonic velocity found, that velocity should then increase upon the introduction of water due to the filling of the pores and microcracks by a better transmitting medium. Evidence supporting this hypothesis is given in figure 3 of MATSUOKA (1984). Results from preliminary experiments utilising sheets of perspex as an analogue for laminated quartz-micaschist (the main rock in the study area) also gave the anticipated response (Fig. 1).

A block of quartz-micaschist from the Antarctic study area, whose properties were known (HALL 1987), and a block of local Cave Sandstone from the Drakensberg mountains of South Africa were oven dried ($\pm$ 105 °C for 10 days) and then left to cool to ambient conditions (19 °C) in the laboratory. High resolution 1 MH_z ultrasonic transducers were then attached, by means of phenyl salicylate, at 3, 4, 5,

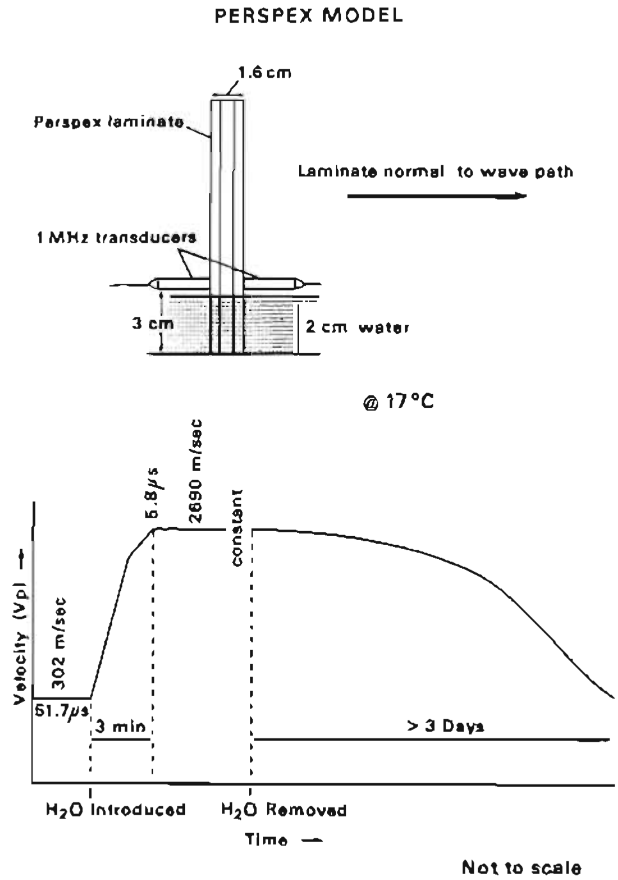

Fig. 1. Laminated perspex model used to test the hypothesis.

7 and 9 cm respectively above the base of the rock. Pulse transit time was measured to an accuracy of ± 0.1 μs by means of a "PUNDIT" apparatus manufactured by C.N.S. Electronics, and printed out, via a microcomputer, together with experiment time at one minute intervals. The blocks were left for three days after attachment of the transducers to make sure all components had equalised to laboratory conditions. Water was then introduced to a depth of 2 cm from the base of the rock in an attempt to duplicate field conditions of rock either partly buried in a wet substrate or situated in a melt rivulet. Pulse time was monitored prior to, during, and after the introduction of the water. Monitoring then took place for up to seven days before the water was removed from the tray in which the rock resided. Pulse time continued to be logged for up to five weeks after water removal. In all, 15 repetitions of the wetting and drying sequence were undertaken on the two rock samples.

PISSART & LAUTRIDOU (1984) and HAMES et al. (1987) have shown that, for some rocks, it is possible for expansion to take place due to water uptake. As an increase in distance between the transducers would cause an increase in pulse transmission time (i.e. a decrease in pulse velocity) it was important to know if any such expansion had taken place. To this end a fiber optic crack-detection system ("Opticat"), manufactured by British Maritime Technology, was utilised. Two fiber optic crack-detection gauges were bonded to each rock and an infrared signal, of known amount, was fed through an optical fibre loop, the attenuation of which was continually monitored. If attenuation were to exceed a preset value then a relay operates which causes the computer to print out which channel (of the two) has failed, the time of failure and the ultrasonic reading. If failure does occur, then a laser can be attached to the optical fibre loop such that visible light is then emitted from the crack(s) on the gauge. This, in its turn, facilitates determination of exactly where cracking took place and whether single or multiple failure occurred. This system, with gauges bonded transverse to schistosity, was thought to be particularly ideal for monitoring expansion of the quartz-micaschist.

Results and Discussion

As no evidence of either expansion or cracking was given by the fiber optic system the effects of volumetric rock changes are not considered in the following discussion. An example of ultrasonic pulse time response during water absorption, at all transducer levels, for the quartz-micaschist, is given in Fig. 2. As the distance between the sending and receiving transducers remained constant, any variation in pulse propagation time must reflect some change in the transmitting medium. Thus increases and decreases in pulse time equate to velocity deceleration and acceleration respectively. For convenience actual pulse velocity has been calculated for a number of key points on each curve. The response of the quartz-micaschist to water removal is illustrated in Fig. 3. During the periods of pulse oscillation (5 and 7 cm levels) the curve indicates the integrated value generated by the computer whilst the bars reflect the actual end values shown by the Pundit. An oscilloscope tied-in to the system showed that only the two end values were being produced and that there were no intermediate readings.

Responses for the sandstone were very similar with only the actual timing and magnitude of the inflections showing any difference.

Adsorption

The original hypothesis was that as water filled the pores so this should be evidenced by a marked pulse time decrease (i.e. an increase in velocity). However, as can be seen from Fig. 2 this was not the case. At the 3 cm level (i.e. 1 cm above the water level) there was a sudden increase in pulse time at nine minutes as the water moved up through the rock. Pulse time then continued to gradually increase until 100 minutes from experiment start, after which point it remained relatively constant ($\pm$ 0.1 μs). This gave an overall *decrease* in pulse velocity of 24%. At the 4 cm level the sudden

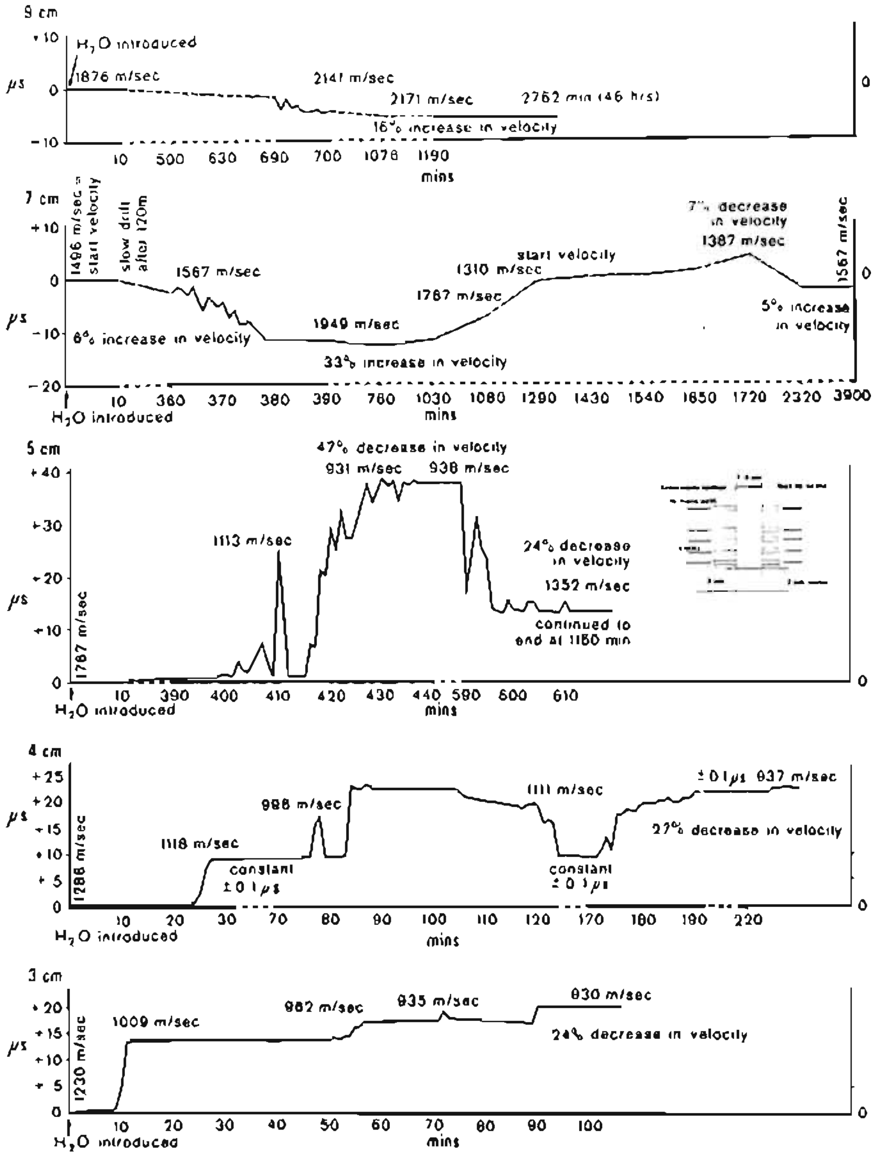

Fig. 2. Graphs of ultrasonic pulse time response for all transducer levels during water absorption. (Note: an increase in pulse time equates to a decrease in pulse velocity)

pulse time increase began 24 minutes after the introduction of the water and then, after several phases of acceleration and deceleration, ended with a constant 27% velocity decrease. Despite its peakedness, the 5 cm level indicates a similar response to that of the 4 cm level but with the inception of the pulse time increase occurring 390 minutes after the introduction of water. Although a stable 24% velocity decrease was obtained by the end there was a marked 47% decrease peak prior to this. Thus, the 3, 4 and 5 cm levels all showed very similar velocity decreases (24, 27 and 24% respectively).

At the 7 cm level, however, a slow decrease in pulse time began after 120 minutes which transformed to a marked decrease after 360 minutes. By 390 minutes there had occurred a 33% velocity *increase* but this was followed by deceleration such that by 1720 minutes there was an overall velocity decrease of 7% compared to the start. However, pulse time then once more decreased such that by 2320 minutes there was an overall 5% velocity increase that remained for the duration of the experiment. At the 9 cm level pulse time gradually diminished after 11 minutes to ultimately achieve a 16% velocity increase that continued for the next 2762 minutes. Thus, at these two levels there was a net decrease in pulse time (an increase in pulse propagation velocity) as suggested by the hypothesis.

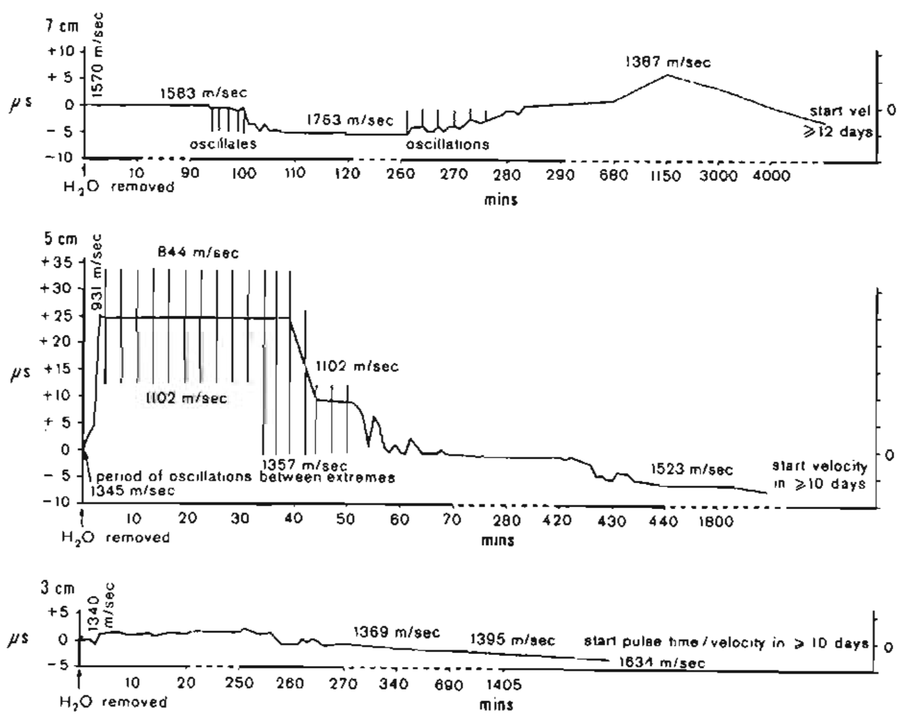

Fig. 3. Examples of ultrasonic pulse time response during the drying phase.

Desorption

During desorption of water (Fig. 3) some very unexpected responses were monitored. At the 3 cm level there was an initial short, decrease in pulse time followed by an increase that persisted for about 260 minutes before a long (≥ 10 days), gradual drift back to near the original (dry) start value. At the 5 cm level there was a marked response to water removal that began almost immediately the water in the tray lost contact with the rock. There was a sudden (within two minutes) 25 µs increase in pulse time followed by 30 minutes during which pulse time oscillated between two extremes some 22 µs apart; no intermediate values occurred. The oscillations were then continued for five minutes as one end of the range moved towards a 1 µs pulse time decrease. The range then diminished and disappeared altogether by 50 minutes, after which there was, apart from a few minor accelerations and decelerations, an overall move back to the start, dry value.

At the 7 cm level, oscillations (4 µs range) began after 94 minutes, for a period of six minutes, but were associated with a general pulse time decrease. This state continued until 260 minutes at which point small oscillations once more began (for 16 minutes) and pulse time began to increase. A peak was reached at 1150 minutes, after which pulse time once more slowly decreased and finally (≥ 12 days) returned to near the original, dry value.

Discussion

Although the results are not yet fully understood, it is apparent that the velocity decrease, found in that part of the rock which takes up the greatest amount of water, is not related to volumetric expansion. The transmission of ultrasound is a function of the density and elasticity of the rock. As density has not decreased, this must mean that the rock elasticity has diminished; it has declined to such an extent that there is a loss of pulse velocity despite the presence of water in the pores and microcracks. Water is, however, known to weaken the bonding strength of rock (MICHALOPOULOS & TRIANDAFILIDIS 1976) and hence reduce the rock elasticity. During the drying phase, elastic strain recovery can be erratic and hysteresis will take place (MICHALOPOULOS & TRIANDAFILIDIS 1976).

Thus, the variations in pulse propagation here presented are thought to be, at least in part, associated with changes in rock elasticity resulting from water causing a softening of the bonding strength of the rock. The wide fluctuations of pulse propagation time are thought to be due to air-water variations concomitant upon water loss during the initial drying phase. The large responses, noted above the 3 cm level, began almost immediately the meniscus left the rock. At that time a small suction gradient would exist, with water exiting the rock at the base and air entering at the top of the rock. BROCH (1974) demonstrated that rocks lose $\pm 25\%$ of their water content within 10 minutes, 50% after one hour, and after one day only 18–40% of the water is left ($\pm 20-22$ °C, 60–65% rel. hum.).

If water is lost at the sort of rate indicated by BROCH, and considering that the block in this experiment was not fully saturated, then there will be considerable internal variation within each individual laminae of the water-air content. This variation will be enhanced by bonding of the water molecules in the "narrower" sections and the presence of more mobile adsorbed water in the "wider" sections. Thus, a situation will exist during initial drainage where there are "zones" within each laminae of more and of less air, with the transducers monitoring a whole range of these. Visual observation does indicate that some laminae are wider than others, and thus variability of internal water drainage could be expected. After the initial phase when the free (or less bonded) water is lost, the ultrasonic oscillations cease and a steady, slow drift to near the original pulse time ensues as water is very slowly lost. The effects of hysteresis preclude the attainment of the original ultrasonic value.

The results here may appear to contradict the information regarding ultrasonic response to varying degrees of saturation as presented by such as BROCH (1974, p. 35), where an increase in P-wave velocity is indicated to occur with increase in water content. However, it is difficult to resolve the differences as the exact methodology of BROCH is not known. In this present study, the actual moisture contents during experimentation are not known as this would have necessitated removal of the transducers, and so direct comparisons are therefore impossible. This study used a quartz-micaschist and a sandstone, neither of which were tested by BROCH. This is not to argue that these two lithologies are somehow different, but rather that comparisons would still have not been possible even had moisture content been known. Finally, in this study there was a distinct moisture gradient from the base to the top of the rock, and so the ascertaining of the "average" moisture content (i.e. as if water were equally disseminated throughout the rock) might well not be the same, in terms of physical effects, as the values of BROCH. Thus, the observations given here must be viewed on their own, whilst recognising that multiple replications of results were obtained thereby implying that the observations are indeed real.

Implications for freeze-thaw

What do these results mean in terms of weathering in general, and for freeze-thaw in particular? The data indicate that in many field situations where only part of a rock is in contact with moisture (i.e. partly buried in the soil or situated within a melt rivulet), then the wetted section will experience a decrease in strength resulting from diminished elasticity due to loss of bonding strength. This loss of strength is reversible but, due to hysteresis, there is a very slow but progressive decrease of rock strength; a form of fatigue due to wetting and drying. Thus, as was suggested by MUGRIDGE & YOUNG (1983), rock failure could occur due to the action of wetting and drying alone. More realistically, it suggests that the process of wetting and drying may help weaken a rock, including the formation of microfractures that further decrease rock strength (CROOK & GILLESPIE 1986), such that lower stresses are required by other weathering mechanisms to cause breakdown. As MUGRIDGE & YOUNG (1983) found, the combination of wetting, freezing, thawing and drying caused more rapid breakdown of a shale than did wetting and drying alone. Thus, the two processes (freeze-thaw and wetting and drying) may work together.

This then suggests, based on the evidence obtained here, that the already complex process of freeze-thaw may be even more complicated than was hitherto thought. Freeze-thaw requires the uptake of water, for it is the water that actually freezes and thaws and in so doing damages the rock. If the presence of that water causes a diminution of rock strength, this would mean that when freezing then took place the effective pressures needed to cause failure would not need to be as large as have been calculated (HALL 1986 b). Rocks are known to have varying moisture contents in the field (TRENHAILE & MERCAN 1984; HALL 1988 a) and thus, by implication, varying resistance to stress. Thus, the moisture status may not only be important for deducing the actual freeze-thaw process, but it may also be pertinent for the determination of the available strength of that rock during the freezing process. In addition, these results indicate that wetting and drying of rock during thaw phases, a phenomena that may occur frequently in many cold regions during the summer months, will continue to weather that rock, further weakening it prior to the return of freezing during autumn-winter period. Breakdown which was presumed to be due to freeze-thaw weathering may owe much to the inter-operation of wetting and drying.

Conclusions

Wetting and drying would appear, based on the evidence so far obtained, to be a component of the total freeze-thaw process. The uptake of water, which will later be subject to freezing, changes the elastic properties of the rock due to its effects upon the bonding strength of that rock. This not only suggests that freezing may take place when the strength of the rock has been diminished, but also that, with multiple repetitions hysteresis will cause progressive weakening, and possible ultimate failure, of the rock. This latter factor is particularly relevant as rocks in many cold environments are subjected to wetting and drying during the warmer, rainy summer period, and so undergo weakening, prior to the onset of freezing during the autumn. In both laboratory experimentation, where samples are often subjected to wetting and drying, and in the field, where this is a frequent, natural occurrence, this element of rock weathering should not be forgotten. More research is needed to fully understand the mechanism(s) and effect(s) of wetting and drying, and thus what is its actual role within freeze-thaw.

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A LABORATORY SIMULATION OF ROCK BREAKDOWN DUE TO FREEZE-THAW IN A MARITIME ANTARCTIC ENVIRONMENT

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ABSTRACT

Results of freeze-thaw simulations on three large blocks of quartz-micaschist are presented. Three types of water to ice phase change were identified from temperature and ultrasonic measurements. It is suggested that the type of phase change results from a particular combination of rock moisture content, solute concentration, freeze amplitude, and rate of fall of temperature. The temperature at which ice thawed inside the rock (−0·7 to −1·9 C) was also found, and this indicates the possibility of freeze-thaw effects without positive temperatures. Approximately 80 per cent of the water that will freeze under natural conditions, in the Maritime Antarctic environment under study, appears to have done so by −6 .

KEY WORDS Weathering Freeze-thaw Simulation Ultrasonic testing Quartz-micaschist Maritime Antarctic

INTRODUCTION

Simulation of frost weathering on rocks has been a major avenue of investigation in the attempts to elucidate the mechanisms and rates of rock breakdown in cold environments (e.g. Brockie, 1972; Fahey, 1983; Potts, 1970; Swantesson, 1985; Thomas, 1938; Wiman, 1963). More recently (Williams and Robinson, 1981; McGreevy, 1982; Fahey, 1985), the implications of saline solutions in freeze-thaw have been considered, as too has the effect of the thermal properties of the rock (McGreevy, 1985). However, all of these experimental conditions have suffered from a number of inadequacies, notably '... experimental conditions are not particularly representative of actual periglacial environments'. (Fahey, 1983, p. 543). Until now true representation of the environment in question has been almost impossible since, apart from inadequate climatic data in many instances, there has been a lack of information on rock moisture content (McGreevy and Whalley, 1985) and the chemistry of that interstitial water (Fahey, 1985, p. 103). In addition, many studies have not considered the physical properties of the rocks that are being used making useful comparisons between material, even from the same lithologic group, very difficult.

In an effort to correct this inadequacy, a series of simulations have been undertaken as part of the British Antarctic Survey 'Fellfield Ecology Research Programme' (Walton and Hall, in press). In this project, based on Signy Island in the Maritime Antarctic, freeze-thaw activity is being studied not in isolation, but as part of a wider investigation encompassing chemical and biological weathering, mineral cycling, plant colonization, micrometeorology and pedogenesis. As a fundamental basis to the whole study there is already information on the microclimate (Walton, 1977, 1982), the geology (Storey and Meneilly, 1985), the field moisture content of the rocks (Hall, 1986a), the chemistry of the interstitial rock water (Hall *et al.*, 1986), the physical properties of

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the rock (Hall, 1987), and a study of rock fracture mechanics of quartz-micaschist (Hall, 1986b). Thus, for the first time, there is an adequate data base to allow representative simulations to be attempted.

As part of these simulations, use was made of ultrasonic testing upon the basis that ultrasonic pulses travelling in a solid material can be effects on the timing and character of the freeze. Whilst the monitoring of exotherms also provides evidence of the timing of the water to ice phase change it will be seen that the ultrasonic data give far more information (see also Hall, 1986c).

METHODOLOGY

Samples of quartz-micaschist were collected from a variety of environmental positions and at various times during the summer months on Signy Island (60° 43'S, 45° 38'W) in the South Orkneys (see Hall, 1986a or Walton and Hall, in press, for details of the study area and field procedures). The field moisture content of the rocks was measured, followed by porosity, microporosity, saturation coefficient, and water absorption capacity (Hall, 1986a). Using the Schmidt hammer, the point load test, and indentor tests, mechanical strength characteristics were established for the rock both normal to and parallel to the schistosity (Hall, 1987). From these data estimates of K_{Ic}, the stress intensity factor, were derived (Hall, 1986b). In addition, analysis of the interstitial rock water chemistry was undertaken to establish both the types and concentrations of salts within the rock (Hall *et al.*, 1986).

In this experiment three large blocks of quartz-micaschist were used. One block (ref. #65a) was totally immersed in water for 48 hrs to allow natural saturation (i.e. not under vacuum). To facilitate unidirectional freezing it was then encased, except for one b/c plane (240 × 600 mm in size), in a 0·3 m thick, waterproof, closed-cell, polyurethane jacket. The placing of a thick sheet of greased polythene around the rock whilst the foam was forming inhibited the foam bonding to the rock or wedging into all the surface irregularities. Thus it was possible to easily extricate the rock once the jacket had solidified thereby allowing for periodic removal during the experiment for weighing and resoaking. The exposed face had its schistosity normal to the direction of freeze penetration, this being the attitude of schistosity on cliff faces in the field. Details of the rock size and properties, together with those of the other samples, are given in Table I. This sample simulated unidirectional freezing in an open system of a continually saturated, or near-saturated, block (saturation here meaning the rock contained as much moisture as it could freely absorb during a 48 hr immersion in water).

A second block of rock (#65b) from the same outcrop was also saturated but, after being encased in a polyurethane jacket and fitted with temperature sensors, it could not then be removed for resoaking. Thin film platinum resistance sensors were placed on the exposed rock face and, in drill holes, at distances of 20 mm and 60 mm behind the open face. The exposed area was 240 × 650 mm with schistosity normal to the freezing plane.

Table I. Details of rock properties and measurements

Property		Rock sample 65a	65b	66
Dry Weight (experiment start)	(g)	4478·1	6857·4	9019·6
Wet Weight (experiment start)	(g)	4491·7	6887·8	9044·6
Change	(g)	13·6	30·4	25·0
Percentage Change		0·30	0·44	0·28
Exposed face	(m)	0·24 × 0·06	0·24 × 0·065	0·345 × 0·09
Porosity	(%)	1·46	1·46	1·93
Water absorption capacity	(%)	1·04	1·04	1·65
Saturation coefficient		0·71	0·71	0·86
Wet Weight (experiment end)	(g)	4483·6	6883·8	9052·8
Dry Weight (experiment end)	(g)	4476·8	6857·5	9019·2
Loss of material	(g)	1·3	gain of 0·1*	0·4*
Percentage loss	(%)	0·03	—	0·004

* Marginal weight gains due to precipitation out of NaCl from introduced solutions (65b and 66 only)

Finally, a third block (#66) was placed, with the face to be exposed to freezing, in water to a depth of 7 mm for 12 hours, to simulate limited exposure to moisture—as on a cliff face. It was then encased in polyurethane foam with temperature sensors on, and at distances of 20 and 60 mm behind, the exposed face. In addition, ultrasonic transducers were emplaced to monitor ultrasonic sound velocity, across the c-axis of the rock, 40 mm behind the 60 mm depth temperature sensor. The transducers, bonded to the rock with phenyl salicylate, comprised ceramic piezoelectric elements in stainless steel cases pulsing 10 times per second at a frequency of 54 kHz. The pulses were both generated and measured, with an accuracy of better than + −0·5 per cent, by a PUNDIT (Portable Ultrasonic Non-Destructive Digital Indicating Tester) manufactured by CNS Instruments (London) Ltd.

Knowing the pulse time and the distance between the transducers (91 mm) then pulse velocity (*Cp*) could be calculated. However, as during the running of the experiment the distance remained constant, any change in pulse time must reflect some change in the state of the rock. The pulses were running continuously but the transit time, generated by the PUNDIT, was logged by the controlling microcomputer at the same time as the temperature and humidity data (Walton and Hall, in press). Thus it was possible to obtain almost simultaneous readings of air and rock temperatures together with pulse time. Prior to the simulation start, the PUNDIT was tested to see if there were any effects due to subzero temperatures. Using the supplied reference bar a maximum deviation of −0·2 μs at −20°C was found. During the actual simulation the transducers were periodically removed for recalibration and the maximum drift encountered was 0·2 μs.

The samples were placed in a simulation chamber (Walton and Hall, in press) in which a computer controlled the climatic cycles to which the rocks were subjected, and monitored the chamber temperature and humidity, the rock temperature sensors, and the ultrasonic pulse time. The sensors were read at various specified rates between 30 seconds and 1 hour, variable throughout the cycle, and then stored on disk and printed out. The climatic cycles used, detailed in Figure 1, were based on micrometeorological data collected from reference sites on Signy Island (Walton, 1977, 1982 and personal communication; Collins *et al.*, 1975). This sequence allowed for comparison with earlier simulations (Hall, 1986c) but with two extra freezes, one with a 'fast', and one with a 'slow' rate of fall of temperature, down to −20°C to both simulate a winter freeze plus allow some comparison with the theoretical considerations of Walder and Hallet (1985). The simulation was run for 1705 hrs which gave 11 full cycles, each of which lasted almost one week, and realized 450 000 data readings. Initially all three samples were only wetted with deionized water but later in the experiment a solution of 0·25 M NaCl was given to rocks 66 and 65b.

RESULTS AND DISCUSSION

Knowing the surface, 20 mm, and 60 mm depth temperatures it was possible to calculate the internal rate of fall of temperature for blocks 65b and 66 for each of the freeze cycles and for different temperature ranges within those periods (Table II). Within the range of freeze amplitudes considered here (−3, −6, and −20°C), the rate of fall of environmental temperature appeared to be less significant than the fixed final temperature of the

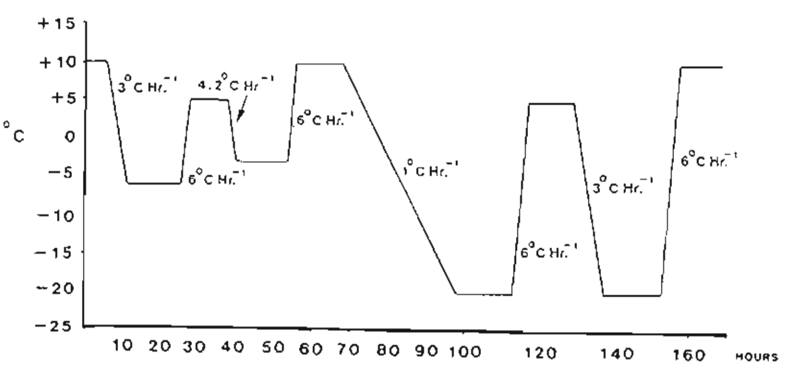

Figure 1. Details of the freeze–thaw cycles, together with the rates of change of temperature, used in the simulations

Table II. The rate of fall of temperature found for different freeze cycles and ranges within them

	Rate of fall of temp (°C hr^{-1})*			
	66		65b	
Range (rate)	20 mm	60 mm	20 mm	60 mm
0 to −3 cycle (4·5°C hr^{-1})	0·20	0·20	0·22	0·23
0 to −6 cycle (3°C hr^{-1})	0·47	0·48	0·42	0·44
0 to −20 cycle (1°C hr^{-1})	0·76	0·80	0·83	0·84
0 to −20 cycle (3°C hr^{-1})	0·98	1·03	0·96	0·95
0 to −3 range of −6°C cycle	0·71	0·72	0·77	0·79
0 to −3 range of slow −20°C cycle	0·94	0·88	0·83	0·81
0 to −3 range of fast −20°C cycle	1·58	1·43	1·00	0·95
0 to −6 range of slow −20°C cycle	0·94	0·95	0·97	0·96
0 to −6 of fast −20°C cycle	1·34	1·58	0·98	0·82
−6 to −20 range of slow −20°C cycle	0·71	0·75	0·79	0·81
−6 to −20 of fast −20°C cycle	0·86	0·92	0·9	1·00

* For cycles with *no* NaCl given

freeze in determining the rock's internal rate of fall of temperature. Consideration of Table II shows that although the −3°C freeze occurred at a rate of 4.5°C hr^{-1} the internal rate of fall of temperature was 0·2°C hr^{-1}, whilst the −20°C freeze with an environmental fall of only 1°C hr^{-1} showed an internal drop at a rate of 0·76°C hr^{-1}. Consideration of the same fixed value (−3 or −6°C) shows that the rate over that range increases with a decrease in final temperature. Even for the −20°C cycle, the 200 per cent difference between the 1°C hr^{-1} and 3°C hr^{-1} rates only resulted in a 29 per cent difference in internal rates of fall of temperature. It was also possible to calculate the variation in rate of fall of temperature, for each negative degree, for any one cycle (Table III).

The rate of fall of temperature, either environmentally (e.g. Battle, 1960) or within the rock (e.g. Walder and Hallet, 1985), has been suggested to be a major control on the form, and rate, of rock breakdown due to freezing. From this simulation neither environmental nor internal rates of change were found to be 'simple' in terms of their controlling influence. As stated above, within the limitations of the cycles used here, it appears to be the amplitude of the freeze that exerts the greatest influence on internal rock temperatures (Table II). The rate of change of temperature within the rock, as a product of the lowest temperature of the freeze cycle, is significant with respect to theoretical modelling. Walder and Hallet (1985, p. 342) state that 'Clearly, crack

Table III. An example of the variation in rate of fall of temperature within the rock for a given freeze cycle

	Rate of fall of temperature (°C hr^{-1})			
	Rock #66		Rock #65	
Range	20 mm	60 mm	20 mm	60 mm
0 to −1°C	0·87	0·80	0·91	0·91
−1 to −2°C	0·66	0·67	0·74	0·78
−2 to −3°C	0·64	0·68	0·68	0·70
−3 to −4°C	0·48	0·52	0·32	0·30
−4 to −5°C	0·22	0·23	0·33	0·40
For same rock with NaCl solution				
0 to −1°C	1·34	1·28	−6°C cycle at rate	
−1 to −2°C	0·71	0·70	of 3°C hr^{-1}	
−2 to −3°C	0·56	0·60	environmental	
−3 to −4°C	0·46	0·52	change	
−4 to −5°C	0·31	0·32		

growth for a fixed final temperature is greater, the lower the cooling rate'. The data from this simulation would tend to suggest that, in reality, the cooling rate of the rock is controlled by that 'fixed final temperature'. Environmental rates would have to be significantly lower than the 1°C hr^{-1} used here to result in the sort of rates (*c.* 0·025°C hr^{-1}), said to achieve the maximum damage (Walder and Hallet, 1985), within the outer *wetted* part of the cliff face. Certainly environmental rates of less than 1°C hr^{-1} are, from available data (Walton, personal communication, 1977), thought slower than any occurring within the present study area. The rate of fall of temperature would, however, certainly decrease with depth into the cliff such that, at some point (the value for which is unknown but from available data appears to be *c.* 0·6 m), the rate would be very slow. But, it is suggested that at this depth (*c.* 0·6 m) there would be negligible, if not non-existent, interstitial moisture. The cliffs have been shown (Hall, 1986a) to have very low moisture contents in the outer, obviously wetter, *c.* 0·2 m and so although rates of fall of temperature of the order required by Walder and Hallet (1985) are found, they must occur in the absence of moisture. This, then, would beg the point made by McGreevy (1982, p. 486) as to whether '. . . conditions favourable for breakdown . . . persist in nature . . .'. Without doubt the field observations point to very limited mechanical weathering of cliffs on Signy Island.

Whilst the rock temperature decreased linearly with time ($r = -0·98$ for -6°C cycle), as was suggested by Walder and Hallet (1985, p. 339), the rate of change of temperature also changes linearly ($r = 0·92$). This means that, for example, during the -6°C cycle, there is as much as a 332 per cent difference in rates (0·31 to 1·34 °C hr^{-1}, Table III) through time. The differential appears to be even larger when salt solutions are given to the rock. Thus, for any given point, in the outer wetter layer of the cliff, there is not found a constant rate of change and certainly the slow rates suggested by theoretical modelling are not attained at any time prior to the water–ice phase change having occurred.

Sample 65a, which continued to be resoaked throughout the experiment, gave a good indication of the loss of moisture through evaporation during the experiment run (Figure 2). In addition, it provided a comparison between 'saturated' and 'non-saturated' samples experiencing unidirectional freezing in an open system. It was

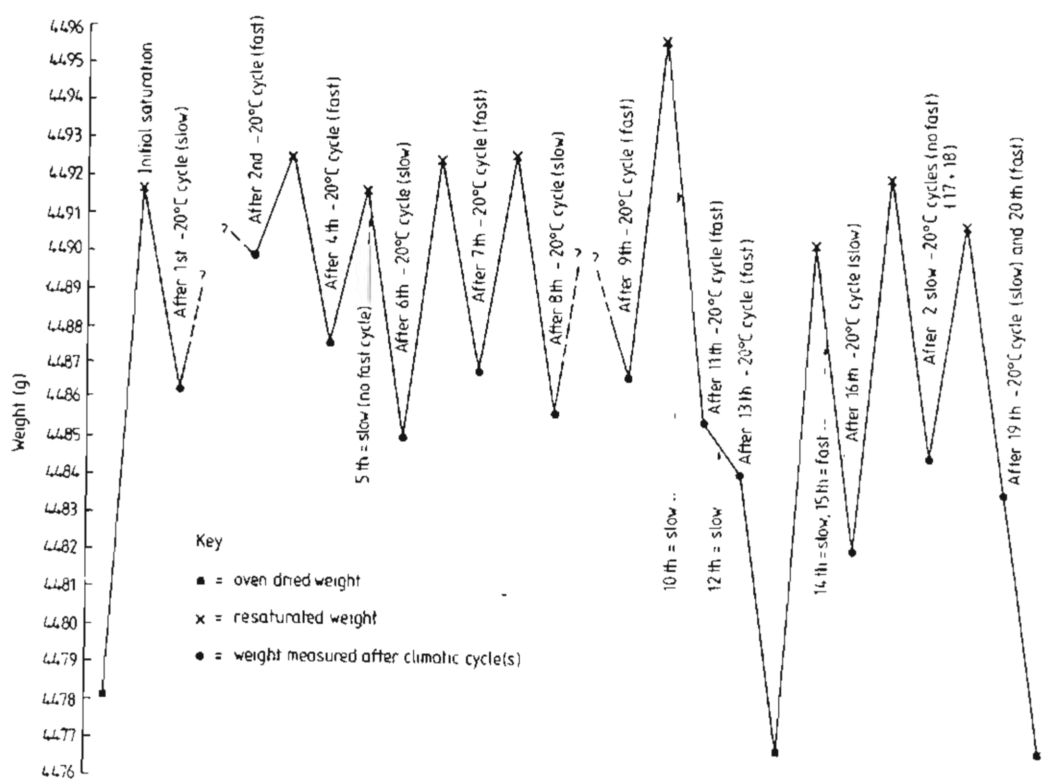

Figure 2. Graphs of weights for resaturated block No. 65a

found that, although block 65a and its exposed face were the smallest of the three, it nevertheless experienced the greatest loss of material (Table I). The actual amount of moisture lost varied, and may have partly been a function of experimental procedure, particularly as laboratory humidity varied considerably. However, it is this evaporative loss which may help explain the change in water mass and/or chemistry that seems to have affected the nature of the freeze in some cycles (see below). In addition, although individual values appear random there is a linear relationship ($r = -0{\cdot}79$) between the number of cycles and the weight of the block; as the number of cycles increased so the weight of the block decreased. This change cannot be related solely to the loss of material due to weathering as Figure 2 clearly shows that after the 13th −20°C cycle the dry weight was the same as after the 20th −20°C cycle. The exact cause for this change is unknown but may be related to the forcing apart of laminae at the rock face which facilitated a more rapid moisture loss.

Certainly, of these three blocks, it was the continually soaked 65a which showed the greatest amount of weathering despite having the smallest exposed face (Table I). This, albeit limited sample, suggests that, even with unidirectional freezing the amount of weathering is largely controlled by moisture availability. However, despite this, and as important as it is with respect to the applicability of simulations to nature (Hall, 1986a), it is most noticeable that the amount of weathered material was very limited (Table I). In addition, no saline solution, which has been argued to increase the efficacy of frost action (Williams and Robinson, 1981), was given to sample 65a. Thus, the degree of saturation is seen as a very important parameter in the control of weathering rates.

Rock samples with temperature sensors (65b and 66) showed exotherms indicative of the water to ice phase change (Table IV). In addition, the change in the extent of supercooling due to the introduction of a 0·25 M NaCl solution can also be seen. The ultrasonic pulse times obtained for sample 66 during these freeze periods also proved of great value. Accepting that there is a decrease in pulse time due to the transformation of water to ice, three major forms of phase change were recognized. The first comprises a rapid, large scale transformation of water to ice, with a large exotherm, subsequent to extensive supercooling of the interstitial rock water (A in Figure 3). This type of freeze was observed for the −6°C cycle and the slow (1°C hr^{-1}) −20°C cycle. The second type of freeze consisted of a small exotherm at the 20 mm depth followed by a (relatively) slow, progressive freeze (B in Figure 3) and was characteristic of the fast (3°C hr^{-1}) −20°C cycle. The final form of transformation (C in Figure 3) occurred for all cycles when 'high' saline conditions were present (*c.* > 0·5 M) and comprised a progressive but relatively rapid freeze (i.e. somewhat intermediate between the other two forms) with *no* sign of supercooling or an exotherm shown by the temperature sensors.

It is suggested that in a simulation of this kind it would not have been possible to discern the detail of what was happening inside the rock, with respect to the phase change, without the ultrasonic readings. For example, in the case of the freeze with the higher saline solutions it would have appeared as if no phase change had taken place as no exotherm was apparent. In the other two examples there certainly would have been a freeze indicated but its nature would not have been discernible. Other techniques such as strain gauges (e.g. Douglas *et al.*, in press) could prove useful (see below) but would not resolve the detail obtained via ultrasonics. The ultrasonic pulse time, together with the parallel temperature readings, also indicated at what temperature the phase change back to water took place during the thaw cycle. In the initial runs (i.e. prior to the introduction of NaCl) this began when the 20 mm depth temperature was *c.* −0·7°C and the 60 mm at *c.* −1·1°C, and at approximately −1·1°C and −1·9°C respectively after the introduction of the saline solution. The recognition of thawing at these temperatures indicates that there need not be positive temperatures for freezing and thawing of interstitial rock water to take place (Hall, 1980).

It was observed that the ultrasonic pulse time at the end of the freeze phase was the same for both the −3°C (when freezing occurred) and −6°C cycles (17·1 μs), whilst the two −20°C cycles also had comparable times (15·9 μs). This would imply that the physical end result in terms of the air–water–ice mix within the rock was the same for each of the two pairs of end temperatures, but that the −20°C group were different to the −3°C and −6°C group. Accepting that shorter pulse times are associated with a greater ice presence (Miles and Cutting, 1974) then this would imply, as might be expected, that more water was frozen at −20°C than at −3° or −6°C. With a linear relationship of ice increase to pulse time decrease, then the amount of water frozen for the −20°C cycle is only approximately 18 per cent more, not a great difference for the decrease in temperature between −3°C and −20°C. In other words, this implies that certainly by −6°C (accepting that freezes

Table IV. Examples of some typical thermal jumps associated with the water to ice phase change monitored in rock 66

1 cycle	2 Air t°C	3 Surface t°	4 Rise of	20 mm depth from	to	Δ	60 mm depth from	to	Δ	time for t to rtn hr:min:sec
−6	5·7	−3·6	+0·1	−2·6	−0·8	1·8	−2·1	−1·3	0·8	1:10:4
slow −20	−7·6	−4·9	+0·2	−3·3	−1·1	2·2	−2·8	−2·2	0·6	0:32:01
fast −20	−12·6	−7·0	+0·1	−2·7	−1·8	0·9	−1·7	−1·4	0·3	0:06:58
−6	−5·9	−3·9	+0·1	−3·3	−2·5	0·8	−3·0	−2·6	0·4	0:28:01
−6	−5·9	−2·8	+0·1	−2·6	−1·1	1·5	−2·5	−1·8	0·7	1:21:33
slow −20	−8·2	−6·2	+0·2	−3·2	−1·4	1·8	−3·0	−2·3	0·7	0:34:05
−6	−5·7	−4·0	+0·1	−3·0	−1·2	1·8	−2·8	−2·2	0·6	1:02:29
slow −20	−7·8	−5·2	+0·3	−3·5	−1·3	2·2	−3·1	−2·6	0·5	0:30:00
fast −20	−8·8	−6·4	—	−1·5	−1·1	0·4	—	—	—	0:05:50
−3	−2·6	−1·4	*0·1	−1·1	−0·7	0·4	−1·0	−0·7	0·3	1:21:45
slow −20	−6·7	−4·7	+0·2	−3·3	−2·6	0·7	−2·9	−2·2	0·7	0:44:10
fast −20	−6	−3·0	—	−1·0	−0·7	0·3	—	—	—	0:04:05
−3	−2·6	−1·5	+0·1	−1·0	−0·7	0·3	−1·0	−0·7	0·3	0:25:01
After introduction of 0·25 M NaCl solution										
−6	−6	−4·6	+0·3	−3·4	−1·7	1·7	−3·3	−2·7	0·6	1:07:14
slow −20	−8·0	−5·6	+0·1	−3·9	−2·1	1·8	−3·6	−3·0	0·6	0:34:51
fast −20	−9·5	−4·8	—	−2·0	−1·7	0·3	—	—	—	0:03:12
−6	−5·7	−5·3	+0·2	−3·9	−2·8	1·1	−3·9	−3·3	0·6	0:52:06
slow −20	−8·0	−5·1	—	−4·1	−3·4	0·7	−3.9	−3·4	0·5	0:20:58
−3	−2·9	−2·5	+0·1	−2·2	−1·4	0·8	−2·0	−1·8	0·2	1:52:09
slow	−7·0	−4·7	+0·1	−2·5	−1·3	1·2	−2·1	−1·9	0·2	0:19·57

occasionally did not occur for the −3°C cycle) then *c.* 80 per cent of the water that is likely to freeze in any normal environmental condition on Signy Island (i.e. temperatures are rarely below −20°C) has frozen. How much this is of the total water within the rock is not known but it is highly likely that some tightly bonded water remains unfrozen.

Another interesting observation regarding the end pulse velocities is that they are achieved some time before the end of the freeze cycle: 1 hr 14 min for −3°C, 2 hrs 30 min for −6°C, 6 hrs 55 min for the fast −20°C, and 11 hrs 45 min for the slow −20°C. In addition, the actual temperature at optimum pulse time is higher than the final temperature achieved (i.e. optimum conditions are attained *before* the coldest stage is reached): that for the −3°C cycle being *c.* −2·4°C, the −6°C cycle being *c.* −5·0°C, the slow −20°C cycle being *c.* −18·7°C and the fast −20°C being *c.* −17°C. These data imply that after a certain point is reached no more freezing of water takes place despite the continuation of negative temperatures. Thus, at −3°C and −6°C, where it is known more water *is* available to freeze (as shown by the subsequent −20°C cycle) it in fact cannot do so at this temperature. Thus, the suggestion of Walder and Hallet (1985) that ice growth will continue for sustained temperatures in the range *c.* −4 to −15°C, if ample water is available, does not appear to operate here. This is not to say that their hypothesized mechanism does not function but, more likely, that there is not enough water in the Signy rocks and that which is available is relatively tightly bonded thereby requiring the lower temperature before it can freeze. In turn, this again implies that by *c.* −6°C almost all the water that is available

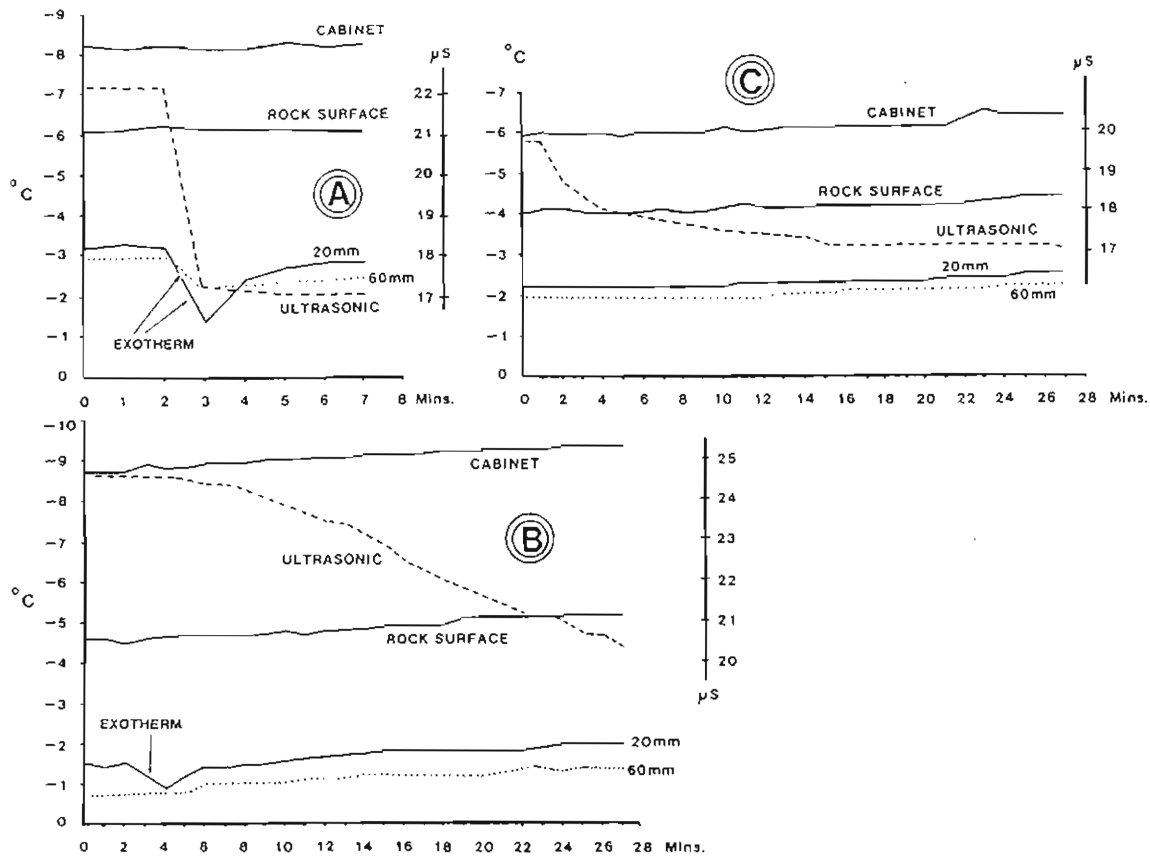

Figure 3. Graphs of actual readouts to show the three forms of phase change found during the simulations. A exemplifies the rapid large-scale phase change, B the relatively slow, progressive freeze, and C is characteristic of high saline conditions

to freeze has done so and that by c. − 18°C all the rock water that will freeze under natural conditions on Signy Island is in a frozen state. These results are similar to the findings of Thorpe (1983) where, in freshly placed concrete, at − 3°C +/− 90 per cent of available moisture had transformed to ice and a drop to − 45°C only increased the amount to 97 per cent.

Ultrasonic pulse times were obtained for the dried rock samples prior to the experiment start (41·0 μs), after the initial wetting of the exposed face (37·0 μs) and then at the experiment end when the rock had once more been dried (40·2 μs). Subsequent to the 9·7 per cent decrease in pulse time after the initial introduction of water, there was a further diminution in pulse time, to 28·6 μs, later in the experiment after more water had been given by spraying and, presumably, interstitial water was pushed further into the rock closer to the transducers. With the transformation of the water to ice, and the filling of more void space due to the volume change, pulse time decreased significantly (at − 6°C = 17·1 μs and at − 20°C = 15·9 μs). However, what may be important is that the start and end dry rock pulse times show a slight (1·95 per cent) decrease rather than, as was first anticipated, an increase. It was thought that internal damage to the rock would produce more pore space which would therefore increase pulse times; the hope being that the ultrasonics would show evidence of otherwise undetected internal disruption to the rock. But, as stated, rather there was a decrease in pulse time which implies a 'diminishing' of voids within the rock. The only explanation for this, apart from the obvious possibility of operator error, is that of the introduced salts being precipitated out, during drying, within the interstices of the rock which thereby caused the marginally slower pulse transmission times. This explanation is considered probable in the light of the very limited weathering found for all of the large blocks, both in the laboratory and the field. The overall implication of the evidence is that considerably more freeze events are required before noticeable damage to the rock, internal or on the rock face, is obtained and that this simulation, in that respect, was of too short a duration. The corollary of this, though, is that in the field it would be a considerable length of time before damage to cliff faces is effected: an observation suggested from other lines of study as well (Hall, 1986a, 1986b, and Hall *et al.*, 1986).

The ultrasonic indication of a rapid phase change (A in Figure 3) during the slow rate of fall of temperature towards $-20°C$ is of great interest. Davidson and Nye (1985) have shown, by means of a photoelastic study of ice growth in simulated rock blocks, that the temperature environment of the block affected the way the ice grew. With pre-cooling to $-10°C$ followed by surface cooling they found that this resulted in supercooled water that suddenly froze, a situation very analogous to the slow $-20°C$ freeze of this simulation. A rapid freeze in their experiment gave results very similar to that of the fast $-20°C$ freeze in this simulation (B in Figure 3) during which there was progressive freezing which Davidson and Nye (1985, p. 146) describe as '. . . a smooth ice front which progressed evenly down the slot . . .'. This recognition of a sudden freeze of water within the rock is pertinent insofar as it conflicts with the hypothesized model of Hallet (1983) which argues for progressive freezing with a slow (rock) rate of fall of temperature but, at the same time, offers evidence in possible favour of Hodder's (1976) theory of cavitation-induced nucleation of ice.

Hodder (1976) suggested that where homogeneous nucleation were possible then a shock wave would be produced which could be the cause of rock damage, rather than the actual phase change. This is a novel idea and one which, to the author's knowledge, has never been considered in any simulation study. The possibility of this mechanism had not been foreseen within the present simulation but was suggested rather by the data during the course of the experiment, and so no sensors for evaluation of this hypothesis had been built in to the programme. However, by means of a crystal pick-up held in contact with the rock, and linked to a y/t recorder, it was possible to obtain some preliminary results (Figure 4). Although no acoustic emission concomitant upon phase change was detected, two peaks that could not have been due to electrical disturbances were recorded some time (2 hrs 7 min and 6 hrs 31 min) after the freeze took place. Their significance is unknown, but experiments employing acoustic emission sensors and fibre optic crack detection equipment are currently in progress to further evaluate Hodder's hypothesis.

Walder and Hallet (1986) argue very strongly that it is not the volumetric expansion upon phase change that causes rock breakdown but rather water migration to a freezing front such that ice growth in excess of the volume change occurs, and it is this which causes crack propagation. Without doubt some of the ultrasonic

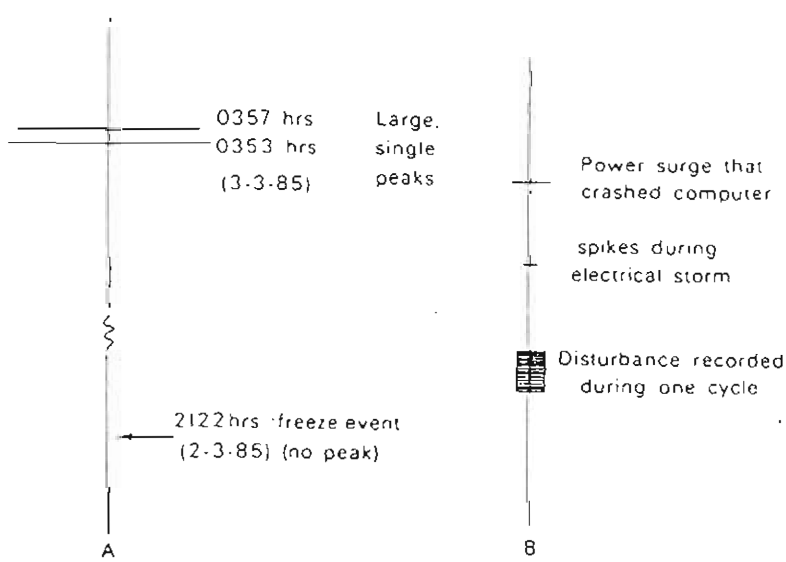

Figure 4. Examples of readout from y/t recorder showing typical peaks. Output A indicates two peaks possibly indicative of rock damage during a slow -20 C cycle. Output B gives examples of electrical disturbances (to same scale as A) showing their relatively small amplitude

evidence supports this model (B and C in Figure 3). However, two lines of evidence from this study qualify the general use of Hallet's (1983) model. Firstly, the less-than-saturated state of the rock appears to have produced far less weathering than was found for the saturated sample, a possibility that Walder and Hallet (1986) do recognize. However, moisture levels in the rock during this simulation were higher (> 50 per cent) than are found under natural conditions (Hall, 1986a) and freezing certainly occurred in a progressive manner, and yet weathering products were minimal. Secondly, the ultrasonic velocities do not appear to indicate significant internal disruption, and, whilst fully accepting the occurrence of progressive freezing, it must be recognized that rapid phase changes are also found. Thus, although the y/t record has failed to so far indicate any such event, the possibility of crack growth by, what Walder and Hallet (1986) term, a series of bursts associated with each freezing episode, must still be considered possible.

The measured rate of change of temperature at depth will be affected by the nature and duration of the freeze. Depending upon the temperature gradient, the rock experienced either a slow, progressive freeze with minimal supercooling prior to the initiation of the phase change, or there is 'severe' supercooling followed by a sudden, massive freeze. In the latter case there is a 'plateau' period (see Hall, 1986c, Figure 4) the length of which is a product of the amount of water and the steepness of the temperature gradient. These two types of freeze occurred in the first part of the simulation when only distilled water was given to the rock. However, and this is a point other simulation studies have frequently not appreciated, this is by no means intimating a 'pure water condition'. In fact, available data (Hall *et al.*, 1986) would suggest that there could be an initial NaCl molarity of anything up to 0·5 M due to the presence of the field salt status being preserved in the rock as a result of precipitation out during drying. There was a freezing point depression, as would be expected, upon the introduction of 0·25 M NaCl to the rock in the order of 0·8°C which is very close to the calculated depression of 0·9°C (freezing point depression = Van't Hoff Factor × [Molality × Cryoscopic Constant] = 2 × [0·25 × 1·85]).

Information on the − 3°C cycle is less definite. On some occasions an exotherm is recognized (Table IV) and the ultrasonic reading indicates a progressive phase change. However, for an apparently otherwise similar event there is sometimes no freeze indicated by either temperature or ultrasonics, the appearance being that of supercooled, unfrozen interstitial water. The occurrence or not of a freeze may be a product of the amount of water available in the rock for, as has been noted, there was loss due to evaporation. Remembering that it will not be a pure water system, then whether a freeze does or does not occur will depend upon a suitable amount of water, without a high salt content, being present. If there is a very large amount of water present then there will be a great deal of latent heat to be removed and a freeze to only − 3°C may not be sufficient to allow ice growth. Equally, if there is a high salt concentration then the freezing point depression may preclude a phase change. Freezing will only occur when the mass of water and its salinity are within certain boundaries.

The − 6°C cycle, on the other hand, responded in a very similar manner to the slow − 20°C freeze, including the change in response with the introduction of the saline solution. Temperatures in this cycle, as compared to the − 3°C cycle, went sufficiently low to allow supercooling, although it was noticeable that air temperatures were sustained at *c.* − 6°C for over five hours prior to any freeze and the rock surface was at a negative temperature for a similar duration. In other words, a period of time at the fixed end point (in this case 5 hours) was required prior to freeze initiation. If this time were available, then a gradient slowly developed that produced a rapid, extensive freeze once phase change began.

It is felt that the results of this simulation have important implications with respect to both field and laboratory studies. Most importantly, the frequent criticism that the experiments do not reflect the conditions operative in nature (McGreevy, 1982; McGreevy and Whalley, 1982) and that adequate cognisance of internal controls such as rock properties and moisture content (Douglas *et al.*, 1983; McGreevy and Whalley, 1982; Trenhaile and Mercan, 1984) are not met, is not valid here. The experiments are based on a knowledge of field conditions and so the simulation results are applicable to the Signy Island environment, and, possibly, to general theory.

The limited degree of weathering corresponded to that observed for cliffs in the field (Hall, 1987). The weathering rates of these cliff-analogous, unidirectional freeze penetration, open systems were orders of magnitude less than was obtained for small rock pieces subject to omnidirectional freezing (Hall, submitted). Thus, it is suggested, that the bulk of simulations to date reflect only the weathering of loose blocks and are not applicable to bedrock. More important, though, is thought to be the information on the timing and nature of the freeze. There has been much discussion (see McGreevy and Whalley, 1982) on the temperature at which freezing takes place in rocks and building materials. This has largely been seen to be the product of pore size, with the effects of salts and/or clays also being recognized. Unquestionably pore size has a strong control but the simulations here suggest that solute content exerts an influence on both the temperature at which the freeze occurs *and* the type of freeze.

Consideration of field and laboratory studies to date indicate that rocks are considered to freeze at certain 'critical' temperatures (McGreevy and Whalley, 1982) and that duration of temperature at a certain level may be important (Lautridou, 1971). However, it has been shown here that the lowest temperature of the freeze cycle exerts a powerful influence on the nature of the freeze. It is not always clear from earlier studies what the freeze amplitude was when the thermal peak denoting the water to ice phase change appeared (e.g. Douglas *et al.*, 1983; McGreevy and Whalley, 1982). The results obtained here may help to explain why Lautridou (1971) suggested the duration of negative temperature to be significant (-5°C for 10 hours), for consideration of the present data shows that for a freeze of this amplitude a negative temperature needed to be sustained for *c.* 6 hours before there was a sufficient thermal gradient to initiate freezing. What is not known from earlier studies is the actual nature of the freeze, and that, via ultrasonic readings, is here seen to vary as a function of freeze amplitude, rate of fall of temperature and solute content. Thus, it is suggested that previous considerations may have been too simplistic.

A factor which is now thought to greatly complicate experimentation, and which may help to explain some of the variations in results of previous experiments, is that of solute content. Many studies have used natural rocks obtained from the field. Deionized water has then been used in the experiments and the implication is that the results are associated with 'pure' water. However, as already noted, this need not be so. Solutes trapped within the rock during drying are remobilized upon saturation and the system is no longer 'pure'; lack of data on solute content to date has negated any quantitative appreciation of type and amount of solutes that might be present. As has been shown here, solute content can have such an effect on both the timing and nature of the freeze that the results of many previous simulations should be viewed with some caution. This is particularly the case with small amplitude freezes (*c.* -3°C) which have been considered controversial in their geomorphic significance (McGreevy and Whalley, 1982), which are seen here, in some combinations, to be effective. In addition, with the higher solute contents in this simulation the sensors did not record an exotherm and, particularly as many other experiments have used similar techniques, it is possible that without the aid of ultrasonic data these experiments may well have missed the actual temperature and timing of the phase change.

It is so far not clear what precisely is the actual mechanism causing rock weathering, since the same rock may experience different types of freezing as a result of particular combinations of freeze amplitude, freeze rate and solute content (pore size being seen as a constant as only one rock type is being used). Broadly there is the dichotomy of a slow progressive freeze and that of a sudden, extensive freeze. The two are quite different and may or may not exert different effects on the rock. The sudden freeze is such that the idea of cavitation-induced nucleation of ice (Hodder, 1976) must be considered but the initial results within this simulation are by no means conclusive. However, both types of freeze may satisfy the theoretical model of Walder and Hallet (1985). The slow phase change certainly appears to offer some evidence that the type of progressive freeze they describe should be most destructive, albeit at rates of change of temperature greater than the intimated optimum. On the other hand, the slow rate of freezing of any water not initially frozen in the case of the rapid phase change may also be conducive to Walder and Hallet's model. Both, though, are probably constrained by the amount of water present, although long-term small stresses may induce as yet unquantified effects of rock fatigue. However, the experiments on the three blocks used here indicated that greater weathering rates were found with higher moisture contents for the same conditions. All of the simulation results are meaningless unless they can be applied back to the field. In this instance, it is known that the cliffs experiencing unidirectional freezing have low moisture contents (Hall, 1986a) and so, whatever the actual destructive mechanism, the rates of breakdown found for samples 65b and 66 are representative of the field situation.

Ultimately, it must be the applicability to the field which justifies a simulation of this kind. Whilst there is some value in the standardization of methods and experimental conditions, this may not reflect the field situation. In fact, it may well be that because most simulations have not been founded on a firm data base of field conditions, their meaning and potential comparisons are not viable. It is here suggested that, except for experiments with deliberate simplifications seeking to understand physical processes by varying only one factor whilst holding the others constant, the laboratory simulation should be a close approximation to nature and that as many parameters as possible should be considered. Whilst this might initially lead to a vast amount of, apparently, unrelated simulations, the ultimate results may show some recognizable themes or factors and so our knowledge of this complex topic will grow in a realistic manner. Standardization of methods and conditions (McGreevy and Whalley, 1982), whilst it may appear scientifically sound, may not help solve what happens in nature, with its multitude of variations, but simply what happens in standardized experiments.

CONCLUSIONS

A number of major points emerge from this investigation of freeze–thaw effects on quartz-micaschist in the Maritime Antarctic environment of Signy Island. It may be that the results are applicable to some other rock types and to other cold environments, but until further experiments are carried out the points noted here are meant only to reflect the Signy situation.

For the same final temperature (-20°C in this instance) the rate of fall of temperature will affect the timing and nature of the phase change, but the freeze amplitude is more important than the environmental temperature change where different end temperatures are concerned. Whilst some freezes are slow and progressive, others are rapid and extensive such that progressive freezing need not always occur with sustained subzero conditions. Ultrasonic data indicate that, of the water that will freeze at -20°C, 80 per cent has done so by *c.* -6°C. In addition, it also shows that the ice to water phase change takes place between $-0{\cdot}7$°C and $-1{\cdot}9$°C. In the case of low amplitude freezes (to *c.* -6°C), there is a need for the temperature to be maintained for a period of time before freezing of the water will occur. If that period of temperature maintenance is not achieved then the phase change to ice will not take place.

In all instances, including that of a unidirectional freeze in an open system, the degree of rock saturation was found to affect the amount of weathering that will take place. In the case of unidirectionally frozen quartz-micaschist the weathering rate is very slow. Although the exact mechanism causing the rock breakdown is still uncertain, it is recognized that it will vary for the same piece of rock as a function of the interaction of freeze amplitude, environmental rate of fall of temperature, moisture content, and solute concentration. Although certain combinations of conditions give rise to phase changes that appear conducive to the hypothesis of cavitation-induced nucleation of ice, the preliminary evidence to-date does not indicate its occurrence. However, some acoustic emission may take place subsequent to freezing. For the length of the simulation used here, ultrasonic evidence suggests that only extremely limited internal damage to the rock could have been effected.

It must be repeated that the observations reported here are specific to quartz-micaschist within a Maritime Antarctic environment. However, it is thought that many of the broad concepts, if not the exact values, derived from this study may be generally applicable. An attempt has been made to overcome many of the shortcomings of earlier simulations but, whilst this has been largely achieved, there is need for further refinement. Continued field data collection, which is running concurrently with the laboratory simulations, allows the characterization of the complex process of natural weathering in more realistic terms than hitherto.

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WEATHERING BY WETTING AND DRYING: SOME EXPERIMENTAL RESULTS

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ABSTRACT

A series of experiments on sandstone and dolerite was undertaken in an attempt to better understand the wetting and drying weathering process. As rock samples are frequently subjected to wet–dry cycles within the simulation of other weathering mechanisms (e.g. freeze–thaw), three common methods of moisture application were used and the influences of these evaluated. It was found that the method of moisture application could affect the nature of the weathering products resulting from wetting and drying. It was also observed that there were changes in the internal properties of the rock (e.g. porosity/microporosity) and that these could influence the synergistic operation of other weathering processes. Although not all of the observations could be explained, it is apparent that wetting and drying has both a direct and an indirect effect on the weathering of rock that has not been taken into account in simulations. Greater cognizance needs to be given to the role of this process both in the field and in laboratory simulations.

KEY WORDS weathering; wetting and drying; simulations; sandstone; dolerite

INTRODUCTION

Temporal and spatial variability of rock moisture in cold regions are recognized as major factors affecting the nature and extent of rock weathering, (Hall, 1986, 1988a, 1991, 1992; Thorn, 1988, 1992). In addition to exerting a direct control on mechanical and chemical weathering processes, the fluctuations in rock moisture can themselves cause weathering (Pissart and Lautridou, 1984), so-called 'wetting and drying'. The process of wetting and drying is not well understood (Nepper-Christensen, 1965; Ollier, 1984; Hall, 1988b) but is recognized as affecting a whole range of rocks including those that do not have a clay component (Felix, 1983). A number of studies, on a variety of rock types, have suggested that moisture fluctuations cause weathering as a result of the rock expanding during take up of water and its inability to return to the original dimensions upon losing moisture (Nishioka and Harada, 1958; Nepper-Christensen, 1965; Venter, 1981; Felix, 1983; Pissart and Lautridou, 1984; Hames *et al.*, 1987). In addition, it has been shown that high moisture contents diminish rock strength (Brock, 1979; Dube and Singh, 1972) and that, through time, wetting and drying cycles can decrease the bonding strength of the component minerals such that, ultimately, there is a decrease in overall rock strength and possibly even failure (Pissart and Lautridou, 1984; Hall, 1988b). Furthermore, recent studies by Hall (1991, 1992) have suggested that fluctuations in moisture content, without the rock ever drying fully, result in wetting and drying taking place in a zone below the surface, where the effects may not be immediately apparent. This, in turn, can lead to rock failure that has previously been ascribed to other processes (e.g. freeze–thaw), or it can operate synergistically with other weathering processes, both facilitating their operation and abetting their overall effect. Recent studies by Haneef *et al.* (1993a, b) have reported on the applied aspects of wetting and drying interaction with pollutant acids and gases plus the significance of different rock associations. Finally, although it may be that wetting and drying is a relatively slow process, nevertheless the number of wet–dry repetitions in many areas,

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even polar areas, can be very high and may far exceed the number of 'effective' freeze–thaw cycles (Hall, 1992).

Although laboratory studies have attempted to elucidate the processes involved in wetting and drying, little information is available regarding actual rates of breakdown. These data are, however, very important for two reasons. First, information is needed regarding the rate of weathering by wetting and drying in its own right together with an understanding of what the nature of that rate is (i.e. linear, positively or negatively exponential, etc.). Second, the information is of paramount importance for assessing the true effect of other processes, such as freeze–thaw. Such processes are dependent upon the presence of moisture for their operation, and clearly moisture content varies through time or as a function or experimental procedures in laboratory experiments, so it is necessary to understand the effects of these moisture fluctuations if the role and rate of the process under study are to be ascertained. This second point is particularly important with respect to laboratory studies when frequently the aim is to investigate a single mechanism by attempting to filter out all other effects.

For cold regions, freeze–thaw is frequently cited as the major operative mechanism and large numbers of laboratory experiments have been undertaken in an attempt to ascertain either the mechanism involved or the rate of breakdown. Common to almost all of these experiments has been the non-natural apportionment of moisture. In other words, in the absence of field data on rock moisture content (McGreevy and Whalley, 1985) the rocks have been placed in trays where they were fully covered by water, partially covered or simply wetted (i.e. by spraying or having a limited amount of water poured over them). The rocks are then subjected to freeze–thaw cycles of various kinds with frequent drying events for reweighing, etc. Clearly, the rocks are also being subjected to wetting and drying, the effects of which are not known either in terms of process or rate of breakdown.

In this study, a series of experiments was undertaken in an attempt to elucidate the effects due to wetting and drying resulting from the apportionment of water to rock samples in the manner used in many freeze–thaw and/or salt weathering laboratory studies. In addition to indicating the weathering effects with respect to laboratory procedures, the resulting data also have applicability to field situations in which the manner of rock wetting is analogous to that used in the laboratory (see below).

EXPERIMENTAL PROCEDURE

Sandstone and dolerite samples used in the experiments were from areas where frequent moisture changes to the rocks were observed to be operative (sandstone from Drakensberg Mountains, South Africa; dolerite from Livingston Island, South Shetland Islands, Antarctica). Three procedures, common to many frost and salt weathering experiments (see McGreevy and Whalley, 1985), were utilized for providing moisture to the rock specimens: the covering of the rock sample by deionized water, half-covering the sample with water, and spraying the sample with a fixed amount of water (in this instance 50 ml). For each of these moisture allocation procedures, three samples of each rock type were used; all samples were roughly oblong with dimensions of approximately $8 \times 4 \times 3$ cm. The samples were thus reasonably comparable with respect to size and shape within and between rock types. Prior to beginning the experiment, all the samples were saturated by immersion in water until a stable mass was reached; it took two days to obtain this saturated mass for the dolerite and one day for the sandstones. Following saturation, the samples were dried for 48 h at 105°C and weighed again. Knowing the saturated and the dry mass of the samples it was possible to monitor changes in both mass and percentage saturation throughout the experiment. The covered and half-covered samples were weighed in the morning, put in their trays of water and then left for 3 h. They were then removed, weighed and left to dry at room temperature (*c.* 19°C) until the following morning, when the procedure was repeated The sample to be sprayed with water was first weighed, then sprayed with 50 ml of water and left for 10 min in a closed glass container prior to being weighed again. The wetted sample was left in the closed glass container for 3 h and then subjected to the same procedure as the other two samples, described above. The 3 h wetting period was used partly for logistical reasons but is was also felt that a 3 h wetting followed by a 21 h drying is possibly realistic with respect to many field situations, e.g. as could be produced by the wetting of rocks at the margins of meltstreams during peak flow, the melting of snow that settled on warmed rock and then evaporated, or short rain events followed by sun or wind that caused moisture loss. At weekends the sample was left to dry for three days (Friday to Sunday) and so a lower mass is seen on the graphs after every four day (Monday to Thursday) period.

RESULTS AND DISCUSSION

The results for each of the three moisture regimes can be considered under four main headings:

(1) material weathered free from the rock as expressed by changes in sample dry mass;
(2) weathering of the rock affecting the internal structure as shown by changes in water holding ability;
(3) the influence of variation in experimental procedure duration on the degree of saturation; and
(4) the significance of the findings with respect to weathering rates and processes in both the field and the laboratory.

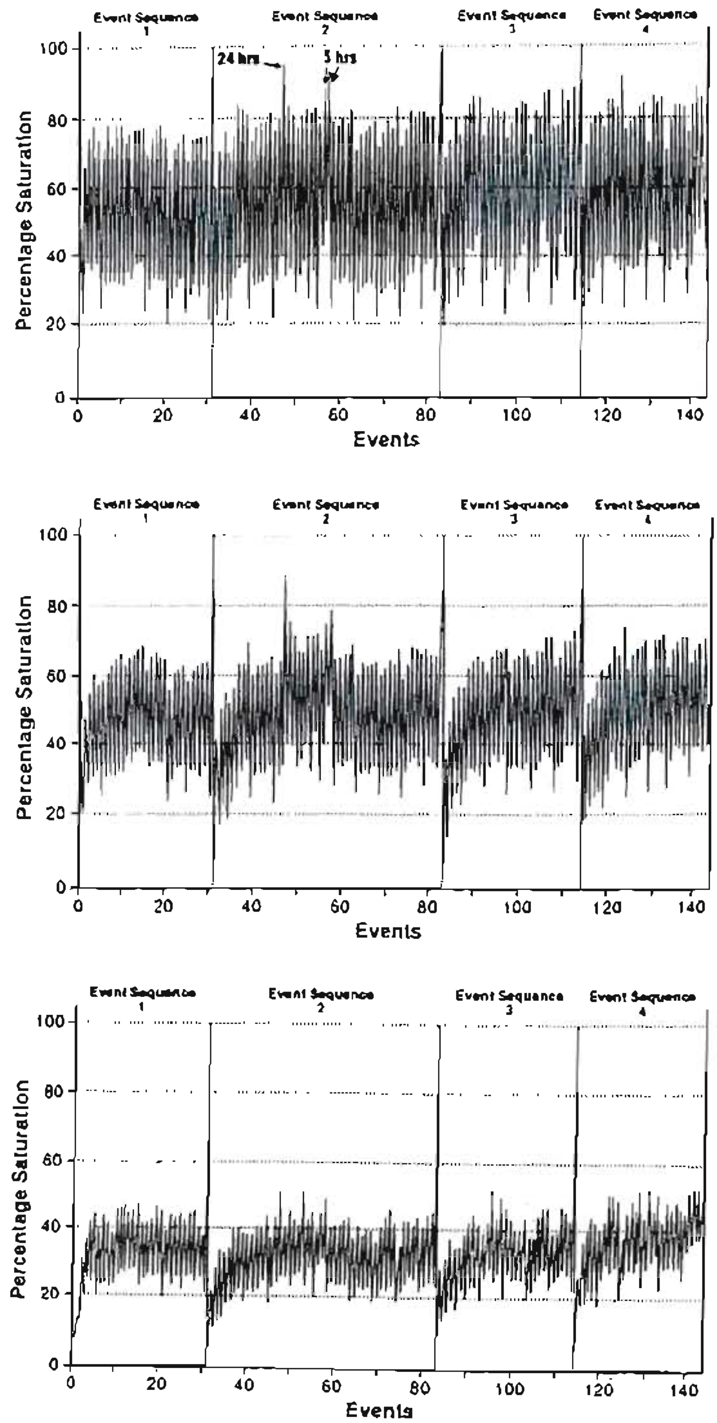
Responses with respect to percentage saturation between the three samples in each of the moisture allocation procedures were very similar indeed. Correlation values within each moisture set were all greater than $r = +0{\cdot}9$; values ranged between a low of $r = +0{\cdot}9088$ (samples 1 and 2 of Elliot sandstone, totally covered) and a high of $r = +0{\cdot}98893$ (samples 1 and 3 of dolerite, totally covered). Thus, the data presented below are for sample 1 in each of the three rock types in each of the moisture allocation procedures as, with such consistently high correlation values, these can be considered representative of the sample group in each instance.

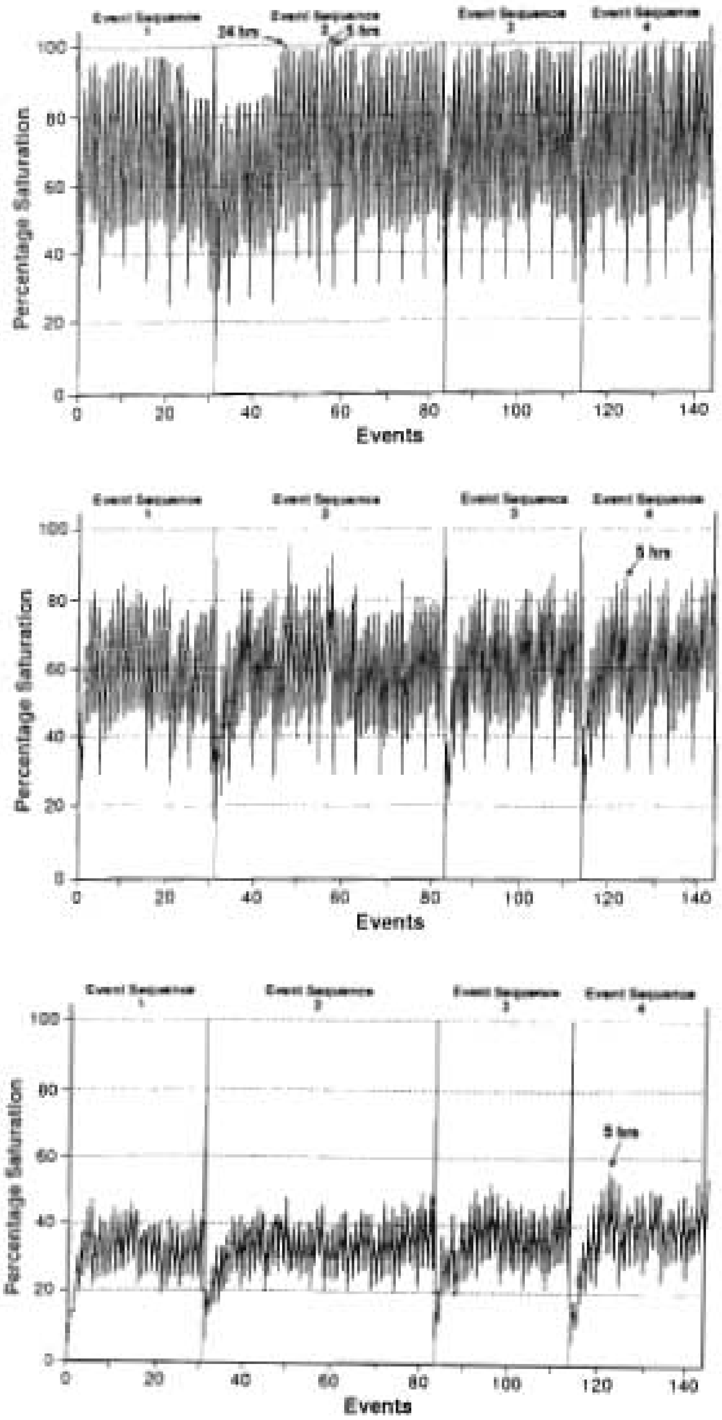
No 'control' experiment was run for three main reasons. First, to leave a sample of each rock type in water for the duration of the experiment would have been of dubious significance, with any changes reflecting, primarily, chemical weathering effects of long-term immersion, a factor not being considered in this experiment. Second, and perhaps more importantly, leaving a sample in water would have replicated only one of the sample wetting procedures; there would have been no relationship to the sprayed sample. Third, the effects of long-term saturation had already been considered by Felix (1983). His findings were mainly that it enabled the separation between '..water-sensitive facies, generally prone to rapid weathering, from partially, to non-sensitive ones' (Felix, 1983, p. 310). What had not hitherto been done was to monitor the changes to rock moisture content and from that to deduce the changes to rock properties and hence the weathering effects. Thus, in effect, the experiments undertaken here go someway towards answering the questions raised by Nepper-Christensen (1965) and Felix (1983) regarding the effects of, and changes to, the pore size distribution and their effects, direct and indirect, upon mechanical weathering.

In the presentation of the results (Figure 1) the terms 'events', 'event sequence' and 'percentage saturation' are used. 'Event' simply refers to daily measurements, in other words the daily weighing, wetting and weighing cycles, i.e. each wet–dry event. The event sequence is the series of events between times when the samples were dried and saturated again in order to monitor changes in water-holding capacity and to enable the mass loss to be determined. In determining saturation, the sample was again placed in water and allowed to stand until a stable mass was obtained; the time taken to achieve this was noted as it was found to take longer than the one or two days needed at the start of the experiment. Percentage saturation is the amount of water that was held compared to how much the sample could hold, presented as a percentage. Data were corrected throughout as a function of mass loss (i.e. if mass were lost then the original, experiment start values would no longer be valid) and as the water-holding capacity of the rock changed through time. The percentage saturation values are thus meaningful as they were corrected for changes in sample properties and mass during the experiment rather than the more normal procedure of obtaining only start and end values, which does not allow for this on-going correction to be undertaken.

Details of mass loss

Details of the mass lost for the two sandstones and the dolerite after 140 cycles are shown in Table I. The fact that the covered and half-covered samples with the exception of the Elliot sample, responded in such a similar manner may reflect the availability of water rather than the degree of cover. In other words, although only half submerged, the rock was able, via capillary forces, to continue to take up water and thus responded in a similar manner to the totally covered sample. Even in the case of the Elliot sample, if the large (20·5 g)


Event Sequence 1
Event Sequence 2
Event Sequence 3
Event Sequence 4
24 hrs
5 hrs
Percentage Saturation
Events


Event Sequence 1
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24 hrs
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Percentage Saturation
Events
5 hrs
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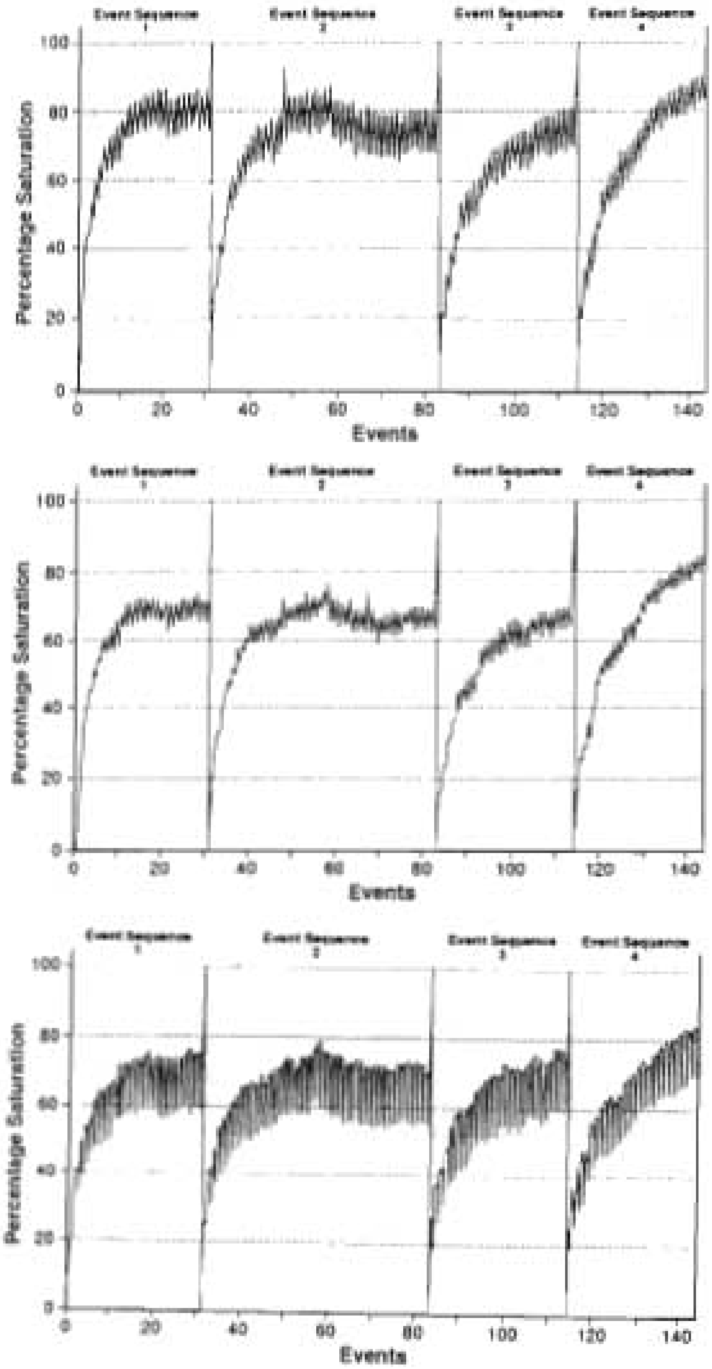

Figure 1 Graphs to show percentage saturation of the three rock types (Clarens sandstone, Elliot sandstone and dolerite) for the three methods of moisture application (covered, half-covered and sprayed): (1a) covered Clarens sst; (1b) half-covered Clarens sst; (1c) sprayed Clarens sst; (2a) covered Elliot sst; (2b) half-covered Elliot sst; (2c) sprayed Elliot sst; (3a) covered dolerite; (3b) half-covered dolerite; (3c) sprayed dolerite

Table I. Percentage of mass lost after 140 cycles for the three rock types subject to the three modes of moisture application

	Covered	Half-covered	Sprayed
Clarens SST	0·26	0·26	0·11
Elliot SST	4·7*	0·47	0·14
Dolerite	0·47	0·45	0·32

* Pieces > 0·5 g lost from the covered Elliot sst = 0·7 g after event 14, 20·5 g after event 24 and 7·5 g after event 44

piece weathered free is ignored, it would appear that the mass loss for both the totally covered and half-covered samples is about the same. However, both the covered and half-covered samples show a much greater mass loss than those samples that were only sprayed with water. Surprisingly, it was the dolerite from the South Shetlands that showed the greatest percentage mass loss for the sprayed samples as, subjectively, it was thought the sandstones might be more prone to breakdown. Furthermore, all the samples (both sandstones and dolerite) lost small 'splinters' of rock from the covered and half-covered samples but no obvious 'splinters' occurred from the sprayed samples, which only experienced granular disintegration.

The 20·5 g piece of rock that weathered free from one of the covered Elliot sandstone samples was the largest piece produced during these experiments. However, the Elliot sandstone consistently lost larger pieces than the other lithologies with, for the sample identified here, a piece of 0·7 g lost after event 14, the 20·5 g piece after event 24 and a piece of 7·5 g after event 44 (Table I). Thus, it may well be that the Elliot sandstone is particularly prone to this form of weathering. It is also worth noting that the pieces were lost from the covered Elliot sandstone sample near the start of the experiment and so if percentage of weight of sample remaining compared to original weight is plotted against the number of cycles on the x-axis, this produces the reversed 'S'-shaped graph discussed by Yatsu (1988, p. 115), said to be indicative of a slow rate of breakdown.

Although it is generally perceived that mainly shales and tuffs, or rocks with a high clay content, experience the expansion–contraction of wetting and drying, several studies have shown that a whole range of other rocks can also be affected (e.g. see Nishioka and Harada (1958) for details regarding sandstones, schists, limestones, marbles, cherts, granites, basalts, talc, obsidian and a number of other rock types, and Nepper-Christensen (1965) for basalt and flint). In fact, Nepper-Christensen (1965, p. 554) states 'Many, if not all, rocks are sensitive to variations in the relative humidity of the surrounding atmosphere since they shrink and swell as they give off or absorb moisture', and adds that clay minerals are but one factor influencing the process, '...mineral composition and pore structure are probably important factors' (p. 555). This conclusion is reiterated by Felix (1983, 1984, 1987) based on studies of sandstone, '...the swelling is, therefore, strongly influenced by the pore size distribution...water sensitivity appears...as a complex combination of mineralogical (clay minerals) and structural factors, yet to be defined' (Felix, 1983, p. 310). Felix also noted (p. 309) that several studies shown '... the strong effect of water sorption on the expansive properties of rocks not containing clay minerals'. Thus, contrary to many general perceptions, wetting and drying can have an expansion–contraction effect upon a whole range of rocks including those that have no clay minerals. The actual cracking of the rock is thought to be most active during submergence rather than during desiccation (Yatsu, 1988), the cause being related to the swelling or expansion strains that are developed in the rock when it takes up moisture (Hames *et al*, 1987; Pissart and Lautridou, 1984). This, then, would clearly help explain why the submerged and half-covered samples showed significantly greater degrees of breakdown than did the sprayed samples. What is not entirely clear is why, with the exception of the Elliot sandstone, the fully submerged samples did not exhibit a greater mass loss than the half-covered samples.

It is the nature of the 'materials contained in rocks, tensile strength and the structure of the rocks, and the pore size distribution that are all important' in wetting and drying (Yatsu, 1988, p. 115). Whatever the cause of the difference in effects between the Elliot and Clarens sandstones, and between the sandstones and the dolerite, there are clear implications for laboratory simulations. By covering, or even half covering, rock samples with water and then subjecting them to freeze–thaw cycles during which water loss and replacement occurs within the experiment, then a component of weathering due to wetting and drying affects the final result. The role of wetting and drying has two elements. First, there are the direct effects in terms of mass loss due to the wetting and drying process itself, and in some rock types this could be significant. Second, it operates synergistically by generating large cracks (Yatsu, 1988) that are then available to be exploited by the frost processes. Without the generation of these cracks the efficiency of freeze–thaw would be reduced. In other words, the wetting and drying effects abet the freeze–thaw mechanism.

With respect to salt weathering, the same principles apply. However, the rate of breakdown due to wetting and drying with a saline solution can be slower than with fresh water for a number of reasons (Yatsu, 1988): there is less water penetration due to the large surface tension and higher viscosity of salt water; desiccation is slowed as a result of surface coating by salts and of the attractive forces between clay grains being intensified thereby strengthening the rock. Nevertheless, there is still a wetting and drying effect to be taken into account. The same is the case for studies of thermal stress and fatigue such as those of Griggs (1936), Blackwelder (1933) and Birot (1960), all of whom concluded that the addition (by immersion) of water to the rock samples greatly accentuated weathering rates. As more recent studies (e.g. Yong and Wang, 1980) have shown, microcracking of granite can occur with temperatures of 72°C or greater, so the combination of this with the effects of wetting and drying (caused by immersion in water followed by heating) could be expected to exacerbate the rate of breakdown.

Effects upon the internal structure of the rock

The second topic, namely changes to the internal structure of the rock as expressed by changes in rock properties, complements the above discussion, particularly with respect to the effects of wetting and drying abetting other processes. Here, it is the change in the internal structure of the rock, as indicated by changes in the water-holding property of the rock (i.e. the percentage saturation), that is considered. Figure 1 indicates that for the dolerite there is an increase in percentage saturation with time for all three sample procedures but with the overall percentage saturation being: covered > half – covered > sprayed. There must be an increase in the size of pores/microfissures and/or an increase in their numbers within the rock to explain this increased percentage saturation. Five other interesting elements can be seen from the figures: (1) the diurnal fluctuations on the rising (wetting) limbs; (2) the drop in percentage saturation later in the event sequences; (3) the unusual 'low' level in the third event sequence; (4) the sprayed samples show a greater diurnal variation in percentage saturation; (5) an increase in overall saturation.

In all three sample procedures it is very clear that, for the dolerite, after each drying event the rising limb of the graph becomes less steep and, at the same time, shows the effects of diurnal changes earlier and with greater frequency. For reasons not yet understood, there was a continuous gain of moisture, with no diurnal loss, on the rising limb of event sequence 1 (i.e. cycles 1–32 through to saturation and total drying). In event sequence 2, the rising limb again indicated a continual mass gain but the effects of diurnal variations began much earlier, whilst in event sequences 3 and 4 the diurnal responses became apparent earlier and occurred with ever-increasing frequency. Quite why these changes should occur is unclear, but the fact that the rising limbs are seen to become more gentle suggests that it is taking longer for the dolerite to absorb water. This may partly reflect the effects of the increasing diurnal variability (i.e. by events 3 and 4 there is a greater diurnal change than was found in event 1). However, the dolerite, which took only two days to fully (i.e. 100 per cent) saturate at the experiment start, took five days to fully saturate after event 4; thus other changes must also have been occurring. One explanation is that the pores near the rock surface are getting bigger and that concurrently there is an increase in the micropores and fissures inside the rock. Thus the gain and loss of water from the larger, surficial pores is the cause of the diurnal variability seen on the rising limbs, but the increase of micropores means it takes longer for the whole rock to saturate. Indeed, data regarding porosity and microporosity, as determined by the methodology of Cooke (1979), do show marked change for the dolerite, as has been argued above (Table II). However, no detail regarding the actual size of the 'micropores' is available and so, until these experiments can be repeated with detailed porosimetry at selected intervals during the experimental sequence, the suggestions must be given in generalized terms. Interestingly, the

Table II. Changes in porosity and microporosity for the dolerite after the three different modes of moisture application, compared with average initial values

	Porosity	Microporosity
Initial values	2·86	27·31
After spraying	3·64	85·38
After half immersion	5·0	95·29
After full immersion	4·98	96·59

sandstones did not show these changes and the time for total saturation (24 h) at experiment start was the same after event sequence 4.

It can be seen from the graphs (Figure 1) of both sandstones and the dolerite that the rising limbs reach a 'plateau'. However, in event sequences 1 and in particular, 2 there can be seen to be a *decrease* after the plateau high is reached. Again, quite why this should occur is not certain but it may reflect a change in porosity/microfractures such that it is easier for more water to be lost (i.e. the voids have been enlarged). If this is the case then it might help explain why there is a general rise in percentage saturation from event sequence 1 through to event sequence 4, e.g. for the dolerite from *c.* 86 per cent saturation in sequence 1 to *c.* 93 per cent saturation in sequence 4 for the fully covered sample; from *c.* 73 per cent in sequence 1 to 84 per cent in sequence 4 for the half-covered samples; from $\leq$ *c.* 77 per cent in sequence 1 to *c.* 85 per cent in sequence 4 for the sprayed sample. Thus, there does seem to be an increase in water-holding capacity, as shown by the general increase in percentage saturation, and this must reflect an increase in porosity/microfissure size and/or numbers to accommodate this. The same general changes are evident in the two sandstones but the change is most evident in the sprayed samples.

Event sequences 1 to 4 would show a consistent linear increase in percentage saturation, as discussed above, were it not for event sequence 3. Here, as the graphs clearly show, there was a *decrease* in percentage saturation for all the sample procedures. Again, the cause is not clear and the only common factor is that event sequence 2 was particularly long (49 rather than 32 events). However, the longer event sequence exhibited a post-plateau decrease in percentage saturation (as noted above) ending, for the dolerite, at *c.* 80 per cent saturated, which is the plateau attained by event sequence 3; post-event 2 highs were greater in the two sandstones although they did not achieve the levels of event 2. This, then, begs the question as to why event sequence 4 should show such a marked increase in percentage saturation? Clearly the effects of wetting and drying are anything but simple!

Finally, the data for the dolerite show not only the degree of saturation which the samples could attain under the experimental conditions in use (which may have parallels with some field situations), but also the diurnal variability as a function of the daily wetting and drying events. Even here some unexpected results were obtained. Not surprisingly, perhaps, the daily range (i.e. the difference between the 'wet' and 'dry' levels) for the covered sample was greater than that for the half-covered sample. However, the daily range for the sprayed sample was almost twice as great as that of the water-covered sample, with a daily range of nearly 20 per cent. Why such a sharp contrast? One possibility, which explains both the high daily range and the generally lower overall degree of saturation of the sprayed sample, is that, owing to limited water being available and the short period of application, it was not possible to fill the smaller pores and/or fissures. Thus, on the sprayed sample only the large voids took in water and these would more readily and rapidly lose it during the drying period. In the more saturated samples, once the smaller pores were filled (as shown by the gradual overall increase in percentage saturation) they would not lose moisture so rapidly. Also, the overall degree of saturation was higher in the samples located in trays of water and, recognizing that the water must have penetrated substantially to produce a > 80 per cent saturated sample, it was less easy for water deep inside the rock to migrate to the margins within the time frame of the drying periods used here. Daily differences between the wet and dry states were much greater in the sandstones compared to the dolerite, as might be expected owing to their greater porosity.

The influence of variation in experimental procedure

With respect to experimental procedure, it is found that this can have a marked effect. For all samples, th relative degree of saturation was: covered > half – covered > sprayed. However, the differences between th applications were very small for the dolerite (as discussed above, Figure 1), greater for the Clarens sandston and greatest for the Elliot sandstone, where the sprayed level of saturation was approximately 50 per cent o that for the sample fully immersed in water. Further, the Elliot and the Clarence sandstones both attaine similar levels of saturation by spraying, but the Elliot was marginally higher than the Clarence in the half covered procedure (80 per cent vs. 60 per cent) and also in the fully covered sample (> 90 per cent vs. 80 pe cent). This, then, implies that for the Elliot sandstone, an almost 100 per cent saturated condition is attaine with only 3 h submersion in water. Thus, the properties of the individual rocks play a crucial role in deter mining the effects of the various means of applying moisture. This implies that rocks should be tested wit respect to this as an integral part of any test sequence for frost, salt or other form of water-based weathering

The duration of wetting and/or drying, not surprisingly, was shown to exert an influence on the degree o saturation. However, as obvious as this may be, it is a factor *not* quantified in frost of salt experiments. Fo example, the Elliot sandstone, if given water for 24 h shows little or no difference from the level of saturatio attained after only 3 or 5 h. For the Clarens sandstone, a 5 h wetting period produces a level of saturatio about 10 per cent lower than does a 24 h. wetting; for the dolerite there is an 11 per cent difference (8 per cent vs. 97 per cent). Thus, even a relatively short (i.e. 3 or 5 h) period of wetting will attain high level of saturation and this, in turn, may have important ramifications for freeze–thaw experiments. This is par ticularly so as saturation levels of 80 per cent or greater imply that, as a result of the moisture gradient withi the rock, a substantial portion of the outer part of the rock must be at, or very close to, 100 per cent satura tion; such conditions are required by many of the hypotheses and models of frost action.

Significance of the findings with respect to weathering

Although the implications of some of the findings reported above are uncertain, it is clear that wetting an drying of rock samples during laboratory experiments can have an influence on the nature, degree and rate o breakdown. That there is debris loss as a result of *just* wetting and drying implies that this effect must influ ence the results of freeze–thaw or salt weathering experiments. Furthermore, Letavernier (1984) and Leta vernier and Ozouf (1987) have suggested that any index of frost susceptibility should not be based upon th amount of the original rock sample left after testing but rather on a 'coefficient of comminution' (coefficien d'amenuisement) based on the particle size distribution of the resulting fragments between 0·5 and 2·5 cm This is certainly important for it recognizes the fact that particles detached from the original block are them selves subject to breakdown. However, from this present experiment it would seem that the manner of moist ure application helps determine the character of the resulting debris produced by wetting and drying (covered and half-covered samples give 'splinters', sprayed samples give granules) and this could have an influence on the particle size distribution. Furthermore, the debris would themselves experience satura tion–drying cycles that would result in breakdown additional to any frost effects and so influence the result ing granulometric curve.

Thus it would seem that in addition to the laboratory experiment of frost or salt weathering there should be concurrent assessment of the weathering due to wetting and drying effects of moisture application. The 'norm' in most laboratory experiments of frost or salt weathering has been to use samples in distilled water as a control (and even this has varied greatly from experiment to experiment (McGreevy and Whalley, 1985)) However, this approach has only been used to see what degree of weathering (i.e. mass loss) would have taken place as a result of the moisture conditions in the absence of the salt or frost effects. What these con trols have not done (in the instance where they have been used) is to tell the effects, in terms of changes to pore size distribution and water-holding capacity that, as a result of the wetting and drying, have had an effect upon the frost or salt weathering process; rather, they have simply documented the amount of weath ering (as identified by mass loss) that took place. The data presented here show that the wetting and drying alone can influence pore and moisture conditions as well as effecting weathering in its own right. Thus, the results of such a concurrent experiment would not only indicate the amount and character of debris that could be produced but would also provide information of changes in rock properties that might enhance or inhibit the effects of frost or salt weathering. With so little known regarding the role and influence of wetting and drying, combined with the variability, both within and between lithologies, of rock properties, there is a need for more information from laboratory experiments.

With respect to the field, the information presented here indicates that greater cognizance should be accorded to the role of wetting and drying as both a weathering mechanism in its own right and as one that acts synergistically with other processes. Apart from obvious locations such as tidal areas and the margins of lakes or rivers, where wetting and drying can occur frequently, there are situations, for example in association with snowbanks, where this process may be common. Bedrock and debris near ablating snowbanks may be subject to frequent wetting and drying cycles as water is produced during warm, dry periods and then evaporates during the cooler, frequently windy, periods. Such a concept could help explain the breakdown of material in association with a snowpatch (i.e. in nivation or cryoplanation).

In other areas, as has been argued for the South Shetland Islands by Hall (1993), rain can occur frequently during the summer. This results in wetting and drying of bedrock taking place during a time when other mechanical processes are not active. As a consequence, rock breakdown continues during this summer period and the rock properties are altered such that winter processes (e.g. freeze–thaw) are more effective. As a result of what can, in some locations, be large numbers of wetting and drying events (Hall 1993), this form of weathering could be more effective than the more usually cited freeze–thaw.

CONCLUSIONS

It would seem that the process of wetting and drying is even more complicated than was hitherto thought. Whilst a number of the findings reported here are unclear and the explanations of others speculative, it is certain that wetting and drying operates as a weathering mechanism in its own right and that the nature of the weathered material is influenced by the manner of wetting. Furthermore, the process of wetting and drying has an effect upon the internal characteristics of the rock and this, in turn, can influence the nature and degree of other weathering processes operating synergistically with wetting and drying. The results presented here indicate three important future avenues of research. First, investigation of wetting and drying as a process in its own right and the need to understand exactly how it operates. Second, field data collection of the frequency and degree of saturation of rocks in different environments. Third, and perhaps of major importance, the testing of rocks for wetting and drying effects within the context of freeze–thaw, salt and water-based chemical weathering experiments to ascertain the role that this process plays.

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Antarctic rock weathering simulations: simulator design, application and use

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Abstract: The design of a computer-controlled climatic simulation cabinet used for mechanical weathering studies on Antarctic rocks is described. It is argued that if the results of simulations are to be applicable to field situations they should be firmly based on field environmental conditions. Some weathering results from a laboratory simulation based upon microclimatic data collected from the maritime Antarctic are presented and it is shown that they could not have been obtained from field measurements alone. Further simulation studies are required for the Antarctic and it is argued that the nature of Antarctic research is such that it is particularly conducive to this type of approach.

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Key words: rock weathering, simulator application, simulator design, weathering simulation.

Introduction

Although researchers in the Antarctic have recognized the role of weathering in landform formation, they have not undertaken simulations as an aid to determining either process operations or rates. Despite their apparent simplicity, weathering simulations are fraught with difficulties and practical problems. As Thorn (1988, p. 13) points out '... laboratory experiments are ultimately only worthwhile if it can be shown that they mimic natural conditions successfully' and that 'the general validity of laboratory experiments is open to question because of the scale of these experiments'. Thus, although the usage in freeze–thaw simulations of the idealized broad climatic themes of 'Icelandic' and 'Siberian' cycles (e.g. Tricart 1956) facilitates comparison of results, they may be of little significance if these cycles do not reflect the field situation. As a consequence it may be misleading to attempt to apply laboratory-derived results back to the field (McGreevy & Whalley 1982). Ultimately, it is the deficiency of field data on interstitial rock moisture content, and its chemistry, together with knowledge of rock temperatures, that precludes the running of satisfactory simulations of temperature-controlled mechanical weathering processes (McGreevy & Smith 1982, Kerr *et al.* 1984, Thorn 1988).

In an attempt to overcome many such problems in the maritime Antarctic, a combined field and laboratory investigation of mechanical weathering processes was initiated (Hall 1986*a*) as a part of the 'Fellfield Ecology Research Programme' undertaken by the British Antarctic Survey (BAS) on Signy Island, South Orkney Islands (60°43'S, 45°38'W). To ensure that the simulations would accurately replicate the field conditions, information was obtained on the geology (Storey & Meneilly 1985), the field microclimate of individual rocks (Walton 1977, 1982, unpublished) rock moisture content and the rock properties affecting thi (Hall 1986*a*), interstitial rock water chemistry (Hall *et al* 1986), and rock strength (Hall 1986*b*, 1987). In addition continuing field experiments, including the daily monitorin of changes in mass of a rock tablet as an indication o moisture content changes and weathering rates, provide base-line information for the continued refinement of th simulations (Hall 1988*a*). In the laboratory, the utilizatio of relatively large rock samples (up to 25 kg) partly reduce problems arising from the use of only hand-size specimen (Thorn 1988). The monitoring of internal rock temperature overcomes the problems identified by Jerwood *et al.* (1987 of how well the external temperature cycles are reflected i actual rock temperatures.

The simulator

Laboratory simulations have used a great variety of equipment, some very sophisticated and others based on domesti appliances. Often only limited information has been give about the equipment used (Table I); yet the equipment itsel may well be a major determinant of the type of simulatio possible (Jerwood *et al.* 1987). In addition, the lack o equipment specifications makes it difficult for others t build simulators which have some degree of compariso with those already in use. The present study used a purpose built simulation cabinet.

The basic design specifications required that:

a. it could work in the temperature range +70°C t –50°C,
b. it would allow sequences of temperature cycles of a wide range of duration, magnitude, frequency an rate of change,

Table I. Simulator specifications.

	Wiman 1963	Pous 1970	Brockie 1972	Steijn 1979	Fahey 1983	Lautridou & Ozouf 1982	Swantesson 1985	Pécsi 1987	Jerwood *et al.* 1987	Matsuoka 1988	Hall 1986*c*
Adapted commercial equipment		?	✓		✓		?	?		?	
Purpose built	(✓)	?		✓*1		✓	?	?	✓*1	?	✓
Automatic	✓	?		(✓)	?	✓	?	?	✓	?	✓
Manual		?	✓		?		?	?		?	
Rate of change of temperature control	?	(✓?)	✓	?	?	(✓)	?	?	✓	(✓)	✓
Cabinet temperature sensor	(✓)	✓	✓	?	?	✓	?	(✓)	(✓)	✓	✓
Sample temperature sensors	✓	(✓)	✓	?	?	✓	?	?	✓	✓	✓
Automatic sensor logging	✓	(✓)	✓	?	?	✓	?	?	✓	✓	✓
Manual sensor reading				?	?		?	?	✓		
Precision of equipment	?	±1.0°C	±0.75°C	?	?	?	?	?	±0.5°C	?	±0.25°C
Humidity sensor	✓	—	—	?	(✓)	?	(✓?)	?	?	?	✓
Humidity control	(No)	—	—	?	No	?	No	?	?	?	(✓)
Moving air conditions	(No)	—	—	?	✓	?	?	?	?	?	✓
Infra-red lamp facility	—	—	—	?	(✓)	?	?	?	?	?	✓
Ultrasonic testing	—	—	—	?	—	✓	?	?	?	(✓)	✓
Other testing	—	—	—	?	—	?	?	?	?	✓	✓

*1 the name of the equipment is specified; manufacturers could be consulted for details
(✓) by implication or similar specifications
? does not imply that facility does not exist but that no information is available

c. data output would be displayed on both VDU and printer and stored on disk
d. data could be read at varying pre-specified intervals (range 5 s to 99 min) throughout the cycle sequences,
e. facilities would be available for a range of sensor types,
f. a feedback system continuously checked the programmed sequence and instituted corrective action where necessary,
g. programming (in Basic) was very simple,
h. the machine needed minimal attention,
i. mechanical and electronic equipment protection be built-in,
j. data output contained experiment time, programmed status and actual status together with other sensor data outputs.

Based upon the above broad specifications, a cabinet has been built (Fig. 1) utilizing a Commodore 64 PC which, via additional hardware, acts as the central control unit (Fig. 2). The PC also provides the facility for the writing, storing and input of the main control programmes. The programmed temperature is updated every 10 s, compared with the actual cabinet temperature and then a power controller adjusts the heating or cooling.

Output from the six thin film platinum resistance temperature sensors are sequentially selected and converted into digital values which are stored in the PC. Information from the humidity sensor is converted to a linear analogue output and switched into the analogue to digital converter at the appropriate time. A humidity controller is also available but is not at present in use. Output from the 'Pundit' ultrasonic test equipment is converted into a linear pulse width voltage signal and switched into the analogue to digital converter. Signals from the 'Opticat' fibre optic crack detection system are routed back into the PC input port and checked every 10 s for any change. A variable infra-red lamp controller to supply surface heating can be activated via software control. If an 'out of range' condition should occur for any reason for a period greater than 3 min then the whole system is automatically shut down.

The cabinet itself is constructed with an outer body of 'Zintex' and an inner shell of Type 316 stainless steel (with argon welded seams to make it watertight), with 100 mm of expanded polystyrene foam between the two as an insulator. Refrigeration is provided by a 1-HP compressor utilizing 1 kg of Freon 502 as refrigerant. The liquid 502 is divided into two streams (Fig. 3), one passing through an expansion valve into 6 m of 10-mm diameter copper tube, supported in wooden slats, that rests upon the base of the box and provides the ground cooling. This is supplemented by the second stream that gives, aided by a fan, air cooling. Defrosting of the air chiller is achieved by hot gas injection and this process is initiated by the PC every 6 h for a period of 7 min. Heating is provided by a commercially available heating element whose power output is regulated by the PC and which can be balanced against the cooling system.

In this way all of the original design requirements have been achieved and an example of a programme sequence, together with a typical print-out, are shown in Fig. 4. The incorporation of ultrasonic equipment facilitated continuous,

Fig. 1. View of simulation cabinet identifying the main components.

non-destructive monitoring of changes in certain rock properties (e.g. elasticity, cracks, moisture status) during the simulation (Hall 1988*b*). The fibre optic crack detection system not only showed when failure occurred but also, by means of a laser beam passed down the optic fibre, where and in what multiples cracking took place (Hall 1988*c*). Overall, the cabinet overcomes a number of the shortcomings of many previous units. Not only can any number of highly complex cycles be replicated (driven if required by data directly logged in the field) and data collected during the run at suitable acquisition times, but the read-out indicates how well the machine replicates the programmed cycles. In addition, the ability to heat samples by means of infra-red lamps overcomes problems associated with thermal fatigue simulations (Cooke 1979) and proves especially valuable for the investigation of solar heating effects on rocks at low temperatures. By running the simulations with large samples at field moisture and solute status, and from data logged in the field, it is possible to offer a close approximation to a natural environment

Results and discussion

Antarctic weathering studies have largely concentrated on the cold and very dry ice-free valleys and mountains of the continent (Campbell & Claridge 1987). As a counterpoint to this, the present studies aimed at investigating weathering processes in the warmer, wetter and biologically active maritime Antarctic. Process investigations under controlled conditions are an essential component of this, complementary to the field experiments. Indeed, simulation studies can produce data which would be impossible to obtain in the field.

Initial simulations were directed towards the influence o schistosity upon freezing, the effects of different moistur and solute contents, the interaction of salt weathering an wetting and drying with freeze–thaw, the influence of omni and unidirectional freezing plane penetration, and the natur and rate of production of weathered material. Data were als obtained on the rate of change of temperature inside the roc during different warming and cooling conditions, the temperature at which the ice into water phase change took place and the occurrence of acoustic emissions during rock freezing. Although some of the results of these simulations hav already been reported in detail elsewhere (Hall 1986*c* 1988*b*, 1988*c*) they are summarized here to illustrate th value of process studies of this type in investigations of pola and alpine weathering.

Internal rock temperature data show that not only is th rate of temperature change largely controlled by the ampli tude of the freezing event but that in schistose rocks it can b up to five times faster when schistosity parallels the freezin front than when it is normal to it. It would thus appear tha the attitude of schistose rocks and the amplitude of th freezing event are both important and interrelated factors This in turn leads to the distinction in weathering rates foun for unidirectionally (open system) and omnidirectionall (closed system) frozen rocks. The latter, which also ofte attain a high moisture content (Hall 1986*a*), may experienc different responses to freezing (e.g. undergo hydrofracture than in the open system freezing, and thus simulation utilizing omnidirectionally frozen material should not us the results as an analogue for the cliff-type open systems Weathering rates for unidirectionally frozen rocks wer found to be as much as 5800% slower than for th

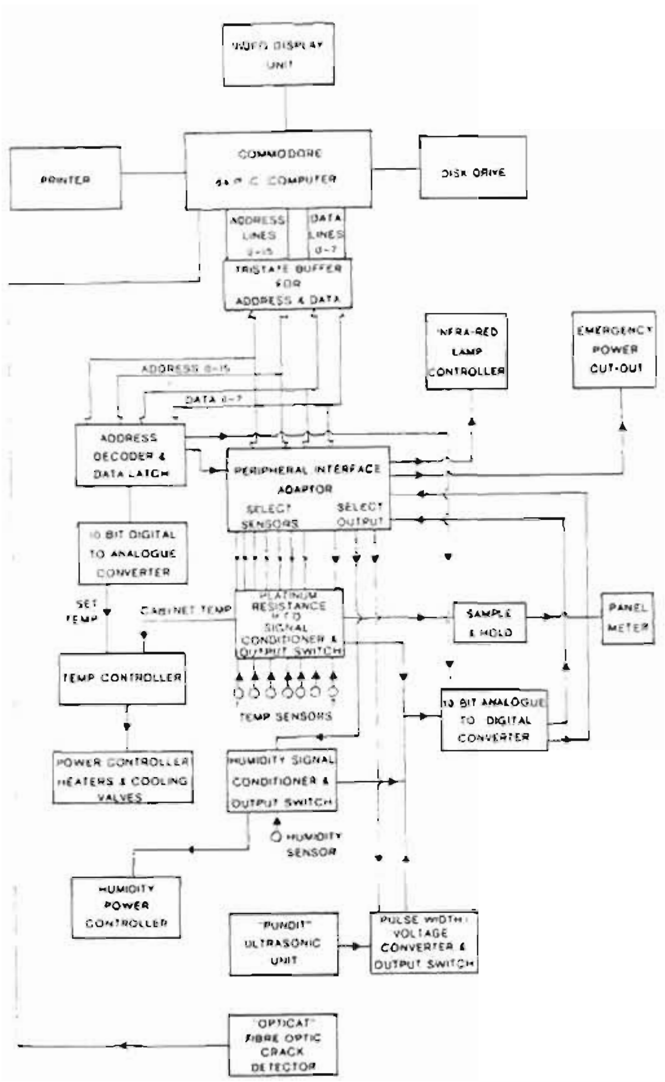


Fig. 2. Circuit diagram of simulation cabinet control system.

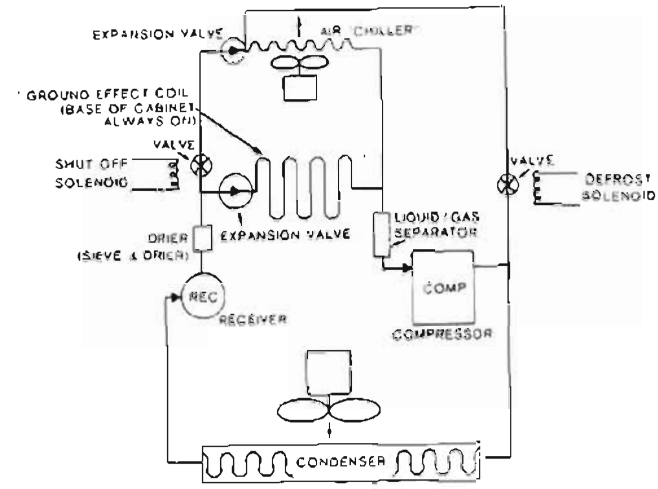
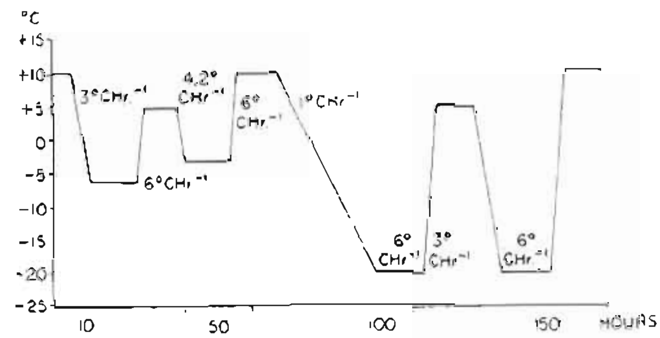


Fig. 3. Refrigeration system of simulation cabinet.



Time	Set	Cab	S.1	S.2	S.3	S.4	S.5	S.6	Hum.	U/Sc	Row
	°C	°C	°C	°C	°C	°C	°C	°C	%	µs	
Col. 1	2	3	4	5	6	7	8	9	10	11	
02:36.00	-6.0	-5.6	-0.7	+0.0	-0.7	-0.3	-4.5	-4.6	36	23.7	240

Where Col. 1 = actual time, Set = programme temperature (°C), Cab = actual cabinet temperature (°C), S1–S6 = temperature sensors 1–6 (°C), Hum = cabinet humidity (%), U/Sc = ultrasonic velocity (µs), Row = data row number

Fig. 4. A typical programme sequence together with a line of output data.

omnidirectionally frozen, wetted rock subjected to the same temperature fluctuations.

The actual nature of the water to ice phase change that takes place inside the rock was found to vary in response to particular combinations of rock moisture content, solute concentration, freeze amplitude and rate of change of temperature. Three types of phase change were monitored:

a. a rapid, large-scale (≥ 80% of the moisture that would freeze) transformation,
b. a slow, progressive freeze, and
c. a relatively rapid, progressive freeze (intermediate between a and b).

The first type was encountered during relatively slow rates of fall of temperature, the second during higher rates of change, and the third when high saline rock moisture conditions existed irrespective of the nature of the thermal change. It was also found that, for the smaller amplitude freezes (i.e. down to *c.* -6°C), it was necessary for the maximum subzero condition to be maintained for at least 10 h before a phase change took place in the rock. This data is not only valuable with respect to interpretation and application of field microclimatological information but it also has application for theoretical models of weathering (Hallet 1983) in which certain responses are hypothesized to result from specific climatological conditions. Salt content, and thus the potential for combined freeze–thaw and salt weathering (Williams & Robinson 1981, McGreevy 1982), was found to influence the rate of rock breakdown as well as the freezing point depression and the temperature of the ice to water phase change (–0.7°C to –1.9°C).

Within the present technological and logistical constraints of Antarctic research, little, if any, of the laboratory findings briefly cited above could have been obtained by means of field investigation. However, by using simulations firmly based on field conditions, it is possible, particularly with field experiments to act as controls, to apply many of the laboratory results back to the field.

Future use

Why should simulations be so valuable to Antarctic studies? Firstly, the slowness with which many weathering processes operate due to the paucity of moisture and the long periods of subzero temperatures makes some field studies impractical. The time-compression ability of the simulation thus takes on a special significance. Secondly, simulation also permits investigation of the potential geomorphic effects of particularly severe (and thus potentially hazardous) climatic conditions. Thirdly, by means of simulations it is possible to investigate the detailed relationships between the various weathering processes. A particularly good example is the role of thermal fatigue initiated by solar heating of rock surfaces during times of severely subzero air temperatures, and the associated thermal strains introduced when that heat source is instantaneously removed by shadow effects. Finally, logistical constraints often preclude frequent visits to particular field sites; simulations backed up by continuous unattended field experimentation can provide a strong framework for weathering studies in remote regions.

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Chapter 4

Processes

"Sir, I have found you an argument. I am not obliged to find you an understanding"
Samuel Johnson

As a result of the field data and the laboratory simulations it was possible to undertake some evaluation of the processes themselves. Much of the information pertaining to findings regarding processes is also found in both Chapters 2 and 3 (e.g. Hall, K. 1988b. discussed in Chapter 3) and thus here only material that deals directly with 'process' is presented. Clearly no definitive answers are provided; rather the material offers some additional insights into how the individual processes, or at least some component of the process, may work.

There have been many studies regarding processes, both theoretical and practical (e.g. Grawe, 1936; Aguirre-Puente, 1978; Nye and Davidson, 1982; Hallet, 1983; Fahey, 1983; Davidson and Nye, 1985). The applicability of the findings are frequently in question as the conditions used may be relatable to few, if any, situations on our planet. Rather, in the sense of the discussions introduced in Chapters 2 and 3, the situation is now one of testing and verifying the existent hypotheses based upon the recorded field data. As the latter are still limited, so is our ability to test theory. Equally, as new field data are made available, so new theory can be developed and tested.

Papers that are presented within this section, in order of discussion, are the following:

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The material presented in this Chapter begins with an 'Introduction' to a special issue of *Permafrost and Periglacial Processes* dealing with cryogenic weathering. The special issue was the outcome of a meeting held in France of which this author was one of two organizers. Detailing the papers in that special issue, the Introduction presents some background to the relationship between process and field data. The next four papers cited deal with the questions surrounding freeze-thaw weathering and its use as a basis for the origin or development of many cold region landforms. These are followed by a paper (Hall, 1992c) that also discusses these same problems but puts the argument within the context of debate regarding Quaternary and present day processes in southern Africa. Apart from the inadequacy of data upon which to make determinations regarding the the occurrence or not of freeze-thaw, the argument is made that there must be a process synergy. The very factors (e.g. thermal changes) that drive freeze-thaw must also drive other processes (e.g. thermal stress fatigue). Through time (diurnal, seasonal or longer time scales) processes will change; in fact, process combinations or process domination can be changed by the landform itself as it develops. Thus, it is most unlikely that any one process alone occurs and is responsible for landform development. With that premiss, and considering the range of climatic conditions involved through the diversity of cold environments, it is naive to state that freeze-thaw 'predominates' within cold environments (which has been a recurring theme in the literature as clearly shown in Chapter 1).

The next four papers discuss weathering by wetting and drying and thermal stress fatigue in terms of process, process application and its historical perspective. Again, rather than actually answering questions related to process, these papers attempt to define the questions we should be considering. These papers outline how the processes of wetting and drying and/or thermal stress fatigue are very probable weathering processes in cold regions. Indeed, in cold, arid environments a good argument can be made that, rather than freeze-thaw, both thermal stress fatigue and thermal shock are likely significant processes. Equally, in the more maritime cold environments, a strong case can be made for the action of wetting and drying - a process that *must* occur in parallel with freeze-thaw weathering. Salt weathering is not discussed here, although clearly a significant process in cold regions; information on this can be found in papers within the preceding chapter. The next paper (Hall, K. and Otte, 1990) deals with a study of weathering by algae on a nunatak in Alaska. Biological weathering, particularly in cold environments, suffers from a paucity of studies and yet, as shown here, it can be a major factor. Much more needs to be done to investigate the role of micro-organisms in the weathering of rock in cold regions, although the biological studies of Nienow (1987), Friedmann and Weed (1987), and Friedmann *et al.*, (1987) clearly show the weathering potential.

The next two papers in this section report what could fairly be described as an entirely new weathering mechanism based on observations in the McMurdo dry valleys of Antarctica. Material blown by wind is well documented and studied with regard to its effects in terms of transport, erosion and sedimentation. Here, though, the argument is made for wind-blown particles acting as weathering (rather than erosion) agents. The packing of wind-blown material into a pre-existing crack operates akin to salt or ice crystal growth - exerting an expansive, tensile stress on the crack. That extraneous material was found in cracks of rocks in the dry valleys suggested the hypothesis. The simple calculations of the possible forces involved justified the hypothesis, by showing that forces could be sufficient to cause crack propagation and rock failure. This new weathering mechanism adds a further dimension to the consideration of weathering in cold, hyper-arid environments, including those of the inner planets in our solar system.

The last paper in the above list (Hall, *et al.*, In Press b) is an expression of the relevance of the weathering work presented in this thesis plus it also shows the wider significance of the studies within the current understanding of periglacial processes and landforms. As part of the International Geographical Union Working Group on "Periglacial Processes and Climatic Change" activities, a group of papers are planned that will deal with major advances in our knowledge and pertinent questions or issues associated with a number of key topics. As part of that undertaking I have been asked to be involved with two papers, the first is the one pertinent to this chapter (the other, Thorn and Hall, In press*. is cited in Chapter 5) and deals with the issues related to cold region weathering that are the very essence of this thesis. The paper will deal with the many presumptions that comprise much of the foundation of periglacial weathering concepts together with a discussion of the many advances that have been achieved.

Attention is, however, drawn to the following papers, presented in other chapters, that have within them details regarding how certain processes may operate:

- Hall, K. 1986b. The utilisation of the stress intensity factor (K_{IC}) in a model for rock fracture during freezing : an example from the maritime Antarctic. *British Antarctic Survey Bulletin*, 72, 53-60.
- Hall, K. 1986c. Freeze-thaw simulations on quartz-micaschist and their implications for weathering studies on Signy Island, Antarctica. *British Antarctic Survey Bulletin*, 73, 19 - 30.
- Hall, K. 1988e. A laboratory simulation of rock breakdown due to freeze-thaw in a maritime Antarctic environment. *Earth Surface Processes and Landforms*, 13, 369-382.
- Hall, K. 1988b. The interconnection of wetting and drying with freeze-thaw : some new data. *Zeitschrift für Geomorphologie, N.F. Suppl. Bd.*, 71, 1-11.
- Hall, K. 1991a. Rock moisture data from the Juneau Icefield (Alaska), and its significance for mechanical weathering studies. *Permafrost and Periglacial Processes*, 2, 321-330.
- Hall, K. 1992b. Weathering processes on the Byers Peninsula, Livingston Island, South Shetlands Islands, Antarctica, *In* Y. Yoshida, K. Kaminuma and K. Shiraishi (eds.): *Recent Progress in Antarctic Earth Science*. Terrapub, Tokyo, 757-762.
- Hall, K. 1993c. Enhanced bedrock weathering in association with late-lying snowpatches: evidence from Livingston Island, Antarctica. *Earth Surface Processes and Landforms*, 18, 121-129.
- Hall, K. 1993a. Rock temperature data from Livingston Island (Maritime Antarctic): Implications for cryogenic weathering. *Proceedings of the 6th International Permafrost Conference, Beijing*, 1, 220-225.

- Hall, K. 1993b. Rock moisture data from Livingston Island, (Maritime Antarctic) and implications for weathering studies. *Permafrost and Periglacial Processes*, 4, 245-253.
- Hall, K. 1997a. Rock temperatures and implications for cold region weathering: I. New data from Viking Valley, Alexander Island (Antarctica). *Permafrost and Periglacial Processes*, 8, 69-90.
- Hall, K.1998a. Rock temperatures and implications for cold region weathering: II. New data from Rothera, Adelaide Island (Antarctica). *Permafrost and Periglacial Processes*, 9, 47-55.
- Hall, K. 1998c. Some observations and thoughts regarding Antarctic cryogenic weathering. *Proceedings of the 7th International Conference on Permafrost.*
- Hall, K. and Hall, A. 1991. Thermal gradients and rates of change of temperature in rock at low temperature: New data and significance for weathering. *Permafrost and Periglacial Processes*, 2, 103-112.
- Hall, K. and Hall, A. 1996. Weathering by wetting and drying: Some experimental results. *Earth Surface Processes and Landforms*, 21, 365-376.

The material presented here, in conjunction with those dealing with process(es) in the other chapters, provides some insight in to cold region weathering processes as well as raising questions regarding those processes. In some ways, at this stage, it is as much the questions as the answers that are important. Historically it has usually been the occurrence and effects of freeze-thaw weathering that have been *assumed* to constitute the question and research has been structured with respect to this. Within the papers cited here this assumption has been questioned and arguments made, and backed up by findings in a number of instances, that suggest alternatives. As Thorn (1988, p. 13) states with respect to this issue "...it should be seen as closing the book on the ill-founded but oft-cited notion that freeze-thaw weathering is ubiquitous in periglacial regimes". Thorn indicates that rock breakdown "...is certainly likely to stem from processes other than freezing and thawing in many instances" (Thorn, 1992, p. 11). The bigger problem of changing this "ill-founded" notion still remains and is shown to be a major issue in Chapter 5, with respect to landform genesis and evolution in cold climates. Thus, the discursive, rather than data-rich, papers presented here are important for they go to the center of the very questions that have dogged periglacial geomorphology for some considerable time. While trying not to belabour this point, it is hoped that the extensive citations given in the Introduction substantiate the need for this re-evaluation of the cold region weathering paradigm. By defining the right questions it is possible that we may then go some way to finding the right answers, rather than perpetuating the repetitive assumptions regarding cold region weathering processes.

CRYOGENIC WEATHERING:

Proceedings of a Workshop on Mechanical Weathering held in Caen, France, April 29–May 2, 1991

Organised by Kevin Hall and Jean-Pierre Lautridou

PREFACE

An understanding of the processes by which weathering occurs under freezing and sub-zero temperatures constitutes important and basic information relevant to permafrost conditions and periglacial environments. The Editors of *PPP* are pleased, therefore, to present an issue devoted solely to this topic. We hope that *PPP* can be the forum for further thematic issues upon other topics of interest to the permafrost and periglacial communities.

H. M. French, E. A. Koster and A. Pissart
December, 1991.

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Introduction—Cryogenic Weathering

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Cryogenic weathering refers to the combination of mechanico-chemical processes which cause the *in situ* breakdown of rock under cold-climate conditions. Yet the question often arises as to whether or not it is well understood. For example, Ives (1973) cited '... the efficiency of freeze-thaw in the role of bedrock disintegration ...' as one of the four main areas of deficiency regarding our knowledge of arctic and alpine processes. Eight years later French (1981) reiterated this inadequacy. As a result, a number of studies conducted during the 1980's either questioned the basic foundations of certain cryogenic weathering processes or offered alternative explanations (e.g. Konishchev, 1982). Studies such as those by McGreevy and Whalley (1982, 1985) and Hall *et al.* (1986) examined the availability of data on rock temperatures, rock moisture content and rock moisture chemistry, respectively. Without such data it is difficult to deduce what processes take place in the field, and the applicability of laboratory simulations is brought into question. Subsequently, Thorn (1988) critically reconsidered the role of freeze-thaw weathering in association with late-lying snow patches. Progressing in a different direction, Hallet (1983) proposed a theoretical approach. While not actually solving the problem of where and when freeze-thaw weathering operates, he nevertheless produced a timely reassessment of the somewhat simplistic assumptions previously made.

During this time of questioning, other investigators had been following alternative approaches. For example, laboratory simulations and modelling, notably in France (e.g. Lautridou, 1982; Ozouf, 1983; Letavernier, 1987), emphasized the granulometry of many frost-weathered lithologies and compared them to the particle size distributions obtained for various Quaternary sediments. Elsewhere, particularly in Japan, South Africa and England, attempts were made to investigate the nature of freezing within rocks via the use of techniques such as ultrasonics (e.g. Matsuoka, 1988; Hall, 1988), and photoelastic investigations utilized Perspex models of rock cracks (e.g. Davidson and Nye, 1985). Thus, information on the fundamental weathering controls of temperature and moisture expanded in both number and geographic distribution (e.g. Francou, 1988, in South America; Whalley *et al.*, 1984, in the Himalayas; Miotke, 1982, in Antarctica).

Several weathering processes other than freeze-thaw are now recognized to be significant under cold-climate conditions. Among these are (i) the recognition of the interaction of freeze-thaw with salt weathering (e.g. Williams and Robinson, 1981); (ii) the role of wetting and drying (e.g. Hames *et al.*, 1987; Pissart and Lautridou, 1984); (iii) thermal fatigue (Hall and Hall, 1991); and (iv) the role of chemical (e.g. Dixon *et al.*, 1984) and biological (e.g. Broady, 1981) weathering. In some instances it appears that processes *other* than freeze-thaw constitute the main cause(s) of breakdown.

Against this background a meeting was held at the Centre de Géomorphologie (Caen, France)

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during April, 1991, under the auspices of the International Geographical Union (IGU) Commission on 'Frost Action Environments' and the International Permafrost Association (IPA) Working Group on 'Periglacial Environments'. A number of papers presented at that meeting are published in this issue of *Permafrost and Periglacial Processes*. All have been subject to normal peer review. As such, they constitute but the bare bones of the questions and discussions which occurred.

Many aspects of cryogenic weathering are covered in this collection of papers. They include the laboratory investigations of weathering mechanisms, regional studies, problems of field data acquisition and theoretical models. It is hoped that these papers may provide a stimulus to all interested, either directly or indirectly, in cryogenic weathering.

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FREEZE-THAW WEATHERING: NEW APPROACHES, NEW ADVANCES AND OLD QUESTIONS

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1. INTRODUCTION

In a 1973 review of notable gaps in our knowledge pertaining to arctic and alpine geomorphology, Ives (1973, p.1) identified four major areas of which one was "...The efficiency of freeze-thaw processes in the role of bedrock disintegration...". The hope of Ives was that by identifying key questions, it might stimulate research in that direction. However, in 1981, French (1981, p. 267) in a further review, reiterated the continued inadequacy of our knowledge regarding rock weathering under cold conditions. McGreevy (1981), in the same year, in the first of his series of key reviews on weathering processes (McGreevy and Whalley, 1984; Whalley and McGreevy, 1985) detailed the state of knowledge regarding freeze-thaw from the point of view of experimentation, mechanisms and field observations. Once more our lack of understanding is emphasised, but now it is clearly noted that it is the lack of base line field data, as the basis for experimentation or mechanism determination, that is the problem. At that time of writing the only studies that contained the essential field data in the English language (i.e. work may well exist in such as Russian, e.g. Konischev and Rogov, 1978), were those of Gardner (1969), Thorn (1979, 1980), Thorn and Hall (1980) and Hall (1975, 1980).

Subsequently, in two further key papers, McGreevy and Whalley (1982, 1985) identified and summarised the state of knowledge regarding the importance of rock temperature variation and rock moisture content respectively, in freeze-thaw weathering. For the former they concluded, amongst others, that there is a need for field data acquisition and that field and laboratory experiments need to be refined. With regard to rock moisture status, there is a call for data as none exist and that cognisance of this should be taken in simulations if they are to be of any meaning. To further stimulate 'thought' and mitigate against simplistic field judgements, Williams and Robinson (1981), McGreevy (1982) and Fahey (1985) all showed that the saline nature of the water which is subject to freezing indicates that both 'frost' and 'salt' weathering are operative, and not just the former. The actual relationship and inter-operation of 'frost' and 'salt' is, though, still unclear.

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Many questions have been asked, areas of data inadequacy specified, and new mechanisms hypothesised, but have any advances been attained? In short, yes, enormous advances have been made in the context of theoretical modelling, data acquisition, simulations and the application of new technology over the last five years. With respect to the investigation of weathering in South Africa little cognisance has been taken of these advances, and judgements still tend to be both subjective and qualitative (Hall, 1988). Presented here are some of the new techniques, approaches and findings regarding freeze-thaw, with particular emphasis on work that has been undertaken in South Africa as part of a joint investigation with British Antarctic Survey.

2. NEW APPROACHES

One new approach, which has grown in importance in recent years, is that of theoretical modelling. Hallet (1983), and later Walder and Hallet (1985, 1986), suggested a theoretical model, to explain the breakdown of rock due to freezing, using the well known theory of frost heaving in soils as an analogue. Hallet argued that the pressure exerted by ice growth is not primarily the result of volumetric expansion, concomitant upon the water to ice phase change, but rather that "... the induced pressure is assumed to arise thermodynamically because mineral surface effectively decrease the chemical potential of water in the close proximity...". Thus, there is unfrozen water existing during subzero conditions and this water flows towards the mineral surface and exerts a pressure; really a form of hydration shattering. Ultimately, this led to the formulation of a simple model based upon the stress intensity factor K_{IC} (the strength of the singularity in the cracktip stress field that tends to produce opening mode failure). Idealising rocks as isotropic linear elastic media, the suggested model is:

$$K_{IC} = \left(\frac{\pi\ell}{2}\right)^{\frac{1}{2}} (P + \sigma)$$

where ℓ is the length of a two-dimensional crack, P is the pressure applied inside the crack, and σ is the "applied" normal pressure perpendicular to the crack plane.

The model recognised a number of factors that exert an influence upon its operation, namely lithology, thermal regime, moisture content and moisture chemistry. However, it was predicted that the most rapid breakdown will occur in the temperature range -5° C (Hallet, 1983), with slow rates of cooling being the most conducive to destructive pressures (Walder and Hallet, 1985) but with less-than-saturated conditions decreasing the efficacy of the process (Walder and Hallet, 1986). Ultimately their findings led Walder and Hallet (1986) to suggest the general applicability of their model. However, despite its wholehearted adoption and further development by some workers (e.g. Tharp, 1987) there are still some problems. Whilst the model has introduced a major new approach, it is argued (Hall, 1986a) that its full application is inhibited by the lack of field data pertaining to the controls upon the model. For instance, as will be discussed in more detail below, the recent findings on rock moisture content (Trenhaile and Mercan, 1984; Hall, 1986b), rock moisture chemistry (Hall et al., 1986), the effects of rock anisotropy on freeze penetration (Hall, 1986c), the rate of fall of temperature in the outer, wetter part of the rock and the subsequent nature of the freeze (Hall, 1987a) do not all well agree with the generalised hypotheses regarding these assumed factors within the model. However, despite these reservations, this model of Hallet (1983) and its further development by Tharp (1987) introduces an exciting new approach which augurs well for the future.

Other innovative approaches are associated with the application of new technology to the study of freeze-thaw weathering. One recent approach that yields information direct relevance to the model of Hallet (1983) is the photoelastic study of ice pressure in rock cracks (Davidson and Nye, 1985). A new technique to measure the change in the stresses as water by means of photoelastic tchniques, utilising a photometric approach with digital processing of the resulting signals, was developed. The approach uses circularly-polarised light and a rotating analyser which allow for the fast and precise measurement of the lines of constant principal stress difference (isochromats) and the lines of constant principal stress direction (isoclinics). Upon freezing of water in a slot cut into a perspex block, the progression of the ice front could be monitored and the pressures exerted calculated. Two regimes were distinguished; (1) where an ice plug extrudes and so pressures are less, and (2) where the ice plug is fixed and grows *in situ*. The association of freezing front progression and the manner of cooling give very similar results to those found by Hall (1987a) using ultrasonic techniques.

Non-destructive ultrasonic testing is a technique that has been available for some time (e.g. Ide, 1937; Hornibrook, 1939; Timur, 1968; Fukada, 1971; New, 1976) but which has been employed in recent years, as was suggested by Aguirre-Puente (1978) and Fahey and Gowan (1979), to gain new insights into the breakdown of rock (e.g. Filonidov, 1982; Zykov, et al., 1984; Mak, 1985; Crook and Gillespie, 1986; Hall, 1987a). Variation of ultrasonic pulse propagation through rock or building material can be used to indicate changes in the quality of that medium through time. As the ultrasonic velocity in air is c.330 m/sec, in water it is c.1400 m/sec, and in ice c.3000-4000 m/sec (Press, 1966), it has been possible to discern a number of factors related to freeze-thaw by means of ultrasonic testing. For instance, data on the following have been obtained; the timing and progression of the water to ice and ice to water phasechanges (Hall, 1987a), calculation of moisture content and its variation resulting from freeze-thaw cycling (Matsuka, 1984), estimation of the amount of interstitial water that is frozen (Hall, 1987a), the effects of anistropy upon freezing (Zykov et al., 1984; Hall, 1986c), and the effects of water adsorption and desorption during the thaw phase (Hall, in press). Thus, this technique has offered a great deal of new information regarding freeze-thaw (see below).

A further new approach offers great potential for the determination of rock moisture content in the laboratory situation, this is time domain reflectometry (TDR). Originally a technique developed for the analysis of moisture content in soils (e.g. Patterson and Smith, 1981) it has now been applied to rocks (Hare, 1985). TDR is a type of pulse-reflection measurement with a broad band pulse travelling down the transmission line of the TDR from which, knowing the start and end points of the transmission line, the horizontal trace length can be measured. From the then known time of the pulse in the sample, it is possible to calculate the dielectric constant of the material. The dielectric property of rock or building material is a function of such factors as constituent materials, their structure and density, the presence of water and ice and salt content and temperature (Hare, 1985, p.89). The dielectric constant of most rock forming minerals is between 2 and 12 whilst that of water is between 80.1 and 87.7 (Hare, 1985). Thus, the measurement of the relative permittivity of the rock can provide a good indication of volumetric water content. Whilst suitable for use in soils, it is nevertheless difficult and time consuming when applied to rocks. Other techniques that can be utilised for determining rock content are such as neutron moderation (Bundey, 1982), differential thermal analysis (Mellor, 1970), suction-moisture content tests (Keune and Hoekstra, 1967) and dilatometry.

Dilatometry involves the measurement of volume change of material. Davison and Sereda (1978) developed a technique for monitoring the linear expansion of a brick due to freezing, whilst Pissart and Lautridou (1984) and Hames et al.(1987) undertook a similar approach to determine volumetric changes in materials due to the uptake of moisture. Yet another very new technological development which offers a whole new approach to the study of freeze-thaw weathering is that of optical fibre sensors (Hale, 1984). Somewhat akin to the use of strain gauges (Douglas et al., in press) the fibre optic crack detection system offers a whole new insight. Fibre optic crack detection gauges are bonded to the rock and infrared signal of known amount is fed through an optical fibre loop, the attenuation of which is continually monitored. If and when attenuation exceeds a preset amount a relay is triggered which can operate recording equipment. This technique is ideally suited to the monitoring of volumetric change upon water uptake (Hall, in press) and for use during freeze-thaw experiments (Hall, in prep a). One advantage of this system is that, upon failure the infrared light source can be uncoupled and a laser attached such that visible light is then emitted from the failure point(s). This allows exact determination of where failure occurred and whether it was at one or more places.

Finally, the last new advance that is readily available, and in use, is that of the application of microcomputers for the running of simulations and the simultaneous multiple channel monitoring of a variety of sensors. With the ever increasing power and accessibility plus the decreasing costs of microcomputers, they are ideal tools for the controlling, monitoring and data manipulation of weathering simulations. As part of a joint weathering project, in the Maritime Antarctic, between the University of Natal and British Antarctic Survey, a computer-controlled simulation cabinet was constructed. By means of purpose-made hardware, the microcomputer may be programmed for temperature cycling of any length and complexity based upon data logged in the field. During running, the computer continually monitors the cabinet temperature, compares it against that which was programmed and initiates corrective action, if required. Data from six temperature sensors, a humidity sensor, ultrasonic transducers and the fibre optic crack detection system are read, stored, displayed and printed at intervals varying between 10 seconds and 99 minutes (together with the actual time) dependent upon what was chosen for that particular part of the simulation. Thus, a very sophisticated system is available that is able to undertake a variety of functions with great precision, for long periods of time, without the need of continous operator presence, and that can handle enormous amounts of data.

3. NEW ADVANCES

The new advances are largely as a result of the information derived from the application of the technology detailed above. However, one realm in which highly pertinent new progress has been made is that of fundamental field data acquisition. As was stated above, with regard to the constraints upon the use of hypothetical models, the lack of field data on such as temperatures, moisture content and chemistry, rock properties, and natural weathering rates, all serve to inhibit our understanding of freeze-thaw. Rock moisture content, as noted by McGreevey and Whalley (1985), was a largely unknown factor. However, the recent studies of Trenhaile and Mercan (1984), Hare (1985) and Hall (1986) have all begun to make available, albeit to a limited degree, data on field moisture content of rock. Three main points emerge from the available data. First, that, contrary to White's (1976) contention regarding rocks being greater than 50 % saturated and subject to freezing, a significant number of samples have been obtained (Hare, 1985, Table 4.2; Hall, 1986b, Table vi) that were in excess of 50% saturation. Secondly, that samples obtained from the faces of cliffs show very low moisture contents (Hall, 1986b). Thirdly, that the degree of saturation used in simulations is a poor representation of field conditions (Trenhaile and Mercan, 1984). All the data to date are, however, with respect to the 'averaged' moisture content of a sample and show no respect for moisture gradients within that sample (see below).

As was noted by Hallet (1983), and was a major inadequacy of the 'frost' and 'salt' weathering experiments of Williams and Robinson (1981), McGreevy (1982) and Fahey (1985), there is an almost complete absence of data pertaining to interstitial rock water chemistry. Prior to the development of utilisation of a new technique by Hall et al.(1986) there was only one analysis available, that of Kinniburgh and Miles (1983). However, the development of this new, relatively simple technique offers the potential for more data acquisition. Evidence available to date, from the Maritime Antarctic environment under investigation, indicates NaCl molarities between 0.34 and 0.57 ($\bar{x}$ = 0.47), values very small to those suggested by McGreevy (1982) to be the most efficaceous for rock breakdown.

Another aspect of field data where new advances have been made, is with respect to the physical properties of rocks (Hall, 1987b). As part of the study of weathering in the Maritime Antarctic, the following rock properties were determined; compressive strength, indentor penetration, porosity, microporosity, absorption coefficient, saturation coefficient, rock mass strength and the size range of the weathering products. This gives fundamental background data for comparison with other studies, whilst the size range of the weathering products allows for direct collation with that from simulations. A further development of this is the long-term study of rock tablets in the field and the daily monitoring of a large tablet. In the former, a large number of small blocks whose properties had been determined, were placed in the field close to data loggers that monitored the climatic factors to which they were subject. A number of tablets are retrived each year and the properties reassessed to give some idea of weathering effects and rates (Hall, in prep). The large tablet, on the other hand, was weighted daily for a whole year and the climatic conditions noted. This gave data on daily changes in moisture content plus a standard regarding the amount of weathering against which simulation results could be compared (Hall, submitted). This is what is, perhaps, most significant in that, in addition to the providing of base-line data, the information is highly pertinent to the running of real world simulations.

The use of the computer-controlled cabinet together with the application of ultrasonics and fibre optics, has led to many new advances in our knowledge of freeze-thaw processes and their controls. The sort of new information derived from these simulations include such as the effects of rock anistropy on freeze penetration and freeze mechanism (Hall, 1986c), that the rate of fall of temperature appears to be less significant than the final temperature to which the freeze is going (Hall, 1987a), and that the rate of fall of temperature in the outer, wetter part of the rock is faster than that suggested to be most effective for breakdown by Walder and Hallet (1985, p. 342). Another important finding (Hall, 1987a) was that the cooling rate of the rock, which is largely controlled by the fixed final temperature, affects the nature of the phase change,with some being rapid and extensive whilst others are slower and progressive. Further, for low amplitude freezes (to c.-6^{o} C) it was found (Hall, 1987a) that there was a need for temperatures to be maintained for a period of time before a phase change would take place. Finally, amongst the other specifics noted in Hall (1987a), it was also found that ice, during the thaw phase, returned to water between -0.7^{o} and -1.9^{o}C., and that during freezing c.80 % of the water that will freeze under natural conditions had done so by c.-6^{o}C.

Other experiments have shown that small, omnidirectionally frozen samples of rock, with high moisture contents, do not replicate large, undirectionally frozen blocks (Hall, in prep b). This result therefore questions the applicability of the majority of earlier laboratory situations to many field situations. Yet other new evidence regarding changes in the elastic properties of rocks during water absorption and desorption, together with associated hysteresis effects, suggests that wetting and drying, like the saline solutions, is an intimate part of the freeze-thaw mechanism (Hall, in press).

The above comprise, albeit extremely briefly, but a part of the many new advances with respect to our understanding of freeze-thaw. Much new information on fracture mechanics is becoming available in engineering texts, and applied aspects are ever increasing. Details of all past and present information, together with future prospects regarding 'pure' and 'applied' aspects of weathering in cold climates will shortly be provided by Hall and Walton (in prep).

4. OLD QUESTIONS

In spite of the many new techniques and advances in our understanding, some of which have been cited above, the fundamental questions with respect to freeze-thaw remain essentially the same. In truth, the question of Ives (1971) given in the introduction still remains; we do not yet know the efficacy of freeze-thaw. Really what has happened is that rather than answering questions, we have been making a start with respect to filling-in the gaps in our knowledge that are required before the questions themselves can be addressed. Concomitant with this is the fact that our data base is still pitifully small. There are inadequate data, from a variety of environments and for any length of time, on such as the thermal regime the rocks are subject to, their moisture content and the chemistry of that moisture. Other problems such as the question of moisture gradients within rocks still remain to be solved.

At the same time , the number of questions have now increased due to the recognised interaction of saline solutions, wetting and drying and thermal effects within, or parallel with the freeze-thaw process itself. We have yet to fully understand these other mechanisms before their rôle within freeze-thaw needs can be assessed. Biological activity, particularly endolithic and chasmolithic bacteria and lichens, is yet another unknown that may well exert an influence on freeze-thaw.

Thus, whilst we have made major steps forward so our vistas have increased. Considering the depth and complexity of the topic, the number of active workers are few, the longevity of most research programmes too short and the cost of the technology becoming an inhibiting factor for many. What we are able to do, though, is to refrain from making the old, simplistic qualitative comments that have become so glib regarding 'freeze-thaw'. Just because, for instance, the Drakensburg are high mountains where it is cold for part of the year, then we can no longer justifiably say "freeze-thaw takes place". We must now rather ask the (same, old) questions regarding such factors as how cold does it become, is there any water in the rock when it freezes, is there sufficient moisture to effect damage, what other processes are operative and is it perhaps not these that actually cause the breakdown? The exciting thing is that the questions are still there to be answered and that many are being investigated within this country.

5. SUMMARY

A review is given of some of the advances that have been made with respect to freeze-thaw weathering, with particular emphasis on work undertaken in South Africa. Much of the recent findings result from the application of new technology. It is shown that 'freeze-thaw' is a far more complex process than may have been hitherto thought, and that despite our advances many of the original questions remain.

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2 Weathering

K.J. Hall

Weathering is the sum of physical, physico-chemical and biochemical processes that detrimentally alter the composition, state and properties of rocks in the upper part of the earth's crust (Priklonskij, 1955, quoted in Ondrasik, 1976). It operates strictly *in situ*; that is to say, it is distinguished from the other destructive processes by the fact that it operates "in place" without the involvement of transport (Bloom, 1978). It is frequently the precursor of mass-wasting and erosion. Thus, weathering is not synonymous with erosion (thereby precluding within its boundaries such processes as abrasion) and, acting alone, does not itself produce landforms (Bloom, 1978). Rather, weathering produces altered or broken rock and it is the subsequent action of erosion and transport upon this material that generates landforms.

Weathering processes can be divided into two main groups: mechanical (sometimes termed physical) and chemical, together with a subset of both, namely biological (or biotic) weathering. Although biological processes can be considered on their own, they are not really a separate entity, but rather comprise *biologically induced* chemical or mechanical agencies. Although weathering is usually discussed under major group headings, it is important to realise that the mechanical, chemical and biological processes operate together both in parallel and in series. Rarely is rock affected by a single process only: rather, several are operative simultaneously (in parallel), and through time there is a sequence of processes that take over one from another (in series).

It is far from simple to determine what weathering process acted to cause the formation of any given landform. The type of weathering *currently* active may be in the process of destroying, rather than forming, that landform. Equally, the landform must be considered in the context of time, as climatic variation in a single year may produce a series of processes so interconnected that it is almost impossible to separate one from another in terms of their significance. That same problem must also be considered on a larger timescale, in which the overall climate has changed during recent geological time. As a result, the type and rate of processes may change. Even then, as already stated, weathering by itself does not produce the landform but rather prepares the material for action by mass wasting and other processes. Just as changes in climate have adjusted the weathering processes, so too may they have altered rates of mass wasting, thereby slowing down or speeding up the final development of the landform.

FACTORS CONTROLLING WEATHERING

Any discussion of weathering processes requires a consideration of those factors controlling the type and rate of weathering. Lack of knowledge of these controls inhibits our understanding of what is taking place in the landscape. Though there is a substantial body of theory on weathering processes, much cannot be applied due to the paucity of data on such controls as rock properties, rock moisture content and chemistry, time, climate and biological factors. In fact, despite significant advances in our understanding of weathering processes (e.g. the use of the stress intensity factor K_{IC}: Hallet, 1983) and the application of new technology to its study (e.g., the photoclastic investigations of Davidson and Nye, 1985), the results cannot be applied to the real world because of inadequate field data (McGreevy and Whalley, 1985).

As already noted, time is a factor that cannot be ignored in any consideration of weathering. For instance, a cliff that shows very limited deterioration may only recently have been exposed, by faulting or mass movement: it is not necessarily a result of ineffective weathering agents. Conversely, an area that has been subjected to weathering for a prolonged period of time and yet exhibits only limited breakdown points to the slow rate of action of those processes (as in the dry valleys of Antarctica). The time that a feature has been exposed to the actions of weathering is therefore important. The type and emphasis of the weathering process will also change as climate changes through time. In the short term, there are seasonal variations in climate, while in the longer term, climatic changes must be taken into account.

Many landforms change with time, and change in the landform may in turn modify the weathering processes that act upon it. Processes that are operative at the present may be different to those that acted in the past. This may be due less to climatic variation than to the landform itself inducing change. Special care must be exercised with the concept of time. For example, weathering processes are rarely monitored in the field for long periods. In the southern African context, where currently there seems to be a sequence of several wet years followed by several dry years, results would not be representative if data were obtained from only one of these phases. Long-term data acquisition in geomorphological studies is extremely rare, so care must be taken in extrapolating from the very short-term record to the long-term (Gardner, 1982).

Climate is the major driving force behind many weathering processes, the term "weathering" being derived from its association with the "weather" (Ollier, 1982). The main climatic controls are precipitation and temperature. Precipitation is the prime source of moisture supply in most instances (sea spray, fog and dew usually being minor components) and so its character in terms of form, frequency, duration and amount is very important. Various combinations of these four factors will promote different weathering processes; other controls, such as rock properties, being equal. Some reactions will operate best with an equable distribution of rain, whereas others will be enhanced by seasonal changes (Ollier, 1982).

Associated with the availability of moisture are the effects of temperature. Chemical processes accelerate with increases in temperature, with, roughly speaking, a doubling in intensity for every 10°C rise. Hot, humid regions will experience extensive chemical weathering processes, and regions with low temperatures may experience diminished chemical weathering rates, but accentuated mechanical processes. However, even in regions such as Antarctica, where air temperatures may rarely exceed 0°C, it is possible for snow on rocks to melt under the heating effects of direct solar radiation. The effect of radiation in causing heating of rock can also play a major role in promoting thermal fatigue in rocks. In this regard, aspect may also be important, since one side of a feature may experience shadow effects, and so be very cold, while the other side is being warmed by the sun. On the larger scale, as in a valley, weathering processes on the two valley sides may be markedly different.

Other factors, such as wind, which may promote cooling or desiccation and inhibit plant growth, can also be important, as can solutes in the precipitation. Rain or snow in maritime regions usually has a high NaCl content, and man-made pollution may have introduced extra carbon dioxide and sulphur dioxide or sulphur trioxide, all of which will greatly accelerate chemical weathering. Thus, proximity to industrial environments may cause an adjustment both to weathering processes and to their rates as a response to the introduction of pollutants.

Many European workers regard climate as so important that climatic boundaries to different types of weathering zones have been suggested (e.g., Peltier, 1950). In the southern African context, Weinert (1965, p. 41) went so far as to state "... climate is the most important factor in weathering... ". Climate is intimately connected with time, and its effectiveness, in any one setting, is intermeshed with rock properties and biological activity. General climatic data, however, are usually inadequate for serious research, and more detailed microclimatic data on, in and about the features under study, are required.

Biological activity can exert a strong influence on the type, timing and rate of weathering. Details of the extent of involvement of biological agents in weathering are still poorly understood. However, certain aspects do stand out clearly. Vegetation (e.g., turf-making grasses) may aid water retention and so increase the potential for a number of chemical weathering processes. Vegetation may also act to diminish erosion and thus the removal of weathering products, resulting in a reduction in the rate of weathering. The vegetation cover has a marked affect on the microclimate, and may affect weathering processes. The actual rate of weathering would depend upon the balance between weathering-increasing factors and weathering-reducing factors (Ollier, 1982).

Plants may introduce organic acids to the rocks, may take up certain ions and cause chelation, may encourage or inhibit the activity of micro- and macro-organisms, alter the microclimate, and even physically grow within the rock itself. Inside the rock, organisms such as bacteria can constitute a prime weathering agent (even if not a particularly fast or intensely active one): the endolithic and chasmolithic bacteria often present in Antarctic rocks are an example. It is well known, for example, that certain bacteria can break-up and absorb specific minerals. The effects of these bacteria in the landscape are largely unknown at present.

Changes in plant type and density may have serious implications in terms of altering the type or rate of weathering. These changes to the biological world may be a result of plant succession or climatic change, or may be due to human action. The removal, burning or introduction of plants may alter the type and rate of weathering. It has already been noted how the introduction of pollutants into the atmosphere can affect rates of chemical weathering. In many urban environments weathering is accelerated by the direct application of salts and other chemicals as de-icing agents. These salts can cause extensive salt weathering of concrete, bricks and other building or road materials. Thus mankind is a direct, as well as an indirect, biological agent affecting weathering processes.

The varying properties of rocks exert a great influence on the type, form and rate of weathering that will take place. These properties include such factors as rock chemistry, rock strength, colour, size and thermal properties of the constituent minerals, bedding and jointing, pore size, porosity, microporosity, permeability, water absorption coefficient, saturation coefficient, isotropy, and inherent residual stresses within the rock.

Rock chemistry plays a very important role in chemical processes. However, with the exception of a few monominceralic rocks, such as some limestones and quartzites, most rocks are composed of a mineral assemblage of varying proportions and so the presence and abundance of certain minerals will, to some extent, aid or inhibit the action of chemical agencies (see Spears, 1986 for details). Those same minerals will also play a role in mechanical weathering, for their intrinsic albedo and thermal characteristics, taken together with their relative abundance and size within any given rock, plus the texture that assemblage may impart, can greatly affect the passage of heat and the generation of thermal stresses (McGreevy, 1985; Berg and Esch, 1983; Kerr *et al.*, 1984; Johnson and Parsons, 1944).

The tensile strength of the rock is also an important parameter, at least for mechanical weathering. Strength depends on mineral assemblage insofar as it is determined by the type and size of minerals and the way that they associate with each other, but it is more than the sum of the individual minerals, for the strength of the rock may be dependent on the manner in which the minerals are arranged. Some rocks (e.g., granite) have their minerals tightly interlocking and with no preferred orientation or association and so, particularly as the component minerals are relatively strong in themselves, the rock has a high tensile strength. Conversely, other rocks may have their minerals arranged with a preferred orientation or distribution (e.g., the laminar assemblage found in such rocks as gneisses, schists, slates and shales). That laminar attribute dictates that the rock has inherent planes of weakness; i.e., the rock is weaker parallel to the laminae than transverse to them (Hall, 1987).

Another factor that affects the strength of rock, but is also important with respect to such attributes as permeability, is bedding and jointing. These lineaments within the rock provide channels for the ingress and movement of water within the rock mass. Thus their occurrence is important as they increase the permeability of the rock and the presence of water reduces overall rock strength (Broch, 1979). Permeability, a measure of the fluid-transmitting capacity of a rock (Curtis, 1971), is important to both chemical and mechanical weathering processes *and* to the removal of weathering products from within the rock, either in solution or suspension.

Although bedding and jointing are major avenues abetting permeability, the pores of the rock may also hold and transmit water. The amount of water that can be moved through the pores depends on their connectivity with other pores and their size. Some rocks (e.g., schists) may have very small pores, within which molecular attraction holds any available water in such a way that it is highly immobile. Thus permeability, porosity and pore size, water absorption coefficient (a measure of the amount of water that can be absorbed in a specified time) and saturation coefficient (the amount of water absorbed in 24 hours as a fraction of the available pore space) are all major factors to be considered (Cooke, 1979; Hall, 1986a) when assessing the role of rock properties.

Details pertaining to rock strength, permeability and porosity can be found in such texts as Goodman (1980), Bell (1983) and Roberts (1981), and descriptions of the methods of determining these parameters are given by Brown (1981).

The amount of water that enters the rock will depend upon both the properties of the rock and the climate. Once inside, it exerts a major influence on many chemical and mechanical weathering processes. With respect to interstitial rock water, the main questions that need to be asked are, first, how much water is in the rock? (i.e. how saturated is it?), second, how does this amount vary with time?, third, what is the distribution of that water within the rock?, and fourth, what is the chemistry of the water? In view of the importance of the above four factors to weathering, it is surprising that there are negligible data on any of them!

Measurements of the amount of interstitial rock moisture in field situations are extremely rare. For example, despite its importance in the process of freeze-thaw, there are only three studies in which attempts to quantify this parameter have been made (Ritchie and Davison, 1968; Trenhaile and Mercan, 1984; and Hall, 1986a). These studies have done little to substantiate the qualitative judgements used in laboratory studies, where attempts are made to deduce processes. Trenhaile and Mercan (1984) point out that most laboratory experiments use saturated, or near-saturated, conditions and that these are a poor representation of field conditions. Data on rock moisture distribution *within* the rock are even rarer, especially in field situations. One may concur with McGreevy and Whalley (1985, p. 338) that "... under natural conditions the surface layer of a rock is likely to contain most moisture", and that a moisture gradient from the surface inwards will be found, but data to substantiate this are rare.

The only information known is that of Roth (1965) from the Mojave Desert, where a block of rock was blown up with dynamite (!), samples taken and their moisture content ascertained by weighing, drying and then reweighing. This study did, in fact, indicate a moisture gradient, which was asymmetric in form due to aspect. The amount of water that *could* have been held was not established, and thus the variation in the degree of saturation was not measured.

Rock water chemistry is also a major problem, despite its significance to many processes. Kinniburgh and Miles (1983) undertook the extraction and analysis of water from a single sample of chalk, and Hall *et al.* (1986) developed a new technique for analysing the solute content of interstitial rock water and applied it to samples from the Maritime Antarctic; otherwise no data are available. The effects of saline solutions on, for example, the lowering of freezing points, on causing chemical reactions, and facilitating salt weathering are largely unquantified. The problem is perhaps best exemplified by the recent pointed debate between Goudie and Cook (1983) and McGreevy and Smith (1983) regarding the nature of the salts found in rocks in hot, arid environments.

Many factors exert control on the type, form and rate of weathering. Once more it is emphasised that these factors interact, aiding and abetting each other, and that they may well change through time either as a result of external influences (e.g., climatic change) or self-induced effects. The weathering processes can only be interpreted if there is adequate knowledge of the controlling factors. Generalisations and unverified assumptions should be avoided, and it is necessary to gather pertinent data before identification of weathering processes is possible.

WEATHERING PROCESSES

Details of weathering processes can be found in a variety of textbooks (e.g., Ollier, 1982; Selby, 1982; Trudgill, 1986; Colman and Dethier, 1986; Hall and Walton, in press), and the journal *Progress in Physical Geography* has frequent reviews of new information or approaches on almost every facet of weathering. The understanding of weathering processes is currently in a state of flux, partly because of the recognised inadequacy of data on controls, but also as a result of the input of new approaches and technology. Before considering weathering in a southern African context, it may serve some purpose to note some of the advances in technology that are being used, and to consider briefly a few of the inadequacies of some of the well-entrenched concepts.

Technology has advanced considerably since that which was available to earlier researchers such as Blackwelder (1925) and Griggs (1936), who investigated thermal fatigue, or those like Potts (1970), who attempted to simulate freeze-thaw weathering. Non-destructive ultrasonic testing allows the continued monitoring of changes within the body of a rock (Fahey and Gowan, 1979) and can detail such factors as the timing and nature of the water to ice phase change (Fukada, 1971; Hall, 1986b, in press a), the effects of anisotropy on freezing (Hall, 1986b), internal damage to a rock (Hall, in press a), and the growth of the volume content of ice in a rock (Zykov *et al.*, 1984). It can also be used to monitor changes in the elastic properties of a rock caused by the absorption of water during wetting and drying (Hall, in press b) and highly sensitive transducers have been used to record the volume change taking place in a rock during the wetting phase (Pissart and Lautridou, 1984). Photoelastic techniques have been used to monitor the form and rate of phase change during freezing of water in a laboratory model (Davidson and Nye, 1985), and time domain reflectometry (TDR) has been applied to rocks to determine volumetric water content and time of freezing (Hare, 1985). Strain gauges have been utilised to monitor changes in rock during weathering in laboratory simulations (e.g., Douglas *et al.*, in press), and the newly developed "OPTICAT" fibre optic crack detection system has also been successfully employed (Hall, in press b). Use is now made of computer-controlled climatic simulation cabinets (Walton and Hall, in press), which not only give a high degree of accuracy and flexibility of use, but can also be "driven" directly by micrometeorological data logged in the field. In chemical weathering, the application of the plasma atomic absorption spectrometer has allowed the detailed analysis of the solute chemistry of interstial rock water (Hall *et al.*, 1986).

The techniques cited above are examples of those now available and being employed in both the field and laboratory, and are providing insights that were unobtainable 10 or 15 years ago. Concurrent with these advances in technology has been the recognition and application, by geomorphologists, of work and approaches developed by engineers. A classic example is the introduction of rock fracture mechanics and fracture toughness testing (Ouchterlony, 1980) into weathering studies (Hallet, 1983). The application of the stress intensity factor (K_{IC}) to freeze-thaw weathering (Hallet, 1983; Walder and Hallet, 1985, 1986) has generated a great deal of interest (Hall, 1986c), as too has the computer modelling of moisture flow within a rock as a means of explaining the selective occurrence of weathering in the formation of tafoni (Conca and Astor, 1987).

Despite these new approaches and the injection of new techniques, much still remains unsolved or unresolved. Within the body of theory many concepts become entrenched by repetition in the literature rather than by enforcement due to experimental duplication. A number of examples will now be cited as a basis for encouraging new experimental investigation and the questioning of often blindly accepted statements. For instance, thermal fatigue, so often mis-named "insolation weathering" ("insolation" *cannot* "weather") is frequently cited as an ineffective process (e.g., Clark and Small, 1982), based upon the studies of Blackwelder (1925) and Griggs (1936). However, as far back as 1896, Branner showed that the degree of expansion that could be expected for large masses of gneiss in the climatic environment of Brazil would be sufficient to cause exfoliation. More recently, Rice (1976) initiated a series of letters, under the banner "Insolation Warmed Over", that clearly showed the potential for the operation of thermal fatigue on dry rocks. Bauer and Johnson (1979) and Yong and Wang (1980) demonstrated that granites experienced micro-cracking when subjected to temperatures in excess of 72°C. Kerr *et al.* (1984) have shown that distinct temperature gradients occur within rocks in hot deserts, and Miotke (1982) showed a significant temperature differential between the radiated and shadowed sides of a rock in the Antarctic. Williams (1986) even illustrated how thermally induced spallation can be an effective form of bedrock drilling! Thus, rather than blindly following the conclusions of Griggs or Blackwelder, it might be wise to consider the statement of Brunsden (1979, p. 127): "... there is a complex series of temperature variations in the near-surface zone of rocks capable of creating differential expansion forces, but whether these are large enough to cause fatigue remains to be demonstrated by stress-field analysis and experiment."

In discussions of the chemical weathering of limestone, it is frequently noted (e.g., Derbyshire *et al.*, 1979) that limestone (calcium carbonate) reacts with carbonic acid to produce calcium bicarbonate, which is soluble in water, and that the amount of weathering is directly related to CO_2 content. However, as Gunn (1983) points out, this cannot be the way in which limestone weathers, as there is no evidence of calcium bicarbonate molecules in solution. In a similar vein, there is much discussion (cf. McGreevy and Smith, 1983; Goudie and Cook, 1983) on the chemistry of the salts involved in salt weathering of rocks in hot deserts. Different salts have different crystallisation and thermal properties and so a knowledge of which salt is actually operative in any given situation is of major importance.

It is suggested that in many high-altitude or high-latitude situations freeze-thaw is "... the most widespread type of pure physical weathering" (Clark and Small, 1982, p17). This is a simplistic presumption and does not begin to tackle the question of the actual mechanism or mechanisms involved. The mechanism of breakdown may be constrained by the 9 per cent volume expansion concomitant upon rapid fall of temperature (0,1°C min^{-1}), as suggested by Battle (1960), or the pressure of unfrozen water being pushed ahead of the freezing front (the hydrofracture of Powers, 1945), or unidirectional crystal growth of ice (Connell and Tombs, 1971), or cavitation-induced nucleation of ice (Hodder, 1976). Alternatively, it might be none of these: freezing may not take place, and rock breakdown may be due to hydration shattering (White, 1976). Such problems can be resolved only when field data are available, but, nevertheless, the actual mechanism can exert an effect on the form and rate of weathering. Salts that are in the water that freezes can also exert an effect (Williams and Robinson, 1981; McGreevy, 1982; Fahey, 1985). Thus, the interpretation of "freeze-thaw" is very far from easy, despite its frequent casual application by many authors.

Finally, there are two mechanical weathering processes that may be very important but whose mechanisms are so poorly known that they are hardly ever considered, namely wetting and drying, and dilatation. Very little is known of the effects of wetting and drying on most rocks (Ollier, 1982) and yet the very taking up of water for many other weathering mechanisms might actually be exerting a deleterious effect. Pissart and Lautridou (1984) have shown that a rock can physically expand upon absorption of moisture and that repetitions of this may cause fatigue. Hall (in press b) has found evidence to suggest that moisture uptake can affect the elastic properties of the rock. The actual role of absorption and desorption of water, despite its importance in many other weathering processes, is still unknown. Equally, dilatation (the expansion effects on unconfined rock) can be significant, with linear expansions of 0,1 per cent being recorded (Bloom, 1969). This process is not directly controlled by such factors as moisture, temperature and vegetation, but is the result of mass movement and erosion. Its effects are known from glacial environments (e.g., Lewis, 1954), but its role in the development of weathering features is still unquantified.

The foregoing discussion indicates the two main issues that should be borne in mind in the ensuing consideration of weathering in southern Africa, namely the inadequacy of many statements that have all too easily become accepted, and the need for detailed investigation utilising the new technologies that allow for non-destructive testing and insights into mechanisms and rates hitherto not possible. This is not to denigrate earlier investigations, but rather to suggest they should be viewed with a degree of caution.

WEATHERING STUDIES IN SOUTHERN AFRICA

Despite the size of the subcontinent, the diversity of climates and rock types, the variety of vegetation and the possible effects of human interference in a number of areas, Kent's comment of over 20 years ago (in van der Merwe, 1962) on weathering studies is still pertinent: "... in fact, relatively little on this topic is to be found in South African geological literature." There are a few papers, notably from engineering and geology, that deal directly with weathering, or weathering processes (e.g., Hawkins, 1978) but the bulk of information is somewhat secondary in nature, being but a part within a broader geomorphological study. As will be shown, with a few notable exceptions, much of what is available is presented either by implication or as a generalised, unquantified and unverified statement.

There are a number of general observations pertaining to weathering, such as that by Botha (1968), where, in a discussion of the Karoo System, he notes "positive weathering" and spheroidal weathering in Red Bed sandstones. "Positive weathering" refers to ferruginous concretions that project above the weathered surface, but the actual type of weathering is not noted. Likewise, it is stated (p. 109) that the Cave Sandstone weathers "... to form smooth dome-shaped hills, but the weathering is very irregular locally, and pillars and buttresses, up to 300 feet in height, are then present ..." These are but observations and no detail is included. In a similar manner, Frankel (1952) describes some "interesting weathering features along the Natal coast". Examples of honeycomb weathering in Ecca sandstone and Ecca shales are noted together with the occurrence of hollows and pits in spheroidally weathered dolerite. However, Frankel does (p. 385) suggest that the breakdown of the dolerite is due to "... chemical reactions caused by spray" aided by the abrasive action of blown sand, and that most of the pitting occurs on the northern side of outcrops. No evidence as to why it should be some form of chemical reaction rather than salt weathering is given, which is surprising considering coastal sites are prime locations for this form of mechanical weathering. Despite their obvious inadequacies of detail and measurement, these observations are valuable insofar as they do record the direct occurrence of weathering features. Numerous comments of this type are to be found in the early descriptions of the geology of South Africa (e.g., Hall, 1905, 1918, 1920; Rogers and Schwarz, 1898, 1900; Mellor, 1905).

There have been a number of studies relating to the "Pleistocene" and periglacial landforms of South Africa that have referred to weathering and nivation (e.g., Sparrow, 1967a, 1967b, 1971; Marker and Whittington, 1971; Linton, 1969; Nicol, 1973). For instance, Sparrow (1971, p. 809) suggests a whole range of landforms that are associated with freeze-thaw, notably nivation hollows. It is said that the Pleistocene environment was one of "light snowfall" and frequent diurnal oscillations about the freezing point. The source of moisture for freezing and thawing is not mentioned, unless by default it is meant to be the light snowfall, and the amplitude and duration of the oscillations about freezing point are not quantified. Even if temperature does frequently cross freezing point (is this 0°C or the actual temperature at which freezing occurs within the rock?) the water still need not freeze or thaw if there is not sufficient longevity of each phase. In an earlier publication Sparrow (1967a) suggests that although the climate was too dry for glaciation there was, nevertheless, extensive periglacial activity characterised by freeze-thaw. In some instances (p. 554) "frost-riving" is said to have produced broken material found at higher altitudes. In another publication (Sparrow, 1967b), within which snowfalls are considered to be relatively heavy, frost action is once again cited as "severe". It is stated (p. 9) that "Evidence of former frost-shattering is common in almost all areas above 6,000 feet and takes the form of a veneer of angular boulders that covers but does not mask the underlying topography." The ambiguity regarding the amount of snow is difficult to resolve, particularly when he refers to "nivation", which is a group of processes, including weathering, associated with snowpatches (Thorn and Hall, 1980).

The problem of the possible occurrence of nivation features, and the associated weathering, is further highlighted by the work of Nicol (1973) who suggests the occurrence of nivation hollows on the south-facing slopes of the Little Caledon valley in the Orange Free State. These hollows are said (p. 59) to be the product of frost weathering beneath snowpatches and on the backwall above the snow, and movement of the weathering debris over the snow surface is said to have produced protalus ramparts below the "roll zone of basalt fragments" produced by "current weathering processes". As an aside, it might be worth considering that if nivation hollows of the magnitude suggested by Nicol could exist in the altitudinal range *c.* 1 974 m to 2 163 m, then why should the top of the Drakensberg not have supported glaciers? Equally, if snow could accumulate to such an extent, is it practical to envisage permafrost-related ice wedges at the altitudes suggested by Lewis and Dardis (1985)? However, two points with respect to weathering emerge from this study: first, that some form of weathering is said to be currently active, and second, that freeze-thaw is argued to have been the main cause of large landforms. The main question that must be asked is, are these features really related to nivation? When Nicol undertook his study there existed no detailed investigation of nivation processes. Later, the work of Thorn (1975, 1976), Hall (1980, 1985), and Thorn and Hall (1980) showed that the subjective evaluations of weathering and transport associated with nivation were in severe need of revision. Freeze-thaw weathering was found not to be as active as had been argued, and to be rarely operational beneath snowpatches; in fact, Hall (1985) argues that nivation sites may well be characterised by increased transport rather than enhanced weathering. Certainly landforms consisting of nivation-like hollows exist in the Drakensberg, but in the light of more recent quantitative studies these features described by Nicol and others are, as suggested by Butzer (1973, p. 3), in need of further work.

Something that does emerge clearly from studies in the Drakensberg and other parts of southern Africa is that freeze-thaw weathering is frequently suggested to have been a major process during the recent past. In addition to the references already cited, freeze-thaw is noted by Sparrow (1964), van Zinderen Bakker (1975), Alexandre (1962), van Zinderen Bakker and Butzer (1973), Harper (1969), Linton (1969) and Butzer (1973). These studies have postulated frost action from the high Drakensberg to the Cape at Nelson Bay and near Knysna, to Montagu on the edge of the Karoo, to Grahamstown and north into the Transvaal. One must also consider it as a process possibly active at higher elevations today. However, likely as this process seems to be in either the past or the present, it must be pointed out that no quantitative study has yet been undertaken to characterise any of the controls that exert an influence on freeze-thaw processes.

The foregoing discussion on temperatures and amount of snowfall exemplifies a major problem in any weathering study: until there is a better understanding of climatic variation in the last few million years it is going to be difficult to evaluate the type of, and changes in, weathering mechanisms and rates. While this is not the place to enter into a detailed consideration of the past climate, particularly in the light of the recent work on climatology (Tyson, 1986), a few points are worthy of note. Geomorphologists appear to make vague, unsubstantiated generalisations about climatic conditions during the recent past. Sparrow's vagueness regarding "heavy" or "light" snowfall illustrates this point. Nicol's (1973) attempt to suggest the altitude of the snowline, based on a relationship to the altitude of what he interprets as protalus ramparts, and finding it located "... some hundreds of metres ..." lower than that indicated by climatic reconstructions (e.g., Harper, 1969), indicates the dangers of ignoring the placing of features within the larger framework into which they must ultimately fit. There are, though, several other studies (e.g., van Zinderen Bakker, 1975, 1976; van Zinderen Bakker and Butzer, 1973; Butzer, 1973), based on climatic reconstructions, which are more wary in their interpretation of processes.

Harper (1969), in his climatic reconstruction, details the temperatures that could have been expected, and notes (p. 97) how the "intense diurnal freezing and thawing" during winter months could produce the basalt stepping observed in the Drakensberg. Pertinently, he notes (p. 97) that although some idea of temperature might be obtained, it is "... impossible to give quantitative figures on precipitation", although it is reasonable to recognise that the lower temperatures would have produced a greater amount of solid precipitation. As noted by Butzer (1973), Harper's interpretation of the landforms and processes within this climatic framework are extremely ambiguous. Ultimately, one might well concur with Butzer (1973, p. 10) when he states that whilst climatic conditions have been ascertained fairly rigorously by a multiplicity of approaches, "The only problem is that of imprecise geomorphological observation and reasoning."

It is worth noting that Fitzpatrick (1978, p. 475) suggests that "... pedological characteristics provide a more reliable means of determining periglacial weathering than the commonly used geomorphic method ..." Certainly Fitzpatrick produces good evidence in favour of an environment characterised by freezing and thawing in the

Drakensberg region above *c.* 2 280 m, but whether this can be extended to suggesting the presence of permafrost is doubtful (Boelhouwers, pers. comm.).

The consideration of the importance of climate, as a major control upon weathering mechanisms and rates, leads directly into a number of applied studies that investigate this as their central theme. Weinert (1961, 1965) undertook an investigation of the effect of climate on the weathering of Karoo dolerites from the point of view of their possible use for buildings and road construction. Weinert (1961) classified the dolerite into four field-descriptive groupings, namely "fresh", "weathered", "badly weathered", and "soil". He did not only use field-determinant criteria for classification (e.g., colour, response upon being struck by a hammer) but was also able to show that secondary mineral percentages increased with increasing degrees of weathering. It was noted that engineering tests indicated a regional distribution of results, with the fourfold classification showing that in the western part of South Africa the state of "weathered" was not passed. This led Weinert to suggest that disintegration prevailed to the west, whilst decomposition took place to the east, of a roughly NE line that intersects the coast near Port Alfred, crosses the O.F.S. to the east of Bloemfontein, and continues towards Rustenburg. This line was considered to broadly divide two main climatic areas, with water being the prime controlling factor (Weinert, 1961, Fig. 4). The amount of decomposition increased eastwards in accordance with moisture availability. He concluded by noting (p. 329) that "... the amount of water available ... and also a combination of air and soil temperatures may thus provide a means for defining the probable type of weathering ...".

In a later paper, Weinert (1965) undertook a highly detailed analysis of the climatic factors affecting the weathering that he had previously described. Weinert distinguished between what he termed *passive* weathering factors (i.e., parent rock, topography and time), which do not cause the rock to weather but do exert an influence on the final products; the *partly passive* and *partly active* factors, such as plants and animals (including humans), and the *fully active* factor of climate. He stated (p. 32) that "Either disintegration or decomposition will predominate, depending on the climatic conditions." Weinert then generated the "N-value", which is a numerical expression for the balance between certain climatic factors (mainly precipitation and evaporation for the month of January). Using South African climatic data he was able to draw an N-contour map for the present climate; pertinently he noted that "The effects of a different past climate may occasionally mask the results of weathering which would occur under recent conditions." He found that an N-value of 5 is the most important climatic indicator, because values greater than 5 equate with regions of disintegration and values of less than 5 denote areas of decomposition (Fig. 2.1). With N-values between 2 and 5, basic rocks generate the clay mineral montmorillonite and acidic rocks produce kaolinitic clays.

It is interesting that Fair (1947), in a study of slopes in Natal, also noted that dolerite was prone to decomposition in the presence of moisture. It would certainly be interesting to investigate the processes operative to cause "distintegration" and "decomposition" in the light of the recent developments in technology and theory. Van der Merwe (1962, 1964) undertook a very detailed investigation of not only dolerite but also andesite, diabase, norite, basalt and granite. The tests consisted primarily of X-ray diffraction analyses, but he also considered Atterberg limits and potential expansion of the resulting clays. Van der Merwe (1964, p. 222) concluded that the main factors in determining which clay minerals would be produced as a result of chemi-

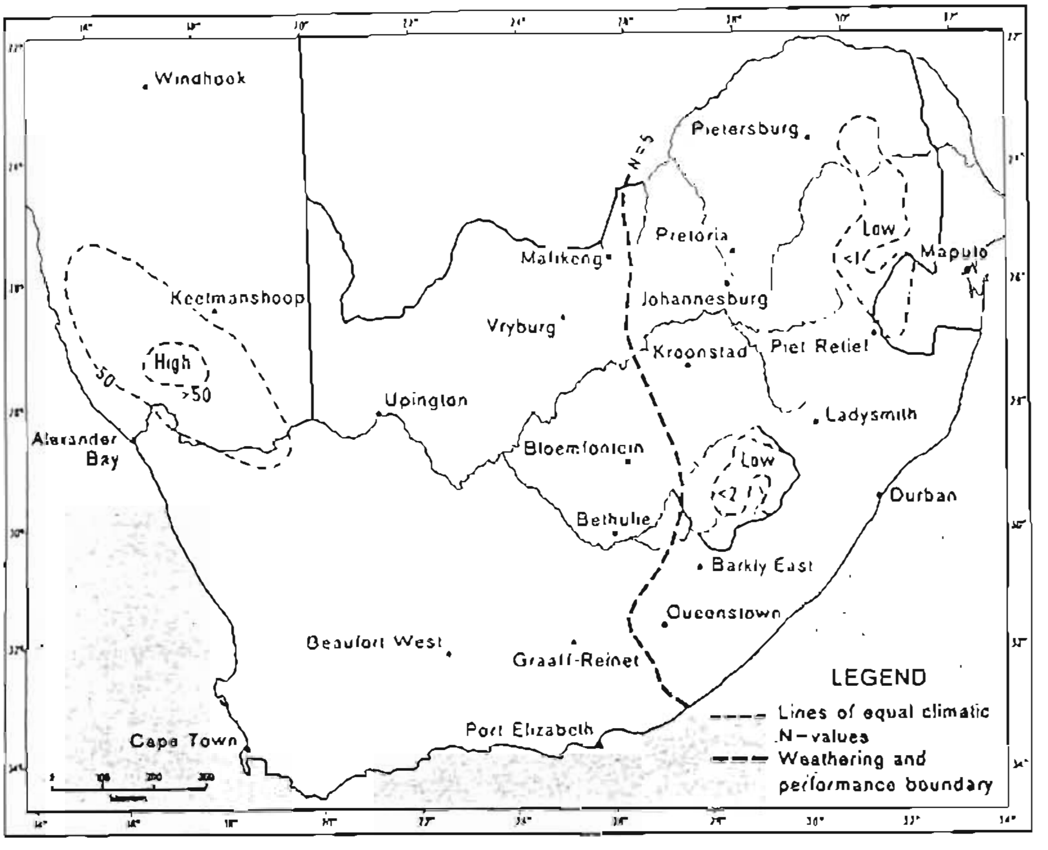

Fig. 2.1: The N=5 weathering boundary, and zones of high and low N-values (after Weinert, 1965).

cal weathering are "... apart from climate and rainfall ... [those of] ... local topography and internal drainage ...".

More recently, Orr (1979) examined six occurrences of dolerite that exhibited rapid weathering upon exposure to the atmosphere. Once more, this was an applied weathering study, based upon the observation that what appeared to be sound dolerite aggregates decomposed rapidly, frequently while still in stockpiles. Analyses of the N-value for the six sites showed that all except one (Venterstad, N=7,5) were located in areas with N-values less than 5. Petrographic, X-ray diffraction, ethylene glycol soak and slake durability tests were undertaken on the original bedrock at these sites where the rocks were known to break down within months or years of initial atmospheric exposure. It was shown that the dolerites had suffered a deuteric alteration (i.e., alterations in igneous rocks as a direct consequence of the consolidation of the magma or lava), which resulted in the formation of the swelling clay smectite. The swelling pressures exerted by the clay during expansion, upon water absorption, causes tensile rock failure. Pressures up to 500 kPa have been measured (Orr, 1979), although values of only 14 kPa are thought to have been needed to produce rock fracture. Many microfractures had significant concentrations of this clay, as did some joints within the rock mass. The swelling of the clay effected the observed weathering, upon the uptake of moisture, with the microfractured variety breaking down to sand- and gravel-sized particles and the swelling in joints loosening joint blocks. This led Orr to suggest that the dolerites with a high percentage of smectite-filled microfractures would break down the most rapidly, and that they should not be used for construction purposes.

Two recent papers on ancient South African paleosols in which weathering conditions are considered are worthy of note. Retallack (1986) reappraised a Precambrian palaeosol near Waterval Onder, and by means of chemical analyses was able to show how weathering might have taken place. Grandstaff *et al.* (1986) undertook similar studies of other palaeosols in the Transvaal. Using clear deductive reasoning Retallack shows how, at the time of formation during the Precambrian, although oxidising could take place, the atmosphere was one in which oxygen was much less abundant than at present. As plants had not yet developed, the probable source of weathering acids is likely to have been atmospheric carbon dioxide. Based upon the available evidence, Retallack was able to state that the climate was probably semi-arid to sub-humid, seasonally dry and temperate. Grandstaff *et al.* (1986) calculated the partial pressures of oxygen and carbon dioxide during the Precambrian, and showed that that of oxygen was 0,2 to 0,5 per cent of the present level whilst CO_2 was 5 to 30 times higher than present levels.

There are a number of studies in which the actual mechanical properties of rock are considered in the context of weathering. For instance, Cooks (1981, 1983) considered the compressive, tensile and shear strength, Young's modulus, Poisson's ratio and the seismic wave velocity of five rock types, and related the findings to gemorphic response in terms of landform development. Although the studies dealt with unweathered rock, the data do, nonetheless, provide the background information essential for weathering studies. In a similar manner, the rock mass strength (RMS) studies of Moon (1984) (described elsewhere in this volume) also indirectly involve weathering, insofar as the degree of weathering requires rating as one of the factors controlling rock slope development. Other data required for rock mass strength determination, such as intact rock strength, joint properties and groundwater flow, are all interactive with weathering and thus provide useful data. Equally, the sort of data required in rock mechanics (e.g., Stacey and Page, 1986) and for the assessment of rock durability (Olivier, 1979) are fundamental to all weathering studies. Olivier considered some of the weathering processes that could be expected in response to the various parameters measured and showed how these mechanisms operate in response to the controls. Dilatation (also noted by Marker and Whittington, 1971) and the swelling of clays, frequently affected by the anisotropy of the rock, were cited as operative.

Karst, and the weathering associated with carbonate rocks, are dealt with in detail elsewhere in this volume and so shall not be considered here.

From this brief review it can be appreciated that weathering studies in southern Africa are somewhat rare and that, considering the diversity of the country, they are limited in scope. Geomorphological approaches have tended to be very qualitative, with a great number of unsubstantiated statements, and rarely, if ever, tackle the problem with any degree of sophistication or depth. On the other hand, applied investigations by engineers or geologists have been somewhat more rigorous and quantitative and have made greater use of standard test procedures. However, even these applied studies frequently fall short, particularly with respect to mechanical weathering, of investigating the detail of the actual processes involved.

A further basic problem appears to be the failure to consider weathering in both spatial and temporal contexts. Certainly there is a lack of knowledge of the nature of the climate during the last few million years, but this cannot excuse the almost total neglect by some authors of the information that is available. Landforms associated with weathering cannot be considered in isolation, but must be put into the greater environmental context within which they must exist. For instance, the possibility of nivation hollows developing needs to be considered not only within the framework of possible snowfall and temperature conditions, but also with a view to whether such an environment would then imply glaciation at higher elevations. Equally, the suggestion of ice wedge polygons, with their requirement of a cold, relatively dry, permafrost environment, might be at odds with the wetter, warmer situation more conducive to active nivation. In a similar vein, there appears to be an almost total lack of modern data capture with regard to the microclimatic factors that exert an influence on weathering. Though broad-based meteorological station information is suitable for the derivation of broad zones within which certain rocks can be classified as to their weatherability for engineering purposes, it is hardly sufficient for the detailed study of process.

At the moment the potential for work is almost boundless. Apart from more detail on past climates, there is the need to obtain data on present-day microclimatic conditions at a variety of locations, to establish data bases on rock properties, investigate moisture content and chemistry, and their variability through time, and to simulate processes based upon these field data. There are multitudes of weathering forms visible throughout the country, from the coast to the Drakensberg, from the Kalahari to the Bushveld, from the Cedarberg to Timbavati; they are ubiquitous! Even their accurate detailing and measurement would be of great value. In many ways, weathering studies have yet to begin in southern Africa and there is a wealth of information waiting to be gathered.

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THE ALLOCATION OF THE FREEZE-THAW WEATHERING MECHANISM IN GEOCRYOLOGICAL STUDIES : A CRITICAL COMMENT

KEVIN HALL

ABSTRACT

In many cold region geomorphological studies weathering and weathering products are often ascribed to freeze-thaw. It is questioned whether such judgements can be made in the absence of firm data to justify process allocation. The ramifications of such unsubstantiated claims are considered and it is suggested that greater care should be taken in qualitatively allocating process.

Introduction

In a review of notable gaps in knowledge pertaining to arctic and alpine geomorphology, Ives (1973) identified four major areas of deficiency, one of which was "... the efficiency of freeze-thaw processes in the role of bedrock disintegration ...". French (1981), reiterated this continued inadequacy, while McGreevy in the same year, in the first of a series of key reviews on weathering processes (Whalley and McGreevy, 1983, 1987; McGreevy and Whalley, 1984), clearly showed that it is the lack of base line field data that inhibits meaningful experimentation or mechanism determination. In two further key papers, McGreevy and Whalley (1982, 1985) identified and summarized the state of knowledge regarding the importance of rock temperature variation and rock moisture content with respect to freeze-thaw weathering. Both papers clearly indicated the urgent need for field data acquisition and the running of more realistic laboratory simulations. In the same vein, Thorn (1988) in his recent discussion regarding the, as Tricart (1970) put it, somewhat nebulous topic of nivation, noted that a lack of quantitative information regarding freeze-thaw makes it difficult to accept such a process as a central tenet of nivation. In fact studies in both polar and alpine situations (Thorn and Hall, 1980) have shown that freeze-thaw is unlikely to predominate at most snowpatch sites: transport, rather than weathering, may be the dominant process (Hall, 1985). It is also improbable that cirques can grow from nivation hollows (Thorn, 1974), and yet many workers continue to repeat the unsubstantiated, qualitative judgements regarding the role of freeze-thaw in nivation and of its role in the developmental sequence of hollow to cirque (see similar comments in Sparks, 1986).

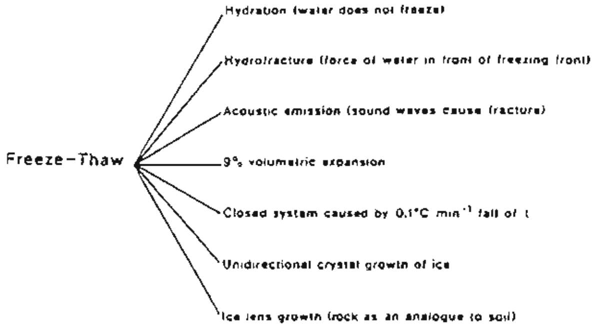

FIGURE 1: Some of the various mechanisms that constitute "freeze-thaw"

To further confuse matters recent research (e.g. Hall, 1988b; Matsuoka, 1990) has clearly shown that freeze-thaw weathering is far more complex than the oft-repeated nine per cent volumetric increase concomitant upon the water-to-ice phase change (e.g. Clark and Small, 1982; Sparks, 1986; Rice, 1988). First, freeze-thaw does not constitute a single mechanism, but rather there are a range of potential means by which it can take place and all of these fall under the umbrella term of 'freeze-thaw' (Fig. 1). These mechanisms are extremely diverse in terms of their mode of operation, in their controlling conditions and in their effects with respect to the form of breakdown and ensuing sediments. Amongst the available mechanisms are hydration (White, 1976) and hydrofracture (Powers, 1946), within which it is *unfrozen* water that causes rock damage during freezing conditions. In the case of acoustic emission (Hodder, 1976) it is not the force generated by the nine per cent volumetric change that effects breakdown, but rather the pressure from sound waves that are emitted as a result of the instantaneous freezing of interstitial water. Unidirectional crystal growth of ice (Connell and Tombs, 1971) does not rely on the water-to-ice nine per cent volume change, but rather on the growth of individual ice crystals (by water migration to the freezing centres) that grow within cracks or pores and generate tensile stresses as they impinge upon the opposing crack or pore wall. However, volumetric expansion alone may cause damage if ice extrusion is inhibited by the growth of an ice seal over the underlying, as yet unfrozen, water to create a closed system: Battle (1960) hypothesized that a rate of fall of temperature of 0.1°C min $^{-1}$ would generate such an ice seal.

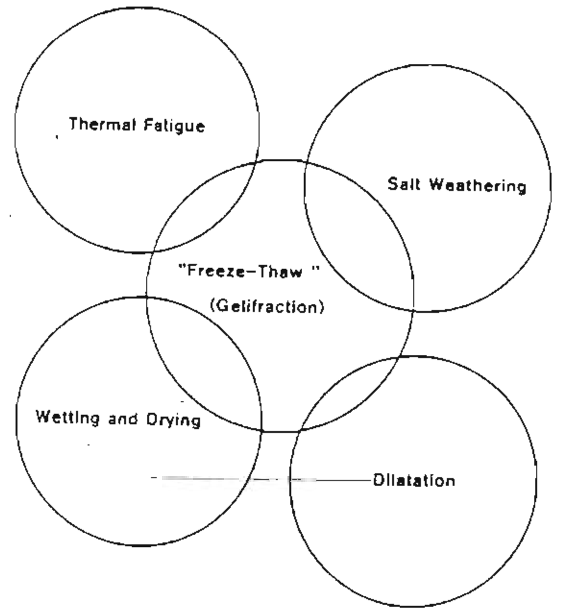

FIGURE 2: The inter-connection of various mechanical weathering processes within freeze-thaw.

Second, freeze-thaw rarely, if ever, operates alone but is usually an intimate combination of several weathering processes (Fig. 2). Wetting and drying, a little known mechanism (Pissart and Lautridou, 1984; Hames *et al.*, 1987; Hall, 1988a), must operate within freeze-thaw as the rock is wetted during the thaw phase and then, subsequent to freezing, dried during the warming that produces the ensuing thaw. The temperature cycles that induce the freezing and thawing may also cause thermal stress fatigue. In many cases the rock upon which these processes operate is likely to have been affected by either macro- or micro-dilatation, the resulting failure planes from which are exploited by these other processes. The presence of salts within the water which is subject to freezing means that salt weathering must also be taken into account in any consideration of the frost weathering process (Williams and Robinson, 1981; McGreevy, 1982; Fahey, 1985). It is also likely that chemical and biological processes are operating synergistically with freeze-thaw, the expansion and contraction of chasmoendolithic algae at 3 mm depth *inside* granitic bedrock observed on nunataks in Alaska being a typical example (Hall and Otte, in press).

Discussion

When it is stated that freeze-thaw weathering is or was operative it begs the question as to the basis for making such an emphatic judgement. If researchers *are* in a position to state clearly that freeze-thaw *does* (or did) take place then they should also have sufficient data to specify what form of freeze-thaw occurred and what its interrelationship to the other operative processes is (or was). However, it would appear that with *very* few exceptions researchers who cite the operation, past or present, of freeze-thaw have done so in an unsubstantiated manner. How can such a statement be justified when so many authors categorically state freeze-thaw to be a causative process within their study? First, there is very little data regarding rock temperature, particularly temperatures at depth within the rock (McGreevy and Whalley, 1982). Failing this it is impossible to know what the temperature regime of the rock is and whether temperatures conducive to freezing actually occurred; screen data regarding air temperatures show no relationship to rock temperatures.

Second, even should rock temperature data be available they are of little consequence without information regarding interstitial rock moisture (Hall, 1986). Unless data on the degree of saturation, the moisture gradient and the chemistry of the water are available there is insufficient information to say whether freeze-thaw did or could take place. For, unless it is known that water *was* present then it is impossible to say that freeze-thaw weathering took place, irrespective of how cold the temperatures may have been. In addition, the absence of information on the moisture gradient precludes determination of the breakdown mechanism. As these data are absent from all but three or four studies (e.g. Hall, 1986; Matsuoka, 1990), it is clear that the majority of undertakings that cite the operation of freeze-thaw weathering are, in fact, making unsubstantiated judgements. It would be ideal if field instrumentation were adequate to determine whether freezing *did* actually take place via temperature exotherms or ultrasonic information in the manner that has recently been undertaken in the laboratory (Hall *et al.*, 1989; Matsuoka, 1990). If that level of information could be obtained, particularly via ultrasonics which can discern the nature and extent of the freeze (e.g. Hall, 1988b), then valid judgements could be made. Third, and finally within the frame of this brief synopsis, the recent attempts at generating hypothetical physical models of rock breakdown due to frost action (e.g. Hallet, 1983; Tharp, 1987) are of limited help to the researcher as they also fail due to the lack of empirical data upon which they can be based.

Despite many of the shortcomings noted above there have been enormous strides forward in the understanding of the freeze-thaw process. During the last 10 years there has been a shift away from qualitative judgements towards addressing the real issues. The laboratory studies of Lautridou (1985), Hall (1986) and Matsuoka (1990) have quantified many of the controlling factors and the manner in which they operate. Gardner (1983), Trenhaile and Mercan (1984), Matsuoka (1984), Hall (1986, 1987, 1988b), Fahey and Lefebure (1988), Francou (1988) and a number of others have all obtained an ever-growing body of information regarding field conditions from a variety of geographical locations. In some instances (e.g. Francou, 1988) these data are used to help explain landform development in a meaningful way, and in other instances the data are integrated with chemical, biological and pedogenic information to attempt a first synthesis of a weathering regime. Although the range of quantitative information appears to be increasing there is an enormous range of cryogenic environments (see Hall and Walton, in press) and a multitude of different rocks, and much more information on rock, temperature and moisture attributes is still required. This is particularly the case where an attempt is made to associate landform or sediment development with the action of freeze-thaw.

All the above begs the question as to why it is so important to know whether freeze-thaw actually does, or did, take place. That, in its turn, leads to the requirement of an answer to the question that if it *did* take place then in what form, at what rate did it operate, and what were its results? Many landforms and sediments, for example nivation hollows, cryoplanation terraces, rock glaciers, and grèze litées, are said to be associated with mechanical weathering and with freeze-thaw in particular. In many instances these features are now fossil and are considered to represent some former environment and/or process association. If the arguments regarding the main causative process (freeze-thaw) are wrong or misinterpreted then incorrect judgements may be made regarding the palaeoenvironmental situation. For instance, angular fragments of rock in cold or formerly cold mountain environments are usually argued to result from frost action, but how is it known that they were not the result of (say) wetting and drying or biological weathering instead? Wetting and drying occurs in many cryogenic environments but probably operates at a different rate to freeze-thaw. With respect to biological processes, Hall and Otte (in press) found that on some nunataks of the Juneau Icefield in Alaska the major cause of angular debris is biological weathering. In addition, Eichler (1981, p. 422), from work on Ellesmere Island in the Arctic, suggested that "Not the frost but the insolation appears to be the main agent in the High Arctic temperature weathering." A more detailed discussion regarding the (under-estimated) role of thermal shock and thermal stress fatigue in causing rock breakdown in cold regions is in preparation. The problem is simply that whereas angular material may well be produced by frost action, this is not the only possible causative process available. Would there be any way of discerning the relative contributions of the various processes, all acting in both series and parallel, which could produce such angular fragments? Thus, not only would incorrect estimates of rates of formation be made, but environmental conditions different to those that actually existed may be envisaged.

Conclusions

Since the original paper of Ives (1973) enormous strides forward have been made with respect to the understanding of freeze-thaw weathering. However, many of the original questions regarding its efficacy, its place of operation, its relationship to other processes, and its role in landform development still remain unanswered. Thus, although it is reasonable to recognise cryogenic weathering as having taken place, it is rarely justified to appropriate process. It is suggested that within future geocryological studies in southern Africa care should be taken regarding casual allocation of process. Rather, an attempt should be made to look for solid, valid evidence to substantiate any such claims.

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FREEZE-THAW WEATHERING: THE COLD REGION "PANACEA"[1]

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Abstract: Freeze-thaw weathering is commonly cited as a major agency of landform development in high latitudes and at high altitudes. This is, however, largely an unsubstantiated qualitative judgment. The reality is a paucity of data regarding key factors, such as rock temperature and interstitial rock moisture content data, necessary for the correct assumption of freeze-thaw rock weathering. The problem is compounded when it is presumed that angular clasts in cold regions are the result of freeze-thaw weathering and then this argument is used as a basis for the interpretation, and possible paleoclimatic reconstruction, of Quaternary environments. In reality, angular clasts can be produced by a variety of weathering processes and there are no criteria that can identify a clast as being the product of freeze-thaw weathering. Even the term "freeze-thaw" is really a collective noun, for it encompasses a range of individual mechanisms, each requiring different controlling conditions and, potentially, generating different weathering effects. This problem is discussed and some of the fallacies outlined.

INTRODUCTION

Almost all texts on Arctic weathering processes and/or landforms cite "freeze-thaw" weathering (or any of the many synonyms) as the major process. Yet, at the same time, we can read ". . . what periglacial geomorphologists need more than any other single item is a way to determine in the field whether or not bedrock fragments have been frost weathered" (Thorn, 1992, p. 11). This contradiction is not new, but it is one that has plagued periglacial geomorphology for the better part of a century. The heart of the problem, as Thorn (1992, p. 11) explains, is that "the concept that freeze-thaw weathering dominates cold regions gained respectability long before there was the ability to test it in the field. As angular rock fragments are common in cold environments they were assumed to be the product of the dominant process, namely freeze-thaw weathering. Today it is common to assume that angular rock fragments are definitive evidence of frost

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weathering," and that ". . . the majority of periglacial researchers believe that freeze-thaw weathering of bedrock is an established fact . . . ." Thus, "the story is one of casual empiricism gathering respectability by repetition until it attained the stature of an article of faith" (Thorn, 1992, p.10) or, as Thorn (1988, p. 12) described it earlier, ". . . the status of a sacred cow within periglacial geomorphology"!

First, perhaps, it is best to be clear as to what is meant by "freeze-thaw" weathering—or any of the many synonyms (e.g., frost wedging, gelifraction, frost shattering, frost riving, etc.). The term implies the mechanical disintegration, splitting, or break-up of rock by the pressure of the freezing water in cracks, crevices, pores, joints, or bedding planes in that rock (van Everdingen, 1994). This definition generates a number of clear assumptions. First, that it is *not* a chemical or biological weathering process. Second, it requires *water*—the presence of water *within* the rock is necessary for it to operate, and thus without water freeze-thaw weathering *cannot* take place. Third it requires sub-zero [°C] temperatures of adequate duration and sufficient magnitude to make the water within the rock actually *freeze*—if this does not happen *within the rock*, then this process cannot be presumed to have occurred. Last, it requires that, by some mechanism, the freezing of water within the rock does actually effect some *damage*—i.e., freezing and thawing of the water could occur *without* effecting any damage. It also should follow that the expression of this breakdown due to freeze-thaw is *angular debris* and that any damage can *only* produce debris that is angular in form.

It also must follow from all the above that the angularity of the clasts could not, within the broad climatic conditions under consideration (i.e., in this instance a "cold" climate) be created by any other mechanism such that, *de facto*, the finding of the angular clasts *must* indicate freeze-thaw weathering. A corollary of this is that other than angular clasts *cannot* be the result of freeze-thaw weathering. If it were considered that "rounded" clasts could also be produced by freeze-thaw, then the situation would become sufficiently confused that clast shape could no longer equate to a specific environment. Furthermore, it must follow that the use of the term "freeze-thaw weathering" (or any of the synonyms) is, of itself, unambiguous as to the *exact* conditions required for its operation. In other words, it cannot hide within itself a variety of conditions of sufficient magnitude that would indicate other than a singular climatic inference.

DISCUSSION

The assumptions and/or constraints outlined above are not new. As early as 1897, Merrill noted that in the absence of water the effectiveness of freeze-thaw would be minimized. Then, in 1936 Grawe complained about the presumption of freeze-thaw and its effectiveness in the absence of confirming data. He explained that the frequently cited pressures that could be generated by frozen water (2115 kg cm^{-2}) were hypothetical insofar as no rock could constrain them. Furthermore, he noted that such pressures were *only* possible if the rock was saturated (i.e., no air was present) and that the water (which had to be pure) was held within a closed system during a period when the temperature in the rock was -22° C *and falling*. Without all of these conditions, the pressures could not be attained even if the rock could withstand them. The reality is that most rocks do not have a tensile strength much above 250 kg cm^{-2}—almost an order of magnitude lower than the theoretical maximum pressures. Despite these warnings, authors continued to cite freeze-thaw weathering as operative and, even today, texts repeat such values (e.g., Renton, 1994, p.156)—and often inaccurately!

In 1973, Ives (1973, p.1) once again questioned the role of freeze-thaw when, as part of a list of four subjects most in need of research, he included "the efficacy of freeze-thaw process in the role of bedrock disintegration." White (1976) undertook a review similar to this present one and even begged the question as to whether it was actually hydration shattering that was operative rather than freeze-thaw. Again, despite White's outlining of the many problems, noting the need for empirical data, and the suggestion of an alternate mechanism, the majority of writers continue to assume freeze-thaw weathering as dominant in cold regions. The inadequacy of temperature and moisture data were discussed by Mcgreevy and Whalley (in 1982 and 1985, respectively), but this still seemed not to engender any greater rigor in Arctic studies. As a result of the almost total absence of rock temperature and moisture data, laboratory experiments are brought into question, thereby denying that avenue of investigation. As early as 1914, Warren (p. 413) warned that it would be unsound ". . . to assume that the results of a certain experiment must also be produced by natural agencies, without evidence that similar conditions exist in Nature to those employed in the experiments." As the bulk of experiments employ samples that first are saturated, are standing in water, or are submerged in water when subject to freezing, these are ". . . certainly unnatural conditions affecting the results" (White, 1976, p. 3).

This absence of rock temperature and moisture data continues to be highlighted by a few authors (e.g., Thorn, 1988, 1992; Matsuoka, 1990; Humlum, 1992; Hall, 1993) but freeze-thaw continues to be relied upon to explain many cold-region landforms and sediments. Even the hypothetical models of rock breakdown by frost action (e.g., Hallet, 1983) suffer the inability to be tested because of this absence of data. What compounds the issue is that freeze-thaw is considered ". . . an acceptable premise upon which to base many secondary concepts (e.g., cryoplanation)" (Thorn, 1992, p. 11). Thus we see explanation of the existence of such controversial landforms as nivation hollows and cryoplanation terraces having as a central tenet the action of freeze-thaw, without recognition of the inadequacy of our understanding of this process or even any proof as to whether the process actually *is* operative.

This is not even a new problem but, sadly, one that has been ignored for a very long time. In 1914 Warren confronted the very same problem when he entered into the controversy of whether a number of flints were made by humans or were the product of Nature. Essentially the case was the same—were they the product of weathering or of human action, and, if the former, then what form of weathering? The bottom line was that the clasts themselves do not offer any evidence as to their origin, there being no characteristic or diagnostic feature indicative of a particular weathering process.

For example, angular clasts are produced by, among other processes, salt weathering, thermal stress fatigue, wetting and drying, and biologically induced mechanical processes (e.g., the mucilage expansion and contraction of endolithic algae). As all of these processes can be active in cold climates, why should it be presumed that the angular clasts are the product of "freeze-thaw"? Furthermore, in arid cold climates, it is very likely that processes other than freeze-thaw are active (there, by definition, being limited water), but angular clasts may still be produced. Even allowing that water can be available for limited periods in arid polar areas (e.g., water crystallization onto rock at dew point in the early morning and then melting from the early morning sun, etc.), it is unlikely to be either sufficient or frequent enough to cause freeze-thaw to exert a major role, but it could be sufficient to aid other processes (e.g., hydration of salts, chasmoendolithic biological activity, etc.). Thus, the presumption of frost action based upon angular clasts may incorrectly deduce that there was moisture available to freeze in what may have been an arid climate. Put simply, at the present time there is no way of deducing the origin of the clasts by their form. However, authors do continue to use the angularity of clasts as an indicator of frost action: "Because the dominant characteristic of the stratified deposits at Sonskyn is the angularity of the clasts, it is likely that frost action is the primary agent for their derivation," and "the preponderance of angular clasts . . . suggest[s] that this unit has accreted under alternating freeze-thaw conditions" (Hanvey and Lewis, 1991, p.35). To compound the problem, Hanvey and Lewis then generate a whole model for the development of these sediments based on this singular (fallacious) argument! It should be noted that Hanvey and Lewis are far from the only ones to still associate angularity with freeze-thaw (e.g., see Czudek, 1993; Heine, 1994; Coltorti and Dramis, 1995 as random examples, taken from a single journal, where freeze-thaw and angularity are presented as "facts").

To further confound this issue, recent studies (Hall, pers. obs.) in an arid region of the Antarctic found that a dark-colored, coarse-grained sandstone produced angular clasts, while on exactly the same surface, with the same exposure, etc., only 1 cm away across a lithologic junction a light-colored, coarse-grained sandstone produced rounded clasts. Chemical weathering was absent, and freeze-thaw could take place only where water was provided by melting snow, which itself was limited by the aridity of the area, and yet everywhere, on horizontal and vertical exposures, on all aspects, the light-colored sandstone produced rounded forms. A discussion of why this occurs is not appropriate here, but the important question is how would the rounded clasts have been interpreted and how would the angular have been perceived if found in a Quaternary sediment? Equally, without the clear juxtaposition of the two shapes in this area, how would the angular or the rounded clasts, if found independently, have been interpreted? Considered independently, I would have to argue that two very different scenarios, with very different climatic conditions, would have been generated—and yet here they are contemporaneous, adjacent to one another, and experiencing the same basic conditions. In a recent discussion, Ballantyne and Harris (1994) indicate much the same finding when they state: "The effects of granular disintegration are apparent from the rounded appearance of exposed rock and clast surfaces, especially on granite and sandstone mountains. Such rounding contrasts strongly with the angularity of buried clasts and bedrock." They proceed to argue that it is the granular disintegration by microgelivation that produces rounding of exposed clasts, whereas protected, buried blocks remain angular. Although I do not necessarily disagree with Ballantyne and Harris as to the explanation for the clast differences they discovered it also must be noted that the two different forms of weathering can occur simultaneously, as found here, and that it need not be only frost action that operates in this lithologically constrained differential fashion.

If the other possible weathering processes are considered, then it can be seen that those factors that control freeze-thaw also control them. The major controls on weathering are rock temperatures (i.e., range, extremes, variability, and rate of change through time and with depth), interstitial rock moisture (i.e., chemistry, distribution, amount, and state, together with their spatial and temporal variability), and rock properties (i.e., permeability, porosity, tensile strength, thermal conductivity, albedo, etc.). All of the weathering processes are affecting rock, and so the properties of the rock exert an influence on what processes can occur and at what rates. Temperature and moisture thus are the dominant factors in influencing nearly all (dilatation being the major exception) weathering processes—chemical, mechanical, and biological. There really is no one factor that is unique to freeze-thaw other than the actual freezing of water; that which constrains and controls whether there is water to freeze and whether it does freeze (and subsequently thaw) thus also exerts an effect on the other processes. Also, it must be recognized that there is a "series" of processes occurring at any one point as a function of diurnal, seasonal, and/or annual changes in temperature and/or moisture together with the synergistic effects of these combinations. Thus, there is no region where a single process operates in isolation. Therefore, even with the data to actually prove it, to say freeze-thaw is the main or even dominant weathering process in any given cold region is to hide and obscure the synergistic relationships that occur and have facilitated and/or enhanced the role of that freeze-thaw activity. To date, however, those data are absent and so the statement cannot be made.

The two "keys" here are temperature and moisture—for any one given rock type, the properties of that rock can be considered a constant. Only when considering between rock types does the importance of individual rock properties need to be added to temperature and water. In the case of both temperature and water, this refers to rock temperatures and interstitial rock water. Air temperatures are irrelevant—they are not a surrogate for rock temperatures (e.g., Thorn, 1992). The many freeze-thaw cycles cited for so many locations based on air temperature are of no consequence, as the rock may not experience any of them if covered by snow. Conversely, in high polar or altitudinal locations where the air temperature rarely rises above 0° C, the rock may experience large diurnal variations as well as shorter-term oscillations across the freezing point as a result of radiative heating. Thus, unless the available temperature data pertain to rock temperatures, they are meaningless. With respect to water, it needs to be known how much water there is, how it is distributed, what its chemistry is and, for freeze-thaw, whether it actually froze—this latter as detected either by actual monitoring or by calculation based on a knowledge of pore size, moisture distribution, rock temperature and duration, and the calculation of the freezing-point depression as a function of moisture chemistry (Van't Hoff Factor).

At this point it is worth briefly noting the following:

(1) the presence and variability of rock moisture, even in the absence of any chemical weathering, will effect weathering by wetting and drying, which also can play a synergistic role;

(2) in the presence of interstitial salts, both moisture and temperature variability will cause salt weathering;

(3) temperature oscillations alone can produce thermal stress fatigue and even thermal shock;

(4) as a result of the presence of water and the occurrence of high rock temperatures, chemical weathering could play a role, even in high latitudes and altitudes; and

(5) the presence of moisture and high rock temperatures within a porous rock could produce an ideal ecological niche for organisms (endolithic and chasmolithic colonization (Viles, 1995), which can cause both mechanical and chemical weathering).

Thus, the deduction that in cold regions it is freeze-thaw that causes rock breakdown is far from a simple presumption.

All of the information on rock temperature and moisture content is necessary, not only for deduction of which processes are operative, but also because there are a variety of freeze-thaw mechanisms and each has different constraints regarding the amount of water required, the rate of change of temperature, and the amplitude of freeze. Not only is there no one singular mechanism that constitutes "freeze-thaw," but the variety of processes that fall within this collective noun can have different effects in terms of their debris production. This also presumes the absence of any other weathering process, acting synergistically, that also could strongly influence the nature, extent, and character of the weathering.

At this point, it is useful in describing the reality of the situation to return to Thorn's (1992, p. 11) observation above that "the concept that freeze-thaw weathering dominates cold regions gained respectability long before there was the ability to test it in the field. As angular rock fragments are common in cold environments they were assumed to be the product of the dominant process, namely freeze-thaw weathering. Today, it is common to assume that angular rock fragments are definitive evidence of frost weathering. Nevertheless, the angularity of comminuted bedrock must always be strongly influenced by lithology and is certainly likely to stem from processes other than freezing and thawing in many instances." What we have is, largely, the enforcing of the concept by unqualified repetition and a distinct absence of quantitative testing.

Three examples are presented here to illustrate the preceding statement. First, in many high-altitude and high-latitude locations, water is the limiting factor. Rock temperature fluctuations are frequent and can be quite large. The problem has been the absence of data on rock temperatures (as opposed to those of the air) recorded at sufficient frequency to facilitate meaningful analysis.

Recent studies (Hall, pers. obs.) with temperatures collected at 20-second to 1-minute intervals have shown that rates of change of temperature are quite close to or exceed the 2° C per minute threshold for thermal stress fatigue. The result is fracturing of the rock, sometimes explosively (as with thermal shock), and the production of highly angular fragments. In some instances, as in the high Andes or parts of Antarctica, water can be considered nonexistent for all practical purposes. Thus, the angular debris clearly is not an indicator of freeze-thaw—which requires water—but rather of a dry environment, with sub-zero air temperatures but high radiative inputs that can generate rock temperatures on the order of 30° C. Environmental ramifications with respect to climate, vegetation, processes, etc. follow and would *not* conform to that of a "freeze-thaw" environment.

The second example is from the Juneau Icefield area of Alaska, where the destruction of granites on nunataks has been perceived to be the result of freeze-thaw. Detailed studies, however, demonstrated that moisture availability was minimal and only penetrated a few millimeters into the rock. During this same period, there were no freeze-thaw events. Rather, chasmolithic algae were found in the granite below an outer shell of 1 to 2 mm. The available moisture caused swelling of the algal mucilage such that the algae are the main cause of rock breakdown and produce angular flakes to the extent of, at some sites, more than 1 kg m^{-2} yr^{-1}. Again, angular debris is produced, but not as a result of freeze-thaw and from a very different environment than in the preceding example.

The final example is from the maritime Antarctic, where freeze-thaw is so often cited as the main process. Here there can be large freeze-thaw cycles in late autumn through the end of spring, but there is such extensive snowfall that most bedrock is insulated and does not experience thermal oscillations until the snow cover ablates. Through the summer the air temperatures are low (ca. 2° C), with rare, low-amplitude and often short-duration freeze-thaw events in the air, but rock temperatures remain above zero. There is, however, a great deal of precipitation in the form of rain, together with strong winds and intermittent sun. The end result is extensive weathering by wetting and drying. Rock fragments are angular, and while probably not *solely* produced by wetting and drying, this is, nonetheless, the major process. So, again, angular debris is produced, but in another, entirely different environment.

CONCLUSION

The argument presented in this paper is that angular clasts are *not* caused solely by freeze-thaw and therefore are not an indicator of the occurrence of this process. Greater care should be used in Quaternary and modern-day interpretations of process and in the reconstruction of climate based on such criteria. Freeze-thaw is *not* as ubiquitous as we often think and many other processes may be at least as effective. Freeze-thaw has been cited wrongly in many modern studies, and therefore its use in any paleo-interpretation is even more doubtful. Freeze-thaw is not the panacea (universal remedy) so frequently considered. We now should attempt to overcome the psychological need for citing freeze-thaw and find the actual cause—only then can we progress to a meaningful diagnosis.

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A DISCUSSION OF THE NEED FOR GREATER RIGOUR IN SOUTHERN AFRICAN CRYOGENIC STUDIES

KEVIN HALL

ABSTRACT

Recently there has been a marked increase in the number of studies pertaining to both the past and present geocryology of southern Africa. Although there is certainly a great need for a better understanding of the cryogenic processes and landforms of this region, it is suggested that a number of recent contributions lack rigour with respect to both terminology and interpretation. In a number of instances qualitative presumptions have been made and then deterministic hypotheses built upon them. If great care is not taken it may be that these hypotheses will become established by repetition in the literature without ever having been truly tested. A plea is made for more detailed work and greater exactitude in future studies, combined with a better understanding of the cryogenic literature and greater care in the use of its terminology.

Introduction

In recent discussions, Hall (1991, 1992) and Thorn (1992) have been critical of the simplistic presumption, so often made, that angular clasts found in present or former cold regions are the result of freeze-thaw action. It is pointed out that angular material can also result from thermal fatigue, wetting and drying, salt weathering, hydration shattering or biological weathering. Equally, these processes rarely work in isolation; rather, breakdown is as a result of a combination of weathering processes working both in series and parallel synergistically to produce angular material. Whilst this in itself may seem little more than semantic argument, the problem is, in fact, much greater, for upon this evidence of 'angular clasts', authors frequently erect a whole palaeoenvironment within which this material is said to equate to a climate where freeze-thaw, with an adequate moisture supply, prevails. Having then 'established' the climate, a complex morphological/process/sedimentary reconstruction is built up. The edifice, though, is dependent upon the presumption that angular clasts result from freeze-thaw weathering. The whole is perhaps best expressed by Warren (1914, p. 546), in a somewhat comparable discussion regarding the human versus natural formation of flint 'tools', when he states "we must surely make our theories accord with the facts as they are, not with imaginary facts which might conceivably be but which are not."

In marginal periglacial environments such as South Africa, it is very difficult to discern what combinations of processes have been operative in the recent past. Great care and rigour must be applied to any situation otherwise a completely misleading picture may result that could confuse and/or prejudice subsequent studies. Analogous to the concern regarding the presumption of freeze-thaw being the cause of angular material is the uncritical use of terminology. There are a range of terms which help to describe clearly a landform or process such that another when seeing this term should know, or be able to elucidate from a glossary of terms, exactly what is meant. Thus, when applying a term one should be certain as to both its meaning and its applicability to the instance in question. Many 'casual' users of terms such as 'freeze-thaw' might be surprised to realise how much we still need ". . . a more sophisticated conceptualization." of what the term really means and implies (Thorn, 1991, p.16). Equally, the doubts and/or controversy regarding some terms should also be known; the problem of utilising the term 'nivation' as discussed by Thorn (1988) epitomises this point.

Discussion

South Africa has a range of environments, processes, landforms and sediments, both contempory and fossil, that are associated with cold climates. It is fair to state that knowledge regarding the periglacial environments of the area is in its infancy. It is equally fair to state that this is a very difficult area to study (due to both its marginality and the complicating effects of post-glacial climate) but, at the same time, one with potential for providing a valuable contribution to the understanding of the world's periglacial environments (Corté and Hall, 1991).

Recently there have been a number of publications (Willcox, 1989; Marker, 1989, 1990, 1991; Le Roux 1990; Boelhouwers, 1991; Hanvey and Lewis, 1991, Lewis and Hanvey, 1991) dealing directly or indirectly with periglacial conditions in South Africa. Although these papers provide some new and valuable information they nevertheless suffer from a general slackness with respect to terminology and unjustified presumptions. Nearly twenty years ago Butzer (1973, p.1), in much the same vein, expressed similar feelings regarding a number of earlier publications pertaining to the periglacial history of southern Africa: ". . . there is a serious problem about many of the geomorphological observations or their interpretation.". It is worth considering aspects of some of these papers cited above from the point of view of both their individual arguments and their general representation of the state of periglacial science. Although this may involve partly a direct criticism of a paper, it should also be seen as a matter for reflection or an alternative view point for consideration by the southern African periglacial community, and as motivation to all to consider possible alternatives.

Direct criticism, even within a small scientific community like South Africa, is not necessarily bad. Neither is it an unusual procedure for, as shown recently by, amongst others, Richter and Haendel (1989), Walder and Hallet (1986), McCarroll (1989, 1991), and the sequence of letters between McGreevy and Smith (1983) and Goudie and Cooke (1983), the taking to task of data, hypotheses, methodology, or the inadequacy of background information, is a common practice when and where the need arises. Nearer to home, Marker (1990) and Le Roux (1990) recently entered into a debate regarding many of the problems pertaining to the periglacial of South Africa that are outlined here. By this present discussion it is hoped that the great potential contribution that southern Africa has to offer to the periglacial community at large might be achieved, and the workers in this region may be seen as competent, discerning scientists.

First, weathering due to frost action as being indicated by angular material should be considered. Boelhouwers (1991) cites Sänger (1988) as stating that frost shattering is operative at the present day in the western Cape mountains based upon the occurrence of angular clasts of up to 0.2 m in diameter. The question arises as to how, in the absence of any empirical data, it can be determined that frost action is the causative agent? To a lesser extent, the size threshold is also intriguing as no explanation is apparently presented to explain this. Is it process controlled or is it a function of a factor such as joint spacing? If the latter then this should be demonstrated, but if the former, then a detailed explanation is required as to how and why there is a process constraint upon weathering product size. Beyond all of this there is still the perception that 'angular clasts are produced by frost action'. What of other angular-clast producing mechanisms such as thermal stress fatigue, wetting and drying, biological processes, and others? Data must be provided to justify a conclusion: unquantified subjective judgements are not 'good science'.

Following on from this last point, it is noticeable that Hanvey and Lewis (1991), in an otherwise excellent presentation, solely consider frost action. Having found angular clasts in their sediment then *ipso facto* they are due to frost action : "The preponderance of angular clasts . . . . suggest that this unit has accreted under alternating freezethaw conditions. It is envisaged that the conditions were severe enough to cause frost shattering . . ."(p.35). Further, "Because a dominant characteristic of the stratified deposits at Sonskyn is the angularity of the clasts, it is likely that frost action is the primary agent responsible for their derivation" (p.35). But why should this be? Do salt weathering, thermal fatigue, wetting and drying, hydration shattering, plus some forms of both biological and chemical weathering *not* produce angular material? However, having presumed frost action *then* a whole model is evolved.

The model itself, without addressing the many pertinent questions raised by Thorn (1988), has a number of fundamental errors. It is stated (pp.35-36) that other factors can affect the action of frost upon the rock. Amongst these it is cited that thermal stress can predispose the rock to being split by frost. Whilst it is true that the synergistic role of thermal fatigue can aid the action of frost, two things must be remembered. First, that this is a *separate* process and so could cause angular fragments *on its own* and under, potentially, very different climatic conditions to those envisaged by the authors. Second, that it can synergistically operate equally well with any or all of the other weathering processes to cause angular material. The other factors mentioned (tectonically induced fracturing and petrographic composition) *also* increase the propensity for the production of angular clasts by other weathering mechanisms. Thus, again, angular clasts have no unique association with frost action.

It is also interesting to note that the petrology of the sandstone is thought to be predisposed to angular clasts (as found) whilst that of the basalt is not (and hence the observed rounded, basaltic clasts). This too is spurious argument on the preconceived notion of angularity equates to frost. Certainly, from my experience at a variety of locations, it has been observed (e.g. Hall, 1981) that in cold climates basalts as well as sandstones can produce angular clasts. Where basalts produce more rounded clasts, even in cold climates, is where there is a greater amount of water together with high rock temperatures that then produces the extensive chemical weathering to which basalts are predisposed, under warm, wet conditions.

One other point calls for clarification. In a discussion regarding angular, rhythmically bedded slope deposits it is stated (p.35) "The prerequisite conditions is *(sic)* a high frequency of thermal oscillations around 0°C . . ." One must ask what is the significance of 0°C? Freezing does *not* take place at 0°C except under ideal laboratory conditions, and thus it has little meaning with respect to freezing in rock or unconsolidated sediments. Second, does this imply that frequent oscillations between +0.1°C and -0.1°C would be geomorphically significant? Patently the authors did not mean this, but then neither did they define numbers or thresholds and so the reader is left with some uncertainty as to what sort of values *are* meant. In the same vein, the assertion by Marker (1989, p.148) that " . . . 1,0 cc of water becomes 1,9 cc as ice" is nothing but incorrect. Water has a 9%, rather than 90%, volumetric increase upon phase change to ice: 1,0 cc of water becomes 1,09 cc as ice.

Confusion abounds with respect to terminology. Willcox (1989, p.14) states that the term 'periglacial' is used to describe " . . . the effects of meltwater at the edges of snow banks" and that this is really a misnomer as the term implies 'around' a 'glacier' and, as, in his opinion, glaciers did not exist in the Drakensberg so the term is meaningless. The reality is that whilst the term was originally coined by Lozinski (1909) to describe the climate and climatically controlled features adjacent to the Pleistocene ice sheets, the term, through time, has come to be associated with non-glacial processes and features of cold climates *regardless of age or proximity to glaciers* (Washburn, 1979). In fact, the present definition is "The conditions, processes and land-forms associated with cold, nonglacial environments" (Harris *et al.*, 1988, p.63).

In the same vein, Willcox correctly relates cirques to glaciers but then produces the term 'periglacial cirques' which must, semantically at least, be a contradiction in terms. The collection of processes he associates with the features, namely freeze-thaw weathering and solifluction, are in reality usually presumed to be a function of nivation, and hence the features would be 'nivation hollows' — but here again the warnings of Thorn (1988) should be heeded before ascribing either this process association or the land-form term. Willcox (1989, p.14), in his description of processes, states that solifluction is ". . . the washing down of fine material by flowing water". This could not be further from the truth as transport by water *is not* a component of solifluction! Solifluction is the "Slow downslope flow of saturated unfrozen earth materials" (Harris *et al.*, 1988, p.78) and, following the original definition, need not require either a frozen substrate or even freezing and thawing (Andersson, 1906). Where these cryogenic components are involved the term 'gelifluction' might be better used: "The slow downslope flow of unfrozen earth materials on a frozen substrate" (Harris *et al.*, 1988, p.39). Marker (1989, p.152) also incorrectly describes the mode of solifluction movement when she states that soliflucted material *slides*; equally it is most unlikely (p.152) that boulders carried on top of the soliflucted material would "float by gravity". In addition, Marker (1989) cites frost action as acting *with* nivation rather than more correctly as being a component *of* nivation, and states that ground frozen solid in winter is permafrost, whereas permafrost is ground that "remains at or below 0°C for at least two years" (Harris *et al.*, 1988, p. 63).

On a different tack, a paper regarding cirque glaciation in Lesotho by Marker (1991) has a number of statements that cause concern, particularly for such an important topic. First, following comments regarding the possibility of both ice fields (like the ice field ranges of the Alaska panhandle?) and ice caps in the southern African mountains, it is suggested that if there were enough snow for an ice cap than there must have been enough for cirque development (p. 29). There are, I believe, two problems with this presumption. First, that there were any pre-existing hollows (of whatever origin) in which the cirques could develop. Second, and perhaps more significant, that this being a marginal area, there was *time* for cirque development if there was rapid ice cap growth and decline. As in many other locations, there is no reason why a growing ice cap would not have covered incipient cirques such that they never had time to develop. Thus, whilst such a scenario as proposed by Marker is not impossible, neither *without evidence to the contrary* need it be the only possibility.

However, if we do accept sufficient snowfall for cirque development upon the argument used that this must be so if an ice cap could develop, then why should cirques not have a southfacing preferential orientation? The bulk of the perceived cirque forms are north-facing and this orientation is justified upon the basis of a snowfence effect, but with such an adequate snowfall then the shaded, and hence better protected, colder south-facing hollows should surely be at least as prevalent? Last, whilst glaciation of the higher southern African mountains is such a big question, with so many climatic-periglacial-botanical ramifications, great care must be taken in arriving at any judgements. Whilst the purely mathematical approach *is* valid, nonetheless coalescence of form shows that similar forms can have radically different origins. Thus, the mathematical approach provides indicators but *not* answers. The answers together with their supporting evidence have still to be found. Until they are, great care should be taken in extrapolating from a map-based exercise to a real world scenario.

Conclusion

All of this may seem over critical and destructive particularly when the periglacial community of southern Africa is so small. Rather it would be better viewed as a constructive criticism as, despite the noted errors or over-simplifications, the cited papers *all* contain significant contributions to the periglacial history of southern Africa. Where care must be taken is with respect to the unqualified presumptions and, what could be called the monomaniacal assumption of frost processes. It is highly likely that a strong association with frost processes will be found, but it *cannot* be presumed and the whole edifice built upon these imaginary facts. If care is not taken the situation will revert to that described by Butzer (1973, p.1) nearly twenty years ago: "it cannot be disputed that southern Africa has experienced cold, glacial-age climates, but there is a serious problem about many of the geomorphological observations or their interpretation." The final word must also go to Butzer (1973, p.9), for in this statement he clearly defines the worries that I have attempted to discuss above: "The re-evaluation of 'periglacial' phenomena in southern Africa suggests that . . . concepts have sometimes been vague or even erroneous, and that too much interpretation has been based on high latitude preconceptions."

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SPECIAL ISSUE

28[TH] BINGHAMTON SYMPOSIUM: CHANGING THE FACE OF THE EARTH - ENGINEERING GEOMORPHOLOGY

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CONTENTS

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GEOMORPHOLOGY

The role of thermal stress fatigue in the breakdown of rock in cold regions

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Abstract

Many geomorphological studies have ignored the role of thermal stress fatigue. Engineers have long maintained that thermal stress is an effective process. Physicists and the ceramics industry have provided a theoretical foundation to explain the nature and mode of operation of thermal breakdown. Recent geomorphological studies have shown the importance of thermal stress in cold regions by means of high-frequency rock temperature data which can identify rates of temperature change in excess of $\geq 2°C\ min^{-1}$. The occurrence of weathered material in cold but dry regions can be explained by thermal stress/shock. Differential radiation receipts on slopes of varying orientation and gradient can exert a very strong effect.

Keywords: cold regions; weathering; thermal stress; building materials

1. Introduction

Most discussions of weathering in cold regions are dominated by "freeze–thaw" and its many synonyms (Hall, 1995). Thorn (1992) maintained that much of the literature relies on conjecture and that there is no clear way of determining the origin of weathered material in the field. The general presumption is that the products of freeze–thaw are angular and thus any angular debris in a present or former cold region must therefore, de facto, be a result of freeze–thaw. All of this is, though, a circular argument based on some unproven and untested assumptions and thus "the story is one of casual empiricism gathering respectability by repetition until it attained the stature of an article of faith" (Thorn, 1992, p. 10). Nevertheless, it is hard to find a discussion of cold region weathering, present or past, that is *not* dominated by the role of freeze–thaw. As an example, Hall (1995) cites a range of examples where freeze–thaw is used as the sole basis for the explanation of landforms and sediments based primarily on the finding of angular clasts; consideration of any run of recent articles will continue to show this (presumed) dominance of frost action as the cause for rock weathering. Temperature data, usually collected at 15 min to 1 h intervals, are often used to indicate the impact of freezing and thawing; e.g., Ødegård and Sollid (1993) collected data at hourly intervals and, despite an absence of moisture data and the mention of the possible role of thermal stress, used the ice segregation model of frost weathering of Hallet (1983) to 'explain' rock

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breakdown. Ødegård et al. (1993) presented temperature data at 15-min intervals and gave greater consideration to the possible role of thermal stress but, no conclusions were drawn. Matsuoka (1991) developed a model for the rate of frost shattering that may well miss *other* weathering effects as temperature data were collected (for logistical reasons) at 2 to 4 hourly intervals. The key is if data are not collected at 1-min intervals or less it is not possible to evaluate the role of thermal stresses or, in fact, to determine the freeze–thaw mechanism.

In essence, any weathering study has three main attributes: rock properties, moisture conditions and thermal conditions. The spatial and temporal variability of these factors are also critical. Although not used to the extent that it should be, information about strength, porosity, permeability, thermal conductivity, etc., is available for a whole range of rocks and minerals. There are also a range of tests and equipment for measuring these parameters, e.g., porosimeters for determining the pore size distribution, point load test units for rock strength determination, ultrasonics for measuring changes to rock strength, etc. McGreevy and Whalley (1982) undertook an evaluation of rock temperature conditions but this was done within the context of frost action. Subsequently, McGreevy and Whalley (1985) evaluated rock moisture conditions but again in the context of frost weathering. Since then a number of papers have looked at thermal conditions and moisture conditions but primarily in the context of either, again, frost weathering or with respect to the role of weathering by wetting and drying. What is almost absent, but urgently required, is an analysis of thermal data in the context of weathering by thermal stress fatigue together with a discussion of the possible role of this process. An attempt to undertake this evaluation and discussion will be presented here.

2. What is thermal stress?

Although physicists, ceramics scientists and engineers generally use the terms 'thermal stress fatigue' and/or 'thermal shock', in most geomorphological texts they are frequently (and erroneously) referred to as 'insolation weathering'. Simplistically, thermal stress fatigue is where material is subjected to a series of thermally-induced stress events, each less than that required to cause immediate failure, that, collectively with time, cause the material to fail. Thermal shock is where the thermally-induced stress event is of sufficient magnitude that the material is unable to adjust fast enough to accommodate the required deformation and so it fails. It is very possible that the continued use of the term 'insolation weathering', coupled with certain specific laboratory experiments of the 1920s and 1930s, has tended to (subjectively) favour 'frost' action as the dominant weathering process in cold regions; sun, and thus heat, seemingly contraindicated in a cold environment. A typical definition of insolation weathering might read, it "is the result of alternate warming and cooling of rock surfaces under the direct influence of solar heating" (e.g., Selby, 1985, p. 191). The reality is that insolation does not 'weather' but rather is, in most instances, the driving force. The actual mechanism of breakdown is thermal stress fatigue *resulting from* 'warming and cooling' and this could be induced by wind not just sun. Many authors consider this form of weathering as potentially operative in *hot* deserts (e.g., Ollier, 1984; Selby, 1985) although Smith (1994, pp. 39–44) provides a detailed discussion outlining some of the questions and problems with this presumption. Nevertheless, conceptually, it has been easier to conceive of the action of thermal stress in hot desert environments as compared to cold polar regions. The key is, as Smith (1994) argues for hot deserts, the lack of adequate field temperature data from the actual rocks. The problem has been compounded by the studies of Blackwelder (1933) and Griggs (1936) who, based upon their laboratory experiments, concluded that thermal stress did not work. Their findings have subsequently been repeated to the extent that, as is the case with freeze–thaw weathering, "it attained the stature of an article of faith" (Thorn, 1992, p. 10). Even relatively modern texts continue to reiterate these findings, for example, "Laboratory experiments on rock samples heated in furnaces have often failed to produce splitting and have caused some workers to doubt the existence of the process" (Selby, 1985, p. 191).

From a geomorphological perspective, Ollier (1984), one of the proponents of the thermal stress process, notes that the experiments of Blackwelder and Griggs had a number of limitations such that they were not an abstract of reality and therefore the findings are not applicable to the field situation. Among the limitations are that the specimens were polished (which increases their strength), cooling was by tap water (Griggs, 1936), the samples were very small (thereby limiting the effects of different crystal sizes, albedos and thermal coefficients), and the samples were not confined. This latter is very important, for an unconfined sample that is heated simultaneously on all surfaces can expand uniformly and becomes isothermal very quickly. Thus, there is an absence of confining pressures that exacerbate the expansion of a single face — often termed 'buttressing' (Folk et al., 1982; Winkler, 1984). Further, the laboratory experiments have really dealt with 'thermal shock', whereas 'thermal fatigue' operates in a quite different way and the laboratory experiments have not attempted to simulate this. Thus, the findings pertaining to the former (shock) provide no insight into the latter (fatigue).

The engineering and geomorphological perspectives are very different. A number of early workers (e.g., Bartlett, 1832) ascribed great efficacy to the role of thermal stress fatigue and undertook relatively sophisticated experiments to determine the effects of thermal expansion and contraction on a range of rocks used as building stone. The experiments were undertaken due to the finding of fissures that grew in the cement between blocks of stone after only a few weeks as a result of the "ordinary variations in atmospheric temperature" (Bartlett, 1832, p. 136). In other studies, Branner (1896, p. 281) argued that "The unequal contraction and expansion of the minerals composing the rock tend to disintegrate the entire mass, while the even annual and diurnal changes... cause the rock to spall off in layers of even thickness...". Branner also noted that different minerals expand and contract at different rates and that this varies as a function of the orientation of the crystals axes such that "however slight the (thermal) changes may be, they cannot go on among heterogeneous materials... without causing... disintegration of the entire mass" (p. 283). Merrill (1906) provided calculated rates of expansion for several rock types and commented that, as "slight as these movements may seem, they are sufficient to in time produce decided weakening and afford a starting-point for other physical and chemical agencies" (p. 159). These calculations, based on available rock temperatures, suggested that a sheet of granite of 100 ft (30.48 m) would produce a lateral expansion of 1 in. (2.54 cm) and that this "must tend to lessen the cohesion and tear the upper from the deeper layers..." (p. 160). Merrill (1906) also refers to the fracturing of granite due to "constant expansion and contraction from temperature changes" but that as the rock was bounded on the sides the only way it could find relief was in "an upward direction where resistance was least" and that this gave rise to "dome or roof shaped forms"(p. 232). Merrill considered this process to be particularly prevalent in cold regions and saw the role of thermal stress, accentuated by buttressing, as a major factor causing the breakdown of rock and building material.

The findings outlined above predate the experiments of Griggs and Blackwelder and, to a greater or lesser extent, reflect the thinking of geomorphologists. The division comes *after* the work of Blackwelder and Griggs. Geomorphologists subsequently moved away from the concept of thermal stress (e.g., Reiche, 1950, p. 10) but engineers did not. In fact, (from the literature I have been able to find) engineers did not even *refer* to the studies of Griggs or Blackwelder. Johnson and Parsons (1944) studied 123 specimens, including crystals of calcite, quartz and feldspar, over the temperature range $-20°C$ to $+60°C$ with a heating rate of $0.5°C\ min^{-1}$. They found the orientation of crystals had an influence on expansivity: $\Delta L/°C$ for calcite parallel to the *c*-axis being positive while normal to the *c*-axis it is actually negative. For quartz, $\Delta L/°C$ normal to the *c*-axis is greater than that parallel to the *c*-axis. So, if a rock "has more of its component crystal grains orientated in one direction than in any other, the coefficient of expansion in that direction will be different from those in other directions" (Johnson and Parsons, 1944, p. 117). Thus, where there are several mineral types in a rock then different crystal orientations will lead to a substantial variation in expansivity. Johnson and Parsons (1944, p. 121) also observed that in cold regions "the materials expand on heating until temperatures between 0°C and 10°C... with further heating they either expand at a lesser rate... or contract until temperatures between

Fig. 1. Examples of fracture patterns, thought to be due to thermal stress, found on a rock at 4000 m altitude in the Argentinean Andes.

15°C and 30°C...when expansion is resumed". Some of the rocks tested by Johnson and Parsons (1944, p. 125) "have sufficiently different expansivities to warrant some question as to their satisfactory use under severe weather conditions" as building material. Hockman and Kessler (1950) undertook a comparable series of experiments and concluded "It is probable that the repeated thermal expansion and contraction of granite resulting from natural weather conditions, is one of the many factors contributing to the eventual disintegration of the stone" (p. 409).

The 1955 'Symposium on Thermal Fracture' contained a number of papers that discussed the role of thermal stress and thermal shock on the breakage off materials (e.g., Buessem and Bush, 1955; Kingery, 1955; Manson and Smith, 1955). As Kingery (1955, p. 3) pointed out, with a change in temperature "no stresses arise providing that the body is homogeneous, isotropic and unrestrained (free to expand)"; more or less the conditions used in the experiments of Griggs or Blackwelder. The reality is, how many rocks *are* homogeneous, isotropic *and / or* free to expand in all directions? When rock does not meet these ideal conditions then "stresses will arise due to difference in expansion between crystals..." and these may lead to "serious weakening or fracture" (p. 3). In large rock masses, temperature gradients can give rise to thermal stresses as a result of temperature not being a linear function of dimension. This non-linearity results in the expansion of each volume element and thus the separation of the elements (Kingery, 1955). Both tensile and compressive stresses occur at the same time: "On cooling, the maximum stress is the tensile stress on the surface and the centre is in compression. On heating, the maximum stress is the compressive stress on the surface, and the centre is in tension" (Kingery, 1955, p. 4). There is also a shear stress component equal to half the difference between the principal stresses in both the heating and cooling modes. Significantly for cold regions, during cooling the most potentially destructive stresses are tensile, but during warming phases either shear or tensile stresses are likely to be greatest (Kingery, 1955). The main problem is that to calculate the thermal stresses generated in a rock body it is necessary to know the temperature distribution, and these are the very data not sufficiently available in most field studies.

Studies of fracture patterns developed in ceramics as a result of thermal shock for different values of ΔT show great similarity to those found in rocks (see Fig. 7.1d of Selby, 1985 or Fig. 2.17 of Ollier, 1984 for example). Marovelli et al. (1966) and, more recently, Bahr et al. (1986) show diagrams of fracture patterns for ceramics and basalt as formed by thermal changes equating to values of $\Delta T = 72°F$ (22.2°C) through to 450°F (232.2°C). Marovelli et al. (1966, p. 1) point out that thermal stress has been used historically for rock removal and that in Scandinavia even mining was undertaken by means of 'fire-setting' rather than blasting. The fracture patterns produced in the basalts of Marovelli et al. (1966, p. 14) are very similar to the patterns, in

Fig. 2. Fracture patterns of a similar type to those in Fig. 1 and from the same area. Note that fractures are on all faces, do not follow the obvious lines of weakness (i.e., the bedding) and that there is a heirachy of fracture sizes.

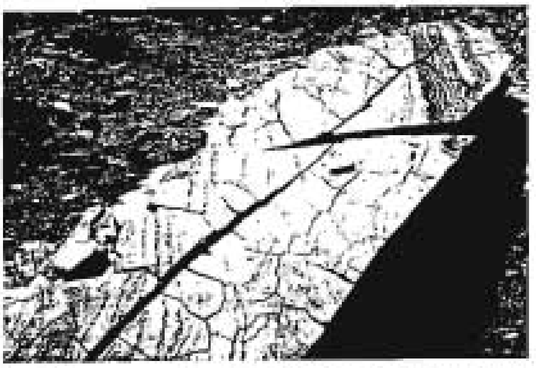

Fig. 3. Fractures on a large block from the same area as Fig. 1. Note that, like in Fig. 2, there is a heirachy of fractures and that they do not follow the obvious lines of weakness.

ceramics, generated by Bahr et al. (1986, p. 2717), and *both* look very similar to the fracture patterns seen on rocks in field situations (e.g., Ollier, 1984; Selby, 1985). The connection that seems *not* to have been made is that the patterns seen in the rocks from *hot* deserts can look very similar to those found in rocks of cold, dry regions that experience large thermal variations (Figs. 1–3). As Bahr et al. (1986) note, stresses are greatest at the rock surface and crack density increases with increased values of ΔT. There appears to be, at least in the experiments undertaken by Bahr et al. (1986), a hierarchal order of cracks which "stands out clearly against the random background"(p. 2719) something *very* noticeable in some crack patterns found in the field on rocks thought to have been broken by thermal shock (Fig. 3). Very importantly from a field perspective, the relationship between the thin ceramic discs used in the experiments of Bahr et al. (1986) to that of large rocks is that the value of ΔT necessary to induce cracking *decreases* as the rock body *increases* in volume. Thus, large rock bodies do not require such a large value of ΔT as used in the laboratory experiments on the thin discs to induce comparable cracking patterns.

3. What temperature data are required?

As obvious as it may be, the temperature attribute that is the least pertinent to rock weathering studies is that of the air. Too many studies have used, as a basis for their weathering determinations, air temperatures and, as a result, have made incorrect decisions regarding both processes and the timing of breakdown. In many instances it has given a very false picture of the freeze–thaw regime. Freeze–thaw cycles in the air have *no* impact upon rock where it is insulated by snow cover and yet many authors have used the strong thermal cycling shown by air temperatures during the spring as indicative of an active weathering environment. Conversely, despite continued sub-zero air temperatures many rocks attain temperatures in excess of +10°C as a result of warming by insolation and *do* experience freeze–thaw cycles that are not indicated by the air temperature. As Thorn (1992) pointed out, air temperatures are *not* a surrogate for rock temperatures.

Recognizing that it is *rock* temperature data that are required then the first component, and perhaps the 'easiest' to obtain, is that of rock surface temperatures. Preferably measurements should also be obtained at depth within the rock. Such data here presented were obtained by, in the laboratory, drilling blocks of local rock from their 'base' to varying depths below (what will become once placed back in the field) the 'top surface'. The rock was then positioned in the field such that unidirectional heat exchange occurred via that (undamaged) top surface. In the present studies, surface temperatures were obtained by drilling to within a millimetre of the rock top surface and then to depths of 5, 10, 15, 30 and 40 mm below the top surface. The data from different depths are valuable for calculating thermal gradients and for determining ΔT at different depths. It is also very important to obtain these data for different aspects and for different slope angles on each aspect. Aspect and slope greatly influence the thermal conditions such that great variability can occur at any given site or between sites (see below). If rock or building materials are not uniform (e.g., lithology, surface characteristics, constituent materials, etc.) then it is important that this is taken into account in the monitoring as the variability in albedo, thermal conductivity, etc. between materials can greatly influence the resulting data such that one material type may not reflect the conditions on another. Failing that, then so long as some form of 'representative' data could be obtained then tables of thermal conductivity (e.g., Clark, 1966) could be used to improve calculations of the thermal effects for depth based on the range of values as given by the tables.

4. What temperature data are available?

As noted above, McGreevy and Whalley (1982) give a fairly detailed account of temperature data from cold regions in the context of a discussion regarding freeze–thaw weathering. In their presentation, the main (thermal) argument centred on freeze amplitude and, although clearly germain to analysis of the actual freeze–thaw mechanism, little was said about the rate of change of temperature. In fact, McGreevy and Whalley (1982) state, "It is proposed here that freezing rate is not as important as has been suggested" (p. 158). In fairness to them they do go on to say that "The question as to whether or not freezing occurs rapidly or slowly is less decisive than whether or not water within a rock actually freezes" (p. 158). Certainly, within the context of freeze–thaw, this (the actual freezing of water) *is* the crucial question and one that is overlooked in most weathering studies. However, it can be argued, particularly in the light of more recent models (e.g., Hallet, 1983) which are so dependent upon the value of $\Delta T/t$, that the rate of change of temperature *is* a crucial component and that, as with monitoring of rock moisture content, it has not been sufficiently quantified in field studies. For example, as noted above, Ødegård and Sollid (1993), upon the basis of temperature data obtained at 1-h intervals, assumed the operation of Hallet's model which requires a $\Delta T/t$ in the order of 0.1°C h^{-1} (Walder and Hallet, 1985). They could *assume* this rate of $\Delta T/t$ from hourly data but this must hide within it (potentially, as shown below) enormous variability that would preclude such a generalised result. Thus, the values of $\Delta T/t$ *are* extremely important for freeze–thaw as well as thermal stress fatigue.

Coutard and Francou (1989) summarized available cold region thermal data for the period after the work of McGreevy and Whalley and present data of their own. Again, though, their data are at 2 hourly intervals which, although giving some reasonable indication of daily maxima and minima, are of no use in determination of $\Delta T/t$. With few exceptions outside the laboratory, this lack of detail regarding the rate of change of temperature seems to be the general outcome for most geomorphological and engineering studies (e.g., McGreevy, 1985; Matsuoka, 1991; Matsuoka et al., 1990; Shiraiwa, 1992) although Warke et al., (1996) did use 1-min record intervals in their laboratory simulation of stone breakdown in urban environments. Rather, it has been from *biological* studies that, until recently (see below), the most detailed have come (e.g., Kappen et al., 1981; McKay and Friedmann, 1985; Friedmann et al., 1987; Nienow, 1987). Many of these studies recorded rock temperature with data intervals of 1 min or less for up to 3 years as part of an investigation of cryptoendolithic microbial environments in Antarctica. Moisture, humidity and radiation were also measured at the same time interval. Although it was mainly rock surface temperatures and at depths to ca.4 cm that were measured, the data were also obtained for different aspects. All data were continuously transmitted via satellites thereby overcoming storage problems. Of great value, at least in the context of surficial weathering, was the modeling of thermal changes as a function of temperature, wind speed and internal light gradients. These models were created as temperature maxima were found to occur 1 cm *below* the rock surface while the most rapid fluctuations were taking place *at* the rock surface — factors that could have implications for several types of weathering. The models showed that temperature oscillations can be caused, in the top 4 mm of the rock, by wind fluctuations of short duration. Thus, thermal changes may occur in the rock when air temperature and radiation are constant and these fluctuations will not be indicated by longer-interval rock temperature records.

Available data show, while rock temperature may mirror changes in ambient temperature, rock (particularly internal) temperatures are considerably higher than those of the air (McKay and Friedmann, 1985). Rock temperature data acquired at 20 second intervals also show that during sunny days with moderate winds, when rock–air temperature differences are high, there can be *very* rapid temperature oscillations at the rock surface. These data clearly show thermal changes ≥ 2°C min^{-1} at the rock surface on at least (from Fig. 6 of McKay and Friedmann, 1985) 12 occasions during a 45-min record and several are ≥ 3°C min^{-1}; 11 changes

$\geq 2°C\ min^{-1}$ also occur at 4.3 mm depth. This recording of $\Delta T/t$ values $\geq 2°C\ min^{-1}$ is extremely important for, as Yatsu (1988, p. 131) clearly explains, "For heating rates $> 2°C/min$... new cracks and hence permanent strain are developed... the irreversible change is most likely due to the creation of cracks along grain boundaries." This value of $\geq 2°C\ min^{-1}$ may well be lower with large bodies (Bahr et al., 1986), and is certainly affected by the nature, size and orientation of component minerals, but it stands as a working threshold (Yatsu, 1988). Further, as shown by Thirumalai (1970), from studies on basalt, quartzite and granites, thermal spalling occurs when there is a high thermal gradient and that the spalling is very dependent upon the thermal expansion and shear-strain characteristics of the rock (Yatsu, 1988). Thus, by obtaining high frequency data is it able to be shown that values of $\Delta T/t$ are such that thermal shock could occur and that these rapid thermal changes could also cause a very high thermal gradient that can facilitate thermal spalling. Consequently, not only are there frequent thermal changes to facilitate thermal fatigue but also values commensurate with thermal shock — and this is only from consideration of one 45-min record!

These biological studies also show that there are not only large differences between the air and the rock temperatures, with rock temperatures being above freezing for 10–12 h day^{-1} when air temperatures are subzero, but also that the differences are greatly affected by aspect. The greatest air–rock difference (18°C) was on the northern exposure while the lowest (7°C) was on the southern exposure. These aspect differences can have large implications with respect to weathering both in terms of deducing the 'weathering environment' and/or the varying impact on different sides of a building or other structure. Such thermal differences may also facilitate stress differences (including buttressing) that were not taken in to account in the laboratory experiments. Thermal conductivity of the sandstone in this Antarctic region allowed penetration of the diurnal temperature wave to a depth of ca. 20 cm and thermal gradients of 4°C in 23.7 mm were recorded (Fig. 7 of McKay and Friedmann, 1985). All of these data indicate that the sort of thermal environment required by models such as that of Hallet (1983) or Tharp (1987) may not be as prevalent as the longer interval temperature records would appear to indicate. Rather, it is highly likely that thermal stresses may play a role far in excess of that which has been generally considered until now. This is all the more so as field considerations of the freeze–thaw mechanism indicate that the constraining factor is not thermal but rather moisture availability (Hall, 1995). Rejmánek (1971) also refers to information regarding rock temperatures and their importance in the context of the thermal behaviour of rocks and the implications of these for the distribution of plant species and the thermal balance of ecosystems.

5. Data from engineering studies

The breakdown of rocks and building materials due to thermal changes has been recognised by engineers for some time. As Marovelli et al., (1966) point out, "thermal energy has been used for primary rock removal" and that "thermal rock fragmentation was still in use 100 years ago in systematic underground mining operations in Scandinavia" (p. 1) the method only being abandoned there in 1885. In the 1940s, Union Carbide in the United States started developing thermal spallation for use in the mining of taconite ore and this process has produced more than 40 million feet of shallow holes for blasting! Williams (1986) refers to a standard piece of equipment, that has been in use for more than 30 years, for quarrying granite whereby a hand-held spallation burner is used to cut slots in granite. Much more recently, Williams (1986) suggested thermal spallation as a drilling process for hard rocks and showed that not only could it work but that it was very cost efficient. Although engineers have accepted failure due to thermal stresses and thermal-based mining techniques are still in use, little actual data on thermal conditions are available. The difference has been in the recognition of the role of thermal stress fatigue and the calculation of its role based on available empirical data. Those data have related more to the rock properties than the actual temperatures the rock were subjected to, these latter being assumed and a range of values used (e.g., Bartlett, 1832; Mitchell, 1953; Lecznar, 1962; Freeman et al., 1963; Wolters, 1969; Rodriguez-Rey and Montoto, 1973; Richter and Simmons, 1974;

Aires-Barros, 1975; Aires-Barros et al., 1975). Some interesting observations and data are, though, available. Branner (1896) investigated the difference in rock temperature between the sunny and shadow sides of a rock body and found differences of 27°C (50°F) and he anticipated values in excess of 55°C (100°F) *but* as he did not actually measure these considers that "We are obliged to confine ourselves, for the present, to the data" actually measured (p. 285). Branner (1896, p. 286) presented rock temperature data measured in Brazil by various authors and bases his calculations and deductions on these. Merrill (1906) makes numerous references to actual rock and/or building material temperatures. The key here is that these early workers seemed to all measure the actual material temperatures and to not use air temperatures as a surrogate. Although not exemplified by actual measurements, Merrill (1906) pertinently discusses the breakdown of rock in cold regions versus warm regions and, in the context of this present discussion, states "The sharp contrasts of temperatures on mountain peaks bring about excessive exfoliation and disintegration" (p. 265). Thus, the perception of thermal stress operating in cold regions is not new but rather has been ignored by most subsequent workers.

Schaffer (1932) discusses the role of temperature changes in effecting the breakdown of natural building stones in a British monograph of the Building Research Station. Again, though, despite referring to a number of studies explaining how and why failure occurs, no actual field data of temperatures are presented. Arni (1966) refers to, in the context of resistance of building materials to weathering, the effects of thermal stresses but actual temperature data are absent. In the context of cold regions and engineering, Corté (1969) provides a discussion document but refers only to ground (unlithified sediments) thermal conditions and only in the context of frost weathering, not thermal stress. Rautureau et al. (1991), in a well-named booklet (*Fragile like Stone*), explores the processes seen to be the main causes of the breakdown of stone in buildings and monuments in France; arguably the country with the greatest expertise in this field. In this they indicate the effects of aspect, with monitored temperature differences between north and south as high as 50°C. However, rather than seeing thermal conditions, particularly thermal gradients, as facilitators of stress fatigue, they suggest the main thermal attribute is in causing circulation of water in the stone. It is then the water that does the damage by mechanical and/or chemical (or in conjunction with biological) means. Detailed studies of building stone thermal conditions *have* been undertaken (e.g., Manté and Coutard, 1988) but the consideration of the observed temperature fluctuations have centred on the subzero component and its role in frost weathering. The attribute of observed day-time temperatures of up to +25°C were not considered in respect to effecting thermal stresses. Detailed fourier analysis of thermal cycling was undertaken but the role of ΔT in facilitating a thermal stress appears not to have been investigated.

Warke et al. (1996) investigated the thermal response of building stone to darkening of the surface by particulate deposition. Dark particles on the stone surface change the albedo and can thereby accentuate rates of surface temperature change. Although a laboratory undertaking using a variety of heating conditions, the main attribute is that temperature data were collected at 1-min intervals. This facilitated a very real appraisal of the thermal effects on the stone. The data showed that surface temperatures were raised as a result of soiling, that $\Delta T/t$ was increased when incident radiant energy was interrupted, and that this all resulted in increased thermal stresses. The overall argument is placed in the context of hot environments, for which such thermal changes are said to be potentially highly destructive if there are albedo changes to the rock or building material. I believe it could rather be suggested that, in cold environments with high radiative inputs the effects could be even more pronounced as the difference in temperature between the (radiation) heated rock and the cold air is much larger than in hot deserts. Thus, upon removal of the heating source, the value of $\Delta T/t$ will be much greater as Newton's Law of Cooling states that the rate of change of temperature is dependent on the difference in temperature between the two bodies and will be greatest where this difference is largest. So, in a cold environment if stone is darkened by pollutants or biological activity (e.g., lichens) such that an albedo change facilitates both increased rock temperatures and higher values of ΔT then the role of thermal stress and/or thermal shock may be increased.

6. Recent data from geomorphological field studies

Three recent studies (Hall, 1997a,b and In Prep.) two from the Antarctic and one from the Canadian Rockies, utilized high frequency temperature data acquisition and, in so doing, gave new insights into cold region weathering. The first Antarctic study, detailed in Hall (1997a), collected data at 2-min intervals and showed the occurrence of $\Delta T/t$ events in the range of 2°C min^{-1}, and thus the potential for thermal shock and/or fatigue as a cause of rock breakdown in this cold, but arid, environment. Aspect was also shown to exert a strong influence on the weathering regime. This study was complemented by an undertaking at a different site where data were then collected at 1-min intervals. In the Canadian study (Hall, 1997b), data were collected at 30-s intervals. In each undertaking, data were obtained from different aspects and for different depths. In summary, the data indicate a number of generalisations that apply to cold region weathering of rocks and building material. First, air temperature is not a surrogate for rock temperatures. Second, there is a marked aspect effect on the nature of the thermal environment experienced by the rock. Third, that it may well be erroneous to presume freeze–thaw weathering just because it is a 'cold region'. Fourth, for any meaningful judgement of process there is a need for temperature recording at very frequent intervals.

Temperature data collected at 1-min intervals for a (roughly) 24-h period at a Southern hemisphere location (Rothera Station: 67°34′S, 68°07′W) on the western side of the Antarctic Peninsula are illustrated in Fig. 4. These data clearly illustrate the importance of aspect in cold regions. For example, although the western aspect is the coldest during the 'night' pe-

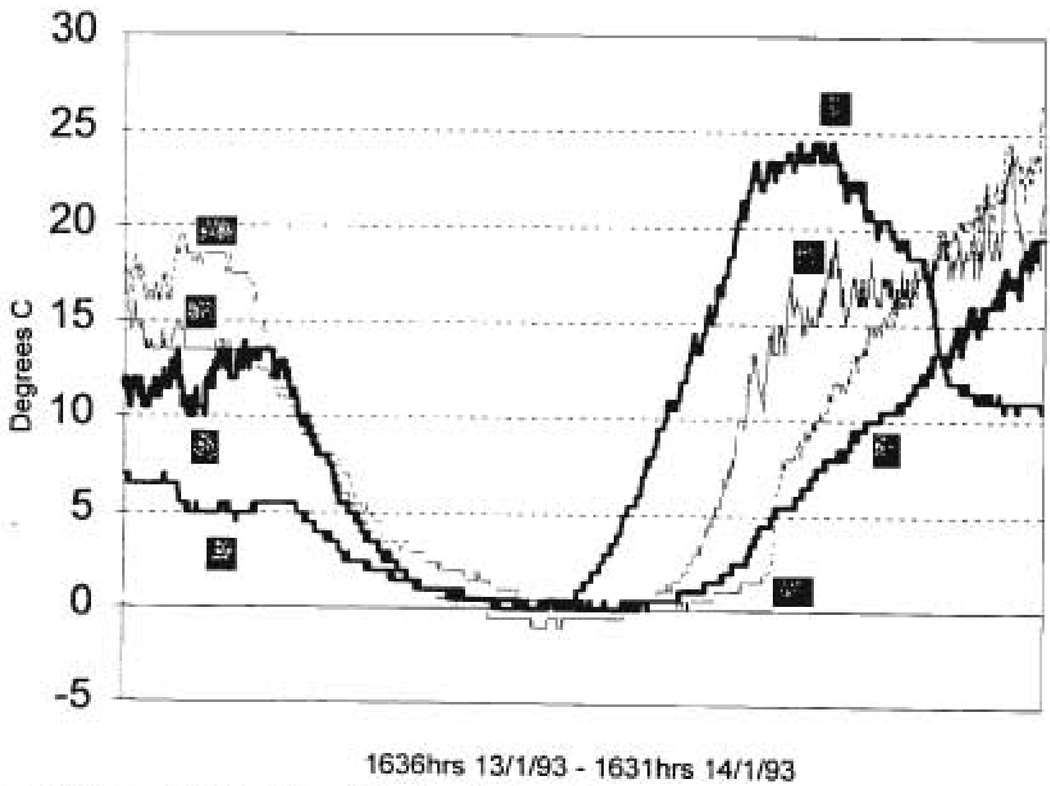

Fig. 4. Rock temperature data (at 1-min intervals) from Antarctica for the four cardinal directions over a 24-h period in summer to show the distinct influence of aspect.

riod, the west has the highest temperature during early evening, achieving (almost) a temperature of 20°C at a time when the eastern aspect is at 5°C; colder even than the southern aspect (10°C–13°C). The corollary occurs the next morning, when the eastern aspect is at almost 25°C while the west is only between 1°C and 2°C. The northern aspect, in this record, never achieves the highest temperature, in fact the western aspect has the highest value (at the record end: 26°C). Equally, the southern aspect does not have the lowest temperature in this record, rather it is the northern aspect (at ca. −1°C, while south is 0°C to −0.2°C)! The reason for the higher temperature on the southern aspect is thought to be, remembering these are rock temperatures, that this aspect still receives some low angle radiation in the summer at midnight which warms the rock while the northern aspect experiences the cold air with no radiative input. Conversely, when the northern aspect is receiving radiation during the morning, the air temperature is high and this raises the temperature of the southern aspect. Thus, the presumption of (in this hemisphere) the southern aspect being colder than the northern is simply not substantiated by the data: at least for part of the year. This temperature record also indicates that blocks of rock may be subject to thermal stresses as a function of the large thermal differences between aspects. Recognizing the aridity of many polar and high altitude locations (e.g., the Andes) and yet weathered blocks of rock are still common, the propensity for frost action is extremely limited. The thermal conditions, as illustrated here, offer a far more likely cause for rock breakdown.

Detail of part of a daily temperature record for the north-facing aspect (for the 16th January, 1993: 1523 to 1603 h) from this site at Rothera is shown in Fig. 5. During the 40-min record, there is an event with $\Delta T/t$ of 2°C min^{-1} and one event where a value of

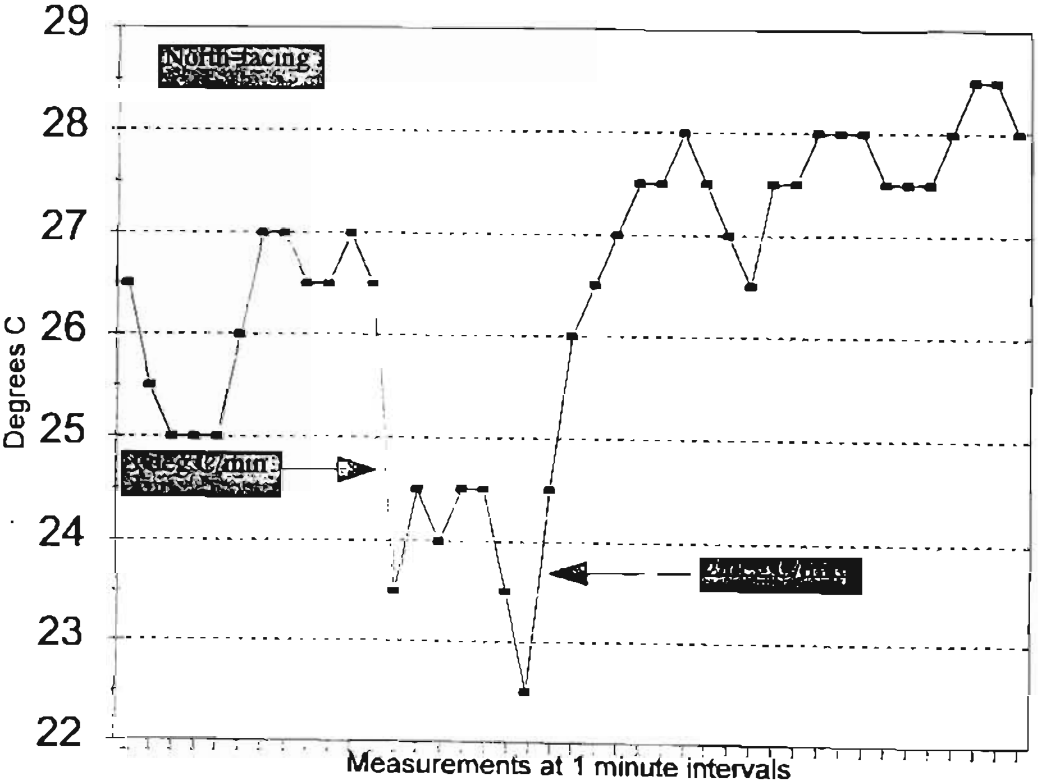

Fig. 5. Detail of the Antarctic data for a 40-min period with observations every minute on the north-facing exposure. Note the two events indicated where $\Delta T/t$ was ≥ 2°C min^{-1}

3°C min^{-1} is attained. Both these values achieve the threshold for possible thermal shock, and both would certainly facilitate thermal stress. Neither event could have been identified if observations had been taken at the more common record intervals of $\geq$ 10; 30 min or hourly observations would have shown no indication that such events could even take place. These are *not* isolated events but can be found for every record day (also see below for similar occurrences at a Northern hemisphere site).

An example of the 30-s record interval data for a single day from the Northern hemisphere Canadian Rockies (Fig. 6) shows the marked differences between aspects and indicates the likelihood of high $\Delta T/t$ values. The record, that runs from 0729 h on 26 July, 1996 through to 1039 h on 27 July, was taken in mid-summer from a rock block located above the tree line at an altitude of ca. 1750 m (see Hall and Meiklejohn, 1997 for a description of the area). Data were collected from the rock surface and at ca. 2 cm depth on each of the cardinal aspects. The eastern and southern faces have the highest temperatures (37.4°C on both) while the north (30.1°C) and west (27.8°C) are lower. The lowest temperature for that record period (8.3°C) is found on the northern aspect, and this is followed by west (9.3°C), south (10.6°C) and east (11.8°C), respectively. A good example of the loss of information that parallels the increase in record time interval is shown in Hall (1997a, Fig. 8). The most obvious loss with time interval increase is that of the shorter term variations — those most important to thermal shock

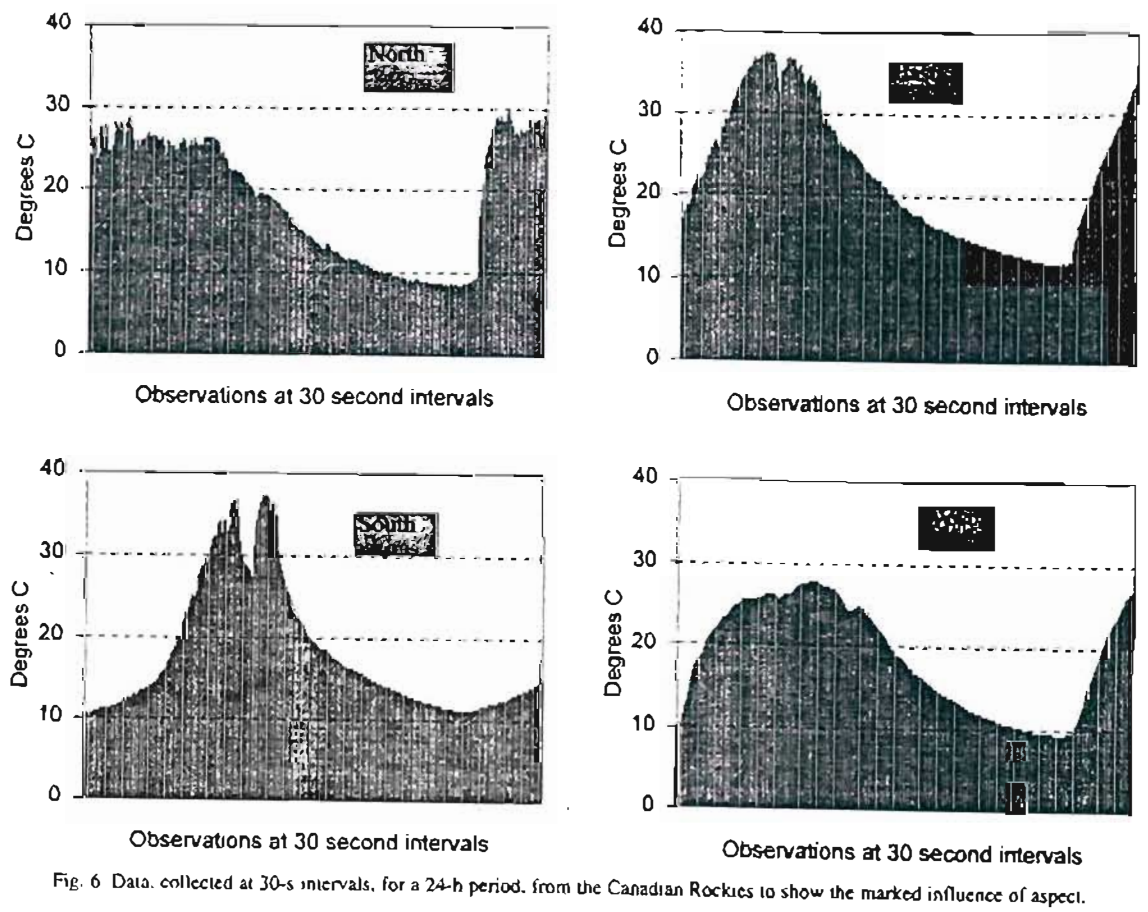

Fig. 6. Data, collected at 30-s intervals, for a 24-h period, from the Canadian Rockies to show the marked influence of aspect.

and/or stress. Even the maximum recorded temperature drops by 4°C from a record interval of 2 min (20°C) to that of 1 h (16°C); even at a 10-min interval the maximum value drops to 18.5°C (Hall, 1997a, Table 7). Thus, our knowledge of rock thermal conditions must be highly suspect unless high frequency record intervals are used. Examples of $\Delta T/t \geq 2°C\ min^{-1}$ were found at this site (Hall, 1997a, Fig. 11) with such events during both warming and cooling phases.

The east-facing record from the Rockies (Fig. 7) exemplifies the thermal difference between the surface and 2 cm depth. Maximum temperatures at 2 cm depth are some 12°C lower than those recorded for the surface. Considering the relatively short distance, such a thermal gradient could be expected (e.g., Kingery, 1955) to be creating tensional and compressive stresses within the rock. During the cooling phase, there is a cross-over such that the 2 cm depth temperature becomes higher than the rock surface. Although both the surface and the interior are cooling, the surface cooling is much faster than at depth and so this, too, would create a stress field. In other instances, data clearly show that a warming pulse continues to penetrate the rock for a short period while the surface is cooling there by creating two differing stress fields. Clearly, more data regarding temperature at depth is required before any detailed calculations to ascertain the stresses and their location can be attempted. In the interim, the data *do* suggest that failure-inducing stresses can occur parallel to the rock surface at depth (probably within the upper few centimetres).

The detailed record, fully utilizing the 30-s record interval, provides important information that would not be available from the time frames used in most field-based data recording. A 25-min record (1613 to 1638 h on 26 July) from the south-facing exposure

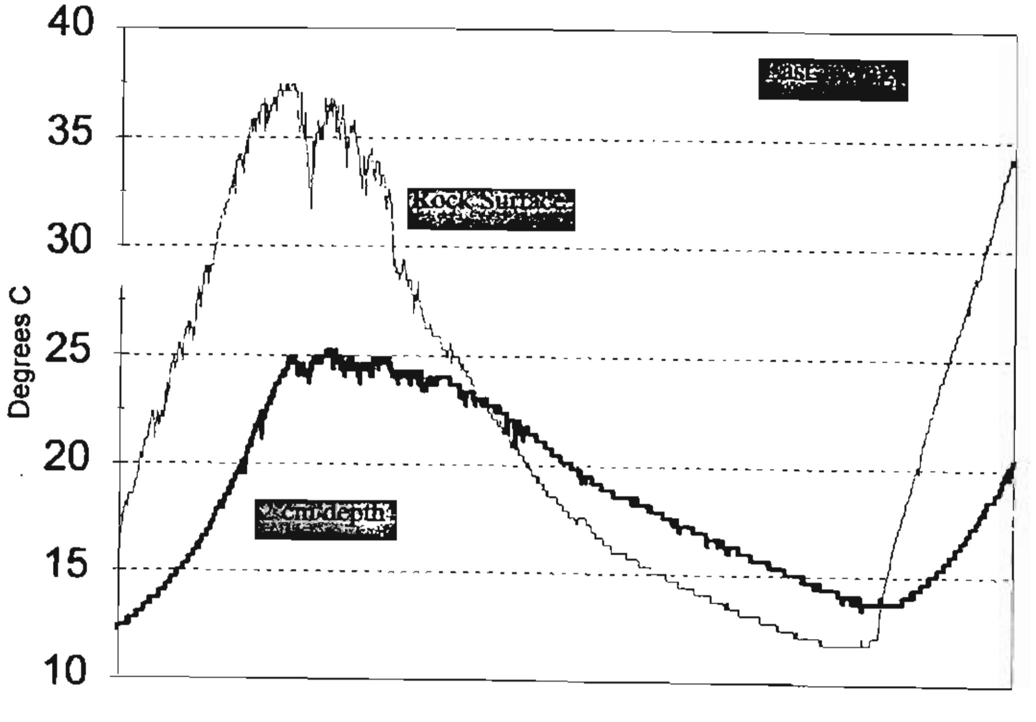

Fig. 7. Detail from the east-facing record of Fig. 6 to show the thermal difference between the rock surface and at 2 cm depth.

(Fig. 8) shows the occurrence of three events with a $\Delta T/t \geq 2°C\ min^{-1}$. It would not have been possible to discern such events from temperature records of even 5-min intervals, let alone 30 min or 1 h — within which records all these events would have been 'hidden'. Thus, in terms of the weathering regime, the potential for thermal stress fatigue and even thermal shock must now be considered. Another detailed record for a 25-min period (0908 to 0933 h on 26 July) for the north-facing aspect also shows three events with $\Delta T/t \geq 2°C\ min^{-1}$ (Fig. 9). One occurs on a warming limb, one on a cooling limb, and one is a combination of cooling and then heating. Thus, thermal stresses related to heating, cooling, and a combination of both are found within this record period; the stresses related to a rapid cooling followed by a rapid warming being particularly large. Further, this record (Fig. 9) also shows two occasions when the value of $\Delta T/t$ was $\geq 1°C$ over 30 s but did not, in conjunction with preceding or subsequent 30-s records meet $2°C\ min^{-1}$. It does, though, beg the question as to the cumulative effect of such short term, large $\Delta T/t$ events? Multiple such events must add to rock fatigue.

The overwhelming conclusion that comes from the above (extremely) brief outline of the sort of data that are becoming available, is that thermal stress and/or thermal shock is a weathering process in cold regions that has not been sufficiently considered. The nature of the existent temperature data is such that, with few exceptions, it is impossible to adequately analyze the potential for thermal stresses and thus their role has yet to be fully quantified. Equally, the perception of freeze–thaw as the "explanation" for cold region weathering, particularly in arid or semi-arid regions, is both qualitative and subjective. Thus the assumption of angular clasts being due to frost weathering and accordingly frost

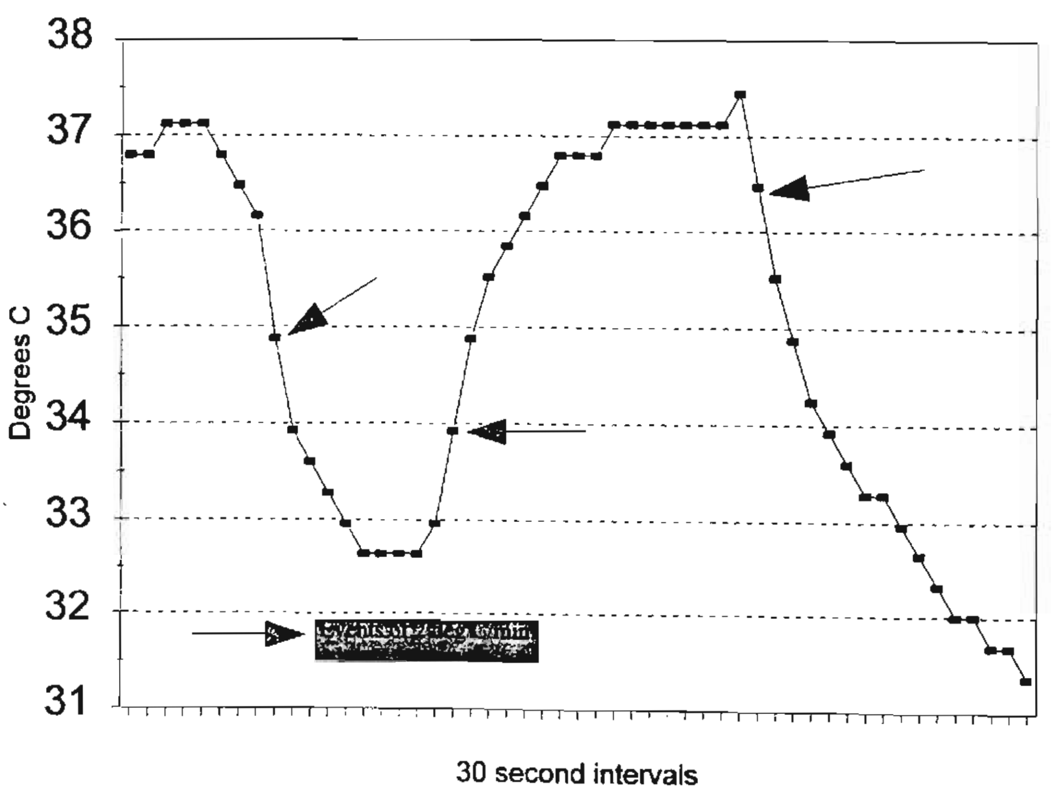

Fig. 8. A 25-min detail from the south-facing record of Fig. 6 indicating three events with $\Delta T/t$ was $\geq 2°C\ min^{-1}$.

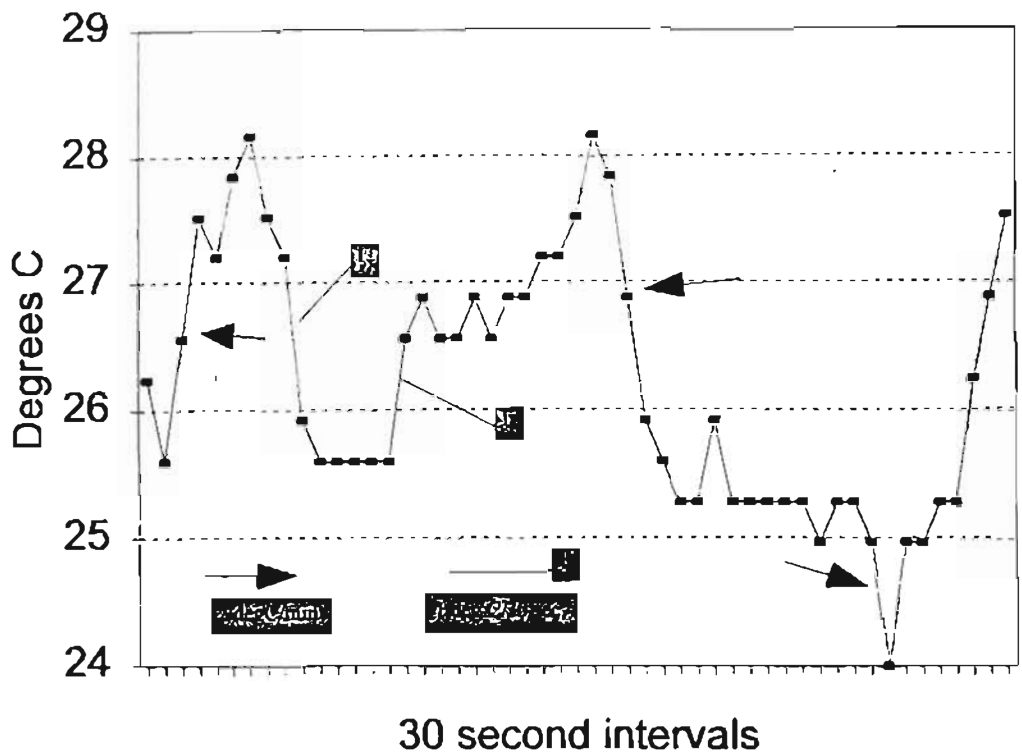

Fig. 9. A 25-min detail from the north-facing record of Fig. 6 to show events with $\Delta T/t$ was $\geq 2°C\ min^{-1}$ and where $\Delta T/t$ was $\geq 2°C\ min^{-1}\ 30\ s^{-1}$.

weathering being "proved" as operative is both circular and false. The studies that use thermal data obtained at time scales in excess of 5-min intervals (upon which many "frost weathering" judgements are based) have completely missed the possible role of thermal stress and thus the determinations of the weathering regime may be (substantially) in error. This is important not only for the understanding of present conditions but also for the use of the present as an analogue for the past, particularly for the interpretation of Quaternary sediments. This need to move from our obsession with freeze–thaw as the agent of destruction in cold regions is very much in accord with the views recently expressed by McCarroll (1997, p. 1) where he suggests: "We do not improve our theories or models by admiring them, or by proclaiming how well they seem to fit our observations. The only way to improve them, and therefore to make progress, is actively to seek conflict between our models and the real world." We need data to see if our conceptual models of weathering in cold environments do really function and to achieve that we need better thermal data (as well as moisture, but that is argued elsewhere: Hall, 1993; Hall and Hall, 1996).

7. Conclusion: implications for engineering

It would seem that the considerations outlined above regarding geomorphological interpretation of landform or sediment origin have application for engineering in cold regions. Unless cognisance of the potential role of thermal stresses is taken into account a major contributor to deterioration and/or a synergistic component of the final breakdown is missed. It is possible that in some cold environments, particularly the more arid ones, the role of thermal stress may predominate. Another consideration with respect to anthropogenic structures is that even where thermal stress does not operate at a particularly fast rate, its effects may be exponential. That is to say, the weathering effects may be initially minimal but once they start to manifest themselves, the material is already badly damaged and breakdown will then be rapid. Concurrent with this is the synergistic role. A rock or building material that is subjected to slaking or freeze–thaw tests for its durability may not respond according to the test data as these have not taken into account the role of thermal stresses. Even where the outcome of those stresses is not failure they may change such as the porosity and/or permeability to the extent that the material will then respond in a quite different way to that suggested by the test results. That material which was deemed non frost susceptible may have its porosity changed by thermal stresses such that it becomes highly susceptible. Equally, the changes to the rock structure may facilitate the operation of biological weathering that were initially precluded such that this, in conjunction with all the other weathering processes, may exacerbate the rate of breakdown. While much of this is unlikely to cause, in the time scales of buildings, roads or bridges, significant structural damage, it may, nevertheless, cause additional costs, monetary or aesthetic, in terms of surficial or unsightly damage that may result, particularly if any fine detail work is used as surface dressing.

In terms of assessment of rock or material durability, it is necessary to obtain a quantitative assessment of the environment where construction is to take place. Based upon this information (e.g., rock temperatures and rock moisture content together with their temporal and spatial variability) to undertake weathering durability tests that include, where appropriate, that of thermal stress. Finally, to assess the durability of materials, such tests should not be 'discrete', i.e., a slaking test on one sample, a freeze test on another, etc., but rather a number of samples should be subject to the suite of processes they may be exposed to such that the synergistic component of the different weathering processes is taken into account. With any form of global warming which may extend the summer period, and hence the period of radiative heating, in cold regions, so cognisance of the role of thermal fatigue may become all the more important.

Acknowledgements

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Wetting and Drying Weathering

Rocks and building materials in many locations experience, at least for part of the year, periodic wetting and drying (humidfication- séchage) as a result of rain, snow melt, wave splash, tidal events, high river discharge, splashing of road water on to buildings and pavements by cars, wind-driven waves on lakes or dams, etc, etc. Many of these wetting events are short term and are followed by drying, or at least partial drying, only to be followed by a further wetting. As most chemical weathering processes are water-driven and a number of mechanical processes also require (e.g. frost action), or are enhanced by (e.g. hydration of, or precipitation out of, salts), the presence of moisture, so these wetting events facilitate the operation of various weathering processes. Wetting, and drying, is obviously highly pertinent to chemical weathering. Water taken in to some minerals provides a "...bridge or entryway for hydronium (hydrated hydrogen) ions to attack the structure" (Buol, *et al.,*1989, p. 101); commonly referred to as 'hydration'. Further, the association of that water or hydroxyls with aluminium or silica at broken edges can be the first step in hydrolysis (Buol, *et al.,* 1989). Thus, clearly there is a strong relationship to chemical weathering processes but it is the physical (mechanical) weathering attribute that is discussed here for, the very action of 'wetting and drying' of the rock (or building material such as brick or concrete) does, itself, cause weathering. Further, but likely interconnected, some rock properties (particularly strength) change dramatically as a result of wetting such that this too can influence weathering. This process is often referred to as "slaking" but care should be taken with this term as it has a strong association with processes requiring the presence of clay minerals (as shall be outlined below) but, in fact, *all* rocks can be affected by wetting and drying which may be an active, but underestimated, agent in their breakdown.

Slaking, an integral part of wetting and drying where clay minerals (or minerals such as mica) are present, involves molecular pressure from 'ordered water'. Water, a 'polar liquid', has positively charged hydrogen atoms at one end of the molecule and a negatively charged oxygen atom at the other end. Upon wetting of rock electrostatic bonding takes place such that the positively charged end of the water molecules are attracted to the negative charge of the crack or pore wall and subsequent water molecules, like small bar magnets, will be attached to the exposed negative charge of the first 'layer' of water. In this manner an adsorbed layer of water is built up and, as this grows, so a swelling pressure can grow within the crack or pore, thereby creating strain, that can, ultimately, lead to failure. In particularly small pores (<40Å) the like poles of adsorbed water on opposing sides all meet in the middle where 'like poles repel' - to create tensile stresses that could lead to failure, particularly through multiple replications inducing fatigue. This attribute can be particularly important in freeze-thaw weathering where sub-zero temperatures enhance the electrical charge of the water thereby increasing the tensile forces.

Wetting and drying has a range of other effects. Not the least of these is that a wetted rock increases in volume, and this may result in failure due to fatigue after multiple cycles. In some instances expansion due to wetting may not be reversible upon drying and in other instances hysteresis effects occur such that, although shrinkage upon drying takes place, the rock does not return to the original dimensions. Data regarding the elongation that a variety of materials experience upon wetting has been available for some time (e.g. Nishioka and Harada, 1958; Nepper-Christensen, 1965), beyond that associated solely with clay-rich materials (e.g. Venter, 1981a & b). Rocks such as sandstone, talc schist, limestone, chert, pumice obsidian, basalt, granite and flint were all shown to increase in volume as a result of wetting. Nishioka and Harada (1958, Table 2) were even able to show the relationship between the rate of elongation and the rate water absorption for a medium-grained, micaceous sandstone and an andesite from dry through saturation and back to a dry state (Fig. 1). A number of more recent studies (e.g. Felix, 1983, 1984, 1987; Pissart and Lautridou, 1984; Hames, *et al.*, 1987) have clearly shown that elongation takes place as water is absorbed by a range of rocks and that this can result in permanent elongation and breakdown. Felix (1983, p. 310) makes the valid point that there is a distinction between short- and long- term responses; the former only demonstrate the degree of swelling that may occur whilst the latter shows what rocks will be more prone to weathering as a result of the wetting and drying. He suggests that the swelling is as a result of "capillary pressures" (p. 310) set up in the pore system and cracks and thus the degree of swelling is greatly influenced by pore size. Felix (1984, p.1) notes that the linear swelling of the sandstones he studied varied from 0.1 to 4.0 mm/m, dependant on the petrographic facies, but that the water-sensitive facies were *not* those absorbing the greatest amount of water. Hames, *et al.* (1987) showed that rock samples increased in length from 1 to 18 μm/cm and that, in some instances, the rock remained elongated. Again, pore size was found to play a role as it was the adsorbed water in the finest pores that provides the destructive force. Pissart and Lautridou (1984) also found an extension of 5 μm/cm for a Bathonian limestone and that (p. 111) the changes in "...length are proportionally larger when the amount of water in the sample is low". They also found that, as the sample increased progressively in length after multiple wet-dry cycles, this was a form of mechanical weathering.

Wetting and drying has other effects on rocks that are likely linked with the overall weathering impact. For example, wetting of rock is known to decrease the strength of that rock (e.g. Dube and Singh, 1972; Broch, 1974). Dube and Singh (1972) found a decrease in tensile strength of sandstone from 11 to 48 per cent under fully saturated conditions such that, overall, tensile strength decreased with an increase in humidity and porosity. In much the same way, Broch (1979) showed that rocks experienced a decrease in tensile strength with increasing degree of saturation and that anisotropic rocks were especially responsive to changes in water content below 25 per cent. Broch (1979) suggested that the decrease in strength was as a result of the water causing both a reduction in surface energy and a reduction in internal friction. In a similar type of study, but on radically different materials, Venkatswamy, *et al.* (1987) found that wet-dry cycles caused a reduction in the strength of the midribs of coconut leaves. Weathering as a result of the wetting and drying caused the ultimate tensile strength (UTS) to decrease by about 56% whilst the breaking load decreased from 747 to 620 N after 25 cycles. Thus there are two factors. The rock is weaker when wetted and it decreases in strength as a result of the wetting and drying cycles. This led Hall (1988a), based on a number of laboratory experiments, to suggest that the bonding strength between minerals decreased during water uptake to the extent that the rock was more prone to other water-based weathering processes as the tensile strength they need to overcome is less than anticipated by the dry-strength data. Further, the experiments, using ρ-wave ultrasonic data, showed that hysteresis effects produced a progressive strength reduction that could, ultimately, lead to rock failure. Hall and Hall (1996) also showed that manner of wetting and drying (at least of a sandstone and a dolerite) could affect the character of the weathering. In their experiments the rock samples were submerged, half submerged or sprayed with water: samples subjected to the latter showed granular disintegration whilst the first two approaches produced "splinters" of rock. The results also showed that there was a progressive change in the pore size distribution with pores increasing in size with wet-dry events plus an increase in the number of pores within the rock. Both of these findings indicate that the synergistic role of wet-dry weathering with other weathering processes will change with time.

Whilst wetting and drying is a weathering process in its own right, and may well be more effective than has been frequently considered in areas where numerous wet-dry events occur (see Hall, 1988b, 1991,1993), it also plays an important synergistic role with other water-based weathering processes. In all environments where water-based chemical weathering processes affect the rock there is likely an interaction with mechanical breakdown due to wetting and drying as the water condition within the rock will rarely be constant but rather will fluctuate with daily, seasonal or annual climatic conditions. The impact upon pore size and number may help increase permeability and hence chemical processes whilst, at the same time, causing rock breakdown and loss in its own right. In the same manner and due to the same responses, weathering by wetting and drying will operate synergistically with several mechanical weathering processes. Before identifying some works that have shown this interaction, it is worth noting that the effects of wetting and drying may often be masked by the operation of the other (often presumed) processes. Where a rock is wetted and then experiences dry*ing* but not total dryness, the zone of wetting and drying is then sub-surface (Hall, 1991) such that the changes in pore size plus the effective breakdown in a zone *below* the surface may be masked or not identified as a result of the presumption or identification of a more apparent process (e.g. freeze-thaw weathering).

Mugridge and Young (1983) detailed the enhanced breakdown of shale as a result of the interaction between wetting and drying and frost action along a river in Manitoba (Canada) They suggest that breakdown by wetting-freezing-thawing-drying is more rapid than breakdown by wetting and drying alone, although alone it did still cause destruction. In truth, consideration of the process shows that freeze-thaw can, in fact, *never* operate without some aspect of wetting and drying being involved for the rock must be wetted prior to freezing and thawing will involve some water loss. This is an aspect *not* taken in to account in laboratory experiments of freeze-thaw (as detailed and investigated in Hall and Hall, 1996). Moss, *et al.*, (1981) also discuss this inter-operation, in a river environment, of wet-dry cycles with frost action whilst Fahey and Dagesse (1984) identify this connectivity as possibly operative in cold climates. More recently, Prick, *et al.* (1993) discuss the variation in length of a rock cylinder as a result of frost action. They show the intricate and non-isotropic variations in rock dimensions resulting from migration of non-frozen water - water content and porosity being important. Thus, directly and indirectly wetting and drying has an impact on freeze-thaw weathering.

Wetting and drying is thus a complex processes that has many attributes, the totality of which can lead, via a number of mechanisms (not well understood), to the breakdown of rock and/or building materials. It is a process that certainly works synergistically with other weathering processes but can, in its own right, lead to rock breakdown. The role of this process in rock breakdown in many climates and environments is likely underestimated.

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Caption:

Fig. 1. Graph to show the elongation and contraction (with hysteresis) of an andesite based upon the data given in Nishioka and Harada (1958, Table 2).

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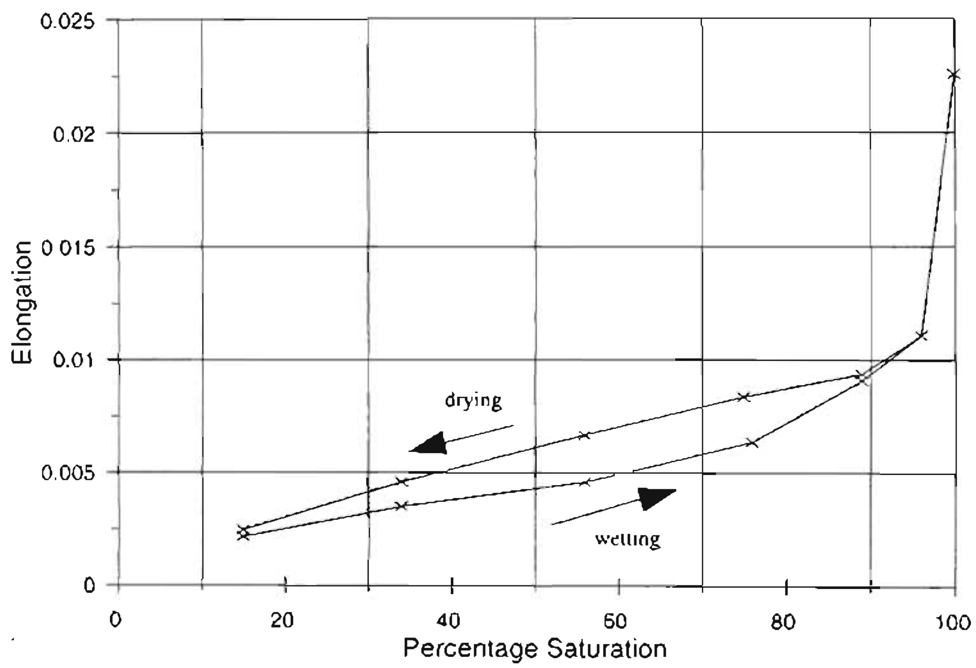

0.025
0.02
0 015
0.01
0.005
0
Elongation
drying
welling
0
20
40
60
80
100
Percentage Saturation

Thermal Stress Fatigue ('Insolation Weathering')

Geomorphological texts generally term the breakdown of rock due to (solely) thermal changes as 'insolation weathering' (e.g. Ollier, 1984; Bland and Rolls, 1998). Selby (1985, p.191) provides a typical definition of insolation weathering: breakdown "..is the result of alternate warming and cooling of rock surfaces under the direct influence of solar heating". Although the driving force behind the thermal changes is, in many cases, insolation it is evident that it not the 'insolation' that weathers but rather the thermal changes brought about as a *result* of variations in radiative input. Thus, despite its common usage, it is suggested that the term 'insolation weathering' is misleading and that the terms 'thermal stress fatigue' and/or 'thermal shock', as used by physicists, ceramics scientists and engineers, are semantically and scientifically more correct. The use of the term 'insolation weathering' may have suggested to many researchers that, in other than hot regions, the sun (and associated heat) is contraindicated and so the likelihood of this process being operative was not sufficiently considered.

Thermal stress fatigue is where rock (or other material) is subjected to a series of thermally-induced stress events , each less than that required to cause immediate failure, that, collectively, cause the material to fail (Hall and Hall, 1991; Hall, 1999). Thermal shock is where the stress event is of sufficient magnitude that the material is incapable of adjusting fast enough to accommodate the required deformation and so it fails. Simplistically, thermal stress fatigue can work at two levels: that of grain to grain and/or that resulting from thermal gradients. Stresses can occur between adjacent grains as a result of thermal differences caused by such factors as differing albedos, grain sizes, grain roughness, crystal orientation, thermal conductivity, etc. In other words, any combination of attributes that will cause differences in thermal expansion and/or contraction between two adjacent grains may lead to stresses that ultimately cause failure ("granular disintegration"). As the rock surface is that which is first warmed or cooled so there will be a temperature gradient from the surface in to the rock (see Hall, 1998, Fig. 5 for examples). Where that gradient is of sufficient magnitude so spalling will occur. Essentially, a spall will release the maximum strain energy, generated by the thermal gradient, where (Rossmanith, 1983),for an infinite slab, the strain energy in a unit volume is given by:

$$U = \frac{\sigma^2(1-\mu)}{E}$$

and the depth of spall should be proportional to:

$$E/(1-\mu)\sigma^2$$

where U is the strain energy in a unit volume, σ is normal stress, μ is Poisson's ratio, and E is Young's modulus of elasticity.

As the residual temperature gradient in the spalled piece will be nearly linear, essentially all of its strain energy will be removed upon spalling (see Thirumalai, 1970 for details regarding studies on basalt, quartzite and granites). Thus, 'exfoliation' can also occur as a result of stresses, tensile, compressive and shear, induced by a thermal gradient (see also Fig.3 of Hall and Hall, 1991). When the rate of change of temperature (ΔT/t) exceeds the threshold of a materials ability to deform sufficiently fast to accommodate the required dimensional changes, so that material must then fail: thermal shock. Studies (e.g. Yatsu, 1988; Bahr, *et al.*, 1986) have shown that the threshold value of ΔT/t is $\geq 2°C\ min^{-1}$.

Historically, many early workers recognised that thermal stresses were the cause of failure in rock and building materials and, in some instances, this knowledge was used for economic exploitation of mineral resources. In Scandinavia mining was undertaken by means of 'fire-setting' rather than blasting until 1885 (Marovelli, *et al.,* 1966). The Union Carbide Company in the United States developed thermal spallation for use in the mining of taconite ore in the 1940's and has produced more than 40 million feet (12,192,000 m) of shallow holes by this method. Williams (1986) identifies a standard piece of equipment now used for more than 30 years for quarrying granite by means of a hand-held spallation burner and provides information regarding a drilling process for hard rocks based on thermal spallation technology. Others, such as Bartlett (1832), Branner (1896) and Merrill (1906) recognised the efficacy of thermal changes in breaking down rocks: "The unequal contraction and expansion of the minerals composing the rock tend to disintegrate the entire mass, while the even annual and diurnal changes....cause the rock to spall off in layers of even thickness" (Branner, 1896, p. 281). Merrill (1906) claculated the rates of expansion for a number of rock types and, based on available temperatures, showed that a sheet of granite 30.48 m (100 ft) long would expand laterally by 2.54 cm (1 inch) and that the effect of this would be to (p. 160) "lessen the cohesion and tear the upper from the deeper layers".

Branner (1896) observed that expansion and contraction rates differed from mineral to mineral and that, significantly, these vary as a function of crystal orientation. Johnson and Parsons (1944) found that crystal orientation had a great impact on expansivity. For example the rate of change in length per degree C ($\Delta L/°C$) for calcite was positive parallel to the c-axis but normal to the c-axis is actually negative. Conversely, for quartz $\Delta L/°C$ is greater normal to the c-axis than parallel to that axis. Thus, where a rock has a variety of crystals there can be substantial crystal-to-crystal stresses set up as a result of the varying $\Delta L/°C$ components associated with the axis orientations. The experiments of Johnson and Parsons (1944) also showed that materials at sub-zero temperatures (p. 121) "...expand on heating until temperatures between 0°C and 10°C....with further heating they either expand at a lesser rate..or contract until temperatures between 15°C and 30°C...when expansion is resumed". Thus, the reality is that thermal changes induce very complex reactions on a grain-to-grain level such that granular disintegration should not be too unexpected. These findings were reiterated by Hockman and Kessler (1950) such that they considered thermal expansion and contraction to be a major cause of stone disintegration.

With respect to spalling, large rock masses will inevitably experience temperature gradients that can produce thermal stresses as a result of temperature not being a linear function of dimension. Such non-linearity produces the expansion of each volume element and this can cause element separation (Kingery, 1955). This breakdown occurs because both tensile and compressive stresses occur at the same time. As Kingery (1955, p. 4) points out, "On cooling, the maximum stress is the tensile stress on the surface and the centre is in compression. On heating, the maximum stress is the compressive stress on the surface, and the centre is in tension". Data such as that shown in Hall and Hall (1991) and Hall, (1998) indicate that a rock can experience (for example) surface cooling whilst the interior is still subject to a warming pulse; thus, multiple stress fields can occur approximately parallel to the rock surface. When the temperatures, at any level, exceed the $\Delta T/t$ threshold of 2°C min^{-1} so the further factor of thermal shock is added. Substantiation for the effect of thermal shock can be seen in the shock patterns produced in the laboratory (e.g. Marovelli, *et al.,* 1966, p. 14; Bahr, *et al.,* 1986, p. 2717) that are also found in rocks in the field (e.g. see photographs in Selby, 1985; Ollier, 1984; Hall and Hall, 1991). From a practical viewpoint, in many hot and cold dry environments it is impossible to find an explanation other than thermal shock for the observed fracture patterns and when those patterns closely approximate to that produced experimentally in the laboratory by thermal shock then the case is made stronger.

Considering the above, it is surprising to see the results of studies by such as Blackwelder (1933) or Griggs (1936a) repeated, without questioning or testing, time and time again such that geomorphologists, with few exceptions until recently, have tended to discount the efficacy of thermal stress fatigue. Typical of the non-acceptance of the thermal stress fatigue concept would be "Rock disintegration due to daily temperature fluctuations therefore may be questioned as a mechanism of physical weathering" (Renton, 1994, p. 161). Substantial discussion is frequently given (e.g. Easterbrook, 1999, p. 20) to the findings of both the work of Griggs (1936a & b) and Blackwelder (1927,1933) whilst ignoring the extensive material available from engineering studies. Preceding the work of either Griggs or Blackwelder, engineers and geomorphologists were in reasonable agreement regarding the role of thermal stresses but subsequent to their studies the two groups diverged (e.g. Reiche, 1950, p.10); but it is noticeable that in *none* of the engineering/ceramics studies was there any reference to either of these geomorphological undertakings. Whilst the majority of geomorphologists moved away from the concept of thermal stress fatigue, a few (e.g. Ollier,1963, 1984; Gray, 1965; Rice, 1976) continued to support the idea as they could not perceive of what other process could possibly have been operative. Further, Ollier (1984) noted the limitations of the experiments undertaken by both Blackwelder and Griggs and suggested that the outcome was *not* relatable to most field situations. Amongst these limitations were the small size of the samples (that will become isothermal very rapidly, the fine grain size, the heating procedure (on all sides) and that some samples were polished which will increase the sample strength). Of great significance was also that, owing to the small sample size, it was obviously not confined and thus the exacerbating factor of "buttressing" (Winkler, 1984) could not occur, as would happen in many large blocks or in bedrock, and so the stress effects produced by this were not replicated. Buttressing is where the unaffected (or subject to a lesser heating) faces of a block or unheated bedrock inhibit lateral expansion of the heated face that must then be accomplished by outward expansion and accompanying fracturing (exfoliation). Expansive forces resulting from buttressing are significant (see Winkler, 1984; Folk, *et al.*, 1982) and can lead to multiple failures parallel to the relieving face. Finally, the studies dealt primarily with thermal shock rather than thermal stress fatigue and so provide no insight in to the latter. Essentially, the case is one where (most) geomorphologists have continued to cite unproven and untested assumptions until (Thorn, 1992, p.10) "...the story is one of casual empiricism gathering respectability by repetition until it attained the stature of an article of faith" (Hall, 1999). There is a very strong parallel to the above arguments provided by Warren (1914 a & b) where there is extensive discussion regarding whether certain Eolithic flints were the function of weathering or human action. In that discussion, Warren (1914a, p.418) considers the role of rapid temperature changes and (p.414) argues that the flaws found in the flints are "..most commonly the result of molecular strains set up by rapid changes of temperature (thermal fracture)". However, the significant arguments made by Warren (1914a, p.413) are with respect to the meaning of experiments, where he states: it would be unsound to "...assume that the result of a certain experiment must also be reproduced by natural agencies, without evidence that similar conditions exist in Nature to those employed in the experiment". This seems to be exactly the case with the majority of thermal stress experiments.

Thermal processes have been accepted, to some extent, as responsible for weathering in hot desert: "Although experimental work suggested that heating and cooling is not very important, the results are not conclusive, and there is some evidence to suggest that the process may be significant in hot deserts where there are large diurnal variations in temperature" (Trenhaile, 1998, p. 47). Smith (1994, p.40-44) provides an excellent review of the role of thermal stress within hot desert environments. Perhaps surprisingly, Smith (1994, p. 40) is not as strongly in favour of its operation as might have been thought: :High rock temperatures may therefore not only be unnecessary but undesirable for many desert weathering processes". Nonetheless, the *potential* for thermal stress fatigue exists and the key issue, as pointed out by Smith (1994), is the need for detailed, accurate rock temperature data against which to assess process. The point needs be made that *cold* deserts may be an even better location for this process as they may be drier (precipitation is in the solid form) and, due to the low air temperatures couples with high radiative inputs, so thermal variability may be even greater. A significant factor here is that several studies (e.g. Nienow, 1987) have shown that wind can greatly affect rock temperatures such that the largest variations occur beneath the surface, even during times of constant radiative input and air temperature.

The key factor in thermal stress fatigue is rock temperature data. As pointed out by such as McGreevy and Whalley (1982) and Smith (1994) rock temperature data are neither as extensive nor as detailed as we frequently assume. Thorn (1988) has noted that air temperature data are not a surrogate for rock temperatures and so the very minimum required would be that of the rock surface. Even then such data should be for different aspects and angles of rock surfaces (see Hall, 1998, Fig.2). Ideally, temperature data should also be recorded at different depths (see Hall, 1998, Fig. 5) so that temperature gradients can be calculated. Further, all of these data should be collected at a frequency that allows for meaningful calculation of $\Delta T/t$; a measurement frequency of every minute has been suggested as optimal (Hall, 1997, 1998). Few studies have measured rock temperatures, especially for any duration or variety of locations, at this frequency (see Hall, 1997) but where such data have been obtained (e.g. Hall, 1997, 1998) it has been demonstrated that events equal to or greater than the threshold for thermal shock do occur and there are multiple events that are likely to cause thermal stress fatigue and spalling. Significantly, analysis of one-minute temperature data show the effects of different measurement intervals (Hall, 1998, Fig.3) such that even a five-minute measurement interval cannot resolve all of the important events. This becomes all the more significant when the impact of wind fluctuations on rock temperatures are considered (e.g. Nienow, 1987). Nienow (1987) modelled thermal changes in rock as a function of air temperature, internal light gradients, and wind speed and showed that short-term temperature fluctuations to a depth of *c.* 4 mm were caused by fluctuating wind speeds over the rock. Interestingly, Nienow (1987), complementing the work of McKay and Friedmann (1985), showed that maximum rock temperatures occurred not at the surface but rather at 1 cm depth. Such an occurrence could readily explain spalling due to generated stresses. The study on Nienow (1987) failed to show any impact of light penetration through translucent minerals upon the thermal regime but this may be a consideration with some mineral assemblages that has yet to be fully evaluated.

It may be that the study of thermal stress fatigue, from a geomorphological perspective, is now going full circle such that available data are suggesting this process may well be active in a number of environments. Indeed, in some instances it is difficult to find an alternative mechanism to explain the observed breakdown. Certainly, there is a need for better data collected from, and for greater application of linear elastic fracture mechanics theory to, geomorphological situations. Considering the potential stresses, both grain-to-grain and along thermal gradients, it should not be too surprising to find that thermal stress fatigue/thermal shock play a significant role in rock breakdown in a variety of climates.

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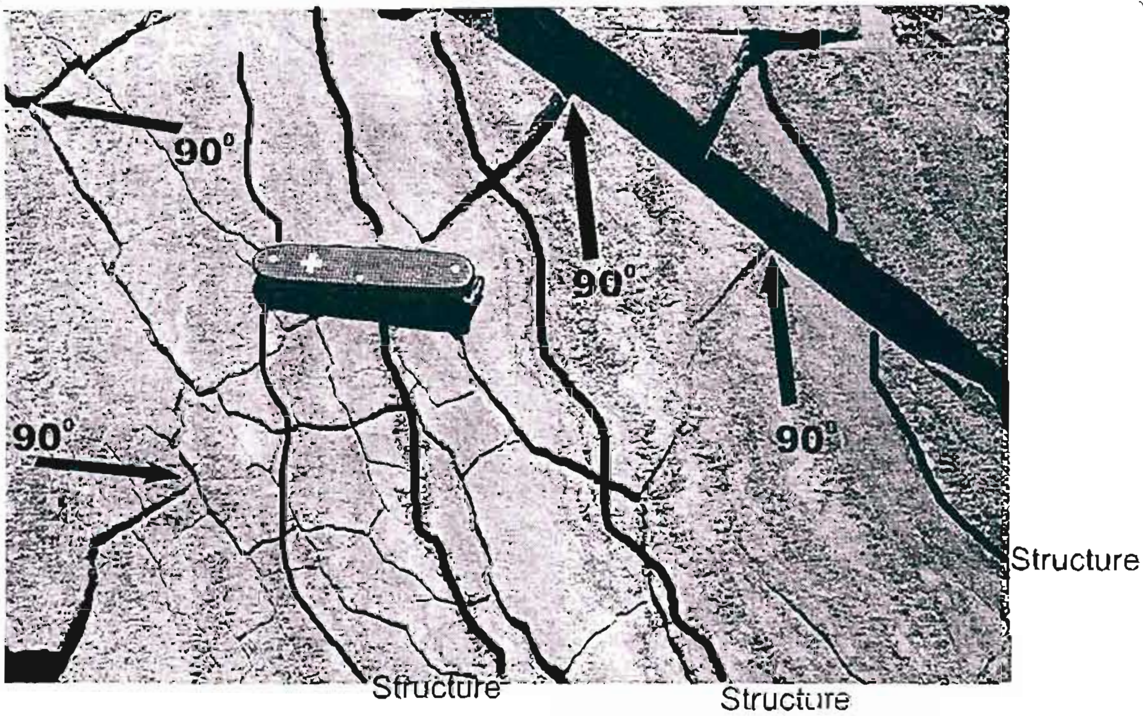

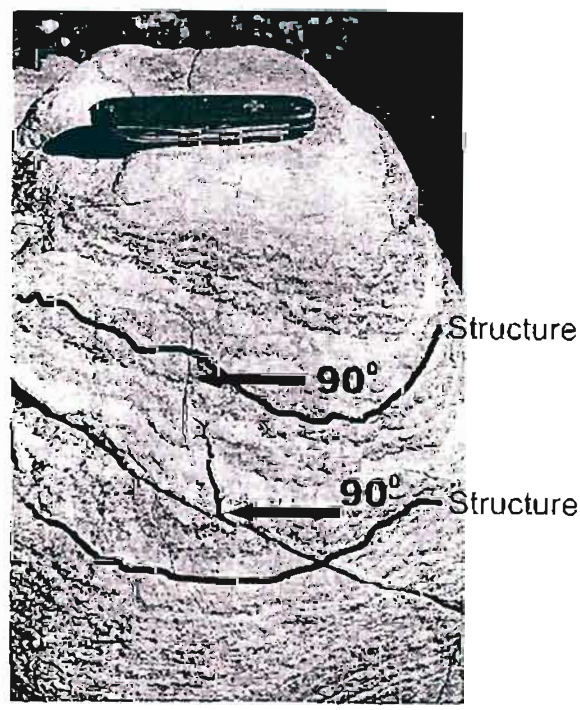

Examples of rock fracture due to, it is thought, thermal stress fatigue or thermal shock. Note that the fractures do not follow the structures within the rock(outlined by black lines) and that there is both a hierarchy of cracks and that they frequently join at right angles (some examples are shown) - all factors observed in laboratory studies of thermal fracture.

INTERNATIONAL UNION FOR QUATERNARY RESEARCH
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INQUA 87

CANADA

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July / juillet 1987

FROST WEATHERING IN THE MARITIME ANTARCTIC : SOME NEW INSIGHTS.

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Frost weathering is cited as an active agent in periglacial regions and is considered central to a number of processes (eg nivation). The bulk of studies have taken place in the Northern Hemisphere and yet frost weathering is a potentially active agent in the Antarctic where an extensive assemblage of periglacial forms are found. Despite its somewhat casual application, little is really known of the frost weathering processes, and even less, in terms of quantitative data, about the controlling factors.

Results of a current long-term study in the Maritime Antarctic have provided extensive data on both controlling factors and the actual processes involved. Data has been obtained from the field on rock moisture content, interstitial rock water chemistry, compressive and tensile rock strength, the stress intensity factor, weathering product size range, and the climatic environment, whilst laboratory tests have ascertained rock porosity, microporosity, saturation coefficient and water absorption capacity. Utilising a computer-controlled climatic simulation cabinet, large rock samples have been subjected to field conditions and it has been found that the orientation of laminae, in anisotropic rocks, affects freeze penetration, and that this can affect the weathering process.

Ultrasonic techniques have shown that during water absorption, very large changes take place within the rock. Also, ultrasonics have indicated that the nature and timing of the phase change is a function of a complex relationship between rock moisture content and its distribution and chemistry, the freeze amplitude and rate, and rock properties. Results of current experiments will be presented.

INTERNATIONAL SYMPOSIUM ON THE QINGHAI-TIBET PLATEAU

ABSTRACTS

Formation

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JULY 1998

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Evolution

Sustainable Development

July 21st ~ 24th, 1998

Xining, CHINA

MECHANICAL WEATHERING IN COLD REGIONS: THERMAL STRESS FATIGUE, A FORGOTTEN FACTOR

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Most studies regarding present or past cold regions emphasise the role of freeze-thaw weathering. Although this form of weathering does likely take place in many areas the problem is that many studies regarding it as the dominant or, in some cases, the only form of weathering. Whilst temperature data often appear to substantiate the role of freeze-thaw weathering the limiting factor is really the availability of moisture. At high altitudes and latitudes water is limited and thus the action of freeze-thaw weathering is constrained to locations where water is present. However, at these elevations and latitudes, thermal conditions are such that they alone may cause rock breakdown as a result of thermal stresses. The majority of studies have ignorred the role of thermal stress fatigue and/or thermal shock in causing mechanical weathering. One of the main reasons for the ignorring of this weathering process is the lack of adequate data as temperature record intervals need to be one minute (or less) and few studies have achieved this for any length of time. Recent studies in the Antarctic and in the canadian Rockies have collected thermal data at 30 second and one-minute intervals and these show that both thermal fatigue and thermal shock take place and may be a better explanation for rock breakdown. Fracture patterns found on rocks at high elevations in the Andes mountains are comparable to those produced in thermal shock tests in the ceramic industry. Examples of temperature data exemplifying thermal stress and thermal shock will be shown together with the character of patterns produced artificially and naturally as a result of thermal changes. It will also be shown that fracture patterns need not be angular and, in fact, rounded weathering forms can result. This latter may have serious implications for the analysis of Quaternary deposits in former cold regions.

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Short Communication

A Note on Biological Weathering on Nunataks of the Juneau Icefield, Alaska

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ABSTRACT

Observations on a number of nunataks of the Juneau Icefield indicate that chasmolithic algae play a major role in the breakdown of granitic rock. Expansion and contraction of the algal mucilage, caused by wetting and drying episodes, results in the surface flaking of the rock. Available data suggest that the average mass of material lost per year from 1 m² of rock could be as high as 562 g. It is suggested that biological weathering may be a major destructive mechanism of the granitic lithologies.

RÉSUMÉ

Des observations sur plusieurs nunataks du glacier Juneau indiquent que des algues chasmolithiques jouent un rôle capital dans la rupture des roches granitiques. L'expansion et la contraction du mucilage algaire, dues à des périodes de sécheresse et d'humidité entrainet la détachement de plaques superficielles de roche. Les données disponibles suggèrent que la masse de matériau perdue par an sur une surface de 1 m² de roche peut atteindre une valeur aussi élevée que 562 g. Il est suggéré que cette altération biologique peut être un mécanisme de désagrégation capital pour des roches granitiques.

KEY WORDS Biological weathering Algae Nunataks Juneau Icefield

INTRODUCTION

The physical and chemical role of lichens, algae and bacteria in rock breakdown has been recognized for some time and for a variety of environments. Muntz (1890) was the first to suggest that bacteria play a role in rock breakdown, while at about the same time Bachmann (1890, 1892) was investigating the effects of lichens on the weathering of different rock types. In addition to their chemical role (e.g. Branner, 1897), the mechanical effect of these organisms has long been recognized (e.g. Fry,

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1924, 1926). The role of these various biological agencies has been considered as significant in deterioration of building material (Mellor, 1922; Palmer, 1989) as well as for their effects on rock breakdown in various environments, particularly hot deserts and Antarctica (e.g. Friedmann *et al.*, 1967; Friedmann, 1971; Broady, 1981; Kappen *et al.*, 1981; Friedmann and Weed, 1987). Algae have received less attention than either bacteria or lichens but in recent years their effects in desert areas, particularly the Antarctic, have been recognized (Friedmann, 1971; Friedmann and Ocampo, 1976; Broady, 1981; Vincent, 1988). According to Friedmann and Gulun (1974), the so-called 'lithophytic algae' (i.e. rock algae) occur in four broad groups: epilithic (on the exposed rock surface), chasmolithic (in fissures), endolithic (in the internal air spaces of the rock fabric) and hypo- or sublithic (on the undersurface of light transmissive stones found in soil).

Studies to date on the Juneau Icefield have emphasized the role of mechanical weathering processes (Hamelin, 1964; Shenker, 1979; Klipfel, 1981; Linder, 1981) and, to a lesser extent, chemical weathering (e.g. Dixon *et al.*, 1984). However, during the summer of 1989 observations on a number of nunataks of the Juneau Icefield indicated that biological agencies, primarily algae, were a major contributor to rock breakdown.

STUDY AREA

The Juneau Icefield (Figure 1), a relict of the great Cordilleran ice sheet, covers an area of approximately 4000 km^2 along the Alaska-Canada Boundary Coast Range (Marston, 1983). Some 30 glaciers drain the Icefield, which is situated within a maritime environment along the southern and western edges of the Coast Range but which becomes more continental inland towards the east. Specific details regarding the glaciers, climate and mountains of the Icefield can be obtained from Miller (1964). The studies were undertaken at two main sites, designated as C-10 and C-18 (Figure 1). C-10 (Camp 10) is on a nunatak that rises to a height of *c.* 1555 m a.s.l. and approximately 426 m above the surrounding ice. Observations were obtained from the west-facing side of the nunatak. Additional data were obtained from the C-18 (Camp 18) region on nunataks just to the south of the Alaska-Canada border, in the region of the Gilkey Glacier at an altitude of *c.* 1700 m a.s.l. Sampling took place from glacier level to *c.* 100 m above the ice.

METHODOLOGY

Observations regarding extent of rock damage were made on randomly identified 1 m × 1 m or 0.5 m × 0.5 m areas. In these squares the area of either the damaged or undamaged surface was obtained by measuring (to the nearest millimetre) the main axes of the affected parts. In some instances it was only possible to measure the areas involved, while at other times samples of rock material removed were immediately weighted and then later, if possible, dried in an oven at 105 °C for 24 h and weighed again, so that both the mass of rock removed and its moisture content could be calculated. Equally, at some sites the mass of the removed material was obtained but the area it was derived from was not calculated. Samples of the active organisms were collected and returned to the laboratory, where they were cultured for identification. General field observations regarding the distribution of biologically weathered bedrock, its altitude and aspects together with lithological associations were made.

RESULTS

Biological activity was only apparent on granitic rocks in the study areas. No evidence of biological weathering could be found on the dark, gabbroic lithologies, where a distinct relationship of other mechanical weathering processes to aspect was observed, with the south-facing sides of outcrops showing severe granular disintegration, while the north-facing sides were resilient and unaffected. The granitic rocks showed no such effect of aspect but rather had a 'pock-marked' appearance as though hit by a geological hammer (Figure 2). Thin bands of green-coloured algae were also often present and were particularly evident along the edges of the rock after a flake had been removed. It was very noticeable that while these flakes could be removed with ease during and immediately after wet weather, no amount of prising with steel knife blades could dislodge them subsequent to several days of hot weather. The removed flakes (Figure 3), of which large numbers could be found at the base of outcrops, were all in the region of 2-4 mm in thickness but varied greatly in their areal expression (measurements from 4 cm^2 to 600 cm^2 were recorded).

Measurements regarding the biologically affected area, the area of additional material that could easily be removed, the total affected area and the weight of material removed are given in Table 1. In

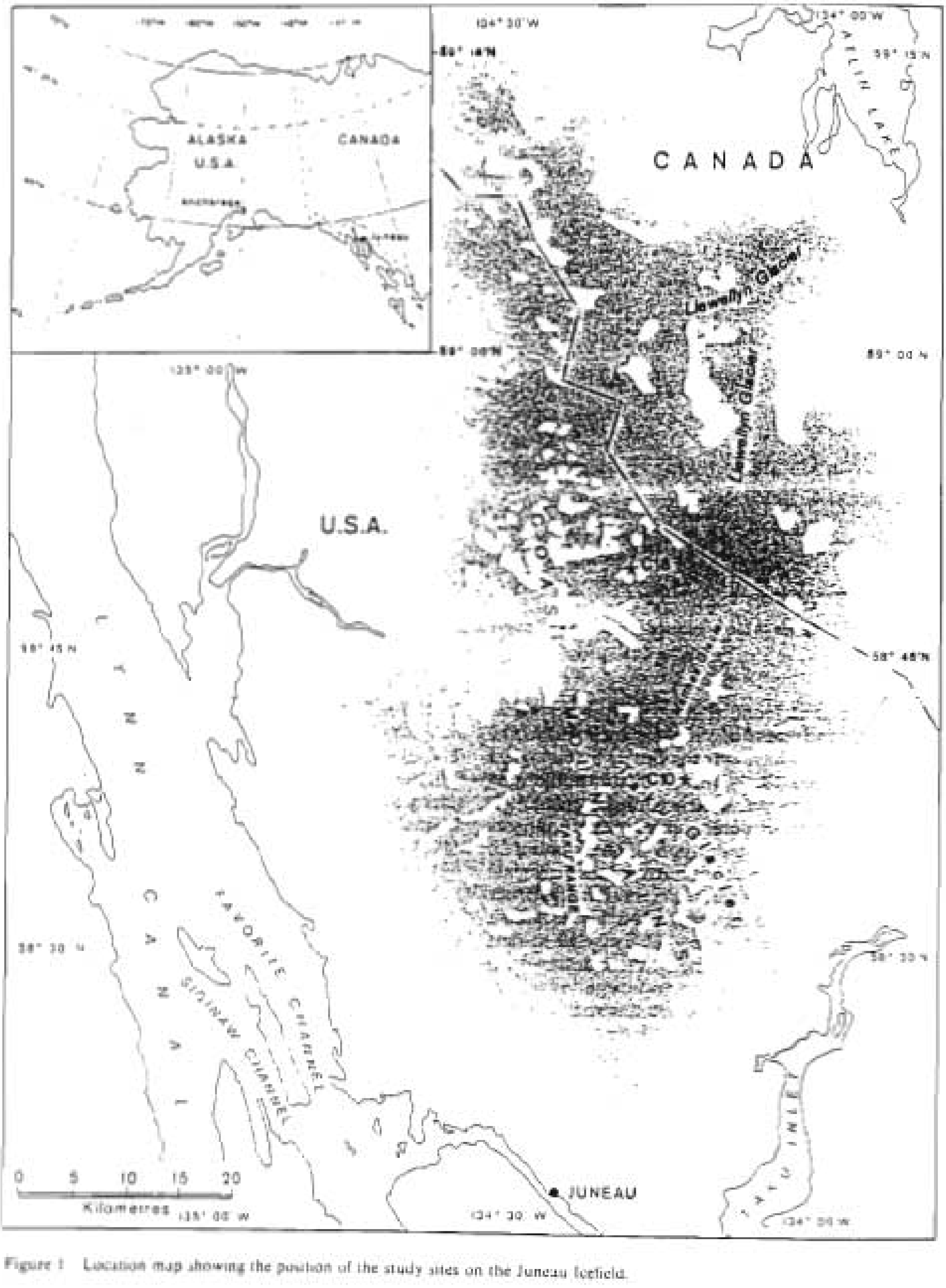

Figure 1 Location map showing the position of the study sites on the Juneau Icefield.

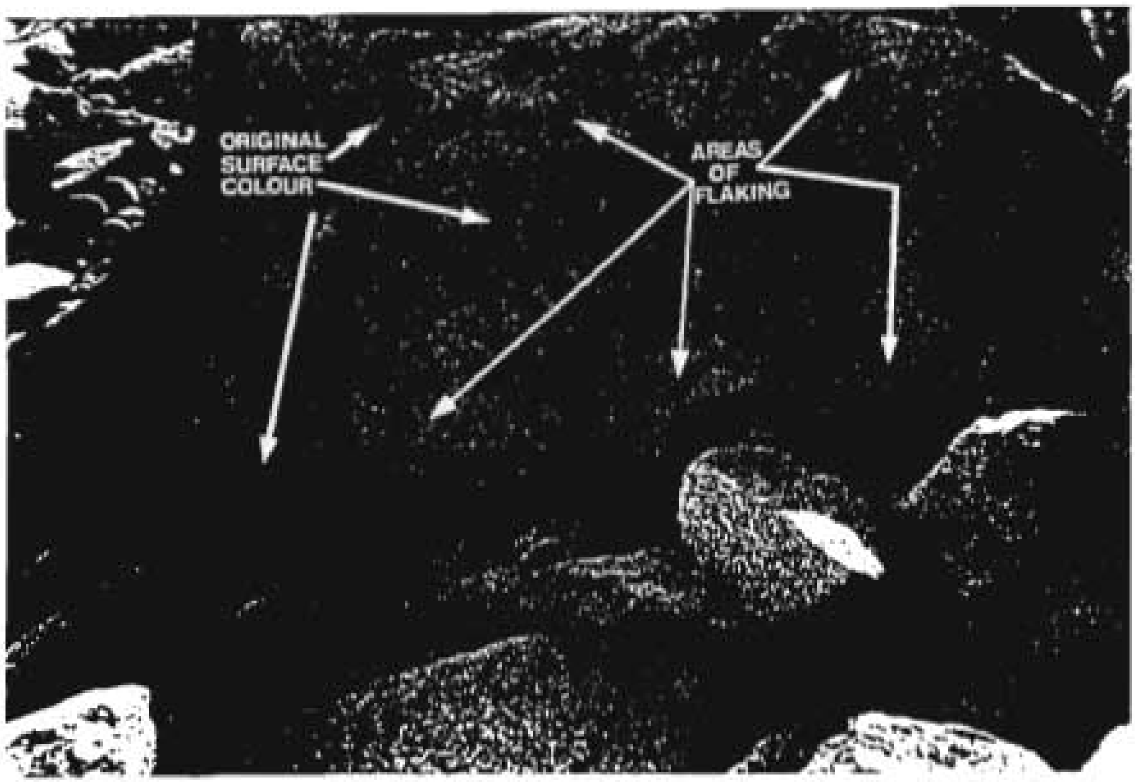

Figure 2 Flaking due to algal mucilage expansion and contraction on one face of a granitic boulder at the C-10 site.

addition, it was found that at several exposures there were inclusions of gabbroic rock within the granitic pluton which were unaffected by biological weathering and thus protruded above the surrounding surface. As the edges of these small (c. 5 cm × 5 cm) protrusions were stepped, it appears that they give some indication of the amount of material that has been removed; measurements ranged between 2.5 mm and 15 mm. Finally, it was most noticeable that the areas where active flaking appeared to be operative were relatively lichen-free, while surrounding unaffected lithologies had an extensive cover.

Cultures of the organisms from the rock indicate the presence of two species of chlorococcalean green algae, one filamentous form (single cells in a common mucilage) and a fungus. The chlorococcales are green in their early stage of development but can change to a red colour in later stages or when under stress from high temperatures and/or light intensity. It is thought that the fungal growth lives on the algal photosynthates or on the dead algae.

DISCUSSION

The algae are probably of the chasmolithic type (i.e. living in fissures or cracks within the rock), utilizing microcracks parallel to the rock surface that were created by dilatation. Although Friedman (1971, p. 419) states that alpine lithophytic algae are less well-researched than those of hot and cold deserts, it is considered that the algae live in the rock as a means of avoiding stressful environmental conditions (Vincent, 1988). Certainly the Juneau Icefield provides a range of stressful environments, from very cold in winter to very hot and dry in summer. During the study period rock surface temperatures (Hall, in preparation) attained levels as high as 38 °C and values of 20-25 °C were common on many days. That the chlorococcalean is a green alga but was seen mostly as red-coloured may also reflect stressful conditions for high light intensities and temperatures cause the chlorophyll to degrade, thus causing a change in colour from green to red. In other words, the red colour of the (green) alga indicates a green alga under stress.

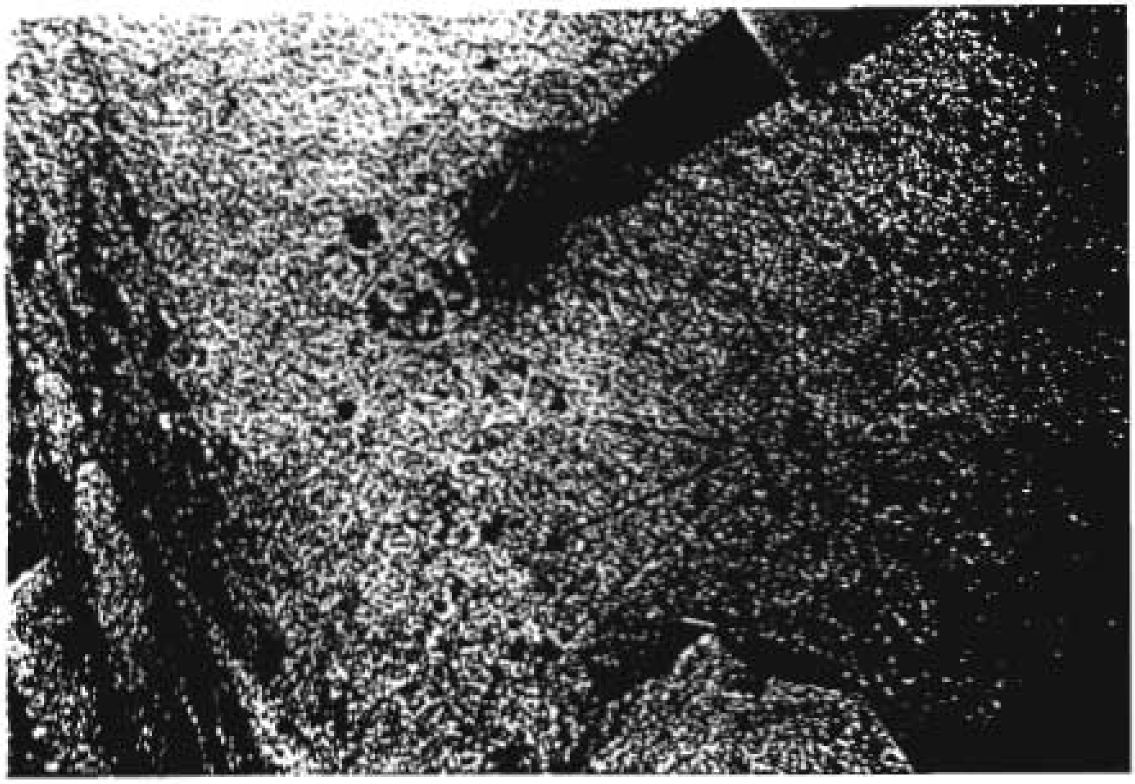

Figure 3 Granitic flake removed by hand with algae attached.

In order to retain their moisture during dry phases algal cells will first lose water from their surrounding polymer sheath or 'mucilage'. The latter is secreted by the algae for the protection of the inner cell and the amount of mucilage is partly controlled by environmental conditions: the more stress the more mucilage produced. This mucilage is hygroscopic, and it expands when water is available and contracts as moisture is lost. Expansion can be substantial (twenty-fold according to Correns, 1898) and during drying it exerts a very strong bonding force with a very high tensile strength.

It is this expansion and contraction that is seen as the damaging force. Friedmann (1971, p. 426) suggested that the 'shrinking and swelling of gelatinous cell sheaths ... may physically dislodge rock particles ... in a similar way to the activity of freezing and thawing'. The observation that plates of rock, with layers of algae up to 1 mm thick still attached (Figure 3), could be lifted free with the utmost ease during, and for a short time after, rain, while no amount of effort, even with a steel knife blade, could remove flakes after several dry days showed the effectiveness of the wetting and drying of the mucilage. The piles of rock flakes found at the base of outcrops clearly demonstrated the efficacy of this process in causing surface flaking of the rock. The flakes then break down into individual sand grains, probably mainly owing to the process of wetting and drying but also aided by thermal fatigue and frost action. Thus, the nunataks show a scene of rock outcrops in various stages of flaking with accumulations of small flakes at their base and the whole area with a covering of unconsolidated sand grains.

The findings of Broady (1981, his Figures 4 and 5) with respect to rock flaking and the extent of algal growth in 50 cm squares in the Antarctic are very similar indeed to the findings on the nunataks of the Juneau Icefield. Broady (1981, p. 263) found chasmolithic algae a few millimetres below the rock surface '... below thin flakes more or less parallel with the rock surface'. The surface flakes in the Antarctic were between 1 mm and 6.5 mm thick, while in the present study they were found to be 2–5 mm thick. Although Broady does not comment

Table 1 Information regarding the damage caused by algae as measured in the field.

Aspect	Angle (°)	Affected area (cm^2)	Extra material easily removed (cm^2)	Total area affected (%)	Dry weight removed by hand (g)	Moisture content (%)
SW-facing	40	780	?	7.8	?	?
Horizontal	0	8500	250	87.5	?	?
Horizontal	0	2800	74	28.7	?	?
Horizontal	0	1943	333	22.8	?	?
Horizontal	0	1723	100	18.2	?	?
SW-facing	90	3792	6	38.0	?	?
Horizontal	0	6572	986	75.6	580	?
N-facing[1]	90	2000	396	24.0	?	?
Horizontal	0	2014	110	21.0	?	?
Horizontal	0	1620	294	19.0	?	?
Horizontal	0	1438	126	15.6	?	?
E-facing[2]	15	?	398	?	579	2
SE-facing[2]	5	?	1251	?	15576	2
N-facing[3]	3	?	?	?	266	?
W-facing[3]	20	?	?	?	114	?
Horizontal[3,4]	0	?	?	?	709	?
S-facing[3,4]	70	?	?	?	456	?
Horizontal[3,4]	0	?	?	?	490	?
		$\bar{x}$ = 3016.6	267.5	32.6	596.3	
		s = 2401.4	282.2	25.5	438.5	

[1]Surface exposed from under snow cover by digging.
[2]Samples collected during wet weather but affected area not measured.
[3]Material collected but area not measured.
[4]Surfaces had varying gabbroic rock component.
? = data not available, owing to various reasons (i.e. balance or oven not available).

upon the areas exhibiting recent flaking (i.e. areas from which rock had been recently lost), he does note the widespread occurrence of the living algae, they being found in 75 % of the 61 km squares examined. Thus, the role of algae is significant over a large area and it is argued that on many nunataks of the Juneau Icefield they are equally, if not more so, prevalent and active.

From the available data (Table 1) it is difficult to discern whether any particular aspect of the rock is more prone to biological activity. In Antarctica Broady (1981) found growths restricted mostly to the W through SW to SSW-facing sides of outcrops and boulders, mainly on the downwind side of the exposure. However, the sandblasting associated with the powerful winds and hyperaridity of Antarctica is far in excess of anything taking place on the Juneau Icefield. Present data are insufficient to quantitatively show any effect of aspect, but qualitative observation failed to show any obvious orientational bias. Certainly observations about a number of large blocks (c. 2 m × 2 m × 2 m) at C-10 failed to show any apparent difference in flaking between any of the faces.

Detailed climatic data for the rock areas do not exist but available information (Hall, in preparation) suggests that during 34 days of observations there were a minimum of 16 wetting-drying cycles. In addition, on a number of nights when snow, rain, drizzle or fog did not occur, there was a relative humidity of ≥90%. Thus, the potential number of times that the mucilage polymer sheath could expand and contract during the spring to autumn period is relatively large (probably > 50) and so the destructive effect of the algae would be expected to be significant, particularly when the degree of expansion that can occur is considered. This same destructive force has been suggested by Palmer (1989) to be the cause of damage to historical churches in northern Germany. One of the churches was completed in 1893 and is already heavily weathered, with large pieces of brick having fallen away, while the other was completed in the eighteenth century and large chips of rock can easily be removed from this (Palmer, 1989). In the first case it is only for 96 years that the bricks have been exposed and in the latter for less than 300 years. In the case of the nunataks the rock may have been exposed for a much longer period. While no valid data are available, observations showed that *Rhizocarpon* sp. lichens in the same areas as the flaking rock suggest that the rock has been ice- and snow-free for 1000 years or more. Thus, considering the potential efficacy of this process, it is not surprising that the upstanding, unaffected gabbroic nodules indicate at least 1-2 cm of material lost over the whole rock surface.

The mass of material that could be removed from 1 m^2 was substantial (Table 1), such that, although no quantitative data are available, qualitative observation of the extent of algal activity suggests that this process far exceeds any other in causing rock breakdown. Certainly, on the basis of the amount of loose material ($\bar{x}$ = 562 g m^{-2}: Table 1) on rock surfaces and the accumulation of algal-derived flakes at the base of outcrops, the effectiveness of biological weathering seems to be substantially faster than that reported by other workers (see Introduction) for either mechanical or chemical weathering. Although *actual* weathering rates are not known, the indication by the upstanding gabbroic inclusions of the loss of 5-20 mm of material over the *whole* rock surface during the last *c.* 1000 years gives some idea of overall mass lost. However, this is still a conservative estimate in so far as it presumes the top of the inclusion to have equated to the original rock surface and this need not have been the case.

Once the flake is detached, then the exposed alga dies. The flake is then broken down by other weathering processes (probably a combination of freeze-thaw and wetting and drying) to produce the grus that abounds on these nunataks. The exposed rock surface is then subject to all but biological weathering until algae can recolonize, at which point the whole sequence is repeated. Thus, any granitic exposure shows all stages of biological weathering activity.

CONCLUSIONS

Until this present study, weathering of nunataks on the Juneau Icefield has been considered to be primarily mechanical, although the role of chemical processes has also been documented. This present study shows that the granitic rocks, in particular, are subject to very dynamic biological weathering. The expansion and contraction of the mucilage of chasmoendolithic algae that accompanies wetting and drying cycles effects a flaking of the rock surfaces. Although it is highly probable that chemical weathering by the algae also takes place, no data are yet available regarding this. However, the mechanical effects of the algae are such that several hundred grams of rock per square metre may be lost each year. The granitic outcrops examined showed the effects of this biological weathering over their entire surface, with large amounts of debris visible at the foot of vertical exposures. The inclusions of unaffected gabbroic rock stand as much as 10 mm or more above the surrounding surface.

It is probable that the role of algae in causing rock breakdown is far more complex than indicated by this preliminary note. However, algae are certainly a major weathering agency on the nunataks investigated and are likely to be operative on other nunataks in this region. More detailed studies, particularly regarding weathering rates and the role of biologically induced chemical weathering, are needed.

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A NEW WEATHERING MECHANISM SUGGESTED UPON EVIDENCE OBTAINED FROM THE DRY VALLEYS OF ANTARCTICA

Based upon observations made in the ice-free Taylor Valley of the Victoria Land region, Antarctica, a new weathering mechanism is suggested. The Taylor Valley, like the other dry valleys of this region, is a severely cold (winter minima below -60°C), arid location subjected to very powerful down-valley winds that may attain velocities of up to 300 km hr^{-1} (MIOTKE 1985). These winds are recognised to move unconsolidated surficial material such that sand dunes occur in some localities (CALKIN & RUTFORD 1974), abrasion by blown sand is frequently cited (SELBY et al. 1973), observations of particles 10 mm in diameter being lifted to a height of 2 m (BULL 1966) have been reported and a sand accumulation in the order of 3 Kg hr^{-1} for the height range of the ground to 40.5 cm recorded (MIOTKE 1985). In addition to these recognised abrasive and accumulative actions of moving sand it is now hypothesised that it can also act as a weathering agent.

The basis for this hypothesis is the multiple observation of particles of an extraneous origin packed into cracks on the windward side of rocks and under loosened slabs on the tops of boulders. The extraneous nature of the material was recognised by the presence of lithologies different to that of the host rock. It is suggested that the weathering action of the windblown material is somewhat akin to the volumetric growth of ice or salt within pre-existing cracks.

It is proposed that three possible inter-related processes are involved: (1) the direct pressure of the wind and material, (2) the hammering effect of moving particles upon those already wedged in place, and (3) the prevention of the closure of cracks opened by other forces. Although it appears very simple to quantify the forces involved, in reality the effects of turbulence, the size and form of the bearing surfaces, the role of Bernoulli forces and the lack of data regarding the velocity and direction of the wind immediately after impact preclude any meaningful calculations. This is why in an American study of the effects of wind pressure on rock obstacles (DOWDING et al. 1983) actual field measurements had to be collected.

However, as a rough guide it has been calculated that pressures of up to 62 Nm^{-2} could be generated by the wind and that these may be doubled if the wind 'rebounds' upon contact. In addition, the hammering effect of moving particles upon those already in place may produce pressures in the order of 202 MNm^{-2}. The role in precluding crack closure could not be quantified. The pressures involved thus vary between those that might cause fatigue and those in excess of the rock strength.

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Wind-blown Particles as Weathering Agents? An Antarctic Example

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Abstract

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Observations from Taylor Valley, Antarctica, suggest that the wedging effect of sand particles, packed by powerful winds into cracks in rocks, may act as a weathering mechanism. The strong winds and their accompanying solid material may cause stresses to the rock in three ways: the direct pressure of the wind and material, the hammering effect of the particles, and by preventing the closure of cracks opened by other forces. Although crude estimations of the forces involved have been calculated, actual field measurements are needed. This is a process which may operate in other dry environments on our planet and possibly on some of the hyper-dry inner planets of our solar system.

Introduction

Since their discovery during Scott's 1901–1904 expedition, and subsequent study by such well known scientists of the "Heroic Age" as Priestly, Armitage and Griffith Taylor, the dry valleys of Antarctica have been known for the effects of blown-sand on both rock and man (Markov and Bodina, 1961). Studies to date have concentrated on either the abrasive effects of the blown sand, clearly demonstrating its role in denudation and the formation of ventifacts (e.g. Selby et al., 1973; Miotke, 1982; Malin, 1985) or its role in the development of sand dunes (e.g., Calkin and Rutford, 1974; Miotke, 1985). Apart from studies in Victoria Land (e.g., Calkin and Cailleux, 1962; McCraw, 1967; Cailleux, 1968; Robinson, 1982) there are also reports regarding the effects of blown-sand from the Darwin Mountains (Whitney and Splettstoesser, 1982), the Vestfold Hills (Zhang and Peterson, 1984), Enderby Land and Dronning Maud Land (Sekyra, 1969).

Although Zhang and Peterson (1984, p. 23) suggest that the formation of cavernous ventifacts by blown-sand, snow and ice particles is a type of weathering, they are, in reality, describing abrasion. However, recent observations in Taylor Valley (Fig. 1) indicate that blown-sand packed into cracks under the force of the wind may cause rock breakdown in a manner akin to ice or salt crystal growth, i.e. a volumetric growth of an extraneous agency within a pre-existing crack. If this is so, then "sand wedging" would constitute a weathering mechanism. This mechanism has been referred to briefly by Campbell and Claridge (1988, p. 54); "sand wedging is also an important process in bedrock disintegration and the formation of felsenmeer surfaces". However, they do not de-

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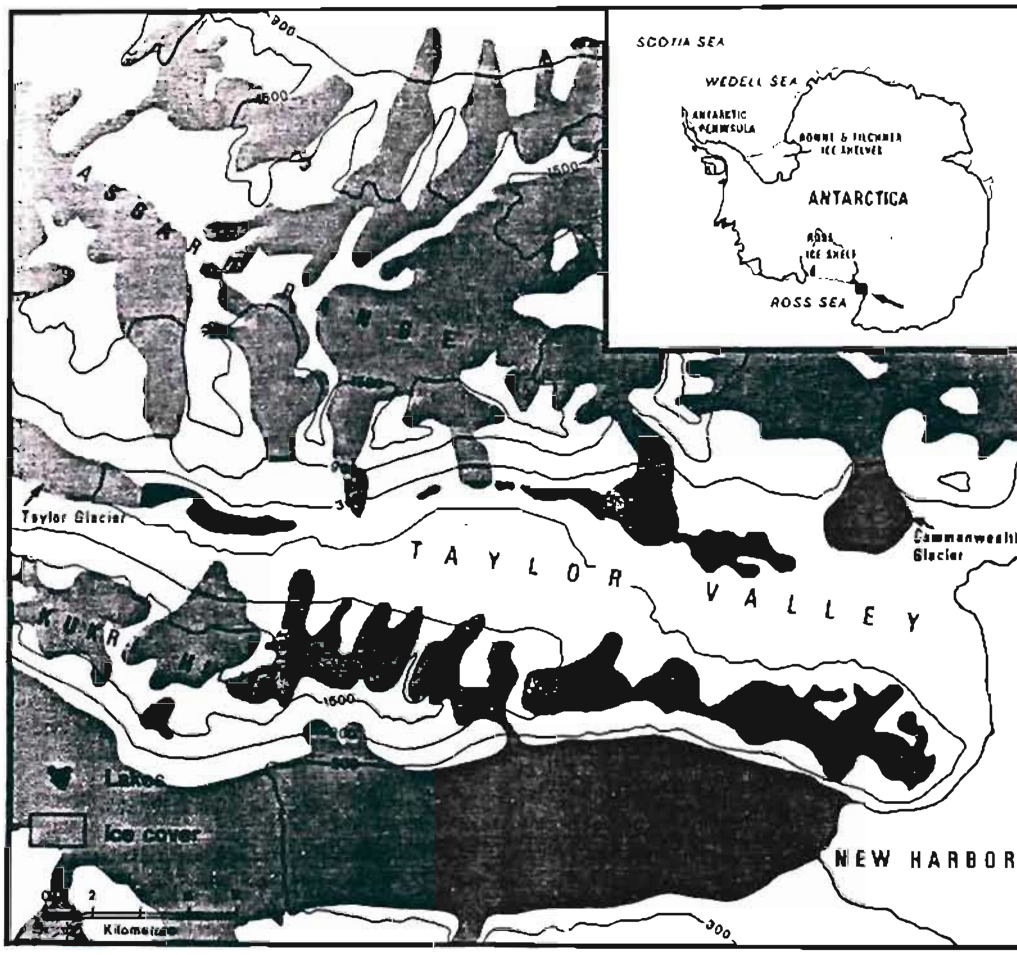

Fig. 1. Map of the study area.

scribe it in their recent major work on Antarctic weathering processes (Campbell and Claridge, 1987) and so their perception of its operation is unknown. The mechanism of "dirt cracking" described by Ollier (1965, 1984) for the hot desert environment of Australia appears to be a somewhat similar process. Thus, observations and data are presented in the hope that they may promote further investigations in other dry environments where blown-sand occurs. It is possible that this process might operate on other inner planets (e.g. Mars) for which this Antarctic dry valley region has been used as an analogue (e.g. Vishniac and Mainzer, 1973).

Study area

Taylor Valley (77°33′S, 163°25′E) is an ice-free valley situated in the mountainous South Victoria Land of Antarctica (Fig. 1), on the western side of the McMurdo Sound. The roughly east–west valley is bounded at its inland-end by the terminus of the Taylor Glacier (alt. 140 m) approximately 30 km from the coast. A number of glaciers margin both sides of the valley, whilst there are several perpetually frozen lakes along the valley floor. Detailed information regarding the climate of this valley is scarce (Miotke, 1982) as no recording station exists. There are however a number of short-term measurements associated with projects undertaken in this area. The best summary of climatic conditions is that of Bull (1966) in which the predominance of westerly winds is noted. One westerly, recorded by Bull, at the eastern end of Taylor Valley reached velocities of more than 70 km h^{-1} and was sufficiently strong to lift stones 10 mm in diameter to a height of 2 m. Miotke (1985) in the nearby Victoria Valley, cites velocities of up to 300 km h^{-1}! Temperatures are extreme, with winter minima below $-60°$C and the mean monthly temperature from April to September below $-30°$C (Thompson et al., 1971). Summer temperatures may reach $+4.6°$C but can rapidly drop to below $-20°$C (Miotke, 1984). Although long-term climatic data are not available, it is thought that precipitation is extremely limited (Bull, 1966, table 14) with Thompson et al. (1971) recording 82 mm in 1969 and 7 mm in 1970 in the nearby Wright Valley. Most of the snow is ablated by sublimation. The overall extreme aridity of this area is said by Nichols (1961) to be demonstrated by the dry kettle holes, small saline lakes, surface and subsurface efflorescences of salt, calcite veneered fragments and undrained lakes that have very small areas in comparison to their basins.

Most of the following observations were made across the width of the valley at its eastern end, from the coastal plain at New Harbour to the Commonwealth Glacier. A few observations were also obtained at the western end, around the margins of the Taylor Glacier (Fig. 1).

Observations and discussion

During several visits to the Taylor Valley, many cracks in rocks and the undersurfaces of loosened, but still in situ, slabs on the tops of boulders were found to be packed with coarse-grained material. Initially these were thought to be weathering products remaining in situ, but then it was observed that many of the crack surfaces were smoothed, sometimes with a patina, and lacked signs of granular disintegration. In addition, many of the particles were of a different lithology to that of the host rock (e.g. quartz or granitic granules were found in dolerites or basalts), and there was a tendency for granular material to occur more frequently in cracks on the windward (westerly) sides of the boulders. Thus it appears that this was largely allochthonous material blown into the cracks during periods of strong, westerly winds.

Textural analysis of a sample collected from beneath three detached slabs on top of a doleritic boulder (> 2 m a-axis) at a height of 0.5 m from the ground surface shows that it is predominantly of "medium sand" size, with 27% in the granule and pebble sizes. The five largest particles in the sample have a-axes of 20, 19, 18, 16 and 12 mm, respectively, and a maximum weight of 1.5 g, which suggests very powerful wind activity. In the nearby Victoria Valley, Malin (1985) calculated free-stream wind speeds of about 250 km h^{-1} and showed sand transport rates of between 10 and 280 g cm^{-2} at 0.5 m above the ground surface. In the same valley, Miotke (1985) measured velocities of up to 300 km h^{-1} and found that at 15 m s^{-1} (54 km h^{-1}) nearly 3 kg h^{-1} of sand was collected between ground level and a height of 40.5 cm. These data, together with Bull's (1966) observation of 10 mm stones being lifted to a height of 2 m, point to the powerful driving force of the wind in the dry valleys.

In the Vestfold Hills on the other side of the Antarctic continent, Adamson et al. (1984) refer to audible wind-generated shock waves during maximum gust velocities in excess of 50 m s^{-1} (180 km h^{-1}). They state (p. 187): "Considerable energy must be dissipated when these large, high velocity pulses of air hit stationary cliff faces". This suggestion is substantiated by the study of Dowding et al. (1983) with respect to the response of rock pinnacles to air blast pressures. They found that low frequency, subaudible, pressure waves are effective in exciting structures and that large pinnacles (3.2 Gg) responded to wind gusts at 48–64 km h^{-1}. They concluded that wind induced effects were not insignificant and that fatique and loss of strength resulted from large winds that produced vortex shedding of gusts.

Such powerful winds, and the granular material moved by them, may be able to operate as other than a solely abrasive medium. There are three (interconnected) mechanisms which wind and blown-sand may cause weathering. First is the direct pressure effect of the wind and the accompanying material that may be exerted upon pre-existing cracks. Second is the force due to grain impact, particularly the hammering on grains that are already wedged. Third is the possibility of material being wedged into cracks that were opened by some other process, such as thermal effects, which then prevents closure of those cracks upon relief of the initiating stress.

Wind-blown material forced into a crack will exert a pressure upon the bearing surfaces. In addition, as more material continues to be packed in, the force of the wind may cause it to operate in a dilatant manner (i.e. an expansive change in volume) and so exert even more stress. Although it appears theoretically simple to calculate the pressures exerted by various wind speeds, to do so satisfactorily requires knowledge of factors such as the velocity and direction of the wind immediately after impact, the effects of turbulence, and the role of Bernouilli forces. These controlling factors are all influenced by the size, shape and surface roughness of each individual rock, and by the size and shape of the specific cracks and their bearing surfaces. It thus seems reasonable to regard every rock as unique and any calculation of the applied pressure as extremely crude. In fact, this is why in the study of Dowding et al. (1983) they had to undertake actual field data collection of the pressures involved in order to solve the problem. However, as a *rough* guide, a 10 km h^{-1} wind will generate a pressure of 9.9 Nm^{-2} rising to 62 Nm^{-2} at 250 km h^{-1} (the highest free flow speed suggested by Malin, 1985) if the wind is presumed to rest upon impact. These estimates may be doubled if it is assumed that the wind "rebounds" upon contact with the rock. Taking turbulence and gusting into account would make them even higher.

The above is with respect to the direct pressure exerted by the wind and the material it transports. But, that material itself may exert a force upon impact, particularly if it hammers upon material already wedged in the crack. Again, this force would appear easy to calculate but, like the wind pressure, in reality the estimation is far from simple. The main problem is knowledge of impact duration. However, crude estimates can be made which indicate that a 100 km h^{-1} wind, carrying particles of 1 mm diameter with a specific gravity of 2.65 (quartz), an impact area taken as one-quarter of a sphere, and impact durations of 10^{-3} and 10^{-6} s, could impose stresses of between 0.20 MN m^{-2} and 202.65 MN m^{-2}. Obviously this is a large range, with values more than doubled for an increase of windspeed to 200 km h^{-1}. Patently, the lower values would have little effect whilst those towards the higher end would be in excess of the tensile strength of all rocks. Towards the higher end of the scale the problem is compounded by the possibility of the crushing-upon-impact of the particles in transit (a potential source of fine material?). However, the calculations do indicate the viability of this mechanism.

Finally, the possibility arises as to the material being wedged into pre-existing cracks that are open due to some other process and that their presence would then prevent closure upon relief of the initiating stress. Such a process would aid fatigue and thus contribute to the ultimate failure of the rock. The physics of this are uncertain, but qualitatively it could be envisaged that widening of a crack due to thermal considerations may occur and that if that crack were then packed with wind-blown material it would not be able to close when the crack-opening forces were relieved.

It is recognised that these observations pose more questions than they answer, and yet there is no question as to the occurrence of strong winds and the presence of extraneous particles in cracks. Actual measurement of the forces provided by wind and blown-sand upon rocks in the field, following the approach of Dowding et al. (1983), is required. However, in the light of another report regarding this mechanism (Campbell and Claridge, 1988) and the analogous process of "dirt cracking", it does seem that blown-sand may operate as other than solely an abrasive agent. In addition, this is a mechanism that may well operate on the hyperarid surfaces of other inner planets where particle transport by wind takes place (White, 1979).

Conclusions

In environments such as the dry valleys where other weathering processes are limited (Friedman and Weed, 1987), particularly in the severely cold, dry winters, the type of weathering described above, if operative, could be important. Thus, it is envisaged that the wedging effect of wind-blown sand can produce weathering which, operating in both series and parallel with other mechanisms, exploits lines of weakness generated by those other mechanisms. Whilst Antarctic conditions appear to be ideal for its operation, it may well be that similar activity could take place in hot deserts during periods of strong winds (Lancaster, pers. commun., 1987) and on some of the hyper-dry and windy planets of our solar system. Certainly, this is a process in need of more detailed investigation and it is hoped that this note might induce those fortunate enough to work in these regions to find information either in favour or against the suggested mechanisms of sand wedging.

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WEATHERING IN COLD REGIONS : SOME THOUGHTS AND PERSPECTIVES

"We do not improve our theories or models by admiring them, or by proclaiming how well they seem to fit our observations. The only way to improve them, and therefore to make progress, is actively to seek conflict between our models and the real world"
(McCarroll, 1997, p.1)

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Abstract

Weathering in cold regions has primarily focussed on the notion of 'cold', such that process and landform theory have generally used this both as the developmental criteria and as the outcome of palaeoenvironmental reconstructions based on landforms or sediments. As a result of this approach, the process focus in terms of weathering has been that mechanical processes predominate, with freeze-thaw weathering as the prime agent, and that chemical processes are temperature-inhibited, often to the point of non-occurrence. Here a reconsideration of the whole conceptual framework of weathering in cold environments is undertaken. It is shown that, contrary to popular presentations, weathering, including chemical weathering, is not temperature limited but rather by moisture availability. Indeed, summer, and oft-times even winter, rock temperatures are more than adequate to support mechanical *and* chemical weathering if water is present. Where water is available it is clearly shown that chemical weathering can be a major component of the weathering regime. The argument is made that there is no zonality to cold environment weathering as none of the processes or process associations are unique to cold regions; indeed, many cold regions show similar weathering assemblages to those in hot arid regions. Process-form

relationships are also questioned. The assumption of angularity with weathering in cold regions is questioned, all the more so as hot arid studies identify exactly the same angularity of debris form. Further, that all forms *have* to be angular is shown by field examples to be no more than an artefact of original unquestioning and oft-repeated assumptions, now over a century or more old. The argument is made that there is a strong need for the reconsideration of the nature of weathering in cold environments, that current theory should be questioned and challenged, and field observation undertaken within this revised frame of reference.

Introduction

In general terms, the perception of weathering in cold regions identifies three basic tenets. First, that weathering is dominated by mechanical processes. Second, that the predominant mechanical process is freeze-thaw. Third, that chemical weathering is not a significant element of cold region processes due to low temperatures. Upon these tenets are then built a number of secondary concepts. Among the most important are that angularity of coarse debris always reflects freeze-thaw weathering, and that mechanical processes, particularly freeze-thaw weathering, is the fundamental cause of most bedrock landforms. By and large these concepts have dominated cold region geomorphology for the past century. Consideration of most texts shows use, in some form, of these ideas with the unspoken implication being that they are well tested and proven. The concepts are 'comfortable' and appear, to a large extent, intuitive. However, as Thorn (1992) has clearly pointed out, these are all only assumptions that lack substantive corroboration, let alone unequivocal 'proof'. Indeed, careful consideration of most investigations suggest that experiments, in either field or laboratory, are set up to 'prove' the operation of, for example, freeze-thaw rather than designed to test it.

A major focus of this discussion centres on the perceptions associated with the prefix 'cold' in 'cold region weathering'. It is argued that this prefix serves to misdirect perceptions as well as directions of investigation. In fact, in many cold regions processes must be driven by warmth rather than cold. As paradoxical as this may seem, the reality is that unremitting 'cold' such as may be associated with permafrost profoundly limits, if not precludes, geomorphic, biological or pedogenic processes. It is the input of heat that raises the temperature, mobilizes water, etc. thereby initiating weathering processes or, upon removal, allows the subsequent cooling and freezing of the water to effect the weathering. Temperature changes, in most cases above and below freezing, are implicit in almost all weathering processes associated with cold regions. Without the rise in temperature little or no weathering would occur. Thus, it will be argued, that in considering warmth we may interrogate our system (i.e. what and how we measure) in a very different way and may, thus, gain the benefits of a different perspective.

A further argument focusses upon zonality. Are 'cold region' weathering processes truly zonal, might they not be azonal? Repeating the argument made above, would a new (i.e., an azonal) perspective illuminate afresh our understanding of the operative processes and their potential synergy? In reality, chemical weathering processes and the majority of mechanical processes, e.g., salt weathering, wetting and drying, thermal stress, and dilatation can occur in any climate, very specific physical limitations notwithstanding. The real issue in the overwhelming majority of instances is not presence or absence, but rather a question of absolute frequency and magnitude, and also relative intensity. For example, the intuitive assumption is that freeze-thaw weathering is indeed zonal, but examination of *hot* desert texts (e.g. Abrahams and Parsons, 1994; Thomas, 1997) indicates that freeze-thaw weathering is identified as a viable component of the weathering suite. Thus, the only fundamental zonal limitation is that sub-zero temperatures occur, and such a situation extends the role of freeze-thaw action well beyond polar or alpine zones. An important corollary to azonal freeze-thaw action is the operation of thermal stress fatigue within cold regions. Again, hot desert texts include much discussion of the role of thermal stresses, but cold region texts (e.g. French, 1996) address the topic sparingly or not at all. Nevertheless, thermal stress fatigue/shock is ideally suited to cold regions and may, indeed, be more plentiful there than in hot deserts! In the absence of field data, the perceived importance of such processes is largely controlled by the preconceived notions that the terms 'cold' and 'hot' trigger.

What is measured and within what temporal framework also profoundly color our characterization of 'cold' or 'warm' . As Thorn (1992) has pointed out, measurement of air temperature is no surrogate for that of the rock. Unfortunately, a substantial portion of environmental and weathering literature is founded on the plentiful availability of air temperatures: as such it is wholly misleading. In recent years direct measurement of rock temperature, which has clearly shown a lack of close correlation with that of the abutting air (e.g. McKay and Friedmann, 1985; Engelskjøn, 1986; Ødegård and Sollid, 1993), has grown, but a great deal more is needed. While logistical and technical constraints are readily apparent to any periglacial field researcher, one of the results of researchers attempting to 'prove' the presence of freeze-thaw (e.g. Ødegård, 1993; and Gardner, 1992) has been the creation of temperature measurements that are of low frequency and inadequate spatial resolution. Such data may serve to characterize the freeze-thaw environment of rock, but they are of no value in characterizing its short-term thermal variability (Friedmann, *et al.*, 1987; Hall, 2000). Comparison of one-minute interval data with that obtained from longer time intervals (Hall, 1997a: Hall and André, 2001) clearly demonstrates that the latter fail to produce an adequate perception of the thermal fluctuations a rock experiences. Measurement of rock temperature at depth is equally important because it permits determination of several other important attributes, for example, thermal gradients, thermal fluctuation at depth, and derivation of stress fields in the z-axis. Such measures are extremely important: for example Nienow (1987) has shown that the largest temperature fluctuations in a rock are sub-surface; while Kingery (1955) and Thirumalai

(1970) have addressed the role of vertical temperature distribution within rock with respect to spalling and a 'spalling index'.

Aspect and slope also play fundamental roles in thermal variability. These factors become highly significant at high latitudes where thermal conditions can run counter to intuitive assumptions. For example, Hall (1998a) has shown that poleward rock exposures can experience, for a small window of time, higher temperatures than the equator-facing exposures; in fact, the latter may even experience sub-zero temperatures when the poleward face does not. Thus, in addition to careful consideration of monitoring interval, considerable thought must be given to spatial microvariability if field installations are to provide the opportunity for realistic appraisal of weathering environments.

Last, some rethinking must be given to the characteristics of weathered material. Must mechanical weathering products in cold regions be angular? Indeed, paradoxically, do we not already identify some mechanically weathered cold region landforms as exquisitely 'rounded'? The question must also be asked about how much we actually 'filter' what we see based on what we expect to see? Just as the case is growing (see below) in favour of a far greater role for chemical weathering in cold environments, so too might be the one for non-angular, mechanically weathered products (e.g. Ballantyne and Harris, 1994; Hall, 1997a). Certainly, rock fracture mechanics have never suggested that linear fractures must be the only outcome of mechanical weathering; curvilinear fractures can certainly occur in some rock types (Rossmanith, 1983). So, despite the fact that field observations show the majority of rocks to fracture in an angular fashion, perhaps there should be greater effort directed at investigating situations where the rock does *not* crack in this way; a classic example of this might be taffoni. Taffoni are distinctly rounded forms (Fig. 1) and yet the bulk of the literature deems them to be the product of some form of mechanical weathering. Even though the rounding is considered by some to be the result, in some instances (but not all), of granular disintegration and thus rounding is not so unexpected, this juncture of 'roundness' with 'mechanical weathering' seems never to have been extended to other weathering products. Perhaps our preconceived notions of 'frost action' equates to 'angular clasts' has been so ingrained in our thought processes that we filter out anything that does not fit? Indeed, many field researchers appear to invert cause and effect by identifying mechanical weathering purely on the basis of the presence of angular fragments.

The preceding sketches highlight the principal themes to be pursued in detail in the remainder of this paper. Frequently the issues will overlap, but the primary objective is to separate presumption from substance, distinguish weakness from strength, and in the process develop penetrating questions that can truly be tested and thereby improve the science of weathering in what we loosely term 'cold regions'. The aim here is *not* to undertake a review *per se* although it is inevitable that, in order to address the issues noted above, substantial recourse to literature is required. For reviews of the literature, the excellent undertakings in this journal by McGreevy (1981), McGreevy and Whalley (1984), Whalley and McGreevy (1983, 1985, 1987) and, for the French literature in particular, Lautridou and Ozouf (1982) are suggested. Pictures and/or data could be presented but the problem is the 'context' in which to present them. As we hope to show, many studies have assumed context and, as a result of this, "proved" the initial hypothesis but this is not the same as having "tested" the hypothesis. The same can be argued for showing data: it is the context in which it is derived and/or the end to which it is used that are the problems, not the data itself. Thus, the need is for greater consideration of our questions, that which underlies what we do and how we do it, rather than trying to support one argument or another by the presentation of data.

Chemical Weathering

Łozińsky's (1912) definition of 'periglacial' invoked freeze-thaw weathering as a necessary, but not sufficient, attribute. Consequently, it is hardly surprising that periglacial geomorphologists have always embraced freeze-thaw weathering as a dominant and given theme in the discipline. Two unfortunate circumstances have stemmed from this posture. First, because the universal efficacy of freeze-thaw weathering is taken as a given, there has been inadequate research into this allegedly pervasive process. Second, again because of the presumed dominance of freeze-thaw weathering, other weathering mechanisms have not been investigated thoroughly. Specifically, the combination of presumed dominance by freeze-thaw weathering and cold air temperatures in periglacial climates has promoted an entrenched belief that chemical weathering is all-but-absent in cold climates. There is, in fact, a long - albeit scanty - literature supporting the notion that chemical weathering is significant in cold climates.

Laboratory work by Tamm (1924) indicated that, at the temperatures commonly experienced at the Earth's surface, chemical weathering in cold regions would not be arrested. To this may be added William's (1949) observation that carbon dioxide is more soluble at colder temperatures. Other scattered commentary, such as Boch's (1946) comments concerning the role of chemical processes in nivation, may also be found (see elsewhere in this volume). However, it is really Rapp's (1960) comparative study of transport processes in Kärkevagge (Arctic Sweden) that staked a truly fundamental claim for the role of chemical weathering in a periglacial context.

Using the metric 'vertical tonnes moved', Rapp calculated that solution was the single most important transport mechanism within the glaciated U-shaped trough (now periglacial valley) of Kärkevagge: see Caine (1976) for a similar study reaching the same kind of conclusion. As transport in solution must necessarily be preceded by chemical weathering this claim merited great attention. Interestingly, while Rapp (1960) is considered a seminal paper within periglacial geomorphology, this particular aspect of his paper received relatively little attention.

Glacial geomorphologists appear to have more readily adjusted their thinking concerning the role of chemical weathering within their domain than have periglacial geomorphologists. Work by Reynolds (1971) is an early example attesting to the (potentially advanced) nature of chemical weathering in glaciated mountains. More recently, glacial geomorphologists have taken up this issue within the context of global chemical cycles (e.g., Anderson *et al.*, 1997). This work has led to the conclusion that glaciated terrains may experience chemical weathering rates that are quite high, but by virtue of their water discharge totals rather than because of weathering intensification *per se*. Chemical signatures of glacial waters may also be distinctive due to inhibited silicate weathering but increased weathering of potassium and calcium (Anderson *et al.*, 1997). Interestingly, Anderson *et al.*, (1997) specifically call attention to the need for greater scrutiny of glaciated areas presently beyond glacial margins - clearly, an environmental context that encompasses much of the periglacial regime.

Research into the chemical weathering regime of presently periglacial regions has been fairly focused geographically. Among those building a strong regional picture has been Caine (e.g., Caine,1989; Caine and Thurman, 1990) who has demonstrated the significance of chemical weathering in the relatively mild continental periglacial environment of the Front Range of Colorado, U.S.A. In Norway, McCarroll (1990) has been able to show that chemical weathering merits much greater attention than it has received hitherto. Indeed, the modern trend to emphasize chemical processes in cold environments is not limited to the milder end of the cold region spectrum, Elberling and Langdahl (1998) found high rates of weathering of sulphide minerals occurring around Citronen Fjord at a latitude of 83°N in northern Greenland. Equally, Balke *et al.* (1991) noted the only limitation to chemical weathering during the Antarctic summer was moisture, not temperature; a sentiment also expressed in Meiklejohn and Hall (1997).

In recent years a multi-year project in Kärkevagge (summarized in Thorn *et al.*, in press) has been undertaken for the specific purpose of elucidating Rapp's (1960) original claim for the preeminence of solute transfer in this relatively mild, Arctic environment. While this work has shown that Rapp's original estimate was perhaps a little high, the general picture emergent from the recent work has strongly supported his original claim. The fundamental underpinning of the importance of chemical processes in Kärkevagge is derived from the presence of some pyrite-rich rock units, weathering of the pyrite produces sulphuric acid which, in turn, fuels rapid chemical weathering. In addition to determination of rock weathering attributes, both directly through electron probe and SEM observations (Dixon *et al.*, 1999) and indirectly through the study of surface waters (Darmody *et al.*, 2000a), soil development has also been examined (Darmody *et al.*, 2000b). The presence of secondary interlayered minerals in soils and their pattern of vertical distribution within soil profiles (Allen *et al.*, in press) has clearly demonstrated that chemical weathering is progressing in a fashion, but probably not rate, typical of the mid-latitudes, and appears to precede distinctive horizonation on the older parts of the landscape.

There is now sufficient evidence to state, categorically, that chemical weathering occurs at a magnitude that renders it significant among the suite of weathering mechanisms in many, if not most, periglacial environments. There is also sufficient evidence to indicate that the chemical processes that are occurring are those that would be anticipated given a knowledge of mid-latitude weathering regimes. What then are the important remaining issues? The most important scientific question, given the present state of knowledge, is 'What is the unifying concept, if any, that underpins modern findings?' The answer is that traditional conceptual frameworks for periglacial geomorphology have overestimated the importance of temperature in the range at which it occurs at the surface of the Earth, and have underestimated the importance of moisture. Simultaneously, traditional frameworks have made both of these mistakes in terms of air climates, while the true focus of attention should have been, and should be, ground climates. Pope *et al.* (1995) provide a comprehensive explanation of this issue. Given the present dearth of chemical weathering research within periglacial environments it is essential that the topic become a primary focus of future research in periglacial geomorphology.

Several distinct threads needs to be pursued within the general rubric of cold region chemical weathering. First, the absolute rates of chemical weathering should be determined. Second, the relative rates of chemical weathering processes should be determined. The latter issue needs to be taken up in two ways: chemical weathering rates relative to periglacial mechanical weathering rates, and periglacial chemical weathering rates relative to global chemical weathering rates. In the case of periglacial chemical weathering rates vis-à-vis periglacial mechanical weathering rates it is worth noting the emergence of at least some research that suggests recent periglacial mechanical weathering rates are, in fact, slow (André, 1995, 1996). Finally, the interaction between mechanical and chemical weathering needs to be investigated as there are data (Berrisford, 1992; Hoch *et al.*, 1999) that strongly suggest that mechanical processes may intensify chemical weathering rates; however, as might be expected, the chemical component of this research is well-founded while the mechanical component tends to be circumstantial. There is an obvious re-emergence of an important general theme in a specific study of periglacial chemical weathering - periglacial environments have much in common with mid-latitude, and probably many other, environments - the thread of commonality is the dominant influence of ground moisture availability. Periglacial geomorphologists cling to their entrenched, and poorly substantiated, belief in the profound difference of periglacial weathering regimes, founded overwhelmingly on thermal concepts, at the expense of making real progress in their understanding of cold region landscape development.

'Cold' Regions and Mechanical Weathering

A) Introduction

If weathering regimes are to be subdivided into a zonal climatic system it must be a functional one. As with any subdivision of the natural continuum, identification of core areas will be easier than demarcation of 'sharp' boundaries. But, in either case, air temperature does not afford an acceptable metric and geomorphologists must turn to 'ground climate' if they hope to unravel the intricacies of weathering (Thorn, 1992; Pope *et al.*, 1995). This necessary step is an enormous emotional, as well as intellectual one, for the geomorphic community: classic climatic classifications (e.g. Köppen, 1923) embrace air temperature, albeit indirectly through vegetation, and so do most notions of 'periglacial' or 'cryogenic' zones (e.g. French, 1996). Intuitively, high altitudes and high latitudes are 'cold' for not only do we have the visual evidence of snow and ice, even in summer, but there is also the clear impact on all life forms, as well as the strategies they employ to survive, if not flourish, in such regions (e.g. Marchand, 1996). However understandable, preoccupation with cold has produced a scientific imbalance in which the impact of heat has been underappreciated. While temporally constrained heat is not only present, it is present for much longer periods at the relevant ground surface than seasons derived from air climate temperature data would suggest. In reality, ground climate summers at high latitudes and high elevations in lower latitudes may be quite lengthy.

Examination of 'cold environments', in terms of rock temperatures, is not only fundamental, it has wide-ranging implications. In the realm of chemical weathering Balke *et al.* (1991) have clearly demonstrated that even on the Antarctic continent summertime rock temperatures are more than adequate to facilitate chemical weathering - if water is available. Indeed, multiple Antarctic studies have shown that even on the 'frozen continent' rock temperatures are frequently substantially above 0°C even when the air temperature is -10°C or lower (Kappen, *et al.*, 1981; McKay and Friedmann, 1985; Engelskjøn, 1986; Hall, 1997a). Several studies (e.g. Van Autenboer, 1964a & b; Hall, 1997a) have shown summer rock temperatures of +30°C or higher and that the period when, at least for part of the day, rock temperatures exceed 0°C extends well into the period when air temperatures remain below 0°C. For example, Hall (1997a, Fig. 5) showed air and rock temperatures for an Antarctic site for the January to March period for two years during which air temperatures were largely sub-zero but the rock experienced diurnal fluctuations across zero with positive temperatures sometimes reaching 20°C. Thus, while air temperatures at this site suggest a return to 'winter' conditions from about mid-January, the rock continues to experience not only daily positive temperatures but also a daily freeze-thaw cycle that might be highly effective if water were available, right through to mid-March. From mid-March onwards both rock and air temperatures are sub-zero and fairly comparable: although even then the rock occasionally shows sharper, more distinct sub-zero fluctuations indicative of both the impact of radiation receipts and the absence of a snow cover on the surface, as opposed to the air. Nienow (1987) also presents rock and air temperature data for several years from an Antarctic site at 77°36'S in which he demonstrated that rock experiences above zero temperatures (for at least part of a day) into March and starts to do so again in October, whereas the air only starts to experience temperatures above zero in December In short, from a thermal perspective, the Antarctic weathering season for all mechanisms *potentially* stretches from October to March - 50% of the year! Water availability, not temperature, is the limiting factor!

Considering snow-free rock surfaces, and accepting the need for moisture availability, it must be recognized that chemical and/or water-based mechanical processes (e.g. freeze-thaw) may occur over a much greater portion of the year than air temperatures suggest. Indeed rock temperatures of 20° to 30°C would greatly facilitate active chemical weathering, as well as salt weathering and wetting and drying in the presence of moisture. Both salt weathering and wetting and drying may be more effective closer to ice-free coasts and with decreasing latitude. While salt weathering is relatively well understood (e.g. Goudie, 2000), wetting and drying remains more enigmatic with respect to its mechanism and synergy with other processes (Hall and Hall, 1996; Prick, 1999). However, in many maritime environments, even at high latitudes, wetting and drying can be highly significant (Hall, 1991, 1993) and a better understanding of its role and synergistic interactions within the weathering suite is needed.

B) Wetting and Drying

The question of the effectiveness of wetting and drying as a weathering process in cold regions has not yet been greatly investigated or widely considered. Generally speaking, the effectiveness of the weathering process itself, and the mechanical causes of hydration weathering (for example when no mineral clays are present in the rock pores), have not been sufficiently documented (Prick, 1999). The destructive effect of wetting/drying has been, so far, linked to adsorption processes, which have been studied both theoretically (Sidebottom & Litvan, 1971; Feldman & Sereda, 1963) and experimentally (Fahey & Dagesse, 1984; Félix, 1984). Rock swelling when in contact with water (liquid or vapour) has been only known in recent decades and documented by only a few dilatometry studies (Nepper-Christensen, 1965; Félix, 1983; Pissart & Lautridou, 1984; Hamès *et al.,* 1987; Schuh, 1987; Delgado-Rodrigues, 1988; Weiss, 1992). These studies showed the impact of rock microporosity (Hamès *et al.,* 1987), of clay mineral content (Hamès *et al.,* 1987; Schuh, 1987), and moisture distribution (Delgado-Rodrigues, 1988) in the dilation response to hydration. In the future, such studies could lead to a better understanding of the processes involved (e.g. the linkage between dilation and pore volume increase, *cf.* Sidebottom & Litvan, 1971) and possibly to recognition of the fatigue response.

For hot deserts, wetting and drying has been given some recognition as a possible weathering mechanism (Griggs, 1936; Smith, 1994) where one of the processes involved could be a thermal dilation of the water contained in the rock pores, which builds up fatigue stress (Birot, 1968). Hudec and Sitar (1975) determined that adsorption sensitive rocks dilate as much during an hydration event at constant temperature as during a 78°C temperature change when dry. The occurrence of weathering by thermal shock has been suggested for cold deserts (Hall, 2000) but there is limited information pertaining to the possible interaction between hydration and thermal stress and that which is available deals only with hot deserts. However, following the arguments in this discussion, including recognition that water is the limiting factor in both hot and cold deserts, potential for interaction between hydration and thermal stress in cold environments must be considered.

Wetting and drying action in 'cold regions' can be analysed by looking first at the process itself, the way it acts at positive temperatures and how its operation will be influenced by the cold, and secondly by considering wetting/drying action against what is generally considered as the major weathering process in cold areas, i.e. frost shattering. The shrinking of a rock sample during a wetting/drying cycle is greatest during the last stage of the drying process, i.e. when the moisture content of the rock is lowest (Pissart & Lautridou, 1984; Hamès *et al.,* 1987). This means that atmospheric humidity variations (or even dew formation) could be sufficient to produce weathering-effective wetting and drying cycles. It also means that fatigue stress can be better induced under climatic conditions offering deep drying phases, i.e. climates with a dry season (like the Mediterranean climate, as underlined by Hamès *et al.,* 1987), or, by implication, some cold environments. If we adapt Hamès' conclusions to cold regions, a task that can only be undertaken conceptually in the absence of any pertinent field data, two kinds of climatic condition may meet the deep drying requirement. First, more continental climates would fit this requirement, allowing a deep seasonal drying, but reducing drastically the occurrence of liquid water and the number of wetting/drying cycles per year. Second, climates with wet summers would allow a greater occurrence of hydration processes, but in this case, drying would have to depend on the freezing process : ice-lensing (see below: Walder & Hallet, 1986; Matsuoka, 1988; Prick, et al., 1993; Prick, 1996 & 1999), and the induced water migrations towards the already formed ice lenses, would cause local (and certainly temporary) drying of some parts of the porous media. The lack of climatic data, in particular of data about moisture content within the rock (Hall, 1995) in various cold environments, makes it difficult to reach any firm conclusions regarding this idea.

Recent dilatometry research (Prick, 1996 & 1999) has shown that sandstone and limestone samples undergo anisotropic length variations during wetting and drying, and may even experience some shrinkage after 20 wetting/drying cycles. This complex behaviour results from water migration within the sample, during both the wetting and the drying, in an attempt to achieve a balanced moisture distribution within the rock. However, the experiments referred to were carried out using oven drying (at 50°C) which results in a redistribution of this water in the rock according to internal thermal and hygrometrical conditions rather than complete removal of the adsorbed water. Thus, a large number of cycles might be necessary in order to induce any weathering effect but this approach may be a good replicate of nature.

The effect of wetting and drying cycles at positive temperatures has been compared with the action of other weathering processes that depend on the presence of water (salt weathering, frost shattering, chemical weathering). While most researchers consider that hydration processes alone are unable to shatter material they do concede that wetting/drying interacts with the other weathering processes and, thereby, contributes to rock failure (Schaffer, 1932). However, Hall (1993) has suggested that as the zone of effectiveness for weathering by wetting and drying is sub-surface, that is within the same zone that frost action, salt weathering and thermal stress operate. If this is the case, then it is possible that the cause of breakdown has been wrongly assigned in some studies. Hydration as a mechanical weathering process has been considered either less effective than frost shattering (McGreevy, 1982) or as potentially as effective as frost action (Dunn & Hudec, 1972; White, 1976; Fahey & Dagesse, 1984). The real significance is that, the combination of these two weathering processes is recognised as more effective than each process taken separately (Mugridge & Young, 1982). Thus, in the wetter cold regions, or zones of frequent water availability in drier zones (e.g. below snowpatches or alongside perennial water bodies), the operation of wetting and drying and frost action may be a particularly effective combination.

Adsorption of liquid water or of vapour on the rock pores surface has been assessed to be more destructive than frost itself for shales and clay-rich limestones tested between -20 and -40°C (Dunn & Hudec, 1972). According to these results, in cold climates, some carbonate rocks could be weathered by repeated adsorptions and desorptions of water molecules on the clay mineral surfaces during periods of wetting and drying cycles, because of repeated swelling and shrinking. Fahey & Dagesse (1984) confirmed these results by testing more clay-rich limestones and conclude that the fatigue induced by adsorption within these rocks is as effective as microgelivation for the production of fine materials on debris slopes in periglacial environments. For unsaturated fine grained rocks, frost action could be negligible in comparison with wetting/drying effect.

The significance of ice segregation processes and unfrozen water migration during freezing of unsaturated porous rocks has been recently highlighted (Walder & Hallet, 1986; Matsuoka, 1988; Prick, *et al.,* 1993; Prick, 1996 & 1999) but was first articulated by White (1976) in his paper "Is frost action really only hydration shattering?" In that paper, White (1976, p. 3) suggests that the large tensional forces generated by adsorbed water (up to 2000 kg cm^{-2}) are sufficient to break down rocks and form the block fields of Arctic and mountain areas without the need for ice formation. However, the question of liquid water availability in cold environments is often raised as an impediment to the possible effectiveness of hydration weathering. In answer to this, in dilatometry experiments the largest stresses were produced by only slight moisture content variations (Hamès *et al.,* 1987). Moreover, as discussed elsewhere in this paper, freezing cycles measured in the air need not be reflected in rock temperature (i.e. the rocks may not reach freezing point) and, moreover, adsorbed water has been shown to still be liquid in pores and cracks at temperatures much lower than 0°C (Sidebottom & Litvan, 1971; Weiss, 1992; Remy, 1993). Thus, the idea of unfrozen water being present within the rocks and available for migration, phase changes, and adsorption / desorption cycles all must be recognised as being likely in many more cases than has hitherto been considered. Further, liquid water availability can be observed temporally and spatially, i.e. during snowmelt. Water distribution has been measured in the spring in sound limestone and dolomite blocks lying on dry soil or in contact with a melting snow patch in a continental mountain climate (e.g. in the Front Range of the Canadian Rocky Mountains by Harris & Prick, 1997). Very high moisture contents were observed in the superficial part of the blocks where they are in contact with a water source; this water can migrate inwards at a rate that is, partly, a function of rock permeability.

C) Other Mechanical Weathering Processes

As discussed elsewhere in this paper, the presence of shattered rocks or angular fragments in cold regions does not necessarily indicate the predominance of freeze-thaw weathering, but may rather reflect another physical weathering process or combination of processes. The identification of the responsible process(es) thus requires evidence other than form. One line of direct evidence can be obtained from field monitoring of dilatation or cracking in bedrock while also monitoring contributing environmental factors. Such an approach has been applied in recent years with regard to freeze-thaw weathering.

Widening of rock joints has been monitored by means of an "extensometer" fixed to a rock face (Matsuoka *et al.*, 1997; Košták *et al.*, 1998; Matsuoka, 2001). Significant widening occurred in some joints subject to abundant rain or snowmelt water during freeze-thaw periods. For instance, a rock joint in the Japanese Alps experienced frequent widening events in autumn, concurrent with short-term (diurnal or cyclonic) temperature oscillations across 0°C (Matsuoka, 2001). A large part of individual widening was recovered upon subsequent thawing, but the repetition of these widening events resulted in inter-annual, permanent widening. These results demonstrate that frost wedging is active in jointed bedrock situated in an optimal moisture regime and subjected to at least shallow freezing. The monitoring technique is designed to detect the effect of freeze-thaw action, but it is applicable to the evaluation of other weathering processes by using short record intervals (for thermal fatigue/shock) or by recording in combination with rock moisture (for hydration shattering).

Dilatation of frozen bedrock was also monitored at Jungfraujoch (Swiss Alps) during an assessment of the influence of construction activities on the thermal regime of bedrock permafrost (Wegmann, 1998; Wegmann and Keusen, 1998). Several thermistors and extensometers were installed in boreholes to a depth of about 10 m below the rock surface. The temperature of the uppermost part of permafrost rose to the melting point during summer and fell again to a few degrees below the melting point during winter. Correspondingly, annual cycles of 'summer contraction' and 'winter expansion' of bedrock were recorded within permafrost, probably in response to partial melting and refreezing. As a result of these annual cycles, permanent expansion of bedrock within the permafrost took place; Wegmann (1998) attributed this to the opening of several joints across the borehole.

D) Weathering with depth

As these field measurements indicate, the depth at which rock breaks is a function of the type (diurnal vs annual) of freeze-thaw cycles. The corollary is that the type of freeze-thaw cycle influences both the timing of rockfall and the size of fragments produced (Matsuoka *et al.*, 1998). Monitoring of rock temperature at different depths suggests that short-term (diurnal and cyclonic) freeze-thaw action prevails within the top 20 cm of bedrock but rarely reaches 50 cm or deeper (e.g. Matsuoka, 1994; Matsuoka *et al.*, 1997). Such short-term cycles are probably responsible for producing pebbles or finer debris. These cycles are effective where temperature oscillations across freezing point occur in the zone close to the rock surface for a short period of a year; even in regions well beyond the normally identified 'periglacial zone'.

The depth reached by annual freeze-thaw action is spatially variable, depending on several factors including the freezing (or thawing) index at the rock surface, moisture distribution in the rock, thermal conductivity of the rock and direction of freezing (downward or upward). The depth is estimated to be commonly 1-2 m and as large as 5 m near the border between permafrost and non-permafrost regions (Matsuoka *et al.*, 1998); recent PACE results suggest depths may, at some locations, reach to 20 m. The definition of the temperature range for active cracking (cf. Anderson, 1998) also affects the perceived zone of activity. This estimation of active depth means that annual freeze-thaw action can release boulders or larger blocks from steep rock slopes; in fact, such rockfalls have often been observed during thawing periods in a number of cold regions (e.g. Rapp, 1960; Church *et al.*, 1979). Because thaw penetration is time-dependent, the major rockfall activity lagged by 1-2 weeks behind the start of thaw at the rock surface (Douglas, *et al.*, 1992; Matsuoka and Sakai, 1999). The occurrence of metre-scale rock detachment by annual freeze-thaw action is likely to be characteristic of many periglacial zones.

Where permafrost is present, downward migration of unfrozen water in the active layer can induce ice segregation near the permafrost table, both in the base of the active layer and the top of permafrost. This process, which is well recognized in soil freezing (e.g. Chen, 1983; Mackay, 1983), is likely to be significant also in rock weathering (Murton *et al.*, 2000). The products of weathering may differ between the active layer and permafrost. Weathering profiles in the English chalkland, for instance, show that fine debris prevails in the former active layer, which experienced the repetitive freezing and thawing, possibly in association with other weathering processes. In contrast, the formation of thick ice lenses reflecting long-term ice segregation would have resulted in brecciation to blocks a few centimetres in diameter in the uppermost part of the former permafrost (Murton, 1996). Ice-rich layers within bedrock permafrost are actually observed in the present-day continuous permafrost regions (e.g. French *et al.*, 1986; Mackay, 1999). A recent laboratory study simulating two-sided freezing of chalk has showed that frost heave takes place during both freezing and thawing phases; the heave amount being larger during the latter phase (Murton *et al.*, 2000). Akagawa (1993) also reported significant frost heave during thawing of porous tuff. These results imply that ice segregation near the permafrost table occurs during summer thawing and autumn freeze-back periods (cf. Mackay, 1983).

This new recognition of ice-rich layers in bedrock permafrost is of great significance with respect to ongoing global warming. Melting of ice-rich layers at the top (and possibly bottom) of permafrost would destabilize rockwalls in mountain permafrost regions, possibly releasing rock blocks to a depth of a few metres to as much as a hundred metres (Dramis *et al.*, 1995; Haeberli *et al.*, 1997). Such a large rockfall may be a serious hazard to human activities in alpine valleys (Haeberli, 1992). Numerical modelling of thermal fields in bedrock is useful for predicting near-future destabilization of permafrost slopes (Wegmann *et al.*, 1998; Wegmann and Keusen, 1998).

E) Thermal stress/shock

An additional component of mechanical weathering, namely thermal stress/shock, stems from the juxtaposition of cold air conditions, but heated rock. Several recent studies (Hall, 1997a, 1998, 2000; Hall and André, 2001) have begun to investigate the role of thermal stress fatigue by integration of both temperature data and thermal crack patterns (see Figs. 1-3 of Hall, 2000). Such ideas are not actually "new" as workers such as Bartlett (1832), Branner (1896) and Merrill (1906) clearly showed the impact of thermal stresses, with Merrill (1906, p. 232) even noting the process' particular prevalence in cold regions! However, early interest in thermal stress was damped by the negative findings of Griggs (1936) and Blackwelder (1925, 1933) and investigation of this process subsequently waned (see Hall, 2000 for a discussion). Nevertheless, the fracture patterns observed in many rocks at a number of polar or alpine sites show very specific fracture patterns, for example, fractures normal to each other (Fig. 2), a crack hierarchy much akin to 'ordered rivers', and cross-cutting of pre-existing lines of weakness, bedding, jointing or even minerals by cracks. Such patterns show enormous similarity to those identified by thermal fatigue/shock studies within the ceramics industry (see Bahr, *et al.,* 1986; Buessem and Bush, 1955). Convergence of fracture pattern, ceramic studies, and temperature measurements in the field led Hall and André (2001) to argue that thermal stresses can be a major cause of rock breakdown in cold environments. Indeed, cold environments are likely more efficacious than hot ones, as tensile stresses are twice as great during a cooling event than during a heating one (Marovelli, *et al.,* 1966). Considering the alleged lack of moisture, or at least the great spatial and temporal variability in its availability in many cold environments (French and Guiligimin, 1999; Hall, 1993; Humlum, 1992), mechanisms not dependent upon water are required to explain breakdown in some environments: in such circumstances thermal stress merits serious investigation. Meaningful examination of thermal stress necessitates creation of detailed rock temperature data sets; a task that geomorphologists have rarely undertaken. Indeed, it seems appropriate to place the onus squarely on geomorphologists because of the very extensive engineering literature that shows a wealth of information regarding the role of thermal stresses as causal agents in pavement/road breakdown in cold regions (e.g. Janoo *et al.,* 1993; Joseph, *et al.*, 1987; Hunt, 1972; Jokela, 1983;Kanerva, *et al.,* 1992; Chantelois *et al.*,1997). Scrutiny of the pavement literature also shows great similarity in both microscopic (see Hussein *et al.,*1998, Fig.7) and macroscopic (Zubeck *et al.,*1996, Fig. 2) crack patterns with those identified by Hall (2000) and Hall and André (2001) on natural bedrock surfaces, as well as with those found in the ceramics industry (e.g. Bahr, *et al.,* 1986). Given the willingness of geomorphologists to import fundamental concepts from other disciplines to bolster development of a well-founded theoretical underpinning of freeze-thaw weathering, the omission of similar behaviour with respect to thermal stress research appears neglectful, if not perverse.

The points in the preceding paragraph may be sharpened using data from within the discipline. Concepts and data in this instance come especially from the work of, Soleilhavoup (1977,1978) and Bertouille *et al.* (1979), and to a lesser degree from Royer-Carfagni (1999). Soleilhavoup (1977) created a classification of boulder and pebble fracture typology. He generated a 'genetic and petrological' classification based upon the interaction of crack shape, physical and chemical characteristics of the rock, and climatic impact (Soleilhavoup, 1997). Comparison of Soleilhavoup (1977, Fig. 10, or Figs. 36 & 37) and Hall (2000, Fig. 1) is truly instructive. Remarkably, Soleilhavoup's work is all from the Sahara desert! Mendacity aside, the match between the desert and cold region fracture patterns is undeniable. In fact, Bertouille *et al.*(1979) later extended Soleilhavoup's classification to other climates while retaining 'thermoclasty' (thermal stress fatigue/shock) as the main cause of fracture. Royer-Carfagni (1999, Fig. 2) showed micrographs of marble that has experienced thermal cycling and this too shows fracture patterns comparable to those of Soleilhavoup (1977) or Bahr *et al.* (1986). Royer-Carfagni (1999) extended the arguments regarding thermal effects to that of the grain to grain contact as minerals can have a $\Delta L/°C$ that varies by axis (e.g. $\Delta L/°C$ for calcite is positive parallel to the c-axis but is negative normal to the c-axis: Johnson and Parsons, 1944). Thus, grains that fit well together may no longer do so when temperature rises and this anisotropic effect, although the same as that produced under high stress (for the minerals may still be anisotropic with respect to stress) requires only "...a few degrees temperature increase" to produce the same strains (Royer-Carfagni, 1999, p. 119). Royer-Carfagni (1999, p. 123) also showed that in a granular rock such as marble, stress concentrations are located at grain boundaries as a result of 'uniform temperature changes being most critical at points at the intergranular interface'. Thus, thermal fluctuations can explain both the major fracturing of the rock as well as its further disintegration into grus: - a necessary combination when considering the bigger picture which must embrace the complete breakdown of bedrock to its finest components.

Thus, a critical question may be legitimately posed - 'Are cold deserts any different from hot deserts in terms of weathering regime?' If an answer embracing thermal stress is attempted then the simple answer is 'no'; but the qualified answer is 'yes', with the unexpected coda that 'cold' deserts are, in fact, even more dynamic regimes than are 'hot' deserts. Indeed, the closer ones looks, the more one is impressed by the similarity of deserts (a fundamentally moisture-related concept) than one is by the differences reflected in the words 'cold' and 'hot'. In his discussion of 'weathering processes and forms' in desert environments Smith (1994, p. 40) lists six issues identified by Cooke *et al.* (1982) as a guide to understanding weathering in deserts, of which two are pertinent:

"(a) Weathering processes (and presumably forms) are likely to be distinctive because of distinctive and seasonal temperature and relative humidity regimes.....

(d) Physical processes are probably significantly more important than elsewhere, but the role of chemical should not be ignored."

Does this differ from the broad general perception of weathering in 'cold regions'? Smith (1994, p.40) adds the caveat, 'Laudable as these observations are, they are, like all generalizations, open to question and qualification. It is these qualifications that will identify directions of future research.' Smith (1994, p.55) then adjudges those directions to include 'In particular, it is clear that the weathering environments...are numerous and varied. We need much more information on the range of environmental conditions experienced before rock responses can be fully understood. Moreover, the distinction must be made between conditions as measured for meteorological purposes and the temperature and moisture regimes experienced at the rock-atmosphere interface.' Sentiments worthy of any periglacial geomorphologist! Lastly, it is worth highlighting Smith's integrated approach to desert weathering, he emphasized (Smith, 1994, p. 56) the need to understand different weathering mechanisms acting simultaneously and in sequence: 'These mechanisms should include pressure release, limited but varied chemical alteration, and temperature variations - alone and in the presence of moisture and salts - which combine frost weathering with other mechanisms and diurnal cycles with shorter-term fluctuations.....most breakdown occurs in conjunction with thermally induced expansion and contraction.' This, and the other parts of the discussion by Smith (1994, p.56-57) are *identical* to the points raised here. Why then are we so preoccupied with our 'cold' or 'hot' prefixes and blind to the shackles they impose upon us? Why are we not at least equally preoccupied with the aridity or wetness of periglacial regimes, not only at macroscales, but also at meso- and microscales?

F) Angularity (or not) of weathering products and landforms

Lastly, let us return to the preoccupation of weathering form. In most considerations of weathering in polar or alpine environments, present or past, the argument generally is that the material will be angular. This argument is used to the extent that, in Quaternary studies, angular clasts are used to exemplify a former frost climate when found in sediments (e.g. Hanvey and Lewis, 1991, p. 35). However, this argument is frequently both circular and a self-fulfilling prophecy as it is only used when it is known that the former climate was indeed cold. This is important, for as Thorn (1992) has stated, there are no characteristics that actually indicate a clast was produced by frost action. With the exception of Ballantyne and Harris (1994, p. 171) and (Berrisford, 1992) arguments are almost exclusively that rocks weathered in cold regions will be angular and thus the finding of angular clasts in (known) former cold environments indicates the previous activity of frost action. Cooke and Warren (1973, p. 56) in their discussion regarding weathering in (hot) deserts, state that 'Debris resulting from spalling, splitting and flaking is characteristically angular and coarse...' and so, again, we see the convergence between hot and cold environments in terms of resulting form although causative mechanisms may differ in perceived efficiency (see Cooke and Warren, 1973 or Abrahams and Parsons, 1994 for discussion). Recognition of the reality that hot desert weathering generates products of comparable form to those found in cold deserts undermines the oft-cited one-to-one relationship between angularity and frost action. Most importantly it undermines the inversion of cause and effect that permits the widely used casual empiricism in the field that claims angularity (effect) as proof of freeze-thaw weathering (cause). Interestingly, consideration of hot desert texts indicates a greater flexibility in terms of potential candidates (frost action, thermal stress, salt, wetting and drying, chemical processes, and biological weathering) for rock breakdown than do most cold region texts. An appropriate question, for both cold and hot environments, might well be 'Must mechanically weathered material be angular in form, or is it merely possible that it may be angular?'

To anyone who has experienced high mountain or high latitude environments it is clear that the overwhelming sense is that rock debris is angular, often extremely so. Thus, the issue is not to deny this but to ask if it is exclusively so? Ballantyne and Harris (1994, Fig. 9.9 and p. 171) show a photograph of a rounded tor in the Cairngorm Mountains resulting from cold region weathering, and also discuss the rounding of clasts found in Scottish mountains. Whilst discussion could ensue regarding the actual mechanism(s) causing this rounding, the issue is not so much the cause (clearly identified as cold-based mechanical weathering) as the resulting form - rounded. Ballantyne and Harris (1994, Fig. 9.11) show both angular and rounded sandstone 'frost-weathered debris' - the angular showing the anticipated form but the rounded clearly at odds with this general perception. Hall (1997a, p. 88 and Hall, 1997b, Fig. 7) shows a photo of rounded sandstone outcrops in Antarctica plus discusses the observation of highly rounded light-coloured sandstone next to highly angular dark-coloured sandstone (Fig. 3). The point is made that the light-coloured sandstone exhibits not only rounded clasts but also rounding on bedrock outcrops whilst, literally, centimetres away the dark coloured sandstone is highly angular and yet weathering 'potential' (i.e the available mechanisms) are identical. As Hall (1995, p. 82) comments regarding these rounded clasts, '...how would the rounded clasts have been interpreted....if found in a Quaternary sediment?' In other words, we have a 'filter' that would accept angular debris but would, it is here argued, look for an alternative explanation to mechanical weathering under a cold climate for the rounded debris. If this is so, how often do we 'miss' the rounded in our observations or mis-interpret them when found in Quaternary (or older) sediments?

The paradox that may show some light on our perceptions is that of taffoni (also spelt 'tafoni': a "Corsican dialect term for honeycomb cavities developed in the vertical faces of crystalline rocks.." Termier and Termier, 1963, p. 414) and their recognition within cold environments, coupled with the general perceptions of the role of mechanical weathering in their formation. Why a paradox? Mainly because there has never been any objection to, or questions raised regarding, the roundness of form observed in taffoni being derived from mechanical weathering processes. Consideration of any photographs of taffoni indicates that they are far from angular in form and yet, irrespective of climate, mechanical weathering processes of salt, frost, wetting and drying and thermal attributes (see Abrahams and Parsons, 1994, French, 1996) are all considered as major agents in their formation. Thus, if such rounded forms as taffoni, in a variety of rock types, can occur, then why should rounded clasts not be equally likely? Again, this is not to question the common occurrence of angular clasts but rather to instigate a reconsideration of our basic tenet regarding the universal or invariant form of mechanically weathered material.

G) The influence of lithology and of inherited forms

Two issues arise that suitably complete this section. First, the argument that the rounded attributes are linked to lithology and hence to process, and, second, inheritance of features. Clearly lithology can exert a strong influence on the processes that may take place, particularly that granular disintegration is likely a result where such as fine-grained sandstones occur, and that this can result in a rounded form. Thus, rounding can indeed be due to mechanical processes. This, however, misses the overall point that the 'general perception' in cold regions is "mechanical weathering equates to angularity"; any recourse to lithologic exceptions avoids the general argument. In addition, the application of lithology hides within it a multitude of problems, notably why is it that *only* granular disintegration occurs where such lithologically-appropriate outcrops occur; why should other forms of fracture not still take place to produce larger, angular forms? Indeed, one of the authors (KH) has observed extensive flaking, rather than granular disintegration, in developing taffoni in the Antarctic and thus the question arises as to how this creates a rounded rather than angular form, or is it that, once developed, the form influences the character of the weathering? Thus, recourse to lithology can help answer some problems but, in doing so, can also raise other questions. However, the attribute of form influencing weathering, leads to the second of the major issues, that of inherited features. This is an issue that is likely dealt with to some extent in the discussion regarding blockfields (van Steijn *et al.*, this series). However, it is pertinent to raise the concern here. Simplistically, landforms found in cold regions are assumed to be the product of that environment. Recently, more and more, features (e.g. cryoplanation forms as argued in Hall, 1998b) are being identified that must be of an age that indicates they have experienced a range of climates, from (perhaps) glacial to temperate during their time of existence (see the wider discussion on this issue by André, 1999). With this in mind, the problem arises as to the inheritance of features that owe little or nothing to the present environment: tors and blockfields being possible classic examples. In terms of form, chemical processes may have helped create a form under a preceding climate that has yet to be modified or, as per the argument above with respect to taffoni, the form controls the subsequent character of weathering and thus 'maintains' itself. Clearly these attributes of lithology and antecedence are complex issues that require more extensive argument than can be dealt with here, but suffice to say that both can exert an (perhaps as-of-yet unrecognised) impact on present-day weathering and resulting forms in cold regions.

Biological Weathering

While biological weathering has been identified in a variety of cold climate environments (e.g. Webley, *et al.,* 1963; Friedmann, 1982; Walton, 1985; Hall and Otte, 1990; Weed and Norton, 1991; Walton, 1993) it has clearly yet to receive the level of attention it merits or has received in other environments. Organisms involved range from bacteria through to mosses (see Walton, 1993), and weather by exerting physical stress and/or through chemical processes such as oxidation, reduction or chelation (see Viles, 1988; Vincent, 1988). In extreme cold environments organisms frequently use sub-surface rock sites as a locations of environmental protection - from extreme cold, high rock surface temperatures, wind, desiccation and/or UV radiation. Chasmolithic (in fissure or cracks of rocks), sublithic (undersurfaces of rocks), or cryptoendolithic (in the internal air spaces between crystals) environments are themselves limiting on life by the very shortage of physical space that they impose (Vincent, 1988) and are also constrained by rock type. Not only must the rock have the available space (i.e some lithologies are too pore-deficient) but also they must have some degree of light transmisivity (which also constrains lithology). In high polar environments, aspect exerts an influence as the life forms require the heat associated with the equator-facing, or at least locally sun-exposed (i.e based on local topography that may influence exposure to the sun), side of rock outcrops (Vincent, 1988). Other sites may be more influenced by moisture availability or the impact of predominant winds and thus spatial variability may be the result of multiple factors. Whatever the cause of distribution, at the sites of activity the organisms can effect weathering - especially in terms of deriving inorganic nutrients from the rock substrate (Friedmann, 1982). Where lichens can survive there is not only the physical stresses induced at the rock surface but also the mobilisation of various minerals such as ferric iron, alumina, silica and phosphorous (Walton, 1985). Indeed, Weed and Norton (1991) have shown that organisms can cause exudation of oxalate, the formation of thin siliceous crusts, and the silicification of porous quartz sandstones. Thus, the results of organic activity, in terms of chemical processes, can both remove minerals as well as precipitate minerals and, in so doing, change the rock properties (Fig. 4). The creation of impermeable coatings (up to several centimetres thick) may have important ramifications on other, non-biologic, water-driven weathering at these sites (Arocena and

Hall, Subm.). All in all this is a complex attribute of weathering in cold environments that has not yet been fully integrated into weathering studies. Here attention is simply drawn to the possible role of micro-organisms in rock breakdown and/or influencing other weathering processes (e.g. even the colonisation of a rock surface by a dark lichen could change the albedo such that the rock thermal regime is altered). It is quite clear that weathering studies have reached a point where periglacial geomorphologists must give greater cognizance to the role of organisms throughout the range of cold environments and their effects integrated, in both space and time, with that of the more "conventional" weathering processes under consideration.

Conclusion

Interestingly, the whole issue regarding weathering in cold environments might be summed up by the concluding remarks of Smith (1994, p. 56) addressing the issues confronting desert geomorphologists. Smith suggests that 'There has been a long-term tendency in studies of desert weathering to isolate individual weathering mechanisms and to see understanding of weathering phenomena through segregation rather than integration' and yet '...ultimately such mechanisms must be set within wide temporal and spatial contexts to understand the landscapes within which weathering occurs'. Smith goes on to suggest four attributes that need to be dealt with to facilitate this integration, those same four would serve periglacial studies equally well (p. 56-57):

> integration must occur:
>
> a) Between different mechanisms acting simultaneously and in sequence.
>
> b) Between observations of how mechanisms operate and the landforms they produce.
>
> c) Between weathering landforms and landscape, and
>
> d) Between field and laboratory studies.

Clearly, periglacial weathering environments are highly varied and cognizance must be given to this, some are cold and dry, others wet and cool, and many have experienced a variation through the Quaternary, all of which will be reflected in process activity and landscape evolution. Continued use of the term 'periglacial' in an unqualified manner obfuscates rather than clarifies many of the weathering issues that are now being addressed. In this context, a further comment by Smith (1994, p. 55) is highly pertinent, 'Moreover, the distinction must be made between conditions as measured for meteorological purposes and the temperature and moisture regimes experienced at the rock-atmosphere interface'. As has been demonstrated, once interrogation of rock field conditions is undertaken then the character of periglacial regimes, in terms of weathering, becomes less zonal. Bearing this reality in mind, periglacial geomorphologists must approach investigation of periglacial weathering regimes with an open mind , indeed with the objective of determining what is present, rather than with the purpose of vindicating their preconceived notions. To come full circle, only by embracing McCarroll's (1997) sentiments and Platt's (1964) principle of 'strong inference' will we improve our understanding of periglacial weathering - in short we need to challenge our theory, not admire it nor seek data to fit it.

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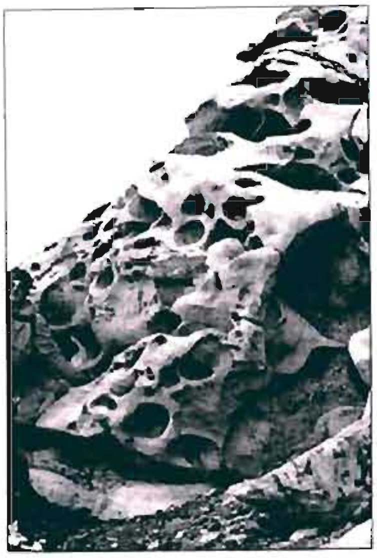


Fig. 1: Taffoni development on a sandstone outcrop on Alexander Island, Antarctica. Note the highly rounded nature of the forms.

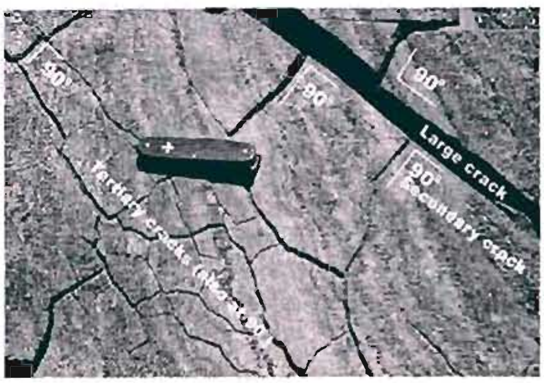
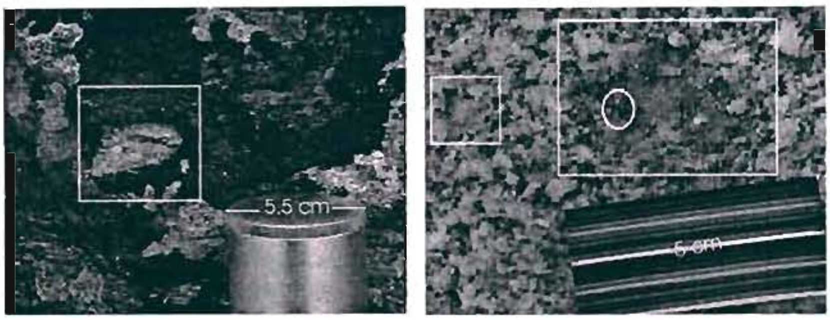


Fig. 2: Crack patterns developed in rock by thermal stresses in the high Andes. Note the crack hierarchy, that the cracks are at (or close to) 90°, and that the cracks cut across the bedding within the rock.

Fig. 3: Light and dark coloured sandstones on the top surface of a nunatak in Antarctica. Both rocks are *in situ*, very close to the lithologic junction between the two sandstones, and are exposed to the same general environmental conditions. The light-coloured sandstone has weathered to rounded forms while the dark-coloured has resulted in angular forms



A B

Fig. 4: Two pictures of lichen growth on rocks in a permafrost area at 4900 m in Tibet. In (A) the area in the box shows where material has flaked away beneath (the now lost) lichen cover. In (B) the circle is around an area that, in colour is green, identifies now exposed algae that was growing beneath the surface of the rock. The square boxes identify areas where the rock has flaked away as a result of algal expansion and contraction, and which now shows a chemically affected area (as a result of biolo-chemical activity).

Chapter 5

Weathering Associations with Landforms

"It is a capital mistake to theorize before one has data. Insensibly one begins to twist facts to suite theories, instead of theories to suit facts."
Sir Arthur Conan Doyle

The perusal of almost any text dealing with landforms of cold (periglacial) regions will indicate the overwhelming use of freeze-thaw weathering as the basis for landform development and evolution. This is the case whether the discussion be either a 'general' one or very landform specific. A typical example of the former might be that of French (in French and Slaymaker, 1993, p. 148), where in a discussion regarding Canada's cold environments, he states "The most influential bedrock-weathering process has generally been assumed to be frost wedging, a physical process relying on repeated freeze-thaw activity. Features attributed to frost wedging in northern Canada (Fig. 7) include extensive areas of angular bedrock fragments (blockfields), scree and talus deposits, and irregular bedrock outcrops termed tors". With respect to more detailed texts on specific landforms, the discussion by Priesnitz (1988) regarding cryoplanation is fairly typical where he states (p. 56) "Rock disintegration seems to result mainly from frost weathering".

The same is true of many papers where it is simply stated, but untested, that weathering is due to frost action. For example, Ballantyne, *et. al.* (1997), in a discussion regarding former nunataks in northwest Scotland, state that there is "...a thick cover of frost debris..." (p. 228), "...frost shattered terrain..." (p. 230), and (p. 233) a "...landscape dominated by frost-shattered rock and frost debris". No actual unequivocal "evidence" that frost action was the cause is presented, rather it is just "stated", and thus assumed. Whether this has any bearing on the reconstruction within the cited paper is not significant here. The point *is* that in some instances different cryogenic weathering processes may indeed impact upon the final hypothesis/reconstruction. There is thus a danger in accepting as the central tenet of secondary hypotheses the premiss that the causative process was, or is, freeze-thaw weathering. For example, mechanical processes *other* than freeze-thaw can cause angular rocks in a cold, dry climate. Assumption of freeze-thaw in such a palaeoreconstruction may introduce more snow/moisture than was the case and so the whole scenario is in error. Equally, the production of *rounded* weathered material by cold-based weathering processes, including possibly freeze-thaw weathering, would *not* have been recognised as such and so, again, the wrong reconstruction generated.

The overall problem is simply that our answers are generated by our question - if the question is in error then so too are the answers. If our question is (simplistically):

"What is the effect of freeze-thaw on landform development?",

which it intrinsically is if we assume freeze-thaw, then clearly we will "find" evidence to answer the question and, in so doing, support the assumption. If, however, our question was:

"what are the weathering processes affecting landform development?"

then we may well find a different answer, as no assumptions are being made beyond that of the expectation that weathering *is* playing a role.

Fig. 7

Scree slopes in the Canadian Rocky Mountains. The material comprising the scree is angular and usually considered to be the result of freeze-thaw weathering

The same is the case with respect to the assumption of angular clasts. If our questions were to become:

"What processes are causing the angular clasts?"

and/or

"Are, in fact, only angular clasts being produced by mechanical weathering in cold regions?"

then the answers may be significantly different to those produced by the assumption of angularity due to frost action. This becomes all the more significant when dealing with Quaternary sediments where angularity is used as a proxy for the former occurrence of frost action.

Papers that are to be considered within this chapter include:

- Arocena, J. and Hall, K. Subm. Calcium Phosphate Accumulation on the Yalour Islands (Antarctica): Surface Coatings and Geomorphic Implications. *Arctic, Antarctic and Alpine Research.*
- Bockheim, J. and Hall, K. In Press. Periglacial processes and landforms of the Antarctic continent: A review. *South African Journal of Science.*
- Boelhouwers, J. and Hall, K. In Press. Periglacial and permafrost landforms and processes of the Southern hemisphere: A Synthesis. *South African Journal of Science.*
- Corté, A.E. and Hall, K.J. 1991. Geocryology of the Americas (IGCP Project No 297): Introduction. *Permafrost and Periglacial Processes*, 2, 3.
- Grab, S. and Hall, K. 1996. North-facing hollows in the Lesotho/Drakensberg mountains: hypothetical palaeoenvironmental reconstructions? *South African Journal of Science*, 92, 183-184.
- Hall, K.J. 1973a. An investigation of nivation processes at a late-lying snow-patch in Austre Okstindbredalen, *In* R.B.Parry & P.Worsley (eds): *Okstindan Research Project, Preliminary Report for 1972,* 23-24.
- Hall, K. 1991b. The significance of periglacial geomorphology in southern Africa: A discussion. *South African Geographer*, 18, 134-137.
- Hall, K. 1992a. Mechanical weathering in the Antarctic: A maritime perspective, *In* J. Dixon and A. Abrahams (eds): *Periglacial Geomorphology.* Binghampton Series in Geomorphology, John Wiley, 103-123.
- Hall, K. 1992b. Mechanical weathering on Livingston Island, South Shetland Islands, Antarctica, *In* Y. Yoshida, K. Kaminuma and K. Shiraishi (eds): *Recent Progress in Antarctic Earth Science.* Terra Scientific Publishing, Tokyo, 757-762.
- Hall, K. 1993c. Enhanced bedrock weathering in association with late-lying

snowpatches: evidence from Livingston Island, Antarctica. *Earth Surface Processes and Landforms*, 18, 121-129.

- Hall, K. 1994a. Cutbacks in the Natal Drakensberg - an hypothesis on their origin. *South African Journal of Science*, 90, 263-264.
- Hall, K. 1995b. Cutbacks in the Natal Drakensberg: Some further comments. *South African Journal of Science*, 91, 285-286.
- Hall, K. 1995c. Cryoplanation terraces: Some new information. *The 25th Arctic Workshop.* Centre d'Études Nordiques, Université Laval, 71-73.
- Hall, K. 1996a. "Mountain benches" or "cryoplanation"?: A tautological conundrum. *Abstracts, 28th International Geographical Congress, The Hague*, 170.
- Hall, K.1997c. Observations regarding "cryoplanation" benches in Antarctica. *Antarctic Science*, 9,181-187.
- Hall, K. 1997d. Did South Africa really look like this during the Quaternary? Questions regarding some recently suggested periglacial landforms and sediments. *Abstracts, Southern African Society for Quaternary Research, XIII Biennial Conference*, 5.
- Hall, K.1998b. Nivation or cryoplanation: different terms, same features? *Polar Geography*, 22, 1-16.
- Hall, K. 1998c. Some observations and thoughts regarding Antarctic cryogenic weathering. *Abstracts of the 7th International Conference on Permafrost*, Yellowknife, Canada, 151-152.
- Hall, K. 1998d. Nivation or cryoplanation: Is there a difference? *Abstracts of the 7th International Conference on Permafrost*, Yellowknife, Canada, 149-150.
- Hall, K. Subm a. Periglacial landforms and processes: Southern Alexander Island, Antarctica. *Permafrost and Periglacial Processes.*
- Hall, K.In Press' c. Present and Quaternary periglacial processes and

landforms of the Maritime and sub-Antarctic: A Review. *South African Journal of Science.*

- Hall, K. and Bühmann, D. 1989. Palaeoenvironmental reconstruction from redeposited weathered clasts in the CIROS-1 drill core. *Antarctic Science*, 1, 235-238.
- Hall, K. and Meiklejohn, I. 1997. Some observations regarding protalus ramparts. *Permafrost and Periglacial Processes*, 8, 245-249.
- Hall, K. and Walton, D.W.H. 1992*. Rock weathering, soil development and colonisation under a changing climate. *Philosophical Transactions of the Royal Society*, B, 338, 269-277.
- Hall, K., Arocena, J. and Smellie, J. In Press a. Analysis of weathering rinds and reconstruction of palaeoenvironmental conditions from weathering rinds on clasts in the Cape Roberts drillcore. *Terra Antartica.*
- Lautridou, J-P., Francou, B. and Hall, K. 1992. Present-day periglacial processes and landforms in mountain areas. *Permafrost and Periglacial Processes*, 3, 93-101.
- Siegmund, M. and Hall, K. 2000. A study of valley-side slope asymmetry based on the application of GIS analysis: Alexander Island, Antarctica. *Antarctic Science,* 12, 471-476.
- Smellie, J.L. 1999. Cape Roberts Drilling Project. Personal Communication.
- Thorn, C. and Hall, K.J. 1980. Nivation: an Arctic-Alpine comparison and reappraisal. *Journal of Glaciology*, 25(91), 109-124.
- Thorn, C. and Hall, K. In Press*. Nivation and cryoplanation: The case for scrutiny and Integration. *Progress in Physical Geography.*
- Hall, K., Thorn, C., Matsuoka, N. and Prick, A. In Press. K. Weathering in Cold Regions: Some thoughts and perspectives. *Progress in Physical Geography.* (Provided in Chapter 4, p.412-464)
- Walton, D.W.H. and Hall, K. 1989. Rock weathering and soil formation in

the maritime Antarctic: An integrated study on Signy Island. *Geoöko plus ①*, 1, 310-311.

- Walton, D.W.H. and Hall, K. Unpubl. Rock weathering and soil formation in the maritime Antarctic: An integrated approach for Signy Island. (The detailed text of the above Conference Abstract)

Inevitably there is also information applicable to landforms in some of the papers cited in earlier chapters (e.g. Hall, 1985 in Chapter 2 that deals with nivation landforms) but the main discussions are those given above. Work here encompasses not only the Antarctic but also parts of the Arctic, southern Africa and Canada, as well as mountain areas in general (e.g. Lautridou, Francou and Hall, 1992). Much of the work cited, directly or indirectly (e.g. protalus ramparts), deals with the concepts of nivation and/or cryoplanation. Both are concepts which have as a central tenet the action of weathering, specifically freeze-thaw weathering. However, as Thorn (1988, p. 24) discusses, the problem is one of "... a discipline split between a traditional focus on landforms and a present emphasis on geomorphological processes." The work here has attempted to consider the concepts (nivation and cryoplanation) themselves as well as to derive process data pertaining to the landforms. The outcome of these studies on both nivation and cryoplanation, which provides some of the first process data from such features (see Thorn, 1988), was the new suggestion that rather than 'distinct' landforms the two (nivation and cryoplanation) are end members of the *same* process/landform suite. As both nivation and cryoplanation have been used as palaeoenvironmental indicators (with, simplistically, nivation equating to a snowy environment with frequent freeze-thaw cycles and cryoplanation with a cold, arid climate and the presence of permafrost) so this new approach brings in to question any such reconstructive attempts.

There are, then, a number of papers that deal specifically with landforms in the

Antarctic. The papers cited here are complemented by material given in Chapters 2 and 4. A whole range of landforms are dealt with, many of which are not germane to the discussions here (e.g. patterned ground), although they do pertain to the overall picture of the periglacial region and, in some instances (e.g. the unusual type of patterned ground reported in Hall, In press a), there is a relationship to weathering. Discussions deal with the whole of the Antarctic area (Hall, In Press, Bockheim and Hall, In Press) as well as for the Southern hemisphere in general (Boelhouwers and Hall, In Press). In the context of the Southern hemisphere there are a number of papers that deal with present and past conditions in southern Africa. For southern Africa, particularly the Kingdom of Lesotho and the Drakensberg mountains of South Africa, there are questions regarding the occurrence or not of Quaternary glaciation as well as the character of the present cold-based processes. Due to the physiography, and the warm post-glacial climate that has caused severe weathering of the basalts in this region, no glacial evidence has yet been unequivocally presented. The corollary is that, if the region were not glaciated then what sort of periglacial conditions existed, including the possibility of permafrost, and what features exist to reflect this? Similar questions have arisen for Australia, New Zealand and parts of South America. The material presented here covers some of the questions and problems with respect to Quaternary conditions in southern Africa.

Three joint authorship papers from Antarctic work extend the basic concepts on weathering into a broader perspective. Hall and Bühmann (1989) provide a new approach to investigating past terrestrial environmental conditions. Using X-ray diffraction analysis of weathering rinds from glacially transported, ocean-deposited, clasts it was possible to deduce some idea of the palaeo-weathering conditions occurring on the land. By this means it was possible to gather some knowledge of the climate, via the weathering environment, for an area for which no direct terrestrial evidence is available. This was a completely new technique, applied with some surety, to material collected as part of an ocean drilling programme aimed at investigating the Cenozoic history of the McMurdo region of Antarctica. Although not dealing directly with landforms it does provide a means of investigation that may be applicable to some landform studies. This method of investigating terrestrial condition via an analysis of weathering rinds on clasts deposited off-shore by glaciers has now been extended to a detailed study of the Cape Roberts drill core. In this new undertaking a study is made of a number of clasts from the core to try and derive a picture of the variations in terrestrial conditions (Smellie, pers. comm. 1999; Hall, *et al.,* In Press a). Arocena and Hall (Subm.) is an analysis of surface coatings, produced by penguin guano, on the rocks of the Yalour Islands, Antarctica. In addition to the discussion regarding the origin and nature of these coatings, some discussion is also given to their geomorphic impact: decreasing of water infiltration which, thereby, decreases the potential for all water-based weathering processes but increases surface run-off, the change in albedo that influences the impact of thermal stresses that are driven by solar radiation, and an in crease in hardness that helps protect the rocks from abrasion. Although weathering rinds and, to a lesser extent, surface coatings are known from the Antarctic little has been said about their potential geomorphic impacts and so this paper opens new avenues of study.

Siegmund and Hall (2000) deals with valley asymmetry as deduced from 'Geographic Information Systems - Digital Elevation Models' (GIS/DEM) analysis. Asymmetric valleys are a relatively common feature of periglacial regions and are certainly associated with aspect-controlled weathering (see French, 1996). Here, valley asymmetry is proved, via GIS analysis, and the shallower north-facing slopes explained by weathering differences influenced by aspect-controlled micro-climate. The other joint paper, Hall and Walton (1992), does not deal with landforms specifically. Rather, as part of a wider work on "Antarctica and Environmental Change" (Drewry, *et al.,* 1992), it integrates cold region weathering with soil development and plant colonization to help understand the manner in which some Antarctic environments might adjust as a result of climatic change, particularly global warming. While not dealing with specific landforms it does, nevertheless, relate to landscape development in cold regions and the role weathering, in the broadest sense, plays in this. In a like manner, the unpublished report by Walton and Hall deals with similar aspects of weathering and soil formation but, in this instance, is specific to one maritime Antarctic location (Signy Island).

Thorn and Hall (In Press) represents recognition of the work presented in this Chapter. As in Hall, *et al.* (In Press b) cited in Chapter 4, this paper is at the request of the International Geographical Union's Working Group on Periglacial Processes and Climatic Change and is to constitute a "state of the art" discussion regarding the concepts of nivation and cryoplanation. The contentious issue relating to these two concepts, as outlined in several of the papers cited above, are dealt with and recommendations for future work presented. The paper is noted because it helps identify the recognised significance of some of the research presented.

Thus, here, I present some discussion regarding the association between weathering and landforms, both present and fossil. This association is very important for, as has been shown by various quotes both in this chapter and in Chapter 1, freeze-thaw weathering is almost ubiquitously cited as the primary cause of many periglacial bedrock landforms. As recognition of these landforms is frequently used to deduce palaeoenvironments, if the tenets of landform formation are in error, the reconstruction will be false. The issues in this regard with respect to 'nivation' and 'cryoplanation' exemplify the greater problem. It is suggested that, with more care and concern for empirical data, weathering studies may have more to offer than has frequently been the case to date. With a reassessment of weathering there may be a need for reappraisal of landforms and sediments in present and past cold regions.

1

Calcium Phosphate Accumulation on the Yalour Islands (Antarctica): Surface Coatings and Geomorphic Implications.

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Abstract

Samples of a coating found over extensive areas of andesitic rock on the Yalour Islands, Antarctica, shows a ~25μm thick, white, shiny, relatively hard layer of hydroxylapatite with traces of calcite and quartz. Based on scanning electron micrographs, x-ray diffractograms, Fourier Transform Infrared spectra, *in situ* chemical composition of the coatings. This predominantly calcium phosphate coating is formed mainly through the decomposition of guano from nearby penguin rookeries and subsequent precipitation from solutions containing high amounts of calcium and phosphorus. The change in permeability and albedo of the andesite resulting from the presence of the calcium phosphate coating can have significant moisture and thermal effects in terms of geomorphic processes.

Introduction

Calcium phosphate minerals are known to occur in various environments ranging from lava melts to soils to oceanic floors (Lindsay et al. 1986). Calcium phosphate minerals can also be found as accessory minerals in igneous, sedimentary and metamorphic rocks and are known to accumulate from animal remains as well as from the chemical precipitation of seawater (Klein and Hurlbrut, 1999). Currently, there are over 40 different calcium phosphate minerals identified, due, most probably, to the high affinity of tetrahedral PO_4^{3-} for Ca^{2+} (Lindsay et al. 1986). Some of these minerals are economically important because they are the raw materials in the manufacture of phosphate fertilizer (Rutherford et al. 1995). The formation of calcium phosphates in soil environments with high pH is known to significantly reduce the availability of phosphorus to plants. Calcium phosphate minerals are also the main components of teeth and bones.

In Antarctica, calcium phosphate minerals are known to accumulate in ornithogenic soils derived from guano of krill-eating penguins (Tatur and Kreck, 1990; Tatur and Myrcha 1989; Tatur, 1989). Calcium phosphate minerals form from precipitation of P released from the decomposition of penguin guano and its reaction with Ca. Apatite, $Ca_5(PO_4)_3(F,Cl,OH)$ is the dominant phosphate mineral identified in these soils as well as in the guano layer. Tatur and Kreck (1990) estimated that apatite content in a leached guano layer ranges from 30 to 60%. Keys and Williams (1981) also reported the crystallization of calcium phosphate on penguin or skua rookeries established on andesitic outcrops. Other calcium phosphate minerals identified in Antarctic soils are brushite and $CaHPO_4{*}2H_2O$ (Tatur and Kreck, 1990). In phosphatized rock zones, Tatur and Kreck (1990) identified a number of aluminum phosphate minerals such as leucophosphite, $[K_{0.31}(NH_4)_{1.69}]_{2.0}(Fe_{2.66-3.74}Al_{0.26-}$

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$_{1.34})_4(PO_4)_4(F_{0.31}OH_{0.69})_2{*}4H_2O$, minyulite, $KAl_2(F_{0.82}OH_{0.18})(PO_4)_2{*}4H_2O$, taranakite, $[K_{1.4-1.9}(NH_4)_{1.1-1.6}]_3(Fe_{0-0.6}Al_{4.4-5.0})_5(PO_4)_8{*}4H_2O$ and other amorphous aluminum phosphate minerals. Understanding the properties of phosphate minerals in Antarctica has profound implications for the establishment of vegetation because P is one of the six major elements (the others are N, Ca, Mg, S and K) needed for plant growth. The impact of climatic change on vegetation establishment in cold climates will certainly be dependent on the availability of P in the growing medium.

Another important occurrence of phosphate minerals is in surface coatings of rocks. Surface coatings ('patina', 'rock varnish', or 'desert varnish') are generally associated with arid regions, although, as Summerfield (1991, p. 136) notes, they can also sometimes be found in more humid environments. These coatings may take substantial periods (thousands of years) to form but environments like that here (the Yalour Islands - see below) may be more conducive to development owing to the high moisture input from snow melt and sea spray coupled with extensive periods of aridity (i.e. time of water loss leading to precipitation of the minerals) during times of low air temperatures but high radiation inputs that produce high rock temperatures. In a discussion regarding rock varnish, Oberlander (1994) notes that most varnish constituents are derived from sources external to the rock and that they are characteristically less than 300 µm in depth. Importantly, Oberlander (1994, p. 108) also identifies that most rock varnishes have a clay content of 60 to 80% as well as "...hydrous non-crystalline oxides of Mn, Fe and Si, each of which can attain values of 50% or more". Although work by such as Oberlander

(1994) or Dorn (1994) consider in great depth the origin of the varnish, its use as an environmental indicator, for dating, or as evidence for climatic change, little is overtly discussed regarding the impact the varnish has for subsequent geomorphic processes (e.g. weathering); Merrill (1898) being an exception. Hardness of the varnish may also be useful in protecting the rock from abrasion (Allen, 1978). Although it might be argued that an Antarctic "coating" is quite different from a "desert varnish" this is more semantics than applicable science. First, there is absolutely no reason why they should not respond, geomorphologically, in the same manner should both create a change in hardness, albedo and/or permeability. As Antarctic information on either coatings or varnish is relatively limited, as too is any extensive discussion regarding the impact of such chemical changes on resulting geomorphic processes, it does seem suitable to identify hot, arid desert information here.

The objective of the paper is to elucidate the origin of the calcium phosphate coating found on the andesite on the Yalour Islands, Antarctica. Specifically, we characterized the mineralogical and chemical composition of the calcium phosphate coating. We also suggest the implications for geomorphic processes of the presence of the rock coating because it can influence the character of reflected solar radiation and emitted long-wave radiation (Fortescue (1980) cited in Dorn (1998)).

Study area

Rock samples were collected from the Yalour Islands (65°14'S, 064°10'W), a small group of islands and rocks 2.41 km in area, located to the west of the Antarctic Peninsula (Fig. 1) on the east side of the Penola Straight (Hattersley-Smith, 1991). Extensive searching through Antarctic bibliographies, data bases and libraries failed to elicit almost any scientific information pertaining to these islands (C. Phillips, Pers. Comm., 2001). Charcot (1911) refers to the Jallour Islands (alternate spelling) and that landings were made but no references to geology or life were made; it is noted, though, that the islands were in open water much of the time when other nearby water was still ice covered (Charcot, 1911, p. 61 & 66). Croxall and Kirkwood (1979) provide three references pertaining to penguin counts, with the value of 10,400 nests (Smith, 1958) being considered a minimum overall estimate. Although the local rock is volcanic and exhibits extremely well developed *roches moutonnées* (the shapes of which seem to be mirrored by the penguin colonies, see Fig.1 and, for detail, Fig. 4.1.1 of Croxall and Kirkwood (1979), no petrographic investigations appear to have been undertaken. The geological studies of Curtis (1966, Fig. 8) identify the Yalour Islands ('Jalour Islands' in Fig. 8) but provide no information whatsoever despite their obvious ease of access. From geological studies in the surrounding areas (e.g. Elliot, 1964; Curtis, 1966), coupled with the mineral analysis of the bedrock, it appears that the Yalours' comprise a metamorphosed andesite.

Gremmen et al. (1994) detail the epilithic macrolichen vegetation of this area and show that where concentrations of phosphate are relatively high communities of *Mastodia-Rinodina* are common; they also note that these communities are characteristic of sites strongly influenced by animals, such as close to penguin colonies. Specifically, Gremmen et al (1994, p. 466) identified *Mastodia-Rinodina* complexes over 100 m inland on the Yalour islands, in the areas influenced by penguins. Patches of the Antarctic grass *(Deschampsia antarctica)* were seen by one of the authors

(KH) and are also noted by Gremmen et al. (1994). Climatic data from the nearby Argentine Islands (van Rooy, 1957) indicate a mean annual air temperature of -5.7°C; mean summer (Dec., Jan., Feb.) air temperatures are -0.3°C but with a standard deviation of 2.33. Summer relative humidity values are high, 84-87% and mean relative sunshine is also fairly high (van Rooy, 1957). Given that rock temperatures are a product of radiation input, rather than air temperatures (Hall, 1997; Hall and André, In Press), the relatively high incidences of radiation indicate that summer rock temperatures will be positive with values ≤30°C expected.

Methods

A) Sample collection and X-ray diffraction analysis

Three types of samples were separated from the andesite fragments collected from Yalour Islands for detailed investigation of their mineralogical and chemical composition. These samples are (1) andesite (the predominant rock type in the study area), (2) the white coating, or 'veneer' of white materials, on the surface of andesite, and (3) flakes or patches of translucent whitish materials adhering to rock cavities characterized by the presence of abundant root growth. To minimize contamination from the andesite, samples of white coating used in x-ray diffraction and infrared analyses were separated by microdrill of about 1.0mm in diameter operated at about 10,000 rpm under a dissecting microscope. Samples of flakes were taken by peeling them from the cavities of the rocks under the dissecting microscope using fine forceps and tweezers.

The mineral composition of the andesite and the white coating were determined using a Rigaku X-Ray Diffraction system. Samples of flakes were not analyzed using x-ray diffraction primarily due to their very small quantities. X-ray diffraction analysis was conducted on a sample that was randomly mounted on a silicon sample holder using acetone. The powder mount was scanned from 3 to 90 °2Θ at ambient condition using Cu Kα radiation generated at 50 kV and 150 mA. Collection of the diffractogram was done in a step-scan mode of 0.05 °2Θ and at a dwell time of 2 seconds per step. Identification of the minerals in the samples was based on the following x-ray reflections (in nm): (1) quartz - 0.425, 0.334, 0.245, 0.228 and 0.182, (2) feldspars - 0.635, 0.404-0.420, 0.315-0.3S25, (3) hydroxylapatite - 0.281, 0.278, 0.272 and 0.871, (4) calcite - 0.304, 0.229, 0.210 and 0.386; (5) chlorite - 0.707, 1.41, 0.354 and 0.252, and (6) mica - 0.3.32, 0.995, 0.257 and 0.201. Estimate of the relative amounts of plagioclase feldspars was based on the separation between -132 and 131 and between 111 and 1-11 x-ray reflections as suggested in Huang (1986).

B) Electron microscopy and chemical analysis

The submicroscopic surface morphology of the samples was examined using a Philips XL30 Scanning Electron Microscope equipped with EDAX energy dispersive system (SEM-EDS). Samples of the white coating, andesitic fragments and the flakes were mounted onto an aluminum stub using double-sided tape, then sputter-coated with gold (Au) for 60 seconds prior to analysis. Once Au-coated, the morphology of the samples was examined under SEM while the semi-quantitative chemical composition was determined using standardless EDS technique. The chemical composition was conducted on ten sub-areas (5-10 μm^2) in each sample. The mean corrected (Z - atomic number, A -absorption, and F - fluorescence factors) elemental composition was determined in an energy dispersive spectrum collected for 200 seconds from 0 to 20 keV. Due to the near overlap of the Au and phosphorus (P) energy peaks in the SEM-EDS system, the chemical composition of the white coating was conducted on samples without the Au coating. The spatial distribution of Ca, P, Si, Al and C in coating, flake and andesite samples were determined using the x-ray mapping capability of SEM-EDS at a setting of 1024 x 800 matrix with a dwell time of 20 milliseconds.

C) Fourier transform infrared analysis

Infrared analysis of the samples was conducted using Perkin Elmer 200 Fourier Transform Infrared Spectrometer (FTIR). A pellet composed of approximately 1:100 ratio of sample and KBr was prepared by pressing the mixture using a manual press. The KBr pellet was subjected to IR beam for about one minute and the interferrogram was recorded from 400 to 2000 cm^{-1} wavenumber. Identification of the molecular group corresponding to major regions of IR absorption bands (in cm^{-1} wavenumber) was based on the following: (1) Si-O-Si = 950-1080; (2) CO_3^{2-} = 870, 1430 - 1439, (3) PO_4^{3-} = 1037 - 1086. The IR interferrograms were also compared to a library of IR patterns for common minerals (Kodàma 1985).

Results

A) Andesite

Andesitic rocks appear to predominate on the Yalour Islands (Fig. 1a, b) although, as noted above, no detailed studies have been undertaken of the geology. The mineral composition of the andesite consists of plagioclase feldspars (albite (Ab), $(Na,K)Al_3SiO_8$, anorthite (An), $CaAl_2SiO_8$), quartz (Qt), SiO_2; muscovite (Ms), $KAl_2(Si_3Al)O_{10}(OH,F)_2$, and chlorite (Ch), $(Mg,Al,Fe)_6(Si,Al)_4O_{10}(OH)_8$. The presence of plagioclase feldspars (Fd) is based on x-ray reflections at 0.281, 0.278, 0.272 and 0.871 nm (Fig. 2), IR absorption bands at 533, 589, 780, 1034-1084 cm^{-1} wavenumber (Fig. 3) and the presence of K, Na, Al and Si (Table 1). Based on the separation of -132 and 131 between 0.267 and 0.285 nm reflections and 111 and 1-11 between 0.375 and 0.385 nm x-ray reflections (Huang, 1986), the calculated ratio of An/(An+Ab) is ~0.20. The higher proportion of Ab compared to An is also reflected in the higher amount of Na compared to Ca (Table 1). Quartz is identified based on x-ray reflections at 0.425, 0.334, 0.245, 0.228 and 0.182 nm (Fig. 2), IR absorption bands at 514, 798, 1034-1084 cm^{-1} wavenumber (Fig. 3), and the ~20 percent Si content (Table 1). The presence of Mg, Fe and x-ray reflections at 0.707, 1.41, 0.354 and 0.252 (Fig. 2) indicate the presence of chlorite, probably a clinochlore. Chlorite is also indicated by IR absorption bands at 647 and 1034-1084 (Fig. 3). The presence of x-ray reflections as 1.0 nm and the corresponding d_{002} reflections at 0.5 nm (Fig. 2) show that muscovite is the species of mica present in the andesitic rock. The presence of 3 % P indicate the presence of phosphate containing minerals in andesite. Electron micrographs of the andesite show some degree of dissolution pits into the fine-grained rock-forming minerals present in the sample (Fig. 4a, b).

B) The white coating

Fig. 1a and b show the accumulation of a white coating on the surface of the andesite. This white (10YR 8/1) coating is ~25 μm thick, shiny and relatively hard (hardness ~ 5 because it can be scratched by a knife blade), impervious, and fills most of the small indentations on the surface of andesite. In some parts of the coating, cracks filled with whitish materials are also recognizable (Fig. 4b, e). Remnants of biological activity are also indicated by the presence of fungal hyphae and spores (Fig. 4c, d).

The mineral composition of the white coating on the surface of the andesite is dominated by hydroxylapatite (Hp), $Ca_5(PO_4)_3OH$ with traces of calcite (Cc), $CaCO_3$ and quartz (Qt), (SiO_2). The presence of Hp is based on x-ray reflections at 0.281, 0.278, 0.272 and 0.871 nm (Fig. 2), IR absorption bands at 576, 605, 960 and 1037-1086 cm^{-1} wavenumber (Fig. 3) and the predominance of Ca and P in chemical composition (Table 1). Calcite is identified in the white coating based on x-ray reflections at 0.304, 0.229, 0.210 nm (Fig. 2) and 0.386, IR absorption bands at 870, 1430-1439 cm^{-1} wavenumber (Fig. 3) and the presence of Ca and C in the samples (Table 1). Some of the x-ray reflections for Cc are masked by the strong reflections for Hp and shown in x-ray reflections that are designated as Hp/Cc (Fig. 2). The relatively higher intensities of the x-ray reflections for Hp compared Cc indicate higher amount of Hp than Cc. Scanning electron micrographs show a "smooth" appearance that may indicate that Hp and Cc are intimately mixed in the white coating (Fig. 4b). This is also shown by the spatial distribution of Ca, P and C in white coating (Fig. 5). Identification of quartz is based on x-ray reflections at 0.425, 0.334, 0.245, 0.228 and 0.182 nm and the IR absorption bands at 514, 798, 1034-1084

cm^{-1} wavenumber (Fig. 3). Small amount of Si also shows the presence of quartz.

C) Flakes

The chemical composition of flakes is quite similar to that of the coating and is dominated by Ca, P, C and O (Table 1). The mineral composition is also similar to coating and consists of Hp, Cc and traces of Qt. Hydroxylapatite is indicated by IR absorption bands at 576, 605, 960 and 1037-1086 cm^{-1} wavenumber (Fig. 3) and the one-to-one distribution of Ca and P (Fig. 5). Calcite as a discrete entity is shown by adsorption at 870, 1430-1439 cm^{-1} wavenumber (Fig. 3) and the distinct distribution of carbon in the sample (Fig. 7). These flakes have abundant evidences of biological activity such as plant roots (Fig. 4g), spores, and mycelia (Fig. 4d, f, h).

Discussion

A) Formation of calcium phosphate coating on andesite

The formation of calcium phosphate minerals on the Yalour Islands could be attributed mainly to biological activity associated with penguin rookeries. For instance, krill, the main food source for penguins, contains high amounts of Ca and P (Everson, 1987) necessary for the formation of hydroxylapatite. The formation of rock phosphate as insular guano was attributed directly or indirectly to bird droppings (Cook 1983). Although, the andesitic rocks in the area contains P, the amount must be small (< 4% P) in order to account for the extensive distribution of Hp coating on the island. The abrupt boundary between the andesite and the calcium phosphate coating (Fig. 4b) also indicates the crystallization of Hp on the surface of andesite and precludes the formation of Hp from the breakdown of andesite. The accumulation of calcium phosphate from penguin rookeries has been reported earlier from other parts of the continent, particularly from the maritime Antarctic areas (Tatur and Keck 1990; Tatur and Barzuck 1985, Tatur 1989). The precipitation of Hp on andesite from the breakdown of penguin guano must have resulted from the favorable environmental conditions, such as sufficient amount of rainfall coupled with evaporation, by sun, wind or freezing and thawing, to superconcentrate the contents of Ca and P in guano solutions (Tatur and Keck 1990) and exceed the solubility product of Hp.

A1) Developmental Sequence

From the information presented in this paper, the stages in the formation of an Hp coating involve decomposition of penguin guano and the subsequent precipitation of Hp on andesite. First, an adequate amount of water, combined with sufficiently high rock temperatures, are needed to initiate the breakdown of penguin guano. Like any other organic matter, the breakdown of penguin guano leads to the formation of CO_2, which can aid weathering of the andesitic surface due to its tendency to form carbonic acid upon reaction with water (Stevenson 1994). The presence of pits on the rock surface suggests that weathering of the andesite may have resulted from the organic acids released by the decomposition of penguin guano (Fig. 4a). Conversely, the pits may have formed by subaerial weathering prior to their being coated, but this may have still been a result of organic acids from the penguin guano. However, whatever the cause, once pitted, the rock surface will be conducive to the precipitation of Hp because the pits retain the guano solution containing Ca and P until it reaches saturation, exceeding the solubility product of Hp, and so initiating the formation of an Hp coating on the andesite (Fig. 4b). The presence of P on Hp may trigger the growth of organisms such as fungus on the Hp coating as shown by the presence of hyphae (Fig. 4c) and spores (Fig. 4d) within the coating. Epilithic macrolichen is also known to grow on phosphate rich environment (Gremmen et al., 1994). The production of organic acids from these organisms can initiate subsequent weathering based on the breakdown of precipitated Hp and the formation of pseudomorphosed hyphae and spores (Fig. 4c, d, f, h). Other weathering process acting on the Hp coating could be purely abiotic (e.g., thermal shock) as indicated by the weathering cracks (Fig. 4e) filled up with secondary Hp. The product of the weathering processes (biotic and abiotic) acting on Hp coating results in the formation of "Hp flakes" as well as secondary accumulation of Hp on the other parts of andesitic surface. Similar to the Hp coating, Hp flakes are also conducive to the growth of organisms as shown by the vigorous root growth (Fig. 4g). From our results, there seems to be a continuous breakdown and formation of Hp on Yalour islands that started with the breakdown of penguin manure on both former and present-day rookeries.

B) Implications for geomorphic processes

The deposition of the phosphatic coating has several geomorphic implications. First, it's impervious nature means that where the andesite is covered so water penetration is extremely, if not totally, inhibited. Within this environment the ramifications of this are that: (a) any potential for chemical weathering is negated, (b) freeze-thaw is inhibited, as too are (c) wetting and drying, and (d) salt weathering. The creation of this coating thus significantly impacts future weathering, essentially inhibiting any further activity while the coating is in place. Second, the coating will, while it survives, help protect the rock from any wind or water-driven abrasion. Third, indirectly, the impervious nature of the coating will also serve to increase surface water run-off from the areas where it occurs. This may have ramifications for water-based redistribution of penguin guano from the rock surfaces which, in turn, may influence biological activity.

In addition to the direct impact on weathering processes outlined above, the coating will also influence the thermal condition of the rock as a result of changes in albedo. As Warke and Smith (1998) have discussed, temperature is a significant control on rock breakdown and albedo greatly influences this. The very dark, andesite is, by the coating, transformed to a shiny, significantly lighter-coloured, higher albedo, surface. The increase in albedo might lower the temperature of the andesite and inhibit thermal expansion of minerals (Dorn, 1998). Indeed, calculations of radiative exchange (using the METEONORM® software) resulting from a change in albedo from 0.2 to 0.3 for a horizontal surface at the Yalour Islands clearly demonstrate that even such a small albedo increase can have a significant impact on radiation budget (Table 2). As rock temperature will be greatly influenced by net all-wave radiation balance the small increase in albedo may have significant thermal responses. While not affecting the net loss for the winter months (May to August) the change in albedo decreases the radiative heating significantly for the summer months (December = 16% less radiation absorption, January = 17.5%, February = 21% and March = 44% less). Further, the timing of radiative warming is off-set at the end of winter as September experiences a net loss and October has 23.5% less absorption for the higher albedo surface. Thus, the albedo change has the potential to exert a significant influence on the thermal regime of the rock. The lowered temperatures will not, though, help with processes such as freeze-thaw due to the generally impervious nature of the coating prohibiting water being available for any freeze events. However, Paradise (1993 as cited in Dorn, 1998) reported that where rock coating formed in fractures this may increase the friability of weathering rind and so localized water penetration, and hence frost weathering, might be possible where water is available.

Although obviously untested, identification of possible geomorphic responses to the presence of the coating are germane to any understanding of their significance. That coatings should exist, and a developmental sequence postulated, is valid in and of itself, but it is important that some cognisance be given to their impact on the landscape as they exist within a dynamic, changing framework and they, themselves, affect how that develops. Although the geomorphic responses postulated here are obviously speculative they are based on our understanding of processes coupled with supportive radiation data calculations. Clearly this is a field that needs much more research, but also in the wider context beyond the present situation.

Conclusions

Our observations provide new information regarding the properties and formation of rock coatings and for the Yalour Islands in particular. An exposition of the developmental sequence of the coating from decomposition of penguin guano to formation of rock coating is provided together with the geomorphic implications of that coating. Whilst the chemistry and origin of the coating seem clear, the same cannot be said as to exactly why this took place on the Yalours but, from visual appraisal, no where else in this region. This needs further investigation both spatially (to determine the extent of coatings in this region) and site specific (i.e. why did it occur here?) to better understand causative mechanisms. Further, the geomorphic implications of the coating are in need of far more study and data. While the investigation as to the origin of rock coatings seems well developed, the same is not the case for the subsequent geomorphic implications. It is suggested, particularly as there are biological connections, that rock coatings are in need of greater study in the Antarctic.

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The opportunity to visit the Yalour Islands and to collect the rock samples was made possible (to KH) by the "Students on Ice" undertaking led by Geoff Green. We are extremely grateful for making this opportunity available. Funding for the work was provided by NSERC award #185756 and the Deans Office of UNBC (for KH), and NSERC award #21027 for JMA. We also thank Dr. O. Omotoso of CANMET for x-ray diffraction analysis, Dr. D. Dick of UNBC for FTIR analysis and Ms. Christine Phillips of BAS for undertaking a library search for information on the Yalour Islands.

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Table 1. Mean (and standard deviation) of elemental composition (weight %) of crust, flakes and basalt as determined by an energy dispersive system. Each value is a mean from 10 measurements.

	Crust	Flakes	Andesite
Carbon	10.0 (2.34)	17.5 (3.60)	8.2 (1.32)
Oxygen	52.6 (8.4)	29.3 (7.43)	45.6 (3.46)
Sodium	2.2 (0.39)	0.64 (0.46)	3.2 (2.0)
Magnesium	1.3 (0.37)	0.74 (0.26)	1.1 (0.30)
Aluminum	0.92 (0.70)	0.60 (0.34)	12.7 (1.01)
Silicon	1.6 (0.73)	2.3 (1.48)	19.6 (1.29)
Phosphorus	13. 6 (2.75)	18.8 (1.97)	3.4 (1.32)
Potassium	0.32 (0.18)	0.46 (0.17)	2.4 (0.83)
Calcium	17.0 (6.52)	28.7 (4.31)	0.86 (0.25)
Titanium	0.08 (0.02)	0.33 (0.17)	0.12 (0.05)
Iron	0.42 (0.19)	0.73 (0.34)	2.8 (1.11)

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Table 2. Mean net all-wave radiation balance (W/m^2) for the andisitic rock using METEONORM® for bare rock (albedo: 0.2) and that with the coating (albedo: 0.3).

Month	Albedo = 0.2	Albedo = 0.3
January	143	118
February	76	60
March	16	9
April	-11	-13
May	-22	-22
June	-23	-23
July	-21	-21
August	-16	-17
September	1	-5
October	68	52
November	143	117
December	178	149
Year Mean	**44**	**34**

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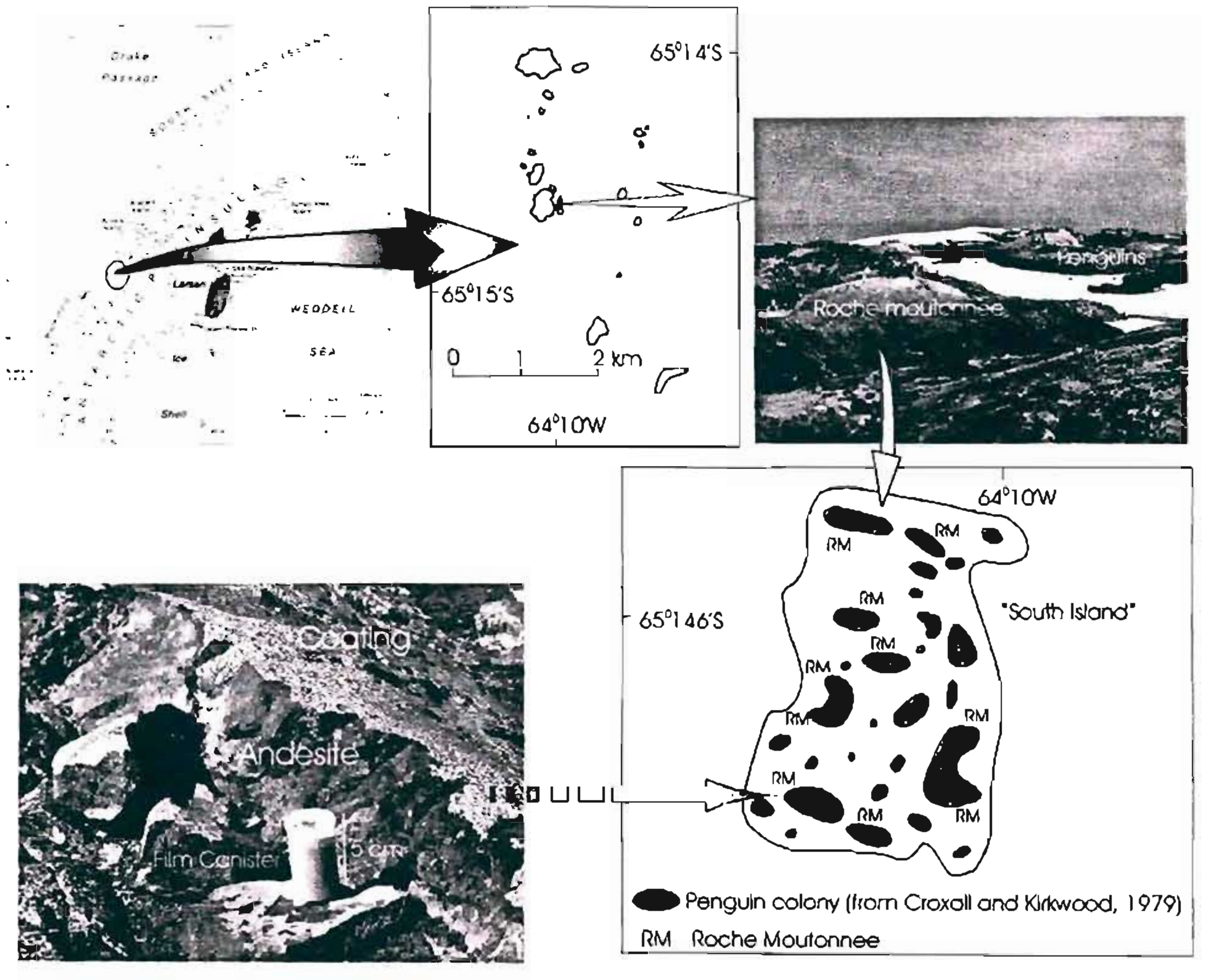
Fig. 1.To show the location of the study area, the study site itself and an example of the coating on the andesite. (The map of the Antarctic Peninsula is from Handy Atlas, 1995)

Fig. 2. X-ray diffractograms of coating and andesite samples. Hp – hydroxylapatite, Qt – quartz, Cc – calcite, Ch – chlorite, Ms – muscovite, Fd – feldspars

Fig. 3. Fourier Transformed Infrared spectra of flakes, coating and andesite samples. Hp – hydroxylapatite, Qt – quartz, Cc – calcite, Ch – chlorite, Ms – muscovite, Fd – feldspars

Fig. 4. Scanning electron micrographs of calcium phosphate coating on andesitic rocks from the Yalour Islands (Antarctica). (A) pits on the surface of andesite, (B) sharp boundary (arrow) between calcium phosphate coating (Hp) and andesite, (C) and (D) pseudomorphosed fungal hyphae and fungal spores, respectively, (E) cracks on rock coating filled with calcium phosphate, (F) rounded calcium phosphate present in flake samples, (G) abundant root growth on flakes, (H) another rounded accumulation of calcium phosphate in flake samples.

Fig. 5. Distribution of silicon (Si), aluminum (Al), calcium (Ca), phosphorus (P) and carbon (C) in calcium phosphate coating on andesitic rock of the Yalour Islands (Antarctica).


Drake Passage
WEDDELL SEA
65°14'S
65°15'S
0 1 2 km
64°10'W
Penguins
Roche moutonnee
64°10'W
65°146'S
"South Island"
RM
Penguin colony (from Croxall and Kirkwood, 1979)
RM Roche Moutonnee
Coating
Andesite
Film Canister
5 cm

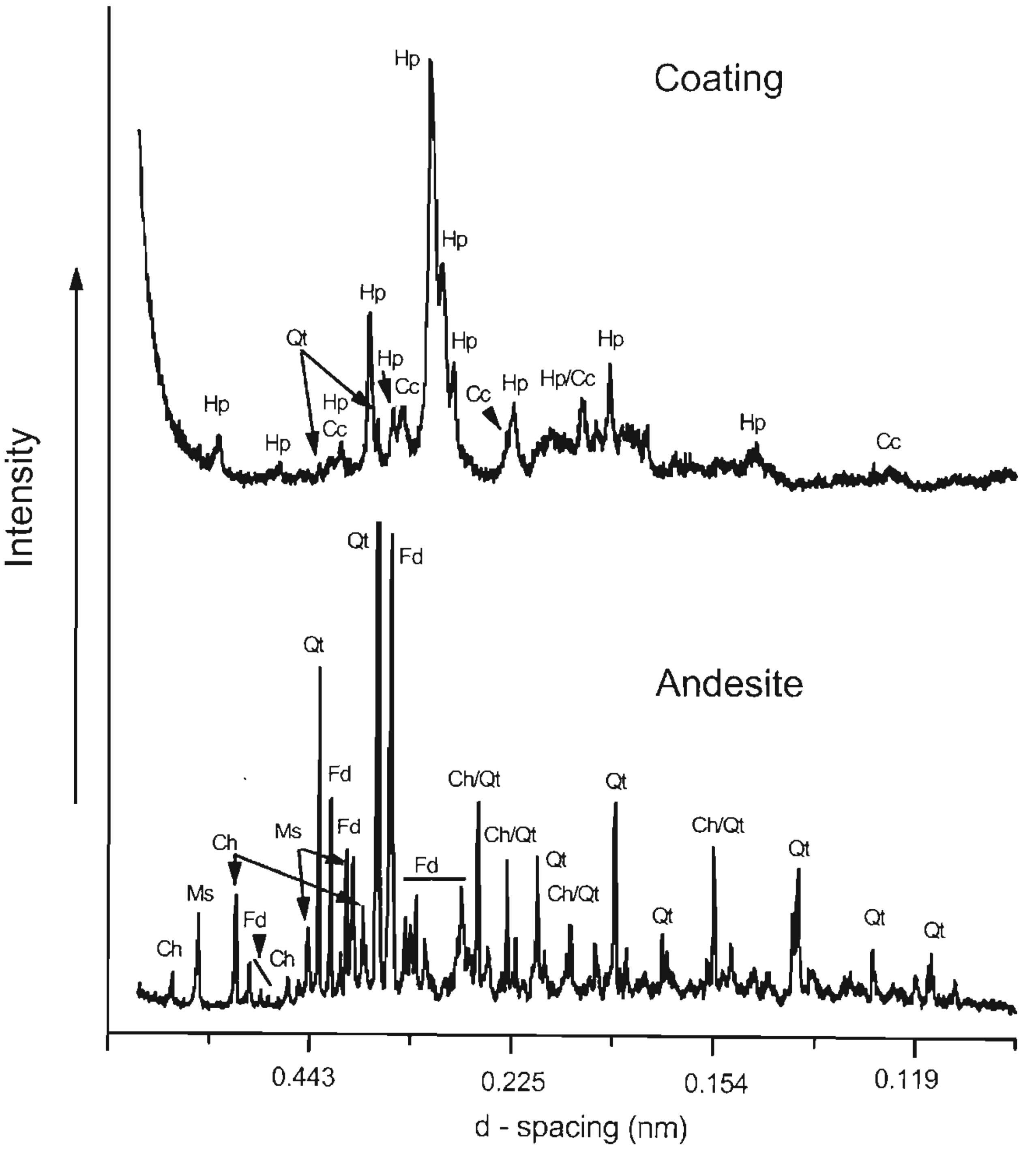


Fig. 2 - Arocena and Hall

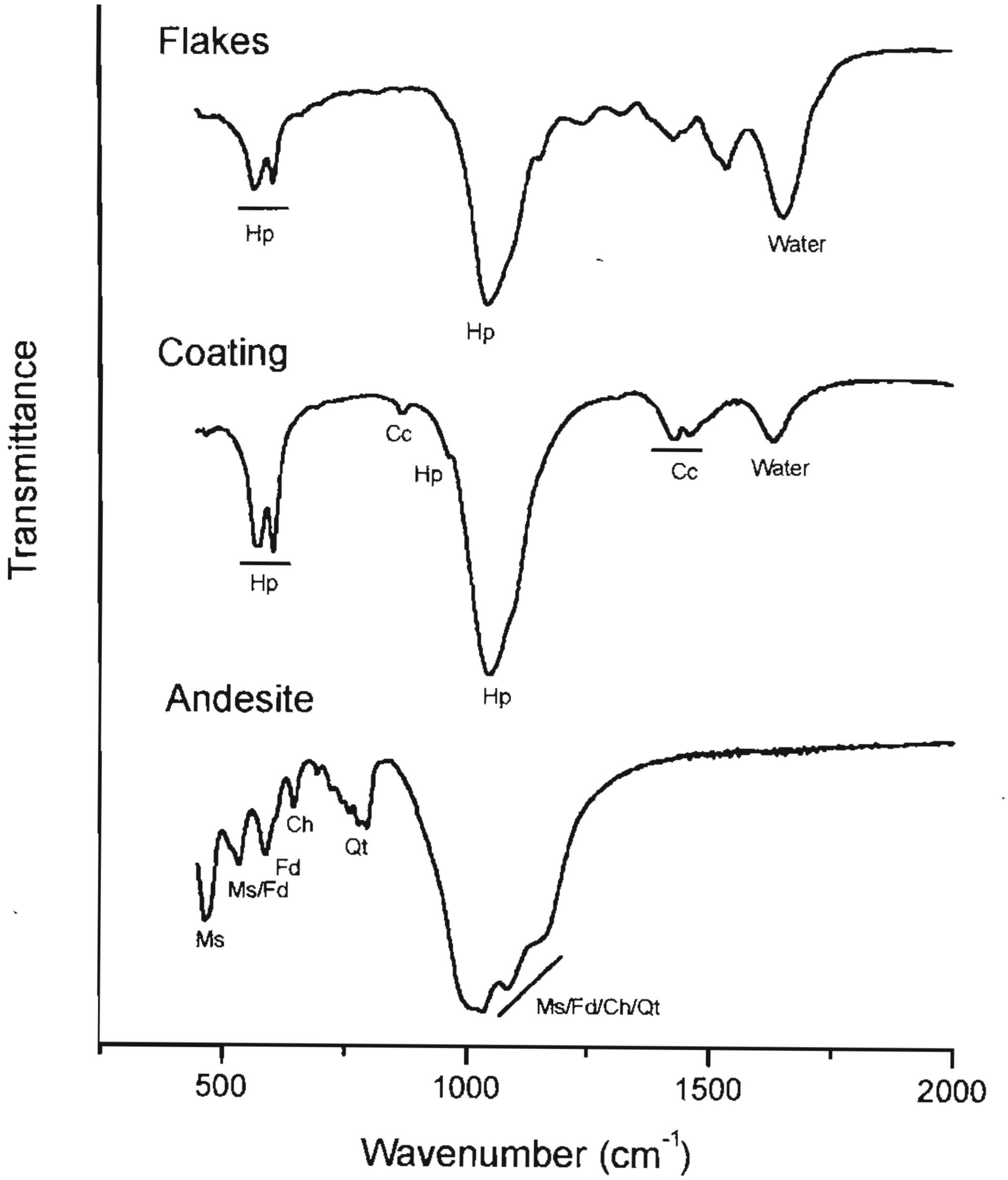


Fig.3-ArocenaandHall

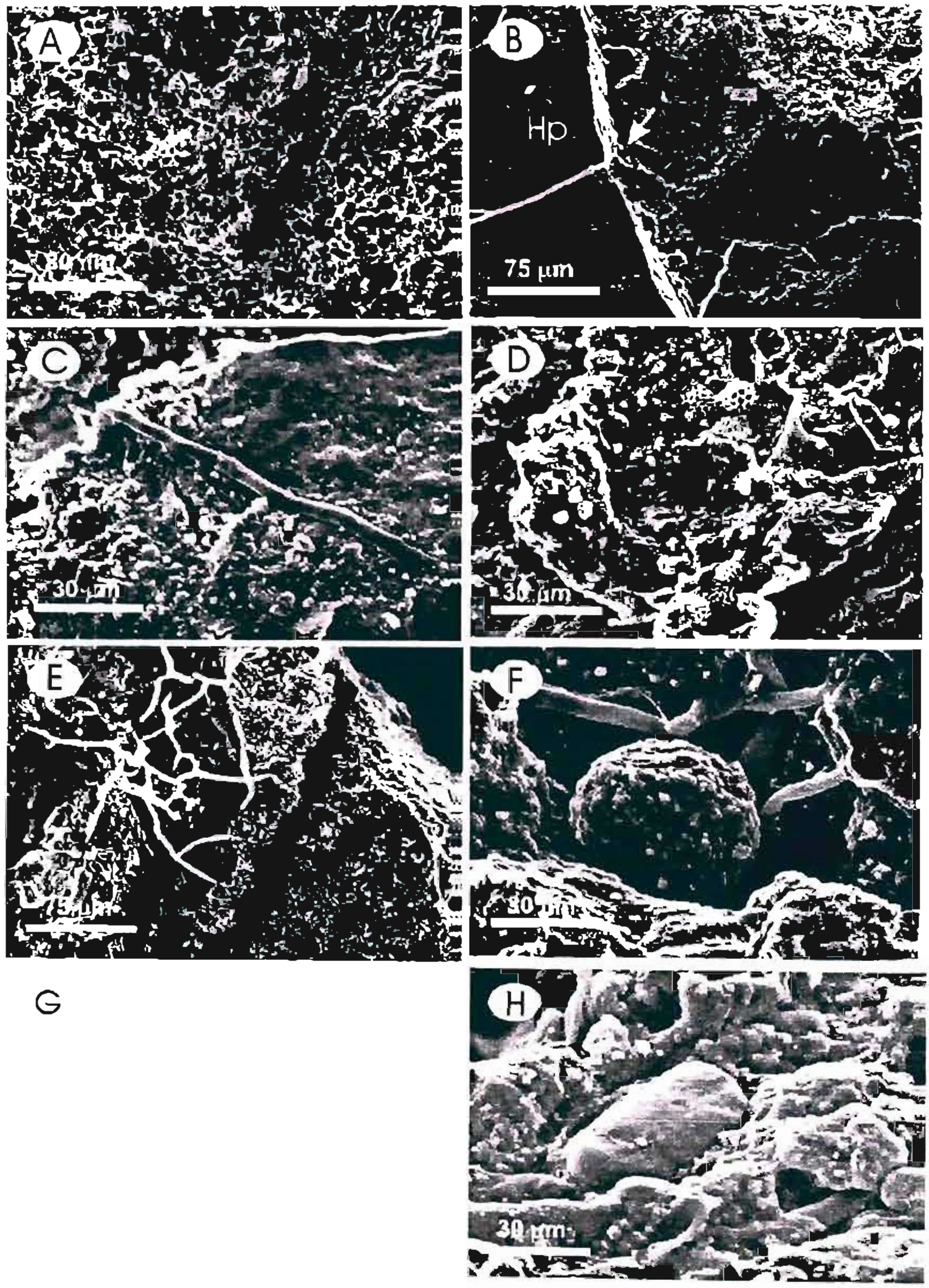


Fig.4-ArocenaandHall

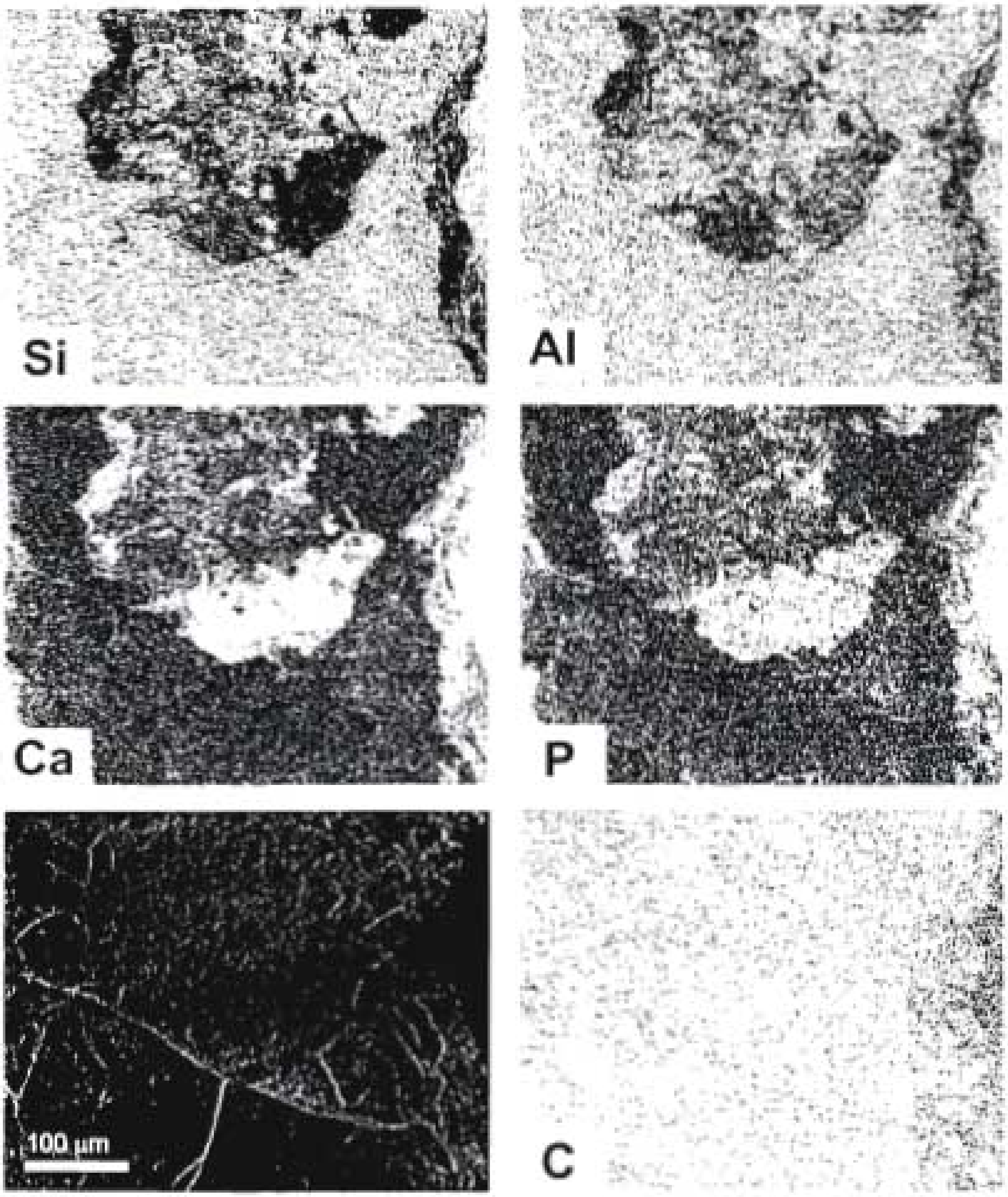


Fig. 5 - Arocena and Hall

Permafrost, Active Layer Dynamics and Periglacial Environments of Continental Antarctica

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ABSTRACT

Active-layer dynamics, permafrost and ground-ice characteristics, and selected periglacial features are summarized from recent published literature and unpublished data by the authors for three eco-climatic regions of continental Antarctica: the Antarctic Peninsula and its offshore islands (*c.*61-72°S), maritime East Antarctica (*c.*66-71°S), and the Transantarctic Mountains (*c.*71-87°S). Active-layer thickness and depth to ice-cemented permafrost are related to regional climate, proximity to glaciers, and albedo of surface rocks. In the McMurdo Dry Valleys (MDV), the active layer is commonly underlain by dry permafrost which can be detected only from frost tubes or temperature measurements. Permafrost thickness ranges from zero near thermally stratified saline lakes in dry valleys and beneath parts of the Antarctic Ice Sheet to ≥1000 m elsewhere. Permafrost temperature measurements are scant and range between -14 and -24°C at a depth of 50 m. Ground ice is present as rock glaciers along the polar plateau and in upland valleys, ice wedges near the coast, and in ice-cored moraines and buried ice throughout the MDV. Cryoplanation terraces occur along the Antarctic Peninsula and are important for interpreting fossil periglacial features in temperate environments. The potential effects of humans on permafrost and periglacial environments are briefly reviewed. Recommendations are made for future work in the region.

INTRODUCTION

Antarctica is the fifth largest continent with a summer area of 14 million km^2, possesses the highest mean elevation (~3000 m) of any continent, and experiences the lowest mean annual air temperature (-40^oC according to Weyant, 1967). Although technically the Antarctic region includes all oceans and land south of 60^oS, 'Antarctica' as used here refers solely to the Antarctic continent and islands offshore of the Antarctic Peninsula (Fig. 1). Antarctica is commonly divided into two parts that are roughly bisected by the Transantarctic Mountains (TAM): East Antarctica, a high-elevation land mass capped by the massive East Antarctic Ice Sheet (EAIS), and West Antarctica, an archipelago buried by ice from the West Antarctica Ice Sheet (WAIS) (Fig. 1). East and West Antarctica contain 10.2 and 2.3 million km^2 of glacial ice, respectively, which represents 90% of the world total (Denton, *et al.,* 1991). Whereas the EAIS has remained comparatively stable during the Pleistocene, the WAIS has periodically disintegrated, resulting in global increases in sea-level on the order of 6 m (Hughes, 1975). Moreover, the behavior of the WAIS correlates well with Northern Hemisphere glacial/interglacial cycles.

Less than 1% (55,000 km^2) of Antarctica is ice-free. Whereas the terrestrial ecosystems of Antarctica have a low biodiversity (Llano, 1965), the Southern Ocean has among the most productive marine ecosystems on Earth (Arntz *et al.*, 1997). This is due to the Antarctic Convergence, a zone that features upwelling of nutrient-rich sediments due to mixing of warm subtropical waters and cold sub-antarctic waters. Antarctica acts as a heat sink and exerts a strong influence on the earth's atmospheric and cryospheric systems (Weyant, 1967). Antarctica contains 37% of the world's permafrost (French, 1996). However, much of the land mass below the massive EAIS is above the pressure melting point and is unfrozen (Herterich, 1988). Realistically, 25% or less of the Antarctic region contains permafrost (Bockheim, 1995).

Ground ice is common in Antarctica, especially in the form of rock glaciers (Hassinger and Mayewski, 1983) that have potential for reconstructing past environments of the region (Gillespie *et al.*, 1997). For example, ice buried 50 cm below the ground surface in Beacon Valley (77°51'S; 160°35'E) may be ≥8 My old (Sugden et al., 1995). Human effects on the active layer and permafrost have received recent attention (Bockheim, 1993; Campbell, *et al.*, 1994).

The objective of this paper is to summarize relevant data on active-layer dynamics, permafrost and ground ice, and periglacial features and environments of Antarctica, with an emphasis on data published since 1994 (or missing from other review papers) and unpublished data by the authors.

ECO-CLIMATIC REGIONS OF ANTARCTICA

Ice-free areas of the Antarctic continent may be divided into three eco-climatic regions: the Antarctic Peninsula and its offshore islands (*c.*61-72°S), maritime East Antarctica (*c.*66-71°C), and the Transantarctic Mountains (TAM) (*c.*72-87°S) (Bockheim, 1995) (Fig. 1). The McMurdo Dry Valleys (MDV) have been sub-divided from environmental factors into sub-regions, including coastal, inland valley floors, inland valley sides, upland valleys, and the plateau fringe (Campbell et al., 1998b). The TAM, which extend from northern Victoria Land to the Pensacola Mountains (Fig. 1), comprise the largest proportion of the ice-free area of Antarctica at 55%, or 30,400 km^2. The Antarctic Peninsula and its offshore islands comprise about 14% (8,000 km^2) of the ice-free area of Antarctica and include the South Orkney Islands, islands to the west of the Antarctic Peninsula (the South Shetland Islands, the Palmer Archipelago, the Biscoe Islands, Adelaide Island, and Alexander Island), islands northeast of the Trinity (Antarctic) Peninsula (e.g., Joinville, Snow Hill, James Ross, and Seymour Islands), Thurston Island, Peter Øy, and the Balleny Islands (Fig. 1). Marie Byrd Land contains 12% of the ice-free area (6,600 km^2), primarily in the Ford Ranges and the Executive Commitee Range. Lesser proportions of ice-free areas exist in Queen Maud Land (4.8%, or 2,600 km^2, including the Wolthat Mountains and the Sør Rondane

Mountains), maritime East Antarctica (4.6%, or 2,500 km^2, including the Schirmacher Hills, Enderby Land, the Vestfold Hills, the Queen Mary Coast, Wilkes Land, the Adélie Coast, and the Wilson Hills), the Prince Charles Mountains (4.4%, or 2,400 km^2), and the Ellsworth Mountains (3.1%, or 1,700 km^2).

These regions and sub-regions differ not only in climate but also in vegetation, soils, permafrost characteristics, and periglacial processes and rates (Blümel and Eitel, 1989). Blümel and Eitel (1989, p. 213) provided geoecological divisions of Antarctica based on periglacial features, suggesting that the lower latitude maritime Antarctic system "...can be used as a model for the ecodynamic in western mid-latitudes during certain Pleistocene phases." The mean annual air temperature (MAAT) along the Antarctic Peninsula and its offshore islands commonly ranges between -2.7^{0}C and -3.7^{0}C (Table 1). In maritime East Antarctica, MAAT ranges from -9.4^{0}C to -11^{0}C whilst in the TAM, MAAT ranges from -15^{0}C to -20^{0}C in coastal areas to colder than -35^{0}C along the polar plateau. Mean annual precipitation ranges from 400 to 1000 mm yr^{-1} in the Antarctic islands to <100 mm of water-equivalent precipitation along the polar plateau (Table 1).

Bliss (1979) classified vegetation of continental Antarctica in the 'Polar Desert Biome'. Major subdivisions included grass-herb-fellfield communities in the Antarctic islands and maritime Antarctica, maritime moss communities in maritime Antarctica and coastal areas of the TAM, and lichen barrens in maritime Antarctica and in the Transantarctic Mountains. Aleksandrova (1980) provided much the same geobotanical division, although she grouped the islands south of the Antarctic Convergence with the polar deserts of the continent which may underestimate the biodiversity and productivity of these islands compared to the continent. To some extent Aleksandrova (1980) rectified this concern by generating the "northern subregion of the Antarctic polar deserts" (the southern subregion being the continent proper) in which some of the maritime Antarctic islands are then considered.

Antarctic soils are classified as Gelisols, i.e., soils with permafrost within 100 cm of the surface or soils with gelic materials within 100 cm of the surface and permafrost within 200 cm of the surface (Soil Survey Staff, 1998). Gelisols are divided into three suborders: Histels (soils in which ≥40% of the upper 50 cm contain organic materials), Turbels (mineral soils that have one or more horizons showing cryoturbation), and Orthels (other soils). Histels are infrequent in Antarctic and occur primarily in small bogs in the Antarctic islands and maritime East Antarctica. Soils in the Antarctic islands and maritime East Antarctica lacking anhydrous conditions and abundant salts are primarily Psammiturbels/Psammorthels (if there are less than 35% fragments that are >2 mm) or Haploturbels/ Haplorthels (if there are more than 35% fragments that are >2 mm). Soils of the TAM have anhydrous conditions (i.e. are exceptionally dry) and are dominantly Anhyturbels and Anhyorthels (Bockheim, 1997). Spodosols developed in relic ornithogenic materials have been reported at Casey Station in Wilkes Land (Fig. 1) (Beyer et al., 1997).

ACTIVE-LAYER DYNAMICS

The concept of an active layer applies to the Antarctic Peninsula and its offshore islands and to maritime East Antarctica but is less meaningful in interior Antarctica. The main reason for this is that the active layer and near-surface permafrost in interior Antarctica have exceptionally low moisture contents, i.e., less than 3% (Black and Berg, 1963; Campbell, *et al.*, 1998a, b). A gravimetric moisture content of at least 5% is generally required for coarse-textured soils of Antarctica to be cemented by ice (Campbell, *et al.*, 1994). Therefore, much of the permafrost in interior Antarctica is "dry" (Bockheim and Tarnocai, 1998).

The maximum active layer thickness in Antarctica is dependent on eco-climatic region and albedo of the ground surface materials. In the comparatively mild Antarctic islands, the active layer commonly ranges between 40 and 150 cm (Table 2). Maritime

East Antarctica has a cooler but drier climate with active-layer thicknesses of 60 to 150 cm. However, active layer depths in East Antarctica may be overestimated as dry permafrost likely occurs there. In the TAM, the active layer thickness is dependent on proximity to the coast and elevation, ranging from 30-60 cm in coastal areas to 15-20 cm along the polar plateau (Table 2). A recent study by Guglielmin *et al.* (1997), based on sampling 20 sites in the northern foothills of northern Victoria Land, showed a thin active layer of 5 to 30 cm. Gragnani *et al.* (1998), also working in north Victoria Land, suggested that seasonal melting-refreezing occurs to depths of 30-90 cm.

Surface albedo is an important factor with regards to active-layer thickness and dynamics in Antarctica (Campbell *et al.,* 1998b). Wilson and Bockheim (unpublished) monitored soil temperature and moisture on six pairs of plots with dark- and light-colored desert pavements in the MDV. Plots with an abundance of dark-colored, mafic rocks had significantly ($p \leq 0.05$) lower albedos, greater soil temperatures, and greater soluble salts in the upper 30 cm than adjacent plots with a light-colored surface (Table 3). Wójcik (1989) measured active-layer temperatures in the Bunger Oasis (66^{0}18'S, 100^{0}43'E), showing that daily fluctuations are greatly influenced by diurnal temperatures and external factors such as cloud cover. Partial cloudiness, full cloudiness, and full cloudiness and snow all affected the vertical temperature profile, with vertical temperature gradients being greatest during cloud-free days. Temperature gradients decreased as autumn approached. Temperature fluctuations in the bedrock active layer were highly conducive to mechanical weathering (Wójcik, 1989). In coastal areas of the MDV, maximum soil surface temperatures are mainly radiation controlled whilst minimum temperatures are strongly linked to air temperatures, thereby generating the greatest temperature ranges (of soil surface temperatures) in the highest and coldest environments (Campbell et al., 1998b).

Active-layer temperatures and dynamics vary with eco-climatic region in Antarctica. Deep active layer and extensive cryoturbation occurs along the Antarctic Peninsula and its offshore islands. Working in the South Shetland Islands, Chambers (1967) emphasized the dual nature of the active layer. Solifluction, ice segregation, and frost sorting were all concentrated in the upper 40-60 cm of the active layer. The lower 60 cm of the active layer played no apparent role in the formation or current activity of patterned ground. In maritime East Antarctica, MacNamara (1969) reported active layer thicknesses of 100-150 cm. Suspended materials were carried downward in moisture as films on soil particles and in soil macropores. Sharp decreases in soil moisture occurred during thawing and increases during freezeup. In inland areas of the TAM, solifluction and debris flows are absent (Marchant and Denton, 1996) and cryoturbation occurs only to a limited extent. In the Asgard Range of the TAM, McKay et al. (1998) reported an active zone (the depth in which melting could occur since temperatures exceeded 0°C) of about 12.5 cm by interpolating instantaneous temperatures between the surface and subsurface, below which was a 12.5-cm zone of dry permafrost overlaying ice-cemented permafrost at 25 cm. There was a net flux of water vapour from the ice to the atmosphere, resulting in a recession of the ice-cemented ground by about 0.4-0.6 mm yr^{-1}.

PERMAFROST DISTRIBUTION AND PROPERTIES

Permafrost is defined here as soil and/or rock that remains below 0°C for at least two consecutive years (ACGR, 1988). Moisture in the form of water or ice may or may not be present. Accordingly, two kinds of permafrost exist in the cold deserts of Antarctica. With the exception of the inland valley floors and sides of the MDV, most of Antarctica contains ice-cemented permafrost. Figure 2 was prepared from 482 soil pits in the MDV examined by Bockheim during the period 1975 to 1987. The data show that dry permafrost is prevalent in the region, occurring in 41% of the pits examined, with ice-cemented permafrost being restricted to (1) Ross Sea drift 12-20 ka in age in coastal areas, (2) alpine drift <74 ka in age in the dry valleys, (3) sediments at elevations above 2,000 m in upland valleys and along the edge of the polar plateau, and (4) below the dry permafrost at depths exceeding 100 cm.

Compared with the circumarctic region, very little is known regarding the thickness, properties, and age of permafrost in Antarctica. The most extensive database for permafrost in Antarctica is from the Dry Valley Drilling Project (DVDP) (McGinnis, 1981) (see Fig. 3). During this study 15 boreholes were drilled in the MDV to depths ranging from 4 to 381 m. Descriptions of the strata were taken, including notes of whether or not the sediments were wet-frozen (Mudrey and McGinnis, 1975). Electrical resistivity was used to estimate permafrost thickness (McGinnis and Jensen, 1971; McGinnis *et al.*, 1973). Subsurface temperatures were measured at discrete points in the hole with thermistor probes (Decker and Bucher, 1977). Because the boreholes were located near lakes and have filled with water, they are no longer suitable for monitoring permafrost temperatures (G. Clow, U.S. Geological Survey, personal communication to Jerry Brown). Three boreholes have been established at the Italian research station in North Victoria Land (Guglielmin *et al.*, 1997). Permafrost characteristics have been studied in northern Victoria Land (Guglielmin *et al.*, 1997; Gragnani, *et al.,* 1998), Seymour Island (Fournier *et al.*, 1987; Fukuda and Corté, 1989), and on King George Island in the South Shetlands (Govorukha *et al.,* 1975; Chen, 1993; Zhu *et al.,* 1993).

Non-relict permafrost thickness can be estimated crudely as the product of mean annual air temperature and a lag rate of 33 m/^{0}C (Gold *et al.*, 1972). Therefore, in ice-free areas permafrost thickness would be expected to be least in the Antarctic islands and greatest in interior Antarctica. On King George and Seymour Islands, permafrost is between 20 and 180 m thick (Table 4). We were unable to locate data for permafrost thickness in East Antarctica. In interior Antarctica permafrost may approach 1000 m in thickness (Decker and Bucher, 1977). Based on limited data, the temperature of Antarctic permafrost at a depth of 50 m ranges from -13 to -24^{0}C (Table 4). However, radio-echo soundings of the East Antarctic ice sheet, where the ice thickness is in excess of 3,000 m, suggest that the presence of saline lakes indicate the lack of subglacial permafrost (Oswald and Robin, 1973; Herterich, 1988).

Environmental factors related to permafrost in the circumantarctic region were reviewed by Bockheim (1995) and include air temperature, relief, vegetation, hydrology, glacier and snow cover, and soil and rock, and age of geomorphic surface. Bockheim (1995) provided a preliminary map of permafrost distribution in the Southern Circumpolar Region (Fig. 1) that related the northern occurrence of permafrost to the -1^{0}C isotherm for mean annual air temperature. Subglacier permafrost was determined from a map by Herterich (1988) that was based on a three-dimensional model of the Antarctic ice sheet. Since the preliminary permafrost map was published, virtually no new data for the distribution of permafrost in Antarctica have been published.

GROUND ICE IN ANTARCTICA

As might be expected there is extensive ground ice in Antarctica, primarily in the form of pingos, rock glaciers, ice-cored drift and buried glacial ice, and ice wedges. Pingos were reported by Pickard (1983) from the Vestfold Hills (68^{0}40'S, 78^{0}00'E), but Fitzsimons (1989) later refuted the interpretation. Rather than 'growth' features indicative of a pingo, the 4 m high by 12 m diameter features were argued to be residual landforms resulting from the decay of an ice-cored moraine, i.e., a form of thermokarst. Although the features are pingo-like in form, the "...environmental conditions necessary for pingo growth preclude their origin as pingos" (Fitzsimons, 1989). Grigor'ev (1965) also refers to the occurrence of thermokarst features on the moraines in the Bunger Hills, an area similar to but to the east of the Vestfold Hills. The lack of pingos in Antarctica is likely related to the absence of suitable sites combined with the lack of suitable conditions (Pickard, 1983).

Nichols (1966) refers to large areas in the MDV covered by ice-cored moraines. The buried glacier ice in this area produces a topography characterized by "knobs and mounds" as well as "valleys and kettles" and "inverted topography" where what were pond sediments now occur on ridge tops due to protection of the underlying ice from melting (Nichols, 1966). Nichols (1966, p.27-28) cites ice-cored moraines that "cover hundreds of acres", that are composed of ice *c.*16 m thick and that may be of "some antiquity"; photographs of these features were also provided (Figs. 30-34.) Based on resistivity data, Guglielmin *et al.* (1997) showed the occurrence of subsea permafrost in raised beaches of northern Victoria Land as well as permafrost with a very high ice content. Gragnani *et al.* (1998), also working in northern Victoria Land, recorded a ground ice layer with a thickness >60 m. In the ice-free areas of the Bunger Hills, Grigor'ev (1965, p. 175) observed "ground ice inclusions in the form of ice-cement, segregation ice, and multiveined ice.." within the Quaternary deposits. The predominant form of ice, by volume, was multiveined ice which, when melting occurred, produced "settling and sink phenomena". However, thermokarst development was inhibited by multiveined ice below the depth of seasonal thawing (1.5 - 1.7 m) that was only *c.* 0.4 m. Sugden et al. (1995) reported the existence of glacier ice lying beneath 50 cm of sediment in Beacon Valley in the MDV that may be of Miocene age. Hindmarsh *et al.* (1998) provided a detailed analysis regarding sublimation of the ice through the overlying sediments and questioned the 8 Ma age of the ice.

Rock glaciers are cited by a number of authors for different areas in Antarctica. Barsch *et al.* (1985) referred to the occurrence of active 'talus rock glaciers' on King George Island (South Shetland Islands) in an area of shallow but continuous permafrost. These rock glaciers are moving with an average speed of 30 cm yr^{-1} and are crossing a raised beach that was formed *c.*500 - 1000 yrs BP (Barsch and Mäusbacher, 1986). Mayewski (1979) and Mjagkov (1980) studied rock glaciers in the TAM. Guglielmin *et al.* (1997) described rock glaciers in northern Victoria Land and measured geoelectrical properties of the rock-glacier ice. Based on resistivity measurements, they were able to distinguish between ice-cemented rock glaciers and those with an ice-core (considered to be probably glacier ice). Humlum (1998) and Strelin and Sone (1998) reported rock glaciers on James Ross Island (64^{0}S, 58^{0}W) on the eastern side of the Antarctic Peninsula. The island is in an area of continuous permafrost and numerous rock glaciers are found in the (glacier) ice-free northwest corner of the island. Humlum (1998) suggested that most of the rock glaciers are glacier-derived but that there are also a few talus-derived forms. Strelin and Sone (1998) refer to protalus lobes' and 'protalus ramparts' as landforms not associated with glacier ice but rather may have ice-cemented interiors.

'Ice-cored talus aprons' were described in the Thiel Mountains of interior Antarctica (Ford and Andersen, 1967) as well as associated processes that cause debris movement and sorting. The talus aprons, located near to and south of latitude 84^0S, have an underlying ice core that, when subject to sudden temperature changes, causes "shaking" of the surface such that sorting of material takes place. This area experiences a mean annual temperature -36^0C but the ground can undergo rapid temperature changes as a result of variations in radiation input or winds (Ford and Andersen, 1967). Not only is this one of the few descriptions of massive ice at this latitude but also it is used to explain a most unusual form of sorting, one never yet evidenced in the Northern Hemisphere.

Further, Ford and Andersen (1967) suggest that in addition to the features observed as a result of debris movement due to shaking resulting from ice contraction there are what appear to be protalus ramparts. This unusual sorting mechanism is seen to produce stripes of surficial debris averaging 3 m long but which can be as great as 10 m, and are uniformly spaced 3 m apart. From their observations it is not clear whether the ice within the apron is aggrading or degrading, but there are certainly rock particles within the ice (Ford and Andersen, 1967). Ford and Andersen (1967) identify several other authors who refer to the "snapping and popping" associated with thermal changes to ice and ground in Antarctica resulting from shadow falling across previously warmed ground at a time when air temperatures are substantially sub-zero. The thermal contraction that produces tensional stress in the surface layer, said here to be the cause of the sorting, may also help explain the pseudo-sorting of thermal contraction cracks cited by Hall (1997a).

In Figure 4, we present a preliminary map of ground-ice features in the MDV. Rock glaciers and buried ice are prevalent in upland valleys and along the edge of the polar plateau such as Arena and Beacon Valleys (Mayewski, 1979; Mayewski and Hassinger, 1980) and elsewhere in the TAM (Mjagkov, 1980). The rock glaciers and buried ice in Arena and Beacon valleys contain less than 1 m of drift over pure ice (Hassinger and Mayewski, 1983). Ice-cored drift is common in lower Wright Valley on Holocene drift from the Wright Lower Glacier and on alpine moraines in Taylor and Wright Valley that are of Holocene age. Buried ice that is possibly 75Ky in age occurs in central Wright Valley (Bell, 1966). Ice- wedge polygons are common throughout the dry valleys, particularly in coastal areas (Black and Berg, 1963; Berg and Black, 1966; Black, 1973)

Cryoplanation/Nivation

Part of the problem with the consideration of 'cryoplanation' is its intimate interaction with 'nivation' such that it is almost impossible to consider one without the other (see Hall, 1998a). Thus it is pertinent to consider references to both process and their association within the Antarctic region. One of the first clear references to either process is that of Taylor (1914, p. 555) where, with respect to East Antarctica, he cites the creation of "...small cwms by "nivation"....". Taylor (1916, p. 175) also refers to nivation ("erosion by thaw and freeze") in his explanation of landscape development in the Royal Society Range. Taylor (1914, Fig. 16; 1916, p. 374) observed processes (1914, p.555) that appear very similar to what Groom (1959) would later term "niche glaciers" - forms that relate very strongly to "longitudinal nivation hollows" (Lewis, 1939). Souchez (1966, 1967) also notes the role of nivation in creating hollows, below which are found stratified deposits (grèzes litées). Nichols (1963) and McCraw (1967) discuss the occurrence of "nivation cirques" in the MDV. Nichols (1963) describes hollows which still had snow present that were 8.2 m long, 6.7 m wide, and 0.76 m deep, with a surface gradient of *c.*10^0. Nichols (1963) estimated that as much as 1.52 to 1.83 m of sediment had been removed. In the nearby Taylor Valley, Nichols (1963) observed 'nivation cirques' up to 30 m wide and 30 m long, and near Marble

Point he found similar cirques cut in bedrock that were "...hundreds of yards wide..".

Markov *et al.* (1970) noted that in the mountains of Queen Maud Land nivation process are currently active and the formation of cirques in this region may be due to nivation during warmer periods when these processes were more vigorous. Bardin (1963) also referred to the role of nivation in the mountains of Queen Maud Land. All workers noted the influence of aspect with the north-facing slopes being preferential for the action of nivation. Bardin (1964a) referred to nivation within the Schirmacher Oasis whilst Sekyra (1969) specifically mentioned the occurrence of cryoplanation terraces in the Schirmacher Oasis and the processes (primarily frost weathering) associated with these forms were seen as major contributors to the development of the present day landscape.

In his review of permafrost and periglacial processes in the Southern Circumpolar Region, Bockheim (1995) summarized data on cryoplanation terraces in Antarctica. Hall (1997a) added further references regarding cryoplanation. Previous reports suggested that cryoplanation terraces occur along the east coast of the Antarctic Peninsula (Corté, 1983) and in maritime East Antarctica (Sekyra, 1969; Vtyurin, 1986; Aniya and Hayashi, 1989). More recently, Hall (1997a, 1998a, 1998b) examined cryoplanation benches on Alexander Island (71°50'S, 68°21'W) near the Antarctic Peninsula. The terraces on Alexander Island were 2-12 m in width, 6-200 m long, with risers 0.8 - 2.0 m in height, whilst treads were at an angle of 1-10° (Hall, 1997a). Although Hall (1997a) found a distinct structural control, terraces showed a clear aspect influence with a preference (despite equal opportunity of exposure) for the southwest through north to northeast sector. Of importance to the broader concepts was the observation on the cryoplanation terraces of certain lithologies weathering to well rounded, rather than angular, forms (see below). Jordan and van der Wateren (1993, p. 375) refer to the mechanical weathering of bedrock in northern Victoria Land that can result in "subrounded to well rounded" rock fragments. Although no discussion was provided, several photographs and drawings by Taylor (1916, opp. p.

380, opp. p. 353) show granite boulders said to be *in situ* that are well rounded rather than angular in form. Thermal stress is likely a, if not the, major process associated with terrace formation (Hall, 1997a, 1998a, 1998b). In mountainous areas of Enderby Land, "relic" cryoplanation features occur (Sekyra, 1969), although this same author (p. 286) notes that amongst the processes causing the greatest remodelling of the relief in areas such as the Schirmacher Oasis are those of "cryoplanation processes". Jordan and van der Wateren (1993) suggest the occurrence of cryoplanation terraces in the ice-free area of Litell Rocks in northern Victoria Land. The terraces are in the order of 2-6 m in width, 20-60 m in length, and with a tread slope of a few degrees on the wider features to >10^0 for the narrower terraces. Risers at the back of the terraces range from almost vertical on the smaller terraces to *c.*60^0 for the wider terraces. The formation and location of the terraces may be structurally controlled. These forms appear, from the descriptions, to be extremely similar both in form and origin to those described by Hall (1997a, 1998a). Derbyshire (1972) identified tors (see below) which he considered to be developed above cryoplanation surfaces in southern Victoria Land.

Forms that could be considered associated with 'nivation-cryoplanation' processes include tors, blockfields and "rock furrows" for which some literature exists. Running water is more prevalent in the Antarctic than is often recognised. Although it is temporally and spatially constrained, largely to where radiative heat from rock outcrops causes snow to melt during the short summer, nivation can play a role in not only debris transport but also in creating erosional features (Brook, 1972). Several Russian authors have identified a peculiar form of "rock furrow" that they attribute to the melting of snow on north-facing slopes that, by a combination of fluvial action and freeze-thaw weathering, generate parallel furrows *c.*0.5 m wide, 0.3 m deep, and tens of meters long (Bardin, 1964b, 1964c, 1966; Markov *et al.*, 1970). Bardin (1966) notes that although the period when meltwater is active is very limited, the presence of the water facilitates intense frost weathering such that distinct "furrows" are created on vertical granite walls on the sunny north-facing slopes; this process of frost weathering and running water, to help remove the debris and re-wet the rock, is said to be a daily occurrence in summer. Bardin (1966) also notes that on horizontal surfaces the availability of melt-water causes "potholes" - again the origin being the frost weathering of the rock as a result of water being available. More recent studies (Balke, *et al.,* 1991) show that in such situations it is very likely that there is active chemical weathering taking place. Markov *et al.,* (1970) provided data regarding water temperatures (≥30⁰C) for the Bunger Oasis which would certainly facilitate active chemical weathering in these spatially restricted sites.

There are several references to the occurrence of tors (e.g. Sekyra, 1969; Selby, 1972; Derbyshire, 1972; Zhang and Peterson, 1984) that relate to the periglacial environment and the operation of both mechanical and chemical weathering - often associated with water from snow-melt. Taylor (1916, facing p. 275) showed the development of small 'tors' developed in weathered kenyte boulders. Zhang and Peterson (1984) identified what they consider to be frost shattered bedrock summits with remnant upstanding blocks about 1 m high, whilst those found by Derbyshire (1972) were 3-4 meters high and those reported by Selby (1972) were as high as 20 m. In the light of the above discussion regarding cryoplanation and the shape of the material on the terraces by Hall (1997a & 1998a), it is worth noting that Derbyshire (1972, p. 96) suggested that the observed summit tors of well rounded rock blocks standing above a "...surface of cryoplanation poses a problem in interpretation". The issue is that, as argued above, most workers assume that mechanically weathered debris should be angular and thus the finding of rounded material is in conflict with a cryogenic origin. As discussed earlier, without the assumption that the mechanically weathered material *must* be angular the problem does not exist and there is no conflict in finding rounded summit blocks above a cryoplanation surface - the rounding being a function of mechanical breakdown that results in curvilinear cracks rather than chemical weathering (see Hall, 1998a, 1998b). Sekyra (1969, p. 284) referred to the development of tors ("castle- and tower-shaped forms") in massive gabbro-granites but unfortunately provides no measurements of the features. Tors cited in the above references occur in dolerites, granites, gneisses and sandstones - the largest being in granites, the smallest in gneiss. Whilst Selby (1972) argued for salt weathering as the dominant process, Derbyshire (1972) and Zhang and Peterson (1984) suggested frost weathering, and Hall (1997a, 1998b) found the cause of weathering to most likely be thermal stress fatigue.

With respect to blockfields (felsenmeer) and block slopes, there are a number of references that indicate such forms exist on the tops and/or sides of many nunataks. Zhang and Peterson (1984) refer to the occurrence of block slopes in the Vestfold Hills, whilst Brook (1972) observes mantles of *in situ* weathered sandstones and siltstones on cliff tops in the Theron Mountains. Baroni (1987, p.201) suggested that the surface of the hills near Terra Nova Bay "...is shattered and produces blockfields (*Felsenmeer*)....Block sheets occur on steeper slopes." Nichols (1966) noted that although felsenmeer are found in the McMurdo Sound region they are not common. Nichols (1966) suggested that felsenmeer result from a combination of frost shattering and removal of fine material by the strong winds. In some areas there may be confusion between tills and felsenmeer but Nichols (1966) suggested that the latter may discerned by their more angular debris, that it is matrix deficient and that, in some places, the fragments can be fitted back together again showing that they were developed *in situ*.

HUMAN EFFECTS ON PERMAFROST AND PERIGLACIAL ENVIRONMENTS IN ANTARCTICA

Humans have had significant direct impacts on terrestrial ecosystems of Antarctica (Campbell and Claridge, 1987; Campbell *et al.*, 1994; Chen and Blume, 1995; Vincent, 1996; Harris, 1998). These impacts include contamination of soils and vegetation, disturbance of wildlife, and import of alien organisms. Campbell *et al.*, (1994) refer to the impact of human-induced disturbance of permafrost by runway construction and active layer removal. Where active layer removal occurred, warming of the exposed ice-cemented permafrost took place causing melt-out and surface subsidence (Campbell *et al.*, 1994). However, an active layer depth was re-established after a few days but with volumetric moisture contents higher than at undisturbed sites.

The moisture content of ice-cemented permafrost was significantly lowered on cut surfaces at Marble Point (77°25'S; 163°41'E) in the MDV (Campbell *et al.*, 1994). No significant re-establishment of icy permafrost has occurred in disturbed soils 30 years since land disturbance. Other manifestations of the release of moisture from the active layer and permafrost during construction activities includes the formation of intermittent streams, thermokarst, and salinization (Campbell *et al.*, 1994).

The polar regions are especially sensitive to global climate change (Maxwell and Barrie, 1989). Bockheim (1993) discussed the potential effects of climate change on terrestrial ecosystems of Antarctica, including changes in ice-free areas, changes in active-layer thickness and moisture content of permafrost, shifts in vegetation and microbial populations, and changes in rates of soil-forming processes such as salinization.

RECOMMENDATIONS FOR FUTURE WORK

The following lines of future investigations are recommended for continental Antarctica:

(1) Establish an observational network of active layer monitoring sites in Antarctica and interface with the Circumpolar Active-Layer Monitoring program (CALM) (Nelson *et al.*, 1996).

(2) Develop a protocol for measuring active-layer thickness and differentiating it from dry permafrost using frost tubes and thermistors.

(3) Develop site-specific descriptions of the relation between active-layer thickness, albedo due to lithology, and climatic parameters.

(4) Investigate and analyze short- and long-term variations in soil temperature and

soil moisture at selected sites.

(5) Measure the thickness of dry permafrost using ground-penetrating radar and vertical electrical soundings (VES) (Guglielmin *et al.*, 1997) and relate it to landform age (see McKay *et al.*, 1998).

(6) Drill and instrument boleholes for long-term measurement of permafrost temperatures.

(7) Monitor soil and rock temperatures to understand the rates of weathering, patterned ground formation, and soil creep (Hall, 1997b).

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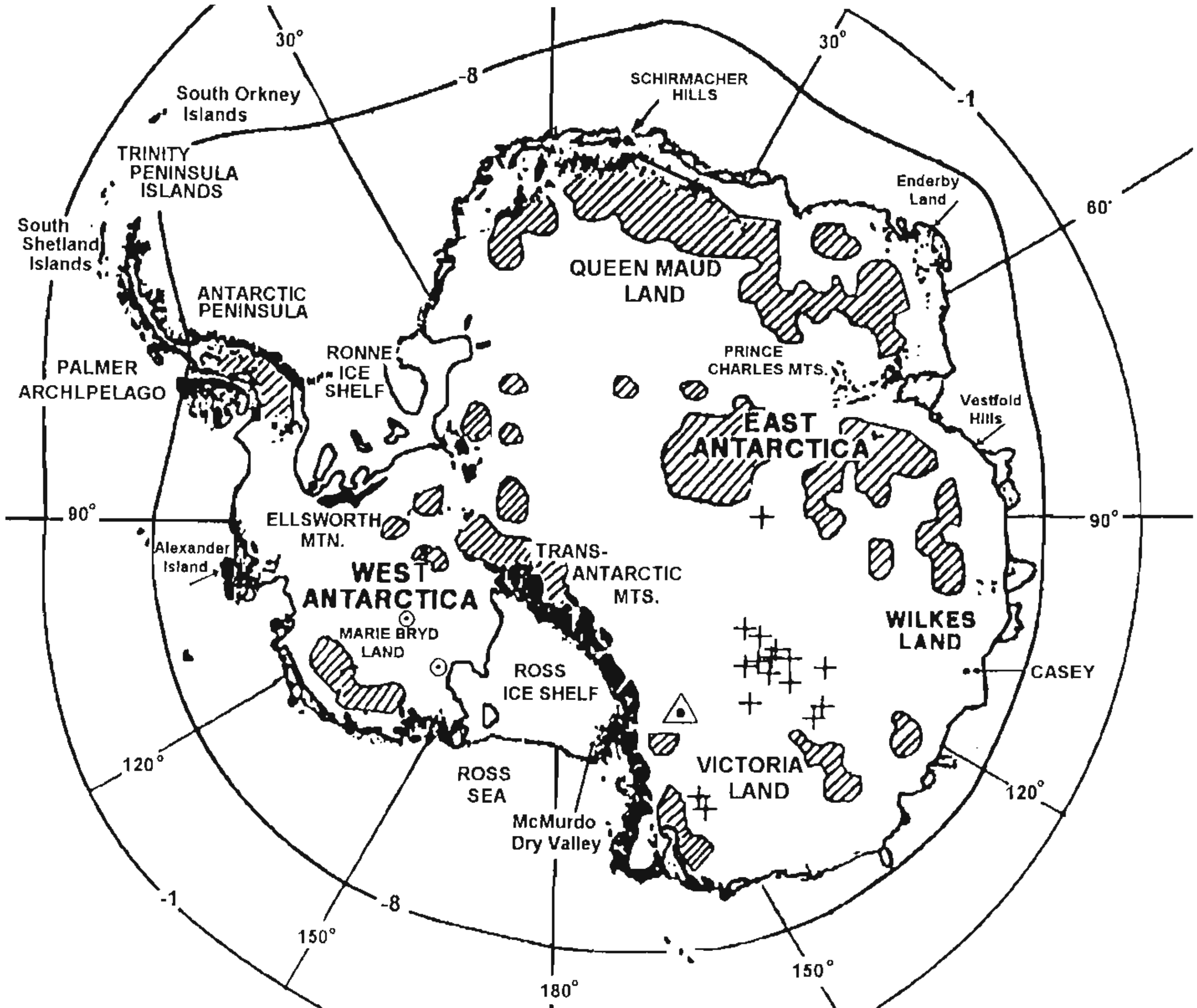
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PALMER ARCHLPELAGO
RONNE ICE SHELF
SCHIRMACHER HILLS
30°
-1
Enderby Land
60°
QUEEN MAUD LAND
PRINCE CHARLES MTS.
EAST ANTARCTICA
Vestfold Hills
90°
ELLSWORTH MTN.
Alexander Island
WEST ANTARCTICA
TRANS-ANTARCTIC MTS.
MARIE BRYD LAND
ROSS ICE SHELF
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ROSS SEA
McMurdo Dry Valley
VICTORIA LAND
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150°
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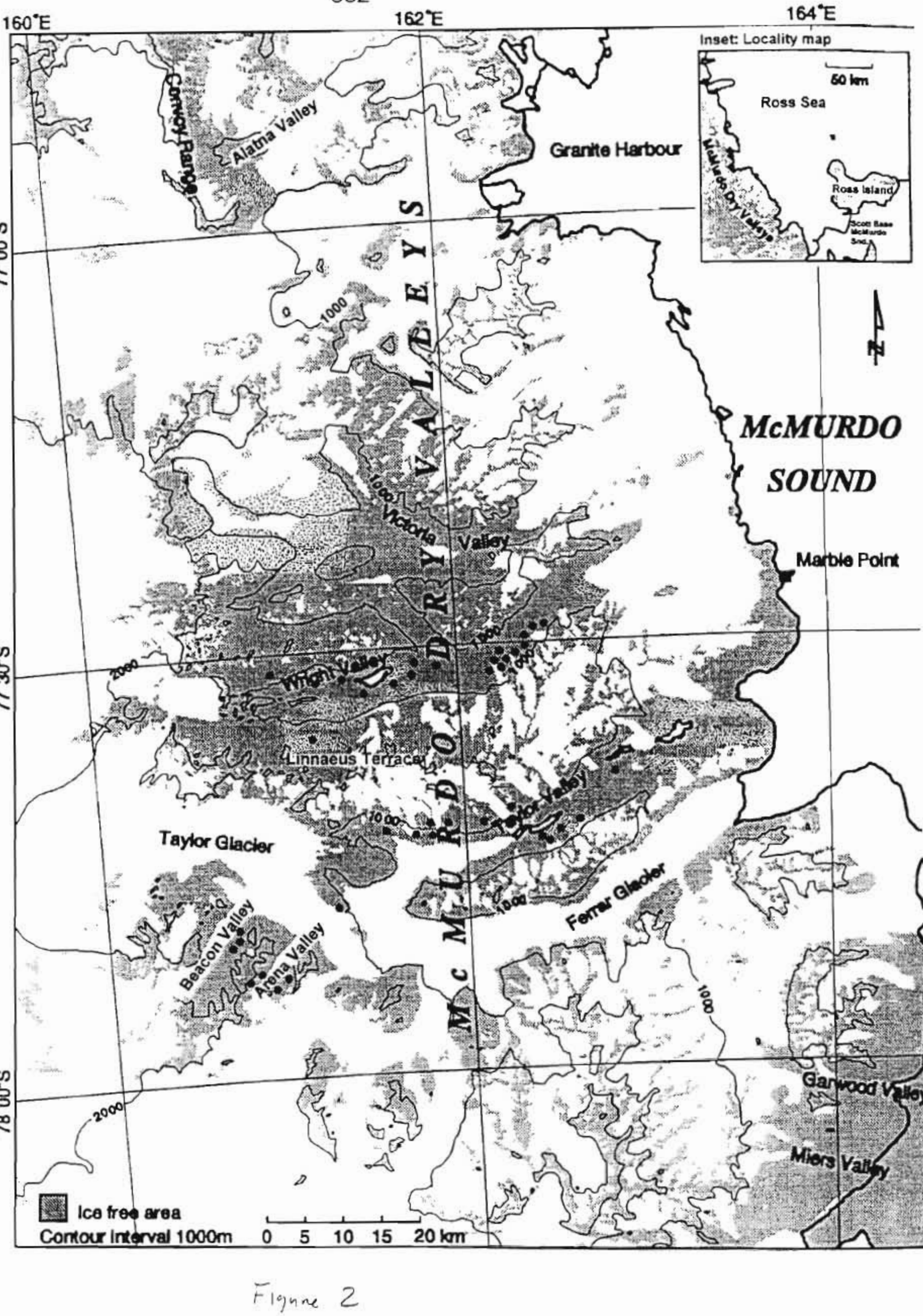
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Granite Harbour
77°00'S
1000
McMURDO
SOUND
Victoria Valley
Marble Point
77°30'S
2000
Wright Valley
Linnaeus Terrace
Taylor Valley
Taylor Glacier
Ferrar Glacier
Beacon Valley
Arena Valley
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Garwood Valley
Miers Valley
Ice free area
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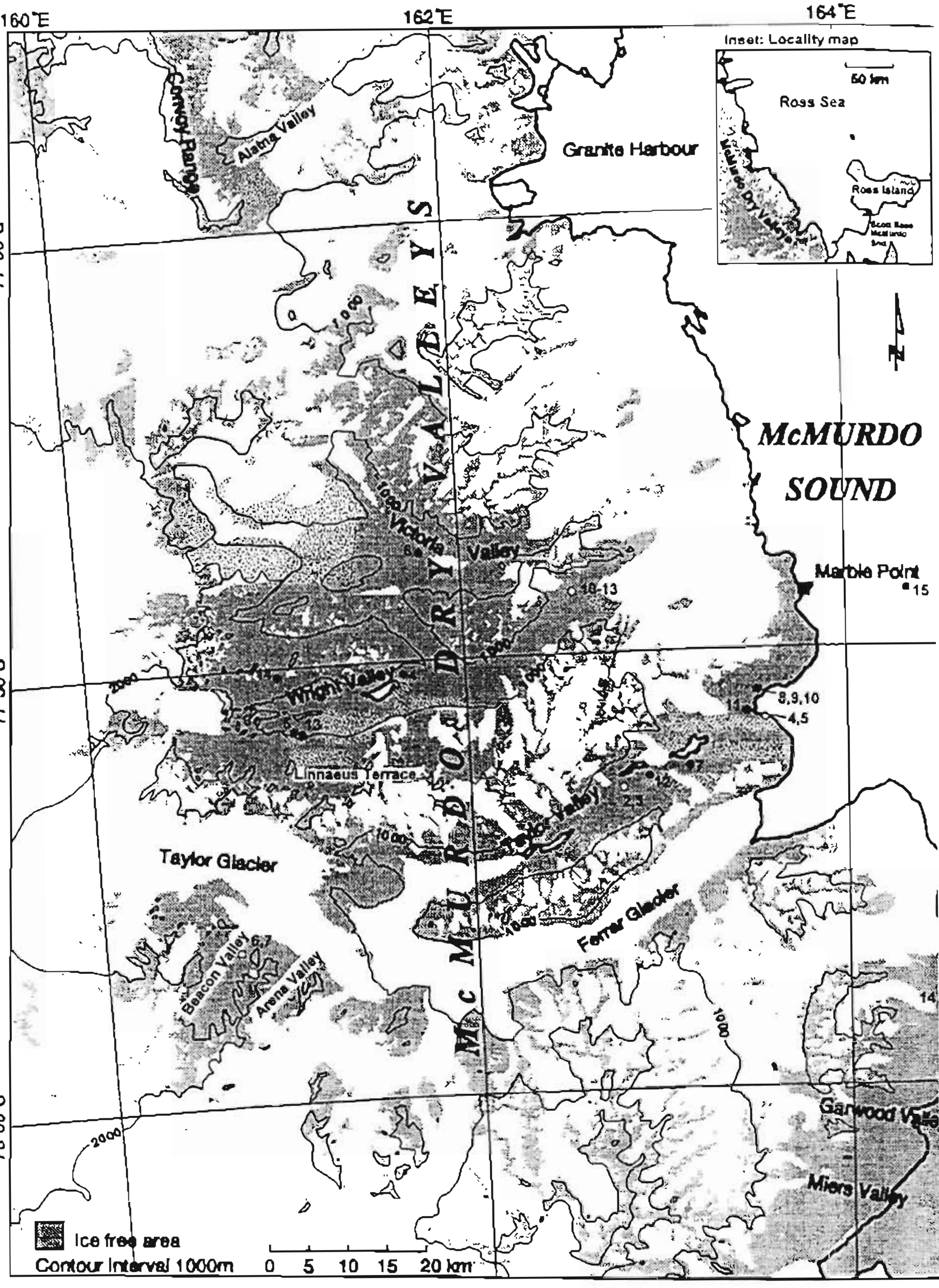
Figure 2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Alatna Valley
Granite Harbour
McMURDO
SOUND
DRY VALLEYS
McMURDO
Victoria Valley
Wright Valley
Linnaeus Terrace
Taylor Glacier
Taylor Valley
Beacon Valley
Arena Valley
Ferrar Glacier
Marble Point
Garwood Valley
Miers Valley
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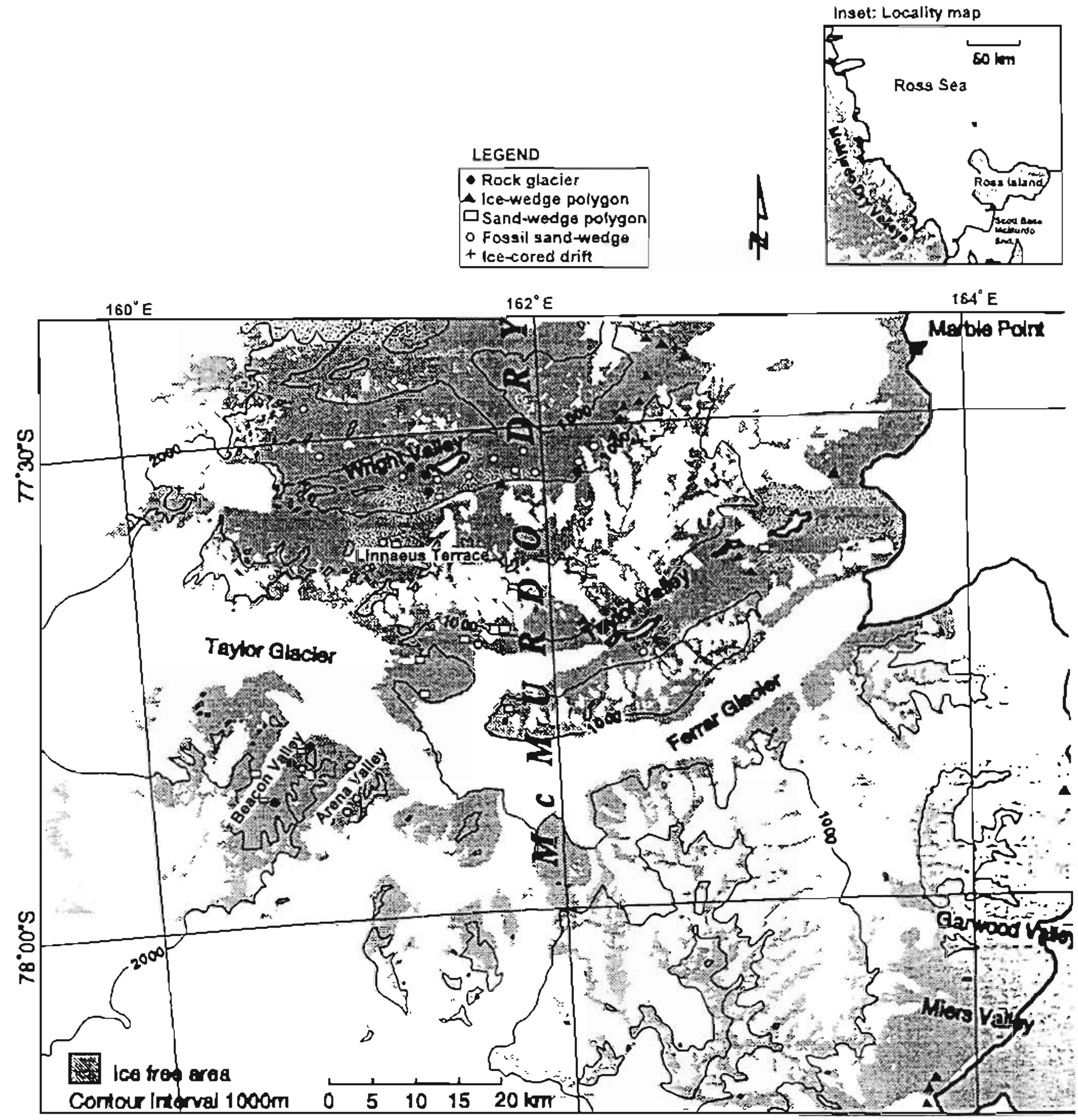
Ice free area
Contour Interval 1000m
0 5 10 15 20 km

Figure 3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Sand-wedge polygon
Fossil sand-wedge
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160° E
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Marble Point
77°30'S
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McMURDO DRY
Wright Valley
Linnaeus Terrace
Taylor Glacier
Beacon Valley
Arena Valley
Ferrar Glacier
Garwood Valley
Miers Valley
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Ice free area
Contour Interval 1000m
0 5 10 15 20 km

figure 4

Table 1. Eco-climatic regions of the Antarctic climate.

Eco-region	Latitude (°S)*	MAAT (°C)*	MAP (mm)*	Vegetation§	Soil taxa
Antarctic Peninsula & islands	61-72	-2.7 to -3.7	400-1077	Grass-herb-fellfield	Haploturbel, Haplorthels, Psammiturbels, Psammihaplels
Maritime East Antarctica	66-71	-9.4 to -11	190-850	Grass-herb-fellfield, maritime moss, lichen barren	Haploturbel, Haplorthels, Psammiturbels, Psammihaplels
Transantarctic Mountains[a]	71-87				
Coastal		-15 to -20	150-200	lichen barren, continental moss	
Inland valley floors		-20 to -25	100-150	lichen barren	
Inland valley sides		-20 to -25	100-150	lichen barren	Anhyturbels, Anhyorthels
Upland valleys		-25 to -30	100-150	lichen barren	
Plateau fringe		-30 or colder	<100	lichen barren	

* Phillpot, 1967; Weyant, 1967; Orvig, 1970.

§ Bliss, 1979.

[a] Campbell et al., 1998b.

Table 2. Permafrost characteristics and ground-ice features in Antarctica.

	Permafrost	Permafrost		Active layer		Ground-ice
Eco-climatic region	distribution	form	moisture content (%)*	thickness (cm)	Moisture content (%)*	features§
Antarctic Peninsula & islands	discontinuous	wet	16-20	40-150	5.8-21	icd, pi, rg, iw
Maritime East Antarctica	continous	wet, dry (?)	6-22	60-150	1.5-28	iw, icd, th, pi
Transantarctic Mountains						
Coastal	continous	wet	6.1-77	30-60	1.0-10	iw, icd
Inland valley floors	discontinuous	dry	1.6-9.5	20-40	0.3-3.0	iw, sw, rg, icd
Inland valley sides	continuous	dry, wet	--	20-30	0.3-2.0	iw, sw, rg, icd
Upland valleys	continuous	dry, wet	--	20-25	0.3-3.0	iw, sw, rg, icd
Plateau fringe	continuous	dry, wet	1.7-30	15-20	0.3-4.0	iw, sw, rg, icd

* Campbell et al., 1998b.

§ Ground-ice features: iw, ice wedges; sw, sand wedges; rg, rock glacier; pi, pingos; icd, ice-cored drift (after Bockheim, 1995).

Table 3. Surface characteristics, temperature, and salt and moisture contents of soils on paired dark and light plots in the McMurdo Sound area. All temperature and albedo readings shown were recorded under cloudless skies (Wilson and Bockheim, unpublished).

Location	Plot no.	Color	Dark:light ratio	Albedo (footcandles)	Surface soil temp. (^{0}C)	Salts, 0-30 cm (mg/cm^2)	Soil moisture 0-30 cm (g/cm^2)
Rhone Platform	1	dark	19	28	17.4	2976	75.7
(77^{0}40'S, 162^{0}25'E)	2	light	0.7	45	5.6	739	45.7
		difference	18.3	17	13.7	2494	30
Hughes Glacier	5	dark	1.7	60	9.6	609	102
(77^{0}42'S, 162^{0}30'E)	4	light	0.5	72	7.4	564	74.0
		difference	1.2	12	2.2	46	28.0
Rhone Platform	6	dark	19	30	20.4	1978	81.5
(77^{0}40'S, 162^{0}25'E)	8	light	1.4	50	15.1	346	33.4
		difference	17.6	20	5.3	1632	48.1
Arena Valley	9	dark	21.7	45	16.6	1594	138
(77^{0}52'S, 160^{0}58'E)	10	light	0.4	72	9.0	912	57.3
		difference	21.3	27	7.6	682	80.3
Arena Valley	11	dark	9.6	45	14.2	420	24.6
(77^{0}52'S, 160^{0}58'E)	12	light	2.1	57	9.7	208	25.5
		difference	7.5	12	4.5	212	-0.9
Sollas Glacier	14	dark	1.1	36	10.5	634	128
(77^{0}41'S, 162^{0}35'E)	13	light	0.6	50	8.7	398	88.5
		difference	0.5	14	1.8	236	39.1
		Probability level*	0.015	0.032	0.040	0.075	0.082

* Comparisons of values within a column for dark- vs. light-colored surfaces, based on ANOV and Fisher's PLSD.

Table 4. Permafrost characteristics in continental Antarctica.

Area	Latitude (0)	Longitude	Eco-climatic region	Permafrost thickness (m)	Permafrost temperature (℃)*	Reference
North Victoria Land	74°45'S	164°E	Transantarctic Mtns.	3-20		Guglielmin et al., 1997
Seymour Island	64°15'S	56°45'W	Antarctic Peninsula	48-180		Fukuda & Corte, 1989; Koizuma & Fukuda, 1989; Fournier et al., 1987
King George Island	62°12'S	58°57'W	Antarctic Peninsula	20-100		
Dry Valley Drilling Project						
McMurdo (#3)	77°51'S	166°40'E	Trans.-coastal	440-500	-15	Decker and Bucher, 1977
Lake Vanda (#4)	77°31'S	161°32'E	Trans.-inland valleys	0§		Decker and Bucher, 1977
Lake Vida (#6)	77°23'S	161°48'E	Trans.-inland valleys	800-970	-24	Decker and Bucher, 1977
New Harbor (#8, 10)	77°35'S	163°31'E	Trans.-coastal	240-310	-14, -15	Decker and Bucher, 1977
Commonwealth Glacier (#11)	77°35'S	163°25'E	Trans.-coastal	405	-18	Decker and Bucher, 1977
Lake Leon (#12)	77°38'S	162°51'E	Trans.-inland valleys	360	-13	Decker and Bucher, 1977
Don Juan Pond (#13)	77°33'S	161°10'E	Trans.-inland valleys	0§	-16	Decker and Bucher, 1977
North Fork (#14)	77°32'S	161°24'E	Trans.-inland valleys	350-360	-16	Decker and Bucher, 1977

* 50 m depth; measured using electrical resistivity or electromagnetically.

§ Thermally stratified lake (Cartwright et al., 1974).

Periglacial and permafrost landforms and processes of the Southern hemisphere.

Jan Boelhouwers and Kevin Hall

Introduction

This special issue of the *South African Journal of Science* contains a set of review papers to provide an overview of current understanding of present-day and Quaternary permafrost and periglacial environments in the Southern Hemisphere Working Group of the International Permafrost Association. This working group was established in 1998 with the primary aim to facilitate information exchange between researchers within the Southern Hemisphere, as well as with their counterparts in the Northern Hemisphere. In addition, the WG aims to synthesize and make more accessible existing information on this topic from the Southern Hemisphere.

International awareness of Southern Hemisphere permafrost and seasonally frozen ground is very limited. However, extensive areas of permanently or seasonally frozen ground exist in the Southern Hemisphere, including the Antarctic, the Andes and Patagonia and the New Zealand Alps, as well as the mountain summits of Irian Jaya and Papua New Guinea. Evidence for Quaternary periglacial activity is also documented for the mountains of southern Africa and Tasmania, with limited current activity. Reasons for the poor recognition of these areas may be in part, as in the case of South America, because the research may be published in a language not easily understood by the English-speaking world, or be inaccessible by publication as 'grey' literature. In addition, the research itself may have progressed in isolation from the science development and debates in the NH, where most of the permafrost terrain, and resources for their study, are concentrated. In the case of the Antarctic, including both the continent and peri-Antarctic islands, permafrost science has seldom been advanced through dedicated programmes. Rather, information is found scattered through the life and earth science literature and not easily found. An important limitation on the output of Southern Hemisphere permafrost science is the small capacity, both in terms of human and financial resources in many countries, relative to the Northern Hemisphere.

The outcome of this separate development is a general lack of awareness of issues pertaining to SH permafrost and periglacial science. Environmental conditions in which frost processes occur are, in some instances, very different from those encountered in northern high latitude environments. This poses new questions to process understanding and the palaeoenvironmental significance of resulting forms, as exemplified by the blockstreams of Tasmania and southern Africa. The non-periglacial interpretations of these forms in Tasmania have stimulated reconsideration of blockfields and blockstreams origin in northern high latitude environments. Similar examples can be forwarded for diurnal soil frost processes at low- and mid-latitudes, and weathering in hyper-arid Antarctic environments. Because of the differences Southern Hemisphere perspective have, the potential to substantially contribute to the scientific development of understanding of basic driving mechanisms and boundary conditions in permafrost and periglacial processes.

The present collection of reviews offers a first comprehensive access point in English to the literature on Southern Hemisphere permafrost and periglacial research and the issues contained therein. As such, it offers a frank assessment of the current status of periglacial research in various regions. From the different papers it will be clear that big differences exist in the level of scientific understanding that has been achieved to date. This can be easily understood, taking cognisance of the specific context in which scientists in various regions have developed their work, as outlined in general above. Even so, these papers offer an important contribution in that they emphasize the existence of past or present periglacial environments, raise directly or indirectly the problems concerning research in these areas and, as such, point at the potential contribution of research in these areas.

The significance of raising awareness for SH permafrost goes beyond the purely scientific. Climatic warming impacts are expected to be very significant in the permafrost regions of the world. Already, permafrost thawing has led to major landscape changes and terrain instability and has triggered widespread interest from the research community, such as in the EU-funded PACE project. To date, this work is largely focused on the NH. Attention is drawn here to the fact that the SH possesses equally large and ecologically important and sensitive areas of permafrost and periglacial terrain. Current changes in permafrost in these areas are largely unknown as are the potential responses in landscape dynamics. Thus, there is a real need to understand current changes in permafrost conditions, given the potential impact this may have on erosion rates, natural hazards and ecosystem functioning in sensitive areas such as the Andes, New Zealand Alps and the Antarctic. Already, for the Antarctic, a dedicated permafrost-monitoring project has been started and closer collaboration between the permafrost community and SCAR is being developed. International participation in the IPA/Global Terrestrial Network for Permafrost is being sought to extend permafrost temperature monitoring in the Southern Hemisphere.

It is hoped that this special issue will contribute to raise awareness in permafrost and periglacial issues in the Southern Hemisphere and highlight the scope for meaningful research in these areas.

Jan Boelhouwers
Kevin Hall

Permafrost and Periglacial Processes, Vol 2: 3 (1991)

Introduction

Geocryology of the Americas—IGCP Project No 297

While there is a large store of information pertaining to active or past geocryological conditions and processes from the Northern Hemisphere, the same is not the case for the Southern Hemisphere. However, that information which does exist not only enhances the general body of knowledge but also adds information either unique or specific to the Southern Hemisphere. For instance, from Antarctica there is information regarding the formation of sand wedges, and in the Puna de Atacama, South America, at 23°S latitude, ice islands in shallow salt lakes have been reported. In Patagonia, a complete glacial chronology has now been correlated with cryogenic events in the eastern Andes at 33°S latitude. In addition, the filling of wedge casts with powdery calcium carbonate is considered to be a cryogenic feature of dry environments.

The International Geological Correlation Program Project No 297 is being supported by UNESCO and the International Union of Geological Sciences. During the first meeting of Project No 297 in Mendoza, Argentina, 16-20 October 1989, researchers from South Africa and the Americas agreed to undertake research regarding geocryogenic characteristics on both sides of the Atlantic (i.e. in South Africa and Argentina). Both areas contain the same Gondwana rocks located at similar elevations and at about the same latitudes. Accordingly, a second meeting of Project No 297 was held in South Africa between 5-17 September 1990. Hosted jointly by Rhodes University, Grahamstown and The University of Natal, Pietermaritzburg, the meeting included nine days of field excursions in the Drakensberg Mountains. Papers were presented over four days and a field guide to the geocryogenic features of the Drakensberg region, edited by Patricia M. Hanvey, was used as the basis for the field discussions. This volume of papers is the result of this international symposium.

It was agreed that geocryogenic processes are currently active at high altitudes in southern Africa and that there is evidence for their more widespread activity during the Quaternary. However, the type and intensity of the processes operating during the Quaternary needs further analysis.

The Organizing Committee consisted of the Chairman, Colin A. Lewis (Rhodes University) together with H. Beckedahl (University of Transkei), J. Boelhouwers (University of the Western Cape), Kevin Hall (University of Natal), P. M. Hanvey (University of the Witwatersrand), and M. Marker (University of Fort Hare). Overall, the meeting was a well planned event for which the Organizing Committee must be sincerely thanked.

We hope that the following papers will provide additional useful information with respect to cryogenic events and processes in South Africa, South America and elsewhere

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North-facing hollows in the Lesotho/Drakensberg mountains: hypothetical palaeoenvironmental reconstructions?

Sir. — The conspicuous hollows on the slopes of the eastern Lesotho plateau and adjacent highland regions have been the focus of much discussion.[1–6] Nivation and cirque glaciation have both been proposed to explain the origin of these hollows. The prevailing use of the nival/glacial-related hollow origin in southern African palaeoenvironmental reconstructions should be of concern to Quaternary scientists as these hypotheses have not been proven. The recent construction of a Pleistocene periglacial gradient in southern Africa,[7] based notably on these hollows, calls for an urgent reconsideration of such landforms, particularly as they may imply a different suite of evolutionary and palaeoenvironmental events in the landscape from what has been commonly postulated. We outline here some of the numerous problems of the nivation hollow origin and recommend an alternative hypothesis.

The term nivation has been used either as 'a collective noun to describe a suite of geomorphological processes associated with late-lying snow, or as an adjective to identify the origin of the ensuing landform' (ref. 8, p. 5). Thorn,[9] however, concluded that 'in essence, it is a term which has served its purpose, but now clouds rather than clarifies the issue' (p. 424), and more recently suggested that the term be abandoned owing to its lack of an acceptable operating definition and our lack of understanding regarding some of its central tenets (notably freeze-thaw weathering).[8] Thorn[10] calculated a slope degradation rate of 0.0074 mm yr^{-1} for a nivation hollow site and later concluded[9] that perennial snowpacks probably decrease the absolute denudation rates. With the Lesotho hollows being up to 800 m wide and 1207 m deep,[1] a nival origin would require the presence of perennial snowpacks for several hundreds of thousands of years!

Dyer and Marker,[4] and Marker[5] argued that the hollows in the eastern Lesotho highlands are restricted virtually to land above 3000 m and that 75% of these are north-facing. This suggestion of a predominance of hollows facing equatorwards immediately questions the proposed nivation or glacial origin. It seems unlikely that snowpatches should survive longer on the warmer north-facing slopes than on the cooler south-facing slopes, unless a snowfence effect occurred with substantially greater snowfalls than at present; even then, south-facing hollows would be expected due to this greater snowfall. The idea of large quantities of snow drifting into leeward positions also raises problems. Contemporary anticyclonic winter snowfalls on the plateau are frequently accompanied by strong winds and cold snow conditions where daily maximum temperatures of −3°C have been recorded, yet there is little, or sometimes no, snow drifting into leeward positions (personal observation from 1992 to 1995). Further, the steep south-facing slopes which are characterized by rock scarps near their summits are not conducive to leeward drift, but rather to trapping snow as drifts at the base of the south-facing rock scarps, as is frequently found during recent winters.

It is important to recognize that late-lying snowpatches occur in pre-existing hollows and are unlikely to have produced the hollow in which they occur. The Lesotho/Drakensberg hollows developed over a long period, so it is difficult to record the relative contribution of glacial or periglacial processes when unequivocal evidence (for example, glacial striations or protalus ramparts) is absent. Unless it can be shown quantitatively that nival/glacial-related processes were the dominant palaeo-weathering and/or palaeo-transporting agents at the hollow sites, then such landforms cannot be referred to as nivation or glacial hollows.

The idea of 'bog cirque' development was proposed by Dzulynski and Pekala[11] for landforms in north-eastern Mongolia and is supported by observations and measurements of active processes forming the hollows. Most of the climatological and geomorphological attributes described by Dzulynski and Pekala are strikingly similar to those encountered on the Lesotho plateau, and include: 1, summer rainfall and winter drought; 2, low mean annual air temperatures and a scarcity of snow; 3, intense diurnal and seasonal freeze-thaw cycles; 4, hollow morphometry resembles that of so-called nivation hollows and are prominent features of the landscape; 5, the hollows are irregularly spaced, have arcuate rock benches and basin-like depressions; 6, the hollows develop preferentially on the sun-exposed slopes; 7, the floors of hollows are boggy and host peat-forming vegetation; 8, the hollows commonly merge into trough-like or flat-floored valleys; 9, the hollows occur on plateau-like summit surfaces at high elevations.

Owing to shading on south-facing slopes, there would (temporally) have been considerable snow cover, relatively few temperature fluctuations, and consequently, restricted freeze-thaw oscillations on such Lesotho/Drakensberg slopes.[12,13] Conversely, the restricted snow cover and high radiation receipts on north-facing slopes has permitted considerable temperature fluctuations on these slopes.[13] Consequently, at seepage sites (associated with areas of high joint density) mechanical and chemical weathering has been enhanced and mass movement promoted. It would thus seem that the prominent occurrence of north-facing hollows on the Lesotho plateau is a result of a set of synergistic lithological and climatological factors. The bogs which have emerged within many north-facing hollows display diatomaceous earth,[14] peat and gravel horizons[14–17] and are possibly the key to the understanding of hollow evolution. The presence of diatomaceous earth and peat at several hollow sites is an indication that not only mechanical weathering, but chemical breakdown of the basaltic parent material had been an active process at these sites for considerable periods during the Quaternary. The presence of bog vegetation would have increased groundwater acidity and thus enhanced chemical weathering. Biological weathering components would also have helped increase overall weathering rates. The gravel units may attest to phases of increased solifluction and debris removal.

The nival/glacial hypotheses are faced with significant problems when applied to north-facing Lesotho hollow development and should therefore be abandoned until unequivocal evidence is available. An alternative hypothesis of ground seepage at lithologically favourable sites and consequent enhanced weathering and transportation processes would overcome most of the reservations associated with the nivation or glacial hypotheses but still produce such hollows as are presently found. This calls for a re-evaluation of high altitude palaeoenvironmental theory in southern Africa. Because prevailing conclusions about the Pleistocene periglacial gradient in southern Africa[7] are highly reliant on the Lesotho/Drakensberg hollows, they should be revised, based on less ambiguous evidence.

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Alternative approach to the treatment of addictive withdrawal

Sir, — Professor Pick highlights the importance of alternative and novel approaches to health care in the new South Africa.[1] This notion should have wide applications. Unfortunately, history shows that this is easier said than done. Over the centuries mankind has a very poor record of readily accepting and adopting new approaches to traditional problems. As O'Dowd has written, '... the pre-eminently persecuted minority in history has been the minority of the innovators'.[2] Although a recurrent and tiresome problem for the innovators, it goes much further, because it retards human progress in a more general sense. A good current example is the failure of any academic department of psychiatry or provincial hospital in South Africa either to test or apply analgesic nitrous oxide therapy for addictive withdrawal states, although the treatment was pioneered and researched here. Moreover, the originators have been willing to assist or cooperate in any appropriate manner.[3] Instead, the work has been actively and improperly impeded by those very people who should be the most open-minded.[4,5] Abundant evidence indicates that the therapy is rational, safe and effective. It has been tested by various investigators, in many thousands of individuals suffering acute withdrawal from alcohol and other substances of abuse, with well over 15 000 having already benefited in South Africa and Finland. Apart from the unrivaled speed of recovery, following its use, it is also safer than currently accepted techniques because it drastically reduces reliance on addictive sedatives, such as the benzodiazepines, and provides an efficient screening test.[6] The screening test has important economic advantages because it enables the majority of withdrawing patients to be treated as out-patients, who can continue their economic activity while receiving therapy. The procedure also has the potential of saving the country millions of rands by preventing bed occupancy by those who can safely be treated as out-patients.

Although not yet used by any academic department in South Africa, the therapy is being successfully employed at numerous SANCA (South African National Council Against Alcoholism and Drug Addiction) clinics, in private practice and at the largest non-governmental alcoholism treatment facility in the country (Rand Aid Association, Wedge Gardens, Edenvale, Gauteng). The therapy has been researched by SABRI since 1979 and has been officially recognized by the Medical Association of South Africa, who have granted it an official tariff in their guide to fees. Consequently, patients who use the treatment can claim refunds from their medical aid societies should they so wish.

While no one should dictate a physician's choice of clinical approach, it would seem inexcusable that a new, effective and economical primary health care therapy is not even put to the test by those who caste aspersions at it,[4,5] especially in a country that is prepared to limit the use of advanced medical procedures in favour of funding primary health care.[7]

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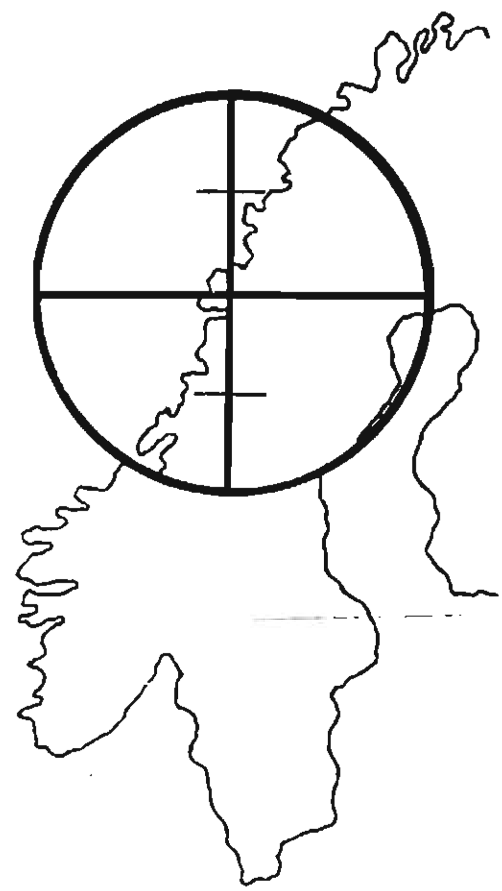
Okstindan
Research Project
1972

Preliminary Report

Reading University



AN INVESTIGATION OF NIVATION PROCESSES AT A LATE-LYING SNOW-PATCH IN AUSTRE OKSTINDBREDAL

by

Kevin Hall

ABSTRACT

An investigation has been embarked upon in an attempt to assess the processes responsible for nivation in the Okstindan area of northern Norway. Detailed instrumentation of a late-lying snow-patch site has been undertaken in such a manner that a continuous recording of various parameters within the area throughout the year might be obtained. A continuous record of air, ground, and rock backwall temperatures has been attempted, and a record of the type and amount of ground movement over the year is being recorded. It is suggested that nivation is probably an active process in the Okstindan region.

1. INTRODUCTION

Nivation is the general term to describe the weathering and disintegration of the ground where a snow-patch persists into or through the summer and the subsequent removal of that broken material so as to create a distinct hollow. It is a process which can occur in cold environments ranging from the arctic to regions of high mountains in low latitudes. It is a term often misunderstood or misquoted and has thus come under criticism, (see, for example, Tricart's (1963) comment .... "nivation; a useful word which explains nothing") but which nevertheless still has a place in periglacial geomorphology.

One of the first workers to study this, though he did not suggest a term for it, was the Russian I.P. Tolmachev, in 1899. The actual name 'nivation' was suggested by Matthes in 1900 after his investigations of cirque-like landforms in the Big Horn mountains of Wyoming. Matthes suggested that temperature fluctuations across the freezing point were greater and more frequent in the moist zone of contact between the melting snow mass and its periphery and thus frost shattering and the comminution of material was at its maximum in that zone. This work was followed by studies made by Ekblaw in Greenland. Ekblaw (1918) stated that nivation was ...... "the process by which quiescent névé affects the disintegration of rocks, and the destruction of some landforms and the formation of others. In this process the snow itself produces very little, if any, effect; it is the water from the gradual melting of the snow that does the work". He went on to describe three main forms of snow-patch, the dome-shaped, wedge-shaped, and piedmont, and suggested that here it was solifluction that was the main agent of debris removal from the depressions that resulted from the different types of snow-patch.

1.1 Nivation Processes

The dominant process involved in nivation appears to be that of repeated freeze-thaw, and the first detailed studies of this in connection with snowpatches were those of W.V. Lewis (1936, 1939) in Iceland. Lewis defined three types of snow-patch, basically the same as those of Ekblaw, a circular, transverse (to lines of drainage), and longitudinal form (Figure 1). The transverse was considered to occur at the break of slope between rocks of different hardnesses; the longitudinal was associated with a stream course or some other factor, such as a fault, which gave a depression running downslope; the circular was best develop on a gently sloping surface where there was no pronounced variation of geological structure. Lewis attempted to study in detail the temperature fluctuations acros the freezing line and the effects of these on the area around and under the snow-patch. He concluded that although snow is basically an insulator of the underlyi ground from temperature changes, it was possible that meltwater tunnels or caves beneath the snow might provide an access for cold or warm air from outside. He went on to suggest that the large amount of meltwater running beneath the snow was capable of transporting material up to the size of medium sand even when the gradient was small. However, it was stated that these melt runnels were not capa of erosion of the bedrock but only able to move loose material.

McCabe (1939) working in Spitsbergen as a continuation of Lewis's work found that the streams flowing out from beneath the lower ends of the snow-patch "were n observed to be moving any material". Botch (1946), working in the northern Urals observed that from the lower margins of snow-patches issued streams of "yellowish mud" which formed a network over the "pavement" below the lower snow margin. However, below these and possibly occurring under the snow, solifluction was considered to occur so long as the slope angle was less than 40 degrees. Frost action was found to occur on the rockwall at the back of the patch and many angula blocks were seen to be lodged in the gap between the snow and the backwall, fines here being absent due to movement away by meltwater. As the patch became smaller the coarse material was left free to slide over the top of the snow, leading to th accumulation of protalus ramparts at the base. He felt that water movement under the snow might cause temperature changes, and suggested that the ground under the snow thaws out thus enabling mass movement to occur even though the air temperatur above the snow remained below 0°C. As a consequence, movement seemed to be more continuous here than on the margins; and it was this action which he saw as being the cause of the growth of a hollow (Figure 2).

Since freeze-thaw activity is thought to be of major importance within the nivation process, a number of experiments undertaken on this topic help to throw some light on the facts and problems involved with the process. Tricart (1956) a Wiman (1963) attempted a quantitative evaluation of the disintegration of rock by repeated freezing and thawing. Rock samples were subjected to what were called "Icelandic" and "Siberian" climatic fluctuations in the laboratory. The Icelandic was characterized by a slow fall of temperature to just below zero (+6°C to -7°C over 36 hours), whilst the Siberian was a rapid fall to well below zero (2 hours from +10°C to -30°C). Dry, humid and wet conditions for both were investigated. Dry conditions were found to produce no weathering whatsoever. Martini (1967) experimented further on the Icelandic type and concluded that the rocks most susceptible to frost weathering were sandstones, porphyraceous granites and schists and the most resistant were the crystalline limestones and eruptive rocks. The actual amount of disintegration was considered to be a function of temperature fluctuations, structure, amount of chemical weathering and water availability. In connection with this are the interesting field observations of Gardner (1969) who analysed the influence of a snow-patch upon the temperatures on the rock backwall it was in contact with. He found that it lowered both the maximum and minimum daily means and that freeze-thaw occurred more frequently in a restricted zone at the edge of the snow than on the rockface above. The result o this was to create a "niche" on the backwall (Figure 3).

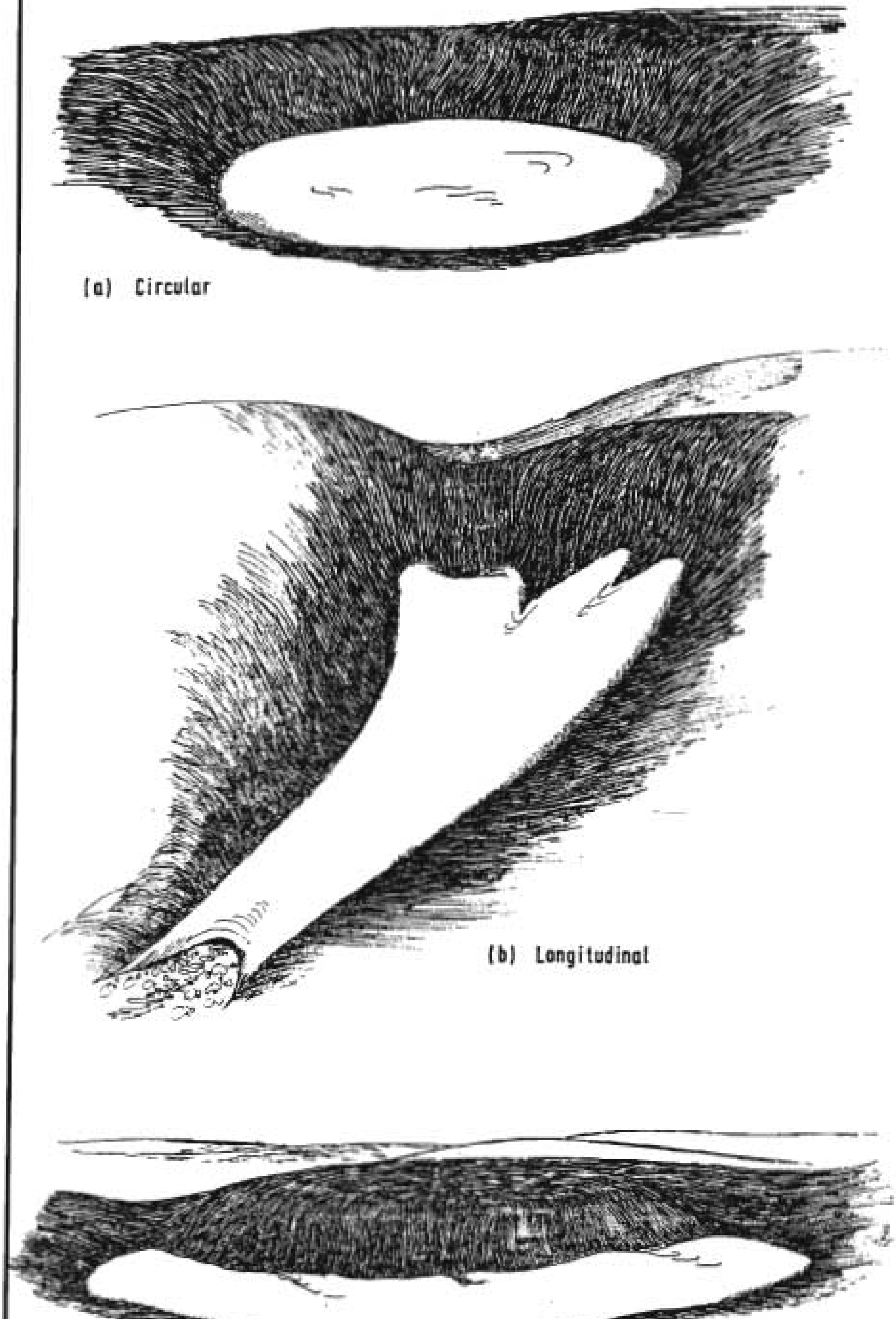


Figure 1

SKETCHES ILLUSTRATING THE SNOWPATCH CLASSIFICATION PROPOSED BY LEWIS (1939)

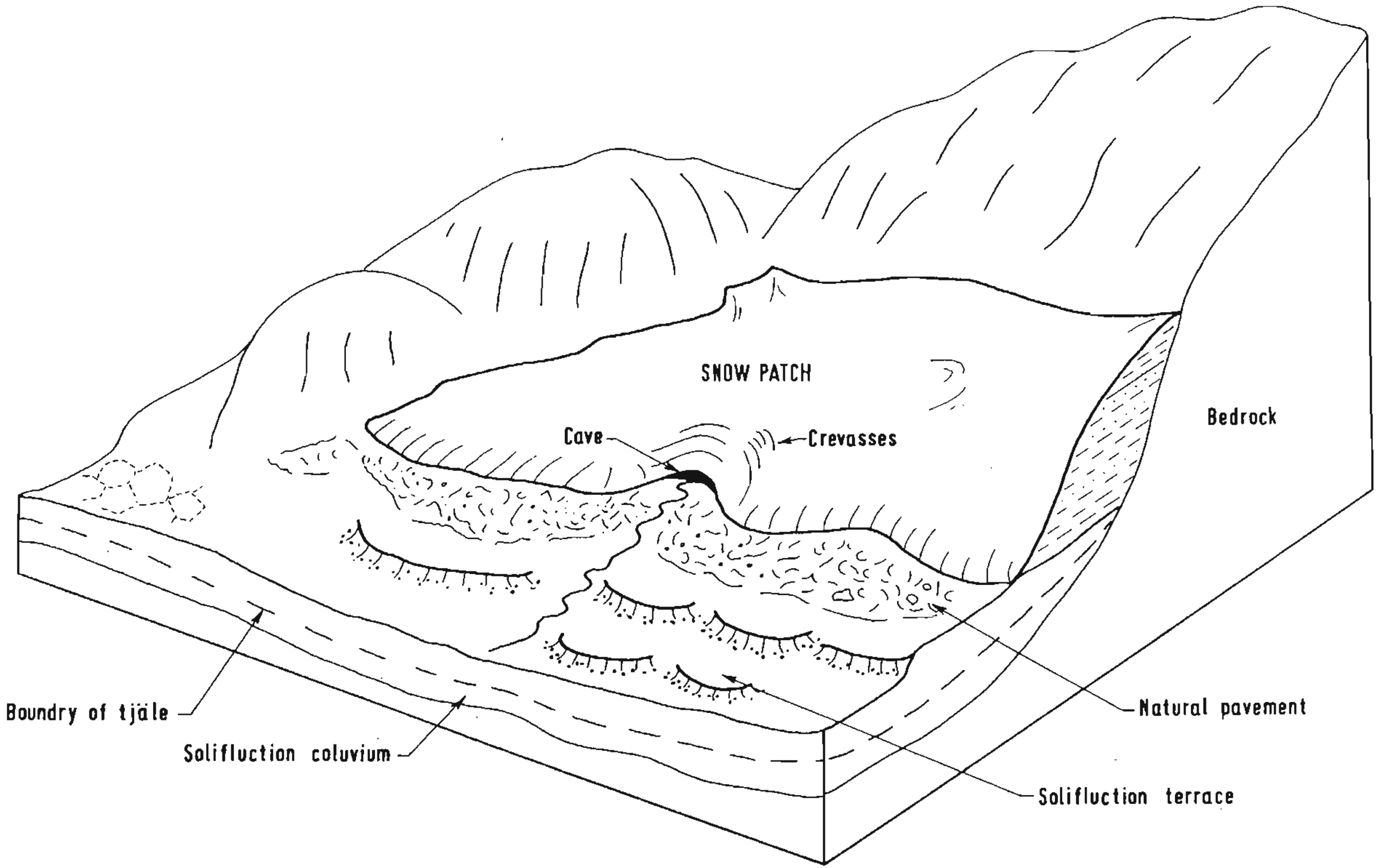


Figure 2 SNOW-PATCH IN MELT SEASON (AFTER BOTCH 1946)

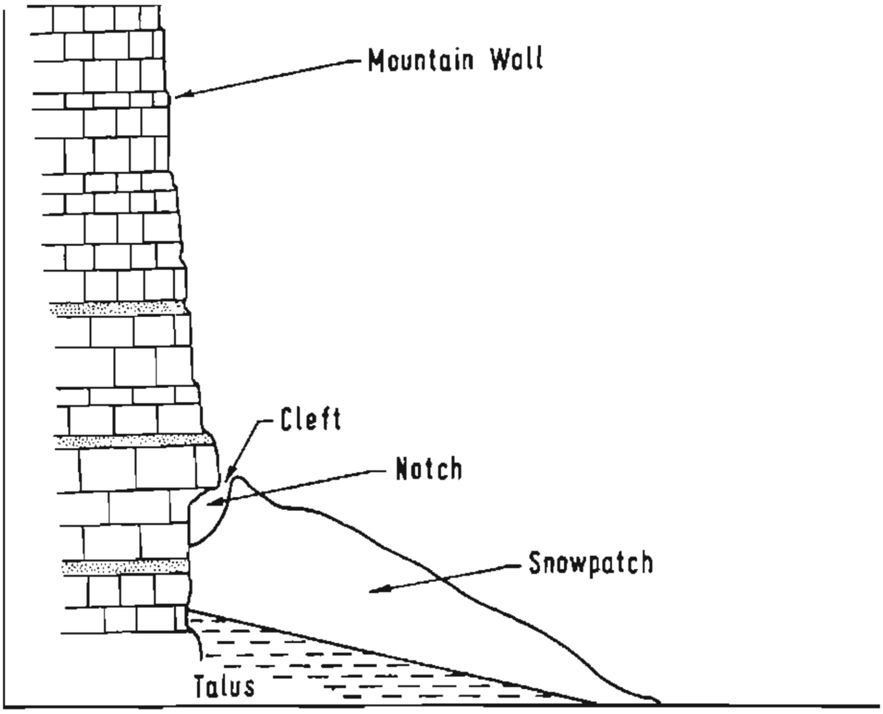


Figure 3 THE RELATIONSHIP OF THE SNOWPATCH TO THE MOUNTAIN WALL
(AFTER GARDNER)

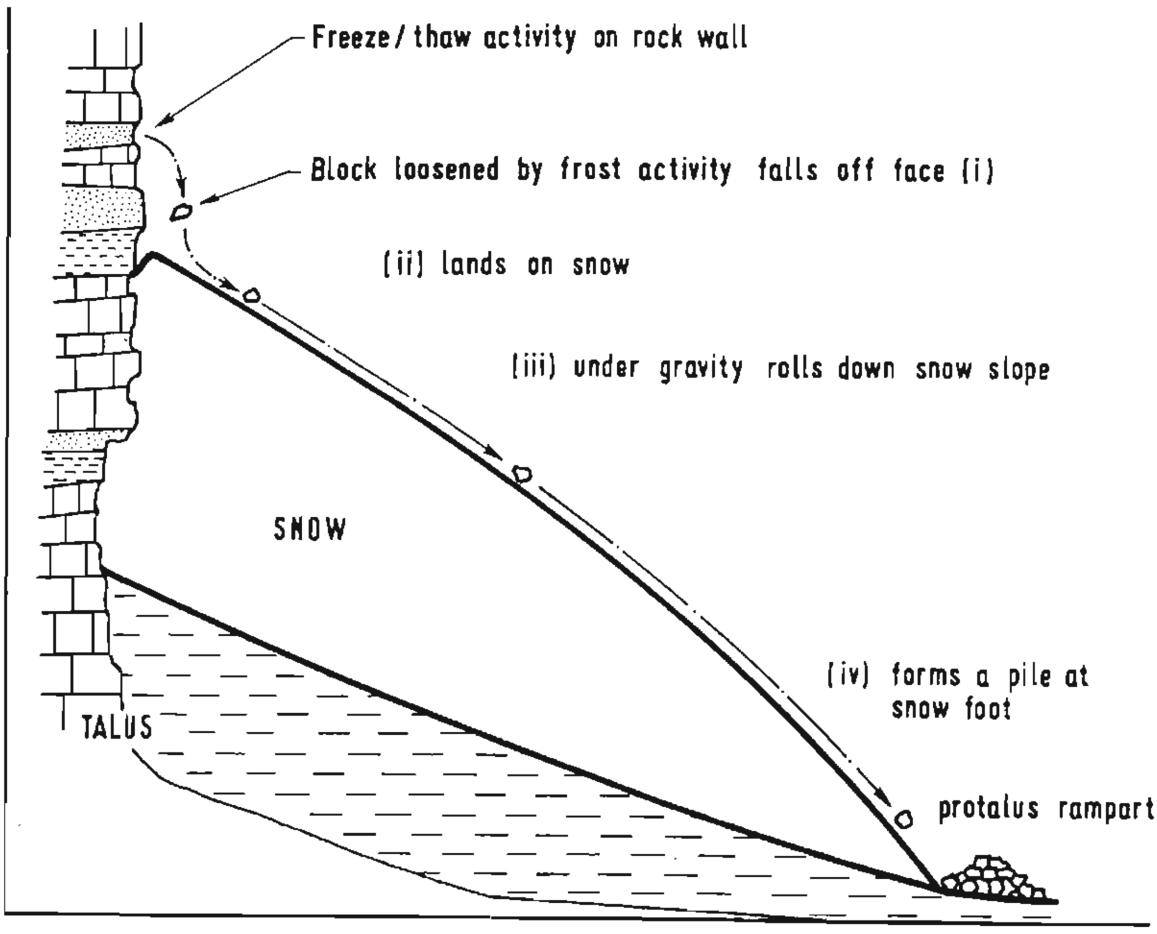


Figure 4 FORMATION OF 'PROTALUS RAMPART' *(AFTER BRYAN, 1934)*

Hewitt (1968), in his study of the climatic environment of the Karakorum Himalaya, expressed very clearly the importance of variations of temperature around $0^{o}C$. He showed that an analysis of freeze-thaw cycles requires consideration of:

(a) absolute or mean number of frost shifts in given period;

(b) frequency of cycles comprising
  (i) wavelength,
  (ii) recurrence intervals or "density" of cycles;

(c) intensity of freeze-thaw, depending upon
  (i) amplitude of temperature change,
  (ii) slope of temperature curve;

(d) scale relations of freezing and thawing phase;

(e) the problem of what constitutes the "effective" temperature shift for a given process and what is the best measure of it.

The problem of what constitutes an effective temperature shift was something also studied by Cook and Raiche (1962) in field studies in Canada.

1.2 Nivation Landforms

Most landforms associated with nivation are encompassed within the classification of Lewis (1939), noted above. The transverse snow-patch and associated hollow, is usually structurally determined, and it lies along the contours of a slope. Snow-patches of this type are thought to be responsible for an altiplanation terrace form (Demek, 1969). They widen notches and shelves on mountain slopes (Ahlmann, 1919) and often pick out the strike of sediments, giving this altiplanation terrace-like form, as noted in Spitsbergen by Waters (1962).

The longitudinal forms often follow fluvially eroded gullies and here it is a problem to determine which is the dominant process - fluvial or nival? It is a problem made more difficult by the fact that a proportion of the water running in the gulley is a direct consequence of the melting of the snow. However, many gully forms which have snow-patches for a significant period of time are often widened out into a conical form in the upper reaches, and frequently the water falls over a very large weathered step which is not determined by structure. This step is often far wider than the water channel, and suggests action of the snow-patch at this point. Ahlmann (1919) termed special forms of this type "raaskars" and Groom (1959), working in Spitsbergen, mentions forms of this type which contain ice maintaining angles of up to 40^{o} which she terms "niche glaciers".

Circular patches appear to be independent of any structural or fluvial control and a possible reason why they are round is the tendency of the snow to present the least periphery per unit area and thus it tends towards a circle. The result is to create circular hollows or arcuate scars.

Often associated with the snow-patch is a depositional form known as the "protalus rampart" (Bryan, 1934), which is a debris ridge formed at the bottom of a snow slope by movement of the broken material over the surface of the snow to its foot (Figure 4). Solifluction is also associated with nivation and this, as is well known, causes distinct landforms. On slopes of less than 40^{o}, many solifluction lobes can be seen associated with the snow-patch area (Botch, 1946). Other features of slope failure, such as terracettes, may also be seen in the snow-patch area, together with areas of fine material caused by frost action, termed "aleurite" (Kachurin, 1959). These have been washed out from under the snow to lie in distinc channels or fans.

2. FIELD WORK 1972

2.1 Background

The summer field season was planned to be used mainly for the installation of the instruments at a chosen snow-patch site. If time allowed after the instrumentation of the patch, then a number of general observations of the snow-patches in the area were to be made. However, a number of instrumentation problems were encountered and thus this took longer than was anticipated. Although the field station was opened in early June, the instruments did not arrive until late June and consequently no measurements of the ablation during the early period were possible.

The snowfall for 1971/1972 was greater than usual and most of it fell late in the winter and thus there was a thicker than normal covering of snow in the hollows. However, the weather in early May and June was unusually hot and dry and thus although there was a large amount of snow it melted very quickly. By the beginning of August all but the higher snow-patches had melted away; even some of the lower ones which appear normally to last throughout the summer had melted completely. Thus, although the period during which nivation was active was much reduced, it did allow the whole snow-patch area to be instrumented without the problem of late lasting snow covering a proportion of the area.

2.2 Site Description

In 1972 the instrumentation was concentrated in a single site, since it was considered best to monitor one area adequately. The site chosen was on the north facing side of the valley Austre Okstindbredalen.

The upper margin of the snow-patch lay approximately at 850 m, whereas at its lowest point it was at a height of approximately 710 m. However, the snow-patch under study was not a simple form but a complex one consisting of an upper transverse patch with a longitudinal drift running from the lower edge of the transverse patch, along a gully, down into Austre Okstindbredalen. The main form investigated was the upper transverse patch, and it is this which persists the longest. It was some 280 m long at its largest extent at the start of the ablation period, and its width (excluding the gully portion) was approximately 120 m at its widest point. The gully, and its associated drift, was a prominent feature but was mainly due to fluvial processes. This stretched from the backwall, where a small melt-stream had created a waterfall, down to the valley bottom, a distance of some 300 m (Figure 5).

The upper edge of the area was bounded by a rock backwall up to 10 m high formed of various types of gneisses and schists. The edges of the area were bounded by a number of rock outcrops, running parallel to the backwall, up to 2 m high. The lower lateral edges at the foot of the area were expressed by the junction between the stone pavement and the vegetated area (Figure 5).

2.3 Aim of the Work

The aim of the work is to attempt to find the amount and type of weathering occurring within the area on the bedrock outcrops and the ground surface. Upon t basis of this it is also hoped to find the mode of transport of broken material ou of the area and how much is in fact moved. Thus the snow-patch site has been instrumented to attempt to find the local meteorological conditions and the micro-meteorology of a number of points, both on the ground and rock, when the snow is n present. Thus it is hoped to find the amount of weathering occurring on the grou and on the rock outcrops, and the amount and mode of transport of this broken material away from the snow-patch area.

Eventually it is hoped to throw light on such problems as:

1. The effect of a snow cover on the ground beneath.

2. Whether the top 50 mm of the ground beneath a snow-patch melts, thus allowing solifluction to occur.

3. Whether the frequency of frost cycles is sufficient to explain the observable weathering by freeze-thaw alone.

4. The relative role of chemical as against physical weathering.

5. The mechanism of transporting the broken material away from the snow-patch.

6. Whether nivation, within the strict limits of its definition, is an important process within the area.

It is hoped that the results can then be compared with those of other workers. The similarities and differences can be noted and perhaps explained, and thus hopefully some further light will be thrown on the little known process of nivation

3. WORK UNDERTAKEN IN 1972

3.1 Meteorological Observations

For a consideration of the amount and type of weathering and for an estimation of the amount and rate of snow loss, it was felt necessary to have a good meteorological record for the area under study. Thus a small meteorological station was set up on the edge of the snow-patch. A record was also made at the field station, about 2 km distant. This started on 17th June whilst that at the snow-patch was started on 28th June, both continuing until 14th September. At the snow-patch site a standard Stevenson's screen was set up housing a thermohydrograph, Piche evaporimeter, wet and dry bulb, and maximum and minimum thermometers. Next to this were set up a Pluvius raingauge and a Campbell-Stokes sunshine recorder. A 50 mm and 100 mm ground thermometer were placed in the ground next to the screen and a maximum and a minimum thermometer was placed on the snow surface. Wind, cloud cover and type were also noted. Readings were taken at 1200 hours and 1800 hours, those at the hut being taken at 0900 and 2100 hours.

It is possible to calculate the convective heat transfer, that is the amount of heat absorbed by the snow, from the results of the daily mean snow surface temperature, mean daily air temperature and mean daily wind velocity. This can be derived from the graph prepared by the "Snow, Ice and Permafrost Research Establishment", from which the absorbed heat can be read off if the other factors are known (SIPRE Data Sheet 1). In conjunction with this, the amount of radiation heat supply can be obtained from the nomographs of Gerdel, Diamond and Walsh. Thus, knowing the amount of radiation heat supply in any one day and the amount of heat absorbed by the snow, these can be correlated against the effective snowmelt and condensate from the snow-patch. The amount of snowmelt and condensate is found from Light's equation:

$$D = U\,(0.00184(T-32)10^{-0.0000156h} + 0.00578(e-6.11))$$

where
- D = effective snowmelt in inches per 6 hr period
- U = average wind velocity in mph
- T = air temperature in degrees F
- e = vapour pressure in millibars
- h = station elevation above sea level in feet

Thus from the meteorological station record it will be possible to find the amount of snow loss each day and a value for the parameters which cause this snow loss.

3.2 Snow Retreat Map

In conjunction with the results from the observations mentioned above, a map of daily snow retreat was drawn. This was done by measuring the distance from a number of fixed points on the ground to the edge of the snow. Thus, having mapped the position of the markers it was possible to plot the successive positions of the snow margin through time. This will help to show the relevance of the temperature records from the thermistors in the ground under the snow-patch in terms of those areas which receive maximum protection from the snow and those areas which are clear of snow for the greatest period of time. The knowledge of this will be of value in attempting to evaluate the types and amount of ground movement and in assessing the time and cause of its occurrence.

3.3 Frozen Ground Map

In connection with the loss of snow mapping, some measurement of the depth to the frozen layer with distance away from the snow edge and the change of these values through time was undertaken. Some eight lines of section from eleven to seventeen metres in length were taken from the snow edge downslope and the distance to the frozen layer was found at 0.5 m intervals with a metal probe. This will be of interest in determining whether the run-off was surface or sub-surface and the effects of this on subsequent ground movement. It also helps to answer the question of whether the ground under a snow-patch is frozen or not and whether solifluction could occur underneath or not.

3.4 Snow Density

The density of the snow was taken at various times, using a steel sampling tube of known volume and a weighing bag and balance. The knowledge of snow density was needed as an aid in computing snow loss by radiation and convection; and it is also a prerequisite for calculating the effectiveness of the snow in protecting the ground beneath from temperature changes in the air, since the denser the snow the better it transfers heat. The density of the snow can have a great effect on heat loss or gain from or to the ground under a snow cover. It was also needed for studies of the ablation hollows which are seen to occur on the surface of melting snow. Samples were taken at the edges and the centres of the ablation hollows and the results showed a very marked difference. The mean for the snow as a whole was found to be 0.603, whilst that of the centres was 0.590 and that of the crests was

0.623. Extremes were from 0.779 for the crests to 0.55 for the centres. For a experiment, samples of only the top 50 mm were taken for density sampling and the were found to have an even greater difference between crests and centres and valu were much higher.

3.5 Thermistor Temperature Records

Two Grant model 'D' temperature recorders with some 32 thermistor probes wer emplaced in August. One recorder measured 4 probes at 2-hourly intervals whilst the other recorded 28 probes at a 6-hourly interval. The two recorders were pla in protective bags within their waterproof boxes, then the whole was placed in a large heavy duty polythene bag and covered by a stone shelter.

The thermistors were placed at various depths in pits excavated in the ground so that the speed of penetration of the freezing plane might be measured at shallo depths in the ground in the area of the snow-patch to attempt to find the conditio underneath the snow and compare them with the temperatures above the snow, and the changes that occur as the snow melts away. Also, probes were placed on the rock backwall and on the outcrops of rock on the snow-patch edges. It is hoped that t positioning of the probes will show the movement of the freezing plane and the thawing planes at the start of winter and summer respectively, and the ground condition during the winter down to a depth of 1 m. The positioning of probes in the stone-pavement area (Figure 5) i.e. that area which will be covered with snow for the greatest length of time, will show the effect of the arrival of the snow cover and the insulating effect that this should give, and also whether there is a melting of the top few centimetres of the ground under the snow. The probes plac on the rock should be able to show the number and kind of freeze-thaw cycles that the rock is subjected to, so that an idea of the amount of physical weathering can be estimated.

For a study of nivation it is important to know the frequency of the freeze-thaw cycles, their lengths, the length of the freeze cycle, and its amplitude, the length of the thaw cycle and its amplitude, the rate of change, and the density of the frost shifts. It is hoped that the thermistors operative on the backwall and the rock outcrops will give this information.

A pit was dug to a depth of 1 m just below the stone-pavement and in the wall of this were placed type C probes, 10 x 125 mm in size, at depths of 50, 100, 250, 500, 750, and 1000 mm. A rod the same size as the probe was initially pushed into the wall to avoid possible damage from obstructions. The probes, installed at depths of 500, 750 and 1000 mm, are being recorded, purely on the basis of convenience, at 2-hourly intervals; those at 50, 100, 250, and another at 500 mm depth at 6-hourly intervals. A type C probe, in a protective housing, is to measure ground surface temperature, and another, similarly protected, is to measure air temperature for comparison with the ground and rock records. Some seven type C probes were placed in various positions within the stone-pavement area, from top to bottom, at depths of 40 mm. Three probes were placed in the niches on rock outcrops on the margin of the stone pavement. A further three probes were placed in the gully to find the variations in temperature due to the meltstream that flows there under the snow at the start of the ablation season. The rest of the probes, mainly type E disc probes, were placed in various positions and at different heights on the rock backwall. It is hoped that these 12 probes will give an indication of the temperature conditions along the whole length of the backwall at different heights and controlled by the microrelief of the rock.

Thus it is hoped that all aspects of the temperature effects on the processes of nivation will be monitored. A computer programme has been written to cope with the vast amount of data and this will compute all the aspects of the freeze-thaw cycles as mentioned above, for each of the 32 points. In 1973 another site will be monitored in a similar way and it is hoped that this will be a site with a southerly aspect. Then it will be possible to show the differences between the north and south facing slopes and the variations in the processes as a function of this. As an aid to the understanding and computation of the amount of physical weathering, a detailed field geological analysis of the rock in the area of the probes has been undertaken. From the thin sections of the rock at these points it might be possible to evaluate the kind of weathering, and a mineralogical analysis may make it possible to estimate the potential of the rock for chemical or physical weathering.

3.6 Measurement of Ground Movement

Having attempted to establish the type and amount of weathering in nivation it is necessary to try to find how much and by what method comminuted material is transported away so as to enlarge or create the hollow. The various conflicting ideas about transport methods have been mentioned in the introduction. It seems to be accepted that over-snow transport is ubiquitous, and this has certainly been observed at Okstindan. It is thus a question of whether solifluction or water transport is a dominant transport medium.

To attempt to find if the ground is moving by some form of mass-movement, several experiments have been initiated to measure the speed. Within a periglacial regime temperature is bound to exert some control, and thus there are a number of thermistors measuring the temperature conditions at depth and on the surface. Thus it is hoped to find when the ground is frozen and to what depths it is frozen. It will also show when the ground begins to thaw and to what depth the thaw penetrates. Assuming that (a) the surface layer becomes saturated during thaw with water released both from the thawing ground and snow melt, and (b) that the frozen layer precludes percolation at depth, then it is possible that some sort of mass movement will occur. This is likely to be accentuated by any subsequent freeze-thaw cycles of short duration which will help upset the stability of the ground. As an aid to the understanding of this, soil samples were taken which have been subjected to an analysis to find their physical properties and their susceptibility to freezing.

Four main methods were used from which it is hoped to establish movement of the ground. These are:-

(1) Tubes
(2) Rudberg pillars
(3) Stakes
(4) Painted lines

Some 16 high pressure polythene tubes of 750 mm in length and 20 mm diameter were inserted in the ground. They reached to depths varying from 450 - 750 mm, depending on where stones or bedrock were encountered. At each site the hole for the tube was made with an auger and the tube inserted with a metal rod stiffener (the material from the hole being kept for mechanical analysis). Each tube was numbered and its position fixed by triangulation and by tape and compass survey from a number of fixed points. The tubes cover the whole of the lower edge of the snow-patch site (Figure 5), and are on slopes from 5 to 34°. In two areas a line of 3 and 4 tubes respectively have been placed in a line downslope so as to find the form of velocity profile with distance downslope. Other tubes were situated in terracettes, on a stone-banked lobe, below the stone-banked lobe and in conjunction

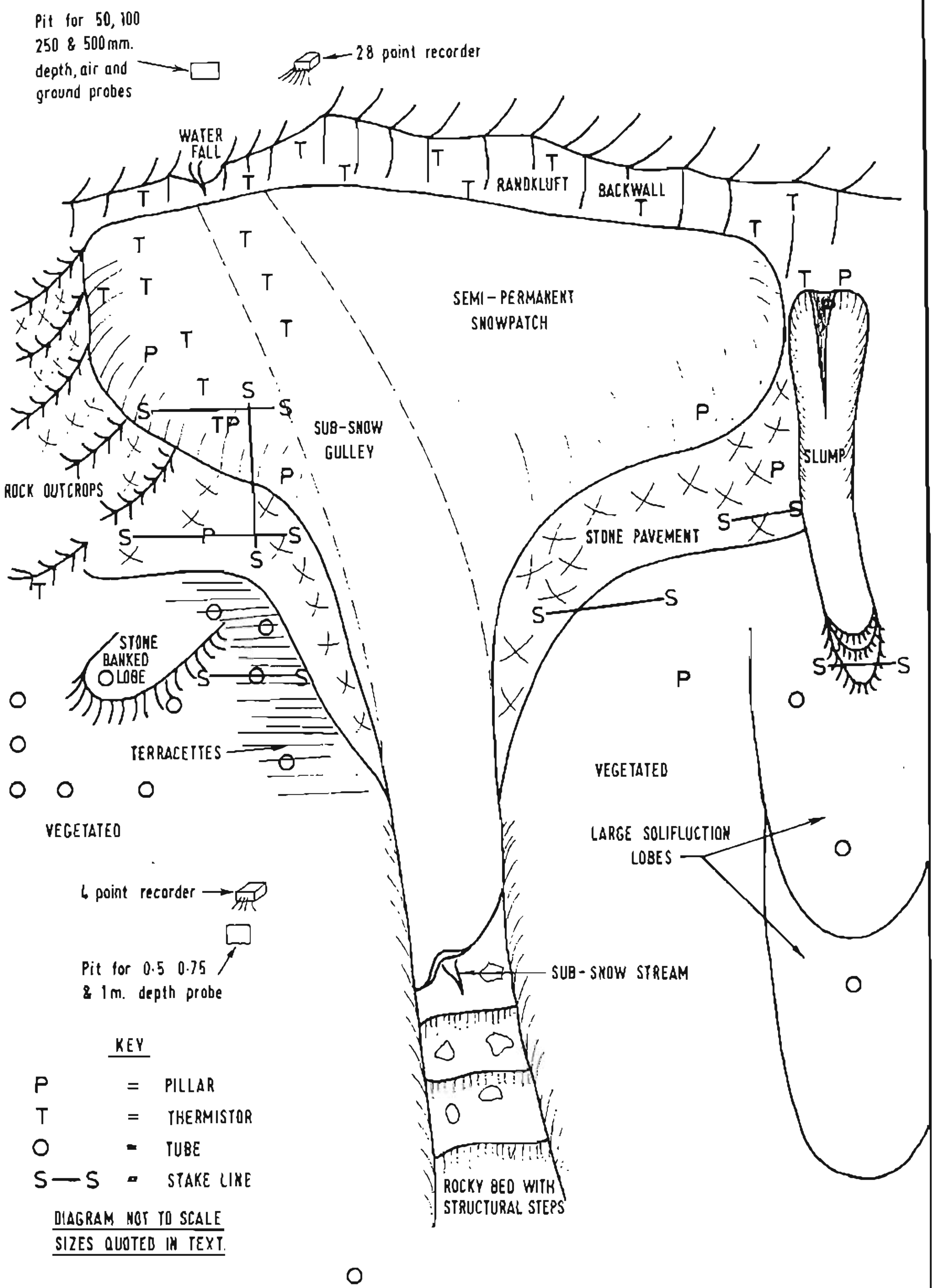
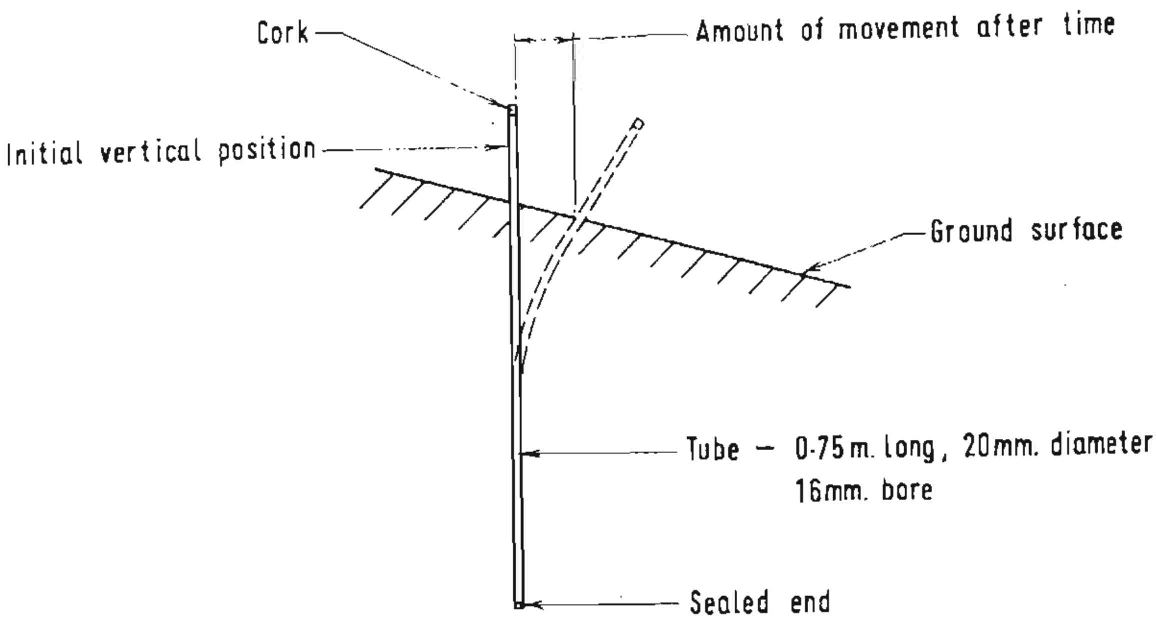
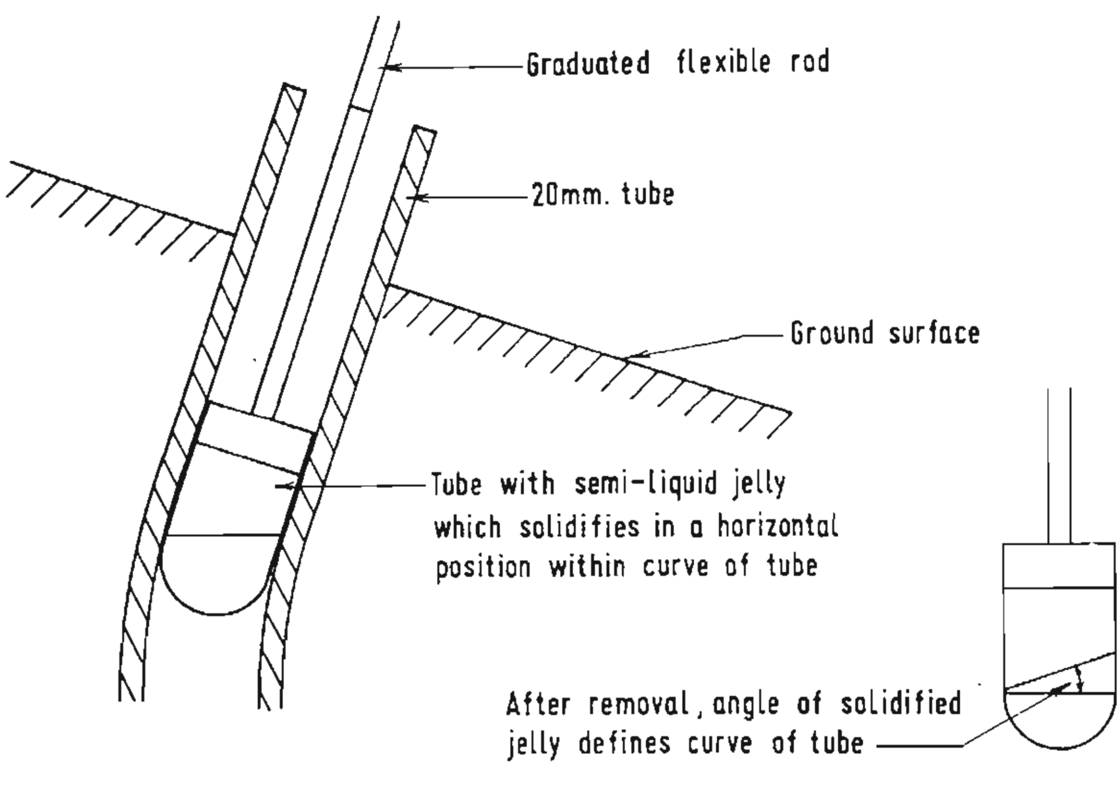


Figure 5. SCHEMATIC DIAGRAM TO SHOW SALIENT FEATURES OF FIELD AREA AND LOCATION OF INSTRUMENTATION.



(a) To show bending of a vertical tube, located on a slope, due to mass movement.



(b) Method of measuring displacement of tube through time without disturbing the tube.

Figure 6. THE 'TUBE METHOD' OF MEASURING GROUND MOVEMENT.

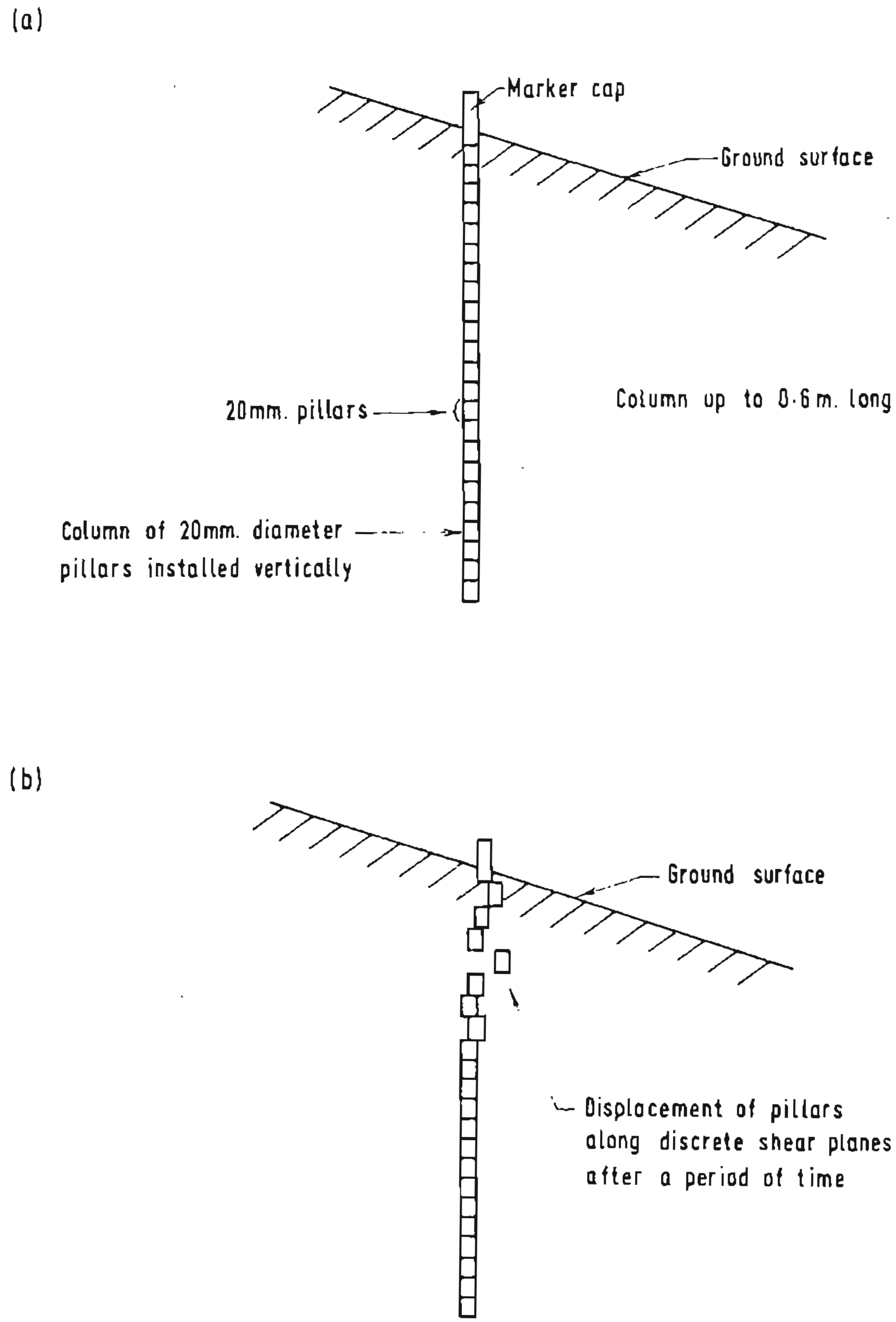


Figure 7. THE 'RUDBERG PILLAR' METHOD FOR DETECTING SMALL BANDS OF DISCRETE MOVEMENT NOT EXPRESSED BY THE TUBE METHOD.

with the other measuring devices. The advantage of these tubes is that they can measured *in situ* with no damage to the tube position and thus measurements can be made over a long time period. The method to be adopted is that of inserting a small glass tube containing a quick setting jelly which will show the angle of the tube at the depth it is left to set (Figure 6), (Harris, 1971 - personal communication).

The tubes, however, can only show a generalized movement and obviously cannot show discrete shearing. Therefore, in the stone-pavement area some 8 sets of Rudberg pillars (Figure 7), made from the same material as the tubes but cut into 20 mm lengths, were inserted. The surface material here was probably very thin s it was advantageous to use pillars as the tubes would have necessarily been cut ve short. The length of the columns varied from 280 (14 sections) to 660 mm (33 sections). The slope here was much steeper, 30 to 50^{o}. It is hoped that th combination of methods for registering movement will enable a general picture of tl whole area to be built up. The main problem with the use of Rudberg pillars is that they have to be exhumed to find their amount of movement and thus can be satisfactorily used only once.

In addition to the tubes and pillars some six lines of wooden stakes, 500 and 250 mm in length by 6 mm diameter, were placed in the ground at various points to attempt to help check the measurements of the tubes and pillars. The lines of stakes are in both the stone-pavement area and in the vegetated area below. Each line was surveyed and the positions of each stake was noted. Thus it will be possible to find how much they move from year to year. In one line was placed a tube and in another was set a column of pillars. As the stake lengths were 250 or 500 mm this will also help show the variation with depth. Five of the lines were placed across the slope and one was set with the slope, running from one transverse stake line high up on the stone pavement to one low down on the vegetated area, and passing near two pillar sets and one tube. Besides helping to find the amount of downslope movement, they may also be useful in finding the amount of heave as they were all installed with 100 mm protruding above the ground surface, so if they are measured from year to year the amount over 100 mm that is sticking out of the groun shows a minimum measure of heave.

Finally, to attempt to find the total amount of surface movement downslope, which might be considerable as the slope is so steep, two lines of painted surface stones were laid out. A taut wire was strung out between two marked points and a line was painted along this over all the stones and ground between the two end markers. One line was red and the other white and so it is hoped to see how much the surface material is moved downslope, either by gravity or possibly by snow-creep.

3.7 Sectioning of Rock Outcrops

The whole of the rock backwall and the bands of bedrock which outcrop on the edges of the snow area were sectioned by vertical lines and each area given a number Each section was then photographed with a scale in an attempt to record any changes that occur from year to year on the outcrops. The position of thermistors in relation to each section was also noted in the aim of finding the conditions encountered along the outcrops, and by designating the walls into areas this was mad easier. It is also useful in the making of geological maps, as artificial boundaries are created which make location easier and a position can be fixed for the origin of any blocks which are found to come off and move over the snow surface to the bottom of the slope.

4. CONCLUSIONS

It is hoped that a single snow-patch area has been monitored in such a way that it will be possible to establish the processes occurring within the area. From the results, an idea of the importance of the different processes should be gained and it will be possible to make correlations between these and the findings of other workers in different parts of the world and perhaps explain the differences.

However, in the coming year to make the work of more value some additional experiments are to be undertaken and also another site will be monitored. Snow depth throughout the thaw period will be measured and at the same time ground water conditions will be studied with tensiometers. A bimetallic actinograph is to be used to try to find a better way of calculating the energy needed to give the snow-melt. A number of undisturbed samples are to be taken for triaxial shear testing and bulk density measurement. Work will continue with the mapping of the area and the collection of geological samples.

With the addition of another snow-patch it will be possible to study variation of process at different altitudes and possibly with different aspects. In addition to the full study of the new patch, measurements of sizes and angles between floor and backwall of various patches at all altitudes and aspects will be made and subjected to statistical analyses.

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THE SIGNIFICANCE OF PERIGLACIAL GEOMORPHOLOGY IN SOUTHERN AFRICA: A DISCUSSION

(1) COMMENT BY PROF KEVIN HALL*

A recent interchange of opinions (Marker 1989; Le Roux 1990a) and book review (Le Roux 1990b) in the *South African Geographer* reflect, I believe, the general lack of awareness by Southern African scientists of the nature and significance of periglacial phenomena in this region. Sadly, Butzer's (1973: 9) remark of 18 years ago, when periglacial studies in Southern Africa were in their infancy, still seems to be valid: "The re-evaluation of 'periglacial' phenomena in Southern Africa suggests that... concepts have sometimes been vague or even erroneous, and that too much interpretation has been based on high latitude preconceptions." A lack of knowledge regarding current thinking with respect to periglacial processes compounded by an inadequate first hand experience of these features, in a variety of environments, only serves to exacerbate the problem when attempting to, directly or indirectly, enter this field.

In a review of *The Geomorphology of Southern Africa* Le Roux (1990b) states, with respect to the chapter by Lewis on Periglacial Landforms, "However, one may well ask why this very minor aspect of Southern African geomorphology was allocated a whole chapter." Upon what basis are the periglacial processes and landforms of Southern Africa deemed "minor"? On the contrary, particularly in the light of the discussion regarding possible periglacial features at Golden Gate (Le Roux 1990a) it would appear that the question might well be restated: "Why is it that periglacial phenomena in Southern Africa are so poorly understood and so little researched?"

The fact is that firm steps have been taken since 1988 to broaden the scope of periglacial studies in Southern Africa. Following from the first meeting of the International Global Correlation Project 297 (Geocryology) held in Mendoza, Argentina during 1989 (which three Southern African scientists attended) the second meeting was held in South Africa during 1990. Why South Africa? Partly because so little *is* known about the subcontinent and so few people are actively working here a *quid pro quo* could be achieved by having geocryological specialists from Argentina, the USA, Russia, France, Poland, and Germany visit the country. They would gain first hand knowledge of a new, largely unknown region and, in turn, could offer their expertise towards a better understanding of the features observed. A special issue of *Permafrost and Periglacial Processes* (2,3) was dedicated to the papers given at this meeting and, as stated in the Introduction (Corté & Hall 1991: 3): "It was agreed that geocryogenic processes are currently active at high altitudes in Southern Africa and that there is evidence for their more widespread activity during the Quaternary."

As a result of that meeting, a greater awareness and interest has been taken

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by the international community of the potential reservoir of periglacial information from this region. This is important, for by means of such features it is possible to gain a better understanding regarding global conditions during the Quaternary. Here again, though, rigorousness and a deeper understanding of periglacial landforms, sediments and processes by South African workers are required. Contrary to the suggestion of Meadows (1988), it is still not clearly known whether the mountains of South Africa were glaciated. If they were, then certain periglacial features (particularly such as proposed ice wedges: Lewis and Dardis 1985) are *highly* unlikely. Equally, or perhaps even more importantly, if these mountains were *not* glaciated then they offer one of the few truly "periglacial" areas (as opposed to para- or pro-glacial) on our planet, i.e. one which has *not* been glaciated but subject to cold-based processes *without* the pre-conditioning or provision of debris by glaciers. Thus, we may have on our doorstep one of the few areas where it is possible to study the role cryogenic processes can play in the development of landforms without glacial interference. The fact is that most "classic" periglacial features such as sorted patterned ground are found in areas that have been glaciated and may be uncommon or non-existent in non-glaciated areas (French, personal communication 1989).

Leaving the discussion of the importance of periglacial studies in South Africa it is pertinent to now consider the Le Roux's comments (1990a) pertaining to the earlier paper of Marker (1989). Firstly, I must wholeheartedly agree with Le Roux with respect to the dangers of "...inferring causes from outcomes." This has been a major problem with South African periglacial studies and somewhat universally with respect to inferring freeze-thaw action from angular clasts (Thorn 1992). Second, I am most sympathetic to Le Roux's arguments that the features at Golden Gate could owe their existence to processes other than periglacial. My problem is with the comments of both Le Roux (1990a) and Marker (1990) within their discussion as to the possible periglacial origins of the various features and some of the terminology used.

With respect to an alternative explanation for asymmetric valleys Le Roux suggests that this may be explained by enhanced chemical and frost weathering under the present climate. Whilst not doubting for one moment the plausibility of the argument this begs two questions. First, if frost action is enhanced on north-facing slopes today then surely this was even more likely during the last glacial? If so, then this is hardly an 'alternative' explanation but rather it explains the maintenance of these features - a different proposition altogether. Second, if freeze-thaw (i.e. a cold-based weathering process) is a *major* contributor to the present-day formation of these features then why can this not be considered a (marginal) periglacial area? In the recent *Glossary of Permafrost and Related Ground - Ice terms* (Harris 1988: 63) periglacial is defined as 'The conditions, processes and landforms associated with cold, non-glacial environments." All periglacial environments are said to be dominated by *frost action* processes. As an aside, with respect to the potential for freeze-thaw weathering being operational today, it might be pertinent to ask whether rock temperature data, rock moisture data and clear evidence of interstitial rock water freezing are actually available to support this contention? In the same vein one is tempted to ask upon what basis Marker (1989: 148) can state: "The number of frost cycles when temperatures fluctuate above and below 0°C are critical?" The 0°C threshold is of no significance to water freezing in either soil or rock due to the effects of such factors as impurities in the water and pore size, both of which depress the freezing point. Equally, it has been shown by many workers that unless temperatures attain -5°C for approximately 10 hours or more then no rock damage will accrue as a result of frost weathering.

Marker (1990) states that the nivation niches are believed to be snowpatch hollows due to their orientation and elevation. However, by considering nivation in this fashion the author goes against recent quantitative findings regarding nivation (e.g. Thorn & Hall 1980; Nyberg 1991) and the present trend of thought on this nebulous process (Thorn 1988). To quote Thorn (1988: 3): "The integration of weathering and transport processes in a single term, plus their variable interaction and the difficulty of defining 'intensification', all preclude a viable operational definition of nivation as a process term. As a landform genesis term, it is rarely possible to assess the degree to which nivation has modified the location of a contemprary snowpatch; therefore operational definition of a nivation landform is also rarely possible. It is therefore recommended that the term nivation should be abandoned..." Thus, with or without potential structural controls, the whole concept of nivation is one best ignored until some rigorous control can be attributed to this multi-facetted term. Stress should rather be placed on workers being less casual in attributing hollows in former or present-day periglacial environments to a term for which there is no common working definition and for which the operation of many of the processes thought to be involved are themselves largely unknown. Thus it is of even greater concern to see the original paper (Marker 1989: 149) statements such as "Nivation and frost acting together can erode back into the hill..." or (1989: 151) "The persistent snow patches would result in *nivation* by freezing and thawing." By all working definitions (see Thorn 1988) frost action/freeze-thaw *are* a component of nivation and so statements such as those above exemplify why Thorn argues for its abandonment! In addition, reference to recent nivation literature (Thorn & Hall 1980; Thorn 1988; Nyberg 1991) would show that it is more likely that transport rather than weathering is enhanced at a snowpatch site.

The other factors discussed by Le Roux (1990a: 130), namely "Terraces", "Boulders" and "Slopes", are impossible to consider without hard evidence. Either argument, or neither, could be valid for, and this is the central problem; no *real* evidence exists to justify any exclusive cause, periglacial or otherwise, and thus both sets of arguments are superfluous. In fact, the study of periglacial forms in Southern Africa is bedeviled by this very problem: a lack of evidence as opposed to unsubstantiated opinion. Thus, equally one could rewrite the statement of Le Roux (1990a: 130) in the Conclusion as "The contemporary hypothesis becomes a credible alternative only when all hypotheses based on periglacial conditions are exhausted." However, for clarity it must be pointed out that contrary to the statement of Marker (1989: 150) ground that is frozen solid in winter is *not* a "...form of *permafrost*". Permafrost is "Ground (soil or rock) that remains at or below 0°C for at least two years" (Harris 1988) whilst the ground that is frozen in winter and thaws in summer (i.e. *does not remain frozen)* is the active layer.

Lastly a few small points regarding the implications of the terminology used. Marker (1990: 131) implies that the glacial phases may have been arid and hence not of a periglacial status. This may not be what is meant but certainly it is implicit in the statement "...the best explanation is that of a periglacial landscape created during Pleistocene cold conditions. Although glacial phases have been accepted as arid in Southern Africa, the evidence on which this view is based..." Be it bad understanding or bad expression, nevertheless there is nothing contradictory in having an arid periglacial environment (e.g. the dry valleys of Antarctica). Marker (1989) suggests that the climate of Golden Gate during the recent geological past may have had a subpolar climate akin to that of Marion Island or northern Norway. However, these two areas have *significantly* different climates and if it were like Marion Island one would *not* expect to find the range of features discussed as the climate is not conducive to their development! Additionally, during solifluction material does *not* "slide" (Marker 1989: 152) as solifluction (Harris 1988) is a form of slow downslope *flow* or *creep*.

To conclude, periglacial processes and landforms *are* an integral part of Southern Africa's recent and present-day geomorphic assemblage. They have been poorly, and largely qualitatively, researched and are in need of a more rigorous perspective. The subcontinent has great potential for periglacial studies and scientists should to become aware of their importance and be more involved in their study. In the mean time authors ought to be much more careful in their use of terminology and refrain from speculative, qualitative judgements that serve only to confuse and misdirect the less experienced.

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5 Mechanical Weathering in the Antarctic: A Maritime Perspective

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Abstract

Considerations of mechanical weathering in the Antarctic are usually limited to the continent and to the dry valleys in particular. However, the oft-unrecognized maritime Antarctic differs climatically from the continent, particularly in terms of moisture availability, and consequently experiences a markedly different weathering regime. Data are presented on mechanical weathering processes, with special emphasis on freeze-thaw, from a maritime Antarctic location. The interrelationship of the various weathering mechanisms is shown and the manner in which those factors controlling freeze-thaw can exert an influence on other processes is demonstrated. For the first time an attempt is made to integrate a combination of field and simulation data to deduce the actual freeze-thaw mechanism causing rock breakdown. It is shown that, compared to the continent, despite the potentially more dynamic maritime weathering environment, weathering rates are still slow.

Introduction

As a consequence of factors such as the distribution of population, the large numbers of centers of research, and the relative ease of access to study areas, the bulk of periglacial and cold region weathering investigations have been conducted in the Northern Hemisphere. In general, periglacial texts (e.g., Embleton and King, 1975; French, 1976; Washburn, 1979) as well as specific review papers (e.g., Lautridou, 1988) have paid little attention to the mountainous and cold regions of the Southern Hemisphere and to Antarctica in

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particular. In fact, it has been suggested (C. Thorn, personal communication, 1986) that while those working in Antarctic regions find a need to keep abreast of northern studies, this is often not mirrored by their northern-orientated counterparts with respect to research in the Antarctic.

As a generalization, the Antarctic appears to be perceived as a great ice-covered continent. Even for those geomorphologists who recognize its existence, the notion of extremeness of climate tends to put it in a unique category—one that has no parallel elsewhere. Partly as a consequence of this alien perception of Antarctic regions and partly as a result of the poorly disseminated, and often obscure, nature of research publications, very little is generally known regarding weathering studies from Antarctica. However, in reality there is a substantial body of Antarctic information pertaining both to mechanical weathering processes and to their association with certain landforms. In fact a number of original hypotheses have originated from Antarctic studies, and in several instances, observations and data bases pertaining to weathering controls exceed their northern counterparts. Central to the whole problem is the realization that although Antarctica is very cold and arid, there are gradations, both in time and space, in the severity of these elements. Consequently, mechanical weathering is neither as simple nor as extreme as it is so often portrayed. Indeed it may often be comparable to that operating in the Arctic and Sub-Arctic today and possibly over a more extensive area during the Quaternary.

Climatic considerations

Campbell and Claridge (1987, p.4) state that "... the climate in Antarctica is unique, by virtue of the very low temperatures and the aridity." Although technically true, this statement hides within it two important factors with respect to weathering. First, Antarctica does not constitute a single climatic entity (Blümel and Eitel, 1989); and second, although air temperatures may be low, rock temperatures can be relatively high owing to warming by direct radiation. This rock warming may facilitate the presence of unfrozen water in or at the rock margin for short periods of time.

Four major climatic divisions are recognized for the Antarctic, namely the interior Antarctic plateau, the Antarctic slope (i.e., the steeply dipping ice margin between the plateau and the coast), the Antarctic coast, and the oceanic or maritime Antarctic (Weyant, 1966; Holdgate, 1970). Without considering the effect of direct insolation on rock, there are still some significant differences between these four zones. For instance, although no rain falls on the continent proper, this form of precipitation can occur on some of the maritime Antarctic islands as well as on parts of the Antarctic Peninsula (Loewe, 1957). Snowfall is also greater within the maritime regions. On the Peninsula and some of the islands not immediately adjoining the continental coast, air temperatures may frequently rise well above freezing during the summer, while on the plateau they may reach only −10°C (Campbell and Claridge, 1987). However, the higher temperatures in the oceanic zone are usually offset by low incoming radiation receipts owing to the high incidence of cloud cover. Toward the continental coast air temperatures may still be low. However, these temperatures can be greatly affected by the amount of exposed rock, which can cause them to reach as high as +9°C at rocky locations (Phillpot, 1985).

Although there is relatively detailed information available regarding the general meteorology and climatology of the Antarctic (van Rooy, 1957; Weyant, 1966; Phillpot, 1985; Dolgin, 1986), this is not the case with respect to the conditions actually experienced by the rock. When rock temperature and moisture regimes are considered, the perception of the weathering potential changes. Considering temperature first (Hall and Walton, 1992), data indicate substantial periods above 0°C. For instance, Sekyra (1970) referenced rock surface temperature ranges in the order of +30° to −35°C at the coast and +10° to −60°C at the plateau margin. Jonsson (1985) measured temperatures at depths of 2 to 3 cm inside rock cracks and found values of +20° to +30°C when the air was −7°C. Both Jonsson (1985) and van Autenboer (1964) noted that when the heat source (i.e., the sun) is cut off, temperature changes very rapidly, up to 49°C in only three hours in the outer shell of rock. In addition, rock albedo exerts a strong influence. Kelly and Zumberge (1961, Table 1) showed that a black rock (biotite schist) was 21.1°C, while a nearby white marble was only 12.8°C, a 65% difference. Even in the extreme-climate dry valleys of Victoria Land, Friedmann, McKay, and Nienow (1987) found that in spring, while air temperatures fluctuated between −45°C and −10°C, the rock surfaces could achieve temperature of +5°C for short periods. In summer, when mean air temperatures is −5°C, the rock surface could be as much as +10°C. In fact, McKay and Friedmann (1985) showed that rock surface temperatures could be above 0°C for 10 to 12 hours per day and that diurnal freeze-thaw could occur for part of the year.

In the milder maritime Antarctic, air temperatures are higher and freeze-thaw cycles relatively frequent. For the islands of the South Shetlands group, Simonov (1977) reported a mean annual temperature of −2.9°C but showed that rock surfaces experience frequent oscillations through 0°C. Blümel (1986) monitored 122 days with freeze and thaw (Frostwechseltagen) during 1979. The high incidence of cloud cover in the maritime region limits frequent rock heating by incoming radiation, but on clear days rock temperatures of +20°C have been recorded (Simonov, 1977). Similar results have been found for the South Orkneys (Chambers, 1967; Walton, 1982), where data also clearly indicate the influence of aspect (Hall and Walton, 1992).

Thus temperatures conducive to both freeze-thaw and thermal stress fatigue and suitable for thermal changes to salts occur within the Antarctic regions. However, for freeze-thaw to occur moisture must be present. Yet the general perception of the climate would appear to preclude this, even though it is recognized that rain can fall during summer both on the Peninsula and on some of the islands.

Although water is relatively sparse, it is far from nonexistent. Along much of the ice-free coastal margin of both the continent and the islands, loss of sea ice during summer facilitates tidal wetting of the coastal zone. Exposure of the wetted rock to subzero air temperatures means the potential for freeze-thaw weathering is high. In the same vein, wetting of rock can take place around the short-lived rivers, lakes, and ponds in the ice-free areas of both the continent and the islands. Shallow ponds, fed by snow melting in contact with heated rock, can form at the base of cliffs (Armitage, 1905). A number of authors have cited the formation of water as a result of snow coming into contact with rock warmed by the sun (Priestley, 1914; van Autenboer, 1962; Bardin, 1964). Friedmann, McKay, and Nienow (1987) also showed that water exists for short periods during the summer in the surface layers of rocks in the dry valleys of the McMurdo region. At the other end of the scale, Debenham (1921) found streams up to an altitude of ca. 100 m that grew to become rivers closer to the coast. More recently, Mosley (1988) gave details of some of these rivers in Victoria Land, such as the Onyx, which may flow for up to 90 days each year. In some locations, such as the Vestfold Hills, the area can become ". . . awash with water, soggy and inaccessible by foot because of the rapidly flowing streams in summer" (Pickard, 1986, p. 344).

Thus Antarctica, although experiencing an extreme climate, does produce conditions conducive to a whole range of mechanical weathering processes. Certainly the area in which these processes operate is limited, and many processes are constrained to a short time period. However, despite these limitations, mechanical weathering is operative, produces a number of distinctive features, and has some modes of operation that are optimum under these climatic conditions.

Recent texts that cover weathering in the Antarctic (e.g., Campbell and Claridge, 1987) provide relatively extensive detail with respect to the continent but almost completely ignore the maritime component. There is, however, a substantial body of work from this region (see Hall and Walton, 1992) which offers a different perspective on mechanical weathering. For example, although a number of authors (e.g., Souchez, 1967; Sekyra, 1970; Selby, 1971, 1972; Campbell and Claridge, 1987) argue strongly against the action of freeze-thaw weathering on most of the continent, the same is not the case for the oceanic islands. As the maritime Antarctic experiences more precipitation than the continent and has a milder climate with frequent oscillations through 0°C, freeze-thaw weathering (gelifraction) is generally considered to be very active (Holtedahl, 1929; Olsacher, 1956; Corté and Somoza, 1957; Dutkiewicz, 1982; Blümel, 1986). Thus the aim of this paper is

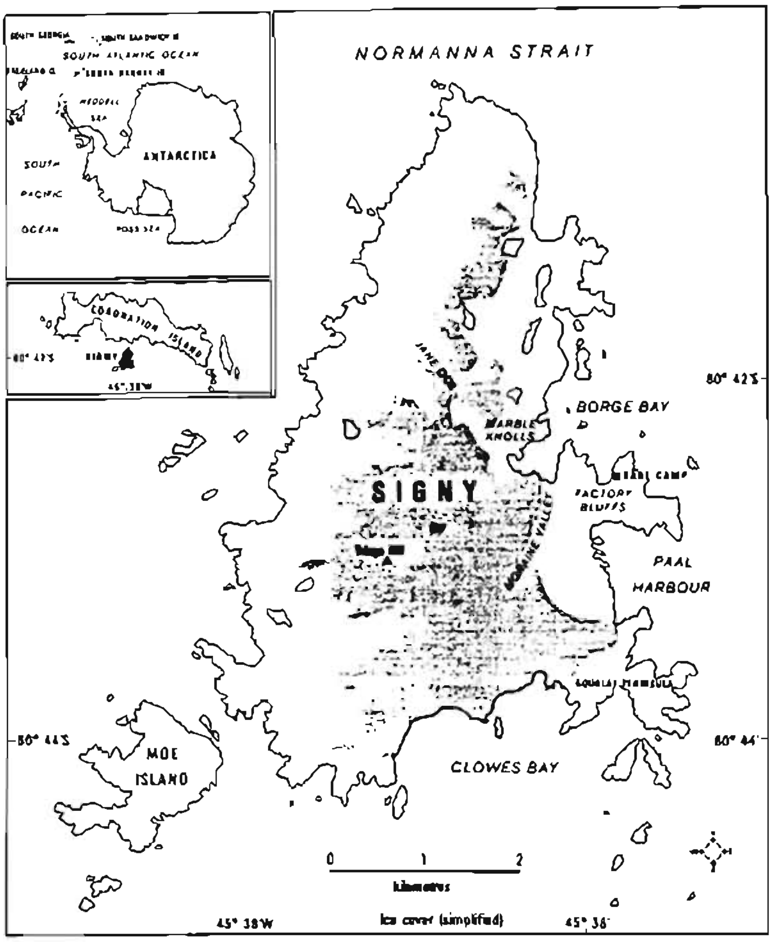

Figure 5.1. Location map of Signy Island

to offer a synthesis of recent studies from Signy Island (60°43'S, 45°38'W), one of the smaller islands in the South Orkney group (Figure 5.1). This island is a representative maritime Antarctic location and the site of a major British scientific station. It is studied here as a means of illustrating the nature of physical weathering within the maritime Antarctic. In addition, an attempt is made to identify the possible mechanisms causing rock breakdown. This is important for although many advances have been made with respect to our understanding of freeze-thaw weathering, we are still unable to make judgments regarding the actual mechanism of breakdown involved (H. French, personal communication, 1989).

Signy Island: background

Early studies on Signy Island (e.g., Mathews and Maling, 1967), as in most of the rest of the Maritime Antarctic (Araya and Hervé, 1972), clearly viewed freeze-thaw as the major weathering process. However, although freeze-thaw does constitute a process in its own right, it usually operates synergistically with several other mechanical processes (Figure 5.2), Singleton (1979) being one of the few to note this. In addition, although most researchers have clearly stated that the observed weathering is due to "freezing action," rarely, if ever, have data been supplied to corroborate this judgment. In fact, process determination requires background information on the controlling factors (Figure 5.3), and without this information it is virtually impossible to determine either the spatial and temporal operation of the various weathering processes or their interoperation with freeze-thaw.

In an effort to overcome these limitations, and so derive a meaningful estimation of process interaction plus weathering rates for Signy Island, background data pertaining to the controlling field conditions were first obtained. These data related to geology (Mathews and Maling, 1967; Storey and Meneilly, 1985), rock temperatures (Walton, 1977, 1982, personal communication, 1984), rock-moisture content (Hall, 1986a), rock moisture chemistry (Hall, Verbeeck, and Meiklejohn, 1986), the physical properties of the local quartz-micaschist bedrock (Hall, 1987), and thermal gradients and rates of change of temperature within the rock (Hall and Hall, 1991). On the basis of these field data it was possible to perform computer simulations

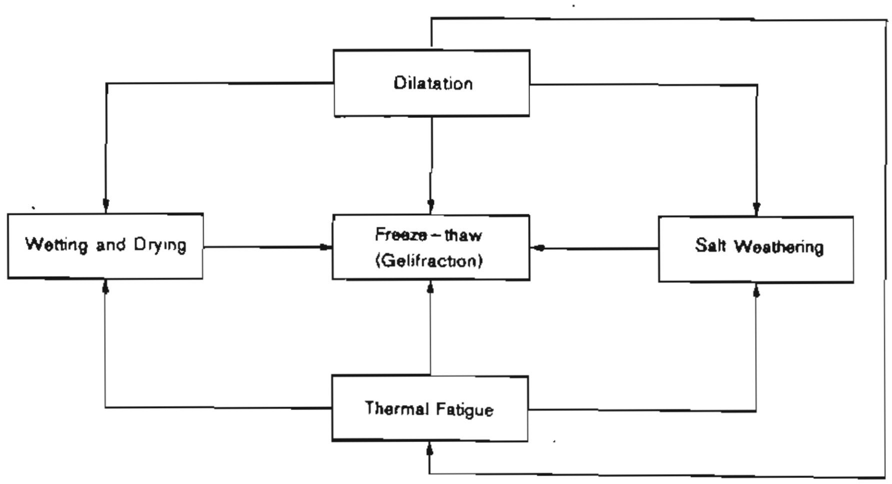

Figure 5.2. A flow chart showing the relationship of freeze-thaw weathering to the other mechanical weathering processes

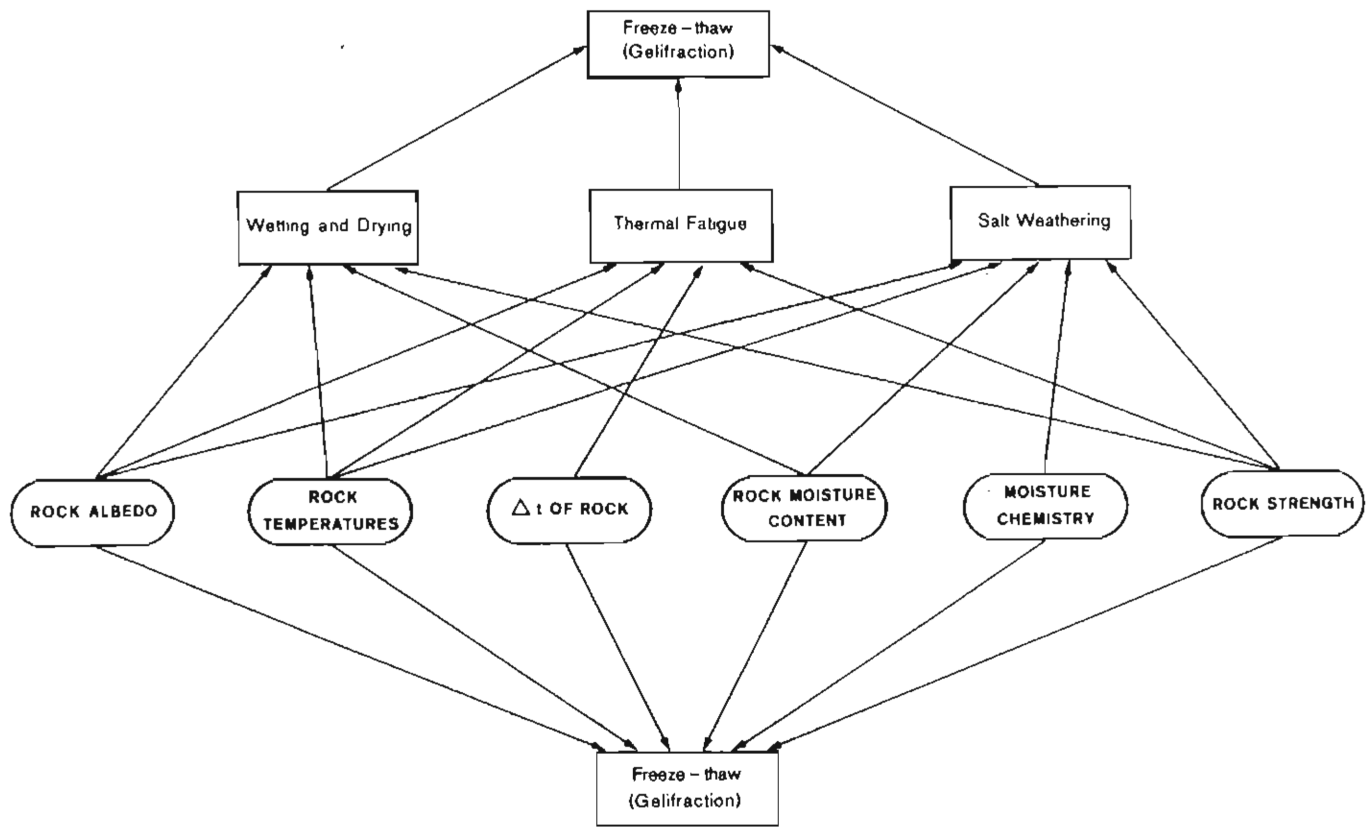

Figure 5.3. A flow chart showing the role of the controlling factors with respect to freeze-thaw and the other mechanical weathering processes

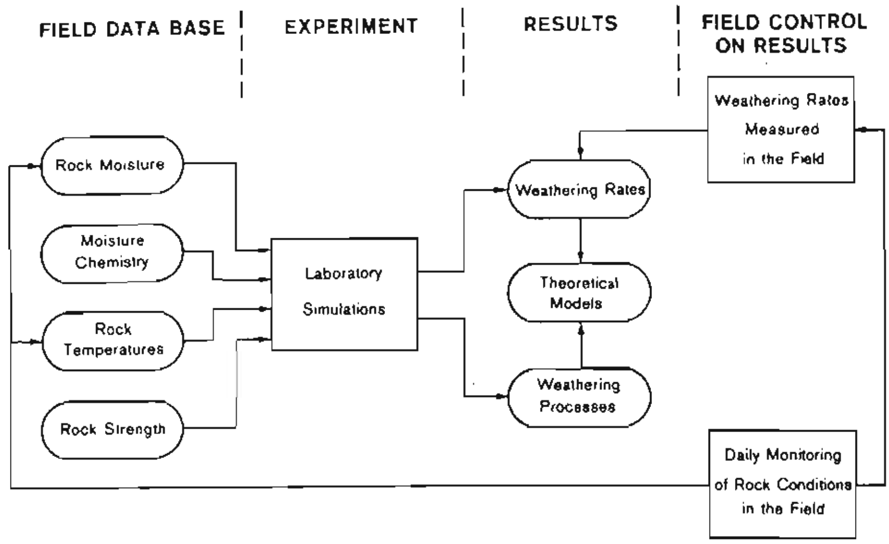

Figure 5.4. A flow chart showing the framework within which the Signy Island studies were undertaken

(Hall, Cullis, and Morewood, 1989) that replicated real world conditions and so provided information on both weathering processes and rates (Hall, 1986b, 1988a,b). As a further control on the simulations and as a verification that the derived weathering rates were of the correct order of magnitude, two long-term field experiments were initiated. The first monitored daily changes in rock moisture content, mass loss, and environmental conditions for one year (Hall, 1988c), while the second consisted of retesting rock tablets left in the field for varying lengths of time (Hall, 1990). The framework within which these various approaches were integrated is shown graphically in Figure 5.4.

In summary, the field data describe an environment with a small amplitude of mean monthly temperatures; the mean of the warmest month is greater than 0°C and that of the coldest month is −9°C (Collins, Baker, and Tilbrook, 1975). Radiation receipts are diminished by the extensive cloud cover but can be high on cloud-free days. Permafrost is present over most of the island, but the active layer is usually more than 1 m thick (Chambers, 1967). Temperature oscillations across 0°C are fairly frequent (>40 yr^{-1}), and on days with radiative heating rock temperatures can exceed 30°C. Although precipitation occurs fairly frequently, the annual water equivalent is small. However, precipitation can occur in the form of rain and, as humidity is high and summer temperatures not frequently below 0°C, much of this water becomes available to the rock. This moisture availability was reflected in data on interstitial rock moisture content. These data showed that from a total of 155 rocks sampled, 40.4% had moisture contents in excess of 50% and 17% in excess of 90% saturated. The degree of saturation was largely constrained by physical location, with the highest moisture contents occurring in positions close to melting snow or associated with snow-melt rivulets. Situations away from such favorable conditions had commensurately lower moisture contents. Data on daily changes in moisture content indicate that it can be high during time of freezing temperature; thus the potential for freeze-thaw weathering does exist. The maritime location of the site is reflected in the chemistry of the interstitial rock moisture, which showed a mean NaCl molarity of 0.47 ($s = 0.1$).

The rock itself, quartz-micaschist, is strongly anisotropic, and this is reflected in rock strength data. The mean compressive strength normal to schistosity is 2.3 MN m^{-2}, while that parallel to schistosity is only 0.50 MN m^{-2} (i.e., 78% weaker). As tensile strength (i.e., that applicable to most weathering forces) is approximately 20% of the compressive strength (Szlavin, 1974), the rock is classified as "low" to "medium" in strength on the engineering rock strength scale of Broch and Franklin (1972) when measured normal to schistosity but "low" to "very low" in strength when stressed parallel to schistosity. In a similar vein, calculations (Hall, 1986c) of the stress intensity factor K_{IC} (an index of fracture toughness) showed low values, varying from 1.4891 (normal to schistosity) to 1.2000 (parallel to schistosity).

Processes

With respect to freeze-thaw weathering, the derived values of K_{IC} allow indirect calculation of the pressures required for crack propagation due to ice formation. Knowing that ice pressure in saturated rock increases with negative temperatures at a rate of 1.14 MPa deg^{-1} (Hallet, 1983), it is possible to calculate the temperature required to generate any given pressure. These values (Hall, 1986c, Table II) indicate that for saturated rocks, cracks in the size range 0.1 m to 0.005 m with an overburden of between 1 m and 150 m of bedrock are able to be affected by frost action on Signy Island. Propagation of smaller cracks requires temperatures lower than −29°C, which are rare in this region. It should be noted that these calculations are valid only for saturated rock; the determinant equations do not strictly apply to nonsaturated rock. Nonetheless, a maximum possible threshold for potential frost action has been established.

With respect to the action of freeze-thaw, it is the combination of temperature, direction of freeze penetration, moisture content, and moisture chemistry together with the anisotropic nature of the rock that define the manner and rate of operation of this process. There is a clear distinction, often not explicitly recognized by many workers, between omni- and undirectionally frozen rock. A saturated omnidirectionally frozen block lost more than 57 times more mass than a saturated unidirectionally frozen block subject to the same temperature changes. The greater mass loss for the omnidirectionally frozen block reflected not only a "closed" versus an "open" system but also the influence of anisotropy.

The orientation of the plane of schistosity with respect to freeze penetration affects the manner in which freezing takes place and hence the rate of breakdown. When schistosity is normal to the freezing front, two possibilities arise. First, if the rock has a high moisture content, water can be forced away from the freezing centers into the rock to generate high hydrostatic pressures (the "hydrofracture" of Powers, 1945). Second, if moisture contents are low and the freezing rate is favorable, water may be drawn to the freezing center to produce strong tensile forces (Hallet, 1983). Both these possibilities may produce extensive breakdown of the rock. On the other hand, if schistosity is parallel to freeze penetration, then breakdown is limited to situations in which the moisture content along any laminae is greater than or equal to 91%. This high moisture content is necessary, as water migration between laminae is not possible. Consequently, as the laminae are frozen consecutively, only those that have ice growth in excess of the volume available experience damage. Thus an unconstrained block will be affected by omnidirectional freezing that can generate large internal stresses (Hallet, 1983). In addition, the freeze penetration will be asymmetrical owing to the effects of schistosity.

Irrespective of schistosity, rock damage can occur only if temperatures are conducive and water is present. In the absence of suitable temperatures for an adequate length of time and the availability of water to actually freeze, breakdown cannot occur by means of the freeze-thaw mechanism. Temperature data from the field indicate that in the presence of moisture, freezing will occur. However, it has been found that the nature of the temperature conditions will influence the manner in which freezing takes place. To date three main forms of freezing have been identified. First, with a relatively slow decline in environmental temperature (ca. 1°C h^{-1}) there occurs a rapid, large-scale (ca. 80% of the water freezes) transformation of water to ice subsequent to extensive supercooling of the interstitial water. Second, with a more rapid fall in temperature ($\geq$3°C h^{-1}) there is a slow, progressive freeze from the outer margin of the rock inward. Third, when the salinity of the interstitial rock moisture is high ($\geq$0.5 M NaCl), a progressive but rapid freeze (intermediate between the other two forms), with no sign of supercooling, takes place. Constraining these three forms of freezing are the requirements (1) that temperatures be less than or equal to −3°C, and (2) that these temperatures be maintained for at least 10 hours. These requirements are similar to those suggested by Lautridou (1971).

Thus with schistosity normal to freeze penetration and a slow rate of temperature decline, it is possible for hydrofracture or cavitation-induced nucleation of ice to take place (Hodder, 1976), dependent upon the actual rate of phase change. With a more rapid decline in environmental temperature, ice lens may grow, as proposed by Hallet (1983). With schistosity parallel to freeze penetration, either the simple 9% volumetric growth concomitant with phase change or frost bursting (Michaud, Dionne, and Dyke, 1989) can occur. The rate of temperature decline and the degree of saturation control whether either of these two processes can take place.

Moisture exerts a powerful influence on the nature of the freeze-thaw mechanism. It is not just the degree of saturation but also the distribution and chemistry of that water that are influential. Although the available data from Signy Island show that rocks in opportune locations may be greater than 50% saturated, so far it has not been possible to determine the distribution of the water within the rock. Ultimately this is of major importance, for the concentration of moisture in the outer margin may produce saturated conditions in that zone while the rock as a whole is well below 91% saturated. As a consequence of this moisture gradient, the actual freeze mechanism may differ from that presumed upon the basis of total rock moisture content.

With respect to moisture chemistry, the presence of salts in solution has a direct effect upon freeze-thaw in addition to their role in salt weathering. As already discussed, the nature of the freeze is directly affected by high (≥0.5 M NaCl) saline levels. In addition, the presence of salts depresses the freezing point (by 0.9°C for 0.25 M NaCl and 1.9°C for 0.5 M NaCl). Thus temperatures must be much lower for freezing to occur, and thawing takes place at solution temperatures below 0°C. This then means that freeze-thaw can take place without the temperature of the interstitial rock moisture ever going above 0°C.

All of the above is synthesized in Figure 5.5, which shows the various combinations of temperature, moisture, and rock conditions together with the resulting possible freeze-thaw mechanisms. Thus for the Signy Island situation it is now possible to make, for the first time, a reasonable judgment as to the possible form that freeze-thaw weathering may take based upon the extant environmental and rock conditions. However, as stated at the beginning, freeze-thaw does not work on its own, and thus it is now pertinent to consider how the elements shown in Figure 5.5 fit within, and integrate with, the other mechanical weathering processes.

The maritime location of Signy Island results in relatively high levels of NaCl in the interstitial rock water. This has a number of repercussions with respect to both freeze-thaw and salt weathering (Figure 5.6) in addition to the effects noted above. In freeze-thaw itself the role of salt is complex and little understood, although it has been shown that "... the extent to which salts enhance or inhibit frost weathering ... varies with both the type and concentration of the salt, the intensity of the freeze regime, and the rate of freezing and thawing" (Jerwood, Robinson, and Williams, 1990, p. 619). With low moisture levels and high concentrations of salts it is possible that the freezing point depression may facilitate a longer period of water mobility and

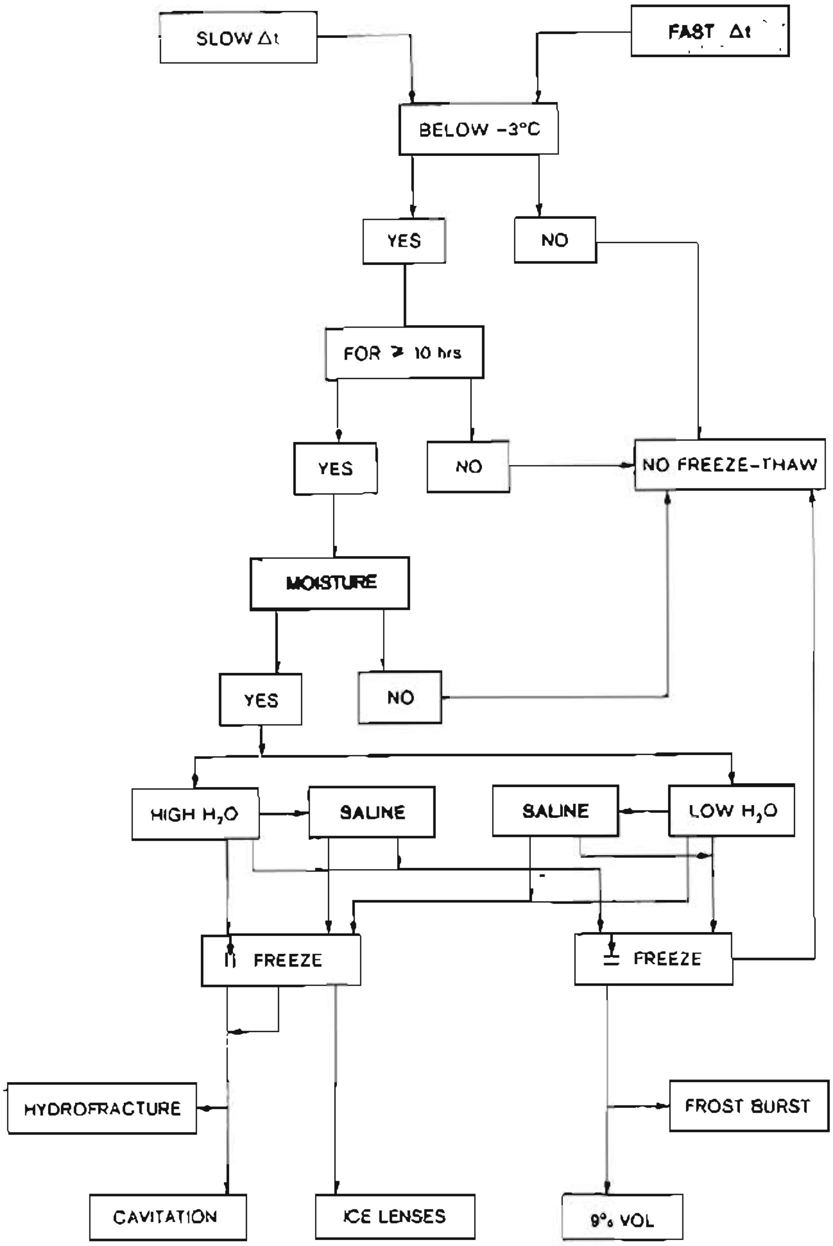

Figure 5.5. A flow chart of freeze-thaw weathering that allows deduction of the possible freeze-thaw mechanism. The shaded boxes (Saline, Fast Δt, and Moisture) are the starting points for the more detailed flow charts referring to salt weathering, thermal fatigue, and wetting and drying presented in Figures 5.6, 5.7, and 5.8. The symbol = indicates where freeze penetration is transverse to schistosity, and the symbol ll where freeze penetration is parallel to schistosity

thus enhance the growth of ice lenses in the quartz-micaschist (Figure 5.5). With high moisture levels, according to Jerwood, Robinson, and Williams (1990), intense freezing regimes ($\leq -30°C$) and low salt concentrations result in the eutectic temperature being reached ($-21.1°C$ for NaCl), at which point a solid cryohydrate is formed and high internal rock stresses are created. With high salt concentrations and low intensity freezes (ca. $-10°C$), destruction is limited, as not all the water freezes and there is little or no crystallization of salt. Although no detailed simulations have been undertaken on the quartz-micaschist, the findings of Jerwood, Robinson, and Williams (1990) appear to agree with some preliminary tests which show that for non-saturated samples subject to omnidirectional freezing down to $-19°C$ the greatest amount of breakdown occurred to the sample in a nonsaline solution. The next greatest amount of damage was found for a 1.0 M NaCl solution (the others being 0.25 M, 0.5 M and 0.75 M NaCl). However, the damage in this case may have been caused by salt crystallization resulting from drying of parts of formerly wetted rock during the thaw phase.

Salts brought into the rock in solution can precipitate out during warm periods, particularly when the rock is heated by incoming radiation, such that crystallization pressures may cause rock breakdown (Figure 5.6). Once the salts are present in the rock, usually close to the outer margin of the rock from whence the water is evaporated, they can cause further damage either by thermal expansion during times of high radiation receipts or by hydration resulting from high humidity (Figure 5.6). Thus many of the elements that affect the freeze-thaw process (such as rate of change of temperature, freeze amplitude, duration of freeze, heating during thaw phase) can also directly or indirectly affect saline rock moisture, and thereby produce salt weathering during the thaw phase or create an element of salt weathering within the freeze-thaw process itself.

The changes of temperature, both positive and negative, which are necessary for freeze-thaw to take place and which also affect salt weathering, can themselves cause rock breakdown (Figure 5.7). If the rate of change of temperature Δt is rapid ($\geq 2°C\ min^{-1}$), then the rock is subject to thermal stresses. Although this rate is very high, far in excess of that usually considered with respect to freeze-thaw, two situations can favor its occurrence. First, during subzero air temperatures the outer shell of a rock may be subject to intense warming by incoming radiation. This can cause a steep temperature gradient (and thus tensile stresses) and high values of Δt ($\leq 9°C\ min^{-1}$) for short periods of time (ca. 3 to 4 min) at the rock margin. The heating, and hence the expansion, will not be uniform about the rock, and so heated faces trying to expand will be buttressed by unheated surfaces, thereby accentuating tensile stresses. Second, when the heat source is rapidly removed, the rock is subject to extremely rapid cooling and, thus, compressive stress. During these conditions rates of temperature change ranging from $2°C\ min^{-1}$

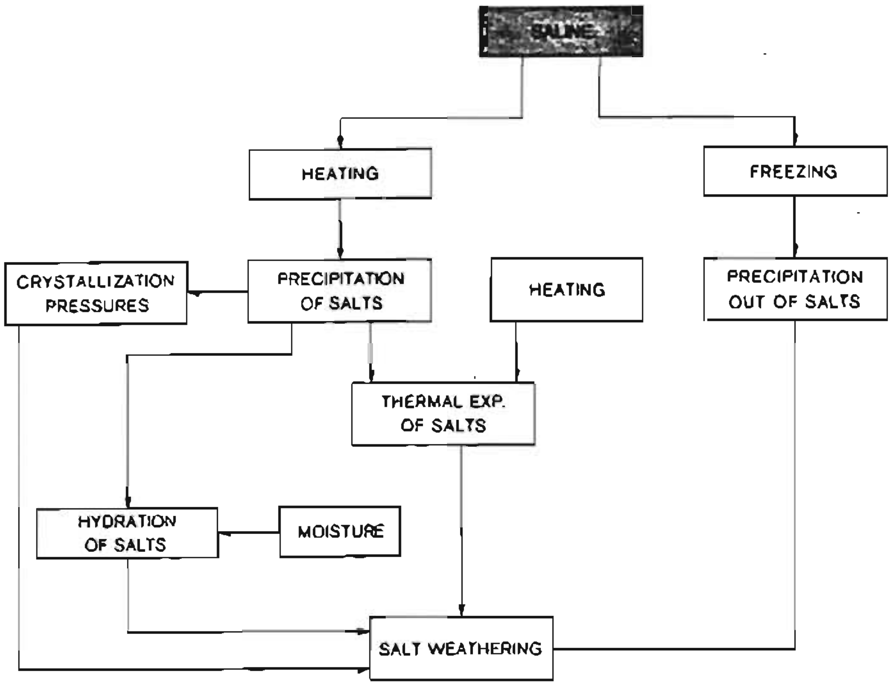

Figure 5.6. A flow chart showing the possible mechanisms associated with salt weathering

to 7°C min^{-1} have been recorded for four minutes. In these cases, the rate of thermal change is equal to, or in excess of, that suggested to be the threshold for the initiation of thermal stress fatigue (Richter and Simmons, 1974; Yatsu, 1988).

Although radiative heating of the rock during times of subzero air temperatures does not occur very often on Signy Island owing to the high incidence of cloud cover, available data suggest that when it does take place it affects the outer 2 cm of rock. Data from the laboratory simulations indicate that during these times of high positive Δt, above-zero temperatures penetrate to a depth of ca. 1 cm. This then means that, should water be present in the outer margin, freeze-thaw weathering will take place within the zone affected by thermal stress fatigue. Thus freeze-thaw weathering and thermal stress fatigue will operate both independently and synergistically such that it may be difficult to discern the role played by either in causing surface cracking and flaking of the rock.

The presence of water and, more particularly, fluctuations in water through time can cause mechanical weathering. The variations in moisture content that rock undergoes during the wetting → freezing → thawing → drying

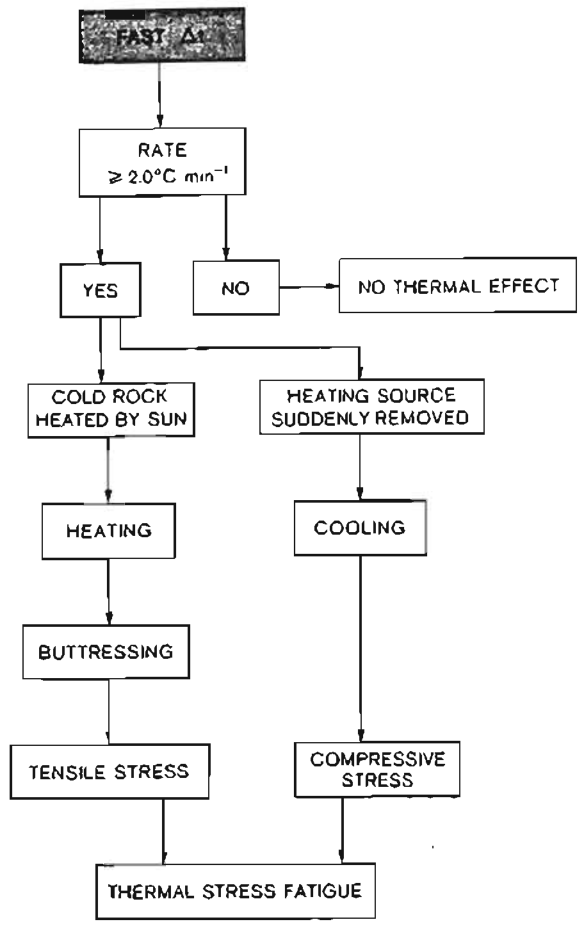

Figure 5.7. A simplified flow chart of the thermal fatigue weathering mechanism

sequence or during snowmelt and evaporation within periods of above-zero temperatures can promote the mechanical weathering process "wetting and drying" (Figure 5.8). Laboratory simulations on quartz-micaschist utilizing ultrasonic monitoring of rock conditions show that the presence of water weakens the bonding strength of the rock and that during drying elastic strain recovery is not always total. Thus, the wetted part of the rock ". . . will experience a decrease in strength resulting from diminished elasticity due to loss of bonding strength" (Hall, 1988b) which can lead to the formation of microfractures. It also means that the strength available to resist failure during freezing is diminished. Again, like thermal stress fatigue, the zone within which wetting and drying occur is the same as that affected by freeze-thaw, and the two work synergistically.

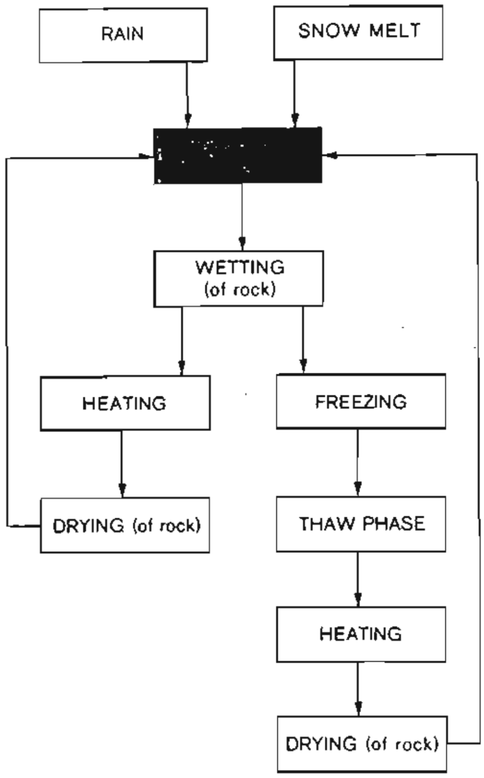

Figure 5.8. A simplified flow chart of the wetting and drying mechanism

Synthesis

Figures 5.5. to 5.8 show that process interaction takes place such that an element controlling one process (e.g., temperature change within freeze-thaw) can exert an influence on another process (e.g., thermal stress fatigue). Thus, as shown simplistically in Figure 5.2, given the presence of the required factors (e.g., moisture, salt, rapid Δt) all the mechanical weathering processes can operate both independently and synergistically. This means that the conditions facilitating freeze-thaw weathering can also cause wetting and drying, salt weathering and/or thermal fatigue. The maritime Antarctic situation, as exemplified by Signy Island, may be ideal for such combinations of mechanisms. Compared to the continent, precipitation is relatively large, can even occur in the form of rain, and is saline. Thus, unlike on the continent, exposed rock is frequently wetted. Although temperatures are not severe, freeze-thaw cycles are frequent, particularly during the spring to autumn period. Winter temperatures, however, approach −30°C and both give rise to a prolonged, severe freeze and exceed the eutectic point of NaCl, thereby allowing the formation of a potentially rock-damaging cryohydrate. Although cloud cover is usually fairly extensive, clear days occur on which radiation receipts are high and rock heating (and subsequent cooling) may be dramatic.

Perhaps the most significant result is that, for the quartz-micaschist, it is now possible, based on knowledge of the controlling factors, to determine the actual form of the freeze-thaw mechanism (Figure 5.5). In addition, based on the same data it is possible to gain some insight into the role of the other mechanical weathering processes (Figures 5.6, 5.7, and 5.8). Thus, for the first time the background data have been obtained and utilized to determine, with some degree of certainty, the actual weathering process(es). Now it is possible to get some idea not only of the temporal and spatial variability in the freeze-thaw mechanism but also of its effect upon, and interaction with, the other mechanical weathering processes.

As discussed above and illustrated in Figure 5.5, the situation is far from simple. Cold temperatures alone need not imply the operation of freeze-thaw; no moisture may have been present or the freeze may not have been of sufficient amplitude or duration. Thus the argument is not that Signy Island experiences a particularly destructive regime, but rather that in any consideration of weathering simplistic judgments should not be made with respect to what is the operative process. In fact, available data on weathering rates on Signy Island (Hall, 1988c, 1990) show them to be relatively slow with cut rock tablets exhibiting an extrapolated mass loss of only 2% per 100 years. This rate for omnidirectionally frozen, relatively wet samples may be as much as 50 times greater than for unidirectionally (relatively dry) frozen bedrock!

Having just argued for the operation of a number of mechanical weathering processes and recognizing that chemical and biological weathering components also aid breakdown, the finding of slow weathering rates may seem contradictory. However, this is not so for several reasons. First, the data base with respect to weathering rates only pertains (so far) to five years and the extrapolation to 100 years was linear, whereas the reality is more likely exponential (but the data base is not yet sufficient to model this). Second, cut rock blocks were used for the weathering experiments, and these may show initially slower rates than naturally occurring blocks. Third, and perhaps most important, the discussions with respect to process interaction were not put within a temporal or spatial framework. The reality is that while the maritime Antarctic may be more dynamic an environment than the continent with respect to weathering, this is still relative, and rates are substantially slower than in most arctic or alpine situations. Also, weathering often operates through fatigue, and it is the repetition of stressful conditions that ultimately leads to failure. In the maritime Antarctic the required combination of conditions (e.g., Figure 5.5) necessary to exert stress does not occur all that frequently, and so weathering rates remain slow.

Conclusions

Weathering conditions in the maritime Antarctic are somewhat different from those of the Antarctic continent with, perhaps, the most significant factor being the greater availability of moisture in the maritime region. Studies undertaken on Signy Island, a representative maritime Antarctic location, have shown the potential lines of interaction between the major weathering processes. More importantly, with respect to the process of freeze-thaw it has been possible to determine which mechanism is actually operative and how the various factors that influence this mechanism also affect other weathering processes. Because the various processes are both temporally and spatially interactive, simplified judgments as to the cause of weathered debris are no longer viable. Despite the potential range of weathering processes, it appears (from the available data) that weathering rates in the maritime Antarctic, although faster than on the continent, are still relatively slow.

Acknowledgments

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Recent Progress in Antarctic Earth Science

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MECHANICAL WEATHERING ON LIVINGSTON ISLAND, SOUTH SHETLAND ISLANDS, ANTARCTICA

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Abstract: Mechanical weathering, particularly the freeze-thaw mechanism, is frequently cited as the predominant cause of bedrock breakdown in the maritime Antarctic. Rarely, however, are data available to justify this presumption. During the austral summer of 1990/91 a study was undertaken of weathering processes on the Byers Peninsula of Livingston Island (South Shetland Islands). Information pertaining to rock temperatures and moisture content, general climatic data, Schmidt hammer rebound values, indentor penetration and weathering rind thicknesses were collected from rock outcrops. Additional data relating to the number of weathered clasts on the different raised beach levels plus the amount of weathered-free material obtainable from rock faces were also collected. From the available data it is concluded that freeze-thaw weathering is not as ubiquitous as is so often thought. Wetting and drying, salt weathering and chemical weathering all play major roles and, in fact may be more active than freeze-thaw.

Key words: weathering, rock properties, rock conditions, aspect, maritime Antarctic

Introduction

Although mechanical weathering is often cited as operative within the Antarctic, actual measurements are extremely rare and in many instances process type appears to have been presumed rather than proved. Central to any consideration of Antarctic weathering processes is the distinction between the continental Antarctic and the maritime Antarctic. The climatic differences between these two regions, not the least being the greater precipitation (often in the form of rain) and the potentially greater number of oscillations through 0°C experienced in the maritime zone, have marked implications for weathering process type and rate (Hall, 1992). It is most noticeable that a perusal of references relating to the maritime Antarctic, particularly to the South Shetland Islands, frequently, sometimes exclusively, cite freeze-thaw weathering as the major operative weathering process (e.g. John and Sugden, 1971; Araya and Herve, 1972; Simonov, 1977; Hansom, 1983; Stäblein, 1983).

Part of the British Antarctic Survey "Fellfield Research Programme" has included a study of mechanical weathering processes in the maritime Antarctic environment of Signy Island (see Hall, 1992; for a review of work to date). As an extension of this project, a study was undertaken during the 1990/91 austral summer of weathering processes on the Byers Peninsula of Livingston Island in the South Shetland Islands. As the basis for this study, data were collected regarding the local climate, rock moisture content, rock temperatures, rock strength, micro-indentor hardness and weathering rind thicknesses. Combined, this information allows some insight into weathering processes on the Byers Peninsula and, although for only one summer, indicates that the general perceived notion of the importance of freeze-thaw weathering may be in need of reassessment.

Study Area

The Byers Peninsula is at the western extremity of Livingston Island (Fig. 1) and is the largest ice free area in the South Shetland Islands (Lat. 62°40'S, Long. 61°00'W). To the east of the peninsula is the Rotch ice dome and behind that an alpine terrain with extensive valley glaciers. John and Sugden (1971) suggest a mean annual temperature of −3°C and an annual precipitation in the order of 100 to 150 cm water equivalent for this area. Further, Thom (1978) indicates that the mean daily temperature for December to March is normally above freezing but that permafrost is present below an active layer of 0.3 to 0.7 m thickness. The Byers Peninsula comprises an extensive assemblage of raised beaches interspersed with upstanding volcanic plugs, that frequently exhibit columnar jointing, and a number of dykes and sills. The geology consists mainly of volcanics interbedded with conglomerates and sandstones representing a sequence of marine, lacustrine and terrestrial environments (Hobbs, 1968; Smellie *et al.*, 1980).

Upstanding solid rock outcrops are relatively rare on most of the Byers Peninsula and those that do exist are primarily volcanic dykes, sills or plugs. However, those that do occur frequently constitute the major relief features of the peninsula, e.g. Chester Cone (193 m) and Usnea Plug (114 m). The rocks of these upstanding blocks are usually of basaltic to andesite composition but may tend towards diorites or microgabbros (Smellie *et al.*, 1980). Due to their constituting a major obstruction to the snow-bearing winds there is substantial lee side accumulation about these outcrops; predominantly on the south-facing sides. It was the weathering of some of these volcanic outcrops that was investigated.

Approach

During the field period data were collected several times each day regarding rock moisture content for the north- and south-facing aspects of a dyke, for the north-, south-, east-, and west-facing aspects of a small boss, and on a horizontal surface (Hall, unpublished data). At these same locations rock surface temperatures were also recorded (Hall, unpublished data) and at the open site measurement

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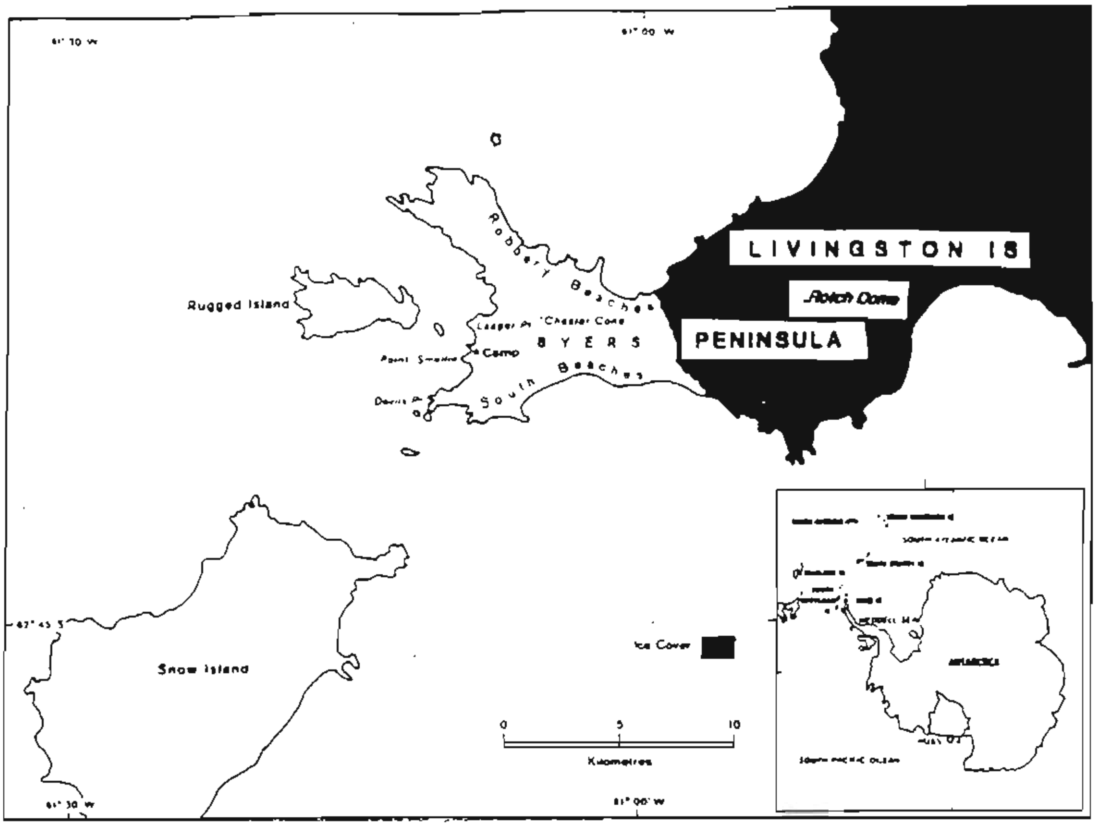

Fig. 1. Location of study area.

was made of wind speed and direction, air temperature, humidity and radiation input. At the dyke, measurements were made regarding the thickness of weathering rinds on the two faces. In addition, measurement of Schmidt hammer rebound and micro-indentor hardness values were obtained at the dyke, the volcanic boss and a variety of other outcrops (Hall, unpublished data). Observations regarding periglacial landforms, together with information from other studies undertaken in this area (e.g. Thom, 1978), help give indirect evidence regarding the cryological regime. Other indicators of weathering were also obtained. Data were collected pertaining to the number of weathered clasts found on the different beach levels as this provides some indication of the effectiveness of the weathering processes. In addition, weathered material was collected from cliff faces at a number of locations to gain some idea of its form.

Fig. 2. Snow accumulation on the lee (south-facing) side of a dyke.

Results and Discussion

One feature immediately apparent on the Byers Peninsula that has a marked affect upon weathering is that of the preferential lee-side accumulation of snow. It was very obvious that east-west trending dykes had snow accumulation on the south-facing side whilst the north-facing side was snow free (Fig. 2). This preferential accumulation was clearly indicated by the distribution of miniature sorted stripes (Hall, unpublished data). Despite sediment properties and slope angles being conducive throughout the area, sorted stripes were preferentially aligned between 270° and 90° (W-N-E) with 75% of all observations within this (relatively) snow-free arc. Although the upper part of slopes in the arc 90° through to 270° (E-S-N) are snow-free during much of the short summer period, they nonetheless rarely exhibit sorted stripes (only 25% of all observations (n = 217) fell within this arc). These periglacial observations provide

Table 1. Example of temperature differences between northern and southern aspects as shown by an east-west trending diorite dyke.

	Percentage of observations in the range:				
	0° to 7.5°C	1.5° to 4.5°C	2.5° to 7.5°C	$\bar{x}$ temperature	temperature range
North-facing	84.4%	60%	74.1%	5.3°C	19°(1° → 20°C)
South-facing	99.3%	46.7%	87.4%	4.6°	9°(0° → 9°)

(Record period: January 5th 1991 to February 19th, 1991)

Table 2. Example of differences in rock properties for north and south-facing* sides of rock outcrops as exemplified by an east-west trending diorite dyke.

	$\bar{x}$ Rind thickness (mm)	$\bar{x}$ indentor penetration (mm)	$\bar{x}$ Schmidt hammer rebound value
North-facing	1.57	0.21	41.9
South-facing	4.34	0.37	35.9
% difference	S > N: 176%	S > N: 76%	N > S: 17%

*The northerly aspect rapidly becomes snow-free whilst snow-lay is prolonged on the southern side.

two important pieces of information relating to the weathering regime. First, as it was only on the tops of south-facing slopes that became snow-free *very* early in the summer (i.e. pre December) that any stripe formation occurred (the 25% of observations) this indicates that during the snow-free period (± end of December to end of February) little freeze-thaw of any significance takes place (see temperature data below). If it *did* occur then, with time, so stripes *would* develop on these slopes as there is no shortage of moisture. Thus, the stripes that are found must develop prior to or after the late December to late February season. Second, all stripes are of the miniature variety and, although often very long (>100 m), have a mean coarse stripe width of only 5.6 cm (s = 2.5) and a mean fine stripe width of 7.4 cm (s = 3.8). Thus, those cycles that *do* affect the snow-free ground are, in the light of the small size of the patterns produced, of small amplitude. Clearly rocks still standing above the snow will suffer some freeze-thaw action until they are covered and it is likely that, given sufficient moisture, this is the main period when rock outcrops are subject to this form of weathering.

Rock temperature data collected during this summer season substantiate the periglacial interpretations given above. A summary of rock surface temperature information from a diorite dyke (Table 1) shows that during January and February no freezing events of significance took place (i.e., temperatures did not go below −2°C). However, whilst the temperatures do not indicate the occurrence of freeze-thaw they are, nevertheless, significant with respect to the weathering that takes place. Although there is only a 0.7°C difference in mean temperature between the north and south aspects of the dyke, it is the difference in temperature range between the north and south sides, and the concentrations of temperatures within these that are significant. As Table 1 indicates there is a 111% difference in temperature range between the north and the south sides of the dyke (19°C *vs.* 9°C) and this greater temperature range on the north side is reflected by 99% of readings for the south side occurring in the range 0° to 7.5°C but only 84% of those for the north side fall in this same range. However, located within the temperature band of 1.5° to 4.5°C are 60% of readings for the north side but only 47% of those for the south. In other words, whilst the north side is marginally warmer (on average) and experiences a larger temperature range, temperatures in general are concentrated in a lower range (1.5° to 4.5°C) than on the south. Thus, although the south does not experience such high, occasional temperatures, it is weighted towards higher general temperatures (2.5° to 7.5°C). This may have repercussions with respect to the nature of the resulting weathering.

A good indication of the degree of weathering is given by the combined data regarding weathering rinds, Schmidt hammer rebound values and micro-indentor penetration (Hall, unpublished data). Data for the east-west trending dyke (Table 2) clearly show the difference between the north and south faces: The weathering rinds (said to be a good indication of chemical weathering: Thorn, 1975) average 176% thicker on the south-facing side and this is mirrored by the greater (76%) penetration of the micro-indentor and the lower by 17% Schmidt rebound values for the south side. Thus, there is every indication that weathering in general, and particularly weathering processes requiring the presence of water, are more effective on the south-facing sides of outcrops. Data from the northern and southern aspects of other outcrops support this argument (i.e. $\bar{x}$ micro-indentor penetration for south-facing sides is 85% greater than for north-facing (n = 10) whilst rebound for north faces is 43% higher than for south (n = 10)).

The rock moisture data for the east-west trending dyke are very complex and show marked variations both for and between the north and south faces of the dyke (Hall, unpublished data). In summary, the data indicate that due to the prevalence of rain-bearing northerly winds so the north face is more frequently wetted and can attain high moisture

Table 3. Mass of material removed from 1 m^2 areas at four locations.

	Location	Amount of material removed (g)	% difference
1	N-face of dyke	4.6	33%
	S-face of dyke	6.1	
2	N-face of outcrop	45.0	562%
	S-face of outcrop	298.1	
3	N-face of dyke	26.0	546%
	S-face of dyke	168.0	
4	Backwall of snowpatch*	312.1	59%
	Wall edge of snowpatch*	196.5	

*Snowpatch faced to the south but main distinction is with respect to site (backwall) that has snow present for a long period *vs.* that (edge of snowpatch) where snow occurs for a shorter period.

levels. Equally, the north-face also dries more rapidly due to the effect of the sun and strong, dry northerly winds. The southern aspect only shows a higher moisture content when rain is brought by a southerly wind or there is snow accumulation on the south side. However, whilst the southern side does not attain high levels of saturation with such frequency neither does it dry so often or as rapidly as the northern side and so it, particularly when associated with melting snow, maintains a relatively wet status. This difference was clearly shown on the 4th of February 1991 when, after snow, south-facing samples at 1000 hrs, 1400 hrs and 2200 hrs measured 100%, 96% and 94% saturated respectively whilst those on the north measured 66%, 62% and 61%. Conversely, during wet, northerly winds moisture levels may be as much as 30% higher on the north face. However, in general, except when snow is associated with the south face, moisture levels are usually within ±10% of each other. Thus, the impact of the northerly winds is to cause frequent wetting and drying but it is the presence of snow that produces the longest duration of high moisture levels.

A further indication of this weathering difference is given by the amount of weathered material that could be removed from 1 m^2 on the north and south sides of outcrops. Material loose enough to be brushed off the rock face with the hand from a 1 m^2 was collected and weighed at four locations (Table 3). A difference of between 33% and 562% was found between north and south (snow-covered) sites. Again, the apparent effects of weathering suggest more weathered debris on the south compared to the north face. In general, the removed material constituted flakes less than 2 mm in thickness but of varying area.

In another attempt to gain some insight into weathering, but this time *not* aspect constrained, data regarding the number of fractured clasts within 1 m^2 on the different beach levels were measured. This showed that on the most recent beach level (*c*.3 m) of 152 broken clasts found in 1 m^2 only 14.5% were volcanics, i.e., it was the sedimentary rocks that broke down fastest. On the next beach (*c*.17 m) of the 110 clearly discernable broken clasts 81.8% were volcanics and on the next (*c*.40 m) 83.7% were volcanics. However, on the 2nd and 3rd levels most of the sedimentary clasts had completely disintegrated and it was only those that were discernable as a complete clast showing fracturing which were counted. In other words, very few volcanic clasts showed sign of breakdown on the first beach but their numbers increased significantly on the older beaches.

When put together, what do these various pieces of information indicate regarding weathering on the Byers Peninsula? First, that whilst *air* temperature records may indicate frequent freeze-thaw cycles during the spring to autumn period the presence of a protective snow cover means that very few actually affect *rock* other than that which is both snow-free *and* wetted. Thus, the extensive and substantial snow cover constrains the main zone of activity to those outcrops which, due to height or position, are late in becoming snow-covered and/or rapidly become snow-free. Second, that it must be primarily the north-facing sides of these rock outcrops which are affected but that, as the freeze-thaw cycles are of low amplitude and moisture contents are likely to be low, the total effectiveness is limited. That weathering rates by frost action are, at least on the diorite dykes, relatively slow is also indicated by the small amount of weathered debris and the limited number of fractured pebbles found on the lowest beach level. Thirdly, it is clear that more weathering takes place on the south-side of outcrops despite the period of longer snow lay. The reason is relatively complex but appears to involve a strong interaction between mechanical and chemical weathering such that although the former is limited its effect upon the chemically altered rock produces a more rapid rate of breakdown than is found on the less chemically altered northern exposures.

In the South Shetlands, due to the extensive almost continuous cloud cover (Simonov, 1977) there is rarely dynamic heating (and subsequent cooling) of the rock due to direct radiation. Thus, thermal fatigue does not play a major role in weathering the north-facing exposures. Salt weathering may play a role but data on interstitial rock water chemistry are not yet available and, whilst efflorescences were not seen, should salts be present, they would likely be active on the southern as well as northern sides. However, in the light of the greater number of wetting and drying cycles on the north side so there may be a greater frequency of salt crystallisation and/or hydration on this side. The effectiveness of this weathering would not normally be visible as it would be concentrated at a depth below the outer rock margin which would be little affected due to usually remaining wet. No clear indications of biological action were found on any of the dykes but as they were composed of dark, fine-grained material it is likely that they were not easy for organisms to penetrate and light transmission into the rock would be limited. Thus, these rocks do not constitute a suitable niche within this environment.

Put together what do these various pieces of information indicate regarding the present day weathering regime on the Byers Peninsula? In the absence of firm data to the contrary, it would appear that thermal fatigue and biological weathering have little or no impact on the landscape. Salt weathering is likely to be operative but, so far, its role can only be surmised. Thus it would seem that freeze-thaw, wetting and drying and chemical weathering (not necessarily in that order) are the dominant processes. Freeze-thaw is temporally and spatially constrained such that its occurrence is limited to wet, snow-free locations during either spring or autumn (no freeze-thaw having been recorded in summer). Within these limitations it would appear that it is the more frequently snow-free northern aspect that will be the most affected, but that effectiveness will depend on the presence of moisture in the rock. Freeze-thaw, will, though, still affect the southern sides of outcrops but may be limited to the wetted zone immediately above and below the ablating snow surface.

The relatively frequent, and often rapid, changes in rock moisture content found for the northern aspect would be highly conducive to weathering by wetting and drying, and, by implication, salt weathering. It can even be that three or more cycles can occur in one day and so the total number of wet/dry cycles per year can be substantial (>100), far exceeding the number of freeze-thaw cycles affecting the rock. What the actual contribution of wetting and drying is to the rock breakdown is unknown. However, wetting and drying alone can cause rock failure and so its role cannot be ignored. Most likely it contributes to rock fatigue such that other mechanical processes are more effective than would otherwise be the case. As the outer margin of the rock often remains wetted and it is within the rock that the fluctuations in moisture content occur, then, (as hypothesised by Hall, 1991), this may indicate that the effectiveness of the wet/dry cycles is at some depth below the surface where it is not visually apparent. As this is a maritime environment and the precipitation is (relatively) rich in salts, so the wetting and drying taking place in the rock may facilitate salt crystallisation and/or hydration. With the frequency of wetting and drying so there may be a commensurate action of salt weathering but, operating as it does at the same depth as wetting and drying, it may be difficult to determine its contribution.

Although the south side experiences fewer wet/dry cycles, the continued presence of moisture, combined with temperatures above 0°C are conducive to chemical weathering as evidenced by the weathering rind thicknesses found for this aspect. As Ballantyne *et al.* (1989, p. 750) state, in paraphrasing Thorn (1975): "... areas of maximum weathering rind thickness appear where free water in the snowpatch is concentrated, and ... that chemical weathering may be enhanced under late-lying snow simply because ... rock surfaces are exposed to prolonged wetting by percolating meltwater under late-lying snowbeds." The saline nature of the water will help enhance the role of chemical weathering. Wetting and drying, salt action and freeze-thaw *will* also take place on this aspect but, mainly due to the presence of the snow cover for long periods, will be much more limited.

In summary, it would appear that on the northern aspect wetting and drying, salt weathering and freeze-thaw predominate and that chemical weathering is limited (hence thin weathering rinds). Conversely, on the southern aspect, chemical weathering is of major importance and is aided by the mechanical processes that predominate on the northern side. Thus, freeze-thaw is not as ubiquitous as is often implied and chemical weathering may play a significant role contrary to most perceptions regarding Antarctic weathering regimes. The temporal and spatial variability of processes is recognized and, in this area at least, may be a major factor in the development of the landscape.

Conclusions

Due to the relatively large number of freeze-thaw cycles recorded for the climate of the South Shetland Islands combined with the occurrence of rain and snowmelt, many workers have suggested freeze-thaw to be the major weathering process for this region. However, from the data collected from diorite dykes on the Byers Peninsula of Livingston Island, it would appear that this judgement is in some need of re-evaluation. Whilst freeze-thaw cycles certainly occur in the air, they can be effective only on snow-free, wetted bedrock. This, it is suggested, may be limited and constrained primarily to northern aspects. The mechanical processes of salt weathering and wetting and drying are recognized as being potentially more active than freeze-thaw. However, all processes operate synergistically and thus it is difficult to discern the contribution of any one individual. Chemical weathering predominates on the southern aspects as is clearly seen by the thickness of the weathering rinds on this aspect. In broad terms, there appears to be an aspect-constrained process differentiation between predominant mechanical weathering, aided by chemical weathering, on the northern aspects and chemical weathering, aided by mechanical weathering, on the southern aspect.

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ENHANCED BEDROCK WEATHERING IN ASSOCIATION WITH LATE-LYING SNOWPATCHES: EVIDENCE FROM LIVINGSTON ISLAND, ANTARCTICA

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ABSTRACT

An indication of the extent of weathering on different aspects of rock outcrops on Livingston Island, Antarctica, was obtained by means of a Schmidt hammer, a cone indenter and measurement of weathering rind thickness. Results show that weathering, particularly chemical weathering, is enhanced on the lee side of outcrops where snow accumulates as a result of prolonged wetting by the melting snow. Rock moisture and temperature data indicate that the south-facing, snow-accumulation side of obstacles have high rock moisture levels and maintain relatively high temperatures. Whilst chemical weathering is greater on the leeward side of outcrops, mechanical processes are greater on the windward side. The presence of late-lying snow thus appears to exert a strong influence on weathering.

KEY WORDS Weathering Snowpatches Antarctica Schmidt hammer Indenter Weathering rinds

INTRODUCTION

The Schmidt test hammer has been employed by a number of workers (e.g. Augustinus, 1991, 1992; Campbell, 1991a; Day, 1980, 1981; Matthews and Shakesby, 1984; McCarroll, 1991a; Sjöberg, 1990; Sjöberg and Broadbent, 1991) as a means of rapidly assessing the degree of rock weathering based upon the premise that, for the same rock type, harder unweathered surfaces will give higher rebound values than softer weathered surfaces (Day and Goudie, 1977). Recently, however, there have been a number of comments regarding problems associated with the use of the Schmidt hammer (e.g. Williams and Robinson, 1983; McCarroll, 1989, 1991b) and other methods have been suggested to give better results (e.g. Allison, 1990, 1991). Despite the greater accuracy provided by techniques such as ultrasonics, the Schmidt hammer still provides a relatively cheap, robust field tool (Campbell, 1991b) that, if used with care, can provide an insight into weathering differences within the same lithology.

By means of the Schmidt hammer Ballantyne *et al.* (1989) investigated boulder weathering at snowpatch and snow-free sites in Switzerland, Scotland and Norway, and came to the conclusion that the presence of late-lying snow enhanced chemical weathering. This chemical weathering was considered to be as a result of '. . . prolonged wetting by percolating meltwater under late-lying snowbeds' (Ballantyne *et al.*, 1989, p. 750). McCarroll (1990) took these authors to task on a number of points with respect to the use of the Schmidt hammer and the meaning of the results that, subsequently, Ballantyne *et al.* (1990) were able to either refute or explain. Nonetheless, two major questions emerged, namely the value of the Schmidt hammer in such studies and the influence of late-lying snow upon rock weathering.

As part of an ongoing study of rock weathering processes in the Antarctic (Hall, 1992) work was undertaken in the 1990/91 austral summer on Livingston Island, South Shetland Islands. The Schmidt

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covered areas were avoided (Ballantyne *et al.*, 1989) and hammer recalibration was undertaken at each site. As measurements were taken only 2–3 m away from each other on opposite sides of dykes, lithology and lithologically constrained surface textures could be considered similar. Measurements were always taken with the hammer in a horizontal position. A minimum of 100 values were obtained for each side but without multiple blows to the same spot; rather they were undertaken in as small an area as was practical according to the constraints set out above.

At two of the study sites, weathering rind thicknesses were measured. Thorn (1975) showed that weathering rinds can reflect the availability of free water and hence chemical weathering, with thicker rinds indicating sites of enhanced moisture availability. A combination of thick rinds and low rebound values might be expected at sites particularly prone to chemical weathering. As a further refinement, use was also made of a National Coal Board (NCB) cone indenter (Szlavin, 1974; Hall, 1987). This indenter, which measures the amount of penetration of a tungsten carbide cone for a known applied load (West, 1991), works at the micro-level. The weakening of a rock surface by chemical weathering will result in greater cone penetration compared with a harder, less weathered surface. The NCB indenter is less prone to the constraints imposed upon the Schmidt hammer (i.e. proximity to joints, etc.) and is particularly amenable to use on fine-grained rocks such as were found in the study area. Between 50 and 100 indenter measurements, according to the procedures set out by Szlavin (1974), were undertaken at each sample site.

Of direct relevance to the study, but dealt with in detail elsewhere, were measurements of air temperatures, radiation receipts, wind direction and speed as well as rock temperatures at north-, south-, east- and west-facing exposures and on a horizontal surface. Combined, these data provide the background information required for an understanding of weathering processes on Livingston Island (Hall, in press). Rock temperature data are critical for the determination of weathering mechanisms and yet data are extremely limited (McGreevy and Whalley, 1982; McGreevy, 1985; Hall and Hall, 1991). In fact, it is the inadequate data-base pertaining to rock temperature and rock moisture content that is the main constraint upon our understanding of weathering processes: 'Foremost among . . . problems has been uncertainty concerning the thermal and moisture regimes which actually prevail within natural bedrock and regolith fragments' (Thorn, 1992, p. 10). McGreevy and Whalley (1985) clearly discussed the importance of rock moisture data in weathering studies whilst Hall (1991) demonstrated how its temporal and spatial variability could greatly affect both the nature and timing of weathering. Thus, the availability of temperature and moisture data to this study provides a firm foundation upon which to base judgements regarding the character of the weathering as indicated by information from the Schmidt hammer and indenter.

Additional confirmatory information regarding the nature of freeze–thaw cycles and the predominance of snow on south-facing exposures is provided by sorted patterned ground. Sorted forms, mainly stripes, were almost exclusive to north-facing slopes and were all of small width (*c.* 7–10 cm) and shallow depth (2–10 cm). Their non-occurrence on south-facing slopes is a result of the longevity of snow-lay on that aspect. In addition, the fact that sorted forms were rare on ridge tops and on the tops of south-facing slopes shows that freeze–thaw cycles are few in summer (and the micromet data confirmed this). Further, the small-scale nature of the features is suggestive of small amplitude freezes. Thus, field observations and collected climatic data do *not* appear to support the occurrence of summer freeze–thaw whilst the thick snow cover and small scale of the patterns do not indicate a more powerful winter cycle, which, with respect to weathering, is also constrained by the absence of rock moisture.

RESULTS AND DISCUSSION

Schmidt hammer data (Table I) clearly indicate the influence of late-lying snow in effecting weathering. Mean hammer rebound values for east–west trending dykes (1–10 in Table I) are consistently higher for the snow-free northern sides compared to the snow-accumulation southern aspect. The variability in mean hammer rebound value between dykes is probably a reflection of the variability of dyke lithology (Smellie, *et al.* 1980) and hence the nature and degree of weathering. The difference between snow-accumulation and snow-free sides of obstacles is also reflected at site 11 where the east face was the one against which snow accumulated. This dyke clearly showed process differentiation between the two faces, with the east side exhibiting gravel to

Table II. Kolmogorov–Smirnov tests of Schmidt hammer data

Site	Median		Interquartile range		DN	Accept/Reject
	N	S	N	S		H_0^*
1	42	36	11	12	0·38	Reject
2	36	28	9	8	0·50	Reject
3	41	27	10	8	0·72	Reject
4	34	27	10	8	0·52	Reject
5	42	28	8	9	0·82	Reject
6	42	29	6	15	0·66	Reject
7	43	38	6	9	0·48	Reject
8	38	29	9	11	0·62	Reject
9	42	29	8	11	0·78	Reject
10	38	32	13	12	0·34	Reject
11	29	16	7	6	0·68	Reject
12	49[†]	38	7	10	0·54	Reject
13	40[‡]	31[§]	5	12	0·68	Reject

* H_0: The data for the north side are so similar to those of the south side such that they can be considered part of the same population. H_a: The data for the north side are so dissimilar to that for the south side that they must be considered two different populations. Rejection level set at 1 per cent ($D = 0{\cdot}326$)

† Due to local topography north side has greater snow accumulation than south side

‡ Rocks at edge of snow (as per Table I)

§ Rocks at snowpatch centre (as per Table I)

granule size material whilst the west side was broken into blocks (*c*. 30 × 6 × 4 cm) that frequently remained *in situ*; the presence of these loose blocks helps explain the low mean rebound value for the western aspect. The influence of snow is also seen at the rock boss study site (12) where, due to the local topography causing extensive snow accumulation against the northern, eastern and western faces, it is the southern face that has the marginally greater mean rebound value.

The influence of snow in accentuating weathering, particularly chemical weathering, is given by the data from sites 13, 14 and 15. At site 15, the mean rebound value from high (± 6 m) on the cliff-face is greater than that for the snow-affected lower (± 2 m) part of the cliff. At site 14, small gulleys occur as a result of exploitation of the closely spaced (*c*. 5 cm) jointing along which enhanced chemical weathering occurs. Here, the effects of chemical weathering were most obvious as the rock not only has a very low mean rebound value (24·1) when compared to that of the between-gully faces (42·4), but also the rock in the gully shows a marked colour change due to chemical alteration. Site 13 is a snowpatch hollow on the lee side of Point Smellie. Values obtained here from rocks on the hollow floor and from the bedrock backwall area clearly show the influence of the snow. The rocks located at the margin of the snowpatch area showed a mean rebound value far greater than those from the centre of the snow accumulation area, whilst the bedrock at the snowpatch margin gave a larger value than was found for the backwall at the snowpatch centre.

Thus, the Schmidt hammer consistently yielded lower values at sites subject to snow accumulation compared to adjacent snow-free locations on rock of the same lithologic composition (Table II). It would thus appear that the presence of snow markedly affects the nature and extent of weathering. The information from weathering rind measurement and indenter penetration tests corroborate the Schmidt hammer results.

Weathering rind measurements (Table III) obtained on dyke 1 (site of Schmidt hammer 1) indicate a mean thickness of 1·57 mm ($s = 1{\cdot}16$) for the north-facing snow-free side, whilst the south-facing snow-accumulation side had a mean thickness of 4·34 mm ($s = 2{\cdot}20$). These values clearly reflect the greater amount of chemical weathering taking place on the south side, thereby confirming that differences in such weathering are responsible for the differences in mean rebound values at this site. The importance of the snow acting as a source of moisture is clearly indicated by the weathering rind thicknesses found at the rock boss study site. Here, as the boss lies in the lee of a slope, snow accumulation is greater against the *north face* than the

Rock temperatures are also important with respect to weathering, particularly in facilitating or impeding chemical weathering. Synthesis of available rock temperature data indicates that, during the summer period, the northern aspect has a marginally (0·7°C) higher mean temperature and a wider temperature range compared to the southern aspect (19°C *versus* 9°C). However, despite the higher mean and larger range the bulk of values for north-facing aspects are concentrated in the range 1·5–4·5°C whilst those for south-facing aspects are in the range 2·5–7·5°C. Thus, during the summer months when the snow is melting, the south-facing rocks do not experience temperatures less conducive to chemical weathering than their north-facing counterparts. During this same summer period, no freezing events of more than a few hours duration or below −2°C were recorded, and so freeze–thaw is deemed not to be as important a process as has often been thought (Hall, in press). With temperatures, at least during the summer, not being a limiting factor it appears that availability of moisture is critical in promoting weathering.

Rock moisture data indicate a complex situation with much spatial and temporal variability. However, three main facts emerge. First, that the north side receives much direct precipitation in the form of rain and thus is frequently wetted, but at the same time rain-free northerly winds rapidly dry the rock. Second, during periods of snow-melt the rocks on the south side are frequently at, or near, saturation. Third, that despite the prevalence of precipitation-bearing northerly winds, the south faces are not significantly drier (mean for north of 50 observations = 57·5 per cent saturated, whilst that for south = 50·7 per cent saturated). The most noticeable factor is that although the north side experiences many frequent wetting and drying cycles, the south remains at a high moisture status for prolonged periods. Thus, as proposed elsewhere (Hall, in press), it is suggested that whilst the northern aspect may experience marginally more mechanical weathering, particularly in the form of wetting and drying, so the south side experiences more chemical weathering.

Returning to the discussions between Ballantyne *et al.* (1989, 1990) and McCarroll (1990), it would appear that first, the Schmidt hammer *does* provide a rapid indication of differences in the degree of weathering and, second, that enhanced weathering of rock in association with snow *can* take place. Certainly care must be taken in the use of the hammer, but the results from both rinds and indenter support the results obtained in this study. That the presence of snow has enhanced rock weathering accords with the findings of Ballantyne *et al.* (1989). The available information indicates that the outermost 10–20 mm of the rock are ≧50 per cent saturated for the summer period, with values exceeding 80 per cent when in direct contact with melting snow. During this time rock temperatures are ≧5°C for sustained periods.

With respect to the Antarctic environment, it is interesting to note the important role played by chemical weathering. The general perception of the Antarctic is of extremely low temperatures and a very limited moisture availability. In the maritime Antarctic environment, however, temperatures are not so severe and moisture, including rainfall, is more generally available during the short summer period. Thus, chemical weathering can play a significant role in rock breakdown. In addition, as few freeze–thaw events were recorded during the summer and much of the rock is insulated by a protective snow cover for most of the autumn to spring period, so the role of freeze–thaw weathering may be more limited than has been thought previously (Hall, in press). Equally, weathering due to thermal fatigue is unlikely due to the extensive cloud cover experienced in this region during summer. Thus, the two mechanical weathering processes with the greatest potential for operation (but still mainly constrained to the summer period) are salt weathering combined with wetting and drying (Hall, in press).

CONCLUSIONS

The Schmidt hammer provides a useful field tool for weathering studies and, for use in remote areas such as the Antarctic, has the advantages of sturdiness and ease of operation that may make it a more viable proposition than more sophisticated tools. Ancillary data regarding both weathering rinds and indenter penetration support the findings indicated by the hammer, suggesting, at least for this situation, that it alone provides a good indication of weathering differences. The presence of snow clearly affects weathering, particularly chemical weathering, which in all instances was greater on the snow-accumulation faces of rock exposures. This conclusion supports the contention of Ballantyne *et al.* (1989) that the presence of late-lying snow enhances chemical weathering. From available information regarding rock temperatures, rock

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Cutbacks in the Natal Drakensberg Escarpment: an hypothesis on their origin

Kevin Hall

Several cutbacks in the basaltic escarpment of the Natal Drakensberg are noticeable for the distinctive channel systems found at their lower ends. These channels and their complementary enclosing borders, or levees(?), are composed of fluvially derived, cobble- to boulder-sized debris. While the fluvial origin of these features, as displayed by both their morphology and sedimentary characteristics, appears obvious, no exposition as to the timing and nature of their formation is available. It would appear that extreme, high-energy events are necessary to produce such deposits and landforms but it is difficult to envisage under what conditions. It is suggested here that meltwater from a rapidly ablating ice body could be responsible; recent studies in the vicinity of the Antarctic Peninsula reinforce this hypothesis. If this is the case, then the origin of the cutbacks and their associated deposits would be at the end of the last glacial rather than a result of present-day fluvial processes and is indicative of some form of ice cover on the Drakensberg.

The distribution channels at the lower ends of cutbacks, such as that at Bannermans Pass, are characterized by high borders, often in the order of 10 m, composed of large boulders. These basaltic boulders frequently show evidence of substantial rounding. The distribution system of channels develops as the pass widens at its lower end and, at an altitude of about 2 100 m, can have a lateral extent of 100 m. Rounding of the boulders is evidently fluvial in nature, rather than due to weathering, as they are often juxtaposed against sub-angular blocks of comparable size. The whole, then, is the product of fluvial action down the short (< 1 km), steep (~ 30°) cutbacks. The question that arises is, what is the source of a sufficient volume of water that could generate such a substantial deposit as well as rounding the boulders during transport within such a limited downslope distance? Even with blocks of basalt readily available to be transported, because of weathering, it would still require an enormous volume of water to move material often greater than 2 m in diameter with sufficient energy to produce a high degree of rounding within a travel distance of barely 1 km. Certainly the steep slope would help increase water velocity, but what is the source of the large volume of water?

An immediate answer is precipitation, particularly from high-energy events such as Cyclone Demoina. However, three factors must be considered. First, that at the top of the escarpment the general downward slope is not towards the cutbacks but rather into Lesotho. Second, the presence of large volumes of water in these channels during contemporary high-energy events does not necessarily mean they could form the features: it may simply be that the water is using a pre-existing low as a drainage line. Third, the high geomorphic influence of events such as Cyclone Demoina is expressed at lower elevations where water has coalesced from a variety of upland and local sources, while at higher elevations catchment areas are very small and therefore the fluvial input is reduced. In other words, one cannot transfer the magnitude of impact of such events at low elevations to their possible effects at high locations.

If, then, one is sceptical of the rain-induced, contemporaneous origin of these features, what possible alternatives are there? Whilst it is tempting to consider debris flows of some sort, features such as boulder streams[1] and Sturzstroms[2] are distinctly different from those observed in the Drakensberg. A debris-flow hypothesis would still require a cause and, not least, one that would explain the rounding of the boulders (not convincing with respect to debris flow *per se*). Further, even if the fluvial action that produced the distribution channels were said to post-date the debris flow, the arguments remain much the same with the need to explain how such dynamic fluvial activity could take place over such a short distance. However, one clearly recognizable source of high discharge that can have a dramatic geomorphic effect in mountain areas, even over very short distances, is that of melting snow and/or ice. Recent discussions[3,4] have argued in favour of both substantial snow accumulation, in the form of nivation hollows, and of glacier ice as evidenced by cirques. With the Drakensberg being a borderline cryogenic area, any such snow or ice would melt very rapidly during the amelioration of climate (as has been argued for the lower latitude, lower altitude environment of Marion Island[5]) and the large volumes of water released by ablation of ice in Lesotho would pour down the incipient cutbacks. If a debris-flow model is preferred, then this water may have either initiated the debris flow and subsequently contributed to the erosion or would have moved down the already existing cutback and incised the pre-existing debris. The presence of snow and ice on the top of the escarpment would have helped negate the drainage of meltwater back into Lesotho. It is envisaged that the situation could have looked somewhat like that depicted in Foto 2e of Meier,[6] in which a small glacier is shown tumbling down a very steep gully on the edge of the Öksfordjökel plateau glacier in Norway. Further, if there were ice on the plateau, then it is highly likely that niche glaciers would have existed in the embayments along the edge of the escarpment. The characteristic form of niche glaciers developed in resistant rocks[7] bears a striking resemblance to the cutbacks and embayments of the Natal Escarpment. With the recent arguments in favour of nivation and cirque glaciation,[3,4] a model could be envisaged encompassing the development of the cutbacks by nival processes, during which debris and debris flows would be produced.[7] The subsequent meltwaters from ablating snow and ice would have then cut in to those deposits to produce the distribution systems we see today.

Although the above hypothesis, as presented here, lacks detail to substantiate it, information on clast size and shape, sediment thickness, clast fabrics, and the form of these deposits is available and will be published in detail (Grab, pers. commun.). Despite studies[8–10] that have produced palaeoenvironmental reconstructions based upon evidence from nearby locations, a main problem is our inadequate knowledge of Quaternary conditions in the Drakensberg. One possible way forward may be by analogy. Recent studies on Alexander Island (73°S, 68°W), Antarctica, indicated features very similar to those observed in the Drakensberg. Deglaciation of Alexander Island has left a number of large V- to cone-shaped cutbacks above the Mars Glacier. Although only a few hundred metres in vertical extent, these cutbacks are very steep (about 40°). At their lower ends, abutting a flat bench along the margin of the Mars Glacier, are to be found distribution channels, levees and rounded boulders akin to those of the Natal Drakensberg. The only major difference is that the features on Alexander Island are somewhat smaller in size.

The present-day snowfall on Alexander Island is small, the term 'oases' being coined for the ice-free areas that are found here.[11] Therefore, the cutbacks and

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their distribution channels were developed during deglaciation and are being maintained by the present snow melt. The cutbacks all face roughly NW as a result of the ice lasting longer on the aspect-protected south-facing side, such that the meltwater ran down the cutbacks to the north–north-west as the Mars Glacier ablated both down and away from the valley wall. Snowmelt during this time would have greatly abetted their growth, and the retreat of the Mars Glacier left an area in which the distribution channels could develop unhindered. The features bear a striking resemblance to those observed in the Drakensberg and an alternative origin (such as one where the effect of glaciers and their meltwater is not involved) is difficult to envisage for Alexander Island. No such cutbacks were observed on the south-facing slopes of the nunataks as they still support an ice/snow cover.

Finally, if such an hypothesis were accepted for the Natal Drakensberg, then it might help to explain the thick sequences of fluvial beds found at lower elevations, such as at Injasuti. Here the present-day river cuts down through earlier fluvial sequences that often show beds composed almost exclusively of boulders. Whilst, as proposed earlier, such sequences could be the product of storm events such as Cyclone Demoina, equally they would be the logical consequence of large glacial or nival outpourings.

A call is made for more detailed research on the Drakensberg. More needs to be known regarding the climatic and geomorphic conditions during the Quaternary; the question of whether the Drakensberg was glaciated or not is important. At present we have more questions than answers, but a number of features, the large cutbacks of Natal being one, appear to have analogues in present-day glacio-nival areas.

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Piscivory in the Comoro Islands flying fox *Pteropus seychellensis comorensis* – a refutation

R.E. Stobbs

The author of two recent publications[1,2] claimed that the flying foxes of the Comoro Islands, *Pteropus seychellensis comorensis* (Megachiroptera; Pteropidae), are facultative piscivores that snatch their living prey from the sea surface in the manner of the African fish eagle, *Haliaetus vocifer*. In the first article, Balon wrote,

'These large bats are normally nocturnal and fruit-eating, but during our expedition in 1987 we noticed that, in addition to being diurnal and crepuscular, the bats catch fish in the sea in a similar way to fish eagles *Haliaetus vocifer*. The bats then fly onto land with the fish in their claws, hang inverted on a tree, and eat their catch (Fig. 19). Although it has been reported that these bats may scavenge when their natural habitat has been disrupted by agricultural developments, we could not find any mention in the literature of this active fishing habit. Are we witness to evolution in the making?'

And, in a second report, he stated,

'.... Instead, large bats were seen soaring over the land and sea during the broad daylight hours; their numbers increased somewhat at sunset. I had read before about these fruit-eating flying foxes, but expected them to be active only at night. Soon we observed .... fruits were not their only food. Those individuals soaring over the sea often splashed, swallow-style into the waves. .... By sheer luck one bat dove into the sea .... and when taking off flew toward me with a twitching silver fish in its claws. It landed on a kapok tree behind me, hung upside down in typical bat fashion, and started to eat the fish. I could not identify the fish, but saw the bat's sharp teeth tearing it .... The bat catches the fish in a fashion nearly identical to that of the African fish eagle *Haliaetus vocifer*, ... I searched in the library but found no mention that the flying foxes are diurnal or fish-eating'.

The article goes on to say that,

'It was obvious that we had discovered another fishing bat in addition to the small one known from the Amazon (which dives like a kingfisher rather than swooping in like an eagle)'.

This article has been written to set the record straight; to refute these misleading and inaccuarate claims.

Eight species of frugivorous flying fox in the genus *Pteropus* occur in the western Indian Ocean, where they inhabit many of the larger islands but not the mainland of Africa.[3–6] In the Comoro Archipelago two species of *Pteropus* and one *Rousettus* are known; the common *P. s. comorensis* (Fig. 1) inhabits the lowlands and coastal regions of all the islands and is found also on the African east-coast island of Mafia, and the rare, endangered *P. livingstonei*, which inhabits the highest, mountainous regions of Anjouan and Moheli and is in urgent need of protection.[3,7]

P. s. comorensis is a common flying fox in the Comoro Islands where it is known as *ndema*.[4] Flying foxes are generally treated as vermin in the Comoros, a situation very different from that prevailing on many Indo-Pacific islands, where they are considered such a culinary delicacy that many species are endangered as a result of the activities of hunters and poachers.[5]

Balon claims that he was unable to find reference to the fact that flying foxes are known to fly during daylight hours. The facts are that *Pteropus* spp. bats are well known to be crepuscular, and often diurnal and there are, indeed, many references to this in both the scientific and popular literature (which were not difficult to find) (refs 3–6 and numerous others).

Pteropus species lack the complex neural and behavioural mechanisms associated with ultrasonic echo-location and must therefore navigate entirely by sight. It follows, then, that since flying foxes are unable to fly successfully in total darkness (although they may be seen flying during moonlight nights or in proximity to well-lit buildings and townships) they are obliged to fly during hours of reasonable visibility. Pteropid bats living in regions where there are few or no predators (such as the Comoro Islands) are able to fly at any time of the day and are often seen flying to their feeding grounds

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Cutbacks in the Natal Drakensberg Escarpment: comments on an hypothesis on their origin

Hall has outlined an hypothesis for the origin of cutbacks along the Natal Drakensberg Escarpment. He postulates on the formation of debris in cutbacks and on the fluvial distribution systems within the debris. These are attributed to the rapid ablation of ice bodies (niche glaciers) within the cutbacks and the melting of snow and ice in Lesotho. While the understanding of the geomorphology of the Drakensberg is still in its infancy and the question of glaciation in the higher mountains in southern Africa is highly relevant, there are points that need to be raised concerning the hypothesis. The purpose of this response is not to provide an alternative interpretation but simply to highlight the problems with the hypothesis in order to prevent misconceptions emerging from the related process interpretations and associated climatic implications.

The hypothesis is based on the presence of 'fluvially derived, cobble to boulder size debris' of which the 'origin ... appears obvious' and that 'extreme, high energy events are necessary to produce such deposits and landforms'. The high-energy events are then attributed to meltwaters. Hall accredits the rounding of boulders as 'evidently fluvial in nature, rather than due to weathering, as they are often juxtaposed against sub-angular blocks of comparable size.' The petrology of basalt, however, as opposed to that of sandstone, is thought to be predisposed to rounded clasts[1] and will weather to rounded clasts[1] particularly in cold climates, where rock temperatures are high and conditions wet.[2] Basalt, prior to movement, can already be highly rounded *in situ* by weathering. Rounded basalt clasts are frequently found on the escarpment, where no fluvial action or clast movement could have taken place. The presence of rounded basalt in a deposit does not then necessarily imply high-energy fluvial transportation. The juxtaposition of rounded and weathered material will militate against the *in situ* rounding of clasts in the deposit but not against rounding prior to transportation and deposition if there was any lateral or vertical spatial difference in debris source location. As a consequence the fluvial nature of the debris cannot be assumed by clast appearance alone, particularly since no quantitative data on the degree of rounding and on the location of rounded and angular clasts in the deposits were cited in support of the hypothesis. If the rounding of the boulders can be explained by some other mechanism, then this immediately questions the basis for the hypothesis, namely that 'a rapidly ablating ice body could be responsible' for the 'extreme high energy events'.

In apparent contradiction to the fluvial-rounded hypothesis Hall later attributes the debris to 'nival processes, during which debris and debris flows would be produced'. He states that the 'subsequent meltwaters from ablating snow and ice would have then cut in to those deposits to produce the distribution systems we see today'. The role of the meltwaters is therefore not clearly defined. Are the deposits fluvially derived, or are they formed by debris and debris flows from outwash, with the subsequent incision to form the channels? If the deposits are from a mass movement event or events as suggested, how can 'The whole' then be 'the product of fluvial action'? Although the later scenario presented appears to be more plausible it must be stressed that in the absence of quantitative data this is only a presumption, yet the associated climatic implications are considerable.

Some further comments on the contribution of meltwaters are appropriate. In negating the effects of precipitation (presumably rainfall) Hall states that 'at the top of the escarpment the general downward slope is not towards the cutbacks but rather into Lesotho', yet later he postulates that 'the large volumes of water released by ablation of ice in Lesotho would pour down the incipient cutbacks'. In explanation, Hall states that 'the presence of ice on top of the escarpment would have helped negate the drainage of meltwater back into Lesotho'. Gradients immediately beyond the escarpment above Bannerman Pass are in excess of 10° sloping into Lesotho. Therefore meltwater contributions would have been negligible unless there was a substantial accumulation of snow/ice above the top of the Escarpment. Such accumulations would, for example, require ice thicknesses in excess of 80 m a distance of 500 m away from the escarpment for the surface of the ice/snow just to slope back into Natal. The ice/snow thicknesses would have to be well in excess of the escarpment height at the top of the cutback for 'large volumes' of meltwaters to have contributed to processes in the cutback. In addition, the area immediately above Bannerman Pass is north facing. The volumes of snow/ice on the shaded, colder, south-facing slopes and in the valley floors must then have been considerable. Some clarification of the scenario is required because the large volumes of snow/ice envisaged would suggest glaciation of the higher regions of Lesotho. While such a situation is not impossible, the geomorphic–climatic implications are nonetheless considerable.

With reference to precipitation and events such as Cyclone Demoina, Hall is correct in stating that 'one cannot transfer the magnitude of impact of such events at low elevations to their possible effects at high locations.' It is also apparent, however, that one cannot transfer the magnitude of events in recent history to those operating in the past. Indeed, based on personal observation of the features referred to in the hypothesis, one may just as readily construct an hypothesis for a warmer, moister climate, with intensified weathering and high-intensity rainfall events, which, in association with structural control of the basalt, have formed the cutback and the features in Bannerman Pass. Such an hypothesis would superficially be equally plausible, implying different processes and climatic conditions, but will be equally as flawed as the meltwaters hypothesis. The point again is that without quantitative data on the channel systems and the debris, process–form relationships cannot be established. Perhaps in recognition of this shortcoming the author refers to recent work by Grab (referenced as pers. commun.) on debris in the Drakensberg. This work will not, however, serve to contribute to explanations of these debris features as Grab concentrates on the high regions of a cutback in the southern part of the Drakensberg where the debris morphology differs greatly from that described in the hypothesis (Grab, pers. commun.).

In the hypothesis Hall attributes the development of the cutbacks to 'nival processes'. Nival processes cannot fully explain the origin of the cutbacks. If 'niche glaciers' were present or 'nival processes' operative in the 'embayments along the edge of the escarpment' then these would have been in pre-existing cutbacks of some form. The hypothesis, contrary to what the article's title suggests, serves rather to explain the origin of the debris and channel systems within the cutbacks, rather than the origin of the cutbacks themselves, as some form of embayment would have to exist for the accumulation of snow/ice.

Finally, analogies are drawn with observations of cutbacks on Alexander Island, Antarctica (lithology not specified). Although Hall mentions 'recent studies in the Antarctic Peninsula', no references to previous or current research are provided to substantiate the arguments. While both areas may have similar morphological features the process of development is disputable unless quantitative assessments are undertaken, and even then may still remain debatable. Although recent publications show Hall to be familiar with the geomorphology of high-latitude areas, observation-supported hypotheses as 'One possible way forward' must be treated with caution. It appears that the situation described by Butzer[4] 21 years ago on 'periglacial' phenomena in southern Africa, that 'too much interpretation has been based on high latitude preconceptions' (p. 9), still exists.

The intention of this discussion was not to provide an alternative hypothesis for the formation of the features but to highlight problems with inferring process and related climatic conditions from qualitative judgements. Research on geomorphology in the Drakensberg may still be in its infancy but this does not preclude sound geomorphological assessment. The arguments presented here are best summarized by a number of recent comments pertaining to geomorphological research in southern Africa. Le Roux[5] highlights the problem of 'inferring causes from outcomes' (p. 129), which Hall[6] in support notes 'has been a major problem with South African periglacial studies' (p. 135). In reference to southern African cryogenic research, Hall[2] notes that 'qualitative presumptions have been made and deterministic hypotheses built upon them' (p. 69) and continues 'Great care and rigour must be applied to *any situation* otherwise a completely misleading picture may result that could confuse and/or prejudice subsequent studies' (p. 69, emphasis mine). Interpretation of these features has great potential for unravelling past climatic conditions, providing avenues for geomorphologists to offer significant contributions to palaeo-climatic research. Hall[2] states that 'whilst glaciation of the higher southern African mountains is such a big question, with so many climatic-periglacial-botanical ramifications, great care must be taken in arriving at any judgements' (p. 71). The closing comments go to Butzer: 'southern Africa has experienced cold, glacial age climates, but there is a serious problem about many of the geomorphological observations and their interpretation' (p. 1), and to Hall:[2] 'A plea is made for more detailed work and greater exactitude in future studies' (p. 69).

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Response

These comments are useful, contribute to the discussion and highlight the inadequacies of my hypothesis. Thus, they are testing the hypothesis and, as with most postulates, finding problems that require the original to be adjusted. I hope that the author will now go out and get some of the required data and adjust the hypothesis accordingly.

A comment is made in the second paragraph regarding the rounding of boulders. I certainly do suggest that, qualitatively, the rounded shape of the large basaltic boulders is indicative of fluvial action and that their juxtaposition with angular material militates against their having been weathered to this shape *in situ*. However, I do not see how a reference to the chemical weathering of basalts, and a mathematical discussion at that, as cited in the response by Sumner, is applied to cold climates. It is true that basalt could be weathered to rounded shapes prior to transport such that no high-energy fluvial transport is required. But, where is the evidence for this? Yes we do see rounded, *in situ*, boulders in the basalt – but are they pre or post the age of the deposit under debate? If they post-date it, then we need to explain their occurrence prior to the deposit. If they pre-date it, this will not explain the close proximity of rounded and angular clasts. However, fluvial and/or mass flow could explain the juxtaposition of angular and rounded clasts? Also, one might note that high-energy fluvial transport would also explain the observed rounded-but-fractured clasts.

I refer to 'nival' processes and the movement of debris as a result of these. I tried to make it clear that there could have been mass movement and then the cutting in to the debris by fluvially transported material '... subsequent meltwaters ... would have then cut in to those deposits ...'. To me it was clear. How is ice of < 80-m thickness at 500 m from the escarpment, so that it could slope back into Natal, calculated? Why could ice simply not end at the escarpment edge, with small lobes down the 'incipient' cutbacks? The geomorphic-climatic implications of ice in Lesotho are significant – but then so are those of not having ice! In other words, we currently have no idea what the situation was.

Sumner goes on to suggest that the channel systems might have been '... formed more gradually by fluvial action'. How does a long period of rainfall create the deep channels in a coarse boulder deposit with an enormous void ratio? I cannot but think there is a problem of semantics in which it is suggested my hypothesis serves to explain the debris within the cutbacks rather than the origin of the cutbacks themselves. However, unless the origin of the debris is not associated with the cutback, then surely this origin pertains to the origin and development of the cutback?

Regarding the relationship of cutbacks on Alexander Island to those in Natal, there may be a telecorrelation problem. However, two researchers from Natal independently observed similar morphological features in the two places. The comments by Butzer and Hall cited by Sumner must be put in context. We may be using high-latitude preconceptions and this may be wrong. However, the real problem with some of the preconceptions by South Africans is that these researchers have no experience of high-latitude regions and so make judgements based on reading alone.

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CRYOPLANATION TERRACES: SOME NEW INFORMATION

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V2N 4Z9

Abstract

Cryoplanation terraces are "A step-like or table-like bench cut in bedrock in cold climate regions" (Harris et al., 1988, p.23). Although there appears to be a substantial body of literature regarding cryoplanation, it is still far from clear as to the actual formative processes involved, the distribution of these features or their significance. Considered by many to be diagnostic of permafrost terrain, these features are, in general, thought to be a function of freeze-thaw weathering in association with snowbanks ("nivation") and to occur in continental periglacial areas of moderate aridity (Harris et al. 1988). A survey of the literature indicates that cryoplanation terraces are cited from two main areas: the Arctic and Central Europe. Those in the former area often being considered active whilst those in the latter inactive: the Arctic often being regarded as the modern analogue for cryoplanation terrace development in Europe during the Quaternary. Recent Arctic studies (Nelson, 1989) have argued that, contrary to earlier suggestions, cryoplanation terraces are not climatically determined phenomena and that they may represent the periglacial analogue to the glacial cirque. However, as Nelson himself states (1989, p.33) "Few climatic data are available for sites with active cryoplanation terraces" and this problem is further compounded by the fact that "The main gap in present-day knowledge of cryoplanation is the deficit in reliable information on the processes involved" (Priesnitz, 1988, p.63). A major problem with the cryoplanation concept is the centrality of nivation to it's initiation and development, and the presumption of frost weathering as the 'foundation' of nivation. A number of recent studies (e.g. Thorn and Hall, 1980; Thorn, 1988; Hall, 1980, 1985; Nyberg, 1986, 1991) have all questioned the role of nivation and of the freeze-thaw mechanism in particular. As Thorn (1992, p.18) states "Both cryoplanation and nivation are examples of concepts that presently defy satisfactory operational definition because they embrace many component parts" and that, contrary to the suggestion of Nelson (1989, p40) who considers that cryoplanation "..could serve to relate periglacial studies more closely to the mainstream of geomorphology..", "It is the unquestioned acceptance of debatable concepts such as these which must be abandoned if the discipline is to progress". Despite all of these shortcomings some authors (e.g. Clark and Hedges, 1992) still feel able to use cryoplanation to explain fossil features and landforms.

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"Mountain Benches" or "Cryoplanation"?: A Tautulogical Conundrum

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Non-glacial benches on mountain sides in present or former periglacial regions are somewhat on an enigma in terms of their formative prosesses and the terminology that surrounds them. Cryoplanation, and a whole variety of synonyms (e.g. altiplanation, periglacial glacis, goletz terraces, etc.), is said to produce bench-like features at various elevations on mountain sides in periglacial regions. The question arises as to what actually constitutes "cryoplanation"? Does it only describe form or does it have process attributes? If there are specific periglacial processes involved, then how do they differ from those that constitute nivation, a term often used almost synonymously with cryoplanation? Recognising that nivation is itself currently being questioned as to it's efficacy and role, then is it really viable to have a process-landform assemblage (i.e. cryoplanation) that depends on an equally enigmatic process collection (i.e. nivation). This is partially a terminological problem and partially one of process recognition. The term "mountain benches" is suggested to be less "fuzzy" as it does not have aby of the process or structural connotations that tend to confuse the various other, more generally used, terms. This problem will be discussed in the context of some recent data from an area of active mountain bench development in Antarctica. Although the data are limited they do offer some insights in to the process attributes and thus the significance of fossil forms found in Europe and North America. It will be seen that the concept of cryoplanation is fraught with confusion and contradictions and may, like has suggested for nivation, be a term best dropped from periglacial use.

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tread. The formation of fines, and the growth of segregation ice as a result of this (and which produced massive bedrock heave in some locations) may be an additional factor that helps explain the occurrence and development of terraces in this region.

The information presented here is not meant to "explain" cryoplanation terraces but it does offer new information and from an area not taken in to account in any of the discussions hitherto. Perhaps this Antarctic data may help throw new light on Arctic perceptions of cryoplanation and to further the discussion of what these features really mean within the periglacial context.

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Considering that, despite the extensive ice cover, the ice-free regions of Antarctica cover a wide range of climatic and permafrost zones, there are very few references to cryoplanation. In fact, of the hundreds of references searched, in a range of languages, few (e.g. Strelin and Malagnino, 1992) categorically refer to "cryoplanation terraces" (for James Ross Island which is a continuous permafrost area with an arid climate) although a number of others allude to cryoplanation but do not name it (including Markov and Bodina, 1960 who refer to both "nival terraces" and "corrasion terraces"). This seems surprising considering the range of present and past climatic zones that occur, many of which are good analogues for much of the Northern Hemisphere during the Quaternary. As a personal observation, despite being able to visit a range of areas from the sub-Antarctic through to nunataks on the margin of the polar plateau, the observations here presented are the first I have encountered regarding the occurrence of cryoplanation in Antarctica.

Studies were undertaken on nunataks in the region of the Mars Glacier (71°50'S, 68°21'W) at the southern end of Alexander Island. This is a region of continuous permafrost with horizontally bedded mudstones, shales and sandstones. The terraces comprised risers of highly weathered material 0.8 to 2 m in height with treads sloping between 1° and 5° orientated primarily north through west to south and of width 2 to 10m and length 6m to several hundred metres. Bench development was weakest on eastern and south-eastern aspects. Detailed summer temperature data (at one minute intervals) are available and year round information at 20 minute intervals for three years). Thus, it is hoped to characterize the air and rock temperature regime. The treads of the terraces showed a form of patterned ground (see below) in the upper part but drainage runnels (c.2cm deep and 3cm wide) in the lower, steeper part of the tread. Detailed studies were undertaken regarding the weathering regime which showed that freeze-thaw weathering was, as in most of the continental part of Antarctica, limited not by temperature (rock temperatures attained values of +29°C and frequently showed diurnal freeze-thaw events) but by moisture availability. Being an arid region, very little snow falls during winter or summer and only small (c.40cm deep) snowbanks were observed at the back of the terraces as spring progressed to summer. Processes other than freeze-thaw may well play the greater role in weathering.

An adjunct observation, that may or may not be pertinent to the development of the terraces, was that of a new form of "patterned ground" on the upper parts of the terraces that consisted of "sorted, non-sorted patterns". Being a relatively snow-free, cold, continuous permafrost zone, ice-wedge development was extensive. However, being developed in mudstones and shales there was also the production of fines such that the non-sorted crack pattern was outlined by sorting of the surficial material, even where the patterns were developed in bedrock. The nature of these patterns will be discussed elsewhere. There importance here may be that, developed as they are in continuous permafrost and in bedrock, they help break up the rock, or at least crack it apart, and so allow for a more rapid breakdown by the weathering processes. As the riser wears back it continuously encounters, or is broken in to by, the wedges. Thus, the weathering back of the riser is aided and perhaps accentuated by the development of ice wedges on the

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Observations on "cryoplanation" benches in Antarctica

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Abstract: A series of benches on nunataks of Alexander Island (Antarctica) are described. An increase in bench size with distance away from the retreating glacier suggests an age spectrum. The benches have thermal contraction cracks (in bedrock) on shallower, upper sections of the risers as well as salt encrusted runnels on the steeper lower section of the tread. The benches also show a distinct orientational preference (orientated to the north through to west) and, from first principles, these seem to be the aspects with optimal freeze-thaw cycles and temperatures conducive to thermal stress fatigue. The extensive dilatation associated with the retreating glaciers is thought to play a significant role in the origin and development of the benches as the combination of extensive jointing and optimal process conditions are thought to constrain where benches begin. The jointing, aided by the thermal contraction cracking, then facilitates extension and continued weathering of the treads. It would appear that these benches are examples of so called "cryoplanation terraces" that have been reported as fossil forms in Europe and North America. The study of such active forms in the Antarctic may provide good analogues for fossil features found in the Northern Hemisphere.

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Key words: Antarctica, cryoplanation, dilatation, mountain benches, patterned ground, permafrost, weathering

Introduction

Although there is a substantial body of literature on cryoplanation (e.g. Demek 1969a & b, 1972, 1980, Boch & Krasnov 1951, Dylik 1957, Pinczés 1968, Reger & Péwé 1976, Evans 1994, Czudek 1995) the actual formative processes involved and the distribution of these features is still far from clear. At present the literature indicates that cryoplanation features comprise gently sloping erosion surfaces (treads) with steeper, weathered slopes (the risers) at their upper edges. They may frequently occur in sequences, one above another, but rarely circle an entire mountain. They seem to be associated with mechanical weathering (freeze-thaw usually being cited) on the risers but with mass movement on the treads; the whole is usually perceived as related to nivation. Geographically, they are frequently located in zones of continuous permafrost, but altitudinally are in zones with adequate snow to provide summer moisture, plus a substantial number of freeze-thaw cycles (although this last point seems in many cases to be based on air temperature data). Based largely on climatic inferences, Nelson (1989) suggests that the cryoplanation terrace is the periglacial analogue of the glacial cirque. A number of recent studies (Thorn 1988, Thorn & Hall 1980, Hall 1980, 1985, Nyberg 1986, 1991) have, however, questioned the general concept of nivation as well as the ubiquitous acceptance of freeze-thaw in cold regions (Thorn 1992, Hall 1995).

The concept of cryoplanation (or any of its synonyms) is fraught with problems of terminology, process and, even its very existence as a separate entity (i.e. as distinct from nivation). There are few climatic data available to understand the process, and since the majority of studies from the Northern Hemisphere are, in the development of broad conceptual models and analogues, there has been no inclusion of the Antarctic. Indeed, the three most recent general discussions regarding the spatial distribution of these features, and their formative processes (Priesnitz 1988, Nelson 1989, Czudek 1995), do not cite any examples from Antarctica.

There is thus a need to include the Antarctic and other Southern Hemisphere information in the development of any general models. Interestingly, despite the variety of climatic and permafrost zones in Antarctica, there are very few direct references to cryoplanation. Strelin & Malagnino (1992) refer to the presence of cryoplanation terraces on James Ross Island, a dry (c. 137 mm annual precipitation), cold (mean annual temperature -13.5°C) island to the east of the Antarctic Peninsula, in an area of continuous permafrost. Transverse and longitudinal nivation hollows are also cited as occurring in this area – forms that are considered associated with cryoplanation. Corté (1983) also refers to cryoplanation terraces along the east coast of the Antarctic Peninsula. Sekyra (1969), working in the dry 'oases' of East Antarctica, also refers to "cryoplanation relief". Although no detail is given, Sekyra (1969, p. 286) states that with respect to remodelling of the relief the greatest influence is glacifluvial action and "cryoplanation processes (origin of frost-riven cliffs, cryoplanation terraces, steep slopes, bounding-valley depressions, etc.)" – features that sound to be related more to 'nivation' than cryoplanation *per se*. Although Bockheim (1995), in a recent review of Antarctic permafrost and associated features, refers to the absence of terraces in the

Antarctic interior, Jordan & Van der Wateren (1993) do identify cryoplanation terraces in the ice-free region of Litell Rocks, in north Victoria Land; a cold, dry continuous permafrost region. Markov & Bodina (1961), in their map of periglacial formations in Antarctica, do not directly describe cryoplanation but they do show "nival terraces" – noted specifically at Cape Burn in Victoria Land, a region of very low mean annual temperatures and continuous permafrost. Myagkov (1979, p. 12) states that in the oases areas of McMurdo Sound, the "final stage of mountain planation is (the) 'deserption pediment' inclined by 30°"; a form which may be related to cryoplanation. Aniya & Hayashi (1989) refer to the "step-like topography" on exposed bedrock 272 m above sea level in East Antarctica and state that (p. 143) "At present, it appears that periglacial processes such as freeze-thaw, gelifluction and nivation are strongly working in this area". In much the same way, Simonov (1977), working in the maritime Antarctic environment of the South Shetland Islands, refers to (p. 227) "The flat-topped hills and stepped ledges... (that) ...owe their origin to nivation, which is extremely intense... (such that) ...nivation and frost weathering are among the present-day relief-forming processes". Other workers refer to neither cryoplanation nor nivation but do cite "terrace-like projections" (Korotkevich *et al.* 1973), "flat surfaces and step-like topography" (Aniya 1988), mountain "flank erosion surfaces" (Stephenson 1966), "block terraces" (Nichols 1960), "stepped slopes" (Miotke 1984) and, in the area of the present study, "stepped topography" (Taylor *et al.* 1979). In the Falkland Islands, Clark (1972) refers to altiplanation benches that are mantled with rubble derived from 'frost-riven crags'. This paper describes in detail benches discovered on nunatacks at the southern end of the Antarctic Peninsula.

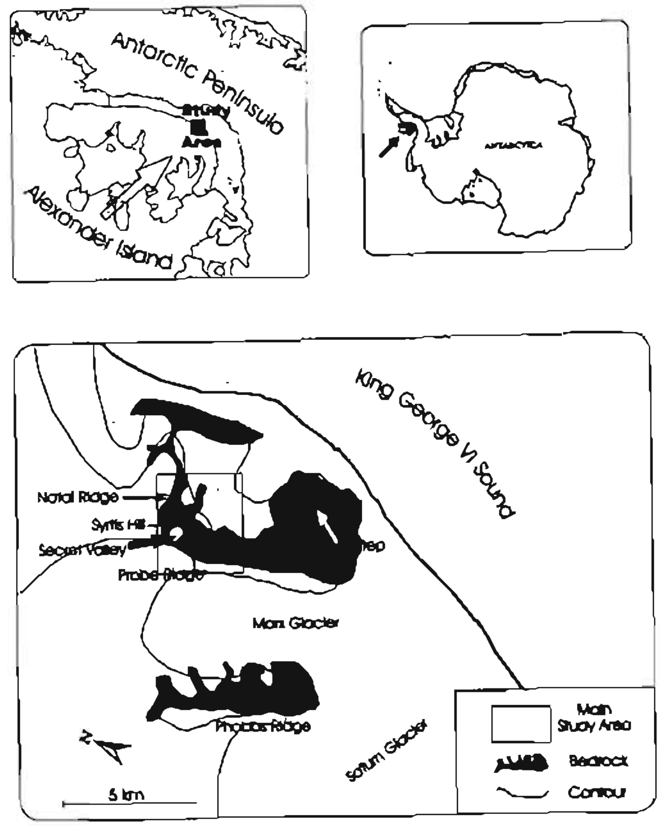

Fig. 1. Maps to show location of study area.

Study area

The study was undertaken (Fig. 1) on nunataks along the northeastern side of the Mars Glacier at the southern end of Alexander Island (71°50'S, 68°21'W). This area has a number of ice-free valleys and ridges, the term 'oases' being coined for such locations on Alexander Island (Stephenson 1938), which experience a cold, dry, continental type climate. Although there are no climatic stations near enough to the study area to be representative, data from loggers located in Viking Valley (Fig. 1) only 200–300 m below the surrounding peak tops have provided some information (Table I). Air temperatures have a summer mean of -2.4°C and winter mean of -11.5°C, with a winter minimum value of -35.2°C and a summer high of +6.0°C. Rock temperatures, however, show a very different picture (Table I), with a mean summer value of +5.9°C (low of -4.4°C, high of +24.4°C) but winter mean, minimum and high only slightly warmer than those of the air. Details of rock temperatures are given in Hall (in press). It is a region of continuous permafrost, with 0°C temperature and frozen ground being encountered at *c.* 0.27 m depth on 13 December 1992 on Syrtis Hill (Fig. 1). The active layer is thought to be in the order of 0.3–0.4 m thickness at the most open, snow-free locations and substantially thinner nearer to longer lasting snow (0°C at 0.15 m depth 1m from snow and 0°C at 10 mm depth at snow margin on same date).

Geologically the area consists of sandstones, conglomerates and argillaceous sedimentary rocks (Taylor *et al.* 1979). Much of the rock comprises arkose sandstone with sub-spherical, post-compaction concretions (Moncrieff 1989), sometimes leading to the name "cannonball sandstone" (Horne 1965). These siliceous nodules, held by a ferruginous

Table I. Temperature data (°C) from Viking Valley for the period 8 December 1992–7 August 1993.

	Summer mean	Summer maximum	Summer minimum	Winter mean	Winter minimum	Winter maximum
air temperature	-2.4	+6.0	-6.1	-11.5	-35.2	+1.9
rock temperature	+5.9	+24.4	-4.4	-11.4	-33.3	+2.6

cement whose age is considered to be post lithification (Thomson 1964), readily weather free. Mudstones, sometimes with lime carbonate nodules, shaley mudstones, and both light-coloured and dark-coloured orthoquartzitic sandstones are also found. These sedimentary rocks are horizontally, or near-horizontally, bedded. Faulting determines the linearity of the coast and the east-west orientation of the glaciers that cut the coast (Taylor *et al.* 1979). All the nunataks were covered by ice during the last glacial maximum, with an ice mass centred on Alexander Island (Sugden & Clapperton 1978). The small glacier that runs from the col between Syrtis Hill and Probe Ridge down in to Viking Valley (Fig. 1) is currently receeding as indicated by the end moraines found in the valley.

Observations

The benches were observed on and between Syrtis Hill and Probe Ridge, along Natal Ridge to the northeast and between Probe Ridge and Two Step (Fig. 1). The benches had risers 0.8–2.0 m in height, comprising weathered bedrock, and treads 2–12 m wide sloping at an angle of 1–10°, composed primarily of weathered debris (Fig. 2). The treads extended laterally 6–200 m. The smaller benches (*c.* 0.8 m riser, tread 2–3 m long and 6 m in lateral extent) were somewhat arcuate in plan and found lower down on the col between Syrtis Hill and Probe Ridge. They were discrete entities but, in some instances, appeared to be close to joining together. From their lower location and their proximity to the retreating Viking Valley glacier it is surmized that they are younger than the larger forms found above them closer to, and around, the ridge tops. The larger forms sometimes showed a break in slope on the tread, with a shallower upper section at an angle of 1–4° and a steeper lower section at an angle of 5–10°. Noticeably, the upper section displayed well formed circular or polygonal patterned ground, (sometimes grading into garlands), whereas the lower part of the tread exhibited well developed drainage runnels. Although both patterns and runnels could be observed in the other sections of the tread they were not well developed there. The patterns were *c.* 1 m in diameter, or cross-tread dimension in the case of garlands, with coarse borders 0.1–0.15 m in width. The borders were not just sorted but also showed a central crack *c.* 0.1 m in width within which the majority of the coarse debris accumulated (see below). The runnels were spaced *c.* 0.3–1.0 m apart, were 0.03–0.15 m in width and 0.02–0.10 m in depth, and were normally marked by either efflorescences of gypsum on the clast surfaces and/or a concentration of coarse debris (due to the removal of the fines) giving the impression of sorted stripes. Snow was observed at the base of the riser on some benches during December but was usually not very deep (<0.40 m) nor laterally extensive. The benches could be seen to have developed along lithologic boundaries.

Fig.2 View along the tread of a typical bench; riser at the back of the bench is on the left-hand side of the photograph (the riser is *c.* 2 m high).

The cliffs of Viking Valley and along the ridge from Probe Ridge to Two Step showed very marked dilatation sheets as a result of the glacial retreat. Although composed of horizontally bedded sediments, the cliffs exhibited extensive vertical cracks parallel to the relieving surface. These vertical joints were not observed on the ridge tops away from the cliff edges, i.e. away from the location of the recently receded ice cover. Jointing was also more closely spaced at the outer cliff face with increasing distance between joints, until their total disappearance, several metres in to the cliff (as viewed on the crumbling cliff tops).

The benches showed a distinct orientational preference. The best developed benches and components of a bench were orientated to the north-northwest. Benches were less well developed or, on the longer benches, phased out towards the south and northeast. The lithologic boundaries could be perceived by a slight break of slope but benches were not developed on the easterly aspects. A bench or succession of benches could be followed around Probe Hill with well developed features facing west and north, the features would diminish in height and width to the south and immediately to the east of north, and not be found at all facing east (Fig. 3). The best developed patterned ground was found on the treads of those benches orientated to the west–northwest. These aspect constraints occurred despite occurrence of the same lithologies and exposure of the same boundaries around the peaks and on both sides of the ridges.

The patterned ground forms are dealt with in detail elsewhere but warrant some comment here as they are developed on the benches and may be related to their formation. The patterned ground appears to be a previously undescribed variety that comprises a 'sorted, non-sorted' form (Fig. 4). The initial feature is believed to be a non-sorted ice wedge since the features are developed in continuous permafrost, exhibit a crack and, significantly, are found in bedrock (Fig. 5). Two mechanisms, that may be complementary, can explain the subsequent sorting. First, the weathering products of mudstones and shales can be sorted once there are sufficient fines to allow the development of segregation ice. The presence of segregation ice is indicated by bedrock heave and

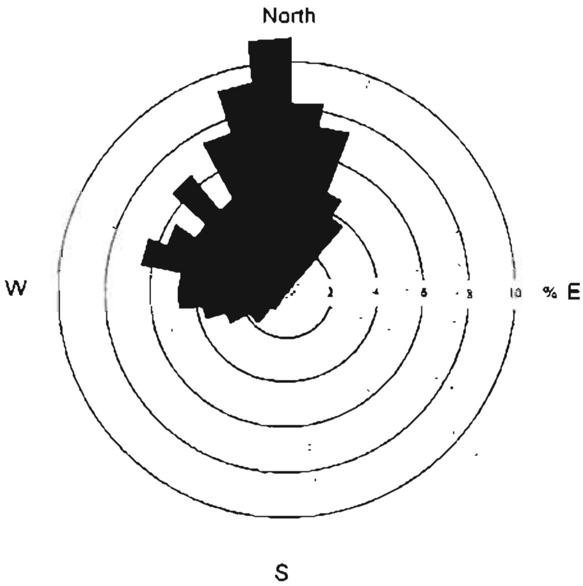

Fig. 3. Rose diagram to show the preferential orientation of benches.

by the occurrence of other sorted features close to the terraces. The second possible mechanism, is that of weathering of the bedrock along the crack junctions allowing coarse material to accumulate along the crack line. Observations certainly support this latter suggestion for, in several instances, with polygons of either raised or non-raised borders, including some in bedrock and not in weathered debris (Fig. 5), the borders are seen to have a distinct crack in which blocky material, of the same lithology as that in which the cracks have developed, has developed *in situ*. As there is no mechanism available for this debris to have been transported and then deposited in the crack, it must be associated with weathering of the rock around the crack. The whole is argued in detail elsewhere and it is only their presence that is significant, in the first instance, here.

Fig. 4. View of the 'sorted, non-sorted' patterns found on the upper part of the benches (features are *c.* 1.6 m in diameter).

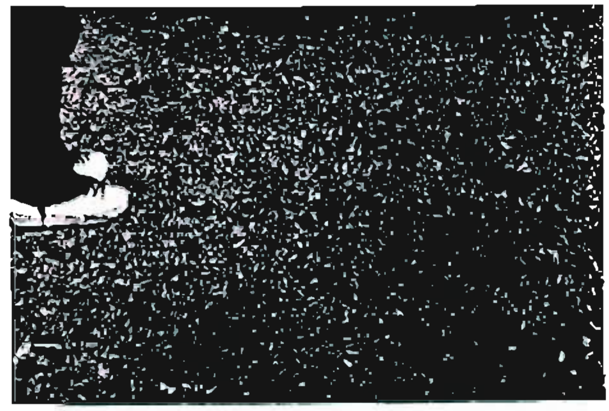

Fig. 5. To show the thermal contraction crack in bedrock.

Discussion

Snow, primarily as a source of moisture, is a major factor in either cryoplanation or nivation. In the study area, however, snowfall is extremely limited. In December, when much of the winter snow still remained on the ground, all but the base of the riser on the terraces were snow-free. Thus, unlike in the cryoplanation model suggested by Nelson (1989), elevation

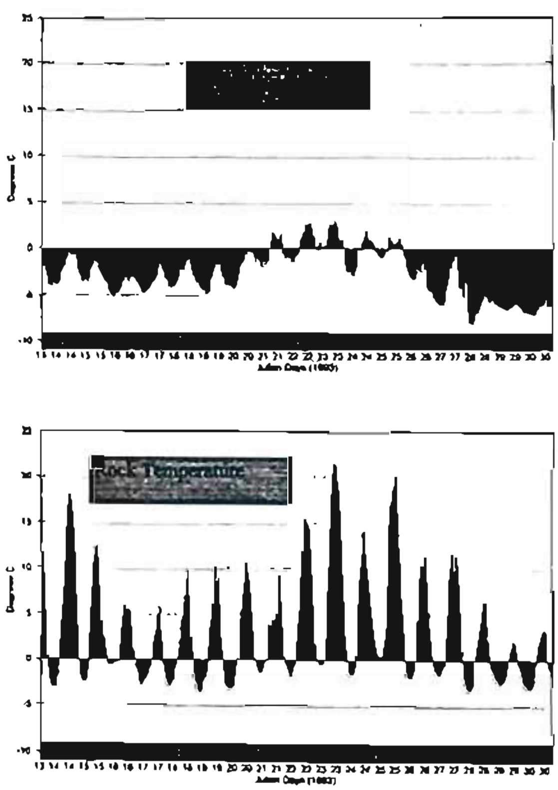

Fig. 6. Air and rock surface temperatures from Viking Valley.

constraints on snow accumulation are not applicable, either in terms of the availability of snow or snow as a protective element. The use of air temperature data in that model is considered misleading. Although the required "very small thawing indices" (Nelson 1989, p. 33) are found in the air temperature data, they do not reflect what happens to the rocks – where the actual weathering takes place. An example of the large difference between air and rock temperatures is shown in Fig. 6 where it can be seen that during the period the air experienced six freeze-thaw cycles (defined here simplistically, as three crossings of 0°C) the rock experienced 14; with diurnal events on all but one day. Rock temperatures clearly followed daily radiation patterns (r^2 = +0.84). The most significant difference between air and rock was, for the period shown, that air temperatures did not go above +3°C but the rock attained a surface temperature of +22°C; the recorded rock minimum was -4°C whilst that of the air was -8°C. A further example is that for the period 13 January–18 March (60 days encompassing "summer") the maximum air temperature was +5°C, the minimum -13°C with 16 freeze-thaw cycles whilst the rock surface had a high of +22°C, a minimum of -10°C and 49 freeze-thaw cycles (Hall in press).

Priesnitz (1988), in his review of literature on cryoplanation, notes orientation with respect to snow accumulation is often more important than orientation with respect to insolation. In the present study, the preferred orientation (Fig. 3) to the north-northwest indicates an orientation receiving a large amount of radiation. However, this orientation need not reflect snow accumulation as a result of wind directions or pre-existing topographic hollows. Rather, it could be that it is the sunward slopes that experience the greatest amount of geomorphic activity. On shaded slopes, even with a large body of snow, geomorphic activity will be low as there is no meltwater to drive it and only a very shallow active layer in which cryogenic processes can operate. On the sunward aspect there is the potential for weathering processes, mass movement, fluvial activity (small scale) and frost action in the ground. Therefore, it is argued that in high latitude situations protective shading could be detrimental to bench development and that it is the sunward slopes that experience the greatest geomorphic activity and thus exhibit benches; the whole being accentuated where the greatest snow accumulation, because of winds and the presence of topographic hollows, is also on the sunward slopes. In fact, measurements of taffoni size and occurrence, coupled with Schmidt hammer rebound values, show that the eastern aspect experiences the least weathering whilst the northern and western exposures have the greatest amount (Hall in press).

A variety of factors could be involved in the initiation and development of benches. In these horizontally bedded sediments it is the vertical dilatation joints that offer a ready means of exploitation commensurate with extension of the bench into the nunatak. It was most noticeable that the joints were being exploited insofar as blocks were seen to be loose and "toppling" along the weathered face of the riser (Fig. 7). The intersection of these joints with the horizontal bedding planes should provide ideal conditions for accentuated breakdown, but since dilatation effects had no orientational preference, other aspect-constrained processes must be involved in producing the observed orientation of the benches.

Whilst thermal contraction cracks cannot explain initiation, they may aid and accelerate growth. A number of authors have referred to the occurrence of both sorted and non-sorted patterned ground on benches (e.g. Demek 1969a) but have not linked it to bench development. Many of the benches at the higher elevations had well developed non-sorted patterned ground that frequently exhibited sorted margins (Fig. 4) and it was clear that these features were developed by the cracking of bedrock rather than in unconsolidated sediments as is usually the case with thermal contraction cracks (van Everdingen 1994). The vertical cracking of the bedrock by thermal stress further enhances the effects of the vertical dilatation joints by introducing an additional cracking component.

Weathering is suggested by all authors to play a major role in bench development and it is almost always freeze-thaw weathering that is perceived as the most active agent. With respect to freeze-thaw weathering, the requirements are (in broad terms) that there be water available in the rock and that it actually freeze. Although no rock moisture data are available, this is an arid region and ultrasonic data collected on the valley floor indicate dry rock conditions (except when wetted by glacier melt running down the valley). On the ridges there is certainly melt (rather than just sublimation) during the summer but the available snow is minimal. However, snow does accumulate at the break of slope between risers and treads of benches and it was here that noticeable undercutting of the bedrock risers takes place. The presence of water seems to be complemented by the necessary temperature requirements as there are certainly many diurnal freeze-thaw events at, and in the top few millimetres of, the rock surface. Thus, the potential for freeze-thaw weathering

Fig. 7. View across the tread of a bench showing the weathered bedrock on the riser.

does occur but would appear to be most effective, given moisture availability, on the western to northern aspects.

The availability of moisture from snow melt allows for weathering by wetting and drying; a slow process but one that exerts an effect in its own right and which also operates synergistically with freeze-thaw (Hall 1988,1993). Gypsum coatings were found on the rocks throughout the area and efflorescences of this mineral were also seen along the runnels on the lower sections of the benches. The occurrence of water also suggests that chemical weathering might occur as has been recently suggested (Balke *et al.* 1991) for Antarctica. However, no sign of chemical weathering (e.g. rinds) were seen in the rocks and so, partly because water supply was so limited and also because of it's low temperature (usually close to 0°C), chemical effects are considered minimal or non-existent.

Temperature ranges on the rocks were significant and here it was noticeable that the western and northern exposures experienced a far greater range than did the eastern – frequently in the order of 50% greater range and 65% higher temperatures (e.g. 19 degree range with maximum of 15.5°C on the eastern aspect and 28.5 degree range and maximum of 25.5°C on the northern and western aspects during 11–15 December 1992). The rate of change of temperature was far greater on the western and northern aspects and close to the 2°C per minute, proposed by Yatsu (1988) for thermal stress fatigue. Thus, with bedrock already fractured by dilatation joints and abetted by thermal contraction cracks, the potential for destruction by thermal stress fatigue must be considered a real possibility.

It is suggested that it may be thermal stress fatigue that first starts to exploit the pre-existing lines of weakness in the bedrock to initiate the benches; with the accumulation of snow this is enhanced by the addition of freeze-thaw, wetting and drying, and possibly salt weathering.

With the largest benches at the higher elevations and the smallest closer to the valley floor in proximity to the retreating Viking Glacier it is suggested that size is closely related to age. Temperature data and observations of such as the patterned ground and runnels in the weathered debris support the contention that these features are still actively growing.

In answer to the question of Priesnitz (1988), the weathering processes appear to work faster on the riser than on the surrounding surfaces. The key may well be that initiation is a combination of close, extensive jointing on the optimal aspect. Thus, contrary to the suggestion of Büdel (1982), it would seem that mountain benches (the result of "cryoplanation" in his terminology) are actively forming at the present time.

Conclusion

The origin of benches on mountain sides in periglacial regions appear to be far from clear and the absence of actual data regarding conditions on active benches has only served to confuse the issue. The forms here reported and referred to as "benches" do appear to fit to the general definitions of 'cryoplanation terraces' and the available evidence certainly indicates that these are active features. The limited data suggest that weathering processes are potentially more varied and effective on the northern to western aspects which could explain the preferred orientation of the benches. It is suggested that bench initiation could well be a function of preferential process occurrence coupled with extensive jointing as a result of dilatation. Once initiated, the banking of snow along the riser/tread junction enhances the weathering effects. The development of thermal contraction cracking on the tread together with on-going weathering and removal of debris by mass wasting and fluvial action (as evidenced by the runnels) both extends and maintains the bench form. Freeze-thaw weathering may certainly play a role but cannot be considered to be the dominant process without supporting evidence; thermal stress fatigue in cold regions with high radiation inputs may play a greater role. Clearly the evidence presented here does not answer the many questions regarding 'cryoplanation' but it does offer new information and indicates that such features are still forming. Finally, it would seem that the Antarctic might be an ideal place for further studies on these features, and hence a potential source of information that could be applied to the re-interpretation of fossil forms found in the Northern Hemisphere.

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SASQUA XIII

DELEGATES, PROGRAMME

AND

ABSTRACTS

OF

PAPERS

DID SOUTH AFRICA REALLY LOOK LIKE THIS DURING THE QUATERNARY? QUESTIONS REGARDING SOME RECENTLY SUGGESTED PERIGLACIAL LANDFORMS AND SEDIMENTS

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During the past decade there has been a resurgence of interest in Quaternary periglacial conditions in South Africa. These new studies have identified a range of periglacial landforms and sediments, upon the basis of which stadials and interstadials have been named. Such as nivation hollows, rock glaciers, ice wedges, protalus ramparts and stratified sediments are said to occur in their fossil form. The correct identification of these features is extremely significant as they can frequently be used as palaeoenvironmental indicators and, as such, provide valuable information to other Quaternary scientists. The geographical position of South Africa means, however, that few local scientists have little idea of exactly what those environmental conditions are. Equally, there is little ability to critically evaluate the suggested occurrence of some of these features. Few may realize the environmental constraints of the actual occurrence of rock glaciers or ice wedges, with the necessity of a mean annual air temperatures of -6°C or lower, continuous permafrost, winter temperatures in the order of -40°C and little snow. At the same time, the naive association of angular clasts being solely associated with frost action and hence "proving" the periglacial origin of the sediment and landform is not recognized by many geomorphologists. Thus, it is easy for landform and/or sediment identification to err, for supposed environmental conditions necessitated by the landform to be at odds with other available information, and for the information to be grossly misleading and, indeed, in many instances to be unfounded.

Based on modern analogues for the supposed South African periglacial climates, an attempt will be made to show the sort of conditions which would be required for such landforms. Further, the fallacy of the "angular clast equates to frost action" will be demonstrated and, to further confuse the issue, examples of rounded clasts formed by cold region mechanical weathering will be shown. The overall aim is to critically contest many of the suggested reconstructions, including showing contemporaneous protalus ramparts forming in the absence of any permafrost (contrary to the supposed permafrost association alleged from South African interpretations), in order that a better understanding of Quaternary periglacial conditions in this region can be arrived at.

NIVATION OR CRYOPLANATION: DIFFERENT TERMS, SAME FEATURES?

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Abstract: Nivation and cryoplanation are considered to produce separate landforms—hence their distinct names. However, despite this duality of terms, both utilize the same basic processes. A review of the literature indicates confusion and overlap regarding nivation and cryoplanation and their meaning. First, nivation is clearly described in its literature as exploiting pre-existing hollows. Thus nivation cannot be the initiator of cryoplanation as is frequently suggested in the cryoplanation literature. Second, nivation requires snow, usually in some quantity, which must experience melt, whereas cryoplanation is said to characterize arid or semiarid regions. Third, many authors associate cryoplanation with permafrost, whereas this association has never been the case with nivation. Fourth, if both utilize the same basic suite of processes, what then justifies this terminological dichotomy? It is here argued that 'nivation' and 'cryoplanation' are two end members of the same feature/process suite.

A transition from nival to arid conditions is accompanied by a decrease in the action of water and, likely, an increase in the potential for permafrost (due to the diminished insulation from snow cover). The "core" processes remain the same as, in essence, does the landform—the transverse nivation hollow/cryoplanation terrace. The use of two terms, particularly if they are associated with climatic assumptions, only serves to confuse. This is particularly the case when paleo-reconstructions are based on these landforms. Is it possible to distinguish between fossil "cryoplanation terraces" and "transverse nivation hollows"? Is there any real difference? If the answers are "no," then we cannot justify two terms. Rather, there is a single process suite that has "wet" and "dry" end members and which produces similar landforms, but probably at distinctly different rates.

INTRODUCTION

There is a substantial body of literature regarding both cryoplanation (e.g., Oreshkin, 1935; Boch and Krasnov, 1951; Dylik, 1957; Richter et al., 1963; Demek, 1964, 1969a, 1969b; Pinczés, 1968; Reger, 1975; Reger and Péwé, 1976; Péwé et al., 1982; Priesnitz, 1988; Nelson, 1989; French and Harry, 1992; Czudek, 1995) and

[1]I would like to thank the British Antarctic Survey for allowing me to work with them on Alexander Island which brought to the fore, for me, the question of "cryoplanation." Dr. Colin Thom, as always, provided stimulating questions on the entire issue of "nivation"/"cryoplanation" and Dr. Jef Vandenberghe highlighted many of the terminological concerns surrounding "cryoplanation." Dr. Toni Lewkowicz is thanked for bringing us all together in the stimulating environment of his excellent field excursion to Ellesmere Island. Support for this work was provided, in part, by NSERC grant 185756. I would like to thank a number of anonymous referees who contributed a wide range of thoughts and concerns on this topic, as well as Dr. Nel Caine, who was extremely helpful in improving the manuscript and its structure. Their input greatly assisted in the rewriting of this contribution, but all failings remain solely those of the author.

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nivation (e.g., Matthes, 1900; Ekblaw, 1918; Lewis, 1936, 1939; McCabe, 1939; Boch, 1946, 1948; Paterson, 1951; Henderson, 1956; Cook and Raiche, 1962; Nichols, 1963; Lyubimov, 1967; St. Onge, 1969; Thorn, 1976, 1978, 1979, 1988; Hall, 1980, 1985, 1993; Thorn and Hall, 1980; Rapp, 1983, 1984; Nyberg, 1986, 1991; Rapp, et al., 1986; Grab and Hall, 1996). Despite this profusion of literature, there still appears to be substantial confusion as to the exact nature and modus operandi of both these groups of processes and, indeed, as to whether they are actually distinct from each other (Thorn, 1988). In both cases, the terms are well entrenched in the literature and are used as both a collective noun and as an adjective. As a collective noun each term is used to identify a particular process suite, whereas, as an adjective, each is used to describe the landforms that result from that process suite (Thorn, 1988).

Part of the problem, and the confusion, arises because both process suites are remarkably similar (Table 1). Furthermore, nivation often is cited as the precursor to cryoplanation (e.g. Czudek, 1995, p. 100) and yet, although cryoplanation is identified as being able to *initiate* a landform (e.g., Richter, et al., 1963), nowhere in the nivation literature is nivation seen as an initiator but rather as an *exploiter* of a preexisting hollow. Thus, the cryoplanation literature identifies terraces as being initiated by nivation, even on "...the original smooth slope..." (Czudek, 1995, p. 102) whereas the nivation literature clearly shows nivation as an agent of "...modification and not initiation," where a "hollow" must *already* exist in which the snow can accumulate (Thorn, 1988, p. 19). In addition to all this, a further issue emerges. As is seen in Table 1, and shall be discussed in detail below, *both* process suites identify "frost weathering" (or any of its synonyms) as *the* major weathering process; indeed, Richter et al. (1963, p. 183) stated that cryoplanation can (in translation) "...be explained only by processes that are typical for the zone of frost shattering." Examination of Table 1 (below) and Table 1 of Thorn and Hall (1980) shows that frost weathering also is identified by all authors as fundamental to nivation. The problem is that "field corroboration is something of a misplaced concept with respect to frost weathering," and that "at present there is no adequate field criterion to establish that bedrock weathering...has been dominated by freeze-thaw weathering" (Thorn, 1992, p. 11). Thus, there are two hypotheses (nivation and cryoplanation) both having as a central tenet an unsubstantiated judgement (frost weathering) with regard to the dominant process.

There also are climatic differences between nivation and cryoplanation. Nowhere in the nivation literature is permafrost identified as a factor, but some cryoplanation authors (e.g., Reger and Péwé, 1976) consider the cryoplanation terrace to be indicative of permafrost. Last, but far from least, is the issue that whereas nivation seems to be a widely used term, cryoplanation has a number of synonyms and more than one may be used, for the same feature, within a single paper. Thus the reader may be left uncertain as to *exactly* what is the feature under discussion and how the same feature can have more than one adjective to describe it. Ballantyne and Harris (1994) discussed this problem to some degree and suggested (p. 245) that, morphologically, the only difference between "...features labelled "nivation benches" and supposed "cryoplanation terraces" would appear to be one of size or maturity of development."

These issues of terminology, processes, required conditions, paleoenvironmental implications, and interaction between (and possible redundancy of) the collective

TABLE I

Basic Conditions Generally Cited as Associated with Nivation and Cryoplanation

Process conditions	Nivation	Cryoplanation
Freeze-thaw weathering	✔	✔
Salt weathering	✔	✔
Chemical weathering	✔	✔
Solifluction	✔	✔
Subsurface wash	✔	✔
Surface wash	✔	✔
Fluvial transport	✔	(✔)
Aeolian transport	✔	✔
Over-snow transport	✔	(✔)
Slumping	✔	(✔)
Solution	✔	✔
Transport by snow	✔	✘
Permafrost	✘	✔
Pre-existing hollow	✔	✘

✔ Process or characteristic usually cited as occurring in association with nivation and/or cryoplanation.
(✔) Process or characteristic that might be seen as occurring in association with nivation and/or cryoplanation but is not generally recognized as a major factor.
✘ Process or characteristic not usually cited as occurring in association with nivation and/or cryoplanation.

nouns as outlined above will now be discussed at length. Although overlap is inevitable, an attempt will be made to look at each of these issues separately. The foundation for the discussion is presented by means of a brief synopsis of the two concepts—nivation and cryoplanation.

NIVATION AND CRYOPLANATION: THE BACKGROUND AND THE BASICS

Nivation

"Nivation" is the term used for "...the combined action of frost shattering, gelifluction, and slopewash processes thought to operate in the vicinity of snowbanks" (French, 1996, p.159). It is generally perceived as an erosive process, driven by intense frost shattering facilitated by the colder temperatures associated with the presence of snow and abetted by the moisture from that snow. The debris transport processes are largely driven by slope wash—accelerated by solifluction—with the removal of the weathered debris leading to the development of a "hollow" (Fig 1). Thus, the term "nivation" is used both as a collective noun to identify the group of processes that operate in association with the snowpatch (and may be intensified by that snowpatch) and as an adjective to describe the resultant landform—the "nivation hollow" (Thorn, 1988). However, as Thorn (1988) has pointed out, nivation results

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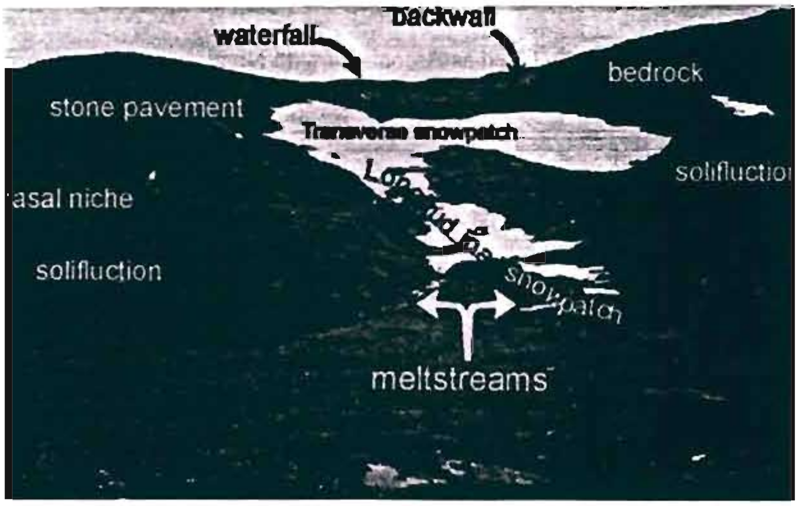

Fig 1. Transverse and longitudinal snowpatches late in the melt season (northern Norway).

from the existence of a hollow that it may modify but that it certainly did not initiate. If nivation is a process of modification and not initiation, then it is clear that a topographic catchment must already exist in which the snow can accumulate—a catchment created by some other process(es), *not* by nivation. Associated with the identified prerequisite of an existent location in which snow can accumulate is the absence of any reference to the need for, or of association with, permafrost. In fact, the very accumulation of snow, and hence the thermal insulation, would tend, in many instances, to contraindicate the development of permafrost. The presence of the snow is obviously central to any perception of nivation. Lewis (1939) classified snowpatches by form—longitudinal (elongated downslope), transverse (elongated across the slope), or circular (transitional between the other two forms) (Figs. 1 and 2). Snowpatch form must, at least in the initial stages, be a function of the shape of the existent hollow in which the snow first accumulates. The most comprehensive discussion regarding nivation can be found in Thorn (1988).

Cryoplanation

Cryoplanation is said to be a "... cryogenic process promoting the low-angled slopes and level bedrock surfaces typical of many periglacial regions" (French, 1996, p. 181). The processes seen as causative for cryoplanation terraces—namely intense frost shattering, solifluction, and, in some references, slope wash (e.g., see Peltier, 1950; Dylik, 1972; French, 1996 for discussion)—are similar in composition to those associated with nivation, although their relative contributions and rates of operation may differ. The major distinction between nivation and cryoplanation is that less emphasis is placed on the role of snow, even though a snowbank can occur at the break of slope between the bench tread and backwall (Figs. 3 and 4). The end result of "cryoplanation processes" is the development of the low-angled bedrock surface. A number of authors (e.g., Reger and Péwé, 1976) consider cryoplanation terraces to be

Fig. 2. Aerial view of a circular and transverse snowpatch (northern Norway).

associated with permafrost—although others, such as Demek (1969a) do not—and for them to exhibit orientational preferences. Furthermore, as an issue that will be discussed in detail later, a number of authors (e.g., Boch and Krasnov, 1943; Demek, 1969a; Reger, 1975; Czudek, 1995) identify nivation as initiating cryoplanation and that, at some later stage, "cryoplanation" takes over from "nivation." Priesnitz (1988) provides a recent synthesis regarding cryoplanation.

TERMINOLOGY

With respect to cryoplanation terraces, consideration of the literature indicates a plethora of terms that describe what may be similar forms, various stages of the same form and/or the same process associations—pediments (e.g., French and Harry, 1992), periglacial glacis (e.g., Beck, 1989), altiplanation terraces (e.g., Boch and Krasnov, 1951), goletz (e.g., Boch and Krasnov, 1951), or golets terraces (e.g., Richter et al., 1963). Reger (1975, p. 11) also provided a number of synonyms based on language groups—"nagornaya terrasa, goltsovaya terrasa, solifiyuktsionnaya (nagornaya) terrasa (Russian); altiplanation terrace, high terraces, goletz, nivation bench, rock-cut bench, goletz terrace (English); haute terrace, terrasse goletz, terrasse d'altiplanation, terrasse de nivation, replat de nivation (French); Nivationsleisten, Golecterrassen, Goletzterrassen, Kryoplanationsterrassen (German); tarasy altiplanacyjne, terasy goliznowe (Polish); an equivalent term in Czech is "kryoplanacni terasa." This multiplicity of terminology for ostensibly the same feature (i.e., the terrace) also is evidenced by the interchangeability within the same discussion. For example, goletz and altiplanation, altiplanation and cryoplanation, pediment and altiplanation, structural bench and altiplanation, altiplanation and nivation bench, nivation and altiplanation, nivation and cryoplanation, golets and altiplanation, and altiplanation and mountain terraces are each examples of two terms being used to describe the *same* feature within the *same* paper.

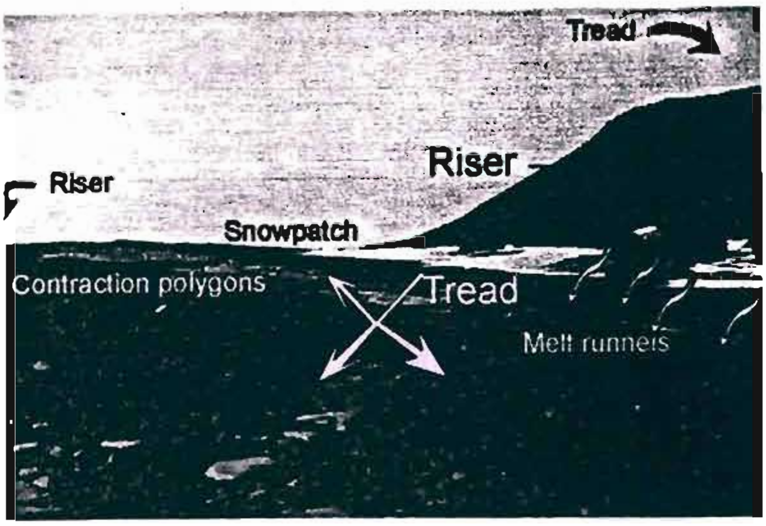

Fig. 3. Cryoplanation terrace in an arid region of the Antarctic. The amount of snow visible in the photograph is close to the annual amount that occurs on the terrace.

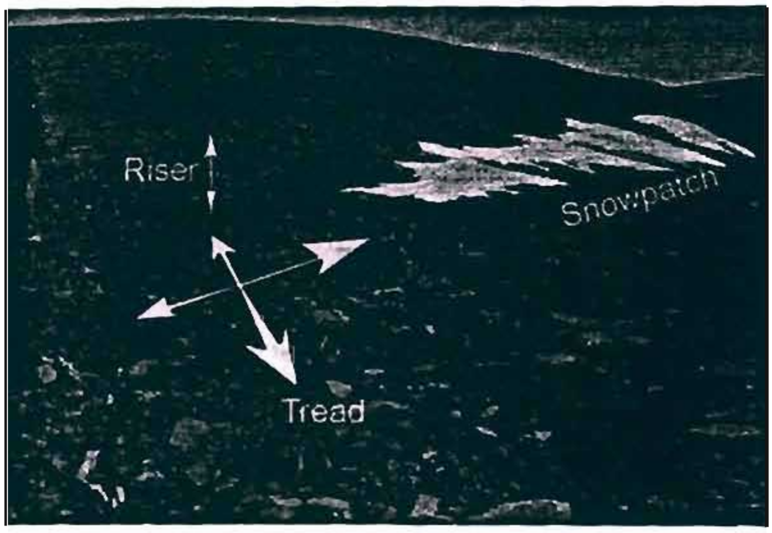

Fig. 4. Cryoplanation terrace in an arid region of the Antarctic. Snowpatch size is close to the maximum areal extent at the end of winter.

From the foregoing, it also can be seen that the terms "cryoplanation" and "nivation" seem to be synonymous for many researchers. Many authors integrate nivation and cryoplanation: "Cryoplanation terraces may start to develop due to nivation..." (Czudek, 1995, p. 102) or, in a discussion on cryoplanation, "The retreat of ascending scarps...is accomplished by nivation..." (Reger, 1975, p. iv) and "Cryoplanation terraces form by scarp retreat as the result of nivation" (Reger and Péwé, 1976). This issue will be addressed in detail in the discussion on processes, but the conceptual foundation for that is created here. The above (typical) quotes beg the question of the relationship between "nivation" and "cryoplanation." If nivation is seen to precede cryoplanation, then what is the threshold where the former (nivation) changes to the latter (cryoplanation)? A further question, to be dealt with later under paleoenvironmental considerations, is, if the two terms are justified and yet they interrelate with respect to operation, how is a fossil nivation bench distinguished from a cryoplanation terrace? This will be shown to be important, as there are perceived climatic associations. This same interrelationship of two distinct terms also raises the major origin-process question regarding the transition between the two, when authors such as Czudek (1995, p. 102) state that "cryoplanation terraces may start to develop due to nivation as initial nivation hollows even on the original smooth slope without any break of slope." This implies that nivation can initiate the terrace. Matthes (1900, p. 182) in his original definition of nivation states that "...snowdrifts do not form except in the presence of favorable topographic hollows" and that "...the effect of their presence is to accentuate these features." Thus, the concept of nivation involves the exploitation of pre-existing hollows, not the creation of them.[2]

The question of terminology becomes even more complex when the collective noun (i.e., nivation and/or cryoplanation) is mixed with process descriptors that are actually encompassed within that noun. For example, Reger (1975, p. iv) noted that "cryoplanation terraces form by a complex of mechanisms acting in unison, including nivation, mass movement, frost action, piping and wind." The problem is that a statement such as this distinguishes "frost action" from "nivation" and yet frost action is central to nivation. Inherent within any discussion of nivation must be the role of frost action (see Thorn, 1988, p. 11-14) and so, as mundane as it may first appear, the question must arise as to how nivation *and* frost action can be cited as separate issues unless frost action is deemed to occur when nivation no longer operates. However, it is this very transformation from frost action within nivation to frost action outside of the context of nivation (i.e., this may constitute the transformation to cryoplanation) that is not explicit in the literature. Rather, authors such as Czudek (1995) continue to discuss the role of frost action in association with the presence of snow (i.e. nivation) as the reason for the development of the cryoplanation terrace. It could be suggested that such casual use of terms is symptomatic of the whole problem.

PROCESSES

Central to the entire discussion on processes must be the role and place of frost action (freeze-thaw weathering) combined with the nature and rate of transport processes acting on the terrace. Descriptions of both nivation and cryoplanation cite frost weathering as a key process. Consideration of the literature (Thorn and Hall, 1980, Table 1) indicates that all authors writing on nivation identify freeze-thaw weathering as enhanced at a nivation site. In many instances, freeze-thaw weathering is—as a result of thermal and hydric conditions associated with the snow—perceived as epitomizing nivation (e.g., Hanvey and Lewis, 1991). In the case of cryoplanation, there is no question that freeze-thaw weathering is a key factor. Grosso and Corté (1991, p. 54) identified two main processes as operating on cryoplanation terraces, the first of which is "frost wedging on the upper erosive zone" and the second "gelifluction on

[2]See Thorn (1988, p. 19) for a more detailed discussion.

the lower accumulation zone." This conclusion is found throughout the cryoplanation literature: Czudek (1995, p. 100) echoed the words above almost exactly when he identified two main processes as causative of cryoplanation terraces: "First they include processes of frost weathering, secondly processes removing the waste material." Some, such as Richter et al. (1963) went as far as to identify cryoplanation terrace formation as an indicator of the "frost-shattering zone." Although all authors on this topic cite frost weathering, Priesnitz (1988, p. 60) summed up the problem when he stated:

> ...the real problem is not whether there happens to be some frost-wedging on the cliff or what importance some authority attaches to this. Rather, the initial challenge is to answer fundamental questions about whether frost-wedging really works faster on the cliff than on surrounding surfaces, whether transport is sufficiently fast to prevent the cliff being buried by its own debris, whether the backwearing rate is high enough to explain a given terrace of known age....

For this discussion, the following questions should be added. First, is it even frost-wedging that is the weathering process? Second, is transport able to remove the debris moved to the cliff by transport across the terrace *above* the cliff? Lastly, do we actually have any firm idea of dates of cryoplanation forms?

With respect to the character of the weathering, how can freeze-thaw weathering be identified as the major cause of rock breakdown when no data are presented to support the contention? In this regard, the arguments of Thorn (1992, p. 11) summarize all of the many problems associated with the presumption of freeze-thaw weathering in cold regions: "It is my belief that what periglacial geomorphologists need more than any other single item is a way to determine in the field whether or not bedrock fragments have been frost weathered." This is not to say that freeze-thaw does not play a role in breaking down the rock, but rather it needs to be proven rather than assumed. With cryoplanation being perceived as operative in arid or semiarid regions, the question arises as to the moisture source required for freeze-thaw weathering. A number of studies (e.g., van Autenboer, 1964; Hall, 1997a, 1977b) question the role of freeze-thaw and suggest that processes such as thermal stress fatigue may be more effective in the arid environments (Hall, 1998). In turn, our lack of understanding regarding active processes leads to questions regarding rate(s) of landform development. Consequently, if a central tenet of the landform genesis is in doubt, then perhaps we do lack a working operational definition (as noted above by Thorn, 1988). It is not questioned that some form of weathering is taking place, but the exact nature and rate of the weathering are unknown.

It is interesting to note that Lauriol et al. (1997) identified extensive chemical weathering on the tread of a cryoplanation terrace that they studied in the Yukon Territory of Canada. A two-event weathering regime is suggested. The first involves chlorite breakdown and leaching of the iron in the moister, acidic conditions at the base of the scarp, associated with a "persisting snowpatch." The weathering results in increased rock porosity. With terrace growth, they suggested that the weathered blocks are left behind on the tread, away from the moisture associated with the snowpatch, so that they then experience arid conditions where the second weathering event occurs. The second event is said to involve iron being introduced to the rocks by percolating water so that a weathering rind is formed; the source of the iron is thought to be windblown dust.

In terms of transport processes, nivation has fewer "problems" than cryoplanation. Although the specifics of what mechanisms may dominate in space or time may vary from nivation site to nivation site, in all instances they are clearly involved with the removal of debris from the nivation hollow. As nivation hollows are discrete and do not occur in sequences, so the debris from one does not usually impact upon another. This differs radically from cryoplanation where, in many instances, terraces are identified as occurring one above another (Priesnitz, 1988). Thus, the transport processes operating to move material downslope on one terrace must bring some, if not all, of that material to the cliff of the terrace below. Therefore, the transport processes operating *on* the cliff must be sufficient not only to remove the products of weathering on that cliff but also the material brought to the cliff from above and which, if not removed, would ultimately bury that cliff. This issue of debris from one terrace impacting on the one below does not appear to be addressed in the cryoplanation literature and yet a debris cascade must occur unless weathered material is removed from each terrace such that it does not move on to the one below. If it does move on to the terrace below then there should be, all other factors being equal, a net accumulation on, and burial of, the lowest terrace. Reger (1975, p.182) argued that the products of weathering are removed down "sideslopes," but such a scenario clearly will not work where terraces are in sequences or where treads do not trend laterally. Lauriol et al. (1997, p. 147), in a discussion of chemical weathering on cryoplanation terraces in the Yukon Territory, noted that the terraces are not "...greatly perturbed by gravity processes" and that there is not an accumulation of debris on the tread, as there is a (p. 148) "constant removal of fine sediments...by wind and water action on the tread." In the terraces these authors studied, the largest accumulation of fine sediments was at the foot of the scarp and diminished toward the riser of the terrace below. Thus, it may be that weathering products are removed completely by wind such that no debris is transported down to impact on the terrace below. There is clearly a need for detailed studies of sediment transport routes on terraces.

REQUIRED CONDITIONS

French and Harry (1992, p. 152) noted that "many features previously identified as cryoplanation terraces are lithologic and/or structural benches" and they cited the opinion of Büdel (1982, p. 78) to the effect that in today's periglacial environment there is not "the slightest evidence of any etchplain creation such as cryoplanation or altiplanation." However, this concern for structural control seems to miss the point, as both Lauriol (1990) and Czudek (1995) emphasized that geological control is a *major* factor in controlling the development of cryoplanation terraces. Clearly, it is not *obligatory* for "cryoplanation" to cut across structures or lithologic boundaries, just as this has never been an issue in nivation. In fact, the definition in the *Multilanguage Glossary of Permafrost and Related Ground-Ice Terms* (van Everdingen, 1994, p. 11) states that "cryoplanation terraces...often lack structural control"; therefore, clearly they would still be cryoplanation terraces if there was structural control. Again, Czudek (1995, p. 101) has stated that "although cryoplanation terraces cannot be considered only purely structural benches, in a lot of cases (but not in all cases) these features are explicitly–structurally controlled"! If such features are developed on horizontally bedded rocks then it would be difficult if not impossible for them to cut across any structural differences and the (horizontal) lithologic boundary would certainly be a site for potential exploitation by weathering. The question therefore is, even if the "terraces" can be seen to exploit a structural or lithologic control (Fig. 4), are the processes any different from those operative on a cryoplanation terrace not structurally or lithologically controlled? Furthermore, as this control may well be a prerequisite for nivation, and nivation is so intimately connected with cryoplanation, then how *can* structural control preclude a feature being considered from having either a nival or "cryo" origin? Grosso and Corté (1991) have argued that rock type does not affect cryoplanation terrace development.

Thorn (1992, p. 18) stated that cryoplanation, like nivation (Thorn, 1988), defies a satisfactory operational definition because it embraces many component parts and many of those parts themselves are poorly defined or understood (e.g., freeze-thaw weathering; see below). The key is that, whereas a "terrace" is solely a form, a "cryoplanation terrace" or a "transverse nivation bench" implies a form with a known origin. Therefore, the use of any such adjective to describe the bench/terrace is inappropriate while we are unable to define a working definition regarding "cryoplanation" or "nivation" (French and Harry, 1992; French, 1996; Thorn, 1992) and we are unable to distinguish between these two forms. Thorn (1988) actually argued for discarding the term nivation altogether. This issue may be further confused, or resolved, by the suggestion that nivation and cryoplanation are two end members of the same process suite and, thus, there is no difference between the processes associated with the two terms or the nature of the resulting two forms.

NEW MODEL

Based on the above, it is proposed that "nivation" and "cryoplanation" (and any of their various synonyms) are not separate entities but rather two members of the same process/landform continuum (see Fig. 5). Moisture is the variable defining this continuum—with "nivation" at the snow-derived, wetter end and "cryoplanation" at the permafrost-based, dry end; both transitional and composite forms also would be possible. The presence or absence of snow, and hence a moisture supply, has an effect on the timing and nature of the processes, but all still fall within the same "process suite." Following the arguments outlined earlier, "nivation" can only exploit a pre-existing hollow, but there is no reason why "cryoplanation" cannot initiate one. This argument is, however, contrary to that advanced by several authors (see above), who viewed nivation as initiating and cryoplanation as exploiting. Within the framework of the argument here, there is no reason why, in a snowy environment, nivation could not exploit pre-existing hollows. Equally, there is no reason why some of the hollows might not have, at some earlier stage, been initiated by cryoplanation. In other words, under a drier climate a bench may develop by means of "cryoplanation" and later be accentuated by "nivation" as the climate became snowier or, very simply, as the hollow facilitated snow accumulation. The end landforms are all the same and simply reflect changes in the water-driven process component and rate of development. Morphologically the "transverse nivation bench" and the "cryoplanation terrace" are identical landforms and may even have a composite history. Where it is clear that only one end of the process spectrum has been operative—(e.g., "cryoplanation" features currently developing in Antarctica (Hall, 1997a)—this landform (Fig. 3) is still part of the same process/landform suite as the "nivation bench" (Fig. 1) currently forming in (for example) the snowy maritime environment of Norway (e.g., Hall, 1985). Only the relative process rates and associations, driven by moisture availability, have changed.

A simple model to show this nivation/cryoplanation relationship is presented in Figure 5. The process groups identified are common to "both" landform end points; only the importance of individual processes within the group varies. The arrows provide a generalized indication of the importance of the respective process to those two "end landforms," but would be considered to have variability dependent upon specific sites and geographic locations. The relative importance of the processes will not only change with location but also will change over time. By viewing the resultant landform as a single entity derived from some variable, and varying, association of the same process suite, the entire terminological/origin problem becomes, at one level, much simpler. It is suggested here that the feature could be termed a "periglacial mountain bench"—it being a *bench* found in *mountain* regions that is of (unspecified with respect to particular process) *periglacial* origin.

Recognizing that there is no known way of identifying the nature of the process that produced weathered debris found in the field, the preoccupation with freeze-thaw recedes and is replaced by "weathering under periglacial conditions." This allows for any and all combinations of cold-based weathering processes. The significance is that the landform *cannot* be used as a proxy to deduce process. It is suggested that this model also negates the confusion regarding "initiation" and "modification" that prevails in the literature and in actuality is a process-based variation of the graphical landform model suggested by Demek (1969a). The main difference is that the confusion associated with those terms and the manner in which there is a transition from one landform term to the other is removed, and the periglacial mountain bench is viewed as the result of a suite of processes that very likely varied through time.

It also is argued that the problem of structural control falls away, as it too becomes irrelevant. It really has no impact if there is a structural/lithologic control, as the processes remain the same, as does the final landform—the bench. The cases where the feature cuts across structure, which actually is all the more difficult to conceive of as having a cryoplanation origin (i.e., how did it initiate and cut back *across* structures?), become a *non sequitur*. The landform may have had a non-periglacial origin that subsequently was exploited by periglacial processes (as in the case in which "nivation" exploits a pre-existing hollow of *any* origin) to produce a periglacial feature—the bench. We have no way of deducing an "origin," only an explanation, for the end result. Thus, any and all landforms described by the plethora of terms outlined at the beginning of this paper would all be considered a "periglacial mountain bench."

Finally, as a result of water being the major factor in determining processes, this also affects the dynamism. At the wetter end of the continuum, process activity is both more rapid and intense, whereas most but not all processes decrease in activity toward the more arid end of the spectrum. The significance here is with respect to the time required for landform development. As dynamism increases so the time for land-

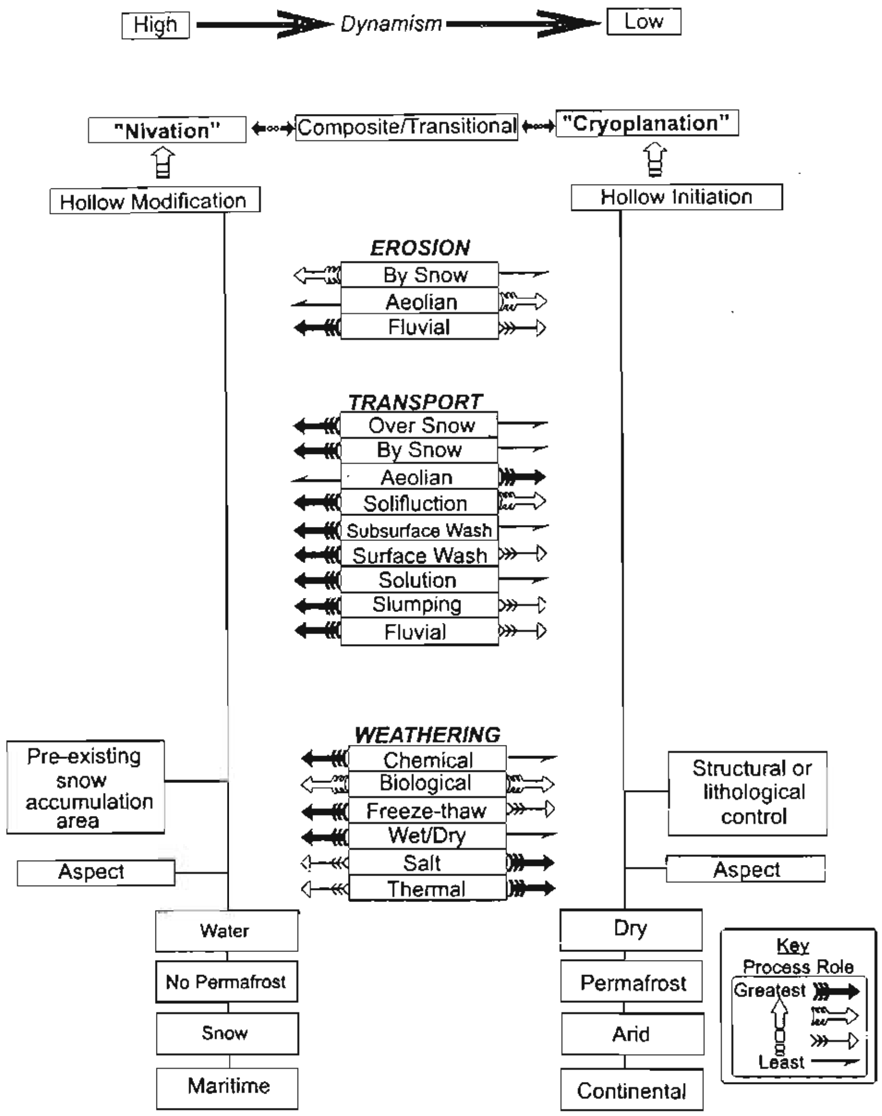

Fig. 5. Simple model indicating the processes associated with both nivation and cryoplanation and how it is only their relative contributions that changes as a result of moisture variability.

form development decreases. This means that landforms associated with more arid environments may be very old, whereas those in snowier environments may be relatively young. In between are composite forms that may have had varying dynamism in terms of development. Unless some idea of paleoconditions is known, it may be very difficult to attribute any idea of age to these benches. This effect of dynamism also may mean that size is not an indicator of age unless it can be shown that only one condition prevailed throughout the perceived "life" of the feature. Larger features need not be "older" if they experienced periods of more dynamic development when rates of weathering, erosion, and transport would have been greater. Such a scenario would mean that the landforms have varying rates of growth depending upon the prevailing conditions; therefore, size may not be a significant indicator of age.

CONCLUSIONS

Following the initial thoughts of Thorn (1988), it is suggested that the two terms "nivation" and "cryoplanation" are part of the same process-landform suite and that the dichotomy has been misleading and confusing. By identifying the resultant landform as of a singular origin, within which the processes may have varied in their individual contributions, the present terminological, environmental, and reconstructive problems fall away. A simple model is presented to identify how and why those processes may vary. As is so often the case, the need is for more empirical data to test this model. Until then, it is argued that we are better served by moving away from the confusion of process-related terms to the singular name suggested here. In so doing, we may assume less and investigate more.

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SOME OBSERVATIONS AND THOUGHTS REGARDING ANTARCTIC CRYOGENIC WEATHERING

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Studies that either directly or indirectly deal with weathering in cold regions generally assume mechanical processes dominate, that the prime process is freeze-thaw and that the resultant products are angular. Recent observations regarding weathering in the Antarctic question these assumptions and the application of rock fracture mechanics indicates that curvilinear, rather than sharply angular, fracture patterns need not be unusual. Rock temperature data collected at one-minute intervals indicate that thermal stress/shock can play a major role in rock breakdown and the steep temperature gradients in the outer few centimetres of the rock would be conducive to spalling. The aridity of the present study area (Alexander Island, Antarctica) argues against freeze-thaw weathering except in very site-specific locations easily identified by visible water. Detailed rock temperature data, including for a significant part of two winters, clearly show an absence of snow (and hence moisture), the occurrence of $\Delta T/t$ values that exceed the threshold for thermal shock, a marked, but varying, aspect influence and steep thermal gradients with significant subsurface fluctuations (Hall, 1997 a & b, 1998).

Classic Griffith fracture theory expresses that a certain combination of excess pressure and crack length (or diameter) is required to keep a crack open or to increase its dimensions. Linear elastic fracture mechanics (LEFM) show very clearly that cracks may curve during propagation in response to a changing stress field; microfissures in the rock can also greatly influence crack direction. In fact, curvilinear (mixed-mode) crack propagation is common in rock mechanics as a crack will propagate in the direction the tensile stress in the crack tip vicinity is maximum. However, LEFM approaches assume that stress-intensity is decreasing with increasing crack length and that cracks may influence each other's stability and trajectory. In essence, there is no reason why curvilinear cracks may not occur as a result of stresses induced by mechanical and/or chemical weathering (particularly as in stress corrosion crack tip propagation). Non-cubic rock forming minerals show thermal expansion anisotropy. When the total linear thermal expansion of a mass becomes equal to the critical crack opening displacement, $\Delta_t = \Delta_c$ (where Δ_t is linear thermal expansion and Δ_c is critical crack opening displacement), the centre of the mass originally subjected to a compressive stress will now be subjected to uniform tension. When those initial stresses are very high so the shape will be selected to minimize stress concentrations. As cyclic variation in temperature induces alternating tensile and compressive stresses, particularly along the boundaries of inhomogeneities, there is no reason why the resultant fractures should not be curvilinear. That being so, why should the assumption be that mechanical weathering will only produce angular forms?

Evidence from the weathering of sandstones on Alexander Island (Antarctica) clearly demonstrates the production of rounded forms and debris as a result of mechanical processes, mainly thought to be thermal stress fatigue. This concept of thermal stress fatigue being the cause of breakdown fits well with the available temperature data plus LEFM theory (e.g., Rossmanith, 1983) and results from artificial weathering studies that investigated physically-induced stress-fatigue microcracking (Blaga and Yamasaki, 1973). Further, despite the frequent association of mechanical weathering processes with taffoni development, it seems never to have been questioned that taffoni are rounded forms. Data collected from the same area on taffoni size and Schmidt hammer rebound values for different aspects show that there is an aspect influence on weathering. It is concluded that, in this area of Antarctica, mechanical processes other than freeze-thaw dominate, that forms other than angular can be produced, and that taffoni development shows a distinct aspect influence.

Key Words: Weathering Thermal conditions Rock fracture mechanics Taffoni Antarctica

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Some Observations and Thoughts Regarding Antarctic Cryogenic Weathering

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PRESUMPTIONS

Weathering in cold regions is considered to be dominated by mechanical processes.
Freeze-thaw weathering is generally assumed to be the predominant weathering process in most studies.
The result of this freeze-thaw weathering is said, in both present and Quaternary studies, to be angular debris and forms.

QUESTIONS

IS IT REALLY FREEZE-THAW WEATHERING THAT PREDOMINATES?
DOES MECHANICAL WEATHERING, OF ANY TYPE, HAVE TO PRODUCE ANGULAR MATERIAL?

STUDY AREA:

Alexander Island (Antarctica)

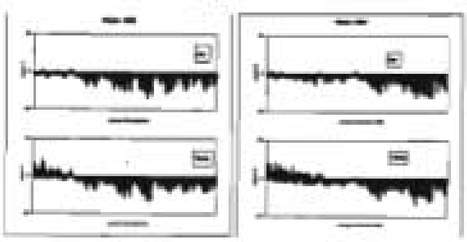
Mars Glacier : 71° 50'S, 68° 21'W

Area of 'Antarctic oases' that experience a cold, dry continental climate.
Available data indicate a summer mean air temperature of -2.4°C and a winter mean of -11.5°C.
Winter minimum temperature recorded is -35.2°C and a summer high of +6.0°C.

Very little snow fall

An area of continuous permafrost with an active layer thickness of 0.3 to 0.4 m.

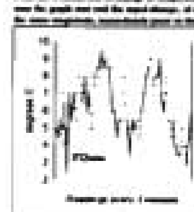
Local bedrock comprises horizontally bedded sandstones, conglomerates and argillaceous sedimentary rocks.



Rock temperature data were collected during the summer at one-minute intervals.
Winter data were collected at 20-minute intervals.

Detailed analysis of the winter data shows rapid fluctuations of temperatures that indicate an absence of snow to insulate the ground.
Field observations at the start of spring also showed a general lack of snow in the area except for a thin accumulation in some declivities.

The absence of snow and the lack of summer snow-fall means that water was extremely limited in this area.

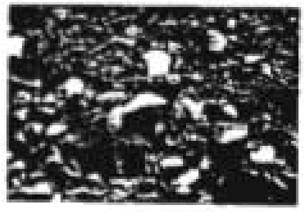


Light- and dark- coloured sandstones on a horizontal surface close to their lithologic junction.

Note that the dark-coloured sandstones are angular whilst the light-coloured sandstones are rounded.

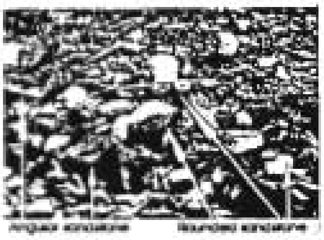
Both are in a dry, cold environment.

Wind effects are discounted as the two lithologies and their respective shapes are both in juxtaposition and are found throughout the area - including on vertical sections.

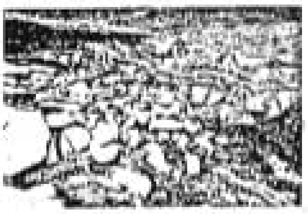


Mechanical Weathering of Bedrock and Clasts

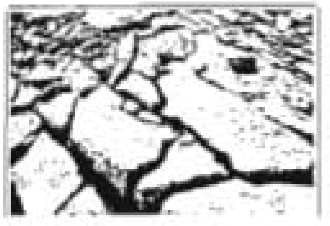
Exposures of two sandstones (one dark-coloured, the other light-coloured) on a mountain bench. The dark-coloured weathers to angular forms whilst the light-coloured weathers to rounded forms.



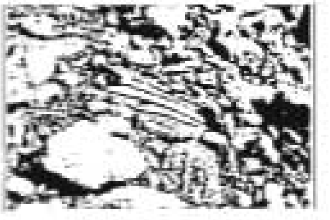
Detail of the weathering forms of the two sandstones close to the lithologic junction.



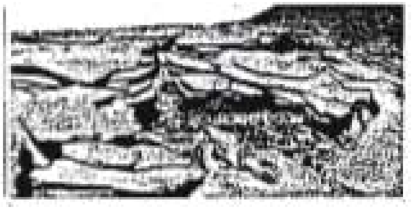
Showing rounding of weathered material on surface and downslope of a terrace in the light-coloured sandstone.



Detail of bedrock weathering in area of light-coloured sandstone to show the rounded forms produced in situ.



Weathered clasts on a debris slope downhill of a terrace showing the rounded forms produced (note split, rounded clast near photo center).



Detail of light-coloured sandstone to show the general rounding of the weathered material. Clearly some breakdown results in angular material, as shown on photo left, but, with time, this must become rounded. This outcrop of light-coloured sandstone is on the edge of a mountain terrace approximately 650 m a.s.l.

Mechanical Weathering of Bedrock to Produce Taffoni

General view of taffoni on a rock outcrop (person on left for scale). Note the distinct rounding of all forms. Exposure of light-coloured sandstone.

Close-up of taffoni, showing the rounded form, together with a detail of the flaking found within the taffoni. In this example it is noteworthy that the taffoni on the left is debris-free whilst that on the right has extensive extraneous debris (these taffoni are some 20 m above the valley bottom).

Detail of taffoni showing the distinctly rounded character. Note the flaking taking place within the taffoni.

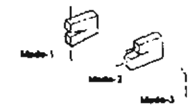
Theory

Fracture mechanics deals with the discrete propagation of an individual crack or cracks and utilizes principles of linear elastic fracture mechanics (LEFM

A **crack** is a line across and/or along which the displacement field exhibits a discontinuity and the crack faces, on opposing sides of this line, may or may not be stress free.

The **stress** intensity factor, **K**, quantifies the intensity of the stress singularity at a crack tip and is directly related to strain energy release, **G**.

There are three modes of crack deformation
mode-1: opening mode (crack surface displacements are perpendicular to the plane of the crack).
mode-2: sliding mode (crack displacements occur in the plane of the crack and perpendicular to the leading edge of the crack).
mode-3: tearing mode (the crack surface displacements are in the plane of the crack but parallel to the leading edge).



The crack will advance when its stress intensity reaches a critical value, K_c; K_{1c} is the critical value of K for mode-1 (opening mode).

For a penny-shaped crack of radius a subjected to uniaxial external compression σ and internal pressure **p**, so the stress intensity factor for the open crack is given by:

$$K_1 = \frac{2}{\pi}(p-\sigma)\sqrt{\pi a}$$

Griffith fracture criterion states that a certain combination of excess pressure ($p_e = p - \sigma$) and crack length is required to increase crack length

Cracks can grow by fatigue and this depends on the cyclic range of the stress intensity factor (K). For a cyclic pressure fluctuation, **p**, so:

$$\Delta K = \Delta(p-\sigma)\sqrt{\pi a}\,Y \quad = \Delta p\sqrt{\pi a}\,Y$$

where Y = 1 for a tunnel crack and Y = $2/\pi$ for a circular crack.

(Above based on *Rock Fracture Mechanics*, 1983 by Rossmanith, H

Thermal stresses are induced by temperature alteration and may result in fatigue or in instantaneous failure.

On **cooling**, the maximum stress is the tensile stress on the surface and the center is in compression. On **heating**, the maximum stress is the compressive stress on the surface, and the center is in tension. There are also shear stresses equal to half the difference between the principal stresses. (Kingery, W.D. 1955. Factors affecting thermal stress resistance of ceramic materials. *Journal of American Ceramic Society*, 38, 3-15)

Instantaneous failure, **thermal shock**, requires $\Delta T / t = \ \geq 2°C/\min$

Thermal-shock crack patterns evade being trapped at some depth where **G** is low by changing to a **spalling** mode of propagation.
This produces **curved cracks** where the crack density is dependent upon ΔT

(See Bahr, et al., 1986. Thermal-shock crack patterns explained by single and multiple crack propagation. *Journal of Materials Science*, 21, 2716-272

A spall will release the maximum strain energy and, for an infinite slab, the strain energy in a unit volume is given by:

$$U = \frac{\sigma^2(1-\mu)}{E}$$

and the depth of the spall should be proportional to:

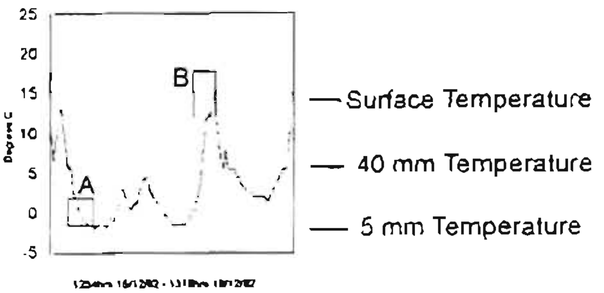
$$E \big/ (1-\mu)\sigma^2$$

(where U is the strain energy in a unit volume, σ is normal stress, μ is Poisson's ratio, and E is Young's modulus of elasticity)

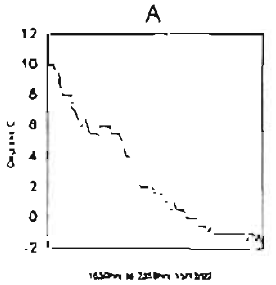
Since the residual temperature gradient in the spalled piece will be nearly linear, essentially all its strain energy will be removed upon spalling.

Lithology obviously plays a very important role in determining the fracture/spalling shape (see Ballantyne and Harris, 1994, p. 171 who discuss rounded clasts resulting from mechanical weathering) and that is likely an important factor in this present study in explaining the difference between the dark-coloured and the light-coloured sandstone weathering forms.

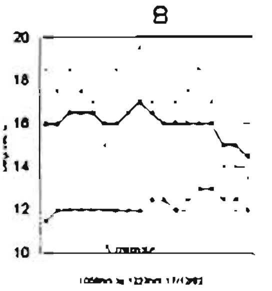
Examples of temperatures at different depths conducive to spalling stresses:



To show temperatures at rock surface and at 40 mm depth over three days. Details of A and B are shown on the right.



Detail A to show that, during the cooling period the rock is warmer at depth than at the surface.



Detail B to show the surface cooler than 5 mm depth at times during the warming phase together with the extent of the temperature gradient.

Comments

The result of mechanical weathering need *not* be angular clasts or angular forms.

Processes *other than* freeze-thaw can cause rock breakdown in cold regions.

The finding of angular clasts in present or former cold regions does *not* indicate the presence of freeze-thaw weathering.

Thermal stress fatigue has been underestimated as a destructive force in cold regions.

Breakdown due to thermal stress is *more* likely than freeze-thaw weathering in cold, arid regions.

The finding of rounded clasts in former cold regions may have been subject to misinterpretation.

The assumption that angular clasts equate to freeze-thaw weathering may have led to misinterpretation of former environments.

Taffoni produce rounded forms and are frequently assumed to be the result of mechanical weathering.

This apparent contradiction seems not to have been questioned.

Data from inside taffoni show rates of change of temperature commensurate with thermal shock and/or thermal stress fatigue.

The observation of spalling inside of taffoni where such temperatures have been recorded (in a dry, cold region) supports the contention that thermal processes may play a greater role than previously thought.

There is clearly a need for more detailed temperature data (at one-minute intervals) at various depths in rock to facilitate mathematical analysis, via LEFM, to discern the potential for thermal processes.

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NIVATION OR CRYOPLANATION: IS THERE ANY DIFFERENCE?

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'Nivation' and 'cryoplanation' are terms indicative of two distinct process-landform associations. In some discussions there is a time-process continuum between the two defining terms but the threshold of transition and the resulting process-landform difference is not clear. For example, many discussions on 'cryoplanation' resort to the role of 'nivation' in the early stages but the actual threshold of transition and the character of process and landform change at this transition are not explained. Some authors suggest the difference appears to be no more than size or maturity. The absence of objective, quantified thresholds with respect to size or maturity prohibit these attributes from distinguishing between features and, at a fundamental level, how can size or maturity be used to justify two distinct terms even if such thresholds could be established? Further, where palaeo-reconstructions are undertaken, is it possible to distinguish between the fossil forms of a 'cryoplanation terrace' and a 'transverse nivation hollow'? Despite this duality of terms, both groups utilize the same basic processes and the resultant landforms may be very similar, if not identical.

Comparison of the literature on nivation and cryoplanation serves only to confuse rather than enlighten. Nivation is clearly described in its literature as exploiting *pre-existing* hollows to produce the nivation hollow or bench. Cryoplanation in its literature is seen as an initiator, the end result of which is the cryoplanation bench. Thus it would *not* be possible for nivation to initiate cryoplanation on an otherwise undifferentiated slope as is suggested in the cryoplanation literature. Further, nivation requires snow, usually in some quantity, which must suffer melt, whilst cryoplanation is said to characterize arid or semi-arid regions. A number of authors associate permafrost with cryoplanation whilst this has never been the case with nivation and so the transition between the two process-landform suites may be problematic. Lastly, as both concepts utilize the same basic suite of processes so the question arises as to what is, then, sufficiently different to justify the respective terms?

With regard to processes, both concepts (nivation and cryoplanation) identify frost weathering as the dominant, if not sole, cause of rock breakdown; some authors even identify the resultant landforms as indicative of a "frost-shattering zone". Conceptually, the questions arise as to why fost weathering, is there any proof for this and, what is the implication if it is *not* frost weathering? To compound the problem, there is no way to determine whether bedrock fragments are the product of frost weathering and recent studies in cold, arid environments suggest that thermal stress may be the dominant weathering process. Failing unequivocal identification of frost weathering, then both concepts lack a working operational definition.

The proposition is made that the two terms and their respective concepts are obsolete. Rather, 'nivation' and 'cryoplanation' are the two end members of the same process-landform continuum. With the transition from nival to arid conditions, so there is a decrease in the action of water and, likely, an increase in the *potential* for permafrost (due to the diminished insulation from snow cover). The "core" processes remain the same as, in essence, does the landform: the 'transverse nivation hollow' - 'the cryoplanation terrace'. Whilst mechanical weathering may predominate, there is no necessity for it to be frost weathering. The continued use of two terms, particularly if associated with climatic, size or age considerations, only serves to confuse. Rather, we have a single process suite that has end members that are 'wet' and 'dry' and, probably at distinctly different rates, produce similar landforms. It is here suggested that all such features could be called "periglacial mountain benches", this having no process connotation beyond the association with a 'periglacial' environment and no form linkage other than it is a 'bench' in the 'mountains' (resulting from unspecified periglacial processes - i.e. it is not a glacial bench).

Key words: Nivation Cryoplanation Weathering Periglacial mountain benches

NIVATION OR CRYOPLANATION: IS THERE ANY DIFFERENCE?

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Canada

'NIVATION' AND 'CRYOPLANATION': TWO TERMS THAT IMPLY TWO PROCESS-LANDFORM CONCEPTS

NIVATION

Nivation is said to exploit pre-existing hollows in which snow accumulates and remains after the surrounding area becomes snow-free.
Nivation processes are a suite of processes enhanced and accentuated by the presence of this snow.
Freeze-thaw weathering is assumed to characterize nivation
The landform resulting from nivation is some form of hollow - characteristically a 'nivation bench'

CRYOPLANATION

Cryoplanation is associated with arid or semi-arid cold regions and may be characteristic of permafrost areas
Freeze-thaw weathering is said to characterize cryoplanation
The landform resulting from cryoplanation is the cryoplanation terrace
Cryoplanation can create a terrace on an otherwise undifferentiated slope

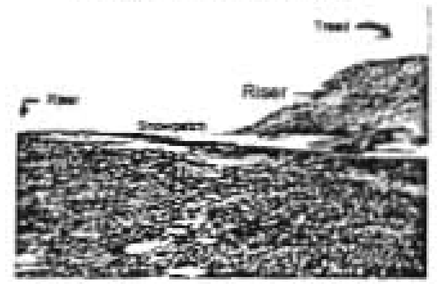
'Nivation' is the "...combined action of frost shattering, gelifluction and slopewash processes thought to operate in the vicinity of snowpatches (French, 1996, p.159)

'Cryoplanation' involves the "...processes of frost weathering.." and "... Processes removing the waste material" (gelifluction and sheetwash) (Czudek, 1995, p.100)

FORM

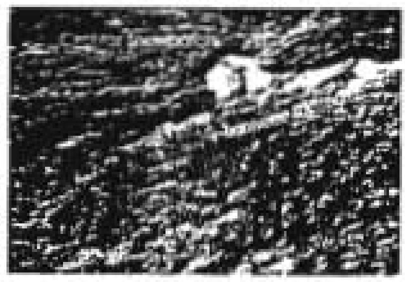
Cryoplanation: transverse (to slope) terrace

'Cryoplanation' in the arid Antarctic



Nivation: circular, longitudinal and transverse forms

'Nivation' in northern Norway



Does a cryoplanation terrace differ from a transverse nivation bench?

PROCESSES

Nivation

Characterised by processes said to be intensified or accelerated by association with a snowpatch in comparison with the surrounding areas that have lesser snow accumulation.

Mechanical weathering, primarily freeze-thaw, is considered the dominant process. Conceptually, many authors characterise a nivation site as one where freeze-thaw weathering is the main cause of hollow development.

Transport processes are largely driven by the water derived from snow melt. Solifluction (gelifluction) and sediment transport by meltwater are considered major processes. Over-snow transport can also be a significant factor.

Significant interaction between weathering and transport characterises nivation and is used to explain hollow growth. Other forms, such as protalus ramparts, exemplify the role of weathering in association with a snowpatch coupled with over-snow transport.

Permafrost is not identified as being significant in nivation. The prime attribute is the presence of snow that persists after the surrounding area becomes snow-free.

Chemical weathering processes and the transport of material in solution are recognised as operative in nivation.

Nivation exploits pre-existing hollows.

Conditions and/or Processes Cited as Being Associated with Nivation and Cryoplanation

Process/Conditions	Nivation	Cryoplanation
Freeze-thaw weathering	✓	✓
Salt weathering	✓	✓
[illegible]	✓	✓
Solifluction	✓	✓
[illegible]	✓	✓
Surface wash	✓	✓
[illegible]	✓	✓?
[illegible]	✓	✓
Over snow transport	✓	✓?
[illegible]	✓	✓?
[illegible]	✓	✓
[illegible]	✓	x
Permafrost	x	✓
Pre-existing hollow	✓	x

✓ [illegible]
✓? [illegible]
x [illegible]

Cryoplanation

Characterised by processes said to be typical of periglacial regions.

Mechanical weathering, almost exclusively freeze-thaw, is considered the dominant process. Some authors go so far as to suggest that cryoplanation characterises the, so-called, "frost-shattering zone".

Transport processes are dominated by gelifluction and sheet wash. Snow is said to be the origin of the water for these processes despite most cryoplanation forms being identified with arid or semi-arid regions.

Transport processes are such that, according to some authors, debris accumulation can inhibit further weathering and/or terrace growth. Transport processes are not as active as in nivation - probably due to the associated aridity of cryoplanation areas.

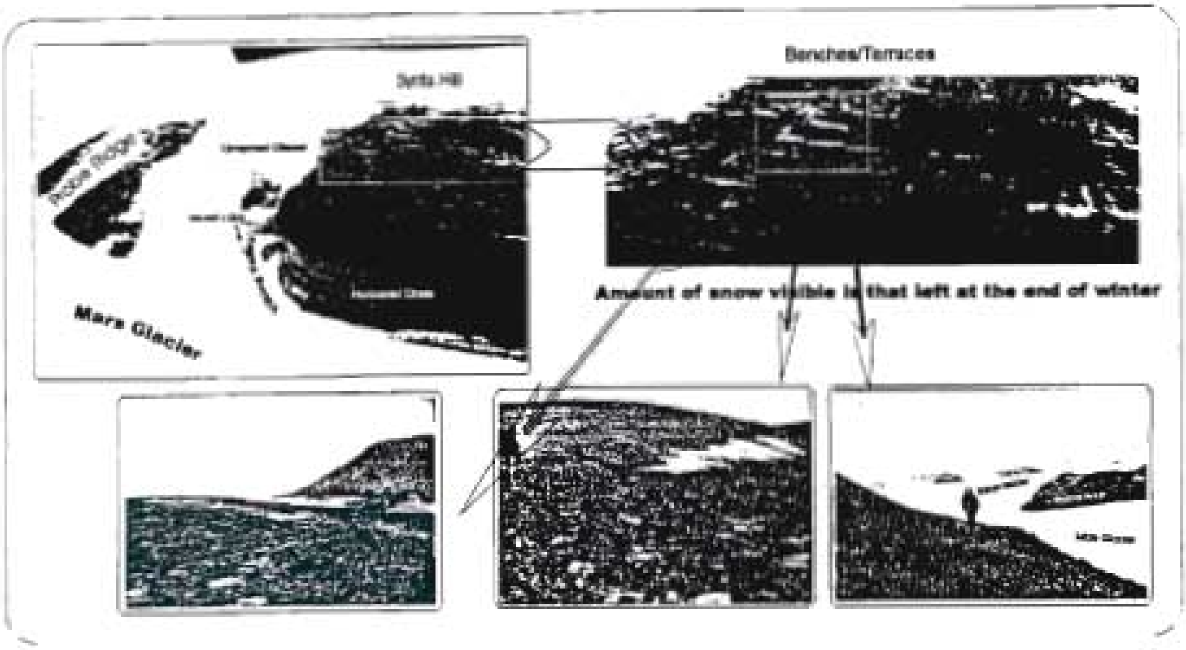
Over-snow transport is not identified as significant.

Permafrost is said, by some authors, to characterise cryoplanation.

Chemical weathering and transport of material in solution are not significant factors.

Nivation is often used as an initiator of cryoplanation, even on an otherwise undifferentiated slope.

QUESTIONS



Are the forms seen here 'nivation benches' or 'cryoplanation terraces'?

What would identify them?
Can they be distinguished by form? If so how?

If processes are the criterion, then are they really any different and are there any measurements (or is it just supposition)?

Even with processes, we may not assume 'cause and effect'. Processes operative today may have nothing to do with those that formed the feature.

Does structure play a role? Clearly structure may be a significant factor in nivation (to help create the initial hollow). Structure is seen by some as significant in cryoplanation. In some instances, cryoplanation terraces are said to cut across structure. This really begs the question of how, then, did they develop? How were processes associated with cryoplanation (namely frost weathering and sheet wash/gelifluction) able to create a form that cuts structure? The features cannot be termed "structural benches" as this explains nothing regarding the process(es) of formation. Further, any such terminology would only add yet another layer of confusion to these forms (a 'nivation bench' - 'cryoplanation terrace' - 'structural bench': how would they ever be discerned?)

How can 'nivation', which exploits pre-existing hollows, be used to initiate 'cryoplanation' on an otherwise undifferentiated surface?
What defines the threshold of transformation from 'nivation' to 'cryoplanation'?
What evidence is there that freeze-thaw weathering is the dominant weathering process?
How can freeze-thaw dominate in arid environments where water is the limiting factor?
If other than freeze-thaw is found to dominate how does this affect the basic concepts - especially where such features are said to characterize a "frost-shattering zone"?
Is there any way to differentiate between a 'cryoplanation terrace' and a 'nivation bench'?
If not then how can they be used in palaeoenvironmental reconstructions (e.g. a 'cryoplanation terrace' equates to the former presence of permafrost)?
If the two features ('cryoplanation terrace' and 'nivation bench') cannot be distinguished by form then are they, in reality, any different?
The difference would have to be related to processes but these are seen to be, essentially, the same
The only difference in processes is the degree of intensification at each site
Recognizing that many nivation and cryoplanation forms may pre-date the last glacial, they may have thus experienced periods with greater and lesser snow than present

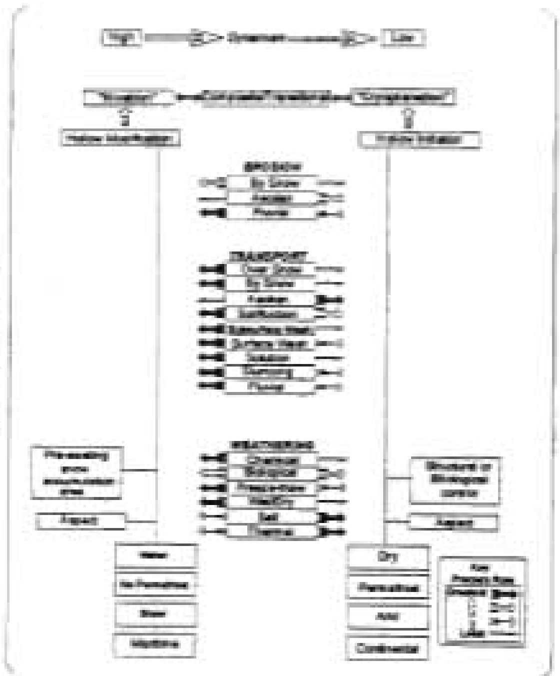
PROPOSAL

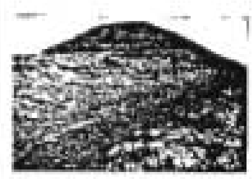
As simplistically depicted in the chart below, it is suggested that 'nivation' and 'cryoplanation' are two end members of the same process/landform continuum.

If the final forms, the 'cryoplanation terrace' and the 'nivation bench', are indistinguishable and the processes of formation are intrinsically the same (differentiated only by degree of operation) then there is no reason to require two terms ('cryoplanation' and 'nivation').

It is suggested that 'cryoplanation' and 'nivation' are distinguished only by dynamism of the processes and that this is largely related to the presence or absence of snow.

It is tentatively suggested that both features should be termed "periglacial mountain benches" thereby describing a form (the *bench*), occurring in the *mountains*, that is of *periglacial* origin.





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PERIGLACIAL LANDFORMS AND PROCESSES: SOUTHERN ALEXANDER ISLAND, ANTARCTICA

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Abstract

Recent studies on nunataks at the southern end of Alexander Island (Antarctica) show a range of periglacial landforms. This is an area of continuous permafrost in a cold, dry climate with limited snow fall. From a periglacial perspective, it is an interesting area with respect to landform development, insofar as it does not have an extensive till cover in which features can develop. Rather, the features described here have developed in either the bedrock itself or the weathered products. A range of features was observed, including taffoni, cryoplanation terraces, sorted and non-sorted patterned ground, and bedrock heave forms. The weathering of the bedrock also exhibited both angular and rounded forms. In addition to describing the features, some information is also provided regarding the formative processes. Notable here is the development of rounded weathering forms (rather than the usually assumed angular forms), on both weathered clasts and the bedrock itself, as a result of mechanical weathering processes. It is also argued that, despite the cold, freeze-thaw weathering is limited owing to the aridity, and that thermal stress fatigue may be the dominant process. An unusual form of non-sorted patterned ground is described where,

as a result of differential weathering of the bedrock, a form of "pseudo-sorting" occurs at the margins of thermal contraction features. In addition to adding to the inventory of landforms and processes for the Antarctic, some of the observations presented here question some of the basic assumptions regarding the character and meaning of periglacial landforms, past and present, in general.

Key words: Antarctic Periglacial Cryoplanation Patterned Ground Weathering Taffoni Asymmetric valleys

Introduction

As the Antarctic encompasses a range of climatic environments from the cool, isothermal and wet sub-Antarctic through to the cold, hyper-arid continent, so there are a wide range of periglacial processes and landforms. This diversity of cold climates may provide climatic-geomorphic analogues for many European and North American locations during the Quaternary. Rarely, however, is reference made to Antarctic work in Northern hemisphere undertakings. The sparseness of information pertaining to the Antarctic region in most major texts results, in part, because sources are disseminated in national reports and within geological descriptions that are not readily available or obvious to many researchers. The problem is further compounded by the fact that most texts available on the Antarctic concentrate on the continent at the expense of other zones, and in so doing not only limit the information but also provide a biassed perspective of the Antarctic as a whole. In an attempt to redress this imbalance and gain a better understanding of the Antarctic periglacial realm, the British Antarctic Survey has facilitated a number of studies at various localities along the Antarctic Peninsula during the past decade. It is the results of one such study, from the southern end of Alexander Island (Fig. 1), that are reported here.

Almost the whole range of periglacial phenomena have been reported from the Antarctic (see Bockheim, 1995 for a recent review), pingos being one of the few major features not so far identified. However, the number of detailed studies are few, particularly when considering the extent and climatic diversity of the area.

Remoteness is a further problem in that it is often impossible to revisit an area. Despite these limitations, the Antarctic does constitute an important geographical region with respect to periglacial studies. The information presented here adds to the general body of periglacial information for this very important region.

Study Area

The present study was undertaken at the south-eastern end of Alexander Island in the vicinity of the Mars Glacier (71°54'S, 68°23'W). This area (Fig. 1) has a number of ice-free valleys and ridges which experience a cold, dry, continental climate. Although there are no climatic stations near enough to the study area to be representative, data from loggers located in Viking Valley approximately 200 to 300 m below the surrounding peaks provide some information. The mean 1993 summer and winter air temperatures are -0.24°C and -9.7°C, with a winter minimum value of -35.2°C and a summer high of +6.0°C. Rock temperature data show a very different environment (Table 1), with a summer low of -4.4°C and a high of +24.4°C (mean value of +5.9°C) but winter mean, minimum and high only slightly warmer than those of the air. Table 1 provides data for several years during which there are notable differences, but in each it is clear that summer rock temperatures are significantly higher than those of the air. It is a region of continuous permafrost, with 0°C temperature, and frozen ground, being encountered at c.0.27 m depth on December 13th (1992) on Syrtis Hill (Fig. 1); the active layer was 0.3 to 0.4 m thick at the most open, snow-free locations on that same date, but substantially thinner near to longer lasting snow (0°C at 0.15 m depth 1m from snow and 0°C at 10 mm depth at snow margin on same date). Aspect plays an important role in controlling ground and rock temperatures, depth of active layer, and landform development (Hookham, 1984; Meiklejohn, 1994; Hall, 1997a).

The local bedrock comprises sandstones, conglomerates, and shales (Taylor, et al., 1979), the most of which are arkose sandstone with sub-spherical siliceous concretions held by a ferruginous cement (Moncrieff, 1989) that lead to the name "cannonball sandstone" (Horne, 1965). Both light-coloured and dark-coloured orthoquartzitic sandstones are also found. These sedimentary rocks are horizontally, or near-horizontally, bedded. The whole area was covered by ice during the last glacial maximum, with an ice mass centred on Alexander Island (Sugden and Clapperton, 1978).

Observations and Discussion

i) 'Cryoplanation' and Asymmetric Valleys

Extensive detail regarding so-called "cryoplanation" terraces are provided in Hall (1997b and 1998a) and on asymmetric valleys in Meiklejohn (1994) and Siegmund and Hall (In Press) and so do not need repeating here. Suffice it to say that features that fit the general description of "cryoplanation terraces" are found on nunataks in this area (Fig. 2). The benches have treads 2 - 12 m wide, sloping at an angle of 1 - 10°, which are 6 - 200 m in lateral extent, and with a riser 0.8 - 2.0 m in height (details are provided in Hall, 1997b). They face west through north to northeast (see Fig. 3 of Hall, 1997b). The whole concept of 'cryoplanation', like 'nivation' (Thorn, 1988), is questioned (Hall,1998a) and so it is suggested that the term 'periglacial mountain benches', a term without specific process connotations, is more suitable. These bench features are so extensive in this area that further investigations are anticipated, especially as other landforms (e.g. contraction features: see below and Hall, 1997a, Figs. 4 & 5) are intimately associated with them. It has been suggested (Meiklejohn, 1994) that asymmetric valleys result from enhanced weathering and transport on the north-facing slopes, thus reducing the gradient of these slopes with respect to their south-facing counterpart. Certainly GIS analysis of the entire nunatak area showed the north-facing slope angle (28°) to be significantly less (at the 99.9% level) than that (32°) of the south-facing (Siegmund and Hall, In Press); in Viking Valley the north-facing slope was 30° and the south-facing 58°. The depth of the active layer was found to exert an influence on slope evolution and a model of asymmetrical valley development was suggested (Meiklejohn, 1994, Fig. 5).

ii) Weathering

It is generally inferred that mechanical weathering, particularly that associated with cold regions, will lead to angular clasts (e.g. Bloom, 1998, p. 311; French, 1996, p.41; Trenhaile, 1998, p.44; McEwen and Matthews, 1998, p.27-28). In fact, in Quaternary studies angularity of clasts is frequently "identified" as resulting from freeze-thaw weathering (e.g. Hanvey and Lewis, 1991; Bloom, 1998). This presumption errs in two ways. First, the angularity may be due to processes other than freeze-thaw (Hall, 1995) and, secondly, mechanical weathering processes need not produce angular forms (Fig. 2b), (Taylor, et al., 1979; Ballantyne and Harris, 1994). This general perception that "angularity implies mechanical weathering" is somewhat naive and, despite the assumption with regard to weathered clasts, is not as universally applied as may be first thought. The classic contra-indicated form is that of taffoni (honeycomb or alveolar weathering) where 'rounded' forms are identified as, largely, the product of mechanical weathering processes (e.g. French, 1996, and see discussion below re taffoni at this site). Taffoni are strongly associated with mechanical processes such as freeze-thaw, wetting and drying, thermal stress and, particularly, salt weathering (Sunamura, 1996; Bland and Rolls, 1998). However, it seems never to have been questioned that these products (in part) of mechanical weathering are frequently exquisitely rounded (Fig. 2 D - F) and thus why should weathered clasts be solely angular and not, in some instances at least, equally rounded? This 'assumption' of angular clasts needs to be questioned.

In the present study, there was a sharp contrast between the weathering of a light-coloured sandstone and that of a coarser-grained, dark-coloured sandstone. Although more detailed studies, based on forthcoming fieldwork, are required, the two sandstones are exposed to an identical weathering environment. The area is dry, with limited snow even in winter, lacking in vegetation but receiving substantial radiation in summer that results in high rock temperatures (Hall, 1997a). Although there are differences in rock temperatures as a result of differing albedos (Hall, 1997a), the operative processes remain similar for both lithologies; only the process intensity differs, largely as a result of lithological differences. Chemical weathering has been shown to occur (Meiklejohn and Hall, 1997), but *only* at highly localised sites inundated by glacier melt water or where radiative heat from rock outcrops causes some local snow melt that ponds against the rock outcrop; this is not the situation for the area described here. Salt weathering is frequently cited for cold, arid regions (e.g. Selby, 1971) but Energy Dispersion System (EDS)analysis of rock samples of both the light and the dark coloured sandstones from this area failed to show the presence of salts. Three spot samples from sub-samples of each sandstone showed negligible amounts of sodium, chlorine, sulphur or magnesium salts. Although the two lithologies are juxtaposed on a (near) horizontal surface, such that their exposure to all weathering elements are comparable, the dark sandstone produces angular clasts while the light sandstone produces rounded forms (Fig. 2B).

While it is highly likely that there will be differences in properties between the two sandstones other than albedo, some of these attributes (e.g. porosity, permeability, etc.) are more linked to moisture and, as this is a dry area, are unlikely to be significant in explaining the resultant weathering forms. Other properties (e.g. strength, internal structure, grain size, etc.) are more germane and likely to offer an explanation for the differences in weathering form; for example, the EDS analysis of the rock samples showed that the light-coloured sandstone is silica rich whilst the dark-coloured sandstone is feldspar rich. Whatever the cause is ultimately found to be, the reality is that under comparable general conditions one lithology produces, even in the bedrock outcrops (horizontal or vertical exposures), rounded forms (Fig. 2C) whilst the other produces angular. Mechanical weathering is likely to dominate (Hall, 1997 a & b) and yet rounded forms can result. Interestingly, in a report on the geology of an area further north on Alexander Island, Taylor, et al., (1979, p. 9) refer to the weathering of sedimentary bedrock into "hemispherical blocks" and weathering is identified as primarily mechanical; certainly there appears to be no evidence for chemical weathering as the cause for the roundness. Further, Clapperton and Sugden (1983, p.134), in a discussion regarding periglacial landforms at Ablation Point, also further north on Alexander Island, refer to rounded debris associated with the weathering of the local bedrock. Thus the association of 'mechanical with angular' and 'chemical (or water-based processes) with 'rounded' can err. This argument has major ramifications for the evaluation of Quaternary sediments in general. The angular dark-coloured sandstone clasts at this site would fit, within the above argument, to the hypothesis of mechanical weathering. However, the observation of rounded, light-coloured sandstone clasts would not, it is argued, normally be considered as mechanically-weathered material produced within a very cold, periglacial environment. The occurrence here of such rounded weathered material indicates that any such judgements could be grossly wrong and thus some Quaternary sediment interpretations may be in error. For example, French (1996, p.236) in summarising evidence for former areas subject to frost action states, "A mid-latitude feature commonly interpreted as reflecting intense frost wedging of exposed rock surfaces is the presence of extensive accumulations of angular boulders...". The argument also becomes circular where, again citing others, French (1996, p.236) notes, ".... in the southern hemisphere....frost-shattered debris are the most widespread of the periglacial phenomena reported.." for, as Thorn (1992) explains, there is nothing that actually identifies a clast as "frost shattered".

Thus, the circularity of the argument is based upon the *presumption* that frost shattered debris *is* angular and thus where it is found it indicates the (former) presence of a freeze-thaw environment. There is a duality to this whole argument: first the expectation of *only* freeze-thaw and second that the result of *only* freeze-thaw is *only* angular clasts. These same issues occur with respect to the angular clasts associated with *grèzes litées* being assumed to be the product of frost action and *only* frost action. It is here argued that these present observations confound these simplistic assumptions. This present study area is a cold region but an arid one. Mechanical weathering is actively taking place but freeze-thaw weathering is highly limited, due to the prevailing aridity, to localised areas of sporadic moisture. The weathered bedrock shows both angular and rounded forms. Thus, both forms can result from mechanical weathering and neither need result from frost action. Hence, any assumption of angularity equating to frost action is totally spurious as too is the assumption that the weathered material should be angular.

There is no question that more studies are needed to explain exactly why there is a difference in the weathering form between the two sandstones. However, those future studies should start *without* the assumption of process, particularly of freeze-thaw, and it is suggested that other present or Quaternary studies should also not presume process, based on false identifiers, but rather look to test it objectively. With equality given to *all* processes, and experiments set up accordingly, it is suggested that the outcomes in terms of weathering interpretations may be quite different (Inkpen, 2000; Viles, 2000). Application of linear elastic fracture mechanics (LEFM) to rock breakdown identifies that there is, in reality, no reason why a crack should not propagate in a curvilinear direction and thus create a more rounded form (Rossmanith, 1983) - the direction is controlled by the interaction of strain release and microproperties of the rock. Internal microstructures will help displace the crack from a straight line (i.e that which would have produced an 'angular' resultant form) as this will be the direction for the most efficient release of the strain energy; in other words, it propagates in response to a changing stress field in the direction where the tensile stress at the crack tip vicinity is maximum (Hall, 1998b). In non-cubic rock forming minerals, the shape of the crack propagation will be selected to minimize stress concentrations, and this can be affected by other cracks exerting an influence on crack trajectory, such that there is no reason why a curvilinear crack might not be as common as a linear one. That the curvilinear weathering form is found everywhere within the light-coloured sandstone, including the formation of taffoni (see below), suggests that the internal microstructure is distinctly different from that of the dark-coloured sandstone. This is now under investigation.

iii) Taffoni

Taffoni are relatively common in the more arid regions of Antarctica where they have been identified since the earliest explorations (e.g. Shackleton, 1909, plate facing page 292). Taffoni, honeycomb weathering, cavernous weathering or alveolar weathering (producing "alveoles") are various terms used to describe the production of both discrete and compound hollows in vertical rock surfaces (Sparks, 1976). Here these forms are all discussed under the term "taffoni". Freeze-thaw, wetting and drying and salt weathering are hypothesised as being the dominant agents of disintegration by many authors (e.g. Selby, 1971). However, Sunamura (1996, p.742) comments that the presence of salts in taffoni does not necessarily mean they were the causative agents. Conca (1984) and Conca and Astor (1987) have suggested that Antarctic taffoni may be due to the combination of the moisture regime of water within the rock and the occurrence of rock surface coatings that determines the flow path of the interstitial water. Taffoni occur where granular disintegration follows lines of equipotential which are concave (Conca and Astor, 1987). Conca and Astor (1987, p. 154) argue that the mechanical weathering, such as salt weathering, that occurs is controlled by the moisture flux. Uzun (1998) shows that taffoni can result from the combination of chemical and physical processes, with dissolution of binding cement in sandstones coupled with the hydration and dehydration of salts can cause cavernous weathering. However, some hollows show flaking rather than granular disintegration and thermal stress has been ignored as a possible causative mechanism in most arguments.

On Alexander Island taffoni are common on the sandstone bedrock exposures (Fig.2D) and measurements of depth, height and width were obtained on the east-, north-, and west- facing sides of a north-facing outcrop (Table 1). It is quite clear that there is a marked aspect difference in the amount of weathering; the taffoni on the north-facing exposure being noticeably larger than on the west which are, in turn, greater than those of the east. There is much less difference in the depths and heights, although those on the east are the least well developed. Schmidt hammer rebound values (which are considered to be a good reflection of the degree of weathering of a rock, e.g. Hall, 1993; Ballantyne, et al., 1989) were obtained from the rock exposures in six areas where taffoni were present (Table 2). A decrease in rock strength, due to weathering, results in a lower rebound value. Data presented in Table 2 show that, with few exceptions, the rebound values for the north-facing exposures are the lowest, those for the west are higher but below those of the east and, where obtained, the south-facing exposures have the highest rebound values (indicating the least weathering). Comparisons of these data (Table 3) show the differences between aspects (using a t-test with rejection set at 99% level and

H_o stating that the two sets of data are so similar they must be considered the same), as outlined above, thereby substantiating the orientation impact upon weathering where the degree of weathering is: north>west>east>south.

With respect to the taffoni, it is worth a brief mention that, as shown in Figure 2E, some display the presence of extraneous material. It is thought that this material was brought in by wind although a glacial origin cannot be discounted; debris shown in Figure 2E includes a clast of a-axis ±10 cm long (see Hall, 1989 for a discussion regarding the ability of wind to move material in the Antarctic). Of the two hollows shown, one has a large amount of external material and the other has not one piece. However, wind cannot be used to "explain" the rounded forms, for nearby clasts of the dark sandstone remain angular (Fig. 2B). Wind may, though, be a significant factor in the removal of the weathering products from the developing taffoni. Last, as shown in the Figure 2F, the surface of the taffoni experiences flaking rather than granular disintegration.

If taffoni are argued to be the product of mechanical processes then why are they not angular in form? In regard to process, the observation of extensive flaking within the taffoni, as well as on rock exposures, must reflect the interaction of process and lithology. This being a dry environment, freeze-thaw, wetting and drying, and chemical weathering are not perceived as significant processes (particularly on the vertical walls where the taffoni occur). However, as Yershov (1998, p. 221) discusses, it may be that the "...impact of wedging out by thin water films", the 'hydration mechanism', may affect the "...ultra- and micro- fractures of rocks.." and cause rock breakdown. As micro-fractures are frequently filled with extremely fine films of bound water so the potential exists for hydration and, during periods of sub-zero temperatures, of 'cryohydration' (Yershov, 1998). Indeed, Konishchev (1982) and Konishchev and Rogov (1993) argue that the freezing and thawing of these micro-films is more deleterious to quartz grains than, for instance, feldspars. As the EDS analysis showed the light-coloured sandstone to be silica-rich and the dark-coloured to be feldspar-rich, so this may exert an influence on the preferential development of the taffoni in the light-coloured sandstone. It is possible that such a process could occur along internal micro-fractures parallel to the rock surface (rather than just within quartz grains) and that this may offer one explanation for the observed flaking. Salt weathering does occur but no evidence of salts within the rock or taffoni could be found (see EDS comments above). Much of the literature has neglected thermal stress fatigue as a factor in cold region weathering but data from this area (Hall, 1997a, 1998c) suggests that it may also play a significant role.

Data from within a taffoni (Hall, 1997a, Figs 10 & 11) show a marked difference between the (interior) bottom and top, with some events recorded in the top equalling the $2^{\circ}C\ min^{-1}$ $\Delta T/t$ threshold required for thermal shock. Yatsu (1988, p. 131) clearly explains $\Delta T/t$ values $\geq 2^{\circ}C\ min^{-1}$ are extremely important, "For heating rates >2°C/min...new cracks and hence permanent strain are developed....the irreversible change is most likely due to the creation of cracks along grain boundaries." The value of ≥2°C min^{-1} may be lower with large bodies (Bahr, et al., 1986), and is also affected by the nature, size and orientation of component minerals. If thermal stress fatigue (and shock) could be accepted then that might help explain the flaking, as the thermal gradients are such that the largest differences are in the surface zone and thus that is where the greatest fatigue will occur (Yershov, 1998). Such a judgement would be in accord with linear elastic fracture mechanics where steep near-surface gradients are expected to produce spalling (Kingery, 1955). Thirumalai (1970) has shown, from studies on basalt, quartzite and granites, that thermal spalling occurs when there is a high thermal gradient and that the spalling is very dependent upon the thermal expansion and shear-strain characteristics of the rock. The need is for more detailed temperature data coupled with an examination of the interaction of thermal stress with rock properties to explain where and when curvilinear failure or linear failure will occur (Kingery, 1955; Hoagland, et al., 1973; Stacey, 1981; Williams, 1986; Tvergaard and Hutchinson, 1988). Kingery (1955, p. 8) clearly explains that thermal spalling can be expected to occur where there is a steep thermal gradient in the rock such that the strain energy in a unit volume is given by:

$$u = \frac{\sigma(1-\mu)}{\mathrm{E}}$$

and the depth of spall would be proportional to:

$$E/(1-\mu)\ \sigma^2$$

where u is the strain energy in a unit volume, σ is stress normal to the face, μ

is Poisson's ratio and E is Young's modulus of elasticity.

Significantly, Kingery (1955, p. 9) notes that thermal cycling can also lead to spalling. Thus, the data obtained for the taffoni, particularly the thermal gradients and cycling of thermal conditions during the summer period (Hall, 1997a), suggest that thermal spalling could well be a significant factor in both taffoni development and bedrock weathering. The observation of spalling inside the taffoni indicates that this is certainly taking place and the measured values of $\Delta T/t$ coupled with the large thermal gradients would seem, particularly in the light of LEFM approaches, to justify the role of thermal fatigue as a potential component causing flaking of the rock within taffoni.

iv) Patterned ground

a) Sorted patterned ground

Sorted patterned ground was not common in this area. The occurrence of sorted patterned ground was limited, largely, it is thought, because of the shallowness of the active layer, the short duration during which it remained unfrozen, the general lack of moisture in this region, and the absence of a till cover over the bedrock in which the features could readily develop. On the flatter areas of Probe Ridge a few small sorted circles were observed whilst on the steeper sections there were some sorted stripes, both developed in the weathered bedrock. Their occurrence is thought to be a result of moisture availability, provided by the melting snow on the ridge, that facilitated segregation ice growth in the surficial sediments during the short summer season.

On the slope to top of Probe Ridge (Fig. 1) a number of sorted forms were found. These were small sorted circles (diam. = *ca.* 1 m) which had, themselves, smaller scale sorting within the fine centres. It was noticeable that these were found in an area wetted by snow melt from a rock step above. At the col between Probe Ridge top and Syrtis Hill (Fig.1) some sorted stripes were seen. These were on an 18° slope facing almost due north. As with the features observed on Probe Ridge, these stripes were found beneath a (15 m x 10 m) hollow filled with snow. Coarse stripes were *ca.* 30 cm in width whilst the fine stripes were *ca.* 50 cm in width. The coarse stripe was in a depression and it appears that melt from the snow above runs along the coarse stripes and removes the fines. Thus, the forms were initiated by sorting but enhanced by water removal of material from the coarse stripes. At other locations on Syrtis Hill melt channels were seen below snowpatches but no frost-induced sorting had taken place. Along the ridge that parallels the Mars Glacier and leads to Two Steps Cliffs (Fig.1) were found what are thought to be plug flows (Washburn, 1997). These are 'islands' of fine material 0.2 to 0. 5 m in diameter isolated amongst otherwise large, coarse debris. Washburn (1997, p. 7) defines plugs thus"...A cohesive, commonly vertical column of gravelly material with considerable fines..." and that "...the surface expression of a plug that breaches the surface is a plug circle or related form of patterned ground...". Associated with permafrost, these are where subsurface sediment masses extend upwards from depth as a result of differential frost heaving in permafrost where the freezing front moves both down from the ground surface as well as upwards from the permafrost table. The fine material from lower down moves upwards through the coarse debris and breaks through at the surface to create this island of fines amongst the coarse debris cover. Some of these features were also observed on a 4° slope where the fines had become elongate downslope (to the south southeast) such that the downslope dimension was 0.9 m whilst the cross slope dimension was only 0.3 m. This form may be what Washburn (1997) defines as a "plug-sorted semicircle where the examples he provides (e.g. his Fig. 25) are on slopes of 5° to 7°. The surrounding clasts had an a-axis dimension ranging from 0.06 m to 0.40 m. It was very noticeable that in all instances the fines had a very sharp boundary with nearly vertical edges that stood isolated amongst the surrounding coarse debris. In other words, the fine material was a distinct 'unit' quite divorced from the surrounding debris and so, in that regard quite different from the relationship observed where sorting produces a fine centre; the material had every appearance of having resulted from a diapiric like action forcing the fines up through the coarse debris cover.

Certainly the forms observed here appear identical to those shown by Washburn (1997) and observed by the author at Washburn's sites; future investigations will involve excavation of these features to better elucidate the nature of the structure.

It was noticeable that the cryoplanation benches did not exhibit any sorted patterned ground on their treads. This is important for it indicates the general lack of moisture, despite the small accumulation of snow at the tread/riser junction. Rather, the treads exhibited, in their lower part (i.e. just above the riser of the bench below) extensive non-sorted patterns that were developed in the thin surficial debris veneer through to the bedrock beneath. Overall, sorted patterns of any type were rare within this area.

b) Non-sorted patterned ground

Non-sorted patterned ground was extensive in this area. Circles and polygons were found developed in the bedrock across the top of Probe Ridge down on to Natal Ridge as well as on Syrtis Hill (largely at altitudes above 500 m a.s.l.) where sandstones and some mudstones are exposed without a till cover; only a few erratics attest to the former presence of ice. The surface weathering layer is largely controlled by moisture availability. Throughout, the surficial weathering layer is shallow, usually less than 0.3 m. Although a thin snow accumulation occurs in topographic hollows, available evidence suggests that the ridges are largely snow-free throughout the winter (Hall, 1997a & b); bedrock weathering is thus generally water-limited. The absence of a snow cover over much of the bedrock does, though, mean that it is exposed to thermal changes that take place throughout the year and this has resulted in thermal contraction features developing in the bedrock (Fig. 3A). The forms tend to appear more circular than polygonal, although the sides comprise a series of linear segments and the intersections between forms are definitely angular.

These forms are all *ca.* 2 m in diameter (Fig. 3B). The visual appearance of these forms is enhanced by the accumulation of coarse debris along the cracks where they are well developed; the accumulation of this debris may also help change the visual impression from polygonal to rounded. The accumulation of coarse debris along the margins gives the impression of a sorted feature but this is not the case. It was also observed that where depressions occurred, in which snow accumulated, the thermal cracks were very narrow just above the snow and grew in size away from the snow area; no cracks could be found under the snow. This indicates that the features are not related to any water-associated process as they increase in size and structure with distance *away* from a water source.

Developed in bedrock, primarily on the dark-coloured sandstone, rather than debris, the coarse border material (10 to 30 cm in width) is thought to result from differential weathering rather than sorting. The margins show a distinct crack, running through the surface material down in to the bedrock, such that the large border clasts are 'balanced' in the boundary crack. The crack width, between the large border stones, was in the order of 5 cm at the top (the total crack width with the stones removed was *ca.* 20 - 30 cm)and was very clear to a depth of *ca.* 10 cm where clasts obscured its total depth. It was a very distinct crack/slot and the coarse material was separated to either side of it. The few observations made showed that the coarse material went to depths of about 14 - 20 cm and that 0°C was measured at depths of 16 to 30 cm; continuous permafrost underlies the whole area. The borders consisted of clasts with 10 to 20 cm a-axis and that clast size decreased with depth, the last few centimetres usually being gravel-sized material. The centres, at the depth of the gravel in the borders was composed of unaffected or slightly weathered bedrock. The centres showed fines, sometimes very clay rich (in the mudstones), at the top to a depth of about 5 to 10 cm followed by broken/weathered bedrock to varying depths with a maximum of about 30 cm.

It is not possible to envisage how sorting could have taken place as this is a water deficient location and the features are developed from bedrock material only. Also, closer to the col between Probe Ridge and Syrtis Hill the polygons comprise mostly just cracked bedrock, with some coarse material at the margins, but almost no weathering of the polygon center. With distance away from here, so the degree of weathering increases as too does the development of the coarse border. This implies a time spectrum with increased weathering away from the col which was only recently uncovered from the ice that is presently retreating down Viking Valley; the ridge tops being the areas first ice-free and thus subjected to the longest weathering/development period. Consideration of Fig. 3A & B shows that there is a surface accumulation of coarse material developed during the time when the bedrock is cracked (B); already a concentration of larger clasts is evident along the bedrock crack. It appears that as weathering progresses the "centres" breakdown to finer material and debris trapped at the boundary crack weathers more slowly such that larger clasts remain to provide the "sorted" appearance (A). The question then remains as to the cause of this differential weathering? It may be that this results from a combination of less (compared to the exposed surface centres) thermal stress fatigue coupled with inhibited water-based weathering (e.g. freeze-thaw). Trapped in the crack, the clasts are exposed to fewer thermal changes compared to the exposed surface rocks. Perhaps more importantly, when snow is available (for instance, a snowier winter or snow blown on to the surface) it can reside on the surface centres but moves down in the cracks. When melting takes place (due to radiation or higher summer air temperatures) the water continues to migrate into the crack but at the polygon surface is available to facilitate water-based weathering. Thus, overall, weathering rates are greatest at the centres and least at the margins, the result of which is that the bedrock in the centres weathers to fines whilst the borders remain coarse. The end result of this is the pseudo-sorted forms observed here which are, in reality, non-sorted patterned ground with the appearance of sorting as a result of differential weathering.

v) Bedrock heave

Bedrock heave is suggested to be a widespread process in permafrost regions. Forms can range from single ejected blocks to dome-shaped accumulations up to several metres in diameter (French, 1996). Dyke (1984) cites possible annual vertical displacements of up to 5 cm. French (1996, p.133) summarizes the available information on this topic and shows that heave is the:

- result of interaction between bedrock and groundwater controls
- upward displacement results from excess pore water pressures in the zone between permafrost table and downward freezing

- bedrock heave is favoured where the water table lies close to the surface
- the active layer may be several metres in thickness
- resistance to heave is the combined weight and shear resistance of the overlying ice-bonded rock mass
-vertical displacement is characterized by progressive and relatively slow movement (up to months).

Interestingly, French (1996), citing Dyke (1978), notes that it is likely that areas of extensive heave indicate either enhanced susceptibility or substantial age, with ages of 5000 to 40000 years being intimated..

Bedrock heave forms were found at one site within the present study area on a north-facing bench at an altitude of 470 m (Fig. 3C & D). The mounds were generally lozenge-shaped with a maximum dimension between 3.5 and 10 m, heights between 0.2 m and 0.45 m and a maximum width of 3 m. It was noticeable that forms all had a longitudinal (Fig. 3D) or transverse (Fig. 3C) crack through them ca. 0.14 m in width and 0.12 m depth, although some were noticeably larger and deeper. It was observed that the bench, comprised of sandstones and mudstones, had a riser to the next bench at its upper end and that, before the riser down to the next bench at the lower end, there was an upward gradient such that the area where the heave mounds occurred was, effectively, a depression. It is thought that this depression facilitated ponding of available water such that, with the continuous permafrost below, freezing of the water in the active layer would produce heave much in the manner outlined above.

Similar heave forms in bedrock (a lapillite-tuff) have been described from elsewhere in the Antarctic (Thom, 1978) where the causative mechanism was said to be the growth and decay of ice lenses.The features described from the South Shetland Islands (Thom, 1978, p. 571) are not dissimilar in size, with lengths in the order of 3 m (maximum = 4.1 m), widths of 2.5 m (maximum = 3.6 m)and heights of 0.3 m (maximum = 0.57 m). In each of these instances the mound had a crack running parallel to the long axis and sometimes another at right angles. Here the mounds were excavated and a debris-free ice lens was found at a depth of 0.45 to 0.6 m below the surface. A segregation ice origin is hypothesised based upon the vertical line of bubbles observed in the ice (Thom, 1978, p.573) whilst the bedrock below was cemented by interstitial ice. The study area comprised permafrost with an active layer thickness of 0.3 to 0.7 m and, although no specific details were given of location, these features appear (Thom, 1978, Fig. 2) to be downslope of melting snowpatches. Interestingly, Thom (1978) suggests that the collapse of the features, due to mass wasting on the sides as growth continues, and the widening of the liner crack, facilitates heat transfer to the interior, so a hollow is produced. As no such hollows have yet been recorded for the study area on Alexander Island this may indicate slower growth as a result of the more moisture-deficient conditions on Alexander Island compared to the South Shetlands.

Discussion

The above observations show a diverse and active cold, dry periglacial environment. Although these observations are preliminary (further work will be undertaken during the 1999/2000 austral season) they do, nevertheless, offer information pertinent to a number of fundamental geomorphological questions and concepts. Recent literature on cryoplanation (e.g. Pinczes, 1997) still describes frost shattering as the dominant process and there still is an apparent confusion between 'nivation' and 'cryoplanation' (e.g. Nelson, 1998) such that the recent argument by Hall (1998a) that the two are end points of the same process continuum seems all the more justified. Further, the orientational observations of 'cryoplanation terraces' from this area (Hall, 1997b), in which most are orientated clearly *towards* the equator (i.e. northwards) is distinctly at odds with the observations of such as Nelson (1998) from the Arctic where, for a mixture of 'nivation' and 'cryoplanation' forms, the preferred orientation is poleward (i.e. northward in the Northern hemisphere). These differences point to the terminological/conceptual problems associated with the term 'cryoplanation' (see Hall, 1998ba) where the relationship of snow and 'degree days' and 'thawing indices', as derived from air temperature data (see Nelson, 1989, 1998) are considered important but which, from available observations seem not to be applicable to the Antarctic situation described here. At this location, the aridity and lack of snow make for an inhibiting of freeze-thaw weathering, a limitation on the role of 'nivation' within the 'cryoplanation' concept, and the requirement of *other* weathering processes (here, mainly thermal stress fatigue) that results in the non-polar orientation contrary to that found in the Arctic. The maintenance of the questionable tenets of the cryoplanation concept explain the problems raised by Nelson (1998) and the presumption of frost weathering (as noted by Pinczes, 1997). Further, those same tenets would fail to explain the origin and maintenance of the forms observed here. Thus, further work on the processes on these active benches is required to help better question or elucidate the cryoplanation concept.

In much the same way, the weathering observations noted here question the assumed tenets. Cold region weathering, certainly the mechanical weathering, is assumed to produce angular clasts. Although observations throughout the world's cold environments *do* show extensive areas of angular material the question must arise as to, based on our presumption of angularity, how often have we been blind to more rounded clasts? Certainly the observations presented here show angular clasts, being produced today, in juxtaposition with rounded clasts, created by the same basic processes - the primary difference being lithology. At the same time, whilst taffoni are frequently identified as the products of mechanical weathering, the resultant form is clearly rounded. Descriptions of 'angular' taffoni are conspicuously absent and yet, if cold-based mechanical processes are deemed to produce angular material, surely there should be some extensive record of such forms? The reality is that rock fracture mechanics do not necessitate resultant products (weathered clasts or bedrock) being angular. Once more, the observations presented here require more detailed studies to attempt to understand what the lithological controls are that produce angular clasts from the dark sandstone but rounded clasts from the light-coloured sandstone. This will be a major part of the future project *but* the observation still remains to beg the above questions.

The peculiar form of non-sorted patterned ground, developed in bedrock, observed here within which weathering differences leave larger clasts at the margins to produce pseudo-sorting requires further study. Although the pattern is produced by thermal contraction, there is a need to understand the development sequence, particularly the weathering, and to determine whether the forms are sand-wedge or ice-wedge or, indeed, composite of the two. As these forms can be found on the benches ('cryoplanation terraces') and show a growth development away from the retreating ice cover, it would be ideal to derive an estimate of process rate as this, together with such as cosmogenic dating, is needed to derive some idea of the temporal framework for the development of all these features. The bedrock heave and plug circles are only important insofar as they are the first reports of these features from the Antarctic. However, as this is a relatively dry zone, they may be significant in demarcating spatial differences in moisture availability. Detailed characterization of the plug circles would certainly complement the study of Washburn (1997) from the Arctic.

Conclusions

The observations presented here provide new information with regard to periglacial concepts and to our knowledge regarding the geographical distribution of forms. Some measurements and observations do not agree with recognised theory (e.g. the expectation of freeze-thaw weathering and the production of solely angular debris) but as that theory has not, in many cases, ever been tested or re-evaluated for decades, the observations here may prove a catalyst for reconsideration of some basic tenets. Whilst some of that presented (e.g. the rounded clasts) cannot be yet explained, nonetheless the observation still stands to question our general assumptions. The information also helps fill the large geographical gap regarding Antarctic observations. It is hoped that the material presented here will encourage more detailed fieldwork and lead to questioning of the (untested) assumptions we so readily accept within our periglacial framework.

Acknowledgements

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Figure Captions

Fig. 1: Location map to show the sites mentioned in the text.

Fig. 2:

A. View across a cryoplanation terrace (riser = ca. 2m high).

B. Close-up of the light-coloured and dark-coloured sandstones close to their lithologic junction to show the marked rounding of the light-coloured sandstone and the angularity of the darker.

C. To show the rounding of the light-coloured sandstone bedrock.

D. View of taffoni developed in the light-coloured sandstone (note the lack of angularity anywhere on the outcrop).

E. View of taffoni in the light-coloured sandstone to show the rounded forms, the absence of extraneous debris in one hollow, and the presence, including relatively large clasts, of debris in the one next to it.

F. Detail of taffoni showing their rounded character plus the scaling at the back of the hollow (scale = 5 cm).

Fig. 3

A. View of the thermal contraction cracks in bedrock (mudstones) with the coarse clasts remnant along the crack margins.

B. Another view of the thermal contraction forms (in mudstone) at an earlier, less well-defined, stage of development but still with remnant clasts along the cracks.

C. One of the heave structures developed in sandstone

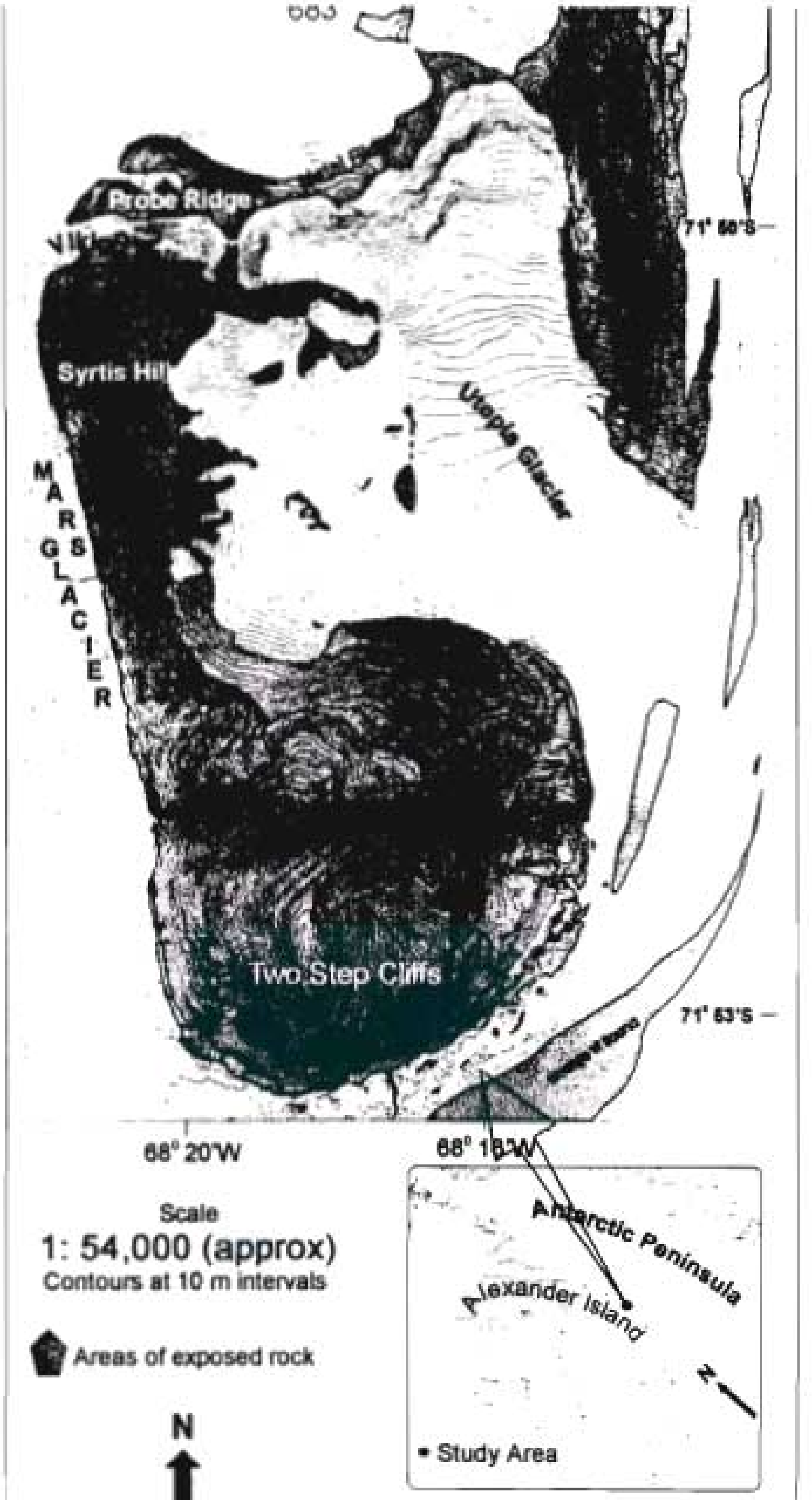
D. Detail of one of the heave structures showing the large crack developed as a result of upward expansion of the bedrock.

Table 1

Temperature data (°C) for the air and rock surface at the study area to show the marked differences between the two.

1993	$\bar{x}$ Summer	$\bar{x}$ Winter	Winter minimum	Summer high	Summer minimum
Air	-0.24	-9.7	-35.2	+6.0	-6.1
Rock	+5.9	-8.7	-33.3	+24.4	-4.4
1994					
Air	-4.9	-9.2	-31.7	+3.7	-14.3
Rock	-1.0	-7.9	-35.1	+21.3	-10.5
1997[1]					
Air	-2.3	-15.3	-44.4	+9.0	-23.1
Rock	-0.9	-11.5	-20.9	+16.0	-17.4

[1]: the 1997 data do not come from Viking Valley (location for the 1993 and 1994 data). The new (long-term) site is located nearby at the foot of Two Steps Cliffs, but it is likely that there may be a minor difference in conditions.



Probe Ridge
Syrtis Hill
MARS GLACIER
Utopia Glacier
71° 50'S
Two Step Cliffs
71° 53'S
68° 20'W
68° 18'W
Scale
1: 54,000 (approx)
Contours at 10 m intervals
Areas of exposed rock
N
Antarctic Peninsula
Alexander Island
Study Area

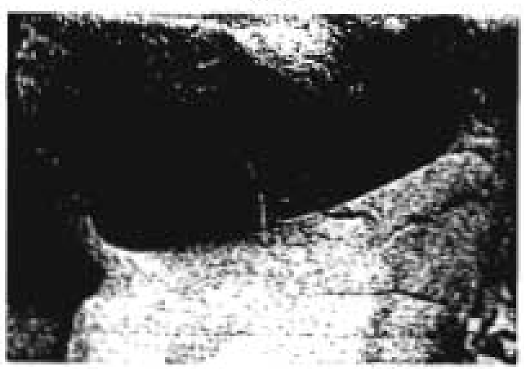
A

B

C

Note both Taffoni and Rounding of Outcrop

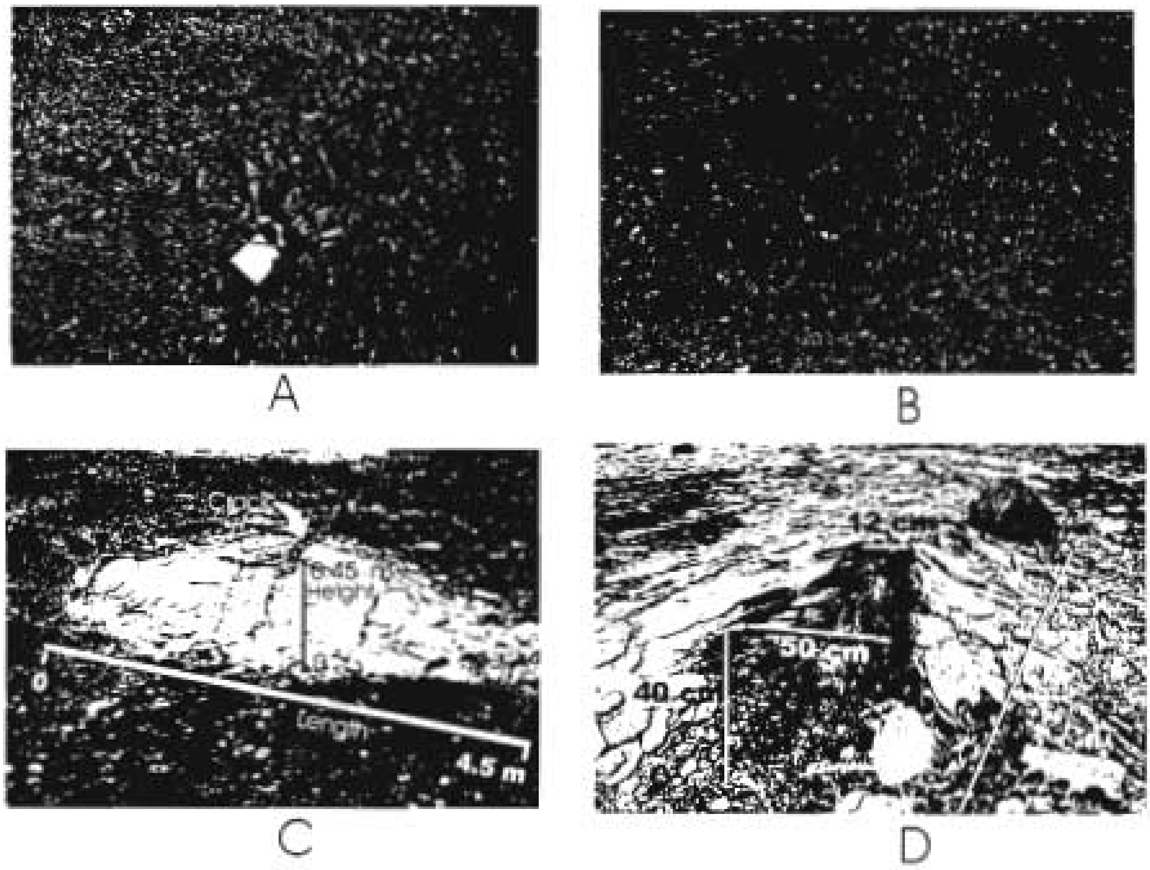
D



E

F

Fig 2


A
B
Crack
0.45 m
Height
0
0
Length
4.5 m
C
12 cm
50 cm
40 cm
D

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Present and Quaternary Periglacial Processes and Landforms of the Maritime and sub-Antarctic Region: A Review.

Abstract

As part of a new initiative by the International Permafrost Association a review is being undertaken of present and past periglacial processes and landforms of the Southern hemisphere. Here details are provided regarding available information pertaining to both the Maritime and sub-Antarctic region. All the islands within this broad zone have experienced glaciation during the Quaternary and many still support ice caps, of varying dimensions, today. These islands currently experience strong westerly winds and high precipitation receipts. Temperatures are less severe than on the continent, with lower values to the south of the Antarctic Convergence and higher to the north. All locations do, however, experience conditions conducive to active periglacial processes and landforms; some of the more southerly locations exhibit discontinuous permafrost. Three reviews (Walton, 1980; Hall, 1987a; Bockheim, 1995) have provided a broad-brush outline of information pertaining to the present landforms for the Antarctic region. The aim here is to give a more up to date account, plus to review the information regarding relict features which are indicative of past, more severe, conditions. Although studies in the Antarctic are, in general, less sophisticated than their Arctic or alpine counterparts, nevertheless there is a extensive body of information that is not well known by most Northern hemisphere workers. A wide range of landforms are currently active, including such as solifluction, sorted patterned ground, protalus ramparts, blockstreams, stone-banked lobes, cryogenic weathering, and nivation. Significantly, the strong winds found throughout this region impact upon landform development, particularly the orientation of some features, to a degree far greater than is recorded for the Northern hemisphere. Here, animals also exert a significant impact within the periglacial regime and comment is made regarding this.

Key Words: Maritime Antarctic, sub-Antarctic, Periglacial Processes, Periglacial Landforms.

Introduction

The many isolated islands of the sub- and maritime Antarctic (Fig. 1) experienced varying degrees of glaciation during the last glacial (Clapperton, 1990; Hall, 1990a). Some, such as Macquarie Island due to its northerly position and low elevations, sustained only a few, very small glaciers. Others, such as Marion Island, had extensive but not a complete ice cover, whilst those closer to Antarctica (e.g. Bovetøya or Heard Island) were totally ice covered (see Clapperton, 1990 and Hall, 1990a for details). Today, ice is still extensive on the islands closest to the Antarctic but those further away sustain only limited ice caps or no ice cover at all. Thus, these islands provide a spectrum of locations that have experienced and/or are experiencing a cryogenic environment of varying severity for varying periods of time. As a result, these islands abound in periglacial landforms, both active and inactive, associated with both permafrost and seasonal freezing. Brief details regarding the periglacial environment of the sub-Antarctic and Antarctic islands have been given in Walton (1980) and Hall (1987a). However, more extensive detail is now available and so it is appropriate to update the information and to put it in the context of recent theory and possible future climatic change.

Climatically, all of the islands, away from close proximity to the continent, are affected by a continuous stream of eastward moving cyclonic depressions. These bring humid air masses, but more temperate, humid air from the north or colder, drier air from Antarctica can also periodically affect the islands (Clapperton, 1990). Climate is severely influenced by the location of any individual island with respect to adjacent land or sea masses as well as its position with regard to the Antarctic Convergence. Land masses can produce a rain shadow effect, as in the case of South America with respect to the Falkland Islands. Islands to the north of the Antarctic Convergence (e.g. the Falklands, Marion and Prince Edward Islands, the Crozet Islands) experience milder temperatures and less snow fall than those to it's south (e.g. Bouvetøya or South Georgia). The islands become increasingly colder with increase in latitude (Clapperton, 1990). Thus, position has a very strong influence on the past and present extent of glaciation as well as the extent and degree of cryogenic activity. It is also significant that not all the islands appear to have been affected by the Little Ice Age but all that were now experience marked glacier retreat (Grove, 1988). Details regarding past and present ice cover for the Antarctic islands can be found in Mercer (1983), Grove (1988), Vinogradov and Psareva (1990), Hall (1990a) and Clapperton (1990). More detailed information and an historical bibliography regarding climatic conditions for this area can be found in Van Rooy (1957), Venter and Burdecka (1966), Dolgin (1986),Gordon and Timmis (1992), Smith, (1993), Appleby *et al.* (1995), Frenot, *et al.* (1997a), and Bergstrom and Chown (1999).

Within this background of past and present ice cover, it is aimed to discuss each island or island group individually. It is hoped that this approach will provide a better picture of the geographical distribution of the islands plus facilitate a more comprehensive understanding of what information is available. One significant limitation of this review, particularly in the light of the title, is that regarding process(es). Sadly the reality is that there has been really very little in the way of true process studies; processes are usually inferred in most reports. This reflects, in part, logistical and instrumentation problems that are only recently being overcome, coupled with the relatively low importance of periglacial studies within most national programmes in recent years. That said, some detailed, instrumented studies have begun (e.g. see the recent work of Boelhouwers, *et al.* (In Press)) in an attempt to investigate processes over long time periods at some locations, but these are few and far between.

The Falkland Islands

The Falkland Islands (51-52°30'S, 60-61°030'W) comprise two main islands, East (c. 5,000 km^2) and West (c. 3,500 km^2) Falkland, and a further 250 smaller islands. The islands are to the north of the Antarctic Convergence and c. 500 km east of the Atlantic coast of southern Patagonia (Fig. 1). They are surrounded by a temperate sea (mean winter temperature at sea level is 2°C), and receive little precipitation (c. 600 mm yr^{-1}) (Clapperton, 1990). The islands experience overcast conditions, with the percentage of available sunshine received being only 20 to 37% of that possible, strong westerly winds, the available precipitation fairly equally distributed throughout the year (Van Rooy, 1957) and only a small seasonal temperature range (Clark, 1972). As a consequence of these present conditions, the climate may not encourage cryogenic activity except on the highest summits (Clapperton, 1990), although Wilson and Clark (1991) have found sorting taking place at 35 m a.s.l. (see below). However, only a small decrease in temperature would be required to initiate a periglacial regime (Clark, 1972).

Clark (1972) provides a detailed synthesis of information available to that date. Features such as blockfields, altiplanation surfaces, tors, block terraces, stone runs, solifluction, thermokarst and scree slopes are all listed and discussed within their climatic context. Clearly one of the most significant works to come from the Falklands is that of Andersson (1906) in which the concept of solifluction is first discussed. Andersson (1906, 1907) utilised the concept to help explain the enormous stone runs that occur amongst the Falkland hills (Fig. 2), features that Darwin (1846) had first commented on and were later discussed by Davison (1889). All discussions regarding stone runs (Andersson, 1906, 1907; Halle, 1912; Baker, 1924; Joyce, 1950; Bellosi and Jalfin, 1984) identify them as of a periglacial origin. Important in this context is that during the Quaternary the Falkland's were glaciated (Clapperton, 1971; Clark, 1972;

Clapperton and Sugden, 1976; Roberts, 1984) but that it was limited to cirque glaciers. Clapperton (1990) identifies only 20 cirques and these are confined to the three highest massifs, and the largest glaciers were only in the order of 3 km in length. Thus, during the Quaternary at the times when the climate was clearly cooler than at present it nevertheless lacked the precipitation for the development of a substantial ice cover as, for instance, took place in Patagonia. This does, though, indicate that the greater part of the Falkland Islands was exposed to a more severe cryogenic environment during the Quaternary and that "...almost the whole archipelago is covered by sediment developed by periglacial mass wasting..." (Clapperton, 1990, p. 232).

Solifluction deposits up to 3 m thick are exposed at some coastal sites (e.g. Bull Valley) where as many as six phases of solifluction have been identified (Roberts, 1984). Available radiocarbon dates (Roberts, 1984) suggest that the last interval of solifluction coincided with the last glacial maximum. However, more recent C^{14} dates (Clark and Wilson, 1992) suggest that climatic amelioration occurred significantly earlier than was previously thought and so ended this period of solifluction activity. The solifluction deposits are now suggested (Clark and Wilson, 1992) to be the correlative facies of periglacial weathering and transport developed in lithologies other than those in which the stone runs are found (e.g. the quartzites and sandstones). Clark, *et al.*, (1998) dated Late Pleistocene organic-rich deposits, enclosed by products periglacial mass wasting, between 36 and 28 ka BP. They found a marked similarity between Late Pleistocene interstadial, Holocene and present-day pollen assemblages and suggest that this reflects the lack of sensitivity of the vegetation to climatic change and/or the lack of climatic variation during this period. Thick accumulations of aeolian and alluvial sands are considered to be derived (Wilson, 1994) from weathering of the local bedrock under periglacial conditions during the Late Pleistocene. In addition to extensive 'head' deposits, Clark (1972) also argues for the former existence of permafrost as evidenced (in the form of lakes) by fossil, degraded pingos and ice-wedge pseudomorphs.

The stone runs comprise "...extensive quartzite blockfields marked by valley axis boulder spreads, almost completely devoid of vegetation, that are attributed to solifluction in a periglacial climate" (Clark and Wilson, 1992, p. 36). Andersson (1906, 1907) suggested that these features resulted from flow-creep of frost-riven quartzite blocks, the quartzose sandstones producing the principal hill features. The hillside debris cover transforms in to stone runs at the lower margins (Fig. 3) where the boulders are several metres in length (that match, in size, the joint-determined blocks of the outcrops), the stone run surface is very irregular and perched boulders are found (Clark, 1972). An important factor, noted by Joyce (1950), is that some runs appear to pass completely over rounded hills but that this may really reflect the juxtaposition of a stone run with an autochthonous blockfield. Strange (1972) makes the point that there is a contrast between the stone runs and the blockfields proper. The runs may have axial gradients as low as 1° with lateral gradients of 6° to 8°. The runs are more extensive than is visually apparent as they both extend below sea level and are often covered by heathland vegetation (Fig. 3); the longest exposed run is *c.* 5 km long; details of this stone run (Andersson stone run) can be found in Bellosi and Jalfini (1984). Bellosi and Jalfini (1984, p.20) found that there was a progressive decrease of block diameter and flatness downstream whilst sphericity and roundness increased but there was only a weak trend in block orientation parallel to the overall flow direction. These authors although not suggestion a rock glacier origin do indicate similarities with rock glacier attributes. Bellosi and Jalfini (1984) suggest that some of the block characteristics were developed after the creation of the stone run as a result of climatic changes at the end of the last glacial. Such ideas are in accord with the recording of deep chemical weathering found by Clark (1976) which has had an impact on recent landscape development.

There is no question that these stone runs remain a controversial feature. Joyce (1950) found it difficult to accept the role of solifluction, in the formation of these features, in other but a minor role. He cites the absence of stone runs on the south slope of Wickham Heights, where quartzites reach their highest point and where solifluction should be operative, as a key factor (Strange, 1972, p. 22). Joyce was also most concerned about the ability of solifluction to actually transport such large material and its ability to do so without jamming of the blocks. Joyce rather suggested that the location of the runs was purely a function of geological structure at the sites but, whilst this can explain the hill-side and near-summit occurrences, it does not account for valley accumulations (Strange, 1972). Strange (1972) also refers to work by Maling, Dodds and other workers who each had different explanations. Dodds (cited in Strange, 1972) found that on summits the loose blocks were still in juxtaposition to each other and showed no sign of movement. Elsewhere, however, there is no question that some form of transport has taken place as quartzite blocks are now across lithologic junctions or resting on other lithologies. Clapperton (1975), based on morphological and internal characteristics ,plus their spatial relationship to glacial features, concludes that they are an extreme form of sorted stripes.

Sorted patterned ground is not only described for the Falkland Islands but is also considered to be actively forming at this time (Wilson and Clark, 1991). At a height of only 35m a.s.l. (Wilson and Clark,1991) found that sorting, in the form of miniature nets and stripes, took place in an area of recent soil erosion. This they attribute to the cool but relatively wet oceanic climate that can facilitate small-scale sorting. Clark (1972) notes frost sorting of the blockfield on Mt Usborne as well as in the debris mantle and stone runs at various locations on the Falklands. He also observes sorting of stony ramparts (protalus ramparts?) enclosing cirque lakes, as well as sorted stripes down to an altitude of 400 m a.s.l. developed in the coarse debris of hillsides. Clapperton (1975), in a detailed discussion regarding the stone runs, concludes that they were, in fact, formed by processes similar to those that form sorted stripes; the stone runs being but an exceptional form at a an extreme scale. Clark (1972) also argues for the former existence of non-sorted patterned ground associated with ice-wedges and uses these as an argument for the previous existence of permafrost in the Falklands. Should this have been the case then this would mean a 12° to 15° drop in mean annual temperature from present.

Marion and Prince Edward Islands

Marion and Prince Edward Islands (Lat 46°48'-46°59'S, Long. 37°35'-37°55'E) are the small (290 km^2 and 40 km^2 respectively) peaks of a submerged volcano (Fig. 1). Located to the north of the Antarctic Convergence the islands experience a cool, isothermal climate with extensive, year-round, precipitation and continuous westerly winds. Although extensively glaciated during the Quaternary (Hall, 1980, 1983a, 1990a) there is presently only a very small (<3 km^2) rapidly receding ice cover at the very top (> 1000 m a.s.l.) of the island (Boelhouwers, pers. Comm. 1998). Based on the reconstruction of glacial maximum equilibrium line altitudes, known lapse rates and from palynological data (Van Zinderen Bakker, 1973; Hall, 1980), a mean annual decrease in temperature of between 3° and 6° is estimated for the last glacial. Such a decrease would give a last glacial maximum (LGM) summer mean maxima and minima of 7.5°to 4.5°C and 2° to -1°C, and winter means of 3° to 0°C and -2° to -5°C respectively. At the present time, frost action at sea level is limited to diurnal frost cycles, frequently associated with needle ice growth (Hall, 1979) whilst at the highest altitudes there is seasonal freezing and possibly permafrost (Holness and Boelhouwers, 1998).

A number of early studies (Van Zinderen Bakker, 1973; Hall, 1979, 1981) identified sorted stripes, stone-banked lobes, miniature sorted circles and vegetation-banked steps (Fig. 4). The stone-banked lobes were, mainly due to their large size and apparent inactivity, considered fossil. The sorted stripes were unusual insofar as they were preferentially aligned parallel to the dominant westerly winds and, in the most exposed locations, were even found on horizontal surfaces. It was thought (Hall,1979) that their origin was related to some form of sorting associated with diurnal freezing, the strong westerly winds and the formation, during calm clear nights, of needle ice. Recently more detailed studies have been undertaken (Holness and Boelhouwers, 1998) that have discerned an altitudinal distribution of features plus identified a number of new forms. Several blockfields have been identified at higher altitudes and further data regarding the large stone-banked lobes has shown that they may have fronts as much as 5 m high, can be up to 20 m in length and several metres wide. Holness and Boelhouwers (1998) suggest that these forms are indeed relict and that they are similar to those described by Benedict (1976) from the Colorado Front Range. Two types of these large lobes are identified, one that develops directly from the weathering of bedrock outcrops (the largest form) and one that develops from platy, clast-rich till. Whilst these larger forms are inactive at the present time, there are also smaller stone-banked lobes that are currently active (Hall, 1981 & 1983a; Holness and Boelhouwers, 1998). It has been shown (Holness and Boelhouwers, 1998, Fig. 8) that there is an increase in lobe size with altitude, with riser heights in the order of 0.1 - 0.15 cm at 200 m a.s.l. rising to *c.* 0.9 m at a height of 500 m a.s.l. At some westerly orientated sites, miniature sorted stripes can be found on the almost horizontal tread of these features.

Sorted stripes are considered presently active on Marion and Prince Edward

Islands (Fig. 4). They occur on a wide range of slopes, from 0° to 20°and generally have the coarse stripe wider than the fine: 140 mm to 84 mm being a typical example. Although Washburn (1973) argued that coarse stripes tend to be narrower than the fine, Hall (1979) found that the coarse were the widest in seven out of twelve areas of study on Marion Island. Holness and Boelhouwers (1998) found that the maximum depth of sorting was between 10 and 15 cm. A full explanation for the association of the stripes with the westerly winds, particularly their development on horizontal surfaces (Fig. 4), or even across slopes, (i.e. parallel to the contours) as was found on Kerguelen (Hall, 1983b), is still needed. Holness and Boelhouwers (1998) also identified *Azorella* terraces (see Fig. 6 in the section on Macquarie Island) as presently active features whilst solifluction terraces are seen as inactive. These two forms are also quite different, the *Azorella* terraces (*Azorella selago* is a cushion plant prevalent in this region) are not always orientated directly across the slope and, although they do show some sorting on the tread, the larger blocks on the surface of some forms appears to be no longer mobile. These forms are likely the product of a variety of downslope movement processes and are seen as presently active by the lack of vegetation on the treads and the movement of clasts on to, or even through, the *Azorella* of the riser. Conversely, the solifluction terraces show vegetation of the riser and the tread plus there is extensive lichen cover on the riser clasts. The similarity of both forms and the likelihood of the solifluction process having played some role in *both* features suggests that more a more rigorous argument is still needed to dichotomize these forms. Long-term movement and other studies are taking place and it is hoped that this argument will soon be explained.

A recent finding is that of fossil large-scale sorted nets at an elevation of 350 m a.s.l. on a south-facing slope of 24° to 28° (Holness and Boelhouwers, 1998). The coarse mesh of these features has a cross-slope dimension of 3 - 3.5 m and a downslope dimension of 1.5 - 2 m. The fine centres are covered with a mixture of grasses and *Azorella* and thus are considered inactive. This combination of larger, but relict, features coupled with smaller, but presently active, forms has led Holness and Boelhouwers (1998, Table 1) to present a vertical distribution of landforms for both the past and the present, with a more active environment, as was also suggested by Hall (1983a), during the early Holocene. Small debris flows, associated with frost-heaved gravel surfaces and the low permeability resulting from frozen ground, have recently been found on one of the volcanic scoria cones (Boelhouwers, et al., In press). These features are short-lived in that it was observed that they were obliterated by subsequent frost heave activity. These observations regarding small debris flows are the first for this region and indicate the potential for significant numbers of such forms in this part of the Antarctic.

Archipel de Kerguelen

This is an extensive archipelago of 300 islands (Lat. 48°27'-49°58'S, Long. 68°25'-70°35'E) located just to the north of the Antarctic Convergence (Fig. 1). The main island, Grand Terre, has an area of *c.* 5,799 km^2 and is about 10 per cent ice covered (King, 1969) whilst the whole archipelago is about 6,200 km^2 in area (Hall, 1990a). Of volcanic origin, Grand Terre experienced substantial volcanic activity during the early Quaternary and still has active fumaroles. The position, just to the north of the Convergence, means that the main island experienced low temperatures (mean = 4.6°C), extensive cloud cover, frequent frosts and strong westerly winds (Weyant, 1967). The bulk of the present day glacier cover (*c.* 750 km^2) comprises the Cook ice cap and its 40 outlet glaciers (*c.*500 km^2) around which peaks rise to 1960 m a.s.l. It is likely that during the last glacial the ice did not fully cover the island (Bellair, 1965) but it must have extended beyond the present coastline. Warming began around 12 ka BP and culminated in major glacier retreat around 10 ka BP (Young and Schofield, 1973). Available information (see Hall, 1990a, p.223) indicates that glaciers are receding, the snowline is rising and that temperatures are increasing on Grand Terre such that more ground is being made available to cryogenic activity at the higher elevations whilst closer to sea level seasonal and diurnal frost effects are decreasing; recent detailed information regarding climate change on Kerguelen can be found in Frenot *et al.* (1997a, 1997b)

Small-scale sorted patterned ground is common on Grand Terre (Troll, 1958; Bellair, 1969; Aubert de la Rue, 1959; Markov, 1971; Nougier, 1964, 1970) and includes polygons, nets and stripes (Fig. 5); at higher elevations some large-scale sorted patterns (Fig. 5) are found (Hall, 1983a & b). The strong stripe orientation parallel to the westerly winds observed on Marion Island (see above) is not so prevelant here as the terrain is far more rugged and dissected, thereby limiting exposure to the westerlies. However, the effects of katabatic winds were seen where sorted stripes were found parallel to the contours along the side of Alouette Valley in western Kerguelen (Hall, 1983b). These stripes cut across a 5° slope orientated to 201° such that the stripe axis was 60° - 240°, roughly parallel to the valley (Hall, 1983b). Elsewhere there was found to be a an increase in stripe width as well as an increase in the fine stripe width (Hall, 1983b, Fig. 5) such that at the lower elevations the coarser stripe was wider but above *c.*200 m a.s.l. the fine became wider. At an elevation of 613 m on Mt Paris 'stripes-within-stripes' were found (Hall, 1983b, Fig. 2) where small-scale stripes were found developing within the fine stripe of a set of large-scale stripes. The larger stripes had a fine width of *c.*1.17 m and a coarse width of *c.* 0.42 m whilst the small-scale stripes were 0.17 m and 0.11 m respectively. It is suggested (Hall, 1983b) that these two sets of stripes result from the combination of large annual freeze cycles to form the large stripes whilst diurnal cycles produce secondary sorting within the larger fine stripe. Small-scale sorting also accounted for the polygons and nets found at a number of locations (Hall, 1983a, Fig. 2). Fine centres varied between 0.64 and 0.21 m ($\bar{x}$ = 0.35 m) in maximum dimension whilst the coarse borders varied between 0.09 and 0.67 m ($\bar{x}$ = 0.27 m) at their widest. It was noticeable that forms developed in a trachyte that weathered to platy fragments showed clasts with their a/b planes vertical at the borders but horizontal at the centres (Hall, 1983a, Fig. 3). Interestingly, forms resulting from solifluction, although not absent (Aubert de la Rue, 1967), are not well reported, nor too are such as the stone-banked lobes found on Marion Island. This may reflect no more than a lack of studies but it may also be due to the longer duration ice cover and the significant regrowth of ice during the Little Ice Age (Nougier, 1970) that may have removed some features. Frenot, *et al.*, (1995) discuss in some detail the impact of freeze-thaw cycles on particle movement and translocation within the context of initial soil development.

Crozet Islands (Îles Crozet)

The Crozet Islands (Lat. 46°- 46°30'S, Long. 50°30' - 52°30'E) are five islands situated roughly half way between Marion Island and Grand Terre (Kerguelen) and to the north of the Antarctic Convergence (Fig. 1). Like Marion and Prince Edward Islands, the Crozets are all of volcanic origin and experience a very similar climate and so it is not unreasonable to expect landforms and processes to be similar to those found for Marion Island. The islands are only 233 km^2 with a highest elevation of 934 m and presently have no permanent snow or ice cover (Walton, 1985). There is little information regarding the nature of Quaternary glaciations on these islands but they may have experienced ice at some time (Hall, 1990a) and Bougere (1992, Fig. 15, p.32) observes moraines, drumlins, cirques, glacial valleys and roches moutonnées. Chevallier (1981) and Giret (1987) suggest that there was a large ice cap on l'Ile de la Possession about 400 ky BP, with ice cover likely also on other islands in the archipelago, and that it was this ice that formed the large valleys. Evidence regarding cryogenic activity is also sparse although Frenot (1987) does provide some information regarding the effects of freeze-thaw cycles on the fellfield above 150 m a.s.l. Small-scale sorted stripes and polygons have been recorded (Philippi,1908; Frenot, 1987) as well as solifluction (Bellair, 1969). The most detailed information available is that by Bougere (1992) who undertook a study of Quaternary history, soil formation and periglacial activity on Ile de la Possession in the Crozet islands. Extensive information regarding solifluction, sorted stripes, sorted nets, frost weathering and even cryoplanation (Bougere, 1992, p. 132-136, Figs. 47- 49) is given. Evidence regarding the action of both pipkrake and segregation ice is presented, both of which are considered important in present day landscape activity. The role of these in the formation of miniature patterned ground is presented including a detailed assessment of patterns, their size and granulometry for 34 sites (Bougere, 1992, Table 13, p. 120). Freeze-thaw weathering is suggested, largely upon the basis of laboratory experimentation, to be operative and details of the granulometry of the weathered basalt is given (Bougere, 1992, Fig. 37). Wind is seen as a major factor in the landscape, causing abrasion of rock outcrops (Bougere, 1992, Fig. 40) as well as deflation hollows (Fig. 39). Bougere (1992) identifies a number of slopes that he interprets as of a cryoplanation origin, with frost action weathering the riser and gelifluction moving material downslope on the tread. A typical example is cited for Plateau Jeannel where the form is at an altitude of 550 m a.s.l. orientated towards 175°. The occurrence of an extensive lichen cover on the rock debris of the terrace suggests it is now a fossil form. Sadly the extensive data and detail regarding the landforms and processes as given by Bougere (1992) are only available in his thesis and so not readily accessible. Frenot (1987) notes that chemical weathering of the basalt is very active, likely a result of the wet climate coupled with temperatures close to, or above, 0°C.

Macquarie Island

Macquarie Island (Lat. 54°37'S, Long. 158°54'E) is a small, ice-free island situated to the north of the Antarctic Convergence and exposed to the full force of the dominant westerly winds (Fig. 1). There has been some controversy regarding the nature and degree of glaciation (see Hall, 1990a for details) but present thoughts suggest that the low altitude of the island could only sustain a few very small glaciers during the Quaternary. The identified cryogenic landforms are very similar to those of Marion Island and the Crozets, namely small-scale sorted polygons, nets and stripes, solifluction features and *Azorella* terraces (Mawson, 1943; Bunt, 1954; Taylor, 1955; Colhoun and Goede, 1974, Löffler, *et al.,* 1983; Selkirk, 1998). Bunt (1954) suggested that the sorted stripes were the result of an interaction between needle ice and water action, with the water removing fines from the coarse stripes, and that the wind may well influence the freezing pattern of the ice needles. Selkirk (1998, p. 491) refers to the occurrence of sorted stone polygons up to 100 mm in diameter.

Taylor (1955) observed that there were differences in solifluction terraces between the windward and leeward slopes. *Azorella selago* inhibited the movement of both when it was able to establish itself (Fig. 6). The leeward terraces were considered to be more stable once established and able to grow laterally to sizes larger than those on the windward side. Although formed in a similar manner, the windward terraces were slowly moving (as opposed to the stabilized leeward ones) and so were not so able to join up laterally and produce terraces as large as those on the leeward slopes. Löffler, *et al.,* (1983) suggested that the size of these terraces was related to solifluction under a former, colder climate but that there was also a strong relationship between vegetation, wind exposure and slope processes. Colhoun and Peterson (1986) argue that during the last glacial the freeze-thaw events on Macquarie Island were more intense even though they may have increased in number only slightly. It is because of the more intense nature of the freeze events that the larger turf-banked solifluction terraces are thought to have developed. More recently, Selkirk (1998) has provided detailed measurements of vegetation-banked terrace movement on Macquarie Island. It is argued (Selkirk, 1998, p. 483) that the terraces are not relict, as suggested by Löffler, *et al.*, (1983), but rather presently active with surface gravel movement in the order of 38-138 mm yr^{-1}. Selkirk (1998) observes that the presence of water, frequently from groundwater seepage, is very important in determining particle movement rates.

Heard Island

Heard Island (Lat. 53°06'S, Long. 73°31'E) is a volcanic cone that is 81% covered by permanent snow and ice (Walton, 1985) situated to the south of the Antarctic Convergence (Fig. 1). Summers are short (Fabricius, 1975) and air temperature fluctuations are limited, with the mean annual temperature (0.5°C) close to zero (Budd, 1964). The snow line is situated at 300 m a.s.l. whilst the highest point rises to 2,745 m and so, with precipitation on 280-300 days per year, the island has little exposed ground. Only at the lowest elevations is there any vegetation or cryogenic activity, and both are mainly found on a series of Pleistocene moraines (Hall, 1990a). The island has been ice covered for some considerable time and thus there has been little opportunity for the development or survival of periglacial landforms. Thus, all periglacial forms found on Heard Island are the product of activity after ice retreat (Colhoun and Peterson, 1986).

Bouvetøya

Bouvetøya is another volcanic island (Lat 54°25'S, Long. 3°21'E) very similar in character and history to that of Heard Island (Fig. 1). Only 50 km^2 in size, it is about 500 km south of the Antarctic Convergence and is some 93% ice covered (Hall, 1990a). Like Heard, any periglacial forms found on the island are thought to be largely the product of processes since glacial retreat. However, above the north coast there is a blockfield (Engelskjøn, 1981) that may or may not be the result of preservation from an earlier period. Surface weathering is also notable in some areas especially as nightly frosts are considered frequent (Engelskjøn, 1981). No other features have been described although there is evidence, in the form of extensive lichen cover, that areas have been ice free for some time and that some of the rock is particularly prone to freeze-thaw weathering (Prestvik and Winsnes, 1981).

South Georgia

South Georgia (Lat. 54°20'S, Long. 36°40'W) is a long, narrow island (160 km long, 5 - 36 km wide) with an axial ridge of mountains, situated just to the south of the Antarctic Convergence (Fig. 1), that experiences a dynamic cryogenic environment in the ice-free areas. At sea level there is a mean annual temperature of 2°C, with a summer mean of 4.5°C and a winter mean of -1.2°C (Clapperton, 1990). Approximately 58% of the island is presently ice covered and most of the ice-free ground is along a coastal fringe below 70 - 110 m a.s.l. (Clapperton, 1990) where, according to Thom (1981), the ground freezes to a depth of 0.5 m for up to 26 weeks per year and permafrost may be present at the higher elevations During the Quaternary the island was completely ice covered except for a few nunataks (Sugden and Clapperton, 1977). Annenkov Island, close to South Georgia also experienced extensive ice cover and periods of periglacial activity that correlates with that experienced by South

Georgia (Pettigrew, 1981). Walton (1980) and Headland (1982) provide bibliographies with respect to South Georgia that include material, to that date, regarding the glacial and periglacial literature.

A primarily descriptive catalogue of the various types of patterned ground on South Georgia was made by Thom (1981) but as a Ph.D. thesis it is not so readily accessible. Stone (1974) refers to the formation of sorted stripes on a number of moraines and screes in northeast South Georgia. Stripes were measured at 5-10 cm in width and in some cases were now overgrown indicating that they were no longer active. Sorted patterned ground was also identified in the form of nets, approximately 1.5 m mesh diameter, as well as large (c. 1 m width) non-sorted stripes on some of the gentle slopes. Some terracing was observed within the areas of sorting. Stone (1975) also refers to an 'unusual' form of patterned ground found at Cooper Bay on South Georgia where lines of the vegetation *Poa flabellata* forms large-scale non-sorted stripes. On some ridges the tussock grass, *Poa flabellata*, occurs in lines approximately 25 cm high and 40-50 cm wide whilst the furrows between are covered with mosses; the whole producing stripes that are about 1 m apart. Air photos of these stripes (Stone, 1975, Fig. 1) show that they are very obvious features down 25-30° south-west facing slopes. Stone (1975, p.197) suggests that although there is no sign of present-day movement the features probably initiated along slurries of fine material that moved over active screes (somewhat similar to the small debris flows of Boelhouwers, et al., In Press).

Heilbronn and Walton (1984) provide more detail regarding small-scale sorted stripes and larger non-sorted stripes, large non-sorted circles and two types of solifluction lobes (one type with a bare terrace and one which is completely vegetated). Heilbronn and Walton (1984) measured the small stripes to have an amplitude of 10-20 cm and a depth of sorting of only 6-7 cm. Spectral analysis of point quadrat data showed that the stripes had three major wavelengths at 120, 55 and 21 cm. The latter (21 cm) stripes were found from near sea level to an altitude of >250 m and generally on slopes of 6-18° with a northerly aspect. They measured an increase in the percentage of material >2mm in both the coarse and fine stripes with an increase with altitude which they attributed to greater downwash or deflation of the fine material from the exposed higher elevations. Large unsorted stripes occurred on a range of slopes up to 30° but mainly on northerly or north-easterly aspects. Crest to crest wavelength was measured at 90-120 cm with trough depth between 15 and 30 cm. All the stripes were completely vegetated with the grass *Festuca contracta* dominating on the drier crests whilst mosses and liverworts were found in the wetter troughs. Large non-sorted circles (⌀ = 1-2 m)were found on an outwash plain (Hestesletten), some of which were completely vegetated and others still had bare centres that exhibit small, sorted nets; needle ice activity was found to be common in the bare centres. Solifluction lobes and benches are described as common on South Georgia. They vary in size and include both turf-banked and stone-banked forms. Pertinently, Heilbronn and Walton (1984, p. 34) note that all the periglacial forms are developed in till and that the general small scale of the forms agrees well with the absence of permafrost but requires the annual, deep freezing event. Regarding the formation of the large, non-sorted stripes, those observed by Heilbronn and Walton (1984) differ from those of Stone (1975), and Thom (1981) could find no convincing explanation for their formation and neither could Heilbronn and Walton (1984). Smith (1960)and Walton and Heilbronn (1983) monitored rates of downslope movement on a variety of slopes. Walton and Heilbronn (1983) found that gelifluction took place to a depth of 12 cm at very active sites but to only 8 cm at most other sites. The majority of sorting was found to take place in autumn, as was also observed by Smith (1960). As in many cases, the observations here regarding mass movement and patterned ground indicate the need for longer-term, more detailed monitoring.

Cryogenic weathering has been a topic well considered for South Georgia. Stone (1974) suggested that the fissile bedrock was particularly prone to frost shattering and that this was the cause of the abundant screes and the formation of felsenmeer (blockfields) at higher elevations (Fig. 7). Gordon (1985, p.45) suggested that weathering has been particularly effective on South Georgia for the past 10,000 years and that the effects of this weathering are most noticeable on the snow- and ice- free mountains below *c.*700 m. Gordon (1985) cites an example, at an altitude of 300 m, where breakdown is proceeding along bedding planes and producing blocks *c.*0.5 m thick whilst, for the same lithology, the foliation is further breaking the rock down in to platy fragments a few millimetres to a few centimetres thick. The propensity for frost weathering, given adequate water, is likely high as, for the year 1975, Thom (1981) measured 214 freeze-thaw cycles in the air at sea level. Even with many of these not being of adequate amplitude or duration *or* being effective on the rock, nonetheless the potential for some effective cycles affecting the rock remains. With this assumption, Gordon (1985) postulates a series of weathering zones that vary altitudinally, spatially and seasonally. The winter maintains continuous freezing conditions in bedrock except at the coast where some cycling can still occur. In summer, frequent freeze-thaw cycles are suggested to prevail in the intermediate altitudes. At high elevations local combinations of insolation, aspect, cloudiness, snow cover and time of day will determine the nature and extent of freeze-thaw cycles. This detailed assessment of the weathering regime of South Georgia by Gordon (1985) was extended to integrate weathering with mass movement as an explanation for the landforms of South Georgia (Gordon and Birnie, 1986).

Gordon and Birnie (1986, Fig. 12) integrate the nature and extent of debris production with the mechanisms of debris transfer and the resulting landforms and deposits for South Georgia via a simple model. As they state (p. 42), they do not "...seek to view individual landforms as unique or special features in mountain geomorphology..." but rather aim to focus "..on the integration of glacier, rock glacier, talus and weathering subsystems." In this regard they show that debris supply and character are determined by the local lithology coupled with available processes and that where lithologies resistant to weathering are present so too are resultant landforms and sediments. This may sound an obvious statement but they actually consider landform distribution within this context. Further, they make detailed observations regarding the rock characteristics and the nature and degree of weathering, including the observation that, despite the cold, chemical weathering does occur. Within their periglacial assemblage of landforms, they identify sorted patterned ground, gelifluction lobes, and rock glaciers that appear to have a glacier ice core. The value of this study is in the identified relationships between bedrock, weathering, transport processes and resultant landform(s). Birnie and Thom (1982) identify two rock glaciers on South Georgia that are thought to occur largely as a result of the debris supply rather than zonal climate. This point is reiterated by Humlum (1998) in that these South Georgia rockglaciers plot outside of the -2°C lower limit of permafrost but may have originated under cooler-than-present conditions and have been maintained by the high debris supply that prevented melt of the ice core. Thus, the relationship of debris supply to landform assemblage is an important one in this region.

South Orkney Islands

The South Orkneys comprise two large islands, Coronation and Laurie, plus two smaller ones, Powell and Signy (60°30'S, 44°25'-46°10'W). With the exception of Signy Island, the islands are extensively ice covered; permafrost occurs in the ice-free areas. All the islands experience a typical cold, oceanic climate with some rain possible in January and February but with snow, which predominates, for the rest of the year; mean annual precipitation is, however, only in the order of 0.4 m yr^{-1}. The mean monthly temperature is *c.* -4°C and low radiation inputs due to the extensive cloud cover and wind speeds average 26 km hr^{-1}.

Available information details the occurrence of cryogenic weathering, sorted and unsorted patterned ground, solifluction and stone streams. The most detailed study on patterned ground from the South Orkneys is that of Chambers (1966a, 1966b, 1967, 1970). Chambers studied sorted polygons, circles and stripes, together with solifluction, in substantial detail; considering temperature regimes, mechanical analyses of the sediments and frost heave measurements. Both short- (Chambers, 1967) and long- term (Chambers, 1970) experiments on these patterned ground were undertaken. Miniature sorted patterns were found to reform within three years but it was found that the mechanism of formation was quite different to that of the large sorted forms - contraction cracks being the major factor in determining the origin and location of the miniature forms. The active layer was found to be in the order of 1.2 m deep but that (Chambers, 1967, p. 19) the top 40 - 60 cm of this was the main zone within which sorting, ice segregation and solifluction occurred. Below this depth, no activity was monitored and, in some instances, a distinct line could be seen dividing these two zones of the active layer. Chambers (1967) also found that at depths below 10 cm only the annual freeze cycle effected movement of stones towards the ground surface and that it is the mass of the coarse borders that helps displace wet, fine sediment plugs up into the fine centres - a major factor in the development of the larger sorted forms. Importantly, Chambers (1967, p. 20) notes that it is not a convective movement that is involved in moving material to the surface but rather the upwelling of fines in plugs - an issue dealt with more recently, and in great depth, by Washburn (1997).

On slopes both creep and solifluction occur and, in the case of miniature sorted forms, creep was considered more significant in moving stones downslope (Chambers, 1967). In the large sorted forms, particularly the large sorted stripes, solifluction plays a major role. Indeed, Chambers (1967, p. 20) suggests that, with respect to large sorted stripes, "The dominant process in bringing about the patterning appears to be streamlining of solifluction in areas unobstructed by large boulders". Large solifluction lobes are also found on Signy Island (Chambers, 1966a) and on Coronation Island (Hall, pers. Obs.). Where the solifluction lobe becomes "...so extended...it takes the shape of a stone stream, winding down the hillside..." (Chambers, 1966a, p. 32). The stone streams have the form of a wide band of coarse stones bordering a small central band of fines, but these features are unlike the large sorted stripes insofar as they do not occur in a series across a broad slope. Flow was found to be, in the stripes, greatest at the surface and in the centre of the fines. These features were considered (Chambers, 1966a) to be related to solifluction of old till deposits.

Weathering of bedrock on these islands was first cited by Grange (in Dumont D'Urville, 1841) and a detailed study of weathering processes was undertaken in the 1980's (Hall, 1986a, 1986b, 1986c, 1987b, 1988a, 1988b, 1990b; Hall and Walton, 1992; Hall, *et al.,* 1986). Detailed information on rock properties, rock temperature, rock moisture content, rock moisture chemistry and the rate of weathering were obtained for quartz micaschist, one of the common lithologies on Signy Island. From a five year study, weathering rates were found to be very slow, with something in the order of $\leq 2\%$ mass loss per 100 years (Hall, 1990b). Whether the rock was a loose block, subject to omnidirectional freezing, or *in situ* bedrock, subject to unidirectional freezing, was found to be significant, with the latter experiencing a rate of breakdown 50 times slower than the loose blocks. Thus, cliffs were considered to be weathering at a *very* slow rate. In this regard, the efficacy of weathering in cold regions may not, everywhere, be as great as is frequently assumed or portrayed in many texts.

South Shetland Islands

A wide range of periglacial landforms and processes have been identified in the South Shetland Islands. The South Shetlands are a mountainous and extensively glaciated group of islands that include (King, 1969) King George Island, Livingston Island, Deception Island, Elephant Island and Clarence Island as well as numerous islets (61° - 63°30'S, 53°30' - 62°45'W). The climate of these islands has a strong maritime influence, with a mean annual temperature between -1°C and -5°C, with a large annual range, frequent precipitation, some in the form of rain, extensive glacier cover and areas of permafrost (Araya and Hervé,1972; Simonov, 1977; Blümel and Eitel, 1989). The observed temperature range coupled with substantial precipitation has led many authors to suggest freeze-thaw weathering to be a major factor in landscape development (e.g. Olsacher, *et al.,* 1956; Araya and Hervé , 1964,1966; Zamoruev, 1971; Simonov, 1977; Dutkiewicz, 1982; Stäblein, 1983; Vtyurin and Moskalevskiy, 1985; Blümel, 1986) and freezing and thawing of the soil to have a major impact on patterned ground forms (e.g. Corté and Somoza, 1957; Araya and Hervé, 1966, 1972; Allison and Smith, 1973:Zhu, *et al.,* 1991; Xiong and Cui, 1991; Hall, 1994). The cool temperatures togther with the substantial snowfall has also led Simonov (1977), Stäblein (1983) and Vtyurin and Moskalevskiy (1985) to suggest that nivation is active on these islands and for Zamoruev (1971) and Dutkiewicz (1982) to identify active solifluction. Zhu, *et* *al.* (1991) also recorded many talus forms including a rock glacier. The combination of cold temperatures, the presence of permafrost and the availability of water has also created argillaceous mounds with a pure, crystalline ice core (Araya and Hervé, 1972). These palsa-like mounds have a heights of 2 to 3 m and a diameter of c. 6 m.

Zhu, *et al.* (1991) state that both the rock glacier and sorted stripes are active during the summer period on the Fildes Peninsula. Sorted polygons comprise penta- and hexagonal polygons 0.5 to 1.5 m in diameter with coarse borders of stones 10 to 20 cm in size (Dutkiewicz, 1982). Some of these forms are interpreted (Dutkiewicz, 1982, p.15) as inactive as they are lichen covered. Xiong and Cui (1991) studied the mechanisms associated with sorted circles on King George Island and found that circles developed well when border material was smaller than 15 cm. Araya and Hervé (1972) observed sorted circles with a centre of fine-grained mud surrounded by a circular ring of angular clasts. The circles were up to 2.0 m in diameter, but there was secondary sorting with small (1.5 to 5 cm) clasts filling polygonal cells within the muddy centre (much as was described for the South Orkney islands by Chambers, 1967). These circles were developed above permafrost measured at 0.05 to 0.25 m below the mud core and 0.1 to 0.15 m below the coarse borders. Some growth of mosses and lichens was observed within the centres, with the vegetation encroaching via the stones within the secondary polygons. This is suggested (Araya and Hervé, 1972, p.107) to indicate that the interior polygons are less active than the unvegetated borders. Araya and Hervé (1972) found that as slopes approached 3° so polygons started to become elliptical and changed in to sorted stripes. Stripe widths are cited as being 5 to 10 cm. Hall (1994) observed that on Livingston Island, despite the presence of permafrost that suggests sorted stripes would be large, the majority of sorted stripes were miniature forms with stripes widths very similar to those described by Araya and

Hervé (1972). A preferred stripe orientation associated with the presence of snowbanks was found: sorting occurring on windward slopes where snow accumulation was least (snow insulating the ground from freeze-thaw cycles on the leeward slopes) . Solifluction was found to be common (e.g. Simonov, 1977) and on some slopes there were seen to be sudden slides of mud and boulders due to an excess of water in the ground (Araya and Hervé, 1972; Simonov, 1977). Hall (pers. obs.) saw slides of muds and boulders, associated with escape of water, that was due to permafrost thaw. Such permafrost thaw slumps may well have been the cause of the features observed by both Araya and Hervé (1972) and Simonov (1977).

The water in the ground was frequently associated with melting of snowbanks that are said (Zamoruev, 1971; Vtyurin and Moskalevskiy, 1985) to play a major role in relief formation. In some instances (Corté and Somoza, 1957) the water from snowbanks is sufficient to cause erosion. Hall (pers obs.), on Livingston Island, observed extensive sediment accumulations along major outwash channels from snowbanks, in some instances the sediments were up to 1.4 m thick. Further, this water availability, coupled with the generally low temperatures, underpins the arguments in favour of freeze-thaw weathering. However, Hall (1993a, 1993b) showed that rock moisture data indicated that weathering by wetting and drying as well as chemical weathering processes are extremely active. Data showed that the southern, lee side of obstacles have high moisture levels during snowmelt and that chemical weathering was enhanced there (as also suggested by Blümel,1986) but that the northern aspects experienced more wetting and drying cycles and so underwent enhanced mechanical weathering. Rock temperature data from the summer showed that, despite the high rock moisture levels, freeze-thaw weathering was not active (Hall, 1993a & b). It is suggested (Hall, 1993a) that weathering due to wetting and drying may be prevalent on northern aspects and that chemical weathering is active on southern aspects. The high moisture contents of the rock do, though, indicate that under freezing conditions extensive damage could occur as a result of freeze-thaw weathering. Blümel (1986) also refers to weathering due to thermal stresses induced in the rock as a result of radiative heating. Interestingly, Simonov (1977) identifies "biogenous" weathering as also active.

Conclusions

The above, somewhat brief, synopsis of the extensive work undertaken throughout this area gives some idea of the extent and depth of the material available and of the work undertaken during the past century. The diversity of climate from the cold but very wet sub-Antarctic region through to the colder and somewhat drier southern maritime Antarctic locations, coupled with the degree and timing of glacial retreat, produces a wide range of periglacial landforms. The extent of Antarctic cryogenic processes and landforms is all the greater when the variations of the continental climate (see Bockheim and Hall, this volume) are also considered. Nevertheless, the actual extent, nature and depth of cryogenic information varies greatly between islands and, in some cases, even within an island itself (e.g. available detail for the east as compared to the west of Kerguelen), such that any meaningful comparisons or future predictions remain, at this stage, unviable. Simply put, so far there is primarily field observation and inference (and even this varies greatly in its quality) rather than hard data that would facilitate any form of synthesis.

The northward shift of the Antarctic Convergence, and the associated severer climate, during the last glacial has left its legacy in the form of fossil landforms, some associated with permafrost, in the more northerly sub-Antarctic islands. Thus, it may also be that many of these Antarctic sites provide an analogue for Northern hemisphere locations during the Quaternary and, as such, are able to offer insights into landform development that are not so readily available now in the North: one such example might be that of active 'cryoplanation' in Antarctica which may offer insights into this little-quantified, with respect to process, landform of the North. Certainly there are detailed studies (e.g. those of Chambers on sorted patterned ground - see above) that offer extent and depth comparable to northern studies but which are little known by many Northern hemisphere workers. However, there is still a great need for more long-term studies and better evaluation of processes within the Antarctic periglacial field.

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Palaeoenvironmental reconstruction from redeposited weathered clasts in the CIROS-1 drill core

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Abstract: The occurrence and nature of weathered clasts in the CIROS-1 drill core from McMurdo Sound was recorded. These data showed both associations of weathered material with particular lithofacies and that certain lithologies were preferentially weathered. XRD analysis of weathering rinds from two lithostratigraphic core units suggest that such weathering data can provide evidence of terrestrial palaeoenvironmental conditions that may be otherwise unobtainable.

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Key words: Antarctica, McMurdo Sound, rock weathering.

Introduction

The CIROS-1 drill hole reached a depth of 702 m in western McMurdo Sound (77°34'55"S, 164°29'56"E), 12 km offshore from Butter Point (Fig. 1). Excellent core recovery (98%) facilitated its division into 22 lithological units (based on marked changes in lithofacies associations) the oldest of which is late Eocene/early Oligocene (Robinson *et al.* 1987). Indications of glacial activity were found throughout the core, including evidence suggesting that ice had advanced close to or over the present drill site on a number of occasions. Evidence for Cenozoic vegetation growth on the Antarctic continent was provided by the occurrence of a *Nothofagus* leaf impression between two glacial beds at 215 m depth (estimated age 30 Ma). Details of the stratigraphy (Hambrey *et al.*, in press), texture (Barrett, in press), clast fabric (Hambrey, in press), clast shape (Hall, in press), and clay mineralogy of the core matrix (Claridge & Campbell, in press) are available.

During the study of clast shape it became apparent that a number of clasts showed clear signs of weathering. Weathering rinds (Fig. 2a) and red-stained cracks in the rocks were observed together with the localized rotting of granites (Fig. 2b); some of the latter displayed the selective loss of feldspars. No data on clast weathering were obtained from either the CIROS-2 (Pyne *et al.* 1985) or MSSTS (Pyne & Waghorn 1980) cores that were obtained from the same area. The data presented here on clasts from two positions in the core add to the knowledge of palaeoenvironmental conditions in the McMurdo region during the Cenozoic. The clasts studied constitute redeposited material which distinguishes this study from most others on weathering rinds (e.g. Chinn 1981, Colman 1981*a*, 1981*b*, 1982, Colman & Pierce 1980, Birkeland 1973) which have focussed on clasts weathered at the site under investigation e.g. in a soil or a moraine. Changes due to diagenesis do not confuse the picture in the present study.

Methodology

Field procedures were confined mainly to visual appraisal of clast weathering seen in the core after it had been cut in half (Hall, in press). At two positions where clast frequency was

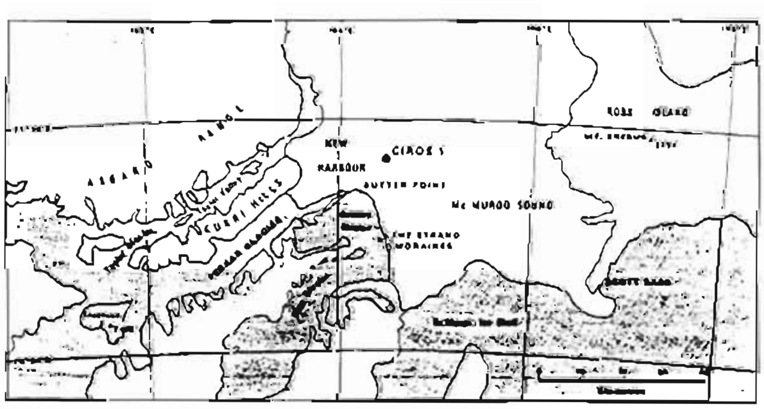

Fig. 1. Map showing location of CIROS-1 drill site.

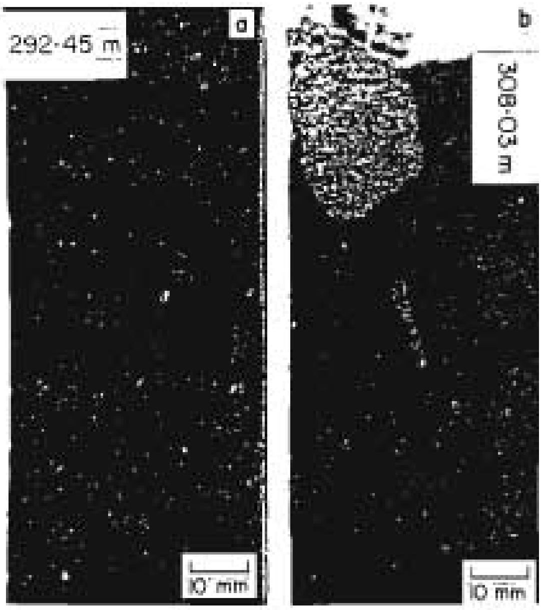

Fig. 2. Two examples of weathered clasts found in the CIROS-1 core: **a.** weathering rind on a dolerite clast, **b.** rotted granitic clast with feldspars weathered out.

high weathered clasts were extracted from the core for mineral analysis of the weathering rinds by bulk analysis X-ray powder diffraction (XRD). Four samples from unit 12 and three from unit 20 were used. The rinds were separated from the rest of the rock on the basis of colour and the crushed samples were packed into the shallow cavity of an aluminium holder so as to minimize preferred orientation. XRD data were obtained from a Philips X-ray diffractometer and Fe-filtered CoKα radiation generated at 40 kV and 40 mA. The specimens were scanned at 1° 2θ/min over the range 2°–75°2θ.

Results and discussion

Although clasts exhibiting signs of weathering were recorded for 14 of the 22 defined units (Fig. 3), this obscures within-unit variability. For instance, of the 215 clasts in unit 17 only 13 showed signs of weathering, and 11 of these were found in a single 3.19-m thick lithofacies. Again in unit 18 (163.29 m thick), although there were only six weathered clasts five were found in the uppermost 10 m of one diamictite lithofacies. Information regarding the lithofacies-specific concentration of weathered clasts that occur in eight of the units are detailed in Table I.

The distribution of weathered clasts in the core should relate to other geomorphological data. The low frequency of occurrence of weathered material in units 3 to 11 (Fig. 3) agrees with the interpretation of these units as distal

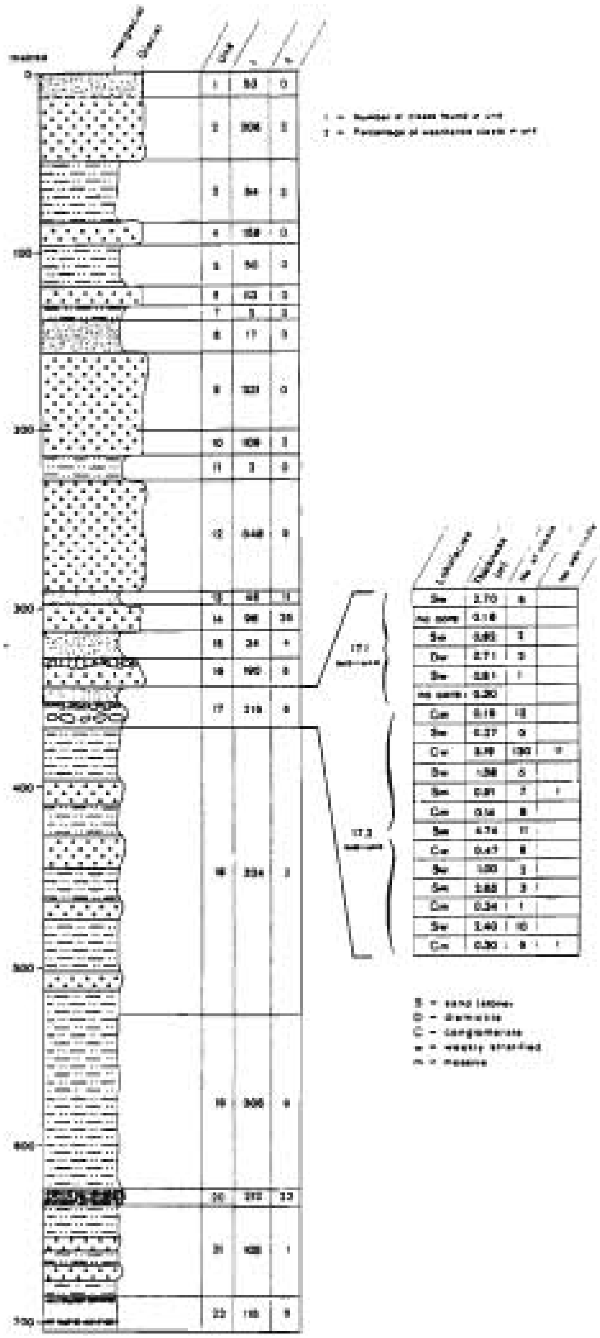

Fig. 3. Generalized data regarding the occurrence of weathered clasts in the 22 core units together with details for unit 17.

glaciomarine or waterlain tills (M.J. Hambrey, personal communication 1987, Hambrey *et al.*, in press). If clast supply to the sediment was distal glaciomarine the probability of finding weathered clasts would be very low. If the debris component of waterlain tills was basally-derived during a time of extensive terrestrial ice cover the clasts would have

Table 1. Occurrence of weathered clasts according to lithofacies.

Unit	Thickness (m)	No. clasts	No. weathered clasts	% weathered clasts	% lithofacies composition of unit					% weathered clasts in each lithofacies				
					S	M	R	C	D	S	M	R	C	D
12	62.25	548	51	9	30.6	14.3	10.2	2.0	42.9	4	0	4	0	92
13	8.18	46	5	11	75	0	25	0	0	100	0	0	0	0
14	13.15	98	25	26	0	50	0	0	50	0	0	0	0	100
16	16.30	190	11	6	50	0	0	25	25	27	0	0	0	73
17	23.58	215	13	6	58.8	0	0	35.3[1]	5.9	7.7	0	0	92.3	0
19	93.90	305	28	9	52.3	46.5	0	1.2	0	60.7	28.6	0	10.7	0
20	6.85	207	47	22	50.0	0	0	37.5	12.5	0	0	0	100	0
22	16.32	115	7	6	38.2	32.4	0[3]	26.5	0	0	0	0	100[2]	0

S = sand(stone), C = conglomerate, M = mudstone, R = rhythmites, D = diamictite

[1] 84.6% of weathered clasts in only one C lithofacies

[2] all weathered clasts in one 0.8 m thick C lithofacies

[3] a breccia comprised the final 2.9%

had little opportunity to weather.

The converse must also be true. Units 12, 16,17, 20 and 22 are all interpreted (Hambrey *et al.*, in press) as being associated with a terrestrial environment within which glacial, fluvioglacial and fluvial conditions conducive to weathering prevailed. Higher concentrations of weathered clasts occurred in all these units. Clear indication that diagenetic alteration did not occur after deposition is indicated by the presence of unweathered clasts residing next to weathered clasts of the same lithology. This is thus quite different to the situation where further change can occur after deposition; as for instance in the weathered talus on rock glaciers cited by Birkeland (1973).

Additional weathering information can be obtained by examining both lithological susceptibility to weathering, and the mineralogy of the weathering rinds. The basalts appear to be preferentially weathered in many but not all units. Weathered material in unit 13, with 51% granitic and 20% basaltic rocks, comprised 66% weathered basalts and 20% weathered granites. In unit 15, where only granites showed signs of weathering, the granites actually comprised only 20% of the clast lithologies present. Quite what the lithological variation in clast weathering indicates is not clear. Basalts generally exhibit more weathering than granites according to the ranking of Gerrard (1988). He ranked the ease of chemical weathering in igneous rocks as: basalt > gabbro > diorite > syenite > granite. Where the granites show the greatest amount of weathering it may be due to mechanical processes. Thus units which show a preponderance of weathered granites (physical weathering) could be indicative of differential environmental conditions to those in which weathered basalts (chemical weathering) predominate. Whatever the case, the imbalance in weathering effects must reflect changes of some sort in the terrestrial environment.

Mineral analysis of the weathering rinds on the basaltic clasts from unit 12 indicated the presence of siderite. Siderite forms in a reducing environment aided by the presence of organic matter (Sokolova 1964). The sedimentological interpretation of unit 12 is that it is proximal glaciomarine to waterlain till, conditions that would produce a reducing environment. In addition, a *Nothofagus* leaf impression found within this unit close (3 m) to where the weathered samples were collected indicates the possibility of some organic matter being present (although the core was not specifically analysed for this). The lack of clay minerals developed in the weathering rinds of these clasts suggests that the environment at that time was not particularly conducive to chemical weathering, although Colman (1982) has shown that clay minerals form only slowly in weathering rinds. Nevertheless, the basalts in the other units show rinds with some clay minerals indicating that environmental conditions for chemical weathering were better at other times.

The rinds from clasts in unit 20 all show the presence of smectite (Fig. 4), normally indicative of chemical weathering within a semi-arid environment. Higher smectite contents in the basaltic rinds than the granitic rinds is not unexpected as the acidic environment generated during weathering of granite inhibits smectite production. The sedimentological interpretation of unit 20 is that it represents a terrestrial environment within which both glacial and fluvial processes were operative. This suggests a cold, but relatively wet subaerial situation, quite unlike that within which the weathering and deposition of the clasts in unit 12 took place. Independent interpretation of the mineralogy of the matrix (Claridge & Campbell, in press) also suggests a terrestrial environment within which podsolized soils formed under forest or scrubland in a cool or cold temperate climate. The smectite found in both the core matrix (Claridge & Campbell, in press) and the clast rims is thought to be allogenic as not only do weathered and unweathered clasts reside next to each other but the absence of smectite/mica interstratification indicates diagenesis must have been low as temperatures could not have exceeded 60°C.

Weathering rinds on clasts from Antarctic offshore drill cores appear to offer direct information on terrestrial palaeoenvironments. Based upon the study of the CIROS-1 core, these constitute three types of data:

a. the nature of the lithofacies in which the weathered

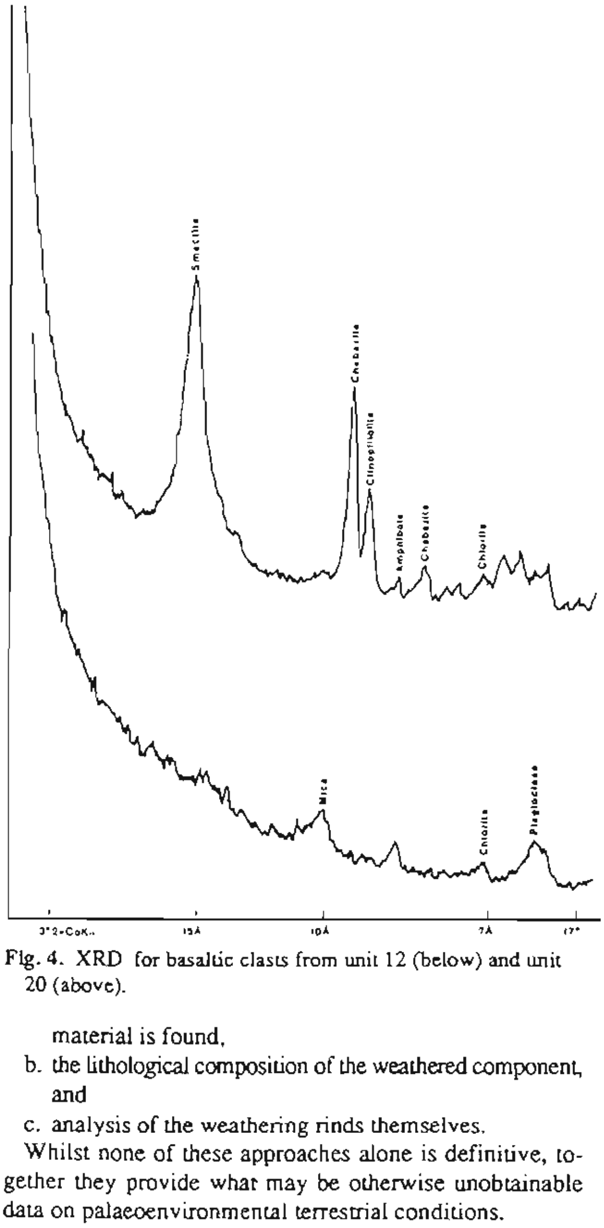

Fig. 4. XRD for basaltic clasts from unit 12 (below) and unit 20 (above).

material is found,

b. the lithological composition of the weathered component, and

c. analysis of the weathering rinds themselves.

Whilst none of these approaches alone is definitive, together they provide what may be otherwise unobtainable data on palaeoenvironmental terrestrial conditions.

Acknowledgements

Analysis of clasts and the weathering characteristics was undertaken by Kevin Hall whilst participating in the joint Victoria University of Wellington, Geophysics Division DSIR and Antarctic Division DSIR (New Zealand) CIROS-1 drilling project in McMurdo Sound. Peter Barrett kindly allowed me (K.H.) to participate in this project and I would like to record my sincere thanks for this. Thanks to P. Birkeland and D. Walton for numerous suggestions regarding references and manuscript improvement. Funding for the field work was made possible by generous financial assistance from the Chairman's Fund of Anglo-American-De Beers and from the University of Natal. Diagrams were redrawn by the Cartography Unit, University of Natal, Pietermaritzburg.

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Short Communication

Some Observations Regarding Protalus Ramparts

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ABSTRACT

A recent study showed that protalus ramparts were self-limiting in that beyond a certain size there was a transformation from stationary firn into a small glacier which would modify or destroy the original landform. Based on this model of protalus development a number of dimensional constraints were suggested. Measurements obtained from a number of protalus ramparts in an area of high debris production of the Canadian Rockies appear to support the model. © 1997 by John Wiley & Sons, Ltd.

RÉSUMÉ

Une étude récente a montré que les moraines de névé étaient limitées par leur propre taille, car au delà d'une certaine dimension, le névé immobile se transforme en un petit glacier qui modifie ou détruit la forme originelle. Sur la base de ce modèle de développement des moraines de névé, un certain nombre de contraintes dimensionnelles sont suggérées. Celles-ci paraissent vérifiées pour plusieurs moraines de névé localisées dans une partie des Montagnes Rocheuses canadiennes où est libérée une grande quantité de débris. © 1997 by John Wiley & Sons, Ltd.

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KEY WORDS: protalus ramparts; slope processes; Canadian Rockies

INTRODUCTION

Ballantyne and Benn (1994) recently suggested a model for the development of protalus ramparts and showed that if the distance between the rampart crest and the talus foot exceeded *c.* 30–70 m there would be a transition from stationary firn into glacier ice with consequent initiation of basal shear and internal creep. As the dimensions of the ice increased and a small glacier developed, so the original protalus rampart would be either destroyed or modified, a 'protalus moraine' being a possible end result. In their model, Ballantyne and Benn (1994, Figure 4) provided a schematic illustration of an idealized protalus rampart together with the attributes that need to be measured for determination of origin. The prime elements to be measured in the field are: the distance from rampart crest to the foot of the talus slope (d), the angle of the talus (β) and the angle of the snow slope (α). From these measurements it is possible to calculate the maximum depth of the firn (h) a factor that is extremely important in the transition from (stationary) snow to (mobile) ice.

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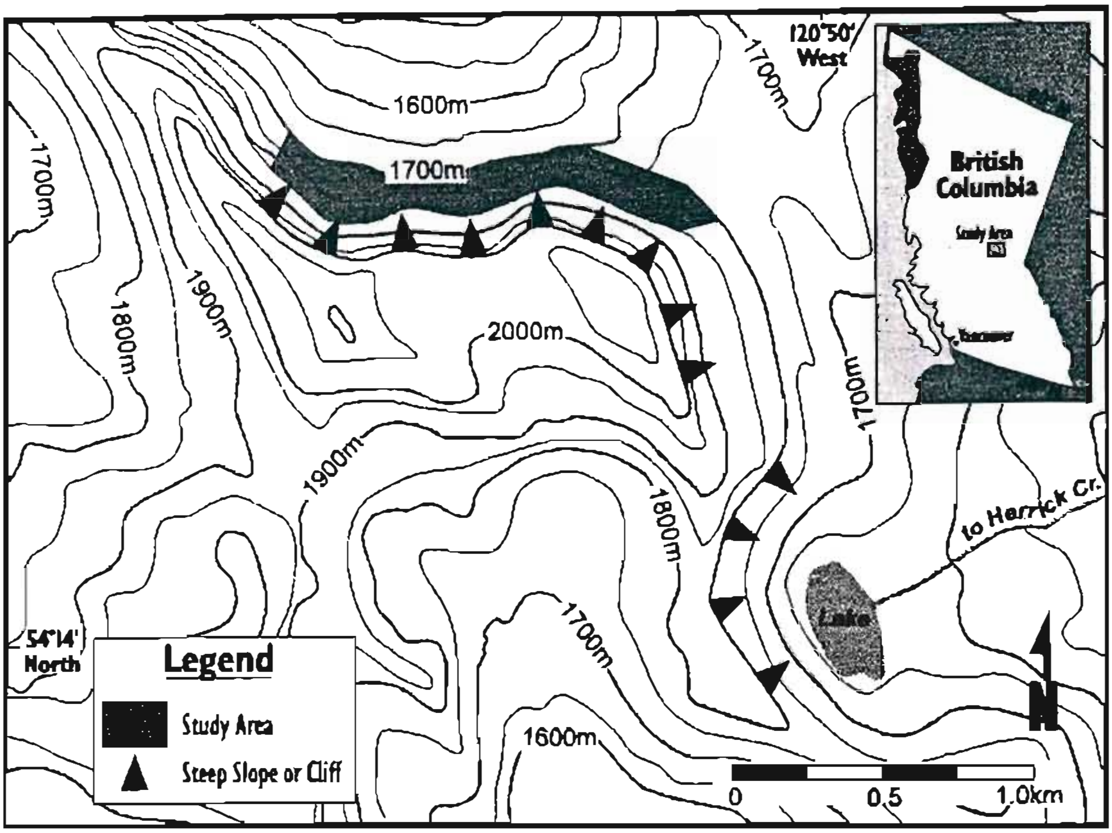

Figure 1 Location of study area.

As part of a study in the Canadian Rockies, a number of measurements of protalus ramparts were made to test the proposed model. These data, together with some ancillary information regarding the observed protalus ramparts, are presented here.

STUDY AREA

The study was undertaken in an area (120°50′W, 54°14′N) just above the local tree line on some unnamed ridges and peaks at an altitude of c. 1850 m in the Canadian Rockies (Figure 1). Lithologically, the area is composed predominantly of limestones but there are also some outcrops of gritstones and low grade schists. The area has been extensively glaciated as shown by glacially smoothed surfaces and erratics; as a result of the glaciation, there are extensive glacially oversteepened valley walls. It is most noticeable that these slopes are highly unstable, with extensive debris spreads over the snow surfaces (Figure 2), large amounts of scree and the sound of debris fall throughout the field period. One reason for choosing this study area was the apparent high debris productivity of the cliffs that results in many associated landforms (Figure 3). It is an area with high winter snowfall and high summer temperatures ($\geqslant 20$ °C). Rock temperatures on all aspects for much of the summer are very high (maximum recorded = +40 °C), even at night, and so the weathering mechanism(s), at least for part of the time when rockfall is active, is other than freeze–thaw; an attempt is being made to investigate weathering processes.

OBSERVATION

Data pertaining to the protalus rampart measurements are given in Table 1. It can be seen that the mean d value is 31.0 m, safely within the bounds of c. 30 to 70 m defined by the model of Ballantyne and Benn (1994). The actual range of d values was

Figure 2 View of the debris on snow slope just upslope of the inner protalus ridge: note the large blocks as well as the fine material.

Figure 3 View of the mountain side in study area to show the extensive scree and resultant accumulation forms.

Table 1 Mean values of the four measured parameters (α, β, d, h) from the protalus ramparts

$\bar{\alpha}$	s	$\bar{\beta}$	s	$\bar{d}$	s	$\bar{h}$	s	n
32.2°	2.8°	28.8°	3.3°	31.0 m	11.3 m	17.4 m	7.0 m	9

from 16.5 m (α = 30°) to 43.0 m (α = 32°) thereby placing all of the observations within the boundaries identified by the model (i.e. for α = 25°, d = 50–80 m; α = 30°, d = 40–60 m; and α = 35°, d = 25–40 m). It would seem that in this area the features that are visually identified as protalus ramparts can be accepted as such owing to their dimensions not being sufficient for glacier ice development. The mean α value of 32.2° is lower than the maximum value (35°) used by Ballantyne and Benn (1994); only one value exceeded their maximum (α = 37°) but the d value of 35 m was within the limitations acceptable for that angle (25 to 40 m). All slope angles were substantially larger (by $\geqslant$10°) than the minimum value of 20° identified by Ballantyne and Benn (1994) as required for movement of debris over firn. That all α angles were $\geqslant$30° may, in conjunction with the ample rock supply, help explain the well developed forms found throughout the study area. The mean β angle of 28.8° was lower than the model value of 35° but, with the sizes of the features in this area, did not conflict with the model expectations.

Correlation values indicate a strong relationship, as might be expected, between values of α and β ($r = +0.97$). The strong relationship between d and h ($r = +0.96$) would also be expected if h were to increase with an increase in d such that as d exceeded the boundary values so there was a transformation from snow to glacier ice. In other words, the data obtained from these highly active protalus ramparts fully support the proposed model and, in so doing, help justify their designation as protalus ramparts.

An interesting observation from nearly all of the protalus rampart sites was that they had a 'double ridge', with an 'outer' ridge beyond the presently active one (Figure 4). Visual observation indicated that the outer ridge was much more subdued, had smaller rock blocks on it (most in range c. 20–30 cm), had a high percentage of fines, and was well vegetated, and that the blocks showed extensive staining from chemical weathering. Conversely, the inner (active) ridge was very 'sharp', had many big blocks ($\geqslant$2 m), with little or no vegetation, and was deficient in fines (there being a high void ratio between the blocks). The blocks were highly angular, many appearing to have been fractured by impact, and showed little or no chemical weathering. Schmidt hammer rebound values taken from the larger blocks on the two ridges at one site clearly indicate a weathering difference: mean rebound of the active ridge is 46.7 ($s = 6.7$) whilst that of the older ridge is 39.4 ($s = 9.2$). A z-test of the means showed that they are significantly different at $p = 0.001$. Cross-profiles of the two ridges were also markedly different. The outer ridge (c. 10 m high) had an

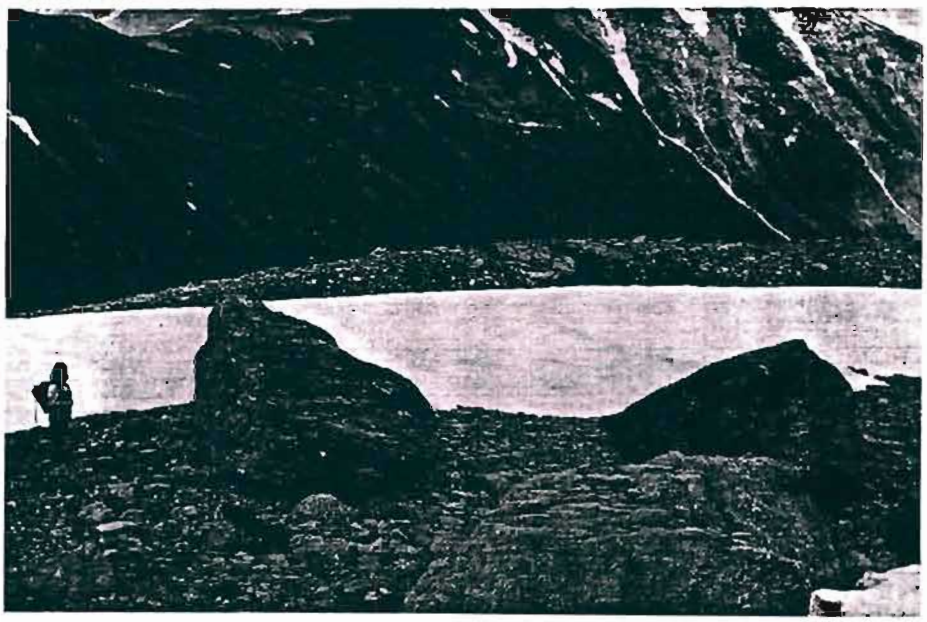

Figure 4 View from the snow slope looking towards the inner protalus ridge (with large blocks and person for scale) and the outer protalus ridge beyond the intervening snow.

outer slope of 25–27°, often covered by solifluction lobes, a 'flattish' (0–4°) top and an inner slope of 6–12°. The inner (active) ridge had an outer slope of 25–35° and an inner slope of *c.* 6°; the top was less distinct, comprising a jumbled crest line. Although no dates, or even relative ages, can be provided, the time frame between the origin of the outer (older) ridge and the inner active one is such that rock breakdown took place, fines filled in the matrix and plants were able to colonize. There must have been a time break between the origin of the outer ridge and the formation of the current one as there is a distinct gap between the two ridges rather than a morphological continuum with two peaks (i.e. the ridge crests). At the site where the rebound values were obtained the distance between the ridge crests was 20.4 m; values for other sites appeared comparable with a range of 20 to 25 m. At the rebound measurement site, the altitude of the inner crest was 6 to 10 m higher than the outer. It was very noticeable that the currently active ridges had many big blocks (often of the order of 4 to 5 m, with 2 m being common) on them and, in some instances, these large blocks had carried into the inner edge of the intervening trough.

Unfortunately, it is not possible to reconstruct accurately the conditions of the outer ridge during the time it was active. The distance to talus foot (d) is unknown and hidden by the inner feature and any difference in the angle of the snow (α) was impossible to determine. However, it seems unlikely that the distance to talus foot was comparable with that of the present d values as this would have necessitated a large amount of talus erosion prior to the development of the inner ridge. As the outer ridge is contiguous and parallels the inner, any such erosion seems unlikely to have taken place otherwise (at least partial) destruction of the outer ridge would have needed to occur in order to allow the erosion and removal of this large volume of debris. Thus, presuming the talus foot to have been in a (reasonably) similar position, the value of d increases to 42 m with an α value of *c.* $\leqslant 31°$ and a β value *c.* 27°. Despite these changes, the model would still, within the assumed parameters, indicate the outer form to be a true protalus rampart and not a 'protalus moraine' of glacial origin. None of this, however, helps explain the timing or the two-phase developmental sequence.

CONCLUSION

Data from an apparently very active region of protalus development in the Canadian Rockies appear to validate the model of Ballantyne and Benn (1994). Fitting of the required measurements into their model confirms the visual interpretation of these features as 'true' protalus ramparts. In this area many of the ramparts have two crests, the outer one (which is still indicated by the model to be of non-glacial origin) being older than the inner. Schmidt hammer rebound values and general observations suggest that the outer ridge is somewhat older than the inner but no evidence is yet available to date it.

ACKNOWLEDGEMENTS

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Rock weathering, soil development and colonization under a changing climate

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SUMMARY

Antarctic continental soils are arid, saline and lacking in organic matter, whereas maritime soils, in a wetter environment, range from structureless lithosols to frozen peat. Two important factors in the development and diversity of their associated terrestrial communities are water availability and the period of exposure since deglaciation. The retreat of ice sheets offers new sites for colonization by microbes, plants and animals.

The interactions between snow lie, freeze–thaw cycles, wet–dry cycles and the length of the summer are considered as critical in determining the extent and rate of localized changes in weathering and pedogenesis. The implications of higher temperatures and differing precipitation régimes are considered in relation to weathering, soil development and the establishment and development of terrestrial communities.

It is concluded that, in the context of decades, most changes will be slow and localized. They are unlikely to be of regional significance, unlike some of those in the Arctic. They will, however, provide a good model of how present soils and communities developed at the end of the last glacial maximum.

1. INTRODUCTION

According to many researchers increasing concentrations of atmospheric greenhouse gases could cause global mean annual temperatures to rise by 2° to 5°C over the next 50–100 years. With respect to the polar regions, Ramanathan (1988) suggests that this temperature increase could be amplified by factors of 1.5–3, thereby producing a warming of 3° to 15°C. The threefold increase will take place primarily during winter and around the margins of the sea ice within the 50° to 70° latitudinal range. The 1.5 factor increase would be a spring-time effect but restricted primarily to the high latitude continents. Due to the melting of ice and snow cover the exposed underlying ocean or land, being much darker, will absorb more solar radiation and thus enhance the initial warming (Dickinson 1986). This 'ice-albedo' feedback could amplify global warming by 10–20% (Ramanathan 1988). In addition, it is possible that near the sea-ice margins the warming could be larger than the global warming by factors ranging from 2 to 4 (Ramanathan 1988). However, Bretherton *et al.* (1990) find that if a dynamic ocean model is utilized then establishment of a large, positive ice-albedo feedback is precluded and the warming may be minimised rather than accentuated. Nevertheless, whatever the magnitude of the actual increase, it will clearly have a significant impact on both polar regions.

The implications for the Arctic terrestrial ecosystems are already under consideration in several countries and will not be examined in any detail here. However there are several valuable indicators from the Arctic which can be used in the development of predictions for the Antarctic. The emphasis in this paper will be more on the maritime Antarctic than on continental Antarctica simply because significant local ice sheet recession has already been detected there.

2. ROCK WEATHERING

One result of the increased temperatures is expected to be a diminution of the polar ice cover, although in the short term the effects will be more immediate and greater for small glaciers and marginal ice sheets. In fact, on the continent there might be glacier growth as a result of increased snowfall. Major ice loss will be more apparent in the smaller and thinner glacial ice of the maritime Antarctic where it may be aided by an increased percentage of precipitation falling as rain rather than snow. This loss of ice will not only affect the terminal regions of the glaciers and ice caps but will also result in a greater exposure of rock on nunataks and valley walls as the glaciers thin. Thus, as the ice retreats so more land, at all altitudes, will be exposed to subaerial processes. As there will be an ever increasing aerial source of rock material and weathering processes are likely to be more active (see below) so the glaciers will be provided with a greater supraglacial debris load. Consequently, as the ice retreats the former glacier-covered area will be mantled by an unconsolidated till deposit of multi-sized, poly-litho-

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logic material subject to weathering at a faster rate than the bedrock (Hall 1986). In turn, this suggests that pedogenic and colonization processes will be be operative sooner and more rapidly than if that till mantle had not been deposited.

The loss of ice cover, particularly if it is rapid, can cause the formation of microfractures in the bedrock as a result of the removal of the constraining overburden. Failure planes, due to tensile forces, result in the splitting of the bedrock parallel to the slope surface (Kawamoto & Fujita 1968 in Yatsu 1988). This fracture system greatly facilitates and enhances weathering by providing a ready means of ingress for the water required for the operation of most mechanical and chemical weathering processes. In addition, Crook & Gillespie (1986) suggest that both granitic and sedimentary rocks can ultimately disintegrate completely as a result of stress relief alone. The fracturing due to stress relief is particularly active on steep slopes (see Yatsu 1988, Figs. 2.2.23 & 2.2.24) and valleys and nunataks undergoing glacial retreat will be particularly vulnerable to weathering, further increasing the debris supply to the retreating glaciers.

Although weathering rates are strongly controlled by climate, micro- and nano-climatic data are not yet adequate to allow comparisons, either temporally or spatially, within the Antarctic. In addition, the results of most laboratory weathering experiments are of doubtful value as indicators of actual weathering rates as the experimental régimes rarely replicate natural conditions (Thorn 1988). Considerations of scale cannot be ignored in any attempt to link ecological and atmospheric models (O'Neil 1988), but the problems of scale at the terrestrial–atmospheric interface are both great and complex (Dickinson 1988). Weathering processes, at both landscape and niche levels, operate at temporal and spatial scales orders of magnitude lower than those proposed for climate change. Although these problems of determining the scale of study and elucidating weathering rates a very real, neither can enter significantly into tl present discussion since we know so little abo Antarctic weathering rates and processes (Hall 199 that we are not yet in a position to deal with the factors in detail. Thus, in this discussion we can onl consider the changes to weathering as a result of warming climate in broad, somewhat speculative terms.

A further ramification of increased temperatures i that they will cause thawing of permafrost. Perma frost, which is ubiquitous throughout the ice-free area of Antarctica (Campbell & Claridge 1987), is ground that remains at or below 0°C for at least two years. Even though the permafrost may not entirely disappear in all areas, although this is a possibility in parts of the maritime Antarctic, nevertheless the zone above the permafrost that thaws each year (the 'active layer') will increase in depth. This is extremely important for a number of reasons. First, weathering, soil development and colonization can only take place in the active layer (Carter 1990). Second, at the present time temperatures are such that there is no active layer (Campbell & Claridge 1987), or it is extremely thin, on most of the continent. Thus, with warming, even the extremely cold areas are likely, with time, to experience the development and deepening of an active layer. Third, the unfrozen moisture present in an active layer facilitates the operation of cryogenic processes such as frost sorting, frost heave and gelifluction all of which interact with weathering, soil development and colonization.

Warming will thus progressively increase the zone available for weathering, pedogenesis and colonization in all spatial dimensions. Clearly, this will not be equal everywhere in the Antarctic, there being a dimensional increase along a transect from the continent interior through the Antarctic Peninsula to the maritime Antarctic (figure 1). Along this transect ice

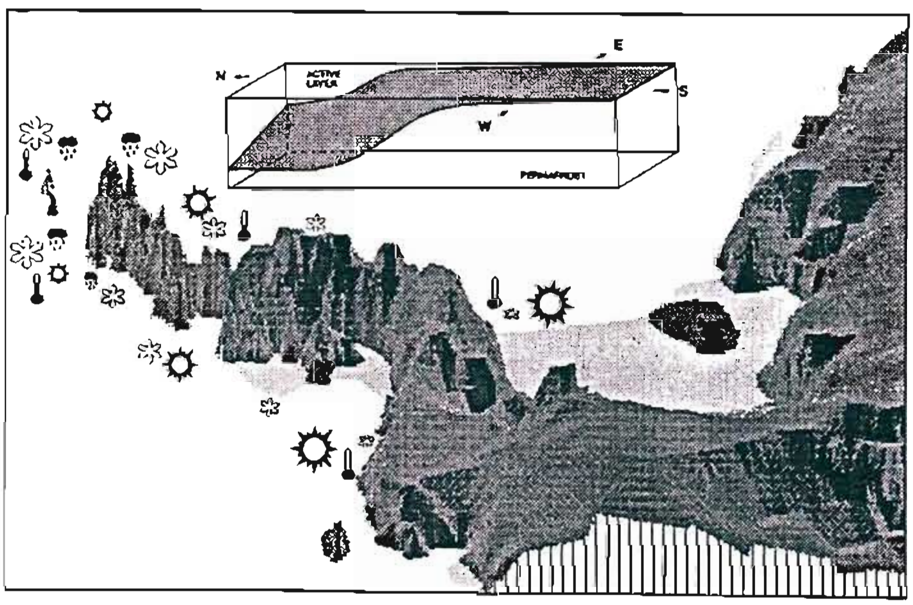

Figure 1. Simplified map to show the spatial variability of permafrost and climatic parameters from the maritime to continental Antarctic.

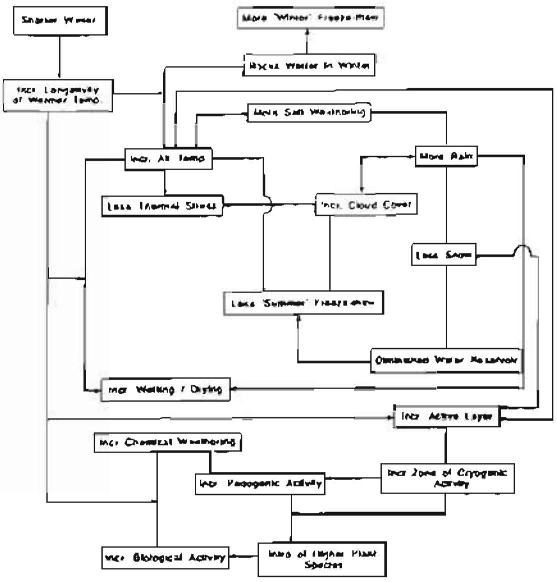

Figure 2. A flow diagram to show some of the effects of climatic warming for the maritime Antarctic environment.

thinning and retreat will be, in the short time frame, least on the continent and greatest at the northern extremity of the maritime zone. Equally, the deepening of the active layer will be most rapid in the maritime zone where present summer temperatures have already melted the permafrost to a depth of over 1 m in places. However, once established on the continent the existence of an annual active layer could have a marked effect on cryogenic and pedogenic processes. A further result of climatic warming would be to increase the temporal dimension with respect to weathering, pedogenesis and colonization. Along the same spatial transect described above there would be a diminution of the longevity and severity of cold conditions and a consequent increase, again greatest in the maritime zone, in the annual period available for active weathering.

3. WEATHERING AND SOIL DEVELOPMENT

Weathering is the precursor to soil development to the extent that many workers recognize a two stage situation rather than a continuum. For instance, Matsui (1969) identifies weathering as the transformation of bedrock to detrital sediment whereas pedogenesis is the change from sediment to soil (the stages of 'geochemical weathering' and 'pedogenic transformation' according to Millot (1982)). However, in spite of this bipartite division the factors that control both weathering and pedogenesis are essentially the same, namely climate, parent rock, topography, vegetation, hydrological conditions and time (Yatsu 1988). Here it will be chiefly the effects of climate that are considered. Both topography and vegetation (due to its spatial and temporal limitation within the Antarctic) are highly site specific. Although hydrological conditions are considered with respect to rock moisture content, the weathering and pedogenic associations with drainage are ignored as these are again constrained by local conditions. Parent rock is important at two levels, those of physical properties and of rock chemistry. Both are briefly considered here but in broad terms rather than in specific detail. Finally time is of particular significance with respect to weathering for, given adequate time, even the chemical effect of the parent rock can be nullified (Chesworth 1973). However, it is the immediate to short time span that is discussed here and so the longer-term effets of time on various attributes of both weathering and pedogenesis are ignored.

Two broad climatic scenarios are considered, both involving a warmer climate but differing in terms of precipitation régimes. The first, which is applicable to the continental environment, involves a higher than present snowfall whereas the second, which is relevant to the maritime Antarctic, entails increased precipitation in the form of rain. These changes in precipitation, in conjunction with higher temperatures, will have varying effects upon the present temporal and spatial disposition of weathering processes. In some instances certain weathering processes will be initiated or enhanced whereas in other instances current processes will be inhibited and succeeded by a new weathering suite. The nature of weathering will change, producing different weathering products which, in turn, will have both direct and indirect effects upon pedogenesis and colonization.

Considering first the maritime end of the spectrum (figure 2). The change here is primarily controlled by a greater incidence of precipitation in the form of rain. If total precipitation remains unchanged the corollary of more rain is less snow. Combined this means that whilst there will be a more frequent and comprehen-

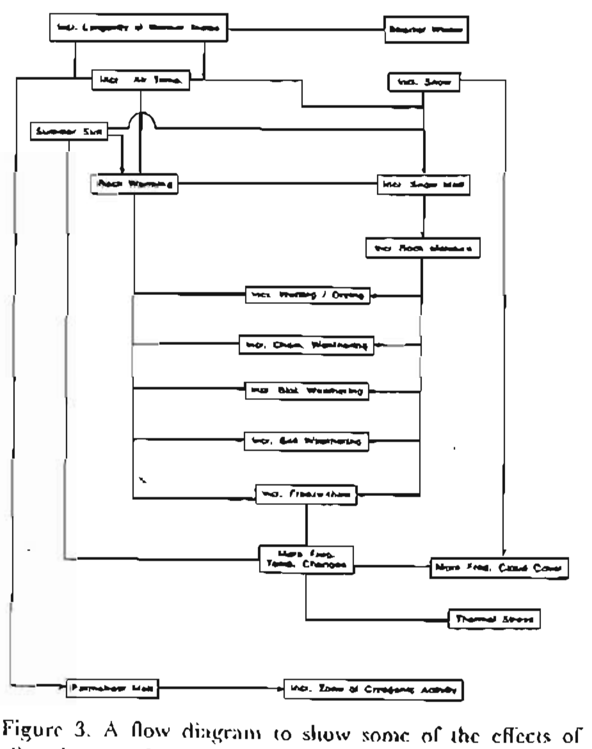

Figure 3. A flow diagram to show some of the effects of climatic warming for the continental Antarctic environment.

sive wetting of the rock, there will be a more limited reservoir of water (i.e. snow) to be released to the rock during times of high temperature. At present, during the summer, cold phases frequently follow warm periods and so freeze-thaw can take place at those locations wetted by snowmelt. With both warming and a greater incidence of rain there is likely to be less freeze-thaw during the main spring to autumn period. Only if rain precedes the freeze at the end of autumn will frost action take place during this 'summer' period. Thermal stress fatigue, although not apparently a major agent in the maritime Antarctic, may be further inhibited by the increased cloud cover.

Conversely, with global warming effectively extending the length of the summer and the rain facilitating extensive wetting of rock, so the weathering process of wetting and drying becomes temporally and spatially more active (figure 2). Salt weathering is also enhanced, with salts being introduced or mobilized during the wetting phase and crystallization occurring during drying. Wetting–drying cycles and salt weathering both operate in the outer shell of the rock, producing small sized debris which is difficult to partition between processes. The more frequent presence of rock moisture combined with higher temperatures will also promote chemical weathering. The combination of more active weathering processes together with their synergistic interaction will result in increased overall weathering rates.

If the higher temperatures are combined with decreased snowfall this will produce a rapid deepening of the active layer (figure 2). As the active layer deepens and more moisture is made available from rainfall so there will be an increase in pedogenic activity. In turn, these changes may facilitate the establishment of vegetation, including, in the more favourable areas, higher species of plants. The combination of greater pedogenic activity, increased moisture availability, higher temperatures and a greater abundance of plant life will generate increased chemical and biological weathering. Although this will not affect bedrock, except perhaps at the base of outcrops, this enhanced weathering will mean more rapid breakdown of glacial debris, scree and other redeposited, non-consolidated materials.

At the continental scale, the rise in temperature combined with increased snowfall effectively moves the present continental-margin conditions inland. The higher incidence of cloud cover associated with an increased frequency of snowfall will produce greater temperature variability at the rock surface (figure 3). The main source of rock moisture is the melting of snow in contact with rock heated by the sun. Now, with increased snowfall and higher overall temperatures there should be a greater amount of melt and so higher rock moisture levels. Thus, freeze-thaw weathering is likely to become a more active and effective process (figure 3). More frequent thermal changes resulting from clouds obscuring the sun may also increase the role of thermal stress fatigue. However, it is the relatively greater presence of rock moisture that is the key factor, for its absence at the present time is the main constraint upon weathering. With more rock moisture so, relative to the present, there will b greater incidence of wetting and drying and of weathering (figure 3). Equally, the potential chemical and biological weathering will also incre as a result of these changes in climatic conditions

Weathering due to chemical, biological, wetti drying and salt crystallization processes are all larg superficial. In fact, it is expected that all but weathering will show only minor and localized creases and, as such, they will have little signific influence on the general scale of weathering. Howev in terms of their singular or combined effect up debris production at any one site the resulting mat ial will be both small in size and overall volun Freeze-thaw may also be primarily a superfic process but if adequate moisture is provided and t rock is highly porous, particularly if in the form microfractures, then the effects may go deeper and tl resulting debris would then be larger. The larg temperature differentials that can occur in Antarctic may allow thermal stress fatigue to produce both larg and small debris in some rock types. The large particles will be generated when the temperatur differentials are sufficient to cause stress at dept within the rock or when the rate of temperatur change facilitates thermal shock and the consequen catastrophic failure of the rock. Overall, the mos important factor is that, in combination, there will be an increase in both weathering processes and weathering rates on continental outcrops of rock.

4. SOILS AND TERRESTRIAL ECOSYSTEMS

In the high polar regions the development of soil is a very lengthy process (figure 4), constrained both by temperature and by the availability of liquid water. Cold soils may be very saline and largely ahumic because of a lack of biological activity (Campbell & Claridge 1987). In lower polar latitudes a much wider range of soils has developed, linked primarily to organism diversity and organic decomposition processes.

Some authors have concluded that the increased temperatures at high latitudes will have different effects at each pole. For the Arctic the climate may get warmer and wetter resulting in the reduction of permafrost and glacial ice as well as the progressive loss of sea ice. In the continental Antarctic increased snowfall would increase continental ice cover, increase iceberg production and intensify the Antarctic Convergence (Houghton *et al.* 1990) although glaciers in the maritime Antarctic would continue to retreat.

In a recent volume which examined the possible effects of greenhouse warming on soils (Scharpenseel *et al.* 1990), there were some useful general reviews of how processes in general might be affected. Many of them are relevant to Arctic soils with their comparatively well developed structure and considerable area. In their examination of boreal and subarctic regions Goryachkin & Targulian (1990) summarized the possible effects of wet warming and dry warming on 16 soil properties in nine soil units. Their general conclusion is that there is no general pattern: the

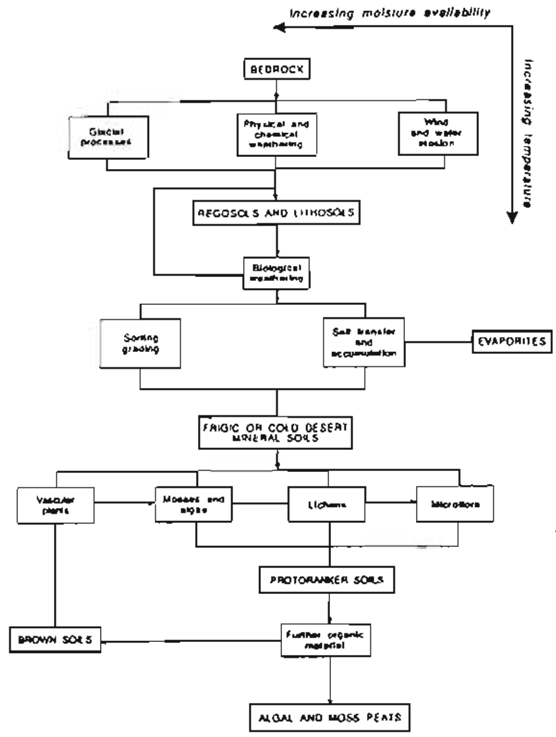

Figure 4. Diagrammatic representation of principal stages in the development of polar soils (after Stonehouse 1989).

different soil properties each have their own characteristic response time and this causes soil properties to change differently in both rate and frequency. They highlight the principal problem of moving to a more quantitative model as a lack of understanding of the links between spatial and temporal change in soil evolution, and a lack of data on time constants for soil processes in each soil type. On present data they conclude that most boreal and subarctic soil will change little in less than 100 years. Peats (Histosols), coarse mineral (Podzols) and calcareous soils (Rendzinas), wet soils (Gleysols, Histosols) and primitive soils (Lithosols, Regosols, Andosols) are especially resistant to change even over a period of 1000 years. The soils expected to change most are those formed from medium textured substrate with developed horizons and active soil processes, none of which occur at present in the Antarctic.

Dregne (1990) comments briefly on the possible effects of climate warming on polar soils in a review of effects on arid soils as a whole. He suggests that warmer temperatures would increase plant growth, soil organic matter and the depth of the active layer in the Arctic. In the Antarctic he predicts a very small increase in soil organic matter. There are, however, likely to be much more rapid changes close to the soil surface with important feedback effects on both vegetation and soils. As with the maritime Antarctic, the overall warming will lead to melting of permafrost and hence to increased cryogenic activity. On the continent the melting will be slow. This is partly due to the severely cold conditions that prevail such that a slight rise in temperature would have a limited effect on the permafrost and it would be some time before this became apparent. A contributing factor to th delaying of permafrost degradation would be th protective effect of an increased snow cover. Howeve as, with time, there will be a deepening of the activ layer so the moisture provided by the melting snov would be extremely important for cryogenic an pedogenic activity. Further, Balke *et al.* (1991) hav shown that chemical weathering currently takes plac in the active zone for about 10 weeks each year i some ice-free areas of the Antarctic continent. With longer summers and additional warming so the potential for chemical weathering and pedogenesis would increase.

Even though most Antarctic soils are largely ahumic there are maritime peat soils and even some small areas under the grass *Deschampsia antarctica* where a simple Brown soil develops. In all these areas and in the ornithogenic soils associated with penguin rookeries nutrient cycling takes place by microbial action. The species involved are not psychrophilic so that increased temperatures could be expected to produce increased breakdown of organic matter at sites where water was not limiting (Wynn-Williams 1990*a*). Thus, in the maritime Antarctic which is likely to have the greatest increases in both temperature and precipitation, there will be a decrease in peat accumulation and an increase in nutrient availability and pedogenic development.

5. COLONIZATION

In a review of the possible effects of climate change on high latitude regions Roots (1989) listed critical questions where knowledge is lacking. Especially pertinent to soils and the terrestrial ecosystems at both poles are the following:

1. What is the relation between changes in high latitude albedo (increase or decrease of net areas of snow and ice) and surface temperature?
2. What is the relationship between average and extreme temperatures, available photosynthetic energy and nutrient supply in limiting biological production?
3. What are the combinations of topography, nutrient supply and microclimate that control biological productivity?
4. What is the relative and absolute role that areas of high productivity play and are these especially sensitive to change?
5. What are the rates and ranges of dispersal, colonization or die-off among key species in polar communities under climate-driven changed environmental conditions?
6. What changes in biological communities or biological succession will signal adaptation to, or disruption by, climate change?

It is not possible to examine the limited evidence for all of these questions in this paper. Attention will be focused on two specific areas: micro-environmental variables, and species diversity and survival.

The change in thermal and moisture status of rock and soil is very important with respect to the distribu-

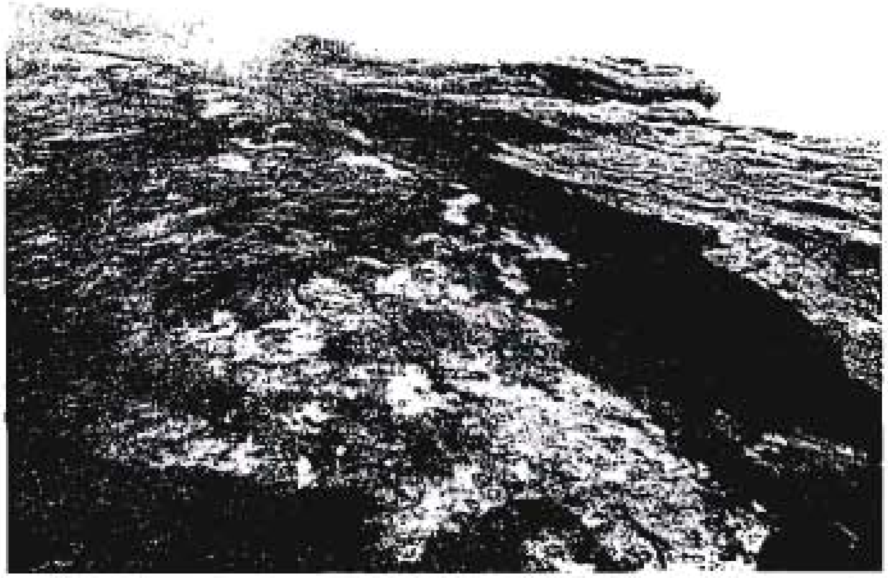

Figure 5. Flora of geothermally heated ground on Deception Island contains many species not found elsewhere in the Antarctic. (Photograph: R. I. Lewis Smith.)

tion of plant species and the overall thermal balance of the ecosystem (Rejmánek 1971). The Antarctic flora shows a clear gradient of climatic tolerance with a decreasing proportion of liverworts and mosses and an increasing proportion of lichens (especially crustose species) as available free water becomes more limited. In the continental Antarctic aspect, topography, wind protection and meltwater drainage patterns are all crucial features defining acceptable niches for all organisms. It is important to remember that under such difficult circumstances organisms may well prefer to colonize the surface or the internal fabric of rocks rather than the surrounding soil, and that as yet we have almost no understanding of the temporal and spatial availability of specific niches to would-be colonisers (Walton 1990).

The evidence that higher temperatures will allow an increased diversity of flora is already manifest in data from geothermal areas. In the Arctic the hot springs in Greenland support species that are either unrecorded elsewhere in Greenland or reach their northern limit in the heated ground (Halliday *et al.* 1974). In the Antarctic both Deception Island and the South Sandwich Islands have specific cryptogamic communities characteristic of heated ground (Longton & Holdgate 1979), with that on the latter islands being considerably richer and more luxuriant than elsewhere (figure 5). Of the plant species recorded from the South Sandwich Islands six of the 13 algae are restricted to fumaroles, as are one of 27 lichens, eight of 30 mosses and seven of 12 hepatics. Many of the species found on unheated ground were also found growing much more luxuriantly in the fumarole communities.

Collins (1969) recorded the presence of a *Funaria* cf. *hygrometrica*, previously unknown from the Antarctic, in a new fumarole on Deception Island. A more detailed survey of heated ground on the island (Lewis Smith 1984) showed that *Leptobryum* cf. *pyriforme*, *Marchantia berteroana* and *Philonotis gourdonii* are all found only in this habitat within the Antarctic although the first two species are widespread in more temperate regions.

There are only two native phanerogams in the Antarctic: *Deschampsia antarctica* and *Colobanthus quitensis*. Although the species flower in most sites every summer the production of ripe seed occurred in only three of the nine seasons assessed (Edwards 1974). Experiments with cloches demonstrated that low temperatures limiting seed development were the cause of this. Climate warming should result in the spread of both these species.

There have also been various attempts to test the limits of survival of species introduced to the Antarctic. In the most extensive trials phanerogamic species from the Falkland Islands (Edwards & Greene 1973) and South Georgia (Edwards 1979) were transplanted to Factory Cove, Signy Island between 1967 and 1973. Out of 23 Falkland species 11 survived for over two and a half years on Signy but only two produced new flowers. The South Georgian species did better. Of 23 species 14 survived for at least one year and eight species produced new flowers some, such as *Phleum alpinum*, every year for 4 years. Seedlings of eight species became naturally established during the summer months but mortality of the seedlings was high during the winter. In general the species most capable of survival were the graminoids and some of the persistent alien species. On the evidence available so far the most likely new colonizer in a warming environment is the almost cosmopolitan alien grass *Poa annua*.

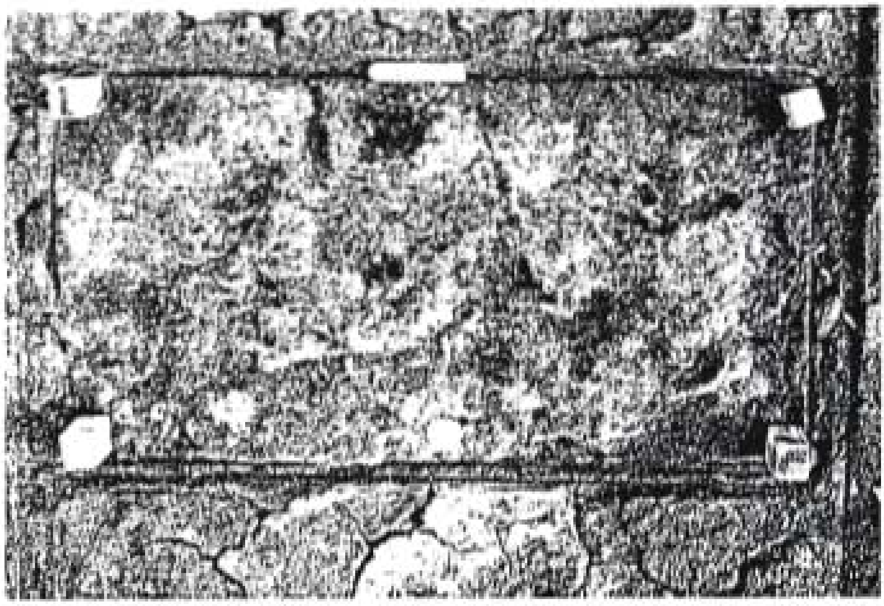

Figure 6. Manipulation of the microclimate with plastic cloches illustrates the potential for colonisation of bare ground by species already present in the soil propagule bank. The 100% cover by mosses and algae shown here has developed under the cloche in three years on the bare sorted centre of a polygon at Signy Island. (Photograph: R. I. Lewis Smith.)

These observations suggest that, if present environmental conditions ameliorate, there will be an increase both in species diversity and in annual growth. The diversity might arise either from an increased opportunity for propagules already present in a soil bank to germinate and grow, or by increasing the probability of incoming propagules finding an acceptable niche in which to establish (Walton 1990). Two approaches have begun to test this.

Lewis Smith (1987) has proved the existence of a substantial diaspore bank in recently deglaciated soils on Signy Island from which a wide range of species can be cultured under laboratory conditions. The next stage was to manipulate the environment to see if these diaspores would establish naturally. Plastic cloches were placed on the unvegetated centres of sorted polygons to increase temperature and humidity, decrease wind exposure and lengthen the growing season (Wynn-Williams 1990*b*). Dramatic changes were evident when the cloches were compared after three years with nearby uncovered controls (table 1).

Table 1. *The response of Antarctic soil algae after three years of improved environmental conditions under a plastic cloche (from Wynn-Williams 1990b)*

	control	cloche	% change
temperature/°C[a]	4.4 ± 1.2	7.6 ± 1.7	+ 73
area colonized (%)	4.8 ± 1.5	73.9 ± 8.0	+ 1440
total length of cells[b]			
(μm 10^{-3} mm^{-2})	3.9 ± 2.0	17.9 ± 7.4	+ 358
cell length/μm	25.3 ± 1.7	78.0 ± 45.2	+ 208
cell breadth/μm	10.2 ± 1.6	7.1 ± 0.9	− 30
cell volume/$μm^3$	1641 ± 479	4719 ± 3744	+ 188

[a] Integrated surface temperature January–March 1988.
[b] All cellular data obtained by television image analysis.

Almost three quarters of the soil surface was now covered with algae under the cloche but only 5% on the control site. The cloches provided a 3.2°C increase in mean summer temperature as well as a continuously humid environment. There are other effects which may also have contributed significantly to the change. The cloches lengthen the growing season by excluding snow, by increasing thermal inertia they reduce the frequency of freeze–thaw cycles and they screen out a high proportion of UV-B. A new experiment has now begun using multiple cloches of various designs to test the interactive effects of each of these environmental variables (D. D. Wynn-Williams, personal communication). Thus the potential is already present at this maritime site to develop more extensive vegetation cover under a warmer and wetter régime. Similar experiments with cloches are now being undertaken at a range of other Antarctic sites.

Two ecological questions remain to be asked. First, how will the existing communities change under a warmer and possibly wetter environment? It is clear from the cloche experiments that changing the environment changes the competitive interactions between species in communities since the pioneer communities under the cloches differ from those found at present on the island. As yet there are no data from cloche manipulations of established communities.

Second, with an increasing snow free area available for colonization will there be a major increase in species diversity due to the establishment of pioneer species from outside the Antarctic? At present little is known about the potential offered by the air flora for new colonizing species. The palynological record in peat banks show clear evidence of the arrival of exotic tree pollens in the subantarctic (Barrow & Smith 1983) and the maritime Antarctic (Kappen & Straka 1989) but there is almost no data on the diversity, likely origin and potential viability of any other aerobiological particles. The new SCAR BIOTAS (Biological Investigation of Terrestrial Antarctic Systems) programme is therefore attempting to address propagule input to both the islands and the continent by a coordinated international effort (Wynn-williams 1992).

The most notable attempts to bring together existing terrestrial data to characterize climatic change have almost all been based on soils in the Dry Valleys or on the ecosystems on Signy Island, South Orkney Islands. Campbell & Claridge (1987) have described climatic changes in Victoria Land over the past 5 million years based on their characterization of patterns and rates of soil development. The key environmental feature of this is the continued aridity of the area, which has effectively limited development of both physical structure and chemical content.

As long ago as the early 1970s Collins (1976) recognized that short-term fluctuations in climate could be identified from vegetation patterns on Signy Island: 'trim lines' alongside glaciers, re-exposure of subglacial peat banks, etc. Lewis Smith (1990) took this approach much further and by indicating the possible changes in ice extent over the past 7000 years has shown how this area at least responded to warmer temperatures in the past.

Future change in the mass of the Antarctic ice sheet will be slow in coming; either from increased precipitation or increased melting. What does seem clear is that at the continental margins and on the offshore islands change is already happening (Fenton 1982) and will continue to gather pace, illustrating the patterns which almost certainly provided the basis of change at the end of the last glacial maximum in temperate regions. In the timescale of 100 years it seems certain that any changes in weathering rates, soil genesis, colonization and community change will all be slow, very localized and unlikely to have any important regional impact or feedback.

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Discussion

D. D. Wynn-Williams (*British Antarctic Survey, Cambridge, U.K.*). Professor Hall mentioned that a warmer maritime Antarctic climate would result in more precipitation. During rock weathering, moisture would penetrate into the fabric of the rock and disrupt it by ice crystal expansion during freeze–thaw cycles. Microbes would penetrate the rock in the water film. At Signy Island, I have observed that most soil and rock microbial colonizers, both phototrophic algae and cyanobacteria, and heterotrophic bacteria, have substantial mucilagenous sheaths or capsules. These expand and contract greatly during wetting and drying and may therefore disrupt the rock structure even further by hydrostatic effects. Does he have any direct evidence of this?

K. J. Hall. Although I have no direct evidence from the maritime Antarctic, work undertaken in Alaska (Hall & Otte 1990) clearly showed that the expansion and contraction of the mucilagenous sheaths of endolithic algae caused extensive weathering of granitic rocks. The expansion was as a result of water absorption during times of precipitation while contraction took place during heating of the rock by solar radiation. Thus, with increased precipitation, particularly in the form of rain, so it could well be expected that this form of biological weathering would become more prevalent in the Antarctic.

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T. Callaghan (*Institute of Terrestrial Ecology, Merlewood Research Station, Cumbria, U.K.*). The measurement of Q_{10} of soils (the rate at which CO_2 is evolved in response to temperature increases) increased in organic soils in a transect from the sub-arctic to the High Arctic. In a lithosol from a High Arctic semi-desert, the Q_{10} was extremely high and could not be accounted for by microbial activity alone. This would suggest a significant evolution of CO_2 due to purely chemical processes.

D. J. Drewry (*British Antarctic Survey, Cambridge, U.K.*). In relation to weathering are there any data or experimental results which can be used to give the sensitivity of weathering processes to climate change: for instance mass loss per unit time per degree C.

K. J. Hall. Unfortunately the relationship of weathering to climate is so complex that it is impossible to provide any simple correlation regarding it, such as mass loss per unit time per degree C, particularly as some processes accelerate as others decrease. Further, at the moment our data base regarding process types and process rates is so small it would be impossible to make any judgements.

W. C. Block (*British Antarctic Survey, Cambridge, U.K.*). We may be slightly misled if we consider only plant or vegetation development as being affected by climate change in Antarctica. These communities have few invertebrate species and a simple structure where food chains are short and competition often very much reduced. Such simple communities may well be more sensitive to environmental changes and their responses faster than plants such as mosses and lichens. Secondly, there are likely to be more subtle changes in, for example, invertebrate physiology than the possible gross, large-scale changes seen in plants with climatic change. An example of this are the results of a long-term study of body water content in a typical Antarctic insect (the collembolan *Cryptopygus Antarcticus*). These data show a marked and significant upward trend in water content with increased water availability in their habitats over several years. This suggests that such species may have been living with body water levels below the optimum physiological level in such arid habitats. Such organisms may be more sensitive indicators of environmental change than has been thought hitherto.

1

Weathering Rinds as palaeoenvironmental indicators: Evidence from the Cape Roberts drill core (CRP3), Ross Sea, Antarctica.

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Abstract

XRD analysis of weathering rinds on ocean-deposited ice rafted clasts provide a snapshot of palaeo-terrestrial conditions prevailing at the time of rind formation. Investigation of weathering rinds found on some CRP3 clasts indicate environmental conditions conducive to the formation of expanding clays, mostly smectites. Although sampling for clasts with weathering rinds was limited, those obtained for the 272 to 300 and 480 to 561 mbsf (metres below sea level) regions show rinds commensurate with relatively wet and warm terrestrial conditions that would have been conducive to enhanced chemical weathering. The occurrence of clasts with weathering rinds within those parts of the core substantiate the palaeoenvironmental reconstructions of other investigators in the Cape Roberts Project.

Introduction

A pilot study undertaken on the CIROS-1 drillcore examined the feasibility of using X-ray diffraction analysis (XRD) of weathering rinds on ice rafted debris (IRD) as a proxy for former terrestrial environmental conditions (Hall and Bühmann, 1989). Rind analysis is based upon the premise that the clay mineralogy/chemistry of the weathering rind is developed under the terrestrial weathering regime and then preserved during subsequent glacial transport and off-shore deposition, it provides a proxy "snapshot" of environmental conditions during the time of rind development. Results from the initial study (Hall and Bühmann, 1989) were able to demonstrate the viability of this approach with, for CIROS-1, the identification of cold, wet subaerial periods with podsolized soils forming under forest or scrubland. An adjunct to the weathering rind analysis used on CIROS-1, but not undertaken on CRP-3, was that of recording clast weathering fluctuations through the core coupled with how those varied as a function of clast lithology. Based upon both the degree of clast weathering and weathering rind analysis, the CIROS-1 data were able to show that rind alteration did not occur after deposition, and the same assumption is used in the interpretation of the CRP-3 rinds.

Analysis of IRD weathering rinds, where no post-depositional change has taken place, should have every expectation of offering indications of former terrestrial environments. What is unknown is time: time *for* formation (a reflection of the degree of warmth and/or moisture) or the actual time *of* formation (as the IRD may be subject to a substantial transport period). Despite these limitations, conditions are otherwise relatively "ideal". Clast transport by glaciers, largely in a subsurface situation when clasts are passively received by the glacier within the accumulation zone, means that ambient temperature and moisture conditions severely limited any alteration or rind formation - chemical weathering being extremely limited if at all active in the englacial situation. Thus, the character of the weathering rind must reflect the nature of the terrestrial environment *prior* to its incorporation within the glacier that transported it to its oceanic depositional situation.

As noted above, there was continuous monitoring of clast weathering throughout the CIROS-1 core coupled with the variability of this as a function of lithologic changes. No such monitoring was undertaken during Cape Roberts Project coring (CRP cores) and thus the information available from this study is an "after the event" investigation with none of the adjunct data to help create a palaeoenvironmental "picture". Rather, the information is based on random samples made available for analysis and this obviously limits interpretation and full application of this approach. Visual appraisal of CRP-2/2A core photos (e.g. fig. 3.2 , especially photo (*i*), Cape Roberts Science Team, 1999, p. 63) suggests that weathered clasts are not uncommon in the drill cores but, in the absence of rind analysis, it is not possible to interpret their value/meaning. However, even the apparent rinds observed would be consistent with a palaeoenvironmental reconstruction comprising (p. 66) "...cyclopels from highly sediment-charged glacial streams.." (Cape Roberts Science Team, 1999, p.66) necessitating warmer, melting conditions conducive to smectite production as found in the CIROS-1 investigation (Hall and Bühmann, 1989). Detailed information, such as "Lithology, Facets and Surface Features" for CRP-3, as were undertaken in CRP-2/2A, will, ultimately, provide adjunct information that may help tease-out more detail regarding the significance of the rind information found here.

Clay mineralogy of the fine fraction from the cores was undertaken in CRP-1 and CRP-2/2A (Ehrmann, 1998;Cape Roberts Science Team, 1999, p94) and CRP-3 (Marinoni and Setti, this volume). However, the relationship between those studies and the detail that would be available from the weathered clasts remains unclear, partly because the interpretation of the fine fraction mineralogy lacks any obvious terrestrial point of reference to help validate the explanation. The smectite may be due to a period of increased terrestrial chemical weathering or it may reflect a change in source area for the sediments (and hence the chemistry). Although smectite was recognised in CRP-2/2A (Cape Roberts Science Team, 1999, p. 95) and identified as being representative of "...hydrolysis under climatic conditions between warm-humid and cold-dry, in environments characterized by very slow movement of water.." the interpretation of the results (Cape Roberts Science Team, 1999, p. 96) lacks consideration of the *in situ* weathering rinds on clasts within the core and the information that it could provide regarding terrestrial conditions. This argument can be justified by consideration of the core matrix clay mineralogy (Cape Roberts Science Team, 1999) that states smectite concentrations in CIROS-1 at *c*.290-320 mbsf (unit 12) may be a result of several causes, one of which may be "..warmer and more humid conditions...resulting in more intense chemical weathering..". This is exactly the region (unit 12) where Hall and Bühmann (1989) recorded smectite concentrations in the weathering rinds of IRD (also for unit 22 where the fine fraction analysis also showed a transition from smectite-rich to illite-rich conditions). Unfortunately, the evaluation of the smectite in the core fine fraction of CIROS-1 reported in the revue by Cape Roberts Science Team (1999, p. 96) takes no cognizance of the findings by Hall and Bühmann (1989), which argued for the warmer, moister conditions, and how those data may have helped resolve the clay mineralogy for the fine fraction. Indeed, in the earlier clay mineral analysis of the CRP-1 core (Ehrmann, 1998, p. 617), it is argued that although smectite is often a result of chemical weathering under warm, more humid conditions, because of the "... evidence of ice being present on the nearby Antarctic continent throughout the time represented by the core, chemical weathering on land is not a likely source of the high smectite content". This does not have to be the case. The presence of ice is not an inhibitor (unless of course all the rock is covered) for, as is the case in the present-day Alaska pan handle, there is a substantial ice cover coupled with extensive forest growth and the *in situ* development of clay minerals. Rather, as Balke *et al.* (1991) have shown for the Antarctic, the main limitation on chemical weathering is not temperature but rather the availability of moisture. As the units for which smectite is identified appear (from the core interpretations) to be ones with enhanced moisture availability, there is no reason why clay minerals could not be produced, given sufficient time, in weathering rinds on rock exposures (later to be ground down by glacial action to help provide the fines). Certainly research has shown that given moisture then Antarctic summer *rock* temperatures (i.e. as opposed to air temperatures)are conducive to chemical weathering (see Meiklejohn and Hall, 1997 for a brief review). Thus, the principle that weathering rinds on IRD surviving glacial transport and deposition in off-shore sediments can, particularly in conjunction with the clay mineral analysis of the fine fraction, offers a valuable insight into palaeo-terrestrial conditions.

In summary, there appears to be justification for the analysis and interpretation of weathering rinds and that these data, if integrated with those for the clay mineralogy of the fine fraction, would be a valuable adjunct in the understanding of former terrestrial conditions. Thus, with this premise the preliminary findings are reported here. Once the clay mineralogy for the core and the palaeoenvironmental reconstructions based on the sedimentological studies are available, the rind information may be refined.

Site

Cape Roberts Project drill core 3 (CRP-3) was obtained from a sea-floor high about 12 km off-shore from Cape Roberts at 77.0106°S, 163.6404°E in western McMurdo Sound. The core was drilled to a depth of 939 m and had a 98 - 100%

recovery. Details of the core division and reconstructed depositional environments are provided by Powell, *et al.*, (this volume).

Technique

Clasts that appeared to show signs of weathering to be used in XRD were collected from the core. Samples of rock were removed from the centre of each clast (here considered to be "unweathered") and from the edges where the rinds occurred (the "weathered" component) using a 0.5 mm drill bit operated at 20,000 rpm. Each sample was then ground to <50 μm in diameter for x-ray analysis. X-ray diffraction analysis was conducted on the samples mounted on glass-slides using acetone. The powder mount was scanned from 2 to 65 °2Θ at ambient conditions at a step size of 0.02 °2Θ and a collection period of 5 - 10 seconds per step. Samples were also solvated with glycerol to determine the presence of expanding clay minerals. Identification of the minerals in the samples was based on the following criteria: (1) quartz - 0.425, 0.334, 0.245, 0.228 and 0.182 nm reflections, (2) feldspars - 0.635, 0.404-0.420, 0.315-0.325 nm reflections, (3) amphiboles - 0.826, 0.324, 0.304 and 0.270 nm reflections. Identification of clay minerals was based on the following: Kaolinite and chlorite were recognised by the reflections at 0.71 nm and 1.4 nm respectively (Barnhisel and Bertsch, 1989) while mica was identified from the 1.0 nm reflection. Expanding 2:1 type of clays were recognised by the reflection at around 1.7 nm in glycerol-solvated samples.

Results and Discussion

Figure 1 shows the preliminary results. In Figure 1, clasts are identified by the depths (mbsf) at which they were collected. Note, of the samples available not all show weathering rinds. In terms of mineral composition of the core, clasts could be classified into two groups of samples (Fig. 1):

Group 1 = clasts at depths 272.11-272.15 (our sample #2), 290.21-290.27 (#3), 297.79-297.83 (#4), 300.0-300.05 (#5), 480.69-480.74 (#11), 537.51-537.54 (#13), and 561.75-561.80 (#14).

Group 2 = clasts at depths 305.71-305.72 (#6), 306.55-306.57 (#7) and 478.68-478.70 (#9).

In Group 1 the samples from the centre of the clasts contain a high amount of weatherable minerals such as feldspars and amphiboles as well as quartz and clay minerals (Figs 1A, C, E, F, G, H and J). In clast no. 5 (Fig. 1G), albite (Na-rich) is estimated to be > 90% using the separation of –132 and 131 x-ray reflections (Smith 1956 in Huang, 1986). Unweathered clasts in Group 1 show the presence of 1.2 nm reflection indicating the occurrence of 2:1 expanding clays in the presence of micacaeous minerals. The x-ray reflection for expanding clays is dependent upon cation saturation, organic solvent solvation and relative humidity such that x-ray reflections could vary 1.2 to >1.5 nm (Douglas, 1986). The samples shown in Fig. 2 are not saturated with specific cation but are solvated only with glycerol. Thus, depending on which species is present (e.g. montmorillonite, beidellite or nontronite), smectite expands at glycerol solvation from 1.5 to 1.8 nm. Vermiculite normally does not expand beyond 1.4 nm upon glycerol solvation but does expand to 1.6 nm upon ethylene glycol solvation. Also, in terms of surface charge, vermiculite has a higher charge (>0.65 charge per formula unit) than smectite (<0.65 charge per formula unit).

Figure 1 shows the minimal differences between the unweathered core and the weathered rind. For example, Fig. 1E shows similar xrd patterns between the unweathered core and the weathered margin of clast no. 4. However, after glycerol solvation, the presence of higher amounts of 2:1 expanding clays (probably vermiculite and smectite) in weathered rind is evident from the shift of the diffraction peak from 1.2 nm to 1.42 and even 1.85 nm regions (Fig. 2A). Clast no. 5 (300.00-300-05 mbsf) also shows XRD patterns very similar to clast no. 4. In Fig. 1A, untreated clast no. 2 shows similar XRD patterns to untreated clast no. 4 (Fig. 1E) indicating similarity in mineralogy. However, Fig. 2B does not have the reflections at higher d-spacings at 1.42 and 1.85 nm that are present in clast no. 4 probably indicative of the differences in weathering environments of formation.

Generally, smectite does not exist in sediments buried deeper than 4 km because Mg, Fe, Al and/or Mg are incorporated in smectite to form mica or chlorite (Borchardt 1986). In terrestrial environment, one of the general pathways related to the origin of 2:1 expanding clays is the removal of potassium from mica or Mg from chlorite in leaching (or humid) environment (Borchardt, 1986; Douglas, 1986). Because our samples were collected from sediments shallower than 4 km, it would appear that the presence of vermiculite and smectite in the clasts of Group 1, were probably formed as a result of terrestrial weathering processes in moist (or humid) environment. The lower intensity of the x-ray reflections at 1.42 and 1.85 nm for some clasts in Group 1 may indicate source areas of less humid environment. In other words, water is less available to remove K from mica or chlorite. Caution must be observed in interpreting the relationships of smectite and moisture conditions because, under restricted environment, smectites may precipitate from soil solution in almost any parent material (Borchardt 1986).

Group 2 clasts have their centres dominated by quartz (likely igneous (primary) derivation as these clasts are from tholeiitic dolerites). Amphiboles and other weatherable minerals are not apparent from x-ray diffractograms; if present, they are in minor quantities. Generally, where weathering rinds are present in this group, no expanding clays (e.g. smectite) are apparent. This could be due to low amounts of feldspars, amphiboles and possibly mica or chlorite as well as the moisture conditions. Removal of potassium from mica and Mg from chlorite are major sources of smectite and vermiculite formation in soils. Thus, those clasts in Group 2 may have originated from areas that are too dry to limit the neoformation of smectite or too wet where the removal of Mg is so intense preventing the formation of smectite (Borchardt 1986).

Although the clast sampling for weathering rind analysis lacks any systematic basis and is not tied directly to other observations (e.g. lithological changes or striated clasts) it can, nevertheless, offer some ancillary information in support of other hypotheses. For example, Atkins (this volume) suggests that clast features and fabrics, in the upper 330 mbsf of the core are consistent with the presence of nearby ice experiencing repeated advance and retreat. Powell *et al.* (this volume) also suggested that the upper 400 mbsf represent shallow, marine, glacially-influenced sediments deposited during several glacial fluctuations. The lithofacies interpretations indicate that a substantial amount of meltwater was present and this, taken with the high rates of sediment discharge, suggest the presence of a warmer climate than at present (Powell *et al.*, this volume). Raine and Askin (this volume), from palynological information, suggest the presence of *Nothofagus* and podocarpus conifers and, at some sites, there may have been low scrub or closed forest and even possibly wetland vegetation. Taken together, this information suggests an environment conducive to chemical weathering in which smectite weathering rinds could occur. The limiting factor for chemical weathering in the Antarctic environment is water (Balke *et al.,* 1991), but the conditions, glacial outwash and possible wetland vegetation, coupled with warmer conditions that would give a longer-than-present summer weathering period, would have been conducive to bedrock/clast weathering and smectite production. Thus, the occurrence of clasts with weathering rinds and the presence of smectite at 272-300 mbsf depths would be in accord with a more conducive chemical weathering environment. The corollary to all this is that it would have been surprising *not* to have found weathered clasts at these depths and so their presence supports such a palaeo-environmental interpretation.

The finding of weathering rinds on clasts at depths of *c*. 480 and 537 to 561 mbsf also suggests conditions suitable to chemical weathering in the terrestrial environment. Analysis of depositional environments for these depths (Powell *et al.*,this volume) suggests deltaic conditions with, possibly, all valleys not being completely ice-filled. It could be envisaged that chemical weathering was viable within valleys having ice-free rock exposures and meltwater available. Palynological evidence for substantial vegetation is lacking although it was conceded that some may have been present (Raine and Askin,this volume). Thus, some localized weathering of rock, constrained by water availability, appears possible.

Conclusion

As a technique, our work on CRP-3 samples further demonstrates the validity of using XRD examination of weathering rinds on IRD as a proxy for palaeo-terrestrial environmental conditions. However, the limited availability of clasts from CRP-3 coupled with the absence of weathering rind evaluation for the whole core limits the applicability with respect to this study. From the available information, it appears that environmental conditions suitable for the development of 2:1 expanding clays, particularly smectites, occurred in some units (e.g. unit 7.5) and, at that time, conditions may have been relatively mild and wet. Recognizing that the limiting condition for chemical weathering on the Antarctic continent is largely moisture (rather than temperature), the presence of such weathering rinds is indicative of warmer, and hence moister, conditions. Had clasts with weathering rinds *not* been found during the reconstructed warmer and wetter times then this would have been worrisome given the palaeoenvironmental interpretation of other CRP-3 investigations. The presence of weathering rinds helps substantiate those palaeoenvironmental reconstructions.

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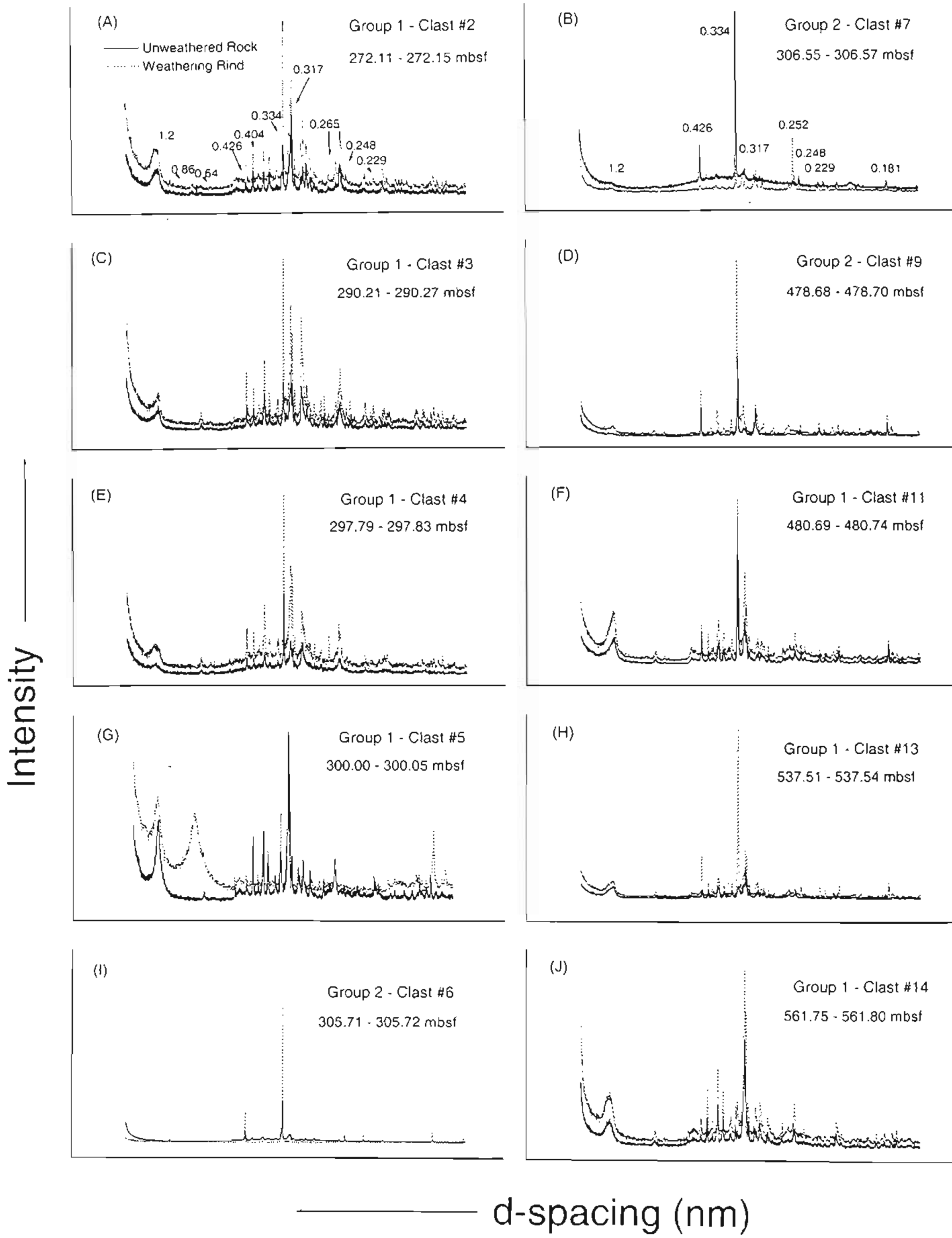
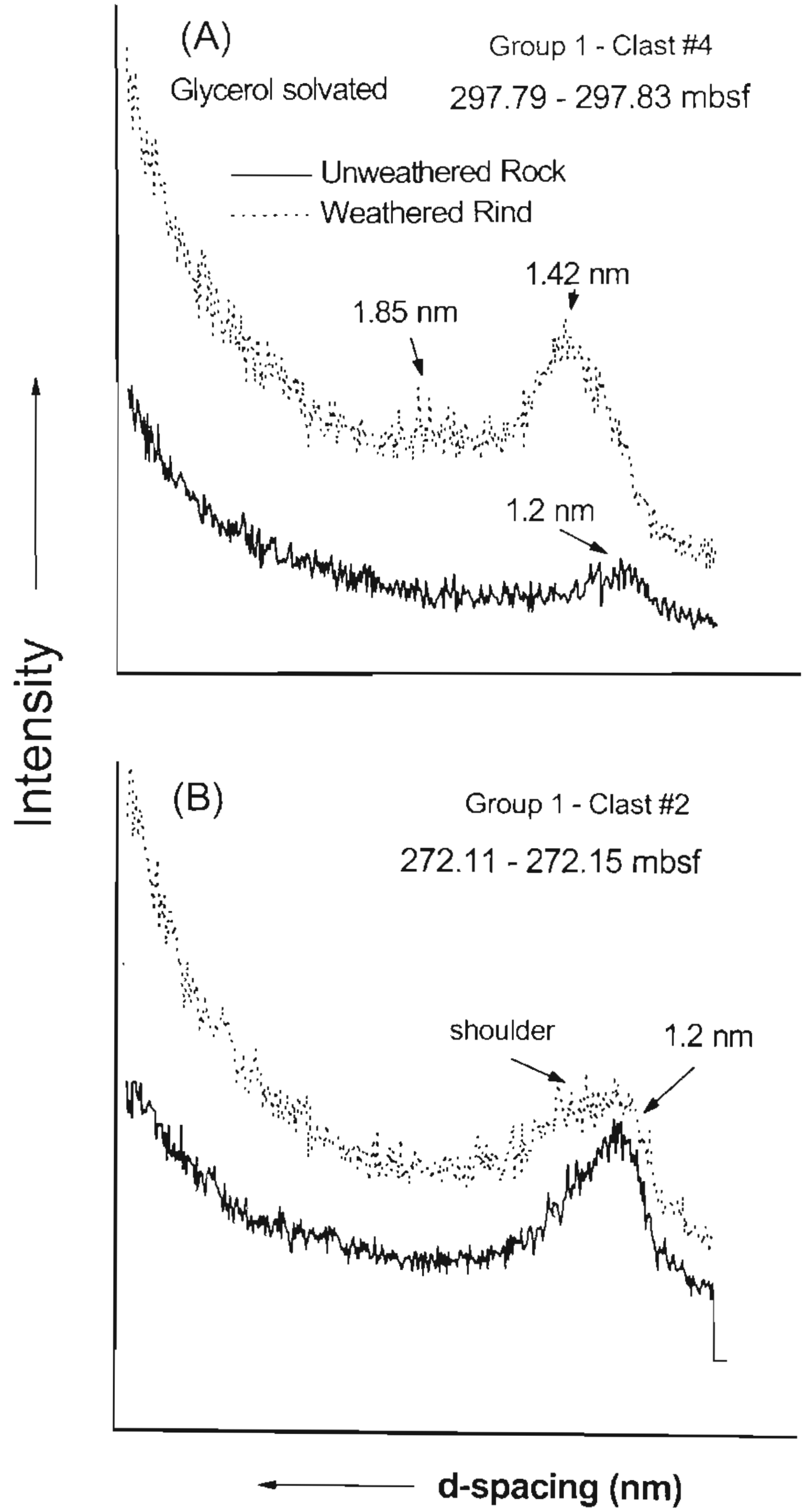
Figure Captions

Fig. 1. Diffractograms for the 10 clasts analyzed using x-ray diffraction. Group 1 – Clast Nos. 2 (A), 3 (C), 4 (E), 5 (G), 11 (F), 13 (H), 14 (J); Group 2 – Clast Nos. 7 (B), 9 (D), 6 (I).

Fig. 2. XRD diffractograms of glycerol solvated samples of (A) Group 1 – Clast no. 4 and (B) Group 1 – Clast no. 2. XRD reflections at 1.42 to 1.85 nm show the presence of 2:1 expanding clays in the weathering rind of clast no. 4.


(A)
Group 1 - Clast #2
272.11 - 272.15 mbsf
Unweathered Rock
Weathering Rind
1.2
0.86
0.54
0.426
0.404
0.334
0.317
0.265
0.248
0.229
(B)
Group 2 - Clast #7
306.55 - 306.57 mbsf
0.334
0.426
0.317
0.252
0.248
0.229
0.181
1.2
(C)
Group 1 - Clast #3
290.21 - 290.27 mbsf
(D)
Group 2 - Clast #9
478.68 - 478.70 mbsf
(E)
Group 1 - Clast #4
297.79 - 297.83 mbsf
(F)
Group 1 - Clast #11
480.69 - 480.74 mbsf
(G)
Group 1 - Clast #5
300.00 - 300.05 mbsf
(H)
Group 1 - Clast #13
537.51 - 537.54 mbsf
(I)
Group 2 - Clast #6
305.71 - 305.72 mbsf
(J)
Group 1 - Clast #14
561.75 - 561.80 mbsf
Intensity
d-spacing (nm)


(A)
Glycerol solvated
Group 1 - Clast #4
297.79 - 297.83 mbsf
Unweathered Rock
Weathered Rind
1.42 nm
1.85 nm
1.2 nm
Intensity
(B)
Group 1 - Clast #2
272.11 - 272.15 mbsf
shoulder
1.2 nm
d-spacing (nm)

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Present-day Periglacial Processes and Landforms in Mountain Areas*

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ABSTRACT

This paper outlines recent advances in our understanding of periglacial processes and landforms in mountain areas, as reported in the literature over the last 8 years. Landforms investigated include rockwalls, screes, avalanche paths, bedded slope deposits and patterned ground phenomena. Processes subject to quantification include cliff recession, rock displacements on slopes, rock thermal and moisture regimes, frost creep, mass wasting and solution. Laboratory experimentation and modelling studies have also been undertaken.

RÉSUMÉ

Le présent article souligne les progrès récents de la connaissance des processus périglaciaires et des formes de terrain dans les régions de montagne ainsi qu'ils apparaissent dans les publications des 8 dernières années. Les formes de terrain étudiées comprennent les abrupts rocheux, les éboulis, les couloirs d'avalanche, les dépôts de pente stratifiés et les sols structuraux. Les processus dont on mesure la vitesse sont le recul des abrupts, les mouvements des roches sur les pentes, les régimes thermiques et d'humidité des roches, le creep dû au gel, les mouvements de masse et l'érosion par mise en solution. Des expériences en laboratoire et des modélisations ont aussi été entreprises.

KEY WORDS: Mountain landforms   Periglacial   Measurement   Experimentation modelling

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INTRODUCTION

This paper summarizes present-day periglacial processes occurring in mountain regions as reported in the literature over the last 8 years. The focus is on slope processes; aeolian and fluvial processes are not considered, while rock glaciers are discussed elsewhere in this volume. A number of periglacial phenomena such as ploughing blocks and icings are not reviewed, because no significant advances have been made with respect to these topics.

ZONATION

Following the noteworthy study of Gorbunov (1969) regarding the Tien Shan, a number of recent studies bring new information with respect to periglacial zonation in mountains. Notable is the synthesis of S. A. Harris (1988), which established a relationship between various phenomena and the type of permafrost. Zonations for the Rockies and Iceland have been produced. A similar study has been published by Sone (1990) for the mountains of Diasetsu, Japan, while in China, Cui and Zhu (1989) have reviewed and completed their earlier work. They, and Gorbunov (1990), place emphasis on latitudinal zonation and the development of the periglacial regime in a dry, snow-free, continental environment which allows intense freezing penetration with, in particular, the development of cryoplanation features. This is also the opinion of Corte (1986, 1988), and of Grosso and Corté (1991), who draw attention to the operation of cryoplanation in association with permafrost at high altitudes (5000 m a.s.l.) in the Andes. Priesnitz (1983) has also described periglacial pediplanation in the Canadian Cordillera.

For the mountains of Mexico, Heine (1983, 1989) proposes a present-day and a Holocene-Pleistocene zonation for the last 35 000 years. Eckelman (1988) has defined the lower boundary of the periglacial belt in the North American Cordillera from Mexico to Alaska, and finds that it corresponds primarily to the position of the 8 °C July isotherm but with a secondary influence of humidity (continentality). Young (1989), however, considers that studies are not advanced enough in the tropical mountain environment (in particular, with respect to those processes which are truly periglacial) to allow a realistic zonation. In an extremely continental arid environment (western Kun Lun, China), the 0 °C boundary was found at 4000 m a.s.l. and so the periglacial zone was inferred to occur there between 4500 and 6000 m a.s.l. (Francou *et al.*, 1990): gelifraction and debris production, patterned ground and, especially, aeolian phenomena (sands and silts) are typical because of the extreme dryness of the winter (Li Shude, 1987; Iwata and Zheng, 1989). Other studies in Qinghai Province (e.g. Wang Jintai, 1987; Wang Shaoling, 1987; Fort and Dollfus, 1989) fix the position of the periglacial zone with respect to temperature and humidity.

Winter aridity limits periglacial phenomena in the mountains of South Africa above 3000 m a.s.l. in the Drakensberg: needle ice, thufa, solifluction features, micro-patterned ground and turf- or stone-banked steps occur (Boelhouwers, 1991a). In the western, more humid, mountains of the Cape other processes occur, such as gelifraction and gelifluction, and stone-banked lobes are typical (Boelhouwers, 1991b).

In New Zealand a list and zonation of processes has been recently established by Soons and Price (1990).

A number of recent publications and theses provide new information from the Andes (Garleff and Stingl, 1983, 1985; Barsch and Happoldt, 1984; Graf, 1986; Trombotto, 1988; Wayne, 1989; Schrott, 1991), from the mountains of Scandinavia (Lindh *et al.*, 1988; Odegard *et al.*, 1988), the Alps (Kaiser, 1987; Rovera, 1990; Veit, 1988) and the Polish Tatras (Kotarba, 1987). In the volume resulting from the Göttingen Symposium (Poser and Schunke, 1983) many papers describe the zonation of frost phenomena and processes in the polar, Scandinavian, European, Mediterranean, African and American mountains (e.g. see papers by Barsch, Schunke, Hollerman, Klaer, Stablein, Jahn, Sting *et al.*, Brosche, Rudberg, Hagedorn, Sponeman and Heine).

According to S. Harris (1988), the zone of periglacial phenomena in association with continuous permafrost is less varied and less common in the alpine zone when compared with the Arctic. This is because topography and a reduced formation of fines, both thought necessary to promote good drainage of water, result in less ubiquitous soil moisture.

MECHANISMS

Mechanical Weathering of Rockwalls

The rockfall process has been studied in detail (alteration, transit, deposition) by Francou (1988) in the central Alps. Although caused by a number of processes, rockfalls are especially frequent in association with water that melts in fissures sealed by ice at high altitudes (above 2600 m a.s.l.). According to Rovera (1990), at Tarentaise, further to the north at a lower altitude, quartzite screes are supplied, especially during spring, by freeze–thaw action. According to Paneza and Ozouf (1988), the detachment of debris takes place in winter in the Swiss Jura.

Screes and Avalanches

A number of authors have studied the complex phenomena of scree formation. An in-depth treatment is by Francou (1988, 1990). He distinguishes a proximal scree zone with a steep gradient (33–41°) associated with accumulation and transport together with a distal scree zone of accumulation only; the two are separated by a dynamic break (a biphase model). A typology of the talus-type forms has been defined together with its associated terminology (Francou and Hétu, 1989; Francou and Manté, 1990). More complex cases with segmented profiles are also observed. A type of scree containing ice has been described in a synthesis by Barsch (1983), and its creep movement studied by Francou (1988). In fact, many so-called 'protalus ramparts' probably have an origin due to this form of movement ('talus rock glaciers') (Haeberli, 1985; Francou, 1988). On the other hand, Ono and Watanabe (1986) have described a protalus rampart in the northern Japanese Alps that is associated with debris flows and landslides. The association of scree formation with avalanches has also been discussed by Luckman (1988) in the Western Cordillera and by André (1989, 1990) for Spitsbergen.

Bedded Slope Deposits

In the Andes, Francou (1988, 1990) suggests that stratified slope deposits are associated with stone-banked sheets moving mainly by frost creep. The loosely consolidated coarse material is then buried by the fine surficial material of the advancing sheet. In addition Van Vliet-Lanoë (1989) emphasizes the downward migration of fine particles as a result of the combination of alternating freezing and thawing together with frost creep. On the other hand, Hétu and Vandelac (1989) have proposed an alternative slope and flow model of stratogenesis in slope deposits for Gaspésie, Canada. Coltorti and Dramis (1987) have also examined stratified deposits in the Apennines with the possible influence of stabilization phases as indicated by palaeosols. Some structures are similar to those of debris flows described by Van Steijn (1986), Van Steijn and Coutard (1987) and Nieuwenhuijzen and Van Steijn (1990).

Needle Ice (Pipkrake)

This process is neither zonal nor azonal (Harris, 1988), given that needle ice has been observed at low altitudes and even in the tropics (Perez, 1984; Heine, 1989; Soons and Price, 1990; Boelhouwers, 1991a). Bernard-Allée and Valadas (1990) emphasize the influence of needle ice on the evolution of a dell in the Massif Central of France, and Hall (1983) has shown that needle ice formation was a major factor in the development of sorted stripes in the sub-Antarctic. Lawler (1988) makes a comprehensive study of the global distribution of needle ice.

Frost Creep

Frost creep has been a subject of micromorphological studies in the Alps (Haute-Ubaye), where frost creep constitutes the main process of movement and of striated soil formation (Van Vliet-Lanoë, 1988). In addition, Harris (1987) has studied this topic from the geotechnical point of view.

In the Tatra Mountains and in Swedish Lapland Jahn (1989, 1991) has undertaken studies of frost creep by means of wooden pegs.

Nivation

The 'nivation' concept has been taken to task by Thorn (1988) because of its imprecision. Thorn proposes that, as a morphological concept, it should be replaced by a more adaptable term such as 'snow accumulation hollow'. Meanwhile, Soons and Price (1990), Rapp (1987), Lindh *et al.* (1988), and Lehmkul (1989) provide evidence regarding the role of the snow patch, and Nyberg (1991) gives extensive details regarding the present-day operation of nivation processes in northern Sweden.

Patterned Ground

Micromorphological studies in the Alps (Van Vliet-Lanoë, 1987, 1988); Norway (Harris and Cook,

1988) and the Colorado Front Range (Harris, 1990) show that miniature polygonal ground is associated with desiccation fissures and frost heave, and that striped ground also forms on slopes in areas of miniature polygonal ground.

Mass Wasting

Although debris flows have often been discussed (Johnson, 1984; Rickenman, 1990; Zimmermann and Haeberli, 1991), not all are necessarily of periglacial origin. In Switzerland, however, periglacial debris flows have been studied in great detail (Naef *et al.*, 1989; Haeberli *et al.*, 1990; Roesli and Schindler, 1990; Zimmermann, 1990). They have also been studied in the Japanese Alps (Ono and Watanabe, 1986) and Svalbard (Akerman, 1984). André (1990, 1991) has also emphasized the importance of slush avalanches in Svalbard. Finally, Eisbacher and Clague (1984) provide a bibliography on mass wasting in high mountain environments.

Heavy Mineral Concentration

Ahumada (1988), from studies in the Andes, has provided evidence of heavy mineral concentrations at the base of the active layer by sorting linked to freeze-thaw.

QUANTIFICATION OF PHENOMENA AND PARAMETERS

Cliff Recession

According to Francou (1988), this depends upon altitude and aspect (lithological parameters being equal) in the French Alps. At two sites wall recession varies from 0.05 mm to 3 mm per year. In Tarentaise, however, quartzite shows rates of 0.01-0.25 mm per year at the altitude of the 0 °C isotherm. For the Swiss Jura Pancza and Ozouf (1989) estimate 1 mm per year and values of 0-2 cm/yr are suggested for the Tatras (Kotarba, 1987). In Spitzbergen the rate of recession of quartzitic rockwalls is estimated to be approximately 0.15 mm/yr, while that for metamorphic rocks is 0.7 mm/yr, but with maxima up to 1.5 mm/yr (André, 1991).

Temperatures and Moisture Content

A number of attempts have been made to quantify these parameters. Temperatures have been measured in the Karakorum (Whalley *et al.*, 1984), in the Andes and Alps (Coutard, 1985; Manté, 1985; Kaiser, 1987; Coutard *et al.*, 1988; Francou, 1988, 1989a; King, 1990), in the Himalayas (Francou, 1989b), in the Qinghai (Wang Shaoling, 1987), in the Rocky Mountains (Harris, 1986, 1988, 1990), in Antarctica (Miotke, 1982; Miotke and von Hodenberg, 1983; Friedmann *et al.*, 1987; Gjessing and Ovstedal, 1989; Matsuoka, 1990), in the maritime Antarctic (Walton, 1982; Vtyurin and Moskalevskiy, 1985; Blumel, 1986) and in Scandinavia (King, 1983). At Aconcagua in the Andes Schrott (1988; 1991) has studied air temperature changes and also those in the active layer. He sets out evidence showing the weak temperature amplitudes and the continuous daily freeze-thaw cycles which affect the ground for approximately 8 months each year in this environment.

Measurements of moisture content are less frequent (Harris, 1986, 1988, 1990; Hall, 1986a, 1988a; Jahn, 1991). In 1990 Harris showed that the movement of temperature is controlled by the effects of water, which, in turn, brings into question models and reconstructions of palaeoclimates.

Frost Creep and Solifluction

These processes have been measured in the Haute-Ubaye area of the Alps (Coutard *et al.*, 1988) at altitudes between 2500 m and 3100 m. Depending upon the altitude and the facies (importance of rocky material, etc.), annual displacements of between 0.4 cm and 9 cm have been measured. In the eastern Swiss Alps movements of 0.2-11 mm per year have been recorded (Gamper, 1983). Elsewhere, movement varies between 0 and 1.1 cm/yr in the Rocky Mountains (Smith, 1988), 0 and 5 cm/yr in Sweden (Lindh *et al.*, 1988), 0 and 2 cm/yr in the Polish Tatras (Kotarba, 1987), 2 and 20 cm/yr in the Vanoise (Kaiser, 1987), 2 and 3 mm/yr in the Sudeten Mountain low meadow zone, and 8 and 9 mm/yr in the Sudeten high meadow zone (Jahn, 1989). In Swedish Lapland the rate is 0-25 mm/yr for frost creep and up to 75 mm/yr for gelifluction, according to Jahn (1991). On average, most movement rates measured are between 10 and 20 mm/yr (Harris, 1987).

Rock Displacement on Slopes

In Svalbard the rate of accumulation due to annual avalanches during spring is low (0.04–8 mm/yr, according to André (1989). This agrees with previous studies in the Rocky Mountains by Luckman (1988). On the other hand, catastrophic slush avalanches mobilize 1300–7000 m^3 of rock debris every 500 years (André, 1990, 1991). In the Japanese Alps debris flows of the order of $10–10^2$ m^3 occur (Ono and Watanabe, 1986).

Chemical Dissolution

The dissolution of gypsum is of the order of 0.5–2 mm/yr in Tarentaise (Rovera, 1990), depending on altitude and topography, and between 1 and 3 mm/yr in the Tatras (Kotarba, 1987). In Svalbard the dissolution of calcareous rocks is of the order of 3 mm per 1000 years (André, 1991).

Biological Processes

According to André (1991), the biological scaling of rockwalls in Svalbard is of the order of 2 mm per 1000 years. In Alaska biological weathering by algae was found to produce surface flaking of granitic rock (Hall and Otte, 1990) and rates of material loss could be as high as 562 g m^2/yr.

LABORATORY EXPERIMENTS

Over the last decade mountain rock facies have been subjected to gelifraction testing at the Centre de Géomorphologie (CNRS) at Caen: limestones from the Jura (Pancza and Ozouf, 1988) and Alps (Lautridou, 1984; Kaiser, 1987; Rovera, 1990), various alpine schists, sandstones, quartzites and granites (Lautridou, 1984; Kaiser, 1987; Francou, 1988), and gypsum and anhydrite from the Tarentaise and the Vanoise (Kaiser, 1987; Rovera, 1990). A scale of frost susceptibility has been established and the importance of the fine fraction has been evaluated. It seems that the fine fraction is common in mountains but cannot be attributed solely to frost action except in the case of chalk or gypsum.

In Japan Matsuoka (1988, 1990) has studied frost action on volcanoclastic and igneous rocks and established a model of frost susceptibility.

The relationship between laboratory experiments and the field remains difficult (Lautridou, 1988); microgelivation tends to be studied in laboratories, whereas in the field macrogelivation operates with other physical and chemical processes (e.g. see Jerwood *et al.*, 1987; Williams, 1988; plus papers from the 1991 Caen workshop on cryogenic weathering published in *Permafrost and Periglacial Processes*, Vol. 2, No. 4). Hall (1986b, 1988b) and Hall *et al.* (1989) provide examples of laboratory experiments that have utilized large blocks of rock (*c.* 5 kg), together with moisture, chemistry and temperature conditions monitored in the field. Several experiments on debris flows and grèzes litées have also been conducted (Van Steijn and Filippo, 1987; Rickenmann, 1990).

MODELS

Mathematical models of temperature and pressure have been undertaken for Arctic environments (e.g. see the *Proceedings of the Fifth Canadian Conference on Permafrost, Québec, 1990*). However, current studies in mountain regions (Sone, 1991) attempt to deduce ground temperatures and the existence of permafrost on the basis of air temperatures.

The action of frost on calcareous rocks (Letavernier and Ozouf, 1987) and on siliceous rocks (Matsuoka, 1990) has been modelled.

Francou (1990, 1991) has proposed two sedimentological models: a scree-rockfall model and a stratogenic model for bedded slope deposits.

CONCLUSION

Significant progress has been achieved in recent years, thanks to a number of detailed process studies and the quantification of both phenomena and controlling parameters. It still remains to better understand the role of moisture in both soil and rock.

Another problem is the relative importance of fossil and present-day processes in mountain regions. This is especially pertinent to tropical mountains and to marginal zones such as South Africa.

Finally, the role of non-periglacial processes in mountain regions is still poorly understood. This is true not only for tropical zones, but also for typical periglacial zones.

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A study of valley-side slope asymmetry based on the application of GIS analysis: Alexander Island, Antarctica

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Abstract: Geographic Information System (GIS) data from southern Alexander Island were used to evaluate valley asymmetry from an area where ground observations had suggested that south facing slopes were steeper than north facing. Using digital elevation modelling (DEM), data were collected from 2° and 10° arcs centred on the four cardinal directions in order to determine average slope angles for a whole nunatak area (Mars Oasis). It was found that south facing slopes were significantly steeper (34°) than the north facing (28°); east and west facing slopes were each 31°. Bedrock in this area is (approximately) horizontally bedded and so valley asymmetry is considered to be due to aspect-influenced periglacial weathering processes.

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Key words: Alexander Island, Antarctica, DEM, GIS, valley-slope asymmetry

Introduction

Observations collected in Viking Valley (Fig. 1) as part of an earlier study (Meiklejohn 1994) had shown that there was distinct valley symmetry, with the south facing slopes being much steeper (58°) than the north facing (30°). Studies of weathering processes and landforms in this same area (Hall 1997a, 1997b) had indicated aspect-constrained weathering processes. The original study by Meiklejohn (1994) was with a very limited sample size and, essentially, dealt with a valley cross section rather than multiple sets of data measurements. Nevertheless, it seemed clear that the asymmetry was not structurally controlled, as the geology consists of horizontally (or nearly horizontally) bedded sedimentary rocks (mainly sandstones and mudstones). Meiklejohn (1994) was able to show that the warming of the north facing valley slope resulted in a much deeper active layer than was measured on the cooler, south facing slope. He argued that the deeper active layer facilitated more extensive weathering and mass movement and that this results in a shallower slope angle when compared to the south facing slope where the active layer thawed later in the summer, froze earlier and was very shallow. The close proximity of the permafrost to the ground surface on southern aspect slopes also helped maintain, by means of bonding the rock material together and increasing the shear strength, the recorded steep slopes (see French 1996, p. 202–203 for a discussion regarding the interaction between permafrost and valley asymmetry).

The aridity of this area suggests that freeze-thaw weathering is spatially and temporally limited to the few locations where water is present, and available data suggest that thermal stress fatigue is likely to play a major role in rock weathering (Hall 1997a, 1997b). Aspect greatly influences weathering within this area (Hall 1997a). Data show that, despite its orientation, the northern aspect exhibits the lowest late summer to early autumn rock temperatures (Hall 1998) but has the highest rock surface temperatures during the summer proper (Hall 1997a). These data (Hall 1997a, 1998) indicate that the influence of aspect at these latitudes is complex and varies with the season. Schmidt hammer rebound values and the sizes of taffoni (Hall 1997a, table 9) indicate that there is substantial weathering on slopes with northern aspects, certainly more than on the eastern and western facing slopes (in that data set information for the southern aspect were missing). Accepting that, in some manner, cryoplanation terraces are the product (in part) of weathering, then the orientation data regarding cryoplanation terraces from this area clearly show the greatest weathering on the north through north-west to western aspects (Hall 1997b, fig. 3) but little or no weathering from east through south to south-west. Meiklejohn (1994, table I) showed the difference between weathering on the north facing slope (greatest) and the south facing slope (least) in Viking Valley and used this, in part, as his explanation for valley asymmetry. Thus, with this evidence of valley asymmetry for Viking Valley and the aspect-constrained effects on weathering for this area, it was decided to test the hypothesis that the whole nunatak complex would show slope asymmetry. The GIS approach offered an ideal method of assessment.

For the Northern Hemisphere valley asymmetry in areas of permafrost frequently exhibits steeper north facing slopes, but with proximity to the pole there is a greater variation in preferred orientation (French 1996). Data from the Antarctic are very sparse and discussion regarding valley asymmetry is rare. French (1996, p. 202) in a discussion regarding valley asymmetry in the Northern Hemisphere, notes that such factors as changes in the inclination of the sun with increasing latitude, direction of snow-bearing winds, impact of active layer thickness, and the role of streams, may all play a role in determining asymmetry orientation. Meiklejohn (1994, fig. 5) also utilised some of these parameters within his model for valley asymmetry development. The GIS analysis presented

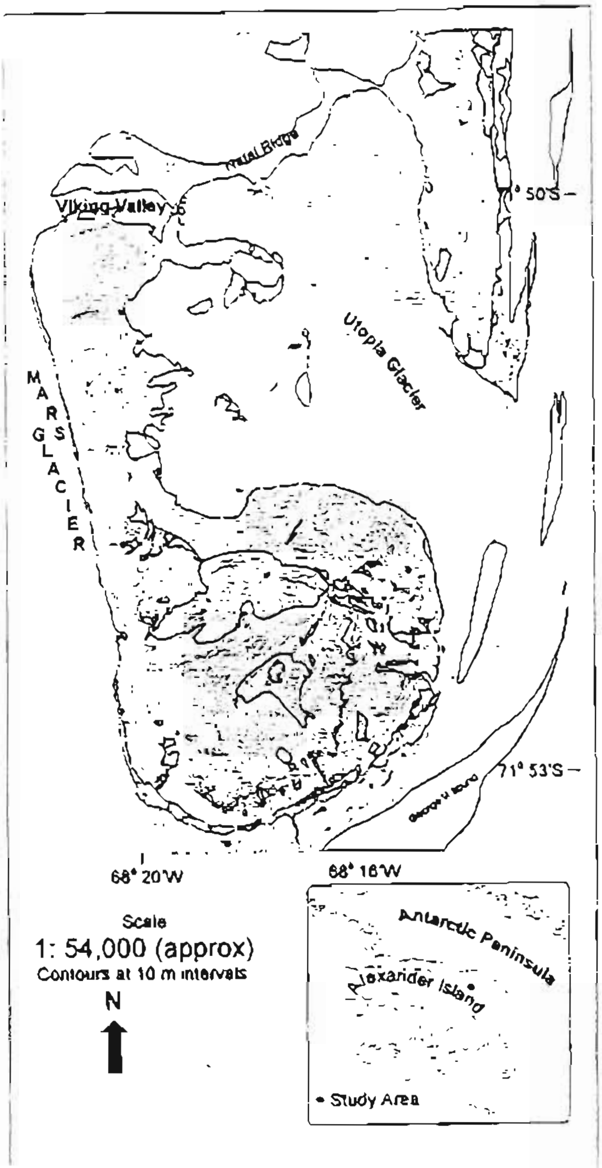

Fig. 1. Location of study area

here can show the occurrence (or not) of asymmetry but cannot explain it. Fortunately, some of the weathering information available for this area may offer some explanation for the asymmetry observed.

Study area

Measurements were undertaken on the nunatak complex between the Mars Glacier and King George VI[th] Sound (Fig. 1) at the southern end of Alexander Island (71°54'S, 68°23'W). Formerly covered by ice from the Antarctic Peninsula (Sugden & Clapperton 1978), deglaciation has exposed the higher rock areas and only a few ice remnants remain (e.g. at the head of Viking Valley and on the western side of Two Step, Fig. 1). The area is one of continuous permafrost with very little precipitation and a summer mean air temperature of -2.5°C and a winter mean of -11.5°C (Hall 1997a, 1997b). Geologically the area comprises argillaceous sedimentary rocks, sandstones and mudstones (Taylor *et al.* 1979). The dominant lithology is an arkose sandstone with subspherical, post-compaction siliceous concretions with a ferruginous cement (Thomson 1964, Moncrieff 1989). There are also mudstones, sometimes with lime carbonate nodules, shaly mudstones and orthoquartzitic sandstones; all are near horizontally bedded, with a 7° dip to the north (Fox personal communication 1999). Linearity along the coast is the product of faulting as too is the east–west linearity of the glaciers that cut the coast.

Methodology

Using existing British Antarctic Survey (1998) digital contour information for the Mars Oasis area of southern Alexander Island a digital elevation model (DEM) in the format of a TIN (Triangulated Irregular Network) was constructed in Arc/Info[R]. The 10 m proximal tolerance TIN was then converted into a polygon coverage that included aspect, slope and surface area values (Fig. 2a). This polygon coverage was built to exclude all areas covered by ice and snow as snow and ice slopes may differ significantly from those of the bedrock beneath, leaving only exposed rock polygons. This 'rock' data set was used to crop the initial aspect and slope DEM to generate a new coverage. From this new layer north, south, east and west information on slope angle was extracted in ArcEdit[R] (Fig. 2b). Topology was created from individual 'packets' of information for each of the aspects. Using the surface area of each TIN triangle as a weighting factor each slope and aspect value has equal significance and thus the values are independent measures of slope. Initially, aspects were constrained to +/- 1° of the cardinal orientation (e.g. between 179° and 181° for the south orientation). Although this approach produced a substantial data set (e.g. 831 north aspect slope values) it was felt that the 2° constraint was too narrow as it did not facilitate continuity down a complete slope as, in many instances, there was more than a 2° orientation distribution on a single slope. The data analysis was thus re-run for a +/- 5° arc and, although this produced significantly more data (e.g. 3659 north slope values), it provided a more realistic assessment of slopes (Table I). For each of the cardinal directions, the two packets of information (aspect and slope for both the 2° and the 10° arcs) were exported to a database where a statistical analysis was performed using one-way ANOVA (with aspect as the independent variable, slope as the dependent, and the whole being weighted as a function of slope with surface area). It would have been possible, but time consuming, to run regressions for all aspects but this was not done for two reasons. First, the size of the data sets that would have resulted. Second, the original focus was the comparison of only north and south slopes as these were the

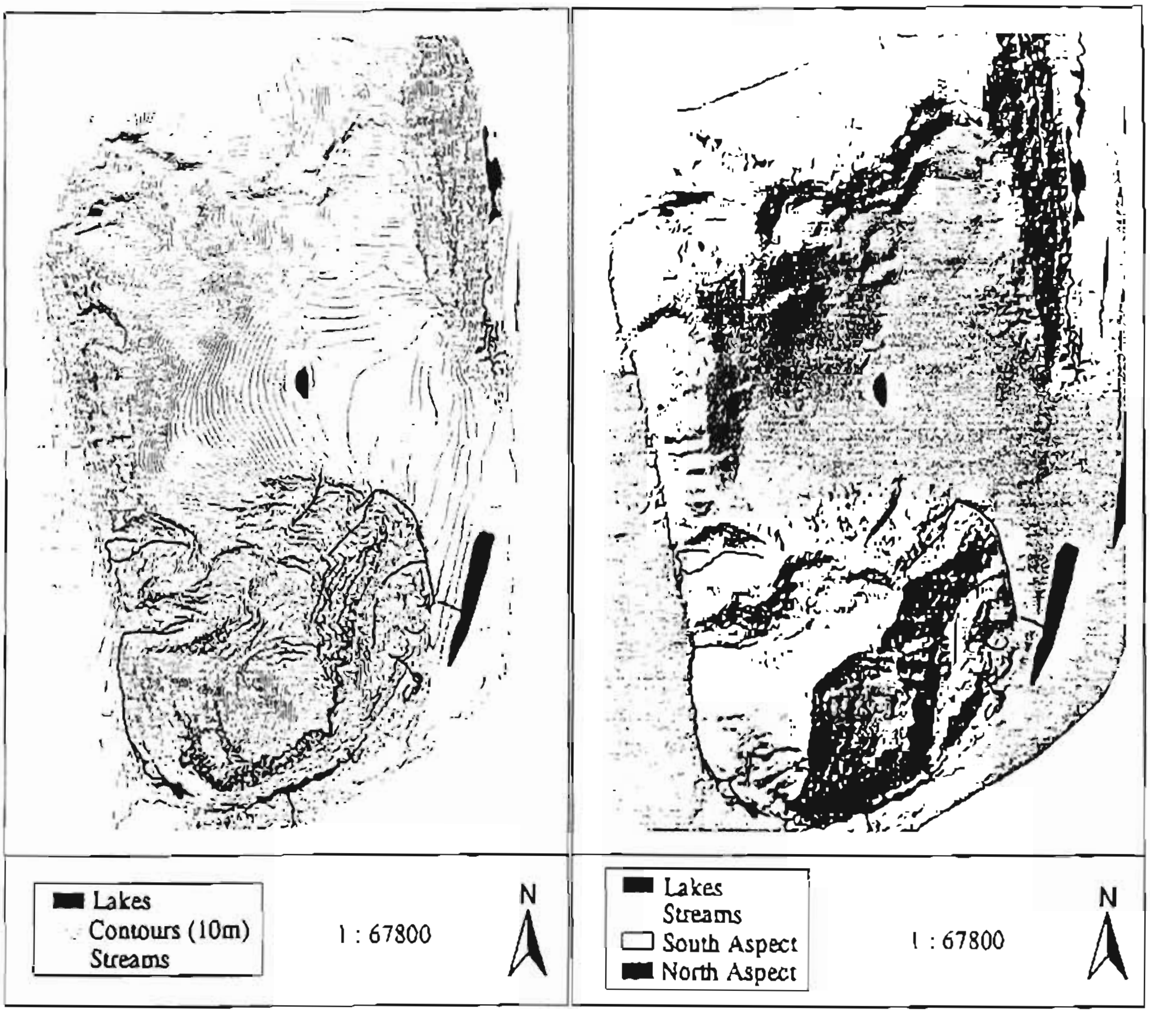

Fig. 2. a. Slope map based on contour information. b. Example of the south and north aspect slope segments extracted by ArcEdit that were used in the statistical analysis.

aspects for which field observations were available. Whilst obviously valuable in the longer term to undertake a more comprehensive evaluation, it was felt that, at this stage, it was solely the original field observation that should be tested.

In the derivation of the contours for the Mars Oasis map, digitized index contours were at 200 m spacing but with a substantial number of height measurements between and throughout the index contours (Fox personal communication 1999). Thus, the overall map accuracy is better than 200 m and the overall inaccuracy of the generated non-index contours is minor since most of the significant and drastic changes in elevation are accounted for in the original sampling. Further, the underlying error of the auto-correlated non-index contours is uniform throughout the area, affecting all aspects equally. Thus, all the values we derive are related in all planes in direct correlation to the index contours and other measured height variables. In addition, the use of TIN drastically reduces the error since the values are not calculated on the two linear values between the index contours. Whilst our significance value is probably on the high side, this was the best DEM generating method for this type of analysis. In the final analysis, the DEM approach is fraught with inaccuracy based upon the derivation of the original data and so the resulting slope values are not necessarily "accurate" in the sense that field measurements would derive the exact same angles (although neither is it impossible). Rather, the derived values

Table I. Summary of slope angle (and number of observations) for the cardinal aspects as a function of the 2° or 10° arc of data acquisition.

Aspect	+/-1° arc Slope angle	n	+/- 5° arc Slope angle	n
north	30°	831	28°	3659
south	34°	631	34°	2164
east	32°	4332	31°	13101
west	30°	1325	31°	4039

are accurate within the accuracy of the original data and the manner in which those were obtained. The argument is made that the approach used here minimizes the inherent inaccuracies such that in terms of bedrock slope the relationship of north to south slopes is a reflection of reality although the specific slope values should be viewed with caution. With these considerations in mind, it is argued that the following analysis of slope asymmetry has validity.

Results and discussion

The results of the analyses clearly show that south facing slopes are steeper than north facing irrespective of whether a ± 1° arc or ± 5° arc around the cardinal direction is used (Fig. 3). In spite of this, the weighted one-way ANOVA indicated, with a 99.999 per cent confidence level, that the data sets were significantly different. The change to a ± 5° arc increased the number of north facing slope observations by 340% (interestingly the south, east and west all increased by just over 200% – 243%, 202% and 204% respectively) and decreased the overall northern aspect slope angle from 30° (in the ± 1° arc) to 28°, but left the south facing slope at 34°. The ANOVA test indicated, with even greater confidence, that the two data sets were different and that, indeed, the south facing slopes were steeper than the north facing for this area. East and west facing slopes were both steeper (31°) than the north facing but less than the south facing (Table I, Fig. 3). Statistically, the east and west aspects are significantly different from those of both north and south. Thus, with a high statistical confidence (99.999%) GIS slope data indicate southern aspects are steeper than the eastern or western, and that, in turn, these are steeper than the northern.

There seems to be clear evidence in support of aspect-related slope asymmetry from this area. The hypothesis of Meiklejohn (1994) that south facing slopes are steeper than north facing is substantiated. This information does not, though, offer any explanation for the cause. Fortunately, the horizontal, or near horizontal, nature of the rocks in this area coupled with the limited suite of lithologies negates any structural or lithological explanation. French (1996, p. 183), states that, for slope evolution in periglacial regions, it is necessary to take into account "...(a) the regional and microclimates, (b) the lithology, and (c) the dominant weathering process". Within this region both climate and structure/lithology can be considered uniform so that the asymmetrical effects are probably due to microclimate and its effect upon weathering and slope processes. This being an area of continuous permafrost, aspect will certainly play an important role in determining the thickness of the active layer; the north facing slope being warmer would have the deeper active layer. As weathering and mass movement are constrained to the active layer it then follows that greater weathering and mass movement can be expected on the north facing slopes. Temperature differences between north and south facing slopes were significant (see table I of Meiklejohn 1994), with the ground surface mean for the north facing being 21.9°C and that for the south 7.1°C. Weathering indicators (e.g. Schmidt hammer rebound values) substantiate the notion of increased weathering on the north facing slopes. Thus, there appears to be evidence regarding aspect-constrained effects on temperatures, active layer thickness and weathering rates that shows a distinct difference between north and south facing slopes.

The majority of periglacial texts consider the weathering attribute associated with valley asymmetry to be that of freeze-thaw weathering, but recently suggestions have been made that this may be an over simplification (e.g. see French 1996, p. 183). The whole concept that freeze-thaw is the ubiquitous and sole weathering process in cold regions is being questioned (e.g. Hall 1995) and this is particularly appropriate here owing to the general aridity of the area. Available information (Hall 1997a) indicates that snowfall is very limited and that the availability of water is highly restricted, both spatially and temporally. If moisture is not present breakdown by the freeze-thaw mechanism is impossible. Hall (1997b, p. 185) argues that in these higher latitudes protective shading is detrimental to geomorphic activity (weathering, transport, water availability, etc.) and so the greatest amount of change takes place on the sunnier north facing slopes. Certainly, weathering data (e.g. taffoni occurrence and size, and Schmidt hammer rebound values) indicate greatest weathering on the northern and western exposures (Hall 1997a), which can also explain the aspect-specific orientation of cryoplanation terraces (Hall 1997b). Rock temperature data (Meiklejohn 1994, Hall 1997a, 1997b) show that thermal stress fatigue and thermal shock are the most likely factors in weathering within this area, and "...the western and northern exposures experienced a far greater range..." and that "...the rate of change of temperature was far greater on the western and northern aspects..." (Hall 1997b, p. 186). Thus, the insolation factor that affects the

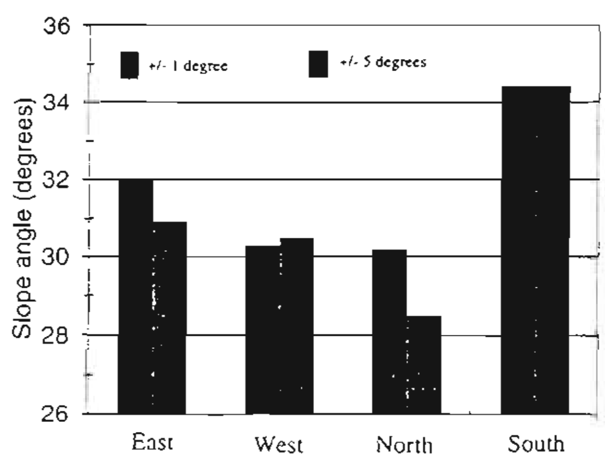

Fig. 3. To show the slope angle differences for the four cardinal directions as indicated by both +/- 1° and +/- 5° arcs of data capture.

thickness of the active layer will, even in the absence of moisture, contribute to enhanced weathering on the north facing slopes. However, water is not totally absent but rather is spatially and/or temporally constrained such that it is on the warmer slopes that it will more readily be in the liquid form and thus able to enhance both weathering and transport. The combination of a deeper active layer and the enhanced transport this can facilitate, together with the increased weathering, can explain the gentler north facing slopes.

The impact of radiation receipts upon the slopes has been qualitatively considered but little has been done to evaluate the quantitative impact, the study of Davidson (1992) being one significant exception. As Meiklejohn's model (1994, fig. 5) and theory (e.g. Fench 1996) suggest, the role of radiation upon ground/rock temperatures can be a significant factor. What is rarely considered is the impact of radiation upon slopes of different angles. The angle which depends upon the time of year (and hence the declination of the sun), can greatly affect solar heating. Bolsenga (1964, p. 1) states "Quantitative information on global radiation is needed in many types of studies including those pertaining to the distribution, accretion, and ablation of ice and snow, the thawing of soils...." and provides tables for daily sums of global radiation for cloudless skies. As invaluable as these are, they still do not facilitate evaluation of different slope aspects, angles and albedos – the very factors needed in slope studies. MeteoNorm™ is a computer programme that allows for the calculation of radiation receipt (global, diffuse and direct) for slopes of any aspect, angle and albedo for non-polar sites. Although it is not possible to run the calculations for Mars Oasis, data from Base Esperanza (63°24'S, 56°59'W, 13 m a.s.l.), the nearest Antarctic station available in MeteoNorm™, does provide information germane to the arguments presented here. Calculations were run to derive irradiation (in kWh m^{-2}) for a horizontal surface and north and south facing slopes of 90°, 45°, 30° and for the mean values (28° and 34° respectively) derived from the DEM analysis (Table II). These show some results that might not have been arrived at intuitively, but which may impact on valley asymmetry. The data for the horizontal surface show the expected pattern, with highest values from the October to March, spring to autumn period, and lowest values during the winter. The slope data are significantly different, with the north facing slope receiving the highest radiation receipt in midwinter on the 90° and 45° slopes – the time when the slopes are nearly normal to the declination of the sun. The 30° slope shows two peaks, one in winter when the sun's rays are nearly normal to the surface and one in summer when, although the sun's angle is much steeper, the radiation input is high. The 28° slope has a winter peak but the highest values are in summer. Conversely, the south facing slopes have low to negative radiation receipts in winter and higher, but still low when compared to the north facing slopes, values in summer. The principle shown by these Base Esperanza data, is that there is a highly significant radiation receipt differential between the northern and southern slopes that should, further influenced by slope angle, greatly impact slope processes. With the availability of moisture, and certainly the propensity for snow melt is higher on the north facing slope, so (following the arguments of Davidson 1992) the north facing slopes should experience more freeze-thaw and wet-dry weathering; the data of Hall (1997a) would suggest that thermal stress fatigue should also be a major factor. Thus, the radiation data indicate that the north facing slopes will be geomorphologically far more active than the south facing slopes, and that the slope asymmetry model of Meiklejohn (1994) appears to be substantiated. The importance of radiation receipt upon slope processes, and the impact that slope angle, aspect and albedo has upon these, is in need of much more quantitative evaluation.

Table II. Data from Base Esperanza (63°24'S, 56°59'W, 13 m a.s.l.) to show the total irradiation (in kWH m^{-2}) of a horizontal plane compared to that of both northern and southern tilted planes (albedo = 0.25) of 90°, 45°, 30° and for the mean slope (28°and 34° for north and south respectively) angles found in the GIS analysis.

	Horizontal	N90°	S90°	N45°	S45°	N30°	S30°	N28°	S34°
Jan	144	101	65	146	93	151	114	151	109
Feb	106	91	44	121	57	121	73	121	68
March	77	96	30	109	37	103	39	102	38
April	39	90	14	85	17	74	17	72	17
May	19	135	4	107	4	82	4	78	4
June	10	300	-1	220	-4	159	-5	150	-4
July	14	210	1	158	-1	117	-1	111	-1
Aug	33	128	9	111	10	90	10	87	10
Sept	66	111	23	116	27	104	28	102	28
Oct	112	114	42	142	53	139	67	138	60
Nov	139	104	60	147	83	151	106	151	100
Dec	154	103	72	153	105	159	126	159	121
Year	915	1582	362	1616	483	1451	580	1423	551

Conclusions

The GIS approach to the evaluation of valley asymmetry provides, where data are available, a very efficient and powerful diagnostic tool. Certainly it facilitates acquisition of far larger data sets than would be possible by non-computer based approaches. The results from Mars Oasis clearly demonstrate that there is aspect-related slope asymmetry, with steeper south facing slopes, that is not the result of structural or lithologic factors. The general aridity of this area mitigates against the ubiquitous action of freeze-thaw weathering but it can occur in spatially-constrained areas where water is available; north facing slopes would be those most likely to experience snow melt where snow is available. Field information suggests that weathering by thermal stress fatigue/shock should also be considered and that this, too, would be most effective on north facing slopes. The impact of solar radiation on the slopes, and the variability of this through the year, is potentially a major factor and more information on this is required. The weathering, coupled with the aspect influence on active layer thickness, helps explain the greater breakdown and debris mobility on north facing slopes and hence the lower angle. The occurrence of a number of weathering related landforms (taffoni and cryoplanation terraces) in this area substantiates this hypothesis. Thus, despite the errors that can be produced from the original DEM data, it appears that there truly is valley asymmetry in this region and that it can be explained by aspect-constrained process differentiation.

Acknowledgements

We would like to thank British Antarctic Survey for making available the digital data for the Mars Oasis area of Alexander Island and Dr David Walton in particular for his help and support in allowing KH to undertake fieldwork in this area. Paul Cooper and Adrian Fox, of the GIS Unit at the British Antarctic Survey, are sincerely thanked for all their valuable advice, information and insights into both the original data and the derivation of the DEM model. Michelle Gray of the British Antarctic Survey is acknowledged as providing the original data. Funding for the fieldwork (KH) was provided by a grant from the Foundation for Research and Development (South Africa) whilst this follow-up work was supported by grant #185756 of the National Science and Engineering Research Council (Canada) to KH. Dr Roger Wheate, Mr. Scott Emmons, and Ms Nancy Alexander are thanked for their help and support with the GIS/DEM production and analysis.

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CAPE ROBERTS DRILLING PROJECT
ONE PAGE SCIENCE SUMMARY

COUNTRY: United Kingdom **INSTITUTION:** British Antarctic Survey

PRINCIPAL INVESTIGATOR: Dr J L Smellie

ADDRESS: British Antarctic Survey, High Cross, Madingley Road. Cambridge CB3 0ET

TELEPHONE: 44- (0)1223 221418 **E-MAIL:** J.Smellie@bas.ac.uk

COLLABORATORS: Prof. K Hall, Dr L. Arocena (North British Columbia)

TITLE: *Using clay mineral analysis of weathered clast rinds to elucidate climatic history*

RATIONALE & JUSTIFICATION: Because different species of clay minerals are formed under different weathering processes (chemical and physical), they are a valuable tool for reconstructing the palaeoenvironmental history of Antarctica. That rationale underpins current studies of detrital clay minerals in mudrocks within CRP. Comparable studies of clast weathering-rinds have thus far been neglected. Although a small pilot study was undertaken for CIROS-1, this type of palaeoenvironmental study is unique at this level of detail. Observations from CRP-2/2A indicate that allogenic clasts with weathered rinds are present in CRP-2/2A, below 580 mbsf at least (pers. comm. from F Talarico). They are predicted to become more abundant as CRP-3 moves to greater depths than its predecessors, when glacifluvial and fluvial conditions conducive to weathering become more prevalent. The rinds formed during contemporaneous terrestrial weathering *in situ* prior to final deposition in McMurdo Sound. Thus, they are a direct proxy for palaeoenvironmental conditions in the source area. Clasts of basaltic composition (e.g. Ferrar dolerites) are particularly reactive to weathering, but granitoids may also show weathering-related rotting. Both clast types are abundant within CRP-1 and 2/2A. The Ferrar types are most valuable in displaying relatively invariant whole-rock compositions, thus largely removing possible clay mineral variations based solely on chemical composition. The persistence of Ferrar clasts down-core is a strength of the proposal, which should increase the likelihood of a successful outcome, but the association with granitoids creates an extra, composition-related dimension to the study, namely additional clay mineral species formed under more acidic weathering conditions. The investigation does not duplicate existing research. It is complementary to current work on detrital clay minerals but is a distinctive research strand that is one more "string to the bow" of palaeoclimate reconstruction.

SAMPLING REQUIREMENTS: Continuous sampling of 1-2 clasts (if available) every 10 m to give a "core-length" picture, and/or selective sampling at key points where rapid palaeoenvironmental changes are suspected from concurrent studies, to fill in the fine detail of the "total core" picture.

EXPECTED RESULTS: Documentation of changing terrestrial palaeoenvironmental conditions in McMurdo Sound hinterland during the Cenozoic.

CRITICAL REFERENCES:

Hall, K. & Buhmann, D. 1989. Palaeoenvironmental reconstruction from redeposited weathered clasts in the CIROS-1 drill core. *Antarctic Science*, **1**, 235-238.

MATERIALS & METHODS: Mineral analysis of weathered clast rinds by bulk-analysis XRD.

PROJECT REQUIRES:

Presence at drillsite	If required
Presence at McMurdo Science & Engineering Center	Yes (JLS)
Samples from core store	Possibly

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NIVATION: AN ARCTIC-ALPINE COMPARISON AND REAPPRAISAL

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ABSTRACT. Nivation is a collective noun identifying a set of geomorphic processes, comprised of an indeterminate number of elements and of unknown relative importance, for which there is little likelihood of ever producing a precise definition. Instead, attention should first be directed towards the relationships between snow-packs and individual geomorphic processes. The relationship between freezing amplitude in the bedrock, snow cover, and aspect at an Arctic site in northern Norway and an alpine site in the Front Range, Colorado, U.S.A. is complex. Comparison of field data and laboratory criteria permit several conflicting interpretations. If oscillations across 0°C, regardless of freezing amplitude, are critical, the alpine site is potentially a more active freeze–thaw weathering regime, with a primary springtime peak and a secondary fall peak. If a freezing amplitude of −5°C is required for effective freeze–thaw weathering then the alpine site is largely inactive and the Arctic active (but with only a single fall peak). Chemical weathering is much more important at snow-patch sites than has traditionally been recognized. Mass wasting at colluvial sites subject to snow patches is dominated by interaction between overland flow and solifluction when the site is unvegetated, and by solifluction when it is vegetated. Given contemporary knowledge of snow and glacial geomorphology, there appears to be no threshold, only differences of intensity. Resolution of the disruptive mechanism associated with bedrock freezing and its constraining temperature and moisture requirements is the most pressing present problem in the field of snow geomorphology.

RÉSUMÉ. *Nivation: une comparaison arctique–alpine et une réestimation.* La nivation est une appellation collective désignant un ensemble de processus géomorphologiques composé d'un nombre indéterminé d'éléments d'importance relative inconnue, pour lequel il est très peu probable qu'on parvienne jamais à une définition précise. On devrait plutôt apporter en premier lieu son attention aux relations entre le manteau neigeux et chaque processus géomorphologique individuel. La relation entre l'intensité du gel du lit rocheux, le manteau neigeux et l'exposition dans un site arctique en Norvège septentrionale et dans un site alpin dans le Front Range, Colorado, États-Unis est complexe. La comparaison entre les données du terrain et les résultats du laboratoire autorise plusieurs interprétations opposées. Si les oscillations autour de 0°C indépendamment de l'intensité du gel, sont critiques, le site alpin peut être soumis à un régime d'alternance gel–dégel plus actif avec un maximum primaire au printemps et un maximum secondaire à l'automne. Si une intensité de gel de −5° est nécessaire pour que se manifeste une dégradation réelle par le gel–dégel, alors le site alpin est surtout inactif et le site arctique actif (mais seulement avec un seul maximum à l'automne). La désagrégation chimique est beaucoup plus importante sur les sites où subsistent des congères de neige qu'il n'est traditionnellement reconnu. La dégradation de masse sur les sites colluviaux sujets aux accumulations de neige est dominée par l'interaction entre les écoulements superficiels et la solifluction lorsque le site n'est pas végétalisé, et par la solifluction lorsqu'il est végétalisé. Les données de la connaissance contemporaine sur la géomorphologie nivale et glaciaire fait apparaître qu'il n'y a pas de seuil tranché mais seulement des différences d'intensité. La compréhension du mécanisme de rupture associé au gel du lit rocheux et ses exigences contraignantes de température et d'humidité est le problème actuellement le plus urgent dans le domaine de la géomorphologie de la neige.

ZUSAMMENFASSUNG. *Nivation: ein Vergleich Arktis–Hochgebirge und eine Neueinschätzung.* Nivation ist ein Kollektivbegriff für eine Reihe von formbildenden Vorgängen; er umfasst eine unbestimmte Zahl von Elementen, deren relative Bedeutung unbekannt ist, weshalb nur geringe Wahrscheinlichkeit dafür besteht, dass sich je eine genaue Definition finden lässt. Stattdessen sollte die Aufmerksamkeit zunächst auf die Beziehungen zwischen Schneeanhaufungen und einzelnen formbildenden Vorgangen gerichtet sein. Die Beziehung zwischen der Amplitude des Bodenfrostes, der Schneedecke und dem Aussehen einerseits an einem arktischen Platz in Nordnorwegen, andererseits an einer Hochgebirgsstelle in der Front Range von Colorado, U.S.A., ist komplex. Der Vergleich zwischen Feldmessungen und Laborkriterien lässt mehrere widersprüchliche Deutungen zu. Wenn Schwankungen um 0°C ohne Rucksicht auf die Frostamplitude massgeblich sind, dann weist die Hochgebirgsstelle ein potentiell aktiveres Frostwechsel-Verwitterungsregime mit einem primären Höhepunkt im Fruhling und einem sekundären im Herbst auf. Wird dagegen eine

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Frostamplitude von $-5°C$ für eine effektive Frostwechsel-Verwitterung benötigt, dann ist die Hochgebirgsstelle weitgehend inaktiv und die arktische aktiv (jedoch mit nur einem Höhepunkt im Herbst). Chemische Verwitterung ist an Stellen mit Schneeflecken weit wichtiger als traditionsgemäss angenommen wurde. Die Massenbewegung an Anhäufungsstellen von Colluvium unter der Wirkung von Schneeflecken wird vom Wechselspiel zwischen Rutschungen und Solifluktion bestimmt, wenn das Gelände vegetationslos ist, von der Solifluktion allein im bewachsenen Gelände. Nach dem derzeitigen Wissensstand in der Schnee- und Eisgeomorphologie scheint hier keine Schwelle zu bestehen, höchstens Unterschiede in der Intensität. Die Klärung des Abrissmechanismus, der mit Frost am Felsbett verbunden ist, und der dafür notwendigen Bedingungen fur Temperatur und Feuchtigkeit ist das gegenwärtig dringendste Problem auf dem Gebiet der Schneegeomorphologie.

Introduction

Traditionally, snow-patches have been distinguished from glaciers by identification of the former as static and the latter as mobile. Evidence of snow-pack mobility (Costin and others, 1973; Mathews and Mackay, 1975) obscures the established glaciological distinction. A parallel problem exists in separating the geomorphic impact of snow-patches and glaciers. In many ways "nivation" (a collective noun used to summarize all snow-pack-derived erosion) confounds the issue.

Matthes (1900) coined the term "nivation" after a reconnaissance study in the Big Horn Mountains of Wyoming, U.S.A. Unfortunately, the term became entrenched in the literature prior to any comprehensive process studies designed to identify individual components and their relative importance. That nivation remains a rather loose concept and lacks a uniform definition is shown by the literature review summarized in Table I.

Table I. Elements of nivation according to most widely cited references

Author	Date	Snow-pack mobile	Frost weathering: At margin	Frost weathering: Under snow	Solifluction dominant	Sheet-wash: Transport	Sheet-wash: Erosion	Chemical weathering	No vegetation
Matthes	1900	No	Yes	No	—	Yes	No	—	Yes
Ekblaw	1918	—*	Yes	—	Yes	Yes	—	—	—
Lewis	1936, 1939	No	Yes	Yes	Yes	Yes	No	—	Yes
McCabe	1939	No	Yes	Yes	—	No	—	—	—
Roth	1944	—	Yes	—	Yes	—	—	—	—
Boch	1946, 1948†	Yes	Yes	No	Yes	Yes	—	Yes	Yes
Paterson	1951	Yes	Yes	—	Yes	—	No	Yes	Yes
Henderson	1956	—	Yes	—	Yes	Yes	—	—	—
Cook and Raiche	1962	—	Yes	No	Yes	Yes	—	—	—
Nichols	1963	—	Yes	No	—	Yes	Yes	—	—
Lyubimov	1967	No	Yes	No	Yes	Yes	Yes‡	Yes	—
St Onge	1969	—	Yes	—	—	—	—	—	—

* Means "not mentioned".
† Boch (1948) specifically corrects some errors in Boch (1946).
‡ Under longitudinal snow-patches only.

Faced with such an amorphous concept, two recent and independent studies (Hall, unpublished; Thorn, 1976, unpublished) have attempted to verify quantitatively the existence and characteristics of nivation. This paper is a *post hoc* comparison of the two studies and provides a comparison between nivation in Arctic maritime (Hall, unpublished) and temperate alpine regimes (Thorn, 1976, unpublished). The individual findings and methodologies of the two studies are not discussed exhaustively, rather the emphasis is upon the comparisons and contrasts and their significance for refining the understanding of nivation.

STUDY SITES

Arctic site

The Arctic site lies on a north-facing slope of the east–west trending valley of Austre Okstinbredal in the Okstindan region of northern Norway (Fig. 1). The north-facing valley wall rises from 725 m, through a series of benches, to a maximum elevation of 1 916 m a.s.l. Only 80 km from the sea, the area is relatively maritime; integration of a partial local climatic record with data from Hattsfjelldal, some 50 km south, suggests a mean annual temperature of approximately −3°C. Mean annual precipitation is 1 120 mm and snow covers the ground for 180 d of the year.

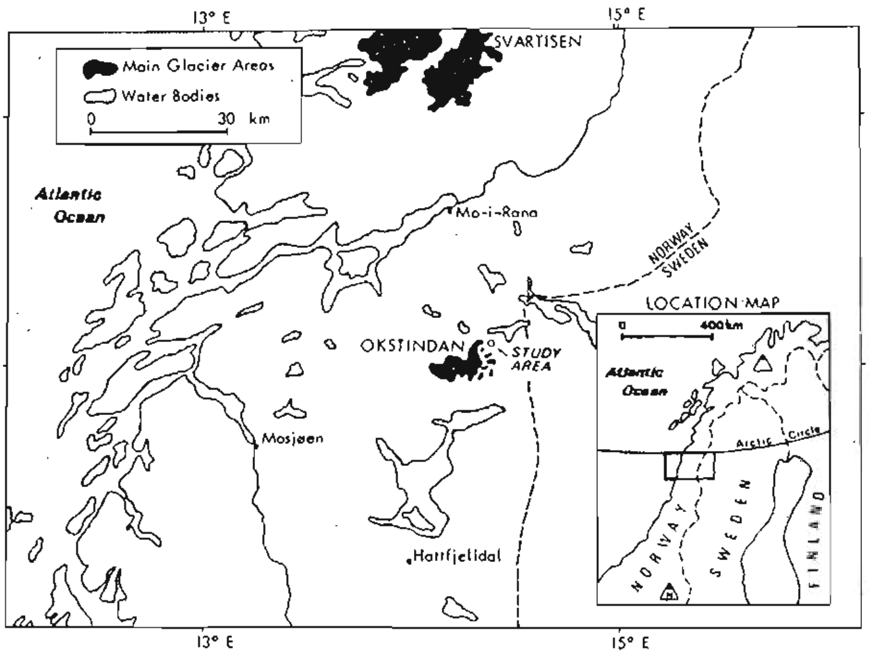

Fig. 1. Regional location of the Arctic research site. The map is modified with permission from The Times atlas of the world. Mid-century edition. Vol. 3. Northern Europe. *London, The Times Publishing Co., Ltd., 1955, plate 52.*

The study site lies mainly at 800 m a.s.l. Overall the semi-permanent snow-patch comprises an upper transverse section and a longitudinal section which drops to the valley bottom (Fig. 2). The transverse section has a maximum width across slope of 150 m, and a maximum down-slope extent of 100 m; the longitudinal section is only 20–30 m wide, but nearly 250 m in length. Only the transverse snow-patch was studied in detail.

Four small-scale elements may be identified within the transverse patch. These are: (1) a near-vertical bedrock back-wall; (2) a stone pavement dissected by rock outcrops, particularly well developed at the eastern end; (3) a lower vegetated area; (4) a gully which bisects the entire site (the gully is headed at the back-wall by a waterfall and below the transverse snow-patch becomes the location of the longitudinal snow-patch). The back-wall ranges from 2–10 m high and averages about 4 m; it is composed of various types of schist and basal niches are prominent. Coarse gravel and cobbles in a matrix of fines forms the stone pavement which is unvegetated and dissected at its eastern end by a series of bedrock outcrops ("rockbands") up to 1.5 m high and parallel to the slope. Vegetation gradually increases down-slope, culminating in a continuous cover, dominated by billberry (*Vaccinium myrtillus*), dwarf birch (*Betula nana*) and *Carex* spp.

Fig. 2. A view looking approximately south-west across the Arctic research site. Note the waterfall at the top of the transverse section of the snow-patch.

Alpine sites

Two snow-patches on the south face of Niwot Ridge, Colorado Front Range, U.S.A. (Fig. 3) were the focus of the alpine study. The bedrock, or "longitudinal", snow-patch (Fig. 4) is approximately 3 590 m a.s.l. It is a seasonal snow-patch which accumulates to depths of 4–5 m against the western wall of a narrow gully dissecting a cliff. The bedrock is gneiss and at the up-slope end of the gully there is evidence of hydrothermal alteration (Thorn, unpublished).

The larger ("Martinelli") snow-patch (Fig. 5) ranges through the tundra–forest ecotone at elevations above 3 450 m a.s.l. Maximum extent is approximately 175 m across slope and 450 m down-slope. Syenite colluvium underlies the snow patch which accumulates along the down-wind edge of a till lobe. As the ablation season progresses the snow-patch divides into an upper, circular snow patch and a lower, longitudinal one; the upper patch rarely ablates totally, but the lower patch is destroyed annually.

Core areas of both Martinelli patches are unvegetated, but peripheral areas are partially vegetated and discontinuous tundra cover occurs beyond the upper basin and sub-alpine meadow beyond the lower basin.

Mean annual air temperature on the crest of Niwot Ridge is -4°C with annual precipitation of 1 021 mm (Barry, 1973). Wintertime wind speeds are high; mean monthly averages from October through March exceed 11 m/s. As a result the landscape is divided into zones of high and low effective precipitation, as snow-fall is rapidly redistributed.

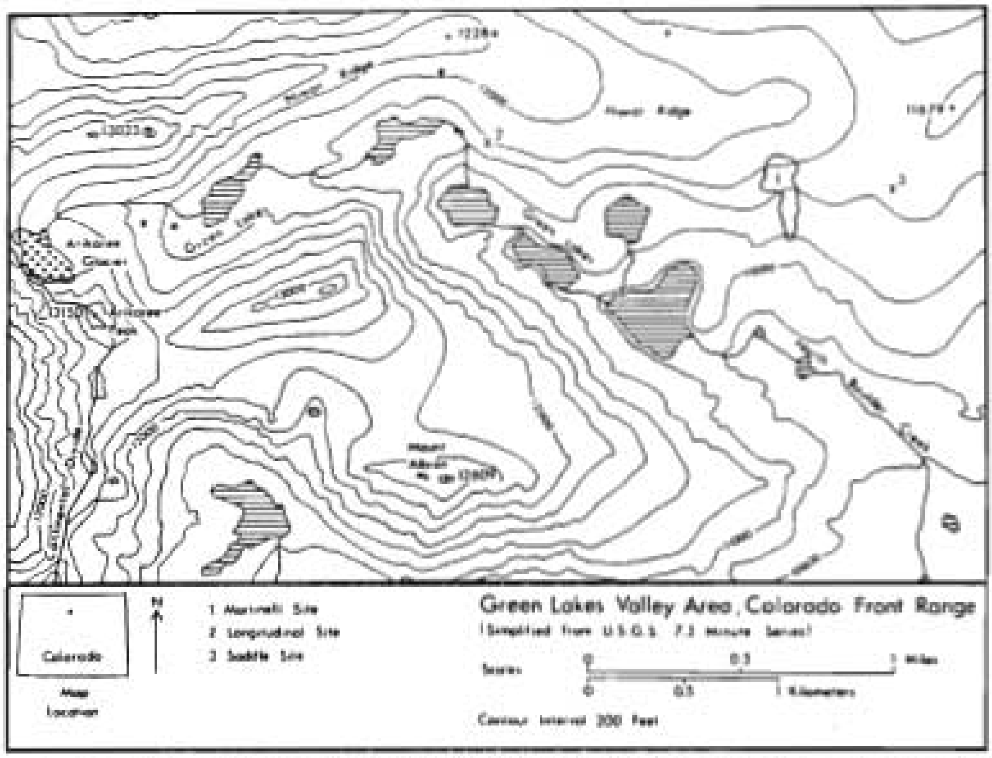

Fig. 3. Location of the three alpine research sites. The longitudinal site is the source of bedrock data; Martinelli site is the primary colluvial site; Saddle site is the subsidiary colluvial site briefly mentioned in the text. Contour interval is approximately 61 m.

Fig. 4. Looking north-west across the longitudinal site. The early June 1973 photograph shows the snow-patch near its maximum extent against the wall. The large snow-free buttress to the right of the person served as the control site.

Fig. 5. Looking north-east across Martinelli snow-patch (the alpine colluvial site). Photograph shows subdivision into upper and lower basins during melt-out.

Methodologies

The two methodologies contain fundamental similarities while differing substantially in detail. Comprehensive methodological descriptions are provided in Hall (unpublished) and Thorn (unpublished); the discussion herein is restricted to essentials.

Both studies attempted to verify nivation by testing the research hypotheses that nivation promotes: (1) mechanical weathering, (2) mass wasting, (3) chemical weathering. Each hypothesis was investigated by monitoring individual geomorphic processes. A fundamental component in the research design for the alpine program was that all measures were comparative. Each snow-patch measurement was matched by identical measurements at adjacent snow-free locations. This approach was intended to recognize nivation as a concept of intensification, rather than one of unique properties.

In both studies a bedrock freeze–thaw cycle (a purely thermal event) was accepted as a surrogate measure for a geomorphologically effective freeze–thaw cycle. In reality, there are important distinctions between the two as the latter is normally considered to be subject to important moisture constraints, in addition many researchers also envisage important freezing-intensity and/or freezing-duration constraints. However, it is not presently possible to monitor bedrock moisture content remotely, and therefore moisture conditions must be inferred.

At the Arctic site, multichannel recorders, with 2 h and 6 h sampling intervals, were attached to thermistors cemented to the bedrock face in a variety of micro-environments. The multichannel recorder at the alpine bedrock site produced spot readings at two-hourly intervals from thermistors inserted into holes drilled into the bedrock to depths of 10, 20, 30, and 50 mm (only those at 10 mm are discussed in this paper). Both records are discussed in terms of Hewitt's (1968) terminology which describes a freeze–thaw cycle and its component parts as a wave form.

Mass wasting at snow-patch sites would appear to be dominated by solifluction, overland flow, and frost creep (Hall, unpublished; Thorn, unpublished). A variety of other processes are possible, or probable, but do not appear to be quantitatively important. Amongst these are the snow-dependent processes of sub-snow creep, produced by a mobile snow-pack, and supra-snow movement, resulting from material free-falling and then rolling or sliding, or being included in surface avalanches.

At Martinelli snow-patch, overland flow was comprehensively monitored using small sediment traps (Thorn, 1976). These traps, with a catching edge of 100 mm, were distributed in groups of five and left at the site during the winter. Direct measurement of solifluction was not undertaken at Martinelli snow patch. A combination of Rudberg pillars and polythene tubes was used to monitor solifluction at Austre Okstinbredal. The hollow tubes bent with down-slope movement. Deflection was measured by inserting test tubes filled with warm jello which was allowed to cool and harden; upon removal, the angle between the jello surface and the horizontal may be used to calculate down-slope deflection. An incremental approach permits reconstruction of vertical profiles (Hall, unpublished, p. 237–46).

Both studies investigated chemical weathering by chemical analysis of snow and melt-water solutes. Arctic samples were subject to analysis by atomic absorption spectrometer; alpine samples to analysis by colorimeter. In addition, rock samples from the Arctic were examined microscopically for signs of chemical weathering, while the spatial variation in weathering-rind thickness was mapped at Martinelli snow-patch (Thorn, 1975).

Mechanical weathering

A geomorphically effective freeze–thaw cycle, that is one which disrupts bedrock, is still an uncertain concept. This is because the precise nature of the disruptive forces remains undetermined (see Hall, unpublished, and Thorn, 1979, for discussions). Most laboratory experiments have consisted of freezing rocks to varying degrees, then thawing them and noting the degree of breakdown, without investigating the mechanics of the breakdown process(es). All laboratory researchers (e.g. Fukuda, 1972; Potts, 1970; Latridou, 1971) identify rock saturation as a prerequisite, but the necessary freeze-phase amplitude remains uncertain. Potts (1970) favors the frequency with which 0°C is crossed, regardless of intensity, as the controlling factor. Most other researchers cite a specific freezing amplitude: Fukuda (1971), −4°C; Fukuda (1972), −5°C; Dunn and Hudec (1966), −6 to −10°C. Latridou (1971), summarizing the results of many workers at Caen, suggests both an amplitude of −5°C and a duration of at least 9–10 h. Finally, Battle (1960) and Mellor (1973) emphasize freezing-rate, Battle stating that a minimum cooling-rate of 0.1 deg/min^{-1} is necessary.

In part this variability reflects uncertainty as to the exact process of disruption. Principal possibilities are pressure due to direct expansion upon freezing (Mellor, 1973), oriented ice-crystal growth (Connell and Tombs, 1971), and the growth of macroscopic crystals in large pore spaces (Everett, 1961). The thermodynamics of the situation is only known in very general terms (Everett, 1961). Hudec ([1973]) has produced a series of papers which provide substantial evidence that in the majority of instances freeze–thaw weathering is, in reality, weathering by wetting and drying. Domination by one process or the other is primarily dependent upon pore size and its distribution within the bedrock. The critical problem is that in the most widely cited geomorphic literature on freeze–thaw weathering the experimental designs do not permit separation of the two weathering processes.

Data

Only the fall freeze and spring melt periods may be compared; even these must be evaluated in general terms as data are for different years, and of uncertain representativeness.

TABLE II. FALL AND EARLY-WINTER FREEZE-PHASE AMPLITUDES AT THE ARCTIC SITE (ROCKBAND 6), 13 SEPTEMBER–9 DECEMBER 1973

Cycle number	*Freeze-phase duration*	*Maximum freeze amplitude* °C
1	21–22 September	−4.4
2	4–8 October	−9.0
3	10–12 October	−6.4
4	12–19 October	−11.0
5	24–26 October	−6.9
6	1–3 November	−7.4
7	4–17 November	−13.5
8	17–28 November	−14.8
9	28 November–6 December	−16.5

TABLE III. FALL AND EARLY-WINTER FREEZE-PHASE AMPLITUDES AT ALPINE MICROSITES, 1971*

Low-amplitude cycles (freeze-phase amplitude ≤ −3.9°C)

Thermistor	*Cycle total*	*Average duration* h
11[a]	26	6.9±5.0
19[b]	26	5.1±4.2
23[c]	5	6.6±3.9
25[d]	8	5.3±4.6

High-amplitude cycles (freeze-phase amplitude ≥ 4.0°C)

Thermistor	*Cycle number*	*Freeze-phase duration* h	*Maximum freeze amplitude* °C
11	1	46	−8.2
	2	13	−5.5
	3	14	−4.1
19	—	—	—
23	1	69	−10.3
	2	16	−5.0
	3	17	−4.6
	4	16	−6.8
	5	41	−8.2
	6	22	−6.2
	7	18	−8.8
	8	19	−11.2
	9	42	−14.2
	10	16	−6.5
25	1	16	−5.0
	2	20	−4.6
	3	19	−6.8
	4	44	−8.6
	5	25	−6.6

* Precise record periods are: 14 September 1971 to 17 October 1971; 24 October 1971 to 27 October 1971; 15 December 1971 to 22 December 1971.

[a] Thermistor 11 is at the foot of the bedrock wall; snow buried in winter, easterly aspect.

[b] Thermistor 19 is at the foot of the bedrock wall in the absolute cove accumulation area; snow buried in winter, easterly aspect.

[c] Thermistor 23, exposed southerly face of buttress; always snow-free.

[d] Thermistor 25 exposed westerly face of buttress; always snow-free.

The clearest contrasts between the two bedrock sites during freeze-up appear to be the higher frequency of low-amplitude cycles at the alpine accumulation site, and the greater freeze amplitude of the late freeze-up cycles in the Arctic (Tables II and III). The former must be viewed cautiously as it is certainly in part a reflection of sampling interval (two hours in the alpine versus six hours in the Arctic). High inter-annual variations in seasonal frequency at alpine accumulation sites (thermistor 19, Table IV) reflect the difference between a dry fall with late establishment of the winter snow-pack (1971) and one with numerous temporary accumulations (1972).

Snow-free alpine sites experience wintertime freeze-phase amplitudes equal to, or exceeding, Arctic fall cycles. This is not generally true of snow-buried sites, where insulation precludes such freezing amplitudes; however, April 1972 was one of record cold, and even snow-buried sites had temperatures as low as −20°C, although without the intervening thaw phases which produce complete cycles.

Comparison of melt-out periods reveals a contrast between the two regimes. In the Arctic, sub-snow temperatures remained just below freezing, upon melt-out bedrock temperatures rose quickly without recording freeze–thaw cycles. Sub-snow bedrock temperatures in the alpine situation exhibited as many as sixty low-amplitude cycles (freeze-phase amplitude $\leqslant -1.6$°C) of brief duration throughout a six-week period preceding melt-out. Those thermistors which melted out in May experienced up to six low-amplitude (freeze-phase amplitude $\leqslant -5.2$°C, commonly much less) diurnal cycles after melt-out; those melting out in June recorded a single post-melt-out cycle at most.

Individually, the alpine record illustrates the insulating role of the snow-pack which, thereby, redistributes the seasonal pattern of cycle occurrence (Table IV). A further facet is that high-amplitude freeze phases are also precluded at snow-pack accumulation sites.

TABLE IV. SEASONAL DISTRIBUTION OF FREEZE–THAW CYCLE FREQUENCY AT THE ALPINE BEDROCK SITE

	1971				*1972*									*1972*		
Month	*Sept.*	*Oct.*	*Nov.*	*Dec.*	*Jan.*	*Feb.*	*Mar.*	*Apr.*	*May*	*Jun.*	*Jul.*	*Aug.*	*Annual total*	*Sept.*	*Oct.*	*Nov.*
Days on record	16	21	0	7	0	6	19	11	19	30	16	9	154	23	22	14
Thermistor 11*	17	12	—	0	—	0	0	0	45	27	1	0	102	10	17	2
Thermistor 19	7	19	—	0	—	0	0	0	7	9	1	0	43	14	48	7
Thermistor 23	5	6	—	4	—	2	17	6	9	1	0	0	50	5	9	10
Thermistor 25	5	8	—	0	—	1	14	6	6	1	0	0	41	n.d.†	n.d.	n.d.

* See Table III for description of thermistor microsites.
† No data from this channel.

Analysis

Evaluation of the relative potential for freeze–thaw weathering at the two sites is dependent upon the thermal criterion selected (accepting the additional critical factors of bedrock porosity and adequate rock-moisture content). If simple oscillations across 0°C, without regard to freezing amplitude, are effective, then clearly the alpine environment is more vigorous. It would also appear to have a strong seasonal rhythm, with a principal springtime maximum, and a secondary fall peak, which is particularly well developed during falls with frequent, temporary snow-pack accumulations.

Selection of any of the cited freezing amplitude criteria appears to come close to eliminating freeze–thaw weathering at the alpine site (Table V). The snow-patch probably supplies the necessary moisture, but snow-pack insulation precludes high freeze-amplitude cycles. Conversely, snow-free sites experience adequate freeze-phase amplitudes, but would seem likely to lack adequate moisture. Thus, the Arctic would be a more effective freeze–thaw weathering regime climatically. Furthermore, the season of optimal conditions shifts from the spring to the fall and early winter. This rather simple contrast is probably subject to substantial modification.

TABLE V. SEASONAL DISTRIBUTION OF FREEZE–THAW CYCLE-FREQUENCY WITH FREEZE-PHASE AMPLITUDE $\geqslant -5.0°C$ AT THE ALPINE BEDROCK SITE*

	Fall†	*Winter*†	*Spring*†	*Total*
Days on record	37	43	65	145
Thermistor‡				
11	3	0	0	3
19	0	0	0	0
23	6	26	0	32
25	5	13	0	18

* Such cycles meet Fukuda's (1972) criterion and, in addition, all met Latridou's (1971) duration criterion.

† Fall is September through November 1971; winter is December 1971 through April 1972; spring is May through July 1972.

‡ See Table III for description of thermistor microsites.

Relationships between a number of factors probably control the location of effective freeze–thaw weathering in both Arctic and alpine regimes. Bedrock porosity is the dominant control, for without suitable porosity either wetting and drying weathering will prevail or rocks will be "sound" (Hudec, [1973]), that is resistant to weathering. Further constraints are imposed by snow-pack distribution in response to prevailing wintertime winds, as without snow melt adequate moisture is unavailable. These are the two fundamental controls which create the presence or absence of freeze–thaw cycle weathering; its intensity is modified by additional relationships.

Freeze–thaw weathering intensity in any region would appear to be dependent upon the interplay between freezing amplitude, snow-pack insulation, and total direct solar radiation; all of which vary seasonally. Snow-pack depths vary the insulation afforded the underlying bedrock; in general, any moderate to deep accumulation will obliterate all diurnal variations and most synoptic fluctuations. Therefore, such sites do not exhibit wintertime cycles; certainly neither site in the present studies did so. Radiation can penetrate snow to a depth of approximately one meter (Geiger, 1961), so even if cold waves penetrate the snow-pack there is no potential for sub-snow melt once the snow-pack exceeds this depth and residual ground heat has been exhausted. Even if totally sub-zero temperature changes can exert stresses on the rock, they will occur very slowly beneath a snow-pack.

Optimum conditions are likely to be associated with deep accumulation sites in fall and/or spring, and with temporary and/or shallow accumulation sites in winter. In either case a southerly or easterly aspect is likely to maximize the weathering regime. The absence of sub-snow cycles during melt-out at the Arctic site is probably due to its northerly aspect, rather than its latitude; conversely, the easterly aspect of the alpine site undoubtedly maximized the frequency of sub-snow cycles. A further point to appreciate is that optimal conditions of high moisture and low insulation also occur immediately down-slope of melting snow-patch sites.

Widespread association between snow-patches and basal niches, such as at the Arctic site, may possibly depend upon wetting and drying cycles, or salt weathering, rather than upon freeze–thaw weathering. Location of such niches at the base of steep faces locates them in zones of maximum snow-pack accumulation, and therefore of maximum insulation and belated melt-out. Both of these factors tend to minimize high-amplitude freezing frequency. In contrast, such locations experience optimal moisture supply as water percolates downward through the rock. It is even possible that the snow produces a perched water table by maintaining freezing temperatures in the abutting bedrock, thereby directing water along an impervious surface. Given the data herein, such an hypothesis compares favorably with the classic idea that such locations experience optimal freeze–thaw weathering regimes. This latter argument would appear to be valid only if cycle frequency, regardless of freezing amplitude, is dominant.

A final point is that comparative studies at both sites indicate the inadequacy of air-temperature data as an indicator of bedrock freeze–thaw frequency, form, or amplitude. Such features in bedrock tend to be dominated by aspect and snow cover, and only the most generalized relationships can be established between air and bedrock temperatures (Thorn, 1979).

Chemical weathering

Traditionally chemical weathering has been assumed to be very low in cold environments. This is based primarily on the assumption that reactions are temperature dependent and therefore slowed in cold regimes. Tamm (1925) presented evidence that between +2 and +15°C reaction rates are invariant. Furthermore, Reynolds and Johnson (1972) propose that water and hydrogen-ion supply, not temperature, are the limiting factors. Most researchers appear to have accepted the temperature-dependency hypothesis and consequently chemical activity has received relatively little attention. Just how misleading this may be was clearly established by Rapp (1960) when he found that in Kärkevagge, Swedish Lappland, solute load was the foremost denudational process.

Data

Snow pH measurements at both sites fell into the range 4.5–5.5, which is a normal measurement for snow (Clement and Vadour, 1968). Comparisons of snow solute concentrations with those in melt waters are given in Table VI (Arctic) and Table VII (alpine). It should be noted that the alpine data are for the colluvial site. Despite the absence of overlap in measures, the data produce a unified picture of rapid solution, but of moderate intensity. Sodium at the Arctic site is assumed to be wholely atmospheric and to result from proximity to the sea. Clearly the high calcium values in melt waters indicate weathering of the calc-schists and impure marbles which outcrop at the site. The three alpine melt-water sites exhibit a trend of increasing solute load with increasing distance travelled underground. This statement, however, is based on careful surficial examination and not upon tracer studies (Thorn, unpublished).

Table VI. Solutes (p.p.m.) from precipitation, snow-pack, and melt waters at the Arctic site

Sample	*Ca*	*Mg*	*Fe*	*K*	*Na*
Rainfall	0	0.006	0	0	0.208
New snow	0.072	0.03	0	0.75	0.729
Snow-pack at 1 m depth	0.192	0.138	0	0	1.66
Waterfall, gully top	4.1	—	—	0.4	0.55
Melt water below rockband 1	7.6	0.72	0	1.8	0.833
Melt water below rockband 2	4.35	—	—	0.7	0.55
Melt water below wall (section 10)	4.8	0.42	0	1.05	0.469

Table VII. Solutes (p.p.m.) from snow-pack and melt waters at the alpine colluvial site

	Aluminum			*Silica*			*Total hardness*			*Total dissolved solids*		
	n	*x*	*s*	*n*	*x*	*s*	*n*	*x*	*s*	*n*	*x*	*s*
Dirty snow	18	0.11	0.14	18	0.10	0.31	17	2.00	3.39	18	5.27	4.03
Zero readings*	9			16			11			0		
Clean snow	16	0.08	0.00	16	0.12	0.33	15	1.20	2.36	16	3.53	2.60
Zero readings*	8			14			11			0		
Melt-water sites												
X1	23	0.20	0.10	23	6.69	1.25	23	20.86	8.23	23	26.54	1.42
X2	19	0.18	0.10	19	4.36	0.75	19	12.94	3.96	19	15.43	1.47
X3	26	0.19	0.10	26	2.03	0.76	26	7.53	5.63	26	6.48	2.34

Note: n = number in sample, x = sample mean, s = sample standard deviation.
* Zero readings indicate the number of occasions when the specified material was absent.

Microscopic examination of thin sections from Arctic bedrock samples showed staining indicative of chemical weathering. Furthermore, surface staining was widely evident in the field. A comprehensive mapping of weathering rind thickness on surficial debris was undertaken at the alpine colluvial site (Thorn, 1975). It revealed a distinct hiatus between the nearby snow-free control site and the snow-patch. Rinds were two to three times thicker within the confines of the nivation hollow, and showed distinct peaks in zones where melt-water concentration was apparent.

Analysis

Overall data suggest that snow-patch sites exhibit high regional relative rates of chemical weathering, although these may be low by absolute world-wide standards. On *a priori* grounds melt waters are likely to be inefficient in comparison to rainfall. Snow-fall is concentrated in a chemically inert state (there being no geomorphic equivalents to sheet-wash and through-flow during concentration of the snow-pack), and then released over a small portion of the landscape. This simple reduction in water–ground surface-area contact is probably the major contributor to reduced chemical weathering-rates. Certainly ground-surface temperatures quickly enter Tamm's (1924) +2 to +15°C range after melt-out.

A final point is consideration of the relative importance of mechanical and chemical denudation. A ratio of approximately one to one was determined at the alpine colluvial site, which suggests that while chemical rates may be low in absolute terms, it is possible that mechanical rates have traditionally been overestimated.

Mass wasting

Traditionally, the mass-wasting component of nivation is assumed to be dominated by solifluction. Overland flow is normally assigned a secondary role, although McCabe (1939) found no evidence of its presence. Unfortunately, the present studies provide little common ground for discussion.

Hall (unpublished) undertook qualitative investigation of the transport role of sub-snow overland flow and rivulets. He found anastomosing networks on unvegetated surfaces, which apparently shifted too frequently to permit entrenchment. On vegetated surfaces small channels became entrenched and exhibited a dendritic pattern. Particles up to coarse sand sizes were observed moving and upon melt-out sinuous ridges of fines were observed; presumably marking the locations of sub-snow channels, or transport on the snow surface.

At the alpine colluvial site, groups of miniature sediment traps were set so that they lay beneath the snow patch during the winter (Thorn, 1976). Sediment totals indicated some sub-snow transport, but a distinct peak occurred shortly after melt-out. Overland flow was dominant within a 5–10 m distance down-slope of the retreating snow margin or for a maximum of 3–7 d after melt-out. Particle sizes up to granules were transported, but competence was commonly limited to coarse sand. Within the nivation hollow sheet-wash transport-rates were one to two orders of magnitude higher than on the nearby snow-free control site; rates appeared to be independent of slope and dependent dominantly upon overland flow frequency. At a subsidiary colluvial site a continuous vegetation mat prevented overland flow, except along a frost crack.

The combined processes of frost creep and solifluction produced down-slope movement which averaged 0.02 m year^{-1} in the top 0.35 m of the Arctic debris surface. Even a sparse vegetation cover depressed the zone of maximum movement to depths of 0.05 to 0.10 m below the surface. Movement of Rudberg Pillars clearly established the presence of discrete shearing and irregular variation of movement with depth. Rates of movement were moisture, rather than slope, dependent.

A variety of other movement processes were examined at one or both sites. Snow-pack creep (Thorn, unpublished) and deflation of the unvegetated area after melt-out (Hall, unpublished) were both verified as present, but are not considered quantitatively significant. Rock fall across the Arctic snow-patch was common, although it is not possible to establish this as a nivation process *per se*.

Analysis

It is not possible to make direct comparison of the relative importance of overland flow and solifluction, either on an intra- or inter-site basis; however, some synthesis of the data may be attempted. Hall (unpublished) postulates a temporal shift in which overland flow or solifluction dominates, with overland flow dominant upon melt-out, but giving way to solifluction as the ground thaws. This may be coupled with the observation from the alpine site that solifluction lobes were generally absent from the unvegetated core area of the nivation hollow, more common at the partially vegetated down-slope margin of the hollow, and most common in a continuously vegetated zone at the extreme down-slope margin of the snow-patch. Such a sequence appears to be a spatial analog of Price's (1974) model of the self-limiting growth of solifluction lobes. Lobes grow thereby providing sheltered sites on their down-wind side which promote snow-patch development, as the lobe continues to extend so does the snow-patch; eventually the snow-patch is so large and its melt-out so belated that vegetation is precluded. At this time solifluction is superseded by overland flow and the lobe eroded.

Both temporal and spatial shifts are envisaged for the interrelationship between overland flow and solifluction. Where a continuous vegetation cover is present solifluction, dominates at all times because overland flow is all but absent. When vegetation is discontinuous or absent overland flow predominates immediately after melt-out but gives way to solifluction when the snow-patch no longer supplies adequate moisture. In the unvegetated, or sparsely vegetated zone, dominance by overland flow or solifluction is dependent upon the duration of overland flow versus the rapidity with which thaw penetrates and thereby limits solifluction. Temporally, overland flow precedes solifluction regardless of the relative importance between them. This entire picture is undoubtedly refined by the texture of the fine material available (a characteristic which may change through time as fines are transported). As overland flow is a more rapid process than solifluction, temporary storage is likely along the transition zone where overall domination by overland flow gives way to overall domination by solifluction. This phenomenon, plus the concavity associated with overland flow as compared with the convexity associated with solifluction, probably accounts for the marked low-angle apron zone immediately up-slope of a distinct convexity which is commonly found at the down-slope margin of nivation hollows.

Conclusions

These two process studies may be contrasted with the largely reconnaissance and morphological studies which produced the traditional view of nivation. As such they indicate that the concept merits further attention and some fundamental reappraisal.

First, a caveat is appropriate: nivation is not a concept which is ever likely to be constrained by a precise definition. It appears that geomorphic processes associated with a snow-patch vary in absolute terms (presence or absence of individual elements), in the relative importance of individual processes, and that both absolute and relative characteristics vary temporally at and between sites. Therefore, the term nivation should be accepted as an imprecise concept and attention focussed upon the role of snow as a driving mechanism for individual geomorphic processes.

Chemical weathering is clearly important at all snow-patches, its relative importance is very high and it may attain significant levels by world-wide absolute rates. Bedrock and surficial colluvial temperatures may be quite high during exposure to melt waters and therefore exhibit rates of chemical weathering which seem inappropriate to what is intuitively considered to be a cold environment. Conversely, snow melt waters would appear to be inherently less efficient than an equivalent amount of rainfall because they are in contact with a smaller portion of the landscape for a shorter period.

Mechanical weathering is probably controlled by bedrock porosity. If porosity is such that the rock weathers by wetting and drying cycles the significance of the snow-patch is reduced. This is because such rocks saturate from high humidity conditions alone, and do not require addition of "bulk water" (Hudec, [1973]). Rock with porosity appropriate to freeze–thaw weathering will experience optimal conditions in association with some sort of snow cover. Available data suggest that the relative mixture of seasonal freezing amplitude, seasonal snow-cover insulation and seasonal radiation receipt may produce a bewildering array of freeze–thaw weathering intensities. Some salient points do emerge from the present studies.

Fall periods which exhibit late establishment of the winter snow-pack preceded by frequent temporary accumulations probably produce optimal freeze–thaw-cycle weathering conditions. A secondary peak is likely to occur in early spring on those sites which have southerly and easterly aspects and melt-out at that time. Aspect is probably as important as snow cover in maximizing freeze–thaw weathering. Indeed, data herein may be interpreted as indicating that shallow snow accumulations, and not deep ones, are optimal for freeze–thaw weathering. The widespread idea that freezing and thawing is frequent immediately beyond a retreating snow-patch margin is not supported by the field data. Combination of available laboratory data and freezing patterns reported herein produce a conundrum. Apparently micro-environments exhibit critical shifts across laboratory defined thresholds; however, until temperature–moisture interaction during freezing is precisely known, freeze–thaw weathering remains a topic for conjecture.

The mass-wasting role of snow-patches is a much simpler topic than their role as weathering agents. A realistic view requires the integration of overland flow and solifluction in a spatially and temporally shifting symbiosis when the snow-patch precludes a continuous vegetation cover; where it does not, solifluction prevails. The transport of fines by overland flow from unvegetated core areas to peripheral, vegetated areas enhances solifluction by delivery of particle sizes most susceptible to the process.

Secondary transport mechanisms are present at snow-patch sites, although they appear to be of limited quantitative importance. The exception, at some sites, is rock fall across the

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snow-patch surface, but it is extremely difficult to establish that rock fall from a high cliff is directly related to the erosive action of the snow-patch at the base.

Snow-pack mobility, with associated basal stresses (Costin and others, 1973), and an appreciation of the limited worth of the nivation concept would seem to place snow geomorphology into closer proximity with glacial geomorphology; while the intensity may vary dramatically it is difficult to identify a threshold which separates the two geomorphically. Perhaps weathering-limited snow-patch sites and glacial sites are more readily distinguished from transport-limited snow-patch sites than anything else. Possibly the glacial environment is geomorphically dominated by mechanical processes, rather than balanced between mechanical and chemical processes, or showing chemical bias, as in the snow regime. It appears that the situation will be most successfully resolved by comprehensive work on individual processes and abandonment of the collective term "nivation"; although "nivation" might be resurrected if a more substantive definition is possible. Certainly the greatest effort must be directed toward identification of the precise mechanism(s) responsible for freeze–thaw weathering.

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Nivation and Cryoplanation: The Case for Scrutiny and Integration

Colin E. Thorn & Kevin Hall

Introduction

Tectonics aside, the absolute and relative availability of water is, perhaps, the single most important issue in geomorphology because moisture fuels the vast majority of weathering and erosional processes. If this proposition is accepted as a first approximation, periglacial geomorphology may be viewed as focussed upon circumstances where the presence of water as snow promotes an unusually high degree of seasonality, as well as spatial fragmentation, to ground moisture regimes. And where, in some but not all circumstances, these phenomena are even more sharply intensified by the impermeability produced by the presence of permafrost. Furthermore, some periglacial microenvironments are also strongly influenced by the phase change of ground and soil moisture associated with freezing and thawing. Viewing periglacial geomorphology primarily from a moisture-temperature perspective, as opposed to the traditional temperature-dominated one, provides a more appropriate conceptual linkage between it and the remainder of geomorphology. An additional, essential step is to examine ground moisture and temperature regimes directly, rather than the associated air climates. The latter are more easily investigated than the former, but are misleading: perhaps nowhere more so than in periglacial environments.

Acceptance of the perspective sketched in the preceding paragraph necessarily highlights the periglacial processes derived from snow-derived moisture, as well as the potentially ensuing landforms. Significant components of such a view already exist, in the form of the traditional periglacial concepts of nivation and cryoplanation, although both are severely debilitated by fuzzy conceptualisation, inconsistent usage, and a lack of telling fieldwork given their near-hundred year longevity. This paper attempts to build upon Thorn (1988) and Hall (1998) while also embracing fieldwork conducted in recent years. The paper is unashamedly skewed towards conceptual issues because they create the framework that steers meaningful fieldwork. This is a particularly important point with respect to nivation and cryoplanation because much (most) published work is founded upon unquestioning acceptance of the process concepts invoked originally and centered upon a (pre)determination to see or find their products in the landscape.

The approach we advocate is splendidly presented by Platt (1964). Platt suggested that some scientific disciplines advance much more rapidly than others and he sought to explain why. His conclusion was that 'strong inference' plays a central role in the successful disciplines. He identified the central elements of strong inference to be: 1) creation of alternative hypotheses; 2) creation of a crucial experiment that will exclude one or more hypotheses; 3) execution of a 'clean' experiment; 4) recycling this procedure to refine results. While Platt's ideas are not foolproof, he raises many substantive issues: not least among them in the context of nivation and cryoplanation is the notion that good ideas are worth challenging. It is the challenge that improves and refines them. By contrast, the ready acceptance and unquestioning application of seemingly self-evident truths fosters poor science and slows scientific progress. Acceptance of Platt's ideas forms the perspective from which we view nivation and cryoplanation, potentially central concepts in periglacial geomorphology, and consequently worthy of being challenged.

I Nivation

1 *A short history*

The initiating and defining article for nivation is that of Matthes (1900), while Thorn (1988) provides a comprehensive review of the research themes and associated problems stemming from the approximately ninety years of usage of the concept. Although Matthes provides a stimulating set of direct observations and inferences in his foundational paper, he does not provide a very strong formal definition of nivation; the closest he comes to a basic definition is:

> The effects of the occupation by quiescent névé are thus to convert shallow V-shaped valleys into flat U-shaped ones and to efface their drainage lines without material change of grade. These névé effects, which are wholly different from those produced by glaciation, I shall for the sake of brevity, speak of as the effects of nivation, the valleys exhibiting them having been nivated. (Matthes, 1900, p.183.)

Truly salient attributes assigned to nivation by Matthes (1900) are the prerequisite of favourable topographic features for the formation of snow patches (p. 182), low level power as an agent of landscape modification (p. 188, and elsewhere), no process continuum from nivation to glaciation (p. 183), a form continuum from nivation hollows to cirques (p. 184), and no specific claim for bedrock weathering by freeze-thaw mechanisms. As reconnaissance-level field observation, Matthes' work would be exemplary even today: as defining and/or hypothesis-testing fieldwork it simply does not match modern scientific standards (nor should it be expected to).

Fieldwork conducted in the 1970s by Hall (1975) and Thorn (Thorn, 1976; Thorn and Hall, 1980) appear to be the first occasions when Matthes' concept was subjected to detailed fieldwork focussed upon process measurement. However, it is important to appreciate that by that time nivation had already become an entrenched and commonly invoked concept within periglacial geomorphology; the veracity of which was unquestioned and stemmed exclusively from Matthes' original publication. Indeed, Matthes' original version of nivation was commonly strengthened by the additional assignment of considerable bedrock weathering power through the freeze-thaw mechanism (e.g., see Hobbs (1910) for an early example), as well as a widespread invocation of nivation hollows as precursors of cirques (e.g., Hobbs, 1910; Flint, 1957).

Hall and Thorn's (1980) results were actually generated in two totally independent studies, one conducted in Arctic alpine Norway, the other in continental, alpine Colorado, U.S.A. Results from the two studies indicated that the efficacy of freeze-thaw weathering at both locations was likely to depend upon the criteria selected to identify geomorphologically effective freeze-thaw cycles. The alpine sites experience more shallow cycles, but the Arctic site experienced a seasonal, winter freeze of greater amplitude. Both authors found the data from their respective sites to raise serious questions about the traditional freeze-thaw role assigned to seasonal snow patches. In both the alpine and Arctic locations evidence for snow patch-induced chemical weathering was apparent. The role of chemical weathering in nivation had originally been suggested by Lyubimov (1967), but his Russian-language work failed to garner any attention in the West, and was later ignored even by his fellow countrymen. Both Hall (1975, 1985) and Thorn (1976) recognized the importance of late-lying snow patches as water sources for intensification of sheetwash and solifluction in a complex pattern largely determined by the interaction between retreating snow patch margins, ground thaw, and vegetation cover.

Several Japanese workers have pursued investigations of nivation, although access to their work is sometimes only available through summary translations. Yamanaka (1979) found various forms of mass movement to be active in nivation hollows in the Iide Mountains of northeast Japan, but believed freeze-thaw weathering to be minimal. Work by Iwata (1980) in the high mountain area of Shirouma-dake in the Japanese Alps led him to emphasize the importance of slow mass movements and channel erosion in association with nivation. Iwata (1989), summarizing earlier work, emphasized the great complexity of mass wasting patterns associated with nivation, found mass transfer rates by slow mass movement increased by as much as threefold in association with snow patches when compared to nearby surfaces without late-lying snow, and highlighted the significance of small-scale variations in surface type versus the more traditional claim of associating geomorphic effectiveness with climatic variability.

A brief note by Ballantyne (1978) merits specific mention. In it Ballantyne highlights the role of snow patches overlying permafrost in raising the permafrost table to the surface beneath perennial snow patches. This has the effect of intercepting interflow and directing it through the upper layer of the snow patch, the lower layers being ice. As a result, mass wasting processes downslope of a snow patch are not driven primarily by snowmelt from the snow patch itself, but by redirected meltwaters from a much larger area.

Nyberg (1986, 1991) conducted work on nivation in both southern and northern Sweden, although emphasis was upon northern Sweden. The research site in southern Sweden was deemed to be largely inactive under the present climatic regime. At an elevation of 1200 m in the Abisko Mountains of northwestern, Arctic Sweden Nyberg found freeze-thaw activity to be very limited. As a result of his work, Nyberg emphasized the slowness of landform development stemming from nivation and increased water availability, rather than increased frost action, as the dominant geomorphic agent within nivation.

In the 1990s two doctoral dissertations (Berrisford (1992) and Christiansen (1996a)) made the most significant contributions to knowledge of nivation; unfortunately, with the exception of Berrisford (1991), the bulk of Berrisford's work remains unpublished. Berrisford's work was pursued in the Jotunheimen Mountains of central Norway, while Christiansen's work was conducted in Greenland, Iceland, Denmark, and Northern Germany; although it is the work in the Zackenberg region of northeastern Greenland (Christiansen 1998a, 1998b) and Western Jutland, Denmark (Christiansen, 1996b) that is the most fully developed.

Christiansen's (1998a) work in the high Arctic regime of northeastern Greenland provides some novel insights into nivation. This is because the Zackenberg region is cold (MAAT -9.8°C) and arid (annual precipitation 223 mm) with almost all precipitation falling as snow. The area is also low-lying and contains abundant unconsolidated sediments. Integration of all of the preceding factors creates an extremely useful 'natural laboratory' in which nivation is 'fuelled' by prevailing snow-bearing winds, winter wind speed, and the amount of precipitation falling by snow, and primarily constrained by topography and lithology. Nivation processes are also underlain by permafrost, and control local vegetation patterns in this region.

At a more detailed level Christiansen (1998a) offers a number of specific suggestions. She suggests (Christiansen, 1998a, 752) that wintertime, snow-bearing wind directions orient the nivation regime more substantially than (topographic) aspect. The overwhelming dominance of precipitation by snowfall provides an unusually clear signal of nivation processes in Zackenberg when compared to the more complex precipitation regimes present in the alpine zones more frequently used for nivation studies. The clarity of some parts of the nivation process suite is further intensified in Zackenberg by presence of weak lithologies – this presumably accelerates the rates of geomorphic development. However, the presence of permafrost within these weak lithologies makes it an important input into the nivation development sequence.

Most of the morphological components found in and around the nivation hollows of Zackenberg are summarized in Figure 1 (Christiansen, 1998a, 759, Fig. 7), and cannot be fully discussed here. Erosion of nivation hollows progresses by retrogressive slope failure at the backwall of the hollows; a mechanism that is activated by the thinning active layer at the upslope margin of the snow patch directing interflow from the snow-free areas upslope to the surface. Christiansen (1998a) postulated, from first principles, that niveo-aeolian transport is favoured in the Zackenberg region, and went on to discuss the role of observed sedimentation patterns within nivation hollows. The role of niveo-aeolian and niveo-fluvial processes, and their potential role as paleo-indicators is further developed in Christiansen (1998b). The pronival zone (Christiansen 1998a) is dominated by the interplay among sheetwash, small channels, and solifluction according to a variety of factors downslope of the snowpatch, and pronival stone pavements occur within the inner parts of some snow patches.

In sum, Christiansen (1998a) found nivation to create a characteristic landform assemblage composed of a nivation hollow containing a perennial or seasonal snow patch, an associated zone of meltwater erosion channels and/or furrows downslope of the hollow, and, sometimes, a depositional fan or accumulation basin downslope of the erosional zone under ideal conditions. However, the complete nivation assemblage may be truncated: for example, Christiansen found that steep slopes downslope of a snow patch preclude formation of secondary forms and nival sedimentation is indistinct; while the presence of bedrock slows erosion by nivation so much that distinct niches do not develop.

Undertaken in an arid low-lying environment, dominated by snowfall, and underlain by weak materials infused by permafrost Christiansen's (1998a) research in Zackenberg stands in contrast to alpine environments underlain by resistant materials, or alpine slopes mantled in till or colluvium that have provided the traditional venue for nivation studies. Consequently, it provides insight, but also lacks comparisons. The work is primarily morphological and sedimentological (or stratigraphic) and founded primarily upon visual observation rather than process measurement. The clarity of the nivation-derived signal clearly attests to the importance of water as a geomorphic agent, because although prevailing snow-bearing winds may orient the moisture source (late-lying snow), the geomorphic work is primarily water-driven, i.e., by snowmelt. Christiansen's emphasis on the channels, small valleys, and alluvial fans derived from snowpatch meltwaters accentuates the meltwater theme presented from other authors working in alpine areas (e.g., Iwata, 1980; Nyberg, 1986, 1991;Thorn, 1988; Berrisford, 1992; Hall, 1998); undoubtedly, regional aridity and soft materials combine to sharpen this theme in Greenland.

The contemporary process work undertaken in Zackenberg serves Christiansen as an analogue for an investigation of the paleo-environmental situation in western Jutland (Christiansen, 1996b). Essentially, the argument is that hollow position and orientation favour an explanation invoking snow patches as reservoirs permitting downslope development of shallow valleys during the Weichselian periglacial regime which was superimposed upon an older, Saalian glacial surface. If the Zackenberg analogue is accepted, the western Jutland landform assemblages make sense; nevertheless, the work, like all morphologic studies, is confronted by the issue of equifinality or convergence in the absence of any established single-process/single-form relationships.

One interesting issue arises from two statements in Christiansen's work, one pertaining to Zackenberg and the other to Jutland. In the Zackenberg context (Christiansen 1998a, 752) prevailing snow-bearing winds are invoked as the dominant factor in orienting late-lying snow patches; while in the Jutland context Christiansen (1996b, 123) relates snow patch size to variation in pre-existing landscape forms. The two statements are not necessarily in direct conflict, but they do raise the issue of the orientation of snowfall delivery, versus the orientation of suitable accumulation sites. By analogy with Graf (1976) who discussed the role of cirques as locations for cirque glaciers, it is interesting, and important, to raise the issue of the relationship between late-lying snow patches and pre-existing topographic accumulation sites, plus the possible role of multiple occupation of nivation hollows by snow patches. Such issues reach critical proportions when discussing whether nivation is a minor or major modifier of landscapes.

Undoubtedly, the most extensive and systematic process study of nivation is that of Berrisford (1991, 1992) who used the insights and flaws of Thorn (1974) and Hall (1975) to undertake a comprehensive investigation of nivation using sixteen snow patches in the Jotunheimen Mountains of Norway. A careful appraisal of the ground freeze-thaw regimes in both snow patch and control contexts led Berrisford to emphasize the role of the annual freeze in combination with the presence of increased moisture as the salient contributing factors to increased mechanical weathering associated with snow patches. Berrisford (1991) used the relationship between truncated weathering rinds and clast angularity to create a surrogate measure of mechanical weathering: he also clearly discussed the limitations of his assumptions. His technique strongly suggests that snow patches enhance mechanical weathering, and the specific patterns of intensification highlight the role of moisture availability. Berrisford also inferred that mechanical processes are, in fact, retarded beneath the perennial snow-ice core of larger snow patches.

Berrisford (1992) found solifluction increased by a mean of four-fold between snow patches and adjacent control sites, and his granulometric data preclude variation in grain-size as a salient explanatory variable. Sediment trap data collected by Berrisford revealed sheetwash and rivulet flow to be the dominant surficial transport processes intensified by snow patches, but these processes exhibited extremely localized variations in sediment yield. Overall Berrisford found snow patch-associated transportation by solifluction to be one to two orders of magnitude larger than that by surficial processes. Chemical weathering was found to be clearly intensified by snow patches, with a pattern that emphasized the role of meltwater inputs and concentrations: rapid increases in dissolved solid concentrations within snow patch meltwaters led Berrisford to conclude that both chemical weathering and transport are enhanced at snow patches. Berrisford also suggested the potential for chemical weathering enhancement by virtue of interaction with increased exposure of fresh surfaces produced by mechanical weathering - a theme recently repeated (seemingly independently) by Hoch *et al.* (1999).

Berrisford (1992) summarized his findings by attempting to derive basic process-response models for seasonally late-lying and perennial snow patches in debris-mantled and bedrock-backwall contexts. One of his salient findings was that the mechanical weathering regime is maximized by the combination of optimal moisture availability and strong seasonal freezing. Consequently, Berrisford believed perennial snow patches are protective. Spatially, this means that they tend to extend landform development at the downslope margin that melts seasonally and downslope of the snowpatch itself. This conclusion is supported more generally by Berrisford's strong emphasis on the role of meltwater in both weathering and transport. In the bedrock backwall context Berrisford placed emphasis on the role of the randkluft at the snow patch-bedrock interface. He also concluded that nivation is a slow landscape modifier in comparison to glaciation, but is a process worthy of much greater attention given its widespread nature in comparison to glaciation. Unlike Matthes, but in consonance with much of the later research, Berrisford suggested a blurred transition between nivation in large snow patches, versus glaciation in small glaciers.

Thorn (1988), Berrisford (1992), and Christiansen (1998) all try to culminate their research by redefining nivation in the light of their findings. Thorn (1988) suggests abandonment of the term, arguing that it reflects so many concepts and has accrued so much uncertainty that it is not functionally definable, and that its existence generates an aura of established knowledge that is misleading. Berrisford takes up the middle ground by suggesting the following definition:

> The modification (intensification or reduction) of geomorphological activity (weathering, transport, and erosion processes) on land surfaces by a seasonally late-lying or perennial snow cover.
>
> Berrisford, 1992, 440.

He goes on to shorten this to:

> The weathering and erosion of land surfaces associated with a seasonally late-lying or perennial snowcover.
>
> Berrisford, 1992, 440.

Such definitions embrace generalization and incorporate a potentially protective role for snowpatches (an important conceptual shift). Christiansen generalizes even more, offering the following definition:

> However, the wide occurrence of snow and the associated nivation processes in periglacial landscapes call for a common concept, parallel to the concept of glaciation (Christiansen, 1996a). Nivation should encompass the many individual forms, processes and sediments associated with and intensified by the presence and disappearance of snow and particularly by perennial and seasonal snowpatches (Christiansen, 1996b). Defined in this way nivation spans large diversity, and is a process association, responsible for the geomorphologic development of large parts of past and present periglacial landscapes (Christiansen, 1996b).
>
> Christiansen, 1998a, 751.

This definition embraces the view that a definition 'does not require a knowledge of individual rates of each single nivation process' (Christiansen, 1996b, 112). Christiansen (2001, personal communication) would also include avalanche activity within nivation.

In effect, Thorn states we know too little about too many diverse things to have a specific term, while Christiansen saves the term by abandoning the (original) concept. Depending upon whether such views are regarded as linear or circular, Thorn and Christiansen are either at opposite ends of the spectrum or very close to each other, being separated by a term rather than a concept. The fundamental question is to ask 'why do we have scientific terminology?' The answer is that we have it as a form of shorthand that permits individual scientists within a group sharing common interests to communicate with each other clearly and quickly. Defined as suggested by Christiansen, nivation becomes the terminological equivalent of 'nival' and is even less specific than when originally coined by Matthes. Equally importantly, it fails to penetrate the research quagmire represented by the present uncertainty concerning the nature, absolute rates, and relative rates of individual constituent weathering, erosional, and depositional processes. Reducing nivation to a generalization equivalent to glaciation may rescue the word, but it does nothing to sharpen our grasp of the issues at hand.

The definitional issue is not trivial or mere semantics, our field research is ultimately steered, knowingly or unknowingly, by our conceptual or theoretical expectations. A large portion of our expectations are embedded in terminological definitions, the sharpness of these definitions reflects the sharpness of our thinking; the sharper our thinking the greater our ability to extract information from what are clearly complex landscapes.

2 *The present situation – a summary*

Nivation is inherently a secondary process term encompassing primary components of weathering, transport, and deposition. As such, it is subject to modification both by virtue of changes in the constituent primary concepts as well as in revision in the way the term nivation itself is defined. Modern research has produced important shifts in the way researchers investigating nivation view it at both the primary and secondary levels.

The transport component of nivation is not particularly contentious. Nivation researchers have not invoked any new primary transport processes. A broad view of geomorphology suggests that much landscape development is controlled by the relationship between available energy and surface resistance. In the specific case of nivation much of this relationship is captured by the notion of abundant moisture availability as the energy source, and variability in surface resistance being in the form of resistant rocks versus soft sediments, and within the latter category between vegetated and unvegetated surfaces. In some instances surface resistance is clearly defined by the absence or presence of permafrost. As nivation is founded exclusively on the presence of late-lying snow (except in the case of Christiansen's radical redefinition of the term) it is always a source of moisture; consequently, it is surface variability that modifies nivation transport most obviously. Resistant rocks weather slowly and produce little material, but soft sediments or colluvium are readily transported. In such contexts vegetated versus unvegetated surfaces provide a sharp contrast. It appears that a vegetation cover all but eliminates surficial transport and solifluction dominates. However, unvegetated surfaces may experience surface wash, even channelized flow in particularly large-scale instances. Using well-accepted periglacial processes, nivation researchers are simply confronted by determination of complex spatial and temporal transport process mixes and their potential significance for landscape development. The variable proportionality of the transport mixes reported in the literature should not be considered particularly problematical. It is almost certain that the process mix varies temporally at a place through the melt season and from season to season; it is also probable that the mix varies spatially at any one time, both within individual sites and from site to site.

The weathering component of nivation is infinitely more contentious than its transport counterpart. This is because primary mechanisms are in question and nivation researchers have contributed to primary research in several instances. The most apparent weathering question centers upon the mechanisms of mechanical weathering to be found in periglacial regions. This topic is taken up at a fundamental level in Hall *et al.* (this volume). One quintessential theme in periglacial geomorphology, to which nivation researchers have contributed, is freeze-thaw weathering. Revelation of actual field conditions in and around late-lying snowpatches has thrown light on a context in which widespread freeze-thaw weathering historically has been invoked. Field data have frequently been found to be at odds with conditions assumed by those relying exclusively on air data, or indeed no data; and field findings have also provided serious challenges to those trying to mimic nature in the laboratory. In general, data from nivation researchers have tended to shift emphasis away from short-term cycles and towards seasonal freezing, while simultaneously placing much greater emphasis upon moisture availability.

Increasing emphasis upon moisture availability and its role, as well as upon ground, rather than air, temperatures has also led nivation researchers to place increasing emphasis upon chemical weathering (see Hall *et al.*, this volume). It is now abundantly clear that chemical weathering is an important contributor to periglacial weathering. This is not the same as saying that periglacial chemical weathering rates are as high as tropical rates, the statement is a relative, not absolute, one.

Landform development associated with nivation processes has also come under scrutiny, and several interesting ideas have emerged. For example, Berrisford (1992) emphasized development of nivation hollows by extension downslope of existing snowpatches, while Christiansen (1998) emphasized headward expansion upslope. Berrisford's work is in a region of resistant rocks with debris-mantled slopes, Christiansen's work is in a region of poorly consolidated materials infused with permafrost. Both ideas may well be correct, and there is certainly no reason to see them as necessarily conflicting. Another extremely interesting possibility has been posited by Rapp (e.g., 1983, 1984); his context is that of the increasing conviction of Scandinavian geomorphologists that much of Scandinavia experienced a regional cold-based ice cover that was essentially protective. Under these circumstances Rapp has raised the idea that interglacial periods of active nivation may, in fact, be the primary erosional and land-forming phases in regions where cold-based ice predominates.

Researchers of nivation are now proposing modification of its definition. For example, suggestions that perennial snow patches are protective and also that there is no sharp break between nival and glacial processes represent profound conceptual shifts, even if they are not yet fully articulated operationally. The rethinking of the relationship between freeze-thaw weathering and late-lying snow is long overdue, but is not yet well defined. It is quite clear that nivation is now being scrutinized in the light of a sound field data base and no longer being presented as a well-understood land-forming agent of unquestionable power – a situation which had developed, but was far from Matthes' carefully constrained initial presentation of the concept.

II Cryoplanation

1 *A short history*

The term cryoplanation was introduced by Bryan (1946) in a paper that was centered upon conceptual clarification embracing etymological niceties: some suggestions faired well, e.g., cryoplanation; some, e.g., congeliturbation, faired poorly. Recognizable versions of the 'cryoplanation' concept may be traced back to Cairnes (1912a, 1912b) and Eakin (1916). Cairnes promoted the concept of 'equiplanation', 'all physiographic processes which tend to reduce the relief of a region and so cause the topography eventually to become more and more plain-like in contour, without involving any loss or gain of material' (Cairnes, 1912b, 76). Cairnes presented equiplanation as a descriptive term that is azonal; he then highlighted (Cairnes, 1912b, 78) the infilling of bedrock hollows with debris that is quickly frozen in parts of the Yukon and Alaska as one illustration. Subsequently, he identified equiplanation as particularly important in semi-arid and arid regions, as well as pointing out that Matthes' (1900) nivation concept appears to embrace equiplanation tendencies (Cairnes, 1912b, 82). Eakins (1916, 77) introduced the terms 'altiplanation terrace' and 'solifluction slope' as readily identifiable landforms ensuing from a broad mix of mass wasting processes present in Alaska. His description of altiplanation terraces (Eakin, 1916, 77-82) incorporated distinctive breaks of slope at the inner edge of the terrace, gradational surface debris characteristics from inner to outer terrace, and an outer edge break in slope. In short, Eakin described a cryoplanation terrace as commonly understood, emphasizing the role of solifluction, but only comments *en passant* (Eakin, 1916, 82) on the weathering regime producing the associated scarps. Bryan recognized Eakin's idea, but dismissed his work summarily from further discussion because it is 'confused' (Bryan, 1946, 638). Bryan (1946, 639) introduced 'cryoplanation' as a climatic accident, or deviation from, Davis' normal Cycle of Erosion (i.e., peneplanation). The formal definition (Bryan, 1946, 640) is: 'cryoplanation = land reduction by the process of intensive frost-action, i.e., congeliturbation including solifluction and accompanying processes of translation of congelifracts. Includes the work of rivers and streams in transporting materials delivered by the above processes.' Bryan failed to cite Matthes in either his 1946 or 1949 papers: consequently, nivation *per se* is not invoked as a constituent part of cryoplanation. By way of summation, using terms with their generally accepted meaning: conceptually altiplanation is identical with cryoplanation, but equiplanation is a generic descriptive term of which altiplanation/cryoplanation is a specific genetic example.

A high proportion of the ensuing work on cryoplanation has been pursued by Czech, Hungarian, Polish, and Russian researchers, their work, published in their mother tongues, is unavailable to us because of our linguistic limitations. Of the work published in English by these groups the comprehensive monograph by Demek (1969) is the benchmark. A more recent summary and review by Czudek (1995), while cautionary and thoughtful, brings little new data to the debate. The title of Demek's monograph is a faithful reflection of its contents 'Cryoplanation Terraces, their Geographical Distribution, Genesis and Development'. Demek presented a largely descriptive, view of cryoplanation, albeit based on enormous personal experience. Cryoplanation terraces are initiated by nivation, involving either headward incision of transverse (Lewis, 1939) snow patches or coalescence of small circular snow patches. Terrace expansion is dependent upon the weathering component of nivation (the additional invocation of frost action appears superfluous) causing retreat of the inner scarp, mass transport on the developing terrace surface is seen as dependent on the entire welter of periglacial mass wasting processes in site-specific proportions. Terraces are not bedrock and/or structurally dependent, but may well be steered by local bedrock and/or structural patterns (e.g., see also Czudek, 1995). For example, some rock types resist erosion and, consequently, may exhibit small and/or infrequent terraces, and, while all cryoplanation terraces are erosional and may cut across rock boundaries, variations in structure may play a role in terrace initiation and/or shaping of terraces. The relationship between terraces and past geomorphological history appears to vary, terraces are particularly abundant in regions which appear to have experienced particularly lengthy post-glacial histories (Czudek, 1995), but may also appear in topographic positions that clearly exhibit post-glacial development by cross-cutting relationships.

Four obvious questions emerge from Demek's monograph with respect to the present focus: 1) is there an independent cryoplanation process subsequent to initiation by nivation? 2) what are the genetic and morphological differences between cryoplanation terraces (or benches) and pediments? 3) what is the rate of development of cryoplanation terraces and pediments? 4) what is the age of cryoplanation terraces and pediments? The answer to the first question appears to be 'no'; the answers to the last three questions are simply unknown.

Despite a steady trickle of papers on cryoplanation very little, if any, progress has been made in either the conceptual framework or the production of long-overdue field measurement of processes. Reger and Péwé (1976) clearly invoked nivation as the driving force throughout the development of cryoplanation terraces. A thoughtful review by Priesnitz (1988) deftly highlighted several of the most pressing issues; while a paper by Nelson (1989) attempted to demarcate the mesoclimatic or regional climatic elements that constrain cryoplanation terrace development. A particularly intriguing comment by Priesnitz (1988, 64) is his suggestion that a periglacial landscape evolves through time and that cryoplanation terraces are indicative of a 'mature' periglacial landscape, this concept constrains cryoplanation terraces temporally as well as climatically. A similarly novel suggestion is the one by Nelson (1989, 39-40) that not only are terraces constrained by snowline, but that they are, in fact, the periglacial equivalent of a cirque glacier, i.e., Nelson suggested that terraces are about an order of magnitude smaller than cirques and form by a different set of processes to cirques, occurring where cirque glaciation is precluded by inadequate snow accumulation to generate a glacier.

Hall (1997) has addressed 'cryoplanation' in what appears to be a presently active environment in Antarctica. The relatively small-scale terraces he examined were developing in bedrock in an apparently paraglacial context. Detailed field inspection, and air and ground temperatures from the immediate locality, provided Hall with an opportunity to assess terrace development with respect to individual weathering processes, e.g., dilation cracking, freeze-thaw, and thermal contraction. While no definitive assessment of the relative importance of individual weathering could be made, integration of weathering observations, terrace size, and terrace orientation provide the most comprehensive scrutiny of apparently active 'cryoplanation' yet. Hall's observations led him to examine prevailing concepts and terminology closely (Hall, 1998). He questioned the separate identity of nivation and cryoplanation, pointing out that nivation is, in fact, the only process suite actually invoked in both nivation and cryoplanation landforms. Nevertheless, nivation is a concept embracing multiple weathering and transports processes, consequently, Hall invoked variable contributions from the suite of individual weathering and transport processes –with variations stemming from position along the dry-wet spectrum. He suggested that the literature appears to invoke nivation in wet contexts and cryoplanation in dry ones. As he pointed out, morphology alone does not preclude the possibility that any individual landform has occupied different positions on the dry-wet spectrum during its development. Overall the rate of development would be expected to correlate positively with increasing moisture availability.

2 *The present situation – a summary*

Cryoplanation as introduced and defined by Bryant can only exist if the researcher accepts a Davisian model of landscape development. However, the term is clearly presently used by many researchers in a truncated fashion to describe two types of landform believed to be found widely in periglacial regions, namely: cryoplanation terraces and pediments. Both terms are morphogenetic, i.e., they purport to describe both form and genesis. However, the only genetic elements invoked are the weathering components of nivation as the initial erosional driving force at the headwall or scarp, and commonplace periglacial mass wasting processes on the surface of the tread. In the case of pediments fluvial processes are also invoked along the downslope margin, but no atypical fluvial requirement is invoked.

The present weathering component in the nivation/cryoplanation mix appears to be a classic invitation to wield Occam's Razor. In the absence of any evidence whatsoever to the contrary the first approximation should be to assume that nivation hollows and benches, and cryoplanation terraces and pediments are erosionally driven at all times (not just initially) by nivation as broadly re-conceptualised in the preceding section. Variation in developed form is not difficult to accommodate, at least conceptually. Presumably, pre-nivation form does much to control the form of nivated landforms. A further control is likely to be the shape of the moisture supply source: by analogy with solifluction a point source is likely to produce a circular (lobate) form, and a linear moisture source will produce a bench (terrace) form. While solifluction is a transport phenomenon and the portion of nivation under discussion is erosional, there is nothing inherent in this contrast to counter the suggested form relationship, despite the fact that solifluction forms with a point source are convex downslope while erosional forms from a point source are convex upslope. Given a single erosional mechanism with variable form explained, the transport elements may be addressed.

Despite a dearth of mass wasting data actually generated upon the treads of cryoplanation terraces the need to move debris across a gently sloping surface in a periglacial context provides no substantive conceptual difficulty. Any number of instances in the literature, including the comprehensive reviews of Washburn (1980) and French (1996), clearly indicate that mass transport on low gradients by solifluction and commonly associated processes occurs widely. The presence of patterned ground on the surface of cryoplanation bench treads and on pediments has often been raised as a potential indicator of contemporary inactivity. This may or may not be true: very slow transport would not preclude patterned ground forms, although it may elongate them (a trait easily accommodated conceptually by comparison of the elongation of stony earth circles to stripes on a steep slope). However, other ideas of pediment development, see Twidale (1983) below, may not even necessitate such adjustment. With respect to pediments specifically, invocation of fluvial removal at the base of the pediment is non-controversial with respect to the fluvial mechanisms *per se*. However, it does raise the issue of the relative rates of headward incision versus trimming at the downslope margin, and it also raises the issue of overall speed of development. Do cryopediments evolve more rapidly than cryoplanation terraces because fluvial removal accelerates throughput?

While transport by periglacial mass wasting across individual cryoplanation terraces or benches appears feasible, the transport of debris across or from sequences or sets of terraces and/or benches is much more problematic. The problem introduced by considering a sequence is the need to move ever-increasing amounts of material across features in the lower portions of slopes or hillsides. Such an issue must arise as the upper features can only transport material to forms developed downslope. Tentative proposals (e.g., Reger and Péwé, 1976) that infer that debris from cryoplanation terraces may be shed laterally do not provide a convincing, potentially universal, explanation of the disposal of transported debris. A potentially universal explanation must presume that the bulk of the debris travels directly downslope. If large, and/or numerous, cryoplanation surfaces are invoked by researchers, as is often the case, then the debris issue must be confronted at a substantial scale; perhaps, as a first approximation, in the same fashion as a glacial geomorphologist would view the relationship between glacial erosion and moraine deposition. At a conceptual level, the only potential explanations that would appear to sidestep the presence of large quantities of debris somewhere in the adjacent landscape would be to invoke enormous time spans for formation (thereby permitting very slow removal) or to envisage removal by chemical solution. All other mechanisms must surely encompass recognition of large debris accumulations spatially distributed in accordance to gravitational forces; unless intervention by some completely unrecorded geomorphic regime is invoked. The possibility of the role of dramatically different geomorphic regimes is raised by the conflict between those that see permafrost as essential for cryoplanation (e.g., Reger and Péwé, 1976), versus those that do not (e.g., Demek, 1969; Czudek, 1995). However, it is difficult to see how permafrost changes the nature, as opposed to the rate, of periglacial mass wasting.

Cryopediments, as opposed to terraces or benches, must be viewed, at the most fundamental level, in the context of the pediment literature: French and Harry (1992) provide an insightful discussion of cryopediments as pediments and the associated conceptual difficulties. This, of course, is so large that it simply cannot be addressed comprehensively here. It is important to note that the pediment literature is primarily a hot desert literature, but this provides no serious obstacle in itself. First, and foremost, the emphasis is upon the notion of 'desert' – i.e., locations where absolute and relative moisture availability is of paramount importance. Certainly, many colder periglacial environments are also clearly deserts, albeit cold ones. It seems likely then that the central issues in cryoplanation pediments fall primarily into the domain of deserts and only secondarily into the periglacial domain.

Twidale (1983) provided a particularly incisive contextual summary of pedimentation. He subdivided pediments (Twidale, 1983, 13-17) into the mantled and rock pediments of crystalline outcrops versus the covered pediments of sedimentary regions. Mantled pediments are fragile and transitory and characterized by veneers that are weathered *in situ*. Rock pediments are stable forms often viewed as the product of scarp recession, but Twidale noted that there is compelling evidence to view many of the scarps as stable and the product of major fracture zones. Covered pediments carry a veneer of clasts delivered from the abutting uplands that simultaneously erode the underlying bedrock and deposit the mantle of debris. Twidale (1983, 15) stated that such forms are azonal and exist whenever weak rocks that are adjacent to uplands (a source of resistant clasts) are subject to a climatic regime experiencing large variations in discharge. He specifically identified springtime snowmelt as one such example. It is quite clear that cryoplanation pediments are readily embraced within Twidale's remarks. An important attribute of such a context is that Twidale only invoked modest erosion of back scarps, such a claim is particularly relevant because nivation is often only seen as a modest modifier of the landscape, yet large cryoplanation pediments and surfaces seem to necessitate extensive erosion and scarp retreat.

III Conclusions

Research of nivation is presently in a significantly healthier posture than at any time since Matthes coined the term: the same cannot be said for cryoplanation. The explanation for the dichotomy is two sides of the same coin, nivation has come under scrutiny, while cryoplanation has not. Scrutiny of nivation has come in the form of direct field measurement of many constituent processes, as well as by expansion of fieldwork into different geomorphic contexts.

Process investigations of nivation must continue to focus upon weathering. The failure of periglacial geomorphologists to determine well-founded general models of mechanical weathering, and generally to continue to ignore chemical weathering, are critical flaws in the subdiscipline at large. Given the pervasive role of snow-derived moisture in periglacial geomorphology, this problem cannot fail to be critical to nivation as well. Once the nature of periglacial weathering processes is determined, it will be critical to make attempts to determine their rates, and the spatial and temporal variability thereof.

Transport processes associated with nivation have never amounted to more than invocation of intensification of widespread periglacial processes.

Refinement has occurred as workers have monitored such processes. In general, there has been increasing emphasis on the role of running water in the absence of vegetation: matched by identification of the minimal impact of surficial water in the presence of a vegetation mat. In situations where soft materials form the surface, aeolian processes appear to be more significant than in the more traditional alpine zones of study.

The addition of low-lying, arid periglacial environments underlain by weakly resistant materials to nivation studies has provided a new slant on the topic. A sharper signal emerges because water drives most surficial processes, and late-lying snow (nivation) provides the dominant water source; plus non-resistant materials permit rapid landform and landscape development. This work has also added a depositional record to this particular context. The critical question at this juncture is what does this simple, clean signal tell us about other nivation contexts? Ultimately, a sound answer will depend upon appropriate field measurement programs in both contexts.

Virtually all recent researchers of nivation has emphasized the pervasive role of snowpack-driven geomorphology. Much of this extends beyond the traditional role assigned to nivation *per se*. However, if valid, this viewpoint raises important topics meriting immediate attention. One theme is clearly orientation of the geomorphic development of a landscape. Such development will be a product of the orientation of prevailing, snow-bearing winds interacting with pre-existing accumulation sites. In short, initiating contexts for nivation need investigation.

While process studies are essential to determine the nature of nivation, the scale-linkage difficulties that pervade all process studies mean that direct upward scaling to assess contemporary landforms will be difficult, probably impossible. In very specific contexts, such as Zackenberg, the depositional record provides some opportunity to assess landform development directly. This will require not only knowledge of sedimentation rates, but volumetric estimates as well. In most locations where resistant materials are present erosion rates will be too slow and depositional signal too complex to permit direct measurement. In these instances, once the nature of constituent processes is known, numerical modeling including sensitivity analysis of rates affords an appealing avenue of advancement.

At the landform scale, initiation and development rates are critical issues in assessing the relative importance of nivation, and/or the seasonal and perennial snowpack more generally, in landscape development. If glacial geomorphologists embrace the notion of the occurrence of both protective cold-based ice and erosive warm-based ice, then there will inevitably be fundamental reconsideration of the history of many glaciated landscapes. Snow-related geomorphology, however termed, will have to be considered in the developmental mix.

Precedence has been given to nivation over cryoplanation in this discussion because those using the cryoplanation concept have overwhelmingly invoked nivation as their causal agent. To deny the fairly widespread existence of features commonly called cryoplanation benches, terraces, and pediments would be foolish: to claim that there is anything approaching an adequate explanation of their origin(s) would be even more foolish. If nivation is a secondary concept, then cryoplanation is a tertiary one. There is little to say about cryoplanation other than the concept needs to be challenged directly. What distinguishes cryoplanation from nivation? Are cryoplanation benches, terraces, and pediments merely form variants, or is there a fundamental genetic difference between them? If the latter is the case how are they embraced by a single genetic concept? There are abundant, nearly flat surfaces in periglacial landscape, snow commonly accumulates at breaks in slope: can cryoplanation take these two realities and offer us a viable explanation of some portion of landscape development? Hall (1997, 1998) represents the beginning of such a challenge, but very much more is required.

Finally, some thoughts on the definitional conflict emergent among present-day students of nivation. Our terminology, as the artifact of our theory, is hierarchical. Pushing nivation as a term up the hierarchy (i.e., generalizing it) does nothing to address the concept(s) it purportedly embraces. As redefined by Christiansen nivation essentially becomes a synonym for nival, that is it encompasses no new concept. Conversely, we are still left with the question 'what are the impacts of late-lying snowpacks?' Without the term nivation we still have the impacts, but we do not have a collective term. The intellectual question is obviously 'are the impacts of late-lying snow sufficiently consistent to constitute and merit a definable term?' A positive answer would seem to require at least clear identification of the nature of the constituent processes; a more stringent approach might also require specification of their rates (at least this is implied by the original use of the notion of intensification). For now the debate itself is a success, it reflects the presence of the scrutiny called for in our title. As for the 'integration' in our title, it is clearly a challenge to students of cryoplanation to scrutinize their concept and validate its independent existence.

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Rock Weathering and Soil Formation in the Maritime Antarctic: An Integrated Study on Signy Island

An integrated approach is described to the investigation of the processes of soil formation and mineral nutrient release in a recently deglaciated area with a polar climate. The study is interdisciplinary and seeks to relate the traditional approaches of geologists, pedologists and geographers to provide a continuum view of soil development in primitive, poorly structured soils. The studies of rock weathering encompass geochemistry, mineralogy, fabric breakdown by physical, chemical and biological attack, and both field and simulation studies of process. Studies on soils and nutrient release incorporate fabric analysis (including micromorphology, chemistry and particle size), clay mineralogy, microclimate and the ionic composition of the soil water. To simplify data synthesis the main studies are concentrated on three quartz-mica-schist sites showing different states of vegetation development.

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Rock weathering and soil formation in the maritime Antarctic: An integrated approach for Signy Island.

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ABSTRACT

An integration approach is described to the investigation of the processes of soil formation and mineral nutrient release in a recently deglaciated area with a polar climate. The study is interdisciplinary and seeks to relate the traditional approaches of geologists, pedologists and geographers to provide a continuum view of soil development in primitive, poorly structured soils. The studies of rock weathering encompass geochemistry, mineralogy, fabric breakdown by physical, chemical and biological attack, and both field and simulation studies of processes. Studies on soils and nutrient release incorporate fabric analysis (including micromorphology, chemistry and particle size), clay mineralogy, microclimate and the ionic composition of the soil water. To simplify data synthesis the main studies are concentrated on three quartz-micaschist sites showing different stages of vegetation development.

KEY WORDS

Maritime Antarctic	Weathering	Soil Formation
Integrated Approach	Synthesis	

INTRODUCTION

The formation of any soil (except peat) is initially dependent on the weathering of parent rock both physically, to provide a range of particle sizes, and chemically, to allow the release of various ions by the formation of new minerals. A concomitant of these processes as far as soil development is concerned is the incorporation of organic matter into the mineral soil system. There are few detailed data available on processes, rates or even the role of particular groups of organisms in these initial stages for polar soils. Yet, in the polar regions glacial retreat is producing a wide range of surfaces (from glacial flour to massive rock) for colonisation. Climatic warming is likely to accelerate this process.

Traditionally most studies of weathering have been undertaken by geographers and have even then usually been limited to the study of a particular process. In the studies on Signy island we wish to address the problem of weathering on a wider front but within a single environment. This paper will present a brief overview of the structure of the investigations and the range of methods used together with some preliminary results.

We wish to establish in a cold, wet climate the types and rates of mechanical weathering of one major rock type (quart-micaschist) both by field measurements and by laboratory simulation. We intend to consider the role of inorganic chemical weathering, particularly in the context of a small polar oceanic island, as well as the significance of biological weathering. Biological attack on rock will be investigated in terms of mechanical disruption by lichens and fungi, as well as chemical changes because of the effects of organic acids.

Data from these fields will be related to the present development of soil structure to provide a preliminary assessment of the rate of soil formation from parent rock, whilst chemical data from analyses of soil minerals and water will be added to this to construct a model of mineral cycling for a primitive polar soil ecosystem.

SIGNY ISLAND

Signy Island (Lat. 60°43'S., Long. 45°38'W), South Orkney Islands, is roughly triangular in outline, 8km long by 5km wide, with a maximum elevation of 276m (Fig. 1). It now has only a small ice cap, although at the glacial maximum it was clearly overridden by the South Orkney ice sheet. More than half the surface of the island is now snow free in summer with large areas of bare rock talus and glacial deposits. Glacial erosion is evident in the landforms with numerous cirques around the margins of the high ground and extensive till and outwash deposits on the lower ground (Holdgate, Allen and Chambers, 1967), the whole underlain by permafrost.

Geologically the island is composed of regionally metamorphosed sediments, principally quartz-micaschist with additional amphibolites and marbles (Matthews and Maling, 1961). Signy Island is part of the Scotia metamorphic complex (Tanner, et al., 1982). The exact tectonic history of the island is at present subject to differing interpretations (Dalziel, 1984; Meneilly and Storey, 1986).

Within the Basement Complex of quartz-micaschists, amphibolites and marbles there is a limited range of minerals (Thomson, 1968). Quartz is virtually ubiquitous throughout the Signy rocks, whilst plagioclase hornblende, chlorite, muscovite and biotite are all very common and widespread. Epidote and garnet are locally common and there are a few accessory minerals (sphene, graphite, apatite, tourmaline, calcite, and haematite). Storey and Meneilly (1985) presented the first detailed mineral chemistry for Signy Island.

Although much of the lowland till cover is probably of local origin it is likely that some is derived from Coronation

Island. Marble fragments occur scattered throughout the till. There are few well-defined morainic series but present recession of the icecap is producing new moraines in some areas. There is widespread evidence of solifluction/gelifluction and periglacial activity.

With the exception of areas close to marble or amphibolite outcrops the soils are derived from direct weathering of quartz-micaschist and from till deposits. On Signy Island frequent precipitation and high humidity throughout much of the summer maintains saturated conditions in the accumulations of mineral soil and rock fragments. It has been suggested that water penetrates along the foliation of the schists and on freezing, breaks the rock into a mass of platy fragments (Holdgate et al., 1967). Sorting of the matrix by solifluction and frost heave produces local concentrations of silt and clay size particles (Chambers, 1966a). Allen and Northover (1967) suggest that although physical weathering is dominant, chemical changes are also significant. In support of this they found evidence for K replacement in the mica lattice by Na, a decrease in the Al/Si ratio from schists to clay and the presence of free sesquioxides. Incorporation of organic matter into the mineral soil is minimal, although highly organic soils do exist on Signy Island (Allen, Grimshaw and Holdgate, 1967).

TECHNIQUES

MECHANICAL WEATHERING

A number of approaches, both in the field and the laboratory, are being used to achieve a data matrix. This approach was considered desirable for two reasons. First, the interactions between processes required a framework for organizing disparate data. Secondly, the dominant rock type on Signy Island, quartz-micaschist, is one for which there is only limited background information available (Deere & Miller, 1966; Fahey, 1983; Myrianthis and Leach, 1978; Brown, et al., 1980). As the characteristics of porosity, microporosity, saturation coefficient, water content and shear strength are very important with respect to rock breakdown it was necessary to investigate these parameters under both field and laboratory conditions (Hall, 1986a and 1987).

Evaluation of compressive strength, giving a derived estimate of tensile shear strength of the rocks, was undertaken by means of a point load tester (Broch and Franklin, 1972). In the case of the anisotropic schists, strength values were obtained both parallel and transverse to the schistosity (Hall, 1987). In the field use was made, where possible, of the Schmidt Hammer (Day and Goudie, 1977) to obtain rock strength values. These values were integrated into Rock Mass Strength (RMS) estimations (Selby, 1980) to gain some indication of the stability of rock faces (Hall, 1987). In the laboratory, microscale relative measurements of strength and degree of weathering on small flakes was undertaken by means of a cone indentor (Szlavin, 1974; Hall,1987).

Rock water content, a factor of very great importance for both mechanical and chemical weathering, was obtained for numerous rocks under a variety of different conditions (Hall, 1986a). The water contents were calculated, after drying for 24hr at 105°C, for samples used in point load tests, together with additional material collected from meltwater rivulets, under snow, vertical rock faces, horizontal rock faces, and from the sea bed. These values of naturally occurring water contents were then used to set realistic conditions in simulation experiments (Hall, 1986b, 1988a). In addition a single rock specimen, kept outside the field laboratory, was weighed daily throughout a year (Hall 1988b).

In the laboratory the porosity, water absorption capacity, saturation coefficient and microporosity of the rock samples were calculated following the procedure of Cooke (1979). These same properties were also calculated for 85 rock tablets cut from the main Signy rock types. The tablets were positioned at three study sites during the 1983-4 summer and several retrieved each year for reassessment of their properties (Hall, 1990). Some of the tablets were aligned with schistosity normal to the ground, in order to investigate whether this physical attribute was of importance to mechanical weathering.

Laboratory simulation of mechanical weathering was undertaken in a computer-controlled climate cabinet designed and built at the University of Natal (Hall, et al., 1989). This enables highly complex cycles to be devised including the ability to produce long, non-repetitive sequences. It also facilitates the exact repetition of any sectors of the cycle which appear to be significant. A 'Pundit' ultrasonic rock-tester is incorporated within the system to provide real time data on internal changes in rock integrity whilst a fiber optic crack detection system is used to monitor for external manifestations of rock failure. The stored data are analysed via separate software on another CBM64 system. Results to date (Hall, 1986b,1988a, 1988c; Hall, et al., 1989) have provided many new insights into the weathering processes and rates for quartz-micaschist. Amongst these are such factors as the influence of anisotropy upon freeze penetration, the role of wetting and drying within the freeze-thaw mechanism, and the finding that the nature of the water to ice phase change is a function of rock moisture content, solute concentration, freeze amplitude and the rate of change of temperature.

INORGANIC CHEMICAL WEATHERING

Chemical weathering of the rock may be an important process. Rock interstitial water is derived from precipitation. Summer rainfall and winter snow were analysed for their anion content using an ion chromatograph, and for elemental content by atomic absorption spectrophotometry. Whilst the presence of sodium, chloride and sulphate are of principal interest in direct weathering processes, the presence of nitrogen and phosphate are significant for indirect weathering because of their importance in plant growth, both on and within rock material.

Using a plasma atomic absorption spectrophotometer an attempt has been made to investigate what salts might be available inside the rock to effect mechanical weathering. This approach has a number of shortcomings but the data do give an indication (for simulation studies) of which salts might actually be available, and the total data are useful for comparison with results from chemical analyses of the soils (Hall, Verbeek and Meiklejohn, 1986).

In the soil the development of a series of clay minerals is fundamental to soil genesis. Clay-sized particles were separated in soil samples from three sites showing different degrees of vegetation development. The particles were analysed by X-ray diffraction to characterise the clays present whilst individual particles were also examined by scanning electron microscope.

To characterise the rates of mineral development at different sites standard mineral bags each containing 1g quartz and 1g of either feldspar, labradorite, phlogopite or vermiculite have been buried 5cm deep. These will be retrieved after not less than five years and analysed for changes in their original mineral content.

BIOLOGICAL WEATHERING

Newly exposed rock surfaces are colonised by a range of micro-organisms and cryptogams. These are capable of causing both physical disruption of the rock fabric (Fry, 1927) and changes in the chemical status of particular minerals (Ascaso and Galvan, 1976).

Schists are particularly prone to physical disruption, especially by fungal mycelia penetrating the foliation, because óf the volume changes associated with wetting and drying cycles in the organic material. To study both the extent of penetration and its direct effects in terms of surface damage thin sections of mica-schist covered by crustose lichens have been used (Walton, 1985).

Observations by scanning electron microscopy can provide direct evidence of the physical effects of weathering on the rock surface (Wilson and Jones, 1983; Oberlies, 1958), identifying etch pits etc. formed during chemical attack by solutions secreted by micro-organisms. SEM observations will also be used to investigate colonization sites to see if those surface texture characteristics which apparently provide favourable microhabitats are correlated with easily weathered mineral types.

WEATHERING OF CLASTS

To assess the physical loss of material from clasts in a natural situation selected specimens were left exposed on a coarse stripe area for 6 years. They were constrained within loose open mesh bags and weighed once each year on a portable balance with an accuracy of 1g. Three lithologies were used - mica-schist, quartz-micaschist and marble - with 5 samples of each ranging in weight from 728g to 1577g. After six years most changes were within the limit of error of the balance and even the largest change for the schist was only a loss of 0.4% (Table 3). Loss of mass, either by physical or chemical weathering, is clearly a slow process for clasts lying out of direct contact with the soil matrix, whose only contact with water is via precipitation and snow melt. An ancillary study by Hall (1990) which re-weighed small rock tablets (and re-tested their properties) after they had been left resting on the ground surface for varying periods of time also concluded that weathering rates are very slow. Based on available data and with the assumption of linear weathering rates (the data base is not yet large enough to model any other form) then a mass change of only -2.35% could be expected after 100 years, and this is a rate for closed system, omnidirectionally frozen samples in environments where moisture is available. Information obtained from simulations (Hall, 1986b,1988) suggest that weathering rates for unidirectionally frozen cliff faces (where moisture content is much lower) could be as much as 5800% slower!

DOWNSLOPE MOVEMENT

Measurements of mass solifluction/gelifluction were initiated in 1981 on Signy Island. Two methods were used to measure soil movement at the surface and one method to measure vertical profile movement at the Moraine Valley site. The site is characterised by large scale relictual sorted stone stripes with soil creep and solifluction. Observations were made over a period of seven years. On the lower part of the slope a line 6.5m long was painted across three fine and two coarse stripes with location posts anchored at each end in the permafrost below the coarse stripes. Measurements were made each February, at 10cm intervals along the line, of movement downslope. Close to this but higher up the slope, beside the rubber tubes measuring profile movement (see below), two transects c. 4m long were laid out across the stone stripes. Each transect crossed three stripes (two fine and one coarse) and across each transect were placed equally placed flat clasts (c. 5cm diameter). Each clast had an arrow painted on it pointing upslope. The movement downslope of each clast and the orientation of its arrow were recorded in February of each year.

Finally, to obtain data on downslope movement through the soil profile thin-walled rubber tubes were inserted vertically into the fine stone stripes in the upper part of the slope. Vertical insertion, checked with a plumb bob, was into holes up to 40cm deep made with a metal tube. A pit was excavated beside each tube after two or three years and the deflection from the vertical measured in 1cm intervals.

For surface movement, Fig 2 shows the positions of the painted line at 1, 3, 5 and 7 years. The annual movement is almost completely confined to the fine stripes; maximum annual movements of the fine and coarse stripes are given in Table 1. There is little consistency from year to year either within or between stripes and this is reflected in the large standard deviations. Comparison of the annual movements with the mean for each stripe shows no discernable pattern affecting all stripes for either good or poor years. This suggests that large and small movements cannot be easily attributed to climatic differences between years but are much more likely to be site functions, probably related to drainage lines at snow melt and of microtopography.

The painted line data principally reflect the movement of the finer particles at the soil surface. The movement of the larger introduced clasts on the surface again clearly indicates the sorted topography of the valley side with rapid movement on the two fine stripes being evident in each transect (Table 2). These stones, each weighing c. 50g, show annual movements of up to 23cm, which is comparable with the most rapid movement of the fines on the lower slope angle. The lack of any significant difference in the overall means is interesting since the slope angles of the two transects differ markedly. The mean movement over the seven years was very similar to the mean for 34 clasts of various sizes on sorted stripes over a six year period on South Georgia (Walton and Heilbronn, 1983). The similarity between the distances moved by surface fines and the clasts suggests that clasts of this size and weight are possibly small enough to be "floated" downslope on the mudflows that often happen during snowmelt. The orientation of the stones change as they are moved (Fig. 3) and there is evidence for overturning on some occasions, probably due to pipkrake formation.

On the upper slope, movement at the surface, as measured by tube deformation, was generally equal to or slightly less than the movement of the line lower down in the valley. The mean annual profile in Fig. 4 suggests that movement is typical of a site with frost creep and surficial freeze-thaw but with little indication of gelifluction (c.f. Benedict, 1970). Whilst the range of movement over the seven year period has been quite large at the shallower depths the standard errors indicate that most velocity profiles lie fairly close to the mean. Movement below 30cm depth is clearly very slow indeed; the permafrost underlies this site at about 50cm depth.

GEOCHEMISTRY AND MINERAL CYCLING

As a consequence of physical, chemical and biological weathering the breakdown of rock fabric and ensuing chemical changes result in the release of soluble ions to the soil water system. Weathering activity is also taking place at the particulate level in the soil. A range of approaches is being used to monitor the rate of release of individual ionic species, their relationship to the parent rock, their effects in the soil system, and the relationships between these processes and other environmental variables.

The elemental potential available both in the unweathered parent rock, and in the soil system, has been assessed on digested rock and soil samples by atomic absorption spectrophotometry. XRF analyses of powdered rock samples will be used to provide a check on the AAS data. Assessments of the extractable content of the soils for K, P, N, Ca, and Na will be made using the methods of Allen _et al_. (1974). Measurements of cation exchange capacity have been made.

To determine the concentrations of a range of elements in the soil water regular water samples have been taken throughout the summer. Constant-tension PTFE water samplers (Riekerk and Morris, 1983) and ceramic cup vacuum samplers were used to collect pore water, and plastic trays inserted into the profile at various depths gathered freely draining water.

A modified version of a chemostat designed to run at relatively low temperatures (Sobek, Bambenek and Meyer, 1982) will be used to investigate the effects of particular fungal and bacterial isolates on biological weathering of specific minerals.

Whilst these data will provide an elemental framework for the weathering process they can only be related to the general rock type and not to mineralogical differences. To provide further detail in this field thin sections of rock will be sampled by point counts for an estimation of mineral frequencies.

It is clear from earlier work that weathering rates vary between minerals by orders of magnitude. The analyses described above are essentially assessments of field rates under uncontrolled conditions. It has been shown that organic acids (Huang and Keller, 1970), humic and fulvic acids (Kodama et al., 1983) and lichen acids (Ascaso and Galvan, 1976) can all have important effects on the availability of some elements in the soil solution. Laboratory studies have been undertaken to investigate their effects on Al and Fe in the soil water. It is known that during the weathering of micas Na often replaces K in the crystal lattice. Long term experiments have been set up to see if this reaction continues, and if so at what rate, in frozen soils.

To provide both the necessary microclimate data for field assessments and laboratory simulations three field sites are equipped with data loggers for year-round collection of rock, soil and air temperature, radiation, wind speed and humidity. More specific investigations of temperatures in isolated clasts have also been undertaken.

SOIL DEVELOPMENT

The present investigations are concerned only with the soils derived from the breakdown of quartz-micaschist. These soils are especially prone to cryoturbation and there is widespread active sorting into small scale periglacial feature (Chambers, 1966a, 1967). Some particle size analyses have been published for various types of patterned ground (Chambers, 1966b) and the degree of frost heave (Chambers, 1967) have both been examined for some sites. Previous work has also established the effects of frost action on particle distribution within sorted features (Chambers, 1966, 1967). Allen and Heal (1970) suggested a classification into four soul categories - mineral, organic, brown earth and ornithogenic. However, a recent study has characterised five mineral fellfield soils in terms of profile structure and particle content (Juneman and Sheil, in prep.). In all samples the stone content was found to be high whilst the organic matter was low (<6%). Clay contents were generally low (often less than 10%) with the soils being dominated by sand-sized particles or, in some cases, by high silt contents. Larger stones in the profiles were normally coated with silt on their upper surfaces, consistent with observations by Washburn (1969) in Greenland of frost-pushed stones.

Nothing is known yet of the relationships between particle size and mineralogy. To investigate these soil samples have been separated by wet sieving to clay, silt and various size fractions of sand. using the 65-60 micron fractions have separated the mineral constituents using magnetic separation and heavy liquid sedimentation (Hutchinson, 1974). Silt and clay size fractions have been analysed using X-ray diffraction (Zussman, 1977). Suspensions of the fine particles have been dried on microscope slides and grain mineralogy examined under an optical microscope.

Of considerable significance in the development of soils subject to cryoturbation is the distribution of pore space within the soil profile. Bulk density, porosity and field capacity saturation have been determined by standard techniques (Black et al., 1965). Kubiena boxes were used to sample a variety of soils on Signy Island for any evidence of differential segregation of particular minerals by sorting processes. The samples were infiltrated using a synthetic resin under partial vacuum and sections prepared to allow the study of micro-morphological features (Fedoroff, 1982).

Measurements of the chemical content of soil water on these sites has provided evidence of localised high concentrations of polyols attributable to leakage from bryophyte and lichen cells caused by freeze-thaw damage (Tearle, 1987). These concentrations (up to 0.5%) correspond well with the substrate requirements of common fellfield bacteria. The part played by bacteria, fungi and algae in the stabilisation of surface silts is at present under investigation. The concentrations of principal inorganic ions in both freely draining and partly bound soil water were compared with precipitation over a three month summer period (Juneman , unpubl.). Signy Island is small enough for all precipitation to be marine-coupled and indeed the ionic composition of summer rain was found to be close to that of seawater. Very significant enhancement was however seen in calcium, magnesium and potassium levels of water passing through the soil.

GENERAL SYNTHESIS

There have been few planned long-term studies which have attempted to provide coherent investigations of the complexities of mineral cycling in an ecosystem. The most important so far has been the Hubbard Brook studies in the United States (Bormann and Likens, 1979) but this, on a forested catchment with developed soils, is of necessity much more complex than any Antarctic ecosystem. It would seem that principal difficulties in implementing work of this kind are the interdisciplinary nature of much of the work, the difficulties in providing long-term funding and the problems of synthesis for complex ecosystems. Despite this an understanding of how mineral cycling is controlled in natural ecosystems is essential for their proper future management.

In our view it is essential to simplify the system under study if basic models are to be achieved in a reasonable time span. The Antarctic provides much simpler ecosystems than are found elsewhere in the world (Bonner, 1980) and an understanding of these, as precursors of the more developed temperate systems, should help in the elaboration of the more complex models needed elsewhere.

Our approach, has been to limit the study to one principal rock type, quartz-micaschist and soils derived from it. The natural vegetation on the Signy Island sites is entirely cryptogamic, and severely limited in species diversity. To introduce a comparative element three sites are being studied to illustrate developmental changes with time since glacial retreat (Smith, 1985). Rates of soil development are expected to be slow at low temperatures. Measurements of ion release by a variety of techniques in both the field situation and in laboratory simulations are essential to ensure that changes close to the limits of resolution of any one technique are detected by other means.

Data transfer between these various lines of research are required to maximise the interdisciplinary value of the studies. One way of formalising this is to use a project matrix (Fig. 5), which highlights for research workers in different fields how their data can contribute to wider scientific questions. For instance, it is clear that microclimatic data can contribute to a periglacial activity project, but it is less obvious that information on soil micromorphology can usefully add to such a wide range of projects.

The synthesis of all these data will be directed towards providing a process description of rock breakdown and soil formation, from the initial weathering of parent quartz-mica-schist through to the loss of ions from the ground water into lakes and streams. In a cold climate many of the processes are slow yet all are relevant to the development of soils in temperate regions after glacial retreat.

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TABLE 1

Maximum downslope movement in cm per year of a painted line across sc
stripes

	Year							mean	s.e.
Position	1	2	3	4	5	6	7		
Fines 1	7	6	7.5	10.5	9	8	13	8.7	1.4
Fines 2	8.5	23.5	13.5	14.5	10	14	11	13.6	1.6
Fines 3	7	4	7	4	8	4	8	6.0	0.7
Coarse 1	2.5	0.5	0	0	5	0	6	2.0	0.8
Coarse 2	0	0	0	0.5	0	0.5	0	0.1	0

(Positions as marked in Fig. 2)

TABLE 2

Mean annual and maximum rates of downslope movement of 26 introduced cl
on two areas of stone stripes over a seven year period

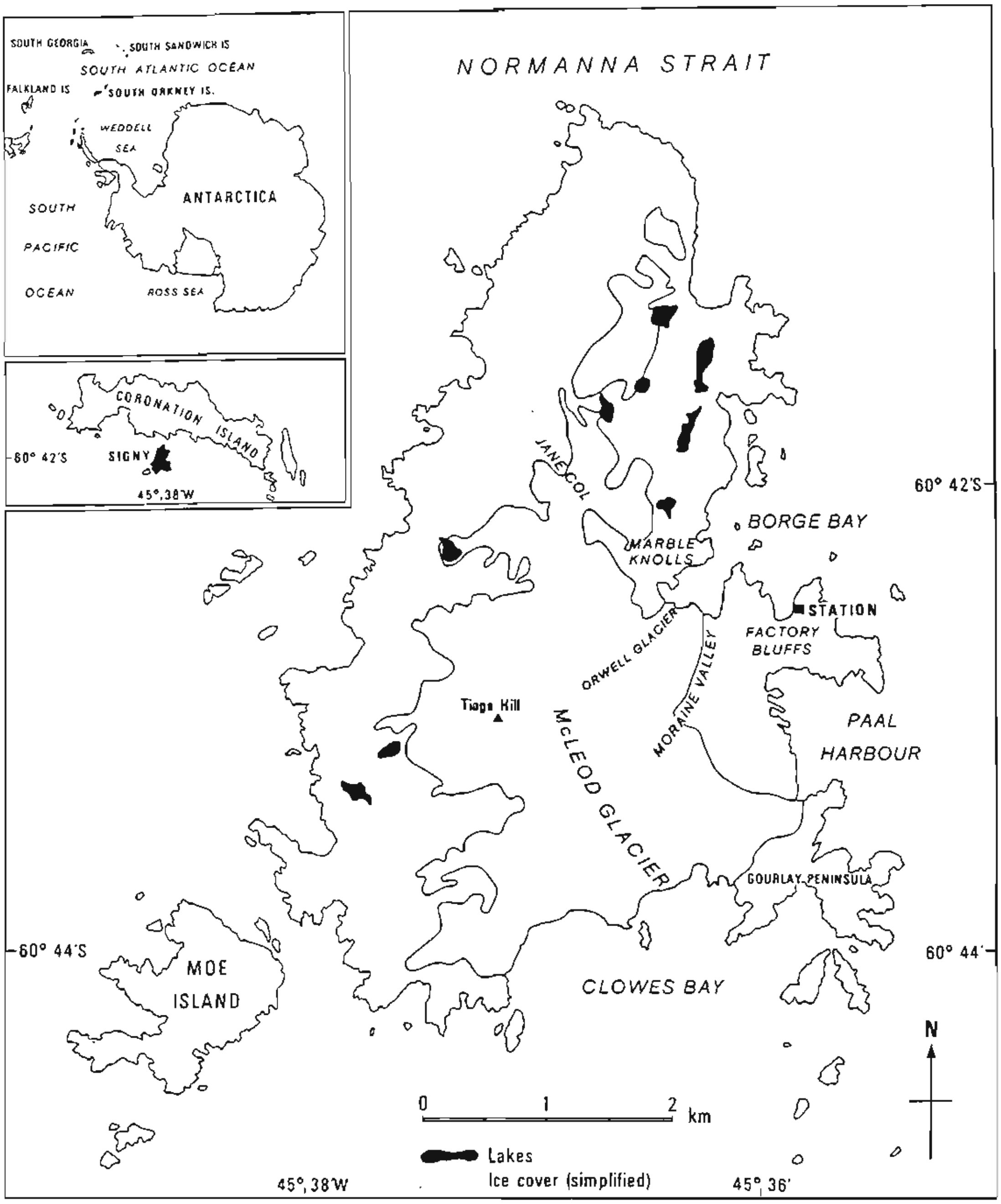
	C1		C2	
Year	max	mean	max	mean
1	10	5.2	9	3.6
2	13	7.4	23	12.4
3	13	7.7	17	9.7
4	9	4.3	6*	6.0
5	6	1.9	15	7.9
6	7*	1.9	9	2.4
7	7	6.9	9	6.9
8	16	8.1	10	4.6
9	15	7.9	10	3.3
10	7*	6.6	5*	3.0
11	6	3.0	12	4.9
12	7	6.0	8	6.0
13	14	6.9	8*	1.6
overall mean			5.7	5.6

*Denotes data set incomplete
C1 is on the upper part of the slope and C2 on the lower part.

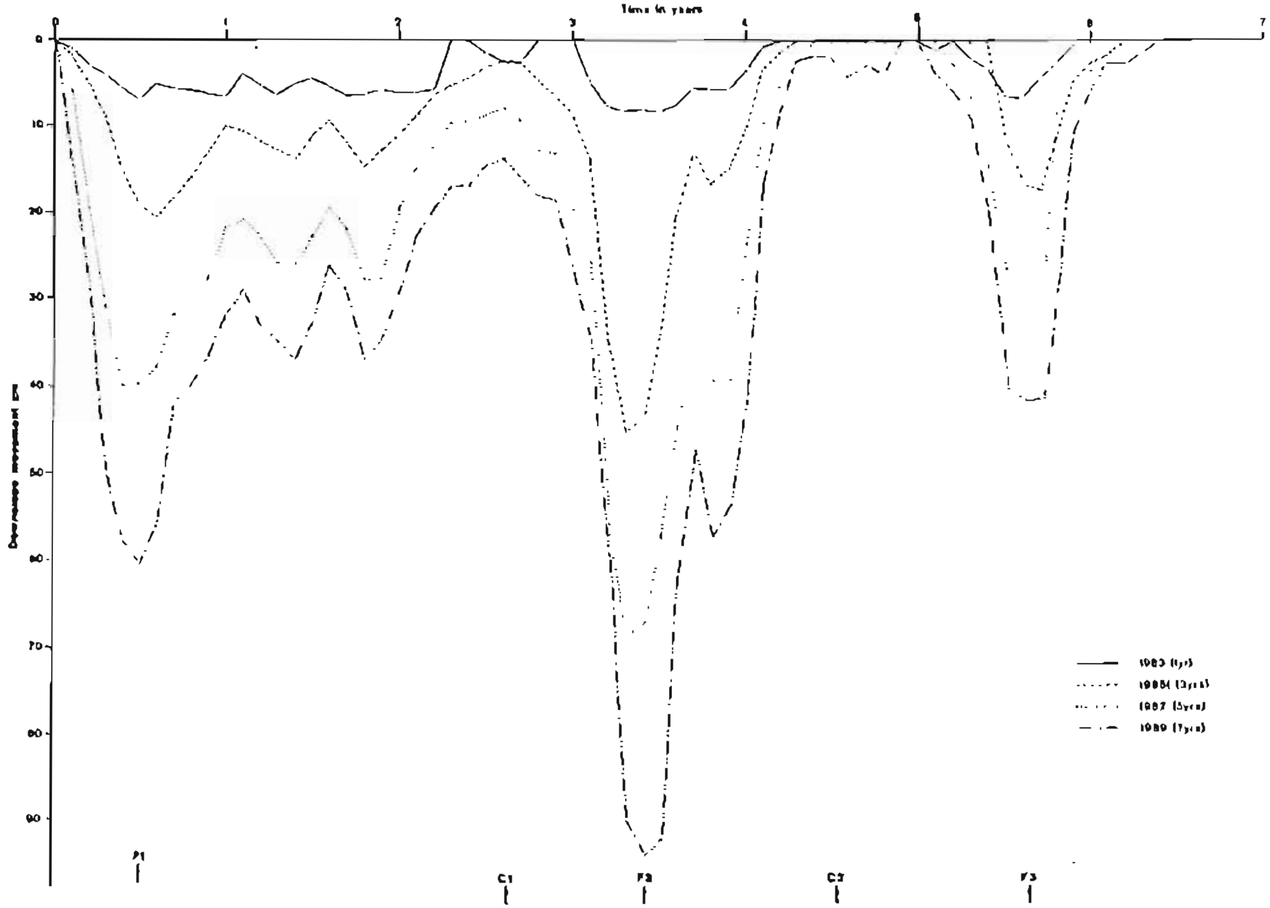
TABLE 3

Loss in weight (g) of clasts of three lithologies over six years

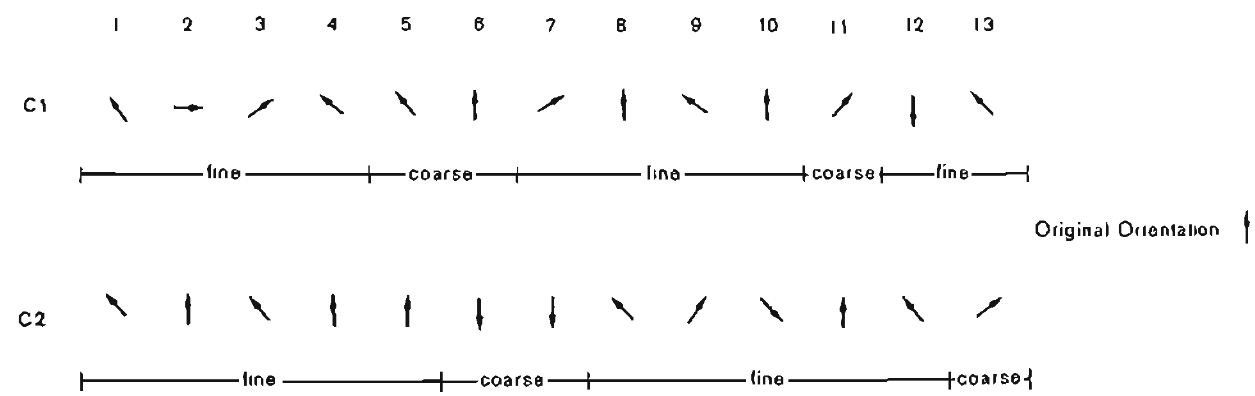
Mica schist					
original weight	762	881	1050	1470	1577
% change	0.2	0.1	0.4	0.1	0.1
Quartz mica schist					
original weight	851	1102	1160	1195	1314
% change	0	0.09	0.1	0	0
Marble					
original weight	728	738	885	1101	1047
% change	0.2	0	0.2	0.2	0.6

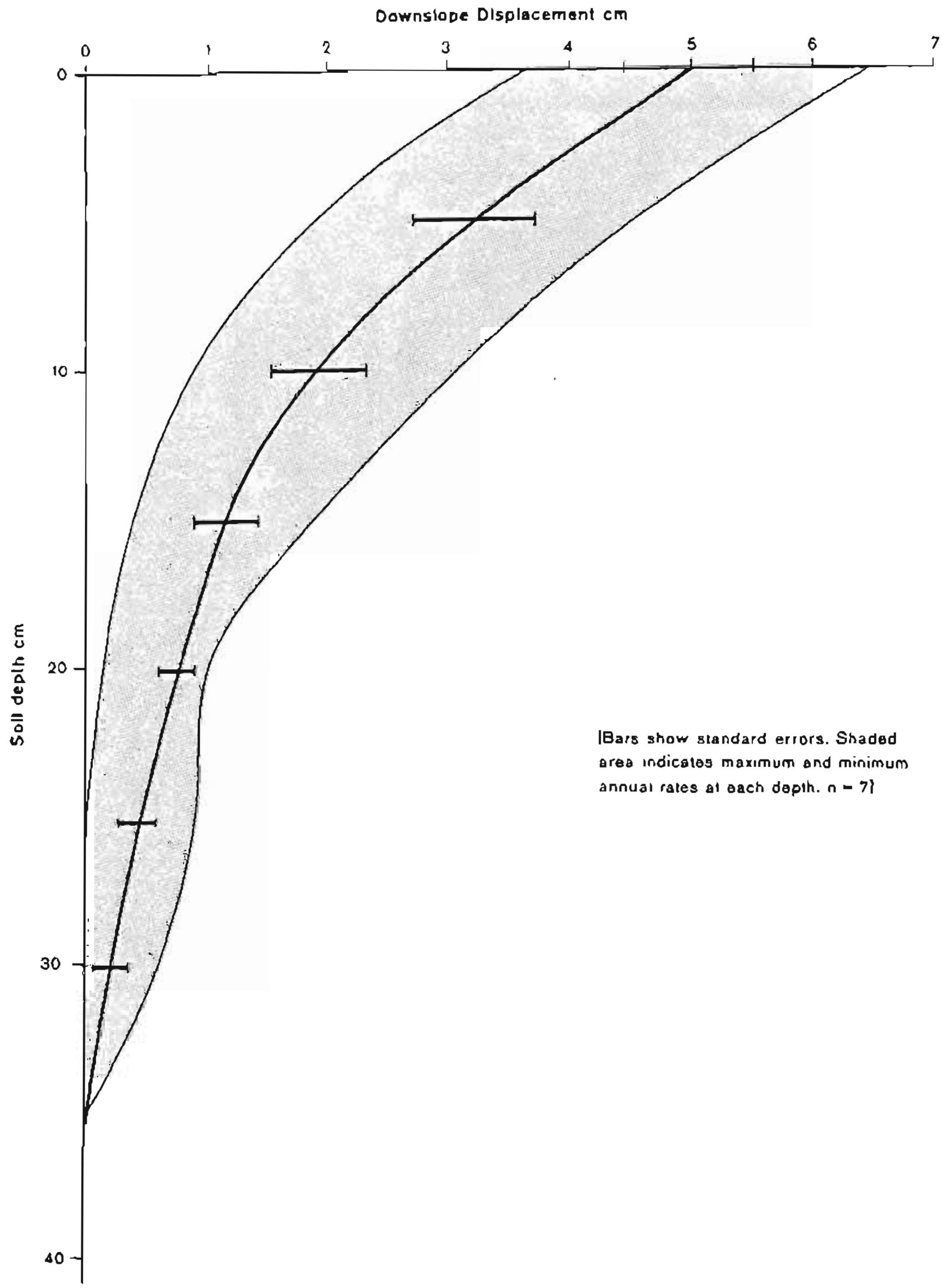


SOUTH GEORGIA
SOUTH SANDWICH IS
SOUTH ATLANTIC OCEAN
FALKLAND IS
SOUTH ORKNEY IS.
WEDDELL SEA
SOUTH PACIFIC OCEAN
ANTARCTICA
ROSS SEA
CORONATION ISLAND
SIGNY
60° 42'S
45°,38'W
NORMANNA STRAIT
JANE COL
MARBLE KNOLLS
BORGE BAY
STATION
FACTORY BLUFFS
ORWELL GLACIER
MORAINE VALLEY
Tioga Hill
McLEOD GLACIER
PAAL HARBOUR
GOURLAY PENINSULA
MOE ISLAND
CLOWES BAY
60° 42'S
60° 44'S
60° 44'
0 1 2 km
Lakes
Ice cover (simplified)
45°,38'W
45°,36'
N



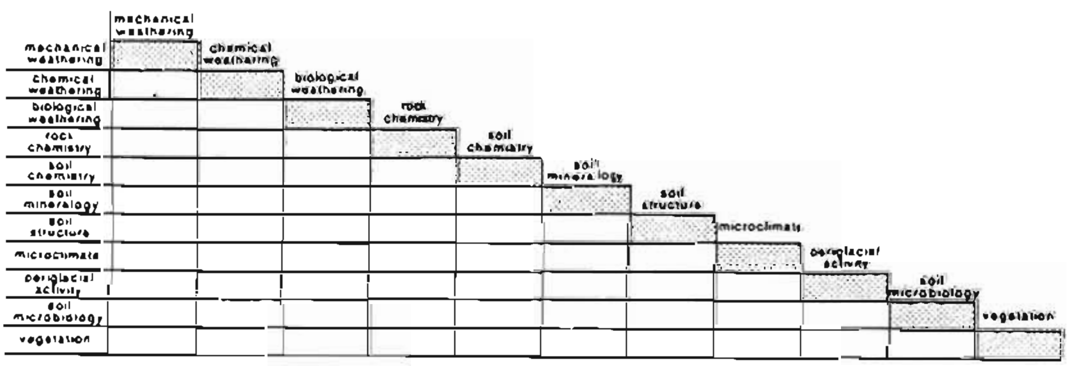
Time in years
0
1
2
3
4
5
6
7
0
10
20
30
40
50
60
70
80
90
Downslope movement cm
F1
C1
F2
C2
F3
1983 (1yr)
1985 (3yrs)
1987 (5yrs)
1989 (7yrs)


1 2 3 4 5 6 7 8 9 10 11 12 13
C1
fine
coarse
fine
coarse
fine
Original Orientation
C2
fine
coarse
fine
coarse



(Bars show standard errors. Shaded area indicates maximum and minimum annual rates at each depth. n = 7)

	MECHANICAL WEATHERING	CHEMICAL WEATHERING	BIOLOGICAL WEATHERING
mechanical weathering	...		
chemical weathering	salt crystallization fracture frequency	...	
biological weathering	rock surface disruption microfracture occurrence	organic acid attack	...
rock chemistry	exposure of unweathered surfaces	pore water chemistry	secondary mineralisation
soil chemistry		cation exchange capacity	chelation index organic acid production
soil mineralogy	distribution of minerals by particle size		preferential mineral weathering
soil structure	particle & clast size	ped size aggregate formation oxide film deposition	production of mucilages
microclimate	freeze/thaw frequency wet/dry cycle frequency thermal fatigue	water availability temperature amplitude temperature duration	free water freeze/thaw & wet/dry freq. & duration
periglacial activity	clast movement		formation of microbial mats
soil microbiology			production of free organic compounds
vegetation	crack propagation by roots	localized pH changes	spp specific interactions with minerals



Chapter 6

Zoogeomorphic Considerations in Cold Regions

"Real knowledge is to know the extent of one's ignorance."
Confucius

Despite the relative ubiquitousness of animals in the landscape, there is very little in the way of information or data regarding the role of animals in periglacial regions. The only available text on zoogeomorphology (Butler, 1995) makes no specific reference to the impact of animals on permafrost or within the periglacial environment. In general, most geomorphological studies consider geomorphic processes to be operating within a 'sterile, non-living void' (Butler, 1995). Although most of the Antarctic continent is animal-free due to the extensive ice cover, many of the coastal margins have substantial penguin or seal colonies for part of the year and the maritime and sub-Antarctic regions abound in animal life. Equally, much of the Arctic sustains substantial numbers of terrestrial animals, whilst many alpine areas are being impacted by domestic animal populations as well as by indigenous species.

Ever since my first studies in northern Norway it seemed to me that animals did influence landscape development; my first observations were the many runs created by lemmings within the tundra vegetation and that these then acted as drainage channels during snowmelt. Subsequent observations in the sub-Antarctic, where penguin colonies of up to one million birds exist, clearly showed a symbiosis between the animals and the landscape: excreta from the penguins provide a nitrogen source that facilitates enhanced vegetation growth and the development of thick peat accumulations, which, as the colony grows, is subsequently eroded by those same penguins (Fig. **). Seals were observed to create distinct pathways and wallows in the soft vegetation and wet substrate (Fig. ""). So great was the impact of seals on the landscape that old, fossil wallows could be used to help reconstruct palaeo-sea levels (Hall, 1977 - Chapter 9). Indeed, in almost all the study areas cited in this thesis it was clear that animals played an varying, but significant, role in the landscape. I could find no reference to the impact of animals on cold region landscape development and so, where possible, an attempt was made to collect observations and data. Although the information presented here is clearly not germane to the main thesis topic it *is* considered pertinent insofar as the impact of animals in cold regions can be significant. It is the, in part, interaction between the animals and the freeze-thaw environment that makes their role in cold regions so important.

If, as I believe will be clearly shown, animals can be recognised as significant factors within the present-day cold environment then, more pertinently, the question arises to what was their impact during the Quaternary? During the Quaternary not only were there larger and more extensive (outside of the glaciated areas) animal populations but there were also more megafauna than are present today: the mammoths, wooly rhinoceros, great elk, etc. If it can be recognised that animals readily impact sensitive-to-change, slow-to-recover, permafrost and seasonally frozen ground to the extent that landforms can be created today (e.g. the dells created by muskox - Hall, 1997e) then how much of our present landscape, created during the Quaternary, owes its origin to animals? No claim is made to even begin to answer this question here, but the point is made that this factor has not been considered in *any* text and that this question needs far greater consideration.

The papers presented within this chapter are:

- Hall, K.J. 1979a. Population pressure in the Roaring Forties. *Geographical Magazine*, July, 707-710.
- Hall, K.J. 1981a. Geology as an aid to climatic-ecological reconstructions : an example from Marion Island. *Lambda*, 3, 18-23.
- Hall, K.1997e. Zoological erosion in permafrost environments: A possible origin for dells? *Polar Geography*, 21, 1-9.
- Hall, K.J. and Williams, A. 1981. Erosion by animals in the sub-Antarctic: some observations from Marion Island. *South African Journal of Antarctic Research*, 10/11, 18-24.
- Hall, K., Boelhouwers, J. and Driscoll, K. 1999. Animals as erosion agents in the alpine zone: Some data and observations from Canada, Lesotho and Tibet. *Arctic, Antarctic and Alpine Research* , 31, 436-446.
- Hall, K. Unpubl. a*. Presentation to the Association of Professional Engineers and Geoscientists of British Columbia, Annual Conference, Prince George, Canada, 1996.
- Hall, K. Unpubl. b*. Zoogeomorphology: A forgotten factor in development? Plenary presentation to the "International Conference on Environment and Development in Africa: An Agenda and Solutions for the 21st Century", Pretoria, South Africa, 1997.
- Hall, K. Unpubl. c*.Animals as agents of landscape evolution in the Arctic: The unquantified element. Presentation at the *Arctic Consortium of the United States* annual meeting, Washington, D.C., U.S.A., 2000. *Twelfth Annual Meeting, Council Notebook*, 6-7.

Fig. 8

Erosion of peat at the margins of a large penguin colony on sub-Antarctic Marion Island.

Fig. 9

Erosion by seals on sub-Antarctic Marion Island

Although the first two papers are more of a 'popular' nature they do, nevertheless, provide some numerical data and highlight the impact of penguins and seals on the landscape. The following paper goes in to greater depth regarding the role of animals on sub-Antarctic Marion Island. Moving to the permafrost region of the Canadian high Arctic, the next paper deals with animals as a possible explanation for dells. Muskox on Ellesmere Island were seen to be following the troughs, associated with ice wedge polygons, that extend down to the rivers. The erosion produced by muskox trampling the active layer and rubbing their coats on the trough sides (to remove moulting hair), caused melting of the ice wedges and thereby facilitated fluvial and mass movement activity that, with time, opened the troughs into dells. The origin of dells found outside of the glacial limits in Europe have long been a matter of speculation (French, 1996). This hypothesis offers a possible explanation for the fossil forms whilst explaining those presently being formed in the Arctic. The next paper provides observations and some limited data regarding the impact of animals on the landscape from three diverse, cold regions: Tibet, Lesotho and Canada. In all instances, animals are shown to have a marked effect on the landscape and, in the cases of Lesotho and Tibet, to have repercussions for animal-based sustainable development.

The final three papers comprise unpublished (but invited) presentations to meetings - one on sustainable development, one on professional geoscience in forest management, and one on Arctic environments. In all instances a major factor discussed was that of animals. Rather than providing "answers", presentations were aimed at making listeners aware of the possible role of animals in the respective undertakings. To that end, pertinent data were given in an attempt to highlight the issue. The contention is that the zoogeomorphic role of animals has not been considered adequately within the geomorphic, planning, or management aspects of environmental issues. This is especially so within cold regions where zoogeomorphic disturbance to permafrost and the synergistic interaction with periglacial processes can both be particularly effective.

Geogra

Population pressure in the Roaring Forties

by Kevin Hall

Marion Island, in the Roaring Forties in the southern Indian Ocean, is host to a variety of bird and mammal species. More than 3,000,000 penguins thrive on its coastal belt which has high rainfall and low temperatures but their population densities create erosional problems

Marion Island is the larger of the two islands comprising the Prince Edwards and is located in the southern Indian Ocean 2° north of the Antarctic Convergence. The island comprises a 290-square-kilometre shield volcano which rises to a height of 1230 metres and maintains a small area of permanent snow and ice. Ideally located close to the rich marine food source of the Convergence, Marion Island is a prime location for the land-based mating and moulting requirements of many bird and mammal species. The extensive nesting areas and large colonies of penguins within a narrow coastal belt combined with the climate, vegetation and geology peculiar to the island result in the penguin being a major agent of erosion.

Climatically, the island is located in the Roaring Forties and so experiences strong, almost continuous winds which average thirty-two kilometres per hour, frequent precipitation of 2500 millimetres per year, low temperatures with an average of 5·3°C and limited sunshine. These climatic conditions, coupled with the isolated position of the island, result in a limited, treeless, plant assemblage. Within the coastal area affected by the penguins there is a herbfield-saltspray complex characterized by *Tillaea moschata* and *Cotula plumosa* with extensive regions of *Poa cookii* tussock grassland and cushions of *Azorella selago*.

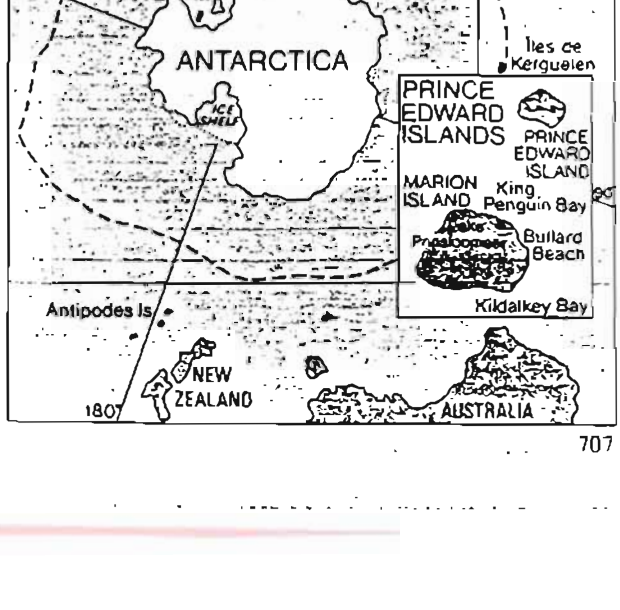
Marion Island is of volcanic origin and essentially composed of two basaltic lava suites: an older, glaciated, grey and a younger black. The grey lavas, dating from 276,000 to 103,000 years ago, are structureless with occasional columnar jointing whilst the black, 10,000 to 4000 years ago, are scoriaceous (with empty cavities), blocky and often friable. Multiple glaciation of the island has resulted in areas of glacial debris and raised beaches.

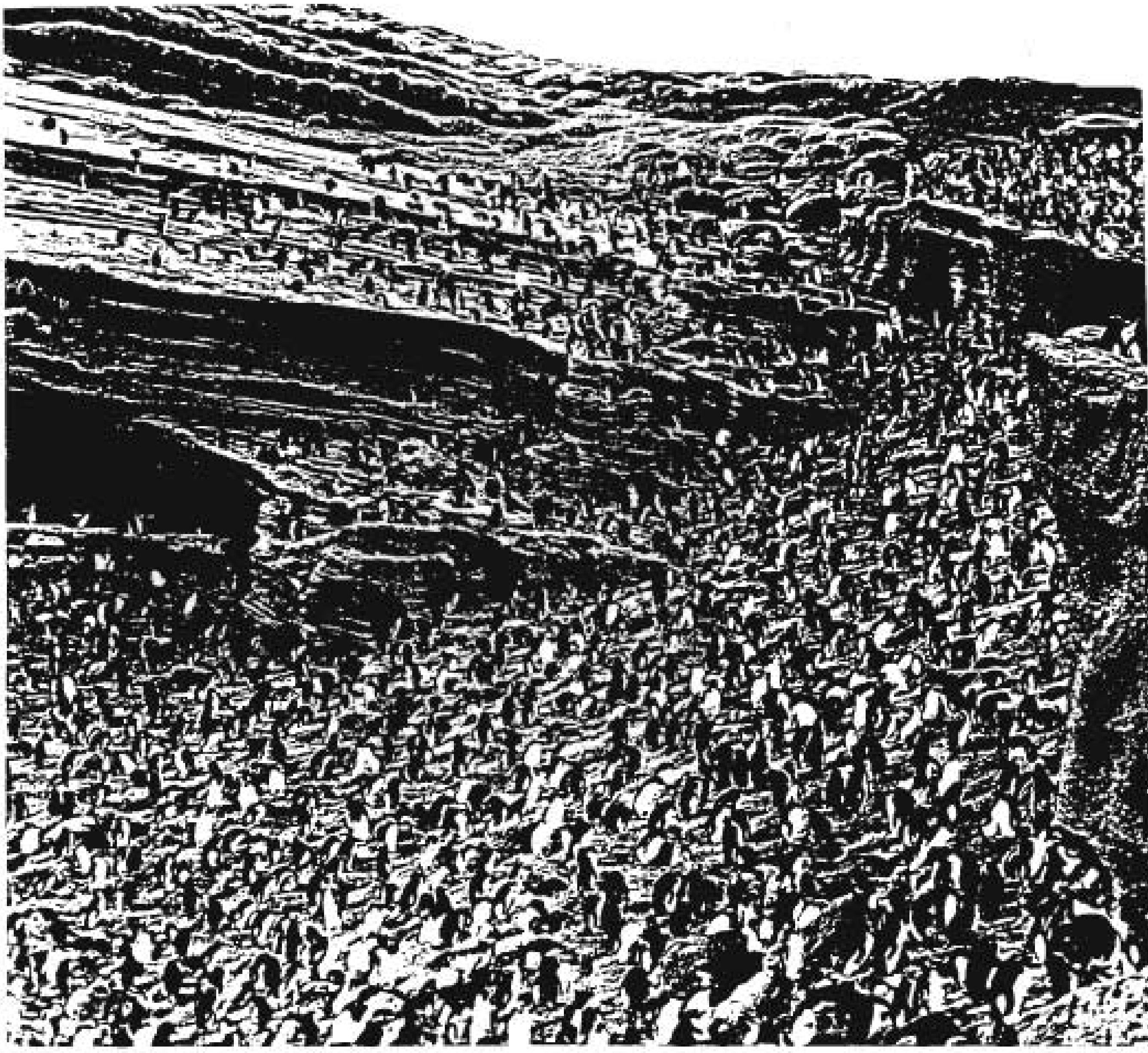
Within this environment are four species of penguin, with a combined population of about 3,400,000, which come to the island to breed and moult: these are the king (*Aptenodytes patagonicus*), macaroni (*Eudyptes chrysolophus*), rockhopper (*Eudyptes chrysocome*) and gentoo (*Pygoscelis papua*). The kings, less agile than the other species, tend to concentrate on present-day beaches and easily accessible raised beaches or areas of glacial deposits; occasionally they may transgress into regions of highly vegetated, low relief black lavas. The aptly named rockhoppers exhibit a preference for unvegetated and rugged black lavas but overlap with the macaronis who tend towards grey lava regions. The gentoo favour coastal vegetated areas which allow rapid escape from predators.

The erosive potential of the penguin can be envisaged when it is considered that if their total population were placed around the seventy-two-kilometre coastline at a density of four birds per square metre - breeding density of kings can be as high as 4·5 per square metre and macaronis 5·3 - then there would be a continuous belt of penguins 11·8 metres wide around the whole island. Whils such an exercise is obviously theoretical it does in fac approximate to the real situation. Much of the coast i: populated by a narrow 'band' of penguins intersperse with barren stretches which are, in turn, compensated fo by several large, extensive colonies.

It is at the large colonies that the full erosional effects o the penguin can clearly be seen. Kildalkey Bay colony, fo example, has an estimated macaroni and king population of 570,000 individuals. The erosive effect of these birds is related to their high breeding density and the number o nest to sea journeys that they undertake for food. A successful breeding king penguin makes fifty-six sea journeys during the breeding period, the unsuccessful eight and the non-breeding four; for the macaroni penguin the numbers are respectively thirty-eight, eight and four. It is estimated that for the island as a whole the king penguins make a minimum of 10,864,000 journeys and the macaronis 24,104,000 journeys each year. The king penguin has an average weight of twelve kilogrammes and the macaroni 4·6 kilogrammes. Thus these two species account for an annual mass movement of 239,170,000 kilogrammes; the rockhoppers and gentoo account for a further 23,300,000 kilogrammes thereby giving a total annual movement of 262,500,000 kilogrammes.

The effect of this movement of birds, concentrated





Inhospitable climate and volcanic rock surface of Marion Island support a limited plant cover. Coastline vegetation is further reduced by the erosive effect of penguins' high nesting densities and frequent trips to the sea. Erosion is greatest in large colonies which extend up to one kilometre inland. 570,000 penguins nest in Kildalkey Bay (left), less agile kings on the lower beach in the background and macaronis on the grey lavas in the foreground. Damage decreases with distance from the colony edges. (Right) peat erosion at the colony edge in King Penguin Bay. (Above) tufts of Azorella survive beyond the upper edge of a macaroni penguin colony on the west coast.

Gentoo penguins (left) are only 6000 strong but cause extensive damage to their nesting sites by trampling and manuring. Erosion is accelerated by almost continuous rainfall which saturates vegetation and removes debris. On Bullard Beach vegetation removal is complete and the jumping and scrambling activities of macaronis polishes and grooves lava (right)

within the moulting-breeding period, is enormous within the colony and its immediate environs. At Kildalkey Bay, 112,750 square metres, and Bullard Beach, 82,388 square metres, vegetation destruction and removal within the colony area is total, the birds now nesting on grey lava or glacial debris. Destruction diminishes with distance away from the colony edge up to a range of approximately fifty-five metres. At these two colonies a peat cover up to four metres in thickness has been completely stripped away. Taking an overall average peat thickness of only one metre, approximately 195,140 cubic metres of peat has now been eroded from Kildalkey Bay and Bullard Beach.

In addition to the stripping of the peat cover macaroni penguins have proved very efficient at polishing and grooving the lava and boulders over which they now must move. The jumping-scrambling action of the macaroni has caused the grey basalts to take on a polished surface pitted with grooves up to ten millimetres in depth. The grooving is caused by the raking action of the claws as the bird fails to jump to a higher point and so slithers down the rock. Once initiated the similar failure of successive birds becomes concentrated along the same line and so the groove is extended and deepened through time. Tens of thousands of birds follow the same route, each year, over the rocks and this number accounts for the size of effect from such small individuals.

The rockhopper is more passive in its erosional role. Being found predominantly in the recent black lava areas the prime effect is to inhibit plant colonization by its movement activity. It is likely that some destruction of the friable, scoriaceous black lava also results from the continued jumping action of these birds. The gentoo, which is the least numerous of the four species, about 6000 individuals, causes extensive erosion within its nesting area by trampling and manuring. However, the limited numbers and consequent small colony size make the effect of the gentoo of only localized importance.

As the colonies currently appear to be growing in size so the area of peat and vegetation destruction is being increased annually. This effect can be seen at Kildalkey Bay where small groups of king penguins can be observed up to one kilometre inland. At the large king colony at King Penguin Bay the penguins can be observed to extend inland, to the Lake Prinsloomeer, across approximately 0·5 kilometres of highly vegetated black lavas. The movement of penguins between the sea and the lake is now beginning to cause extensive erosion over an area of 633,750 square metres.

The destruction caused by the penguins is greatly enhanced by the climate. On average precipitation occurs on twenty-five days in each month, the result of which is a continuously saturated vegetation and peat layer. The wetness of the material facilitates easy removal by the shuffling and jumping of the penguins. The broken down peat is then gradually washed into streams or directly into the sea. The role of the high incidence precipitation in abetting and increasing the action of the penguin is of prime importance.

This action by penguins is not unique to Marion Island nor is it only the penguin species which are active. Other animals which come ashore on these scattered islands also exert a profound influence. Within the coastal area the action of elephant seals in causing large-scale erosion is most marked; often their areas overlap or coincide with that of the king penguins and so the action is even more pronounced. Inland, on the slopes, the nest-building activity of the burrowing petrels, of which there may be millions, accounts for the removal of much material and the accompanying slope instability and resultant slumping.

The penguins, the elephant seals and the petrels are found in comparable numbers on many other sub-Antarctic islands. Some of the other islands including Kerguelen are much greater in size than Marion and thus the effects may not be so obvious, especially when overshadowed by such obvious agents as glaciers. Nevertheless the role of the penguin in the sub-Antarctic in effecting plant destruction and peat removal, together with some rock erosion, is most pronounced and within the narrow coastal strip may be the most active agent present.

Geology as an aid .to climatic-ecological reconstuctions: an example from Marion Island

by Dr. K. Hall, Geography Department.

Marion Island (Lat. 46°S, Long. 37°E) is situated in the great Southern Ocean only 2° of latitude north of the Antarctic Polar Front. Being one of the few land areas within this vast expanse of sea the island is extremely important to the seabirds and ocean-going mammals as a site for breeding and moulting. But, apart from being a wildlife paradise, the island is also of great geological interest. The island is volcanic, as evidenced by the many cones visible about the land surface and by the recent volcanic activity which took place in 1980, and associated with the mid-Indian Ocean Ridge. Despite its past and present, volcanic nature the island is a storehouse of information relating to climatic changes in the most recent (last 300 000 years) part of the present Ice Age. This knowledge is of great importance, for the scarcity of land in this region makes any available site very significant for correlation with the data resulting from the many ocean core studies. The information, as I hope to show is also ultimately pertinent to biological and botanical studies.

During roughly the last 300 000 years, the island has been subject to three glacial episodes, each comprising a series of stades and interstades (ice advances and retreats). Between each placial occurrence there were the so called interglacials when temperatures were comparable to, or even warmer than those of the present interglacial. During these warm phases, when the ice disappeared from this island, there was also extensive volcanic activity. Volcanism has not been found to have occurred at any point during the glacials but only once the ice had melted away.

This information, presented in very broad terms above, is derived form the analysis of features and deposits resulting from glacial-erosional, glacial-depositional, fluvioglacial and volcanic activity, As ice passes over fine-grained bedrock, in this instance basalts, the debris held within it scratches (striates) and polishes the rock in much the same way as a file does a piece of metal. The resulting striations indicate an axis of movement which, in the case of Marion Island, can be turned into a vector as we know the ice flowed from the central highland towards the coast.

As a glacier erodes so it transports. Much of the debris carried by the ice is deposited at the glacier margins to produce moraines. These moraines, ridges of debris often many tens of metres in height and width, demarcate the boundary of the ice at the time of their formation. Thus, by identifying these landforms the positions of the ice edge can be delineated. In addition to the actual moraines the glacier also deposits and plasters a layer of debris (till) beneath itself. During the process of deposition the elongate stones within the till become orientated such that their long axes parallel the direction of flow and they dip up-glacier. This knowledge allows direction of former ice movement to be discerned for the thick sequences of glacial deposits, visible in the coastal cliffs, resulting from periods when the ice extended beyond the present coast. In addition, analysis of the stone orientation, the shape of the stones and other stratigraphic characteristics give information about glacier speed, thermal characteristics and depositional environment.

When the ice melts, during interstades or interglacials, water is liberated and fluvioglacial products result. In addition to the production of fluvial sediments, there is also the formation of landforms such as river channels and outwash plains. At the same time, the ground no longer being ice-covered and temperatures being warmer, soils are formed. Finally in the case of Marion Island, it would appear that the uplift of the land consequent upon the removal of the weight of ice (isostatic readjustment) initiates volcanism and large sections of the island become covered by lava or pyroclasts.

Thus, this information when considered as a totality tells us of the climatic variations, the oscillations of the glaciers, and the timing of volcanic events. This data is then able to be correlated with that from ocean cores to enable a better picture of the Ice Age world to be derived; as in the CLIMAP programme. However, at a more basic level it is pertinent to the study of island colonisation by flora and fauna. The plants and animals each have basic requirements of nutrients and shelter for survival and so the present distribution we currently observe must reflect these necessities. However, the geology and glacier reconstructions tell us that the island has varied greatly in environment conditions over the last 300 000 years. So, with a knowledge of animal and plant requirements, in what order could the island have been colonised subsequent to the last glacial?

The ice on Marion Island diminished during the last glacial with each succeeding stadial but actually began to disappear around 12 000 years ago. As the glaciers melted away, probably very rapidly indeed, so the island began to rise due to isostatic readjustment and volcanism was initiated. Thus, those areas of glacial deposits could not have been colonised until the ice had gone and even then volcanism may have inhibited plant or animal invasion. The lavas would have precluded any colonisation until they were sufficiently cool.

The island supports a large assemblage of flying sea birds who burrow into the thick peats to nest. These then could have only arrived subsequent to the development of a vegetation and peat cover. Nutrients for the plant growth comes, to a large extent, from nitrogen introduced indirectly by penguins. The penguins would have been able to inhabit the glacial, non-volcanic, lowlying coastal areas. Thus ,the colonies observed in these areas, together with the elephant seals that utilize the same localities, were probably the first to reinvade the island. They would promote plant growth in the surrounding areas such that, with time, the burrowing birds could nest in the developing peat. At the same time, birds such as the Great Wandering Albatross which nests on top of flat vegetated areas could also begin to colonise. As the lavas cooled so the macaroni and rockhopper penguins could move into these new rough terrains along the coast and sooty albatrosses could nest on the cliffs. As more animals colonised so more nutrients were introduced and the vegetation cover increased thereby allowing more animals to find living space—and so on.

Thus, in this extremely simple discourse, the means of reconstructing the climatic and geological history of Marion is presented. In addition, the information is used again here, very simply, to show its relevance to the life sciences. Geology is a versatile tool, not bound just to gold, diamonds, coal or economic returns. In fact, somewhere at some point, the geological background will be important to almost any earth science study thereby emphasising the need for our research to be multidisciplinary and not to disappear down narrow short-sighted avenues.

Further information can be obtained from the author (Geography Department, University of Natal, Pietermaritzburg 3200) or from the following publications:

Hall, K., 1979	Popualtion pressure in the Roaring Forties *Geographical Magazine*, July, 707-710.
Hall, K., 1979	Late Glacial ice cover and palaeotemperatures on sub-Antarctic Marion Island. *Palaeo-geography,-climatology*, 29, 243-259.
Hall, K., 1980	Push moraines on Marion Isalnd. *S.Afr.J.Sci.*, 76, 421-424.
Hall, K. and Williams, A.	In Press. Erosion by animals in the sub-Antarctic some observations from Marion Island. *S.Afr.J.Ant.Res.*
Hall, K.	In Press. Rapid deglaciation as an initiator of volcanin activity: an hypothesis. *Earth Surface Proscesses and Landforms*.

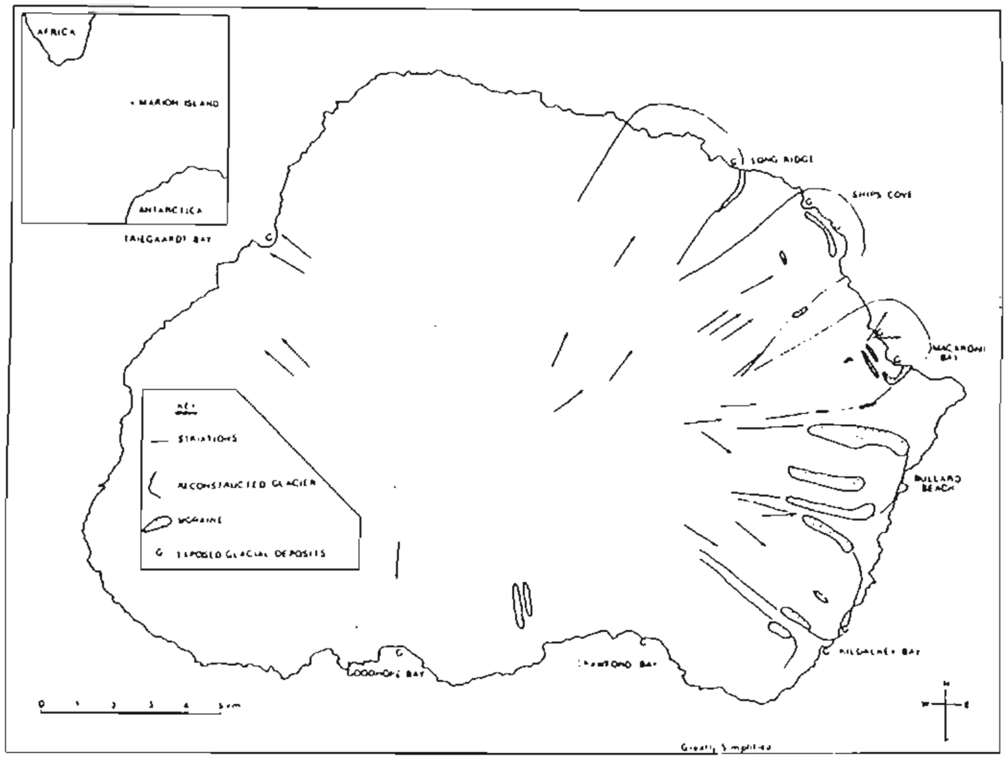


Fig. 1 A reconstruction of the ice cover on the eastern side of the island together with some of the infomation from which the palaeo-glaciers were derived.

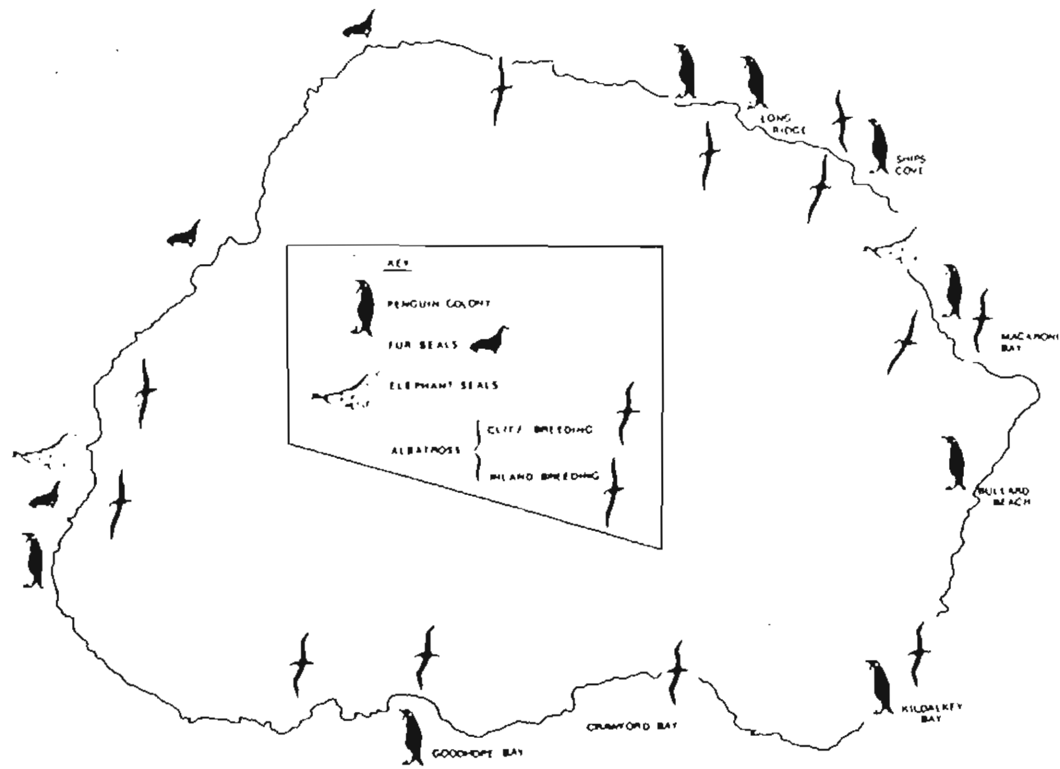


Fig. 2 The location, in a generalised form, of some major concentrations of animals about the island. Compare with the palaeo-glacier cover shown in Fig. 1.

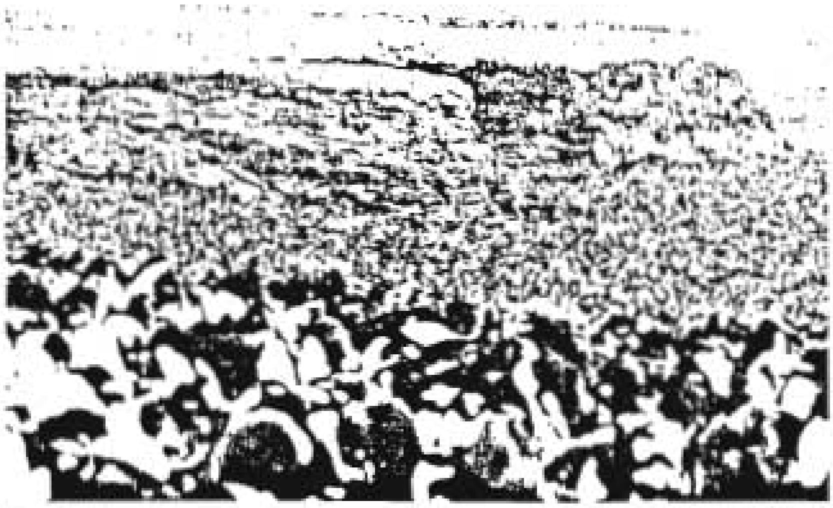


Fig. 3 View across part of the macaroni penguin colony at Kildalkey Bay showing the high nesting density. The peat cliff behind the penguins is 4 m high and is being actively eroded (see penguins on cliff at left) by penguin activity.

Fig. 4 Close-up of nesting king penguins showing remnants of peat (with tussock grass growing on top) which once covered the whole area.

ZOOLOGICAL EROSION IN PERMAFROST ENVIRONMENTS: A POSSIBLE ORIGIN OF DELLS?[1]

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Abstract: In most cold-region studies, animals as geomorphic agents have been considered "curiosities" and thus not in need of study and quantification. Available evidence does, however, indicate that animals can be major causes of erosion and sediment transport. Furthermore, the zoogeomorphic effects may be exploited by other processes that would otherwise not occur. Permafrost is susceptible to degradation as a result of loss of the active layer. Surface processes in permafrost regions are greatly affected by the vegetation cover such that loss of vegetation and erosion of the active layer causes melting of the permafrost, surface runoff, erosion, and debris transport. It is suggested that muskox frequently follow the troughs of ice-wedge polygons as preferential access routes to streams and rivers. In so doing, they cause erosion of the vegetation and active layer, thereby causing permafrost degradation and initiation of secondary geomorphic processes. Based on observations on Ellesmere Island, it is suggested that this form of zoological erosion could explain the occurrence of dells. Key words: zoogeomorphology, permafrost, erosion, ice-wedge polygons, dells.

INTRODUCTION

The effects of animals in the landscape, their impact upon erosion and sediment transport, and their role in landform development has been largely unquantified. The recent, in fact the *only*, volume dedicated to zoogeomorphology (Butler, 1995) makes no reference either to permafrost or, as shall be discussed here, the impact of muskoxen. Despite the lack of extensive data or studies, there is no question that "Mammalian trampling, wallowing and geophagy . . . accounts for large amounts of sediment export and transport . . ." (Butler, 1995, p. 107), as well as for destruction of vegetation cover, leading to enhanced erosion rates. Trampling by large mammals can, in the context of the arguments to be presented here, ". . . lead

[1]Ideas and observations for this paper were obtained during the High Arctic Symposium and Field Meeting on Ellesmere Island organized and led by Dr. Antoni Lewkowicz. (I cannot thank Toni enough for helping fund my participation and for all the work and effort he put into this meeting.) Dr. Lewkowicz also kindly gave the author permission to quote some of his data on Ellesmere climate and permafrost and to submit this paper. Some funding for the author's participation came from NSERC Research Grant #169996 and from the University of Northern British Columbia—both of whom are thanked. Dr Colin Thorn and an anonymous reviewer greatly helped in improving this paper and I am most grateful for their suggestions and advice.

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directly or indirectly to erosion ... direct as when trampling along the edge of a stream . . . indirect when it 'prepares' the soil for erosion by subsequent geomorphic processes . . ." (Butler, 1995, pp. 82-83). Unfortunately, the direct effects of trampling are almost unquantified within the geomorphic literature (Butler, 1995) and the relationship of trampling to landform development has not been considered, at least in the periglacial environment. This trampling effect and its relationship to landscape evolution is likely to have been even more applicable in the Pleistocene, when the numbers of animals, particularly large animals, were much greater. Very pertinently, in the context of the arguments to be presented here, Butler (1995, p. 3) states: "It would be fascinating to speculate on the geomorphic influence of Pleistocene megafauna, such as woolly mammoth wallowing . . ."! Certainly the observations from the sub-Antarctic of the trampling and erosive effects of seals, and even penguins, indicates that in areas of sensitive vegetation, high moisture supply, and freezing conditions zoogeomorphic effects can be substantial (Hall and Williams, 1981).

Based on some simple observations regarding permafrost degradation and muskox trampling on Ellesmere Island (Canadian Arctic), I suggest that periglacial landforms known as "dells" might, at least in some instances, be the result of zoogeomorphic processes. An example of apparent dell production by muskox at the present time will be presented and the argument then advanced to suggest that Pleistocene megafauna might well explain the location and origin of such fossil features in southern England and in Poland, from whence most descriptions of dells emanate.

DELLS

According to Embleton and King (1975, p. 16) dells are ". . . small, shallow valleys with concave cross-profiles, now usually dry except during snow melt or occasional rainstorms . . . which characterize some Pleistocene periglacial areas." These valleys are usually less than 1 to 2 km in length and, although often closely spaced, show no apparent integration into the larger valley systems (French, 1975, p. 267). In terms of formative processes, French (1996, p. 270) refers to ". . . intense physical weathering in combination with meltwater over seasonally frozen ground" and this is said to often occur along (probable) major joint lines. Embleton and King (1975, p. 16) consider them to form under ". . . conditions of impermeable permafrost, sparse vegetation and melting snow . . ." in which surface run-off is encouraged. Dells are reported from southern England and from Poland (French, 1996). Comparable features in France, known as "vallons de gélivation" also are considered not to be the product of stream action but rather the widening of structural weaknesses by frost action with the removal of debris by gelifluction and wash. Of great pertinence to this present discussion, Embleton and King (1975) refer to modern dells in Siberia where (p. 16) ". . . disturbance of the thermal equilibrium induces selective thawing of ice wedges . . ." such that ". . . small channels develop by thermo-erosion, and material moves downslope into them by surface wash and gelifluction as fast as it is removed along the channel . . . ." Nearly all studies (see Ballantyne and Harris, 1994, pp. 152-153 for a discussion) cite rapid frost weathering, niveo-fluvial erosion, and solifluction as the major agencies in dell formation. None of these hypotheses, however, beyond the "assumption" of exploitation of major joints, explains exactly how or why selective thawing takes place at this location. The consideration that the ". . . chalky head . . . that mantles the bottom is the product of frost weathering . . ." (Ballantyne and Harris, 1994, p. 153) must, in fact, remain speculative as there is no known way of identifying frost-weathered debris in the field (Thorn, 1992). Although dells are clearly associated with permafrost and hence periglacial conditions, and gelifluction and wash are evidenced by sediments found on the valley bottoms, their actual origins remain speculative.

MUSKOX AND OTHER LARGE MAMMALS

The range of the muskox (*Ovibos moschatus*) extended in historical time from northern Alaska through to Hudson Bay, on the northern and western islands of the Canadian Arctic, and to Greenland (Nowak and Paradiso, 1983). The same species, or a very close relative, was also present in northern Eurasia during the late Pleistocene. A typical muskox would have a head and body length on the order of 1.9 to 2.3 m, a shoulder height of 1.2 to 1.5 m, and would weigh in the region of 200 to 410 kg (Nowak and Paradiso, 1983, p. 1296); males are typically larger than females. Inside the coarse guard hairs there is an inner coat of fine, soft, hair (*qiviut*). Muskox are gregarious, commonly occurring in herds of about 10 in summer but up to 100 individuals in winter (Nowak and Paradiso, 1983). The herd is highly mobile and usually does not remain in place for more than one day in summer, although movement between summer and winter ranges does not extend beyond 80 km (Nowak and Paradiso, 1983). Dwelling exclusively on the Arctic tundra, muskox prefer ". . .moist habitats, such as river valleys, lake shores and seepage meadows" (Nowak and Paradiso, 1983, p. 1297). During the Pleistocene, when much of these northern lands were ice covered, there is evidence to indicate that muskox occurred far to the south of the great ice sheets (Pielou, 1992).

Therefore, historically, muskox and other large mammals such as the woolly mammoth were living to the south of the ice sheets (Pielou, 1992). Like the muskox, the woolly mammoth also had a soft, woolly undercoat beneath the outer, coarse guard hairs. The mammoth was substantially larger and heavier than the muskox. Mammoths ". . . preferred comparatively open ground, especially tundra . . ." (Pielou, 1992, p. 109), where they chewed coarse tundra vegetation. Evidence from Britain shows the occurrence of elephant (*Palaeoloxodon antiquus*) and rhinoceros (*Dicerorhinus*) as well as mammoth from 130 to 190 ka B.P. and that some of these species also were found in Eurasia (Ballantyne and Harris, 1994). In fact, during the Quaternary many large mammals grew in numbers until suffering a massive extinction near the end of the Pleistocene (Williams et al., 1996). Thus, there were both large animals and relatively large numbers to the south of the ice sheets.

PERMAFROST AND ICE-WEDGE POLYGONS

Only the salient issues regarding permafrost and ice-wedge polygons are presented here. Detailed information and/or clarification appears in French (1996). Because of the long period of winter cold and the short summer, perennially frozen ground that does not rise above 0° C for more than two years is considered to be "permafrost" (French, 1996). The layer of ground on top of the permafrost that thaws each summer is the "active layer." "Continuous permafrost," that geographic zone in which frozen ground is present in all localities—has a southern boundary approximating to the -6° C mean annual air temperature (MAAT). Vegetation can have a profound effect on permafrost, not the least being that it shades the underlying ground from solar heat (French, 1996, p. 66) and ". . . this insulating property is probably the single most important factor in determining the thickness of the active layer"; peaty organic materials also are quite effective in protecting permafrost from atmospheric heat. Loss of vegetation cover and/or organic materials will lead to a thickening of the active layer at the expense of the permafrost. A major feature indicative of permafrost, particularly the growth of permafrost, is the ice-wedge polygon; degrading permafrost is associated with the melting of ground ice and the formation of "thermokarst" (French, 1996).

Ice wedges are a major feature of permafrost landscapes and, simplistically, comprise thermal contraction cracks that are filled with ice. Cracks can be in an orthogonal, hexagonal, random orthogonal, or orientated orthogonal form (French, 1996, Fig. 6.12). The orientated orthogonal cracks are, as in the example presented by French (1996), located around a river with a major set of crack axes normal to the river channel. The melting of ice bodies, such as in ice wedges, leads to local collapse and subsidence of the ground surface and this is termed "thermokarst." According to French (1996, p. 109), "the development of thermokarst is due primarily to the disruption of the thermal equilibrium of the permafrost and an increase in the depth of the active layer. Thermokarst subsidence is then a function of the new equilibrium depth of the active layer and the supersaturation of the degrading permafrost." Of note here is the water that is associated with the permafrost degradation and hence the entire set of geomorphic processes that are associated with its presence. As French (1996, p. 110) explains, the loss of the insulating organic mat results in substantial permafrost thaw before a new thermal equilibrium is reached.

OBSERVATIONS

Observations were made on the Fosheim Peninsula of Ellesmere Island, Canadian Arctic (Fig. 1). This is a region of continuous permafrost that is about 500 m thick (Taylor et al., 1982), with typical active-layer depths hanging from 40 to 90 cm depending on vegetation cover (Lewkowicz, 1996). The mean annual temperature for this area is on the order of -19.7° C, although summer temperatures can be anomalously high (Lewkowicz, 1996). Details regarding ice wedges and climate for this area can be found in Lewkowicz (1994). There are extensive areas of ice wedges, including the orientated orthogonal forms, at the margins of rivers (Fig. 2). It was observed that muskox followed the troughs of ice wedges to gain access to the rivers and, in so doing, caused extensive trampling and erosion by their hooves (Fig. 3). The removal of the vegetation and erosion of the thawed active layer thus caused melting of the permafrost and the slumping plus water transport of debris. To compound the effect, the muskox not only trample and actively erode by their movement over the soft, wet sediments but they also rub their bodies on the trough edges to help remove their molting coats. The latter is evidenced by

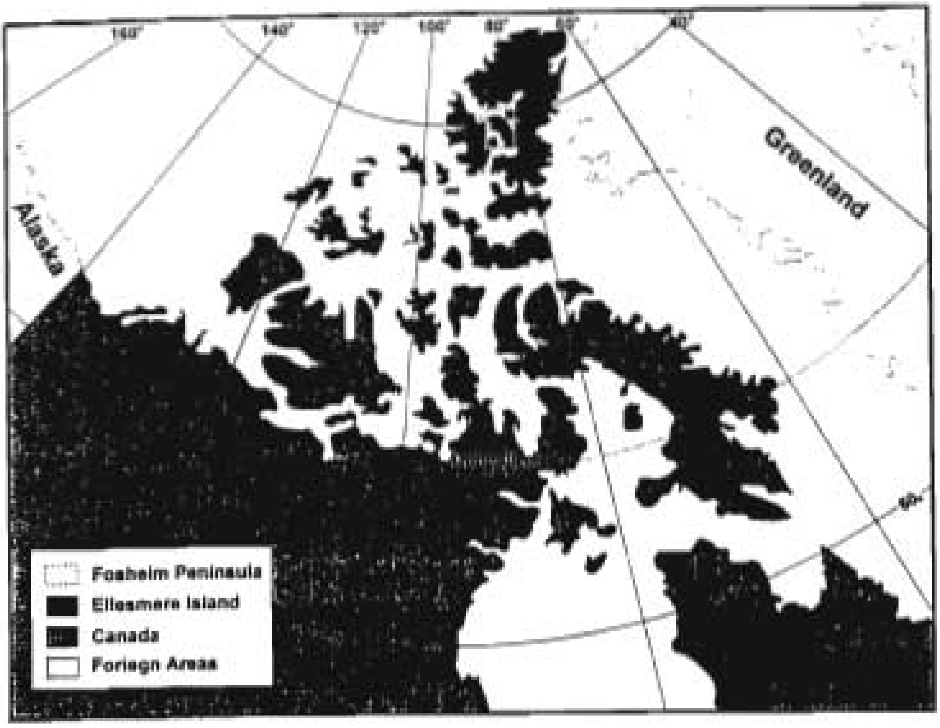

Fig. 1. Location map.

thick clumps of *qiviut* (the soft, undercoat of hairs) found on sediment blocks and trough sides (Fig. 4). Thus, the muskox are encouraging all of the factors outlined above that facilitate permafrost degradation.

Once a route is used by the muskox, a positive feedback situation develops whereby the degradation makes the trough deeper and thus better for rubbing of coats, at the same time that there is more water being concentrated from permafrost melt and so the trough is easier to erode. Wedges that intersect the trough now start to add their melt to that of the main trough and so the net effect is compounded—while continuing to be used by the muskox. The overall effect may be to widen and deepen the access trough until it forms a broad, short valley leading into the river (Fig. 5), which is *not* part of the drainage system per se although it may sustain a small meltstream of its own (Fig. 5).

Returning to the description of dells as outlined earlier, they are short and not integrated into the larger valley system and are said to result from selective thawing of ice wedges. Their growth along *probable* major joint lines or lines of structural weakness (French, 1996) could equally be explained as a result of ice-wedge trough exploitation by animals, as outlined above. This would provide an explanation for the selective thawing of ice wedges as well as for the valley form and its non-association with the river system. This explanation would not conflict with the sediments associated with dells—rubble and muds. As the muds are said to be well sorted and often stratified, considerable meltwater activity is suggested (French, 1996). The melt produced by ice-wedge thaw along the trough, as well as of those wedges that intersect the widening trough, would add to the meltwater

Fig. 2. Orientated orthogonal ice wedges with major axes normal to the river.

and, as the valley develops, would provide a natural drainage line—but *not* associated with the river system. With climatic amelioration and overall permafrost degradation more melt could be expected to follow these valleys. This would not only help explain the sediments but would also imply that other processes, such as freeze-thaw, fluvial erosion, and mass movement, could now take place and add to the overall valley development. If such an explanation has merit, then under paleo-conditions in Europe consideration would have to be given to the impact of the large number of animals active at that time. It would not just be muskox, as in the example cited here, but would also include the mammoth and rhinoceros—both of which had the same soft coat that needed removing during molt. The parallels to the present-day muskox are *very* close and so there is no reason to *not* consider the role of these animals as an explanation for dells.

The geomorphic role of animals, although still relatively little researched, has been clearly shown to cause vegetation destruction and to produce enhanced erosion rates (see Butler, 1995 for a discussion and references). The effects so far documented are from *non*-permafrost regions and yet their impact has been shown

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Fig. 3. Degraded ice-wedge trough and zoogeomorphic trampling and erosion.

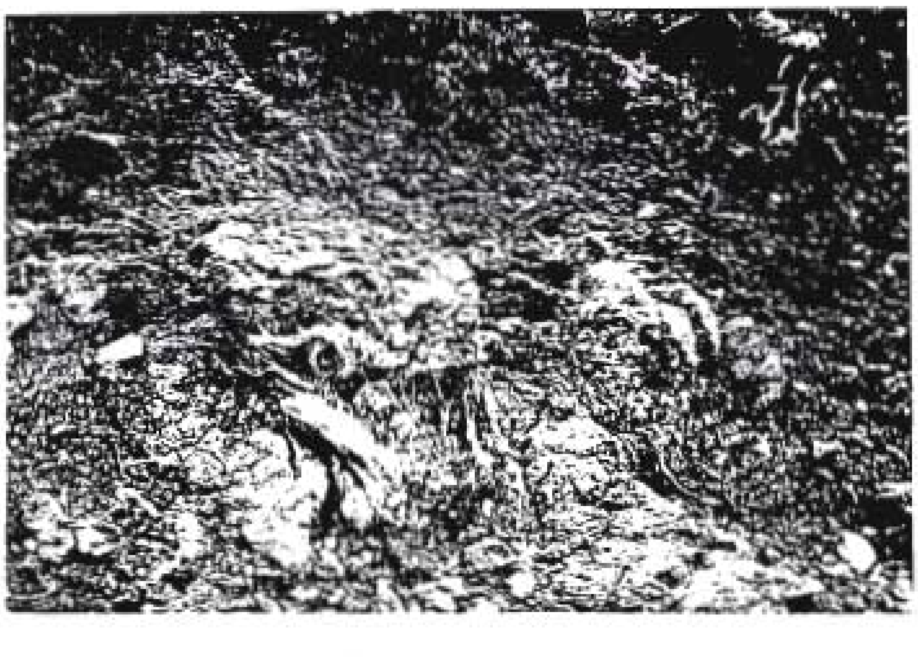

Fig. 4. Detail of eroded trough bank showing *qiviut* on the remnant blocks and in the hands of the person at upper right.

Fig. 5. "Dell" along an ice-wedge axis normal to the river (from where photo was taken) and below the eroded trough shown in Figures 3 and 4. Note the small stream, the wedge troughs intersected by valley widening, and the flat floor derived from transported sediments.

to be considerable. The impact of cows or horses in a muddy field, especially if one has to cross that field and it is raining, is well known by most (e.g., Kondolf, 1993)! The development of animal trails, by animals (e.g., antelope or caribou) following directly behind each other also is well known (e.g., Harper, 1955). The impact of large mammals such as elephant or hippopotamuses on trails and vegetation is also documented (e.g., Haynes, 1991). Thus, considering all of these, together with the highly fragile nature of the tundra and the thermal responsiveness of the underlying permafrost, it should, perhaps, not be surprising to see the geomorphic impact of animals in this region. It is perhaps more surprising, considering that humans are recognized for their impact, that animals have *not* yet been considered, especially with the greater numbers and species that were available in periglacial regions during the Pleistocene.

CONCLUSIONS

Animals remain a geomorphic agency little considered or quantified. The sensitive nature of permafrost coupled with selective vegetation destruction and erosion by large mammals may help explain the formation of dells. The ever-increasing set of animal-initiated geomorphic processes could lead to the development of a small valley, the dell, that does not relate to the main drainage system. The consideration of animal impact in permafrost regions is in need of further study. It might prove invaluable to investigate the exact nature and rate of impact of muskoxen on ice-wedge troughs and to study the valley-bottom sediment characteristics at such a site.

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Animals as agents of erosion at sub-Antarctic Marion Island

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Birds and seals act as agents of erosion on the coastal lowlands of sub-Antarctic Marion Island. Erosion primarily occurs where birds congregate at high densities to breed and is accentuated by the soft-foliage vegetation, easily eroded peaty substrate and wet climate. Different groups of birds - flying surface breeders, flying subterranean breeders and flightless penguins - each have particular erosive effects. Penguins cause most erosion and even affect bedrock. Bird activity causes slumping on steep slopes. The erosive effect of seals is minor.

Voëls en robbe dien as bewerkers van erosie op die kustelike laaglande van sub-Antarktiese Marion-eiland. Erosie kom hoofsaaklik voor waar voëls in hoë digtheid vergader om te broei en word beklemtoon deur sagte loof plantegroei, die maklik verweerbare turfagtige substraat en die nat klimaat. Verskillende groepe van voëls - vlieënd wat op die oppervlakte broei, vlieënd wat onderaards broei en vluglose pikkewyne, het elk 'n spesifieke erosie effek. Pikkewyne veroorsaak die meeste erosie en selfs rotsbeddings word geaffekteer. Die bedrywighede van voëls veroorsaak ineenstortings teen steil hoogtes. Die erosie effek van robbe is van minder belang.

Introduction

Nowhere is the proportion of sea to land greater than in the southern hemisphere between latitudes 40° and 60°S. This vast productive oceanic region supports large populations of seabirds and seals, all of which must return to land to breed and many also to moult. This results in large, seasonal concentrations of animals on the sub-Antarctic islands. The animals affect the islands in two ways: by transferring minerals and energy from the sea to the land (Burger *et al.* 1978, Siegfried *et al.* 1978, Williams *et al.* 1978) and by vegetation destruction and consequent erosion of the substrate. The role of birds and seals as agents of erosion at Marion Island (46°54'S, 37°45'E) is the subject of this paper.

Environmental Situation

Marion Island (area 290 km^2) is the exposed section of a submarine shield volcano (Verwoerd 1971). The island is composed of two basalt lava suites; an older, massive and glaciated grey suite, and a young scoriaceous, blocky and unglaciated black suite. During the last (approximately)

276 000 years the island has been glaciated on three occasions (Hall 1978, 1979a, in press) and the coastal lowlands include areas of glacial debris (Hall 1979a, 1981) and raised beaches (Hall 1977).

Precipitation on Marion Island averages 2 600 mm per year and occurs on 25 days in each month, mainly in the form of rain (Schulze 1971). Due to the extensive cloud cover (average 6 oktas) the radiation received at the ground is very limited, only 20 to 33 per cent of that possible (Schulze 1971) and thus evaporation is relatively low. During the infrequent dry periods the strong westerly winds may cause desiccation of the surface few millimetres of soil and peat. The heavy precipitation, coupled with low evaporation rates, results in continually saturated ground with frequent overland flow. The strong winds which accompany the rain (mean 32 km per hour with gusts up to 198 km per hour, Schulze 1971) may result in further physical damage to surface vegetation and peat, together with the transport of material up to granule size. According to palynological investigations (Schalke & Van Zinderen Bakker 1971) the climate has been very similar to the present since about 11 000 BP.

Marion Island has a typical sub-Antarctic flora with a paucity of species (Huntley 1971, Alexandrova 1980). Many of the plants are soft-foliaged and so are easily damaged by trampling (Gillham 1961). Much of the coastal lowlands, except where covered by the more recent black lava flows, or to a lesser extent, directly exposed to the prevailing westerly winds, has a peat cover which is up to four metres in thickness

Erosion by Seabirds

Twenty-six species of seabirds breed at Marion Island (Williams *et al.* 1979). These birds can be regarded as forming three groups: (1) surface-breeding species which can fly, (2) burrowing or hole-breeding species which can fly, and (3) flightless, surface-breeding penguins. Each of these groups exerts a different effect in terms of the type and amount of erosion caused.

(i) The flying, surface-breeding species

These are generally dispersed over the ground surface and thus, other than plucking vegetation for nest material, they have little erosive effect (Lindsay 1971). An exception occurs where concentrations of albatrosses (*Diomeda* and *Phoebetria* spp.) breed on steep, vegetated slopes. The combination of vegetation removal for nest material and the loading, particularly the impact when landing, of the birds on saturated high angle (35-60°) slopes frequently causes localised slumping. The material on the cliff faces is often beyond the angle of repose and is only maintained by the binding action of the vegetation. Hence, with the removal of some of this binding matter and the shock of landing this may cause the shear strength of the material to be overcome and thus slumping takes place (Taylor 1948). Occurrences of this were observed on the cliffs at Macaroni Bay where units of vegetation and soil in the region of 0,4 x 0,3 x 0,1 m were seen to be removed.

(ii) The flying, subterranean-breeding species

The numbers of burrowing petrels and prions at Marion Island are unknown, but have been estimated at many hundreds of thousands to millions (Van Zinderen Bakker Jnr. 1971, Williams *et al.* 1979). Investigations of petrel and prion burrows by Van Zinderen Bakker Jnr. (1971) show that excavation is extensive and may involve removal of up to 1 m^3 of peaty soil from each burrow. The excavated material is transported away from the burrow by a combination of surface run-off and aeolian action during wet and dry periods respectively. Unfortunately no accurate calculations can be attempted as the total number of birds is unknown and burrow size varies. However a rough estimate of the magnitude of erosion can be seen if each burrow is assumed to necessitate the removal of about 0,2 m^3 of material and a range of bird numbers from 600 000 to 1 000 000 is used. This would suggest the quantity of material removed, to date, to be in the order of 120 000 m^3 to 200 000 m^3.

In addition to the direct removal of material, burrowing has an effect on slope stability. Extensive removal of material, coupled with the frequent saturation of the substrate, causes slumping where slope angles are relatively steep (16° to 35°) due to the decreased shear strength of the soil.

(iii) The flightless, surface-breeding penguins

Their large numbers (the combined Marion Island penguin population is about 3,4 million individuals), high breeding densities (Fig. 2) and relatively high individual body mass makes them the most potent animal agent of erosion at Marion Island. The number of journeys each individual penguin makes annually between the sea and its breeding locality is dependent upon its breeding status and success. Birds which successfully rear chicks must, in order to feed the chicks, make many more journeys between the sea and the colony than those whose breeding attempt is unsuccessful. It is this movement of these birds, to and fro between their nests and the sea, which is the effective agent of erosion.

Each species of penguin has a habitat preference (Van Zinderen Bakker Jnr. 1971, Williams 1978, Hall 1979b). Erosion is most pronounced in colonies of king (*Aptenodytes patagonicus*) and macaroni (*Eudyptes chrysolophus*) penguins, both of which prefer flat or gently sloping terrain where they can breed at high densities (2,3 and 4,5 pairs per m^2 respectively). In colonies of these species, vegetation is entirely destroyed by trampling and manuring. The destruction of vegetation diminished progressively away from the colonies (Fig. 3), and Huntley (1971) estimated that 95 per cent of the vegetation is affected at 5 m from the edge of the colony, 50 per cent at 25 m, and only at 55 m is no effect discernable. Once the surface mat of vegetation has been destroyed the soft, wet peaty subsoil is "clawed" by the penguins and removed either on their feet or by surface run-off. At the two largest colonies, Kildalkey Bay and Bullard Beach (Fig. 1), a peat cover of up to 4 m thickness has been removed from areas of 110 000 and 82 000 m^2 respectively. Assuming an average peat thickness of 1,5 m over these areas, the total volume of peat removed from these two localities amounts to 2 900 m^3. King penguin erosion is also extensive in the Sea Elephant Bay to Log Beach area (Fig. 1) with severe erosion over an area of 630 000 m^2, while adjacent areas become increasingly affected as the population of penguins increases (pers obs).

Erosion is particularly concentrated along the main routeways between penguin breeding colonies and the sea. For example, at Bullard Beach 200 000 pairs of macaroni penguins must pass from a narrow landing beach up a steep valley about 30 m wide, in order to reach their breeding colony. Their passage has not only removed all vegetation and soil, but also grooved the massive grey lava bedrock and polished it to a mirror-smooth finish (Fig 4). The grooves, up to 150 mm long, 10 mm deep and 6 mm wide, occur where birds not successful in their jump from one level to another higher up slither back down, raking the rock surface with

Fig. 1 The locations and approximate areas of the major penguin colonies mentioned in the text.

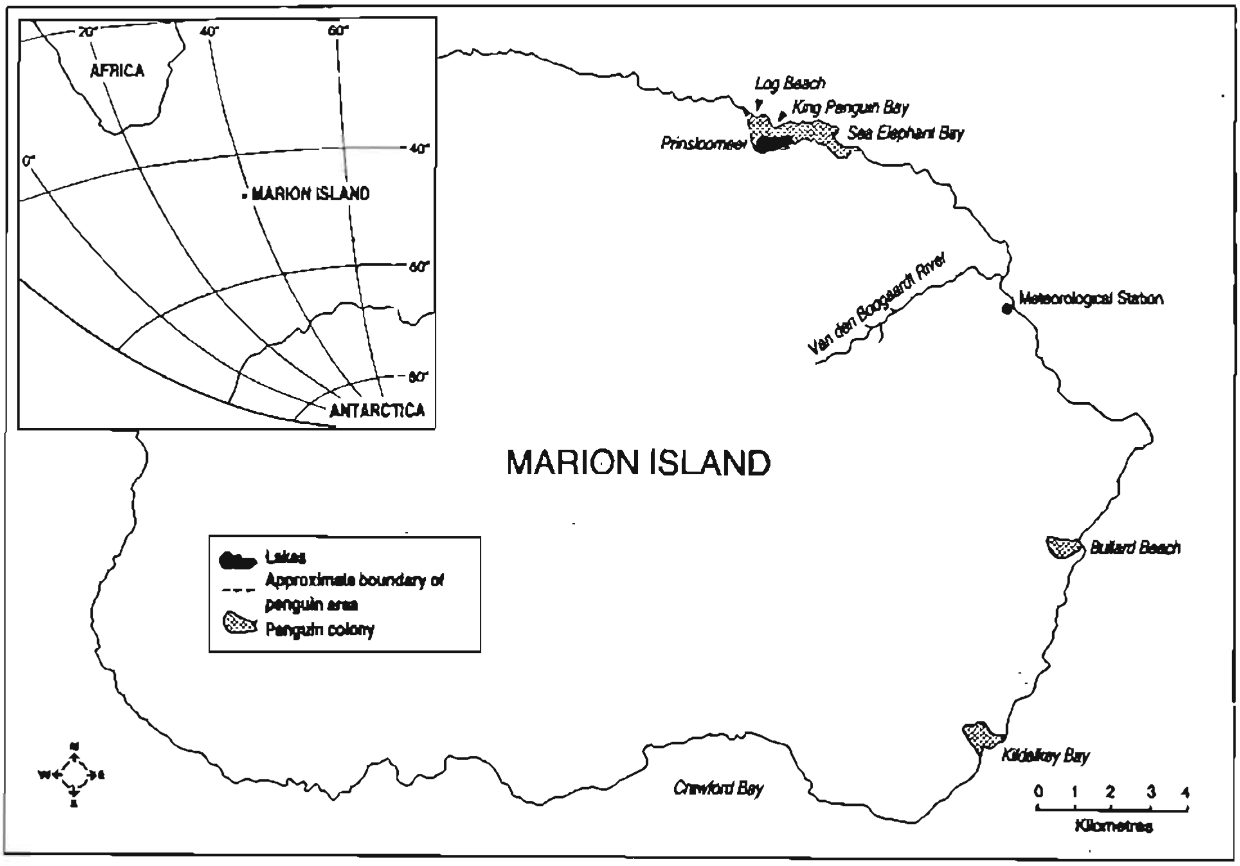

Fig. 2 Part of the macaroni penguin colony at Kildalkey Bay. Note the birds are resting on bedrock having eroded the peat cover (peat cliff in upper left corner is 4 m high).

Fig. 4 Grooving and polishing of a massive basalt along a macaroni penguin routeway at Bullard Beach.

their claws. This effect can also be observed at all the older macaroni penguin colonies on the island.

Rockhopper penguins (*Eudyptes chrysocome*) prefer broken ground, particularly the rugged black lava areas along the the coast. Where colonies spread onto vegetated ground, the effect is the same as at the king and macaroni penguin colonies. Grooving of bedrock by rockhopper penguins has been observed at Gough Island (Broekhuysen 1948) and at the Falkland Islands (Murphy 1936) but not at Marion Island. The absence of grooving by this species at Marion Island is probably a function of rock structure. The black Marion basalts are vesicular, with a clinkery or blocky structure. The friable nature of these lavas forestalls groove formation but does facilitate fracturing. The amount of erosion resulting from this form of destruction is impossible to assess, but freshly exposed rock surfaces have been observed.

Gentoo penguins (*Pygoscelis papua*) are the least numerous of the penguins at Marion Island but make the largest number of overland journeys per individual, because of their habit of returning to the colony each evening throughout the extended breeding season. Gentoo penguins prefer to breed on vegetated areas some distance from the sea but, unlike the other three penguin species at Marion Island, often change the location of breeding colonies not only from year to year but sometimes between the laying of initial and replacement clutches. Gentoo nesting density is also lower than that of the other penguins. Their low nest density coupled with the habit of changing colony location reduces the erosive effects of this species. Gentoo penguins do, however, wear deep tracks through the vegetation and peat surrounding their main landing beaches, and these tracks may form local drainage lines.

Erosion by seals

Elephant seals (*Mirounga leonina*) are only able to haul ashore on stony or sandy beaches (Condy 1977). During the mating season they remain in these areas, but during their annual moult they move inland for distances of up to several hundred metres. Elephant seals do not feed during their three to six week moult and so their body mass upon arrival is enhanced by the large fat reserves necessary to maintain them during the period of fasting ashore (Carrick *et al.* 1962). At the start of the moult, females may weigh about 900 kg and adult males up to 3 600 kg (King 1964).

The population at Marion Island is currently about 4 50C individuals (Condy 1977) but was larger in the past (Rand 1955). As seals move inland to moult, generally in the same localities used year after year, their great bulk flattens and damages the vegetation. The depressions they create in the moulting areas often act as drainage lines which concentrate run-off and initiate small-scale fluvial erosion and transport. Within the moulting area they create hollows which, through time, become deepened into wallows. Typical wallows are elongate with rounded ends and an outward bulge in the middle. Measurement of 12 such wallows indicated an average of 3,07 m^3 of material displaced per unit. This displacement may, in part, be explained by compaction although the effect of dilatancy on the granular mass is, as yet, unknown and may inhibit the amount of compaction which is intuitively thought to take place.

Wallows become full of water with a mixture of sediment and manure in suspension when in use. As a seal enters or moves within a wallow it displaces this mixture, and overland flow, especially during the rainy periods, removes sediment from the wallow site downwards towards the coast.

In areas where many moulting seals congregate, a number of wallows may coalesce to form a "compound wallow", which can vary from a combination of two to many tens of single wallows (Fig. 5). Two compound wallows measured indicated a volume of material displaced of 768 m^3 and 129 m^3. Because the seals moult only in certain months of the year, and may not use the same area for moulting in consecutive years, vegetation in moulting areas is not completely destroyed. Vegetation in the wallows themselves, whilst razed

Fig. 3 The severely eroded peat at the edge of the Kildalkey Bay colony.

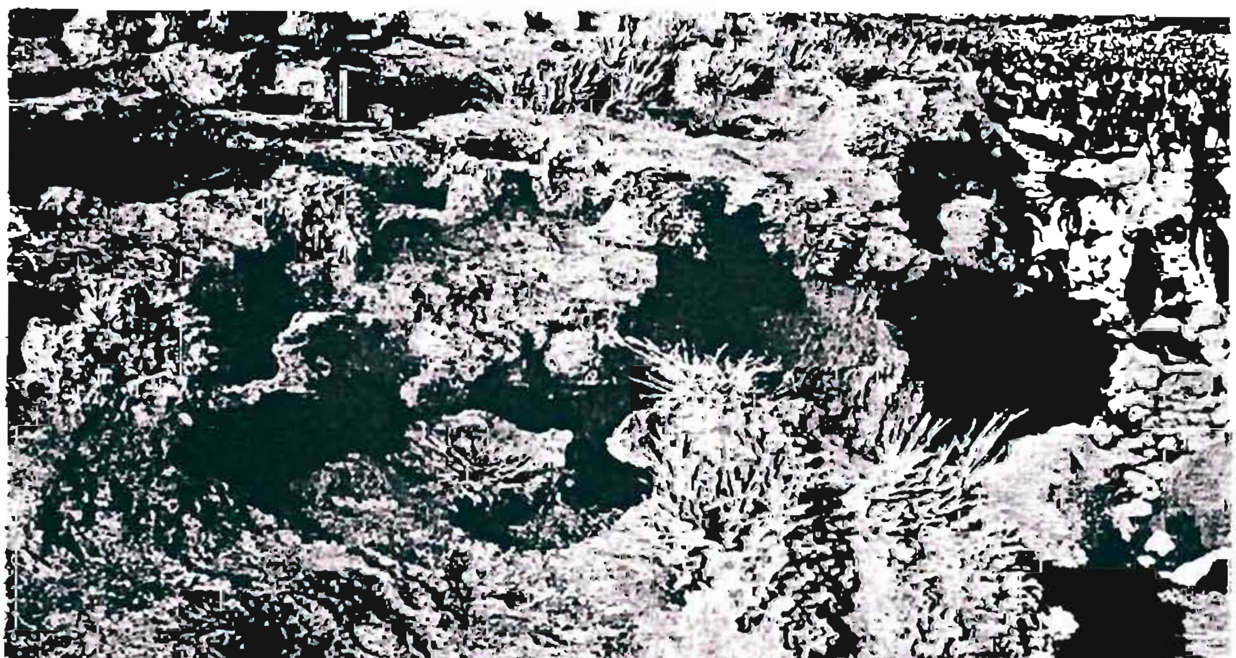

Fig. 5 A compound seal wallow surrounded by coalescing single wallows (near Trypot Beach). Note the damaged status of th vegetation within the general area of wallows and total destruction within the actual wallows.

during seal occupancy, rapidly regrows once the seals have left (Condy pers comm). However, a distinctive topography is developed in which extensive wallow systems are interspersed with vegetated ridges and humps (see Condy 1977; Plate 4).

Two species of fur seals (*Arctocephalus tropicalis* and *A. gazella*) are found at Marion Island. Both are confined to the west and northwest of the island where they breed on boulder and bedrock beaches. They make only limited excursions into vegetated areas and so have little erosive effect, although distinctly "flattened" vegetation has been observed (Condy pers comm).

The time element

Prior to about 17 000 BP the localities where the major present-day penguin and seal colonies are found would have been beneath an extensive ice cover (Hall 1978). However, during the glacial phase world sea level was in the region of 140 m lower than at present (John 1979) and so breeding could have taken place on the low-lying areas beyond the glacier cover. During the main glacier retreat stage, after approximately 14 000 BP, penguins and seals could have colonised the glacial outwash areas and so have begun to move inland as, at the same time, sea levels began to rise due to melting of the world's ice cover. However, on Marion Island (Hall 1982), as elsewhere (Dietrich 1980, Mörner 1978), extensive tectonic movements are considered to have occurred immediately upon the retreat of the ice. The location of Marion Island within a potentially active volcanic area resulted in the tectonics initiating volcanism (Hall 1982). Thus, as animals moved to higher elevations this would have been initially to areas of grey lava or glacial deposits as movement onto the black lava flows could only have take place once they had cooled sufficiently.

There is some evidence to show (Lindeboom pers comn that ammonia released from penguin guano is volatised an where blown onto land adjacent to the colonies, stimulate vegetation growth and hence peat production. Penguins ar thus associated with the production of peat which they ma later erode. C^{14} dates from some thick peats at Kildalkey Ba (Lindeboom pers comm) yielded a basal date of c. 14 000 B and only a short distance above a date of 7 000 BP. Th indicates that the area was certainly ice-free by 14 00 BP. The possibility then arises that the slow growth of pe between 14 000 and 7 000 BP was due to the lack of nutrien introduced by the fauna (penguins in this instance) and th rapid growth after that date results from an increase in th penguin numbers and hence introduced nutrients. Oth factors will obviously complicate this simplistic apprais but, nevertheless, it does fit the present state of knowledg

Other erosion agents at Marion Island

Subjectively it appears that the erosion by animals exceec that of any other agent currently active on the island althoug in the absence of any quantitative data, this is a tenuous statc ment. However, certainly glacial activity can be considere as negligible, since there is only a small area of permaner snow and ice and comparison of photographs taken 11 yea apart indicates no sign of ice movement. Rock breakdow by freeze-thaw activity, whilst undoubtedly active in the pas is at present limited except in the higher parts of the islan Within the coastal lowlands the incidence of freeze-tha cycles is low (Hall 1979c) and the few daily cycles monitore are of small amplitude. The porous nature of much of th bedrock inhibits runoff and stream activity so that, despite the heavy precipitation, there are very few permanent streams on the island. There is little observational evidence to suggest that these streams are currently causing extensive erosion, especially as the main part of their courses occur in the flat coastal lowlands rather than in the steep inland areas. After periods of heavy rain overland flow occurs but, except where aided by another agent (e.g. animals), does not cause any discernable erosion.

Animal erosion elsewhere in the sub-Antarctic

The erosive role of animals at other sub-Antarctic islands has been commented on by a number of workers. Holdgate (1964) notes that many areas on Signy Island are ". . . influenced to some extent by seabirds", particularly the tunnelling prions and that other areas are ". . . markedly affected by wallowing seals". Holdgate *et al.* (1968) found that seals caused extensive trampling and dunging on the beach at Bouvetøya such that it ". . . was almost completely bare . . .". In the South Shetlands, Lindsay (1971) notes the effects of large penguin and seal aggregations and how nitrogen enrichment resulting from the proximity to animal concentrations causes ". . . maximum vegetative growth . . .".

On Elephant Island (South Orkney Islands), Allison and Smith (1973) found that the effects of animals were of only local significance but that ". . . where there were colonies of birds their effects on the vegetation was considerable". Severe erosion at the periphery of a growing chinstrap penguin (*Pygoscelis antarctica*) colony was also observed. In the South Sandwich Islands Holdgate (1963) found that the barrenness of platforms at 15 to 30 m above the sea ". . . was probably accentuated by the presence of abundant penguins". Smith and Corner (1973) observed burrows in a peat bank in the Argentine Islands caused by Wilson's storm petrels (*Oceanites oceanicus*) but considered that ". . . they did not appear to be a cause of serious erosion". Conversely, Wace (1961) found that the major cause of peat instability on Gough Island is ". . . the disruption caused by the burrowing activities of the millions of ground-nesting sea birds . . .". On Kerguelen Island (Hall pers obs) observed that in some areas, particularly valley trains at sea level, burrowing by prions caused localised subsidence over much of the surface.

In one of the few direct investigations on the role of sub-Antarctic animals in modifying flora, Gillham (1961) notes that on Macquarie Island the main cause of damage was trampling and that this was related to animal numbers. Runoff was considered (Gillham 1961) to exploit the initial surface damage, whilst the wet nature of the ground was thought to minimise the amount of wind erosion despite the strong westerlies. At a large royal penguin (*E. chrysolophus schlegeli*) colony, 1,52 m of erosion in peat was measured along a tract to the sea. Finally, Ealey (1954) reconstructed the colonisation sequence, following glacial recession, for Heard Island. The postulated sequence, with consideration for local variations in species and topography, is very similar to that suggested here for Marion Island.

Acknowledgements

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Bryophyte-cyanobacteria associations on sub-Antarctic Marion Island: are they important in nitrogen fixation?

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Five Marion Island bryophyte species containing epiphytic cyanobacteria showed acetylene reduction in the laboratory at ca. 20 °C. Only Ditrichum strictum *exhibited reduction* in situ *at low (around zero) temperatures. This species occurs as a spherical cushion or ball on cold, windswept, rocky plateaux and contains a band of cyanobacteria a few millimetres below the surface of the cushion. The absence of acetylene reduction* in situ *for mire bryophyte species containing epiphytic cyanobacteria is ascribed to low temperatures during the incubation and it is thought that during the warm summer months fixation by bryophyte-cyanobacteria associations may significantly contribute towards the nitrogen status of mire habitats.*

Vyf van Marioneiland se briofietspesies met epifitiese siaanbakterieë het in die laboratorium by ongeveer 20 °C asetileenreduksie getoon. Slegs Ditrichum strictum *het* in situ-*reduksie by lae temperatuur (ongeveer nul grade) getoon. Hierdie spesie kom as sferiese kussings of balle op die koue, winderige, rotsagtige plato's voor en die siaanbakterieë kom as 'n band enkele millimeters onderkant die kussing se oppervlak voor. Die feit dat asetileenreduksie nie* in situ *by moerasbriofiete aangetoon kon word nie, word aan die lae temperatuur toegeskryf wat tydens die eksperimente geheers het. Daar word egter vermo dat stikstoffiksering in die warm somermaande 'n belangri bron van stikstofverbindings vir die moerasagtige gebie moet uitmaak.*

Introduction

Moss-cyanobacteria associations have been found to significant agents of nitrogen fixation in sub-Arctic bo Fennoscandia tundra and Arctic tundra (Granhall & Seland 1973, Granhall & Basilier 1973, Granhall & Lid-Torsvik 197 Alexander, Billington & Schell 1978). This has also been nou for alpine tundra and humid, cool-temperate oceanic isla ecosystems (Porter & Grable 1969, Alexander *et al.* 197 Englund 1976, 1978).

Bryophytes are an important component of the vegetati of sub-Antarctic islands (Taylor 1955, Greene 1964, Hébra 1970, Huntley 1971) and are often closely associated wi cyanobacteria which occur epiphytically on (occasional endophytically in) the leaves. To date, however, no assessme of the possible role of these associations in nitrogen fixati has been made for a sub-Antarctic site. Croome (1973) fou

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Animals as Erosion Agents in the Alpine Zone: Some Data and Observations from Canada, Lesotho, and Tibet

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Abstract

Animals can exert a very strong impact on erosion and sediment transport in the alpine. Although the alpine is recognized for its abundance of animals, animal-soil erosion interactions have been poorly studied. Animals exert a direct influence through their burrowing and digging for food and also indirectly by opening the ground to climatic and geomorphic influences, e.g. rain splash, needle ice, and wind erosion. It is this synergy that is important for alpine erosion. Because the alpine zone is subject to freezing, frost action, and snow melt, exposed sediments and/or the availability of drainage through burrows can have a marked effect on sediment transport and slopes. On the steeper, less stable alpine slopes, the effects of loading can cause failure that produces arcuate slip scars, the exposed faces of which can also be exploited by geomorphic processes. In an attempt to study the effect of animals in the alpine zone, measurements of burrowing and digging based on quadrats (5 m × 5 m) along several transects were made in the alpine zone of the Rocky Mountains of Canada. Results indicated an average of 0.0243 m^3 25 m^{-2} sediment displaced due to digging by rodents and a conservative estimation of sediment removal by rodents varying between 600 and 0.6408 m^3 km^{-2} yr^{-1}. Grizzly bears exerted the greatest erosional impact with as much as 0.4958 m^3 25 m^{-2} being measured. Observations (and limited measurements) relating to the impact of animals on the landscape were also obtained from Lesotho and Tibet. These preliminary findings are presented in an attempt to exemplify the various and interrelated effects of animals, climate and geomorphic process for the alpine. It is suggested that significantly more studies are urgently needed as the situation may be exacerbated by climatic warming and/or by the expansion of pastoralism resulting from attempts at sustainable development in developing nations. The impact of animals is an unquantified factor in many development studies, in geomorphic studies in polar or high altitude environments, and in management plans for national parks or the exploitation of natural resources (e.g. through logging).

Introduction

The geomorphic impact of animals has been recognized for a wide spectrum of species and over a wide geographic/environmental range (Butler, 1995). The impact of domestic, particularly grazing, animals is especially well documented (e.g. Evans, 1998; Trimble and Mendel, 1995), and these authors have shown that animal-induced erosion is widespread throughout the world's rangelands. There appears to also be a significant amount of work on zoogeomorphic effects in hot, arid environments (e.g. Yair and Rutin, 1981; Shachak et al., 1987; Gutterman, 1988; Alkon and Olsvig-Whittaker, 1989; Gutterman et al., 1990; Boeken, et al., 1995; Yair, 1995; Rutin, 1996), but there is much less recognition of the role of animals in cold, alpine regions. Thorn (1978, 1982) has discussed the role of pocket gophers in the alpine zone of the Colorado Front Range and shown that they may be the dominant geomorphic agent there, Price (1971) has also shown gophers to be a major geomorphic agent in the Ruby Mountains of Yukon Territory, Canada. Smith and Gardner (1985), in the Canadian Rocky Mountains, suggest that ground squirrels have a lesser effect than that found by Thorn and Price, but they do not take into account either subsurface debris transfer or the subsequent removal of mound material by other processes. Burrowing, alpine animals do unquestionably have an effect. Marmots (*Marmota*) are also identified as causing erosion and mass movement as a result of their burrowing in the alpine zone (see Butler, 1995). Grizzly bears (*Ursus arctos horribilis*) can, by digging for food and dens, be a major erosional contributor in the alpine zone (Butler, 1992), as can trampling and digging by caribou and mountain goat (Butler, 1995). Grooved-toothed rats (*Otomys* spp.) crop surface vegetation in the same way as grazing livestock (Mahaney and Boyer, 1986), including herded yak which can affect mass wasting, erosion, and soil degradation (Mahaney and Zhang, 1991).

Alpine regions are, despite their floristic richness, relatively fragile insofar as vegetation or surface material disruption may be difficult to restore. At the upper levels of the alpine belt and extending in to the subnival belt, vegetation extent and diversity decreases with proximity to the zone of continuous snow cover (Nakhutsrishvili, 1998). The zone under discussion in this paper is that encompassing the alpine zone, with its low herbaceous meadows, through to the top of the subnival where vegetation is almost absent. In this zone, the impact of animals, both wild

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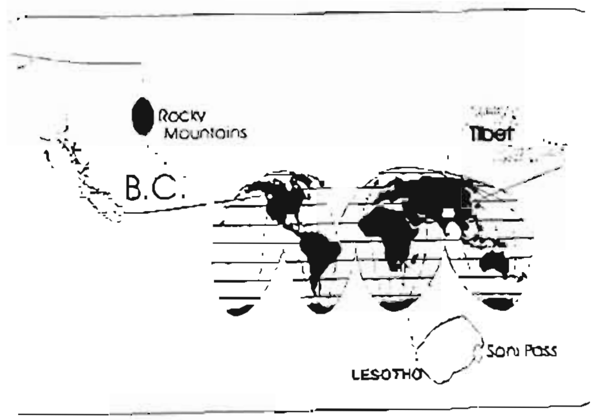

FIGURE 1. Generalized map to show the location of the three areas discussed in the text.

and domestic, can be significant. Happold (1998) has shown that the microclimate of such areas greatly impacts upon the nature and diversity of the small mammal species found there. Thus, not only can the animals themselves impact upon the very microclimate that may facilitate their survival but also climatic change may exert an influence on future fauna and flora composition and their interactions.

The material presented here constitutes a number of initial findings and observations which, although far from conclusive or substantial, we hope will show that animals are, directly and/or indirectly, exerting an influence on the alpine landscape. It is difficult to clearly identify the impact of the animals alone as distinct from the synergistic interaction with nonzoogeomorphic alpine geomorphic agents (e.g. the impact of needle ice). Future measurements and experiments will be aimed at elucidating these interactions. Thus, those given here are with the aim of increasing the awareness of the potential for the role of animals in landscape development and sediment transport within the alpine zones of our planet.

The Canadian Rocky Mountains

This study was undertaken on some unnamed ridges above the local treeline at an altitude of ca.1850 m in the Canadian Rockies (120°50'W, 54°14'N) (Fig. 1). During previous periglacial studies in this area (e.g. Hall and Meiklejohn, 1997) it became very clear that animals were having a significant impact on the landscape. Animal burrows were extensive on certain slopes, and the impact of grizzly bears digging for food was most significant, not the least for the subsequent impact of rain splash and needle ice on the exposed ground. Although this impact was readily observed, it was much more difficult to quantify what the impact really was. Measurements were therefore made using a quadrat approach across one major slope (Fig. 2 A-C) as a means to evaluating impact. The quadrats were 5 m × 5 m in size (Fig. 2C) and were undertaken in two lines (nos 1 to 11 in Fig. 2A, B), one above the other, in an accessible area seen to be used by a variety of animals. For comparison, quadrats 12 to 14 (Fig. 2B) were undertaken in what appeared to be a relatively undisturbed area. The resulting data are here presented.

Butler (1992) has provided a valuable discussion regarding the impact of the grizzly bear as an erosion agent in mountainous terrain and has demonstrated its impact through denning, digging for food, and trampling. The grizzly may weigh 160 kg or more and, with its massive shoulder muscles and claws of up to 10 cm in length, it is a formidable "digging machine" (Butler, 1992, 180). Although these bears roam large areas (Craighead et al., 1982), in their foraging for food, in two seasons (two years apart) we observed groups of grizzly bears following the exact same foraging route across the area in our study area. Such an observation appears to be in accord with other records as detailed in IGBC (1987) and outlined in Butler (1992: 184) where it is cited that "trails" are created as a result of persistent trampling and along which localized erosion may be concentrated. Although no clear trails were evident in our study, the zoogeomorphic impact of the bears did appear to be significant along the broad area through which they continually passed. During the summer season the bears must eat large quantities of foods (including carrion, bulbs, roots, insects, berries, tubers, fish, and rodents) to sustain them through the winter. During the summer months, the bears spend most of their time digging for food above the treeline and, in the opinion of Butler (1992), this is where the bear exerts its most significant geomorphic contribution. In his study Butler (1992) suggested that in Glacier National Park a conservative estimate of sediment removal as a result of grizzly bear activity was in the order of >136,000 m^3 yr^{-1}—more than the other geomorphic processes combined. Significant to this study was the observation that their greatest geomorphic effect appears to be at and above the treeline.

Measurements in the present study (Table 1) showed that the grizzly had the greatest erosional impact of all the animals present in this area. Those quadrats (Nos. 3, 7, and 9) where there was obvious bear activity had significantly greater volumes of material displaced (Fig. 2B), deeper damage (Fig. 2F), and were visually those most obviously affected; much of the impact by the burrowing animals was hidden beneath the dense alpine vegetation and not so readily visible (Fig. 2C, D). At quadrat 3, of the total volume of material removed (0.55667 m^3) 89% (0.4958 m^3) was as a direct and obvious result of grizzly digging; at quadrat 9 only 0.5% of the total volume displaced (0.372 m^3) could be attributed to *non*-bear activity. At quadrat 9 it was possible to discern the digging effect in four instances of a single paw. The average displacement of material was 1778.25 cm^3 (0.00178 m^3), affected an average area of 213 cm^2 and had an average depth of 9.25 cm. The area of the digging by these four events was 852 cm^2 and the area of debris cover resulting from the removal was a further 1585 cm^2 giving a total of 2437 cm^2 (or almost 0.25 m^2). For comparison, in the areas apparently least affected by bear activity but with a similar vegetational assemblage (quadrats 1 or 12 as two extremes), the *total* volume of material displaced was 0.0097 m^3 and 0.0024 m^3, respectively—only ca. 2 and 0.5% of that resulting from bear activity in quadrat 3. In fact, the four bear paw diggings cited above were, alone, equivalent to 87% of the *total* volume removed in quadrat 1 and 350% *greater* than the total volume for quadrat 12. The data presented here are specifically for the 14 quadrats, on the slopes in the area of quadrats 1 to 10 it was clear that both trampling and digging by bears was widespread. Only in quadrats 3, 7, and 10 were there obvious signs of trampling by bear but, again, the broader area on these slopes showed clear signs of trampling by the passage of bears, especially as in the 1998 season it was particularly wet and so it was easier to see by the squashed and muddied plants.

The effects of burrowing rodents, although significantly less than that of the bears, should not be underestimated. The vole (*Arvicolidae*), appears to be the most active burrowing agent in this present study. The arctic ground squirrel (*Spermophilus saturatus*) also exerted a significant impact in terms of burrowing and there is also some evidence of marmot activity (*Marmota caligata*) (Fig 2E) Holes with a diameter of ca. 3 cm, those

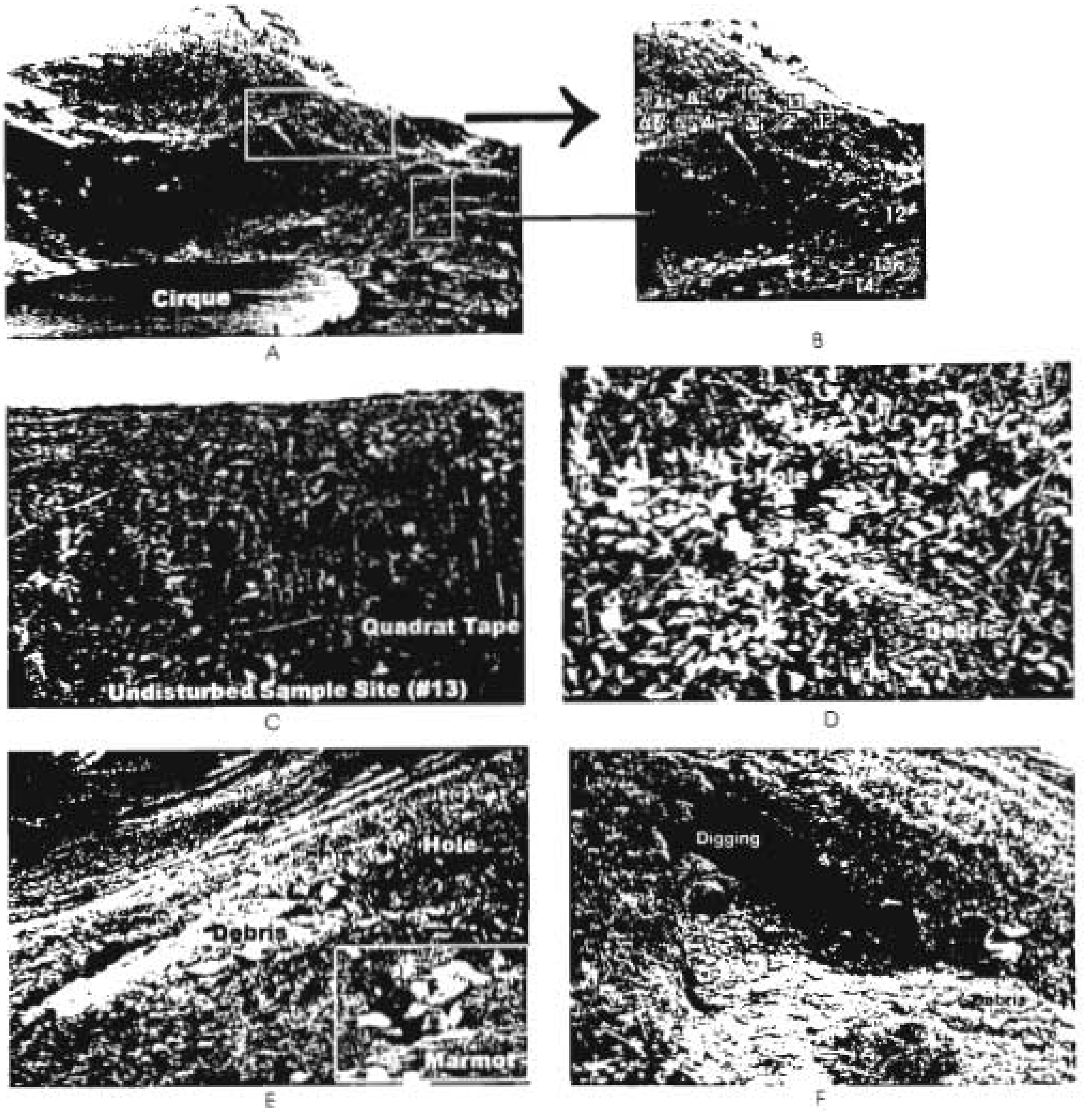

FIGURE 2. Rocky Mountains. A. Generalized view of study area with area of quadrats marked. B. Detail of study area with the approximate location of quadrats shown. C. Sample site no. 13, an 'undisturbed site' showing the dense vegetation cover and the quadrat tape marking a 5 m × 5 m site. D. Hole and debris caused by vole within the dense vegetation. E. Side view of a large digging due to a marmot (with photo of marmot inset) showing the extensive debris over the downslope vegetation. F. Digging due to grizzly bear showing the large excavation plus (part of) the downslope debris.

associated with voles, were the most common (Table 2) (Fig. 2D). Ground squirrel burrows have larger (ca. 8 cm diameter) entrances and can have short simple burrows only ca.1 m in length or, in some species, there can be a maze of galleries 3.5 to 15 m in length, with several chambers and an average of eight entrances (Nowak and Paradiso, 1983). Importantly, as Price (1971) clearly demonstrated, there is usually an aspect influence on burrowing such that the impact varies greatly in distribution and concentration; there can also be seasonal influences (Imeson, 1976). Where detailed studies have been undertaken (e.g. Price, 1971; Imeson, 1976), it is clear that burrowing rodents can be a major contributor to downslope debris supply (Fig. 2D). With a head and body length of 300 to 600 mm and a mass between 3 and 7.5 kg (Nowak and Paradiso, 1983), the marmot may dig

TABLE 1

Detail of measurements for the 14 quadrats undertaken in the Rocky Mountains (Canada)

Site no.	1	2	3	4	5	6	7	8	9	10	11	12	13	14
Slope angle (°)	25	30	32	13	12	10	40	36	38	32	28	34	15	10
Aspect (°)	94	100	100	84	78	90	76	72	86	96	94	64	64	58
Number of holes	5	11	4	3	1	1	2	8	2	2	5	11	2	8
Area of holes (cm²)	152	233	149	233	64	25	76	177	81	25	97	302	8	124
Area of debris from holes (cm²)	360	1149	1196	21	972	V[a]	187	V	V	V	730	1008	V	18
Area of diggings (cm²)	1873	1704	43689	2832	5103	373	19764	416	24073	870	215	405	16	143
Volume of debris (m³)	.0097	.0097	.5568	.0275	.0167	.0012	.2056	.0041	.372	.0014	.0011	.0024	.00006	.0003
$\bar{x}$ depth of digging (cm)	3.7	4.7	7.9	6.6	2.8	2.7	7.9	9.5	11.7	2.8	4.8	4.8	4	2.0
Area of debris from diggings (cm²)	2133	152	19584	868	131	290	4984	V	26702	85	223	1274	V	0
% of debris from diggings that is vegetated	50	85	6.5	69	83.3	88.9	57.1	100	4.8	57	58.3	58.3	100	100

[a] V = area of debris is all vegetated.

TABLE 2

Estimated number of holes resulting from voles and squirrels for each of the quadrats

Quadrat no.	No. vole holes[a]	No. squirrel holes[b]
1	1	4
2	6	5
3	2	2
4	1	2[c]
5	0	1
6	1	0
7	1	1
8	5	3
9	1	1
10	2	0
11	3	2
12	7	4
13	2	0
14	7	1

[a] Vole holes are recognized as being ca. 3 cm in diameter.
[b] Squirrel holes are 5 to 10 cm in diameter.
[c] One hole is of a size (7 × 18 cm) that suggests it might be a marmot.

burrows over 1 m deep (Fig. 2E), but with hibernation burrows to a depth of 5 to 7 m, and with tunnels 10 to 113 m in length (Zimina and Gerasimov, 1973). Some species also have two dens, one for summer and one for winter (Godin, 1977). There is a single mating season each year with a litter averaging 4 to 5 young. The young males are driven out during their second year (Nowak and Paradiso, 1983) thus providing a continuous increase in diggings. The survival rate of the young is said to have increased as deforestation has occurred and large predators eliminated (Banfield, 1974). In fact, the marmot populations have increased so much in some areas that they are considered a serious pest, not least because of the damage their burrows do to farm machinery (Banfield, 1974).

In the present study the area with significant burrowing was on slopes oriented northeast to southeast, only a little different from the preferred southeast orientation of slopes in the study by Price (1971). The erosion is much more than just the digging of burrows as the total number of burrows in the 14 quadrats was only 65, the greater number of these being produced by small rodents (as indicated by the small hole diameter). The total area of holes is, as part of the total erosion, very small indeed (Table 2) even when the observable area of debris from the holes is also included. The total area (note, not the *volume* removed) for holes and debris from holes is only 0.7387 m². This suggests that extensive amounts of debris have been removed and/or revegetated. Thus, despite the number of holes and substantial as the volume of material removed may be, there is no clear evidence of the debris. This finding may be significant for it indicates that much of the material removed to create the burrow has been moved out of the quadrat or dispersed so that it is no longer discernable. Removal of debris brought to the surface is discussed at length by Imeson (1976) who found that the effects of wind, rain splash, and needle ice removed much of the debris and dispersed it downslope. The absence of material in front of many holes may suggest that it has been removed downslope. In addition to digging for burrows by the rodents, there is also other extensive digging, possibly for insects.

For the 14 quadrats the total volume of material displaced by rodent digging amounted to 0.3395 m³, but this varied greatly from site to site. The greatest volume of displaced material for a single quadrat was 0.06087 m^3 for quadrat 3, the smallest was for quadrat 13 where the total volume from diggings was only 0.000064 m^3; the average was 0.0243 m^3. Importantly, in these diggings there was a substantial, but varying, area of the displaced material either absent or revegetated. We were unable to identify any displaced material in quadrats 8 and 13, at other sites this varied from 4.8 to 88.9% (see Table 1). As with the discussion regarding the holes above, this yet again suggests that a substantial amount of the material has been moved away from the site; although it is possible that the sediment has become revegetated. However, with the heavy rainfall, the substantial snow cover in winter and subsequent melt plus the influence of frost action during the fall, the likelihood of sediment removal is very real. A conservative estimate of sediment removal by rodents, based on the data from all 14 quadrats is 200 m^3 km^{-2} yr^{-1}, with a range from 600 to only 0.6408 m^3 km^{-2} yr^{-1}. Values such as these are substantially smaller than those found by Imeson (1976), seemingly less than that suggested by Price (1971) and Thorn (1978) where different measurement units make direct comparisons difficult, and perhaps closer to the findings of Smith and Gardner (1985) farther to the south in the Canadian Rockies.

Lesotho

Observations made in the Sani Pass region of Lesotho (southern Africa) (Fig. 1) at elevations between 2784 and 3200 m (Kotisephola Pass) show significant impact by grooved-tooth rats (*Otomys sloggetti*) (Fig. 3A), the "Sloggetts rat" which is known locally as an "ice rat." With a body length of ca. 20 cm and a mass in the region of 137 g for the male and 121 g for the female, these rodents are found all over Lesotho up to its highest point (3282 m). A study from South Africa (Swanepoel, 1975) showed that females of a similar species averaged two embryos and that they will reach sexual maturity within 3 mo. The species is almost exclusively herbivorous, it feeds on green grasses, tender shoots, grains, seeds, berries, roots, and bark (Nowak and Paradiso, 1983). When occurring in substantial numbers they can be very destructive. These rats build a nest of fine grass at the bottom of a complex burrow system (Fig. 3B, C). Conspicuous to this species are the system of runways and tunnels made through the vegetation to their favourite feeding areas. Such runs were very evident in Lesotho where not only did they occur in the vegetation but also in the lower margins of snowbanks, where the rats had produced runs under the snow. These runways under the snow were seen to act as drainage channels during snow melt. As with many groups within the order *Rodentia*, the grooved-toothed rats can be subject to peak populations that are followed by a crash in numbers when food supplies in a given area are exhausted. Such fluctuations may show a cyclic periodicity and can help explain the extensive areas of now unused burrows found in this area of Lesotho.

As can be seen in Figure 3 (B to F), these burrowing rodents exert a very strong geomorphic impact on the landscape. Not only do they burrow but they also remove vegetation over substantial areas. Individual burrows (Fig. 3B) show an entrance with a diameter in the order of 8 cm beyond which, at the freshly dug holes, there is a spread of debris 10 to 11 cm deep and with an area averaging 0.68 m^2. Their numbers are such that the burrows frequently coalesce to produce compound complexes of entrances, tunnels and excavations (Fig. 3C) where in this area (shown) of 0.41 m^2 there are at least 14 burrow entrances. At another site measurements indicated 12 holes in a 1 m^2 area where as much as 60% of the area had the vegetation destroyed. On slopes the burrows frequently coalesce in a downslope fashion (Fig. 3D) and, as such, provide incipient drainage lines. Meiklejohn (pers. comm., 1998) notes that he has observed water fountaining out of one hollow and continuing downslope along a series of eroded-open burrows; he has also observed frozen water filling a downslope burrow sequence as that shown in Figure 3D. On slopes (Fig. 3E) there can be seen complexes of burrows that join together and coalesce both laterally and downslope. Erosion here is substantial and a number of such eroded areas indicated an average area in the order of 0.679 m^2. On the flatter, open areas there is so much rat activity that not only are there extensive burrow openings (as in Fig. 3B) but there is also massive destruction of vegetation (Fig. 3F). It is likely that the massive destruction of vegetation, compounded by summer use of these areas by goats and cattle, produces periodic fluctuations in *Otomys* numbers. It was very clear in the field that some areas of extensive diggings were now almost free of rat activity.

Animals, mainly *Otomys* (of the indigenous animals), appeared to play a significant role in the landscape of Lesotho. Their effects on drainage and sediment transfer were marked. Not only was material moved by water, as in the development of drainage lines along a series of burrows (Fig. 3D) but also material was removed by wind during the winter (as happens to the material in Fig. 3B where desiccation cracks are already appearing). The alpine vegetation was severely affected to the point of total removal (Fig. 3F), at which point there is likely a cyclic crash of population until sufficient regrowth takes place to facilitate population growth once more. As cattle and goats are moved to these altitudes for summer pasture, the combined effects on the vegetation are compounded such that *Otomys* population decreases may be exacerbated by grazing. There is clearly a need for more detailed studies but the preliminary observations presented here show that *Rodentia* play a conspicuous role in the alpine region of Lesotho.

The Qinghai-Xizang (Tibet) Plateau

The Qinghai-Tibet plateau (Fig. 1), with an average altitude of ca. 4000 m, has an extensive grassland of alpine steppe, that experiences a significant geomorphic impact as a result of animals (Li, 1992). It is an area with an arid to semiarid cold climate, is partly affected by permafrost and experiences extensive periglacial activity (Wang and Derbyshire, 1987; Hövermann and Wang, 1987; Guo and Zhao, 1993; Harris et al., 1998). Such is the periglacial activity that a very wide range of forms and processes have been reported, including pingos, nivation, cryoplanation, blockfields, solifluction, rock glaciers, patterned ground, and tors (Kühle, 1985; Zhang et al., 1989; Zhu, 1986; Guo and Zhao, 1993; Anon, 1998). Solifluction is widely reported as active on slopes across the Qinghai-Tibet plateau (e.g. Kühle, 1985; Wang and Derbyshire, 1987; Zhu et al., 1998). Kühle (1985:183) defines three (vertical) periglacial zones based on solifluction—a lower level of "free and retarded solifluction" then a zone of "bound solifluction" and, at the higher elevations, "free and retarded solifluction" once again. Special note is made here regarding the recognition of solifluction as it will be shown that some forms suggested to be "solifluction lobes" are, in fact, the product of slope failures due to animals. Within this region, animal activity, human-induced and natural, is widespread, of geomorphic significance and "interactive" with some of the periglacial processes/landforms.

Field discussions (Zhu, pers. comm., 1998) indicated that the features observed on many slopes (Fig. 4A, as seen on upper, left center of photo) were considered to be solifluction lobes by a number of the Chinese scientists (see also Guo and Zhao,

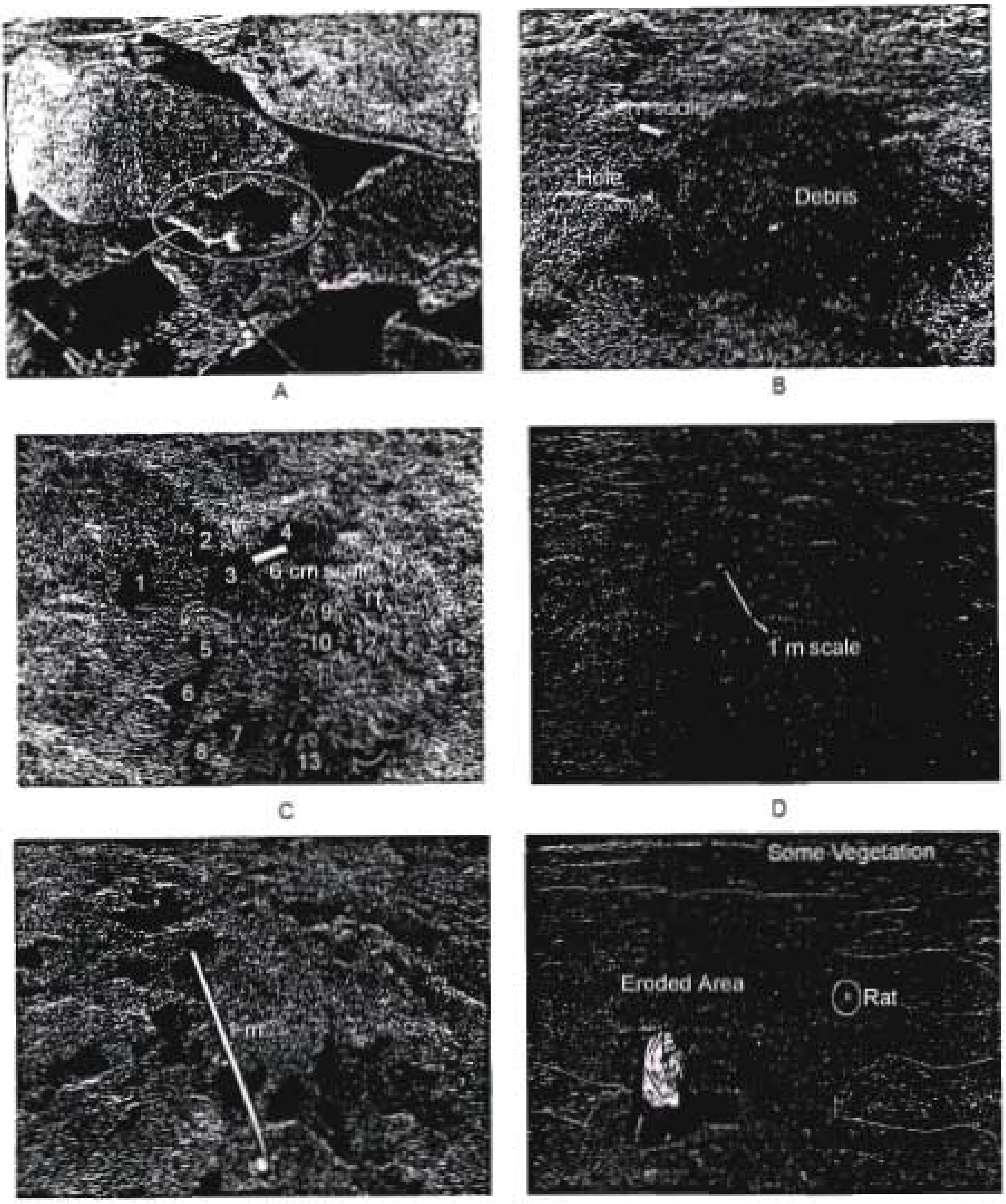

FIGURE 3. Lesotho study area. A. Grooved-tooth rat. B. Detail of a hole showing the spread of debris in front of the hole. C. Complex of holes on a slope showing the spread of debris plus the damage to vegetation. D. View down a gentle slope showing the series of diggings, one above another, that are frequently exploited as drainage lines. E. Detail of a complex of diggings that have coalesced. F. View across a flat surface showing the degree of vegetation destruction (and holes) created by an extensive rat community.

1993). While there is no question that solifluction certainly appears to be occurring on many of the slopes (as evidenced by "classic" lobate forms), the features indicated were, in reality, complexes of intersecting arcuate slip scars that, combined, give an appearance of "lobes." These arcuate failures are due to loading of the slopes by animals, particularly yaks. The forms differ significantly from the classic solifluction lobes found on the hillsides unaffected by animals. Unlike solifluction lobes which, in vegetated areas, exhibit a typical convex-downslope, vegetated lobate terminus (e.g. Fig. 8.13 of Ahnert, 1998 or Fig. 7.4a of Ballantyne and Harris, 1994) these features have an unvegetated, erosional face at the end of an otherwise vegetated lobe-like

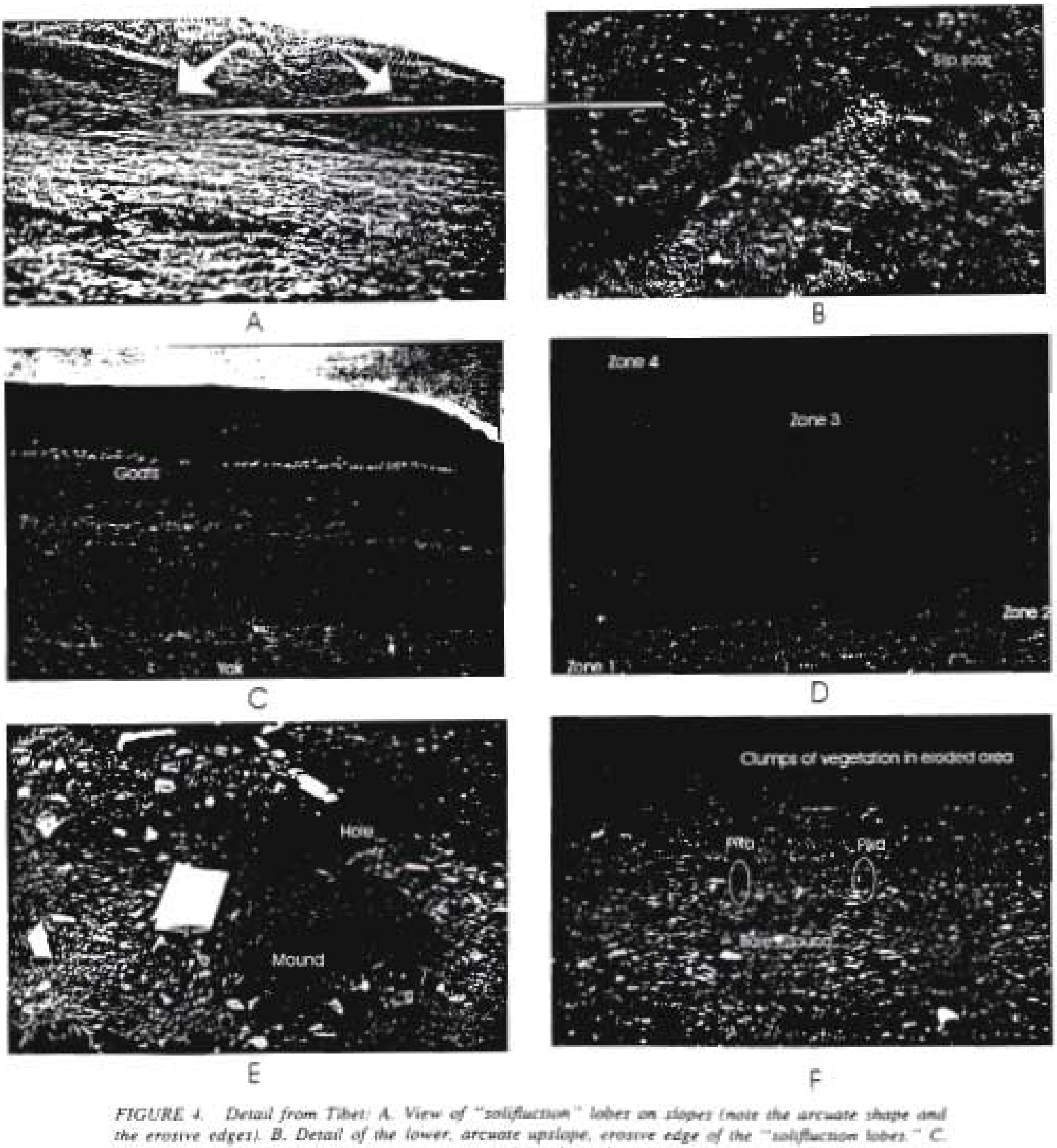

FIGURE 4. Detail from Tibet: A. View of "solifluction" lobes on slopes (note the arcuate shape and the erosive edges). B. Detail of the lower, arcuate upslope, erosive edge of the "solifluction lobes." C. View across a valley showing the extensive use of the slopes by goats and yak together with the erosion and arcuate slip scars. D. Closer view of area in C showing the suggested zonation due to animal erosion plus the typical arcuate-up slope slip scars that have been confused with "solifluction lobes." E. Detail of marmot hole showing the large amount of excavated debris now available for removal by wind and/or rain. F. View across a valley floor greatly affected by extensive pika activity.

form (Fig. 4B) . Further, these forms are found in areas used by pastoralists with herds of yaks and goats and, in most instances, the incidence of these forms increases immediately above herd-eroded lower slopes and then decreases towards the highest valley-side elevations. The yak (*Bos grunniens*) which feeds on grass, herbs, and lichens can weigh up to 1000 kg and congregates in large herds (Nowak and Paradiso, 1983: 1253). This combination of individual mass and herd size leads to the geomorphic impact of these animals such that approximately one third of the plateau is threatened by soil degradation or desertification (Li, 1992). Miehe (1994) notes the impact of yaks, sheep, and goats in both trampling and erosion with the role of yaks, both domesticated and wild, as particularly significant. He notes (Miehe, 1994: 335) that the yaks break the closed turf layer and expose it to continued degradation as a result of the action of needle ice.

There is every indication that although solifluction does occur on the slopes in this area the lobate forms are not themselves due to solifluction but rather are due to intersecting slip scars. As noted above, the argument for this is based on two complementary lines of evidence. First, that all these features have a vegetation-free, vertical, slip-face at their front and that these faces are arcuate in plan, concave downslope (Fig. 4C) (see also Sun, 1994: Fig. 18). Miehe (1994: 336) also refers to the "active cliffs of the turf patches." Such is the density of these forms that they, to the casual observation, create the illusion of lobate forms (Fig. 4A). Second, a view from the valley floor up the valley side indicates five (simplistic) zones associated with the impact of these yak herds (zones 1 to 4 can be seen in Fig. 4 D). On the valley bottom slopes there is complete removal of vegetation (the so-called "black earths" in Chinese literature, e.g. Zhao and Zhang, 1998; Ma, et al., 1998). Above this there is a zone where there is an extremely high impact of the herds but some (little) vegetation remains. This transforms into a broad zone where there is a high density of intersecting slip scars (the so-called "solifluction lobes"). Higher up, where only a few animals go, there are discrete slip scars that rarely intersect and, above this, the vegetated zone not yet impacted by the herds where no such slip scars are seen. Thus, although solifluction is an operative process in this periglacial region, it is clear that some of the forms identified are, in fact, the result of slope-loading, and consequent failure, by herded animals (Fig. 4B). Beyond this misinterpretation, the erosional impact of the yak herds is significant. Not only is there total erosion in the lower area (zone 1 and, essentially, zone 2) but the erosional impact is moved up slope as the yaks move there to find grazing, which thereby causes zone 3. The exposure of the ground in zone 3 along the arcuate scars opens the way to continued erosion by such as needle ice, rain wash, and wind, as can be seen on the lower left of Figure 4B (and also noted by Miehe, 1994). All of this means that, with time, so zone 1 migrates up hill; zone 4 is that just starting to be impacted by this move. The synergy of vegetation loss with periglacial processes (e.g. needle ice) and climatic factors (e.g. wind) that exploits this exposure and inhibits revegetation is what makes the role of herded animals in the alpine doubly devastating. Mahaney and Zhang (1991) show that the impact of yaks is such that aeration of the soil is so affected that plant regeneration is difficult, that compaction by the animals increases soil bulk density so that surface runoff, and hence rill erosion, is initiated and enhanced. Mass-wasting processes in the form of debris slides, debris flows, and earth-slips are said to be a product of overgrazing and, interestingly, they suggest that solifluction progressively increases as vegetation is removed.

On the lower slopes, and particularly in the broad valley bottoms, there is a significant erosional impact by pika (*Ochotona curzoniae*) and marmots (*Marmota caudata*); visually, the whole of the valley bottom is impacted by their diggings. In fact, Smith (1998) regards the pika to be so prevalent that it is a keystone species for biodiversity in this region. Miehe (1994) refers to the destruction of the turf mat by the burrowing action of both pika and marmots where, as with the yaks above, the exposure of the ground surface facilitates continued exploitation by needle ice and other geomorphic processes. Figure 4E shows the individual burrow dug by a pika into the face of an old slip scar generated by yaks (as described above). The volume of newly removed material is clearly seen next to the notebook at the burrow entrance. Figure 4F shows a view across a flat area which is severely affected by pika and marmot burrows (two pika are seen in the photo) such that there are now extensive areas of bare, deflated ground. The breaking of the turf mat exposes the material to the action of rain splash, needle ice and removal by wind (Miehe, 1994) much the same as was described for Lesotho above. There is much similarity between the erosion seen in Figure 3F for Lesotho and that of Figure 4F for Tibet just as the diggings shown for Lesotho (Fig. 3B) by rats is so similar to that by pika (Fig. 4E) in Tibet. Both areas experience the exploitation of the alpine zone by the combined effects of animals, periglacial processes and climatic factors such that regeneration of vegetation is all but impossible and may be made worse under a warming climate by increases (whilst the vegetation can sustain them) of the small mammal populations.

Discussion

Butler (1995) notes that the effects of animals have frequently not been considered in landscape evolution and their effects are rarely quantified and he makes the point that, in many instances, animals have, in the geomorphic context, been seen as no more than "interesting but minor curiosities" (p. 184). Also, it is not just the effect of a single species or genus but rather the synergy of a number of species coupled with their interaction with other geomorphic and climatic processes. In some instances an effect may be localized, in others more widespread, but both may play a role within a particular zone—in this example, the alpine zone. It is suggested that the alpine zone is a particularly sensitive one to disturbance as, once any vegetation cover is removed, the climatic-geomorphic conditions are such that the ground surface and/or available sediments are readily exploited. Further, this zone can be under pressure from the impact of humans through the grazing of herds, the management of wild animals, exploitation of natural resources (e.g. logging at the lower boundary of the alpine zone), or the use of ecotourism in national parks. All of these may be exacerbated by climatic warming and the effects this may have on both animals and geomorphic processes (e.g. through severe erosion caused by climate-induced increases of small rodent populations).

As stated in the Introduction, it would seem that some of the best data and observations regarding the geomorphic role of animals have come from studies in the arid regions of Israel, particularly the Negev Desert. There, not only has the digging effect of such as the porcupine (*Hystrix indica*) been well quantified (e.g. Gutterman et al., 1990) but also the "recovery attributes" of the dug-over and/or removed soil (e.g. Gutterman, 1988); this latter is very important in determining the longer-term impacts of digging. What is surprising is that in the polar and/or alpine zones where vegetation and/or the surface material cover is frequently very sensitive to external influences, so little has been undertaken regarding the zoogeomorphic effects and influences on processes and landscape development. Hall (1997) has shown how the impact of musk ox on ice wedge polygons may influence the location and creation of dells whilst, in the Antarctic, a number of studies (Gillham, 1961; Hall, 1977, 1979; Hall and Williams, 1981; Splettstoesser,1985) have shown the erosive effects of penguins and seals plus the impact they have, via trampling, on the very fragile vegetation of these areas. In alpine areas, the interaction of periglacial processes with the impact of animals seems to be an obvious one and yet so little has so far been identified. Even in his discussion on the impact of grizzly bears in the alpine zone, Butler (1992) does not refer directly to the interaction with cold-induced processes, associated with this zone, as an explanation for the exacerbation of the original zoogeomorphic effect. However, the removal of the vegetation cover and the exposure of disturbed soil significantly changes the microclimate at the site such that, in addition to the direct influence of rain in eroding the exposed sediments, processes such as pipkrake and segregation ice can now be more effective in inhibiting renewed plant growth and facilitating downslope movement of the debris.

Meiklejohn (1992, 1994) has discussed the valley asymmetry found in the high mountains of South Africa/Lesotho from a periglacial processes perspective but, to date, no one has yet considered the possibility of asymmetry arising as a result of zoogeomorphic effects in this alpine zone. As Price (1971) points out, the aspect-specific activity of certain burrowing species can exert sufficient geomorphic influence that it can impact slope development to the extent of creating valley asymmetry. This aspect-specific impact by animals can also relate to herded animals in the alpine zone where the sunny slopes are preferred to the colder, shaded aspect. The use of a specific slope, in this case because of heat differences, could lead to substantial vegetation destruction, increased slope failure, mass movement and sediment transport. Ultimately this may lead to distinctly different slope angles between the shaded and sunny slopes that have, until now, usually been thought to result from nonzoogeomorphic periglacial process differences. Boelhouwers (1988: 915), in considering nonperiglacial origins for the observed asymmetry, suggested that "the valley-side with the highest denudation rates will develop the shallowest gradient, i.e. the north-facing slope." Here, the north-facing slope is the sunny one and where many animals, indigenous plus herded, congregate for warmth and the better grasses. Thus, it certainly needs testing whether the higher denudation rates are as a result of animal activity. Equally, the extensive development of terracettes in the Lesotho/South Africa alpine region may be a result of former grazing practices or of grazing by earlier indigenous herds. In fact, Butler (1995: 88) makes the point that "the existence of terracettes may then offer evidence of the geomorphic influence of natural populations of grazing animals"—eland in the example of southern Africa at these elevations. This too has not been sufficiently considered in the investigation of terracettes in these areas. For information, other authors (e.g. Govers and Poesen, 1998) have identified vertical sorting in scree slopes as a result of animals and so this too might need some consideration in periglacial areas.

It would seem that the alpine zone is a particularly sensitive one to the effects of animals. There is a need for a better understanding of the interaction between zoogeomorphic and periglacial processes in this area and this is all the more so under any idea of climatic change, particularly climatic warming. Heal et al., (1998) show that there will be complex readjustments of both fauna and flora in cold regions. For example, the increase in summer and winter temperatures would affect vegetation significantly which, in turn, will affect forage abundance, the chemical composition and nutritional quality of the forage. For example, if the global warming is driven by increased CO_2 then there will be an effect on the C/N ratio of plants such that there may be a 10 to 15% decrease in protein concentration (Schäppi and Körner, 1997). In turn, this may have a marked effect on biodiversity and animal populations. Such effects would have significant feedback on anthropogenic use of alpine zones for grazing animals, in terms of both on-going use and for restoration of already damaged areas. The effects may not all be obvious, for example the prevalence of water-dependent soil fauna (such as earthworms) may be affected if sites become drier under a warmer climate with less snow available to melt (Heal et al., 1998) and these, in turn, may affect other fauna and flora. Clearly it is a very complex issue, but it will not be possible to consider the geomorphic or anthropogenic repercussions resulting from climatic change in the alpine zone if we do not understand the present-day interactions.

Conclusions

The plant species of alpine regions are generally vulnerable to change, particularly as many are slow-growing, have conservative growth strategies and a low ability to adapt to rapidly changing environmental conditions between generations (Heal et al., 1998). There will be direct and indirect (through impact on food quality) effects of climatic change on alpine animals that can have significant consequences geomorphologically as well as for the economy and culture of indigenous people (Heal et al., 1998). Thus, the preliminary findings and observations presented here fail to answer the many questions regarding the role of animals in the landscape of the alpine zone. However, it is hoped that attention has been drawn to a number of the issues and the potential for further study. In some instances (e.g. Tibet and Lesotho) there may be some urgency in this as the necessity for sustainable development is putting severe pressure on the alpine zone such that, if due consideration is not given to all the factors impacting on the use of this region, then there may be disastrous results. At the same time there is a need to understand the animal-climate-geomorphic interaction so that any adverse impacts of climatic warming can be planned for. On a purely geomorphic note, clearly animals are playing a far greater role in cold-environment landscape evolution and sediment production than has been recognized. Future studies need to pay more attention to the possible role of animals in these regions, particularly if we are to attempt to plan for the future of ecotourism and sustainable development.

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Presentation made to the Association of Professional Engineers and Geoscientists of British Columbia at their Annual Conference, Prince George, 1996.

I think it best that I present my background as this will explain the structure of my talk and, with a high degree of probability, the naivete of some of my observations. I have a doctorate in geology and am currently completing my D.Sc in geography - on the breakdown of rock in cold climates. I worked in the Arctic for 8 years and then in the Antarctic for nearly 20 years plus in the high mountains of Africa and South America - so my main geographic expertise is in high latitude or high altitude locations. My main interest is in process - geomorphic processes in cold regions. In this latter, for my D.Sc, I have been using a range of engineering techniques. However, I am not a practicing P-anything as my interests do not really lie in that direction. I have, however, worked in conjunction with such persons in Britain, France, New Zealand and South Africa - where, mainly, they are professional geomorphologists. Lastly, I am new to Canada, having only been here two years.

So, here I am to talk about hydrological considerations (on which I do not work), in forest areas (whilst I only work where there are no plants, let alone trees) and presenting to a professional body (whilst I am not an accredited professional anything). I thank my Dean for this wonderful opportunity!

So, can I do anything of meaning? It would be precocious of me to say "yes", but I do think I can offer an interesting perception. I took my task as one where, with my background, I was now forced (thanks to my Dean) into undertaking geoscience work within a forest environment. This, all the more so, as I am involved on behalf of UNBC in the new Professional Forest Engineer designation where the two professional bodies will be melded together to create this third, exciting new profession. So, I asked myself, how, as a complete ignoramus, would I start on this and where would it lead me? It is this, within a holistic approach to the topic, that I would like to comment on. Remember, I have NO background for this theme or geographical region but I DO have the geomorphic expertise from which to start.

What did I consider to be my framework that I was working within? As with the **overhead** I saw it simply as a system where there are forests (natural or created) within which, for economic reasons, logging must take place. As an adjunct to that logging there is the need for access - ie roads for vehicles. This use takes place in the context of our climate - and must be considered in the context of a changing climate - and with animal use (which may or may not change). The sum total of these factors affects slopes with respect to slope stability, sediment transfer, solute transfer and nutrient transfer. This transfer occurs as part of a geomorphic cascade of slopes to rivers to lakes where each may have a cascade within itself (i.e. from one drainage basin to the next). There is also a compounding element as each basin also has its own debris transfer system within it , the sum total of which is added to that brought into that basin and the resultant transferred downstream to the next system. Outside of rivers that exit direct to the sea, the likely repository of this sediment/solute transfer system is the lake. The question seemed to be, what is the impact of forestry on this cascade and how would any changes implemented by the new Forest Practices Code impact upon this? Also, are the approaches currently employed adequate to deal with these issues?

As with most research, my starting point was the available literature. Let me see what information is available on any and all aspects of this topic. Opportunistically I use a frequently updated "Cold Regions" CD-ROM which contains a database of some 750,000 references on any and all aspects of cold regions. It has extended abstracts and a sophisticated search system. So, I plugged in various appropriate terms and undertook searches. For me, this was the first interesting finding:

overhead

As can be seen, there are NO hits for the key words "Forests" with "Geoscience" and "Hydrology". "Forests" with "Geoscience" fared no better. "Geoscience" alone produced references to mining alone - very pertinent in the light of later observations from the Canadian Geoscience journal. "Forests" and "Hydrology" not surprisingly gave a multitude of hits - 366: but not as many as might be thought for the whole of our planet, not just Canada. Finally, Forest Resource Management provided 209 hits but none related to the issue under discussion here - primarily dealing with traditional land use, wildlife management or First Nations issues, largely related to landclaims.

This did not, outside of hydrology proper, give me much to go on. So I tried a few other appropriate terms:

overhead

Forests with footpaths gave no hits. Footpaths alone provided 5 but apart from one relating to Yellowknife, the rest were overseas. Roads had a good 6761 hits and Forests 2530, but put them together and I got only 166 which were mainly related to road construction and/or the impact of snowpack on construction. The impact OF construction was almost non-existent and that which were found related primarily to Japan or Russia. Again, not a very good start. Still really nothing to use as a starting point. Remember here, that whilst YOU may know of this or that report, nevertheless as an outsider this IS the sum total of what I could find.

Recognising that as good as this database is, it is highly likely that many items would be missed so I went to the library to see what I could drag out myself from journals, etc. Finding "Geoscience Canada" I thought, what better place to start? Even better, I find a recent issue (vol 22 #1&2, 1995) dedicated to "Future Challenges and Trends in Geosciences in Canada"........**overhead**

This volume included a discussion on the "Status and predicted development trends in resource and environmental industries" but this seemed to deal exclusively with minerals, oil, gas and technology - the word "forests" did not seem to feature at all!

I then thought to look in more detail at the articles to see what it was they hid within them. A matrix of research opportunities .....**overhead** held some interesting components although not one of them dealt with forest geoscience *per se*.
Three sections held within them highly pertinent attributes :
overhead - Global noted "Environmental impact of mining coal" but I thought this could include, or should include, environmental impact of forestry practices.
overhead - Fluids noted analysis of drainage basins, modeling water flow, source-transport-accumulation models and numerical modeling of depositional environments. Absolutely key issues here but NOT identified as germain to forestry *per se.*
Lastly, **overhead - crustal dynamics** has a number of pertinent attributes - landform response to change, sedimentary basin analysis, densifying of soils materials and landslides.

The key words can be identified as : **overhead**
All of which, within a geomorphic context, could be identified as **SOURCE-TRANSPORT-ACCUMULATION MODELS**. This terminology encompass the impact of forest practices on changing sediment sources, the nature and attributes of the transport mechanisms together with the location, nature and impact of the accumulation. In other words, it considers the changes in sediment cascade as a function of changing controls.

Before we look at some of the specifics, it is worth mentioning some other aspects that came forth from the Geoscience Canada report. It notes that the "environmental earth sciences industry is small and is likely to remain so." My question is why? This seems contrary to the rest of the world where environmental geology and environmental geomorphology are growth industries - as an example the dept of Geography at Utrecht (Holland) currently, so I am told by their Chair, produces over 600 environmental geomorphologist types per YEAR! AND they all find jobs!!! As Geology is a waning discipline worldwide at the moment, I also know of several departments who have moved exclusively to environmental geology in order to save their jobs. It also seems to contradict what is said later in that same article where it is stated "Of necessity, Canada has developed an expertise in landslide investigation.....For example, with the passing of the British Columbia Forest Practices Code Act, geoscientists trained or experienced in slope stability are in very high demand, as most areas of future logging will have to be studied in some detail before road building or logging permits are issued". Thus, contrary to their own earlier statement, there would seem to be a demand.

A brief digression is in order here. Slope stability and environmental management - the role of conservation practices. As Hart (1985) in his text "Geomorphology: Pure and Applied" points out - at this point most listeners will greet this in the same way, a feeling that it is old hat, that they have heard it all before. Therein lies the danger! Yes soil erosion is familiar and has been studied for over a century and may have even become less prominent in environmental discussions than it used to be. However, the notion is that, because conservation practices have been developed so the problem is solved. Wrong on three accounts:
1) Conservation practices have NOT always been successful. Mistakes are plentiful and some practices have even made the situations worse not better,
2) In many areas soil erosion conservation techniques are NOT employed, and
3) There remains large areas of ignorance among scientists. Basic data on the irritability of soil is scarce and it is important to recognize the symptoms of erosion at an EARLY stage before damage is irreversible.
As Hart states : "The message is clear. There is much to be done, and the applied geomorphologist must maintain his/her involvement..".

Before we start to look at those aspects which I would expect all here to be highly familiar with, I would like to add another component to the total picture - a component I would argue is usually ignored and certainly not quantified. This component could also take on a very different perspective and importance in logged areas. This is the role of animals. As this component IS unquantified so, until data ARE obtained, it is an unknown factor within our debris production and transfer system. The immediate response, I would hazard, is that it is insignificant - BUT this is a VERY biased and naive judgement! Certainly beaver are a MAJOR player in altering the landscape - and as beavers are associated with trees they are germain to this discussion. I do not wish to dwell on the role of beavers for they, of all the animal species pertinent to forests are the best known. In addition to their cutting of trees they also burrow in soft sediments, they use logs brought down by snow avalanches and even use rocks. Then they have the major impact of their dams. We clearly all recognize these attributes and see their impact, but as I say do not wish to dwell on them - although they are relevant to a later hydrological discussion regarding the geomorphic analysis of forest streams and the role of LWD - large woody debris.

Rather let us briefly consider some of the other forest users. First, why - what do they do?? In fact their interaction is quite complex and can include the removal and/or introduction of nutrients, destabilizing of slopes, active erosion, compaction of soil leading to increased surface run-off, aiding and increasing water access in to the soil, impact upon plants, etc etc. In some areas I have worked the geomorphic impact of animals exceeds that produced by (non-biotic) mass movement, glaciers and rivers combined! As I say, the data are extremely scarce so I will use three examples to exemplify the issue.

Grizzly bears - one of the symbols of the north. The grizzly just happens to be (according to Butler, 1995) a "major digging machine". Digging for bulbs, tubers and rodents they remove material from April through to November. During spring much activity is concentrated along riparian corridors - although this may not have a long-lasting geomorphic effect it does, nevertheless cause "pedoturbation" that has NOT been quantified in terms of its effects on solute production and transfer in to the river. During summer much of the bear's time is spent above the tree line and or/in the alpine ecotone. This is, technically, out of the forest area BUT as higher level forests are logged so the bear will move down into these opened areas. Bear diggings can be as great as 75 m long and 25 m across! Studies in the USA have shown a highly conservative excavation of $0.3m^{-3}$ per year per bear. Recognizing that there are likely in excess of 10,000 bears in northern BC this gives 3000 m^{-3} yr^{-1}. Using a value of 1 m^{-3} per bear per year - which is still, I believe highly conservative based on some measurements from this summer, then we get a value in the order of 10,000 m^{-3} yr^{-1} of debris removed by bears. This is then exploited by rain wash, snow accumulation, needle ice growth and over land flow to increase the amount of sediment removed by potentially the same amount, especially as most excavation takes place on relatively steep slopes. The impact of this erosion, the indirect impact of this erosion and the changes on this that higher elevation logging induces are all unknown values in the overall equation.

Moose are also classic animals of the north and the forests. Conservatively there are probably at least 30,000 of them. They can overgraze to produce degradation and increase erosion. They generate major changes to the soil as result of compaction which, as Lock (1972) showed prevents rapid infiltration of meltwater or rain and thus increases overland flow and thus surface erosion. None of this has been quantified or put into the overall equation. Moose, and to a lesser extent elk and deer, also wallow - the production of wallow pits by their hooves being common. A typical pit constitutes about 0.1 m^{-3} of sediment removal. This they do more than once when they are not in harem and as most moose in the north are NOT in harem a conservative estimate is 5 wallows/animal/year. That would generate 15,000 m^{-3} of sediment removal per year by moose ALONE - make it 10 wallows, a more likely number, and we get 30,000 m^{-3}

Put moose (plus elk and deer) and bear together and we could easily be in the order of 25,000 to 40,000 m^{-3}/yr of sediment erosion - and this DOES NOT take into account the synergistic effects resulting from this initiation.

Lastly, add in the marmots (and the other furry little animals) and they, by their numbers, could exceed all of the above! Marmots have been shown to change pH, chemistry and solutes in the soil. Values of debris removal are estimated at 28 m^{-3} per hectare. Ground squirrels are estimated at 0.44 $m^{-3} yr^{-1}$ $burrow^{-1}$ which can give 1.35 tonnes per hectare per year! We just do not have real figures for these animals NOR do we know what the impact is as forest changes impact the land. Further, they greatly affect water flow, increasing water infiltration by as much as 21% and also making that water accessible at a deeper level in the soils thereby changing the pedogenesis, the soil water chemistry and, ultimately, solute transfer in the forest system.

The end products of all these changes will, ultimately, have some impact on the river basin, the river chemistry and the chemistry of the lakes within the system. What this does to the fish environment is unknown.

When these processes - from moose, deer, elk, bear and burrowing animals - take place within the riparian component of the forest system then their effects might be multiplied many times over. In the temporal sense the effects are almost immediate whereas in the higher forest zones it may be decades or centuries until the response is felt (another issue to consider in our equation - the multiplier effect of time - currently NOT considered in this context in forest practice). As Dr Richardson, at a recent UNBC colloquium, discussed "Do fish eat trees? stream food webs and forest harvest practices?" He noted that small streams are highly dependent upon riparian areas for their supply of organic matter and many stream organisms are dependent on this material. Changes in riparian cover affect the food web and "are difficult to discern". This animal impact is an attribute NOT yet considered in this regard and yet it could have a profound effect - possibly for the better but also possibly with detrimental aspects on fish ecology.

Whilst we as individual scientists can look at any one component in the system there is the need for the system to be considered as a whole. The synergistic effects could be highly significant and if we do not consider this we may, and please excuse the pun, "miss the wood for the trees". In that regard there have been several wonderful studies in BC, studies where an attempt was made to look at the whole system (or at least all the perceived major components) But the problem is that these studies were ALL undertaken in the areas where people (and money?) are found - Vancouver Island and the lower mainland. This is good and is applicable to THERE. Climatically or topographically it is NOT relatable to northern BC - and so it is of limited use in forest practices here - and likely dangerous if we believe, for reasons of financial or economic expediency, to be applicable. Equally, should large sums be spent here to undertake a study the question would still arise as to its applicability to other areas and watersheds. Ultimately what is needed are broader-based criteria to facilitate a simple evaluation backed up by area-specific studies, especially in the context of the COMPLETE system. This is where, I believe, the geomorphologist *per se* has not been used to their fullest potential.

If I turn to the geomorphologist I feel I see much more of direct relevance to the forest situation than is dealt with elsewhere. As an example, using the latest issue of "Earth Processes and Landforms" I found two papers of direct relevance - and papers that introduced ideas that covered that "broader-based" calculation, rather than the area specific form. The first considered the estimation of soil parameters for assessing potential wetness and used a GIS for analyzing model comparisons. The study was undertaken in a National Forest and included soil conservation approaches. The models looked at soil parameters for input into a wetness model, determination of how parameters vary spatially and interesting stressed that it used "alternate literature-based estimates of soil parameters. It showed that wetness responds to how the soil parameters are estimated. The main point is that it was considering how to deal with these broader-based parameters within a forest area, evaluating them testing them and generating outputs which were then tested to see their validity: based on the outcome the models were refined to explain the real world situations. Thus, by this approach, and soil wetness, together with its temporal and spatial variability, is of key importance in mass movement, a whole forest area was characterized. Here, then, is an ideal tool for use in our forests; a tool that is using GIS and DTM already available together with basic soil characterization parameters and showing a methodology for testing and refining this so it can be more meaningful. It should also be relatively cheap and easy to undertake. My question is, I wonder if it is being used and even whether most people involved would have seen it? If the answer is yes it is used and/or is known then my question falls away. However, if it is NOT known and/or used that begs an even bigger question regarding the skills base needed for the professions involved in these undertakings.

The second paper, clearly more at home here, deals with the geomorphic analysis of forest streams and the impact of forest practices on those streams. How much closer to the present question can we get? This uses a multivariate statistical analysis of geomorphic variables to discriminate between pristine streams and those disturbed by land management resulting from timber harvesting. Again, here is exactly the approach we need - the broad based, non-specific approach, into which we can introduce local factors for better discrimination. It would not be appropriate for me to deal with the specifics of the paper but the last sentence of the Abstract clearly exemplify the importance of this type of approach : "Results of this study yield a much needed, objective, geomorphic discrimination of pristine and disturbed channel conditions, providing a reference standard for channel assessment and restoration efforts".

None of this is to say geomorphology has all the answers - quite the contrary.

Rather I would argue it is a case that the professional bodies currently lean too strongly towards the empirical engineering side and the older type geology approaches at the expense of the more system and process response orientated geomorphology - which IS geared to evaluating process response systems, including under changing conditions. As someone who has worked in a large (50 faculty) Geology Dept and in a research institute I know full well that few geologists or engineering types will even consider looking in a "geomorphology" journal ! However, this is where you will find presentation of applied material - the last paper cited being a combination of a forest scientist and a geologist - almost what the new "Forest Engineer" designation encompasses. Clearly there ARE many of you who do read widely but equally many certainly do NOT and thus my concern regarding the current professional designation and the need for the recognition of geomorphology *per se*.

To move on. If animals can affect sediment transport and removal of vegetation can affect sediment transport, what actually is the effect of forest practices on slopes? Part of that sediment transfer system will result from slope failure due to removal of the tree cover. Removal of the trees withdraws their binding effect on the surface sediments plus exposes the ground to increased direct precipitation. Gardner, in his consideration of Canada's north, notes that "Where present as forest cover, vegetation generally acts as a stabilizing factor. Where forest..is removed..stability is lost, thus increasing the potential for certain hazardous processes.

When looking at that treed slope and considering the so called factors of safety it is worth realising that, as Taylor in his "Fundamentals of Soil Mechanics" points out " It must be realised that many types of failure are possible with respect to the system as a whole and also that many types are possible with respect to individual points or individual parts of the system. It thus appears that there is no such thing as *the* factor of safety and that when *a* factor of safety is sued its meaning should be clearly defined". For engineers and professional geoscientists I am sure this is daily thinking BUT when considering a whole forest I, as an outsider, cannot but wonder how much is "generalized" such that the individual parts or points of a system are, in fact, not safe - and failure thus occurs. Seen in the context of the whole undertaking and, on an individual basis, it is seemingly a small component. true. BUT it is the multiplier effect of these small occurrences within the context of the total cascade together with the synergistic resultant in the sedimentary and solute sinks of lakes that should be the worry.

Debris flows are a major hazard resulting from forest removal on steep slopes in environments with large precipitation receipts where the stability of the slope material has not been adequately considered. The next series of three slides clearly show a debris flow resulting from forest practices. Debris flows are rapid downslope movements of water and debris that occur in saturated materials and occur frequently in well defined channels where the water flow can mobilize large volumes of debris. As with the slides shown here the effects of debris flows can be economically very costly where they cut roads. Where that debris, often in remote areas, loads a stream or river system then the effects may be felt much further down in the basin. As Gardner states "The process or its consequences may be translated downslope or far downstream into quite different environments".

In that regard, Roberts and Church looked at the impact of sediment budget as a result of logging. Although the results were not unequivocal in identifying logging as a producer of increased sediment production, it was found that in some basins clear-cutting DID increase sediments into the river system. In part, this is doubtful data insofar as the time component is not included. Returning to Gardner, he makes the point that the time from vegetation removal to failure may vary from site to site, and may take decades or longer in some situations. Thus, as with most of our studies, the time frame is too short for a meaningful evaluation. Certainly we can, as Roberts and Church did, look at short term responses - the longer term are not so easy to quantify but may be more hazardous. The conditions leading to failure may only take weeks or days but others may take decades - what is their impact? Especially when a major failure, that may take longer to initiate (high magnitude low frequency events as opposed to high frequency low magnitude events), could involve landslides of several million cubic meters of debris.

A typical example, as cited by Selby, of failure resulting from exceeding the threshold of resistance is that of a reduction in internal resistance by progressive weathering. Gradually weathering lowers the shear stress that is required to initiate instability. Thus, internal changes can consequently give rise to threshold conditions without the operation of large external stresses. The only requirement is TIME and, as the logging has greatly changed the weathering regime, unless this feedback is understood so failure may occur years or decades later. Selby also points out that a major cause of such extreme geomorphic events is the removal of vegetation cover and can result in 5 to 10 events per century.

The debris from all these events, large and small, is frequently stored, temporarily, in upland catchments. Overall it produces a series of infilling and terrace formation that is a temporary phase in mountain valley development - it is only temporary as the longer term trend must be incision and debris removal. This though is a process already operative but which has now been exacerbated by forestry practices. It is the synergistic outcome that we must be careful of and to start considering on a larger time scale than that of the immediate economic profits. Recognizing that landform change induced by human interference, as a result of changes in vegetation cover, has often accelerated erosion rates by 10 to 1000 times those of normal erosion beneath natural vegetation, so human-induced erosion (as opposed to climatically controlled accelerated erosion) can play a significant role in forest environments.

Ultimately all or a great deal of the sediments and solutes end up in the lakes.

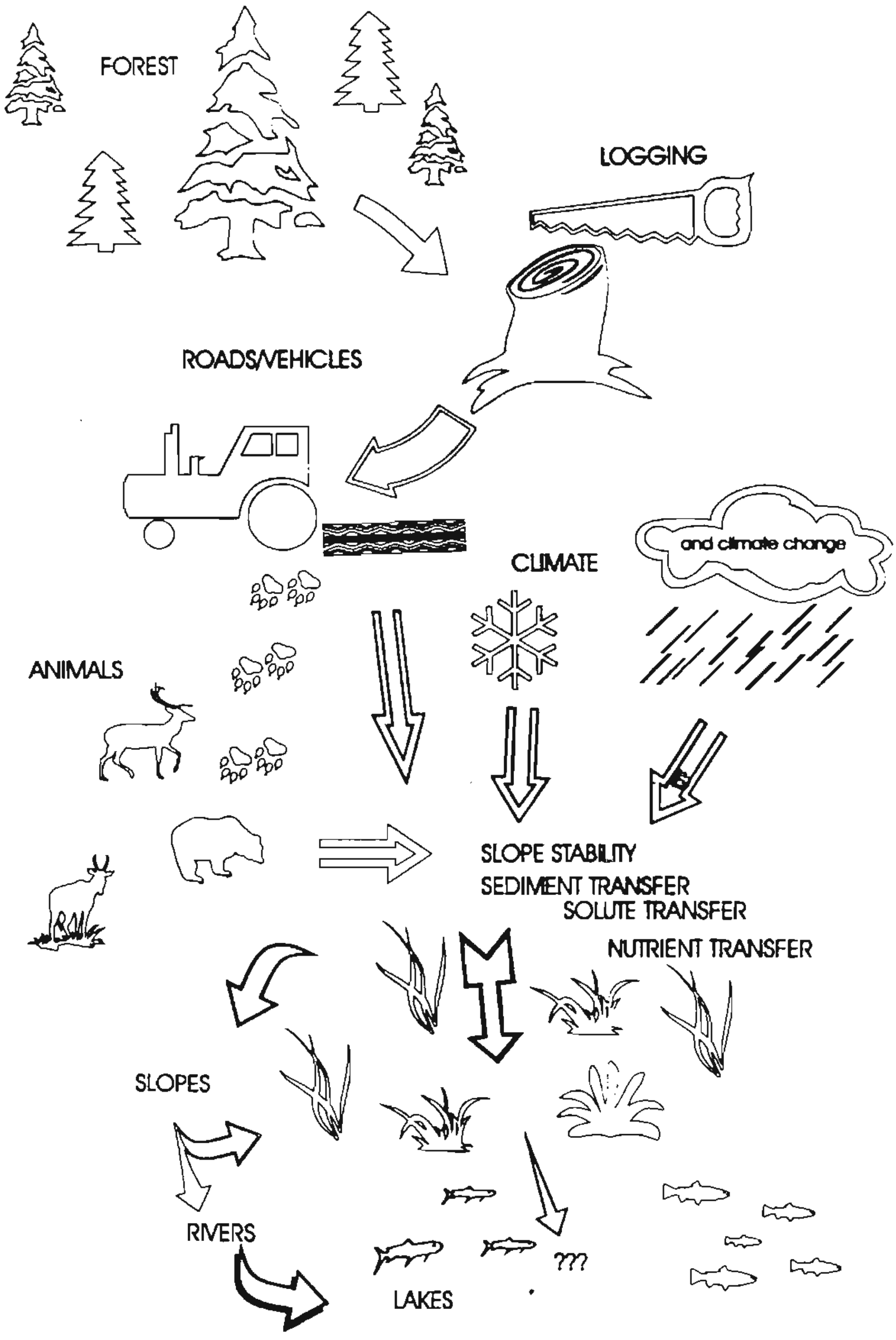
Recent reports on the effects of forest harvest on aquatic ecosystems show significant results. Forest removal shows that total water yield will increase and that this may be temporally unevenly distributed. As yield increases so too does stormflow and peakflow, both of which result in increased erosion. As a consequence of this the suspended sediment concentrations and total yield both increase. Concentration of nutrients increase but the relationship here has been difficult to quantify. In the lake system so those nutrient levels change significantly after logging; initially higher they can change to lower than initial levels after a period of time. Oligotrophic lakes (low productivity) to more autotrophic conditions (boggy, low pH limiting on production). Reduced forest cover will result in lower organic inputs to the lakes as a result of the reduced biomass. As the stream and lakeside vegetation are removed there is an increase in water temperature and this can affect thermal stratification. Tree loss will also increase wind fetch that can result in increased water turnover which will affect, amongst others, the thermocline. These thermal changes, though, will be largely dependent upon individual basin characteristics. Lake water levels, as a result in the flow changes, are likely to increase. All of this can have an effect on fish spawning, with wide, shallow lake basins expected to been impacted the most. This whole interaction and synergistic sequence of events with respect to lakes is poorly understood. It is the effect on spawning success, egg survival and habitat changes that are of immediate concern. Ultimately, it is the efficacy of the Forest Code prescriptions that need to be examined as well as the soundness of the ecological understandings which inform these prescriptions.

For me, as a non-involved scientist, it is this question of our "understandings" that form the basis of the decision making in the new Forest Practices code that is the key issue. We often think we know the relationships but frequently they have, in fact, not been tested or validated. From my own studies of weathering I know how monocentric most engineers and scientists are that it is frost action that causes rock breakdown in cold regions. Equally processes such as thermal fatigue are discounted upon (largely) the basis of a number of experiments undertaken in the 1930's. Frost action is NOT a singular process and has many faces and mechanisms - and is largely unproven as a field process. That is to say we have no true scientific evidence of its operation in the field - only conjecture. Equally, the discounting of thermal stresses is based on the poor experimental procedures of the 30's and have not been replicated (although the ceramic industry has done a great deal of work) but the results are regurgitated ad nauseum without verification. Hence we *think* we know the processes and interactions but in reality we do not and are continually compounding the original errors of science. With situations as large and as important as forests, rivers and lakes, where the interactive nets are so large and complex, I can only wonder how much we *really* know versus how much we presume is known or established without ever verifying it. With this as a foundation for a practices code and no in built checks and balances, the results could be disastrous.

Clearly there *are* many excellent studies undertaken in forest catchments - the Goodwin Creek Catchment study undertaken by the US Dept of Agricultural Sedimentation Laboratory or the Exe Basin study in the UK or the Dinosaur Park Catchment study in Alberta. They are all though, very small scale studies - the Exe being the largest at 1500 km^2 - and with very detailed instrumentation including such as photelectric turbidity meters, specific conductance monitors and purpose-built automatic sampling instruments. Large amounts of new information has come from these studies - information on runoff studies, the role of topography in controlling those processes, the dynamics of saturated areas, the role of piping in runoff production in upland areas, and pathways involved in solute generation and the factors controlling solute yields. Now, as I said earlier, we need a way to integrate these finding to obtain key identifiers we can utilize on the grand scale of the forests in Canada.

To end, I would again like to apologize for the inadequacy of my background to speak to body such as you. Much of what I have said I expect to be well known, well practiced and exemplified in the work undertaken here. However, two points still do arise. First, that the role of geoscientists in forestry areas does not seem to be well reflected in the literature, and secondly that the role of forest geoscientists in Canada does not seem to be identified as a major component *per se.* I would like to add my own third caveat: that I think the role of geomorphologists is under rated. We have excellent engineers - who view the problems from their own perspective. We have excellent geologists who see the geological viewpoint. The forestry issues are dealt with by Foresters. My question is, does the Professional Geotechnics syllabi reflect the needs and issues of the geomorphic component? For instance, does introductory structure and tectonics, mineralogy or petrology serve any useful purpose as courses for a geomorphologist? *Sensu stricto* I, having taken those courses, would have to say - yes it is sort of nice to have and yes it does provide a nice foundation. However, I am prejudiced and constrained by my own experiences and this may not be a good foundation upon which to base my judgements. I can say this on the basis that, for me, it is a sad state of affairs if the discipline has not moved forward in the last nearly 30 years since I took these courses such that they are perceived as required? Certainly I see the foundations of geomorphology having shifted significantly, so I would hope that the 'geology' has as well. The electives are "interesting" but what real purpose do they fulfil? Why would a geomorphologist want economics? Finally, in the Geotechnics specifics, why list applied geophysics and Quaternary geology which may be of no use whatsoever to a hydrologist or Physical hydrology which is of no use to someone like myself who is more concerned with weathering and mass movement processes? The electives are not so bad but fail to facilitate specific directions - like, for instance, people I know in other countries who worked towards coastal processes for beach protection/retention or preservation of buildings, monuments or cave art - like the work in France at Lascaux - all components of geotechnics. Or others, like in

Holland, who work directly towards undertaking Environmental Impact Analysis - and are seen as specialists in this. Sadly I think we are losing and/or watering down a whole field of expertise that could offer something to the problems we currently face. We do need that larger, more integrated picture created by a *range* of specialists who work together to answer the multidisciplinary problem. Perhaps we need to be looking towards a new paradigm in our approach to geotechnics in forest (and other) areas.



FOREST
LOGGING
ROADS/VEHICLES
CLIMATE
and climate change
ANIMALS
SLOPE STABILITY
SEDIMENT TRANSFER
SOLUTE TRANSFER
NUTRIENT TRANSFER
SLOPES
RIVERS
???
LAKES

CD-ROM with 750,000 refs

HITS for:

Forests + Geoscience + Hydrology
0

Forests + Geoscience
0

Geoscience
all = mining!

Forests + Hydrology
366

Forest Resource Management
209
(*traditional land use, wildlife, First Nations)*

Forest + Footpaths

0

Footpaths

5

(1= Yellowknife, rest = other countries)

Roads

6761

Forest

2530

Forest + Roads

166

(mostly impact on snowpack or of snow on road construction)

GEOSCIENCE CANADA
vol. 22 (1&2)
1995

"Future Challenges and Trends in Geosciences in Canada"

includes

Status and predicted development trends in resource and environmental industries

= minerals, oil, gas and technology

nothing on forests

[illegible]

Research Area	A. Understand Processes	B. Sustain Sufficient Resources Water, Minerals Fuels	C. Mitigate Geological Hazards Earthquakes, Volcanoes, Landslides	D. Minimize Global & Environmental Change Assess, Mitigate, Remediate
I. Global Paleoenvironments & Biological Evolution	• Soil development and contamination • Glaciation & its fluctuations • Quaternary record • Recent global changes • Paleogeography & Paleoclimatology • Paleoceanography • Forcing factors in environmental change • History of life • Discovery and curation of fossils • Abrupt and catastrophic changes • Organic geochemistry	• Mineral deposits through time		• Environmental impact of mining coal • Past global change • Catastrophic changes in the past • Solid earth processes in global change • Global database of present day measurements • Volcanic emissions and climate modifications
II. Global Geochemical & Biogeochemical Cycles	• Geochemical cycles: atmosphere & ocean • Evolution of crust from mantle • Fluxes along ocean spreading centers and continental rift systems • Fluxes at convergent plate margins • Mathematical modeling in geochemistry	• Organic geochemistry & the origin of petroleum • Microbiology and soils	• Seismic safety of reservoirs • Precursory phenomena and volcanic eruptions • Volume changing soils	• Earth science/ materials/ medical research • Biological control of organic chemical reactions • Geochemistry of waste management
III. Fluids in and on the Earth	• Analysis of drainage basins • Mineral-water interface geochemistry • Pore fluids and active tectonics • Magma generation and migration	• Kinetics of water rock interaction • Analysis of drainage basins • Water quality and contamination • Modeling water flow • Source transport accumulation models • Numerical modeling of depositional environments • In situ mineral resource extraction • Crustal fluids		• Isolation of radioactive waste • Groundwater protection • Waste disposal • In situ cleanup of hazardous waste • New mining technologies • Waste disposal from mining operations • Disposal of spent reactor material
IV. Crustal Dynamics: Ocean & Continent	• Landform response to change • Quantification of feedback mechanisms for landforms • Mathematical modeling of landform changes • Sequence stratigraphy • Oceanic lithosphere generation & accretion • Continental rift valleys • Metasomatism & metamorphism of lithosphere • State of the crust: thermal, strain, stress • Convergent plate boundary lithosphere • History of mountain ranges depth, temperature, time • Quantitative understanding of earthquake rupture • Rates of recent geological processes • Real time plate movements & near surface deformations • Geological Predictions • Modern geological maps	• Sedimentary basin analysis • Surface & soil isotopic ages • Prediction of mineral resource occurrences • Concealed ore bodies • Intermediate scale search for ore bodies • Exploration for new petroleum reserves • Advanced production and recovery methods • Coal availability and accessibility • Coal petrology and quality • Concealed geothermal fields	• Earthquake prediction • Paleoseismology • Geological mapping of volcanoes • Remote sensing of volcanoes • Quaternary tectonics • Densifying soil materials • Landslide susceptibility maps • Preventing landslides • Dating techniques • Real time geology • Systems approach to geomorphology • Extreme events modifying the landscape • Geographic information systems • Land use and reuse • Hazard interaction problems • Detection of neotectonic features • Bearing capacity of weathered rocks • Urban planning: underground space • Geophysical subsurface exploration • Detection of underground voids	
V. Core & Mantle Dynamics	• Origin of the magnetic field • Core mantle boundary • Imaging the Earth's interior • Experiments at high pressures & temperatures • Geochemical dynamics • Geodynamic modeling			

IV.
Crustal Dynamics:
Ocean & Continent

• Landform response to change
• Quantification of feedback mechanisms for landforms
• Mathematical modeling of landform changes
• Sequence stratigraphy
• Oceanic lithosphere generation & accretion
• Continental rift valleys
• Metasomatism & metamorphism of lithosphere
• State of the crust: thermal, strain, stress
• Convergent plate boundary lithosphere
• History of mountain ranges depth, temperature, time
• Quantitative understanding of earthquake rupture
• Rates of recent geological processes
• Real-time plate movements & near-surface deformations
• Geological Predictions

• Sedimentary basin analysis
• Surface & soil isotopic ages
• Prediction of mineral resource occurrences
• Concealed ore bodies
• Intermediate-scale search for ore bodies
• Exploration for new petroleum reserves
• Advanced production and recovery methods
• Coal availability and accessibility
• Coal petrology and quality
• Concealed geothermal fields

• Earthquake prediction
• Paleoseismology
• Geological mapping of volcanoes
• Remote sensing of volcanoes
• Quaternary tectonics
• Densifying soil materials
• Landslide susceptibility maps
• Preventing landslides
• Dating techniques
• Real-time geology
• Systems approach to geomorphology
• Extreme events modifying the landscape
• Geographic information systems
• Land use and reuse
• Hazard-interaction problems
• Detection of neotectonic features
• Bearing capacity of weathered rock
• Urban planning: underground space
• Geophysical subsurface exploration
• Detection of underground voids

Table 1.2 Matrix of Research Opportunities

Research Area	A. Understand Processes	B. Sustain Sufficient Resources: Water, Minerals, Fuels	C. Mitigate Geological Hazards: Earthquakes, Volcanoes, Landslides	D. Minimize Global & Environmental Change: Assess, Mitigate, Remediate
[illegible]obal Paleoenvironments, Biological [illegible]olution	• Soil development and contamination • Glacier ice & its history • Quaternary record • Recent global changes • Paleogeography & Paleoclimatology • Paleo[illegible] • Forcing factors in environmental change • History of life • Discovery and [illegible] of fossils • Abrupt and catastrophic changes • Organic geochemistry	• Mineral deposits through time		• Environmental impact of mining coal • Past global change • Catastrophic changes in the past • Solid earth processes in global change • Global database of present-day measurements • [illegible] and climate [illegible]
[illegible]obal [illegible]ochemical & [illegible]ogeochemical [illegible]cles	• Geochemical cycles, atmosphere & ocean • Evolution of crust from mantle • Fluxes along ocean spreading centers and continental rift systems • Fluxes at convergent plate margins • Mathematical modeling in geochemistry	• Organic geochemistry & the origin of petroleum • Microbiology and soils	• Seismic safety of reservoirs • Precursory phenomena and volcanic eruptions • Volume changing soils	• Earth mineral materials/medical research • Biological control of organic chemical [illegible] • Geochemistry of waste management
[illegible]ids in and on [illegible] Earth	• Analysis of drainage basins • Mineral water interface geochemistry • Pore fluids and active tectonics • Magma generation and migration	• Kinetics of water-rock interaction • Analysis of drainage basins • Water quality and contamination • Modeling water flow • Source-transport-accumulation models • Numerical modeling of depositional environments • In situ mineral resource extraction • Crustal fluids		• Isolation of radioactive waste • Groundwater protection • Waste disposal • In situ cleanup of hazardous waste • New mining technologies • Waste disposal from mining operations • Disposal of spent reactor [illegible]
[illegible]ustal Dynamics: [illegible]ean & Continental	• Landform response to change • [illegible] of feedback mechanisms [illegible] • Mathematical modeling of landform changes • [illegible] • [illegible] generation & accretion • [illegible] • [illegible] & metamorphism [illegible] • State of [illegible] crust: thermal, strain, [illegible] • [illegible] plate boundary lithosphere • History of mountain ranges [illegible] • [illegible] understanding of earthquake rupture • Rates of recent geological processes • Real time plate movements & near surface deformations • Geological Prediction • Modern geological maps	• Sedimentary basin analysis • Surface & [illegible] ages • Prediction of mineral resource [illegible] • Concealed ore bodies • Intermediate & [illegible] search for ore bodies • Exploration for new petroleum reserves • Advanced production and recovery methods • Coal availability and accessibility • Coal petrology and quality • [illegible] geothermal fields	• Earthquake prediction • Paleoseismology • Geological mapping of volcanoes • Remote sensing of volcanoes • Quaternary [illegible] • [illegible] soil materials • Landslide [illegible] maps • [illegible] landslides • Dating techniques • [illegible] • Systems approach to [illegible] • Extreme events modifying the landscape • Geographic information systems • Land use and reuse • Hazard interaction problems • Detection of neotectonic features • Bearing capacity of weathered rocks • Urban planning: underground space • Geophysical [illegible] surface exploration • Detection [illegible] underground voids	
[illegible]re & Mantle [illegible]ynamics	• Origin of the magnetic field • Core mantle boundary • Imaging the Earth's interior • Experiments at high pressures & [illegible]			

III. Fluids in and on the Earth	• Analysis of drainage basins • Mineral water interface geochemistry • Pore fluids and active tectonics • Magma generation and migration	• Kinetics of water-rock interaction • Analysis of drainage basins • Water quality and contamination • Modeling water flow • Source-transport-accumulation models • Numerical modeling of depositional environments • In situ mineral resource extraction • Crustal fluids		• Isolation of radioactive w[illegible] • Groundwater protection • Waste disposal • In situ cleanup of hazard[illegible] • New mining technologies • Waste disposal from mini[illegible] • Disposal of spent reactor

…earch Area	A. Understand Processes	B. Sustain Sufficient Resources Water, Minerals Fuels	C. Mitigate Geological Hazards Earthquakes, Volcanoes, Landslides	D. Minimize Global & Environmental Change Assess, Mitigate, Remediate
…obal …eoenvironments …Biological …olution	• Soil development and contamination • Glacier ice & it's inclusions • Quaternary record • Recent global changes • Paleogeography & Paleoclimatology • Paleoceanography • Forcing factors in environmental change • History of life • Discovery and curation of fossils • Abrupt and catastrophic changes • Organic geochemistry	• Mineral deposits through time		• Environmental impact of mining coal • Past global change • Catastrophic changes in the past • Solid earth processes in global change • Global database of present day measurements • Volcanic emissions and climate modifications
…obal …ochemical & …geochemical …cles	• Geochemical cycles, atmosphere & ocean • Evolution of crust from mantle • Fluids along ocean spreading centers and continental rift systems • Fluids at convergent plate margins • Mathematical modeling in geochemistry	• Organic geochemistry & the origin of petroleum • Microbiology and soils	• Seismic safety of reservoirs • Precursory phenomena and volcanic eruptions • Volume changing soils	• Earth science of materials/medical research • Biological control of organic chemical reactions • Geochemistry of waste management
…ids In and on …e Earth	• Analysis of drainage basins • Mineral-water interface geochemistry • Pore fluids and active tectonics • Magma generation and migration	• Kinetics of water-rock interaction • Analysis of drainage basins • Water quality and contamination • Modeling water flow • Source transport accumulation models • Numerical modeling of depositional environments • In situ mineral resource extraction • Crustal fluids		• Isolation of radioactive waste • Groundwater protection • Waste disposal • In situ cleanup of hazardous waste • New mining technologies • Waste disposal from mining operations • Disposal of spent ore for material
…ustal Dynamics: …ean & Continent	• Landform response to change • Quantification of feedback mechanisms for landforms • Mathematical modeling of landform changes • Sequence stratigraphy • Oceanic lithosphere generation & accretion • Continental rift valleys • Metamorphism & metamorphism of lithosphere • State of the crust: thermal, strain, stress • Convergent plate boundary lithosphere • History of mountain ranges: depth, temperature, time • Quantitative understanding of earthquake rupture • Rates of Earth surface processes	• Sedimentary basin analysis • Surface & soil isotopic ages • Prediction of mineral resource occurrences • Concealed ore bodies • Intermediate scale search for ore bodies • Exploration for new petroleum reserves • Advanced production and recovery methods • Coal availability and accessibility • Coal petrology and quality • Concealed geothermal fields	• Earthquake prediction • Paleoseismology • Geological mapping of volcanoes • Remote sensing of volcanoes • Quaternary tectonics • Identifying soil materials • Landslide susceptibility maps • Preventing landslides • Dating techniques • Real time geology • Systems approach to geomorphology • Extreme events modifying the landscape • Geographic information systems • Land use and reuse • Hazard interaction problems • Detection of geotectonic features • Bearing capacity of weathered rock	

Research Area	A. Understand Processes	B. Sustain Sufficient Resources Water, Minerals Fuels	C. Mitigate Geological Hazards Earthquakes, Volcanoes, Landslides	D. Minimize Global & Environme… Change Assess, Mitigate, Remed…
I. Global Paleoenvironments & Biological Evolution	• Soil development and contamination • Glacier ice & it's inclusions • Quaternary record • Recent global changes • Paleogeography & Paleoclimatology • Paleoceanography • Forcing factors in environmental change • History of life • Discovery and curation of fossils • Abrupt and catastrophic changes	• Mineral deposits through time		• Environmental impact of mining co… • Past global change • Catastrophic changes in the past • Solid-earth processes in global char… • Global database of present-day measurements • Volcanic emissions and climate modifications

Table 3.2 Matrix of Research Opportunities

Research Area	A. Understand Processes	B. Sustain Sufficient Resources Water, Minerals, Fuels	C. Mitigate Geological Hazards Earthquakes, Volcanoes, Landslides	D. Minimize Global & Environmental Change Waste, Mitigate, Remediate
I. Global Paleoenvironments & Biological Evolution	• Soil development and contamination • Glacier ice & its inclusions • Quaternary record • Recent global changes • Paleogeography & Paleoclimatology • Paleoceanography • Forcing factors in environmental change • History of life • Discovery and curation of fossils • Abrupt and catastrophic changes • Organic geochemistry	• Mineral deposits through time		• Environmental impact of mining coal • Past global change • Catastrophic changes in the past • Solid-earth processes in global change • Global database of present-day measurements • Volcanic emissions and climate modifications
II. Global Geochemical & Biogeochemical Cycles	• Geochemical cycles: atmosphere & ocean • Evolution of crust from mantle • Fluxes along ocean spreading centers and continental rift systems • Fluxes at convergent plate margins • Mathematical modeling in geochemistry	• Organic geochemistry & the origin of petroleum • Microbiology and soils	• Seismic safety of reservoirs • Precursory phenomena and volcanic eruptions • Volume-changing soils	• Earth-science materials/ medical research • Biological control of organic chemical reactions • Geochemistry of waste management
III. Fluids in and on the Earth	• Analysis of drainage basins • Mineral-water interface geochemistry • Pore fluids and active tectonics • Magma generation and migration	• Regime of water-rock interaction • Analysis of drainage basins • Water quality and contamination • Modeling water flow • Source-transport-accumulation models • Numerical modeling of depositional environments • In situ mineral resource extraction • Crustal fluids		• Isolation of radioactive waste • Groundwater protection • Waste disposal • In situ cleanup of hazardous waste • New mining technologies • Waste disposal from mining operations • Disposal of spent reactor material
IV. Crustal Dynamics: Ocean & Continent	• Landform response to change • Quantification of feedback mechanisms for landforms • Mathematical modeling of landform changes • Sequence stratigraphy • Oceanic lithosphere generation & accretion • Continental rift valleys • Metasomatism & metamorphism of lithosphere • State of the crust: thermal, strain, stress • Convergent plate boundary lithosphere • History of mountain ranges: depth, temperature, time • Quantitative understanding of earthquake rupture • Rates of recent geological processes • Real-time plate kinematics & near-surface deformations • Geological Predictions • Modern geological maps	• Sedimentary basin analysis • Surface & soil isotopic ages • Prediction of mineral resource occurrences • Concealed ore bodies • Intermediate-scale search for ore bodies • Exploration for new petroleum reserves • Advanced production and recovery methods • Coal availability and recoverability • Coal petrology and quality • Conductive geothermal fields	• Earthquake prediction • Paleoseismology • Geological mapping of volcanoes • Remote sensing of volcanoes • Quaternary tectonics • Identifying soil materials • Landslide susceptibility maps • Preventing landslides • [illegible] • [illegible] • Systems approach to geomorphology • [illegible] modeling the landscape • Geographic information systems • Land use and [illegible] • Hazard interaction problems • Detection of neotectonic features • Bearing capacity of weathered rocks • Urban planning: underground space • Geophysical subsurface exploration • Detection of underground voids	
V. Core & Mantle Dynamics	• Origin of the magnetic field • Core-mantle boundary • Imaging the Earth's interior • Experiments at high pressures & temperatures • Geochemical dynamics • Geodynamic modeling			

(National Research Council, 1993)

Key Words from
Matrix of Research Opportunities

Environmental impact of...
In situ mineral resource extraction
Preventing landslides
Water quality and contamination
Modeling water flow

Quantification of feedback mechanisms for landforms
Landform response to change

Analyis of Drainage Basins
Modeling of depositional environments
Sedimentary basin analysis

SOURCE-TRANSPORT-ACCUMULATION MODELS

Bears = 1700
Moose = 1793
Marmots = 32
Animals = 5772
Erosion = 5288

Bear + Erosion
0

Moose + Erosion
0

Marmot + Erosion
0

ANIMALS + EROSION
66
(Antarctic, Greenland, Africa, Russia, Iceland; domestic animals; only Canada geese on wetlands)

Grizzly Bear

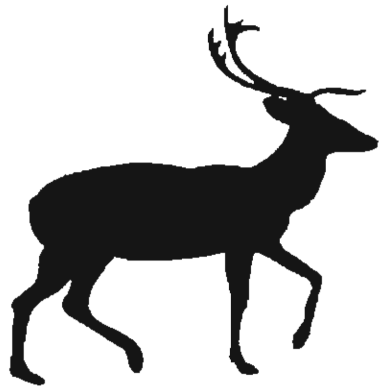
c.10,000 in Northern British Columbia

with
(based on studies in Yellowstone Park)
A minimum of 0.3 cubic metres / year / bear
of sediment eroded.

= 3,000 cubic metres per year

With 1 cubic metre per year

= 10,000 cubic metres per year



MOOSE

Wallow pits from hooves
typical =
1.2m long, 0.6-0.9m wide, 0.05-0.07 m deep
c. 0.1m3

minimum of 30,000 moose
and
5 pits per year

= 15,000 m3/yr sediment removal

Programme & Abstracts

The Society of South African Geographers,
The African Geographers Association,
The International Geographical Union and
The Commonwealth Geographical Bureau

incorporating

The Second Biennial Conference of SOSAG and
The Third General Assembly of the AGA

present

an International Conference on

Environment and Development in Africa: an Agenda and Solutions for the 21st Century

29 June - 3 July 1997
Eskom Conference Centre, Midrand, South Africa

hosted by

The Deparment of Geography, University of Pretoria, South Africa

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- *Potchefstroom University for CHE, South Africa*
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- *Santam, South Africa*
- *CDG, Germany*

ZOOGEOMORPHOLOGY: A FORGOTTEN FACTOR IN DEVELOPMENT?

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To most, animals as geomorphic agents have been considered "curiosities" or, frequently, as "natural" and thus not in need of study and quantification. In southern Africa the impact of domesticated goats has been widely 'recognised' but is still little studied; wild animals have all but been ignored. With development comes a likely increase of animals in rural areas, whilst the monetary returns from animal-centred ecotourism are well established and ever increasing. Any and all development is taking place under a changing climate which may be, in all likelihood, a warming one. That warming may, however, bring with it increased, of differing patterns of, rainfall. whilst the climatic changes may, in themselves be "good" from a development perspective, they do, nevertheless, bring with them a change in geomorphic processes which, if not accounted for, may be detrimental to that developmental growth. A factor *not* taken into account in that geomorphic response to both development and climatic change is the role of animals. Animals impact by such as tramping, wallowing, geophagy, burrowing, bioturbation and pedoturbation, by means of which they can effect significant sediment transport and export. Animals, within the geomorphic context, include *everything from terrestrial* and aquatic invertebrates, through birds to the whole spectrum of mammals. Further, it is not just the direct action of these animals in effecting erosion, which can itself be significant, but it is also the synergistic relationship with other geomorphic processes which we perceive as the detrimental ones are, in actuality, only operating as a result of initiation of conditions conducive to their operation by animals. Also, the geomorphic role of animals must be considered collectively and not singularly by species, for this too is synergistic. It is aimed that the role of animals as geomorphic agents will be outlined, their impacts(s) examined, their interaction with non-zoogeomorphic agents identified and their overall significance within a development policy under a changing climate discussed.

Presentation made at the Environment and Development in Africa conference in Pretoria, South Africa, 1997.

It would be fatuous for me to speak to you about development *per se*. I simply do not have the experience to do that. I am though, as a geomorphologist, able to offer some insights into aspects of development that might be significant - and might not have been thought of.

If "development" is a "stage of growth or advancement" as based on the verb "develop" which means "to become bigger, fuller, more elaborate", then there are terms in here which we must be careful of. The one I would like to focus on is "to become more elaborate" for I would see this as the biggest challenge of development.

We are living in a time which may or may not be unusual insofar as a key issue we are going to have to deal with is "climatic change". That, I could argue, we have always had climatic change is axiomatic. However, when we speak of climatic change today we often, if only subconsciously, link that with "global warming". Even global warming could be an oxymoron as it may precipitate the next Ice Age - hence the preference for the term 'climatic change'. Whatever the realities of this, in the short term, that time frame of 10 to 50 years, during which development is expected, then there can be expected to be climatic variability. If, following the general perceptions, of an anticipated warming resulting from our emission of greenhouse gases, then, from a developmental viewpoint in this region, this may be seen as "good". Why good? Simplistically because that warming could see longer, wetter growing seasons - in other words, there is not just the warming but, in some areas, an increase in, or extension of, the wet season - and this, here in southern Africa, could be seen as beneficial to development.

Again, simplistically, to help set my picture, animals are a significant factor in

Africa. Animals, both domestic and wild, play a variety of roles. There are cultural associations of cattle and/or goats where increase in their numbers not only reflects wealth and status, but also "survivability" or sustainability which, with development, it might be expected to see a significant increase in animal numbers within rural agricultural communities - all the more so if perceptions, resulting from the climatic change, imply greater carrying capacity. Equally, a significant factor in Africa and one related to development, is that of the role of animals in ecotourism. Visitors come from all over the globe to see the wild animals and, in so doing, bring millions of dollars in foreign exchange plus provide thousands of jobs. The recent improvements in economic relationships between National and private game parks with the surrounding communities has shown the way for future development in this sector. With increased demand and increased revenue there will be increased expectations and increased dependability by these communities. Again, the maintenance of these ecotourist centres, and their surrounding, ever dependent, infrastructure, will be constrained by the ability of these centres to maintain their drawability for paying consumers under this changing climate.

All of this is suggests the recognition of what might happen as a response to climatic change and our ability to safely exploit and/or mitigate against the effects. My expedition adage of "Luck favours the well prepared" seems highly appropriate to this scenario. It is the impact of the animals within this future that I would like to draw your attention to.

Why animals? Well, accepting their centrality to traditional and contemporary developmental needs, their impact *must* be considered. That is, not just their current impact but also their impact under a changing climate and the impact that the climate will have on the role of animals. Here I am NOT considering possible negative impacts, driven by climate, on animal numbers but rather anticipating, and within this argument assuming, within the time frame here, number maintenance and probable increase.

As geomorphic agents animals have largely been considered as "curiosities" or, as I heard from Parks Boards, as "natural", and thus not in need of study of quantification. With respect, this is oh so wrong! As an aside, and for you perhaps to consider in the back of your mind as I try to show why, in our future, we need to consider the geomorphic impact of animals, think about the impact of animals in the *past*. I will try to indicate the diversity and complexity of their *current* impact. But, *think* about assessing their role when herds were orders of magnitude greater than today - herds we have removed and/or constrained behind fences. Animals whose diversity of species, particularly of large mammals, has *decreased* since the Pleistocene. How much of the African landscape is, in reality, the product of zoogeomorphic influences? Our lack of knowledge with regard to this only serves to compound our ignorance of the present and the future.

As we now contemplate their current role, we must recognise that we cannot truly assess their significance as geomorphic agents unless we take history into account. As Butler states, when studying animals in the field we have in many cases exterminated the very agents whose geomorphic contributions we seek to understand! A valuable field of research would be, the past impact of animals in creating the present African landscape!

If geomorphology is the study of surface processes and landforms - and considerations of development require that, in some form, we must consider the geomorphic ramifications - then animals cannot be excluded. Animals are a conspicuous element of the landscape and its environmental systems. They are, however, all but ignored. Why? Briefly, Viles suggests the impact of Davision geomorphology is the cause because landscape development focused on landscape history at the regional macro scale rather than landscape *processes* operative at the meso or micro scale. As the Davisian concepts waned after

WWII, so *macro* processes received more attention but it is only in recent years we have begun to move towards the nano-and micro-level - the levels at which animals must *first* be considered. The lack of biological background of most geomorphologists has not helped either. Thus, some of the best work to date has come not from geomorphologists but rather animal ecologists and wildlife managers - and largely from *outside* Africa!

OK, so what do animals do? They, amongst others, trample vegetation and cause soil compaction, directly erode by such as digging or burrowing, they transport sediment, and they influence or change fluvial systems. Perhaps most importantly, they often *prepare* materials for impact by *other* processes. Their impacts are geographically widespread and geomorphologically significant. Each animal species may not only exert an influence in their own right, but they usually operate synergistically with other animals *and* other geomorphic processes. Singularly some species may not have a significant impact but *collectively*, through time, the animals can be *major* agencies; potentially exceeding those we normally consider - rain, rivers,wind, etc. There are some studies that recognize the *local* importance of one or a few species but they fail to recognize the quantity, significance and geographical ubiquity of geomorphic action by animals *collectively as a group*. For example, whilst goats may be seen to displace x-amount of sediment annually, but that this is spatially restricted, what about all the *other* animals working in that same area? Earthworms, rodents and ants, plus, possibly, birds that collect mud for their nests - and humans who also use that area??! In other words, we need to look at impacts collectively rather than singularly!

Further, it is *not* just the synergistic role of the animals but also their interaction with other geomorphic processes. In many instances, other processes of mass movement, erosion and transport *only* take place due to preparation *by* animal agencies. The creation of animal paths with the consequent compaction and vegetative trampling provides routes for water flow; removal of vegetation and exposure of sediments allows for needle ice growth or mass movement. Other geomorphic processes exploit the *combined* animal effects. Thus, this too *must* be considered. This is really the key - recognizing the *initiation* by animals of a geomorphic cascade where it is the end process that does the overwhelming damage.

So, what animals are we talking of here? A brief view of the only zoogeomorphic text, that by Butler (1995), shows impacts by earthworms, termites, ants, crickets, other beetles, bees, spiders, scorpions, snails, crayfish, crustaceans, sponges, fish, amphibians, reptiles, tortoises, crocodiles, a whole range of birds (from penguins to swallows, flamingos, and turkeys), seals, moles, porcupines, hares, rabbits, aardvarks, badgers, wild pigs, boars, elephants, rhinos, hippos, whales, walruses - the list is almost limitless! Some brief examples to exemplify:

In the context of southern Africa it will be extremely obvious to all here that not only do you have a marvellous diversity of animals but that they also play a major role in many aspects of development - and one that may be exacerbated by climatic change. The key, I think, is that this issue has been so little considered that no cognizance of potential impacts, good or bad, has been taken into account. Worse still, there exists no data base whatsoever to act as a foundation - be it in ecotourism, game management or in rural development. As introduced earlier, the whole problem is compounded by the many synergistic relationships. It is this synergistic impact - the compounding of different zoogeomorphic attributes with various other geomorphic processes - and *all* exacerbated by a changing climate.

Recognition of the problem not only assists mitigation but it can also allows for its direct *use* under climatic change - rather than having, at a future date, to overcome its negative impact. Thus it is a two pronged problem.

(1) What are the detrimental impacts?

(2) What can be the *positive* attributes?

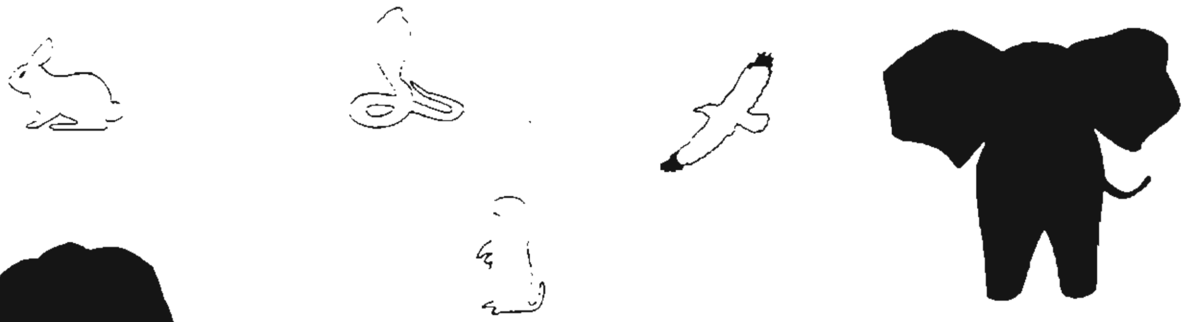
I have outlined the nature of the negatives but we can get positive responses. In much the same way as insects can be used as a biological control of diseases or against other insects, to stop damage without using harmful pesticides, etc., so too can animals be used in a positive fashion. Certain animals are valuable for increasing pedoturbation, aeration of the soil, increasing soil drainage, increasing fertility, or by using one species to mitigate the damage caused by another to change, and thus improve soil chemistry, etc. An obvious, but not directly geomorphic example, would be the Australian use of dung beetles to overcome the problems caused by sheep - this *does*, however, indirectly affect soil chemistry.

Thus, yet again, the recognition of these positive attributes, and how *they* may change under and changing climate, *is* a *major* development tool. This, though, takes me full circle back to the lack of recognition and data. Thus, what to do?

Simply, first is to realize that this *is* a development issue. Even such as the Lesotho Highlands Water Scheme has not identified the sources of sediments that infill the dam(s) insofar as they have not considered the zoogeomorphic attribute(s). Will this increase or decrease and what impact do these have on dam life - particularly with the ensuing development. Once recognized then much of the rest will fall into place - albeit too late.

This is then a major research area - to quantify *both* the individual components *and* their synergistic relationships. Obviously both managerial and cultural attributes must be kept in mind - but these do *not* negate the necessity for understanding. The problem comes rather with implementation. This opens the important door of recognizing, *and* involving, both the natural *and* the social scientists in order to solve what at first may appear a purely geomorphic problem.

Thus, as a development issue, there is the pressing need, at the one level, to acquire data on the geomorphic - both zoo - and non-zoo-geomorphic - attributes. Another level is that of the integration of findings regarding climatic change. This is imperative as these geomorphic processes are being, directly and indirectly, affected by these changes in climate. However, at the third level, impacts of society on the problem *must* also be considered if any meaningful development protocols are to be developed. For instance, it is of no purpose to implement a geomorphic response that impacts in a detrimental, and thus conflict-orientated, cultural requirement. Equally, to detrimentally impact a game park, to the disadvantage of the surrounding infrastructure would be equally disastrous.



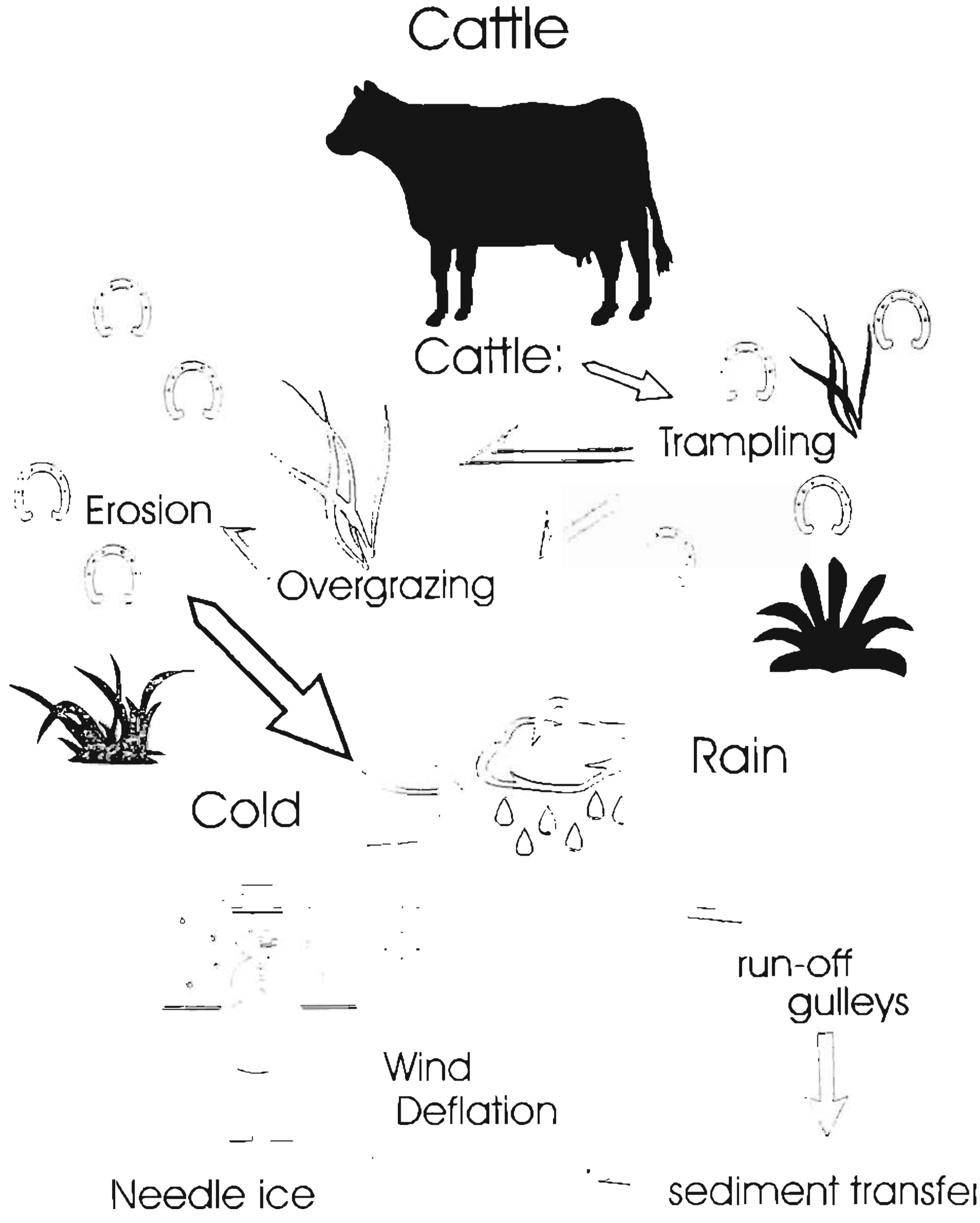
Zoogeomorphology:

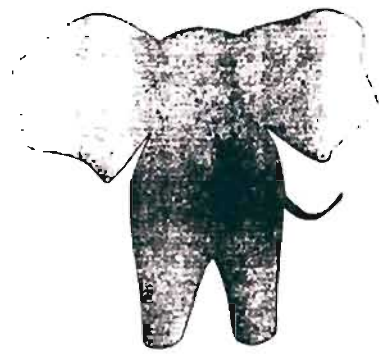
A Forgotten Factor in Development?

ANIMALS DO WHAT (*GEOMORPHOLOGICALLY!*)?

Compact Soil
↓
Decrease Infiltration
↓
Cause Localised Ponding
↓
Increase Overland Flow
↓
Cause Fluvial Erosion and/or Transport
↓
Cause Nutrient Transfer (in solution)
↓
Change Weathering Regime in Bedrock/Soil Below

Animals Actively Erode
↓
Remove/Kill Vegetation
↓
Expose Soil to OTHER Weathering/Transport Processes
↓
Change Microclimate = Weathering/Erosion Effects
↓
Facilitate Water Penetration to Greater Depths
↓
Remove Bonding Components (Selective Erosion-Salts)
↓
Cause a Decrease in Slope Shear Strength
↓
Load Already Unstable Slopes With Increased Mass (debris)
↓
Cause "Bioturbation"
↓
Thereby Changing Weathering/Pedogenesis
↓
Cause Localised Ponding/Water (or Sediment) Transport Paths


Cattle
Cattle:
Trampling
Erosion
Overgrazing
Rain
Cold
run-off
gulleys
Wind
Deflation
Needle ice
sediment transfer



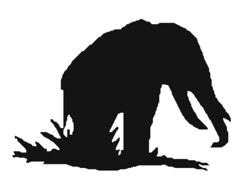
ELEPHANTS

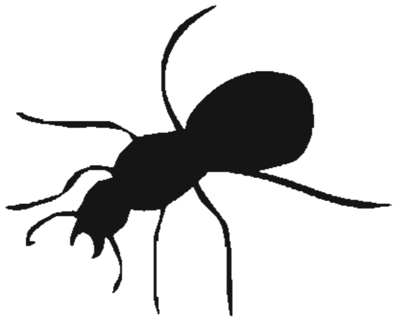
One elephant = 0.3 to 1.0 m^3 removed from a pan EACH time they wallow

Wallows in Kenya = o.45 to 0.59 km^2

THESE ARE THEN AFFECTED BY DEFLATION

Deflation is as a RESULT of preparation by elephants





ANTS

Ant mounds in Brazil: 340 m^3/mound

USA: 1500 mounds/ ha each 0.02 m^3
= 34 m^3 / ha = 1.7% of total area

in Wyoming = > 6 MILLION in Wind River Basin alone
with mean mound = 32l = 192000 m^3

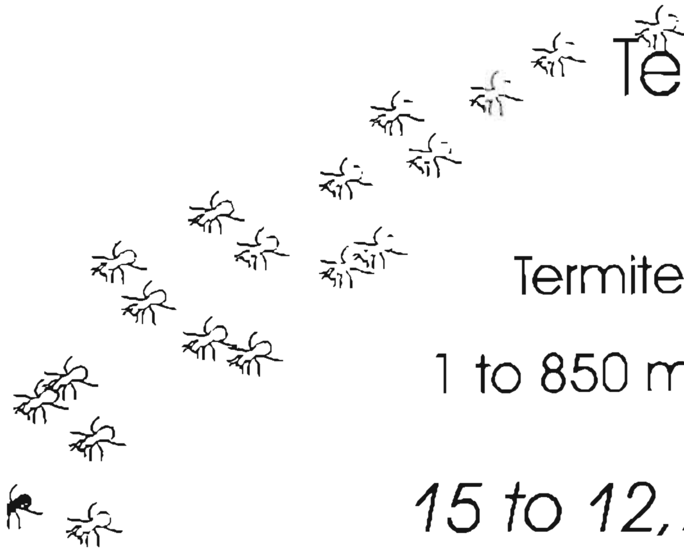
Baltic ants = 40 l/mound with 2500/ha
= 100 m^3 / ha

Branner (1909) cited denudation by ants in Brazil as
3,226,250 kg/ha/100yrs

Darwin suggested 2,598,500 kg/ha/100yrs

Australian studies (1996) showed ant eroding bedrock!

Termites



Termite mounds 15 m^3/mound

1 to 850 mounds per hectare

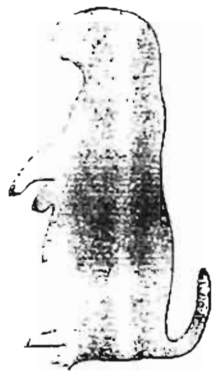
15 to 12,750 m^3/ha

Aardvarks

They erode c. 0.1 m^3/ha

from termiite nests

Wind, rain and frost then remove this material



MOLE RATS

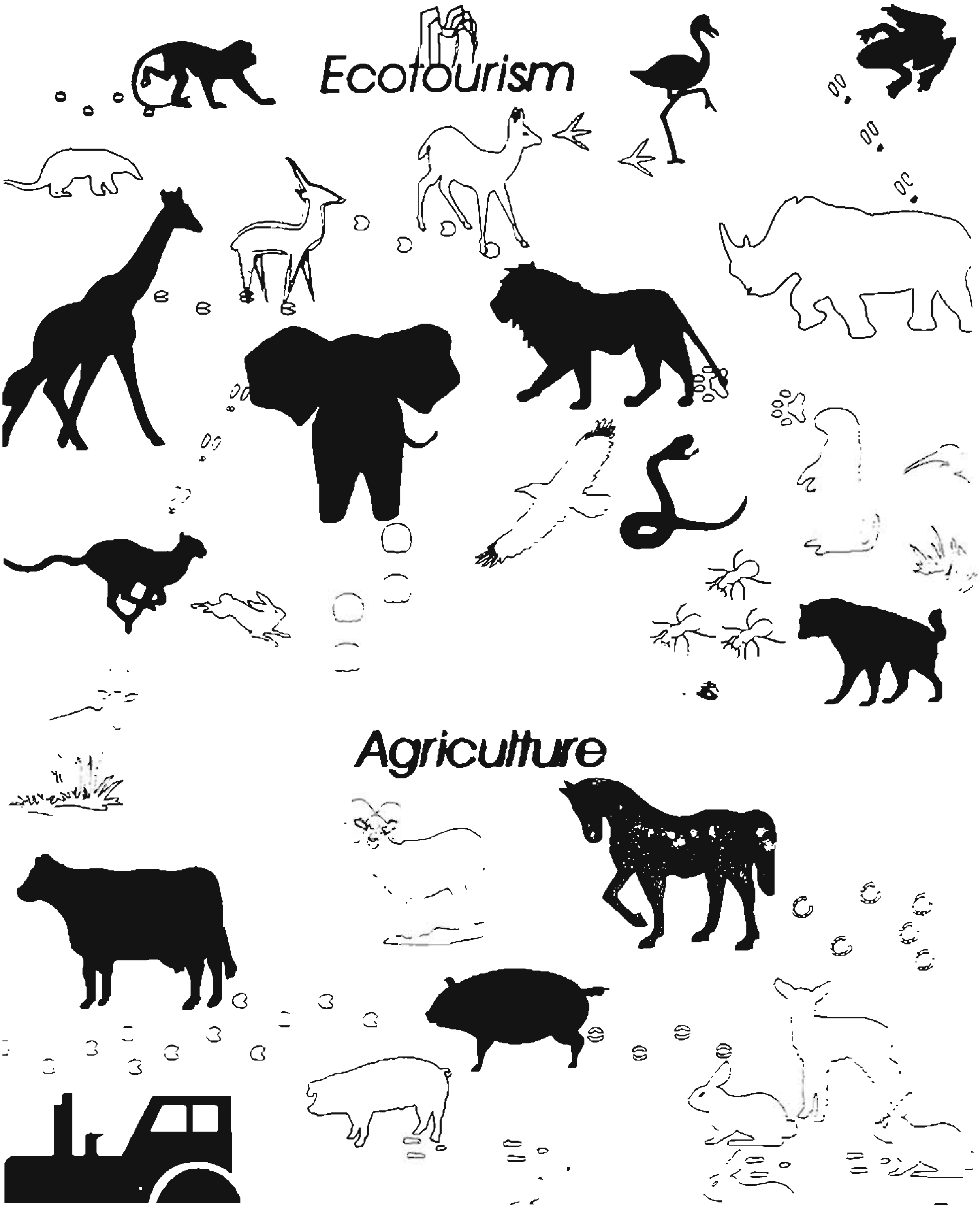
Volume of material = 0.00147 to 0.00728 m^3 / ha

2500 burrows per hectare

Changes nutrient cycling

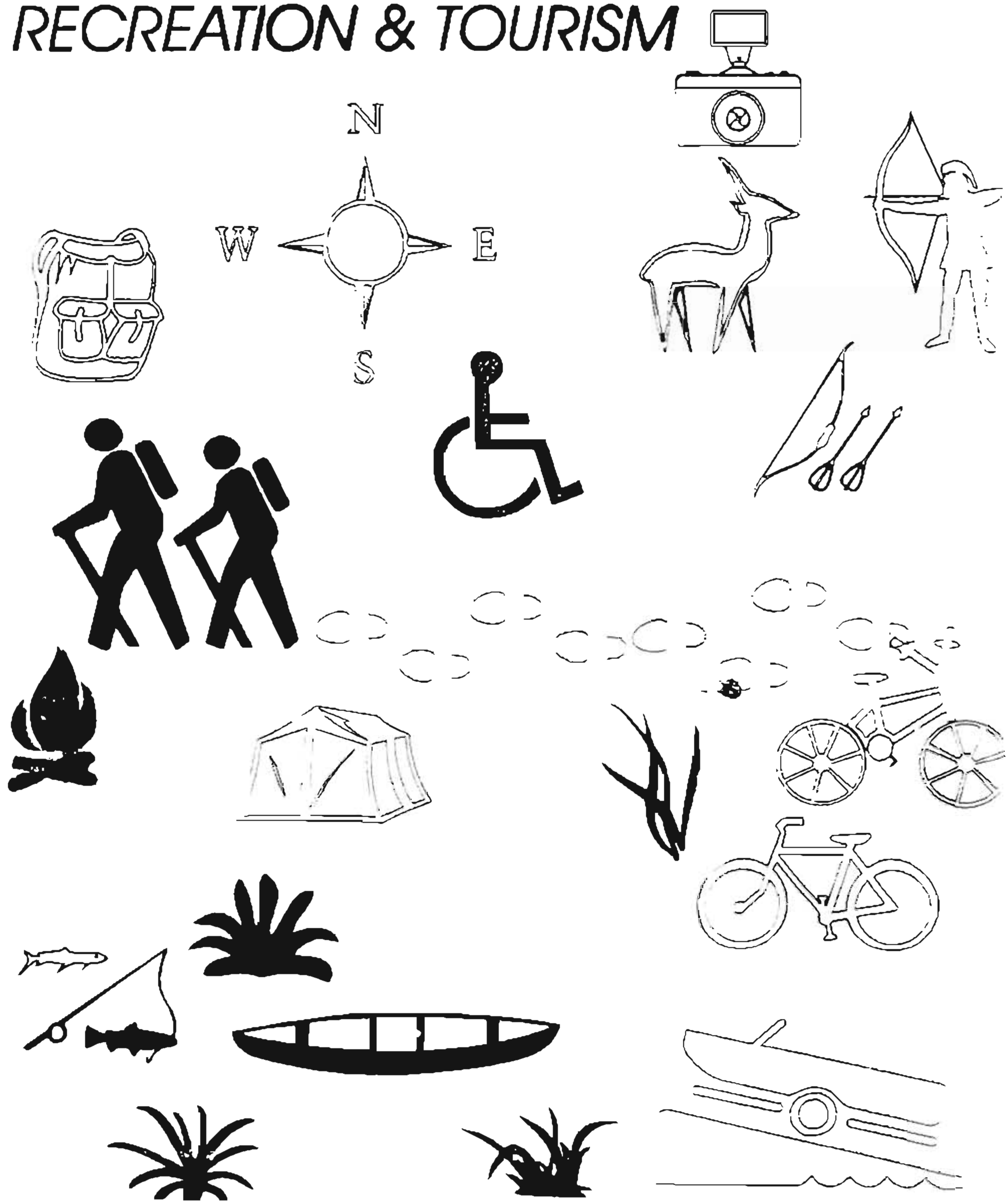
Changes drainage

Changes soil aeration


Ecotourism
Agriculture

ANTHROPOGENIC

RECREATION & TOURISM



ARCUS

Arctic Research Consortium of the United States

Council Notebook

Twelfth Annual Meeting

16 - 19 May 2000

Holiday Inn Capitol (at the Smithsonian)
Washington, DC

Kevin Hall
University of Northern British Columbia

600 University Avenue
Suite 1
Fairbanks, Alaska 99709
USA

Unzoned pyroxene and sodic plagioclase in the dacite indicate that it likely underwent significant crystallization at this depth; highly resorbed anorthitic plagioclase from the andesite indicates that it originated at greater depths and underwent relatively rapid ascent until it reached 3 km, mixed with dacite, and erupted. Diffusion profiles in phenocrysts suggest that mixing preceded eruption of earliest lava by approximately one month. The lack of any compositional gap in the erupted rock suite indicates that thorough mixing of the andesite and dacite occurred quickly, probably due to low density and viscosity differences. Disaggregation of enclaves, phenocryst transfer from one magma to another, and direct mixing of compositionally distinct melt phases were the three mechanisms by which hybridization was accomplished.

Living on the edge: archaeology and coastal dynamics along the Gulf of Alaska coast

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Alutiiq, Tlingit, and Dena'ina peoples and their ancestors have lived along the geologically dynamic coastline of the Gulf of Alaska for 10,000 years, simultaneously at the edge of the sea and on the margin of colliding tectonic plates. Earthquakes, tidal waves, volcanic eruptions, glacial advances, and sinking shorelines are common and sometimes catastrophic occurrences, remembered in Native oral history and evident in the archaeological record of the region. Interdisciplinary field studies conducted in five Gulf of Alaska parks by the Smithsonian Institution and National Park Service (1993–1996) indicate that local Holocene glacial, tectonic, and sea level histories must be reconstructed as a first step toward regional level interpretations of indigenous land use patterns, population movements, and paleodemography.

Animals as agents of landscape evolution in the Arctic: The unquantified element

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The biology of Arctic animals has long been a major research topic as too have been the role of animals within the Arctic ecosystem and within the lifestyle of indigenous peoples. Animals are also recognised to play a significant role within Arctic ecotourism. What has not received attention is the actual role of the animals within the landscape, that they play a geomorphic role (potentially) comparable to other Arctic landform agents such as glaciers, rivers, mass movement, etc. At one level this is surprising. Much of the Arctic is a region of fragile flora coupled with thermally sensitive permafrost, within which occur mobile, large herds of mammals coupled with less mobile small, burrowing mammals. Present day numbers may also be far smaller than in the recent past when other mega-fauna such as mastodon and wooly rhinoceros were abundant. Any disturbance, particularly to permafrost conditions, by the action of these animals (direct erosion, compaction, trampling, overgrazing, etc.) can lead to a whole range of geomorphic responses—from thermokarst to slope failure. The impact of the animals is further exacerbated by other geomorphic processes (*e.g.* needle ice, slope wash, aeolian

erosion, etc.). In many instances, it is the ability of these other geomorphic processes to now operate as a result of animal action that is a major factor in landscape development, and one that is overlooked in landscape evolution. An example of an Arctic landform (dells) being created by musk ox will be presented. The significance of Arctic zoogeomorphology will also be put in the context of the necessity for its understanding for sustainable development or maintenance of Arctic park areas, especially under potentially changeable climatic conditions.

CAPT Michael A. Healy: the man, his ships and the *Healy*

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To commemorate the launching of the U.S. Coast Guard Cutter *Healy*, Mr. Harper collected historical photographs from the Healy family, the U.S. Coast Guard, Georgetown University, the National Archives, builders of the cutter, newspapers and museums. In this slide presentation, Mr. Harper emphasizes one man's impact on Alaska history. His talk covers the history of the Healy family, the naval career of CAPT Healy, and the ships on which he served, including the famous Revenue Cutter *Bear*. Mr. Harper also will discuss the events surrounding the naming of the Cutter *Healy* and present photographs of the vessel's launching.

The Arctic upper atmosphere as a harbinger of global change and space weather

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The Arctic upper atmosphere is an extremely sensitive region that reacts measurably to small changes in chemistry, temperature, and solar inputs. Severe solar eruptions and the resulting solar wind can cause major changes to the Arctic upper atmosphere including greatly expanding the auroral oval, causing polar cap absorption, and the induction of large ionospheric electric currents. Ground-based instruments in the Arctic as well as from spacecraft instrumentation can measure these affects. The changes in ionospheric parameters resulting from the energy input carried by the solar wind is used by modelers to predict the effects of space weather on human activity. Also, the Arctic mesosphere (the coldest place on earth) is the location of polar mesospheric clouds, noctilucent clouds and metallic layers. Chemical changes and material transport, some of which results from human activity influence the existence of these clouds and layers.

Chapter 7

Periglacial Processes and Landforms

"How many legs does a dog have if you call the tail a leg? Four. Calling a tail a leg doesn't make it a leg."
Abraham Lincoln

As an adjunct to the weathering studies, work was also undertaken on a number of periglacial landforms and processes occurring in the same regions. As such, many of these features are frequently associated with some of the forms discussed in Chapter 5 (e.g. polygons and stripes on the treads of cryoplanation terraces). The papers are cited purely in a chronological order and discussion is presented after the list. Although not directly linked to weathering, in many instances the material associated with the landforms cited here may be (at least in part) the result of weathering while the processes associated with the development of the landform (e.g. the sorted stripes) are essentially those (e.g. freezing and thawing) that 'drive' the weathering.

Papers included within this section are:

- Hall, K.J. 1973b. A report on snow conditions in the Okstindsjøen area,

March 1972, *In* R.B.Parry & P.Worsley (eds): *Okstindan Research Project, Preliminary Report for 1972*, 51-54.

- Hall, K.J. 1978a. Ablation hollows on melting snow surfaces: a proposed classification and observations on formation. *Norsk Geografisk Tidsskrift*, 32, 143-146.
- Hall, K.J. 1979b. Sorted stripes orientated by wind action: some observations from sub-Antarctic Marion Island. *Earth Surface Processes*, 4, 281-289.
- Hall, K.J. 1981b. Observations on the stone-banked lobes of Marion Island. *South African Journal of Science*, 77, 129-131.
- Hall, K.J. 1983a. Sorted stripes on sub-Antarctic Kerguelen Island. *Earth Surface Processes and Landforms*, 8, 115-124.
- Hall, K.J. 1984a. Observations of some periglacial features and their palaeoenvironmental implications on sub-Antarctic islands Marion and Kerguelen. *South African Journal of Antarctic Research*, 13, 35-40.
- Hall, K. 1987c. Periglacial landforms and processes of the subantarctic and Antarctic islands. *Palaeogeography of Africa, Antarctica and the Surrounding Islands*, 18, 383-392.
- Hall, K. 1991b. The significance of periglacial geomorphology in southern Africa: A discussion. *South African Geographer*, 18, 134-137. (provided in Chapter 5, p.563-566)
- Hall, K. 1994b. Some observations regarding sorted stripes, Livingston Island, Antarctica. *Permafrost and Periglacial Processes*, 5, 119-126.
- Hall, K. 1999c. Present and Quaternary periglacial processes and landforms of the maritime and sub-Antarctic region: A review, *In* J. Lee-Thorp and H. Clift (eds.): *Book of Abstracts, XV International Congress, International Union for Quaternary Research*, 77.
- Hall, K., Boelhouwers, J., and Driscoll, K. 2001. Ploughing blocks in the

McGregor Mountains, Canadian Rockies. *Permafrost and Periglacial Processes*, 12, 219-225.

- Bockheim, J.G. and Hall, K. 1999. Permafrost, active layer dynamics and periglacial environments of continental Antarctica, *In* J. Lee-Thorp and H. Clift (eds): *Book of Abstracts, XV International Congress, International Union for Quaternary Research*, 26.
- Grab, S., Boelhouwers, J., Hall, K., Meiklejohn, I., and Sumner, P. 1999. *Quaternary Periglacial Phenomena in the Sani Pass Area, southern Africa.* International Union for Quaternary Research, XV International Conference Field Guide, 40pp.
- Van Everdingen, R.O. (Ed.) 1998. *Multi-language Glossary of Permafrost and Related Ground-Ice Terms.* University Printing services, University of Calgary, 162pp.

Other papers, cited elsewhere in the thesis, but particularly appropriate to this section include:

- Hall, K. Subm. a. Periglacial landforms and processes: Southern Alexander Island, Antarctica. *Permafrost and Periglacial Processes.*
- Hall, K. In Press* a. Present and Quaternary periglacial processes and landforms of the Maritime and sub-Antarctic: A Review. *South African Journal of Science.*
- Hall, K. 1997d. Did South Africa really look like this during the Quaternary? Questions regarding some recently suggested periglacial landforms and sediments. *Abstracts, Southern African Society for Quaternary Research, XIII Biennial Conference*, 5.
- Hall, K. and Meiklejohn, I. 1997. Some observations regarding protalus ramparts. *Permafrost and Periglacial Processes*, 8, 245-249.
- Boelhouwers, J., and Hall, K. 1990. The Sani Pass Area, *In* P.M. Hanvey

(ed.): *Field Guide to Geocryological Features in the Drakensberg.* I.G.CP.#297 Guide, University of Witwatersrand, Johannesburg, 57-70.

Hall (1973b) is purely a description of the snow conditions at one of the early study areas but it does relate to Hall (1978b) where a proposal is made regarding the formation and classification of ablation hollows on melting snow surfaces. Although ablation hollows leave no known form in the landscape once the snow has gone, they are a conspicuous feature on melting snow surfaces about which very little has been written. Hall (1979b, 1983a, 1984a, 1987c, and 1994b) all deal with sorted patterned ground, particularly sorted stripes, found in the sub-Antarctic and Maritime Antarctic. A number of interesting relationships between sorted stripes and the environment were recorded. On the islands of the extremely windy sub-Antarctic it was found that stripes were not only aligned parallel with the wind but that they also could occur, under these conditions, on horizontal surfaces. A relationship between coarse and fine stripe widths with altitude was found for Kerguelen (Hall, 1983a) that has also recently been found for Marion Island (Boelhouwers, pers.comm, 1999). Although most of these stripes were associated with needle ice activity, the role of ice lenses was also recognised, especially for Livingston Island (Hall, 1994b). On Livingston Island observations indicated the occurrence of, what appeared to be, ice lenses within the stripes that helped define the boundary between the coarse and fine stripes. This was an "opportunistic" observation as it occurred as the stripes were melting out from beneath the winter snow cover and no sedimentological evidence of the ice lense occurrence could be found further down slope where the lenses had melted.

Papers such as Hall (1981b) deal with stone-banked lobes found close to the sorted stripes on Marion Island, while the papers by Hall (1984a and 1987c) look at the range of features identified for the sub-Antarctic islands. The former of these two papers also attempts to explain the periglacial observations of current and fossil features by recourse to climatic change. Hall (1999c) gives a review of periglacial processes and landforms for the maritime and sub- Antarctic regions that is dealt with more fully in Hall (In Press a), cited in Chapter 5. Equally, Hall (Subm. a), also in Chapter 5, deals with periglacial landforms and processes found in dry valleys and on nunataks of Alexander Island; Alexander Island experiences a drier and colder continental climate compared to the Maritime or sub- Antarctic region. Thus, the range of periglacial landforms and processes found along the climatic gradient from the cool, wet, sub-Antarctic through to the cold, dry continental environment are presented.

Outside of the Antarctic, discussion regarding the role and significance of periglacial landforms in southern Africa is provided by Hall (1991b) and in Hall (1997d) that was presented in Chapter 5. More specific detail regarding some of the southern African features are given in the field guides for two conference field excursions: Grab, *et al.* (1999) and Boelhouwers and Hall (1990) (in Chapter 5). With respect to the Canadian Rockies, Hall *et al.* (2001), together with Hall and Meiklejohn (1997), provide information regarding periglacial landforms and processes observed there. The Canadian studies are a recent undertaking and are still in their infancy.

This Chapter, as noted in the Introduction, is one where many of the papers cited elsewhere in this thesis have components that are applicable to those presented here. The main additional papers have been given as a separate list (above) but many others (e.g. Hall, *et al.,* 1999 which deals with the role of animals and their interaction with periglacial processes in the Canadian Rockies) have attributes that would be suitable for citing here. It would get far to complex to try and cite them all so, rather, it is simply noted that they exist elsewhere within the thesis.

The final paper identified here (Van Everdingen, 1998) is *not* provided in the thesis. This 162 page document is a compilation by a variety of people (as listed) and I am but one of many who helped with definitions, etc. Thus it is seen as unfair and mis-leading to provide the whole document but, at the same time, that I was involved in this undertaking can be identified.

A REPORT ON SNOW CONDITIONS AT OKSTINDAN

by

Kevin J. Hall

1. INTRODUCTION

For some nine days at the end of March 1972, general observations were made in the region of the field station at Okstindan (Figure 1). During the week prior to the observations there were frequent blizzards and a heavy snowfall occurred. During the period of observation the weather varied from cold, clear, sunny days to conditions of extreme sub-zero temperatures and strong blizzards. The observations are of a purely qualitative nature and most were made on a random basis, and thus the ensuing statements and suggestions are not of a definitive nature. Additional observations are being attempted in 1973.

Snow conditions in the Okstindsjøen area were found to be varied both in time and space. A large number of areas were seen to be blown free of snow entirely or to have had all loose snow removed and only a layer of very hard ice left. These areas were on the most exposed points, i.e. ridge tops and the crests of mounds. The antithesis of this were the conditions found in the valley bottoms and the sheltered areas where there was up to a metre of snow overlying an ice layer. This snow was hard (probably due to the effects of the sun) and had been etched by the wind. There was a number of areas transitional between the above two. These were either of drifts of new snow which covered the etched snow forms or of a transitional form between the hard snow and the very hard ice. The latter form was gently rippled and had an icy texture with very little snow adhering to it but was softer than the ice proper.

Thus four main snow types can be distinguished:-

a) hard ice
b) crusty soft ice
c) hard snow
d) soft snow

2. THE DIFFERENT SNOW TYPES AND THEIR FORMS

2.1 Hard ice

The areas of this type were mainly on ridge tops, and especially on those with a southerly aspect, i.e. those faces receiving maximum insolation. The ice was clear but had many inclusions of air bubbles and snow particles. In places it was very slightly rippled and at these points, small patches of snow were sometimes found adhering to the windward side of the ripples. The ice, although very hard, tends to be very brittle. To the touch, the ice was not smooth and glassy but tended to feel 'sticky' and it was consequently very difficult to get the hand to slide over it. As this form was only found in the windy exposed places which faced the direction of maximum insolation it is very possible that it represented a form of regelation ice - almost a water ice - due to melt of the snow by the sun and the refreezing of this water by the sub-zero temperatures the cooling effect of the wind (the air being too cold to absorb much of the . by evaporation or to cause any of its own.)

Crusty soft ice.

Areas of this were of a whiter colour than the true ice areas. They ...ed minor ripple features the same as the true ice but here they tended to more pronounced. This type was not as hard as the ice proper, it was just ...ible to push the heel of a boot through the top surface to a depth of some ...m. Whether the area was in fact a true ice form or a very hard crusted ... was hard to determine by simple observation. It is possible that it ...resented some transitional stage as these areas usually merged into the ice ...as proper and were always found on exposed areas open to sun and wind.

Hard snow

Although these areas were not as extensive as the former two types the ...ms found on this type of snow were of the greater interest. The areas ...urred mainly in hollows or in the lee of ridges where they were reasonably ... protected from a general scouring away by the very strong winds. The ...w was quite hard and formed a well consolidated mass. It usually occurred ...low angle slopes and was considered to have been deposited for some time. ...s helps to account for its hardness and for some of the forms seen to have ...eloped on it. A foot placed on a flat surface would sink in only 30-40 mm.

The snow was hard and granular, the individual crystals being very small in ...e. Many of these grains were seen to be blown by the wind, this being most ...iceable on the ice areas where these blown grains did not adhere readily to ... surface. The snow grains, although hard, were seen in places to impact ...to other grains, but this was not universal and a proportion of grains ...ply rebounded and continued moving, although sometimes dislodging other ...ins in the process. It is possible that there were two grain types present, ... which was too hard to adhere but could dislodge other grains, and another ...ch was soft enough to stick to faces it was blown against; or it could ...e been simply chance which decided whether the snow adhered or not.

The edges of upstanding snow blocks were seen to have a 'stepped' windward ...e where the various harder bands were picked out by the wind (Figure 2). ...se faces were often found to have rounded ends towards the wind and, in some ...es, to have been undercut.

One interesting form seen was that of the growth of a micro-snow dune, ...ngated in alignment with the wind and originating from the footprint of some ...mal, probably a reindeer. It is supposed that the compaction caused by ... hoof makes the snow at that point much harder and thus more resistant to ...sion and ablation, and so in time it stood above the general snow surface ... become a point of origin for the accumulation of snow in its lee.

Another interesting form was seen, for which we might adopt the term ...bular snow dunes'. These dunes were elongated in the direction of the ...d, were rounded and hollow. Often they were found to be on areas of a ...tle gradient. It was thought that these were due to the undercutting of ...ound of snow on one side only (due to local flow lines around obstructions) ... the tendency for this, when the overhang became too large to support ...self, to curl over and join to the ground. Having joined the ground it ...s becomes stabilized and the wind would then blow through the tube thus

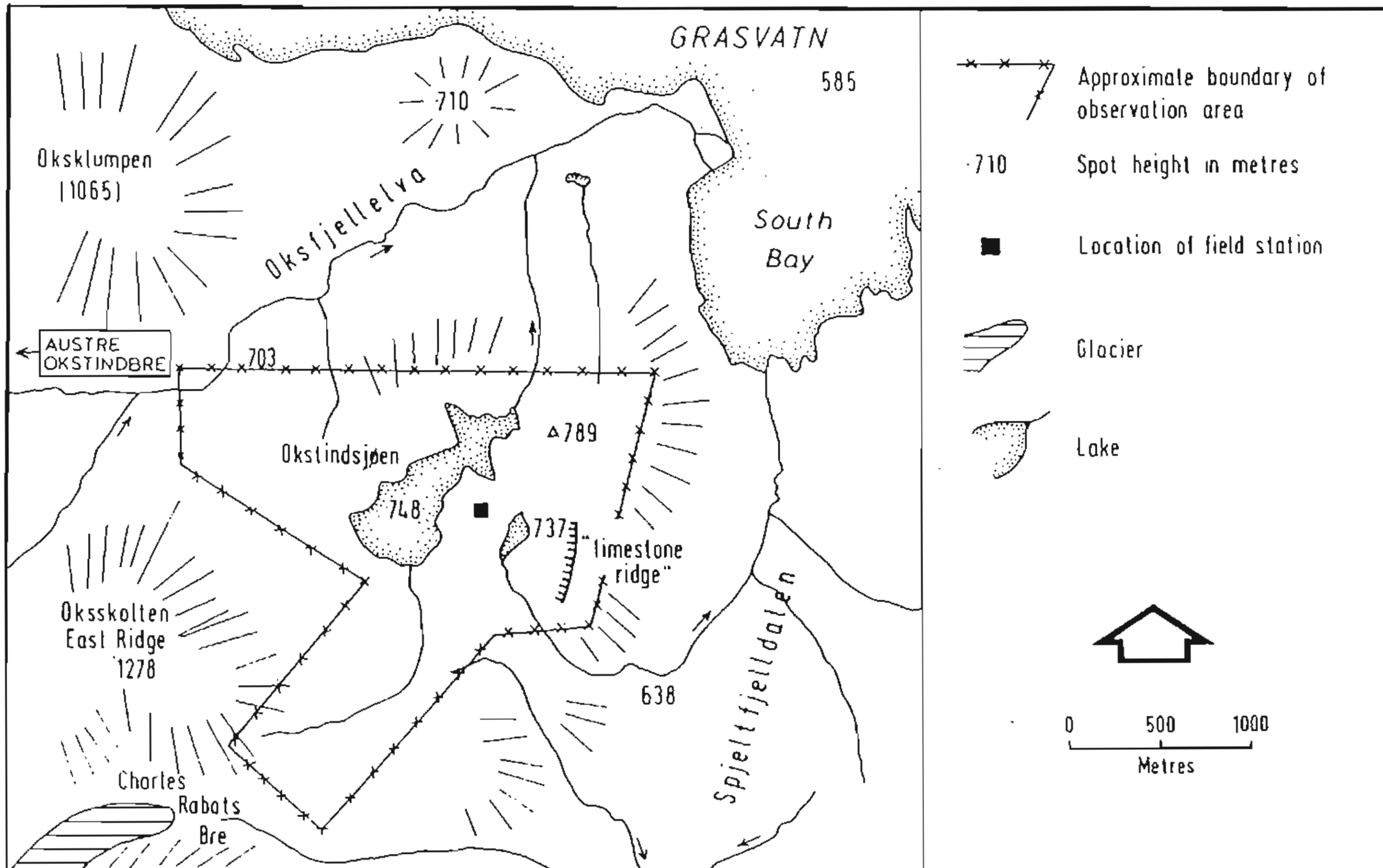


Figure 1. AREA OF SNOW OBSERVATIONS

created, (Figure 3). These were rare, as might be expected from their mode of formation, yet a number of forms were seen which suggested a transition point where the snow was curling over and yet others where this seemed to have occurred but the snow was in such condition that it broke away completely.

Snow was seen to drift into the lee of obstacles, usually harder snow mounds, and to produce a tail to them, (Figure 4). The flow lines of the snow laden wind could clearly be seen by the pattern of snow flakes it was carrying Where two rocks or mounds were close together the deflection caused to the wind by one could clearly be seen by the incutting effect this had on the snow "tail" of the other. This often caused complex forms (Figure 5). Where the obstruction was a rock, an area of water ice was usually seen on the windward side of the rock (Figure 4).

2.4 Soft snow

Here what was observed was that softer snow accreted around the older, hard snow forms, and hence the older forms were transformed such that the new snow covered the old forms to a point where they could not be recognised. Post-deposition saw the old forms re-appearing as the new snow was blown away, or new forms were produced on the new snow over the older ones. Which of the two alternatives occurred was dependent on the meteorological conditions at the time.

3. GENERAL OBSERVATIONS

It was found that the valley at the base of the limestone ridge immediately east of the field station was deeply filled with snow. The main wind direction was found to be from the west, i.e. from the direction of the major exit from the ice cap via Austre Okstindbre, whereas the valley lies at right angles to this. The dominant wind direction explained the presence of a cornice on the top of the low ridge opposite the limestone ridge. Within the valley itself it was found that the wind in fact blew down valley. This was explained by the deflection caused to the wind by the limestone ridge, which channelled it into a north-south direction along the valley. This phenomenon also explained the lack of snow actually on the limestone ridge itself and the large amounts on the slope facing it. The dune forms seen in the snow clearly showed this change in direction. On the flat area on the western side of the ridge the dunes were in the direction of the dominant wind, but with decreasing distance from the ridge the forms changed shape and direction until they showed a preferred orientation north-south down valley. Also to be seen was a snow ridge which ran out from the col in the ridge opposite the limestone ridge. This too was deflected down valley It was also thickest below the col, thinning out away from there. This was thought to be due to the wind coming over the top of the ridge, causing the cornice, and then being deflected down valley by the main flow concentrated in that direction.

Also seen was the clear juxtaposition between the snow-cleared windward face of the limestone ridge, which had only thin sheets of ice in places, and its snow-covered lee face. The etching of the snow-covered lee face of the valley was aligned in a down valley direction Much of this etching was of a 'scoop and ridge' form as was the case in most etched areas.

4. CONCLUSION

The foregoing observations indicate that the snow conditions at Okstindsjøen are both varied and interesting. The observations referred to here suggest that a number of features occur which it would be interesting to study in more depth and it is hoped that this report will be a basis for a more detailed analysis to be undertaken in the 1973 field season.

Figure 2

ETCHING OF SNOW BLOCKS BY WIND

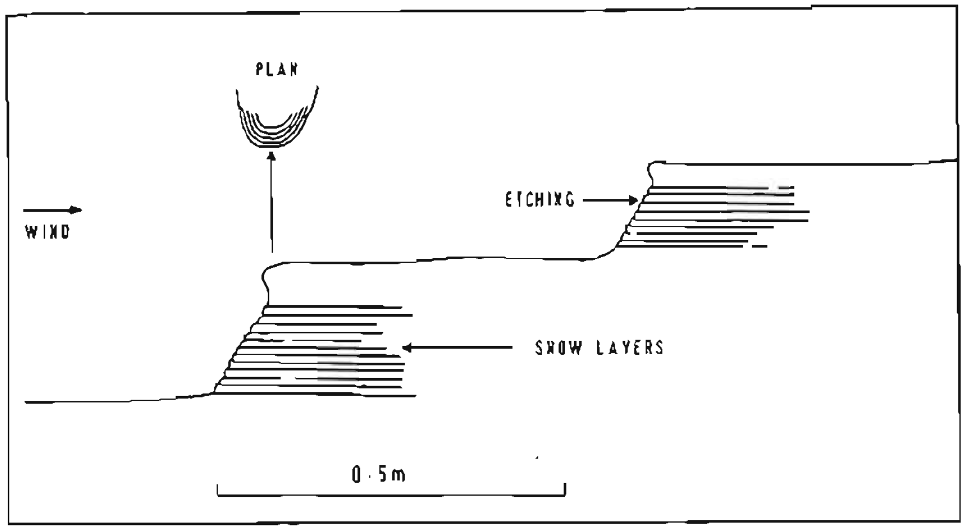


Figure 3

ORIGIN OF TUBULAR FORMS

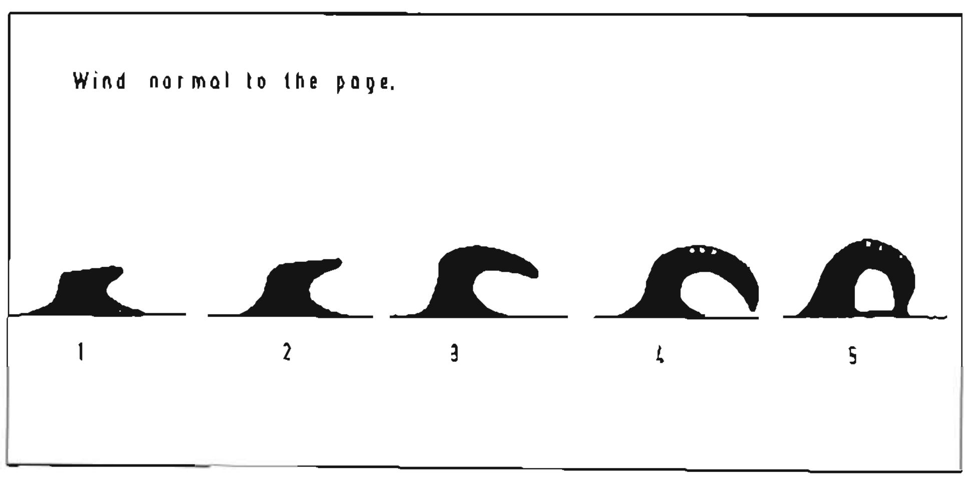


Figure 4

POSITION OF WATER ICE & SNOW TAIL

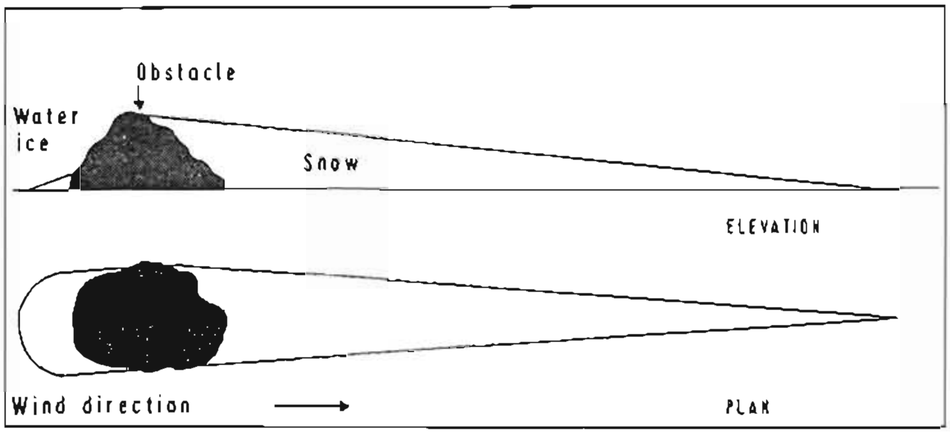
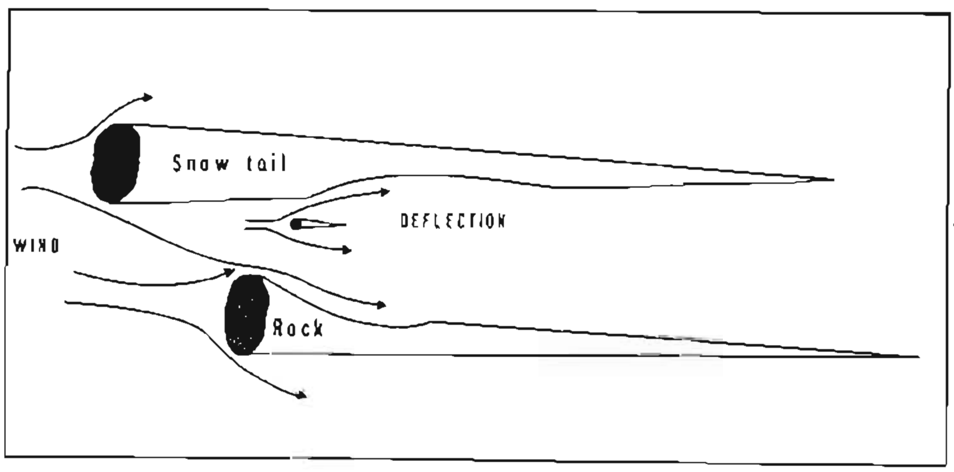


Figure 5

EFFECT OF WIND DEFLECTION ON SNOW TAILS



Notiser og litteratur – *Notes and Reviews*

Norsk geogr. Tidsskr., 32, 143–151

Ablation hollows on melting snow surfaces: A proposed classification and observations on formation

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Observations of ablation hollows on melting snow surfaces at several locations in northern Norway have suggested a tentative classification of these forms together with some indications of their formative processes. In the following note an asymmetrical ablation hollow is described and a brief suggestion as to its mode of formation is put forward.

Ablation hollows are dirt-free cuspate forms with intervening ridges, which may or may not be dirt-capped, that are found on many melting snow surfaces in the arctic or alpine locations of northern Norway. Despite their ubiquitous nature, they are a feature which has been little studied. Jahn & Klapa (1968) describe ablation hollows occurring on flat or sloping surfaces in Poland whilst Leighley (1948) notes hollows which occur on downward facing surfaces. The term 'ablation hollow' is preferred to that of 'sun cup' which is often found in the literature (Post & LaChappelle 1971), as the cuspate form is certainly due to ablation but need not be greatly related to the effects of the sun (as will be mentioned later).

Observations at many locations in northern Norway together with available descriptions from the literature suggest a tentative classification of these forms (Fig. 1). Two major classes are proposed: 'Open' and 'Closed'. The 'Open' class is defined as any location exposed to direct radiation as opposed to the 'Closed' which is not. The slope units are self-explanatory. The 'Angled' and 'Smoothed' divisions refer to the outline of the hollows as these were most noticeable in the field; it is, however, only a simple morpho-descriptive term. The dirt-free crest simply implies an intervening ridge composed of clean, dirt-free, snow. This ridge which intervenes between the cusps is often described as having a '... clearly marked capping of dirt along the crest ...' (Richardson 1954, p. 118); this is the 'Centre dirt' class of Fig. 1. This dirt capping, when present, is always described as occurring along the crest top, but numerous observations have been made where the dirt was found off-set from the actual ridge top by a short distance. In these instances the dirt is seen to parallel the ridge top but to be located some 10 to 30 mm behind the crest to the windward side. Thus a division to incorporate this situation is included: the 'Off-set dirt'.

Thus a simple classification is derived which encompasses all the observed parameters (the features being size transient) but including two forms so far not observed. The classification could be made more complex but as a first approximation it appears to work quite efficiently.

The location of the dirt, which is in fact mainly organic (Jahn & Klapa 1968), on the centre of the crest has been considered by Ball (1954). He suggests that the material, which is disseminated through the snowpack, becomes concentrated at the crest top due to the fact that '... surface grime tends to move along a path normal to the retreating snow surface and therefore follows a trajectory normal to the family of surfaces representing the successive position of the snow surface ...' (Ball 1954 p. 325, Fig. 1). Jahn & Klapa (1968) suggest that the organic material is 'held' to the snow surface by a film of water and thus it follows the normal trajectory rather than moving to the bottom of the cuspate form by gravity. If the surface is sloping then the dirt is off-centred but still associated with the crest top (Jahn & Klapa 1968). In fact it is not the general slope of the snowpack which is the control but that of the micro-elements within the cusps. The line of concentration is effectively along the

OPEN

Angled

	Dirt Free	Centre Dirt	Off-set Dirt
HORIZONTAL (0-5°)	x	x	x
SLOPING (6-80°)	x	x	x
VERTICAL (81-90°)	x	x	not observed
OVERHANGING (> 90°)	x	x	not observed
	Dirt Free	Centre Dirt	Off-set Dirt

Smoothed

OPEN

CLOSED

Angled

	Dirt Free	Centre Dirt	Off-set Dirt
HORIZONTAL (0-5°)	x	x	x
SLOPING (6-80°)	x	x	x
VERTICAL (81-90°)	x	x	not observed
OVERHANGING (> 90°)	x	x	not observed
	Dirt Free	Centre Dirt	Off-set Dirt

Smoothed

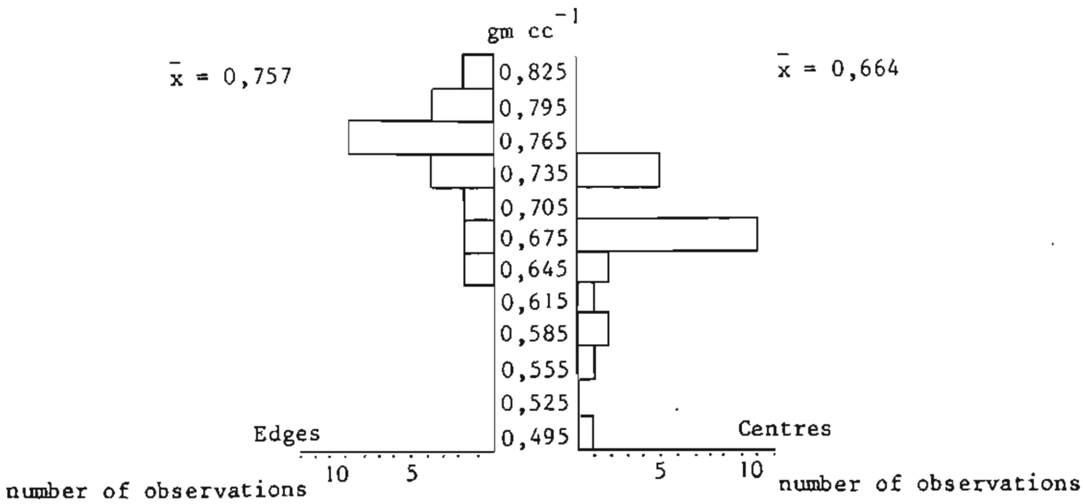
CLOSED

Fig. 1. A proposed classification of ablation hollows.

boundary between two micro-element surfaces inclined in opposite directions. Thus the concentration is symmetrical if the rate of ablation of the two elements is equal but asymmetrical if the ablation is uneven.

However, in the instances described above, no gap need occur between the dirt band and the crest. That the cases where the dirt is found to be off-set from the crest by a gap is not a function of extremity of slope, is shown by the fact that in nearly all instances observed the snowpatch had a general slope of 10° or less. Neither can it be a function of the wind blowing the dirt over the crest to deposit it in the lee, for the dirt band is always found on the windward slope.

A simple hypothesis is suggested to explain the formation of ablation hollows in which the dirt band is found off-set from the ridge crest by



(density values are those for the median value of each class)

Fig. 2. Density samples from 0.05 m depth from edges and centres of hollows.

a gap. It was noticeable that in all situations where this type of hollow occurred there was a strong unidirectional wind (usually southerly) and periods of intense radiation. It is suggested that the shifting of the dirt to the side of the crest is related to a lag effect in the movement of the particles along the normal trajectory. That is to say, due to the strong winds and high radiation input, there is a movement of the hollows and crests, parallel to the wind direction, as the surfaces melt downwards. Thus, the normal trajectory of the particles does not manage to keep pace with the lateral movement of the faces. Therefore, the material concentrates at the point where the melting crest would be under stable conditions but which is now the windward side of the crest slope some distance behind the crest proper. The lower down the crest slope the dirt occurs the greater is the implied lateral movement of the successive snow surfaces.

Chemical analysis of snow from the cusps and the top of the intervening ridges showed interesting results (Table 1). It was found that there occurred a distinctly greater concentration of the measured elements at the ridge top; calcium showing a distinct concentration with almost exactly double that of the centres. It is possible that this concentration results from the same processes that serve to concentrate the dirt particles.

Density samples at 0.05 and 0.25 m depth were obtained from the crest top and the cusp centre. Whilst no noticeable differences could be detected at 0.25 m depth, the 0.05 m level showed a marked dissimilarity (Fig. 2). The crests can be seen to have a higher density than the cusp centres in the 0.05 m samples. This higher density may be a cause of the greater concentration of solutes found at the crests; perhaps due to a melting and refreezing process.

Ablation hollows have not been observed to form during winter when much of the snow is being sculptured by the wind. Even during periods of high intensity radiation input no hollows were observed to form if the air temperature was below 0°C. However, as soon as the air temperature rose above 0°C the hollows began to appear and became most accentuated during warm, dry, windy conditions. Periods of rain were seen as destructive.

Exactly how these observations relate to hollows that occur on the walls and ceilings of sub-snow or sub-glacial tunnels, where there is no

Table 1. Chemical analysis of snow samples from hollow edges and centres (Okstindan, 1973).

HOLLOW	K	Mg	Na	Ca
Edge	0,75	0,03	1,77	0,57
Centre	0,90	0,05	2,24	0,28

radiation and little movement of air, is uncertain. It is suggested that in sub-snow environments other processes are likely to be operative but that they produce the same features.

There would appear to be much scope for the study of the many different types of ablation hollow. It is hoped that the brief descriptions presented here may suggest some lines of investigation which will greatly improve our knowledge of these features.

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The book aims at a presentation of how geography has developed since ancient times. The changes are seen from a philosophical and methodological perspective and it thereby disregards the pragmatical aspects of how geography functions in society. The content and main ideas of geography are followed until the interwar period. In his presentation of the post-war period, Holt-Jensen shifts the focus away from the content of geographical inquiry and leaves more room for methodological issues like quantitative techniques and systems analysis.

Geography is not the only subject dealt with in the book. In addition it contains a brief presentation of some thoughts related to positivism, criticism, Kuhn's idea of paradigms, and the role of values in social sciences research.

One of the merits of the book is its presentation of pre-modern geography. In Scandinavia, as elsewhere, an expansion of geography took place during the 1960s. Interest at leading departments of geography was concentrated on methodological issues. Geographical inquiry to a large extent dealt with processes; diffusion of various features and urbanization. Geographical form was presented in terms of urban structures, communication networks, and other subjects dealt with in systematic geography. The methodological development at departments matched well with planners' need and a pragmatical orientation was irresistable. For those geographers who are raised and matured in this tradition, and there are many of them, Holt-Jensen's book is an important correction and complement. He shows that geography had a life, sometimes a very good one, even before Walter Christaller, and that quantification and planning perspectives are of limited importance.

Another merit is that Holt-Jensen brings philosophical aspects into a geography textbook. Though elementary, they ought to be enough to arouse the reader's interest for such questions; as, for instance, the author's discussion of the influence of Charles Darwin on geographical inquiry and methods of research.

The formulations in the book also give room for some critical comments. The views of a number of geographers are presented in a neutral and correct way. However, a more critical discussion of some of these thoughts in the light of contemporary thinking could have been of use. In some sections it is difficult to distinguish Holt-Jensen's personal view from that of the respective geographers presented.

One important issue in geographical discussion has always been the regional and systematic perspective. Holt-Jensen presents a lot of elements from both perspectives but he gives no clear idea of how these two branches of geography could be defined. 'The traditional view of systematic geography has been that its objective is to formulate laws which could be empirically tested in those areas we study in regional geography' (p. 14, reviewer's translation).

It is probably correct to say that systematic geography has been more attuned to law formulation. But it is misleading to imply that regional geography is a subordinated counterpart in whose laboratory systematic geography gets its hypotheses tested. In his presentation of geographers representing the regional tradition, Holt-Jensen shows that regional geography is a very interesting branch in its own right. A discussion of regional geography today would, however, have been of interest.

Another important issue in geography of philosophical relevance is the distinction between absolute and relative room. It is implied in Holt-Jensen's book and elsewhere that regional

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SORTED STRIPES ORIENTATED BY WIND ACTION: SOME OBSERVATIONS FROM SUB-ANTARCTIC MARION ISLAND

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SUMMARY

Sorted stripes found on the volcanic scoria and glacial deposits of sub-Antarctic Marion Island indicate a distinct preferred orientation. Despite uniformity of slope and material, the stripes are predominantly aligned parallel to the wind. It is suggested that melting of needle ice by the early morning sun is of only limited importance in the sub-Antarctic owing to the almost continual overcast conditions. The effect of the wind is so great that in exposed situations stripes are formed on horizontal surfaces.

KEY WORDS Sorted stripes Wind orientation Needle ice Sub-Antarctic Periglacial

LOCATION OF STUDY AREA

Marion Island (Lat. 46°54′S, Long. 32°45′E) is located in the southern Indian Ocean approximately 2° of latitude north of the Antarctic Polar Front (Figure 1). The island consists of a roughly oval shield volcano of 290 km^2 rising to a height of 1230 m in the central mountain area. During the last 270,000 years the island has been extensively glaciated on several occasions (Hall, 1978) but today there remains only a very small area of permanent snow and ice above 950 m. The island is composed of basaltic lavas with many scoria cones and several areas of till resulting from the last (Würm-age) glaciation. Many areas of sorted stripes occur on both the scoria and the till.

CLIMATIC SETTING

Located in the midst of the 'Roaring Forties' the climate is one of strong westerly winds, high annual precipitation and low but equable temperatures (Schulze, 1971). Outcalt (1971) suggested that a minimum temperature of −2°C is required for the growth of needle ice and the island's meteorological records were analysed with respect to this value. It was found that at the Meteorological Station (24 m a.s.l.) an average of 48 days yr^{-1} occurred with temperatures of −2°C or lower and that the mean freeze amplitude of the 48 occurrences was −3·4°C (Table I). Extrapolation to an altitude of 500 m was undertaken using a decrease in temperature of 4°C 1000 m^{-1} (as obtained from radio-sonde soundings). This yields an average value of 111 days yr^{-1} with temperatures −2°C or lower and an overall mean freeze amplitude of −4·1°C for the 500 m level (Table I). Values are not extrapolated above this height as at higher elevations longevity of snowlay adds an important variable (i.e. a protective influence) and visual inspection suggested a high frequency of stripe occurrence between 350 and 500 m.

At the Meteorological Station winds blow most frequently from the northwest quadrant (60 per cent of occurrences) with an average velocity of 32 km hr^{-1}. Full gales (66 km hr^{-1}) are a monthly occurrence and can last for up to 17 hours and include gusts of 198 km hr^{-1} (Schulze, 1971). Annual precipitation is of the

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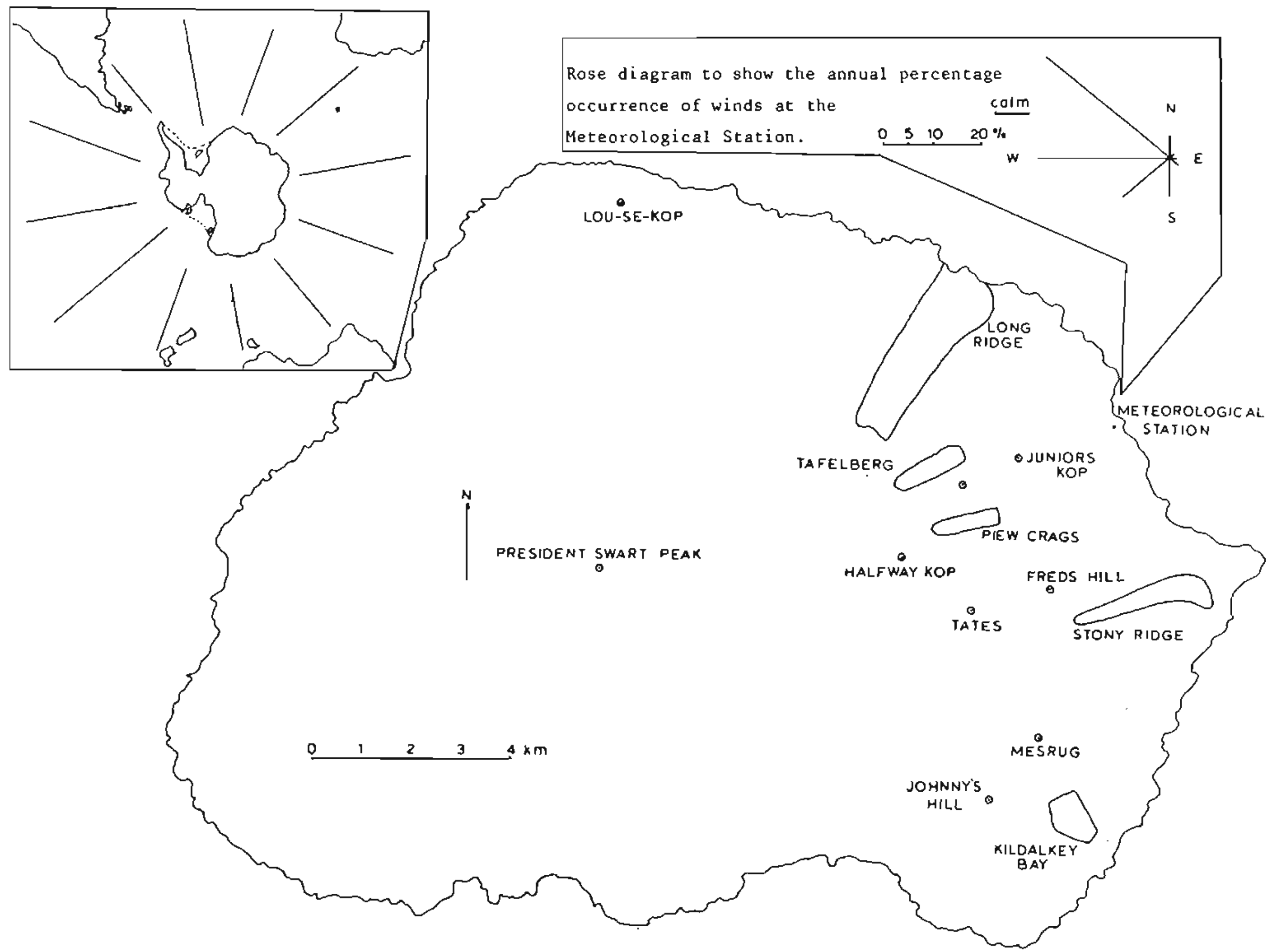

Figure 1. Marion Island and its location

Table I. Annual occurrence of temperatures of −2°C or less at the Meteorological Station (24 m a.s.l.) together with extrapolation to the 500 m level

Year	Meteorological Station			Extrapolation to 500 m	
	A	B	C	A	B
1969	79	−3·7	1·3	145	−4·3
1970	39	−3·2	1·2	108	−4·1
1971	34	−3·2	1·2	104	−4·1
1972	50	−3·6	1·0	101	−4·2
1973	38	−3·1	1·3	96	−4·0
$\bar{x}$	48	−3·4		111	−4·1

A: Number of occurrences of −2°C or lower temperatures.
B: Mean temperature of freezes indicated in A($\bar{x}$).
C: Sample standard deviation (s).

order of 2576 mm with, on average, 25 days in each month receiving some form of precipitation. Radiation inputs are low with average values varying from approximately 500 cal cm^2 day^{-1} in December to 82 cal cm^2 day^{-1} in June (Schulze, 1971). The high degree of cloudiness ($\bar{x}$ = 6·0 oktas) limits the amount of sunshine received to about 33 per cent of that possible in summer and to 20 to 25 per cent of the maximum possible winter value.

STRIPE DESCRIPTION

Sorted stripes are the most widespread periglacial forms encountered on Marion Island and were first described by Van Zinderen Bakker in 1975. According to Washburn (1956), sorted stripes are a form of patterned ground in which alternating lines of coarse and fine stones are orientated down the steepest available slope. On Marion Island these forms can be recognized from 70 m a.s.l. up to the highest point. In the lower areas growth is restricted to vegetation-free locations and at higher elevations to snow-free sites. The most extensive occurrence is between approximately 300 and 500 m.

The material of the coarse stripes consists of pebble (−5·3 ϕ to −2 ϕ) to fine gravel size (−1 ϕ) clasts of scoria or grey lava. The fine stripes consist of approximately 10 per cent clay, 35 to 60 per cent silt and the residue is made up of fine to coarse sand (−0·25 ϕ to +4 ϕ) with a surface covering of granule to gravel size material (−2 ϕ to −1 ϕ). Within the fine stripe, 'flames' of silt and clay penetrate the surface granule-gravel layer (Figure 2).

Measurement of stripes at a number of locations (Table II) shows that the fine stripes are marginally wider than the coarse, but not to the extent suggested by Brown and Kupsch (1974, p. 38). Correlation of coarse stripe and fine stripe widths shows that $r = +0·97$ ($r^2 = 0·94$) which is expressed by the linear regression equation $y = 0·974\,x + 0·009$. It is most noticeable that stripe width increases with altitude and it is suggested that this is a reflection of increase of amplitude and wavelength of free–thaw cycles with altitude which, in turn, gives rise to a deeper penetration of the freezing plane and thus more extensive sorting. Correlation of altitude with coarse stripe width indicates that $r = +0·89$ ($r^2 = 0·79$) and correlation of altitude with fine stripe width yields an r value of +0·95 ($r^2 = 0·90$). The increase of fine stripe width with altitude is expressed by the linear equation $y = (3·10 \times 10^{-4}\,x) + 4·60 \times 10^{-3}$. The coarse stripes are located within linear depressions 40 to 70 mm in depth (Figure 3). In some instances the surface of the fine stripe was seen to be raised 10 to 15 mm above the surface of the coarse material.

An experiment undertaken over a period of one year (1976–1977) showed that there was an accelerated downslope movement within the fine stripes with respect to the coarse. During the period of the experiment, on a scoria cone at an altitude of 304 m, there were 65 freezing events of −2°C or lower, with an overall mean freeze amplitude of −3·5°C. The net downslope movement, on the 18° slope, was 0·14 m in the coarse stripe material and 0·22 m in the fine stripes.

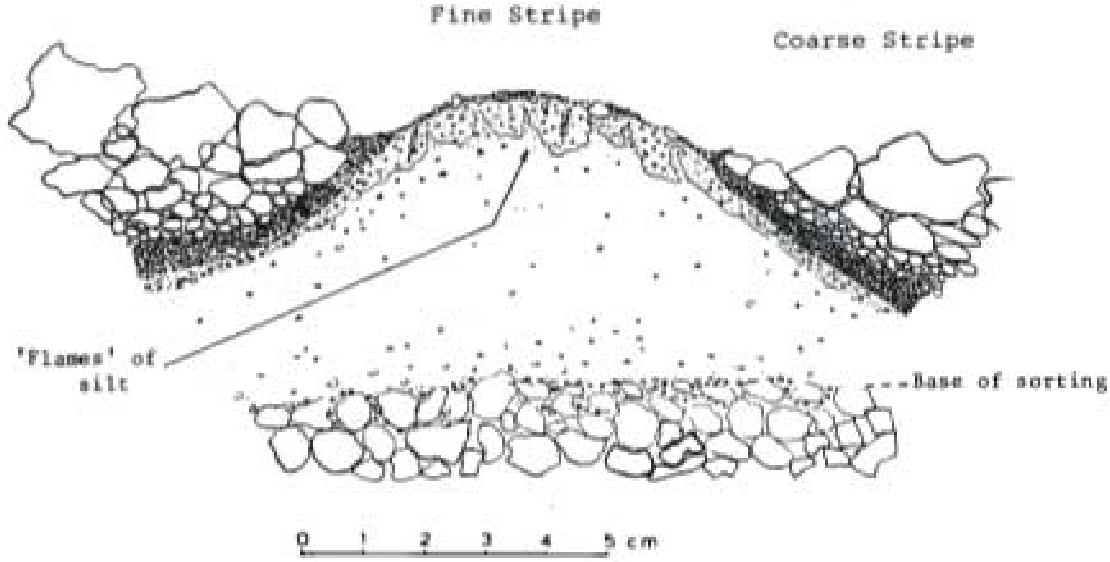

Figure 2. Simplified section to show the sorting of material within the stripes

Lenticular freezing of the ground was rarely observed and then only to a maximum depth of 0·15 m. The brevity and small amplitude of the frequent ground freezing episodes often produces needle ice which, on some occasions, survives the following day and shows a second stage of growth the succeeding night. The maximum needle ice length measured was 51 mm. The needles were more firmly attached to the clasts that they lifted than to the ground beneath. On the other hand, raised fine-grained material was observed to be dry and easily blown away. This attachment of coarse grains and desiccation of fines may, in some manner, be accountable for the sorting.

Table II. Details of sorted stripes from twelve locations on Marion Island

Location	Altitude (m)	Material	$\bar{x}$ coarse width (m)	$\bar{x}$ fine width (m)	slope range	orientation	lengths
Juniors Kop	250	Scoria	0·12	0·11	14–25°	289°	4–30 m
Nr. Freds Hill	350	Glacial	0·05	0·08	4– 6°	330°	5-25 m
Piew Crags	350	Glacial	0·06	0·06	3–14°	311°	1·2– 5·0 m
Nr. Johnnys Hill	300	Glacial	0·09	0·08	4– 8°	280°	2·6– 3·4 m
Kildalkey Bay	150	Glacial	0·11	0·09	3–12°	290°	3–20 m
Nr. Mesrug	300	Scoria	0·04	0·10	4– 6°	290°	3– 8 m
Stony Ridge	300	Glacial	0·11	0·10	5–20°	330°	1·5– 7 m
Tafelberg	350–400	Glacial	0·09	0·10	5–19°	270°	1·4– 8·5 m
Hendrik Fister Kop	300	Scoria	0·10	0·11	12–15°	300°	5– 8 m
Halfway Kop	630	Scoria	0·25	0·23	10–13°	310°	3– 5 m
President Swart Peak	1200	Scoria	0·35	0·38	16–25°	270°	6–10 m
Lou-se-kop	200	Scoria	0.056	0·055	11–23°	278°	1·2– 4·4 m

$a = 295°$ $\bar{x} - 296°$ $H = 10{\cdot}62$ $l = 0{\cdot}94$ $S_A = 6{\cdot}9°$ $s = 20{\cdot}9°$ $R = 11{\cdot}29$
where:
a = angle of mean vector.
S_A = mean angular deviation.
l = length of mean vector.
H = Rayleigh test statistic.
$R = \sqrt{(\Sigma \cos a)^2 + (\Sigma \sin a)^2}$.

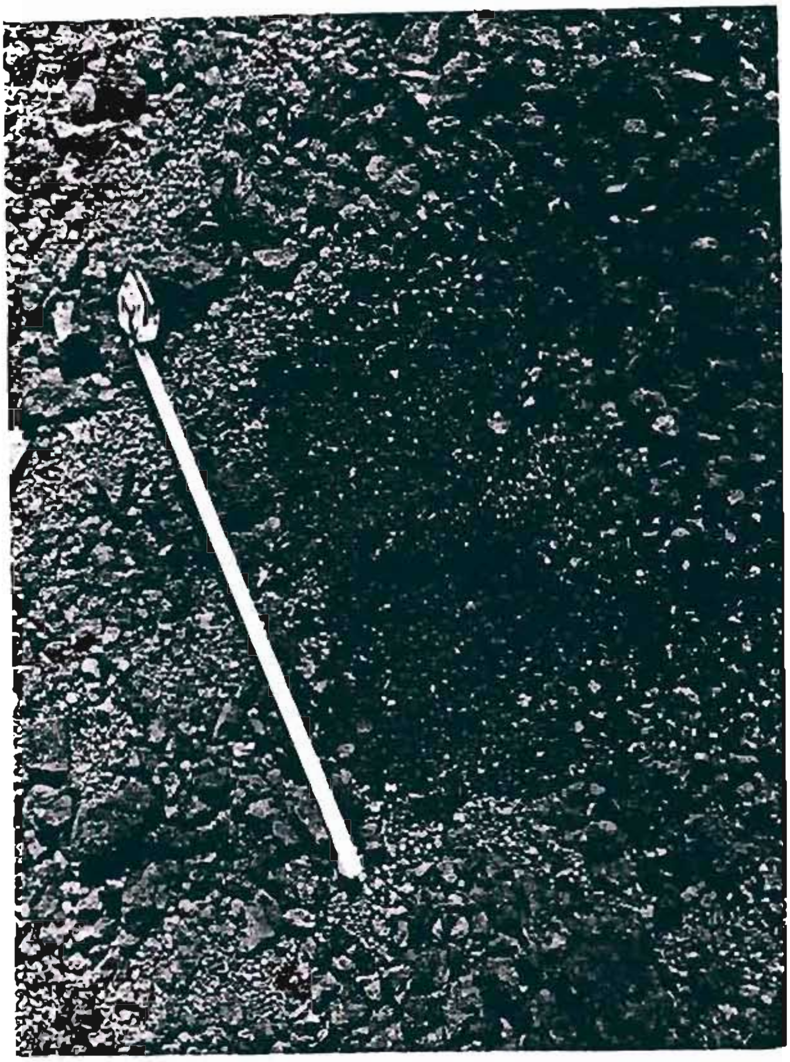

Figure 3. Excavated section across some sorted stripes to show the depression in which the coarse debris is located (tape = 0·5 m)

STRIPE ORIENTATION

The sorted stripes show a preferred orientation parallel to the dominant wind (Table II). The winds in this area are predominantly westerly but are often deflected to northwesterly by island obstructions (e.g. as found at the Meteorological Station). Slopes are steep and material uniform regardless of aspect on the volcanic scoria cones, yet the stripes are found predominantly in the west to northwest sector of the cone. For example, at Lou-se-kop (201 m) located at the northern end of the island and open to the full strength of the westerlies, 80 per cent of the stripes have a westerly orientation with the residual 20 per cent aligned northwest (Figure 4, Table III); no stripes are known on any other sector of the cone. On Junior's Kop (304 m), where the experiment on downslope rate of movement was undertaken, stripe orientation is limited to the range 285° to 330°, that is to say parallel to the northwesterly winds that predominate in this area.

On the glacial debris of Tafelberg, stripe orientation reflects the high percentage of westerly winds whilst at Piew Crags, on similar material, the stripes are aligned parallel to the deflected northwesterlies (Figure 5,

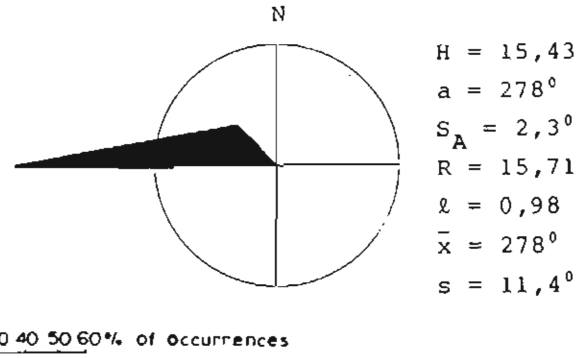

Figure 4. Orientation of sorted stripes on the volcanic scoria cone Lou-se-kop (201 m)

Table IV). At Kildalkey Bay, located at the extreme southeasterly corner of the island, the stripes show a predominant northwesterly alignment but with a small percentage orientated to the south, probably reflecting the incidence of cold southerly winds experienced in this part of the island (Figure 6).

The Rayleigh test (Norcliffe, 1977), otherwise known as von Mises distribution (Till, 1974), indicates that the stripes show a distinct orientational preference. As shown in Figures 4, 5 and 6, the areas of glacial debris and the volcanic cone studied in detail show stripe orientations between 276° and 318° with very low mean angular deviations (S_A). The computed resultant vector for all the observations shown in Table I is located directly within the west-northwest sector. It is interesting to note that the standard deviation (s) calculated for each set of data is misleadingly large compared to the true mean angular deviation (S_A). In all instances H far exceeds the critical value (given in Tables) and so H_O is rejected at the 0·01 level. This implies a strong orientational preference which accords, in this instance, with the predominant wind.

The role of the wind is such that sorted stripes are found on horizontal surfaces that are exposed to the full effect of the wind (e.g. Long Ridge). These stripes are similar in all respects to those occurring elsewhere except that they are found on slopes of less than 1°.

Table III. Details of sorted stripes found on Lou-se-kop (201 m)

Direction	Slope	Length (m)	$\bar{x}$ coarse width (mm)	$\bar{x}$ fine width (mm)
292°	18°	3·2	50	51
287°	14°	4·1	58	55
295°	17°	4·4	59	53
285°	14°	3·7	58	57
284°	20°	3·1	63	56
294°	16°	1·6	52	54
278°	23°	2·2	50	55
265°	19°	1·9	50	53
266°	24°	2·8	62	69
268°	15°	1·5	48	46
283°	17°	3·4	56	57
264°	12°	1·8	67	55
275°	19°	1·4	56	51
284°	11°	2·6	62	60
262°	16°	1·2	45	45
268°	20°	2·4	56	57
$\bar{x}$ 278° s 11·4	17° 3·6	2·58 1·00	56	55

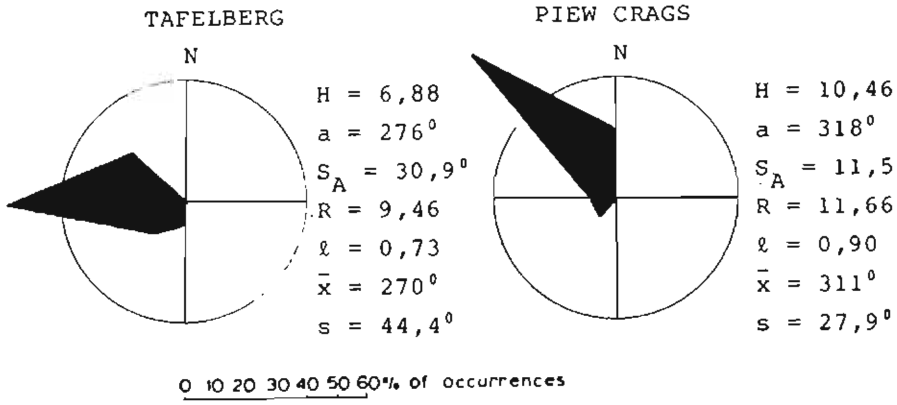

Figure 5. Orientation of sorted stripes on glacial debris at Tafelberg and Piew Crags

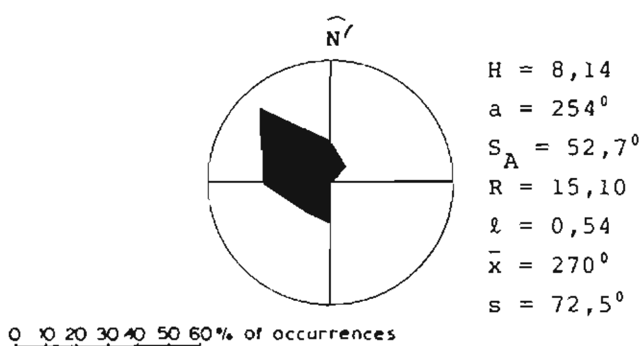

Figure 6. Orientation of sorted stripes on glacial debris at Kildalkey Bay

Table IV. Details of sorted stripes found on Piew Crags (350 m) and Tafelberg (400 m)

Piew Crags							Tafelberg						
Orientation	Slope	Length (m)	Coarse width (cm)	*s*	Fine width (cm)	*s*	Orientation	Slope	Length (m)	Coarse width (cm)	*s*	Fine width (cm)	*s*
343°	9°	2·5	7·8	0·5	7·7	1·0	210°	5°	2·2	6·7	0·6	5·5	1·1
321°	12°	1·8	6·8	0·7	6·8	0·8	226°	7°	3·5	5·2	0·4	5·0	1·5
338°	14°	2·1	7·1	0·6	7·5	0·5	174°	7°	1·8	5·0	0·3	5·3	0·8
306°	8°	1·6	5·5	0·5	5·6	0·7	297°	17°	2·25	8·3	2·5	8·9	2·0
318°	11°	2·2	6·2	0·9	6·0	0·9	278°	16°	1·95	6·3	1·2	5·9	0·2
314°	12°	1·2	5·5	0·5	5·6	0·7	288°	14°	1·48	8·5	1·5	7·4	0·5
298°	14°	2·2	5·6	0·4	6·1	0·6	273°	16°	3·0	6·9	0·8	4·6	0·5
336°	10°	1·9	4·8	0·3	4·7	0·7	268°	18°	5·8	7·6	0·6	5·8	0·5
286°	9°	3·1	4·8	0·4	5·0	0·8	274°	18°	8·5	8·1	1·6	7·3	1·5
294°	11°	2·4	5·5	0·7	5·7	0·7	282°	16°	2·4	6·2	0·7	6·1	0·4
245°	12°	2·9	5·2	0·7	5·2	0·7	278°	19°	2·9	6·6	0·5	6·7	0·9
346°	4°	4·1	6·6	0·8	6·6	0·7	330°	5°	2·0	9·2	1·8	9·8	2·9
300°	5°	5·0	11·5	2·8	9·6	1·4	333°	9°	2·4	10·3	1·7	9·3	1·1

DISCUSSION

Troll (1958) suggested, from observations on Mount Kenya and in southern Africa, that stripes are aligned with the dominant wind. Schubert (1972), in a similar Southern Hemisphere situation, shows that in the Andes needle ice forms parallel to the prevailing wind such that the needles are inclined downwind, lines of debris forming as they topple. In the Northern Hemisphere, Beaty (1974) observed that stripes are aligned parallel to a nightly mini-katabatic wind which flows downslope from an area of snow and Derbyshire (pers. comm.) describes what may be a similar situation on Skaftafell, Iceland. On the other hand the work of Mackay and Mathews (1974) has suggested that stripes are aligned parallel to the late morning sun and are a function of shadow effects during the thaw of the needle ice. They argue that strong wind is inimical to the needle ice growth. On the basis of work in New Zealand they suggest that stripes in the Southern Hemisphere should be preferentially aligned to the northeast.

It is interesting to note that on Iles Kerguelen, in a similar sub-Antarctic situation to Marion Island, Nougier (1970, plate XLIII–153, plate XLIV–156 and 157, and Figure 109) illustrates sorted stripes on horizontal surfaces which appear to be orientated parallel to the wind (plate XLIV–157). Nougier (pers. comm.) also notes an instance where sorted stripes parallel to the wind cut across stripes orientated down a slope. The stripes of Iles Kerguelen are very similar in size to those of Marion Island with widths of 0·08 m to 0·10 m (Nougier, 1970, p. 391).

While it is certainly true that needle ice growth requires cold, calm nights it may be possible that, in some manner, alignment is effected by the wind subsequent to growth. In partial support of this argument it is quite noticeable, especially in winter, that the nights are cloud-free and calm and that in the early morning the sky clouds over and the wind velocities rise. The lack of a northeasterly alignment of stripes is considered to be a function of overcast conditions, the high incidence of daytime cloud cover ($\bar{x} = 6$ oktas) inhibiting the ablation effects of the early morning sun.

Although the precise nature of the relationship between alignment of sorted stripes and dominant wind direction is still uncertain, the coincidence of the two on Marion Island has been established. With respect to sorting, Bellair (1969) argued that the dominant process in the formation of stripes on Kerguelen is the effect of wetting and drying of clay. While this process is widespread, especially in summer, observations on Marion Island suggest that frost action plays a major role.

In summary it may be said that the role of directional melting by insolation, whilst undoubtedly important in some environments is, at best, of minor significance in the sub-polar Southern Hemisphere owing to the extensive cloud cover. The effect of the wind is so great in this zone that sorted stripes are produced on horizontal surfaces but the exact process by which the wind orientates the stripes and effects the sorting is still uncertain. A valuable insight into the respective roles of wind and insolation in the orientation of sorted stripes might be obtained from the consideration of the occurrence, or non-occurrence, of stripes on the slopes of nunataks around a broadly circular ice cap. This would offer a range of slopes affected by either katabatic winds or the early morning sun which would allow direct comparison of the two processes within the same geographical–climatic region. However, it is concluded that the wind orientation hypothesis of Troll (1958) is viable under the specific climatic conditions of strong unidirectional winds and overcast skies.

ACKNOWLEDGEMENTS

This work constitutes part of a programme of multidisciplinary studies undertaken on Marion Island by the Institute for Environmental Sciences, University of the O.F.S. All logistics were provided by the Department of Transport. Dr. E. Derbyshire reviewed and kindly improved the original manuscript both in text and content. He also provided interesting observations of his own pertinent to this paper. Professor van Zinderen Bakker offered constructional criticism and D. Langley helped in obtaining field data.

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Observations on the Stone-banked Lobes of Marion Island

I describe here the stone-banked lobes occurring on the glacial debris of sub-Antarctic Marion Island, through the altitudinal range 70 m to 450 m above sea level (a.s.l.). Correlation coefficients expressing the possible relationships between various lobe parameters have been derived, which suggest that there is an increase in riser height with increase in lobe length and, at the same time, lobe slope increases with increase in field slope angle. From their relationship to past and present climatic conditions, it is concluded that the stone-banked lobes are not a product of the present conditions and that today they are largely inactive. It is tentatively suggested that lobe formation took place during the latter stages of the last glacial and that some were formed immediately upon deglaciation, around 11 000 B.P.

Marion Island (46°S, 37°E) is a shield volcano of some 290 km² (ref. 1), rising to a height of 1 230 m in the central mountain area, and located in the southern Indian Ocean (Fig. 1). It is composed of two basaltic lava suites: a massive, sometimes columnar jointed, pre-glacial grey lava and a scoriaceous, blocky, post-glacial black lava. Associated with the black lava phase are approximately 130 scoria cones dotted about the island. In addition, particularly on the eastern side, there are areas of glacial deposits resulting from extensive ice cover during the last (Würm-age) glacial.[2,3]

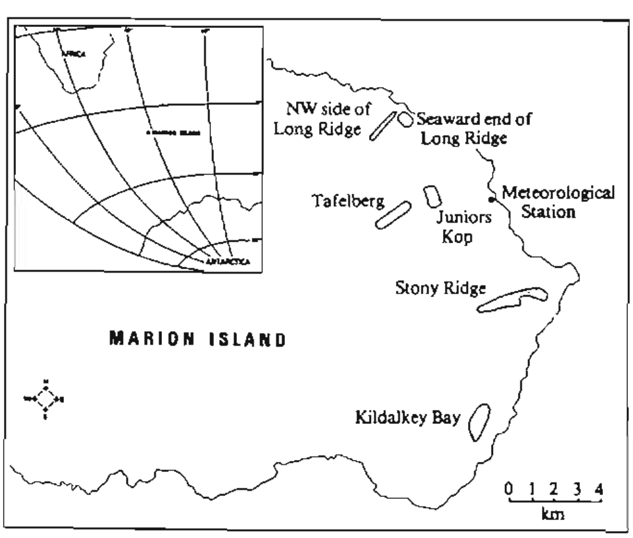

Fig. 1. Marion Island, showing localities mentioned in the text.

Situated within the 'Roaring Forties', the island experiences strong (average speed 32 km/h) northwesterly winds (60% of occurrences) with full gales at least once a month and gusts of up to 198 km/h (ref. 4). Annual precipitation is of the order of 2 580 mm with, on average, 25 days in each month receiving some form of precipitation. Snow can fall at any time throughout the year. Associated with the almost continuous precipitation is an extensive cloud cover (average 6 oktas), which limits the amount of sunshine received to about 33% of that possible in summer and from 20 to 25% of the maximum possible winter value. Thus radiation inputs are low and vary from approximately 500 cal cm² d⁻¹ in December to 82 cal cm² d⁻¹ in June.[4]

The average daily maximum temperature (at 24 m.a.s.l.) is 10.5°C during summer and 6.0°C in winter. The minima are 5.0°C and 1.0°C respectively with extremes for summer and winter (1961 – 1965) being 22.3°C and −6.8°C.[4] The mean monthly temperature range of 4.9 ± 0.5°C indicates the equable climate of the island, which results from the strong marine influence. Meterological Station (24 m a.s.l.) records for grass minimum temperatures (Table 1) indicate an average of 134 days/yr (s.d. =

Table 1. Frequency of temperatures of 0°C or lower as shown by the grass minimum thermometer at the meteorological station.

	1969			1970			1971		
	n	$\bar{x}$(°C)	s.d.	n	$\bar{x}$(°C)	s.d.	n	$\bar{x}$(°C)	s.d.
Jan.	2	−0.5	0.07	2	−0.3	0.07	2	−1.7	0.92
Feb.	1	−1.5	–	2	−0.8	0.35	1	−0.1	–
March	2	−0.5	0.07	3	−0.7	0.44	1	−1.4	–
April	10	−1.1	0.90	9	−0.7	0.45	10	−0.7	0.71
May	15	−1.1	1.19	11	−1.4	0.95	14	−1.4	1.05
June	16	−2.6	1.54	9	−1.6	0.99	19	−1.4	1.03
July	17	−1.9	1.30	14	−2.8	1.75	19	−1.9	1.29
Aug.	18	−3.4	2.22	18	−1.9	1.56	16	−2.4	1.58
Sept.	22	−3.4	2.01	19	−2.3	1.56	18	−1.4	1.30
Oct.	26	−2.2	1.66	16	−1.3	0.92	14	−1.0	1.08
Nov.	17	−2.3	1.43	11	−0.6	0.78	9	−0.8	0.78
Dec.	13	−1.8	1.38	3	−1.0	0.50	3	−0.7	0.75
Total	159			117			126		

22) with temperatures of 0°C or lower. As freezes are of short duration and massive freezing of the ground is rare, it is the action of pipkrake that accounts for much of the present-day periglacial activity.[5] Outcalt[6] suggested that a minimum temperature of −2°C was required for pipkrake growth. This condition is met on an average of 48 days/yr at the Meterological Station (average freeze amplitude −3.4°C), which corresponds to 111 days/yr at an altitude of 500 m (average freeze amplitude −4.1°C) (Table 1).

Embleton and King[7] define stone-banked lobes as comprising 'gelifluction and other deposits confined by crescent-shaped stoney embankments . . .' These lobes have been described by various workers as varying from 8 m (ref. 8) to 30 m (ref. 9) in length with risers from 1 m to 5 m in height.[8,10] Characteristically, lobes are said to occur on slopes of 10° to 25°, whereas the treads of the lobes slope at angles as low as 2° to 3°.[7] Stone-banked lobes are a feature of gelifluction and thus are related to the movement of water-saturated, fine-grained sediment above a frozen layer.[11] During gelifluction, sorting and orientation of the clasts takes place such that the larger clasts become concentrated at the surface and with their *a*-axis orientated parallel to the tread towards the centre of the lobe and normal to the slope at the lobe terminus. Continued movement of the larger clasts towards the lobe periphery, particularly the snout, results in a stone-rich margin with a relatively stone-free tread composed primarily of a sandy silt (Fig. 2).

A 'typical' stone-banked lobe on Marion Island (Fig. 2) has an unvegetated tread, which shows signs of sorting and may even exhibit sorted stripes or polygons on its surface, which is underlain by relatively fine material (sandy silt). The lobes are banked by clasts which vary in (maximum) size from 0.1 m to approximately 1.3 m. The clasts are usually very angular and platy, and always exhibit sorting such that the largest clasts are at the lobe edges; clast size decreases into the lobe until a sandy silt predominates.

Stone-banked lobes are found to occur on areas of glacial debris (largely supraglacial debris resulting from the last glacial[2]) within the altitudinal range 70 to 450 m a.s.l. Lobes do occur above 450 m but are scarcer owing to the lack of glacial debris in the central mountain region and the extensive disturbance caused by post-glacial volcanism. Many of the lobes are found banked against a cushion of *Azorella*. On the basis of observations so far made, a 'typical' lobe on Marion Island (average of 6 locations: Table 2) occurs on a slope of 13°, has a tread angle of 3° to 4°, a tread length of 0.93 m and a riser height of 0.32 m.

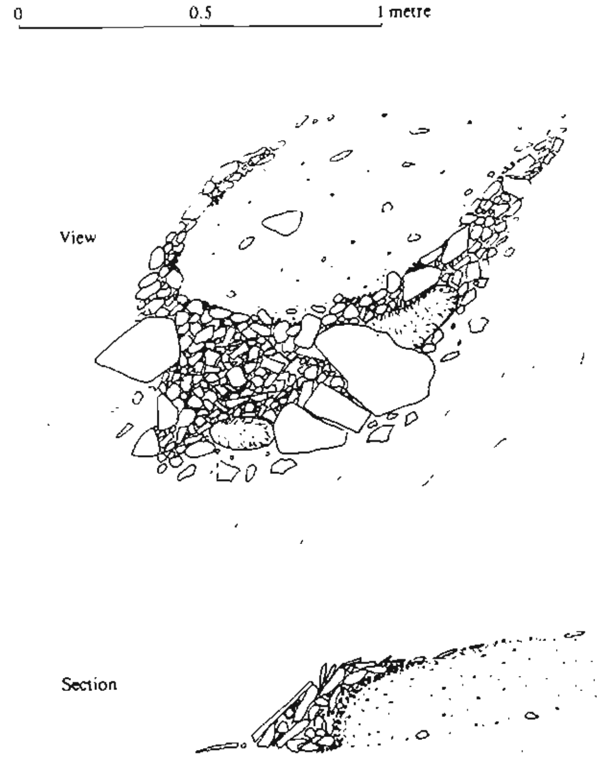

Fig. 2. View and section of a typical stone-banked lobe.

Utilizing the raw data from which Table 2 was derived it is pos ble to correlate the various parameters measured to see if a genetic relationships exist (Table 3). Despite the relatively low cc relation values obtained, a *t*-test of *r* indicated a number of cor binations significant at the 0.05 level. From Table 4 it would appe that there is a positive relationship between the angle of the tre slope and the angle of the field slope, and between the length of tl lobe and the height of the riser. Conversely, there is no apparei relationship between tread length and tread slope, tread length an field slope, and riser height and tread slope. There is an uncertai degree of association between field slope and riser height. It is ir teresting to note that correlation of the mean slope angles wit

Table 2. Measurements of stone-banked lobes at six localities.

Location	Average slope angle (°)	s.d.	Average tread angle (°)	s.d.	Average lobe length (m)	s.d.	Average riser height (m)	s.d.
NW side Long Ridge	23.5	1.6	7.6	3.8	0.79	0.23	0.41	0.12
Seaward end Long Ridge	12.5	4.7	4.1	2.3	0.87	0.23	0.28	0.10
Stony Ridge	10.6	2.7	2.6	1.4	1.01	0.37	0.39	0.11
Kildalkey Bay			2.4	2.1	1.08	0.45		
Tafelberg	12.1	3.9	3.1	1.9	0.76	0.34	0.29	0.16
Nr. Juniors Kop	8.2	3.3	2.5	2.2	1.04	0.37	0.25	0.08
Average	13.4		3.7		0.93		0.32	

Table 3. Correlation values for stone-banked lobe measurements.

Location	A	B	C	D	E	F	*n*
NW side Long Ridge	0.19	0.37	0.27	0.25	0.14	−0.06	57
Seaward end Long Ridge	0.38	0.66	0.55	−0.07	−0.36	0.11	56
Stony Ridge	0.44	0.32	0.20	0.17	−0.12	0.08	89
Kildalkey Bay				−0.05			41
Tafelberg	0.53	0.47	0.25	−0.16	−0.06	0.22	82
Nr. Juniors Kop	0.15	0.43	0.52	−0.01	−0.04	0.34	50

A, Riser height with field slope; B, riser height with lobe length; C, tread slope with field slope; D, tread length with tread slope; E, tread length with field slope; F, riser height with tread slope; *n*, number of pairs of observations.

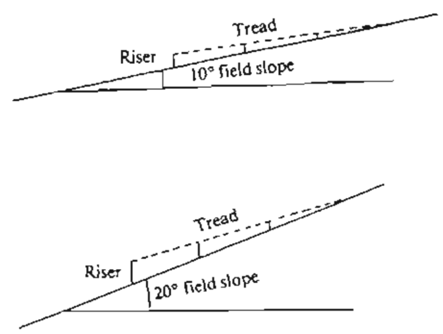

Fig. 3. Illustration of how riser height increases as a function of tread length and field slope.

mean tread angles for the five sites (Table 2) gives a very strong, positive relationship: $r = +0.98$ ($r^2 = 0.96$; $t = 8.53$).

This implies that there is an increase in riser height with increase in length of the lobe: short lobes are low-fronted, long lobes are high-fronted. At the same time tread slope angle increases with an increase in field slope angle. This can be summarized as follows: as field slope increases so does tread slope, but not at the same rate, and this necessitates that riser height should increase for each additional unit of lobe length (Fig. 3).

Discussion

The last glacial is believed to have ended, on Marion Island, around 11 000 B.P. (ref. 12), when the glaciers retreated both rapidly and extensively.[2] As mentioned earlier, the lobes are found on areas of glacial debris and not on the post-glacial scoria or black lava. The absence of lobes on the volcanic debris is thought to be partly a result of inadequate fines to promote ice segregation during the period when lobes were active on the fines-rich glacial debris and partly a result of deposition of the volcanic debris after the lobe-forming period. That stone-banked lobes can be found inland sugests that at least some of them were formed during post-glacial times, for the inland areas would have been the last to lose their ice cover. Those, particularly on the lateral moraines, near the coast may well have been active during the final stadials when that section of the moraine was beyond the glacier terminus. Thus, whilst lobe formation may have taken place over an extensive period of time in the lowland areas, the existence of those on the mountains suggests they were not initiated prior to about 11 000 B.P.

Present-day observations of the stone-banked lobes suggest that they are no longer active. There is extensive growth of vegetation, particularly *Azorella* cushions, over and amongst the clasts of the lobe fronts and many clasts exhibit extensive moss or lichen growth which shows no signs of damage or disturbance due to movement. Other than fine, wind-blown particles the *Azorella* cushions appear to show no signs of covering or encroachment by the larger clasts of the lobe front; rather, stabilization of the lobe has taken place.

Lenticular freezing has not been observed to a depth greater than 0.1 m (ref. 5), and sections through the lobes indicate sorting has taken place to a depth of at least 0.75 m at some time. The absence of present-day lenticular freezing is seen as a result of the frequent freezes, of small amplitude and wavelength, which, under the current cool, wet conditions, result primarily in pipkrake growth.[5] No evidence for temperatures conducive to deep freezing of the ground, between altitudes of 70 m and 450 m, is apparent from meteorological records.

Freezing of the ground to a minimum depth of 0.75 m is not only required to explain the observed depth of sorting but also as a motivating factor for gelifluction. Washburn[11] notes that a frozen layer beneath the thawing surface is required to provide the moisture conditions conducive to gelifluction. Moisture is a prime requisite for gelifluction but, although present-day precipitation is high, there is not an impervious layer to act in a similar manner to the frozen substratum and thereby cause saturation of the ground. In addition, Washburn[11] notes that in wet areas moisture will offset the binding action of vegetation and so promote movement. The lack of evidence for movement of lobes banked against *Azorella* cushions suggests that the factor that initiated movement (i.e. the frozen subsurface) is no longer present. Thus, the stone-banked lobes must have formed prior to the onset of the present climatic regime.

The absence of lobes on the recent (11 000 to 4 000 BP) black lavas or volcanic scoria is thought to be a result of two factors. First, these areas are not generally as rich in fines as the regions of glacial debris and so frost action cannot be so effective.[11] Secondly, as the volcanics are post-glacial in origin it is possible that they formed after the period of lobe production.

It is therefore suggested that the stone-banked lobes observed on Marion Island are largely a product of a former climate when frost action was more severe. At present there is little activity except, perhaps, on the unvegetated tread where there is some growth of pipkrake. The presence of gelifluction features is an indication of a colder climate, but not necessarily permafrost conditions, prior to the establishment of the present climatic regime.

This work forms part of a study of the periglacial and glacial history of Marion Island initiated by Professor van Zinderen Bakker. All logistics were supplied by the S.A. Department of Transport.

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Table 4. Correlation values, shown in Table 3, accepted as significant at the 0.05 level.

Location	A	B	C	D	E	F
NW side Long Ridge	–	x	x	–	–	–
Seaward end Long Ridge	x	x	x	–	x	–
Stony Ridge	x	x	–	–	–	–
Kildalkey Bay		No data			No data	
Tafelberg	x	x	x	–	–	x
Nr. Juniors Kop	–	x	x	–	–	x

x = accepted at 0.05 level of significance by t-test of r; – = not accepted at 0.05 level. Codes A – F as for Table 3.

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SORTED STRIPES ON SUB-ANTARCTIC KERGUELEN ISLAND

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ABSTRACT

Observations of sorted stripes at nine different localities on sub-Antarctic Kerguelen Island are presented. The study sites range from 25–600 m above sea level and are of varying aspect. It is found that there is a strong relationship between stripe orientation and the wind direction, to the extent that at one locality stripes were orientated across a slope. Stripe width was seen to increase with altitude and, in addition, the coarse stripe was dominant at lower elevations and the fine stripe at higher elevations. At one site a secondary set of stripes was observed to occur within the fine stripe of a primary larger set. On a trachyte plug there was evidence of lateral squeezing in the formation of the stripes. A number of minor observations pertaining to the possible mechanism of stripe formation are given.

KEY WORDS Sorted stripes Effect of wind Needle ice sub-Antarctic Periglacial

INTRODUCTION

Kerguelen Island (Lat. 49° 21′S, Long. 70° 12′E), with an area of approximately 6000 km^2, is the main island within the Kerguelen archipelago (Figure 1). About one tenth of the island is ice covered, with the Cook ice cap (500 km^2 and rising to an altitude of 1100 m) being the largest of the ice masses (Mercer, 1967). The firn line varies between 600 and 900 m a.s.l. and all glaciers are composed of temperate ice (Mercer, 1967). The island is composed primarily of plateau basalts but with a number of plutonic igneous bodies interspersed; the oldest date obtained is in the order of 33 MY (Nougier, personal communication). Extensive faulting and exploitation of the faults by glaciers has resulted in a deeply indented coastline, numerous valleys, lakes and ridges, with the only non-mountainous terrain being that of the till and outwash plain to the east: Peninsule Courbet (Carte de Reconnaissance, 1968).

Situated just to the north of the Antarctic Polar Front (Mercer, 1967) the island experiences a typically oceanic climate with only a small temperature range about the annual mean of 4·6°C (Weyant, 1967). Lying within the belt of mid-latitude westerlies the island is subject to strong winds predominantly within the southwest to northwest sector, with very few calm days (Weyant, 1967). There is an extensive cloud cover, 60 per cent of all observations indicating six oktas or more, and a high incidence of precipitation (approximately 324 days yr^{-1}) (Weyant, 1967). At ground level Aubert de la Rue (1959) monitored 200 frosts during a period when only 120 were recorded in the air. Troll (1958), analysing the data of Meindarus (1923), suggested ground frost on 238 days yr^{-1} with no freezing at 0·05 m but later (Troll, 1960) obtained freeze–thaw cycles at the soil surface on 236 days yr^{-1} with a maximum penetration of 2 in (0·05 m).

SORTED STRIPES

According to Washburn (1956, p. 836) 'Sorted stripes are patterned ground with a striped pattern and a sorted appearance due to parallel lines of stones and intervening strips of finer material orientated down the steepest

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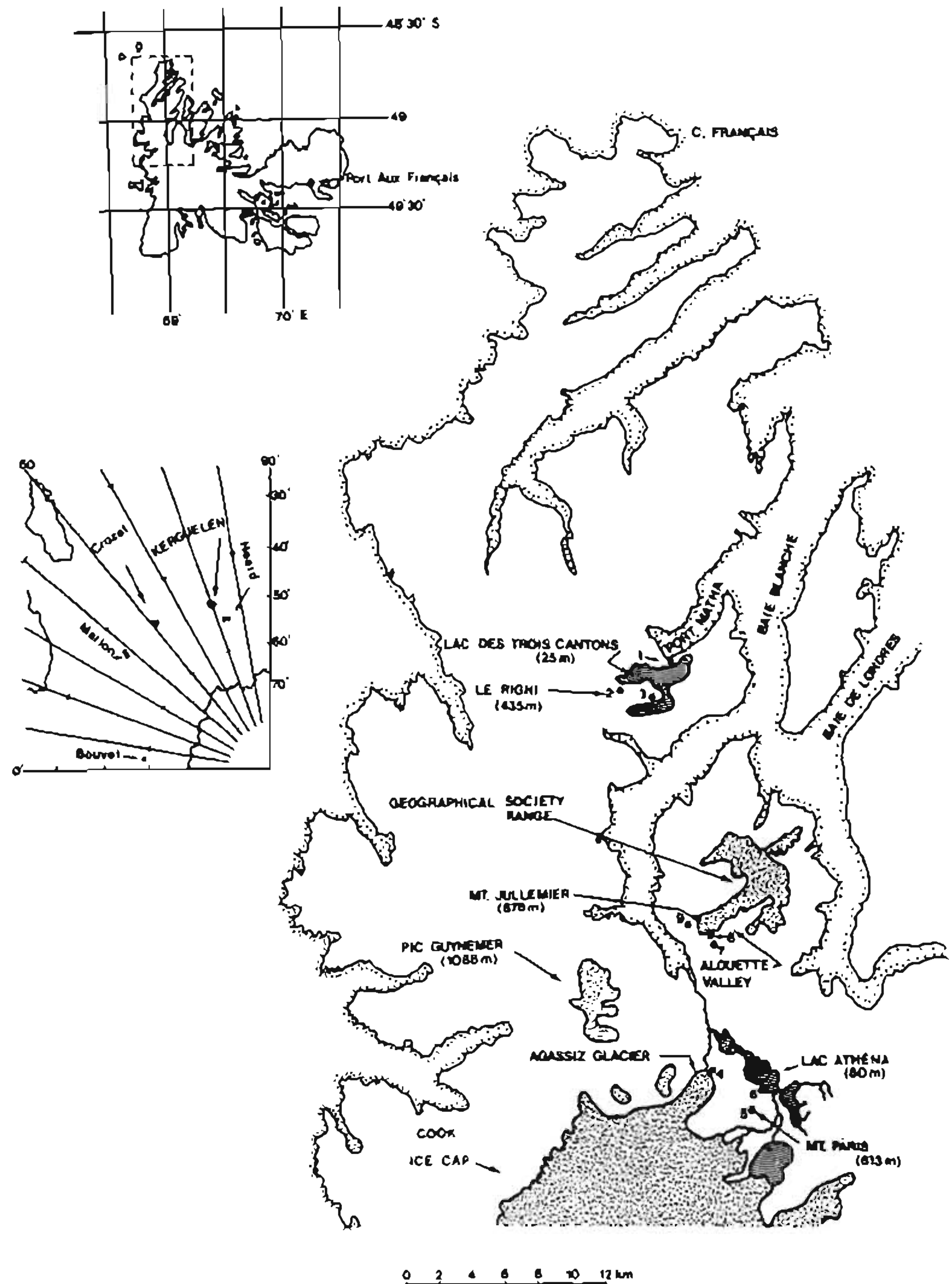

Figure 1. The location of Kerguelen Island together with the position of study sites noted in Table I

available slope'. A review of information pertaining to sorted stripes in general can be found in Washburn (1979) whilst details with specific reference to the sub-Antarctic are given in Araya and Hervé (1972), Aubert de la Rue (1959), Bellair (1969), Bunt (1954), Clapperton (1971), Clayton (1977), Hall (1979), Jennings (1956), Markov (1971), Nougier (1964, 1970), Stone (1974), Troll (1960) and Van Zinderen Bakker (1978).

With respect to the sub-Antarctic, recent work on Marion Island (Hall, 1979), following that of Bunt (1954) on Macquarie Island, has tended to suggest that the orientation of stripes is related to wind direction, as postulated by Troll (1958), rather than ablation of pipkrake by early morning sun as hypothesized by Mackay and Mathews (1974). A recent opportunity to visit sub-Antarctic Kerguelen offered the possibility of further observation of these sorted features.

OBSERVATIONS

A general summary of the information obtained from the nine localities studied (Figure 1) is presented in Table I. Additional, specific details will now be given, where applicable, for the individual sites.

Mt. Paris

Mount Paris (613 m) is a basaltic peak situated at the northeast edge of the Cook ice cap between the Chamonix and Agassiz glaciers (5 in Figure 1). On the gently sloping surface of the mountain top an unusual form of patterned ground was observed. It consisted of a secondary stripe pattern existing within the fine stripe of a primary, larger set (Figure 2). The stripes were orientated to the northwest (326°) down a 2–3° slope; details of stripe widths are given in Table I and Figure 3. A section through the primary stripes (Figure 3) showed the coarse material in a trough whilst the fine part of the stripe exhibited a 'corrugated' surface. Within the corrugation depressions was found a coarser material than on the rises, thereby creating a secondary stripe set (Figure 3).

Alouette Valley

Mount Jullemeir is a 878 m peak in the Geographical Society Range (Presq. 1. de la Societe de Geographie) which forms the north wall of the Alouette Valley (Figure 1). Observations were taken at approximately 500 m a.s.l. on the southern flank of the mountain, near the mouth of the valley (8 in Figure 1). In addition to a

Table I. Generalized details of sorted stripes

Location	Location No.*	$\bar{x}$ CSW (m)	$\bar{x}$ FSW (m)	$\bar{x}$ orient (°)	$\bar{x}$ slope (°)	alt (m)	$\bar{x}$ depth sorting (m)
Port Matha	1	0·22	0·16	304	5	60	0·13
Le Righi	2	0·11	0·16	032	2	335	0·15
Le Righi	2	0·09	0·16	010	1	390	0·16
Le Righi	2	0·20	0·14	266	1	435	0·22
Lac Trois Cantons	3	0·10	0·03	061	1	25	0·03
Agassiz Glacier†	4	0·13	0·09	259	4	80	0·12
Mount Paris	5	0·39	0·84	326	2	600	0·29
Mount Paris‡	5	0·11	0·15	326	2	600	0·11
Lac Athena	6	0·17	0·12	288	6	75	0·05
Alouette Valley	7	0·20	0·28	188	4	250	0·13
Alouette Valley	8	0·28	0·84	191	2	500	0·20
Alouette Valley§	8	0·07	0·14	240	0§	500	0·05‖
Mount Jullemeir	9	0·14	0·20	269	4	300	0·13

CSW = Coarse stripe width. FSW = Fine stripe width.
* See Figure 1.
† Stripes on a moraine proximal slope.
‡ Stripes occurring within the fine stripe of larger set (see text).
§ Stripes running across slope.
‖ Depth to bedrock.

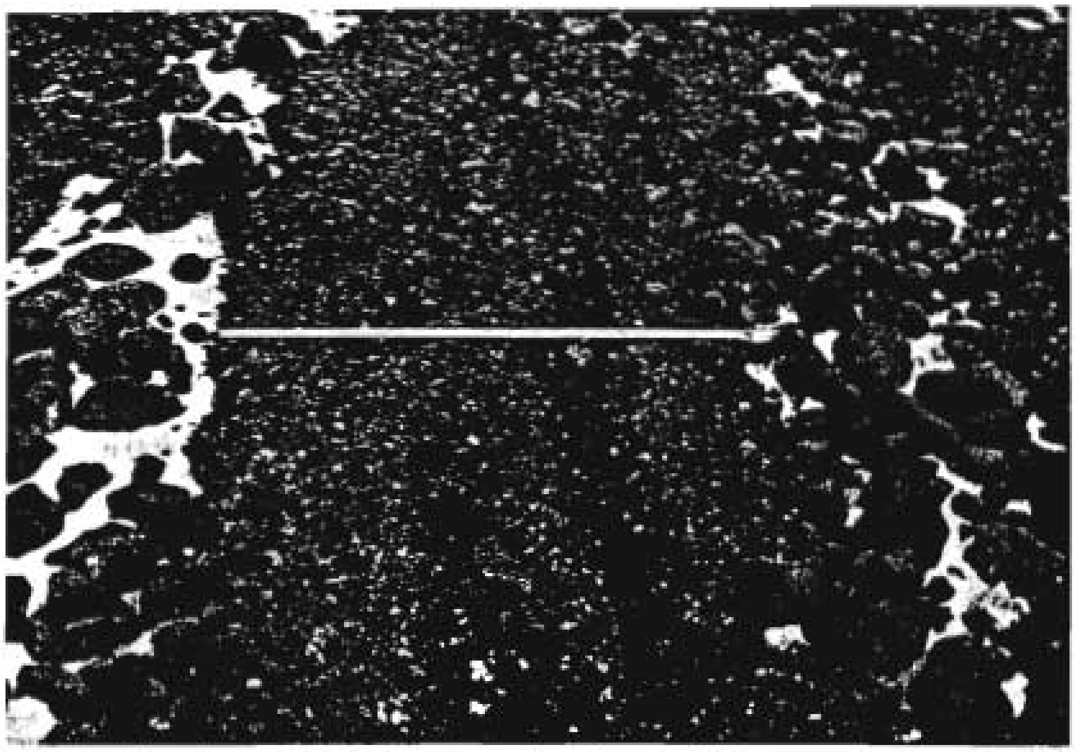

Figure 2. Stripes-within-stripes seen on Mt. Paris. Two coarse and one fine stripe of the larger set are shown (tape = 1 m) whilst the secondary set, within the larger fine stripe, is marked by darker bands (coarse stripes) and lighter bands (fine stripes)

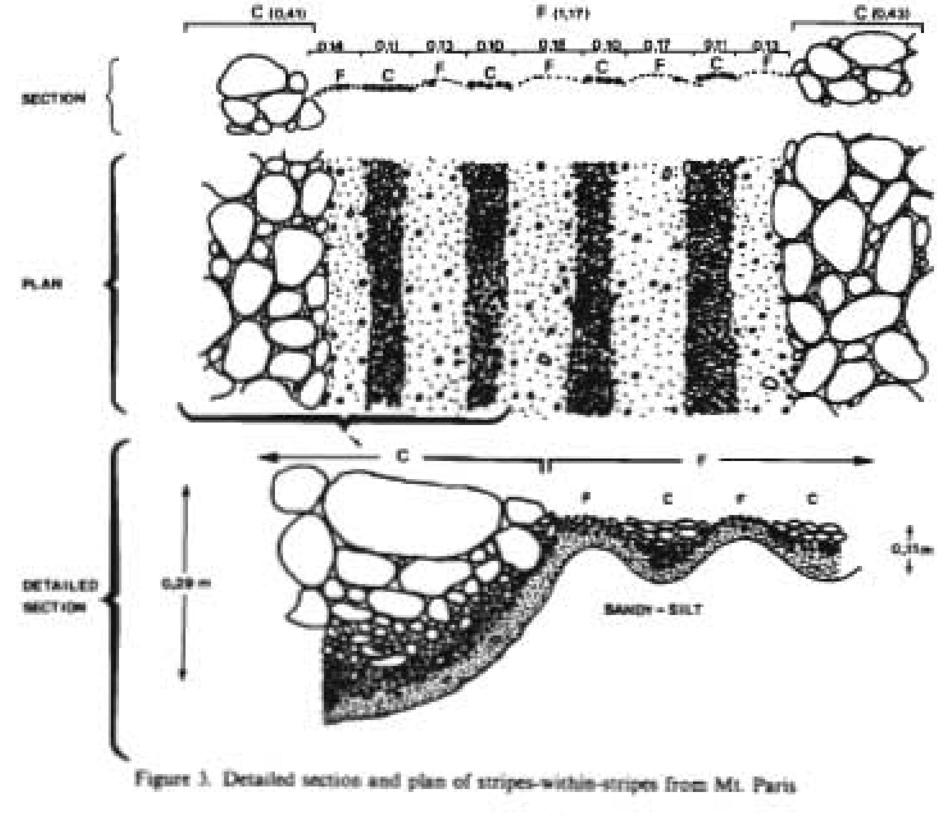

Figure 3. Detailed section and plan of stripes-within-stripes from Mt. Paris

'normal' set of sorted stripes (Table I) there was found a small group of stripes which ran across the slope. These cross-slope stripes cut across a 5° slope locally orientated to the south–southwest (201°) such that their axis (60–240°) was roughly aligned parallel to the Alouette Valley. Development of the stripes, particularly depth of sorting (Table I), was limited by their growth in a thin (0·05 m) veneer of glacial debris overlying a striated bedrock surface.

Port Matha

At Port Matha (1 in Figure 1) there exists a trachyte plug the weathering of which generates angular, platy clasts. Sorted stripes developed in this material (Table I) have clasts within the fine stripe with their *a/b* planes upward facing whilst in the coarse stripe the *a/b* plane is vertical with the *a*-axis orientated parallel to the slope (Figure 4). Measurement of 50 clasts in each of the coarse and fine stripes, of three stripe sets located at various places on the plug, show distinct differences in flatness indices and *a*-axis lengths (Table II). The clasts within the coarse stripe are flatter (especially in sample 1) than in the fine and with distinctly longer *a*-axes. Thus it

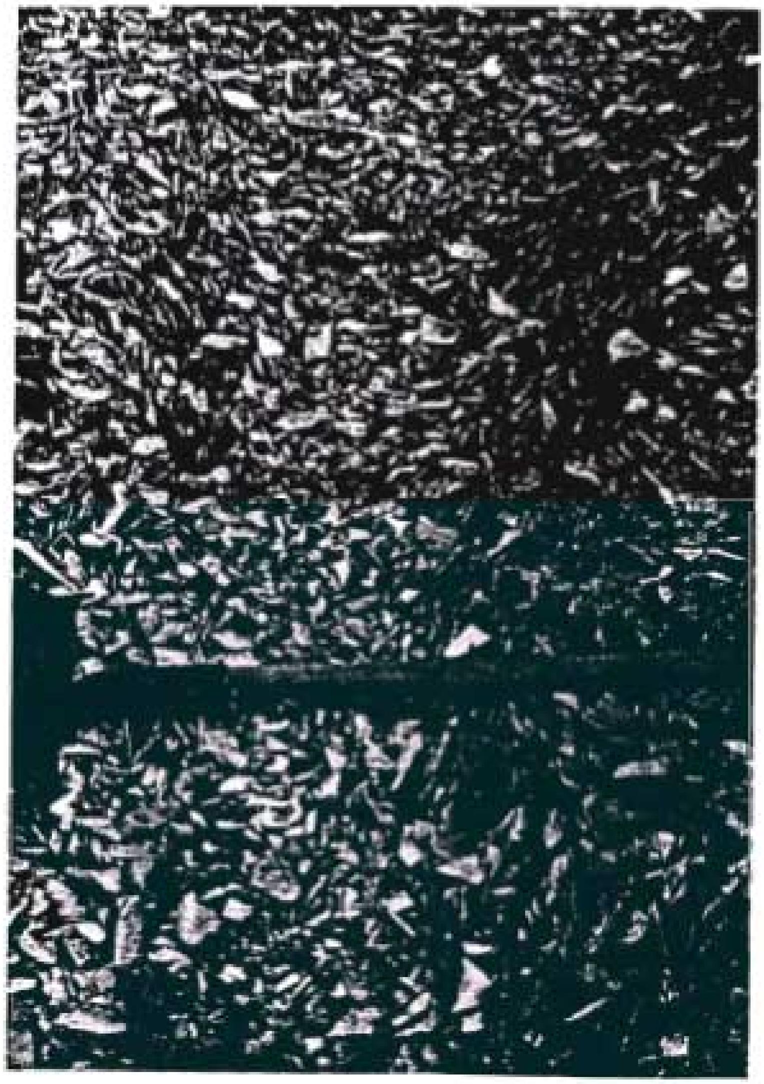

Figure 4. Sorted stripes found on the trachyte plug at Port Matha showing the edge-upwards clasts in the coarse borders

Table II. Clast flatness indices and a-axis lengths for fine stripes at Port Matha

	Coarse				Fine			
	$\bar{x}$ Flatness	s	$\bar{x}$ a-axis (mm)	s	$\bar{x}$ Flatness	s	$\bar{x}$ a-axis (mm)	s
Sample 1	1845	784	86·7	13·3	347	208	37·5	9·7
Sample 2	831	327	66·7	16·1	765	540	29·4	6·9
Sample 3	544	397	49·4	14·5	325	109	16·9	4·9

$$\text{Flatness} = \frac{a+b}{2c} \times 100$$ (Cailleux, A., 1945).

s = sample standard deviation.

would appear that the flatter, longer clasts are located, thin edge upwards, in the coarse stripe whilst the smaller, more blocky clasts, with their greatest surfaces upwards, constitute the fine stripe. With respect to the clast a-axes (Table II) it is interesting to note that a correlation of their mean lengths from the coarse and fine stripes gave an r value of $+0{\cdot}99$.

DISCUSSION

It was most noticeable on Kerguelen, as compared to the earlier findings on Marion Island (Hall, 1979), that there was less direct relationship between the dominant westerly winds and the stripe orientation. This is thought to be due to the more dissected landscape of Kerguelen (Figure 1; Carte de Reconnaissance, 1968) which results in many slopes being completely sheltered from the westerlies. However, at the same time, the role of the wind is still considered to be of prime importance. Subjectively, the majority of the localities where stripe development was observed were in places exposed to winds. This was clearly seen on Le Righi (2 in Figure 1) where stripes were found on gentle-sloping terrain parallel to the local wind whilst steeper slopes, leading into this flatter area from each side were devoid of stripes.

At Port Matha (1 in Figure 1) stripe development was restricted to the windward east, through north, to west sector (plus the top) of the plug. The leeward south–southeasterly slope (along Lac des Trois Cantons), although of the same material, was almost devoid of sorting of any kind. During the study period it was quite noticeable that the slopes along Lac de Trois Cantons was protected from all local winds and was a relative haven compared to working on all other sectors of the plug. Consideration of Carte de Reconnaissance (1968) shows that the southern slopes are protected by surrounding mountains from any winds whilst the other faces are exposed to winds blowing down the Baie du Centre or Val du Thermometre.

The role of the wind was also clearly evident at the Alouette Valley (8 in Figure 1) where stripes were found to cut across the slope rather than, as would normally be the case, to run down the slope. It is suspected, although no evidence is available from the very brief helicopter visit to the site, that katabatic winds from the Alouette glacier (and other unnamed glaciers along the northern top of the valley) are funnelled down the valley such that the stripes are orientated parallel to this. The other set of stripes found at this locality (Table I), running downslope to the south, southwest (191°), were behind a moraine and so in a less exposed position with respect to the katabatic wind. This set of stripes would appear to be of the more 'normal' kind which (Washburn, 1956) develop down the '. . . steepest available slope'.

Thus, as with Marion Island (Hall, 1979), there is development of stripes which appears to be related to the dominant wind direction. Bunt (1954, p. 36) in describing sorted stripes on Macquarie Island stated 'Further, the behaviour of wind currents at the ground surface may also influence the freezing pattern and the size of the ice palisades, as the formations are better developed on windward slopes'. Despite this apparent relationship noted by several workers (Troll, 1958; Schubert, 1972; Beaty, 1974; Hastenrath, 1977; Hall, 1979; Derbyshire, personal communication) the actual mechanism(s) operative is far from clear. Stripes on low angle slopes ($< 1°$) in east Otago, New Zealand, are thought by Brockie (1968) to result from frost action concentrating stones in a sub-parallel network of equally spaced rills that result from small-scale fluvial processes developed as remnant areas of alpine tundra are undercut by frost action and destroyed. This situation is unlikely on the unvegetated slopes of Kerguelen although, as Bunt (1954) suggests, drainage along the depressions of the developed coarse stripes would slowly remove fine material. However, this does not help explain the original development of the stripes but rather their entrenchment once formed.

Hall (1979) noted that fine-grained material raised by the needle ice was in a desiccated state and so easily removed from the needles by wind. Meentemeyer and Zippin (1981), in an experimental study of needle ice growth, showed that there was increased lifting when the soil surface became desiccated. This desiccation allowed the needle ice freezing front to form deeper in the soil and the amount of soil lifted was a function of the freezing front depth. Thus it may be that, in some manner, the combination of undirectional winds, desiccated fine material atop the needles, and variation in amount of lift between fine and coarse material, resulting from non-uniform penetration of the freezing front due to different thermal conditions, results in the initiation of longitudinal sorting. Once sorting has begun it becomes accentuated until stripes are resolved; which can then be maintained by other processes such as postulated by Bunt (1954). Bellair (1969) suggested that on Kerguelen the main process in the formation of the sorted stripes was alternate wetting and drying rather than freeze–thaw activity. Whilst this is certainly active, it is unlikely, particularly in the light of the number of freeze–thaw cycles, to be the main agent responsible. It would be difficult to explain the increase in stripe width with height (see below) or the change from a dominant coarse stripe to a dominant fine stripe with increase in height (see below) by this mechanism. In addition, the arrangement of the clast axes in the stripes at Port Matha (Figure 4) is compatible with the 'squeezing' mechanism suggested by Goldthwait (1976, p. 31) which is a function of the horizontal expansion of the fine stripe towards the still unfrozen coarse stripe during freeze.

Whatever the exact mechanism of sorting, it would appear that the wind-orientation hypothesis of Troll (1958) is applicable to the windy, overcast sub-Antarctic region. Ice needle ablation by the early morning sun, as proposed by Mackay and Mathews (1974), is inhibited by the almost continual cloud cover (Weyant, 1967) found within these regions. Cloud cover is so extensive that the amount of sunshine received may be limited to as little as 20 per cent of the maximum possible (Hall, 1979). The inadequacy of the suns effect is illustrated by the lack of stripes orientated to the northeast, as would be observed if melting were initiated by the early morning sun.

Goldthwait (1976, p. 33) suggests that the '. . . largest clasts in the diamicton control the spacing of the gutters . . .' and so stripe widths should be a function of clast size. However, on Kerguelen, as on Marion Island (Hall, 1979), it was found that individual stripe widths generally increased with altitude (Figure 5). Cubic regression of fine and coarse stripe widths with altitude are given in Table III. The r value for the complete data set was fairly low but increased significantly when the data were considered in two groups: observations from 25–80 m a.s.l. and observations from 250–600 m a.s.l. This splitting of the data is based on the evidence that coarse widths were greater than fine within the first data set whilst the reverse was the case in the second data set (Figure 5). The evident increase in stripe widths with altitude was comparable with that found on Marion Island (Hall, 1979).

It is interesting to note that the plotting of the differences between coarse and fine stripe widths (Figure 5) shows a distinct increase with altitude. Unfortunately no samples were obtained within the height range 81–249 m wherein, at some point, fine stripe width becomes wider than the coarse (Figure 5). Regression of the difference in stripe width with altitude (Table IV) shows that a cubic expression best fits the available data. The reason for the change over from a dominant coarse stripe, at lower altitudes, to a dominant fine stripe, at higher altitudes, is uncertain. However, the general increase in actual stripe width is thought to be a function of more powerful freeze–thaw cycles at the higher altitudes which effect larger sorting cells. Correlation of individual coarse and fine stripe widths at each site (Table V) failed to show the high r value (0·97) found on Marion Island (Hall, 1979). This is thought to be a reflection of the less uniform exposure to the wind and the greater diversity of rock types compared to the more simplistic situation found on Marion Island.

The unusual form of stripe found on top of Mt. Paris (Figures 2 and 3), wherein small stripes occur within larger ones, is said by Goldthwait (1976, p. 33) to be possibly, a result of a bimodal clast composition. He states that '. . . only if two entirely separate sizes of coarse pieces are common at the surface (over 20% each), and

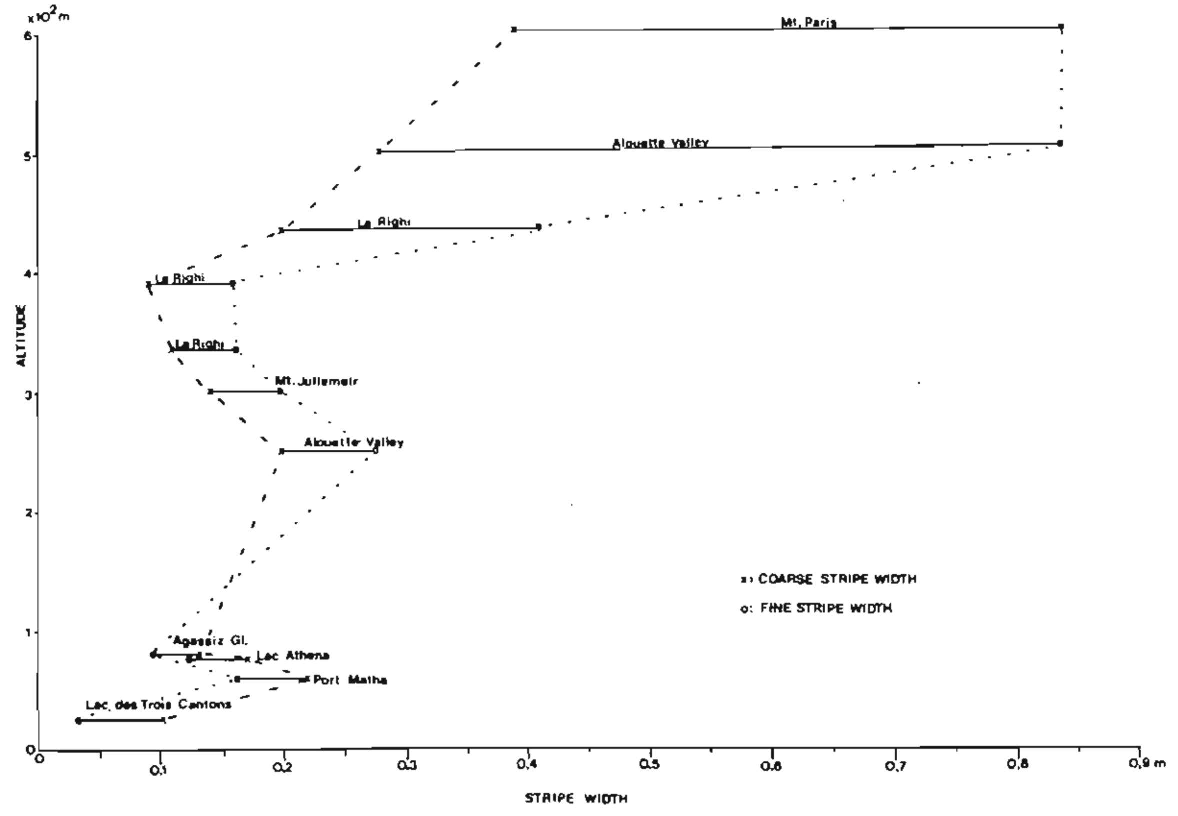

Figure 5. Plot of coarse and fine stripe widths with altitude

only if thaw activity is just right, will such a pattern-within-pattern form'. At Mt. Paris there was no evidence of any bimodal clast distribution for clasts comprising the coarse stripe of the secondary (smaller) stripe set were found in the coarse stripe of the primary set just below the large surface boulders. The cause for this particular pattern is not known but the possibility arises of a secondary set of sorting arising as a function of freeze–thaw cycles less powerful than those forming the dominant stripes. In other words, the large stripes result from larger amplitude and wavelength cycles, of which there may be only a few each year, which effect large sorting cells. The secondary stripes arise as a result of minor sorting of the large fine stripe due to more

Table III. Cubic regression of stripe width with altitude

	FSW		CSW	
	r	r^2	r	r^2
All data	0·68	0·82	0·50	0·71
Sites 25–80 m a.s.l. (Nos 1, 4, 5 and 6)	1·00	1·00	1·00	1·00
Sites 250–600 m a.s.l. (Nos 2, 3, 7, 8 and 9)	0·92	0·96	0·81	0·90

FSW = Fine stripe width. CSW = Coarse stripe width.
Site numbers shown in Figure 1 and Table I.
Individual data units (means) shown in Table I.
Cubic equation of the form $y = a + bx + cx^2 + dx^3$ (see Mather, 1976).

Table IV. Regression of the difference between coarse and fine stripe widths with altitude

	r	r^2	a	b	c	d
Linear:	0·75	0·57	−0·07	7·61e^{-4}	—	—
Exponential:	0·82	0·68	0·03	4·20e^{-3}	—	—
Log:	0·61	0·37	−0·57	0·13	—	—
Power:	0·69	0·47	−1·52e^{-3}	0·77	—	—
Cubic:	0·89	0·80	0·25	−3·52e^{-3}	1·25e^{-5}	−9·80e^{-9}

Equations of the form:
Linear : $y = a + bx$
Exponential: $y = ae^{bx}$
Log : $y = a + b \ln x$
Power : $y = ax^b$
Cubic : $y = a + bx + cx^2 + dx^3$
(see Gregory, 1968, Mather, 1976 and Yamane, 1973).

Table V. Correlation of coarse stripe width with fine stripe width

Location	Map Key (Fig. 1)	$\bar{x}$ CSW (m)	$\bar{x}$ FSW (m)	r	r^2
Port Matha	1	0·22	0·16	+0·85	+0·72
Le Righi	2	0·13	0·24	+0·99	+0·98
Alouette Valley	7	0·20	0·28	+0·94	+0·88
Mt. Jullemeir	9	0·14	0·20	+0·43	+0·18
Total observations	1–9			+0·58	+0·34

CSW = Coarse stripe width. FSW = Fine stripe width.
r = Pearson's product-moment coefficient of linear correlation.
(see Till, 1974, pp. 83–88).

frequent but less powerful freeze–thaw cycles. In fact the major stripes may be due to ice lensing at depth within the ground whilst the minor stripes are due to needle ice activity. Unfortunately no data are available to support this but the hypothesis of Goldthwait (1976) does not appear to fit this case and it is unlikely that the larger set are fossil as their size is compatible with the increase in stripe width observed to occur with height (Figure 5).

CONCLUSIONS

The observations from Kerguelen Island appear to strengthen the case for wind alignment of sorted stripes. The sub-Antarctic climate of almost continual overcast conditions would not be conducive to the ablation of needle ice by the early morning sun whereas most localities are subject to the strong winds. Stripe width is seen to increase with altitude, reflecting the deeper frost penetration. In addition, the coarse stripe is dominant at lower altitudes whilst the fine stripe is more pronounced at higher elevations. Stripes-within-stripes may reflect a response to two types of ice activity (ice lenses and needle ice) and/or two sets of freeze–thaw cycles: one of large amplitude and wavelength and one of more frequent but less powerful effect.

The actual mechanism of stripe development by needle ice and the resultant alignment to the wind is still uncertain. However, desiccation of the fine material, which can then be removed by the wind, plus uneven needle ice growth, as a function of sediment size, may play a part. Considering the ubiquitous nature of these features within the sub-Antarctic, there is a need for detailed field monitoring and investigation to discern their method of genesis.

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Observations of some periglacial features and their palaeoenvironmental implications on sub-Antarctic islands Marion and Kerguelen

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Some quantitative data on patterned ground for the sub-Antarctic islands Kerguelen and Marion are presented. Some features are recognised as fossil and of being a product of cooler than present post-glacial conditions. The present-day sorting is smaller in scale than that produced immediate upon deglaciation. The periglacial features thus record the environmental change since the disappearance of the ice cover.

Sekere kwantitatiewe gegewens oor gepatroonde grond vir die sub-Antarktiese eilande Kerguelen en Marion word aangebied. Sekere kenmerke word erken as fossiel en 'n produk van koeler as teenwoordige na-glasiale toestande. Die huidige sortering is kleiner volgens skaal as die wat direk na deglasiasie geproduseer is. Die periglasiale kenmerke gee dus 'n weergawe van die verandering in omgewing sedert die verdwyning van die ysbedekking.

Introduction

Marion Island (lat. 46°54'S, long. 37°45'E) and Kerguelen Island (lat. 49°21'S, long. 70°12'E) are situated just to the north of the Antarctic Polar Front, within the belt of mid-latitude westerlies (Fig. 1). Both islands are volcanic in origin and have been extensively glaciated in the recent past (Hall 1979b in press, Nougier 1972). Kerguelen (6 000 km^2) has a present-day ice cover of *c* 750 km^2 (Mercer 1967), whilst smaller (290 km^2) Marion Island only maintains a remnant region (<3 km^2) of permanent snow and ice above the 90 km contour. An extensive assemblage of periglacial features is present on both islands (Aubert de la Rüe 1959, Bellair 1969, Hall 1979a, 1981, 1983, Markov 1971, Nougier 1964, Troll 1960) but very little quantitative data pertaining to either the landforms or their formative processes are available. The aim here is to present some new data, on both features and proces-

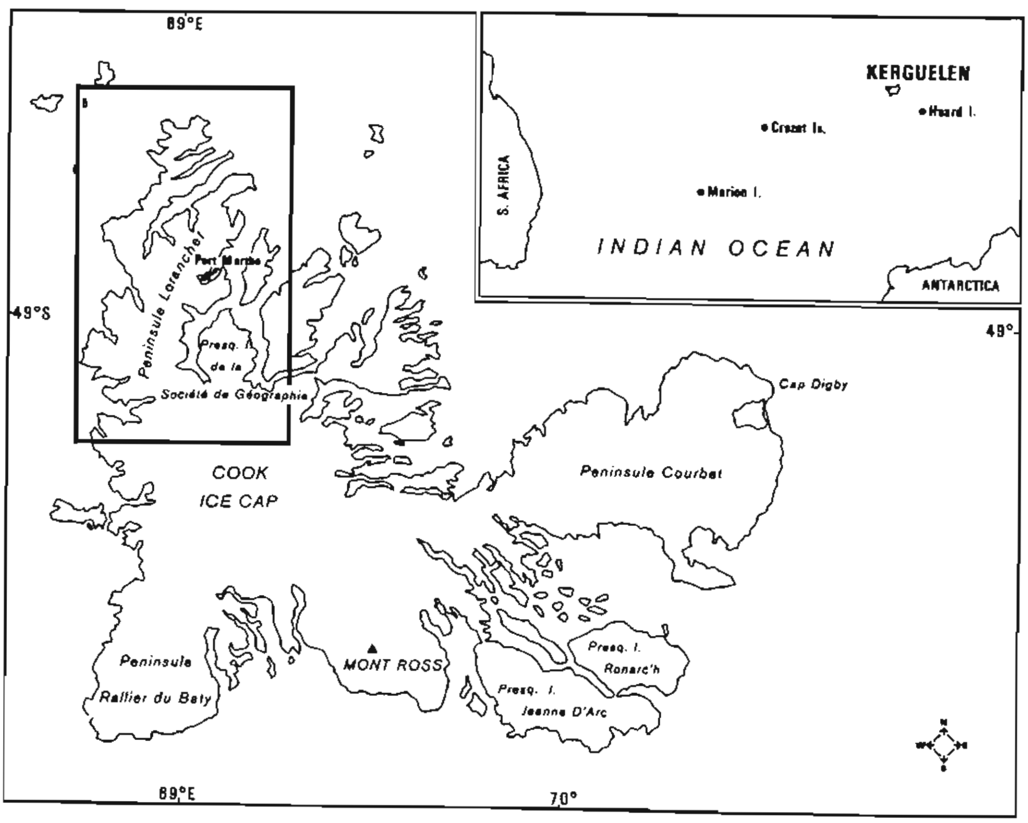

Fig. 1. Map showing the location of Marion and Kerguelen islands together with the study area on Kerguelen.

ses, for periglacial phenomena on both Kerguelen and Marion islands and to use this information to consider environmental changes.

Climatic setting

Both islands experience a typical sub-Antarctic climate of strong winds, extensive cloud cover, frequent precipitation, low radiation inputs and low but equable temperatures with the possibility of frosts any time during the year. Data from Kerguelen indicate precipitation on (approximately) 324 days yr^{-1} with 60 per cent of all observations showing six oktas or more cloud (Aubert de la Rue 1959). During one year Aubert de la Rue (1959) monitored 120 frosts in the air and 200 at ground level, whilst Troll (1944, 1960) recorded freeze-thaw cycles at the ground surface on 238 days yr^{-1} and 236 days yr^{-1}, but with a maximum penetration of 0,05 m.

Marion Island experiences a similar climate with frequent (60 % of occurrences) northwest winds at a mean velocity of 32 km hr^{-1} with, on average, 25 days in each month receiving some form of precipitation (total = 2 576 mm, Schulze 1971). Radiation receipts are decreased by the high incidence of cloud ($\bar{x}$ = 6 oktas), with the amount of sunshine received being cut to 33 per cent of that possible in summer and 20 to 25 per cent in winter (Schulze 1971). Interpretation of meteorological data (Hall 1979b) suggests 48 days yr^{-1} with temperatures of -2 °C or lower at sea level ($\bar{x}$ = freeze amplitude $-3,4$ °C) and 111 days yr^{-1} at 500 m a.s.l., with a mean freeze amplitude of $-4,1$ °C (Hall 1979b, Table 1).

Periglacial features

On Kerguelen observations are restricted to Peninsula Loranchet (Fig. 1) for which area details relating to sorted stripes

Table 1
Summary of clast sizes and shapes for two sorted nets at Port Matha.

		$\bar{x}$ a-axis (mm)	s	$\bar{x}$ flatness	s	$\bar{x}$ OP Index (1)	s	% oblate	% prolate
NET	1								
	Fine centre:	24	8	307	134	$-1,41$	5,93	60	30(2)
	Coarse border:	81	17	1300	659	$-10,68$	29,47	60	40
NET	2								
	Fine centre:	30	7	310	131	$-0,04$	9,03	40	50(2)
	Coarse border:	97	30	1788	490	$-49,39$	37,49	90	10

(1) Oblate – prolate index of Dobkins & Folk (1970).
(2) Residue (10%) constitutes clasts with OP indices of 0,00.

Fig. 2. An area of sorted nets on the trachytic plug at Port Matha, Kerguelen.

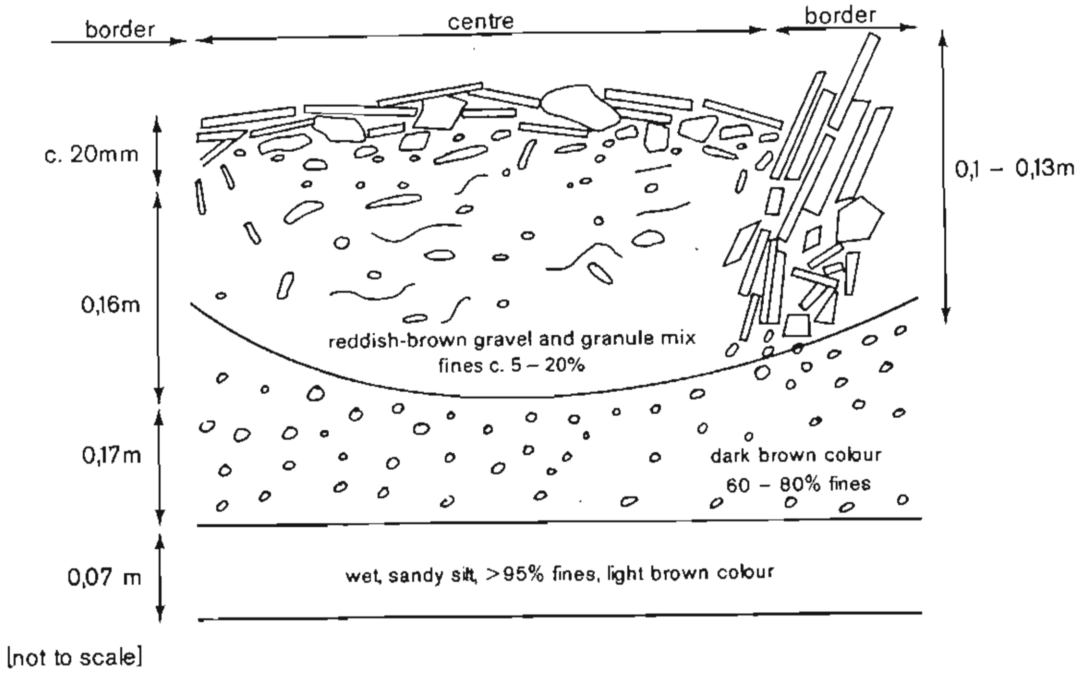

Fig. 3. A section through the sorted nets at Port Matha.

have already been presented (Hall 1983). In addition to sorted stripes a number of observations pertaining to sorted nets were also obtained. Sorted nets, "Patterned ground features occurring in groups whose mesh (interior surfaces) is neither dominantly circular nor polygonal" (Brown & Kupsch 1974), have an extensive distribution on Kerguelen. However, to date they have been described only in qualitative terms (Aubert de la Rue 1959, Bellair 1969, Markov 1971, Nougier 1964, Troll 1960). Quantitative data on sorted nets (Fig. 2) were obtained from an altitude of *c* 60 m on the undulating surface of a trachytic plug at Port Matha (Fig. 1). Weathering, probably some form of freeze-thaw action, breaks the trachyte into angular, platy clasts, the sorting of which generates the nets.

Measurement of the longest axis of net fine centres (n = 25) indicated a variation of between 0,64 m and 0,21 m, with a mean length of 0,35 m (s = 0,12). Coarse border widths varied between 0,09 m and 0,67 m at their widest sections ($\bar{x}$ = 0,27 m). In the coarse borders the clasts reside with their a/b planes normal to the fine centres, whilst in the centres themselves the a/b planes are vertical. Measurement of clast shape and size for the centres and borders indicates a distinct compositional dichotomy (Table 1). The clasts comprising the borders are both larger and flatter than those in the centres. In addition, the border clasts are less well sorted, with respect to size and shape, than are those in the centres.

A section through the sorted nets (Fig. 3) shows that the clasts, in the fine centre, residing with their a/b planes vertical, are in a band only some 20 mm in thickness. Beneath them there is a *c* 0,16 m mix showing vertical sorting. At the borders the platy clasts are aligned normal to the centre to a depth of 0,10 m to 0,13 m, but the number of clasts exhibiting this orientation decreases with depth except at the margin with the centre where it is maintained. Below *c* 0,13 m the border transists into a gravel/fines mix. A distinct transition occurs within the net at about 0,16 m depth, where colour changes to a dark brown and there is a marked increase in the percentage of fines. A further change takes place after another 0,17 m, at which depth a light brown, wet and coarse material deficient zone is encountered. The fine centre of the net tends to show a slight doming and has pockets of sandy silt, some of which break through the surface stone cover.

In the Port Matha area of sorted nets it was observed that some small *Azorella* cushions had been penetrated by flat trachytic plates that had been forced up through them (Fig. 4). The *Azorella* was less than 0,2 m in thickness and showed internal damage due to the upward movement of the clasts. Growth of *Azorella* about the clasts is discounted on the basis of the internal damage and the tenuous position of clasts no longer in contact with debris beneath the vegetation.

What is also noticeable in this area, in terms of their omission rather than inclusion, is that gelifluction features appear to be very scarce. Stone-banked lobes and vegetation-banked mass movement forms are notably lacking compared to Marion Island. *Azorella* is less common in this area but nevertheless major mass movement forms in association with sorting were conspicuous by their absence.

The distribution of recognised periglacial features on Marion Island is shown in Fig. 5. Of these landforms, sorted stripes and stone-banked lobes (sorted lobes) have already been described in detail elsewhere (Hall 1979a, 1981). However, sorted polygons, circles and nets (Washburn 1979) were also observed. Form variety in these sorted features is a result of packing and slope: circles forming polygons due to density of packing and circles/polygons forming nets in areas of gentle slopes (*c* 2-6°). The largest of the observed nets had a fine centre 0,66 m along its greatest axis with a coarser border 0,05 to 0,15 m in width. These sorted features were observed from altitudes as low as 200 m a.s.l. right through to *c* 1 200 m. Circles and nets were also seen developed in a fines-rich till only 0,15 m deep at *c* 300 m on Piew Crags.

In many areas, on both volcanic and glacial debris, "steps" or "terraces" of material banked by *Azorella* risers are found. The bare, sorted tread may be from 1,5 to 10 m in width, with a lateral extent of 3 to 20 m behind a 0,2 to 0,7 m high riser. The vegetation does not appear to be in distress and in only a few instances was material from the tread seen to be moving over

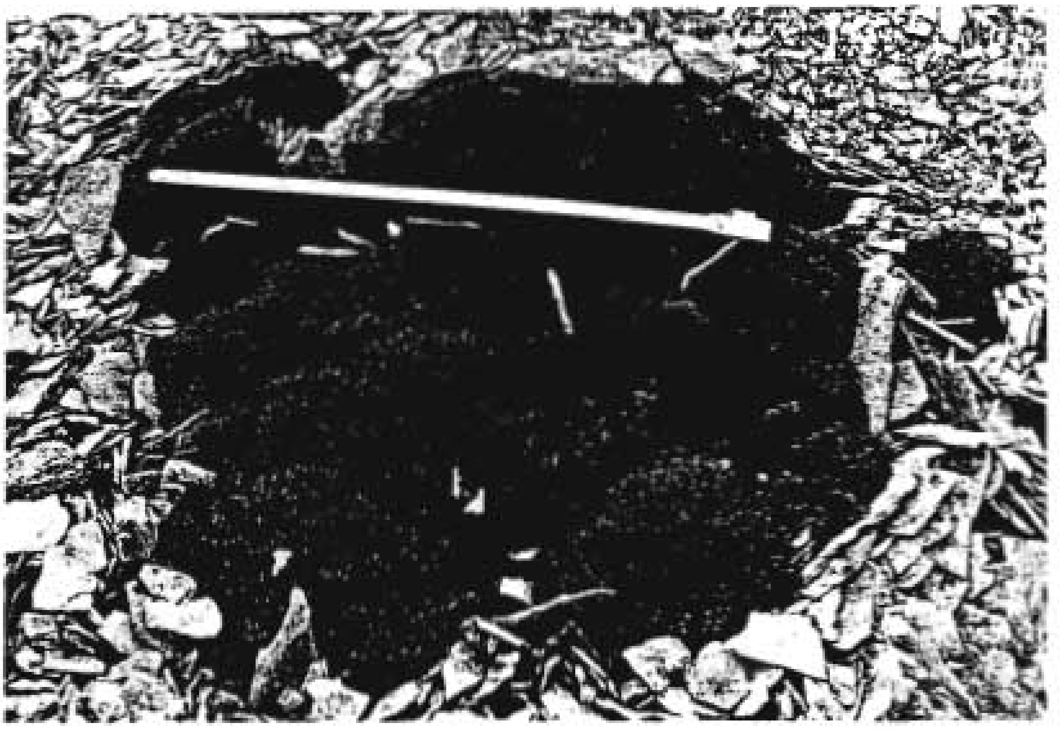

Fig. 4. An example of clasts forced up through an *Azorella* cushion at Port Matha.

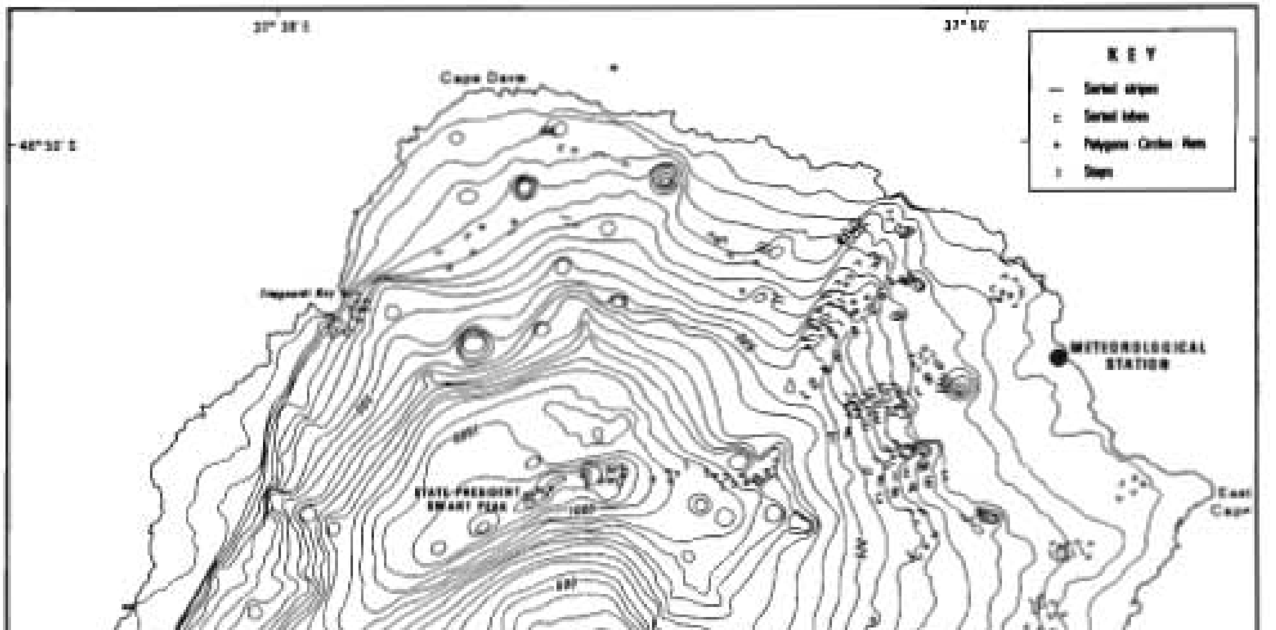

Fig. 5. The distribution of periglacial features observed on Marion Island.

or through the vegetation riser. These steps were found in areas of a 5° to 20° general slope which, subjectively, often appeared to be wet. On the steeper, more debris-strewn slopes beneath rock outcrops, were found mass movement forms characterised by stone-banked lobate snouts. Some of these stone-banked lobes, those with a sorted, fines-rich tread, have already been described (Hall 1981), but there were others which exhibited a surficial covering of rock debris. These were usually larger forms than those which had the bare soil treads and can be in the region of 15 to 20 m long, 4 to 5 m wide and 1 to 1,5 m high. Surface clasts tend to be platy and angular and to lack any preferred orientation. Beneath the one or two clast-thick surficial layer there is a sorted zone with the coarser clasts at the top, fining downwards.

Field observations throughout a year showed the presence of pipkrake at all altitudes above 50 m a.s.l., whilst lenticular freezing of the ground, to a depth of 0,15 m, was observed at heights in excess of 250 m. The pipkrake attained lengths of 50 mm and occasionally exhibited evidence of more than one growth phase. The effects of aspect and shadow were pronounced in prolonging freeze duration and increasing depth of freeze. Basalt outcrops were often seen to show severe fracturing and block disintegration, with large spreads of angular blocky or platy debris mantling their lower slopes. However, thermoclastis under present, or, more likely, past severer conditions, cannot be assumed as the sole cause for strain release or crack propagation forces other than freeze-thaw may be responsible (Whalley *et al.* 1982, McGreevy & Whalley 1982), particularly in igneous rocks such as basalt.

Discussion

The periglacial features recorded here for Marion and Kerguelen islands offer nothing new or exciting in themselves other than their recognition, in new geographical localities, and some quantitative measurements. However, when viewed in terms of their formative processes they are indicators of broad environmental changes from the end of the last glacial through to the present. Seen as a whole, the periglacial assemblage represents a continuum of landforms varying in scale as an adjustment to climatic amelioration.

The glacial-postglacial boundary is suggested to be about 12 000 BP in this sub-Antarctic region (Schalke & Van Zinderen Bakker 1971). However, on Marion Island ^{14}C dates obtained from the base of thick peat sections (Lindeboom 1979, Scott pers. comm) suggest that peat growth at the coast (i.e. the area first ice-free) was only initiated *c* 7 000 BP. This then suggests that there was a time-lapse of several thousand years between deglaciation and development of a vegetation cover. During this phase the ground would have been exposed to temperature variations without the damping influence of a vegetation mat: ideal conditions for periglacial activity. In addition, as the ice-caps waned, even extremely rapidly as in the case of Marion (Hall 1982), the lower regions would have been subject to cold katabatic winds. At the same time, there would have been a greater incidence of cold southerly winds than at present for the retreat of the Antarctic pack-ice would not have been at the same rate as island ice-cap diminution. Thus, the lowland on the islands would have been continuing to experience colder-than-present conditions although the greater ice cover found on Kerguelen (Hall, in prep.) would have retreated at a much slower rate than on Marion Island, so that there would not have been synchroneity of coastal region exposure on the two islands.

Reconstruction of temperatures for Marion Island (Hall 1979b) suggests that mean annual values were only 2 – 4 °C cooler than present, although the greater incidence of cold southerly winds was an additional influence inducing greater cooling. Thus, both thermoclastis of rock outcrops and depth of sorting in the ground were more extensive that at present. The suggested drop in temperature would have given numerous freeze-thaw cycles with a freeze amplitude of −3 °C or lower, a value which McGreevy and Whalley (1982) indicate to be the threshold for effective mechanical weathering by ice action. The greater depth of sorting in the ground would explain the presence of large, possibly fossil, sorted stripes observed on Kerguelen (Hall 1983) and the large, rubble-covered, stone-banked lobes found on Marion Island, for in both cases the depth of sorting is far greater than is presently possible. The enhanced freeze-thaw activity would have produced the large, angular plates and blocks found on the lobes of Marion Island, and may have caused the breakdown of the trachyte in which the sorted nets on Kerguelen then developed as the fines produced allowed ice segregation and hence sorting.

It is probable that the *Azorella*-banked steps found on Marion Island were initiated during the immediate post-glacial cooler phase. Similar steps, or terraces, are described on Macquarie Island (Taylor 1955, Löffler pers. comm.). The "leeward terraces" of Taylor (1955, Fig. 1) appear from the descriptions to be identical to those found on Marion Island and are noted as having (p. 134) their ". . . position and dimensions . . . constant for a long period of time". In fact Löffler (pers. comm.), from studies in 1979 and 1980, considers them "relict solifluction landforms developed during a more severe frost climate". Taylor (1955) suggested that the terraces developed due to banking of material behind established *Azorella* and that solifluction played very little, if any, part. Löffler, however, considers the slopes on which the terraces developed to be insufficient for scree movement and so it is necessary to invoke periglacial solifluction (gelifluction) to explain the downslope mass movement. This then implies that the *Azorella* grew in the sheltered position offered by the terrace riser and thus was a result of, rather than a cause of, the terrace. On Marion the steps/terraces occur on general slopes even gentler than are found on Macquarie, plus the vegetation shows no sign of stress or destruction, and so a similar causal sequence to that suggested by Löffler is envisaged. In fact Löffler (pers. comm.) suggests that the time of development was ". . . at the end of the last glacial immediately upon the disappearance of the ice . . ." and that during this phase temperatures were at least 4 °C lower than at present, a figure very similar to that derived for Marion Island. The apparent lack of these forms on Kerguelen is puzzling and no satisfactory explanation has yet been forthcoming. Certainly, terraces developed close to the glacier margins would have been destroyed during the Neoglacial advance, but this still leaves vast areas where features could have survived and, considering Kerguelen has a severer climate than Marion or Macquarie, may have still been active.

The present-day processes of pipkrake action, wetting and drying, segregation ice to depths of *c* 0,05 m, and the influence of strong, frequent winds (Hall 1979a, 1983) account for the small sorted features and the slow downslope creep of material. Whether the present climate is sufficiently severe at the lower altitudes for thermoclastis is uncertain, due to lack of micrometeorological data. However, the frequent occurrence of pipkrake which require an initiation temperature of at least −2 °C (Outcalt 1971) suggests that some rock destruction due to sub-zero temperature cycling may be operative. Thus the small-scale sorted circles and nets are a product of the present climate, and probably some rock breakdown continues to take place, thereby providing coarse and fine material for sorting.

Conclusions

The periglacial assemblage on Marion and Kerguelen comprises two major groups: one set, now probably fossil, resulting from conditions severer than at present, the other currently active under the present-day climate. The post-glacial climatic amelioration together with the concomitant growth of a vegetation cover has resulted in a diminution in the effectiveness of freeze-thaw cycles, both on rock faces and in the ground. As a result a number of features produced under the severer conditions are no longer active and they have been superseded by smaller-scale processes giving a different landform assemblage. The presence of the fossil features indicates that the climate has ameliorated sufficiently since the end of the last glacial so that a threshold, with respect to development and maintenance of a number of landforms, has been reached. Although presently active patterned ground on Marion and Kerguelen are of roughly the same sizes, it would appear that the marginally cooler climate of Kerguelen produces deeper sorting than is found on Marion. In addition, the greater frequency of temperature cycling about the 0 °C isotherm produces highly active patterned ground on Kerguelen plus the greater potential for thermoclastis.

These generalised observations point the way to a need for detailed research on sub-Antarctic periglacial features on these islands, and a call is therefore made for a multi-disciplinary investigation of the periglacial landforms and history of the South African island, Marion. For such a study, micro-meteorological monitoring on rock faces and about presently active patterned ground, plus detailed measurement of landform size variation with altitude and aspect could prove fruitful. Dating of buried organic horizons (^{14}C method) would help both in establishing the age of fossil features and in estimating rates of gelifluction. All observations to date have been highly simplistic and qualitative when viewed against the depth of periglacial research in the Arctic and alpine regions (c.f. Washburn 1979) and thus a major gap in the pedological, botanical, morphological, climatic and environmental history of these islands is present.

Acknowledgements

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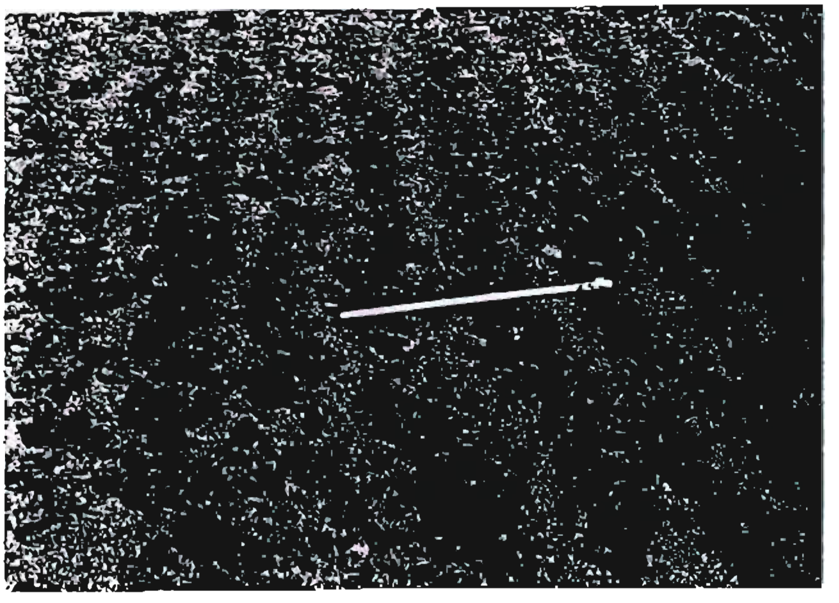

Photo 1. Top: Sorted stripes on Marion Island, bottom: Sorted polygons on Kerguelen

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Photo 2. Top: Stripes-within-stripes on Kerguelen, bottom: An exposure of frost shattered quartz-micaschist on Signy Island

PERIGLACIAL LANDFORMS AND PROCESSES OF THE SUB-ANTARCTIC AND ANTARCTIC ISLANDS

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ABSTRACT

Periglacial research in the Antarctic and sub-Antarctic has not yet achieved the level of sophistication found for Arctic and Alpine regions. Nevertheless, there is a wide assemblage of periglacial landforms on the southern islands and a number of valuable studies have been undertaken, most of which are not that well known by workers in the northern hemisphere. In this species-poor environment a strong relationship is often found between plant life and the patterned ground. As land areas in the vast Southern Ocean are very limited, the palaeoenvironmental and palaeoclimatological evidence provided by fossil and active periglacial features and processes is of the utmost importance. A brief summary of much of the available periglacial literature is presented in an attempt to generate interest in this field of study on the southern islands.

INTRODUCTION

Recent volumes on Antarctica and its surrounding regions (e.g. Bonner & Walton 1985, Walton 1986) have covered such topics as climate, soils, geology and glaciers in some detail but the periglacial system has been relatively ignored. Sugden (1982), in his geographic synthesis of both the Arctic and Antarctic, does show examples of periglacial activity from the Maritime Antarctic, and Walton (in Laws 1984) briefly notes periglacial processes but, compared to the extensive studies in the northern hemisphere (e.g. Washburn 1979, Harris 1986), this is a field comparatively unexplored in the South. However, the periglacial system is important as it is here that many cold region geomorphic processes interact intimately with the biological world. In fact, the relationship between periglacial features and botanical activity is frequently so strong that a study of either requires an intimate knowledge of the other. The periglacial processes operative around the margins of the ice cover take over as the glaciers recede. Several features of the periglacial sys-

Figure 1. Map of the Southern Ocean showing the location of the islands mentioned in the text.

tem are also very useful as palaeoclimatological/palaeoenvironmental indicators and, as such, can be useful tools in our study of the southern islands.

The number of islands, and their areal extent, is very small in the context of the great Southern Ocean (Fig. 1). However, they have great importance for they not only provide accessible land sites, bridging the gap between Antarctica and the surrounding continents, but they also constitute important locations for the acquisition of data that can be compared with those derived from ocean cores. Some of the islands are so extensively covered by ice (e.g. Bouvet Island, the Balleny Islands, Scott Island and Peter I Øy) that they do not support any periglacial activity except perhaps for frost weathering on the few exposed cliffs. Thus the number of islands to be considered is relatively small.

The information which is available on periglacial topics is published in a number of languages and widely spread through various books, journals and reports. The only concise source of information presently available is the annotated bibliography of Walton (1980) which deals with both pedology and periglacial processes in the Antarctic and sub-Antarctic. However, this is probably not as widely known as it should be and much more work has been undertaken in the six years since Walton's volume was completed. Thus, an attempt will be made here to briefly present much of the available information on periglacial features and processes from the southern islands and to comment on their importance.

MARION ISLAND AND PRINCE EDWARD ISLAND

Marion Island and Prince Edward Island (latitudes 46°48'-46°59'S, longitudes 37°35'-37°55'E) are the twin peaks of a submerged volcano (King 1969). The islands are small in size (290 km^2 and 40 km^2 respectively). Extensive studies were initiated here by Van Zinderen Bakker in 1965 (1971) and a bibliography of the research on the islands has been compiled by Siegfried et al. (1979).

Van Zinderen Bakker (1978) observed the presence of fossil and active sorted patterned ground (Washburn 1978) on both Marion and Prince Edward Islands. On Price Edward he recorded exceptionally long sorted stripes (270 m) above 500 m altitude, whilst on Marion it was suggested that permafrost might be present in the central mountain area. The occurrence of needle ice (pipkrake) was thought to account for the origin of moss balls and of some deflated areas. Hall (1979), following the above work, made a number of observations on the patterned ground, particularly that of sorted stripes (Photo 1). It was found that sorted stripes had a preferred alignment parallel to the dominant westerly winds, and that the effect of the wind was so strong that stripes even occurred on horizontal surfaces. The mode of stripe formation was not determined but a strong association with pipkrake activity was observed. Stripe width was found to increase with altitude but the fine stripes were consistently broader than the coarse.

Stone banked lobes (Hall 1981) and *Azorella*-banked steps (Hall 1983a) were recorded but both are thought to be largely inactive at the present time as the depth of lenticular freezing is so limited (c. 0.1 m). Both of these features owe their formation, in part, to gelifluction and, like similar features observed on Macquarie Island (see below), are considered to be related to a more severe climate. It is suggested that *Azorella* associated with the steps grew in that location due to the sheltered position offered by the step riser (Löffler et al. 1983). During this more 'severe climate' it is highly likely that gelifraction was extremely active (Hall 1983a) as there would have been numerous freeze-thaw cycles in what is a relatively wet environment. Gelifraction is probably an active process at higher elevations at the present time but no research has been undertaken there.

ILES CROZET

This group of five main islands resides in similar latitudes (lat. 46°-46°30'S, long. 50°30'-52°30'E) to those of the Marion and Prince Edward Islands and experiences a similar climate, particularly the incessant westerly winds.

Although these volcanic islands are mountainous, have been glaciated, and are sparsely vegetated at higher elevations (Aubert de la Rue 1967), there appears to be very little available information on periglacial activity. Philippi (1908) describes sorted stripes and polygons, as does Bellair (1969) who also noted solifluction activity. Bellair suggests that it is wetting and drying of the ground that is the formative process for stripes and polygons, but the recent work from Marion and Kerguelen (Hall 1979, 1983b) would indicate some freeze-thaw action.

ARCHIPEL DE KERGUELEN

This is an extensive archipelago (lat. 48°27'-49°58'S, long. 68°25'-70°35'E) of some 300 islands, with the main island comprising an area of approximately 6 000 km^2 of which about one-tenth is currently ice covered (King 1969). Situated just to the north of the Antarctic Convergence, the island experiences low temperatures ($\bar{x} = 4.6°$ C), frequent frosts, strong westerly winds and extensive cloud cover (Weyant 1967).

Small-scale sorted patterned ground, particularly sorted stripes, has been reported from numerous parts of the main island (Troll 1958, Aubert de la Rue 1969, Bellair 1959, Markov 1971, Nougier 1964, 1970). Recent work on Peninsule Loranchet (Hall 1983a and b) has detailed sorted nets, polygons (Photo 1) and stripes. Due to the more rugged terrain, stripes do not show a particularly strong relationship to the westerly winds, but some were found (Hall 1983b) orientated across a slope in alignment with a katabatic wind. Stripe width increased with altitude, but below about 120 m coarse stripes were wider than fine stripes whilst at the higher elevations fine stripes were broader than the coarse ones. Several areas of well developed sorted nets and circles were observed (Hall 1983a), and at one location stripes-within-stripes (Photo 2) were found (Hall 1983b; Photos 1 and 2). Surprisingly, gelifluction features, although not absent (Aubert de la Rue 1967), appeared to be relatively rare (Hall 1983a).

The sorting mechanism was said by Bellair (1969) to be wetting and drying of the soil but, although this process is certainly active, the role of ground freezing must be considered. The change of tripe width with altitude and the change from a dominant coarse stripe to a dominant fine stripe with increase in height both intimate frost action (Hall 1983b). In addition, the observation of clasts in the coarse borders of sorted nets and stripes with their a/b planes vertical and in the fine centres with the a/b planes upward facing, argues for ground freezing (Hall 1983b). Numerous weathering features indicative of both chemical and mechanical action are present (Hall, pers. obs.) but no data on the actual processes or their rates are available.

MACQUARIE ISLAND

Situated approximately half-way between Australia and Antarctica (lat. 54° 37'S, long. 158°54'E) this small island is currently ice-free but exposed to the full force of the dominant westerly winds. The island mostly comprises a plateau surface with little land above 350 m (Löffler et al. 1983).

Sorted polygons and stripes, and solifluction features, particularly terraces, have been recorded (Mawson 1943, Bunt 1954, Taylor 1955, Colhoun & Goede 1974, Löffler et al. 1983). Bunt (1954) ascribed the sorted stripes to the action of needle ice and water action and noted that wind currents may well influence the freezing pattern of the ice needles. Taylor (1955), discussing the formation of terraces, observed that different forms occurred on windward and leeward slopes but that both were associated with the establishment of *Azorella selago* which acts as a movement inhibitor with respect to continued downslope transport of material. Löffler et al. (1983) point out, however, that periglacial solifluction was also of great importance in the formation of the terraces and that the large terraces are relicts of a former, colder climate. A very strong relationship between vegetation, slope processes, and wind exposure was noted (Löffler et al. 1983: 234-235).

SOUTH GEORGIA

This long, narrow island (160 km by 36 km), with its central chain of mountains dissected by glaciers, experiences a cold, damp climate (King 1969). The islands position (lat. 54°20'S, long. 36°40'W) locates it as sub-Antarctic but it does, nevertheless, experience a dynamic periglacial environment around the margins of the extensive ice cover.

Sorted stripes, nets, and polygons, non-sorted circles, stripe hummocks, solifluction lobes, stone-banked lobes and frost boils have all been reported (Smith 1960, Clapperton 1971, Stone 1974 and 1975, Clayton 1977, Thom 1981, Walton in Laws 1984). In addition, extensive weathering, particularly by frost action, has also been recognised as a major process below c. 700 m (Stone 1974, Gordon 1985). Weathering is said to have been active throughout the Holocene, to have been abetted by the structural controls of rock bedding and jointing, and to have possibly been aided by dilatation subsequent to deglaciation (Gordon 1985: 49). The debris resulting from weathering is seen as a major factor in landform development and a prime source of material for transport, and subsequent deposition, by glaciers (Gordon 1985; Gordon & Thom, in press).

A strong relationship between patterned ground and vegetation has been recognised (Heilbronn & Walton 1984a and b, Walton & Heilbron 1984). Certain grasses and herbs colonise sorted stripes first and their survival is related to both tiller and root morphology (Heilbronn & Walton 1984a). The rate of downslope plant movement is a function of plant size, and certain plants, notably *Phleum alpinum*, stabilise the soil sufficiently to enable other species to begin colonisation. Differences in microtopography resulting from periglacial activity also generate floristic differences between stripe crest and trough (Walton in Laws 1984). There is a correlation between the effects of frost heaving and plant mortality, and needle ice also plays a role (Heilbronn & Walton 1984a, Walton & Heilbronn 1984, Walton in Laws 1984). In fact, except for the work undertaken on Signy Island in the South Orkneys (see below), the greatest depth of multidisciplinary periglacial studies undertaken on any of the southern islands has probably been achieved on South Georgia.

SOUTH ORKNEY ISLANDS

Situated in the Maritime Antarctic (lat. 60°30′S, long 44°25′-46°10′W) this group consists of two large islands, Coronation and Laurie, and two smaller ones, Powell and Signy. With the exception of Signy Island the islands are almost totally ice-covered. Unlike the sub-Antarctic islands discussed above, the flora of the South Orkneys is very limited and predominantly comprises lichens.

Chambers (1966a and b, 1967, 1970) detailed the patterned ground of Signy Island, describing sorted and unsorted circles, sorted stripes, and stone streams. Solifluction was considered the main agent of downslope movement on Signy Island (Chambers 1967) and forms suggesting solifluction activity were observed on Coronation Island (Hall, pers. obs.). Chambers (1970) found that a disturbed miniature sorted pattern would re-establish itself within three years. In Moraine Valley, on Signy Island, large sorted circles were observed (Hall, pers. obs.) that indicate downslope movement subsequent to their original formation such that they now have an almost 'barchanoid' form.

Weathering of the rock (Photo 2) on these islands was first reported by Grange (in Dumont D'Urville 1841-54) and recently a programme to consider the role of weathering in the Maritime Antarctic environment of Signy Island were initiated (Walton & Hall, in press). Walton (1985) has described the mechanical disruption of quartz-micaschist by lichens, whilst Hall (1986a, b and c, subm. a, b, c and d, and Hall et al. 1986) has detailed the rock water content and chemistry, rock engineering properties affecting weathering, rock fracture mechanics, a series of laboratory experiments on the freeze-thaw weathering of the local rock, and the results of ultrasonic investigations of the freezing mechanism. This programme probably constitutes the largest long-term study of mechanical, chemical and biological weathering in any periglacial environment (Walton & Hall, in press).

SOUTH SHETLAND ISLANDS

This group of mountainous, extensively-glaciated islands (lat. 61°-63°30'S, long. 53°30'-62°45'W) includes Elephant and Clarence Islands, Livingston Island, Deception Island, King George Island, as well as numerous small islets (King 1969). Plant life is scarce with lichens and mosses predominating.

Sorted polygons and stripes together with solifluction lobes have been reported from Elephant Island (Allison & Smith 1973) and from Deception Island (Corte & Somoza 1957) whilst Simonov (1977) describes King George Island as comprising a 'periglacial landscape'. Olsacher et al. (1956), Araya & Herve (1966) and Corte & Somoza (1957) all report the role of frost weathering on rocks. Other periglacial features described include mud streams and palses (Araya & Herve 1972), erosion by water resulting from snowbank-melt and erosion pavements due to wind and water action (Corte & Somoza 1957), permafrost, nivation and 'biogenous' weathering (Simonov 1977). The role of plant life is stressed, with Simonov (1977) noting its effects as a weathering agent and Corte & Somoza (1957) and Araya & Herve (1966) describing the relationship between sorted structures and vegetation cover plus the role solifluction plays in affecting plant growth on slopes.

SUMMARY

Overall it seems that, compared to the northern hemisphere, much of the work on the southern islands is still in its infancy with little emphasis on highly detailed or long-term studies. Nevertheless, there does appear to be a wealth of information from this very important area, much of which still remains relatively unknown to many researchers working in Arctic regions. The islands are extremely important for they comprise the only land areas within the vast Southern Ocean from which detailed information on the relationship between climatic change, geomorphic processes, plant succession and distribution, and biological activity can be obtained. Due to their isolated and frequently exposed positions, the large distances between island groups (affecting plant dispersal), and their particular climatic conditions (notably tne prevailing westerly winds and the overwhelming oceanic influence), the islands might be considered to experience an almost 'special' type of periglacial environment. This may be all the more so in terms of the relationship between periglacial processes and botanical activity as the islands are notably species-poor. In turn, this leads to the realisation of the role periglacial activity plays in the plant colonisation that takes place immediately upon deglaciation, something which botanical studies may not have fully recognised. Thus, the periglacial realm is one which cannot be ignored as it is so intimately part of the past and present environment of this region.

Overall, these wonderful islands suggest an exciting prospect for periglacial research. Sadly, though, this is not a field of study undertaken to any large extent by most Antarctic Treaty nations and, unlike in much of the Arctic and alpine regions, logistical constraints necessitate working through a national Antarctic programme. Until some detailed work on these widely separated locations is initiated, there remains a gap in our knowledge of the periglacial environment of an important part of our world.

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Some Observations Regarding Sorted Stripes, Livingston Island, South Shetlands

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ABSTRACT

Sorted stripes have long been recognized in the sub-Antarctic and the maritime Antarctic. The cold and wet conditions of these regions, when compared with the drier continent, are ideal for their development. In particular, the frequent low-amplitude, short-duration freeze–thaw cycles of the sub-Antarctic favour the development of miniature sorted forms. The maritime Antarctic, however, with its larger amplitude and longer duration of freeze–thaw cycles, plus the presence of permafrost, exhibits a more frequent occurrence of the larger forms.

The Byers Peninsula has an extensive range of sorted features but, despite being situated in the maritime Antarctic and within a permafrost zone, the majority of these features are of the miniature variety. In addition, the sorted stripes exhibit a distinct spatial preference. Their alignment and spatial distribution are not associated with either the sun or the wind, but are a function of the snow cover.

RÉSUMÉ

Des sols striés triés ont été reconnus depuis longtemps dans l'Antarctique maritime et le domaine subantarctique. Les conditions froides et humides de ces régions sont, par rapport aux conditions plus sèches du continent, idéales pour le développement de ces sols structurés. Les gels fréquents de faible amplitude et de courte durée favorisent l'apparition de formes trieés miniatures. Dans l'Antarctique maritime, les gels de plus grande amplitude et de plus longue durée déterminent la présence d'un pergélisol; les sols structuraux y ont des dimensions plus grandes.

La péninsule de Byers présente une variété considérable de phénomènes triés, mais bien qu'étant située dans l'Antarctique maritime et dans une zone de pergélisol, la majorité des sols structuraux périglaciaires y sont de petite taille. En outre, les sols striés triés montrent une répartition particulière; leur alignement et leur distribution spatiale ne sont associés, ni avec le soleil, ni avec le vent, mais ils sont en relation avec la couverture neigeuse.

KEY WORDS: Sorted Stripes Formative Process Snowcover Antarctic

INTRODUCTION

Sorted stripes are 'patterned ground with a striped pattern and a sorted appearance due to parallel lines of stones and intervening stripes of finer material orientated down the steepest available slope' (Washburn, 1956, p. 36). Miniature forms of sorted stripes are those with a

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stripe width ≤20 cm. Despite the frequent occurrence of these features (both small and large) in periglacial areas there is still controversy regarding their origin. For stripes in general, the question involves both the cause of sorting, although explanations for this can be found (e.g. Corté, 1966; Mackay, 1984; Anderson, 1988), and an explanation for their spatial organization (e.g. Krantz, 1990; Hallet, 1990; Werner and Hallet, 1993). Hypotheses regarding the cause of stripe development are varied, and their origin is probably polygenetic (Washburn, 1979). In some areas wind plays a role in orienting the stripe (e.g. Beaty, 1974; Hall, 1979; 1983); in others, vegetation depletion (Brockie, 1968; Wilson and Clark, 1991), rill development (Brockie, 1968; Bunt, 1954) or pre-existing corrugated topography (Muir, 1983) can play a role. Other studies in New Zealand and Canada show that stripes align parallel to the early morning sun (Mackay and Mathews, 1974a; 1974b). These hypotheses, while helping to explain stripe development, do not offer an answer to their spatial organization, i.e. what produces the regularity of coarse and fine stripes. Explanations for this regularity rely mainly on convective or helical flow theories (e.g. Dżułyński, 1966; Krantz, 1990; Ray *et al.*, 1983; Warburton, 1987; Gleason *et al.*, 1989). However, although convective flow can explain the regularity of spatial organization there is the question regarding the efficacy of these mechanisms to actually move the clasts to effect the sorting (Van Vliet-Lanoe, 1989). Thus, questions regarding both origin and spatial organization still need to be resolved.

Sorted stripes, like other forms of sorted patterned ground, have been used as palaeoclimatic indicators. For instance, Goldthwait (1976) suggests that all features with a mesh spacing >2 m are indicative of ice-rich permafrost. Such an environment is then said to be characterized by mean annual air temperatures in the region of −4 to −6 °C and to have moderate or 'adequate' moisture supply. Conversely, smaller sorted forms are said to develop (p. 34) in a 'less vigorous climate ... where ... no permafrost is found by trenching in late summer'. Mean annual air temperatures for miniature forms are said to be between +5 and −2 °C. In much the same way, Wayne (1983) argues that large (2 m) sorted stripes require, in a dry alpine region, an air temperature of −3 to −15 °C whilst small sorted stripes require 0 to −3 °C. By this reasoning, the finding of fossil forms of certain sizes facilitates the determination of palaeo-conditions.

This paper presents evidence that offers another explanation for the occurrence of stripes and also throws doubt on the assumption that small forms are associated with non-permafrost areas.

STUDY AREA

The field observations were undertaken on the Byers Peninsula at the western end of Livingston Island (Figure 1) in the South Shetlands Islands (62°40′S, 61°W) during January and February 1991. This is the largest ice-free area in the South Shetlands and is underlain by

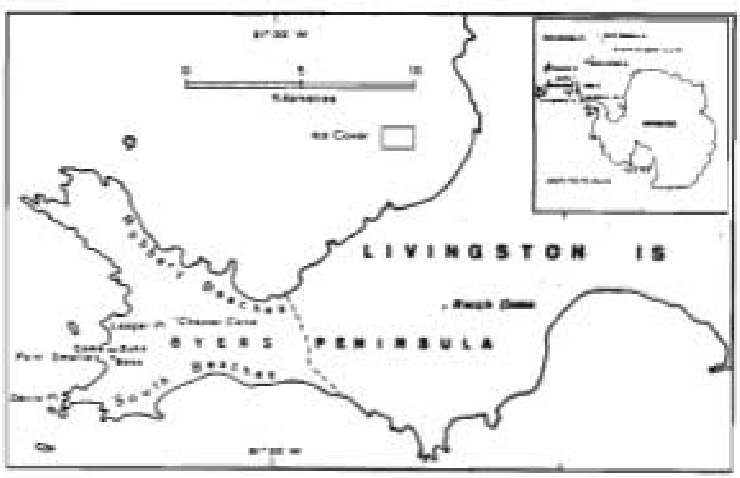

Figure 1 Location of study area.

permafrost with an active layer of between 0.3 and 0.7 m depth (Thom, 1978). The mean annual air temperature is between −3 to −5 °C but during the December to February period the mean daily temperature is above freezing; there are large differences between winter and summer temperatures. Annual precipitation is of the order of 100 to 150 cm water equivalent, with much of this falling as snow from February through to November. In accord with the frequent precipitation, cloud cover is extensive and the amount of direct solar radiation received is limited. The relative humidity is high, usually above 65%, and winds are frequent, with a mean velocity between 40 and 50 km h^{-1}.

While there have been many periglacial studies in the South Shetlands (Araya and Hervé, 1966; Blümel, 1986; Blümel and Eitel, 1989; Corté and Somoza, 1957; Dutkiewicz, 1982; Olsacher, 1956; Simonov, 1977; Stäblein, 1983; Thom, 1978; Vtyurin and Moskalevskiy, 1985) most have centred on King George Island owing to its easy accessibility and the large number of national research stations located there. A variety of periglacial landforms and processes have been reported in the above-mentioned studies. Sorted stripes, circles and polygons are relatively common, as too are solifluction and nivation. Both mechanical and chemical weathering are said to occur. Many of the processes are the result of the large winter snowfall, which ablates slowly during the summer, together with the frequent low-amplitude freeze–thaw cycles that are said to characterize this region.

OBSERVATIONS AND DISCUSSION

A total of 217 observations regarding sorted stripe orientation, stripe widths and length, and slope angles were obtained from the Byers Peninsula (Table 1). Despite the mean width of only 5.6 cm for the coarse stripe, the average *a*-axis length of the spindle-shaped clasts within the stripe was 4.3 cm. The clasts were aligned with their long axes parallel to the stripe.

It was with respect to stripe orientation that a distinct preference was observed. Analysis of the circular orientational data by means of the von Mises distribution indicated a mean vector of 336° with two-thirds of the observations situated within 336° ± 63° (Figure 2). Graphically, the data clearly show a preference for the

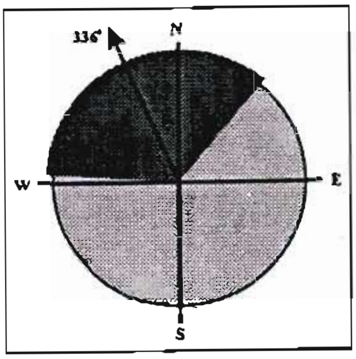

Figure 2. Circular graph showing the results of the von Mises test with the resultant vector (336°) and ± 63° area in which the observations are concentrated.

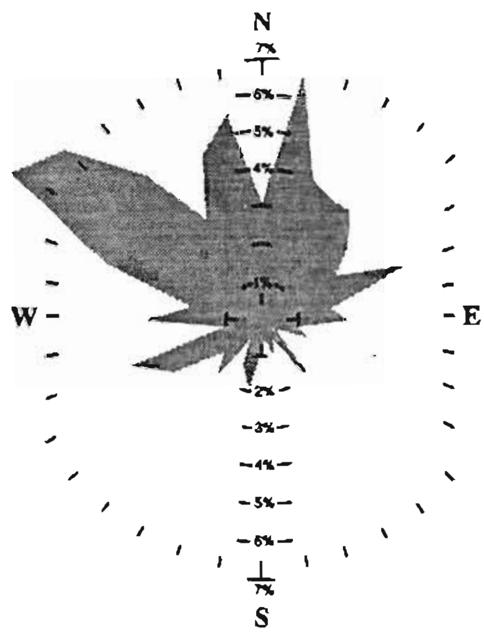

Figure 3. Rose diagram to show stripe orientation distribution.

Table 1 Details regarding sorted stripe observations, Byers Peninsula, Livingstone Island.

	n	$\bar{x}$	s	Min.	Max.
Slope angle (°)	217	7.7	4.7	2	26
Coarse stripe width (cm)	217	5.6	2.5	2	20
Fine stripe width (cm)	217	7.4	3.8	3	35
Stripe orientation (°)	217	336	63		

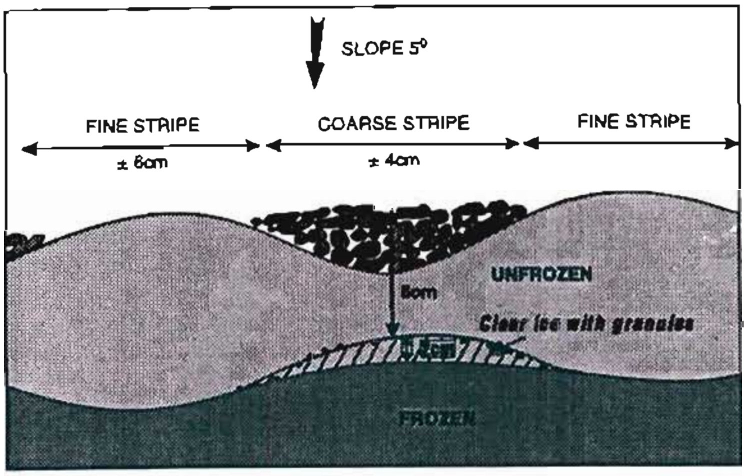

Figure 4 Detail across a sorted stripe to show the location of the clear ice with granules (as seen on 16 January 1991).

northern and north-western sectors with a paucity of observations in the southern and south-eastern sectors (Figure 3). In many instances these southerly slopes were carefully checked for any signs of sorted patterned ground but, despite the slope angles being similar ($\bar{x}$ = 8°), the material being the same and moisture being present, none could be found.

Upon sectioning the stripes as they emerged from beneath the ablating snow cover, it was observed not only that the surface of the underlying frost table was furrowed, as has often been observed before, but that there was an unusual layer of ice on the ridge top beneath the coarse stripe. This ridge of ice, with a maximum thickness of about 3 cm, appeared clear but with inclusions of granular material (Figure 4). Visually, it seemed distinct from the ice of the frozen layer beneath and was only found on the ridge below the coarse stripe. No evidence of the former existence of this ice could be seen further up the slope where melting had taken place. Thus, this appears to be an opportunistic observation for which there is no obvious sedimentological or morphological expression. It may, however, give some indication as to formative processes.

With respect to the stripe orientation, all ridges and upstanding blocks had substantial lee-side accumulations of snow. Because the dominant winds are north-north-westerly, the southern and south-eastern sides have substantial lee-side accumulations, with drifts many metres thick, only exposing the ground beneath by mid January. At the start of January, some 60% of the peninsula is still snow-covered. It becomes snow-free by mid February, shortly after which (i.e. the end of February) snow begins to fall once more. Thus, the southern and south-eastern aspects are covered by a thick layer of snow for the greater part of the year. Although the ridge tops were the first to become snow-free, this was still late in December or early into January 1991. Significantly, during the January to February period of 1991 no freeze–thaw cycles greater than −1 °C were recorded, and even then very few and only late in February (Table 2).

Thus, slopes with southern and south-eastern aspects are insulated from freeze–thaw action by the thick snow cover and, by the time this melts, temperatures are above 0 °C. Therefore, very limited sorting takes place on these slopes despite favourable material and moisture conditions. Sorted patterns are only found on the ridge tops which become snow-free sufficiently early in spring to be exposed to freeze–thaw cycles. Most sorting takes place on the windward slopes where snow accumulation is less and where ablation is enhanced by the effects of direct radiation; the southerly aspects are protected from direct radiation. Further, the sorting that takes place on the northerly slopes is constrained to the early autumn and late spring, i.e. prior to the slope becoming snow-covered and subsequent to disappearance of the snow cover. As mentioned earlier, there is a marked difference between summer and winter temperatures in early autumn and late spring. As a result, the temperature trend is, respectively, towards and away from large-amplitude freezes. Thus, the ground is not exposed for very long, considering the extent and depth of the snow cover. As a consequence, the ground does not undergo large-amplitude freezing but experiences, while in a wet state, diurnal freeze–thaw events. The annual freeze is the only one of any magnitude to affect the ground. Thus, despite the presence of permafrost and air temperatures that would commonly be indicative of large sorted patterned ground, most of the patterned ground is of the miniature variety.

In other areas of the sub-Antarctic (e.g. Marion Island and Iles Kerguelen) miniature sorted stripes have been shown to be aligned to the dominant wind direction (Hall, 1979; 1983). Others (e.g. Mackay and Mathews, 1974a; 1974b) working in different areas (Canada and New Zealand) have shown that needle-ice striped ground is aligned parallel to the early morning sun. On Byers Peninsula, however, stripe orientation is a direct result of longevity of snow cover and so this factor must be taken into account when considering the spatial distribution of sorted stripes. More importantly, despite this being a permafrost area and one which experiences a substantial number of freeze–thaw cycles in the air, the dominant stripe form is miniature. Considering the wet nature of the ground, and its surficial till cover, large stripes would probably develop were it not for the snowcover protecting the ground from freeze–thaw events. This conclusion could be important for palaeoenvironmental reconstructions since miniature stripes are usually associated with non-permafrost, low-amplitude freeze–thaw environments. Inversely, some authors (e.g. Wayne, 1983) have used large fossil sorted stripes as a guide to Pleistocene temperatures and, for a dry alpine region, have differentiated between conditions necessary for their formation and that of miniature stripes. Although the Byers Peninsula is clearly not a dry area, the principle still remains that using size alone as a diagnostic criterion of former conditions can be erroneous.

With respect to the cross-section of the stripes, it is the typical capping of ice with granular inclusions on top of the ridge below the coarse stripe that is of interest. Although Sharp (1942) noted variations in freezing and thawing of stony borders in sorted circles and

Table 2 Simplified temperature data for the Byers Peninsula, January/February 1991.

	6–7 January	8–14 January	15–21 January	22–28 January	29 January to 4 February[1]	5–11 February	12–18 February[1]	19–25 February[1,2]
Max. air temp. (°C)	5.28	6.5	6.6	5.6	5.9	5.3	5.29	2.1
Min. air temp. (°C)	1.52	0.94	0.2	0.7	1.37	0.59	1.4	0.44
Max. ground surface temp. (°C)	16.8	20.6	22.34	21.2	15.68	20.3	9.5	1.1
Min. ground surface temp. (°C)	0.87	0.09	−0.87	−0.5	1.2	−1.3	0.83	0.0
Freeze–thaw cycles in air	0	0	0	0	0	0	0	0
Freeze–thaw cycles in ground	0	0	2	1	0	2	0	0

[1] Incomplete data for this week.
[2] Ground covered by snow for much of the week.

Mackay (1980) discussed the presence of thaw depressions in the active layer of non-sorted patterned ground, this feature has never been previously reported in the context of sorted stripes. The nearest parallel is that of the thawing of saturated fines surrounded by openwork gravel that, on Cornwallis Island (Canadian high Arctic), results, after several days at the start of the thawing regime, in a bowl-shaped depression in the active layer of the central vegetation-free silty soil (Washburn, 1991; personal communication, 1994). Qualitatively, the granular inclusions appeared to be more abundant at the lateral borders of the ice compared with the central top part. The ice is transparent, stands some 2 to 3 cm above the frozen ground beneath, and is up to 6 cm wide. No sedimentological or morphological expression of this ice is apparent upslope in the stripes where it had (presumably?) melted.

The origin of sorted patterned ground, and sorted stripes in particular, has been a matter of controversy and speculation for most of this century (e.g. Dżułyński, 1966). While mechanisms to explain the vertical and horizontal movement of soil have been available for some time (e.g. Corté, 1966) it is only recently that attention has once again focused on attempts to explain the spatial organization of the patterns (e.g. Werner and Hallet, 1993). To this end, convective motion with a number of possible driving mechanisms has been proposed. Although the basic arguments for the development of sorted patterns from convective-like flow have been suggested since at least the turn of the century (Nordenskjold, 1909), they have recently been rejuvenated by a number of workers. Some, such as Dżułyński (1966), consider too much emphasis has been placed upon temperature- and moisture-controlled density currents and would rather associate convection with unstable stratification subjected to freezing. Others, such as Krantz (1990), prefer free convection of water through the thawed soil as the mechanism. A review of the convective arguments has been presented by Van Vliet-Lanoe (1989). Other non-convective hypotheses utilize the role of running water: Brockie (1965) suggests stone accumulation in previously water-eroded rills, and Taylor (1955) the removal of fines by water flowing within the coarse stripe. Further, in the context of sorted stripes, Werner and Hallet (1993) present a two-dimensional computer simulation to explain the development of sorted stripes by needle ice. All in all, the number and range of hypotheses are large. As much as these ideas constitute significant steps forward, one is still confronted by the comment of Albert Pissart (1990, p. 127): 'We need more observations in order to definitely explain the genesis of such structures because the observations made to date do not seem sufficient to define the causes of the movement.'

The observations presented here add another piece to the jigsaw puzzle, although it is not yet certain what that piece is. The problem is that the presence of the ice beneath the coarse stripe may be not a *cause* but rather an *effect*. Considering effect first, the initial impression was that the ice resulted from water flowing down the coarse stripe when it was frozen, and subsequently freezing. In other words, the unfrozen, void-rich coarse stripe channels meltwater from upslope and the water maintains the coarse stripe by removing fines (hence the granular inclusions). Such action is a *consequence* of the presence of the stripe rather than its cause. Also, it is not clear how water concentrates along the ridge 'crest' of the frozen substrate except as a result of the fines freezing first or thawing last, thereby channelling the water along the coarse stripe (Figure 5). Other questions that remain include the morphological form of the ice (i.e. why should it drape over a frozen ridge if it also developed in a trough?) and the concentration of granular material at the margins (a function of the flow regime?). So, the immediate response has a number of major questions that need resolving.

With respect to cause, Krantz (1990), in his

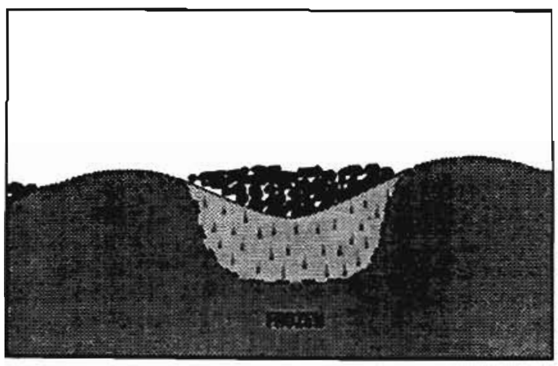

Figure 5 A possible model to explain an unfrozen trough below the coarse stripe along which water could run during the freeze period (but prior to the area under the coarse stones becoming frozen).

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discussion regarding convection cells as a result of unstable moisture density stratification during the thawing of frozen soil, identifies several possible mechanisms whereby sorting can occur. Amongst these he describes 'cryogenic pressure generated during freezing when the descending ice front traps water in the troughs of the underlying corrugated thaw front'. Therefore, during freeze, the corrugated trough may be similar to that depicted in Figure 5. If so, the observed ice may be associated with the 'trapped water'. Interestingly, Krantz shows a *depression* below the coarse stripe, whereas in the Byers Peninsula stripes the observed corrugated surface has a depression below the fine stripe. In the freeze model depicted in Figure 5 the depression *would* be beneath the coarse material. Although it is not clear whether any form of cellular or helical flow takes place, the Byers Peninsula is a location where the active layer *is* very wet. Thus, according to the criteria of Krantz, this is an area where the potential for this form of sorting to occur is great. Nevertheless, the ability of convective mechanisms to effect sorting is still questionable.

Another possibility is that the clear ice represents an ice lens, the thawing of which has been differentially retarded because of the overlying gravel (Washburn, personal communication, 1994). The convex shape could then be explained by its crest underlying the thickest gravel. The bowl-like depression in the frost table is a typical feature below plugs (Washburn, 1991). Unfortunately, no analysis of the nature of the ice was undertaken and it is not possible to test this hypothesis. Also, if this were the case, quite what it means with respect to stripe development is not clear.

CONCLUSIONS

With respect to formative mechanisms, a potentially valuable observation is described in which ice is concentrated below the coarse stripe. The spatial distribution and size of sorted stripes is clearly associated with the longevity of snow cover and the nature of the freeze–thaw cycles to which the ground is subject. The association with snow cover might offer clues as to the reasons for the ice layer. The occurrence of miniature sorted stripes in an area that would normally be expected to support large forms indicates that great care should be taken in palaeoenvironmental interpretations based on the size of such features.

Ultimately, there is a need for more field observations in order to test theoretical models. This suggests the need for fieldwork during early spring and late autumn.

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XV INTERNATIONAL CONGRESS

3 – 11 August 1999
Durban, South Africa

THE ENVIRONMENTAL BACKGROUND TO HOMINID EVOLUTION IN AFRICA

BOOK OF ABSTRACTS

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PRESENT AND QUATERNARY PERIGLACIAL PROCESSES AND LANDFORMS OF THE MARITIME AND SUB-ANTARCTIC REGION: A REVIEW

As part of a new initiative by the International Permafrost Association, a review is being undertaken of present and past periglacial processes and landforms of the Southern Hemisphere. Here details are provided regarding available information pertaining to both the Maritime and sub-Antarctic region. All the islands within this broad zone have experienced glaciation during the Quaternary and many still support ice caps, of varying dimensions, today. These islands currently experience strong westerly winds and high precipitation receipts. Temperatures are less severe than on the continent, with lower values to the south of the Antarctic Convergence and higher to the north. All locations do, however, experience conditions conducive to active periglacial processes and landforms, some of the more southerly locations exhibit discontinuous permafrost. Three reviews have provided a broad-brush outline of information pertaining to the present landforms for the Antarctic region. The aim here is to give a more up to date account, plus to review the information regarding relict features which are indicative of past, more severe, conditions. Although studies in the Antarctic are, in general, less sophisticated than their Arctic or alpine counterparts, nevertheless there is a extensive body of information that is not well known by most Northern hemisphere workers. A wide range of landforms are currently active, including such as solifluction, sorted patterned ground, stone-banked lobes, cryogenic weathering, and nivation. Significantly, the strong winds found throughout this region impact upon landform development, particularly the orientation of some features, to a degree far greater than is recorded for the Northern hemisphere. Here, animals also exert a significant impact within the periglacial regime and comment is made regarding this.

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RELATIVE CONTRIBUTION OF DIFFERENT GRAIN SIZE FRACTIONS IN LOESS AND PALEOSOL TO MAGNETIC SUSCEPTIBILITY

Magnetic susceptibility is increasingly used as a climatic proxy in the study of Quaternary paleoenvironment. The magnetic enhancement of paleosol is observed in the Chinese loess. The origin of this magnetic enhancement is still uncertain. It is, however, a key problem to understand the paleoclimatic significance of changes in magnetic susceptibility, and to transfer the magnetic signals to paleoclimatic parameters. Two main models have been proposed to explain the mechanism of magnetic enhancement in the paleosol, a depositional and a pedogenic model. Together with composition and concentration, grain size distribution of magnetic minerals also plays an important role in the magnetic enhancement of paleosol. Systematic susceptibility measurements are carried out on the samples of the upper part (S0 to top of L2) of three loess sections from Xining, Xifeng and Jixian along the west-east transect in the loess plateau, China. The samples with the highest value of magnetic susceptibility in S1 and the lowest value in L2 of each section are selected as the representatives of three pairs of paleosol (S1) and loess (L2). These representatives were separated into different grain size fractions based on Stokes' law for coarse grains and by centrifuge for fine grains. Measurements of magnetic susceptibility and mass have been carried out on these fractions. Results show that for the loesses magnetic susceptibility changes little in the fractions with different grain size and for the paleosols it increases with decrease of the grain size. The magnitude of changes is bigger in the east (Jixian and Xifeng) than that in the west (Xining). The fraction with the finest particle size in paleosols does not show very high magnetic susceptibility value. It is different from previous work. Taking the mass into consideration, we find a good approach to distinguish the contribution of each fraction to total magnetic susceptibility. The contribution comes mainly from coarse grains (>10 micron) for loess samples. It reaches about 75-[illegible]%. The main contribution comes, however, from the particles with medium size (10-0.5 micron). The very fine grained particle (<0.3 micron), which is considered to be with the pedogenic origin, contributes little to the total magnetic susceptibility, no more than 3% because of their very little amount. This approach provides a sounder basis for the study of the origin of the magnetic susceptibility enhancement in paleosol and of the paleoclimatic significance of magnetic susceptibility of loess and paleosol.

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THE RECONSTRUCTION OF LATE QUATERNARY CLIMATES IN SOUTHEAST AUSTRALIA USING MODERN POLLEN ANALOGUE ANALYSIS

Refined reconstructions of climate in southeast Australia over the last 200 000 years have been made using modern analogue analysis of pollen records from three terrestrial sites and one marine core. Chronologies for the records were established using uranium/thorium disequilibrium and AMS radiocarbon dating techniques and in the case of the marine record, oxygen isotope stratigraphy. The reconstructions indicate that precipitation levels fluctuated within the Holocene and the previous two interglacials. Conditions were generally wetter during the Last Interglacial than during the Holocene, with maximum annual precipitation being up to 300 mm higher. Mean annual precipitation within the Penultimate Interglacial fluctuated between levels similar to those of today and up to 200 mm higher. Driest conditions were recorded during the glacial phases, with precipitation falling as low as 50% of present day levels. The evidence also suggests that seasonality of rainfall varied throughout the last 200

Present and Quaternary Periglacial Processes and Landforms of the Maritime and sub-Antarctic Region: A Review.

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The many isolated islands of the sub- and maritime Antarctic experienced varying degrees of glaciation during the last glacial. Some such as the Falkland Islands or Macquarie Island sustained only a few cirque glaciers while others such as Bouvet Island or Heard Island were totally ice covered. Others, such as Marion Island and Iles Kerguelen had an extensive but not total ice cover. Thus these islands provide a spectrum of locations that have experienced and/or are experiencing a cryogenic environment of varying severity for a varying period of time. As a result, these islands abound in periglacial landforms, both active and inactive, associated with both permafrost and seasonal freezing.

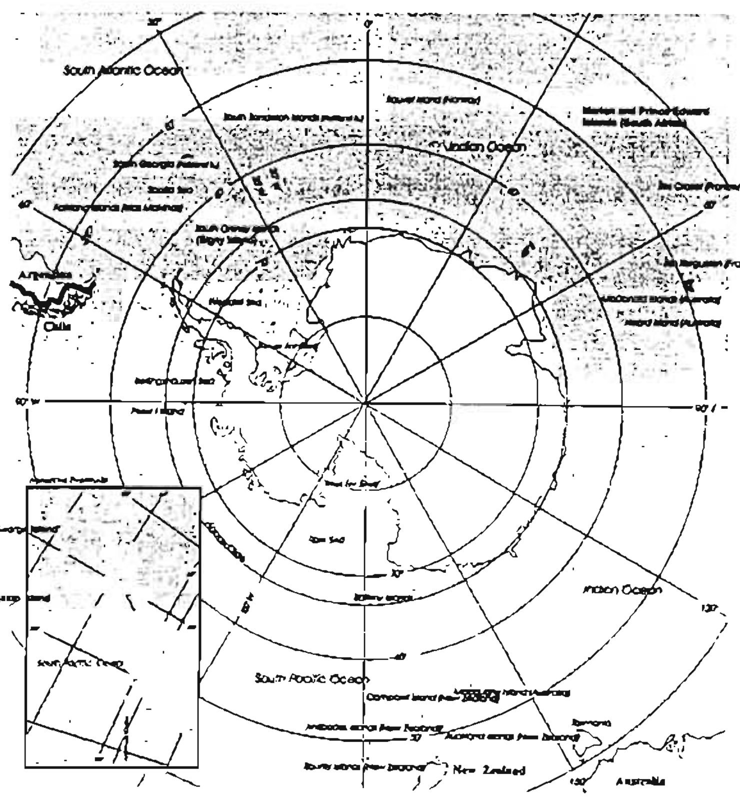
Climatically, all of the islands, away from close proximity to the continent, are affected by a continuous stream of eastward moving cyclonic depressions. These bring humid air masses, but more temperate, humid air from the north or colder, drier air from Antarctica can also periodically affect the islands.

Climate is severely influenced by location of the individual island with respect to the adjacent land or sea masses as well as its position with regard to the Antarctic Convergence.

Land masses can produce a rain shadow effect, as in the case of South America with respect to the Falkland Islands. Islands to the north of the Convergence (e.g. Marion and Prince Edward Islands, Crozet Islands) experience milder temperatures and less snow fall than those to it's south (e.g. Bouvet Island or South Georgia). The islands become increasingly colder with increase in latitude.

Thus, position has a very strong influence on the past and present extent of glaciation as well as the extent and degree of cryogenic activity.

Map of Antarctica to show the islands mentioned in the poster



Permafrost is found mainly on the islands closer to Antarctica although some evidence of present-day permafrost is also evident at higher elevations on the more northerly islands. In the past, during maximum glaciation, it is likely that many of the unglaciated areas were subject to either continuous or discontinuous permafrost. On some islands (e.g. Bouvet Island) the ice and snow cover is so extensive that it is impossible to know whether permafrost exists or not. A further complicating factor is that most of these islands are volcanic in origin and so are either 'hot spots' that inhibit permafrost development or the rapidly changing character of the geomorphology has lost any evidance of permafrost landforms.

It is only close to the continent where the unequivocal existance of permafrost, and a whole assemblage of associated landforms, can be found.

Landforms and climate of the sub- and maritime Antarctic

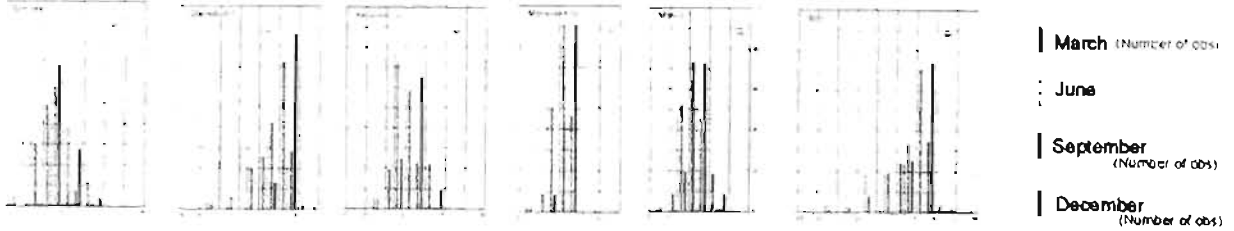
Total cloud amount

Summary of landforms and processes recorded for the sub- and maritime Antarctic

Features	M & PE	Crozet	Kerguelen	Bouvet	FI	SG	Heard/McD	Macq	S. Ork	S. Shet
Permafrost	✓?		✓?	✓	(✓)	✓	✓		✓	✓
Sorted Pat. grnd (small)	✓	✓	✓		✓	✓		✓	✓	✓
Sorted Pat. grnd (large)	(✓)		✓		(✓)	✓			✓	
Non-sort Pat. grnd						✓				✓
Solifluction	✓	✓	✓		✓	✓			✓	✓
Blockfields	✓		✓		✓	✓			✓	✓
Rock Glaciers					(?)	✓				✓
Stone-banked lobes	(✓)		✓							
Tors	✓				✓					
Scree	✓	✓	✓	✓?	✓	✓	✓	✓	✓	✓
Protalus Ramparts						✓				✓
Nivation	✓?		✓?			✓				✓
Cryoplanation		(✓)			✓					
Stone runs										
Solifluction terraces	✓	✓	✓		✓			✓		
Aeolian action	✓	✓	✓		✓			✓		
Cirque gl.			✓		✓	✓		✓	✓	✓
Glaciers/ice caps	(✓)	(✓)	✓	✓		✓	✓		✓	✓

✓ = known to be present ✓? = suggested to be present but not adequately confirmed
(✓) = evidence of existence in the past

M & PE = Marion and Prince Edward Islands FI = Falkland Islands
SG = South Georgia Heard/McD = Heard Island and McDonald Island
Macq = Macquarie Island S Ork - South Orkney Islands
S.Shet = South Shetland Islands



Frequency of occurrence of selected temperature ranges

Temperatures in degrees Fahrenheit as most original data were in this form
Conversion guide:

50F = 10.0C	20F = -6.7C	-9F = -22.8C
40F = 4.4C	10F = -12.2C	-19F = -28.3C
30F = -1.1C	0F = -17.8C	-29F = -33.9C

PERMAFROST AND PERIGLACIAL PROCESSES
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Short Communication

Some Morphometric Measurements on Ploughing Blocks in the McGregor Mountains, Canadian Rockies

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ABSTRACT

Morphology and site characteristics are described for ploughing blocks in a sub-alpine zone of the Canadian Rockies. The ploughing block features (blocks, troughs, levees, etc.) are generally larger than those described for other parts of the world. Good correlations were found between block size and both depression size and mound height. A number of ancillary observations and measurements suggest active block movement in the study area. Copyright © 2001 John Wiley & Sons, Ltd.

RÉSUMÉ

La morphologie et les caractéristiques des sites où se trouvent des blocs labourant sont décrites dans une zone sub-alpine des Rocheuses canadiennes. Les caractéristiques de ces blocs labourant (taille des blocs, importance des sillons, levées, etc.) sont généralement plus grandes que celles décrites dans d'autres régions du monde. De bonnes corrélations sont trouvées entre la taille des blocs et, à la fois, la dimension des sillons et la hauteur des levées. De nombreuses observations accessoires et des mesures suggèrent que des mouvements de blocs se produisent actuellement dans le secteur étudié. Copyright © 2001 John Wiley & Sons, Ltd.

KEY WORDS: Canadian Rockies; morphometry, ploughing blocks

INTRODUCTION

Ploughing blocks (also called 'Wanderblöcke' or braking blocks) are a form of mass movement whereby a rock moves downslope faster than the surrounding material (Washburn, 1973; French, 1996, p. 156). This movement results in a 'mound' on the downslope side of the block and a 'depression' to its rear (Tufnell, 1972). First recognized in Scandinavia (Sernander, 1905; Rekstad, 1909; Högbom, 1914), ploughing blocks are reported for many high altitude and/or latitude environments, as summarized by Tufnell (1972), Reid and Nesje (1988) and Ballantyne and Harris (1994). Tufnell (1969) has claimed that ploughing blocks are amongst the most widespread of currently developing periglacial phenomena, and Ballantyne and Harris (1994, p. 216) state that they are 'the most common manifestations of recent periglacial activity in Britain'. Despite these assertions, ploughing blocks have received little serious attention. Most studies focus on the uplands in Britain (see Ballantyne and Harris, 1994 for a detailed summary) and describe the morphology of the various components of the ploughing blocks (e.g. Tufnell, 1972; Chattopadhyay, 1983; Wilson, 1993; Ballantyne and Harris, 1994; Allison and Davies, 1996).

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Equally, the known movement rates are few and primarily measured at sites in north-western Europe (e.g. Johnson and Dunham, 1963; Tufnell, 1972; 1976; Shaw, 1977; Chattopadhyay, 1983). Outside Europe, Gorbunov (1991) provides movement rates for ploughing blocks in the Tien Shan. In North America, ploughing blocks have been reported by Antevs (1932) for the Mt Washington Range (Maine), while Lyford *et al.* (1963) identify fossil forms in Massachusetts. Surprisingly, no other observations could be found in the North American literature. This study provides the first observations on ploughing blocks in the sub-alpine zone of the McGregor Mountains, Canadian Rockies.

SITE AND METHODS

The study site is located on a north-facing slope in the sub-alpine zone of the McGregor Mountains (54°14′N, 120°50′W) at an altitude of 1660 to 1780 m. The ploughing blocks occur on the slope lateral to, and immediately downslope of, a north-east-oriented cirque lake (Figure 1). The regional geology is one of folded and faulted sedimentary and metasedimentary rocks, mostly Palaeozoic in age, composed primarily of limestone, quartzite, schist and shales (Valentine *et al.* 1978). Fifty ploughing blocks, most of them limestones, were investigated in this study. Vegetation consists of sub-alpine meadows, sparsely populated with clumps of Engelmann spruce (*Picea engelmannii*) within the Engelmann spruce to subalpine fire biogeoclimatic zone (Coupé *et al.* 1991). Stone-banked solifluction lobes occur in some parts of this region (Hall and Meiklejohn, 1997) but, significantly, were not found in the vicinity of the ploughing blocks. Observations indicate that this is an area of very high snowfall and that the snow cover persists late into the summer (mid to late July) with snow accumulation beginning again by early September. Thus, although winter temperatures are cold (−30°C being common and −40°C expected several times each winter for short periods), the thick snow cover inhibits substantial ground freezing.

Various parameters were measured and classified (Figure 2), following the methods and definitions of Tufnell (1972). These parameters included (i) site characteristics (slope angle, slope aspect and deviation of block orientation from maximum local slope gradient or 'alignment') and (ii) block dimensions (long (a), intermediate (b) and short (c) axis, block circumference). The extent of block burial ('block depth') and block height above the surface ('block height') were measured at the rear of the block. Depression dimensions measured included length, width and depth. Mound height at the front of the block was also measured. Depression cross-section and plan form, as well as mound elevation/plan form and shape, were classified according to Tufnell (1972). Block shape was classified following Allison and Davies (1996). However, it was found that some mounds did not have a complete vegetation cover and so frontal/lateral mound types composed of bare ground were classified into a separate category in this study (Figure 2). An additional category was defined for vegetated frontal/lateral mounds where

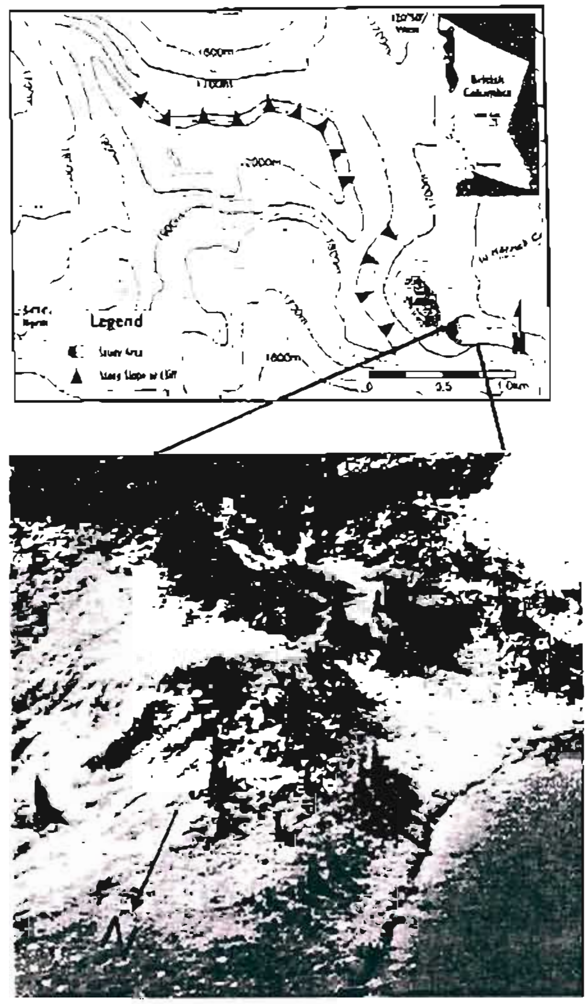

Figure 1 Map to show the location of the study site in the McGregor Mountains together with a photograph taken from a helicopter over the area of ploughing blocks (one large block and trough can be seen between the lone tree just below left centre and the north arrow).

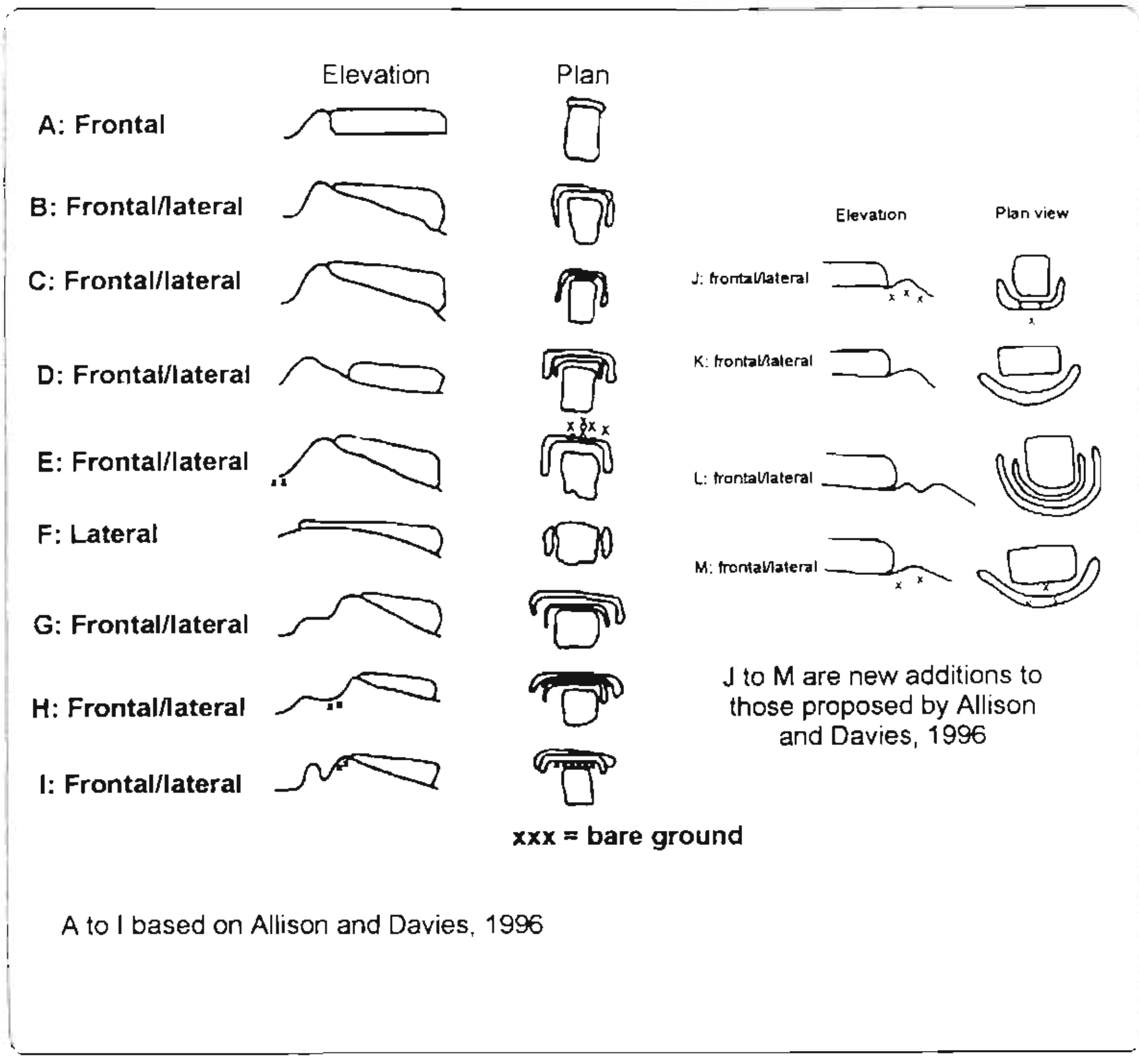

Figure 2 Mound elevation and plan, based on Tufnell (1972), as illustrated by Allison and Davies (1996) with new categories J, K, L and M based on this present study. From. Allison RJ and Davies KC. 1996 Ploughing blocks as evidence of down-slope sediment transport in the English Lake District. *Zeitschrift für Geomorphologie*. Suppl. 106: 199–217. Reproduced by permission of Gebrüder Borntraeger, Stuttgart.

the block has its *a*-axis parallel to the slope and the block rests on the surface, rather than being embedded in sediment. Two final additional categories describe a double-ridged mound with the block resting on the mound surface, and a bare frontal/lateral mound with the block *a*-axis parallel to the slope (Figure 2).

RESULTS

Results from the ploughing block measurements are summarized in Table 1. Ploughing blocks were found on slopes ranging from 1° to 34°, but predominate on slopes from 18° to 24° with a NW to NE aspect. These slope angles are similar to those on which ploughing blocks occur in Europe, but are lower than in the Tien Shan of central Asia (Gorbunov, 1991). The long axes of 82% of the blocks are oriented within 20° of the orientation of the local maximum gradient. Blocks with long-axis orientations deviating significantly from the slope orientation were found to be influenced by local obstructions, such as immobile blocks or local depressions created by other ploughing blocks.

Similar observations were also made by Allison and Davies (1996).

Block dimensions range from 0.53 × 0.44 × 0.15 m to 3.3 × 1.9 × 1.4 m ($a \times b \times c$ axes), in addition to which two exceptionally large blocks were measured at 4.9 × 4.3 × 3.3 m and 4.5 × 4.5 × 3.6 m respectively (Figure 3). Thus the blocks are significantly larger than those observed in the uplands of Britain (Wilson, 1993) and the Tien Shan (Gorbunov, 1991). Their shape is predominantly of the E-type (rectangle/square) or A-type (oblong/elongate), as defined in Allison and Davies (1996). The largest ploughing blocks found in the McGregor Mountains approximate the size of the giant ploughing block of 4.9 × 5.0 × 5.6 m reported by Reid and Nesje (1988) for southern Norway. Blocks are, on average, buried approx. 0.2 m in the soil, but depths range from 0 to 0.7 m.

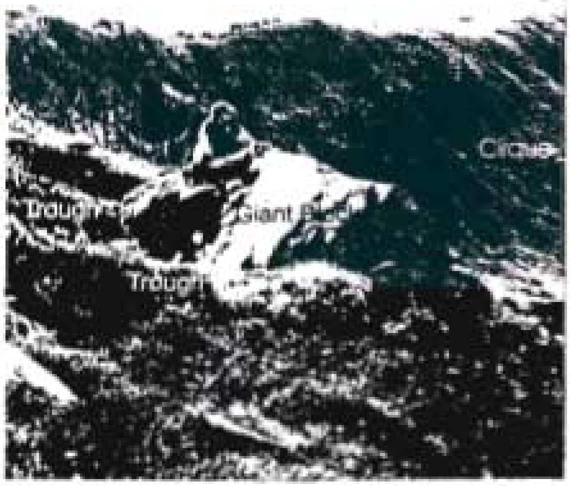

Figure 3 Giant ploughing block at the study area

Figure 4 Young fir tree bent by ploughing block movement indicating block movement followed by stabilization

Depression lengths are highly variable and range from 1.2 m to 33 m, but the 4.9 m block exhibited a depression length of 53 m. These values exceed those typical for Britain (Wilson, 1993; Allison and Davies, 1996) and the Tien Shan (Gorbunov, 1991). The same applies to the depression widths and depths measured in this study. Most depressions are of the A-type (straight) as defined by Tufnell (1972) and have a predominantly B (parabolic), C (U-shape) or

Table 1 Summary statistics of some parameters measured on ploughing blocks ($n = 50$).

Variable	Mean	Median	Min.	Max.	SD
Slope angle (°)	21.4	21.5	1	33	6.41
Alignment (°)	1.2	16	46	30	13.9
Block length (a-axis) (m)	1.6	1.3	0.5	4.9	0.93
Block width (b-axis) (m)	1.3	1.1	0.4	4.5	0.79
Block total height (c-axis) (m)	0.9	0.8	0.2	3.6	0.66
Block depth (m)	0.2	0.2	0.0	0.7	0.16
Block height (m)	0.61	0.6	0.0	1.8	0.39
Block circumference (m)	5.0	4.0	1.6	16.7	2.91
Depression length (m)	10.3	6.4	1.2	53.0	10.13
Depression width (m)	1.55	1.30	0.3	6.6	1.06
Depression depth (m)	0.36	0.30	0.1	2.5	0.35
Mound height (m)	0.42	0.38	0.0	1.7	0.31

Table 2 Pearson correlation (r-values) between various parameters, excluding two outliers (exceptionally large blocks) ($n = 48$).

	Block length	Block width	Block height	Block depth	Block height	Block circum.	Slope gradient	Align-ment	Depr. length	Depr. width	Depr. depth
Slope gradient	−0.371*	−0.293*	−0.268	−0.082	−0.264	−0.307*					
Alignment	−0.015	−0.099	−0.091	0.099	−0.117	−0.056	−0.200				
Depression length	0.542*	0.407*	0.243	0.040	0.454*	0.589*	−0.098	−0.134			
Depression width	0.671*	0.724*	0.518*	0.190	0.688*	0.692*	−0.357*	−0.050	0.382*		
Depression depth	0.509*	0.574*	0.474*	0.152	0.560*	0.607*	−0.046	−0.022	0.297*	0.512*	
Mound	0.575*	0.650*	0.455*	0.185	0.572*	0.534*	−0.155	−0.102	0.192	0.408*	0.500*

* = $p \leq 0.05$.

G (levée) type cross-section form. Mound heights range predominantly from 0 m to 0.8 m with outliers of 1.4 m and 1.7 m for the largest two ploughing blocks. The predominant mound shape is Tufnell's B-type (vegetated frontal/lateral mound with single mound crest) (Figure 2).

CORRELATIONS BETWEEN MORPHOLOGICAL PARAMETERS

Morphological parameters were correlated using the Pearson's correlation coefficient (Table 2). As the two largest blocks form distinct outliers in the frequency distributions, these were omitted from the correlation analysis. All block-size parameters indicate significant positive correlations (at 95% level) with all depression size parameters as well as with mound height. Thus, in the present study, an increase in block size results in an increase in depression size and mound height. Depression parameters also show internally significant correlations, but these may be considered dependent variables. Further, mound height is positively correlated with depression width and depression depth. All these relationships suggest block movement to be responsible for the depression and mound height characteristics. This, in turn, indicates active block movement with, apparently, no other factor(s) significantly determining the depression size and mound height. However, no correlation between slope gradient and depression length was found. This may explain the absence of any correlation between depression length and slope gradient, as the decrease in block size may offset the increase in gravitational force due to the increase in slope angle.

DISCUSSION

Compared with other studies (e.g. Wilson, 1993; Chattopadhyay, 1983; Allison and Davies, 1996) the morphometric correlations found in this study are particularly strong. This may result from the data in this study having been collected from a relatively small area with uniform terrain properties. Elimination of such factors as sediment composition, vegetation type and cover, altitude, and slope orientation results in a clear relationship between the parameters describing block and depression morphometry. This confirms the suggestion by Wilson (1993) that much of the variability found in British studies is explained by site-specific factors. It also provides strong support that gravity is the main process behind ploughing block development. The large depression length, width and depth data measured in this study are also of interest since Gorbunov (1991) suggests the Tien Shan depressions to be particularly long owing to their long survival in a semi-arid setting. Considering the substantially wetter conditions in the present study, this points to potentially high movement rates in the McGregor Mountains, which may be a function of availability of larger blocks in a more humid environmental setting, despite the lower slope angles in the McGregors.

Various field observations indicate that ploughing block movement has occurred in recent years in the McGregor Mountains. First, bare and scarred vegetation on mounds and depression surfaces are direct indicators of recent block movement. Second, a young subalpine fir (*Abies lasiocarpa*) was found in a tilted position due to ploughing block movement (Figure 4). Subsequent stabilization of the block is indicated by growth response of the tree (i.e. in a vertical direction). Indirect indications of activity are provided by the close correlations established between block size and depression parameters. Relict depressions would be expected to be at least partly obscured, thus reducing the strength of statistical correlations between depression data and block size parameters. The strong correlations point to either recent movement, or very low obliteration rates, or both.

The absence of solifluction forms in the study area suggests low obliteration rates; in spite of the heavy snowfall and wetness of the region, trough obliteration is not occurring, as implied in the study of Gorbunov (1991). With respect to movement rates, the 4.9 m block was found with a scarred alpine fir on its levée 5.5 m upslope from the block. The relationship of the scar to tree growth indicators suggests the block moved past the tree about 100 years ago. Assuming movement by continuous creep, this equates to an annual movement rate of about 5.5 cm a^{-1} over the last century, which is much in line with rates measured by Tufnell (1976) and Gorbunov (1991). However, in the absence of more quantitative data, it is unclear whether movement is by single, rapid events or by longer-term continuous (or near-continuous) movement.

In their summary of British ploughing block data, Ballantyne and Harris (1994, p. 217) note three common features regarding their morphological characteristics:

(1) Blocks are all larger than 20–25 cm long axis, suggesting a minimum critical mass is required for movement.
(2) Measurement of boulder orientation shows the great majority aligned directly downslope.
(3) Furrow length is positively related to boulder size, which may reflect either: (a) large boulders leave large furrows that survive encroachment, or (b) large boulders move faster relative to the surroundings than do the small ones.

They also provide data from five British studies (Ballantyne and Harris, 1994, Table 11.2) which give a useful comparison with those presented here (Table 1). In Britain, average movement rates vary from 0.3 to 34.5 mm a^{-1} with rates of movement being somewhat related to slope angle and the timing of movement being associated with spring thaw. Rate estimates from this present study, assuming a linear relationship of movement with time, give a rate of *c.* 55 mm a^{-1}. Boulders in the present study were all substantially larger than those found in the British studies while the relationship of boulder size to furrow length had an r-value of only 0.542. Furrow lengths were substantially longer in Canada than those found in British studies and this may reflect the combination of the larger boulder sizes (in Canada) coupled with the wet conditions during snowmelt. The evidence available from this present study suggests that the furrows survive encroachment and that this may be as a result of the limited solifluction activity in this area (cf. Ballantyne and Harris, 1994, p. 217).

In terms of movement, no specific data are available from Canada. It is thought that movement is related to slope failure resulting from the loading by the boulder on a slope saturated by snowmelt. Such a suggestion does not fit with either the four movement mechanisms suggested by Tufnell (1972), namely heating and cooling, frost creep, freezing and expansion of water upslope of a boulder, or sliding of the boulder over a frozen surface, or that of Ballantyne (1981) which argues solely for gelifluction. The lack of solifluction activity within the area, and certainly the absence of solifluction forms on the slopes where ploughing blocks are found, argues against the effectiveness of this mechanism in this area. The attributes discussed above that limit solifluction would, however, generate high porewater pressures and a lowering of the shearing resistance of the soil. That, coupled with the large sizes of the boulders measured in this study, indicates the possibility of an origin *not* associated with ground freezing conditions.

CONCLUSIONS

The present study confirms the first of two generalizations pertaining to ploughing blocks identified by Ballantyne and Harris (1994, p. 216), namely that they are 'confined to areas of vegetated, frost-susceptible regolith'. However, the character of the local conditions at the Canadian site does not so well accord with the second finding (p. 217), namely that ploughing blocks 'show a remarkably close association with active solifluction sheets and lobes'. Observations throughout this part of the Rocky Mountains indicate a paucity of solifluction.

Ploughing blocks warrant further consideration given the perception that they are regarded, in areas such as Britain, as a reflection of periglacial conditions. If this were the case, why are observations so limited in the high latitude and high altitude environments of North America? Further, despite their recognition and the measurement of their morphology elsewhere, little information is available regarding the timing and rate(s) of their movement.

ACKNOWLEDGEMENTS

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XV INTERNATIONAL CONGRESS

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THE ENVIRONMENTAL BACKGROUND TO HOMINID EVOLUTION IN AFRICA

BOOK OF ABSTRACTS

by successive waves in the tsunami wave train. At tidal marshes within the zone of co-seismic subsidence, the tsunami depo abruptly overlie former marsh surfaces and are overlain by intertidal mud that grades upward into peat. This stratigra succession is indicative of nearly coincident co-seismic subsidence and tsunami deposition, followed by gradual sedin accretion which re-establishes the marsh. In contrast, tsunami sediments in lakes are interbedded with freshwater silt, peat gyttja. Tsunami deposits attributable to great earthquakes at the Cascadia subduction zone and other subduction zones in the No Pacific Ocean have been identified at many marshes and lakes on Vancouver Island. Great earthquakes at the Cascadia subduct zone have a recurrence interval of approximately 500 years; the last event occurred in AD 1700. The most recent large tsunam strike the British Columbia coast, however, was triggered by the great Alaska earthquake in 1964. The physical evidence provi a basis for predicting the effects of future subduction earthquakes and their tsunamis, for example, tsunami run-up, and areas a amount of coseismic subsidence. This information is useful in formulating emergency plans and managing development on Pacific coast.

JG BOCKHEIM & ML PRENTICE

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SOIL RESOURCE DATABASE FOR THE MCMURDO DRY VALLEYS, ANTARCTICA

A soil resource database was prepared for the McMurdo Dry Valleys (MDV) from data collected by J. Bockheim & colleagu during the period 1975-1987. Data collected from 483 sites includes site, soil morphology, and surface boulder weathering featur and chemical, physical and mineralogical soil properties. The data are stored in ARCINFO that is available over WWW and c CD-ROM using ARCVIEW. The sampling localities are shown on 1:25,000 SPOT and LANDSAT schemes superimposed c topographic maps using a WGS84 datum and Lambert Conformal projection. The data are useful for (1) reconstructing the glaci history of the MDV and the dynamics of the West and East Antarctic ice sheets; (2) studying soil forming rates and processes in cold desert environment; and (3) evaluating changes in the physical environment of the MDV due to global change. Examples ar shown of the use of the soil resource database, including regional airmass influence on salt composition of soils, weathering stag and stability of the East Antarctic ice sheet, and a soil map of the MDV based on *Soil Taxonomy* (1998) and the World Referenc Base for Soil Resources (1998) schemes.

JG BOCKHEIM (1) & KJ HALL (2)

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PERMAFROST, ACTIVE LAYER DYNAMICS AND PERIGLACIAL ENVIRONMENTS OF CONTINENTAL ANTARCTICA

Active-layer, permafrost and ground-ice characteristics and selected periglacial features are summarised from recent published literature and unpublished data by the authors for three eco-climatic regions of continental Antarctica: the Antarctic Peninsula and its offshore islands (AP)(ca. 61-72S), maritime East Antarctica (MEA)(ca. 66-71S), and the Transantarctic Mountains (TM) (ca. 71-87S). Active-layer thickness and depth to ice-cemented permafrost are related to regional climate, proximity to glaciers, and albedo of surface rocks. Active-layer thicknesses are 40-150 cm for AP, 60-150 cm for MAE, and 15-50 cm for TM. The gravimetric moisture content of the active layer commonly is 5.8 21% in AP, 1.5-28% in MEA, and 0.3-4.0% in TM. In the McMurdo Dry Valleys, the active layer is commonly underlain by dry permafrost which can be detected only from frost tubes or temperature measurements. Antarctica contains 37% of the global permafrost. However, much of the land mass below the massive East Antarctic ice sheet is above the pressure melting point and is unfrozen so that Antarctica's contribution to the world's permafrost may be considerably less. In ice-free areas permafrost thickness ranges from 0 near thermally stratified saline lakes in the dry valleys to 1000 m. Permafrost temperature measurements are scant, ranging from -14 to -24 °C (50m depth). Ground ice is present as rock glaciers, ice-cored moraines, and ice wedges. Cryoplanation terraces occur along the Antarctic Peninsula, but salt weathering is a dominant landscape process in the Transantarctic Mountains. The Antarctic permafrost environment is especially sensitive to global climatic change. Accordingly, recommendations are given for an active-layer and permafrost monitoring network in Antarctica.

Jan BOELHOUWERS

BLOCKFIELDS AND BLOCKSTREAMS IN SOUTHERN AFRICA: A REGIONAL SYNTHESIS

Blockfields and blockstreams are of widespread occurrence throughout southern African mountains where resistant lithologies are present. These landforms have frequently been interpreted as being indicative of periglacial environments. This paper reviews the characteristics of southern African blockfields and blockstreams and evaluates their palaeoenvironmental significance. Autochthonous blockfields occur wherever blocky weathering and high rates of matrix removal are present. Their interpretation as periglacial landforms relies on a block origin by frost weathering. As no diagnostic criteria for such origins exist, the use of these landforms as indicators for such environments remains problematic.

Sedimentary structures in allochthonous blockfields and blockstreams suggest deposition under slow mass wasting processes on low slope gradients. Analysis of sites suggests that blockstreams in the Lesotho highlands are primarily the result of solifluction, but the result of frost creep in the coarser sediments of the Western Cape Mountains. The landforms offer climatic proxy signals indicative of seasonal frost penetration, but do not require permafrost for their formation. While a temporal framework for the emplacement of these deposits must still be established, the main phase of deposition in the Western Cape is placed around the Last Glacial Maximum under temperature conditions 7-8°C lower than at present.

PERMAFROST, ACTIVE-LAYER DYNAMICS AND PERIGLACIAL ENVIRONMENTS OF CONTINENTAL ANTARCTICA

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1. ANTARCTICA: AN INTRODUCTION

- 14 million km² area
- <1% (55,000 km²) ice-free
- Reportedly contains 37% of world's permafrost but realistically <25% (because of pressure melting below Antarctic ice sheet)
- "Dry permafrost" is common in the McMurdo Dry Valleys (MDV)
- Continental Antarctica has had a cold desert environment since the Miocene

2. IMMEDIATE OBJECTIVES ARE TO SUMMARIZE

- PERMAFROST AND GROUND-ICE FEATURES
- ACTIVE LAYER DYNAMICS BY ECO-REGION
- PERIGLACIAL FEATURES AND ENVIRONMENTS

LONG-TERM OBJECTIVES ARE TO:

- Produce a circumantarctic map of permafrost and ground-ice conditions
- Network with other Quaternary scientists interested in Southern Hemisphere permafrost
- Encourage participation of Antarctic researchers in the Circumpolar Active-Layer Monitoring Program (CALM)

3 ECO-CLIMATIC REGIONS & SUBREGIONS OF ANTARCTICA:

- Antarctic Peninsula (c. 63-72°S)
- Maritime East Antarctica (c. 66-71°S)
- Transantarctic Mountains, TAM (c. 71-87°S)
  - Coastal
  - Inland Valley Floors
  - Inland Valley Sides
  - Upland Valleys
  - Plateau Fringe

[illegible]	[illegible]	[illegible]	[illegible]

4 PERMAFROST DISTRIBUTION & PROPERTIES

- Absent beneath the Antarctic ice sheet where it is thicker than 3000 m
- Continuous in ice-free areas except beneath saline lakes in the MDV
- Dry permafrost is pervasive on older surfaces in inland areas of the MDV
- Thickness ranges from 0 to 970 m
- Temperature at 50 m ranges from -13 to -24°C

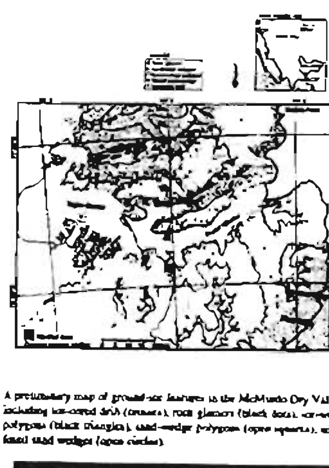
[illegible]	[illegible]	[illegible]	[illegible]	[illegible]	[illegible]

Permafrost exists throughout the ice free areas (shown in black) of continental Antarctica. Subglacial permafrost beneath the Antarctic ice sheet may be restricted to the bedrock areas.

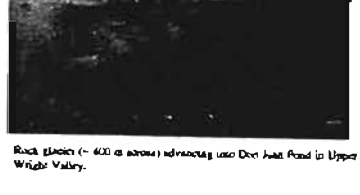
Beacon Dry Valley showing moraines (person for scale) and rock glaciers (background).

5 GROUND ICE IN ANTARCTICA

- Common forms: rock glaciers, ice-cored drift, buried glacial ice
- Less common forms: ice wedges
- Forms absent: pingos



A preliminary map of ground-ice features in the McMurdo Dry Valleys, including ice-cored drift (crosses), rock glaciers (black dots), ice-wedge polygons (black triangles), sand-wedge polygons (open squares), and fossil sand wedges (open circles).



Rock glaciers (~ 400 m across) advancing into Don Juan Pond in Upper Wright Valley.

Sand wedge casts (shown by white arrows) in Pliocene-aged deposit, Arena Valley

6 ACTIVE-LAYER PROPERTIES

- Thickness varies by eco-climatic region
- Exceptionally low (<3%) moisture contents in TAM
- Surface albedo important in TAM

	Active layer	
[illegible]	[illegible]	[illegible]

Dry permafrost is prevalent in the McMurdo Dry Valleys. Areas where ice-cemented permafrost occurs is at depths greater than 100 cm (black dots).

7. PERIGLACIAL FEATURES/ENVIRONMENTS.

- Cryoplanation/nivation are dominant processes in the moist coastal eco-regions (Antarctic Peninsula and Maritime East Antarctica)
- Salt weathering is dominant in the dry TAM

8 CONCLUSIONS:

- Antarctica contains <25% of the world's permafrost
- Dry permafrost is common in the McMurdo Sound region
- Permafrost thicknesses and temperatures are comparable to those measured in the Northern Hemisphere
- The most common forms of ground ice are rock glaciers, ice cored drift and buried glacial ice
- The active-layer thickness is dependent on regional climate and locally on surface albedo

INTERNATIONAL UNION FOR QUATERNARY RESEARCH

Quaternary Periglacial Phenomena in the Sani Pass Area, Southern Africa

XV International Conference Field Guide
12 AUGUST - 15 AUGUST 1999

Leaders: **Stefan Grab**, Department of Geography and Environmental Studies, University of the Witwatersrand, South Africa. **Jan Boelhouwers**, Department of Earth Sciences, University of the Western Cape, South Africa **Kevin Hall**, Geography Programme, University of Northern British Columbia, Canada. **Ian Meiklejohn & Paul Sumner**, Department of Geography, University of Pretoria, South Africa.

With contributions from: **Awie Fourie, Marieta Botha & Gerard Van Weele**, University of Pretoria, S.A. **Steve Holness**, University of the Western Cape, S.A.

Location and Accessibility

The highest and most extensive mountain range in southern Africa is that of Quathlamba (Zulu name for "a barrier of spears"), more commonly known as the Drakensberg (Dutch name for "Mountains of the Dragon"). The Main Escarpment forms a natural border between the Lesotho Highlands and the "Little Berg" in KwaZulu-Natal (Figure 1). The only direct access route for vehicles between KwaZulu-Natal and the Lesotho Highlands is via Sani Pass, cresting at an altitude of 2874 m.

The Main Drakensberg Escarpment which reaches heights of over 3300 m a.s.l., dominates the scenic beauty when driving north from Durban towards Johannesburg. The route from Durban to Sani Pass followed during this excursion is via the eastern Lesotho highlands and takes approximately 7 hours. The Sani Top mountain chalet is located near the upper Sani Pass border post (Lesotho) and overlooks the Main Escarpment. Four wheel drive transport is required for the 8 km long mountain pass which winds its way from the sandstone foothills into the Drakensberg Group amygdaloidal basalts (Figure 2). The field sites visited near Sani Pass are situated west of the Main Escarpment in the Lesotho Highlands.

Environmental Setting

The Main Escarpment

In southern Africa, the Main Escarpment (also known as the Great Escarpment or Drakensberg Escarpment in KwaZulu-Natal) extends approximately parallel to the coast from northern Namibia, south and east through the southern Cape and north through KwaZulu-Natal into Mpumalanga (Figure 1). Maximum altitudes range from below 1500 m on the quartzites of the west coast to in excess of 3000 m on the eastern coast basalts where the escarpment watershed demarcates the national boundary between Lesotho and South Africa (KwaZulu-Natal province).

Several theories for the development of the Main Escarpment have been proposed (e.g. King, 1944; 1962; Ollier and Marker, 1985; Partridge and Maud, 1987; Birkenhauer. 1991). A common aspect of these theories is that the post-Gondwana landscape of southern Africa is characterised by several well-defined landscape cycles (Gilchrist, 1995). This is due to post Late Jurassic episodic tectonic or eustatic landscape rejuvenation around the margin of the continent. Erosion of the downwarped coastal margins and subsequent backwearing of a scarp from approximately the position of the present coastline has given rise to the Escarpment as it exists today.

Notwithstanding the dissected highlands which exist above the oldest African surface (Late Jurassic/Early Tertiary to end of early Miocene; Partridge and Maud, 1987), several erosion surfaces have been suggested for the Little Berg region below the Drakensberg Escarpment (eg. King, 1974), related to the episodic uplift. Although surfaces can be observed in the Little Berg below the Main Escarpment, problems occur when matching erosion surfaces across the Drakensberg (Partridge and Maud, 1987). Due to the nature of the underlying sandstones and basalts, it seems likely that the distinctly stepped landscape is a product of slopes retreating in equilibrium with their rock mass strength (see Moon and Selby, 1983).

The escarpment is characterised by numerous cutbacks and passes. The positioning of the drainage of the Lesotho Highlands and the cutbacks off the escarpment follows major joints and doleritic intrusions through the basalts (Dempster and Richard, 1973). Doleritic intrusions are common to many of the cutbacks. Evidence of the intrusion, characterised by close joint spacing and enhanced weathering at the basalt/dolerite

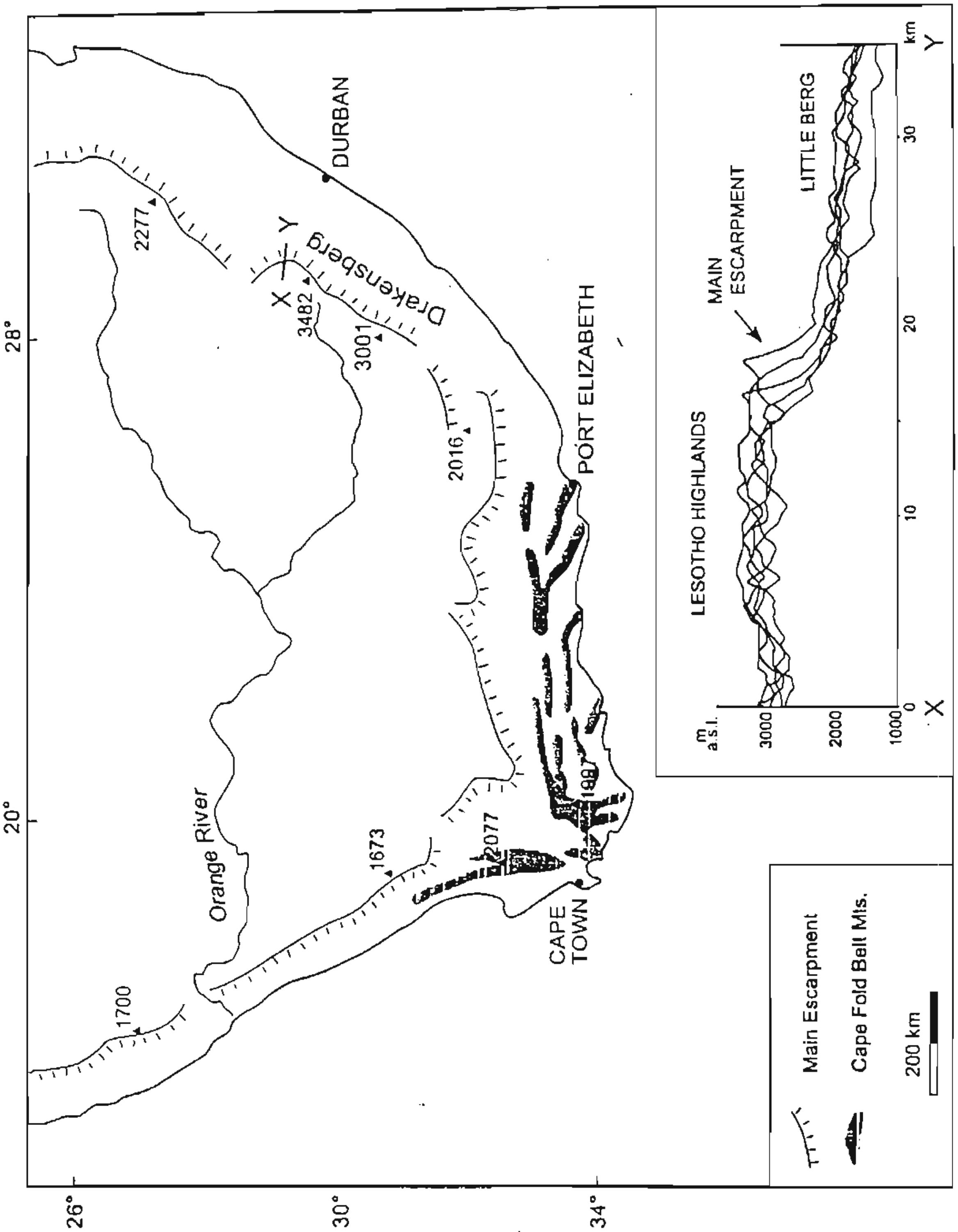


Figure 1. The Main Escarpment in southern Africa (after Marker, 1995a; Boelhouwers, 1992).

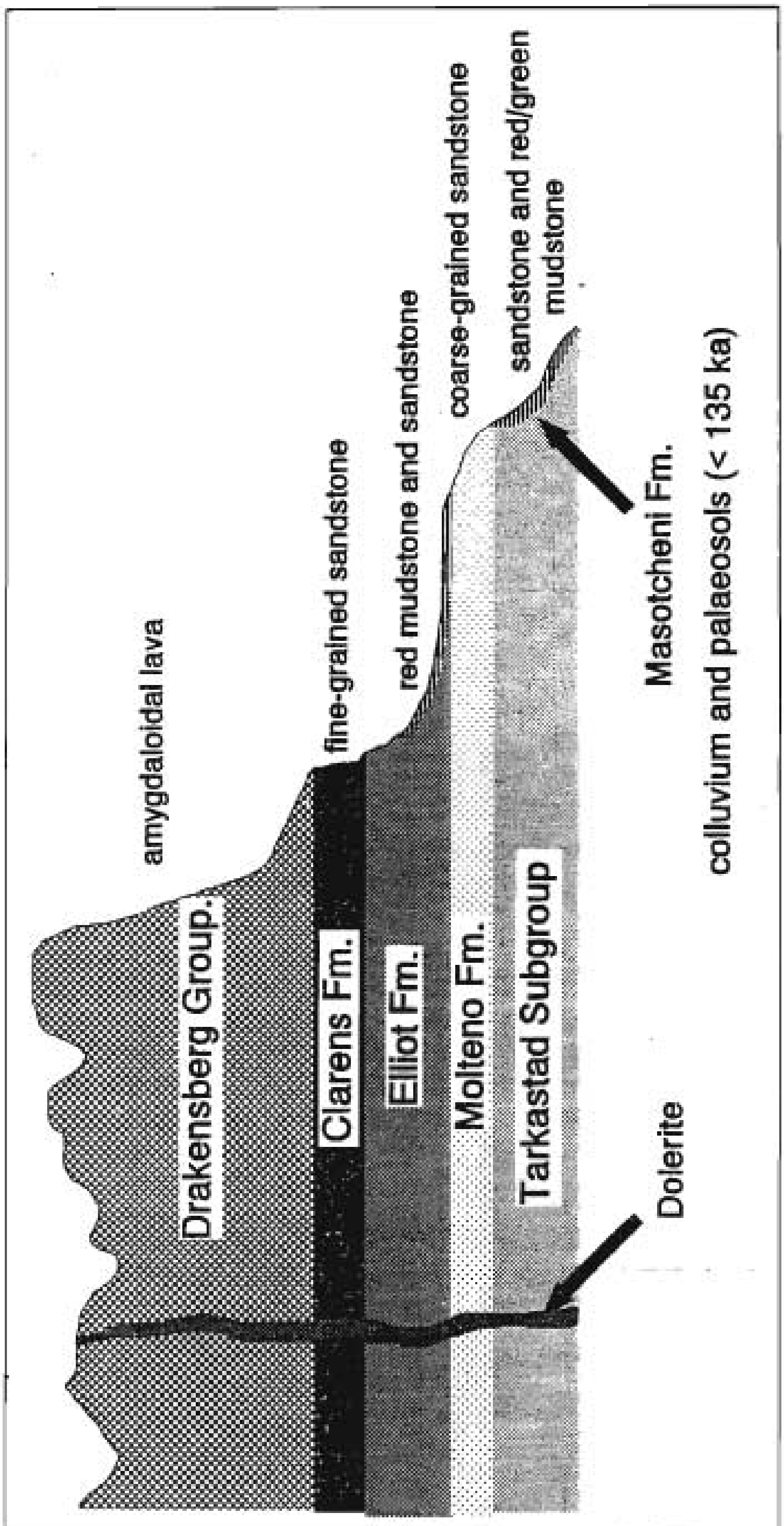


Figure 2. The typical topographical expression of each unit in the geological succession in the Drakensberg (after Pickles, 1985).

contact, can be observed in the upper region of Sani Pass. Preferential weathering and enhanced surface erosion are suggested as the causes for the location of this and other cutbacks. Remnants of other doleritic intrusions can be observed lower in the pass.

Although downwearing and backwearing of the cutbacks is expected to be continuous, major incision occurred during and after an estimated 900 m (in this region) of uplift in the Pliocene (Partridge, 1997). The Escarpment cutbacks appear to have been ideal sheltered sites for the accumulation of snow during the L.G.M. (Grab 1996a). The presence of niche glaciers and nival action in cutbacks north of Sani Pass, has been suggested by Hall (1994) and Grab (1996a). The ideas of niche glaciation are based on depositional landforms found in the cutbacks. Hall (1994) further suggested that meltwater from snow and ice contributed to the development of deposits in the cutbacks. Similar deposits to those discussed by Hall (1994) and Sumner (1995) can be observed adjacent to the river in the lower regions of the Mkhomazana Valley in which the Sani Pass is located.

Soils and Vegetation
The shallow soils of the Lesotho Highlands are mostly residual and colluvial in origin. Small amounts of alluvial material have contributed to the deeper soils in the wide valley-floors. Most of these soils lack a "B" horizon but are frequently underlain by a yellowish-brown cambic horizon at greater depths (Klug *et al.* 1989). Most of the miniature cryogenic features such as sorted and non-sorted patterned ground have developed in mollisols due to the abundance of fine material and soil moisture. Examination of the palaeosols could provide an important contribution to determining palaeoenvironments, such as those recently studied at Tlaeeng Pass (Hanvey and Marker, 1994). Palaeo-histosols have also been examined at a few sites in the Sani Pass region and dated between 13 490 and 2 310 ^{14}C yrs BP (Marker, 1994)(see Day 3).

The montane and alpine vegetation belts have been described by Killick (1963). The Podocarpus forests of the montane belt eventually give way to what has been described as "miniature" and "cushion" plants of the alpine belt (Van Zinderen Bakker, 1981). This alpine vegetation belt is characterized by climax heath communities, in particular woody species of *Erica* and *Helichrysum* (Killick, 1963). The stunted growth of the woody species could be attributed to the shallow soils and high wind speeds encountered along the Escarpment.

Climate
The Sani Pass region has a distinctly seasonal precipitation pattern; 70% falling between November and March and less than 10% falling between May and August (Tyson *et al.*, 1976). The most important source of precipitation over the Drakensberg are orographic thunderstorms. Occasionally, cyclonic weather systems (cold fronts) may bring some precipitation, particularly during autumn, winter and early spring, when they may also deliver snowfalls. The zone of maximum precipitation (about 1800 mm) along the main escarpment is believed to be between 2287 and 2927 m a.s.l. (Killick, 1963). Killick (1963) argues that the decrease in vapour content above this altitude reduces precipitation. The escarpment also produces a rain shadow to the western interior where a marked decrease in precipitation is encountered (Schulze, 1979)(Figure 3).

Estimates of the absolute maximum and absolute minimum temperatures for the Little Berg are +35.0°C and -12.5°C respectively (Tyson et al., 1976), while the higher peaks are thought to have a mean annual air temperature of about 7°C (Schulze, 1979). In the Lesotho Highlands, mean minimum temperatures during June and July are less

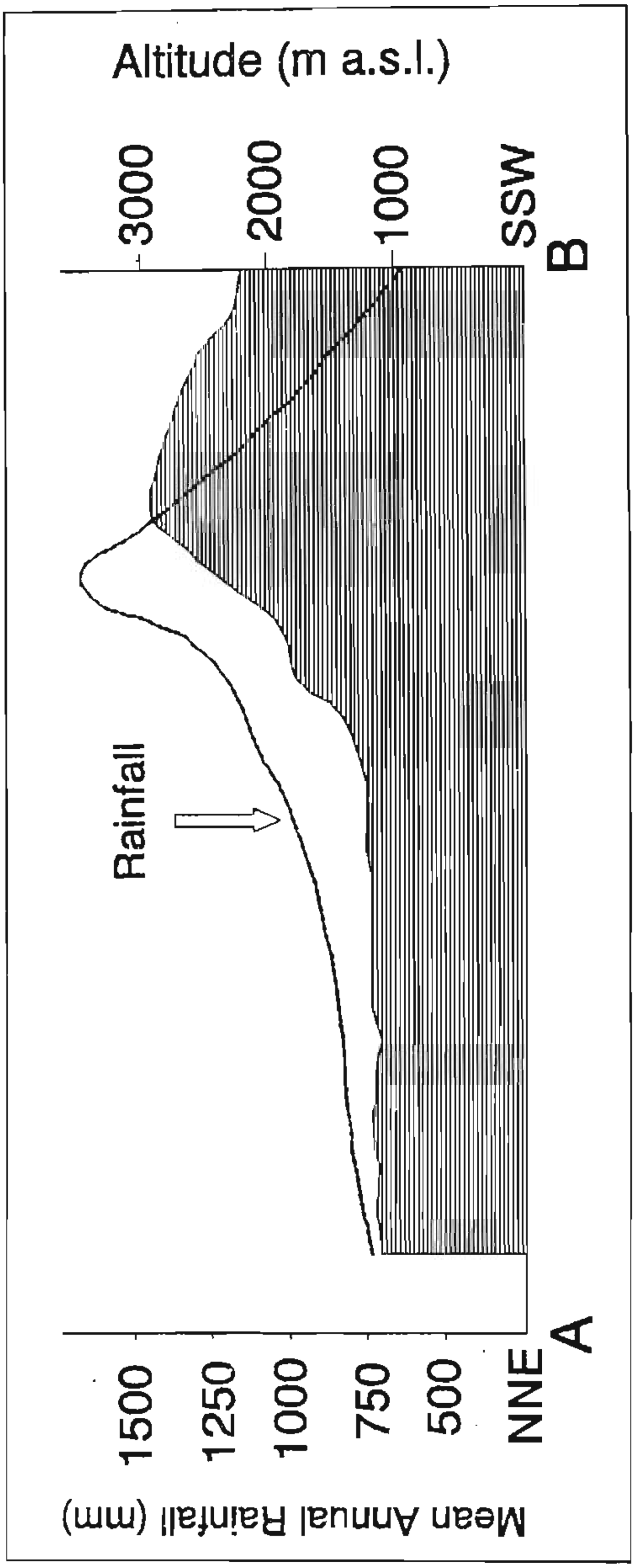


Figure 3. Rainfall : altitude relationship along A (Bergville) to B (Mothelsessane, Lesotho) (after Schulze, 1979).

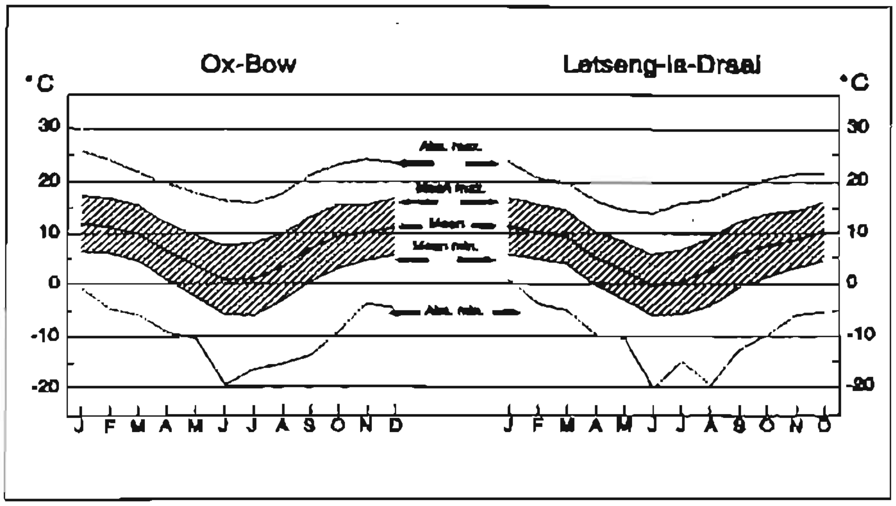


Figure 4. Monthly temperature graphs for Ox-Bow and Letseng-la-Draai (after Grab 1997a).

	January	July	Year
Mean Max	16.6	6.3	11.6
Mean Min	5.5	-6.1	-0.2
Mean	11.1	0.1	5.8
Abs. Max	23.5	15.4	23.5
Abs. Min	0.6	-15.2	-20.4
Frost Days	0	30.3	166.6

Table 1. Temperature data (°C) for Letseng-la-Draai, 3 050 m a.s.l (Average values based on 10 yrs data, after Grab, 1994).

than -6°C and daily minimum temperatures may fall to below -10°C from May to September (Figure 4; Table 1). The lowest temperature yet recorded is -20.4°C at Letseng-la-Draai on the 12 June 1967 and ground level freezing could occur on about 160 or more days per annum above 3000 m a.s.l. (Table 1) (Grab, 1997a).

Present-Day Soil Frost Features

A few varieties of contemporary miniature cryogenic landforms occur in the Lesotho Highlands (c.f. Boelhouwers, 1991) and several of these have been identified in the Sani Pass region (e.g. Boelhouwers and Hall, 1990; Grab, 1994). Although fossil periglacial landforms occur in the Drakensberg, these features are discussed later. Most of the microforms are the product of surficial diurnal freeze-thaw cycles or somewhat deeper (to 20 cm depth) winter freeze. The most widespread features are soil stripes and miniature sorted circles and polygons which are commonly linked to needle ice development, all of which are frequently visible on the slopes and interfluve of Kotisephola pass (Day 2).

Wetlands are visible along the many drainage channels and valley heads. Owing to the abundance of soil moisture in the wetlands during winter, and the presence of deep clay rich sediments, thufur are prominent in the wetlands. Also observable, are the many small terrace fronts with exposed soil, referred to as turf exfoliation by Troll (1973). These terraces have been ascribed to the erosive action of needle ice by Boelhouwers and Hall (1990). Yet, as suggested by Grab (1997b), the cryogenic role in turf exfoliation is still poorly understood. While the occurrence of turf exfoliation will not be explored in this field guide, it is worth exploring the potential cryogenic component whilst visiting the site. Although Gallart *et al.* (1993) argue that terracettes generally "belong to a family of periglacial forms" (p529), others (e.g. Vincent and Clarke, 1976) are not convinced of an "exclusive" periglacial origin. It appears that the terracettes in Lesotho are primarily the result of animal trampling and only slightly modified by cryogeomorphic processes (Day, 1993; Grab, 1997b). Other features in the region include stone-banked and turf-banked lobes, but have yet to be found in the Sani Pass area.

Sorted stripes

Most studies in the Lesotho highlands have attributed the formation of miniature sorted stripes to the growth and decay of needle ice, hence the use of terms such as "raked ground". "striated soil" and needle ice stripes" (e.g. Troll, 1944; Hastenrath and Wilkinson, 1973; Hanvey and Marker, 1992). These so-called needle ice stripes are common in open soil patches where there is sufficient soil moisture available. The stripes usually develop during April/May when diurnal soil freezing occurs but when there is still sufficient soil moisture available. As winter progresses and desiccation sets in at some localities, the soil stripes become less active and eventually form crumbly and "puffy" surfaces of disintegrating stripes. Deflation appears to eventually cause the stripes to decay in July/August. With more abundant moisture during the spring months, stripes frequently re-develop and may become active again until early November. However, on the somewhat wetter sites, or during wetter years, stripes may be cryogenically active throughout winter. Knowledge on the physical processes of such stripe formation still appears to be in its infancy. Troll (1958) suggested that nocturnal winds play an important role in stripe formation and thus may be referred to as "wind-striped frost-heaved soil". However, recent observations have shown that the stripes are predominantly aligned in the direction of the early morning sun, orientated in an east-west direction, irrespective of slope aspect (Grab, 1997b).

Cobble sorted stripes with mean a-axis clast sizes of 7.4 cm and sorted to a depth of 6.7 cm, measuring up to 3 metres in length with coarse stripe widths of 12.6 cm and fine stripe widths of 16.2 cm, are found (Grab, 1996b) about 50 km to the north of Sani Pass, on the high (3 350 m a.s.l.) Mafadi slopes. These stripes appear to be restricted to high altitude slopes but, unlike the needle ice stripes, are perennial features.A number of possible mechanisms such as water induced rill development, needle ice activity, segregation ice development and thermal creep have been suggested as possible formative contributing factors (Grab, 1996b).

Sorted circles and polygons

Miniature sorted circles and polygons may be observed on the Kotisephola interfluve and along dry sections of the Mangaung stream bed. These are usually represented by an accumulation of cobbles and gravel around centres of finer material, primarily the result of needle ice lifting material in the centres (Fig 5). Mean centre dimensions for such miniature sorted patterns commonly range between 11.6 cm and 16 cm. Despite altitudinal differences of 300 - 500 metres, there appears to be little variation in pattern size throughout the plateau region (Grab, 1997c).

Desiccation and thermal contraction cracking appear to be the primary mechanisms for such miniature pattern development at relatively dry sites on slopes and interfluves,while differential swelling and frost heaving are the likely operative processes within saturated stream gravel deposits (Grab, 1997c). Despite the low proportion (<5%) of

Figure 5. Miniature sorted circles depicting fine earth centres surrounded by cobbles and gravels.

clay/silt sediment size fractions, these miniature patterns may become fully developed over very short periods of five to six weeks (Grab, 1997c). Clearly, these miniature patterns are the product of a highly regular and frequent ground freeze-thaw cycle. Many of the patterns at lower altitudes and along drainage lines are destroyed by fluvial processes during the summer, while the better preserved patterns on slopes above 3200 m a.s.l. sometimes host perennial patterns.

Thufur

Several studies have quantitatively examined the morphological and ground freezing characteristics of thufur in the high Drakensberg (Boelhouwers and Hall, 1990; Grab, 1994; 1997d; 1998). Thufur distribution in the Drakensberg is primarily controlled by the spatial occurrence of wetlands. Both active and relic features are present and may be observed on days 2 & 3. Boelhouwers and Hall (1990) found thufur with an average height of 16 cm and average diameter of 70 cm, north of the road towards Kotisephola pass. The thufur are frequently elongated, possibly owing to the steep slope gradients of 9-10°, as the average thufur diameter for this mountain environment is only 50 cm (Grab, 1998).

Up to 50% of thufur in some swarms show signs of disintegration. A variety of factors such as desiccation cracking, solifluction, animal trampling and needle ice activity have been attributed to the turf exfoliation and rupture of thufur (Grab, 1994). However, the recent mild years and relatively dry wetland conditions have had an apparent negative impact on the ice rat (*Otomys sloggetti*) population expansion. The ice rats build their tunnel exits, usually on the thufur sides, where there is the advantageous protection against predation. In addition, many of the thufur have developed within thick organic (sometimes peat) horizons which offer welcome insulation during the cold winter months. Observations from the Mashai Valley (from 1995 to 1998) have indicated that the burrowing activity of the ice rats has increased substantially in recent years and progresses from the wetland fringes towards the central parts of wetlands.

The precise age of the thufur is unknown as age-dating has not yet been undertaken.

However, most of the wetlands are estimated to be between 2000 and 13 000 years old (Van Zinderen Bakker. 1955; Hanvey and Marker, 1994). Better developed thufur frequently occur towards the wetland centres, as opposed to the wetland periphery areas, and may have formed during wet cycles and degraded during drier times (Grab, 1998).

The thermal characteristics of thufur have been studied in the Mashai Valley, approximately 10 km to the south of the Sani Flats (Grab, 1997d). Considerable thermal differentials are found between thufur apexes and their adjoining depressions (Figure 6). The apexes may freeze to a depth of 20 cm and remain frozen from early June to late August while the depression soils remain unfrozen beneath about 3-4 cm depth. The existing micro-topography (e.g. thufur) is an important factor in controlling the development of small scale landforms in marginal periglacial environments by creating "frozen pockets" in an otherwise predominantly unfrozen environment (Grab, 1997d).

SG

Day 1: Travel from Durban to Sani Top Chalet.

A short walk will be offered down the Sani Pass cutback during the afternoon, to observe geological sequences and to examine the weathering of the Drakensberg Escarpment basalt.

Day 2: Cryoplanation Terraces/ Transverse Nivation Hollows (?)

Cryoplanation is said to be a "... cryogenic process promoting the low-angled slopes and level bedrock surfaces typical of many periglacial regions" (French, 1996, p.181). The processes seen as causative for cryoplanation terraces, namely intense frost shattering, solifluction and, in some instances, slope wash are similar in makeup to those associated with nivation, although their relative contributions and rates of operation may differ. The end result of 'cryoplanation processes' is the development of a low-angled bedrock surface. A number of authors (e.g. Reger and Péwé, 1976) consider cryoplanation terraces to be associated with permafrost and for them to exhibit orientational preferences. Many authors (e.g. Demek, 1969; Boch and Krasnov, 1943; Reger, 1975) identify nivation as *initiating* cryoplanation and that, at some later stage, 'cryoplanation' takes over from 'nivation'.

What we see today near Sani are distinct benches cut in the plateau basalts. Each has a vertical, weathered backwall with a tread composed of finer material close to the back wall (although there is some rockfall at the base of the face) and larger debris further away (Figures 7 & 8). In the photos, taken in September 1998, there can be seen to be remnants of snow against the backwall (in the protective shadow) and some ice on the face of the backwall (Figure 8). The benches create distinct steps in the landscape and appear to predominate on the southerly slopes, or at least to occur in greater numbers and as larger forms on these slopes. Whilst there, it is obvious that these *forms* are present, the question arises as to what they actually represent or are indicative of? Consideration of most forms in the high Drakensberg and in Lesotho has, inevitably, led to an association with periglacial conditions. Being able to observe that snow still resides against the backwall and that ice can be found on the weathered face tends to reinforce this association - if conditions are like this today then they should have been even more severe during the glacials. That being so, then a circularity develops such that these forms then *must* be the product of periglacial conditions.

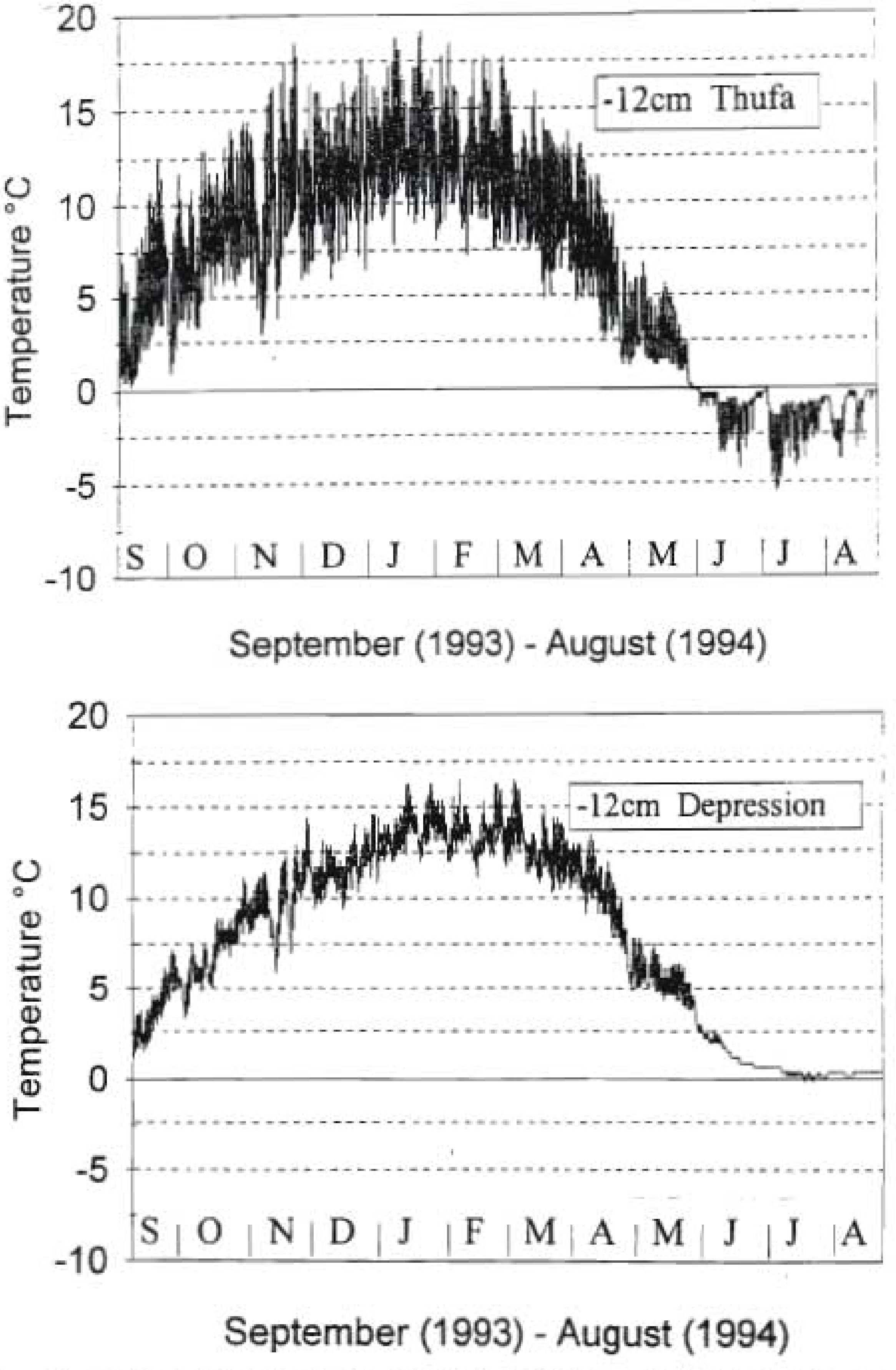
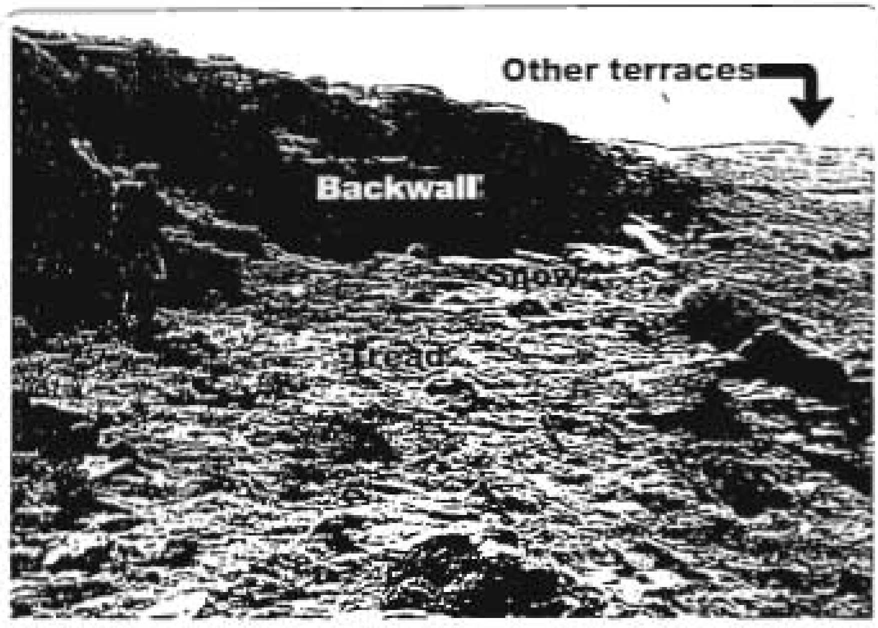
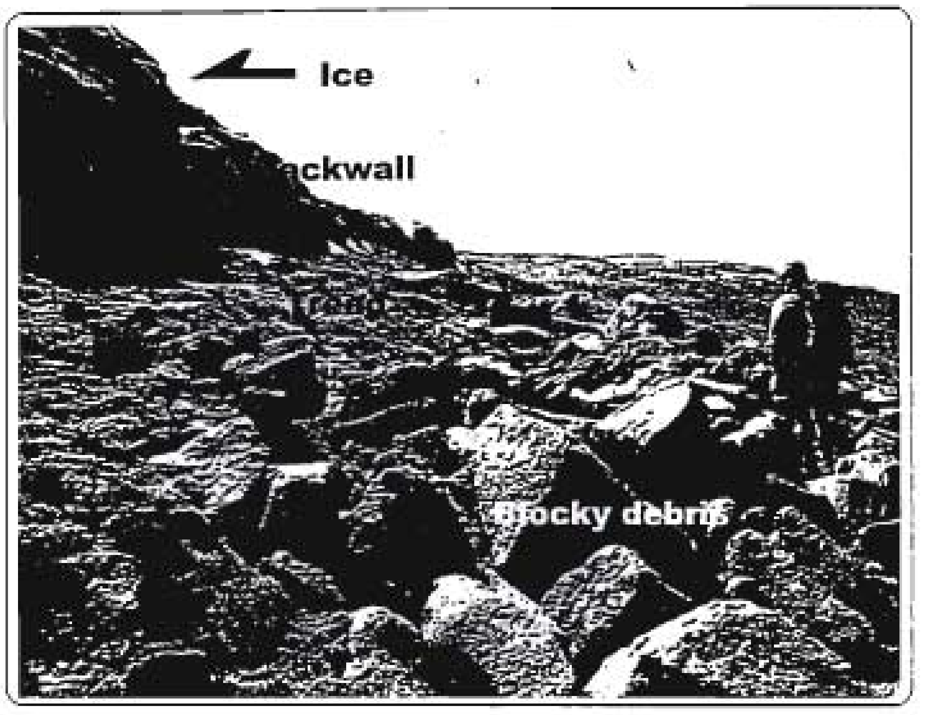


Figure 6. Comparison of apex and depression temperature trends for a thufa in the Mashai Valley, Lesotho (after Grab, 1997d).



View across the front of one of the forms that has the physical attributes of a cryoplanation terrace. The terrace is developed in basalt and faces south.

Figure 7. Basalt benches with scarp backwalls



Closer view of another terrace showing the blocky debris downslope from the backwall. There is a covering of ice on the backwall.

Figure 8. Snow and Ice may be seen along some of the scarp backwalls.

Although this circularity of argument may well not be in error, it still leaves the question as to what do these forms represent?

Nivation has been long associated with landforms in the Drakensberg but, to date, no mention has ever been made of cryoplanation, although Harper (1969) does identify "basalt steps" that he considered to be the product of frost wedging. Certainly as we walk from where the vehicles are left through to the block stream deposit (Figure 9) we will pass a number of features that have all the *visual* attributes of cryoplanation terraces (Figure 7). Further, these features are south-facing (note that the so-called 'nivation hollows' are north-facing) and so certainly experience a cold micro-climate. As you look around you will see many of these benches and there does appear to be a preference for a south-facing orientation. Recognising that as we traverse to the block stream deposits you go around hills with the horizontally bedded basalts and yet the major benches are on the south-facing aspect. Thus, there would certainly appear to be a preference for the colder aspect where snow will prevail longer and temperatures will be lower for a longer period. The thought that these are simply 'structural benches' is a non-sequitur as there still needs to be a suite of processes, apparently aspect constrained, to exploit along the structure - the term describes *only* form and does not consider process. A view across the landscape shows that these benches are certainly a noticeable feature and thus there needs be some explanation within the general framework of the landscape developmental sequence.

If 'nivation' is given some credence for the hollows, then the question would arise as to whether these forms are, in fact, 'transverse nivation hollows'? Certainly, if it can be thought that snow could survive, despite the northerly aspect, in the hollows then it will certainly do so in the lee of these steps - as it does today in winter. Indeed, it may well be that the north-facing hollows are post-glacial (i.e. post cold period) features associated with chemical weathering along joints whilst these benches are, in fact, transverse nivation hollows. Such a judgement would make more sense when aspect is considered. A clear problem here is how to determine *between* a transverse 'nivation hollow and a 'cryoplanation terrace'? The whole issue of form, process and terminology in 'nivation' and 'cryoplanation' is fraught with problems, especially where 'nivation' is seen to initiate 'cryoplanation' and yet cryoplanation takes over at some unspecified point although the processes remain the same! In reality, there is actually nothing that can identify either these benches or the facing hollows as of nivation *or* cryoplanation origin. The whole issue of nivation and cryoplanation - the terminological, process and form problems - is discussed at length in Hall (1998). Here, as we walk across, we can look back to see the 'nivation hollows' and can clearly see the benches - the questions remain as to the origin(s) of *both* features, as to what they tell us about the palaeoclimate (if anything) and what their status is within the landscape development of this area. As significant as these questions are to Lesotho and it's Quaternary history, so these questions need to be posed for all other areas where such forms are thought to exist.

Thus, we are left with a major question mark and all the more so if any argument is made for these to be cryoplanation terraces, for then that may imply permafrost (Reger and Péwé, 1976) and that, in turn, confounds consideration of glaciation or, perhaps, even

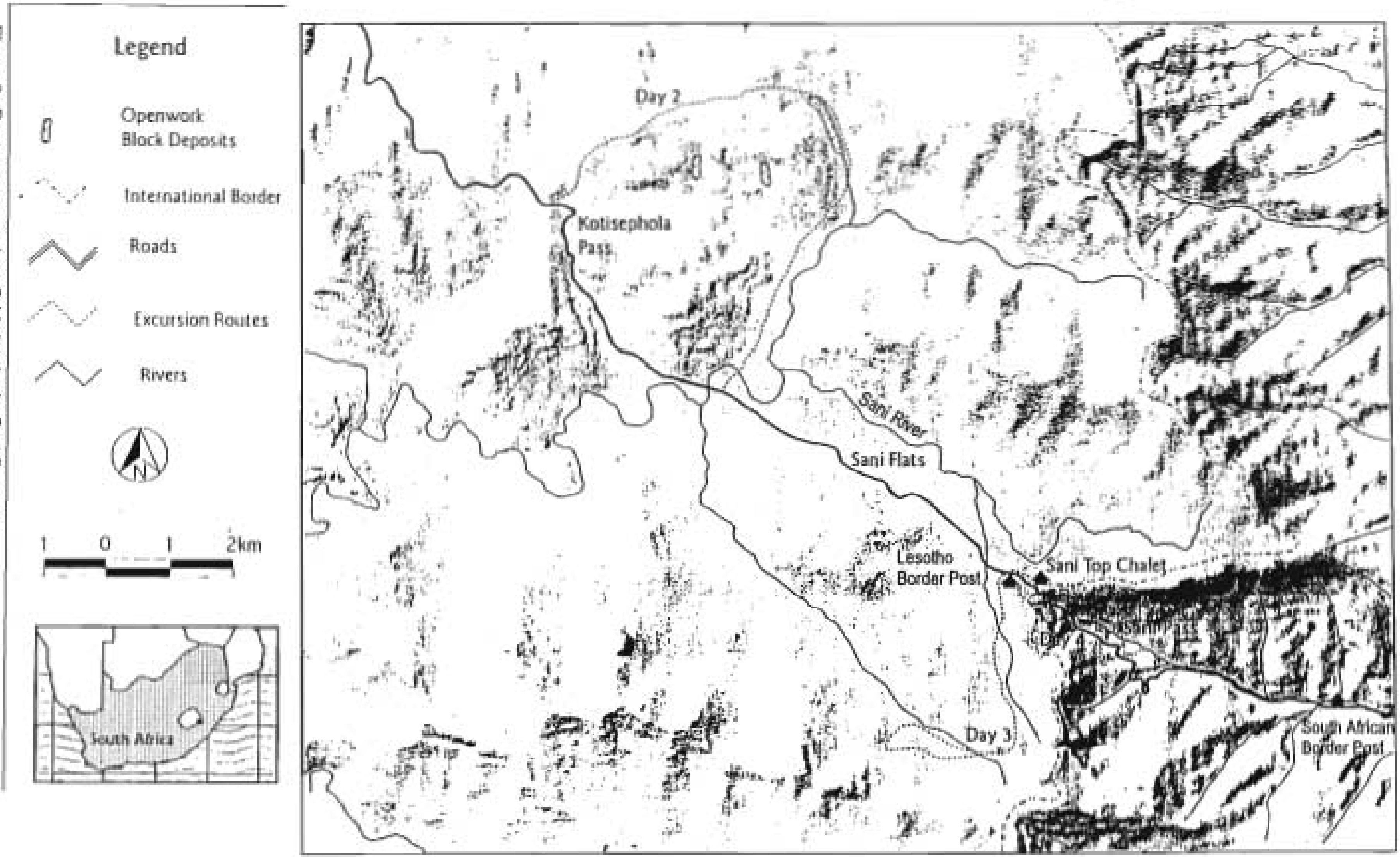


Figure 9. Route map to various field sites in the Sani Pass area.

nivation as both require substantial snowfall for operation. If an argument can be made that these are transverse nivation

hollows then, as they are significantly smaller than the north-facing hollows (considered to be nivation hollows), this may indicate that the latter are more likely a product of post-glacial weathering. Ultimately we are left with the problem of deducing process from form (for both nivation and cryoplanation forms imply specific processes and conditions) and this may simply be too dangerous. Further, this issue regarding the origin of these benches, and the hollows, symbolizes the larger problem of high Drakensberg-Lesotho (indeed southern Africa) Pleistocene conditions and landform development. What can we see that is unequivocal (if anything!) and from this what sort of scenario can we build regarding landform development and developmental processes for this region? **KH**

Periglacial blockstream and solifluction mantles

Location
The blockstream is positioned in a valley due east, and at about one hour walking distance, from the top of Kotisephola Pass (Figure 9). The route follows the ridge crest to the head of the blockstream valley.

Significance
The deposits in this valley are representative of those at many other high elevation sites in the Lesotho mountains. This blockstream was first referred to by Hastenrath and Wilkinson (1973), while Boelhouwers (1994) described relict allochthonous blockfields at Giant's Castle, 30km north from this site, as indicators of gelifluction processes. Similar blockstream deposits are reported for the Western Cape Mountains (Boelhouwers, 1996, in press), with openwork slope deposits also described for the higher regions of the Amatola mountains of the Eastern Cape (Marker, 1986; Sumner, unpubl. data). They must be distinguished from autochthonous blockfields, resulting from *in situ* weathering, which do not necessarily require a periglacial origin.

The allochthonous blockstreams, as observed on Day 2, are suggested to have formed under severe seasonal freezing, but not necessarily permafrost (Boelhouwers, 1994). Such deep frost penetration also implies the presence of a limited snow-cover. None of these openwork block deposits have been absolutely dated, but are considered of Last Glacial age. In the Western Cape, they are most readily associated with severe periglacial conditions during the Last Glacial Maximum (Boelhouwers, in press). As such, allochthonous blockstreams and blockfields of the summit areas of southern African mountains are considered the geomorphological manifestations of the most severe Quaternary periglacial conditions encountered.

Site descriptions
Along the watershed *en route* to the blockstream valley, cryogenic soil surface disturbance can be recognised in the form of needle-ice induced turf exfoliation, small turf-banked steps and micro-polygons. From the ridge crest there are excellent views into both the north- and south-facing valleys. The two south-facing valleys both contain openwork block accumulations in the central part of the valley-floor. Solifluction terraces can be seen along the valley-side and valley-head slopes. The last north-facing valley passed before reaching the blockstream valley contains a thick colluvial mantle that is currently being incised by fluvial erosion. The colluvial mantles are considered to be emplaced by Pleistocene solifluction processes, creating dry valleys with subsequent fluvial incision in the Holocene.

Site 1. Top of blockstream valley
The saddle leading into the blockstream valley offers a good overview of the four small catchments, designated A-D in Figure 10, that converge in the central valley where the main blockstream is located. The catchment is distinctly asymmetrical with tributaries B, C and D feeding into the main valley from the north. Figure 10 offers a geomorphological sketch of the catchment. The upper slopes of the catchment comprise small rock scarps alternating with thin debris mantles which level out in downslope direction on the next rock ledge. Rock scarps on the south-facing slopes are less continuous than on the north-facing slopes and show mechanically-induced fracturing. Rockfall-derived blocks spread over short distances between rock scarps. Downslope from this upper scarp-zone the north- and south-facing slopes develop notably different characteristics. The north-facing slope continues as a rectilinear rock slope with a thin debris veneer less than 1 m thick. Occasional rockbands, less than 2 m high, are exposed at the surface. Clast abundance is low. The south-facing tributaries are slightly concave in profile with deeper debris mantles that have a significantly higher clast content than the opposite slopes. Resistant basalt flows, although not necessarily exposed, create local breaks in slope above which wetlands are frequently present. Thufur are often well developed at such sites. As indicated in Figure 10, deep solifluction mantles have accumulated in these tributaries. In the downslope direction solifluction sheets form terraces with risers of 0.5 m - 1.5 m high and are particularly well developed in catchment (D). Clast content also increases in the downslope direction resulting in occasional openwork block deposits. Where located along preferred drainage lines, openwork deposits may be in a slightly depressed location as a result of matrix removal by suffosion. Lobate terrace fronts are often found where a local decline in slope angle occurs, with thufur immediately downslope from such a front. Rock scarp disintegration with localized displacement of blocky debris also results in small, isolated openwork block accumulations.

The south-facing valley slopes are clearly much longer and wider than the north-facing slopes, thus providing a much larger source area for debris production. Bedrock weathering at the upper catchment scarps also appears more advanced on the south-facing slopes. These two factors are suggested to largely explain the thicker debris mantles feeding into the central valley from south-facing slopes with relatively little sediment input from the north-facing slopes. The explanation for the distinct valley asymmetry in the Lesotho Highlands is discussed elsewhere in this guide. Suffice to point out that the processes discussed here operated in an already asymmetrical valley. Although process dynamics have been very different on opposite slopes during debris production and emplacement of the landforms under discussion, valley asymmetry is considered to pre-date these landforms.

Site 2. The blockstream
The large blockstream that occupies the central valley floor has a length of approx. 1.1 km and a maximum width of about 75 m. This long, narrow openwork block deposit is aligned along the central drainage of the main valley and is positioned at the terminal end of the solifluction mantles from catchments A and B (Figure 10). It extends downslope at a gradient of 8°. Cross profiles in Figure 11 show the block deposit to have an irregular surface, with its highest parts slightly raised above the surrounding valley-floor. The depressions and channels beneath the level of the surrounding slope point at removal of material subsequent to deposition. This provides

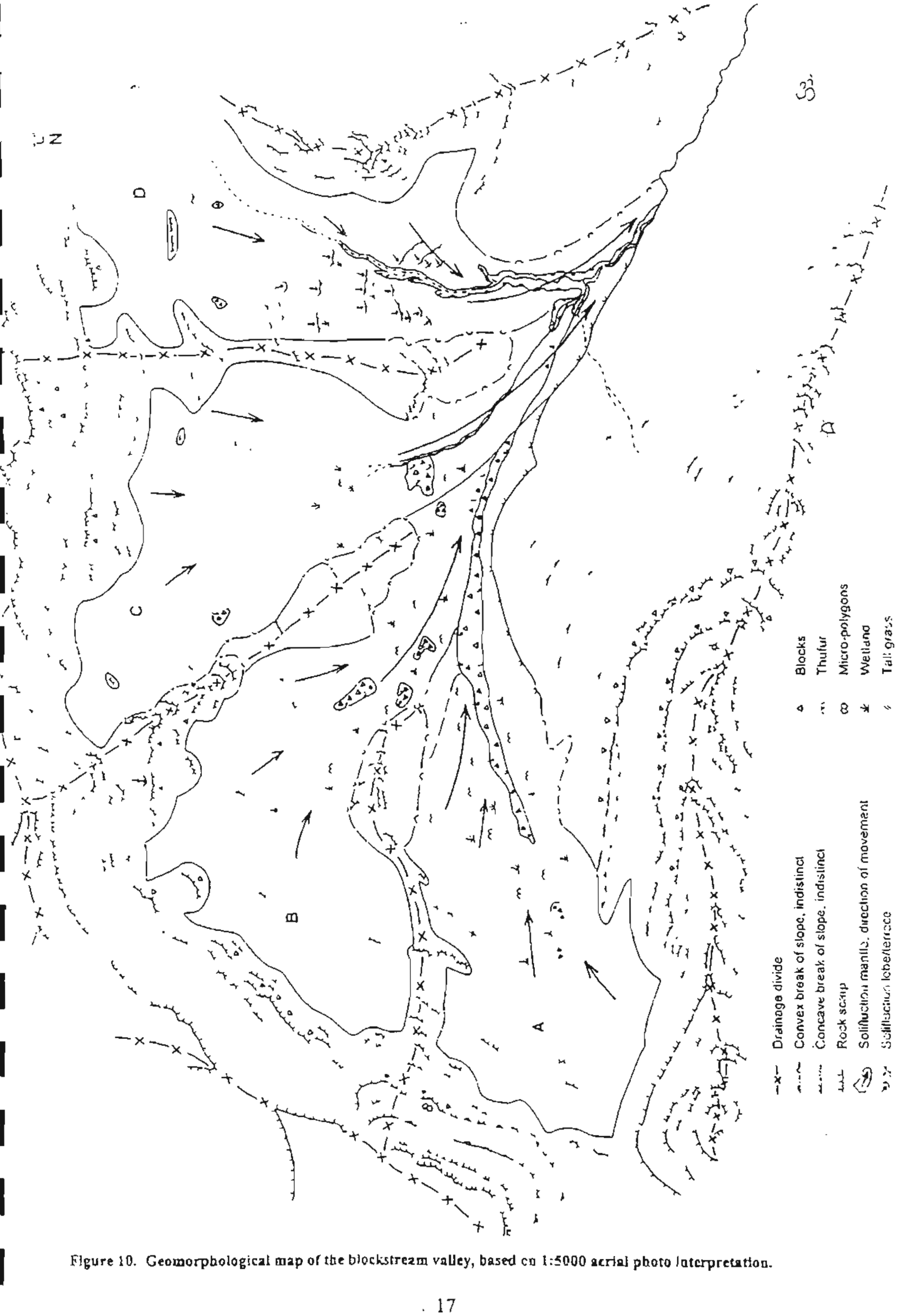


Figure 10. Geomorphological map of the blockstream valley, based on 1:5000 aerial photo interpretation.

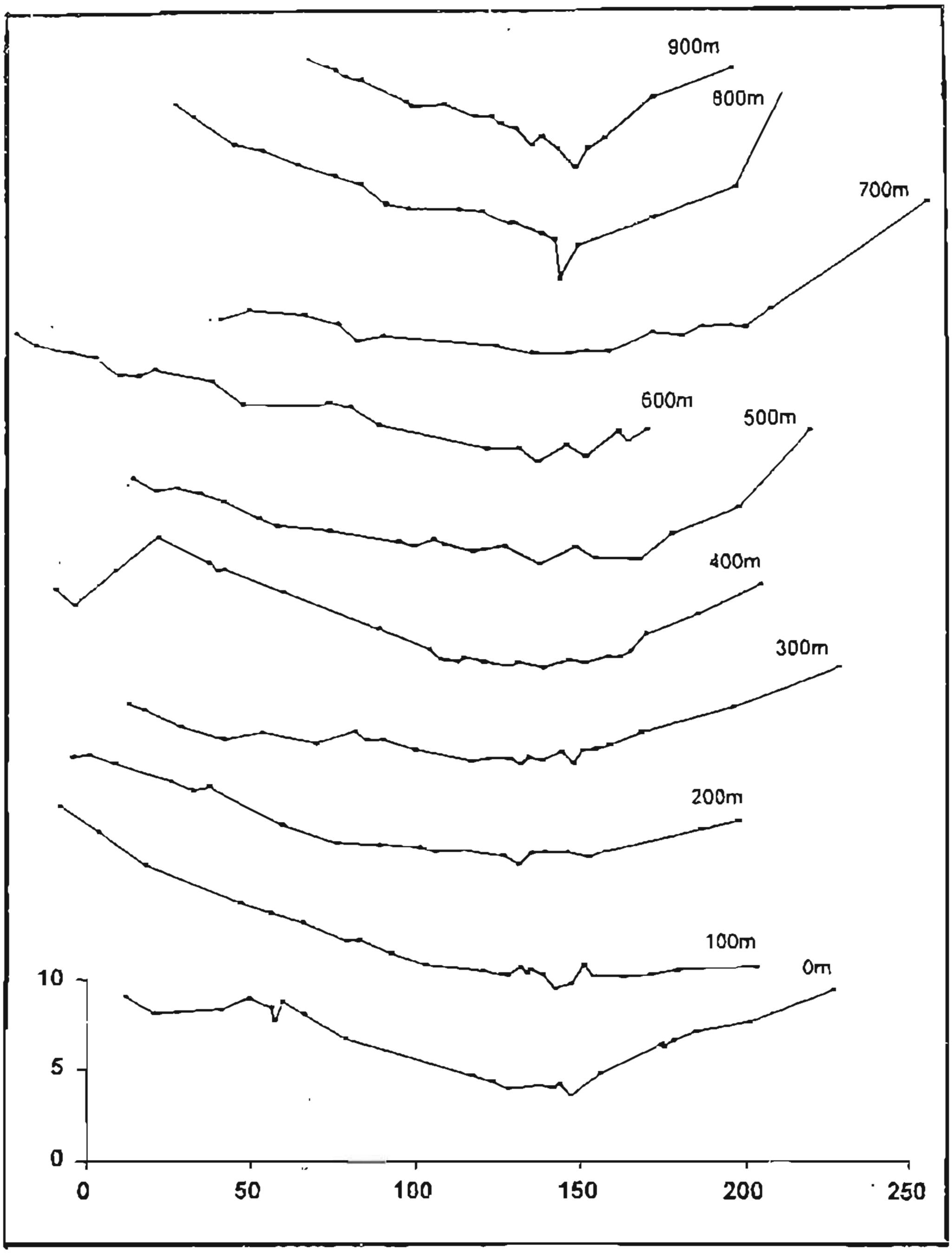


Figure 11. Cross profiles of the blockstream, viewed from its upper end (0 m) in downslope direction. Location of the transects is given in Figure 12.

an important indicator for the presence of a matrix during deposition of the valley-floor deposits, with the blockstream being a residual deposit following washing out of the matrix.

Block size (*a*-axis) data at various distances along the blockstream are summarized in Table 2 and indicate minimal variation in block size composition over the entire length of the blockstream. Fabrics of the *ab*-plane of clasts show well developed bi- or trimodal clast orientation distributions and are strong indicators that the block deposit is not the result of *in situ* weathering but, rather, is of allochthonous origin (Figure 12). The primary alignment of clasts at all sites, except at 500 m, is parallel to the local maximum slope gradient, which supports emplacement of the valley-floor deposits by means of slow mass movement. A secondary concentration of blocks with orientations at right angles to the primary mode is present at all sites reflecting imbricated and often steeply dipping clasts. These may result from local decelerating movement or block displacement by frost heave. A third peak, indicating a preferred north-south alignment, is found on the northern margin of the blockstream at 500, 700 and 900 metres. These sites are located at the confluence with catchments B and C and their more complex (random) fabrics are interpreted as a result of the pressures generated by downslope movement of the solifluction mantles from these catchments. Although difficult to substantiate due to the uniform lithologies in the area, the morphology of the deposits suggests that from 500 m the block material from catchment B is incorporated into the blockstream.

Table 2. Summary statistics of block *a*-axis measurements along the blockstream in downslope direction

Distance (m)	n	Min (cm)	Max (cm)	Avg (cm)	St. Dev (cm)
100	75	28	143	63.7	23.1
300	75	26	141	62.9	25.6
500	75	24	158	69.6	26.9
700	50	21	104	58.9	20.2
900	50	25	112	60.2	24.0

Origin of the blocks

The boulders in the valley-floor blockstream are subrounded with pitted surfaces and weathering rinds less than 2 mm thick. In contrast, the blocks that comprise the smaller blockfields in the tributaries are distinctly angular and surface pits are largely absent. The rock scarps in the upper sections of the catchment have generally rounded faces, but highly fractured, angular rock scarps and tors dominate the south-facing slopes.

Based on the block shape and weathering characteristics, the clasts found in the small blockfields are clearly derived from the rock scarps found in the upper-catchment. The blocks found in the valley-floor blockstream are significantly more rounded and weathered than those in the tributaries and two possible sources for the blocky debris can be envisaged. Besides an origin from surrounding rock scarps, corestones derived from spheroidal weathering mantles in the valley-floor and surrounding slopes could provide significant volumes of rounded clasts. Observations in road cuttings in Sani Pass and Kotisephola Pass, however, indicate that the degree of rounding and advanced chemical decomposition found in these profiles exceeds the rounding, weathering rind thickness and loss of strength observed in blocks from the blockstream. Instead, the blocks found in the blockstream are considered the result of cryoclastic debris production at the scarps.

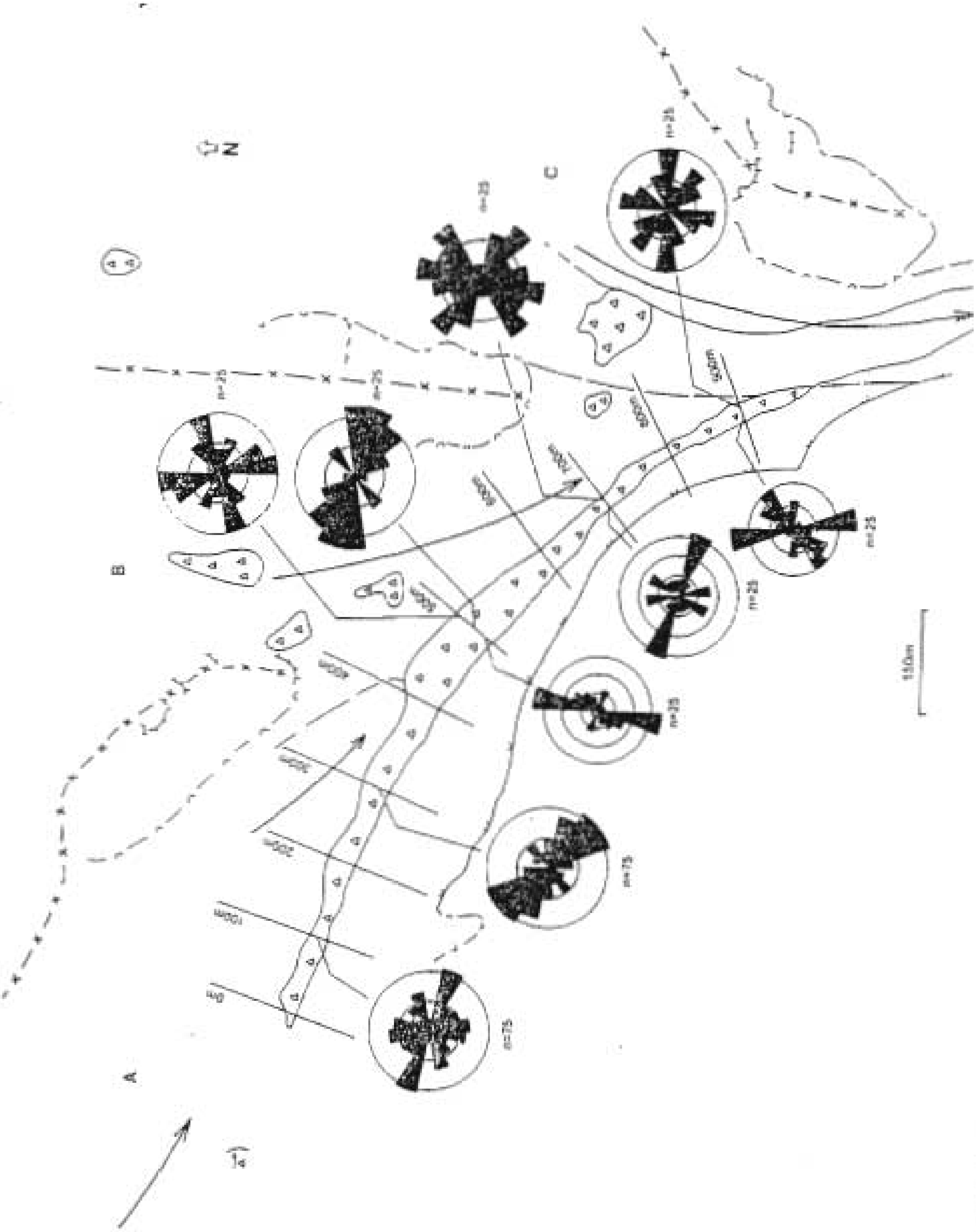


Figure 12. Location of the transects presented in Figure 11 and block orientation (ab-plane) plots. Each concentric circle in the fabric plot indicates a 10% frequency.

The rounding and pitting is interpreted to be the result of subsurface weathering of the diamicton before subsequent washing out of the matrix.

Site 3. Exposure in solifluction mantles
Following the valley in downslope direction a small stream emerges in summer from the terminal end of the blockstream. After another ±500 m (and further openwork material detached from the main body) one reaches the confluence of streams from catchments C and D with the main channel. Stream incision has resulted in exposure of the slope materials that blanket many of the valleys at these altitudes.

Headwall erosion of the central valley-floor deposits reveals a uniform, coarse, clast-supported diamicton at least 4 m in thickness. The washing out of matrix by the stream results in further residual openwork deposits. Geomorphological mapping indicates that this material constitutes the terminal zone of the mass wasting mantle from catchment C (Figure 10).

Exposures in the slope materials from catchment D occur along its stream which runs over bedrock. At the confluence the material is 15-20 m thick and composed of a massive granular loam. In the otherwise unstratified material, occasional thin layers contain a 10-30% abundance of small stones with *a*-axes up to 15 cm. In the uppermost 0.4 m larger boulders occur with *a*-axes up to 1.5 m.

Three hundred metres upstream from the confluence the stream in catchment D cuts through a 3 m thick deposit, including a stone-banked terrace, on its eastern bank. The stone-banked terrace reveals a 1.6 m thick clast-supported diamicton at its 8 m wide front. Clasts have *a*-axes up to 80 cm and dip at 22° in upslope direction. Contact with the underlying material is near-horizontal on a 8° slope. The deposit rests on 1.5 m of granular loam. At 0.9 m below the surface a thin layer of 20% clast abundance with clasts up to 25 cm. The feature is relict with soil development subsequent to deposition.

Blockfield development
The morphology of the openwork deposits, the clast fabrics and the low angled slopes on which they are found, resemble that of allochthonous blockstreams found at other mid-latitude environments, such as those in southeast Australia (Caine and Jennings, 1968), Tasmania (Caine, 1983) and the Appalachians (Potter and Moss, 1968). In all these cases an emplacement of the slope materials by slow mass wasting under periglacial conditions is proposed. Material fed from all four catchments into the central valley where most of the accumulation of blocky material took place. This accumulation took place as a diamicton with matrix present between the blocks as suggested by the weathering status of the blocks and the recessed cross profiles of the blockstream.

Movement mechanisms by slow mass flow require the build up of considerable pore pressure in the debris material. In the coarse materials at an 8° slope angle this is best explained with an impervious layer at the base which here is provided by bedrock. However, the coarse deposits drain readily even when assuming a matrix present between the blocks. Washburn (1980, p. 98) points out that sufficient hydrostatic head may be generated when drainage is impeded by a seasonally frozen surface layer. Such a mechanism has also been proposed by Caine (1983) for the movement of some blockfields in northeastern Tasmania.

Frost creep would provide an additional mechanism of movement of the upper zone, operating in close association with solifluction. Although no estimate of soil frost penetration can be given, Fahey (1974) showed 30 cm heave with seasonal frost to 2 m depth in very wet soil. Under such optimal conditions this would result in 1.76 cm/a (potential) creep on a 10° slope. Based on this mechanism alone, a clast from the upper catchment would travel about 1 km to the blockstream in almost 57 ka. This makes it very unlikely that frost creep is the only transport mechanism. However, the apparent concentration of blocky material in the upper sections of the exposures, with minimal presence of blocks lower down in the vertical profile, suggests that frost heave induced creep played some role in the movement of the material. This is further borne out by the localised, lateral sorting features present in the blockstream and blockfields.

Palaeoeonvironmental implications

The blockstream and related deposits described here most likely developed under a seasonal frost environment, with no permafrost required. Solifluction mobilised a well weathered regolith under conditions suitable for cryoclastic debris production at surrounding scarps. For significant segregation ice to develop sufficient moisture must have been present at the time of freeze up in autumn. Furthermore, sufficient water supply is a necessity for failure under high pore water conditions. On the other hand, to allow for a 1-2 m deep seasonal freeze-up, snow-cover must be largely absent to avoid thermal insulation of the ground surface. Under present-day conditions summer precipitation generates seasonal wetlands in the central parts of the catchment and tributaries. As precipitation amounts decrease in autumn drainage from these wetlands continues, thus providing suitable conditions for the mechanisms outlined above to operate under the right temperature conditions. This suggests that precipitation patterns were not significantly different than at present. This is in agreement with estimates by Partridge (1997).

Concerning the age of the deposits, the regolith material in which the blockstreams and solifluction mantles have developed may date back to the Early Tertairy. The uniform weathering status and angularity of the blockfield material points at emplacement of the, now relict, openwork deposits in the Late Quaternary. Conditions for sufficiently severe seasonal frost would have prevailed only during the Last Glacial, and particularly the Last Glacial Maximum, and it is during this period that the main phase of solifluction and frost creep took place. The apparent dominance of solifluction processes throughout the Lesotho highlands argues against widespread glaciation of the region during this period. Slope materials became largely stable in the Early Holocene with only surficial frost penetration occuring at present. The latter still generates frost creep and solifluction in the upper 0.4 m of the soil. However, low sediment yields due to limited sediment supply and increased vegetation cover in the Holocene result in suffosion and the current phase of stream incision in the colluvial mantles.

JB, SH, IM, PS

Day 3: Valley Asymmetry and Valley-head hollows in the Sani Pass Area.

Introduction

Two major features will be investigated on Day 3, namely valley asymmetry and the valley-head hollows. The study sites investigated are ideal in that they allow for a broad discussion of the Pleistocene palaeoenvironments. The main focus is on an unnamed valley approximately 3 km due south of the Sani Top Chalet. This is typical of the High Drakensberg and Lesotho mountains as it is asymmetrical with a steeper south-facing slope than the north-facing slope (Figures 9 & 13). Further, the north-facing slope of the valley has three large hollows that have been described as being either of glacial or nival origin (Marker & Whittington, 1971; Marker, 1991, 1994). Three sites have been selected for investigation:

Site 1 (Figure 14) is an ideal location for the observation of valley asymmetry, while providing an excellent view of the hollows that will be discussed at Site 2.

Site 2 (Figure 14) is located at the base of a hollow (identified as Hollow B by Marker & Whittington (1971) and Marker (1994, 1995)) where the sedimentary profile will be discussed together with the origin of the observed hollows. The adjacent hollow to the west of Hollow B will be observed *en route* to Site 3 (Figure 14).

Site 3 was chosen as it potentially represents a contemporary analogue from which the depositional environment of the hollows at Site 2 can be interpreted.

Site 1

Introduction

As indicated above, the valley 3 km to the south of Sani Pass is typical of the High Drakensberg and Lesotho mountains. Throughout the region, south-facing slopes have been found to be steeper than north-facing slopes (Meiklejohn, 1992). It has been previously argued that valley asymmetry in the High Drakensberg and Lesotho is the result of periglacial processes (Alexandré, 1962; Sparrow, 1964, 1971). Other research indicated that the observed asymmetry at lower altitudes is not necessarily the result of a periglacial environment (Garland, 1979; Boelhouwers, 1988). Meiklejohn (1992, 1994) used published material suggesting a periglacial environment (e.g. Alexandré, 1962; Sparrow, 1964, 1971; Harper, 1969; Linton, 1969; Hastenrath, 1972; Lewis, 1988a, 1988b), to propose that cryogenic processes may have resulted in the observed asymmetry.

Site Description

At Site 1 the average gradient of the south-facing slope is 20° while that of the north-facing slope is 15° (Figure 13). The north-facing slope of the valley has poorly developed thin soils with very little horizon development; only a humic A horizon can be identified except for the valley bottom (position "a", Figure 13) where a red apedal B horizon with clay illuviation is found. While the south-facing slope is also comprised of similar thin soils, two horizons can be identified where the soil layer is deeper. At location "x" (Figure 13) a cutanic B horizon, 70 cm deep, is found beneath a 35 cm thick A horizon, while a thin yellow-brown apedal B horizon is found at "y" (Figure 13). Saprolite, 70 to 80 cm thick, is found beneath a humic A horizon, of 60 to 70 cm, at locations marked "z" (Figure 13), where the soils are deepest.

The high clay content (Figure 13), presence of saprolite, as well cutanic and apedal B horizons indicate that the basalt weathering is active on south-facing slopes. Weathering on the south facing slopes is likely to be a response to the relatively high moisture conditions (Figure 13). However, the high proportion of clay minerals and saprolite indicate that the south-facing environment is not conducive to the removal of weathering products. It follows that observations on the south-facing slope support the basic premise of Meiklejohn's (1994) model that south-facing slopes are geomorphologically less active than north-facing slopes. Colluvial material near the base of the south-facing slopes (Figure 13) indicates some form of mass movement.

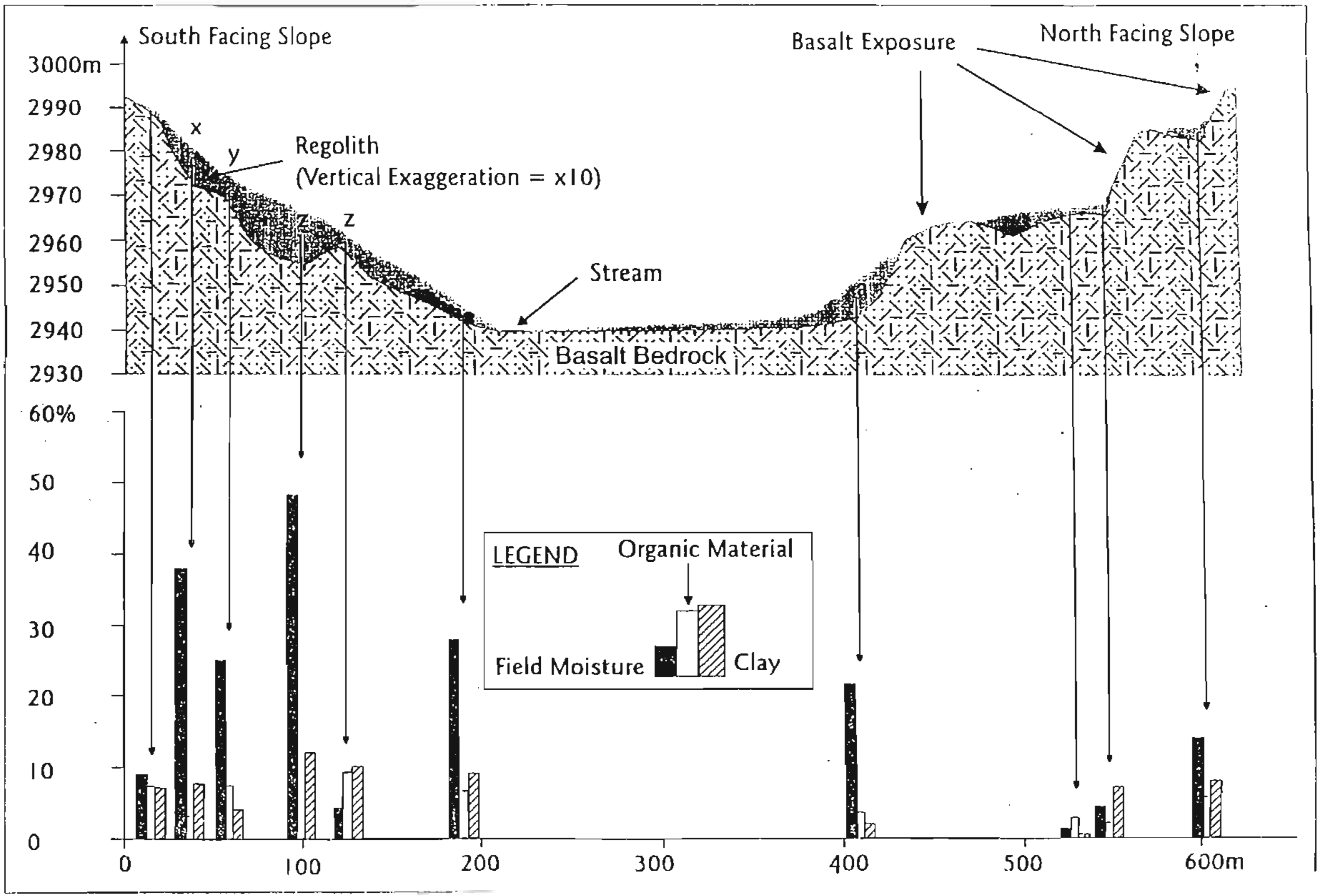


Figure 13. Valley profile and sedimentary analysis for a valley approximately 3 km south of Sani Pass, Lesotho.

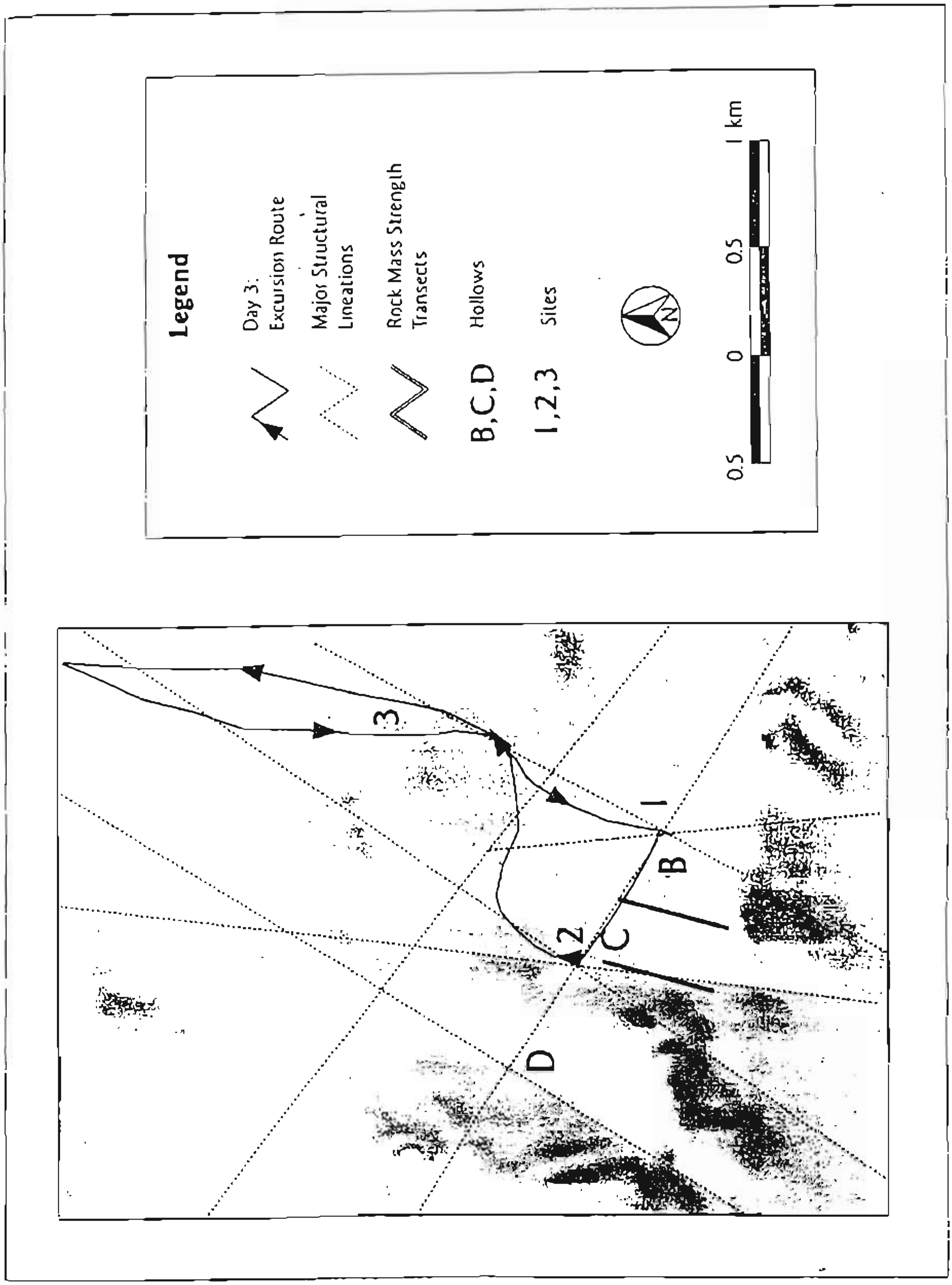


Figure 14. Route map for day 3; Rock Mass Strength Transects and major Lineaments at the study sites.

Given that the valley floor stream channel is located so close to the base of the south-facing slope, it is apparent that the north-facing slope has undergone considerably more retreat than the opposite slope. If the north-facing slopes are more active, then it is surprising that the only evidence of products of mass wasting from the north-facing slopes is the accumulation of sediment at the base of the observed hollows. The implication of this observation is that the environment on the north-facing slope has produced weathered material that is easily removed and transported away. The susceptibility of the Drakensberg basalts to weathering by related moisture processes (Van Rooy and Nixon, 1990; Van Rooy, 1992; Van Rooy and Van Schalkwyk, 1993) supports this, especially as the weathered material is generally comprised of fine, often clay-sized, particles. Further, given the finer size of the weathered product, it is unlikely that any accumulations will be found in fluvial systems downstream.

Palaeoenvironmental Interpretation

Throughout the year, north-facing slopes receive more direct incoming solar radiation than south-facing slopes in areas south of the Tropic of Capricorn. Given that this will result in relatively high temperatures on north-facing slopes, chemical weathering processes will be faster here than on the south-facing slopes. The increased rate of weathering will, in turn, result in more weathering products being available for transport and thus potentially lead to faster denudation of the north-facing slope. This assumption is, however, based largely on the premise that moisture is available for the weathering to take place.

It is apparent, from qualitative visual interpretation, that both the north- and south-facing slopes are currently relatively inactive. Given that soil moisture content is higher on south-facing slopes than on north-facing slopes and that this is unlikely to have been different in the past, it would be logical to argue that weathering and erosion are enhanced on the south-facing slopes. This would result in the south-facing slopes having a shallower gradient than north-facing slopes. However, the opposite is true and weathering processes are more active on north-facing slopes as seen in the sediments in the base of hollows. Two possible scenarios may explain this; first, the moisture availability on the north-facing slopes was more dynamic than on the opposite slope, thus resulting in a relatively high rate of weathering and increased debris production, together with high rates of debris removal. Alternatively, or simultaneously, the moisture on the south-facing slope was in a "non-mobile form"; this may indicate frozen slope conditions. A frozen south-facing slope would mean that the moisture regime was not dynamic and thus relatively inactive in terms of geomorphic activity. Climatic amelioration after the Last Glacial Maximum would have resulted in the south-facing slope becoming more geomorphologically active in terms of weathering, thus, contributing towards the formation of the soils observed on these slopes. It is during this amelioration that weathered bedrock or residuals could be mobilised and result in the observed mass movement features (Figure 13). The clay minerals, saprolite and clay-rich B soil horizons present on the south-facing slope are likely to be a function of the relatively high moisture contents found on this slope.

Site 2

Introduction

Three hollows in a valley 3 km due south of the Sani Top Chalet have previously been investigated by Marker & Whittington (1971) and Marker (1991, 1994, 1995) and it was suggested that they are of nival or glacial origin from lee-side snow accumulations. In each of the three hollows, the basal sediments have been incised, providing an ideal opportunity for description and an investigation into their origin, despite their depths not being determined in this case. The incision of the sediments in the hollows near Sani Pass is likely to have arisen as a result of increased runoff. This could either be due to the impact of overgrazing from cattle and sheep or from increased precipitation (i.e. during a wetter period) or a combination of both.

The argument that the hollows were formed by nival or glacial processes is based on a model proposing that mid-latitude cyclones extended far north during the cold periods of the Pleistocene, resulting in high precipitation in the form of snow (Marker, 1991, 1994, 1995). This in itself is a difficult argument to conceptualise, considering that a pre-existing hollow would be required for snow accumulation, especially in the case of lee-side accumulation on a north-facing slope. The north-facing slope receives more incoming solar radiation than any other slope and snow accumulation would arguably melt first here and rather accumulate on south-facing slopes. In addition, evidence suggests that during the Last Glacial Maximum, precipitation was only 70% of present (Partridge, 1997), making large accumulations of snow on any slope highly improbable. In contrast, Grab and Hall (1997) have proposed that the hollows may be the result of bog-cirque development. If this were the case, then it would be expected that there would not be a sharp contact between sediments and basalt bedrock at the base of the hollows. Moist conditions for a prolonged period at the base of a hollow would result in considerable in situ weathering, especially as Drakensberg basalts are prone to breakdown through related moisture processes (Van Rooy and Nixon, 1990; Van Rooy, 1992; Van Rooy and Van Schalkwyk, 1993). In reality, the hollow sediments have a sharp contact with the bedrock. In situations where the hollow sediments have not been incised by fluvial action, relatively impermeable clay layers would prevent substantial moisture seepage through to the bedrock, thus resulting in a protected rather than an active weathering environment.

Description and palaeoenvironmental interpretation

The sediments at the base of the Sani hollows have been described in some detail (Marker & Whittington, 1971; Marker, 1991, 1994, 1995) and those at Site 2 were further investigated as part of the research for this publication. Essentially, the sediments show a sharp contact with the bedrock; this can be clearly seen downstream from the described profile at Site 2. Further, they show organic peat accumulations, with alternating coarse fluvial and fine aeolian sedimentation in between (Figure 15). The described section in hollow B has two peat layers, while in a gully in the adjacent hollow to the west, up to five peat layers were identified. Downstream from the described section at Site 2, is a layer of large (10-30 cm diameter) clasts within the fluvial sediments in the sequence; this indicates an energetic fluvial regime. Occasionally a relatively impermeable clay-rich layer is located within granular fluvial sediments (Figure 15). From cored samples, a further clay layer is situated beneath the visible section. The formation of ferricretes above the clay layers in the profile (Figure 15) is an indication of a variable water table.

Marker (1994, 1995) has published data regarding material that has been dated from this site. The oldest dates are obtained from organic sedimentation at the base of the hollows are from 13 490 BP and possibly

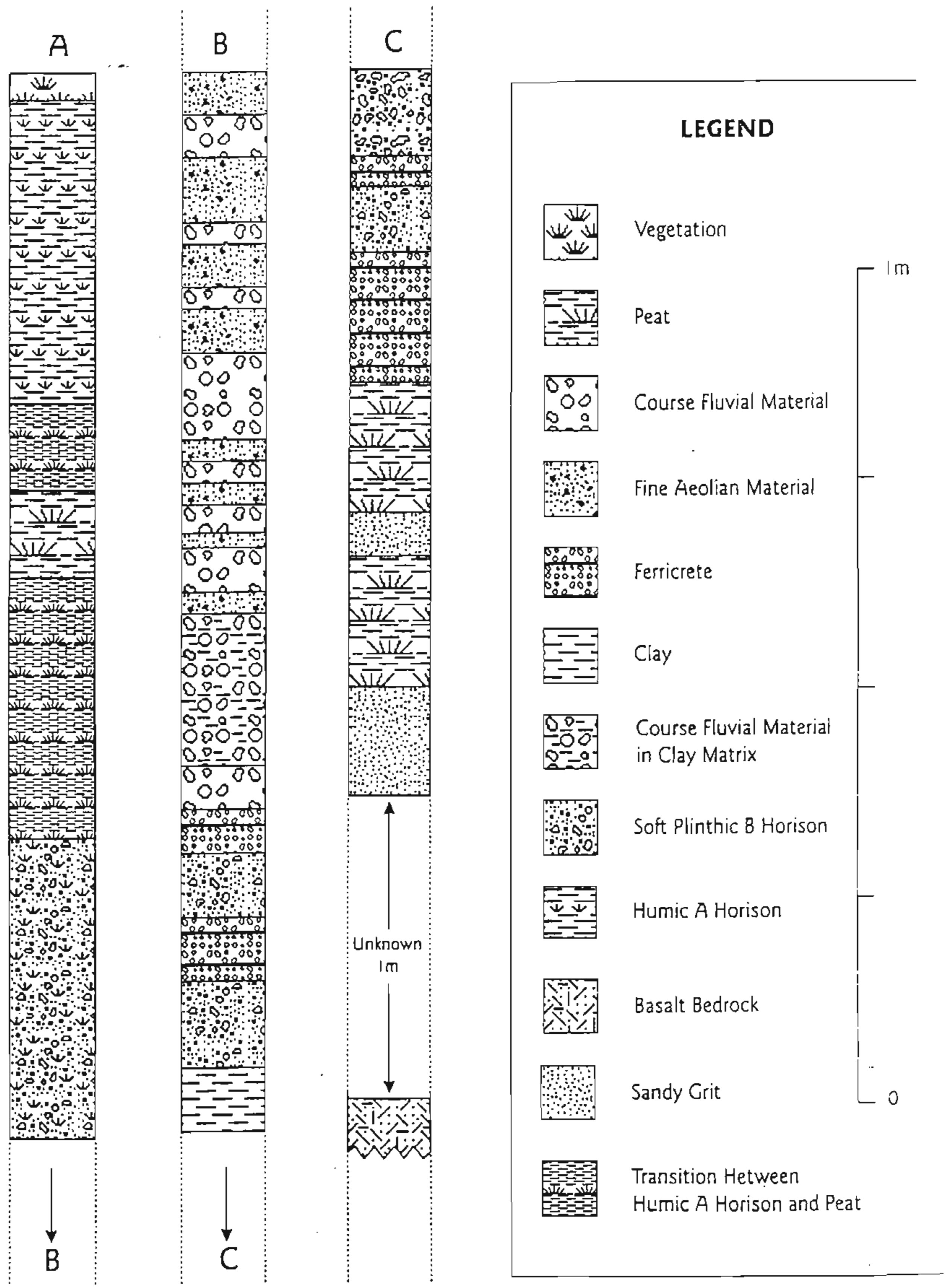


Figure 15. Sedimentary sequence at Site 2.

represent the rapid amelioration of temperature after the Last Glacial Maximum (Marker, 1994, 1995). The upper layer of peat is said to have formed during a warmer and wetter period between 9000 and 5000 BP (Marker, 1994, 1995). In-between, alternating fluvial and aeolian deposition is said to represent a drier period that existed between 5000 BP and 1000 BP (Marker, 1994, 1995). However, while this section may be relatively easy to interpret, the increase in the number of peat layers in the adjacent hollow to the west, indicate that the interpretation of the sediment in the hollows at Sani Pass may be more complex than initially suggested. Similar, but younger sediments have been described at Tlaeeng pass, approximately 110 km north-west of this site (Marker, 1994, 1995).

Currently, no published data are available that suggest that sediments in valley-heads and hollows in Lesotho are older than the Last Glacial Maximum. This either suggests that the area was protected or inactive during the coldest part of the Pleistocene. However, the absence of datable material preceding the L.G.M. may be the result of an inactive geomorphic environment, or an environment with limited vegetation cover. Alternatively, it is possible that increased precipitation during the relatively wet period immediately following the Last Glacial Maximum could have resulted in the removal of much of the material that accumulated during the colder period. Data suggesting that the Last Glacial Maximum was relatively dry and 6-10°C colder than present (e.g. Partridge, 1997) support the argument that this period was relatively inactive. This is especially so if it is considered that much of the in situ breakdown of the basalt bedrock is the result of moisture related processes.

If, as it is argued above, the hollows at Sani and all the others identified by Marker (1994, 1995) do not owe their formation to the action of snow or ice, then another hypothesis regarding their origin should be sought. The answer may lie in the underlying bedrock structure. The rock mass strength technique (Selby, 1980; Moon and Selby, 1983, Moon, 1990) was utilized as it was thought that it may provide some evidence for the location of the Sani hollows. A hollow (Hollow C described by Marker, 1994), and the interfluve between two hollows were investigated (see Figure 14 for the location of transects) using the rock mass strength technique (Figure 16). It was found that while the entire north-facing slope was slightly over-steepened, there was relatively more over-steepening in the hollows than at the interfluves (Figures 16 & 17). This observation possibly indicates recent slope development and that the slopes will still undergo change. However, the qualitative observation made above, that the slopes are currently relatively inactive, does not support this hypothesis. The rock mass strength technique did not, however, prove to be entirely conclusive in this case and further transects are required before any conclusive derivations can be made. The only real difference found between hollows was that the joints in the hollows are more continuous and exhibit a more defined orientation.

Joints, dolerite dykes, and kimberlite dykes in the Drakensberg volcanics show very strong orientations (Dempster and Richard, 1973). From fieldwork, aerial photographic analysis and digital terrain modelling, the positions of the main lineaments in the area of the hollows were plotted (Figure 14). It then becomes apparent that the basal layers of the hollows are exactly where these lineaments intersect

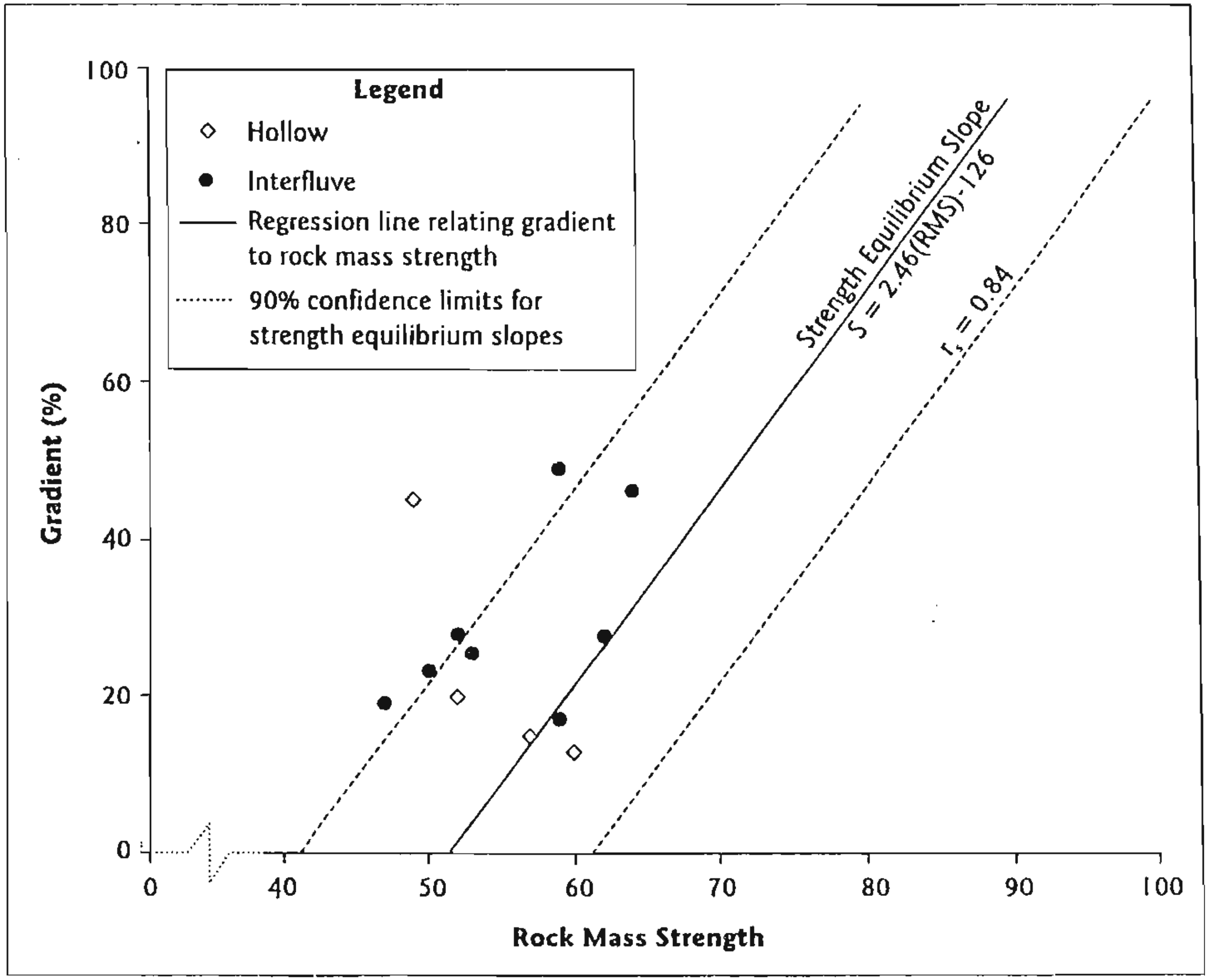


Figure 16. Rock slope classification using rock mass strength and gradient of Hollow C back wall and the interfluve between Hollow B and C (as identified by Marker, 1994).

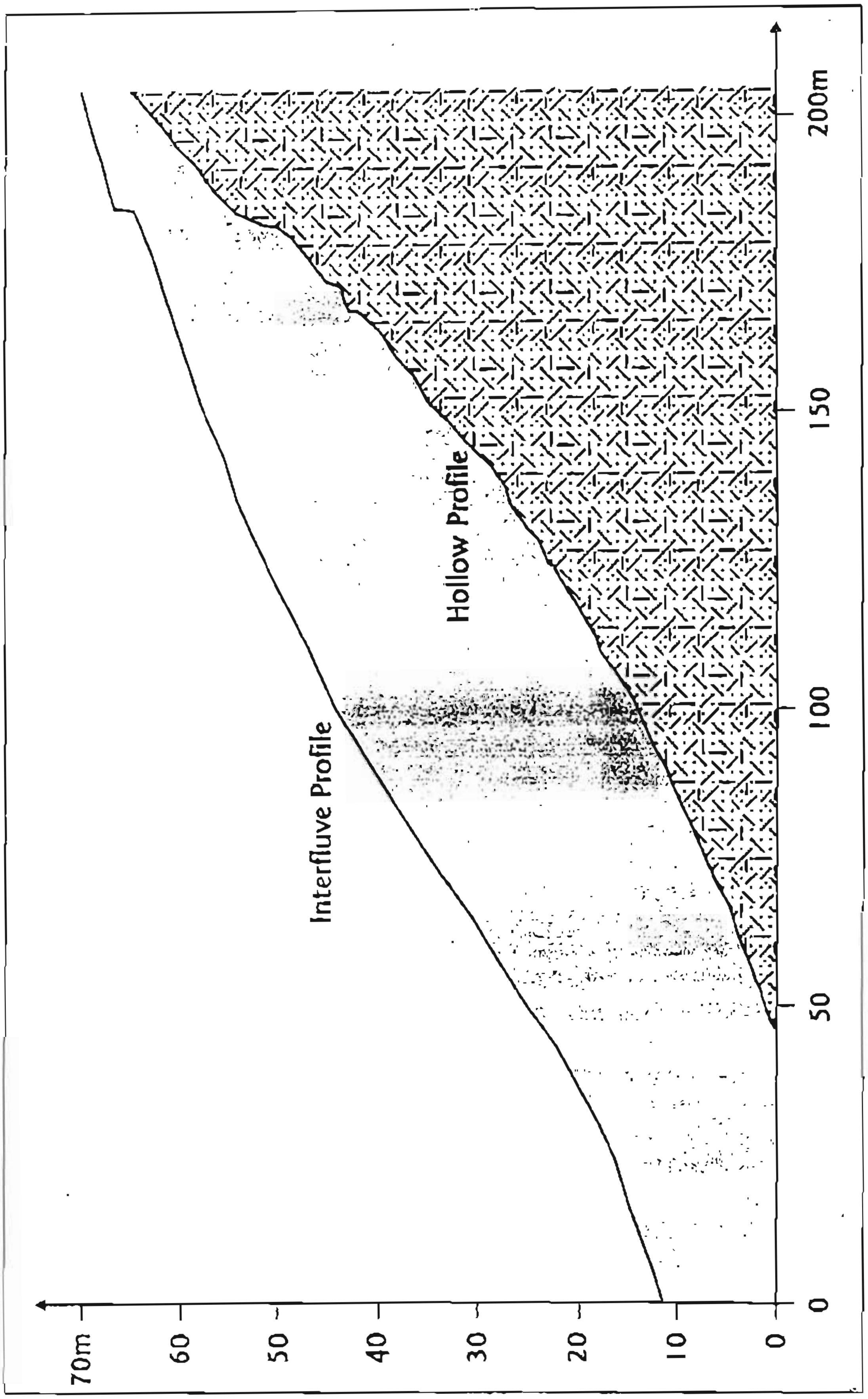


Figure 17. Profiles of Hollow C back wall and the interfluve between hollow B and C (as identified by Marker, 1994).

(Figure 14). Given that these are likely to be major joints, it is possible to postulate that these are avenues of easy moisture access and thus enhanced weathering. While this provides an answer for the location of the hollows, it does not indicate the specific processes responsible for the formation of the hollows.

Site 3

Introduction

A site approximately 1.5 km south of the Sani Top Chalet was investigated as this locality potentially provides a suitable contemporary analogue for the formation of sediments located in hollows at Site 2 and elsewhere in the High Drakensberg and Lesotho mountains.

Description and Interpretation

A core of 3 m was obtained from Site 3, which provided a comparison with the sediments at Site 2. The sedimentary sequence obtained from Site 3 (Table 3) follows a similar pattern to that at Site 2 and other hollows documented from the High Drakensberg and Lesotho, in that coarse fluvial sediments alternate with organic rich sediments. The surface consisted of course fluvially deposited material to a depth of 50 cm. From 50 to 80 cm organic rich sediments comprising a high proportion of clay and silt particles were found. It is proposed that these sediments would eventually form a peat layer. Below 80 cm, to a depth of 2.5 m the sediments were similar to the surface sediments, with some evidence of ion oxidation and ferricrete mineralisation, similar, but less dramatic than Site 2. At a depth of 2.5 m a further organic rich layer of approximately 20 cm was found above coarse fluvial sediments. The difference between this site and Site 2 is that the surface is comprised of coarse fluvial sediments deposited in an "outwash" area rather than an organic rich A horizon. Further, the organic O horizons have not fully developed into peat as in Site 2. In a dynamic system where there are braided stream channels, such as Site 3, the cycle between an area being vegetated and then denuded of vegetation could be as short as a single season in a particular year. In this case, the coarse fluvial material may represent an active fluvial environment rather than a dry period. The layers of peat may thus represent more complex cycles than that proposed by Marker (1994, 1995). More rigorous dating of the sediments is required before any conclusions regarding climatic cycles can be made.

Conclusion

It is apparent that valley forms in the High Drakensberg and Lesotho mountains require further investigation, before they can be used as palaeoenvironmental indicators. Published data currently available provide more questions than answers and often consider isolated features out of context. There is, thus, a need for further research and possible locations for such investigation are the thick, valley-bottom sediments; an example of which can be seen in south-facing hollows to the north of the Sani Top Chalet. From aerial photographic analysis it is apparent that similar sediments are located throughout the high Drakensberg and Lesotho mountains. It is likely that snow would have accumulated here rather than in north-facing hollows, particularly during more arid periods than at present. Future research should thus be directed towards these sediments as they may provide some evidence to the role of snow or ice in the landscape development in the high regions of southern Africa. Research may even indicate that the contribution of snow and ice in this regard is insignificant Until more data are collected and analysed from a broader perspective than past and current research, no definite conclusions will be forthcoming.

IM, AF, MB, GV

Table 3: Description of the Sedimentary Sequence at Site 3, Day 3

Layer	Depth	Description
1	0 - 0.5cm	Coarse fluvial sediments
2	0.5 - 0.8cm	Fine, organic-rich sediments
3	0.8 - 1.8m	Coarse fluvial sediments
4	1.8 - 2.2m	Coarse fluvial sediments with ferricrete mineralisation
5	2.2 - 2.5m	Coarse fluvial sediments
6	2.5 - 2.7m	Fine, organic-rich sediments (more "peat-like" than layer 2 above)
7	2.7m -	Coarse fluvial sediments

Day 4: Road sections in Sani Pass

Location

Exposures of the slope materials and underlying weathering mantles in the basalt are within 20 minute walking distance from the Sani Top chalet at 1.7 km from the top of the pass.

Significance

The exposures in Sani Pass provide the only easily accessible sites to study the slope deposits and basalt weathering profiles in the cutbacks along the Drakensberg Escarpment. Discussion on the conditions in some of the cutbacks during the cold phases of the Pleistocene have centred around the origin of the extensive mass wasting deposits at the base of the cliffs (Sumner, 1995; Hall, 1995) as well as possible niche glaciation (Hall, 1994; Grab, 1996a). No detailed sedimentological work has been carried out on the cutback deposits to date and no ages have been established. The sites described here will hopefully stimulate further debate on these issues.

Site descriptions

The three road sections presented here cut through the steep debris mantles that alternate with rock scarps beneath the cliffs of the main Escarpment. The slopes are orientated towards the east and southeast and have gradients around 32-34°, which must be close to the angle of internal friction of the debris. Sketches of the three exposures are presented in Figure 18.

Site 1a

This section has been subdivided into five units separated by indistinct boundaries. The lowermost unit is comprised of a massive, matrix- to clast-supported diamicton with large clasts. Unit two makes up most of the section and contains a grey-yellow, crudely stratified, matrix-supported diamicton with occasional thin gravel lenses and scattered large clasts. Unit three is a massive clast-supported diamicton, truncated by a matrix-supported diamicton similar to unit two. The dark-grey colour in unit four represent, for local conditions, well developed buried soil horizons. The uppermost unit represents a grey, massive, matrix-supported diamicton with current soil development.

Site 1b

This exposure presents a longitudinal section through similar slope material as at site 1a. Units one and three are massive, grey yellow, diamictons with relative low clast abundance, separated by a layer with high clast content. Pockets of small clasts are present immediately upslope from large blocks in this unit. Unit four represents a buried soil horizon in loam with 20% abundance of small clasts. This unit is superceded by similar crudely stratified

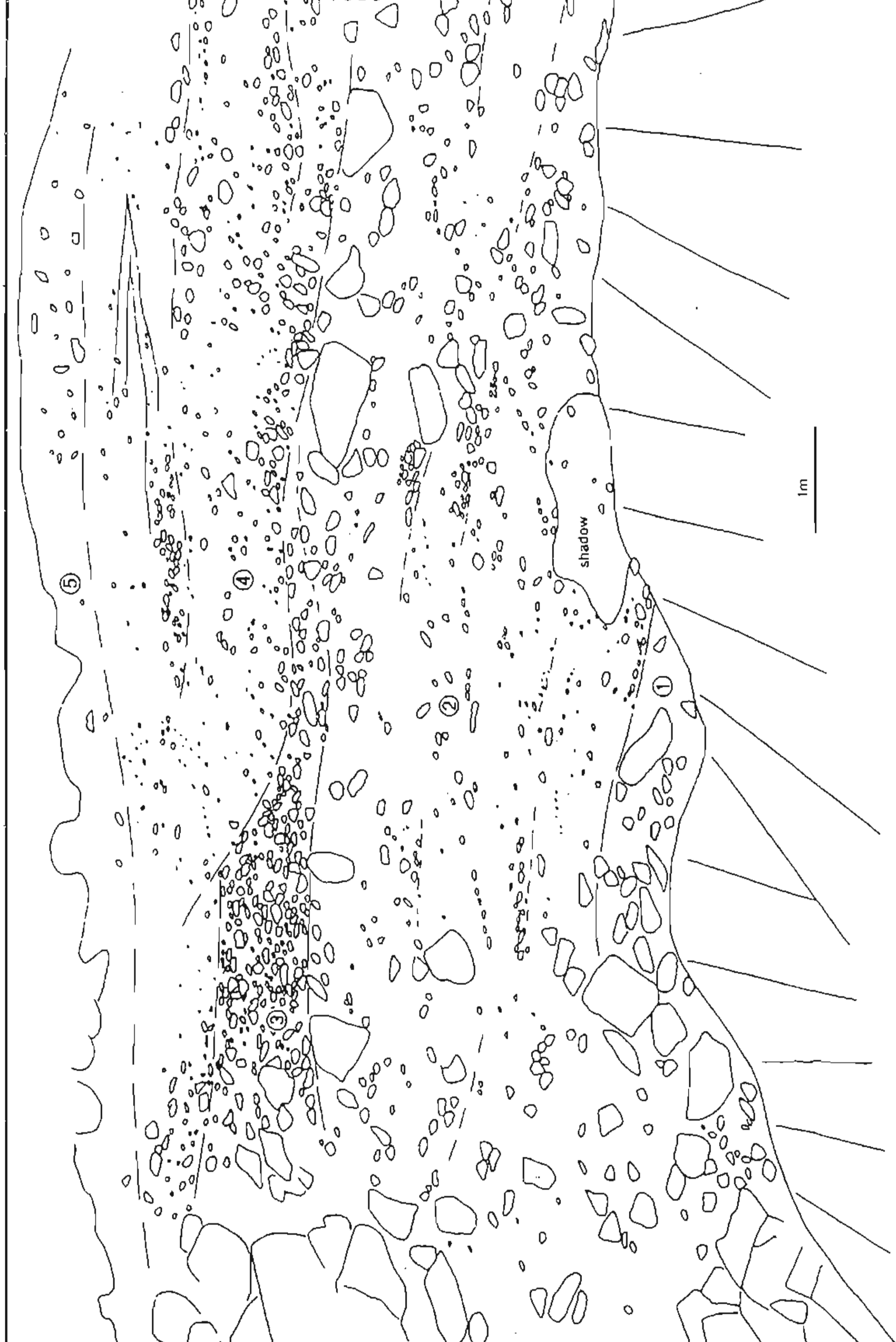


Figure 18. Sketch diagrams of road sections a - b - c, in the Sani Pass.

Figure 18. Sketch diagrams of road sections a - b - c, in the Sani Pass.

material of lower organic matter content. A decrease in clast content separates unit 5 from 6. Clasts contained in the material at this site show a distinct *a*-axis alignment parallel to the slope.

Site 1c
This exposure further illustrates the uniform nature of the slope material contained at all three sites. Four units are distinguished in this 3m high exposure. Units one and three comprise the grey-yellow, crudely stratified, matrix-supported diamicton that dominates all three sections. Clast-supported diamictons create lenses that pinch out laterally (unit 2). These alternate with zones of distinctly low clast abundance. The uppermost unit comprises a pocket of large clasts near the surface.

Interpretation

Dominant in all sections are the matrix-supported diamictons with high percentages of gravelly loams as matrix. Since the deposits are relatively uniform, the units described indicate no major temporal succession in lithostratigraphic facies. Crude stratification is present in the form of lenses of clast-supported diamictons of a metres wide and long and 50 cm thick that pinch out both in lateral and longitudinal profiles (units 1a-3; 1b-2; 1c-2). These coarse lenses are interpreted as small debris flow deposits that redistribute material from upslope (Eyles et al.,1988; Nieuwenhuizen and Van Steijn, 1990; Hinchliffe et al., 1998). The alternating zones of thin gravel lenses and layers of low clast abundance (units 1a-2,4,5; 1b-3,5,6; 1c-1) indicate sedimentation by surface wash, reworking unvegetated slope material, including fines from recently deposited debris flows, during rainstorms (Hinchliffe *et al.*, 1998). The large clasts found scattered in the deposits are considered the result of rockfall, but make only a small percentage of the total debris mass. The buried soil horizon at sites 1a (unit 4) and 1b (unit 4) relates to a phase of reduced sediment deposition conducive to soil development.

Paleaoenvironmental significance

The most outstanding characteristic in these deposits is the large percentage of matrix present in the material. The angularity of the clasts indicates that the bulk of the fines must come from sources other than post-depositional weathering of clasts in the debris mantle. Weathering profiles in the road sections of the pass indicate that a substantial amount of fines can be derived from in situ weathering profiles on the steep slopes beneath the escarpment. A significant percentage of fines must however have been derived from the cliffs themselves. The importance of fines production at cliffs and their contribution to talus slope development has also been stressed by Hétu (1992), Salt and Ballantyne (1997) and Hinchliffe *et al.* (1998). Considering the relative dominance of fines in the deposits it is suggested that chemical weathering processes have played an important, if not dominant role in cliff weathering. On the other hand, the reworking of slope materials exposed here do not allow for statements to be made on the significance of mechanical weathering at the cliffs or any temporal variations in such activity.

Processes implied in the debris production and subsequent deposition point at environmental conditions not significantly different than at present. None of the lithostratigraphic characteristics found in the deposits point at significant periglacial activity at the time of deposition. However, the restriction of buried soil horizons to the upper section of the profiles indicates an increased slope stability by a newly establishing vegetation cover. It is worth noting that the darker colour in the palaeosol at sites 1a and 1b, in comparison to the recent A-horizon, suggests climate conditions amenable to better conditions for soil development than at present. The entire sequence of material is considered of Holocene age, but absolute dating of the palaeosol may prove useful to estimate rates of material accumulation. **JB**

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INTERNATIONAL PERMAFROST ASSOCIATION

MULTI-LANGUAGE GLOSSARY of PERMAFROST and RELATED GROUND-ICE TERMS

in

Chinese, English, French, German, Icelandic, Italian

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Compiled and Edited by:

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Chapter 8

Glacial Landforms and Processes

"Are you really sure a floor can't also be a ceiling?"[1]
M.C. Escher

Material presented in this section relates to, directly or indirectly, studies of glaciers, glaciation and/or glacial sediments - both ancient and modern. Although obviously not directly connected with periglacial processes, there are many components of glaciation that do affect the periglacial realm. Periglacial activity, by its very definition, takes place around the glaciated area and thus a knowledge of ice cover extent can help define where and what may take place. Equally, many of the periglacial landforms, particularly those associated with sorting, occur as a result of the glacial debris, left during deglaciation, in which the cryogenic processes can operate. Indeed, this is a central issue in periglacial geomorphology - the question of effectiveness of periglacial processes in areas

[1]In reference to the findings relating to boulder beds cited in this section (Visser & Hall, 1985)

that have never been glaciated. Such areas are relatively rare and so our cognizance of the periglacial realm is really prevaricated upon that area have been "prepared" by glaciation. Some of the work presented here was undertaken in the context of defining the periglacial area based upon a knowledge of glaciation. Other research, particularly that related to ancient glaciations, was not for this reason but the information is appropriate to this thesis within its broadest context.

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- Hall, K., Arocena, J. and Smellie, J. In Press a. Analysis of weathering rinds and reconstruction of palaeoenvironmental conditions from weathering rinds on clasts in the Cape Roberts drillcore. *Terra Antartica*. (given in Chapter 5, p743-757)
- Robinson, P.H., Pyne, A.R., Hambrey, M.J., Hall, K.J. & Barrett, P.J. 1987*. Core log, photographs and grain size analyses from the CIROS-1 drillhole, western McMurdo Sound, Antarctica. *Antarctic Data Series, Antarctic Research Centre, University of Wellington*, 14, 241pp. [This

document is provided as a separate volume to the thesis owing to its size]

- Scott, L. and Hall, K.J. 1983. Palynological evidence for pre-glacial vegetation cover on Marion Island, sub-Antarctic. *Palaeo -geography, -climatology, -ecology*, 41, 35-44.
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The first six papers all deal with sub-Antarctic Marion Island, the area in which studies were also undertaken of periglacial landforms and processes (see Chapter 7). These papers cover the reconstruction of the ice cover and the climate of the time together with details regarding the glacial landforms and sediments. The penultimate of those six papers deals with a new hypothesis: that deglaciation initiated volcanicity. A very strong relationship between faulting, the associated volcanism, and the reconstructed distribution of glaciers was found for Marion Island. As deglaciation was considered to be very rapid, due to the climatic marginality of the area, it is hypothesised that faulting took place in response to the rapid ice loss and that, as this is an island on a mid-oceanic ridge, so volcanicity ensued. Hall (1984b) attempts a reconstruction of ice cover on Iles Kerguelen, another (but much larger) sub-Antarctic island located closer to the Antarctic Convergence. Scott and Hall (1983) deals with pre-glacial vegetation and climatic conditions on Marion Island as deduced from palynological studies whilst Hall (1990b) is an invited review paper dealing with reconstruction of glaciation throughout the Southern Ocean between Longitudes 0° and 180°. The two papers Hall (In Press d & e) comprise invited contributions (also available on the web: http://members.aol.com/inqua5) to present information and maps (presently available only via the web) that outline the nature and extent of Quaternary glaciations for these two areas. As part of a world-wide undertaking, these two contributions attempt to show the nature and extent of glaciation as indicated by available data and references. As such they are meant to be a 'state-of-available-knowledge' and I was asked to undertake these two areas based on my work in those geographic areas.

The remaining papers deal with interpretation of, and reconstructions from, the Permo-Carboniferous Dwyka tillite of southern Africa, and details of the glaciogenic sediments associated with a drill core (CIROS) obtained in the McMurdo Sound region of Antarctica (Hall, 1989c and Robinson, *et al.,* 1987). The CIROS work was associated with reconstruction of glacial conditions in this part of Antarctica since the Oligocene and was, at that time, the deepest bedrock drilling in Antarctica. The major undertaking of the CIROS work was the detailed logging of the retrieved core together with the preliminary reconstructions (Robinson, *et al.,* 1987). This was a substantial field undertaking as it required both logging and interpretation of the whole core during retrieval under Antarctic conditions. In addition to the work cited here, one paper (Hall and Bühmann, 1989 cited in Chapter 5) undertook a reconstruction of weathering conditions based upon the weathering rinds on clasts from this core; the first application of this new approach developed by Hall. A follow-up to this work was that of Hall *et al.* (In Press a) (also cited in Chapter 5) within which a similar analysis of weathering rinds was undertaken on clasts from the Cape Roberts drill core; the results provide further justification of this technique. It is suggested that this novel approach may offer the opportunity to obtain proxy data for paleo-terrestrial where no other land-based data is available.

The Dwyka papers comprise a number of detailed studies of tillites in South Africa. Two of these are particularly significant. Visser and Hall (1984), by application of some new approaches, provide a totally revised reconstruction of ice flow directions in this part of southern Africa; indeed, the flow directions were reversed. This paper provided new insights into both interpretation of glacial sediments and to palaeo-ice flow directions in Gondwana-time South Africa. Visser and Hall (1985), via studies of boulder beds in the Dwyka tillite, provide a new interpretation of boulder bed genesis and meaning in general (hence the quote at the Chapter start). Via data from the Dwyka tillite it was possible to show that the generalized assumptions regarding the origin and stratigraphic interpretation of boulder beds need not be correct; indeed, a multitude of possible alternatives were provided. This paper provided the foundation of a number of subsequent studies by various authors working on boulder beds of more recent origin. The other two Dwyka papers provide new interpretations of the meaning of clast size and shape with respect to glacial sediments. Here the somewhat generalized assumptions regarding clast size-shape relationships were questioned. By use of data from the South African tillites it was possible to show the oft-stated attributes of shape with size were not only wrong but were conceptually in error. New insights into the size-shape argument with respect to glacial clasts were provided. Again, although the work was derived from the Dwyka tillite, its significance is to glacial sediments in general.

ANTARCTIC GLACIAL HISTORY AND WORLD PALAEOENVIRONMENTS

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Evidence for Quaternary glaciation of Marion Island (sub-Antarctic) and some implications

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CONTENTS

ABSTRACT

Marion Island is at present located only 2° latitude north of the Antarctic Convergence. Besides former palynological and geological proof for a glaciation of Würmian-Wisconsin age, further evidence has been found to indicate that Marion has been subjected to glaciations of both Würm and Riss age. Large areas of drift have been recognised for the first time together with a number of glacial landforms. Work so far undertaken suggests that there was a central ice cap from which a number of glaciers radiated. Till fabric analysis, striation observations and landforms have enabled a reconstruction of the glacier distribution on the northeastern side of the island to be undertaken.

The evidence from Marion is compared with that from other sub-Antarctic islands and from ocean floor sediments. The relationship of the findings to the Quaternary climate of the island is suggested. Evidence for sea level changes is also given and compared to that found on other sub-Antarctic islands.

INTRODUCTION

Marion Island (lat. 46° 54′ S, long. 37° 45′ E) is a roughly oval volcano of some 290 square kilometres, rising to 1 230 m, located in the southern Indian Ocean only some 2° latitude north of the Polar Front (Fig. 1). The radially faulted island is composed of older grey basaltic lavas dating from 276 000 BP (± 30 000) and younger black lavas, and associated scoria cones, dating from 15 000 BP (± 8 000) (McDougall, 1971). Considering the age of the oldest grey lavas Verwoerd (1971) suggested that the island could have been subjected to the southern equivalents of the Riss and Würm glaciations. Pollen analysis by Schalke and van Zinderen Bakker Sr (1967) showed that the island had indeed been subjected to a cold phase which was approximately coeval with the Würm of the Northern Hemisphere. Physical evidence for the existence of glaciers during this cold phase was found by van Zinderen Bakker Jr and Huntley in the form of striated pavements and glacially moulded grey lava outcrops (Verwoerd, 1971). The black lavas show no signs of glacial action and so are considered to post-date the glacial episode. Verwoerd (1971) suggested that any moraines that had been formed were at present below sea level and that the diamicts found at several places on the coast were more likely of volcanic, rather than glacial, origin. Thus although it was known that Marion had been glaciated at least once, the nature and extent of that glaciation was unkown, as too was evidence for the earlier glacial period.

The special research programme to investigate the glacial history of Marion Island was initiated in December, 1975 and the first results give evidence to:

1. substantiate two major periods of ice growth,
2. show the distribution of glaciers for part of the island, and
3. describe the climatic variations experienced in this area during the last glaciations as shown by glacial landforms and deposits.

The post-glacial faulting and lava flows have obliterated all evidence of the glacial episodes from the northern and western sections of the island. The

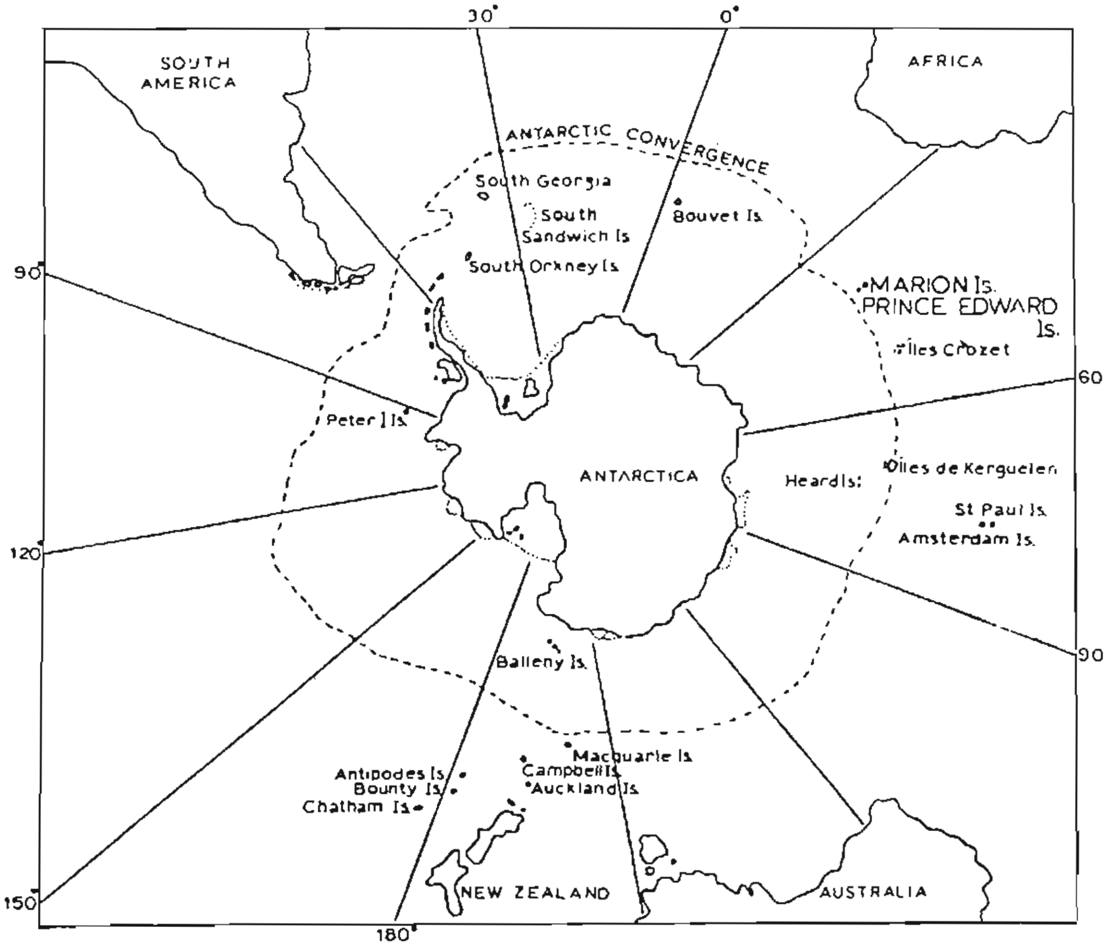


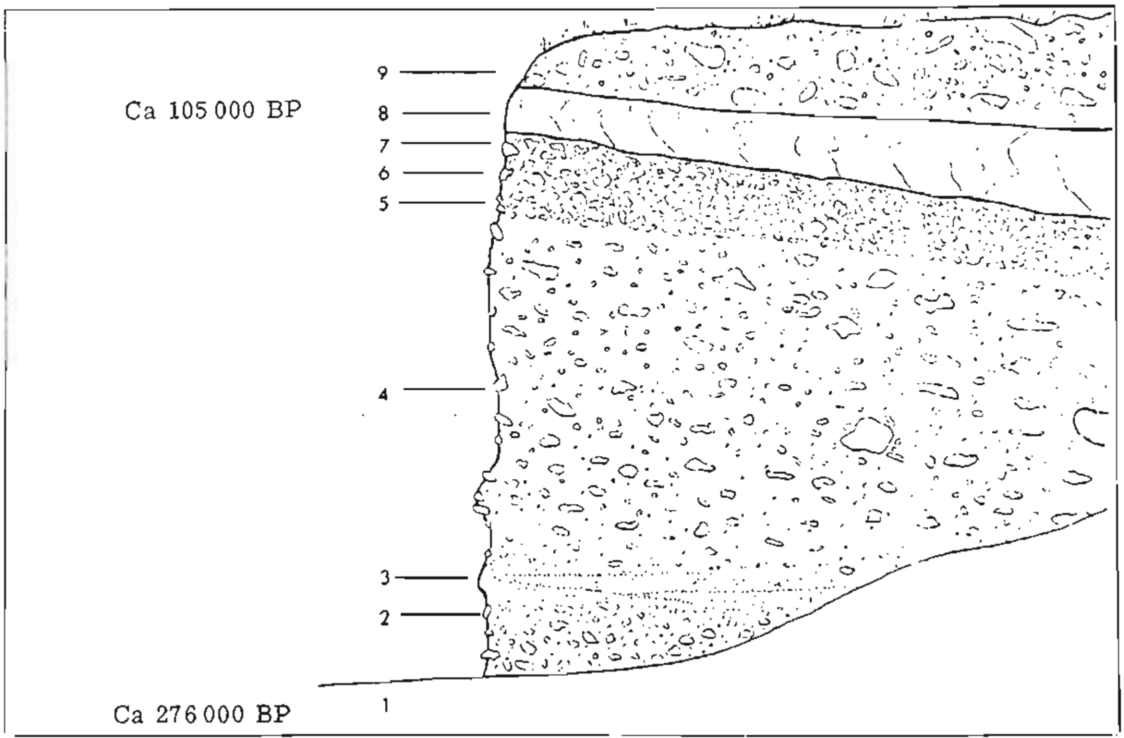
Fig. 1. Locality map, showing the position of Marion Island with respect to the Antarctic Convergence.

main area in which glacial deposits and landforms are exposed stretches from the northeast, at Long Ridge, down to the southern coast at Greyheaded Albatross Ridge. By reference to several specific locations within this area comments on the glacial history of Marion Island can now be given.

TECHNIQUES USED

A number of simple but complementary field techniques were employed in the study. Till fabric analysis was undertaken on many coastal cliff sections and along the banks of streams that were incised in till deposits. A high density of sample points was used with measurement of 50 stones at each point. The orientation and dip of the a-axis of stones longer than 0.02 m and shorter than 0.25 m with a minimum a : b ratio of 2 : 1 were measured. At each area one sample location was used for a detailed study of the stones measured and a number of parameters were monitored. Each fabric was subjected to a Chi-Squared test and accepted at the 95% level. Vertical and horizontal comparisons of fabrics were undertaken by means of the Kolmogorov-Smirnov test and all fabrics were subjected to vector analysis to obtain the resultant fabric vector and strength. At each fabric point the composition of the till with regard to the number of pyroclasts, striated stones and grey lava stones were noted and expressed as a percentage (Table 1). This was used as an aid to mapping of the beds and to noting the occurrence of volcanic phases of the island.

Wherever solid grey lava outcrops occurred measurements of striations and moulded forms were undertaken and the resulting mean directions obtained. Surveyed profiles of the moraines and other depositional landforms were studied as an aid to understanding their genesis and formative ice-flow directions. Surface stone fabrics were also obtained from the moraines for additional information on ice



1 = Lava c. 220 000–276 000 yrs BP
2 = Till, largely composed of pyroclastic material, overlying fluvial sediments (total thickness = 3 m)
3 = Interstadial melt-out till with ablation till capping (in wedge structures due to subsequent shearing)
4 = Main till (11–15 m in thickness)
5 = Capping ablation till disrupted by volcanic bombs (6)
7 = Red pyroclasts
8 = Grey lava c. 105 000 yrs BP of Riss-Würm interglacial age
9 = Till of last glacial (Würm)

Fig. 2. Simplified section along the sea-eroded fault at Ships Cove

flow directions. Samples of till from both inland and coastal locations were collected and subjected to grain size analysis.

Alone each of these simple techniques was of limited value but together their complementary evidence assumes some degree of power. The combination of striated surfaces, surface stone fabrics, and asymmetry of moraines plus the coastal till fabrics allow the identification of the former glaciers with a reasonable degree of certainty.

EVIDENCE FOR AND NATURE OF THE OLDEST GLACIATION

Evidence for the earlier glaciation which Verwoerd (1971) had speculated about has been found at four coastal exposures, namely Kildalkey Bay, Macaroni Bay, Ships Cove and Goodhope Bay. At these locations the occurrence of the earlier glaciation was indicated by a vertical sequence of – till, grey lava, till – of which the upper till was known to be from the last glacial. The lava dividing the two tills had been dated at two outcrops (McDougall, 1971) at about 105 000 BP ($\pm$25 000), thus the underlying till had to predate this lava and postdate the oldest lava (276 000 $\pm$ 30 000) thereby locating the lower till at approximately the same age as the Riss of the Northern Hemisphere. The intervening lavas, where they occurred, made good stratigraphic junctions but additional lines of evidence to denote two till sequences were also found. A location near Kildalkey Bay which lacked the intervening lava flow showed instead a distinct palaeosol developed in the

Table 1

To show the percentage of pyroclasts, striated and grey lava stones in successive tills from the Kildalkey area

Location	a	b	c
Top of till sequence	18	0	82
Top of till sequence	12	1	88
Top of till on top of palaeosol	23	2	77
Intermediate locations	58	4	42
in till on top of palaeosol	82	1	18
Base of till on top of palaeosol	89	3	11
Palaeosol	10	1(?)	90
Base of till below palaeosol	86	0	14

a = % pyroclastic stones
b = % striated stones
c = % of non-pyroclastic stones

lower till. Till fabric analysis for the lower and upper till at this point, and at a similar situation at Macaroni Bay, showed distinctly different preferred stone orientations between the tills. This can be explained by the interglacial outpourings of lava that affected the ensueing ice-flow directions. In addition, at Macaroni Bay, where the oldest grey lavas occur, the striations found on the lower lava agree with the preferred fabric orientation of the till resting directly on top but differ from the fabrics of the upper tills, whose preferred orientation agrees with the striations found on the more recent grey lava outcrops. Finally the lower tills were found to be well consolidated compared to the upper sequences.

At Ships Cove, along a sea-eroded fault, occurs the largest exposure of deposits from the earlier glacial (Fig. 2). The lowest till in the sequence is composed largely of pyroclastic material and only near its top do sub-rounded and angular blocks of grey lava occur and begin to predominate. This reflects the stripping of the unconsolidated surficial pyroclastic cover of the island, by the ice, and then the later incorporation of the frost-shattered underlying grey lava. This till is covered by what appears to be a melt-out till topped by a thin, angular, ablation till. The tills occur in a series of wedges which thin towards the sea. The wedging is thought to result from either slight movement during the melt-out process or due to subsequent overriding by ice of the succeeding stade. The melt-out - ablation till sequence is considered to represent an interstadial for it is covered by a further till. No other major oscillations of this nature have been observed in the 15 m of till but several distinct changes in lithology have been recognised and are still under study. Volcanic bombs can be seen at the top of the till. They disrupt the capping ablation till and gradually phase into a purely pyroclastic layer (very different in composition to the lowest pyroclastic till), upon which the lava rests.

The palaeosol, which was found on the coast where the interglacial lava capping was missing, is some 2 m thick and reddy-brown at the top, grading with depth, back to the grey of the till beneath. There is some evidence to suggest frost-sorting of the surface layers of the palaeosol prior to the return of the ice but no ice-wedge structures, suggesting permafrost, have been found. The depth and degree of weathering suggest that the interglacial may have been warm and humid. Table 2 gives data of the thicknesses of weathering rinds found on stones in the palaeosol and the tills beneath and above thereby indicating the distinct weathered layer in the upper part of the palaeosol. It is considered that the weathered stones found in the lower part of the capping till result from disturbance of the underlying soil and the inclusion of some of the already weathered stones. Samples of organic material have been collected from this weathered layer and it is hoped that they will be useful for pollen analysis which may give indications of the climatic conditions. The evidence for the earlier glacial episode is limited and much of the information is still being analysed but it is possible to state that Marion Island was ice covered sometime during the period after approximately 276 000 BP and before 105 000 BP. The dates of the bracketing lavas locates the glaciation as broadly coeval with the Riss of the Northern Hemisphere. It would appear that there was an interstade early on the 'Riss' sequence and that fans of outwash material were produced which were later

Table 2

Thicknesses of weathering rind observations from palaeosol at Kildalkey Bay

Location	a	b	c	d	
Base of till on top of palaeosol	15	2	0.6	0.57	
Top 0.3 m of palaeosol	15	15	3.14	1.93	n = 24
0.5 m–0.3 m depth in palaeosol	15	9	1.83	0.71	$\bar{x}$ = 2.65 s = 1.69
Top of till below palaeosol	20	0	—	—	

a = number of stones sampled
b = number of stones with weathering rinds
c = mean ($\bar{x}$) thickness of measured rinds
d = sample standard deviation (s) of measured rinds

covered by the till of the re-advance. The ensuing interglacial was a period of volcanic activity with lavas and pyroclasts capping much of the till, but uncovered till areas developed a reddy-brown soil under a possibly warm, humid climate. Prior to their being covered by ice during the youngest glacial the soils were subject to frost sorting but not to permafrost.

YOUNGEST GLACIATION

a) Mapping of the ice cover

Estimation of the area and distribution of the glaciers during the last glacial has been undertaken by means of till fabric analysis from coastal and inland exposures, striation measurements, and the mapping of moraines and other landforms. At the time of writing some 200 fabrics have been completed and striations measured on all known grey lava outcrops. Contrary to the expectations of Verwoerd (1971) large numbers of moraines have been found on the eastern and southern sides of the island. Surveyed profiles have been completed on four moraines and distal and proximal slopes measured on fourteen others; surface stone fabrics have been undertaken on six of the larger moraines. Both 'push' and 'dump' type moraines have been recognised.

At Skua Ridge and Albatross Lakes the ridges of debris are distinctly asymmetrical with shallow (5°-9°) proximal slopes and steep (17°-28°) distals. At both locations the moraine profile is broken by a bench 5-10 m in width on the distal side. This is thought to represent the top of a former moraine which has been partially overriden by readvancing ice thus producing a multiple moraine (Fig. 3). The preferred orientation of the larger surface stones found on the proximal slope of these moraines is at right angles to the moraine crest suggesting additional evidence for overriding ice. The much larger (±50 m high, 200-300 m broad) moraines found in the southeast at Stony Ridge to Kildalkey Bay are of the dump variety, often showing steeper ice-contact proximal slopes.

From the moraines, striations, and till fabrics the

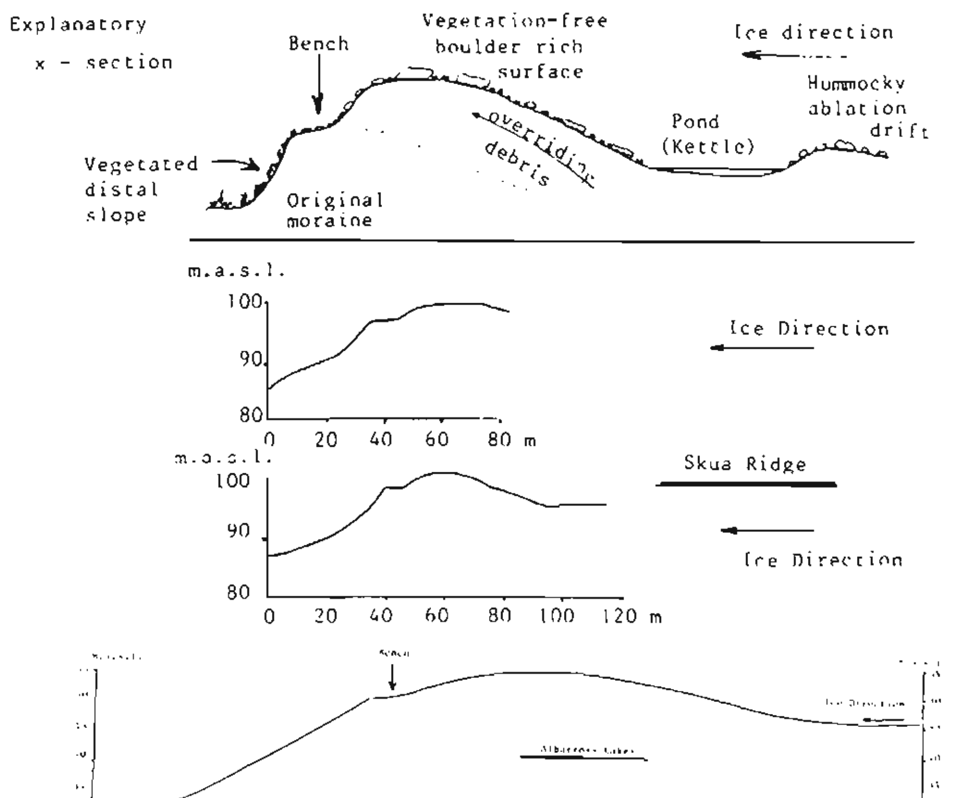


Fig. 3. Levelled x-sections from two locations of moraines exhibiting evidence for minor readvances (with simplistic explanatory diagramme)

locations of five former glaciers have been determined (Fig. 4):

1. to the north of Long Ridge (northward extension uncertain; fed from the Katedraalkrans-Middelman area)
2. Table Mountain to Skua Ridge (being fed from the Piew Crags to upper Table Mountain area)
3. Macaroni Bay-Albatross Lakes Area (being fed from the region of Freds Hill)
4. Stony Ridge to Soft Plume River (being fed from Tates to just north of Hoë Rooikop)
5. Soft Plume River to Kildalkey Bay (being fed from Hoë Rooikop to Beret area)

It was found that many of the post-glacial lava flows follow the intermoraine areas and that these are now the locations of the major stream courses.

The large laterals in this south-eastern area start at approximately 200 m a.s.l. and thus it is considered that this equates to the approximate altitude of the equilibrium line of the glaciers (see Andrews, 1975, p. 54). Using 200 m as the equilibrium line and the width of the glacier as equating to the distance between the main laterals the ablation area of four of the glaciers has been calculated:

1) Kildalkey Glacier $= 6.2$ km^2
2) Stony Ridge Glacier $= 5.9$ km^2
3) Albatross Lakes Glacier $= 5.9$ km^2
4) Skua Ridge Glacier (minimum size) $= 4.9$ km^2

When surveying the profiles across the equilibrium line have been completed it is anticipated that the volume of the ice in the ablation area and the approximate rates of flow can be calculated.

In addition to the moraines areas of disintegration ridges (and kettle holes) and two marginal stream channels have been recognised. Disintegration ridges and kettle holes have been found at three locations (Long Ridge, Skua Ridge and Albatross Lakes) within the moraine sequences. At Long Ridge, far back from the maximum ice extent, two parallel former marginal stream channels, which were formed during the main retreat phase, occur. The significance of these features and the moraines will be put in context after the earlier glacial history, as shown by the till deposits, has been described.

b) The sequence of the last glaciation as shown by the till deposits

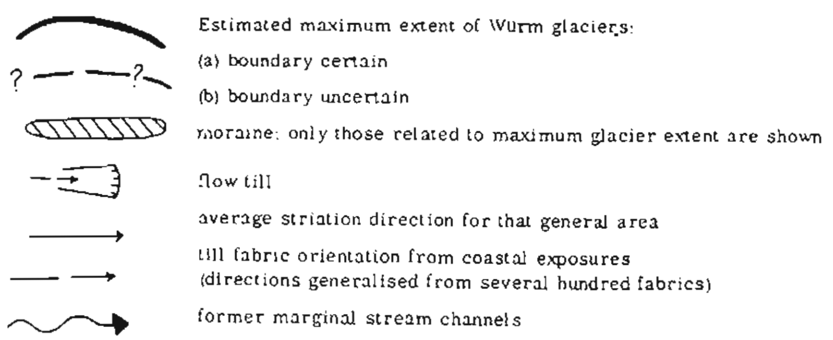
The sequence of the youngest glacial will be described in detail from only one site, that of Long Ridge, but additional data from other locations will be commented on.

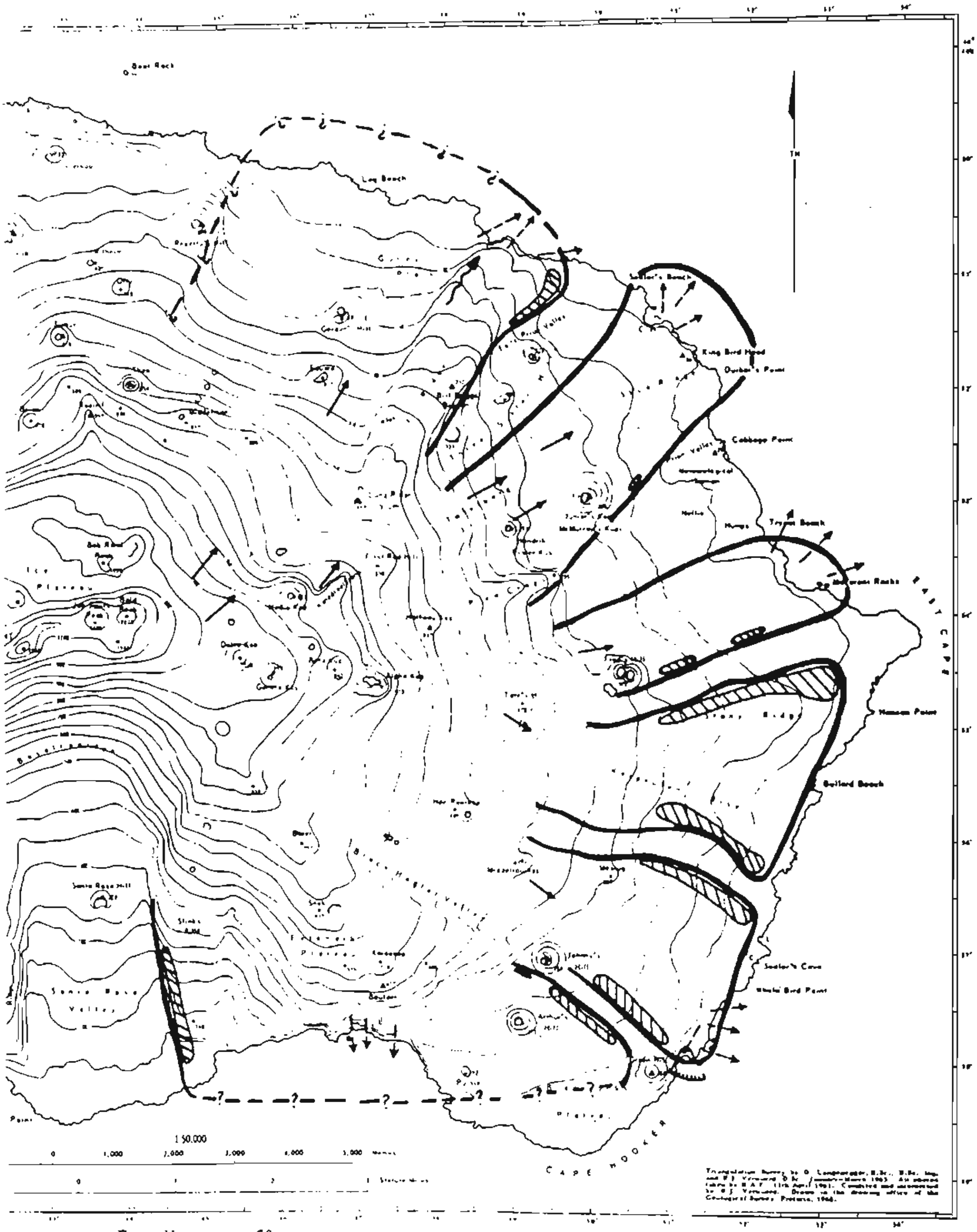
At the seaward end of Long Ridge a clear exposure of glacial debris upwards from a grey lava exists (Fig. 5). Immediately on top of the lava occurs a 2 m thick deposit of bedded fines and gravels which is capped by a distinctive 3 m thick till. The till has a reddish-brown colour and is composed entirely of pyroclastic debris of mainly clay to gravel size with few boulders or cobbles. It is a homogeneous deposit with only one variation occurring at its top where there is a distinct layer of platy, grey lava blocks. This is interpreted as an ice advance and retreat sequence. The advancing ice cleared the surficial pyroclastic debris, which has a few large blocks and weathers rapidly, thus producing the reddish-brown till. The occurrence of angular, platy grey debris at the top of the till suggests a supraglacial or englacial deposit that has been let down, by the melting ice, as an ablation till. If the material had been subglacially derived it would have been more rounded and less angular than was found.

This lower till is covered by a thick (up to 2 m) sequence of sands and gravels (with organic material) which grades upwards into a rhythmite sequence and then back to fluvial beds. This represents an interstade during which the ice retreated extensively

Fig. 4. Topographic map of Marion Island with moraines and glacier positions (etc.) marked

Key





Form lines every 50 m.
The western side of the island is not shown as postglacial volcanics have destroyed nearly all evidence of former glacial activity.
The information is given on the map of Marion Island produced by Langenegger and Verwoerd, 1968.

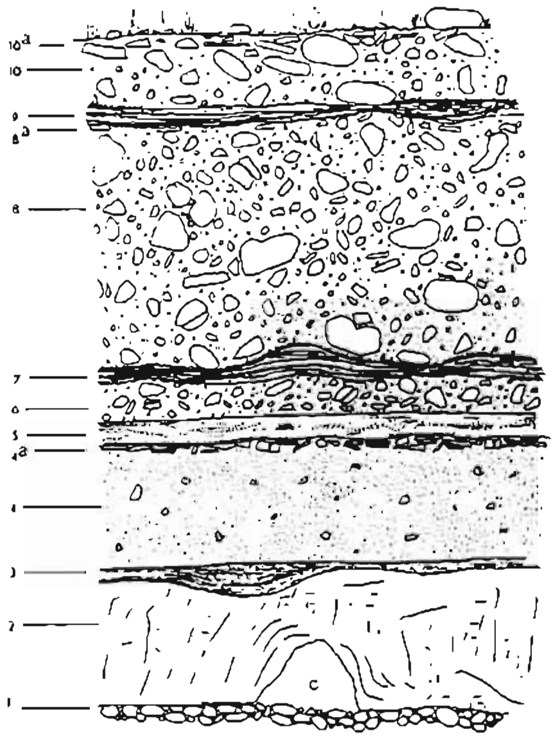


Fig. 5. Simplified section of upper (Würm) till sequence from Long Ridge.
1 = Beach
2 = Grey lava (C = cave)
3 = Bedded fines and gravels
4 = Till, reddish-Brown, composed of pyroclastic debris (3 m)
4a = Platy grey lava blocks
5 = Interstadial fluvial deposit of sand and gravel, containing a rhythmnite sequence (up to 2 m)
6 = Till, grey, c. 0.5 m thick
7 = Rhythmite-with-dropstones sequence, faulted and folded
8 = Till, 10–12 m thick
9 = Rhythmites (c. 1 m) covered by gravels
10 = Homogenous grey till
10a = Platy ablation till
Total height of profile c. 21 m

prior to a readvance. The covering till, which is only (approx.) 0.5 m thick, is grey in colour and has few pyroclast inclusions. Towards the top this till becomes very gravel rich and grades into a rhythmite-with-dropstones sequence. The rhythmites become gravel rich at their upper boundary and show faulting and folding, probably produced by the readvance of the ice which deposited a thick (10–12 m) till on top. This sequence is seen as indicating a glacial advance of relatively short duration followed by a minor retreat during which a proglacial lake was formed. The subsequent stade which disturbed the rhythmites was of long duration and a thick homogeneous till was produced.

The only variation found in the rest of the sequence is a further bed of rhythmites close to the cliff top. Possibly there was a minor retreat and readvance prior to the major retreat sequence and during this oscillation a proglacial lake occurred. A comparable sequence of events can be found at the southern end of the island at Kildalkey (Table 3). It can be seen from Table 2 that the initial stages of the two separate areas are very similar. The thin till is missing from the Kildalkey area which is probably due to the ice not having readvanced to its former maximum position prior to its subsequent retreat. At the southern end a distinct flow occurs which could be attributed to either the earlier or the later interstade. The flow till extends beyond the limits of the Kildalkey ice and is a most distinctive feature. Considering the small size of the island and the enormous oceanic influence it is considered acceptable that the glaciers should be fairly synchronous in their responses to climatic variations.

BRIEF OBSERVATIONS ON FORMER SEA LEVELS

True raised beaches were observed at a number of locations around the island – all of which showed wave-smoothed rocks, wave rounded pebbles and boulders, and macro-cliffs at the back of the beach. At Transvaal Cove and Trypot Beach raised beaches were surveyed and heights of + 3.4 m and + 2.9 m were found respectively. Just to the north of Cabbage Point a sequence of two beaches was found with the lower one at a height of + 3.3 m and the upper one at + 6.1 m. Stone roundness (Pi) (using Cailleux's equation) and flatness (Fi) indices were obtained for both levels:

+3.3 m beach	+6.1 m beach
Pi = 422.9 (s = 125.6)	Pi = 381.2 (s = 121.3)
Fi = 448.1 (s = 61.0)	Fi = 492.9 (s = 117.6)
n = 100	n = 100

The difference in indices between the beaches is further emphasised by the values obtained for the present day beach and the + 2.9 m beach at Trypot where the following values were obtained:

Present Beach	+2.9 m Beach
Pi = 557.5 (s = 120.3)	Pi = 417.1 (s = 119.5)

Thus the three beach levels (present day, ca + 3 m and ca + 6 m) can be differentiated in terms of the stone roundness indices and it is interesting to note that very similar values for Pi were found for the two ca + 3 m levels at Cabbage Point and Trypot Beach (422.9 and 417.1). These levels of ca + 3.0 m and ca + 6.0 m were observed at several locations

Table 3

Comparison of events during the last glacial at Long Ridge and Kildalkey Bay

Long Ridge	Kildalkey area
Ablation till	Ablation till
Thin till	Till
Rhythmites	Sands & gravels
Ablation till	Ablation till
Thick grey till	Thick bouldary till
Rhythmites (+ fluvials)	Flow till sheets
Thin grey till	Thin grey till
Fluvials (+ rhythmites)	Fluvials
Ablation till (grey plates)	Ablation till
Pyroclastic till	Pyroclastic till
Fluvials	Fluvials
Lava	Palaeosol

around the island, notably Macaroni Bay, Water Tunnel Stream, Goodhope Bay, Fur Seal Bay and Cape Davis.

In addition to direct observation of raised beach levels the long profile of a river, the Van den Boogaard, was surveyed and levels extrapolated from the nick points in this. From the long profile obtained a number of distinct nick points were discernable and whilst an exponential curve with a coefficient of determination (r^2) of 0.83 best fitted the whole profile, linear extrapolation was found to best fit the level above each nick point. A line generated for the seaward segment indicated a former level of +5.9 m with an r^2 of 0.98. The inland segment showed an extrapolated level of + 10.86 m with an r^2 of 0.95. Thus the extrapolated level of + 5.9 m is in close agreement with the level of the true raised beaches found around the coast. Whilst several localities show what may possibly be raised beaches at ca + 10.9 m level none have actually been surveyed as such. However, it is very likely that a beach level occurred at a ca + 10.9 m level.

CONCLUSIONS

Combining the information from the glacial landforms and till sequences a first approximation of the glacial history of Marion Island can be made (Table 4). It is hoped that the major events can be dated before long - notably the first interstade of the last glacial. Unfortunately the counting of varve couplets has been precluded by their disturbed nature although a set for the minor retreat at Long Ridge indicate a minimum duration of 78 years. Thus the information obtained, given here in very brief outline, shows that the island was subject to two major glaciations and that each was composed of a series of stades and interstades.

The implication of the glaciations of Marion Island is that the Polar Front moved northwards to encompass Marion and the onset of the glacials is

Table 4

A first aproximation of the sequence of events during the last glacial as shown by the deposits and landforms

Interglacial	Lava flows - scoria palaeosols developed
Interglacial - onset of glaciation	Frost action in soils fluvial sequences
Stade	Pyroclastic tills produced from clearing of surficial deposits Deformation of earlier fluvials
Ice retreat	Distinct ablation till - angular grey blocks capping brown till beneath
Interstade	Outwash sands and gravels capped by rhythmites (thick)
Stade	Deforms underlying rhythmites Thin grey till produced
Minor interstade	Proglacial lake rhythmites Flow tills
Stade	Extensive deformation of rhythmites Thick grey till
Retreat	Distinct level of ablation till
Interstade	Outwash gravels followed by preglacial rhythmite deposit
Stade	Grey bouldary till
Retreat	Capping of ablation till
End of glacial Ice retreating	Ice begins major retreat at end of glacial
Minor readvance	Push moraine formed
Minor readvance	Overriding of earlier moraine
Period of relative stillstand	Dump moraine formed inland Marginal stream channels develop
Ice continues retreat	Ice retreats rapidly Thin till covering of striated pavements inland

seen as a result of this. The northward shift of the Front would lower the mean annual temperature of the sea water by approximately 2° C and in addition put Marion within a zone of more southerly winds which are 3–4° C colder than those experienced at present (Schulze, 1971). Also the precipitation may have increased as the dry anticyclones would have passed further to the north. The increased precipitation together with the lower temperatures would result in a greater annual snowfall with less summer ablation which would account for the growth of the ice-cap and glaciers. Hays et al. (1976) showed that the Polar Front did not move very far north in the region to the south of Africa, so Marion would have only just come within its influence. Thus the island would have been in a sensitive marginal position with respect to the movement of the Front and it can be assumed that there could have been rapid build-ups and losses of ice as the island moved in or out of the influence of the Polar Front.

Hays et al. (1976) suggest that there was a lowering of temperature up to 3.5° C. Using the calculated level of the equilibrium line of the glaciers and from this calculating the approximate height of the snowline, it is considered that the minimum decrease in temperature on Marion was 3.5° C. This gives a mean annual temperature of 1.5° C at sea level. Thus the temperatures calculated from the information available on the island agree very closely with those found by Hays et al. from the ocean floor sediments and van Zinderen Bakker (1973) from palynological studies.

The results do not indicate that the island has been completely ice covered which is in accordance with the suggestions of Schalke & van Zinderen Bakker (1967). They suggest that as certain of the island biota had 'overwintered' the glacial, there must have been ice-free areas, as has also been suggested for Kerguelen by Young & Schofield (1973). The distribution of moraines and the occurrence of localities (not covered by postglacial lava flows) which show no signs of glacial deposits indicate, that a number of sections of the island were indeed ice-free. Clapperton & Sugden (1976) show that the Falkland Islands lacked extensive glaciers during the last glacial. However, the glaciation of Marion is seen as far more extensive than that of the Falklands. The glaciers of Marion are indicated as being 6–7 km long and 3 km wide flowing out from a central ice-cap, whilst Clapperton & Sugden (1976) suggest a maximum cirque glacier length of 2.7 km for the Falklands. The more extensive glaciation of Marion is seen as being due, in part, to the far greater precipitation. Hays et al. (1976) also showed that the Falkland Islands remained north of the Polar Front and were thus not subject to this colder influence.

The raised beach levels found on Marion appear to reflect those described at other Antarctic–sub-Antarctic locations. Nougier (1971) has noted a +3.0 m level on Kerguelen whilst in the South Shetlands Sugden & John (1973) have found levels of approximately +3.0 m and 6.0 m. On Livingston Island Everett (1971) has found levels of +10.6 m and +6.1 m. Thus there would seem to be some evidence from Marion to show that the +3.0 m, +6.0 m and +10.6 m sea levels occurred in the sub-Antarctic region.

The implications of the glaciations of Marion Island are far reaching, for not only do they assume significance for the sub-Antarctic and Antarctic region but also for the African continent. The possible northward shift of the climatic belts, and their oscillations, are pertinent to the understanding of the Quaternary history of Africa. In the light of the work on oceanic sediments to reconstruct the Quaternary climatic conditions the availability of terrestrial results from Marion helps clarify the situation. Much data still awaits analysis and so it is hoped that more detailed information will shortly be forthcoming. It is also hoped that future work undertaken in the Borga nunatak area of Antarctica will provide complementary information to link with that of Marion. Thus this information is seen as a step in the completion of an overall pattern of data collection from Africa through the sub-Antarctic to Antarctica proper from which a better understanding of the glacial episodes in the Southern Hemisphere may be obtained.

SUMMARY

Detailed investigation of Marion Island deposits has shown that the island was subject to two glacial episodes roughly coeval with the Riss and Würm on the Northern Hemisphere. Till deposits have been found at a number of coastal cliff exposures. The results of till fabric analysis together with striation observations has indicated the main ice-flow directions. Mapping and surveying the moraines has enabled the boundaries of the glaciers to be delimited for the southern and eastern sides of the island. Dates from the lava flows interbedded with the tills have shown that there were two major glacial phases which broadly coincide with the Riss and Würm. Where the intervening lava flows are absent palaeosols have been found.

Within both glacial episodes sequences of stades and interstades have been recognised. Evidence has been found for the occurrence of proglacial lakes, flow tills and multiple moraines. Two former marginal stream channels have been recognised together with kettle holes and hummocky ablation drift. Using the altitude of the start of lateral moraines the approximate height of the glacier equilibrium lines

have been calculated together with the ablation area of the glacier. Palaeotemperatures for ± 18 000 BP have been calculated from the estimation of the palaeosnowline, which agrees with evidence from ocean floor sediments and fossil pollen.

A number of raised beaches have been found and surveyed at several locations on the island. Complementary evidence for the raised beach levels has been obtained from extrapolation of a long profile of a river.

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Palaeogeography, Palaeoclimatology, Palaeoecology, 29(1979/1980): 243—259

LATE GLACIAL ICE COVER AND PALAEOTEMPERATURES ON SUB-ANTARCTIC MARION ISLAND

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ABSTRACT

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By means of data from striation observations, till fabric analysis, moraine recognition, surface megaclast fabrics and till stratigraphy, a reconstruction of the ice cover for part of sub-Antarctic Marion Island is attempted. The presence of ice at a number of other locations on the island is noted but former glaciers cannot be recognised. By means of the altitude of lateral moraines a range of temperatures for glacial-maximum conditions are reconstructed. The estimated drop in temperature falls within the ranges suggested by palynological and ocean core investigations.

INTRODUCTION

Marion Island (lat. 46°54'S, long. 37°45'E) is located in the vast Southern Ocean approximately 2° of latitude north of the Antarctic Polar Front (Fig.1). The island consists of a roughly oval shield volcano (Verwoerd, 1971) of 290 km^2, which rises to a height of 1 230 m in the central mountain area and is covered by numerous scoria cones resulting from the most recent volcanic event (± 11 000 to 4 000 B.P.). Geologically, the island is composed of two basaltic suites: an older, glaciated grey and a recent, unglaciated black (Verwoerd, 1971). The hyper-oceanic sub-Antarctic climate of low temperatures, high annual precipitation and low radiation inputs due to an almost continuous cloud cover (Schulze, 1971) result in a small area of permanent snow and ice above 950 m.

Prior to the present study (Hall, 1978a) very little information was available regarding the glacial history of the island. Schalke and Van Zinderen Bakker (1971) showed, from palynological evidence, that Marion Island experienced a cold period that was approximately coeval with the Würm of the Northern Hemisphere and ended at about 12 000 B.P. The finding of striated bedrock surfaces at several locations on the island (Verwoerd, 1971) showed that ice had existed during the cold period described by Schalke and Van Zinderen Bakker (1971). No evidence for glacial deposits was found during the early exploratory work, although a rock consisting of "poorly sorted pyroclasts" (Verwoerd, 1971) was

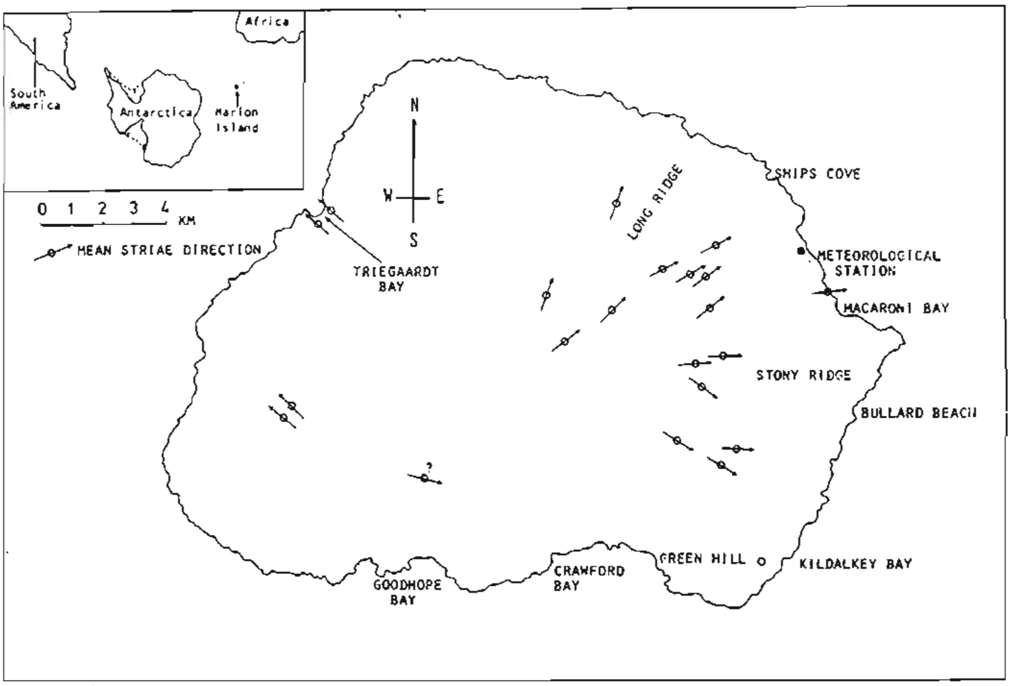

Fig.1. Marion Island and its location. Major areas of striations showing former ice directions.

considered in need of investigation in the light of the knowledge of a former ice cover. Upon the basis of the oldest island lava date so far determined (276 000 ± 30 000 B.P.; McDougall, 1971), Van Zinderen Bakker (1973) suggested that a glaciation pre-dating that of the Würm-age may also have taken place.

Results from a programme initiated in 1975 to undertake an investigation into the glacial history of the island indicate a wealth of information (Hall, 1978a). It was found that Marion had in fact experienced three glacial episodes (Hall, in prep.) that correlate with cycles D, C and B of Kukla (1977). Unfortunately, the information pertaining to the earlier glacials is limited and still under investigation and so only the most recent glacial is considered here.

APPROACH

As a means to determining the extent of the ice cover during the last Glacial a number of simple but complementary approaches were adopted. Inland, areas of the older grey lavas that escaped inundation by recent flows were examined for striations and glacial moulding to determine former existence of ice and its flow direction. At each site several hundred observations were taken and the mean value calculated. Nearer to the coast, glacial-depositional landforms, notably moraines, were investigated. In addition to delimiting the glacier margins, the moraines also indicate the ablation area of the former glacier and mark its fluctuations. At coastal cliff sections and along incised river courses, analysis of the till stratigraphy allows description of the stade—interstade sequence. Till fabric analysis undertaken at these exposures gives further directional information for comparison with that of inland striations. Thus a continuum of supporting evidence is obtained to describe the extent, flow direction and variation through time of the former ice cover.

A crude estimate of palaeotemperatures during the glacial maximum is derived from a reconstruction of the snowline altitude as indicated by lateral moraines. Andrews (1975, p. 54) explains the relationship of the origin of lateral moraines to the equilibrium line altitude (ELA) and that the snowline can be located within a specific altitudinal range of the ELA. By assigning a range of temperatures to the possible maximum and minimum snowline altitudes and adopting a rate of fall of temperature with height it is possible to derive a range of palaeotemperature values. The temperatures so derived agree well with those obtained by other means.

RESULTS

A. Striations

Areas where striations might potentially be found on Marion Island are limited to those outcrops of grey basalts which escaped inundation by the post-Glacial lava flows. Of the areas which did survive many, particularly those which exhibit closely spaced jointing, have subsequently been severely broken-up by frost action so destroying all evidence. Despite these limitations many outcrops of striated grey lavas additional to those noted by Verwoerd (1971) have been found (Fig.1). It is assumed, in the absence of evidence to the contrary, that the ice flowed radially from a central ice cap towards the coast and thus a sense of direction is able to be imparted to the striae.

Consideration of the striae suggests an ice cover over much of the island. The scarcity of observations in the south and west is the result of the almost continuous black lava cover in these areas. The data from the limited exposures available do, however, indicate the former presence of ice in the south and west, which is confirmed by other evidence (see below).

B. Glacial-depositional landforms

Extensive post-Glacial faulting and volcanism (Hall, subm.) have eradicated evidence for glacial-depositional landforms from much of the island. However, despite this disturbance a number of moraines and

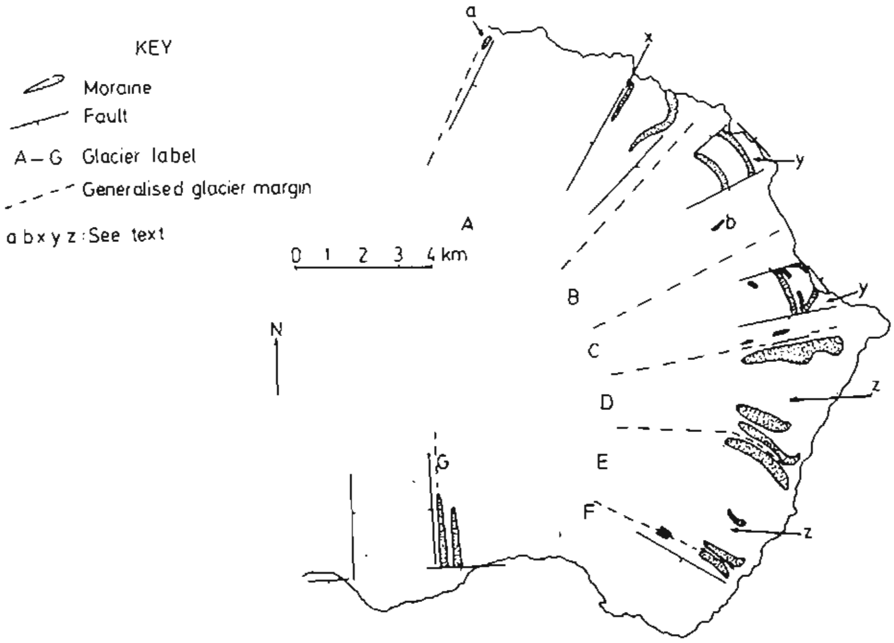

Fig.2. Glacial-depositional features on the eastern side of the island.

associated landforms can still be discerned on the eastern side of the island (Fig.2) contrary to the expectations of Verwoerd (1971) who anticipated any moraines that existed would now be below sea level. Both "dump" and "push" moraines (Andrews, 1975) are recognised, the latter being particularly abundant at Albatross Lakes and Skua Ridge (Hall, subm.).

From the moraine exposures available it is possible to broadly demarcate the positions of five former glaciers and a margin of two others (*A* to *G* in Fig.2). Moraine remnants available for the attempted reconstruction vary such that in some instances a retreat sequence can be discerned, but not the maximum ice extent, whilst others show the reverse; no complete sequence is available except by supposition from the sum of the evidence.

At Long Ridge (*A* in Fig.2) the curving moraine to the southeast probably demarcates the maximum extent of the ice in that direction. The moraine projects up to 40 m above the surrounding area and is approximately 200 m wide at its broadest point. It exhibits a shallow (11°) proximal slope and a steep (21°) distal one. On the inside of the moraine (to the northwest) there is a confused series of small ridges and mounds with some kettles. On the northwestern side of Long Ridge, just above the fault, there occurs a further small lateral moraine along the edge of which is found a former marginal stream channel (*x* in Fig.2). This channel begins, suddenly, approximately 600 m inland from the coast and curves towards the sea with a gradient of 6°, increasing in width from 8 to 35 m. On the other side of the large graben there is a very small outcrop of a moraine (*a* in Fig.2) which, it is thought, is related to those observed on Long Ridge.

The Skua Ridge area (*B* in Fig.2) can be divided into four simplified depositional units: (1) a coastal area with small debris mounds and kettles, (2) a multiple moraine, (3) an undulating till cover, and (4) a broad, high moraine with convex slopes. The outermost area (*y* in Fig.2) presents a much subdued landscape with vegetation-filled kettles and rounded mounds. This area rises inland until it meets the break of slope demarcating the outer edge of the multiple moraine. The moraine is not a continuous feature but rather a series of asymmetrical mounds (Hall, in press a) with intervening water-filled hollows. The mean distal slope of the ridges is 16° (s = 5.5) whilst the proximal mean slope is 8° (s = 3.3). The seaward ridge shows evidence of two advances with some overriding of the original ridge (Hall, 1978b). Megaclast surface fabrics (Andrews, 1975) taken on the distal and proximal slopes indicate a clast orientation approximately normal to the ridge long axis (Fig.3).

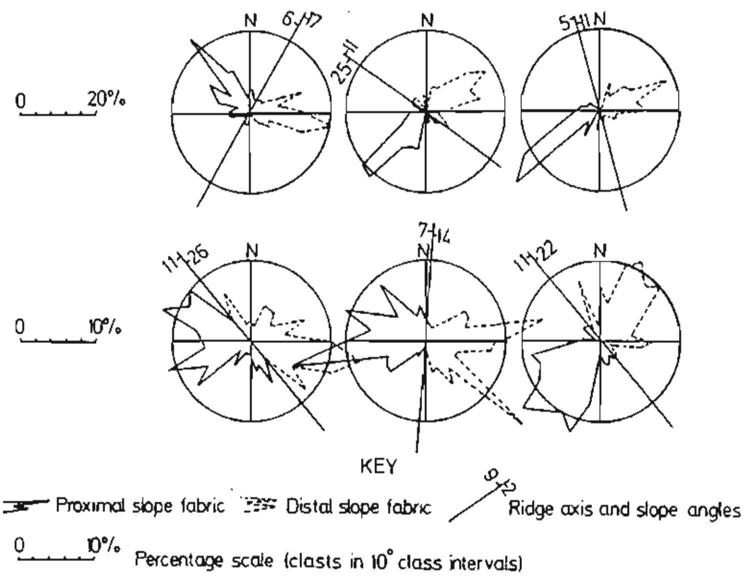

Fig.3. Surface megaclast fabrics for Skua Ridge.

The undulating till surface inland of the moraine exhibits a number of water-filled kettles and asymmetrical mounds (with steeper distal slopes). At the inland terminus of Skua Ridge is found the convex-sloped moraine. This has a steeper proximal (25°) than distal (11°) slope although there may be some oversteepening of the proximal slope due to stream action.

To the south there is found a small outcrop of what may be part of the multiple moraine (*b* in Fig.2). The moraine exhibits a steep (17°) distal slope and a shallow (6°) proximal slope. A surface megaclast orientation fabric

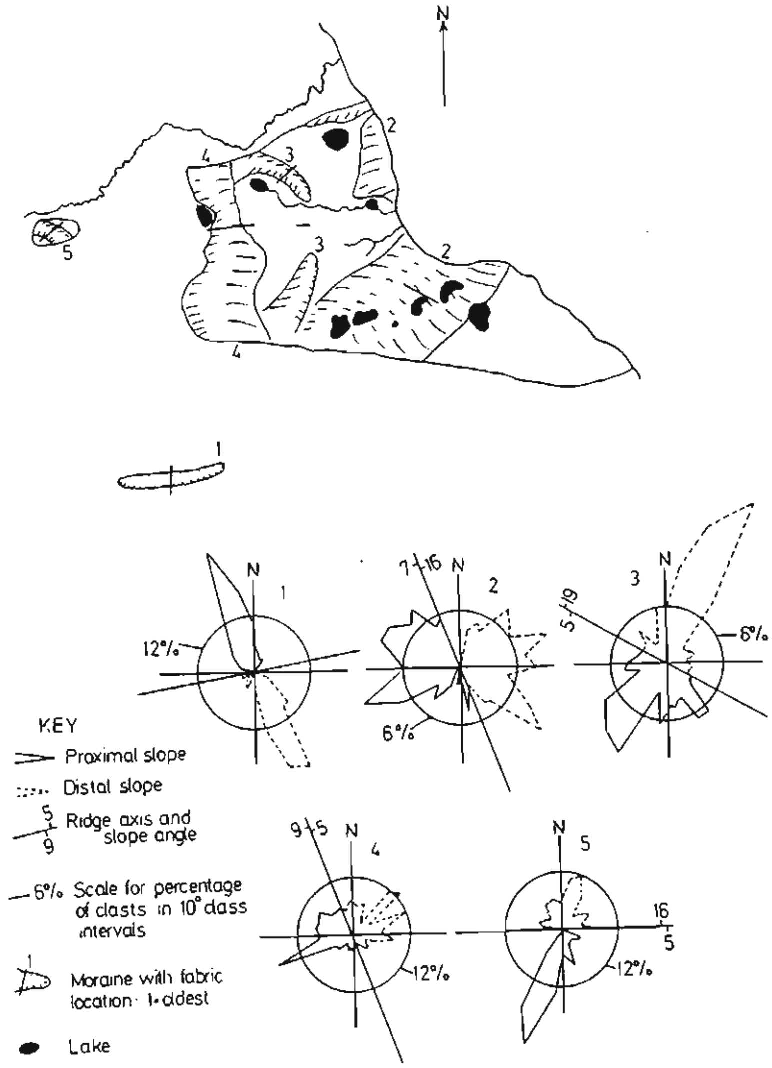

Fig.4. The moraines at Albatross Lakes together with surface megaclast fabrics.

shows a very strong preferential long-axis alignment normal to the strike of the crest (Fig.3).

In area *C* (Fig.2), Albatross Lakes, there is a complex of moraines denoting five former ice positions (Fig.4). The outermost moraine (*1* in Fig.4) is almost completely inundated by black lavas and so it can only be said that, of the ridge exposed, the distal slope appears steeper than the proximal. A surface megaclast fabric indicates a very strong long-axis orientation normal to the moraine crest. Moraine *2* rises to approximately 90 m above sea level (a.s.l.) in the south and reaches almost 1 km in width. It appears to be a multiple moraine with numerous asymmetrical push ridges

(Hall, in press a) with intervening lakes. The ridges indicate a mean distal slope of 15° and a mean proximal slope of 6°.

Moraine *3* (Fig.4), which is best exposed to the north, has a steep (14—19°) distal slope and a shallow proximal one (3—7°). This moraine, like the seaward one at Skua Ridge, appears to be a product of two advances with some degree of overriding during the second advance. Surface megaclast fabrics again indicate an orientation broadly normal to the ridge axis (Fig.4). Moraine *4* abutts against *3* and, contrary to the others, exhibits a steeper (11°) proximal slope than distal (6°). Surface megaclast orientation indicates a broad axis normal to the ridge crest on the proximal side but slightly off-set on the distal side (Fig.4). Finally, Moraine *5* possesses a steep distal slope (16°) and a shallow proximal (5°) with a strong megaclast orientation just slightly off from normal to the ridge long axis.

The area *E* to the south is very similar to that of *D* with large lateral moraines and a confused topography of dumped material between (*z* in Fig.2). For some unknown reason no terminal moraines are present here despite there being no evidence that the ice extended off-shore. Area *F* exhibits a single lateral moraine demarcating the northern boundary of a glacier just to the south of *E*. Finally, in the south there are two lateral moraines at area *G* but their relationship to the former glaciers is not yet known.

C. Stratigraphy

A summary of the deposits, described in detail in Hall (1978a), resulting from the last Glacial is given in Fig.5. The sections are, with the exception of Green Hill, from coastal cliff exposures and thus describe the sedimentary products of glaciers that extended beyond the present coastline. These data, therefore, present a logical extension of that given by the moraines in that they tell the number of times ice extended beyond the present coast and give some idea of the lateral extent of that ice. The converse argument is also used in that whilst coastal till deposits describe a glacier moving seawards the lack of sediments, notably in the Bullard Beach region, indicates that ice did not extend so far.

At Crawford Bay and Goodhope Bay (Fig.1) till resulting from the last glacial is also found, but is not shown in Fig.5, however, as little to no information is yet available regarding its composition.

It is not pertinent to go into specific details of the tills here as it would add little useful information. It is solely the presence of the various units within the stratigraphic columns shown in Fig.5 which will be utilized for reconstruction of former ice cover and its variation through time. Marion Island possesses a relatively "simplistic" till sequence. The glacial deposits are initiated by a "pyroclastic till", that is to say a till with a very high (>40%) pyroclast content. This till results from the ice of the first stade clearing the unconsolidated surficial pyroclastic debris resulting from the

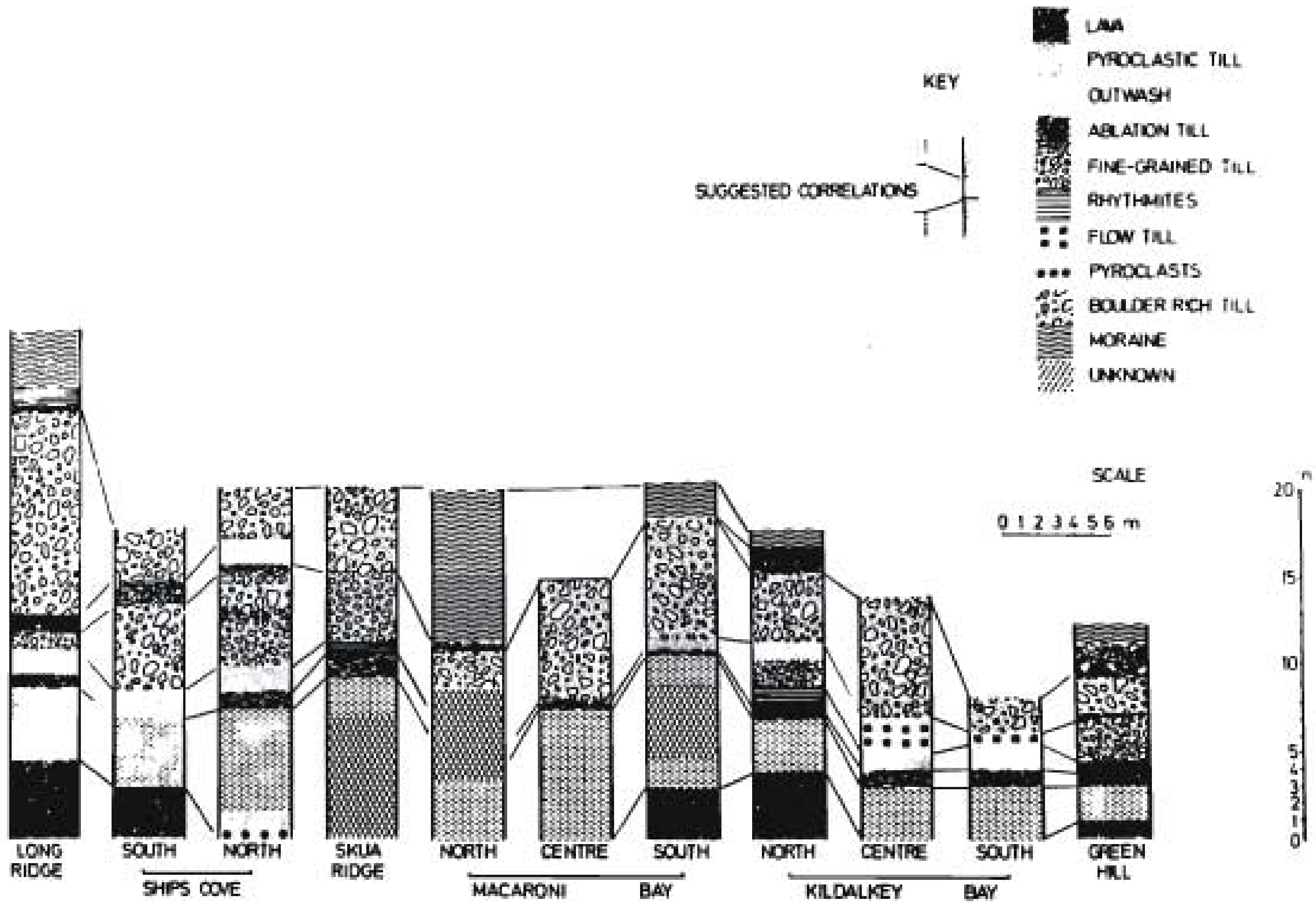

Fig.5. Simplified stratigraphic columns for the major till exposures.

interglacial volcanic event. Pyroclast percentage decreases with height through the till whilst the grey lava content increases, reflecting clearing of the pyroclasts and incorporation of the underlying grey lava debris. The top of the pyroclastic till is frequently marked by a layer, of varying thickness (0.3–1.5 m), of platy, highly angular clasts in a sandy matrix. This is then usually overlain by a fluvialy derived unit. The layer of angular debris is considered to be an ablation till demarcating the start of an interstadial whilst the fluvial deposit represents a later stage when the ice has retreated to some extent. This assumption, as against simple oscillations of the glacier, is justified by the synchroneity of occurrence around the whole island.

Above the initial stade pyroclastic till there occurs a sequence of stadial and interstadial deposits which are a function of that locations position with respect to the ice, i.e., in some instances the ice crossed the present coastline more times than at other places. However, it is interesting to note that a boulder-rich till is often found above the pyroclastic till and that where a moraine is cut by the coast it is seen to be composed of highly angular and platy clasts with a small percentage of sub-rounded blocks. Statistical analysis (Hall, in press b) clearly differentiates between the tills in terms of clast size and shape parameters.

D. Till fabric analysis

Till fabric analysis was undertaken at all coastal and inland till exposures to obtain additional information on palaeo-ice flow directions and their spatial and temporal variations. In all a total of 200 fabrics, with a minimum clast a:b ratio of 2:1, were completed and accepted by the Chi-square test as not being random distributions; diagrams of all completed fabrics can be found in Hall (1978a).

Only the major flow directions, obtained from the till fabric analyses, will be given for use as a complement to striation and moraine data. A simplified summary of the fabrics for each location indicates palaeo-glaciers flowing approximately normal to the present coast (Fig.6). Fabrics frequently show large within-site variation of clast a-axis preferred orientation in both the vertical and horizontal planes. Fabrics with a primary mode transverse to the former ice flow direction are found in areas of deformation and where severe compressive flow occurred due to the ice encountering a lava barrier or such. At terminal positions where the ice was spreading a lateral variation in clast a-axis preferred orientation is observed (Fig.7). In the northern section of Macaroni Bay fabric 1 (Fig.7) was obtained close to the location of the glacier centre whilst fabrics 2 to 8 were obtained successively closer to Moraine *2* (Fig.2). Fabrics 9 and 10 were obtained at the base of Moraine *2* and show a primary mode almost normal to that of fabric 8. Thus, a spreading terminal flow is indicated with primary

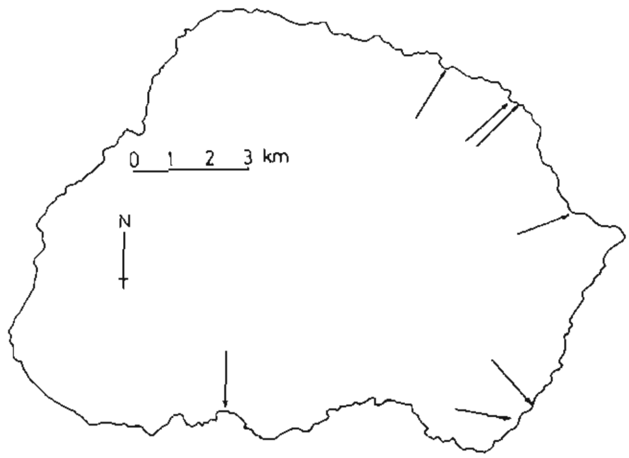

Fig.6. Generalized mean ice flow directions as indicated by till fabric analysis.

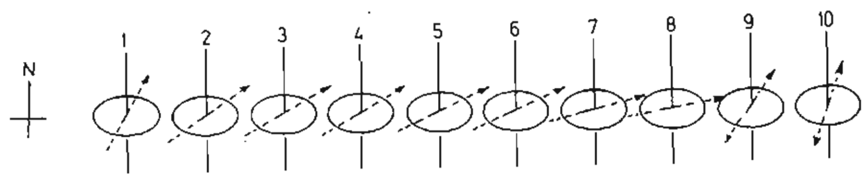

Fig.7. Variation in indicated ice flow direction as shown by clast *a*-axis alignments in a series of till fabrics from Macaroni Bay.

modes normal to flow direction at the moraines; this series is complemented by a mirror image in the southern parts of the bay.

The major flow directions indicated here (Fig.6) take all the local variations into account and show the principal directions of the ice. In Fig.8, where the ice cover is reconstructed, a number of fabrics are indicated for more detail but it is essentially the main flow as shown in Fig.6 which is important.

ATTEMPTED RECONSTRUCTION OF THE ICE COVER

The available information, presented above, provides a continuum whereby an attempt can be made to reconstruct a number of the former glaciers on the eastern side of the island. Inland, the striations give some idea of the broad palaeo-ice flow directions whilst the moraines demarcate the actual boundaries of the glaciers. Surface mega-clast fabrics obtained on the moraines give additional, localised, data on former ice directions. At the coast the occurrence of till denotes ice extending beyond that point and the observed lithological variations and lateral extent denote glacial oscillations and glacier width. Finally, till fabric analysis at these coastal locations give

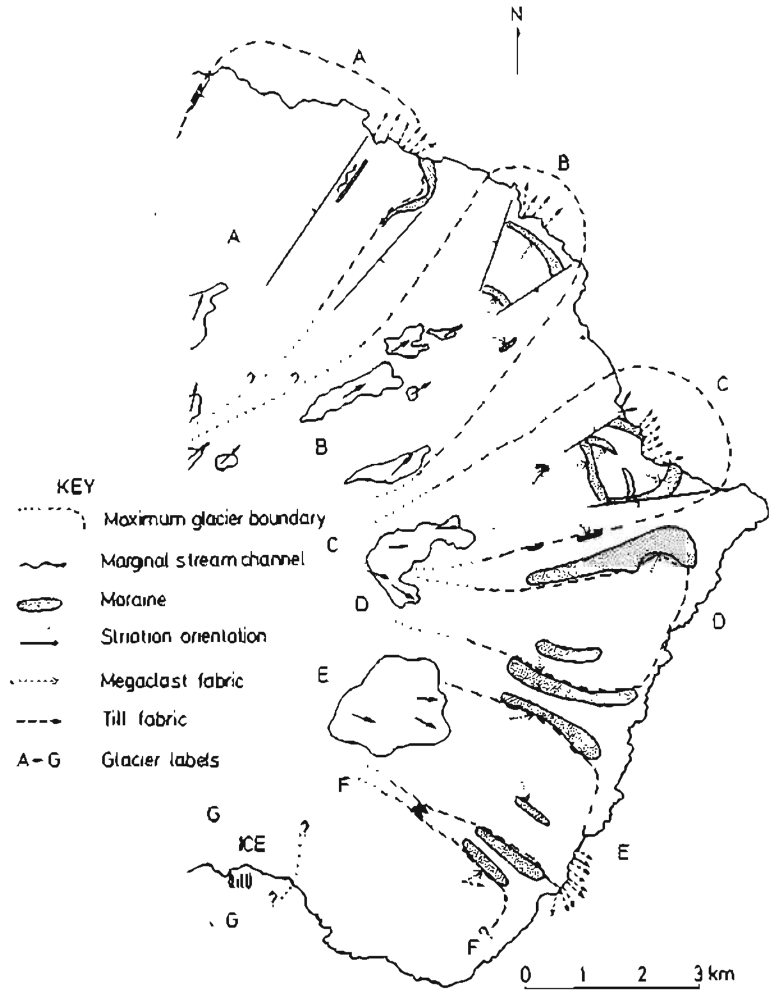

Fig.8. Reconstruction of the glacial maximum ice cover as indicated by the sum of the available information.

further information on palaeo-ice flow directions. The sum of this evidence suggests the positions of several former glaciers (Fig.8).

Whilst glaciers *D* and *E* are reconstructed fairly rigorously, the margins of *A*, *B* and *C* are hypothesised to a large extent. However, from the data available, these margins must approximate to the former true positions (Fig.8). With respect to glaciers *F* and *G* one margin of the former glacier is known but otherwise it is only the presence of ice which can be construed. Ice is also known to have occurred at Triegaardt Bay (striations), in the southwest of the island (striations), at Goodhope Bay (till) and at the western end of Crawford Bay (moraines) but lack of further information precludes any reconstruction.

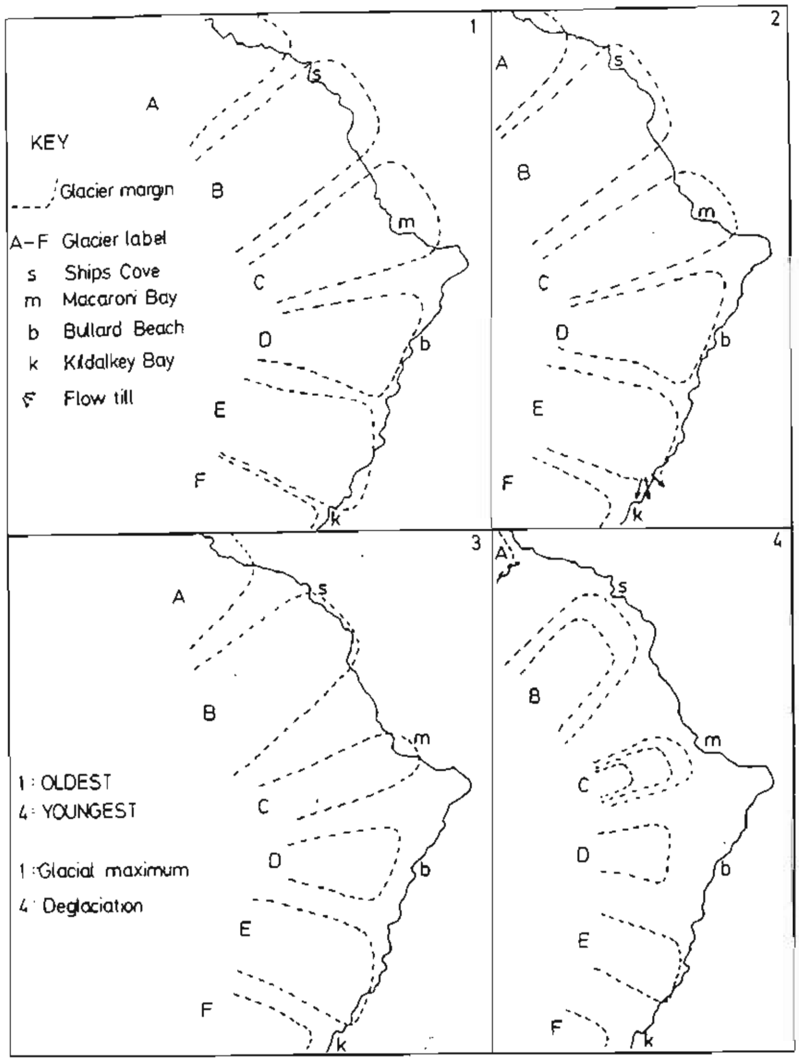

Fig.9. Variation in glacier extent in the series of stadials from the last (Würm-age) Glacial.

The seaward projection of the ice in the cases of glaciers *A*, *B* and *C* (Fig.8) is particularly contentious as there is no information whatsoever from which to derive an estimate. Thus, the seaward extension of these glaciers may have been far more extensive, particularly in the case of glaciers *B* and *C*. For glacier *A* the lateral moraine to the south does appear to be curving towards a terminal position and thus the ice projection may be a good approximation.

Whilst Fig.8 indicates the maximum hypothesised extent of the glaciers, it is also possible to estimate the ice cover during each stadial by combining moraine information and the coastal till sequences (Fig.9). Again, the seaward projection of glaciers *A*, *B* and *C* is little more than a rough guide. It can be seen that the maximum extent of the ice was early in the glacial and that glacier size diminished with each succeeding stadial.

At Albatross Lakes (glacier *C*) a ^{14}C date was obtained (Van Zinderen Bakker, 1973) between Moraines *3* and *4* (Fig.4). An extrapolated date of 15 000 to 17 000 B.P. for the base of the sequence indicates that moraine *3* must pre-date this whilst Moraine *4* post-date it. Tentatively, these dates show some agreement with those of Mercer (1976) from southernmost South America. Mercer found that during the last glaciation the glaciers were most extensive prior to 56 000 B.P. and that at successive cold peaks at 24 000 B.P., 19 500 B.P. and 14 500 B.P. the glaciers were smaller. A pronounced warm interval is considered to be centred on 15 500 B.P. Thus moraine *3* (Fig.4) may equate, in the light of the available ^{14}C date, to the 19 500 B.P. stadial and Moraine *4* to the 14 500 B.P. event. If this assumption were correct then possibly moraines *1* and *2* equate with the earlier stades indicated by Mercer (1976). The successively diminishing glacier extents agree well with the sequences observed by Mercer. The world-wide positioning of the glaciers within their present borders at 11 000 B.P. (Kukla, 1977) fits well with the palynological evidence of Schalke and Van Zinderen Bakker (1971).

The proposed distribution of glaciers fits well with the suggestion of Schalke and Van Zinderen Bakker (1971) in that plant species "overwintered" the glacial on Marion and that it is not necessary to postulate an extra-island refugium. The ice-free areas within the confines of the present island boundary could have allowed the maintenance of plant species in addition to the areas made available by lower sea levels. It is suggested (see below) that temperatures in the coastal ice-free areas would not have been too harsh so as to limit growth.

RECONSTRUCTION OF GLACIAL-MAXIMUM TEMPERATURES

A lateral moraine is formed only along the edge of a glacier in the ablation zone (Andrews, 1975) and thus the highest topographic location of such a moraine equates to the approximate position of the equilibrium line altitude (ELA) of the former glacier. The climatic snowline of a region is suggested (Østrem, in Andrews, 1975) to range in altitude from that of the ELA to 300 m above. Thus, the lateral moraines can be used to approximate the altitudinal range within which the palaeo-snowline is likely to have occurred during the formation of that moraine.

On Marion Island, there occurs a relatively undisturbed series of lateral moraines between Stony Ridge and Green Hill (Fig.2). It is most noticeable in the field that all the outer (those demarcating the maximum extent of the ice) sets of lateral moraines start at approximately 250 m a.s.l. Within this area no faults with a throw of more than 6 m have been found. In addition there is no evidence so far available to show an isostatic rebound of more than approximately 11 m (Hall, 1979). Even assuming a compound altitudinal error in the order of 50 m this is not of sufficient magnitude to complicate the speculative method of value derivation employed here. Thus, the lower possible limit of the palaeo-snowline at maximum glacial conditions is taken as 250 m and the upper limit as 550 m.

Radiosonde soundings have shown (Schulze, 1971) that the winter rate of fall of temperature with altitude is 4.5°C 1000 m^{-1}. This value is used here as a minimum estimation of the temperature gradient under glacial conditions. Finally for the calculation palaeo-temperature values of 0°C, −1°C and −2°C (mean annual) are assigned to the snowline at the two altitudinal extremes in order to obtain the possible range of temperatures.

Using the above assumptions, Table I shows the mean monthly temperatures at present sea level altitude resulting from the combination of different snowline altitudes and temperatures. It can be seen that, for the variables used, the possible temperature depression ranges between 3°C (0°C snowline at 550 m) and 6.4°C (−2°C snowline at 250 m).

From palynological work on Marion Island, Van Zinderen Bakker (1973) suggested that there was a decrease in the mean annual temperature of 3 to 4°C. Hays et al (1976) show that ocean cores in the southern Indian Ocean indicate a depression of 2.5—3.5°C. Thus it would appear that a snowline temperature of 0°C at 250 m or 550 m, or a snowline temperature of −1°C at 550 m would provide mean annual temperatures compatable with other palaeo-temperature evidence.

TABLE I

Reconstruction of temperatures at sea level and snowline at time of maximum glaciation (Würm-age) together with differences from present-day temperatures.

Month	A	B	C	D	E	F
Mean annual temperature at snowline taken as 0°C						
January	7.2	+2.9	4.3	+4.3	2.9	+1.8°C
February	7.8	+2.4	5.4	+4.8	3.0	+2.3°C
March	7.5	+3.3	4.2	+4.5	3.0	+2.0°C
April	6.2	+1.8	4.4	+3.2	3.0	+0.7°C
May	4.9	+0.5	4.4	+1.9	3.0	−0.6°C
June	4.3	−0.2	4.5	+1.2	3.1	−1.3°C
July	3.8	−0.6	4.4	+0.8	3.0	−1.7°C
August	3.6	−0.8	4.4	+0.6	3.0	−1.9°C
September	3.6	−0.9	4.5	+0.5	3.1	−2.0°C
October	4.8	+0.4	4.4	+1.8	3.0	−0.7°C
November	5.7	+1.3	4.4	+2.7	3.0	+0.2°C
December	6.4	+2.0	4.4	+3.4	3.0	+0.9°C
$\bar{x}$	5.5	1.1		2.5		0°C

Fall in temperature 3 to 4.4°C (annual mean)

Month	A	B	C	D	E	F
Mean annual temperature at snowline taken as −1°C						
January	7.2	+1.9	5.3	+3.3	3.9	+0.8°C
February	7.8	+2.4	5.4	+3.8	4.0	+1.3°C
March	7.5	+2.1	5.4	+3.5	4.0	+1.0°C
April	6.2	+0.8	5.4	+2.2	4.0	−0.3°C
May	4.9	−0.5	5.4	+0.9	4.0	−1.6°C
June	4.3	−1.2	5.5	+0.2	4.1	−2.3°C
July	3.8	−1.6	5.4	−0.2	4.0	−2.7°C
August	3.6	−1.8	5.4	−0.4	4.0	−2.9°C
September	3.6	−1.9	5.5	−0.5	4.1	−3.0°C
October	4.8	−0.6	5.4	+0.8	4.0	−1.7°C
November	5.7	+0.3	6.0	+1.7	4.0	−0.8°C
December	6.4	+1.0	7.4	+2.4	4.0	−0.1°C
$\bar{x}$	5.5	+0.1		+1.5		−1.0°C
Fall in temperature 4 to 5.4°C (annual mean)						
Mean annual temperature at snowline taken as −2°C						
January	7.2	+0.9	6.3	+2.3	4.9	−0.2°C
February	7.8	+1.4	6.4	+2.8	5.0	+0.3°C
March	7.5	+1.1	6.4	+2.5	5.0	0.0°C
April	6.2	−0.2	6.2	+1.2	5.0	−1.3°C
May	4.9	−1.5	6.4	−0.1	5.0	−2.6°C
June	4.3	−2.3	6.6	−0.8	5.1	−3.3°C
July	3.8	−2.6	6.4	−1.2	5.0	−3.7°C
August	3.6	−2.8	6.4	−1.4	5.0	−3.9°C
September	3.6	−2.9	6.5	−1.5	5.1	−4.0°C
October	4.8	−1.6	6.4	−0.2	5.0	−2.7°C
November	5.7	−0.7	5.0	+0.7	5.0	−1.8°C
December	6.4	−0.0	6.4	+1.4	5.0	−1.1°C
$\bar{x}$	5.5	−0.9		+0.5		−2.0°C
Fall in temperature 5 to 6.4°C (annual mean).						

A = present monthly mean temperature at sea level; B = calculated temperatures, for snowline at 250 m, at sea level; C = A − B; D = calculated temperatures, for snowline at 550 m, at sea level; E = A − D; F = monthly mean temperature at the snowline; (Present monthly mean temperatures as given by Schulze, 1971, table 13)

The monthly temperatures suggested by the above combinations (Table I) are in accordance with the findings of Schalke and Van Zinderen Bakker (1971) in that plant species "overwintered" the glacials on Marion. In addition, the warm summer temperatures would indicate high ablation rates which would need to be compensated for by a high annual accumulation. This, in turn, would necessitate rapid transference of mass from the accumulation to the ablation zone so as to maintain the glacier. Other data available (Hall, 1978a), particularly the high clast dips observed in the tills, tend to indicate high ice velocities. Thus, there is an indirect line of evidence to justify the computed range of temperatures. Although the approach is rather circular and some of the premises tenuous, the data presented in

Table I do appear to provide a working first approximation of temperatures during the last Glacial maximum and which are in good accord with data from other sources.

SUMMARY

From the data so far available it has been possible to attempt a reconstruction of the glacier cover for the eastern side of the island during the last-Glacial. In addition, it is possible to note the presence of ice at several other localities although the actual glacier form cannot be delineated. The variation in glacier size during subsequent stadials can also be shown and this continuing decrease in glacier size agrees well with other Southern Hemisphere observations. The postulated ice-free areas between glaciers, within the warmer coastal lowlands, would allow the survival of island flora as has been indicated by palynological work. Finally, the hypothesised glacial temperatures not only agree well with those derived by other methods but are also commensurate with the maintenance of island plant life.

This reconstruction of island conditions during the last Glacial provides the first such attempt for the sub-Antarctic. It is hoped that future work will provide more detail which will allow a more rigorous reconstruction, particularly with respect to the off-shore glacier termini and the western side of the island.

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Push Moraines on Marion Island

Recent studies have shown that Marion Island (46°S; 37°E) was subject to multiple glaciation during the last 300 000 years.[1,2] Of the three glacials which are considered to have affected the island,[2,3] the deposits of the first two are exposed only along coastal cliffs or incised stream courses. However, approximately 20% of the island is seen to be affected by either glacio-depositional or glacio-erosional features resulting from the most recent glacial (c. 70 000 BP to c. 11 000 BP).[3] At two localities, Skua Ridge and Albatross Lakes (Fig. 1), a number of moraines, considered to be of the so-called 'push' variety, are apparent. These moraines are thought[3] to be part of a sequence resulting form a series of stades (periods of major ice advance within the glacial), of diminishing intensity, during the last glacial. This paper attempts to describe these moraines and justify their push-type genesis.

A push moraine is considered to result from the advancing of an ice front which 'may bulldoze loose material in its path'[4] and so 'push material into a ridge form'.[5] Whilst the majority of features diagnostic of a push moraine are internal, there are a number of surface criteria which allow identification. The moraine usually exhibits a characteristic asymmetry with 'a long, rather gentle proximal slope with a steep (20° to 33°) distal size'.[6] In addition to the asymmetry, Andrews[6] also notes that on the proximal slope the surface boulders partially reflect the internal till fabric and exhibit a long-axis orientation normal to the ridge crest. Mathews *et al.*[7] also recorded the distinct asymmetry and observed, at the location they were studying, a marked 'saw-tooth' plan outline to the moraines.

Such moraines are known from a variety of environments and areas, for example Baffin Island and Axel Heiberg Island in the Canadian Arctic,[6,8] Vatnajökull and Breidamerkurjökull in Iceland,[9,10] Norrboten in Sweden,[11] Bødalsbreen in southern Norway,[7] and the Himalayas.[12]

The push mechanism is but one of a multitude of means whereby moraine ridges may be formed.[7,10] However, Mathews *et al.*[7] suggest that 'pushing may be an underestimated mechanism in moraine ridge formation generally', and that 'greater consideration be given to this mechanism'. Thus the investigation of the moraines on Marion Island is important, not only to add to the geomorphological knowledge of this area but, if they can be shown to be of push origin, to describe another location where these features occur.

Skua Ridge moraines

At Skua Ridge there occurs a 200 to 300 m wide belt of asymmetrical moraine ridges. The asymmetry of these ridges is quite pronounced, with the mean distal slope (16°) being twice that of the proximal slope (8°). Comparison of paired distal and proximal slope angles suggests that a positive correlation exists ($r = 0.80$) which can be considered significant by the t-test of r ($t = 5.73$; reject H_0 at 0.001 level). The relationship between the distal and proximal slopes is best expressed by the linear regression equation, $y = 5.27 + 1.26x$ (Fig. 2).

In addition to the distinct asymmetry, the outermost (seaward) ridge shows evidence of being a multiple moraine resulting from at least two successive ice advances [1-3] (Fig. 3).

Surface megaclast orientation fabrics show very strong a-axis alignment normal to the ridge crests, particularly in the case of the partially overriden outermost ridge (No. 2 in Fig. 4). In the five megaclast fabric examples shown (Fig. 4), the strong clast orienta-

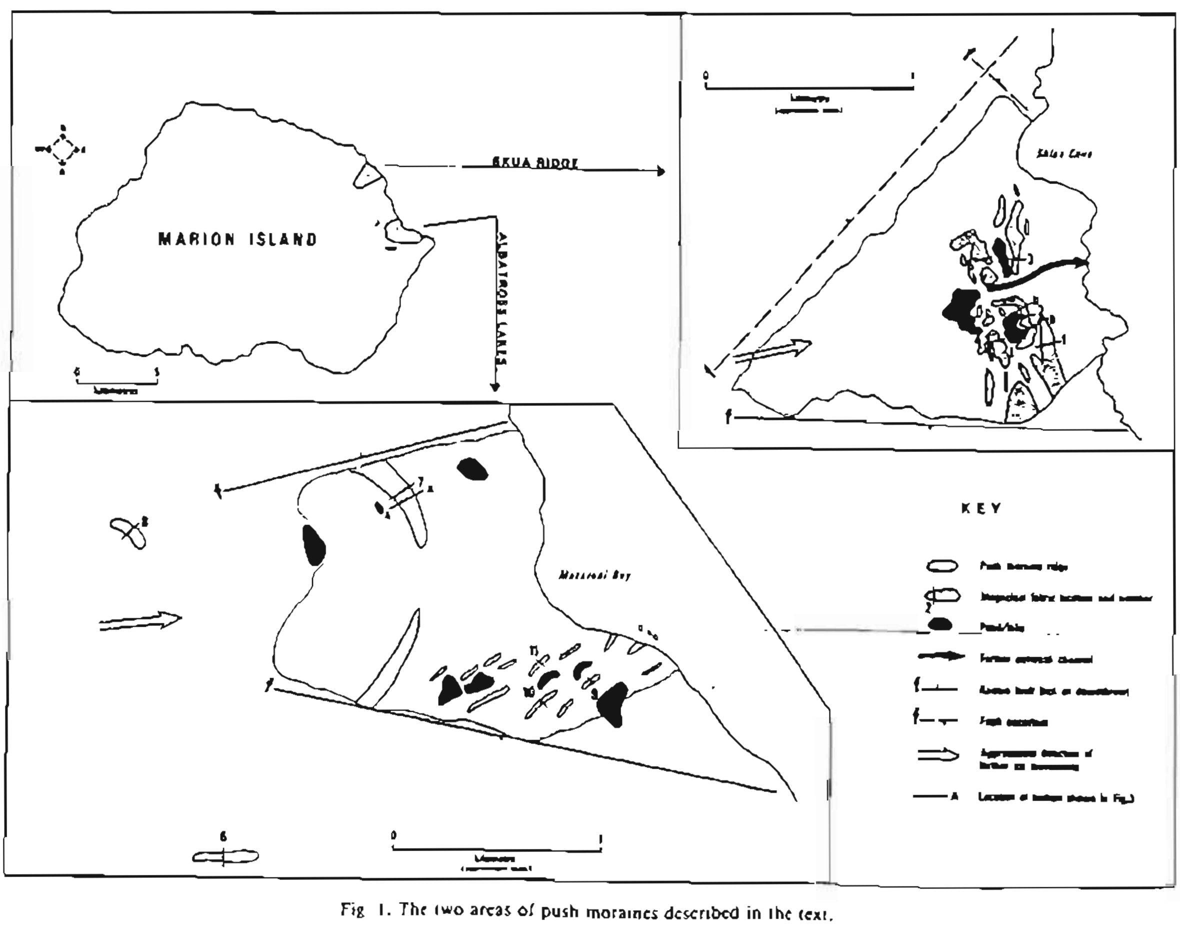

Fig 1. The two areas of push moraines described in the text.

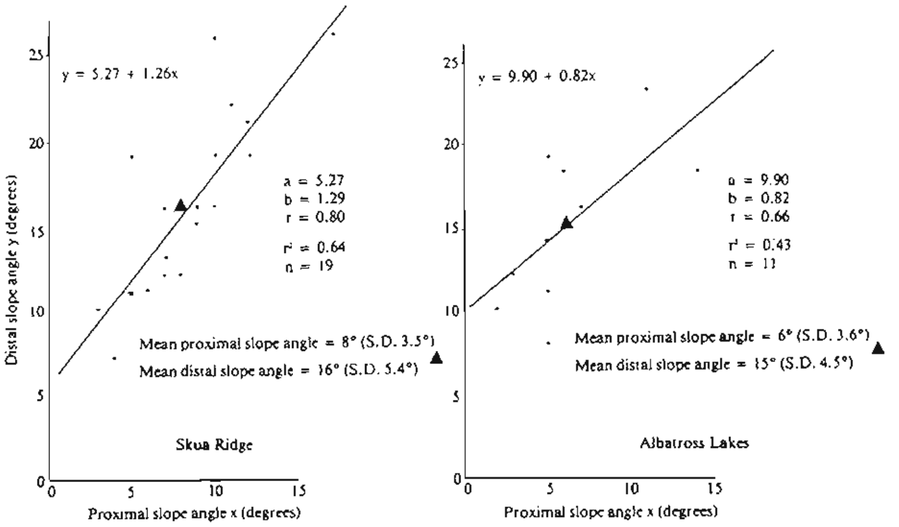

Fig. 2. Proximal and distal slope angles, with regression lines indicated, for Skua Ridge and Albatross Lakes.

tion normal to the ridge axes is clearly seen. It is also evident, as might be expected, that there is a stronger, narrower distribution on the proximal slope, whilst on the distal side there is a weaker, broader distribution.

It is noticeable that many of the moraine ridges have a 'protuberance' on their proximal side, which is approximately normal to the main ridge axis. These secondary ridges are relatively narrow with respect to the main ridge and are at their widest where the two ridges join. They may represent longitudinal crevass fillings at the glacier snout at the time of moraine formation.

Albatross Lake moraines

In this area there is a longer sequence of moraines than is found at Skua Ridge, of which some, but not all, appear to be of the push variety (Fig. 1). Measurement of paired proximal and distal slope angles again shows the characteristic asymmetry with a steeper distal (15°) than proximal (6°) mean slope angle (Fig. 2). Correlation of paired values is less strong ($r = 0.66$) but is still considered significant by the t-test of r ($t = 2.62$). The relationship between the proximal and distal slopes ($r^2 = 0.43$) may also be expressed by a linear equation (Fig. 2): $y = 9.90 + 0.82x$.

It is interesting to note that a multiple moraine, of similar appearance to that of Skua Ridge, is found at this location (Fig. 3). This moraine, which is extremely prominent, also exhibits a strong surface megaclast alignment on the proximal slope and a broader spread on the distal (No. 7 in Fig. 4). This moraine also exhibits some evidence to suggest partial overriding. On the moraine top, towards the proximal side, a large number of severely striated boulders are found. The striations are frequently observed parallel to the a-axis of the boulders. The mean orientation of the striations on the *in situ* boulders was found to be 269° ($s = 29$). This value is very close to that of the preferred orientation of the boulders. Andrews[13] describes a similar situation, in Canada, where 'striations were all parallel to the a-axis and the majority were close to the direction of the preferred orientation. The qualitative impression in the field was that the ice had overriden the boulders.'

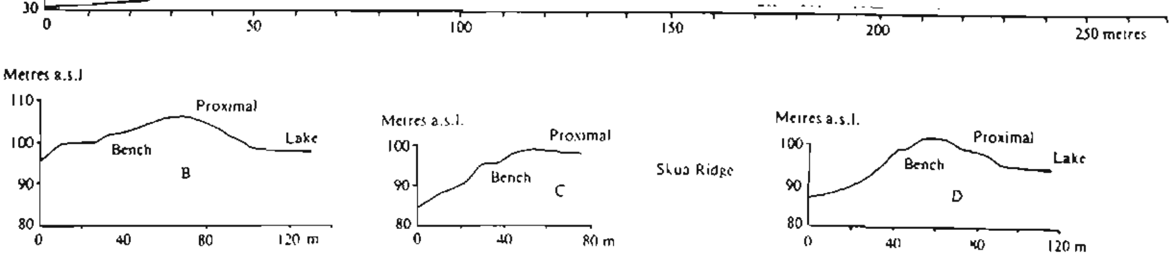

Fig. 3. Surveyed profiles from Skua Ridge and Albatross Lakes to show the bench feature observed on the distal side.

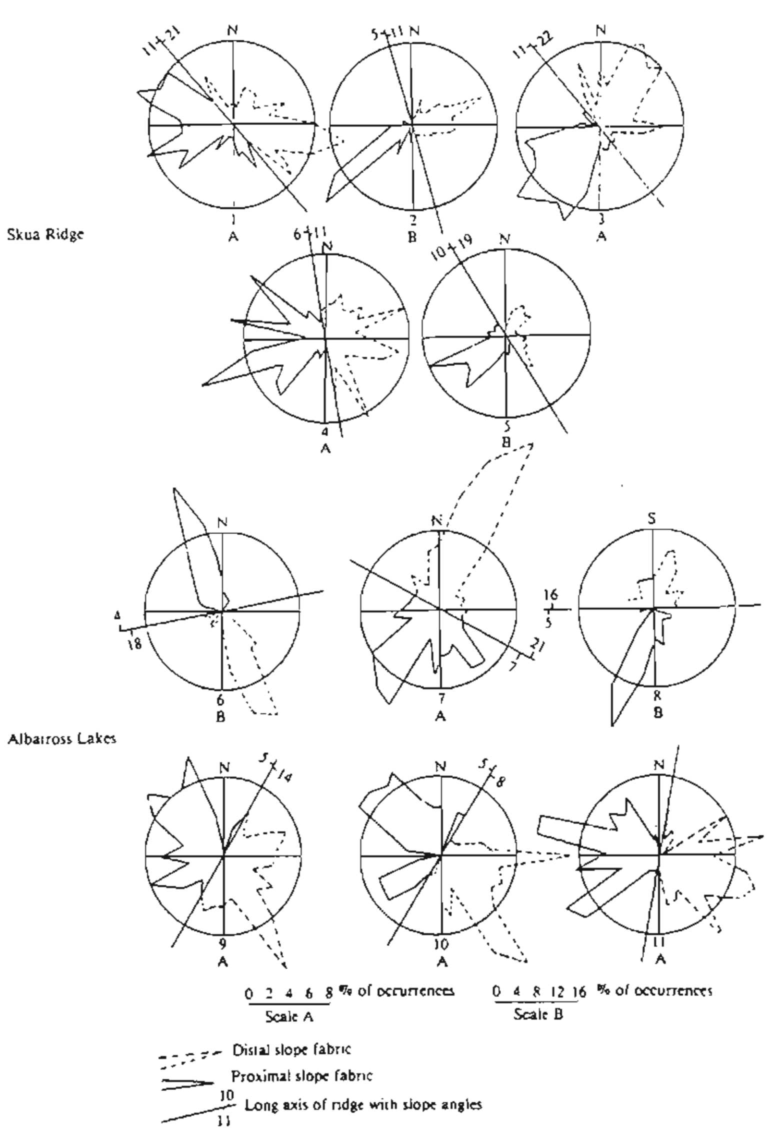

Fig. 4. Surface megaclast orientation fabrics from selected moraines at Skua Ridge and Albatross Lakes.

Megaclast orientation (Fig. 4) varies between extremely strong, narrow distributions (Nos 6 and 8) and rather weak, broad alignments (Nos 9 and 10). With the exception of fabric No. 6, all the distal slopes exhibit a far broader range of a-axis alignments than is found on the proximal slope. Possibly the very strong unidirectional fabric of No. 6 is due to complete overriding by ice, but unfortunately further information is lacking as the majority of the moraine has been inundated by recent lava flows.

Discussion

There appears to be little difference in moraine morphology between the two sites except that the apparent relationship between the proximal and distal slope is stronger at Skua Ridge (r = 0.80) than at Albatross Lakes (r = 0.66). It is noticeable that the distal slope angles at both locations show standard deviations that are greater than for the proximal slopes. The mean of all the distal slope observations was 15° (s = 5.02), whilst that of the proximals was 7° (s = 3.59). Correlation of all sample pairs shows an r value of + 0.75 (r^2 = 0.56) with an overall covariance of 13.5

Whilst the distinct asymmetry and pronounced surface megaclast fabrics are in accordance with a push moraine origin, consideration must be given to other possible mechanisms. Surging of glaciers can generate moraines,[14] but they are generally contorted and chaotic and this does not correspond with the regularity observed on Marion Island. Squeezing of water-soaked till from beneath a glacier snout[15] is a potential mechanism but the Marion Island moraines are too high in relation to the considered ice thickness (as indicated by lateral moraines) for this to be possible. It is possible that shearing or surface melting would bring debris bands to the ice surface to produce moraines in the terminal area but the strong a-axis orientation and evidence of overriding together with the relatively large size of the moraines argue against this. Supraglacial transport of debris which accumulated in 'dump' moraines[6,16] can also be discounted on the basis of the asymmetry, the steep distal slopes (dump moraines tend to have steep proximal slopes[13]) and strong surface megaclast orientation.

The push or bulldozer origin appears to offer the greatest accord with available data. Although pushing is not the only mechanism capable of producing asymmetry, 'Asymmetrical moraine cross-profiles, with relatively steep distal slopes . . . have been attributed by many writers to a push mechanism . . .'[7] The strong megaclast a-axis alignment and evidence of overriding are in accord with a push moraine hypothesis.[6,11] Also, as Mathews *et al.*[7] state, the continuity of ridge outcrop suggests a 'controlled' process of formation which would be of equal effectiveness across the whole glacier front, and this can best be obtained by a push process. Finally, none of the other mechanisms would be so likely, when considered together with the other evidence, to produce such a clear image of the glacier front (including the terminal crevasses).

If the assumptions are correct and the moraines are of the push type, then this is of value in the geomorphological and palaeoclimatological investigation of Marion Island. The mapping of the moraines helps delimit the former glacier distribution and to show the diminishing glacier size during the glacial.[3] It is hoped that future work will produce radiocarbon dates for each of the moraines so that a dated sequence might be obtained. A push moraine implies that the snout of the glacier was composed of temperate rather than cold ice (where the snout would be frozen to the glacier bed and movement would be by internal shear) and this is in agreement with reconstructed temperatures for the terminal areas of the glacier during the glacial maximum.[3]

Thus this information not only adds to the data on push moraines but also helps substantiate and add to the knowledge of conditions during the last glacial on Marion Island. That the data are in accord with the present interpretation of palaeo conditions on Marion Island helps substantiate the push hypothesis and, in addition, support that interpretation.

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Quantitative Analysis of Till Lithology on Marion Island

Investigations[1-3] of the Quaternary glacial geology of Marion Island in the sub-Antarctic have shown that the island was subject to three glacial episodes which appear to equate with the proposed glacial cycles D, C and B of Kukla.[3] Within each glacial sequence analysis of the stratigraphic succession suggests the occurrence of stadial and interstadial conditions. The stadial condition is represented by a lodgement till[4] with subrounded, non-platy clasts which occasionally exhibit striations. The interstadial state is marked by an ablation till,[5] which is characterized by angular and platy clasts with little or no indication of striations, overlain by outwash. Flow till, melt-out till and interfluctuational till have also been recognised but do not occur in sufficient numbers to allow comparison.

Subjective description of the various till units precludes rigorous comparison and assessment of the mode of formation. In an attempt to overcome these problems and facilitate objective description of changes in till lithology, I have measured clast parameters relating to shape and size. Clast shape not only allows between-till comparisons but is also diagnostic, to some extent, of till origin.[6,7] Statistical comparison of the resulting indices provides a quantitative basis for differentiation of tills. Measurements of clast size, percentage of striations, and rock composition afford additional means of comparison.

Techniques

Samples of 50 clasts from different till units were evaluated to obtain the following information: 1) a-axis length (a), 2) b-axis length (b), 3) c-axis length (c), 4) minimum radius of curvature in the principal plane (r), 5) the presence of striations, and 6) whether the sample was pyroclastic or of grey lava (the only two rock types present).

The sample area in each till unit measured 1 × 1 m and consisted of clasts within the size range 0.01 to 0.20 m along the a-axis (most were in the size range 0.01 – 0.05 m) and with an $a:b$ ratio of 2 : 1. The results of measurements (1) to (4) above were used to derive a number of parameters related to clast shape:

(i) Cailleux's[8] index of roundness (P_r)

$$P_r = (2r/a) \times 1000,$$

where the minimum roundness = 0 and maximum roundness = 1000

(ii) Cailleux's index of flatness (F_r)

$$F_r = [(a + b)/2c] \times 100,$$

where the minimum flatness = 100 and maximum flatness = ∞.

(iii) Krumbein's[11] measure of sphericity (S_r)

$$S_r = \sqrt[3]{(bc/a^2)},$$

where minimum sphericity = 0.01 and maximum sphericity = 1.00

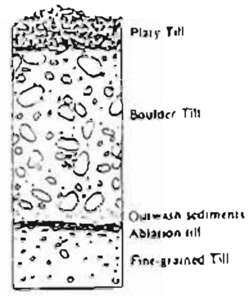

Fig. 1. Schematic stratigraphic column for the till sequence observed at Macaroni Rocks.

The sample mean ($\bar{x}$) and standard deviation (s) were also derived as measures of the clast characteristics for each till. The data for each sample were plotted on arithmetic probability paper to test for normality. The majority of samples approximated to a normal distribution, with a few indicating a slight positive skewness. Successive tills were thus able to be compared, with respect to P_r, F_r and S_r, by means of the F-test,[9] with rejection set at the 0.05 level. Those samples considered to be part of the same population were then more rigorously tested by means of the t-test,[9] where

$$t = \frac{\bar{x}_1 - \bar{x}_2}{S\sqrt{[(1/n_1) + (1/n_2)]}}$$

where

$$S = \left(\frac{n_1 s_1^2 + n_2 s_2^2}{n_1 + n_2 - 2}\right)^{1/2}$$

As a check, particularly in the case of the slightly skewed samples, the data were further tested by means of two non-parametric tests, namely the Wilcoxon-Mann-Witney U-test[9] and the Kruskall-Wallis test.[10] Results from both these additional tests confirmed those indicated by the t-test and so only the results of the latter are tabulated. Conversion of the observed number of striated and pyroclastic clasts into percentages allows direct comparison between tills.

In addition to the above data, measurements were taken of the 10 largest clasts within a 1 m radius of the sample areas. Thus for each till unit, clast shape and maximum size together with the percentage of striated and pyroclastic clasts were recorded. These results provide sufficient basis for comparison of tills.

Results

Analysis of the till succession at Macaroni Rocks

Macaroni Rocks is the name given to the south-eastern extremity of Macaroni Bay and is an ideal site for the evaluation of this method of analysis. A sequence of four tills was observed, and relate to glacial cycle B (Fig. 1). At the base of the exposure there was a fine-grained till which had few clasts of cobble-size or larger ('Fine-grained Till'). This was capped by an ablation till and outwash sediments, above which was a thick till characterized by many boulders and large cobbles ('Boulder Till'). At the cliff top there

Table 1. Measurements recorded at Macaroni Rocks.

	Sample of 50						Sample of 10								Sample of 50
							c						m^3		
Till	P_r	s	F_r	s	S_r	s	$\bar{x}a$	s	$\bar{x}b$	s	$\bar{x}c$	s	$\bar{x}m^3$	s	% striated
Platy	137.7	117.1	390.9	329.5	0.5616	0.1331	0.35	0.05	0.23	0.07	0.12	0.06	0.011	[illegible]	2
Boulder	241.1	90.4	127.1	35.6	0.7186	0.0731	1.20	0.17	0.86	0.16	0.68	0.17	0.743	0.382	12
Ablation	75.5	57.9	430.5	291.3	0.51	0.10	0.41	0.11	0.30	0.12	0.086	0.047	0.14	0.015	0
Fine	281.2	181.7	173.5	30.1	0.7127	0.0598	0.39	0.15	0.25	0.10	0.16	0.10	0.26	0.046	8

Table 2 Statistical comparison of the tills observed at Macaroni Rocks

Platy Till = 1 Boulder Till = 2 Fine-grained Till = 3 Ablation till = 4

F-tests

$H_0 : (s_1^2 = s_2^2 = s^2)$ $H_a : (s_1^2 \neq s_2^2)$

Tills	Units	F	Decision
1 - 2	Pi	1.7	Accept H_0 at 0.05 level, n_1 = 50 n_2 = 50
1 - 2	Fi	86.6	Reject " " " " " " " " " "
1 - 2	Si	3.3	Reject " " " " " " " " " "
1 - 3	Pi	1.3	Accept " " " " " " " " " "
1 - 3	Fi	119.8	Reject " " " " " " " " " "
1 - 3	Si	4.9	Reject " " " " " " " " " "
3 - 2	Pi	1.3	Accept " " " " " " " " " "
3 - 2	Fi	1.4	Accept " " " " " " " " " "
3 - 2	Si	1.4	Accept " " " " " " " " " "
3 - 4	Pi	3.1	Reject " " " " " " " " " "
3 - 4	Fi	93.7	Reject " " " " " " " " " "
3 - 4	Si	1.8	Reject " " " " " " " " " "
1 - 4	Pi	4.09	Reject " " " " " " " " " "
1 - 4	Fi	1.28	Accept " " " " " " " " " "
1 - 4	Si	1.77	Reject " " " " " " " " " "
Mx2 - Mx3	vol	130.4	Reject H_0 at 0.05 level, n_1 = 10 n_2 = 10
Mx2 - Mx1	vol	61.9	Reject " " " " " " " " " "
Mx2 - Mx3	a-axis	1.3	Accept " " " " " " " " " "
Mx2 - Mx1	a-axis	11.6	Reject " " " " " " " " " "

t-tests of till pairs accepted above

$H_0 : (U_1 = U_2 = U, s_1^2 = s_2^2 = s^2)$ $H_a : (U_1 \neq U_2)$

Tills	Units	t	Decision
1 - 2	Pi	5.0	Reject H_0 at 0.05 level df = 98
1 - 3	Pi	6.5	Reject " " " " " " " "
3 - 2	Pi	2.0	Accept " " " " " " " "
3 - 2	Fi	1.0	Accept " " " " " " " "
3 - 2	Si	1.0	Accept " " " " " " " "
1 - 4	Fi	0.54	Accept " " " " " " " "
Mx2 - Mx1	a-axis	11.5	Reject H_0 at 0.05 level df = 18

was a till, associated with a lateral moraine cut by the cliff at this point, rich in angular and platy clasts ('Platy Till').

The measurements taken at this site are recorded in Table 1, which shows the greater percentage of striated clasts in the two lodgement tills (the 'Fine-grained Till' and 'Boulder Till'). Note also the relatively large maximum clast size of the Boulder Till, and the relatively pronounced flatness and angularity of the ablation tills, whereas the Platy Till was characterized by flat, angular clasts, with a large sample standard deviation and less pronounced indices than the ablation till.

The information presented in Table 2 can be summarized as follows:

(i) The flatness, roundness and sphericity of the clasts in the Platy Till differ from those of the Fine-grained and Boulder Tills.

(ii) The three clast indices for the Boulder and Fine-grained Tills indicate the samples came from the same population.

(iii) The clasts from the ablation till differ from those of the underlying Fine-grained Till in terms of all three indices.

(iv) The ablation till and Platy Till differ in terms of clast roundness and sphericity (the Platy Till shows greater overall clast roundness and sphericity) but are similar in terms of clast flatness

Analysis of the till succession at Ships Cove

Ships Cove, which is located just to the north-west of the Meteorological Station, is a far more complex site than Macaroni Rocks and exhibits 14 clearly definable tills resulting from three glacial periods (D, C and B)[1] (Fig. 2). At the base of the succession four tills of cycle D age were found: three lodgement tills separated by fluvial sediments, and an ablation till. The cycle C sequence is initiated by what is termed a 'Pyroclastic Till'.[1] This definition is based upon the visual appraisal of a high percentage of pyroclastic clasts (which give the till a reddish-brown colour), which are the result of the ice of the first stade having cleared the surficial, unconsolidated volcanic debris of the interglacial volcanic event. This is succeeded by an interstadial deposit consisting of a melt-out till and its capping of an ablation till. In addition to a thick lodgement till ('Main Till'), this cycle is completed by an ablation till. The cycle B deposits, of Würm age, are initiated by a Pyroclastic Till above which there occurs a sequence of alternating ablation and lodgement ('Middle' and 'Uppermost') tills.

The measurements for this sequence are shown in Table 3 and the statistical analysis in Table 4. The results of Table 4 are summarized as follows.

Glacial cycle D.

(i) The top of the ablation till differs from the Upper Till in terms of the P_i, F_i and S_i indices.

(ii) The three indices for the Upper and Middle Tills indicate that the clasts are from the same population.

(iii) The Middle Till is similar to the Lower Till with respect to F_i and S_i but differs in terms of P_i.

(iv) The sizes of the clasts in the Middle and Lower Tills are significantly different.

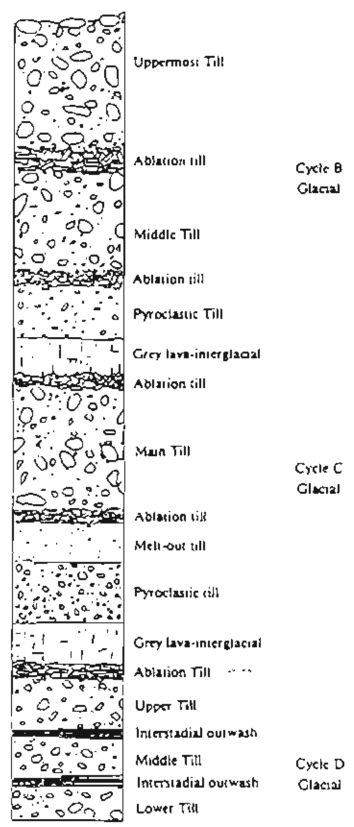

Fig. 2. Schematic stratigraphic column for the three glacial sequences at Ships Cove.

Table 3. Measurements recorded at Ships Cove.

	TILL	Sample of 50						Sample of 10: m						m³			
		Pi	s	Fi	s	Si	s	ā	s	āb	s	āc	s	ām³	s	A	B
┌	Uppermost	209.2	67.5	156.2	26.5	0.7539	0.0799	0.26	0.05	0.21	0.03	0.17	0.03	0.0105	0.0040	6	0
│	Ablation	76.0	58.9	295.0	217.9	0.6180	0.1241	0.41	0.12	0.26	0.07	0.17	0.06	0.0161	0.0146	0	0
│	Middle	249.5	75.3	161.0	27.4	0.7502	0.0844	0.40	0.13	0.33	0.14	0.25	0.10	0.0479	0.0733	6	8
(1)	Ablation	73.2	63.9	325.7	135.0	0.6292	0.0985										0
└	Pyroclastic	219.8	94.3	176.3	66.8	0.7095	0.0835	0.37	0.08	0.30	0.07	0.23	0.04	0.0271	0.0147	1	42
┌	Ablation	104.8	46.6	218.0	56.8	0.6631	0.0891									0	0
│	Main	214.6	93.8	158.6	79.5	0.7241	0.0646	0.39	0.13	0.25	0.10	0.17	0.06	0.0208	0.0183	9	2
│	Ablation	114.8	65.3	287.1	191.9	0.6405	0.1118	0.41	0.14	0.27	0.09	0.13	0.06	0.0187	0.0193	0	2
(2)	Melt-out	224.1	88.1	188.0	45.8	0.6780	0.0914	0.136	0.031	0.087	0.023	0.053	0.015	0.0007	0.0006	2	15
└	Pyroclastic	219.7	103.3	167.8	33.4	0.7337	0.0826	0.24	0.07	0.17	0.06	0.12	0.03	0.0056	0.0043	2	48
┌	Ablation	147.5	84.5	322.9	172.8	0.5759	0.1145										
│	Upper	237.0	85.3	150.6	24.3	0.7744	0.0576	0.29	0.06	0.23	0.06	0.17	0.04	0.0125	0.0086		
(3)	Middle	226.0	90.4	151.4	23.4	0.7741	0.0720	0.28	0.15	0.20	0.07	0.15	0.08	0.0145	0.0282		
└	Lower	316.5	68.9	155.6	25.9	0.7830	0.0772	0.23	0.04	0.18	0.05	0.14	0.05	0.0068	0.0049		

(1) = Cycle B (Kuhla, 1977), Wormage A = % of striated clasts (sample of 50)
(2) = Cycle C B = % of pyroclasts (sample of 50)
(3) = Cycle D

Glacial cycle C

(i) The P_i, F_i and S_i indices of the Pyroclastic and Main Tills indicate material from the same population.

(ii) The Main Till and its capping of ablation till differ widely from all three clast indices.

(iii) The melt-out till and Pyroclastic Till show similar P_i indices but differ with respect to F_i and S_i.

(iv) The melt-out till differs from its ablation till capping in terms of P_i and F_i, and to a lesser extent in terms of S_i.

(v) The 10 largest clasts from the melt-out till differ from those of the ablation till.

(vi) The 10 largest clasts from the Pyroclastic Till differ from those of the melt-out till.

(vii) The 10 largest clasts from the Main Till and the Pyroclastic Till differ.

(viii) The ablation tills show a marked lack of striated clasts with respect to the other tills.

(ix) The Pyroclastic Till indicates a far higher percentage of pyroclasts than do succeeding tills.

Glacial cycle B.

(i) The Uppermost Till differs in terms of the indices P_i, F_i and S_i and clast size from the ablation till beneath.

(ii) The Pyroclastic Till differs radically in all three indices from the overlying ablation till.

(iii) The Middle Till shows a marked difference from its ablation till cover in terms of clast indices. Although a-axis lengths are similar, the greater flatness of the ablation till clasts results in an overall smaller clast size.

(iv) The Pyroclastic Till indicates clasts similar to those of the Middle Till in terms of P_i but not F_i and S_i.

(v) Comparison of the Middle Till with the Uppermost Till shows similar values for the F_i and S_i indices but P_i values differ.

(vi) The ablation tills show a markedly lower percentage of striated clasts relative to the other tills.

(vii) The percentage of pyroclastic material is high in the basal Pyroclastic Till and decreases upwards.

Discussion

The study of clast shape and size indicates the characteristic angularity and flatness of clasts in the ablation tills of Marion Island, and the greater roundness and lesser flatness of clasts from the lodgement tills. Plotting of mean values of roundness and flatness against the standard deviation of samples from lodgement and ablation tills (Fig. 3) indicates the mutual exclusiveness of each group, and thus its distinctiveness. In Fig. 3 the melt-out till can be seen to be incorporated within the lodgement till grouping, whereas owing to the high standard deviation value, the Platy Till sample represents an extreme case within the ablation till group.

In addition to using clast indices to differentiate the tills, they can also help substantiate the proposed mode of origin for the till. Drake[4] suggests that 'crushing and abrasion are the dominant factors controlling the final shape of the pebbles'. He finds that spherical objects, that is, those with a low flatness index, have the most durable shape and that which is most likely to be produced in a subglacial environment. King[11] has shown that clast roundness increases with distance from the ice source, and Mills[7] states that 'Basal and recessional-moraine samples have one or two thirds of their clasts in the subangular to subrounded category, but few are more highly rounded'. Conversely, pebbles in the ablation till are shown[7] to be very angular or angular, the freeze-thaw-shattered source material being little modified during transport. In addition,

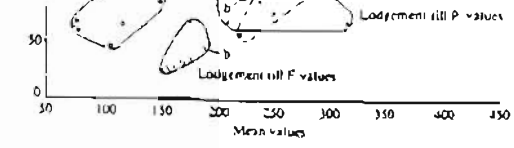

Fig. 3. Plot of the standard deviations of P_i and F_i against the mean values for the tills at Macaroni Rocks and Ships Cove.

Table 5. Statistical comparison of the tills from the three glacial sequences at Ships Cove

Cycle D glacial

1 = Lowest till 2 = Middle till 3 = Upper till 4 = ablation till

F-tests

$$H_0 : (s_1^2 = s_2^2 = s^2) \quad H_A : (s_1^2 \neq s_2^2)$$

Tills	Units	F	Decision
4 + 3	Pi	0.98	Accept H_0 at 0.05 level n_1 = 50 n_2 = 50
4 + 3	Fi	30.37	Reject " " " " " " " " " "
4 + 3	Si	3.95	Reject " " " " " " " " " "
3 + 2	Pi	1.12	Accept " " " " " " " " " "
3 + 2	Fi	1.08	Accept " " " " " " " " " "
3 + 2	Si	1.36	Accept " " " " " " " " " "
2 + 1	Pi	1.22	Accept " " " " " " " " " "
2 + 1	Fi	1.23	Accept " " " " " " " " " "
2 + 1	Si	1.15	Accept " " " " " " " " " "
Mx(3) + Mx(2)	a	6.25	Reject H_0 at 0.05 level n_1 = 10 n_2 = 10
Mx(3) + Mx(2)	vol	10.75	Reject " " " " " " " " " "
Mx(2) + Mx(1)	a	14.06	Reject " " " " " " " " " "
Mx(2) + Mx(1)	vol	33.12	Reject " " " " " " " " " "

t-tests of those pairs accepted above

$$H_0 : (U_1 = U_2 = U : s_1^2 = s_2^2 = s^2) \quad H_A : (U_1 \neq U_2)$$

Tills	Units	t	Decision
4 + 3	Pi	2.81	Reject H_0 at 0.05 level df = 98
3 + 2	Pi	0.44	Accept " " " " " " " "
3 + 2	Fi	0.12	Accept " " " " " " " "
3 + 2	Si	0.20	Accept " " " " " " " "
2 + 1	Pi	3.39	Reject " " " " " " " "
2 + 1	Fi	0.52	Accept " " " " " " " "
2 + 1	Si	0.20	Accept " " " " " " " "

Cycle C glacial

1 = Pyroclastic Till 2 = melt-out till 3 = ablation till 4 = Main Till

5 = ablation till

F-tests

Till	Units	F	Decision
1 + 2	Pi	1.37	Accept H_0 at 0.05 level n_1 = 50 n_2 = 50
1 + 2	Fi	1.88	Accept " " " " " " " " " "
1 + 2	Si	1.22	Accept " " " " " " " " " "
2 + 3	Pi	1.82	Accept " " " " " " " " " "
2 + 3	Fi	17.54	Reject " " " " " " " " " "
2 + 3	Si	1.50	Accept " " " " " " " " " "
4 + 5	Pi	4.05	Reject " " " " " " " " " "
4 + 5	Fi	3.71	Reject " " " " " " " " " "
4 + 5	Si	1.79	Accept " " " " " " " " " "
3 + 5	Pi	1.96	Accept " " " " " " " " " "
3 + 5	Fi	11.40	Reject " " " " " " " " " "
3 + 5	Si	1.57	Accept " " " " " " " " " "
1 + 4	Pi	1.21	Accept " " " " " " " " " "
1 + 4	Fi	1.28	Accept " " " " " " " " " "
1 + 4	Si	1.54	Accept " " " " " " " " " "
Mx(1) + Mx(2)	a	4.5	Reject H_0 at 0.05 level n_1 = 10 n_2 = 10
Mx(1) + Mx(2)	vol	51.36	Reject " " " " " " " " " "
Mx(2) + Mx(3)	a	17.9	Reject " " " " " " " " " "
Mx(2) + Mx(3)	vol	1036.7	Reject " " " " " " " " " "
Mx(1) + Mx(4)	a	3.65	Reject " " " " " " " " " "
Mx(1)	vol	18.11	Reject " " " " " " " " " "

t-tests of pairs accepted above

Tills	Units	t	Decision
1 + 2	Pi	0.23	Accept H_0 at 0.05 level df = 98
1 + 2	Fi	2.52	Reject " " " " " " " "
1 + 2	Si	3.20	Reject " " " " " " " "
2 + 3	Pi	7.05	Reject " " " " " " " "
2 + 3	Si	3.83	Reject " " " " " " " "
4 + 5	Si	2.94	Reject " " " " " " " "

Continued from previous column:

Tills	Units	t	Decision
3 + 5	Pi	0.37	Accept H_0 at 0.05 level df = 98
3 + 5	Si	0.74	Accept " " " " " " " "
1 + 4	Pi	0.26	Accept " " " " " " " "
1 + 4	Fi	0.12	Accept " " " " " " " "
1 + 4	Si	0.64	Accept " " " " " " " "

Cycle B glacial

1 = Pyroclastic till 2 = ablation till 3 = Middle Till 4 = ablation till

5 = Uppermost Till

F-tests

Tills	Units	F	Decision
5 + 4	Pi	1.31	Accept H_0 at 0.05 level n_1 = 50 n_2 = 50
5 + 4	Fi	67.61	Reject " " " " " " " " " "
5 + 4	Si	2.41	Reject " " " " " " " " " "
1 + 2	Pi	2.18	Reject " " " " " " " " " "
1 + 2	Fi	4.08	Reject " " " " " " " " " "
1 + 2	Si	1.39	Accept " " " " " " " " " "
3 + 4	Pi	1.60	Accept " " " " " " " " " "
3 + 4	Fi	63.20	Reject " " " " " " " " " "
3 + 4	Si	2.16	Reject " " " " " " " " " "
2 + 4	Pi	1.18	Accept " " " " " " " " " "
2 + 4	Fi	2.61	Reject " " " " " " " " " "
2 + 4	Si	1.58	Accept " " " " " " " " " "
1 + 3	Pi	5.90	Reject " " " " " " " " " "
1 + 3	Fi	1.60	Accept " " " " " " " " " "
1 + 3	Si	1.02	Accept " " " " " " " " " "
3 + 5	Pi	1.24	Accept " " " " " " " " " "
3 + 5	Fi	1.07	Accept " " " " " " " " " "
3 + 5	Si	1.12	Accept " " " " " " " " " "
Mx(5) + Mx(4)	a	5.76	Reject H_0 at 0.05 level n_1 = 10 n_2 = 10
Mx(5) + Mx(4)	c	4.00	Reject " " " " " " " " " "
Mx(5) + Mx(4)	vol	13.72	Reject " " " " " " " " " "
Mx(1) + Mx(5)	a	2.36	Accept " " " " " " " " " "
Mx(1) + Mx(5)	vol	13.30	Reject " " " " " " " " " "
Mx(3) + Mx(4)	a	1.2	Accept " " " " " " " " " "
Mx(3) + Mx(4)	c	2.8	Accept H_0 at 0.05 level n_1 = 10 n_2 = 10
Mx(3) + Mx(4)	vol	25.2	Reject " " " " " " " " " "
Mx(1) + Mx(3)	a	2.6	Accept " " " " " " " " " "
Mx(1) + Mx(3)	vol	24.9	Reject " " " " " " " " " "
Mx(3) + Mx(5)	a	6.76	Reject " " " " " " " " " "
Mx(3) + Mx(5)	vol	135.8	Reject " " " " " " " " " "

t-tests of pairs accepted above

Tills	Units	t	Decision
5 + 4	Pi	10.32	Reject H_0 at 0.05 level df = 98
1 + 2	Si	2.89	Reject " " " " " " " "
3 + 4	Pi	12.38	Reject " " " " " " " "
2 + 4	Pi	0.15	Accept " " " " " " " "
2 + 4	Si	0.29	Accept " " " " " " " "
1 + 3	Pi	1.74	Accept " " " " " " " "
1 + 3	Si	2.17	Reject " " " " " " " "
3 + 5	Pi	2.82	Reject " " " " " " " "
3 + 5	Fi	0.89	Accept " " " " " " " "
3 + 5	Si	0.38	Accept " " " " " " " "
Mx(1) + Mx(5)	a	2.99	Reject H_0 at 0.05 level df = 18
Mx(3) + Mx(4)	a	0.26	Accept " " " " " " " "
Mx(3) + Mx(4)	c	3.44	Reject " " " " " " " "
Mx(1) + Mx(3)	a	0.57	Accept " " " " " " " "

Mills[7] also suggests that lateral moraine tills have slightly more subangular to subrounded clasts than do the ablation tills.

That the deposits which are considered to be lodgement tills exhibit clasts with high roundness and sphericity indices is in agreement with their considered mode of origin.[4,6,7] Conversely, the let-down ablation till, which has not been subjected to subglacial crushing and abrasion, exhibits the more angular and less spherical shape, which would be expected of clasts little changed from their original state. The moraine origin of the Platy Till is clearly indicated by the landform within which it occurs, yet it would apparently be possible to distinguish it by clast shape parameters alone, for the range of very angular to subrounded clasts gives it a standard deviation greater than that found for ablation tills.

The percentage of striated clasts can also be used as an adjunct indicator of till genesis. Mills[7] has shown that ablation tills exhibit a far lower percentage of striated clasts than do lodgement tills, whilst tills from lateral moraines display an intermediate value (as was found at Macaroni Rocks [Table 1]). This striation information is a reflection of the clast transport mechanism[12] and thus may be indicative of the till mode of origin.

Measurement of the relative percentages of pyroclasts and grey lava debris within a till unit may be an approach only applicable to Marion Island or a site where a similar sequence of events occurred. It is a particularly useful technique for determining the first till of each glacial for, in the apparent absence of any volcanic activity during the glacial, the initial ice advance reworked the interglacial pyroclastic material to give a pyroclast-rich till. Thus, the till which is rich in pyroclastic material acts as a datum for that glacial.

Clast size was not found to be a useful criterion for considering till genesis on Marion Island, although it is suitable as a comparative means of distinguishing between successive till units.

The method of analysis described above, although time-consuming, has been shown to provide objective criteria by which tills on Marion Island may be differentiated, and also permits an assessment of their mode of formation.

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RAPID DEGLACIATION AS AN INITIATOR OF VOLCANIC ACTIVITY: AN HYPOTHESIS

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ABSTRACT

Volcanic activity on sub-Antarctic Marion Island is found to have occurred only during the interglacials. The present volcano distribution is associated with a radial and peripheral fault system, the location of which appears to be related to the former glacier distribution. An hypothesis is presented suggesting that the faulting is a result of deglaciation and that the specific location of the faults is due to the differential stresses occurring between ice-covered and ice-free areas during isostatic uplift. The faulting initiates volcanism due to the location of the island within a volcanic region.

KEY WORDS Deglaciation Isostatic rebound Faulting Volcanic activity Sub-Antarctic

INTRODUCTION

Marion Island (Lat. 46°54'S, Long. 37°45'E) is a volcanic island of 290 km^2 (Verwoerd (1971)) located in the southern Indian Ocean (Figure 1). The origin of the island is related to the Mid-ocean Ridge system, which is located approximately 100 km to the east. Whilst none of the volcanoes are presently active, the volcanic history has been one of basaltic effusions alternating with explosive eruptions from the numerous centres (Verwoerd (1971)). Two series of basaltic lavas, with associated pyroclasts, can be recognized: an older, grey lava with a glaciated surface and a younger, black lava showing no signs of glaciation. The oldest age determination so far obtained for the grey lavas is 276 000 ± 30 000 BP (McDougall (1971)) whilst the black lavas are estimated to have erupted between ±11 000 and 4 000 BP (Hall (1978); Schalke and van Zinderen Bakker (1971)). Prior to the black lava volcanic stage the island was glaciated on three occasions (Hall (1978, 1979)) at times which appear to coincide with glacial cycles D, C, and B of Kukla (1977).

THE INFORMATION

Analysis of the island's fault pattern and recent black lava flows indicate that both postdate the last glaciation for although the faults cut across glacial-depositional landforms the unglaciated black lavas show no sign of disturbance. Many of the eruptive centres are aligned along the prominent faults (Figure 1). If, then, the volcanics postdate (or are even approximately synchronous with) the faulting, this would imply that the magma may well have ascended along the weak zones of the crust caused by the faulting.

Two major sets of faults are present (Figure 1): a dominant radial system and a secondary peripheral group. The radial system breaks the island into a series of horsts and grabens. The black lavas are found, in most instances, within the grabens whilst the horsts consist of till-covered or striated grey lava (Figure 1). The reconstruction of the former ice cover (Hall (1978, 1979)), by means of stratigraphy, till fabric analysis, striation azimuth and landforms, shows that the radial faults are roughly located parallel to the former glacier margins.

The secondary pattern of faults produces a rising, step-like sequence inland. It is most noticeable, with both fault groups, that the amount of displacement increases away from the coast.

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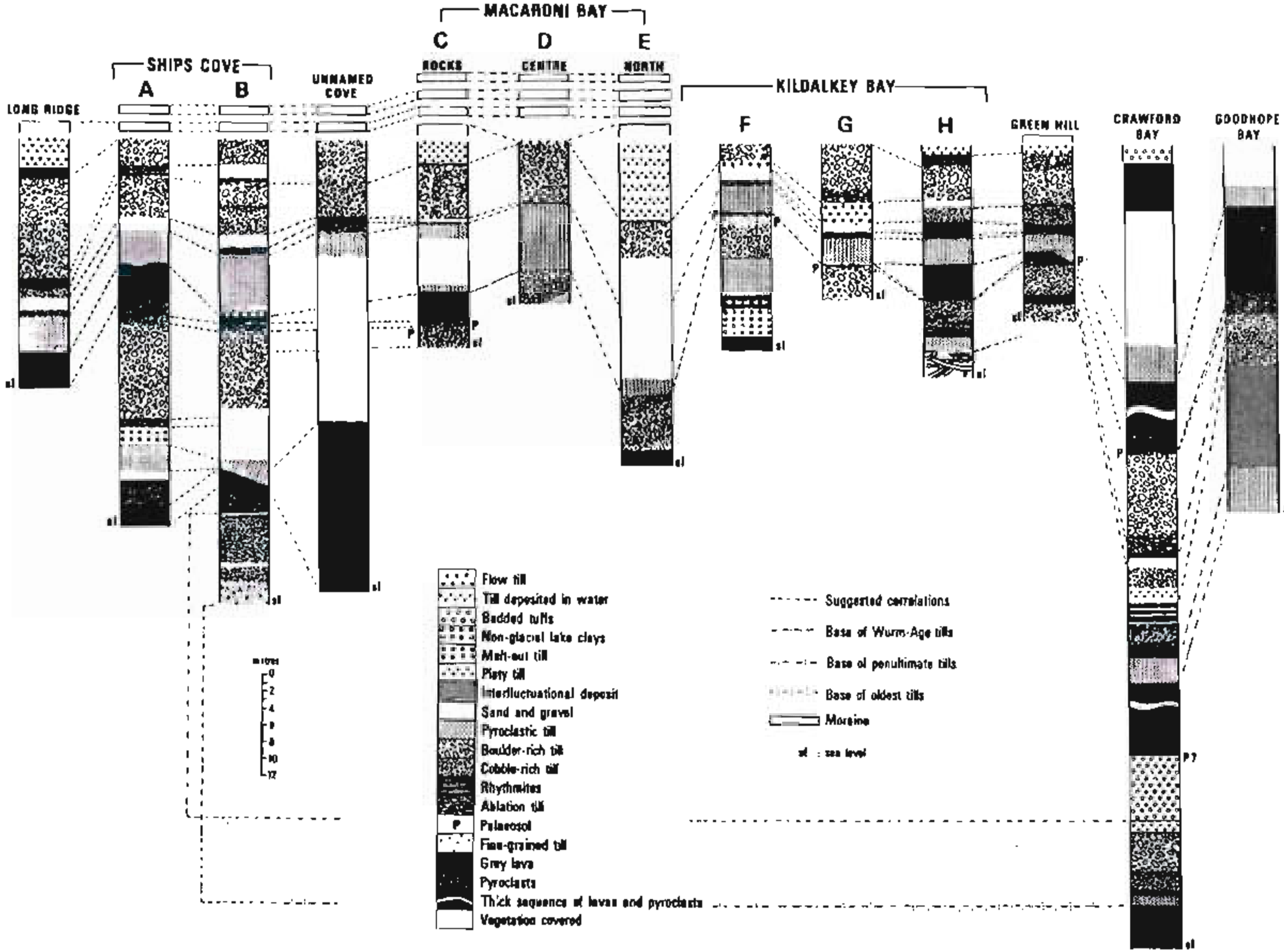

Figure 2. Simplified stratigraphic columns for the major exposures of glacial deposits on Marion Island

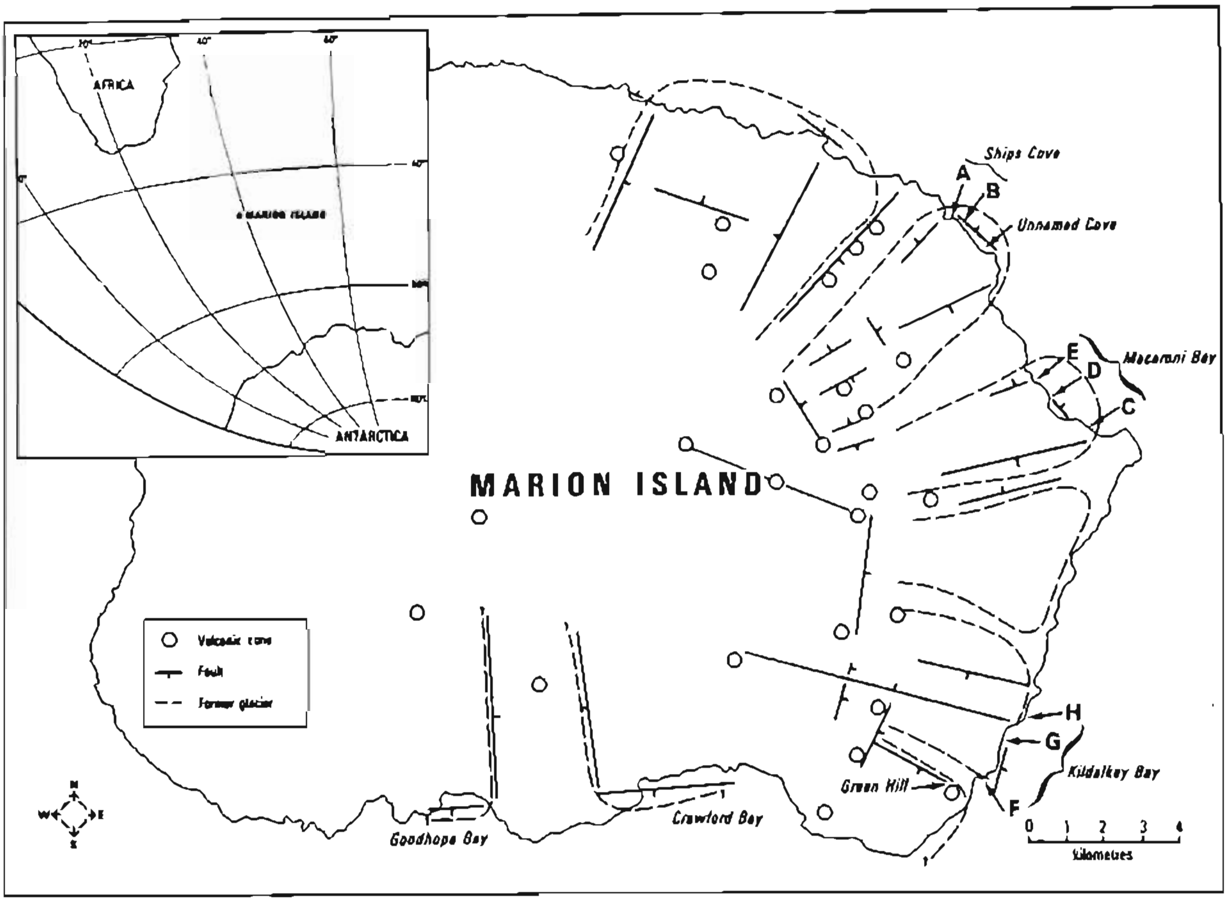

Figure 3. The approximate position, at maximum extent, of the reconstructed glaciers from the last glacial episode superimposed upon the known fault system

why this distinctive faulting should occur in response to isostatic readjustment instead of the more normal uplift of the whole island.

Marion Island is currently about 2° of latitude north of the Antarctic Polar Front and during glacial times the Front was displaced only a short distance to the north in this area, barely encompassing the island (Hall (1978)). Thus, the mean annual temperature of Marion Island fell by only 2–4°C (Hall (1978, 1979)) so that the glaciation of the island was effected by an increase in the amount of precipitation falling as snow. There was extensive ablation at the lower altitudes but annual snowfall was sufficient to maintain the glaciers whilst precipitation continued to occur in the form of snow rather than rain (Hall (1978, pp. 273–279, 1979)). Thus, as the climate moved from glacial to interglacial, the climatically marginal position of Marion Island caused a shift from snow-predominant back to rain-predominant precipitation. The snowline rose rapidly above the top of the island so cutting the glaciers off from their source areas and, at the same time, subjecting them to higher air temperatures, relatively warm rainfall and a lower incidence of cold southerly winds (Hall (1978)). It appears that the ice melted away, perhaps partly *in situ*, at a very fast rate.

With the change in climate towards interglacial conditions there was rapid deglaciation such that the period of 'restrained rebound' (Andrews (1970)) had only just begun when it was superseded by 'post-glacial uplift' (see Figure 55 of Embleton and King (1975)). Thus, superimposed on the isostatic recovery of the island as a whole was a local effect due to the very rapid loss of ice overburden. Mörner (1978) has suggested that deglaciation in itself can initiate tectonism and faulting. This idea is taken one step further and it is postulated, on a limited amount of evidence, that the faulting occurred along a zone

CONCLUSIONS

This hypothesis, essentially an extension of that of Mörner (1978), appears to offer an explanation for the information that is currently available. It is hoped that future fieldwork will enable more data to be collected and so, perhaps substantiate the hypothesis. As many of the other sub-Antarctic islands are of a volcanic nature (and some, e.g. Kerguelen, still have active volcanoes) and have evidence of Quaternary glaciation, it might be that this mechanism has been operative in those locations as well. Certainly, it would be worthwhile to consider the stratigraphic and tectonic information available in the light of this hypothesis.

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This work constitutes part of a study on the Quaternary history of Marion Island initiated by Professor van Zinderen Bakker. Professor J. N. J. Visser kindly read and commented on the first draft of this paper and discussed the evidence in the field. Several useful references and interesting discussion were provided by Dr. V. von Brunn. An anonymous reviewer gave many useful comments on the structure and approach of the paper which were gratefully received. Diagrams were redrawn by Mr. B. Martin.

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A RECONSTRUCTION OF THE QUATERNARY ICE COVER ON MARION ISLAND

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Abstract Evidence in favour of extensive glaciation on sub-Antarctic Marion Island is presented. Available data indicate three phases of glaciation, each comprising a sequence of stades and interstades. The island stratigraphy is shown and the major glacial deposits briefly described. From the available evidence, particularly the moraines, a number of former glaciers are reconstructed. Temperature fell between 3° and 6.4°C at the glacial maximum, with an equilibrium line altitude depression of 650 m and an accumulation area ratio of 0.6. The interglacials are marked by tectonism and volcanism as a response to isostatic readjustment. Temperatures were similar to present with extensive pedogenesis and plant colonisation. High glacier velocities are indicated to compensate for large ablation rates. Glaciation was initiated by precipitation falling as snow due to the northward shift of the Antarctic Convergence.

Marion Island (46°54′S, 37°45′E) is a 290 km² volcanic complex, rising to a height of 1230 m, situated approximately 2° of latitude north of the Antarctic Convergence (Figure 1). Whilst the present day cover of permanent snow and ice is restricted to a very small area (< 3 km²) above 900 m, there is evidence of several extensive glaciations (Figure 1) in the recent past. Geological, geomorphological and palynological information indicate three periods of glaciation, each comprising a series of stades and interstades. Intervening interglacials were characterised by extensive tectonism, volcanism, pedogenesis and plant colonisation. Whilst a number of specific topics have been dealt with in earlier publications (Hall, 1978a, 1978b, 1979) the aim here is to present some new data and give an overview of the past ca 300,000 years.

STRATIGRAPHY AND LANDFORMS

At present the island surface exhibits numerous scoria cones, lava flows and spreads of pyroclasts (Verwoerd, 1971). If ice were to advance over the present day surface then initial deposits would certainly be characterised by a high percentage of pyroclasts mixed with fragments of lava bedrock. Observation of the extensive sections along wave-cut cliffs (Figure 2) shows diamicts 1 to 6 m thick, comprising a mixture of reddish pyroclasts and clasts derived from the underlying grey basalts. Going upsection through these diamicts, the percentage of pyroclasts decreases exponentially as that of the basalt clasts increases (Figure 3). Detailed investigation shows some striated basaltic clasts, and a clast a-axis preferred orientation parallel to striations on underlying surfaces. This, taken together with the upward transition into certain glacial sediments, indicates that the diamicts are true tills. As this "pyroclastic till" usually occurs above a lava or palaeosol (Figure 2) it is interpreted as resulting from the first stadial advance of each glacial.

Pyroclast content does not increase in the till sequence beyond the initial pyroclastic till and this is thought to indicate that no further volcanic activity took place at this time. Further pyroclastic till is only found after a period of extensive interglacial volcanism initiated by tectonism caused by isostatic rebound (Hall, 1982). Interglacial age is also indicated by the presence of ca 2 m thick palaeosols and the development of peat. Palynological evidence (Scott and Hall, in press) indicates a climate similar to the present and that sea level fell towards the end of the peat development; suggestive of oceanic depletion as

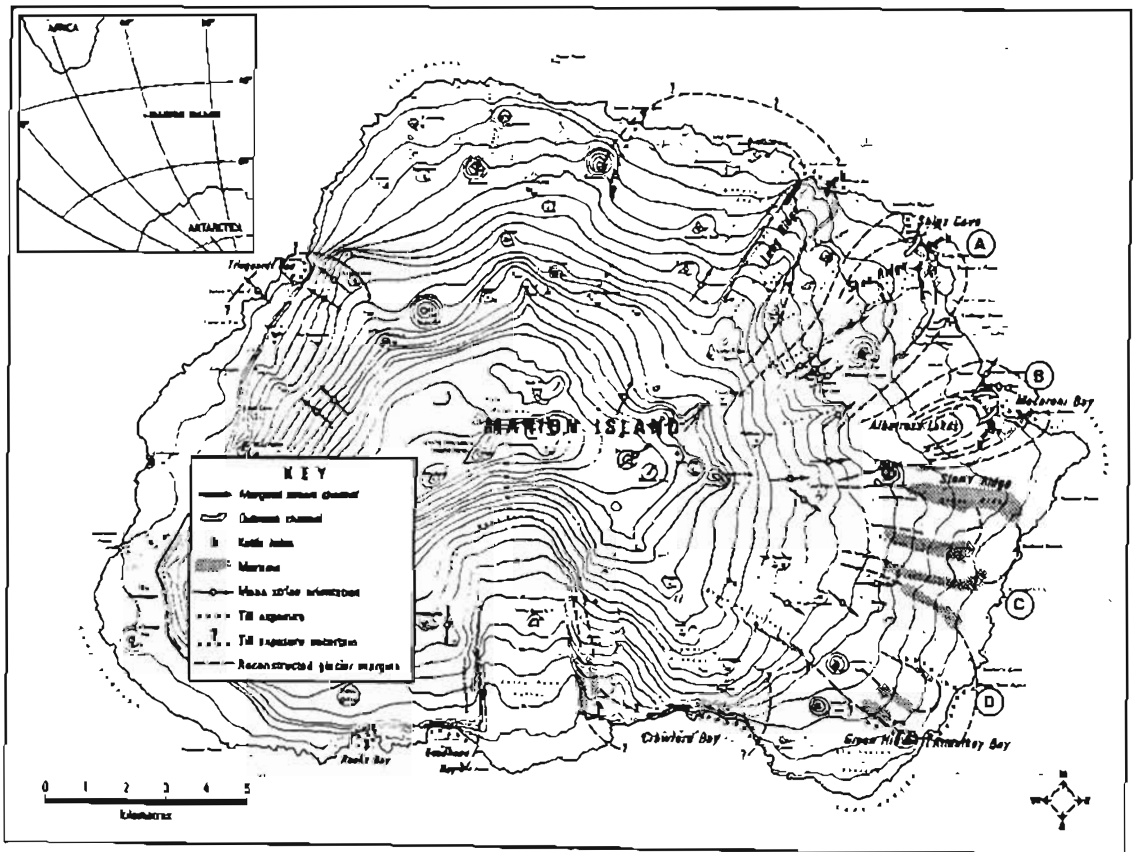

Figure 1. A reconstruction of some former glacier margins together with a summary of glacio-morphological information.

poorly developed fabrics and signs of water sorting and imbrication. Clast size increased with height and the wedges are capped with an ablation till and fluvial sediments. Finally, at Goodhope Bay (M in Figure 2) there is a 12 m sequence of alternating beds of water-sorted material, up to gravel size; and beds of non-sorted angular material, of gravel to small-cobble size, with several distinctive beds (0.3 m thick) of silt and clay. This is interpreted as an "interfluctuational" deposit (Miller, 1975) which is a depositional unit resulting from fluvioglacial and glacial processes operating in a subglacial environment near the glacier snout (Kirby, 1969).

Difficulties in differentiating glacially reworked volcanoclastics from in situ volcanic deposits has led to some interpretation problems on Marion Island (Verwoerd, 1971; Gribnitz, 1981). However, there is adequate corroborative information to substantiate a glacial origin for some of the Marion sediments.

Lateral and frontal moraines, of both "push" and "dump" origin (Andrews, 1975), were observed from the present coastline up to an altitude of 250 m. All recorded moraines relate to the most recent glacial event (Wurm-Weichselian of N. Hemisphere) but many are affected by postglacial faulting and lava flows which makes estimation of their size impossible. Frontal moraines are conspicuously absent in the Stony Ridge to Kildalkey Bay area (Figure 1) but this is in full accord with the argument that glaciers supported by high precipitation (as here—see below) have large transport values, but little accumulation of frontal moraines, due to continuous melt-water discharge. The survival of end moraine remnants at other localities can be explained by the presence of outflow channels cutting through the moraines (Figure 1). Pollen cores obtained between the moraines suggest interstadial conditions ca 17,000 BP and an end to the glacial ca 12,000 BP (Schalke and van Zinderen Bakker, 1971).

RECONSTRUCTIONS AND COMPARISONS

Upon the basis of the foregoing information it has been possible to generate an outline of major events during the past ca 300,000 years (Figure 4) and offer a number of palaeocondition reconstructions. The moraines (Figure 1), particularly the frontal, mark successive glacial termini during retreat from their maximum and so allow reconstruction of former glaciers. The sequence of lateral moraines from Stony Ridge to Green Hill (Figure 1) all begin at ca 250 m and thus this altitude can be equated to the approximate position of the former equilibrium line altitude (ELA). Østrem (*in* Andrews, 1975) argues that the climatic snowline would have been situated somewhere between the ELA and 300 m higher (i.e. from 250 to 550 m in this instance). On this basis it was possible to reconstruct temperatures at present day sea level, for glacial maximum conditions, utilising a range of mean annual temperatures for the maximum and minimum snowline altitudes. This indicated (Hall, 1979, Table 1) that there was a mean annual temperature decrease of 3° to 6.4°C and that temperatures in the former glacier ablation zones were positive for most (66%), if not the whole year. Thus, glaciation of Marion Island is assumed to result from precipitation falling as snow rather than, as at present, rain, with the high annual receipt allowing glacier maintenance despite large ablation rates. Mass transfer from accumulation to ablation zone would have necessitated high ice velocities and the continual ablation would have maintained extensive fluvioglacial outwash systems.

The suggested mean annual decrease in temperature is in good agreement with the 3 to 4°C depression postulated by van Zinderen Bakker (1973) from pollen spectra. It is also within the ranges suggested from various ocean core studies, viz. 2.5 to 3.5°C (Hays, et al., 1976), 3 to 4°C (Prell, et al., 1980), and from other islands, 5°C for Tasmania (Denton and Hughes, 1981) and 3 to 4°C for Macquarie Island (Colhoun and Goede, 1974). The ELA estimation of ca 250 m indicates a fall of 650 m which is similar to that (700 m) recently found for Kerguelen by the writer and within the range of successive depressions cited by Porter (1975) for New Zealand.

With the ELA estimation and the presence of lateral and terminal moraines it is possible to delimit and measure the ablation area of several former glaciers from the last glacial maximum. Four reasonably well-defined glaciers (A to D in Figure 1), which equate to roughly 20% of the estimated ice cover, have a combined ablation area of 22.79 km². Steady-state glaciers in equilibrium have accumulation area ratios (AAR) in the order of 0.6 to 0.7 (Andrews, 1975). Assigning a value of 0.4 to the known ablation zones a combined accumulation area of 34.19 km² is obtained and thus a total glacier area of 56.98 km². Extrapolation to full island cover indicates a total island ice area in the region of 285 km². The present day island is 290 km² but during glacial times would have been marginally larger due to lower sea levels. However, it is known from moraines and stratigraphy that some glaciers did not extend beyond the present coastline, and others did so only marginally, so an AAR higher than 0.6 would generate too extensive an ice cover.

Such an AAR indicates that the Marion glaciation was a product of high annual snow accumulation rates in conjunction with large ablation rates. The transfer of mass required to maintain the glaciers under these conditions would have necessitated high ice velocities. The steep clast dips ($\bar{x} = 23°$) obtained in till fabric analysis, are indicative of just such a flow regime (Andrews and Smith, 1970). The suggested AAR of 0.6 agrees with that found for New Zealand (0.6 ± 0.05) by Porter (1975).

CONCLUSION

This brief synopsis of events of the past 300,000 years on Marion Island is summarised in Figures 1, 2 and 4. The onset of glaciation is a result of a slight northward shift of the Antarctic Convergence. The consequent decrease in temperatures, caused precipitation to fall as snow. The high annual accumulation offset large ablation rates and so the glaciers had relatively high velocities to compensate. A strong positive relationship exists between precipitation, glacial erosion, glacier transport and proglacial runoff such that the Marion glaciers carried a great deal of debris to their margins and there was extensive fluvioglacial activity. Thus, sequences of tills, fluvioglacial deposits and moraines were built-up.

With an interglacial southward retreat of the Antarctic Convergence temperature rose causing precipitation to revert to rain, raising the ELA above the island summit, and causing rapid deglaciation. Accelerated ice-loss initiated tectonism which, due to location in an active area, caused volcanism. During the warm interglacials there was pedogenesis and plant growth. With reversion to glacial conditions sea level dropped, precipitation fell as snow, the ELA was lowered and glaciers advanced over the interglacial volcanic debris and a further sequence of stadial and interstadial sediments accumulated. The final retreat of ice cover left a sequence of moraines, remnants of which are found on the present island surface. Currently, interglacial tectonism, volcanism, pedogenesis and plant colonisation operate once more.

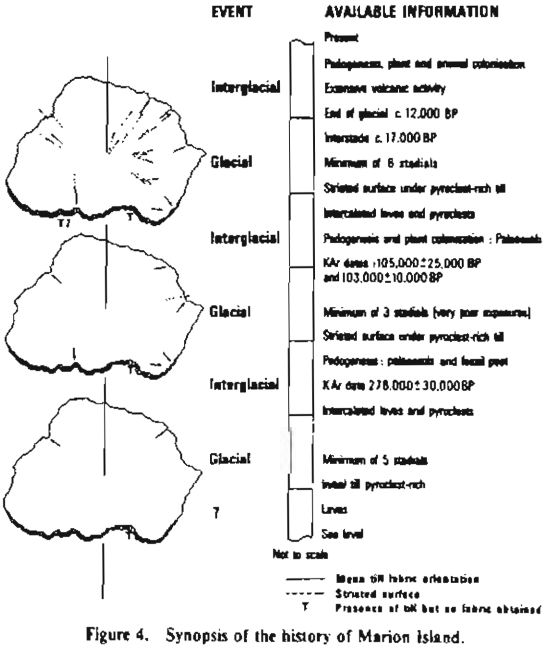

Figure 4. Synopsis of the history of Marion Island.

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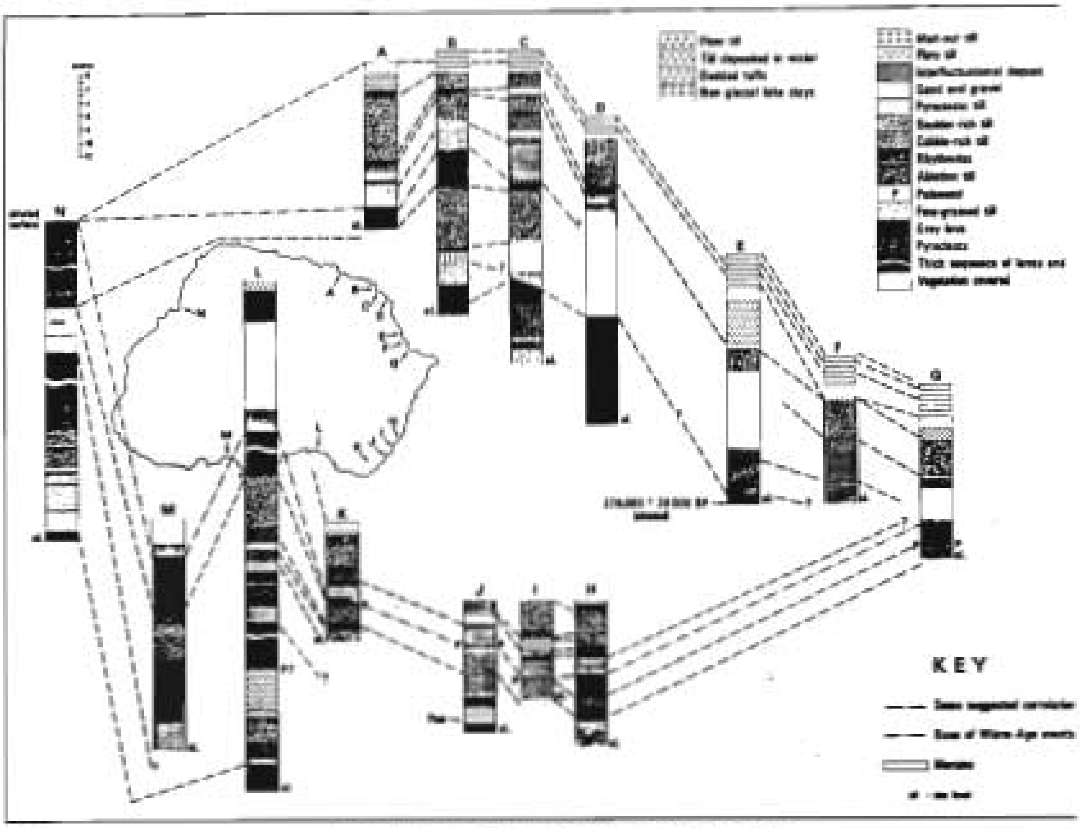

Figure 2. Simplified stratigraphic columns and their locations.

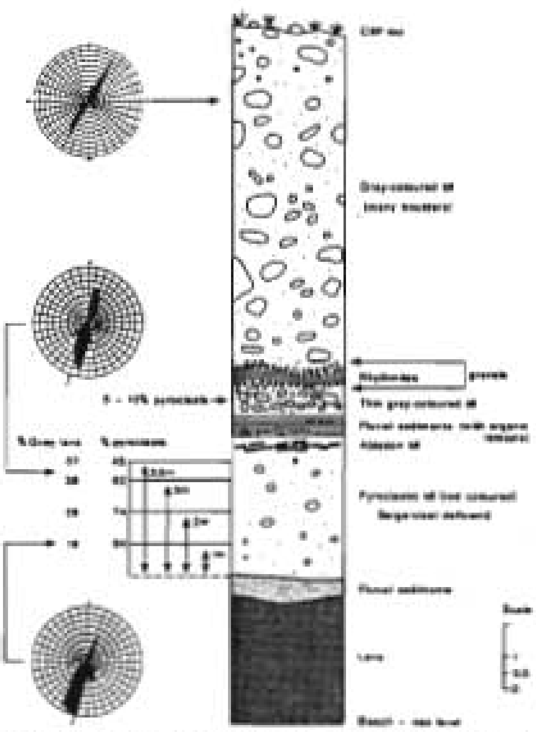

Figure 3. Detail of section at Long Ridge. Pyroclast and grey lava clast exponential variation with height (y = 9.22e[illegible]; r = +0.98) within the pyroclastic till is shown together with similarity of fabric orientation with the lodgement till above.

world ice cover grew as the glacial onset. The lavas, which sho of the features indicative of under-ice origin (e.g. pillow l hyaloclastites), have age determinations which place them within world interglacials. No date is available for the basal l the next in succession has been dated (McDougall, 1971) at 27 30,000 BP. The next lava sequence has dates of 103,000 ± 10, and 105,000 ± 25,000 BP. The present surface lavas are too yo date, but peats on top of the last till give determinations of 1 700 BP, 9500 ± 140 BP (Schalke and van Zinderen Bakker, 19 7120 ± 45 BP (Lindeboom, 1979). Thus the volcanism is interp being of interglacial age and the pyroclastic till to initiate each sequence.

Differentiation of the till sequences into stades and interst based on the presence of ablation tills in association with flu lacustrine sediments. These ablation tills consist of highly a platy clasts exhibiting no striations and lacking a preferential fabric. The overlying lacustrine deposits show rhythmites, ofte dropstones, whilst the fluvial sediments comprise cross-bedded and gravel. At one locality (Long Ridge; A in Figure 2) remains were found within the outwash sequence immediately the ablation till. This, together with synchroneity about the island, argues against these being subglacial deposits unrela interstadial conditions. Periglacial slope deposits are also disc due to the lack of internal sorting and absence of a-axis fabric the intimate association above and below with glacially d sediments.

Other than the ablation type, the majority of tills are lodg with the typical characteristics of such basal tills. In addition, till sequence was recognised at Kildalkey Bay (J and I in Fig consisting of a series of till sheets with rudimentary vertical so exhibiting both thinning and a decrease in clast size with di down flow. Between the sheets are layers of fluvially-sorted san gravel. Clasts are orientated parallel to flow, which is itself norm the main ice direction as shown by striations and fabrics i underlying lodgement till. A melt-out till occurs at Ships Cove Figure 2). This consisting of short "wedges" of subangular debris

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EVIDENCE IN FAVOUR OF AN EXTENSIVE ICE COVER ON SUB-ANTARCTIC KERGUELEN ISLAND DURING THE LAST GLACIAL

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ABSTRACT

Hall, K., 1984. Evidence in favour of an extensive ice cover on sub-Antarctic Kerguelen Island during the last glacial. Palaeogeogr., Palaeoclimatol., Palaeoecol., 47: 225—232.

Arguments to date have suggested that during the last glacial (Würm—Wisconsin—Weichselian) sub-Antarctic Kerguelen Island did not experience an extensive ice cover and that the fjords and glacial valleys are products of earlier events. Recent observations of striation orientations, travel directions of erratics, cirque altitudes, and evidence for isostatic uplift suggest that there in fact may have been extensive ice cover. The equilibrium line altitude (E.L.A.) reconstructed for the cirque glacier stage agrees well with that for sub-Antarctic Marion Island situated to the west. A possible explanation for the lack of glacial deposits and landforms over much of the island is suggested.

INTRODUCTION

Ile Kerguelen (Grande Terre) (Fig.1), the largest (6000 km^2) in the Kerguelen archipelago (48°27'—49°58'S, 68°25'—70°35'E), located just to the north of the Antarctic Convergence (Mercer, 1967) has a present-day ice cover of approximately 750 km^2 (Denton and Hughes, 1981). The Cook ice cap (500 km^2), with 40 outlet glaciers, is the largest single unit whilst the Presqu'île de la Société de Géographie, Peninsule Rallier du Baty, Mont Ross and Pic Guynemer all have a number of glaciers (Bauer, 1963) (Fig.1). Other than glacial deposits associated with the 1750 maximum (Nougier, 1966) and fluvioglacial debris in many valley bottoms, glacial deposits and landforms are said to be almost exclusively restricted to Peninsule Courbet (Bellair, 1965).

Studies of the former glacier cover (Bellair, 1965, 1969; Nougier, 1966, 1970, 1972) suggest that during the last glacial (Würm—Wisconsin—Weichselian) the ice was of very limited extent. Bellair (1969, p. 168) states that it was ". . . impossible that there was a total ice cover in the Würm". This statement is based upon (p. 168) the survival of indigenous flora, the absence of moraines and the lack of any cirques of importance. The many open-ended fjords evident on Kerguelen are said (Bellair, 1969, p. 167) to be the result of an "ancient glaciation". This view is also expressed by Bellair

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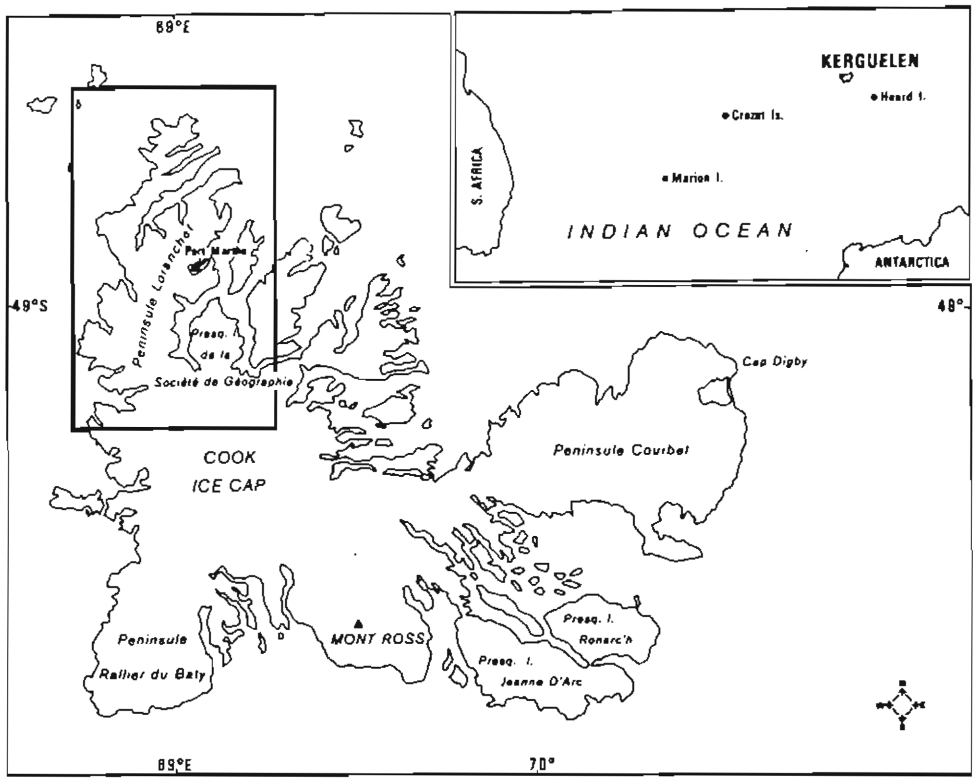

Fig.1. Location of Kerguelen Island together with the major areas mentioned in the text. The boxed area is that shown in detail in Fig.2.

(1965), Nougier (1966, 1972) and, Denton and Hughes (1981). According to Young and Schofield (1973), from palynological work in the southeast of the island, retreat of the ice that did exist in this area, began about 10,000 B.P.

Observations made during 1981, on the Peninsule Courbet and Peninsule Loranchet (Fig.1), suggest that the island may in fact have been almost completely covered by ice during the Würm–Wisconsin–Weichselian. Data, here presented, although of a preliminary nature do appear to offer a reconstruction comparable to that obtained for sub-Antarctic Marion Island situated to the west of Kerguelen (Hall, 1979).

OBSERVATIONS

Striations were found at various localities in Peninsule Loranchet (Fig. 2) but they can only be used as axes of movement due to the lack of additional information to transform them into vectors. Localised differences in weathering result in the striations varying in appearance from "very fresh looking" to "visible, but weathered". However, it is considered that with the extensive weathering observed on Kerguelen all of the striations relate to the last

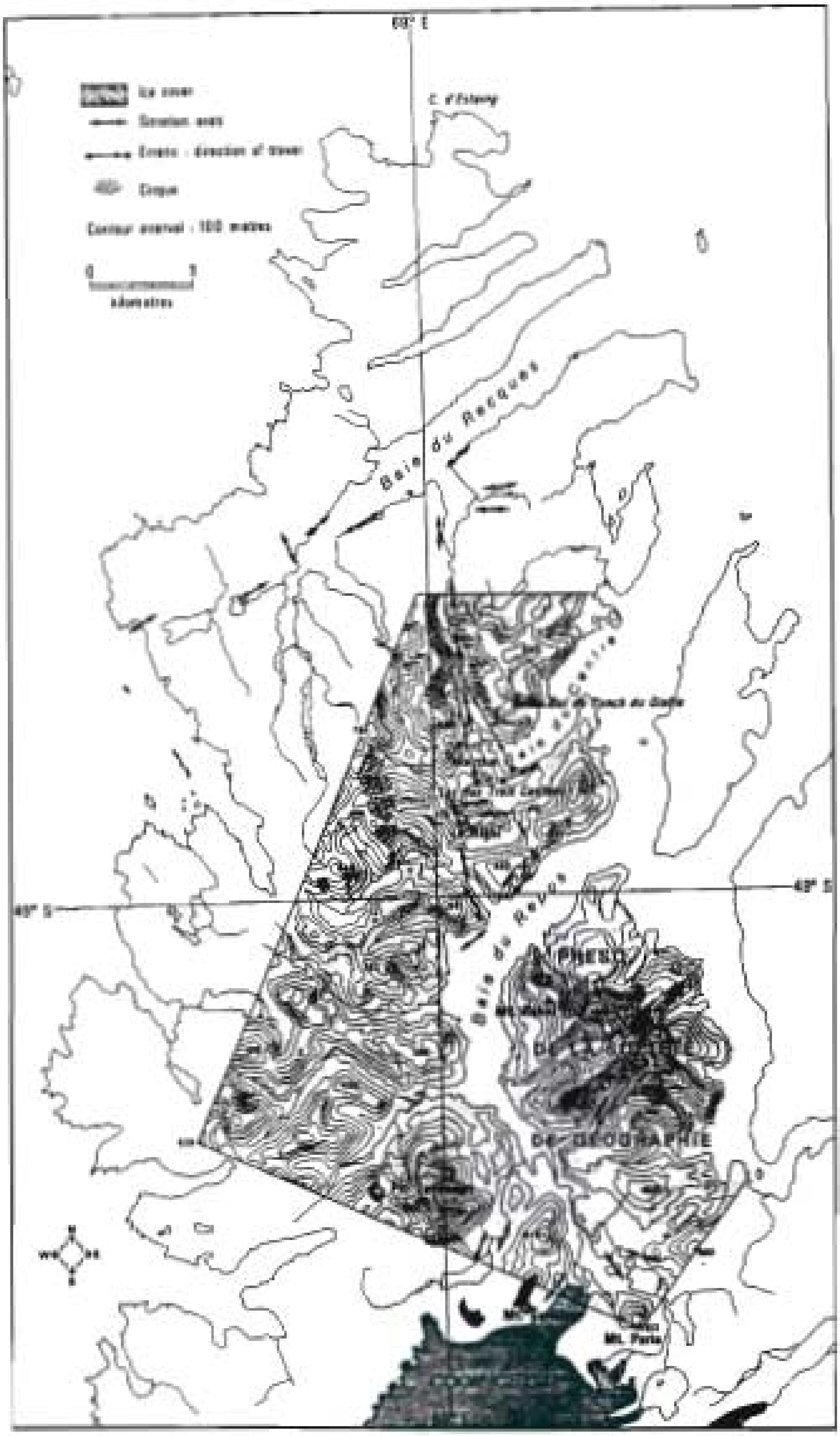

Fig.2. Detail of study area indicating striation observations, cirques and direction of travel of erratics.

glacial. The striations observed on the wet, cooler mountain tops were usually, but not exclusively, of the "visible, but weathered" type. Striations found in the valleys were aligned with the main valley axis but those on the mountain tops varied between parallel and normal to the associated valleys (Fig.2).

Erratics were found on striated surfaces both in the valleys and on the mountain tops. The erratics were mostly "fresh" and did not exhibit a severe degree of weathering; certainly no more than the surrounding bedrock outcrops. Unfortunately, the extensive areas of plateau basalts precluded field reconstruction of point of origin. However, in three cases of non-basaltic erratics these could be traced back to their unique outcrops (Fig.2). This showed that two boulders moved southwards from Vallée du Thermomètre to Lac du Trois Cantons and Port Martha; the other moved eastwards, parallel to the Port Martha coastline (Fig.2). That erratic which came to reside on Lac du Trois Cantons is situated to the lee side (with respect to Vallée du Thermomètre) of a 200 m ridge whilst that which paralleled the Port Martha coast moved in a direction normal to the other two.

An extensive assemblage of cirques, many with associated moraines and/or striated surfaces, with a range of sizes and aspects were observed on Peninsule Loranchet (Table I). Cirque stairways were seen (e.g. on Mt. Rabot) and some cirques exhibited lakes in their overdeepened basins (e.g.

TABLE I

Orientation and heights of some cirques observed on Peninsule Loranchet

Orientation	Height of floor (m)	Height of backwall top (m)
NE	100	300
NW	250	550
NW	250	700
W	200	750
E	200	450
NE	400	650
E	200	500
NE	200	450
E	250	550
SE	150	450
E	200	450
NW	150	400
NE	600	750
Distribution of aspects		
N = 0	S = 0	
NE = 4	SW = 0	
E = 4	W = 1	
SE = 1	NW = 3	$n = 13$

the upper cirque on Mt. Rabot). These were all true glacial cirques, as expressed by their form and the glacial features associated with them and not structural hollows such as Ravin Bol de Punch du Diable, on Baie du Centre (Carte de Reconnaissance, 1968). Of the thirteen cirques noted in Table I, ten are within 50 m of a 200 m a.s.l. mean cirque floor altitude. There is evidence for isostatic rebound (see below) but for reconstruction of the cirque glacier stage the present levels would be appropriate. Major isostatic uplift is suggested by rivers deeply (10 m+) incised in the upper sections of their courses (e.g. Vallée d'Alouette) and by wave-cut cliffs backing wave-cut platforms (e.g. along southern part of Baie de Repos) up to nearly 50 m a.s.l. Minor isostatic rebound is suggested by the incision of rivers into floodplains, valley trains (the valleys have extensive spreads of fluvioglacial deposits) or deltas in their lower sections. In fact, the deltas built out into the fjords at the end of many glaciated valleys may themselves reflect uplift.

The present-day equilibrium line altitude (E.L.A.) is said by Mercer (1967) to vary between 600 and 900 m whilst Watson (1975) locates the snowline between 900 and 1000 m. However, in 1902 and 1929 the snowline was between 650 and 700 m (Watson, 1975, p. 300), some 200 to 350 m lower than at present. In addition, Nougier (1966) describes moraines relating to the "Little Ice Age" which are located several kilometres from the present-day glacier margins. As shall be seen below this information, together with the altitudes of the cirques, is considered of major importance in justifying extensive ice cover during the Würm—Wisconsin—Weichselian.

Finally, to the east of the island, the seaward sector of Peninsule Courbet comprises glacial and fluvioglacial deposits and landforms (Bellair, 1965). Investigation of a sea cliff section near Cap Digby indicates a sequence of a possible beach deposit overlain by till, then a thin (0.4 m) layer of fluvioglacial material upon which occurs the present-day peat and vegetation; it is hoped to obtain a date for the base of this peat (Hall and Scott, in prep.).

DISCUSSION

It is suggested that the above information is indicative of an extensive ice cover on Kerguelen Island during the Southern Hemisphere equivalent to the Würm—Wisconsin—Weichselian. The existence of striations and erratics on the tops of mountains in Peninsule Loranchet points to an ice cap state in which the whole area was under ice. The striations are thought to relate to the last glacial insofar as weathering during the last interglacial and the last glacial, a period of some 120,000 years, would have removed all such traces of the penultimate glacial from exposed mountain tops. In addition, the other evidence in favour of extensive ice cover given below tends to substantiate the probability of striation formation during the last glacial.

The finding of striations on, for instance, Le Righi (435 m) which are normal to those in the surrounding valley bottoms (Fig.2) suggests that the flow direction changed through time. It is hypothesised that with the onset of glacial conditions glaciers grew in the many cirques and, at the same time, any surviving ice caps, such as the Cook, also expanded in size. This "alpine" condition persisted until the valleys were ice-filled and small ice caps had formed in the higher regions: a state much the same as the present Juneau Icefield of southwestern Alaska. With continuation through time, there was a breaching between valleys (glacial transfluence) as evidenced by the many high level cols, and coalescing of some ice caps. This eventually led to the area being completely ice covered and, as a consequence, ice flow was no longer constrained by the valleys and could be normal to them. With degeneration at the end of the glacial, conditions reverted to the alpine state and finally to that observed today.

This hypothesis would explain the striations observed and the movement, normal to each other, of the erratics found in the Lac des Trois Cantons area (Fig.2). It could also explain the positioning of the erratic in the lee of a 200 m high ridge (Fig.2).

That glaciers and a major ice cap, with an E.L.A. varying between 600 and 900 m (Mercer) and a snowline around 900 to 1000 m (Watson, 1975), can exist under present-day interglacial conditions, plus the presence of large moraines relating to the Little Ice Age several kilometres from the present ice margins (Nougier, 1966), suggest that it would not be surprising to expect a major ice cover under full glacial conditions. This suggestion is strengthened by the occurrence of an almost complete ice cover on Marion Island which was smaller, lower, further from the Antarctic Polar Front and further to the north during the last glacial (Hall, 1979) whereas today it only has a very small area (<3 km^2) of permanent snow and ice above 900 m. Additional evidence in favour of extensive ice on Kerguelen is the altitude of the cirques in Loranchet (Table I). It can be seen that, with three exceptions, all of the cirques cited are within 50 m of a ca. 200 m cirque floor altitude. Andrews (1975, p. 53) indicates that cirque floor elevations approximate to the former glacier E.L.A. when small glaciers filled the basins. Hence the E.L.A. of the "alpine stage" (c. 200 m) would have been between 400 and 700 m lower than at present. Glacier growth would have been most marked, as could be envisaged with the present day ice cover if the E.L.A. dropped to ca. 200 m a.s.l. Assuming the former E.L.A., as indicated by cirque floors, to be about 200 m, this then is very close to the 250 m E.L.A. reconstruction suggested for Marion Island (Hall, 1979, p. 255) as derived from moraines. Whilst there could very well be an error of as much as 100 m due to interpretation, mapping inconsistencies or isostatic uplift, this would nevertheless not detract from the basic arguments in favour of an extensive ice cover.

The lack of moraines (Bellair, 1969, p. 167) is explained by few, if any, being deposited in the higher areas, although the author has observed some of unknown age at an altitude of 350 to 400 m in the area close to Presqu'île de la Société de Géographie. Many of those which may have been deposited in valleys were probably either drowned in the fjords or eroded by the valley-concentrated meltwater activity as evidenced by the extensive fluvioglacial deposits found in the valleys.

The existence of glacial deposits in the east (Peninsule Courbet), on land which is sloping gently seaward, would suggest that this was close to one margin of the ice cap. There is additional, indirect, evidence for glacial deposits, in the form of rounded boulders, in the southeast of the island (Presq. I. Ronarc'h and Presq. I. Jeanne D'Arc; Fig.1), found by Young and Schofield (1973) below 3.1 m of organic-rich silts. Hence, other glacial deposits, and possible moraines, might be expected below present sea level on what would have been land areas during glacial times.

The dominant precipitation-bearing winds are westerly, hence the major present-day ice masses are towards the west, and so build-up of ice would have initiated in the west and spread eastwards. An asymmetric ice cover centred towards the west might be envisaged such that Peninsule Corbet was close to its extreme eastern margin.

Vegetation survival is not seen as a major stumbling block to the acceptance of extensive ice cover. As with Marion Island (Hall, 1979), vegetation could survive on intra- or extra-glacial areas which are now below sea level. With ice withdrawal and sea level rise, the plants migrated to the present land area. Qualitatively it appears that vegetation cover is both more widespread and thicker towards the east as might be expected if colonisation was from that direction as the ice retreated westwards.

CONCLUSIONS

There is evidence to suggest that, contrary to earlier reconstructions, the ice cover of Kerguelen Island during the last glacial was extensive and probably covered the whole island. A sequence of alpine → ice field → ice cap → ice field → alpine → present-day conditions is envisaged. Analysis of cirques floor altitudes indicates a possible glacial E.L.A. of ca. 200 m, some 400 to 700 m lower than present. The existence of cirque glacier E.L.A.s at such a low altitude would suggest extensive glaciation and the possibility of growth from a valley glacier stage to an ice cap condition. The presence of striations normal to valley axes and the occurrence of erratics showing travel paths across ridges are probable results of this stage. More detailed fieldwork on tracing erratics, finding evidence of isostatic uplift, measuring cirque floor altitudes, searching for moraines and, perhaps most important, finding dateable events would be most useful. Offshore coring would be a most valuable tool in reconstructing former ice extent.

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SOME QUANTITATIVE OBSERVATIONS OF CLAST SHAPES FROM THE DWYKA TILLITE OF NATAL, SOUTH AFRICA

by

K. HALL

ABSTRACT

Previous studies of clast shapes from the Dwyka tillite have been highly qualitative. Despite the imprecision and subjectivity of visual estimation, a number of size–shape relationships have been postulated which are now becoming entrenched in the literature via repetition. This pilot study was undertaken to attempt to quantify clast shape in terms of roundness, flatness, sphericity, elongation and oblate-prolate index. Additional observations on faceting and striations were also obtained. Clasts measured were within the size range 0,015 m to 0,387 m. The percentage of striated clasts from the 11 sample points varied between 14,29 and 29,55 whilst the mean number of facets were within the range 1,52 to 3,07. Mean shape indices were relatively uniform but standard deviations often varied substantially between sites. The results show that the hitherto qualitative size–shape relationships suggested for the whole of the Dwyka could not be substantiated. Those relationships which were found to exist are difficult to explain by an approach of this type and are, in themselves, worthy of detailed study.

CONTENTS

I. INTRODUCTION

Studies of the Permo-Carboniferous Dwyka Formation glaciogenic deposits of southern Africa have been, with only a few exceptions (e.g. Visser, 1982), largely uninfluenced by recent developments in glacial sedimentology and glacier dynamics. Clast shapes have been dealt with in a highly qualitative form although recent studies (cf. Boulton, 1978; Domack *et al.*, 1980; Humlum, 1981; Sharp, 1982) have shown that they may be indicators of glacier transport mechanisms. Despite visual estimation of clast shapes being highly imprecise and really "... should be considered only as a superficial field statement." (Orford, 1981, p. 88) no quantitative study has yet been undertaken. A review of recent studies on the glaciogenic deposits of southern Africa (Von Brunn, 1980; Bond, 1981; Martin, 1981a, b; Visser *et al.*, 1978) shows that clast shapes are presented in highly qualitative terms. Stratten (1968), cited in Von Brunn and Stratten (1981), attempts a percentage breakdown of clast shapes but the placing of clasts into shape categories appears to be based solely on qualitative judgement. In addition, Stratten (1968, p. 28) states that "The smaller pebbles are generally sharply angular whereas larger ones are usually rounded to varying degrees". This idea is reiterated in Tankard *et al.* (1982, p. 366) where it is stated that "... small-pebble clasts tend to be subangular, but the degree of rounding improves with increasing size, so that large pebbles and boulders are commonly subrounded to rounded with moderate to high sphericity".

The subjective nature of the studies is, in part, due to the problems of working with such highly lithified sediments. However, the potential of clast shapes as a tool in aiding interpretation of sediments cannot be ignored. In addition, the observations of Lister (1981), for both cold and warm valley glaciers, that there is an increase in rounding with a decrease in clast size, and of Drake (1972) who showed that clast shapes, ultimately were independent of clast size, appear to contradict the apparent findings in the Dwyka. In fact Martin (1981b, p. 69) summed up the present situation by noting that "... statistical methods had not yet been applied".

Upon this basis, a pilot project was undertaken in southern Natal to attempt to quantify clast shape and other factors. Throughout this study the warnings and comments of Barrett (1980), pertaining to the apparent ease with which measurements can be obtained being in contrast to ascribing a meaning to them, were kept in mind. However, the basic premise is to quantify clast shape apropos the qualitative descriptions used in preceding studies.

II. STUDY AREA

Two outcrops of Dwyka tillite, said to result from post-depositional erosion following block-faulting and monoclinal crustal flexture, roughly parallel the Natal coast (Von Brunn and Stratten, 1981, p. 72). Study was undertaken at four exposures of the coastal outcrop and one of the inland outcrop in the Mkomazi Valley (Fig. 1).

III. APPROACH

Although the tillite is highly lithified it is possible to remove clasts by means of a geological hammer and chisels. At each site a random sample of clasts was collected with no maximum size constraint but, for practical measuring purposes, a lower size limit of 0,01 m a-axis was selected. Clasts that showed signs of fracturing were discarded. The presence of striations and the number of facets (if any) could then be recorded. The longest (a), intermediate (b) and shortest (c) orthogonal

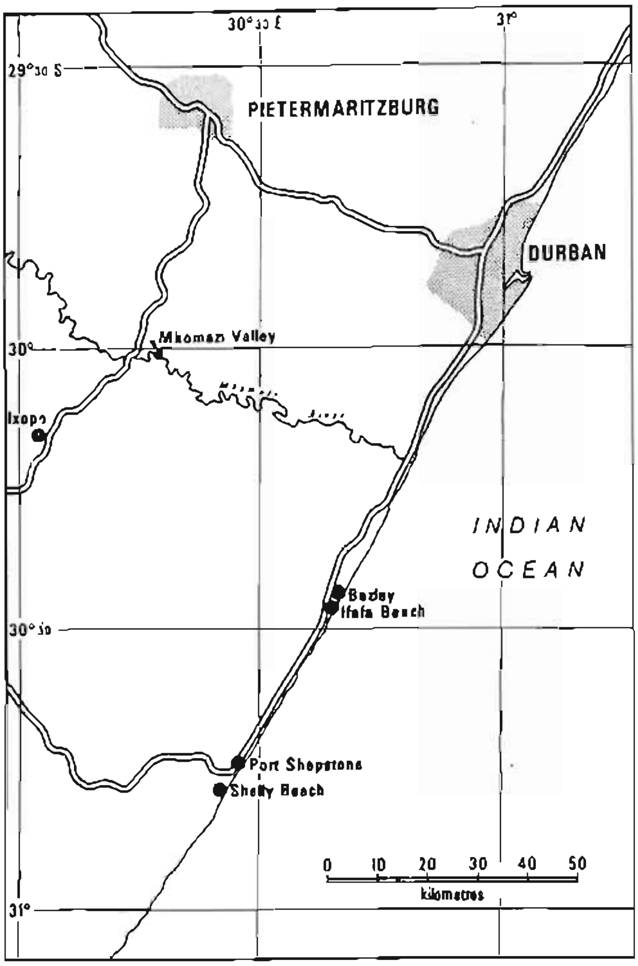

Figure 1
Locality map showing the four coastal and one inland outcrop sites of Dwyka tillite studied in this paper.

axes were measured by means of calipers, and the radius of the circle fitting the sharpest corner (R_1) and the radius of the largest inscribed circle (r_i) were obtained by means of comparison against a nest of concentric circles (see Dobkins and Folk, 1970). From the data (a, b, c, R_1 and r_i) the following shapes were calculated upon the recommendations of Barrett (1980):

1. Cailleux's (1945) flatness index (where a value of 100 = a perfectly equidimensional particle through to ∞ as flatness increases).
2. Modified Wentworth roundness (Dobkins and Folk, 1970) (values from 0 = highly angular to 1 = perfectly round).
3. Sneed and Folk's (1958) maximum projection sphericity (values from 0 to 1 where 1 = a perfect sphere).
4. Elongation index (Lister, 1981) (values range from 0 to 1 with an increase in elongation shown as values approach 0), and
5. the $\bar{O}$ $\bar{P}$ index (the oblate-prolate index of Dobkins and Folk, 1970) (values theoretically run between $-\infty$ and $+\infty$ where negative values indicate disc shapes, positive values indicate rod shapes and 0,00 is a perfect blade).

(For a general discussion on the derivation and use of shape indices see Briggs, 1977 and Goudie, 1981).

Thus, a quantified measure of flatness, roundness, sphericity, elongation and oblate-prolate attributes were obtained for each clast.

Having obtained data on individual clasts then the mean, standard deviation, skewness, kurtosis, maximum and minimum values for each sample were obtained. In the case of the $\bar{O}$ $\bar{P}$ index the percentages in each clast were found, as were the previous measures cited. In addition, the percentage of striated clasts and the mean number of facets for each sample were determined. These data quantify the clast shapes, thereby satisfying the "statistical" need noted by Martin (1981b) and, in so doing, suggest a distinctly different interpretation to that given by the qualitative approach.

In all of the southern African studies noted in the Introduction, no cognisance is made of lithological control on shape. Inherently this is a failing (Visser and Hall, in press; Hall and Visser, in prep.) but in order, as a first step, to facilitate comparisons between previous studies and this pilot programme the role of lithology was *deliberately* ignored in deriving shape criteria for the various study sites.

IV. OBSERVATIONS

Details of the information derived from sample points at the five localities are given in Table I. Sizes of clasts ranged between the extremes of 0,015 m and 0,387 m. Clasts indicate a low flatness index, with highly variable standard deviations (21,2 to 101,68), a slightly, usually positive, skewed distribution, and with a positive kurtosis. Roundness is low ($\bar{x}$ = 0,22) although the original data indicate some highly rounded material (e.g. a value of 1,0 found at site 1, Mkomazi Valley) with 14 per cent of the clasts having values of 0,4 or greater. This roundness index shows the positive skewness and highly variable kurtosis said to be characteristic of this measure (Barrett, 1980). Sphericity is fairly uniform between the samples ($\bar{x}$ = 0,76) with, as was also found by Barrett (1980), distributions very close to normal with a slightly negative kurtosis. Elongation shows no great variation and has an overall mean of 0,78. The $\bar{O}$ $\bar{P}$ index, on the other hand, shows variations from site to site in terms of percentage occurrence, means, skewness, kurtosis and maximum values observed within the oblate/prolate divisions. With respect to clast surface features, it was found that 22,9 per cent of the clasts are striated with a mean of 1,98 facets. Only 12,6 per cent of the clasts showed no signs of either striations or faceting.

There appears to be limited correlation (Table II) between size and shape. The sphericity–size correlations that were obtained are the reverse of what should have been expected according to Tankard *et al.* (1982). Only two a-axis/roundness correlations (Shelly Beach) were obtained and, conceptually, are difficult to explain by a study of this kind.

V. DISCUSSION

According to Barrett (1980, p. 293) "Shape parameters should be independent of size . . " and this is certainly the case in the glacial context where tillite may contain clasts of numerous lithologies which have undergone different travel distances via a variety of glacial, and perhaps non-glacial (e.g. meltwater) transport pathways (e.g. en-, supra-, sub-glacial and their combinations). This multiplicity of factors affecting clasts does not appear to have been considered in the bulk of Dwyka studies, for the very statement of a size–shape relationship should beg the question — what process(es) could cause such an effect? Even if a size–shape correlation did exist then, as Ehrlich *et al.* (1980, p. 480) found, it probably would not be a continuous function but more likely a step-wise progression. If this statement were correct, then, as the shape descriptions to date are not size constrained, they are of little meaning.

Conceptually it is difficult to envisage how or why, on such a large scale, the relatively simplistic size–shape relationship cited by Tankard *et al.* (1982) could come about. For example, as Drake (1972) found, the more

TABLE I
Summary of Clast Shape Data

	Bazley 1	Bazley 2	Ifafa 1	Ifafa 2	Port Shepstone	Shelly Beach 1	Shelly Beach 2	Mkomazi Valley 1	Mkomazi Valley 2	Mkomazi Valley 3	Mkomazi Valley 4
% Striated	21,15	27,78	18,72	29,55	28,00	18,18	26,67	17,24	14,29	24,24	26,32
x̄ Flatness	152,26	169,22	151,47	156,70	156,30	148,80	156,77	155,83	207,41	155,73	162,01
Stand. dev.	28,53	70,86	30,57	28,80	28,67	28,03	34,24	21,20	101,68	27,23	31,02
Skewness	0,75	3,65	1,12	0,67	1,10	0,67	1,82	−0,27	1,29	1,31	−0,51
Kurtosis	3,07	18,83	4,00	2,89	4,21	2,50	6,68	2,13	3,28	5,01	4,32
x̄ Roundness	0,25	0,23	0,18	0,17	0,20	0,24	0,16	0,33	0,22	0,19	0,24
Stand. dev.	0,19	0,22	0,17	0,17	0,17	0,16	0,10	0,29	0,23	0,15	0,20
Skewness	1,16	2,06	2,93	1,83	1,44	0,82	0,79	0,97	2,29	0,78	1,66
Kurtosis	3,55	6,79	12,93	6,19	4,49	3,04	2,67	2,89	7,57	3,80	5,31
x̄ Sphericity	0,78	0,75	0,77	0,74	0,76	0,79	0,76	0,75	0̇,68	0,77	0,76
Stand. dev.	0,09	0,10	0,08	0,07	0,08	0,09	0,09	0,13	0,17	0,07	0,08
Skewness	−0,23	−0,30	−0,07	0,04	−0,20	−0,44	−0,31	0,24	−0,03	−0,53	−0,21
Kurtosis	2,41	3,49	2,19	2,94	1,96	3,10	2,33	3,38	2,78	3,57	2,69
x̄ Elongation	0,78	0,76	0,78	0,73	0,75	0,81	0,84	0,78	0,74	0,79	0,78
Stand. dev.	0,13	0,12	0,12	0,09	0,12	0,10	0,10	0,11	0,11	0,08	0,11
% Oblate	40,38	27,78	31,48	20,45	32,00	39,39	53,33	55,17	50,00	45,45	50,00
% Prolate	53,85	63,89	68,52	70,55	68,00	51,52	40,00	41,38	50,00	36,36	44,74
% Blades	0	8,33	0	0	0	9,09	6,67	3,45	0	18,18	5;26
M_x Oblate	−8,37	17,89	−7,82	−5,59	−4,76	−6,55	−6,67	−6,10	−14,67	−6,88	−7,14
M_x Prolate	+11,28	+8,00	+7,65	+7,31	+7,14	+9,12	+4,90	+7,48	+7,89	+5̇,71	+6,90
x̄ Oblate	−3,02	−3,37	−4,00	−3,11	−3,17	−3,25	−4,69	−2,63	−5,33	−2,04	−2,80
x̄ Prolate	+3,97	+3,46	+3,50	+3,72	+4,15	+3,48	+2,00	+4,18	+4,72	+3,70	+3,88
Skew. Oblate	−0,51	−0,24	0,18	0,36	0,77	0,12	0,20	−0,50	−0,92	−2,43	−0,76
Skew. Prolate	−0,05	0,56	0,53	−0,17	−0,18	0,77	0,72	−0,43	−0,08	−0,29	0,42
Kurt. Oblate	2,18	1,48	2,22	1,48	2,92	2,00	1,39	2,12	2,35	8,91	3,82
Kurt. Prolate	4,29	2,28	2,13	−1,74	1,74	2,93	2,17	1,58	2,35	1,49	2,27
x̄ No facets	2,04	1,97	3,07	2,05	1,52	2,03	1,53	1,62	2,29	2,21	1,50
Stand. dev.	1,37	1,11	1,83	1,24	1,00	1,05	0,99	1,15	1,33	1,27	1,08
x̄ a-axis (cm)	9,14	8,55	7,36	6,48	4,25	5,35	3,87	4,62	5,18	5,08	3,83
Stand. dev.	6,17	6,78	5,56	4,92	2,83	2,70	1,59	3,07	2,30	2,75	1,81
n	53	51	60	51	46	53	46	59	44	58	58

Mx = Maximum value observed

TABLE II
Correlation (r) for Shape Indices with Clast a-Axis Length

	Bazley 1	Bazley 2	Ifafa 1	Ifafa 2	Port Shepstone	Shelly Beach 1	Shelly Beach 2	Mkomazi Valley 1	Mkomazi Valley 2	Mkomazi Valley 3	Mkomazi Valley 4
a-axis with roundness											
r =	+0,02	+0,17	+0,22	+0,04	+0,10	+0,48	+0,63	+0,17	+0,03	+0,14	+0,06
r^2 =	0,000 4	0,03	0,05	0,001 7	0,01	0,23	0,40	0,03	0,000 9	0,02	0,003
accept/reject (0,05 level)	×	×	×	×	×	√	√	×	×	×	×
a-axis with sphericity											
r =	−0,43	−0,52	−0,20	−0,45	−0,10	−0,18	−0,02	+0,28	+0,08	−0,12	−0,18
r^2 =	0,184 9	0,270 4	0,040	0,202 5	0,010	0,032 4	0,000 4	0,078 4	0,006 4	0,014 4	0,032 4
accept/reject (0,05 level)	√	√	×	√	×	×	×	√	×	×	×

As flatness indices approximate to the inverse of sphericity (King and Buckley, 1968) the a-axis/sphericity correlation coefficient also gives an indication of the flatness shape–size relationship.

durable lithologies (e.g. quartzite) may travel a long distance with very little change and because there is a variety of rock types (Ehrlich *et al.*, 1980) all parts of the shape range will not be affected by abrasion at the same rate. Thus, the cumulative mixing of rock types can introduce a variety of shapes and sizes. Further, it is possible (Lister, 1981) that all clasts of the same lithology, irrespective of size, may be subject to rounding due to the removal of asperities but this will have little effect on particle size, save its slow reduction. Thus, clasts of all sizes of any given lithology within the tillite, could display similar roundness indices. In fact it is the very variation in rock type which would be expected to produce a polymodal–polydimensional shape frequency distribution. The polymodality may well be provenance-induced process-sorting by shape, i.e. different lithologies will generate different shapes and, in addition, the processes they are subject to will have an effect and may cause sorting (Ehrlich *et al.*, 1980; Hall and Visser, in prep.).

So in essence, the suggestions of Stratten (1968) and Tankard *et al.* (1982) that size–shape relationships exist for Dwyka tillite clasts appears contradictory to general theory. This is not to say that under certain circumstances localized sorting may not produce a size–shape relationship, but to extrapolate this to the whole of the glaciogenic deposits of southern Africa is untenable. More importantly, the results of this study do not support these qualitative impressions.

The data exhibit a clast shape variety throughout the measured size range (0,015 m to 0,387 m) but with no consistent size–shape relationships evident (Table II). The mean values cited resemble those obtained from studies of Cenozoic sedimentation in Antarctica by Barrett (1975) and Glasby *et al.* (1975). The mean roundness indices are very similar (their observations give means ranging from 0,17 to 0,25) whilst their sphericity measurements are marginally lower (means of 0,60 to 0,71) than those recorded in Natal. The problem, noted by Glasby *et al.* (1975, p. 615), that the relatively low roundness indices seem to contradict the abundant striae which suggests that most debris moved subglacially, also arises for the Natal data. However, one possible explanation may be, as Domack *et al.* (1980) have suggested, that the region was one in which lodgement processes were not as active, in respect to modifying clast shapes, as they are in other regions. Alternatively, the relatively high standard deviations found for the roundness indices (0,10 to 0,29) suggest a large spread of shapes about the mean, and so may indicate that the tillite at the sample points is a product of more than one entrainment and/or transport mechanism, or, more likely, a mixing of lithologies with an overall tendency towards lower rounding values.

The two samples from Shelly Beach do, with respect to roundness, satisfy the suggestions of Stratten (1968) and Tankard *et al.* (1982). Sphericity–size relationships, however, show negative correlations and are contrary to the intimations of Tankard *et al.* (1982). Correlations for this study (Table II) suggest that generalized size–shape relationships cannot safely be assumed for the whole of the Dwyka. It is thought (Hall and Visser, in prep.) that the situation is far more complex and that found relationships may often reflect lithology and/or travel distance. The attributing of cause to observed shapes could be a useful reconstructional tool but is far from simple (Barrett, 1980). Whatever the cause of site-specific size–shape correlations, it does not detract from the present study in which the aim is *solely* to investigate non-lithologically controlled shape quantification. If anything, the number of questions this study has begged shows the inadequacy of the qualitative approach.

VI. CONCLUSIONS

From the pilot study it would appear that the hitherto simplistic, qualitative judgements of clast shape, surface characteristics and size–shape relationships are in need o revision. Clast shapes show the variety which might be expected in glaciogenic deposits whilst surface striation and/or facets appear to be more abundant, at least in this area, than is suggested by Stratten (1968) and Von Brunn and Stratten (1981). Perhaps most important is that the size–shape relationship cited by Stratten (1968) and Tankard *et al.* (1982) appears unfounded. In fact, should any quantitative size–shape relationship become apparent in a future investigation it should constitute a study within itself; for such a situation requires explanation. The data obtained in this pilot study appear to call for more rigorous analysis of clast shapes in future research programmes. In the meantime the qualitative judgements regarding clasts in the Dwyka tillite should be viewed with some circumspection and care should be taken in using any such information for interpretive purposes.

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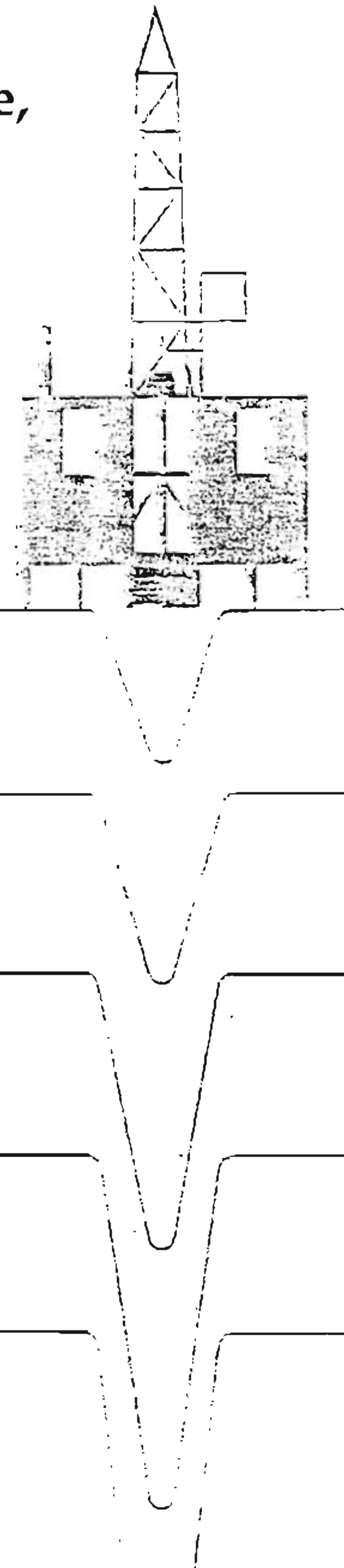

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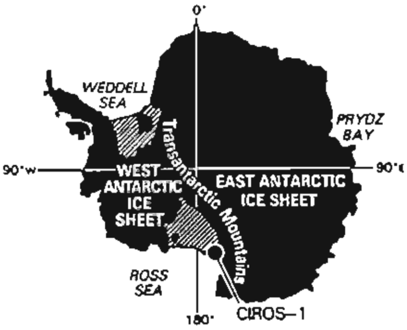

Clast shape

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Abstract

Throughout its length the CIROS-1 core contains clasts with shapes and surface features characteristic of subglacial transport, though they occur in a range of facies from diamictite to deep-water mudstone. Striae and/or facets were found on 60% (206) of the whole pebbles examined. Bullet-noses are a persistent feature in the upper part of the core. Roundness was measured on over 3000 clasts and varied little from a subrounded average. Some clasts also show weathering rims that suggest exposure on land prior to transport offshore.

Keywords Antarctica; McMurdo Sound; CIROS-1; Cenozoic; clast shape; striae

Introduction

Shapes of clasts have been used in many studies of conglomerates and diamictites both ancient and modern to help distinguish those of glacial from those of nonglacial origin (Fisher & Bridgland 1986) and to differentiate between different glacial facies (King & Buckley 1968; Barrett 1975; Domack et al. 1980; Dowdeswell et al. 1985; Hall & Visser 1984). This report describes the shapes of clasts from the CIROS-1 drillhole, which provided almost 700 m of nearly continuous core of early Miocene to early Oligocene age (Hambrey et al. this volume). The core has a glacial aspect throughout its length, making the documentation of clast shape features particularly important. A total of 3290 clasts were examined. Most were seen only in section in the split core face, but 10% (343) were removed from the core and 6% (204) were whole (i.e., uncut). A value for roundness of all clasts was obtained using the visual roundness chart of Krumbein (1941). Where possible clasts were extracted from the core and examined for evidence of glacial transport such as striae, facets and bullet shapes (Boulton 1978; Sharp 1982). Whether the clast showed signs of breakage during transport was also noted. At a few key points a section of whole core was broken down to extract whole clasts so that three-dimensional indices such as sphericity, O-P index and flatness could be determined. In addition clasts showing signs of weathering were noted because of their value in paleoenvironmental reconstructions (Bridgland 1986).

Results and discussion

A summary of observations for each of the 22 stratigraphic units is shown in Fig. 1, and some idea of variation between facies and within units is shown in Fig. 2 for Units 10 and 17.

The number of observations through the core clearly reflects the clast content of the different units with many more taken in 'glacial' than in 'interglacial' units (Fig. 1). The proportions of striated, faceted, broken and bullet-nosed clasts are in most cases greater in the 'glacial' units, although the continual supply of such material from icebergs during interglacial periods has maintained their presence throughout the core. Weathered clasts are mostly associated with nearshore and possibly terrestrial environments with few in deeper water and more distal glaciomarine environments. A more detailed analysis of the chemistry of the weathering rinds is in progress (Hall & Bühmann, in press).

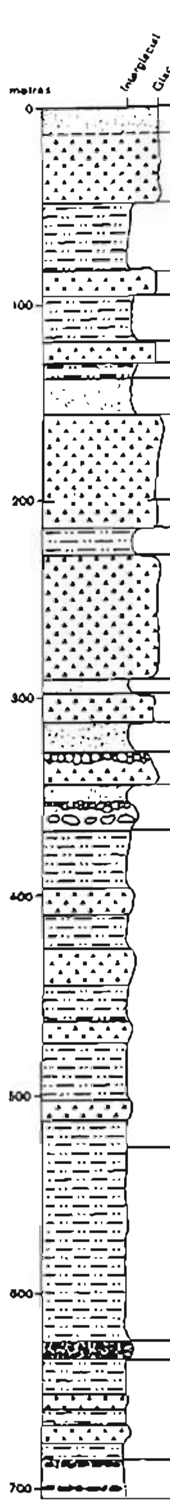
Roundness

Mean roundness for each unit (Fig. 1) varies little throughout the core, with most values ranging from 0.40 to 0.50. Higher values are found in only two places, Unit 16 (0.53), which is believed to be fluvial (Hambrey et al. this volume) and Unit 20 (0.56), a re-deposited deep water conglomerate. The roundness of the latter might reflect fluvial rounding prior to redeposition. The more detailed data available for Unit 10 (Fig. 2) show a considerable spread in mean values from 0.30 to 0.52, with diamictite beds tending to have the higher values. This is also the case in Unit 17, though the 0.60 for the diamictite is based on only 3 clasts. An analysis of the whole core by lithofacies confirms these differences. Mudstone and sandstone facies each average 0.44 with diamictites averaging 0.45 and conglomerates 0.48 (s = 0.07 to 0.08).

Whole clasts were broken out of the core for two 15 cm sections of Unit 20, providing 72 undamaged clasts for more comprehensive analysis. The mean roundness is comparable with that obtained from the sections of clasts in the core. Sphericity and two other form indices have also been calculated. Fig. 3 compares sphericity and roundness of clasts from Unit 20 with fields established by Boulton (1978) for modern glacial debris. The comparison suggests that Unit 20 clasts have experienced basal glacial transport.

Clast composition

Clast composition through the core is reviewed along with petrographic and geochemical data by Roser & Pyne (this volume). Most are granitoids from the late Precambrian and early Paleozoic basement or dolerites from the Jurassic sills intruding the Beacon Supergroup in the adjacent Transantarctic Mountains. White vein quartz is moderately



Unit	No. of Obs.	% Striated	% Faceted	% Bullet Nosed	% Weathered	% Broken	Roundness $\bar{x}$	Roundness σ
1	53	13	13	0	0	13	0.34	0.14
2	206	6	9	7	3	10	0.40	0.16
3	84	0	6	3	3	3	0.40	0.15
4	168	13	13	8	0	6	0.43	0.16
5	50	4	4	2	0	4	0.44	0.14
6	113	2	3	0	0	6	0.47	0.15
7	6	0	0	0	0	0	0.50	0.20
8	17	0	0	0	0	0	0.41	0.12
9	331	2	4	1	0	3	0.42	0.16
10	109	4	8	n	2	5	0.48	0.14
11	2	0	0	0	0	0	0.40	–
12	648	2	7	5	9	7	0.47	0.14
13	46	0	2	0	31	4	0.43	0.20
14	98	0	3	6	26	3	0.42	0.13
15	24	0	0	0	4	8	0.40	0.11
16	190	0	2	3	6	7	0.53	0.16
17	215	0	0.5	1	6	9	0.43	0.14
18	324	5	3	0.	2	4	0.45	0.16
19	306	1	0	0	9	1	0.49	0.10
20	212	7	1	0.5	22	0.5	0.58	0.13
21	105	4	0	0	1	0	0.46	0.17
22	115	1	0	0	5	1	0.52	0.15

Fig. 1 Core log for CIROS-1 showing the various shape characteristics of clasts from each unit.

common, ranging from 1 to 20%, but it occurs : frequently in Units 16 to 21. Cenozoic basalt is moder. common in Units 2 and 3 but rare below this. Clasts well-sorted quartz sandstone, presumably from Devonian part of the Beacon Supergroup, occur infrequ and mainly in the lower part of the core (Units 12, 17 tc

Striae and faceting

Striae and/or facets were found on 60% (206 of 343) o whole clasts examined. This compares closely witł proportions found by Domack et al. (1980) for gl marine sediments on the George V continental she Antarctica (57% striated, 80% faceted). Even the redepo conglomerate in Unit 20 had 21% of the clasts striated 4 % faceted (based on 72 whole clasts, Table 1). Fig. 1 much lower percentages of clasts with striae and f because only a small proportion (10%) were removed the core for examination. Bullet-shaped noses (Bou 1978) are a persistent feature of clasts in Units 2 to 17 are virtually absent below this. This may reflect a gr influence from subglacial transport for clasts in these t

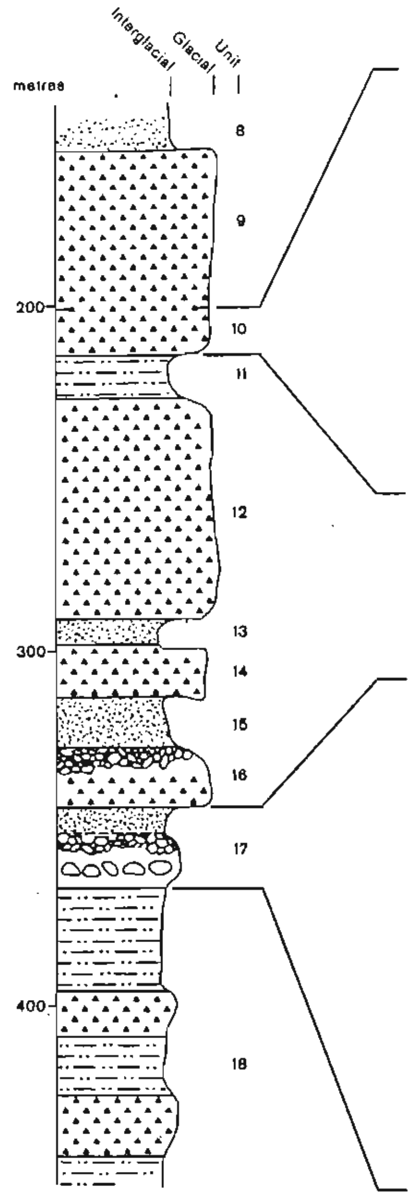
Weathered clasts

A number of clasts have a rim 1-3 mm thick that is erent in colour from the core. The feature is most com in dolerite, but was also found on some granitic clasts considered to result from terrestrial weathering. Weath rims are found throughout the core, but are generally abundant in the conglomerates, where they form (Unit 17.1, Cw) and 22% (Unit 20). The relationship tween weathering and lithology is complex. In Unit 1 only weathering rims are on granites but in Units 13, 1 19, and 21 a higher proportion of dolerites than gra have weathering rims. Some granitic clasts are so weath as to be friable throughout. Others show varying deg of red staining suggestive of oxidation. In a few insta cut sections show cracks into the clast (due to weatheri which are followed by red stains deep into the c Studies on these features are under way to establish conditions under which the weathering took place.

Summary

The surface features of clasts from the CIROS-1 indicate subglacial transport for a significant propor and show that ice was either grounded at the site or cal at sea level throughout the time represented by the The clasts are on average subrounded, which is consistent with such an origin. Even the redeposited water conglomerate at 630 m has clast features, inclu striae and facets, that indicate subglacial transp Weathering rims on a number of clasts indicate expo on land, perhaps on an outwash plain, with subseq transport by glacial ice, or in the case of Units 14 and 1 an advancing outwash plain.

Fig. 2 Partial core log for CIROS-1 showing the shape characteristics for the subunits of Units 10 and 17.



Subunit	Lithofacies	Thickness (m)	n	R*	Rind	Broken	Striae	Facetted	Bullet
10.1	Ds	0.40	4	0.30	0	0	0	1	0
	Ss	0.10	1	0.20	0	0	0	0	0
	Ds	0.42	2	0.45	0	0	0	1	0
10.2	S	0.36	0						
10.3	Ds	1.74	25	0.45	0	6	1	2	0
10.4	Sm	1.34	0						
10.5	Ds	0.68	9	0.42	1	2	0	0	0
	Sw	0.18	2	0.30			0	0	
	Dw	0.22	1	0.50	0	1	0	0	0
	Dm	1.40	3	0.40	0	0	0	0	0
	Ds	0.27	3	0.47	0	0	0	1	0
	Mm	0.61	4	0.40	0	0	0	0	0
	Sw	0.60	2	0.45	0	1	1	0	0
	Dw	1.67	18	0.51	1	1	0	2	1
10.6	Mw	0.44	0						
10.7	Dm	1.77	35	0.52	0	2	2	0	2

Subunit	Lithofacies	Thickness (m)	n	R*	Rind	Broken
17.1	Sw	2.70	6	0.43		
	no core	0.16				
	Sw	0.92	2	0.30		
	Dw	0.71	3	0.60		
	Sw	0.61	1	0.50		
	no core	0.30				
17.2	Cm	0.19	12	0.65		1
	Sw	0.27	0			
	Cw	3.19	130	0.51	11	17
	Sw	1.58	5	0.40		1
	Sm	0.91	7	0.43	1	
	Cm	0.14	8	0.39		
	Sm	4.74	11	0.35		1
	Cw	0.47	8	0.50		
	Sw	1.00	2	0.30		
	Sm	2.63	3	0.30		1
	Cm	0.34	1	0.40		
	Sw	2.40	10	0.49		
	Cm	0.30	6	0.45	1	

* R = visual roundness index

Table 1 Mean and standard deviation of several shape indices for whole clasts from Unit 20. STR - striated, FAC - faceted, BUL - bullets, VIS - visual roundness of Krumbein (1941), MW - modified Wentworth roundness of Dobkins & Folk (1970).

Lithology	No	(%)	STR	FAC	BUL	Roundness VIS	Roundness MW	Cailleux Flatness	Folk Sphericity
Basaltic	42	(58%)	13	2	1	0.50 ± 0.17	0.50 ± 0.19	186 ± 54	0.69 ± 0.11
Granitic	28	(39%)	2	1	-	0.45 ± 0.17	0.45 ± 0.19	160 ± 21	0.74 ± 0.07
Quartz	2	(3%)	-	-	-	0.60 ± 0.10	0.52 ± 0.02	237 ± 64	0.58 ± 0.11
TOTAL	72		15 (21%)	3 (4%)	1 (1%)				

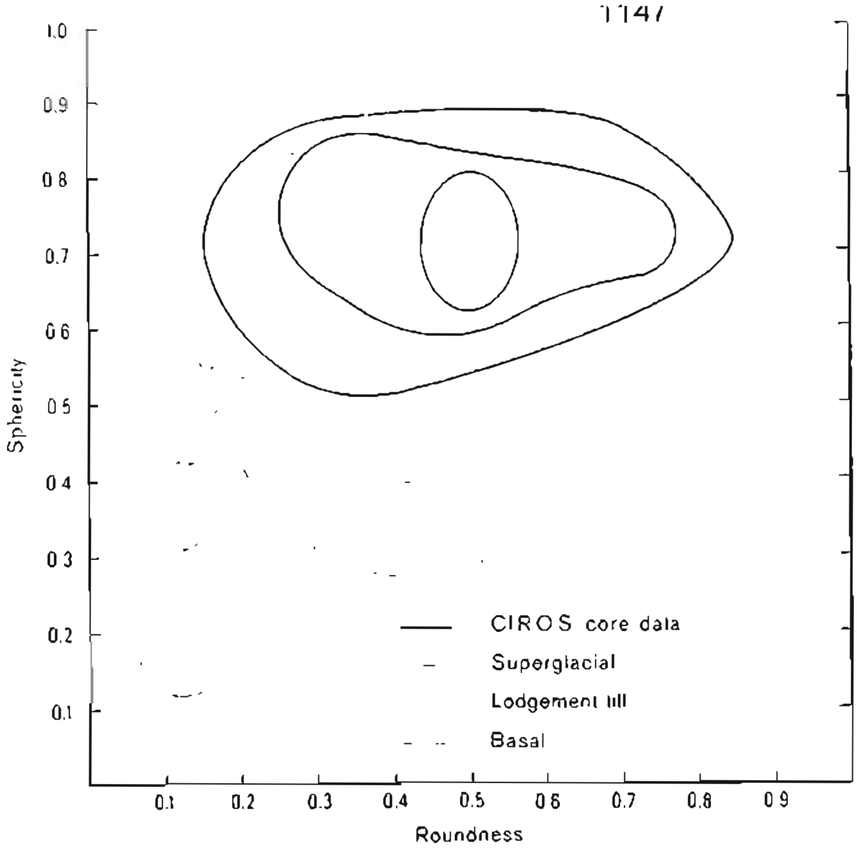


Fig. 3 A comparison of roundness and sphericity for clasts from CIROS-1 with clasts from known modern glacial environments (Boulton 1978).

Acknowledgments

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QUATERNARY GLACIATIONS IN THE SOUTHERN OCEAN: SECTOR 0° LONG.–180° LONG.

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Isolated islands in the Southern Ocean east of 0° longitude are all of volcanic origin. As some were active both volcanically and tectonically during the Quaternary, glacial features may be masked and confused by lava flows and faulting.

The location of the islands close to the Antarctic Convergence is potentially important for studying fluctuations in its position; northward migration cools the islands and causes glacier expansion.

Intercalation of glacial sediments and lava flows on Marion, Kerguelan and Heard Islands provides some evidence of glacier fluctuations during the Mid-Late Quaternary, but much of the data are equivocal. The most detailed dated Quaternary record is from Marion Island where the oldest glaciation (stage 8) had five stadials, the penultimate glaciation (stage 6) had three stadials, and the last glaciation had ended by 12 ka BP.

Glacier expansion is closely related to northward migration of the Antarctic Convergence, when lower temperature converts the high rainfall into snowfall.

INTRODUCTION

Within the vast Southern Ocean, stretching from the coast of Antarctica right up to the region of the Antarctic Convergence, the only land consists of a few scattered islands (Fig. 1); many currently support glaciers and others show signs of recent glacial activity. Although the combined land area is extremely small (9115 km^2 for those beyond the immediate Antarctic coastal margin), it nevertheless constitutes the only possible source of terrestrial records of Quaternary environmental change for the greater part of this region. These islands also span a substantial latitudinal zone and afford the possibility of detailing the nature and timing of glacial events between Antarctica and the surrounding continents. In addition, their longitudinal spread, particularly within the Atlantic–Indian Oceans sector, helps to detect east–west variations in the timing and extent of glaciation.

As yet, the Quaternary glacial history of only a few of these islands has been studied in detail. This is not surprising in view of their inaccessibility and the severe weather conditions that impose serious constraints on field investigations. In addition, the islands are usually perceived to be of lesser importance than the Antarctic continent for national scientific programmes. Thus many of the islands have very little scientific documentation about their Quaternary glacial history; anecdotal records are often the only information available.

Islands close to the Antarctic Convergence are, nevertheless, potentially very important sites for studying the response of glaciers to north–south fluctuations of the Antarctic Convergence during the Quaternary. The Convergence is not fixed, but is known to shift latitudinally on both short (seasonal) and long (Quaternary) time scales. The decrease in temperature associated with a northward movement of the Convergence will cause a larger proportion of the high annual precipitation of these islands to fall as snow. Such increased snowfall, combined with decreased ablation (due to lower temperatures and a higher incidence of cold southerly winds), initiated or increased glaciation on these southern islands during the Quaternary (Denton and Hughes, 1981; Mercer, 1983). Thus, unlike the Antarctic Continent, the sub-Antarctic islands often provide records of all scales of glacier fluctuations, including those of the later Quaternary and Holocene (Grove, 1988). In fact, Mercer (1983) was aware that the greatest glacial–interglacial environmental changes, occurred on islands presently lying north of the Convergence but which were located south of it during glacial periods.

The sub-Antarctic islands (Fig. 1) are situated north of the pack ice limit and are within, or very close to, the Antarctic Convergence (Walton, 1985). There is a gradation in present day ice cover, from extensive ice caps on islands south of the Convergence (e.g. South Georgia, Heard Island, Bouvetøya), to the smaller ice fields on islands like Îles Kerguelen situated on the Convergence and to those islands just north of the Convergence (e.g. Crozet, Prince Edward, Macquarie) that have extremely limited or no permanent ice at the present time. Most sub-Antarctic islands have mild summers with mean monthly temperatures above 4°C, and cold winters with mean monthly temperatures a little below 0°C. Much lower temperatures occur at higher elevations where permafrost may exist. As the ameliorating influence of the ocean produces a fairly equable climate near sea level there may be an extensive vegetation cover (e.g. tussock grasses on the lower ground). At higher elevations the cold-tolerant cushion plant *Azorella* is found whilst higher still, where climatic conditions may become almost 'Antarctic', only lichens can survive.

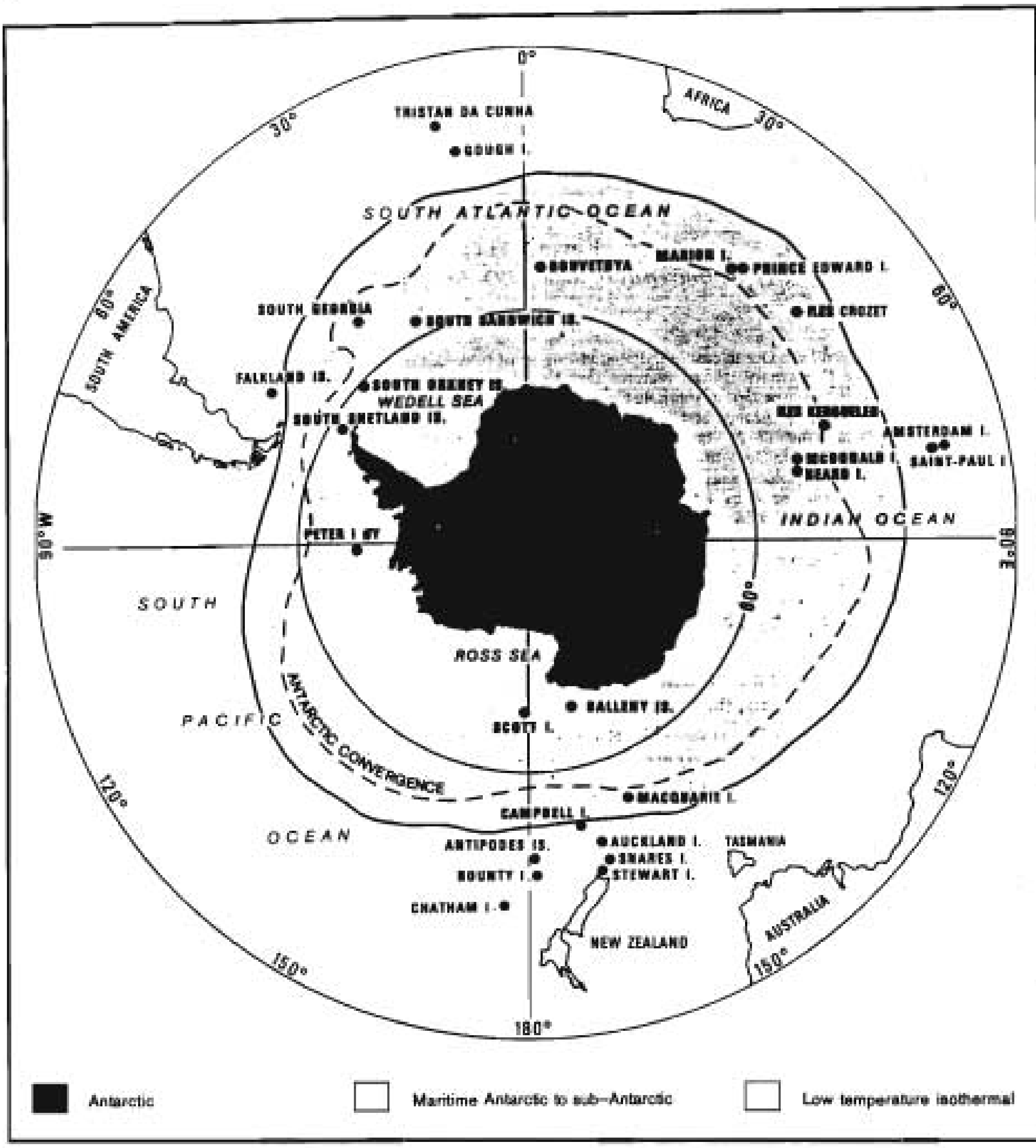

FIG. 1. Map of the Southern Ocean indicating the position of the Antarctic Convergence and the location of the islands mentioned in the text.

Owing to their situation within the west wind belt, the sub-Antarctic islands usually experience strong, westerly winds and high precipitation. Although temperatures are not particularly cold, these islands receive little direct radiation due to the high incidence of cloud cover. Thus, with conditions already conducive to the maintenance of a snow and ice cover on several islands and very close to the glaciation threshold on the others, the sub-Antarctic islands are potentially good reflectors of changes in glacial conditions resulting from fluctuation of the Antarctic Convergence.

MACQUARIE ISLAND

Macquarie Island (54°30′S, 158°55′E) is situated approximately half-way between Australia and Antarctica, and is to the north of the Antarctic Convergence (Fig. 1). It is 34 km long, 5 km wide, and is composed mostly of basaltic rocks but with some dolerite dykes, gabbros and minor peridotites (Crohn, 1986). The island is aligned along the Macquarie Ridge, which is a zone of seismic activity that extends south–southwest from South Island of New Zealand right through to the

Balleny Islands (Crohn, 1986; Selkirk *et al.*, 1986). Faulting is a major influence (Christodoulou *et al.*, 1984), and the resulting structural lineaments and block faulting are an important feature of both the ridge and the island (Ledingham and Peterson, 1984). At present neither glacier ice nor perennial snowbanks exist on the island (Colhoun and Goede, 1974).

Based upon notes taken in the field by Blake, Mawson (1943) presented the first geomorphological–geological description of Macquarie Island and proposed the idea of former glaciation. Mawson suggested that the entire island had been covered by ice that originated from the broad, shallow off-shore shelf area to the west which had been exposed by the eustatic lowering of sea level. He also referred to the recognition by Blake of such glacial features as till, striae, roches moutonnées, moraines and fluvioglacial deposits. Ivanac (1948) concurred with the hypothesis of glaciation but suggested that the ice had originated in the southwest and over-rode the island towards the northeast. Ivanac also proposed that subsequent to the maximum ice cover, during which ice on the east coast extended down to 125 m altitude, retreat took place in a series of stages, the final one being cirque glaciation only. Subsequently, Gwynn (in Law and Burstall, 1956) questioned the initiation of ice on the exposed shelf area and suggested that glaciers had originated on the plateau surface of the island. In addition, Gwynn argued that parts of the island (mainly headlands) and the exposed shelf had not been ice covered and that these areas acted as botanical refugia.

Colhoun and Goede (1974) also recognised many landforms and deposits of glacial origin on the island. They found that although the glacial imprint was widespread on the northern and eastern parts of the plateau surface, it was absent from higher parts of the mountains and from the lower hills (Fig. 2). In addition, the east coast margin below the plateau contained little evidence of glacial action. Following a detailed investigation of part of the island, Colhoun and Goede concluded that the ice had originated from basins, valleys and hollows of the main plateau rather than from an off-shore position and that there had not been a complete ice cover; centres of ice growth were determined by a combination of the pre-glacial topography and the wind-drifting of snow. Thus small ice caps, valley and cirque glaciers developed contemporaneously on different parts of the island and covered approximately 40% of the present island area. Glaciation was thought to have been initiated by northward shift of the Antarctic Convergence which depressed the mean annual temperature by 3–4°C and so caused the $\geqslant$1000 mm year^{-1} precipitation to fall mainly in the form of snow. Deglaciation was synchronous about the island and resulted from the southward movement of the Antarctic Convergence.

Subsequent investigations by Löffler and Sullivan (1980) suggested that a far greater proportion of the island had experienced glaciation. Although evidence of ice action on parts of the island is sparse, they argued that it was hard to accept one part of the plateau being glaciated whilst another part, at the same elevation and with the same exposure to the prevailing winds, was not. Ultimately, they concluded that although certain coastal and shelf areas escaped glaciation, the plateau had been almost totally ice covered and that this fed small valley glaciers that flowed (mostly) eastwards onto parts of the shelf (Fig. 2). That many glacial features are poorly developed in certain areas was explained by the small size of the glacier catchments and the relative thinness of the resulting ice. Only in the valleys was the ice thick enough to have eroded the landscape. The absence of glacial landforms above 200–250 m (also reported by Colhoun and Goede, 1974), was explained by the masking effects of periglacial mass movements subsequent to glacial retreat.

Ledingham and Peterson (1984) have recently produced a completely new interpretation. They investigated the structural history of the island together with the raised beaches that occur up to elevations of 270 m a.s.l. The study concluded that Macquarie Island recently underwent a stage within which ". . . rapid uplift associated with block-faulting and other tectonic factors played a much more important role in landform evolution than was previously recognised" (Ledingham and Peterson, 1984, p. 234). It appears that landforms interpreted by others as glacial are examples of "convergence of form" and that in reality they were *not* produced by the action of ice. Further, as uplift of the island to the snowline altitude occurred *after* the last glaciation, very little land has experienced glacial activity. Ledingham and Peterson thus concluded that only a few small glaciers formed during the Quaternary.

Although Ledingham and Peterson acknowledged the need for more data, particularly regarding dates for the raised shorelines, their hypothesis has substantial merit and fits the available evidence. Thus, although a northward shift of the Antarctic Convergence during the Quaternary would have depressed temperatures such that more precipitation fell as snow, very little of the island was high enough to be above the snowline. Only a few small glaciers (e.g. on the south side of Boot Hill and on the hill north of Square Lake) were able to develop and it is at these locations that striated material is found. What were considered by earlier workers to be deepened glacial lakes and melt-water channels are now recognised as structural features. Areas suggested by Colhoun and Goede (1974) and Löffler and Sullivan (1980) to have been glaciated, exhibit extensive raised beach deposits; this unconsolidated material is unlikely to have survived glaciation with removal and/or burial, and has been elevated tectonically.

Thus ideas appear to have gone from one extreme to the other — from full ice cover to that of but a few, very small glaciers. More data are certainly required to test both hypotheses.

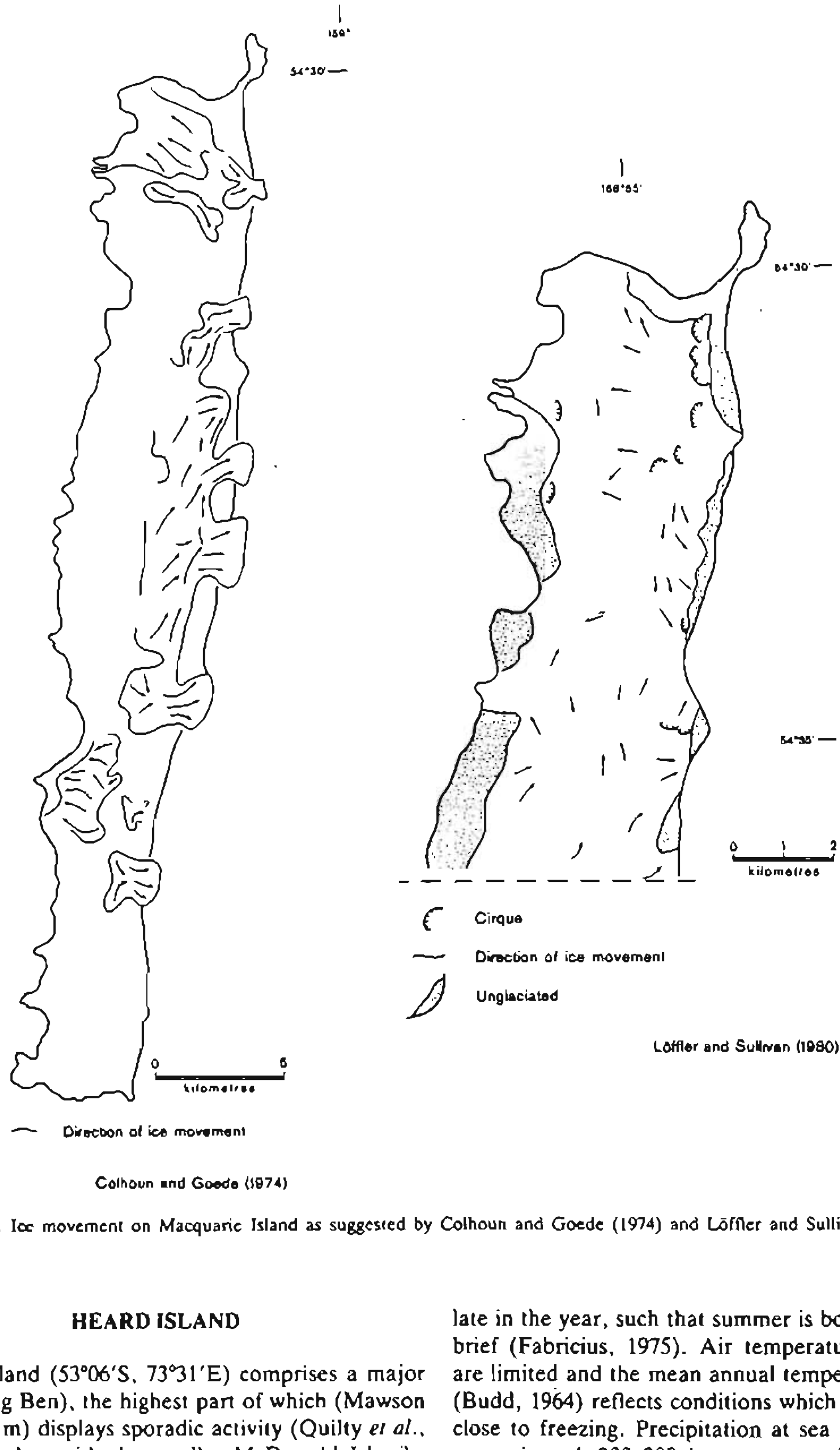

FIG. 2. Ice movement on Macquarie Island as suggested by Colhoun and Goede (1974) and Löffler and Sullivan (1980).

HEARD ISLAND

Heard Island (53°06′S, 73°31′E) comprises a major volcano (Big Ben), the highest part of which (Mawson Peak, 2745 m) displays sporadic activity (Quilty *et al.*, 1983). Together with the smaller McDonald Islands, Heard Island comprises an area of 380 km^2, of which about 81% is covered by permanent snow and ice (Walton, 1985). Situated to the south of the Antarctic Convergence, the island is under the influence of cold Antarctic air which maintains winter conditions until late in the year, such that summer is both delayed and brief (Fabricius, 1975). Air temperature fluctuations are limited and the mean annual temperature of 0.5°C (Budd, 1964) reflects conditions which are continually close to freezing. Precipitation at sea level occurs on approximately 280–300 days per year, of which snowfall comprises 21% of occurrences and snow mixed with hail or liquid precipitation a further 45% (Loewe, 1957). At higher elevations snow predominates and the firn line is estimated to be at 300 m (Mellor, 1959). Frequent snowfall combined with limited ablation resulting from the low temperatures and small radiation receipts (average sunshine is 11.9% of that possible) produces a large net annual accumulation such that many of the glaciers are very active and terminate in the sea (Fig. 3). Lundqvist (1988) described glacier conditions during 1987 and the presently-forming tills and moraines.

Whilst the present-day equilibrium line altitude (ELA) lies between 200 m (Allison and Keage, 1986) and 350 m (Colhoun and Peterson, 1986), it has clearly risen during the last three decades, explaining the glacier thinning observed by Lambeth (1950). Proof of glacial retreat during the past 100 years is a trim line situated 30 m above the level of the Baudissin Glacier (Fig. 3). That the Baudissin Glacier was even thicker at an earlier time is indicated by the presence of old morainal material above the trim line along the flanks of the bordering Mount Drygalski (Mellor, 1959). Evidence obtained in 1963 (Mercer, 1967) indicated that Fiftyone Glacier had shrunk and that Winston Glacier had retreated several hundred meters since 1954. In addition, the Stephenson and Brown Glaciers on the north coast, which both terminated in coastal ice cliffs in 1954, had retreated by 1963, forming a lagoon and shingle beach in their wake. Allison and Keage (1986) cited trimlines as much as 90 m above the present glaciers, indicating this degree of thinning over recent decades. Radok and Watts (1975) stated that glaciers receded slightly between 1948 and 1954 but then extensively during the subsequent ten years. According to Colhoun and Peterson (1986) Winston Glacier decreased in thickness by 90 m, Baudissin Glacier by 60 m and Vahsel Glacier by 30 m. However, whilst some glaciers retreated others subsequently re-advanced and these glacier variations are thought to be due to changes in the proportion of depressions passing Heard Island (Radok and Watts, 1975). Significant changes in the number of depressions on the equatorial side of the island precede monthly mean temperature anomalies. These changes in temperature may partly explain why some glaciers are advancing whilst others have retreated. As a result of glacier retreat and the extensive debris cover of some glaciers, a sequence of moraines (the Dovers Moraines) has developed in front of Stephenson Glacier on the east of the island (Fig. 3).

The Dovers Moraines, which are up to 1.6 km in width and attain heights of 150 m, comprise three groups representing successive moraines of different ages (Stevenson, 1964), all of which are probably late Pleistocene (Budd and Stevenson, 1968). The oldest moraines comprise low vegetated hills, the intermediate group (occurring in several parallel lines) form higher vegetated ridges with steep eastern slopes, whilst the inner, most recent moraines are unvegetated and ice-cored (Stevenson, 1964).

Although the Big Ben volcano developed only a few hundred thousand years ago (Clarke *et al.*, 1983) there is evidence that glaciers were active before that time. Tills (or tillites) of the Drygalski Formation (originally called the 'Drygalski Agglomerates': Lambeth, 1952) are of late Miocene–early Pliocene age (Quilty *et al.*, 1983). The Formation is 250 m thick in places and is composed of glacial sediments interbedded with lavas, some of which may have been formed below sea level. Evidence of glacial activity comprises striated and faceted clasts, laminates, glacial outwash sequences and roches moutonnées formed in the lavas underlying the sediments (Stevenson, 1964). Fabrics obtained within the sediments indicate that the topography during early glacial times was significantly different to that of the present. Some of the glacial sediments are likely to have formed from ice melting in the sea as they are associated with pillow lavas (Stevenson, 1964). However, Quilty *et al.* (1983) found freshwater diatoms in some of the sediments and suggested that the sporadic fragments of marine species were possibly windblown; the freshwater diatoms are probably associated with a meltwater source. Although glaciation may have begun in the Miocene, the island is nevertheless considered by Stevenson (1964) to be 'marginal'

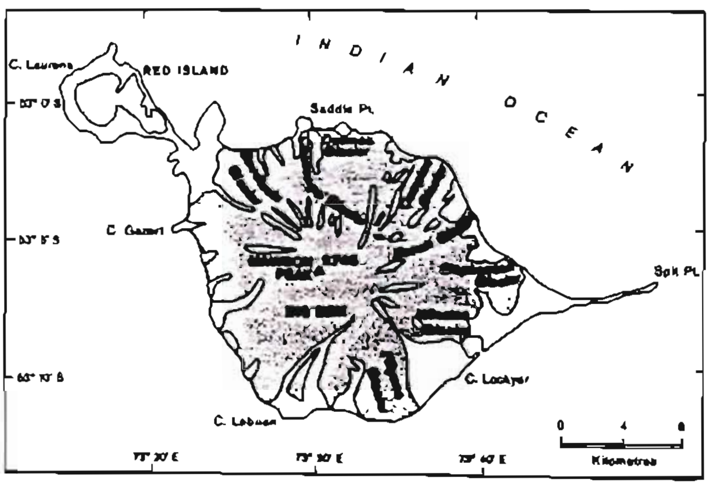

FIG. 3. The ice cover (shaded) of Heard Island. (After Mercer, 1967.)

insofar as slight north or south shifts of the Antarctic Convergence would have marked effects upon the climate and hence upon the glaciers and the resulting sediments.

Ealey (1954, p. 106) suggested that down-faulting of Heard Island must have accelerated glacier recession at the northern end of the island and that this may have initiated the volcanic activity that produced pyroclastic cones in this area. On purely speculative grounds, based upon studies on Marion Island, Hall (1982) suggested that the faulting may have been in response to a loss of the ice overburden due to deglaciation initiated by northward movement of the Convergence. None of the literature on Heard Island contains data on the chronology or sequence of Quaternary glaciations.

BOUVET ISLAND (BOUVETØYA)

Bouvet Island (54°25′S, 3°21′E) is of volcanic origin and, situated in the South Atlantic Ocean, has an area of only 50 km^2. Located approximately 500 km south of the Antarctic Convergence, Bouvet experiences a polar maritime climate. Snow can fall at any time during the year and the average net accumulation is 30 cm water equivalent (Mercer, 1962). Daily and seasonal temperature variations are very small, with a mean January temperature of +1.3°C and a mean July value of −2.7°C at sea level (Vinje, 1981). As cloud cover is extensive (80–90%), radiation receipts are often low; Vinje (1981) cited values ranging from 238 cal cm^{-2} in February to 23 cal cm^{-2} in June. Bouvet Island rises to a height of 780 m and has steep cliffs on all but the eastern side (Orheim, 1981). It is 93% ice covered, with an ELA of 300–350 m on the northern and eastern sides, 50 m lower on the south and 100 m lower on the west; glaciers flow radially down to the coastal margin (Orheim, 1981).

Hotedahl (1929) suggested that the firn line was only just above sea level, but in 1955 more snow-free areas were recognised than in 1929 (Mercer, 1967). Further, according to Mercer, there was little change in the ice cover on the south coast during the 30 years 1899–1929; but on the east coast there appears to have been an increase in the area of exposed bedrock. Comparison of photographs taken from 1927 to 1930 with those obtained during 1976 and 1979 indicate no change in ice cover for the east and northeast of the island (Orheim, 1981). On the southern coast a retreat of 10 to 40 m along a 1 km stretch of coastal ice plus an approximate 80 m of retreat of the Rustadkolleen glacier are recognised. On the western coast glaciers have retreated 50 to 100 m, whilst on the north coast one glacier has *advanced* 150 m into the sea. Although overall ice loss is calculated as 0.12 km^2 this is balanced by the 0.12 km^2 growth of the glacier on the northern coast. Despite this rough balance, Orheim is of the opinion that there has been a slight overall retreat since 1927. However, as the frontal positions of the glaciers are strongly related to topography, and many terminate in calving ice fronts or in the sea, their frontal positions are probably not particularly sensitive to small climatic fluctuations.

The detailed research of Orheim indicates that information exists only for the ice cover of the last 80 years. No data relating to ice conditions prior to 1900 appear to be available. In fact, it has been suggested (e.g. Fricker, 1904) that Bouvet may have actually disappeared in the recent past as it lies in an active volcanic and tectonic setting (Lovlie and Furnes, 1978); fumaroles presently exist on the island. Historical records indicate that two small islands (Liverpool and Thompson) once existed close to Bouvet and have now disappeared. As the famous explorers Cook and Ross both failed to find Bouvet in spite of extensive searching (Ross, 1982) the island may not have existed during the 18th century. Thus the Quaternary record for this island is limited to details of glacier activity since the start of the present century.

ÎLES KERGUELEN

Îles Kerguelen (48°27′–49°58′S, 68°25′–70°35′E) comprise a 5799 km^2 main island (Grand Terre) and a further 300 islands, giving a total area of 6200 km^2. Owing to the insignificant size of most islands only Grand Terre will be considered here. Situated on the Kerguelen Plateau, the island is of volcanic origin and continues to support active fumaroles. During the early Quaternary, volcanism was characterised by the formation of stratovolcanoes with parasitic cones and, more recently, by the eruption of pumice and trachytic tuffs (Nougier, 1972a). The island is situated just north of the Antarctic Convergence and experiences low temperatures (mean of 4.6°C), frequent frosts, extensive cloud cover and strong westerly winds (Weyant, 1967). At higher elevations climatic conditions can be severe, with frequent snowfall and strong katabatic winds that drain from the ice caps into the valleys below. The topography is very broken, with many large valleys, mountains rising to 1960 m, ice caps and glaciers in the west, and extensive flat plains to the east.

The present-day ice cover is approximately 750 km^2 (Denton and Hughes, 1981), the bulk of which (ca. 500 km^2) comprises the Cook ice cap and its 40 outlet glaciers (Fig. 4). Other major ice caps and glaciers are found on the Presqu'île de la Société de Géographie, péninsule Rallier du Baty, Mont Ross and Pic Guynemer (Fig. 4). The largest ice body, the Cook ice cap, rises to 1100 m a.s.l. and has an ELA at 600–900 m a.s.l. (Mercer, 1967). Information regarding the present glacier cover, together with the history of recent changes, are given in Mercer (1967), based upon a detailed study initiated in 1961–1962 by Bauer (1963a) and air photographs taken during 1963 (Bauer, 1963b).

The gross morphology of Kerguelen largely predates the Quaternary, whilst features of glacial erosion and deposition are all thought to be of Quaternary age. Tertiary volcanism produced the basaltic lava flows so characteristic of the island (Aubert de la Rüe, 1962;

Nougier *et al.*, 1983) whereas tectonic activity at the end of the Tertiary produced the major structural lineaments that were then exploited by Quaternary glaciers (Nougier, 1970, 1972b). Simplistically, glaciation appears to have been in two stages: an 'ancient' glaciation approximating to the Villafranchian (early Pleistocene) and a recent one of Würm (late Pleistocene age: Nougier, 1970). However, there is no clear evidence supporting this assertion. Werth (1908) believed that the early glaciation was of Alaskan type supported by mountains in the west of the island, but Aubert de la Rüe (1932) concluded that many centres of glaciation had coalesced as they grew, finally producing an ice cap like that of present-day Greenland. During this early episode ice covered all of the island and exploited the tectonic lineaments to form the extensive glaciated valley system. Nougier (1970, pp. 381–385) discussed the relationship between pre-glacial tectonic activity and valley orientation and morphology but no reference has been made to sediments deposited by this early glaciation.

Glacier fluctuations would have occurred on the island throughout the Quaternary, but Nougier (1970) believed that most evidence of earlier glaciation was destroyed during the last glaciation. Thus it is difficult to establish a chronology of events. This problem is compounded by the lack of glacial sedimentary sequences older than the last glaciation and by the scarcity of dated material. For instance, it is possible that the sequence of moraines in the Vallée de la Longue Attente may not be of last glaciation age and that the sandur below the cirque on Mont Velain is likely to comprise fluvioglacially-reworked material produced by an earlier glaciation (Nougier, 1970).

During the last glaciation ice may not have completely covered the island, but as terminal moraines are absent (Bellair, 1965), it must have extended beyond the present coastline in a number of places. Bellair also argued that the last glaciation must have been of much smaller extent than that of the early Pleistocene as some areas lack tills (e.g. péninsule Loranchet), and others show no sign of isostatic rebound due to the loss of a thick ice cover (e.g. péninsule Courbet). Nougier (1970) suggested that a number of major centres of ice cover developed during the last glaciation in the massif Galliéni, la Société de Géographie, Rallier du Baty, Mont Ross and the present Cook ice cap, whilst in other areas (e.g. péninsule Jeanne d'Arc) there were cirque glaciers and/or nivation hollows. The Holocene retreat of many of these glaciers is marked by sequences of moraines (e.g. see Fig. 106 of Nougier, 1970) and fluvioglacial outwash plains. More recently, Hall (1984) has suggested that the glacier cover may have been more extensive than was hypothesised by Bellair and that it grew from the main centres of ice cover noted by Nougier (1970). Calculations indicate a glacial ELA some 400 to 700 m lower than at present and this would be commensurate with an extensive ice cover.

At the end of the last glaciation, warming began around 12 ka BP and culminated in a major glacier retreat at 10 ka BP. This is deduced from ^{14}C dates and pollen analyses of two cores from péninsule Jeanne d'Arc and péninsule Ronarc'h (Fig. 4) by Young and Schofield (1973a). One thousand years of low temperatures followed and this was succeeded by a warming trend that attained temperatures greater than at present, peaking at around 5 ka BP. Core 1, collected from the northern part of péninsule Jeanne d'Arc, (Young and Schofield, 1973b) has a basal date of 12.5–12 ka BP and rests on a till-like sediment, probably of glacial origin. The other core does not contain glacial sediments but is considered to be of late-glacial age at the base. The site of core 2 may have remained unglaciated during the last glaciation (Young and Schofield, 1973b).

Frontal moraines, often ponding lakes behind them, characterise an advance presumed to be that of the Little Ice Age (Nougier, 1970). It appears that the number of moraines is constant within each of the mountain areas, thus indicating local synchroneity between glaciers. Some moraines are curvilinear whilst others are scalloped, and there is usually a sandur extending beyond the outer moraine. It is possible of course that some of the moraines could have formed in the Neoglacial interval prior to the Little Ice Age. The retreat that has taken place since the Little Ice Age is associated with the ameliorating climate (Mercer, 1962; Nougier, 1970) as indicated by the change in snowline elevation, estimated at 600 m in 1902, 650–700 m in 1932 and 1000 m in 1952; today it is said to vary between 950 m in the west and 1050–1100 m in the east (Bellair, 1965). In addition, no floating ice has been recorded in baie Norvégienne since 1952, icebergs no longer reach the shores of Kerguelen, and a number of small cirque glaciers have disappeared since 1902 (Bellair, 1968; Mercer, 1967). At Port au Française, temperatures have increased by 1.5°C between the mid 1960s and the mid 1980s (Allison and Keage, 1986).

ÎLES CROZET

Îles Crozet (46°–46°30′S, 50°30′–52°30′E), located approximately halfway between Marion Island and Kerguelen, comprise five islands. An eastern group consists of Île de la Possession and Île de l'Est and a western comprises one main island (Île aux Cochons) and two sets of islets: Îles des Pingouins and Îles des Apôtres. Situated on the Crozet Plateau, the islands are all of volcanic origin and appear to have evolved in five major stages from the Miocene (Chevallier *et al.*, 1983). Climatically the islands are very similar to Marion Island (although specific data are scant) with dominant strong westerly winds, frequent precipitation (often in the form of snow), low radiation receipts and cool but equable temperatures. The highest point on the islands, which cover a total area of only 233 km^2, is 934 m and the islands maintain no permanent snow or ice cover (Walton, 1985).

Information about the glaciation of these islands is

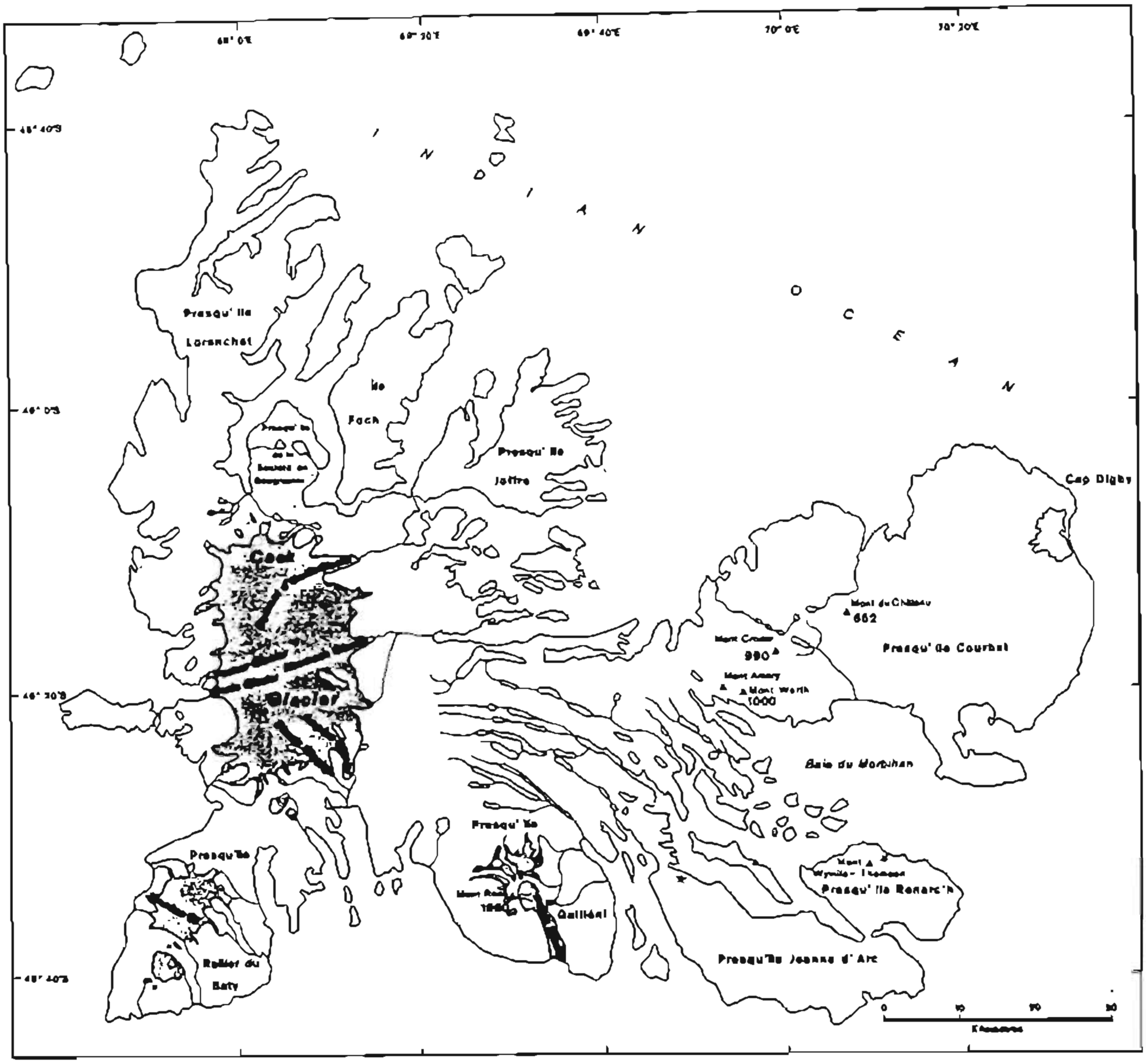

FIG. 4. Grand Terre, Îles Kerguelen, showing the main glacier area (shaded) together with the location of the two core sites (*) of Young and Schofield (1973b). (After Mercer, 1967.)

scarce, particularly as some have yet to be investigated properly and because volcanism may have been active in recent times. However, Bellair (1964) stated that there are no traces of recent glaciation (equivalent to the European Riss or Würm) on Île de la Possession, the largest and most likely to have been glaciated. Some valleys have a U-shape, but as there are no moraines, Bellair argues that if the valleys are of glacial origin, they were formed during an "ancient" event (i.e. pre-Riss).

MARION AND PRINCE EDWARD ISLANDS

Marion Island (46°54'S, 37°45'E), the larger of the two islands, lies approximately 2° of latitude to the north of the Antarctic Convergence. The somewhat oval island has an area of 290 km^2 and rises inland from the coast to a maximum elevation of 1230 m near its centre. Situated on the southeast flank of the mid-Indian Ocean Ridge, Marion is of volcanic origin and continues to experience minor eruptions (Verwoerd *et al.*, 1981). The volcanic activity has produced two 'stages' of trachybasalt and alkali basalt lavas. The first volcanic stage produced gray-coloured lavas that have been glaciated whilst the second stage is represented by black-coloured lavas and scoria cones that show no signs of glacial action. Available K–Ar ages (McDougall, 1971) for the gray lavas range from 270 ka ± 30 ka BP to 48 ka ± 25 ka BP, whilst those for the black lava are in the order of 15 ka ± 8 ka BP. Radial and peripheral faults, along which are found many of the recent eruptive centres, divide the island into a series of stepped horsts and grabens. The grabens are largely filled with the recent black lavas whilst the horsts comprise the glaciated, older, gray-coloured basalts (Hall, 1982). Raised beaches are present around the margin of the island at ca. +3 m and ca. +6 m (Hall, 1977).

Snow is normally recorded on Marion Island during every month of the year but permanent ice presently occupies only a very small area of the summit plateau (Verwoerd, 1971).

Although Prince Edward Island exhibits both the older gray and the younger black lava sequences no signs of glaciation have yet been found. Verwoerd (1971) suggested that the lack of evidence for glaciation on Prince Edward Island may be due to an absence of suitable exposures (i.e. subsequent volcanic activity masked any evidence of former glacial action) or that the combination of small size and low elevation inhibited the growth of an island ice cover. It is also possible that the very extensive erosion of the western side of the island by the strong seas (Menard, 1986, p. 114) has removed all evidence of glacial activity.

Evidence of a former, extensive glaciation on Marion Island was first reported by Verwoerd (1971) and consists of glacially smoothed gray lava outcrops that exhibit striations, chattermarks, crescentic gouges and roches moutonnées. These features were found at various locations around the island thereby demonstrating that glacial activity had affected a substantial area. As no glacial deposits were recognised it was suggested that they may lie on the adjacent sea floor, where they probably accumulated during the lowered glacial sea level.

More recently, Hall (1978, 1980a, 1983) found evidence of two, or possibly three, glacial episodes on Marion Island. Variations in the nature and extent of the ice cover were recognised for each glacial interval, whilst extensive volcanic activity, initiated by faulting resultant upon isostatic recovery during deglaciation (Hall, 1982), marks the end of each glacial phase. Whether or not a glacial 'phase' had the status of a full glaciation interval remains to be determined. In addition to the evidence of glacial erosion cited by Verwoerd (1971), Hall (1980b, 1981) found tills and moraines in positions which agree with ice flow directions deduced from striations (Fig. 5). The a-axis orientation of moraine clasts and of those in the tills concurred with postulated former glacier flow paths, while the presence of striated and faceted clasts helped substantiate a glacial origin for some of the deposits. Using this glacial information, attempts were made to estimate the former ELAs, accumulation area ratios for the glaciers, palaeo-temperatures and, in conjunction with palynological data, to describe the nature of the interglacial vegetation.

Sea level temperatures were estimated for glacial maximum conditions on the basis of the relationship between the altitude of the point of origin of lateral moraines (ca. 350 m), which corresponds roughly to the ELA, and the climatic snowline (Hall, 1980a, Table 1). A mean annual decrease in temperature of 3°–6.4°C was apparent, and agrees well with the 3°–4°C of depression estimated by Van Zinderen Bakker (1973) from palynological data. A decrease of this order, initiated by the northward movement of the Antarctic Convergence, would have caused much more of the annual precipitation to fall as snow; nevertheless, it would have still maintained positive temperatures for a greater (66%) part of the year in the former glacier ablation zones. Accumulation area ratios (AARs) calculated for a number of the palaeo-glaciers suggest values in the order of 0.6 (Hall, 1983) which would be in accord with a situation within which large snow inputs were balanced by extensive ablation. Such glaciers would have required large mass transfers to maintain their position and this would have necessitated high ice velocities. The steep clast dips (mean = 23°) found in a number of beds may reflect these high ice velocities (Andrews and Smith, 1970). During the interglacial southward shift of the Antarctic Convergence, precipitation once more fell as rain rather than snow, thereby initiating an extremely rapid deglaciation of the island (Hall, 1982).

A variety of glacial sediments (e.g. lodgement, ablation and flow tills, rhythmites with dropstones, fluvioglacial sediments, etc.) are exposed in a number of the sea cliffs and along some of the river valleys. Striated gray lava clasts are found in some of the tills (see Table 3 of Hall, 1981) whilst till fabric analyses indicate former ice flow directions that complement those suggested by bedrock striations (e.g. Fig. 8 of Hall, 1980a). Clast shape analyses (Hall, 1981) also agree with the deduced sediment genesis, with lodgement tills exhibiting higher degrees of roundness and sphericity (and the occurrence of striated clasts) than in presumed ablation tills (which also failed to show any preferred a-axis clast alignment). Many of the moraines showed distinct cross sectional asymmetry indicative of a push origin (Hall, 1980b) whilst fabrics undertaken on the (often striated) surface clasts showed long-axis orientations normal to the ridge crests (see Hall, 1980a, Fig. 4 or Hall, 1980b, Fig. 4).

A small exposure of peat situated on top of an undated gray lava and overlain by glaciogenic sediments of at least one glacial sequence provided palynological information for an interglacial interval (Scott and Hall, 1983). Although conditions appear to have been fairly similar to those at present, a number of new palynomorphs were identified, one of which may belong to an extinct taxon. Interestingly, the pollen spectra clearly show that distance from the sea increased at the onset of the following glaciation, presumably due to sea level lowering (Scott and Hall, 1983). As the underlying basalt has not been dated this interglacial information cannot be set within a time framework.

Because Marion Island is volcanic there is a limited range of material available for reworking by the glaciers. Many of the tills have an extremely high content of pyroclastic detritus as this was the material first removed by glaciers spreading over the island during the Quaternary. In fact, Hall (1983) loosely

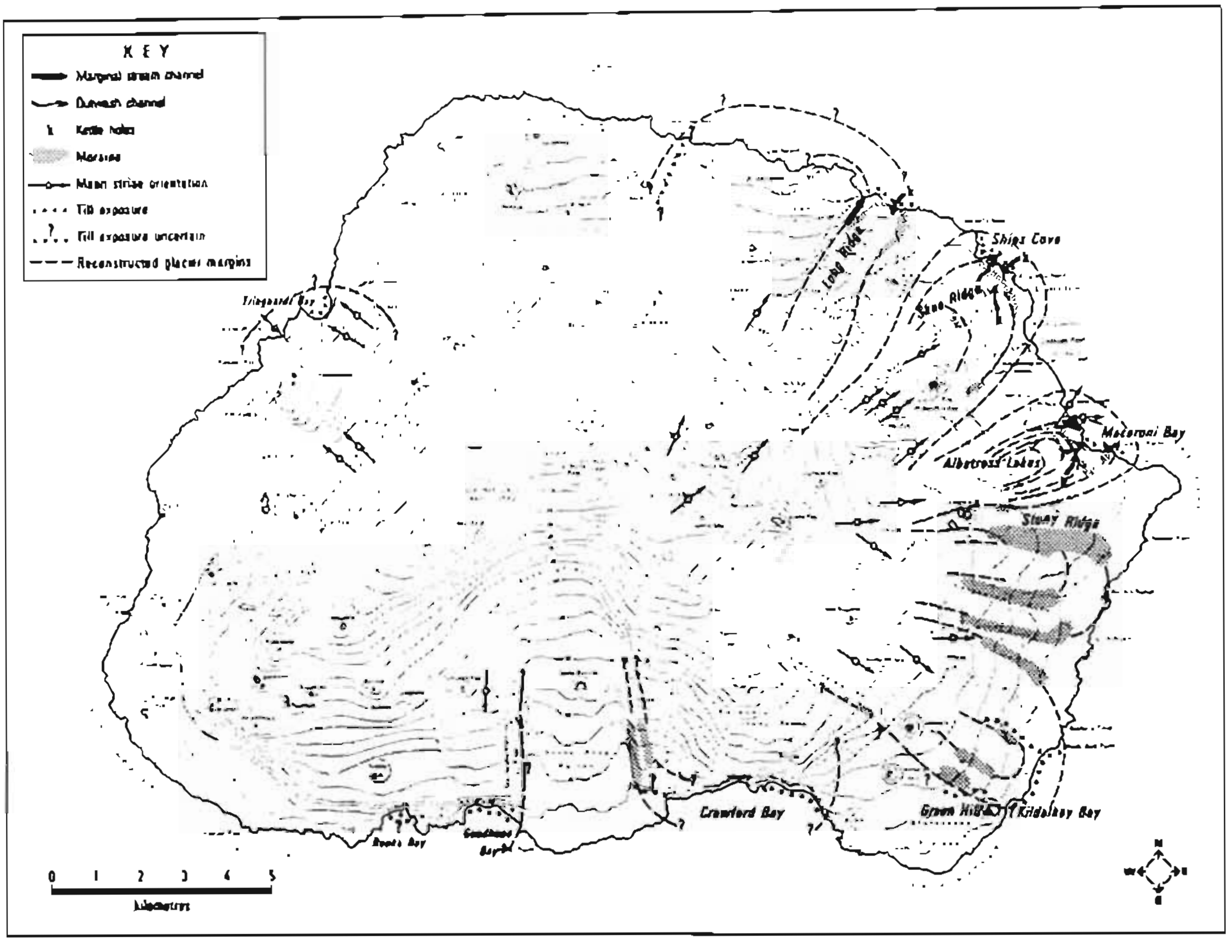

FIG. 5. Reconstruction of ice cover during the last glaciation for Marion Island based upon moraines, striae and till exposures. (After Hall, 1983.)

defined some deposits as "pyroclastic tills", but even these show an upward decrease in pyroclasts and an increase in clasts of striated gray lava.

It is probable that many sediments observed in the cliff sections are of non-glacial (interglacial?) origin, as lahars have been found at a number of localities (Hall, 1977, *pers. obs*). In addition, there are volcaniclastic sequences, frequently with material fused together by heat, that patently have nothing whatsoever to do with glacial activity.

This overwhelming volcanic influence upon the sediments of Marion Island led Kent and Gribnitz (1983) and Gribnitz *et al.* (1986) to suggest that all the sediments interpreted by Hall as glacial are in fact volcanigenic. These authors describe some interesting, but often contradictory evidence, noted during only two very brief visits to a limited part of the island. Although they agreed that the island had been glaciated, they did not explain why no glacial deposits other than moraines exist.

CONCLUSIONS

In spite of the lack of detail it is apparent that the sub-Antarctic islands experienced glacier fluctuations during the Quaternary. These volcanic and tectonically active islands have but a limited variety of materials that can be reworked by glacial action and they are subject to extensive outpourings of lava and to faulting, all of which can mask and/or confuse the glacial geological record. Nonetheless, these islands are very important in that their geographical location bridges the vast ocean between Antarctica and the surrounding continents and they may be particularly good reflectors of changes in glacial conditions owing to their marginal position. This marginality results from their location with respect to the Antarctic Convergence, for glaciation is largely dependent upon the cold associated with its northward migration.

It is difficult to derive any clear picture of the nature and extent of the Quaternary glaciation on these Southern Ocean islands as the information presently available is so patchy and lacking in dates. However, it would appear that some glacial sediments may be southern hemisphere equivalents of the North American Kansan glaciation (Marion Island) and Illinoian glaciation (Marion, Kerguelen, Heard Islands); but with the possible exception of the sediments from

Marion Island information is very sketchy and uncertain. All the islands show some evidence of glacial activity during the last glaciation, with the greatest detail available for Heard Island and Marion Island. Crozet and Macquarie appear to have been little affected; the ice cover of Bouvet would have been total; the detail from Kerguelen is unclear, but appears to indicate a substantial ice cover except for the far eastern part of the island. One of the main reasons for the lack of detail regarding pre-last glaciation events is that the nature and topography of the islands change substantially over relatively short periods of time as a result of volcanic and tectonic activity.

Ultimately what can be said about the glacial history of these islands? First, that they all appear to experience glaciation as the Antarctic Convergence moves northwards during glacial intervals. Second, that glaciation is as much a result of heavy precipitation (rainfall becoming snowfall) than particularly cold temperatures. Third, that Quaternary glaciation was extensive on many of the islands and extended beyond the present coastline in many cases. Fourth, if the record from Marion Island is correct, then the last glaciation ended at ca. 12 ka BP, following an interstade at ca. 17 ka BP; there were at least 6 stadials (Hall, 1983). The penultimate glaciation had a minimum of 3 stadials and the oldest recorded glaciation had a minimum of 5 stadials. All events are intimately tied with the movements of the Antarctic Convergence and a greater number of dates are necessary before any form of inter-island correlations can be attempted. The islands need a higher priority on the National Antarctic programmes of the relevant nations before we can obtain the valuable information that is potentially available.

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The sub-Antarctic

Much of the available information for the sub-Antarctic islands is speculative and lacks the rigorous data that would allow for detailed reconstruction of former ice cover. There are several reasons to explain the lack of information. First, the islands are highly scattered and access is, in some instances (e.g. Bouvetøya), infrequent. Second, Quaternary glacial/periglacial studies are not high on national Antarctic research programmes and thus work is limited and, in some instances, opportunistic rather than planned. All of this is further influenced by the high-cost implications for the off-shore information necessary to better understand both the nature of glaciation and it's extent. The location of these islands within the "Roaring Forties" and "Filthy Fifties" west-wind belt is not overtly conducive to off-shore coring even if it were seen as scientifically and financially viable. Lastly, as noted in the following synopses, these islands are highly active tectonically and volcanically such that their areal and altitudinal extent has changed dramatically during the Quaternary. This further confounds our ability to identify much beyond the most recent glaciation which, in some instances, may be the Little Ice Age event or, at best, the Last Glacial Maximum. Thus, both the information and the maps are limited in detail but they do indicate the potential for some very exciting research areas!

These islands are an important scientific resource as they fill a latitudinal gap between the southern continents and Antarctica itself and, as such, given adequate dated information, may help unlock the timing of glaciation along both a south-north gradient as well as along an east-west traverse (Hall, 1990). Glacial information from these islands might also be significant in understanding movement of the cold-influencing Antarctic Convergence as some are located just to it's north (e.g. Marion and Prince Edward Islands) while others are almost on it (e.g. Kerguelen), or just to it's south (e.g.

Bouvetøya). The islands may also be important as they express a physical response to smaller-term and Holocene fluctuations better than can easily be resolved for the Antarctic continent (Mercer, 1983; Grove, 1988).

Thus, although the information from these islands may have special significance, the maps and the synopsis of available material may be inadequate for many readers. Sadly, until more detailed sedimentological analyses, sub-sea topography, off-shore coring, dating, and palaeo-reconstructions of the palaeo-geography are available, information will be speculative. The produced maps must be seen within this framework; former ice margins are shown in a highly illusionary manner and should not be viewed otherwise.

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Bouvetøya (54°25'S, 3°21'E)

This island of *c.* 50 km^2 has a present ice cover of *c.* 46.2 km^2 (Orheim, 1981); an ice-free area of only 7.6%. Although there has been, based on observations from 1898 (Aagaard, 1930) and again in 1978 (Orheim, 1981), some retreat of ice margins in recent years the amounts are still relatively small (10-100 m only); conversely, some glaciers have advanced by as much as 150 m. A complication recognised for Bouvetøya but applicable to all of the sub-Antarctic islands, is that of recent (Quaternary/Holocene) changes to topography as a result of volcanic/tectonic gain or loss of land. Details regarding recent changes in ice cover are shown in Orheim (1981, Fig. 4). During the Quaternary it is probable that the island experienced extensive ice cover, as now (see Sheet 1 of Mercer, 1967), during interglacials although this may have been influenced by volcanic events. Available dates for the volcanic rocks (Prestvik and Winsnes, 1981) put the maximum age for the island as *c.* 4.5 - 5.0 Ma with recent volcanic events dated at 0.11 Ma (±0.10). Further, large-scale geomorphic events such as the landslide recorded for Nyrøysa, located in the north-west corner of the island, may have also influenced glacier extent during the Quaternary. During the glacials, with the associated fall in sea level, it would be expected that the ice cover extended beyond the present coastline. There is, however, no detailed sub-sea information that allows for determination of actual former glacier extent.

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Heard Island (53°06'S, 73°31'E)

Heard Island is a volcanic island, still displaying sporadic activity (Quilty, *et al.*, 1983), with an area of *c.* 380 km^2 that, at present, is approximately 81% covered by permanent snow and ice (Walton, 1985). Nearly all the glaciers currently end at sea level (see Sheet 1 of Mercer, 1967) although there is extensive evidence for glacier thinning and retreat over the past fifty or more years (see Hall, 1990 for a details). As with Bouvetøya, there have probably been substantial physiological changes to the island during the Quaternary as a result of volcanism/tectonics as well as erosional/depositional changes resulting from geomorphic action of mass movement and sea (e.g. significant land extension as a result of spit growth). That the island is presently significantly ice covered with glaciers ending at present sea level, Quaternary glacial episodes would have, even without comparable topography (there being till fabric evidence that topography *was* significantly different at times in the past: Stevenson, 1964), likely have experienced glaciers extending below present sea level. Interglacials may have been more complicated as this was the possible time of, due to unloading during deglaciation, glacio-isostatically driven volcanism (Hall, 1982) that may have significantly affected ice cover and Ealey (1954) has provided some evidence for this pertaining to the end of the last glacial. Whatever, the nuances resulting from tectonic/geomorphic events, the location of Heard Island assures substantial ice cover throughout the Quaternary.

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Macquarie Island (54°30'S, 158°55'E)

Located within a zone of seismic activity but lacking the present-day volcanic activity of either Heard or Bouvetøya, Macquarie experiences substantial faulting that influences the form of this ice-free island (Crohn, 1986). Equally, Macquarie Island lacks the height given to many other sub-Antarctic islands by means of their volcanoes and there is no compelling evidence that this has changed significantly any time during the Quaternary (Ledingham and Peterson, 1984). Thus, at the present time Macquarie is ice-free and there is now growing evidence that this may have been the situation throughout the Quaternary. A number of early studies suggest that there was extensive ice cover (e.g. Löffler and Sullivan (1980) plus a substantial number of glacially related forms (Colhoun and Goede, 1974) but that the ice did not extend, to any significant extent if at all, off-shore. However, there were always ambiguities in these studies, notably that no glacial imprint was observed above 200-250 m a.s.l. but, with the premiss that the island *must* have been glaciated so explanations were always created. Ledingham and Peterson (1984) reviewed the tectonic history of the island and suggest that uplift was recent and that features considered to be glacial in origin were, in fact, incorrectly identified with the misinterpretations based on the assumption of glaciation having taken place. They suggest that, although more data, particularly dates, are required, the evidence suggests that throughout the Quaternary Macquarie Island was simply too low to take advantage of the depressed snow lines and the increased precipitation in the form of snow to create widespread glaciation. Only a few small, localised glaciers were able to develop in opportune places where sufficient snow could accumulate in protected sites (e.g. on the south side of Boot Hill). Thus, effectively, evidence at this time suggests that Macquarie Island was not glaciated to any significant extent at any time during the Quaternary.

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Îles Crozet (46°-46°30'S, 50°30'-52°30'E)

This is a group of five islands, with a total area of 233 km^2, of volcanic origin reaching to a highest point of 934 m. At the present time there is no permanent snow or ice (Walton, 1985). As with Heard and Bovetøya, the Quaternary volcanic history may have greatly affected glaciation in a variety of ways, notably whether, like Macquarie, there were elevations conducive to snow accumulation. Although Bellair (1964) suggest that there are no traces of recent glaciation, Bougere (1992, Fig. 15, p.32) provides information regarding moraines, drumlins, cirques, glacial valleys and roches moutonnées. Bellair (1964) also notes that some valleys appear to have a 'glacial shape' but, if of glacial origin, must be a result of an event early in the Quaternary. Bougere (1992) also provides evidence for a variety of periglacial forms, notably 'cryoplanation terraces' which, by their size, may have taken some length of time to form. An example is cited for Plateau Jeannel where a terrace occurs at an altitude of 550 m a.s.l. but an extensive lichen cover on the rock debris suggests it is now a fossil form. If this is correctly interpreted then, in the absence of cold-based (protective) ice, it would suggest that glaciation was not widespread, even at this altitude, during the last glacial and thus, as Bellair (1964) has argued, significant glaciation must have been earlier in the Quaternary. Unfortunately, no reconstructions are available in this (relatively) little researched group of islands.

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Îles Kerguelen (48°27'-49°58'S, 68°25'-70°35'E)

An archipelago of 300 small islands and one major island, Grand Terre (5799 km^2), that are of volcanic origin and on which some fumarole activity still takes place. Grand Terre, which presently supports an ice cover of *c.* 750 km^2, primarily on the Cook ice cap and its associated outlet glaciers (*c.* 500 km^2) while there are several other ice caps and glaciers, notably around Mont Ross (see Mercer, 1967) in the west; Mercer (1967) provides details regarding recent glacial changes for Îles Kerguelen. Although the gross topography of Grand Terre predates the Quaternary (Hall, 1990), significantly Grand Terre experienced volcanism in the early Quaternary that resulted in a number of volcanic cones (Nougier, 1972). Tertiary structural lineaments were exploited by Quaternary glaciation (Nougier, 1970) but there is controversy regarding the character of glaciation through the Quaternary (see Hall, 1990 for a discussion). It appears that during the early Quaternary the whole of Grand Terre, at least as it was then, was ice covered (Nougier, 1970). However, the effects of the most recent glaciation have eradicated much of the evidence for earlier events (Nougier, 1970) even though it did not, it is thought (Nougier, 1970), completely cover the island even though, in some instances the ice was thought to have extended below present sea level (Bellair, 1965). The lack of tills in some areas and the absence of isostatic rebound noted by Bellair (1965) also argues the case for incomplete ice cover during the last glacial; some evidence provided by Hall (1984) does indicate that the glaciation may have been somewhat more widespread than originally thought but it does not contradict the main arguments. Thus it seems that there may have been a some major centres of ice growth in the higher areas to the west with a number of locations of only cirque glacier growth. A key component in the consideration of Grand Terre is the localised changes to topography induced by volcanic and tectonic history through the Quaternary and the influences that this may have had on both glaciation and preservation of earlier glacial imprints. A further issue is rigorous dating of volcanic, tectonic, and glacial events to facilitate meaningful reconstruction of ice distribution - in the absence of this sort of information it is difficult to infer the character of ice distribution variability through the Quaternary.

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Marion and Prince Edward Islands (46°54'S, 37°45'E)

Both Marion and Prince Edward are volcanic islands, but the only available information regarding glaciation pertains to the larger of the two, Marion Island; it is possible that Prince Edward Island, like Macquarie, was too low to sustain an ice cover. Roughly 290 km^2, Marion Island rises to a central volcanic complex at 1230 m a.s.l. where there is a small area (2-3 km$^{2)}$ of permanent snow and ice. Volcanism has been active throughout the Quaternary (McDougall, 1971) through to the present (Verwoerd, *et al.*, 1981) and has likely changed the physiography of the island during this time. Equally, much of the volcanism has been associated with faulting, some of which has been in response to isostatic adjustment resulting from rapid deglaciation; the island exhibits horst and graben structures with volcanoes aligned along radial and peripheral, step faults (Hall, 1982). Lava outpourings, associated with volcanicity resulting from ice retreat at the end of the last glacial, have removed some evidence of ice cover predating that time. It is suggested that evidence exists for at least three major glacials on

Marion Island (Hall, 1978, 1980, 1983) but it is not possible to reconstruct the extent of any but the most recent although, where sedimentary evidence is found it clearly shows ice extended beyond the present coastline at that point. Information from striations, moraines and till sequences allow for the reconstruction of ice cover for parts of Marion during the last glacial and it appears that ice extended a short distance beyond the present coastline (Hall, 1983).

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The Falkland Islands (51-52°30'S, 60-61°30'W)

The Falklands consist of two main islands (East Falkland: *c.*5000 km^2; West

Falkland: 3,500 km^2) plus an additional 250 smaller islands. The snowline is presently about 500 m above the highest summits (Roberts, 1984) with a mean annual winter temperature 2 to 2.5°C (Moore, 1968) and relatively low precipitation (600 mm at sea level) as a result of being in the 'rain shadow' of South America. Peaks are not high (<750 m a.s.l.) and so the Falklands, despite their southerly latitude, responded in a similar fashion to that of Macquarie Island and experienced only limited glaciation; the impact of low mountains was further exacerbated by the rain shadow effect. Clapperton (1990) provides a detailed account regarding the knowledge of glaciation of the Falklands although, following Roberts (1984) there may be some confusion regarding the distinction between "nivation cirques" and "glacial cirques". Even if these are both considered to have, at some time, some glacial component, then there are still only a maximum of 27 small cirque glaciers at the height of the Last Glacial Maximum. Ice existed exclusively around the three highest massifs: Mt. Adam- Mt. Robinson and Mt. maria on West Falkland and Mt. Usborne on East Falkland (Clapperton, 1990).

During the Quaternary it would appear that there are at least two intervals of cirque glaciation - partly evidenced by cirque-within-cirque growth (Clapperton, 1990). Roberts (1984) argues that although there may have been two or three episodes of glacial activity there impact was not particularly large in terms of erosional or depositional landform development. Periglacial activity and peat growth were far more significant that glacial action in the Falklands (Clapperton, 1990; Hall, In Press).

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South Georgia (54-55°S, 36-38°W)

This 300 km long but only 5 - 40 km wide island presently supports a substantial ice cover (about 58% according to Clapperton, 1990) with large glaciers descending to sea level at a number of locations (e.g. Moraine Fjord). Situated within the track of precipitation-bearing cyclonic depressions and with the high (reaching to 2960 m at Mt. Paget) spine of mountains along the length of the narrow island, so it is an ideal location for glacier growth. This is so much so that during the Last Glacial Maximum evidence (Birnie, 1978; Clapperton and Sugden, 1980; Clapperton, *et al.*, 1989a & b) indicates that "... a continuous ice cap centred over the axial mountain range extends across the surrounding shelf as global temperature and sea level fall" (Clapperton, 1990, p.234). Outflowing ice from the major fjords have cut glacial troughs almost to the outer edge of the continental shelf while the inter-trough areas have also experienced extensive glacial action (Clapperton, 1990). Sugden and Clapperton (1977) suggest that the outer edge of the ice cap was close to the edge of the continental shelf at a (present) depth of *c.* 200 m. However, even during maximum glaciation, some peaks above 1500 m were exposed as nunataks. Clapperton (1990, Table 2) shows five identified glacial stages for South Georgia since about 18000 BP. Evidence for earlier events, that likely mirrored those of the last glacial, are not yet identified and likely, considering the extent of the Last Glacial Maximum, may not have survived. This does, though, suggest that major glacial erosional forms such as the fjords and troughs across the shelf are the product of multiple events. Detail regarding the fluctuations, since 18000 BP, of individual glaciers is clearly described by Clapperton (1990).

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Glaciation in southern Africa

In many ways the higher areas of southern Africa (Lesotho, the Natal Drakensberg and parts of the Eastern Cape mountains) appear highly likely candidates for Quaternary glaciation. Present day winter temperatures can be severely cold (temperatures frequently down to -15°C or lower) and, when it does snow, falls can be substantial. The problem regarding glaciation is not one of cold but rather one of precipitation. Although the Eastern Cape mountains fall within the winter rainfall region, and, as a consequence, experience snowfall (with an average of 31 days yr^{-1} of snow cover) the other areas all fall within the summer rainfall region and thus snowfall is rare. Even though Quaternary temperatures were several degrees (3-6°C) colder than present and the lower sea levels would also have made for even colder temperatures at higher elevations (Talma and Vogel, 1992), large sections (although not all) of the last glacial were much drier than present (Partridge, *et al.*, 1990; Dollar, 1998). As discussed by Dollar and Goudie (2000, p. 47) the period 40,000 to 30,000 BP was 3-4°C colder than present but may have been wetter except in the east (this is significant as most of the hypothesised glaciers were in the east - see below). From 30,000 to 21,000 BP temperatures were 4-5°C cooler and conditions drier in the east and then from 21,000 to 17,000 BP (incl. the Last Glacial Maximum) the whole area was 5-6°C cooler than present but also significantly drier; sea level was *c.* 130 m lower than present. Warmer and wetter conditions prevailed from 17,000 to 12,000 BP and this, in broad terms, continues through to the present. Thus, throughout the Quaternary, the issue with respect to glaciation in southern Africa is not one of temperature but rather of available snowfall to accumulate into ice. Certainly some areas, parts of highland Lesotho for example, would be ideal areas for snow accumulation as they have plateau surfaces but evidence to support wetter (i.e. snowier) conditions seems to be in question.

Another way to approach the problem is by means of finding direct evidence of glacial activity in the form of moraines, tills, or glacial erosional features, or, as a corollary, the finding of unequivocal evidence of periglacial landforms that, by their presence, would preclude glaciation. A major attribute to these directions is that of dated sediments or landforms to locate them firmly within the Quaternary temporal framework. It is at this point that problems arise. Some authors (e.g. Lewis, 1996) identify what they believe are striated rock surfaces, moraines, fluvio-glacial sediments, cirques, and 'trough-like valleys' (see Table 13.1 of Sumner and Meiklejohn, 2000). This information has led to speculation that plateau glaciers, cirque glaciers, and/or niche glaciers all may have occurred at the time of the Last Glacial maximum (Borchert and Sänger, 1981; Sänger, 1988; Lewis, 1996; Hall, 1994). Lewis (1996) has also suggested that during a very cold period prior to the Last Glacial Maximum that valley glaciers existed in the Eastern Cape Drakensberg. The problem is that the identification of these features is highly questionable plus they are difficult to relate to the palaeo-climatic information. For example, the 'cold period' recognised by Lewis (1996) requires a temperature drop at *c.*40,000 BP in the region of 19°C and this is contrary to all other available palaeo-climate evidence (e.g. see Partridge, 1997). Equally, for much of the Lesotho/Drakensberg area, even if there were glaciers little evidence would have been generated as there are no peaks to stand above the ice as nunataks and much of the ice would have fallen over the edge of the Great Escarpment thereby leaving no erosional or depositional signatures. Thus, at present, the broader consensus is that southern Africa was never glaciated during the Quaternary (Tyson, 1986; Preston-Whyte and Tyson, 1988; Deacon and Lancaster, 1988, DeVilliers, 2000).

The likelihood is that many of the features ascribed to glaciation are, in fact, either completely misinterpreted or are the products of non-glacial cold weather processes; cold not being the limiting factor but rather precipitation.

Sänger (1988) identified, from air photos only, a number of features he ascribed to glaciation. Subsequent field investigation of the 'moraines' showed some to be erosional remnants of bedrock while 'outwash fans' lack any evidence of a glacial origin, and striated pebbles observed at higher elevations are products from the Pakhuis tillite of Silurian age rather than Quaternary events. Equally, many of the sedimentary sequences described by Lewis (1996), and assumed to be of glacial origin, lack any definitive evidence of a glacial signal and can be explained by other processes (e.g. periglacial mass movement). Lastly form is not a good indicator of process in regard to glacial troughs. Where glaciation is assumed so form can be directed towards a glacial origin but the argument is circular - in the absence of the assumption of glaciation is there anything that unequivocally indicates that the form is glacial in origin? If not, then the case is both circular and unsubstantiated. Indeed, Lewis (1996, p. 113) himself uses phrases such as "trough-like" and so it is more a case of trying to fit evidence to an hypothesis rather than actually testing an hypothesis. In respect to the gulleys and their associated deposits, anastomosing channels, and levees observed along the face of the Great Escarpment of the Drakensberg it has been argued that these may reflect the presence of niche glaciers (Hall, 1994). However, this too is unsubstantiated and needs some clear indicator of the former presence of ice (Sumner, 1995).

The maps presented here *must* be viewed in the context of their unproven and highly controversial background. No detail can be given as to glacier distribution as, simply, none exists. Rather, these maps show the generalised location of where glaciers may have existed in southern Africa. The present conclusion is, though, that in the absence of any irrefutable evidence to date, it is probable that no significant ice cover existed although small, short-lived, cirque or niche type ice may have occurred in highly favourable locations.

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OBSERVATIONS ON THE RELATIONSHIP BETWEEN CLAST SIZE, SHAPE, AND LITHOLOGY FROM THE PERMO-CARBONIFEROUS GLACIOGENIC DWYKA FORMATION IN THE WESTERN PART OF THE KAROO BASIN

by

K.J. HALL and J.N.J. VISSER

ABSTRACT

Clasts from platform and valley facies provinces of tillite in southern Africa were considered with respect to their size–shape–lithology relationships. Generalized shape values, with respect to lithology, indicated a difference in shape, particularly roundness, between the two facies. A lithologic control of clast size was noted and size–shape relationships were seen to vary greatly with lithology. A variety of glacial transport paths and the added effects of reworking further complicate analysis, particularly in the platform facies where the final deposit may be the product of several glacial stages plus interglacial subaerial processes. At one site a knowledge of the sediment genesis was needed to help interpret the shape indices. Detailed analysis of clast size and shape with respect to lithology, although time consuming, is considered a useful tool in the investigation of tillites and far more work is needed on the glaciogenic sediments of southern Africa. The results show that, with respect to previous studies on Dwyka sediments, the suggested qualitative size–shape relationships are not valid and that much more rigorous approaches are needed in future.

CONTENTS

I. INTRODUCTION

Studies of clast shapes from the Permo-Carboniferous Dwyka tillite of southern Africa (e.g. Bond, 1981; Martin, 1981a, b; Stratten, 1968; Von Brunn and Stratten, 1981) have been highly qualitative in approach. Shapes were visually appraised, in most instances, without reference to size constraints, cognizance of lithology, or stratigraphic controls. Despite these inadequacies a number of size–shape relationships, notably an increase in rounding with an increase in clast size, have been suggested as pertinent to the whole of the Dwyka Formation (Stratten, 1968; Tankard *et al.*, 1982). However, a recent study (Hall, 1983) in which clast shapes were quantified (but again without reference to lithology or stratigraphy) showed that, in the main, these size–shape generalizations were not valid.

Clast shapes are relatively easily quantified, but much care must be taken when attributing causes (Barrett, 1980). However, a number of recent studies (e.g. Boulton, 1978; Domack *et al.*, 1980; Humlum, 1981; Ballantyne, 1982; Sharp, 1982) have shown that clast shape may reflect transport history. Whilst shape can, with great care, be visually estimated (Olsen, 1983) it is safer, particularly if assessing size–shape relationships, to attempt a quantified approach. To this end, detailed observations were undertaken in the western Karoo during a study of boulder pavements (Visser and Hall, 1984 and in press) and results pertaining to size–shape–lithology relationships were obtained.

II. APPROACH

At each sample point as many clasts as possible were collected (this being constrained by time and the degree of lithification of the tillite), in a random fashion. The minimum clast size measured was 10 mm (a-axis) for it was considered impractical to accurately measure material smaller than this in the field. Randomness was achieved by throwing a hammer, sitting where it landed and then observing clasts within approximately 1 m^2 of that point, or by selecting a fixed area on a face and taking all clasts within the boundaries of that area. The following parameters were then noted: the lengths of the orthogonal a-, b-, and c-axes, the radius of the circle fitting the sharpest corner and that of the largest inscribed circle, the lithology, and whether the clast was striated and faceted. From this it was possible, following the recommendations of Barrett (1980), to calculate for each clast Cailleux's (1945) flatness index, the Modified Wentworth roundness (Dobkins and Folk, 1970), the $\bar{O}$ $\bar{P}$ index (the oblate–prolate index of Dobkins and Folk, 1970), Sneed and Folk's (1958) maximum projection sphericity, and the Elongation index (Lister, 1981). Thus, the following information could be obtained for each sample:

1. the percentage of each lithology,
2. the mean (and standard deviation) a-axis of each lithology,
3. the means (and standard deviations) of roundness, flatness, and sphericity for each lithology,
4. the percentage of striated clasts,
5. the mean number of facets,
6. correlation between a-axis and shape indices,
7. the percentage of discs, blades, and rods (plus the mean and standard deviation of each), and
8. the mean elongation.

In addition, the range of straight line travel distances from rock source to the depositional site for some clast associations was based on the presence of indicator rocks. The data allowed consideration of total-sample size–shape relationships, lithology-constrained size–shape relationships, size and shape related to possible travel distance and genetic facies, lithological control of the clast size spectrum, percentage of striated and faceted clasts, and between-sample comparisons. Analysis of these results gave, in the light of earlier Dwyka studies, a more detailed picture of clast size–shape–lithology relationships.

III. STUDY AREA

Visser (1983, pp. 678–679) subdivided the Dwyka Formation into a valley and a platform facies association. The valley facies association is characterized by a variable thickness and lithology, and proximally derived clasts. The platform facies association consists of a great thickness of predominantly massive diamictite and distantly derived clasts.

Data on size, shape, and lithology were collected from each of the facies associations in areas where clasts could be easily removed from the bedrock. The valley facies was sampled in the Prieska area where the Dwyka Formation attains a thickness of 70 to 80 m in a south-east-striking valley on the farm Fonteintjie (Fig. 1A). The lower third of the sequence consists of clast-rich diamictite with interbedded water-reworked and conglomeratic sections. Clasts with a diameter of up to 3 m were recorded. The upper part of the sequence consists predominantly of faintly bedded argillaceous diamictite with three thin interbedded boulder beds, one of which represents a boulder pavement. Samples were taken on the farm Fonteintjie from the upper diamictite, about 40 m above the base of the sequence, and from the Stofbakkies Quarry on the boundary between the farms Brakpoort and Stofbakkies (Fig. 1A). The quarry is located in faintly bedded argillaceous diamictite near the top of the Dwyka Formation.

The platform facies association was sampled in the vicinity of Elandsvlei (Fig. 1B) where it consists of about 585 m of predominantly massive diamictite with locally developed sandstone bodies which are deformed and shale units up to 25 m thick. Faintly bedded diamictite and clast-rich massive diamictite alternate towards the top of the sequence. Clasts of up to 2 m in length were recorded in the diamictite. Samples were taken near a boulder pavement at Kruitfontein (about 18 m above the base of the sequence) and at the Drift boulder bed site (about 520 m above the base).

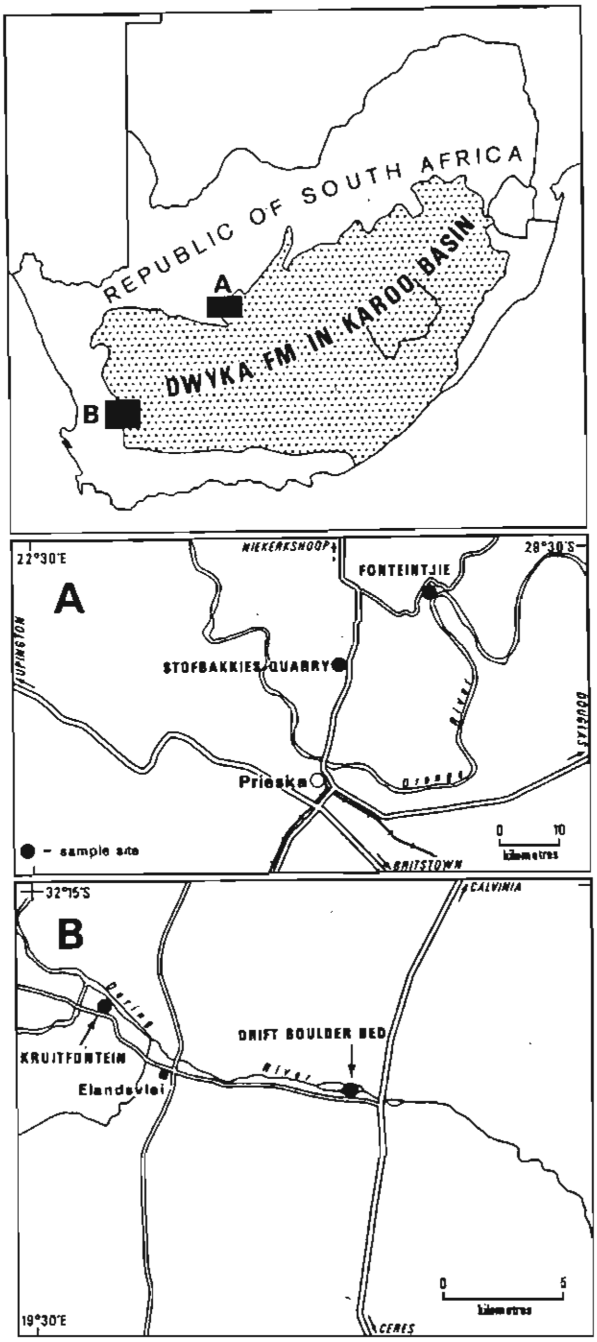

Figure 1
Locality map of the sample sites in the western part of the Karoo Basin. A = sites where valley facies association was sampled. B = sites where platform facies association was sampled.

IV. RESULTS

Table I shows shape indices, together with percentages of striated clasts and their mean number of facets, for the six study sites, irrespective of lithology. In Tables II and III there is a breakdown of shape indices by lithology together with mean sizes and correlation of shape index with a-axis length. Figure 2 shows the correlations of shape indices against size for the various lithologies in a graphical form. Table IV gives the percentage of striations and mean number of facets for the clasts in each lithology whilst, finally, Table V indicates the travel distances possible for the various lithologies. Where observations of any particular lithology were two or less percentages or resultant indices are shown in parenthesis or with the n-value indicated.

The information in Table I constitutes a generalized quantification of clast shape parameters without respect to lithology, size, or stratigraphic control. Noticeable, is the difference in roundness indices between the platform facies association of the Elandsvlei sites (higher values) and the valley facies of the Prieska sites (lower values). This format gives far more detail than the subjective observations of previous studies and yet hides within it many within-sample variations which make interpretation and the generation of size–shape relationships extremely hazardous. Consideration of Table II indicates that the three shape indices shown and clast size vary, sometimes substantially, as a product of lithology. For instance, when considering samples of three or more clasts then at Fonteintjie the mean roundness of dolomite and quartzite clasts is twice that of banded iron-formation (BIF), and the mean a-axis of the lavas is 41,94 per cent greater than that of the quartzite. Again, within-sample clast size variation is clearly seen at the Drift boulder bed (boulders from boulder bed) where 25 per cent of the clasts (dolomite) are between 54 and 41 per cent larger than the bulk of the sample. At other localities (e.g. Stofbakkies Quarry) there is less variation with respect to roundness (indices from 0,23 to 0,25), but a relatively large variation in flatness (values from 173,91 to 269,58).

Consideration of the percentage of blades, discs, and rods by lithology, as generated by the oblate–prolate index

TABLE I
Values for Total Data, Irrespective of Lithology

	ELANDSVLEI AREA Platform Facies Association: Drift Boulder Bed. Clast from Matrix*	ELANDSVLEI AREA Platform Facies Association: Drift Boulder Bed Boulders	PRIESKA AREA Valley Facies: Kruitfontein	PRIESKA AREA Valley Facies: Fonteintjie	PRIESKA AREA Valley Facies: Stofbakkies Quarry	PRIESKA AREA Valley Facies: Stofbakkies Quarry Boulders Only
% Blades	0	6,25	0	0	1,96	0
% Discs	46,15	43,75	56,25	34,0	47,06	30,77
% Rods	53,85	50,0	43,75	66,0	50,98	69,23
x̄ Disc	−3,0	−3,31	−4,53	−4,53	−4,48	−0,91
	(s=2,37)	(s=3,05)	(s=2,29)	(s=4,81)	(s=3,92)	(s=0,78)
x̄ Rod	+5,07	+5,81	+3,20	+3,84	+3,53	+4,49
	(s=5,07)	(s=3,54)	(s=2,06)	(s=2,72)	(s=2,48)	(s=2,45)
x̄ Flatness	186,46	164,15	184,8	204,58	185,95	166,25
	(s=40,94)	(s=30,3)	(s=31,72)	(s=64,37)	(s=55,17)	(s=15,57)
x̄ Roundness	0,34	0,44	0,48	0,28	0,23	?
	(s=0,18)	(s=0,15)	(s=0,2)	(s=0,20)	(s=0,16)	
x̄ Sphericity	0,68	0,73	0,69	0,66	0,69	0,73
	(s=0,08)	(s=0,08)	(s=0,09)	(s=0,11)	(s=0,10)	(s=0,05)
x̄ Elongation	0,71	0,73	0,78	0,70	0,75	0,69
	(s=0,13)	(s=0,16)	(s=0,1)	(s=0,12)	(s=0,11)	(s=0,09)
% Striated	30,77	25,0	81,25	34,0	27,45	76,92
x̄ Facets	0,58	0,31	1,63	1,20	1,33	?
	(s=0,58)	(s=0,79)	(s=0,81)	(s=0,86)	(s=0,91)	
n	26	16	16	50	51	13

* At the Drift boulder bed all stones smaller than approximately 0.1 m (a-axis) were taken as part of the matrix.

TABLE II
Clast Shape Indices with Respect to Lithology

Lithology	%	(cm) x̄ a-axis	x̄ Roundness	r Roundness vs a-axis	x̄ Sphericity	r Sphericity vs a-axis	x̄ Flatness	r Flatness vs a-axis	n
PRIESKA AREA									
Fonteintjie									
Dolomite	44	6,27	0,34	−0,37	0,67	−0,09	197,17	+0,14	22
		(s=2,2)	(s=0,2)		(s=0,11)		(s=69,7)		
Lava	14	7,13	0,31	+0,34	0,62	+0,17	214,86	−0,14	7
		(s=3,8)	(s=0,29)		(s=0,09)		(s=43,13)		
Chert	14	4,67	0,23	+0,76	0,71	−0,29	180,31	+0,19	7
		(s=1,35)	(s=0,18)		(s=0,10)		(s=45,41)		
BIF*	14	5,76	0,15	−0,25	0,57	−0,64	256,11	+0,75	7
		(s=2,88)	(s=0,08)		(s=0,13)		(s=82,67)		
Quartzite	10	4,14	0,34	−0,98	0,69	+0,56	182,37	−0,54	5
		(s=1,17)	(s=0,18)		(s=0,09)		(s=38,95)		
Jasper	2	[4,0]	[0,06]	—	[0,70]	—	[170,45]	—	1
Granite	2	[6,5]	[0,08]	—	[0,55]	—	[250,00]		1
Stofbakkies Quarry									
Lava	45,10	5,97	0,23	−0,20	0,67	−0,27	191,16	+0,32	23
		(s=4,54)	(s=0,16)		(s=0,10)		(s=43,04)		
Dolomite	35,29	5,63	0,22	−0,25	0,70	+0,48	173,91	−0,44	18
		(s=3,65)	(s=0,12)		(s=0,06)		(s=21,75)		
Quartz Porphyry	5,88	7,50	0,25	−0,59	0,60	−0,98	269,58	+1,00	3
		(s=6,64)	(s=0,34)		(s=0,23)		(s=181,39)		
Diabase	3,92	[3,95]	[0,09]	—	[0,61]	—	[221,47]	—	2
Quartzite	3,92	[12,2]	[0,51]	—	[0,82]	—	[141,51]		2
Chert	1,96	[3,4]	[0,08]	—	[0,79]	—	[142,86]	—	1
Gneiss	1,96	[10,5]	[0,28]	—	[0,88]	—	[121,71]	—	1
BIF*	1,96	[3,2]	[0,33]	—	[0,63]	—	[208,33]	—	1
Stofbakkies Quarry: Only Boulders									
Dolomite	53,85	75,14	—	—	0,73	+0,10	164,99	−0,12	7
		(s=24,56)			(s=0,066)		(s=22,75)		
Lava	38,46	80,80	—	—	0,73	−0,29	166,95	+0,17	5
		(s=9,63)			(s=0,042)		(s=15,01)		
Quartzite	7,69	[81,0]	—	—	[0,71]	—	[170,51]	—	1

* BIF— Banded iron-formation.

(Dobkins and Folk, 1970) (Table III) shows a great variation about the generalized values of Table I. For instance, at the Drift boulder bed (clasts from matrix) Table I shows an average of 46,15 per cent discs and 53,85 per cent rods whilst the details indicate the dolomite and the diabase plus norite to have far greater percentages of discs than rods. At the Drift boulder bed (boulders from the boulder bed) the lava and the diabase plus norite have greater percentages of discs than does the dolomite and yet the averaged values indicate almost parity between the two

TABLE III
Blades, Discs and Rods with Respect to Lithology

Lithology	% Blades	% Discs	% Rods	$\bar{x}$ Disc	$\bar{x}$ Rod	Mx Disc	Mx Rod	Mn Disc	Mn Rod
ELANDSVLEI AREA									
Drift Boulder Bed: Clasts from Matrix									
Quartzite	0	27,27	72,73	−5,72 (s=2,87)	+4,91 (s=2,86)	−8,52	+8,16	−2,87	+0,40
Lava	0	37,50	62,50	−1,94 (s=1,94)	+4,85 (s=3,74)	−4,10	+8,45	−0,37	+0,34
Dolomite	0	66,67	33,33	n=2 −2,61	n=1 [7,42]	−4,09	—	−1,12	—
Diabase + Norite	0	100	0	−2,33 (s=0,9)	—	−3,0	—	−1,31	—
Gneiss	—	n=1 [100]	—	[−0,92]	—	—	—		—
Drift Boulder Bed: Boulders from Boulder Bed									
Diabase + Norite	14,29	42,86	42,86	−3,07 (s=2,96)	+3,76 (s=3,49)	−6,48	+7,70	−1,16	+1,04
Dolomite	0	25,0	75,0	n=1 −0,94	+8,80 (s=2,91)	—	+12,15	—	+6,91
Lava	0	66,67	n=1 33,3	n=2 −5,48	+3,21	−8,41	—	−2,55	—
Quartzite	0	n=1 50	n=1 50	−0,81	+5,61	—	—	—	—
KRUITFONTEIN									
Quartzite	0	75,0	25,0	−4,95 (s=2,66)	n=2 +3,56	−9,33	+5,54	−2,12	+1,58
Dolomite	0	n=1 50	n=1 50	[−4,55]	[+1,46]	—	—	—	—
Diabase	0	n=1 50	n=1 50	[−2,25]	[+6,64]	—	—	—	—
Lava	0	0	n=1 100	—	[+2,55]	—	—	—	—
Chert	0	0	n=1 100	—	[+2,81]	—	—	—	—
Gritstone	0	n=1 100	0	[−4,23]	—	—	—	—	—
Gneiss	0	0	n=1 100	—	[+1,83]	—	—	—	—
PRIESKA AREA									
Fonteintjie									
Dolomite	0	27,27	72,73	−4,71 (s=7,0)	+3,80 (s=2,46)	−18,68	+8,16	−0,63	+0,36
Lava	0	42,86	57,14	−2,98 (s=2,25)	+4,91 (s=3,46)	−5,0	+9,09	−0,55	+0,61
Chert	0	28,57	71,43	n=2 −5,03 (s=3,03)	+5,14 (s=3,94)	−7,17	+9,45	−2,89	+0,22
BIF*	0	14,29	85,71	n=1 [−12,90]	+2,21 (s=1,84)	—	+5,0	—	+0,43
Quartzite	0	60,0	40,0	−2,93 (s=3,16)	n=2 +3,64 (s=1,61)	−6,50	+4,93	−0,41	+2,50
Jasper	0	n=1 100,0	0	[−4,04]	—	—	—	—	—
Granite	0	n=1 100,0	0	[−4,04]	—	—	—	—	—
Stofbakkies Quarry									
Lava	4,35	56,52	39,13	−4,08 (s=2,87)	+2,94 (s=2,49)	−11,36	+6,06	−1,56	+0,24
Dolomite	0	44,44	55,56	−3,56 (s=2,59)	+3,46 (s=2,88)	−8,67	+7,92	−0,71	+0,50
Quartz Porphyry	0	n=1 33,33	66,67	n=1 [−1,56]	n=2 −11,44 (s=10,01)	−18,51	—	−4,36	—
Diabase	0	0	n=2 100,0	—	n=2 +3,64 (s=2,56)	—	+5,45	—	+1,83
Quartzite	0	0	n=2 100,0	—	n=2 +5,04 (s=3,51)	—	+7,52	—	+2,55
			n=1		n=1				

(Continued)

TABLE III (continued)

Lithology	% Blades	% Discs	% Rods	x̄ Disc	x̄ Rod	Mx Disc	Mx Rod	Mn Disc	Mn Rod
Chert	0	0	100,0	—	[+1,87]	—	—	—	—
			n=1		n=1				
Gneiss	0	0	100,0	—	[+5,00]	—	—	—	—
			n=1		n=1				
BIF*	0	0	100,0	—	[+5,00]		—	—	—
Stofbakkies Quarry: Only Boulders									
Dolomite	0	42,86	57,14	−1,3 (s=0,79)	+3,84 (s=3,16)	−1,94	+7,12	−0,37	+0,59
				n=1					
Lava	0	20,0	80,0	[−0,25]	+5,28 (s=2,14)	—	+7,74	—	+2,99
			n=1						
Quartzite	0	0	[100,0]	—	3,96	—	—	—	—

*BIF = Banded iron-formation Mx = Maximum value Mn = Minimum value

TABLE IV
Striating and Faceting of Clasts with Respect to Lithology

Lithology	n	% Striated	x̄ Facets
PRIESKA AREA			
Fonteintjie			
Dolomite	22	45,45	1,55 (s=0,8)
Lava	7	57,14	1,45 (s=0,79)
Chert	7	0	0,57 (s=0,52)
BIF*	7	28,57	0,57 (s=0,53)
Quartzite	5	20,0	1,40 (s=1,14)
Jasper	1	0	—
Stofbakkies Quarry			
Lava	23	30,43	1,43 (s=0,99)
Dolomite	18	22,22	1,39 (s=0,78)
Quartz Porphyry	3	0	1,0 (s=1,0)
Diabase	2	0	—
Quartzite	2	50,0	—
Gneiss	1	100,0	—
BIF	1	0	—
Stofbakkies Quarry: Only Boulders			
Lava	5	100,0	Not measured
Quartzite	1	0	Not measured
Dolomite	7	71,43	Not measured
ELANDSVLEI AREA			
Drift Boulder Bed: Clasts from Matrix of Boulder Bed			
Quartzite	11	27,27	0,55 (s=0,52)
Lava	8	37,50	0,75 (s=0,71)
Dolomite	3	0	0,33 (s=0,58)
Diabase + Norite	3	66,67	0,67 (s=0,58)
Gneiss	1	0	—
Drift Boulder Bed: Boulders from Boulder Bed			
Diabase + Norite	7	28,57	0,29 (s=0,49)
Dolomite	4	0	—
Lava	3	66,67	1,0 (s=1,73)
Quartzite	2	0	—
Kruitfontein			
Quartzite	8	87,50	1,63 (s=0,52)
Dolomite	2	0	1,50 (s=0,71)
Diabase	2	100,0	2,0 (s=1,41)
Lava	1	100,0	—
Chert	1	0	—
Gritstone	1	100,0	—
Gneiss	1	100,0	—

*BIF = Banded iron-formation

shapes. Table IV indicates that even the percentage of striated clasts and the mean number of facets can vary as a function of lithology.

Table II and Fig. 2 clearly show that correlation of clast shape with clast size fails to produce the size–shape relationship suggested by earlier workers. With respect to roundness there appears, if anything, to be a decrease in roundness with an increase in a-axis length. But, again, this varies from lithology to lithology with, at Fonteintjie, chert and lava showing positive relationships (+0,34 and +0,76) whilst those of dolomite, BIF and quartzite are negative (−0,37, −0,25 and −0,98). This size–shape response is not lithologically constant either, for the negative quartzite value at Fonteintjie (−0,98) is opposed by positive values at the Drift boulder bed (+0,40) and Kruitfontein (+0,48). Other lithologies show similar between-site size–shape variations.

V. DISCUSSION

Where till has a multi-lithological clast component, particularly if a range of travel distances are involved, then a polymodal–polydimensional shape frequency distribution could be expected. This, and any form of clast quantification, appears to have been ignored in previous studies of the Dwyka (Hall, 1983) and so a number of untested generalizations have evolved. At the same time, the potential of clast shapes for diagnostic purposes has been ignored despite the multitude of problems regarding glaciogenic sediment genesis and differentiation in the Dwyka.

With the large range of rock types present in a till all parts of the shape range will not be affected by abrasion at the same rate (Ehrlich *et al.*, 1980). For example, Drake (1972) showed that the more durable lithologies (e.g. quartzite) may travel relatively long distances without change, whereas Lister (1981) noted that within any one lithology all the clasts, irrespective of size, may be subject to rounding by the removal of asperities, but with little change in particle size save its slow reduction. These responses are well illustrated by this study. For instance, it has already been noted that there is a range of shape indices, when taken lithology by lithology, within the sample. Then, if we consider quartzite in Table II, and note its travel distances from Table IV, it can be seen that the further travelled (300–500 km) platform facies Drift boulder bed and Kruitfontein clasts have higher roundness indices (0,4 and 0,47) than the shorter travelled (0–50 km) valley facies Fonteintjie clasts (value 0,34). The lower (quartzite) values at Fonteintjie can be explained by the BIF bedrock which fractures into flat, angular slabs along its bedding planes and, due to the extremely short transport distance, is not subject to rounding. The shorter travelled clasts have a very strongly negative correlation of roundness with size ($r = -0{,}98$) whereas the further travelled material exhibits positive correlations ($r = +0{,}40$ and $+0{,}48$); the value of $-0{,}98$ is shown to be significant at the 0,05 probability level by the t-test of r but the two positive correlations are not.

Correlation of shape with size shows a number of other variations. For example, at the Drift boulder bed the

TABLE V
Approximate Travel Distances of the Various Lithologies Found at the Study Area

Locality	Lithology	Source Area	Approx. Transport Distance
Stofbakkies Quarry and Fonteintjie	Quartzite Dolomite Banded iron-formation Diabase Lava	Ghaap Plateau	0–50 km
	Ventersdorp lava Igneous intrusives	Areas to N, NE and E of Prieska Basin	100–200 km
Drift Boulder Bed and Kruitfontein	Dolomite Banded iron-formation Jasper Diabase Lava	Ghaap Plateau	450–500 km
	Quartzite Schist	Langeberg Range, Namaqualand Metamorphic Complex	300–550 km
	Gneiss Granite	Namaqualand Metamorphic Complex, Upington area	100–400 km

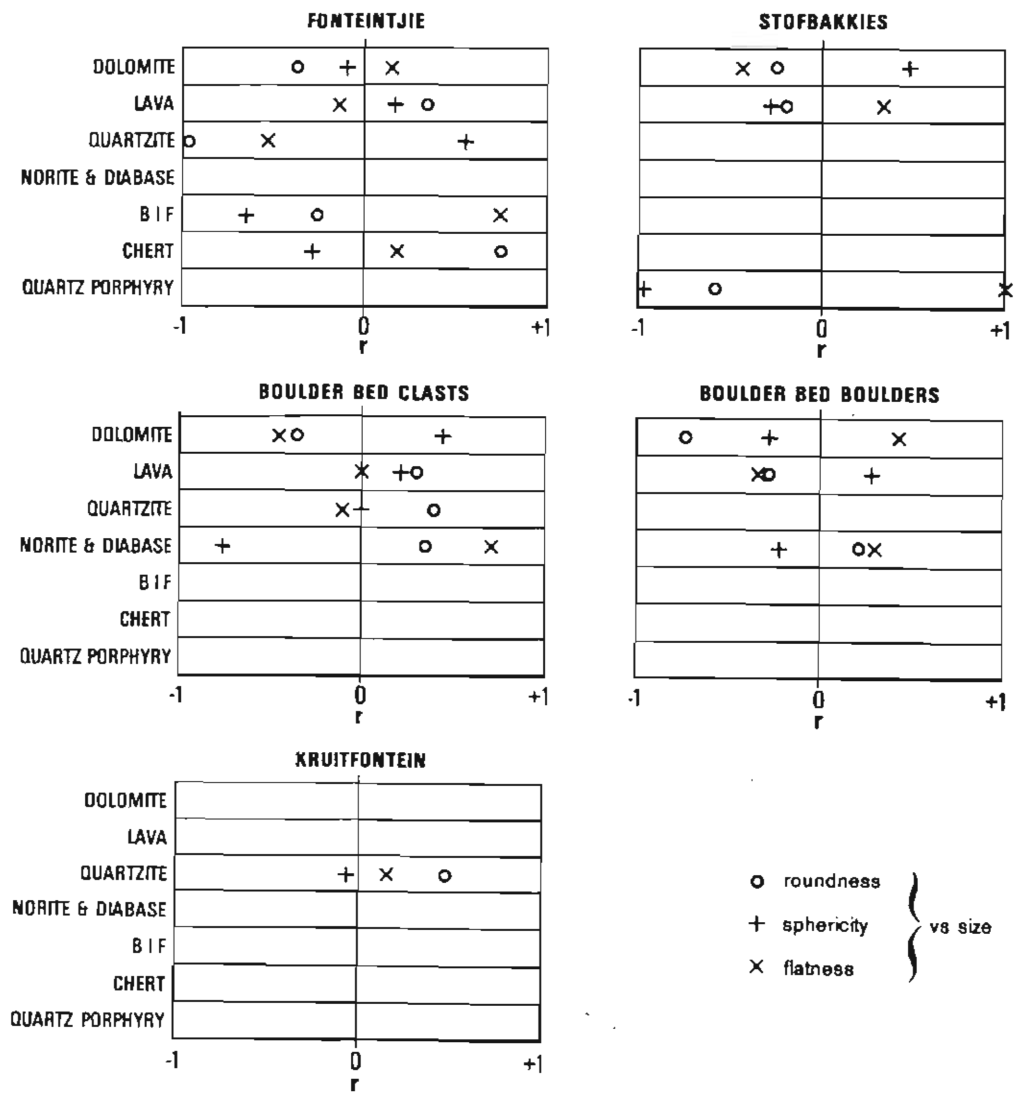

Figure 2
Graphical representation of the correlation between clast shape indices and size for the various lithologies at the study sites.

smaller lava clasts from the matrix show a positive (+0,30) correlation of size with roundness (Table II; Fig. 2) whilst the lava boulders exhibit exactly the opposite (r = −0,30). The other lithologies have correlations with the same signs, but to varying degrees (e.g. for dolomitic pebbles r = −0,35 whilst for boulders r = −0,74). A similar situation is found for sphericity, with dolomitic pebbles showing a positive correlation (r = +0,44), but boulders a negative one (r =

−0,29). For the same site, Table III shows the lava pebbles to be predominantly prolate (rod-like) whilst lava boulders are mainly oblate (disc-like). Conversely dolomitic pebbles are predominantly oblate whilst boulders are prolate. At the same time Table IV shows variation in the striating and faceting of material as a product of lithology. None of this is evident from the generalized details of Table I and yet, whilst the depth of detail is certainly somewhat confusing, it must reflect, at least in part, the transport history of the material.

It was thought that certain lithologies would, due to the physical properties of those rocks, respond in a similar manner with respect to resistance to comminution and that this would be shown by clast size. However, the data available (Table II) show that responses sometimes vary. The boulders and the clasts in the matrix at the Drift site are almost the antithesis of each other, with clasts from the matrix indicating (in terms of size) lava > quartzite > dolomite > diabase, whilst the boulders show dolomite > diabase>quartzite>lava. At Fonteintjie lava>dolomite> BIF > chert > quartzite is found, with a similar situation at Stofbakkies Quarry: quartz-porphyry > lava > dolomite. These observations partly reflect the relative complexity of transport paths and of the lithologies involved. In the Prieska area transport paths were relatively simple as only valley glaciation was involved, whereas in the Elandsvlei platform facies more complex paths with probable reworking by subsequent glacials could be expected. The quartz-porphyry is virtually indestructible, so that large clasts would be expected whilst the dolomite and lava were massive and thickly bedded thus giving relatively large clasts. The thinly bedded form of the BIF and chert, plus their brittle nature, gives small clasts. The Elandsvlei sites showed (Table II) a greater percentage of quartzite clasts than did the Prieska sites. This is thought to reflect the nature of the rock at the two source areas, with the Namaqualand origin of the Elandsvlei material having massive and thickly bedded quartzites which are able to withstand transport comminution better than the well-bedded and brittle quartzites available for the Prieska area. That the actual quartzite clast size is almost the same for the Prieska and the Elandsvlei regions does not contradict this hypothesis because the transport distance (Table V) for the Elandsveli material is as much as an order of magnitude further.

Consideration of the percentage lithologic composition at the Elandsvlei sites suggests that the dolomite and the diabase are more durable and hence they have the greatest component and largest size within the boulders, whilst the less durable lithologies constitute a smaller component and a-axis size. The greater amounts of quartzite and lavas in the clasts from the matrix, together with the observation that they are also the largest in size, may be due to them reaching some optimal size beyond which comminution is slower. This may reflect process variation in so far as the larger boulders of quartzite and lava may be prone to fracturing under subglacial stress situations or to weathering whilst carried as a supraglacial load, whereas the broken fragments are less prone to further breakdown: the converse applying to the other lithologies.

A possible explanation, at least in part, for the apparently unusual situation at Elandsvlei may be the derivation of the material. The Drift boulder bed is thought (Visser and Hall, 1984) to comprise a reworked beach deposit. Thus, the boulders were originally glacially transported, then abraded and rounded by high-energy wave action, and subsequently reworked by ice to produce the boulder bed. At the same time as the reworking of the boulders there would also have been the introduction of material already in the ice from further up glacier. Thus, the final deposit actually reflects a polygenetic, multi-transport path, mixed origin which might explain why the size–shape–lithological-striation composition is so different from the other localities. The boulders from the Stofbakkies Quarry form part of the glacial sequence and hence the difference between them and the smaller clasts is less marked.

Thus, as with shape, the response of size to lithology is not simplistic and great care must be taken in interpretation. The importance of deposit genesis in comprehension of the size–shape component is shown by the Elandsvlei samples. In fact, here genesis is required to interpret the shapes rather than the case of using shapes alone to determine genesis; the information from the shapes enforcing the hypothesized origin.

VI. CONCLUSION

Clast shape can be a useful tool in the interpretation of glaciogenic sediments and it is certainly far better than subjective appraisal, particularly when venturing in to size–shape relationships. However, it would appear that in the Dwyka Formation, and presumably most glacial deposits, the utilization of shape information is far from simple and great care must be taken. Gross sample data are useful, but hide many variations as a function of the lithologic composition. Size, shape, and lithology all interact and may be further complicated by variations in transport distance; reworking of sediments can cause greater complexity. Unfortunately it appears, at least in the areas studied, that no simple relationship can be generated, except perhaps in the generalized form where an improved roundness is indicated for the platform facies association due to debris being transported much further with more reworking than is the case for the valley facies. Further work is required to extend the present findings to other Dwyka depositional environments, but in the meantime it is suggested that workers should ignore the intimated size–shape relationships and attempt to obtain more objective, quantitative data.

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PALYNOLOGICAL EVIDENCE FOR INTERGLACIAL VEGETATION COVER ON MARION ISLAND, SUBANTARCTIC

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ABSTRACT

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Peats and organic-rich sediments found intercalated with glacial deposits on Subantarctic Marion Island yielded pollen spectra indicating vegetation and climatic conditions similar to present. These deposits are considered to have formed under interglacial conditions. A few palynomorph types of plants not recorded on the island were found in the samples. These could have been introduced through long-distance dispersal, but one of them might belong to an extinct taxon. Zonation in the pollen profile suggests that during the interglacial phase the ocean shore was close to its present position and that it moved some distance away at the onset of the glacial stage.

INTRODUCTION

Marion Island is located in the southern Indian Ocean (46° 54'S, 37° 45'E) approximately 2° of latitude north of the Antarctic Polar Front (Fig.1). During the Quaternary the island experienced a series of distinct glaciations separated by interglacial episodes characterized by palaeosol development and extensive volcanic and tectonic activity (Hall, 1978, 1979, 1982). A number of K/Ar dates have been obtained for the basaltic lava flows (McDougall, 1971) but none are from the present study area and are not able to be integrated stratigraphically with any certainty. The dates from these lava flows fix interglacial conditions at around 276 000 ± 30 000 B.P. and 103 000 ± 10 000 B.P., but no rigorous estimation of the onset of glaciations can be given. Palynological investigations have been undertaken from what are considered to be Würm age and Holocene peat cores (Schalke and Van Zinderen Bakker, 1971) and it has been suggested that during the late glacial the lowland areas were dominated by *Azorella* vegetation which implies that temperatures were 2° to 3°C lower than at present. Since approximately 10 000 B.P. conditions appear to have been fairly constant, with the vegetation

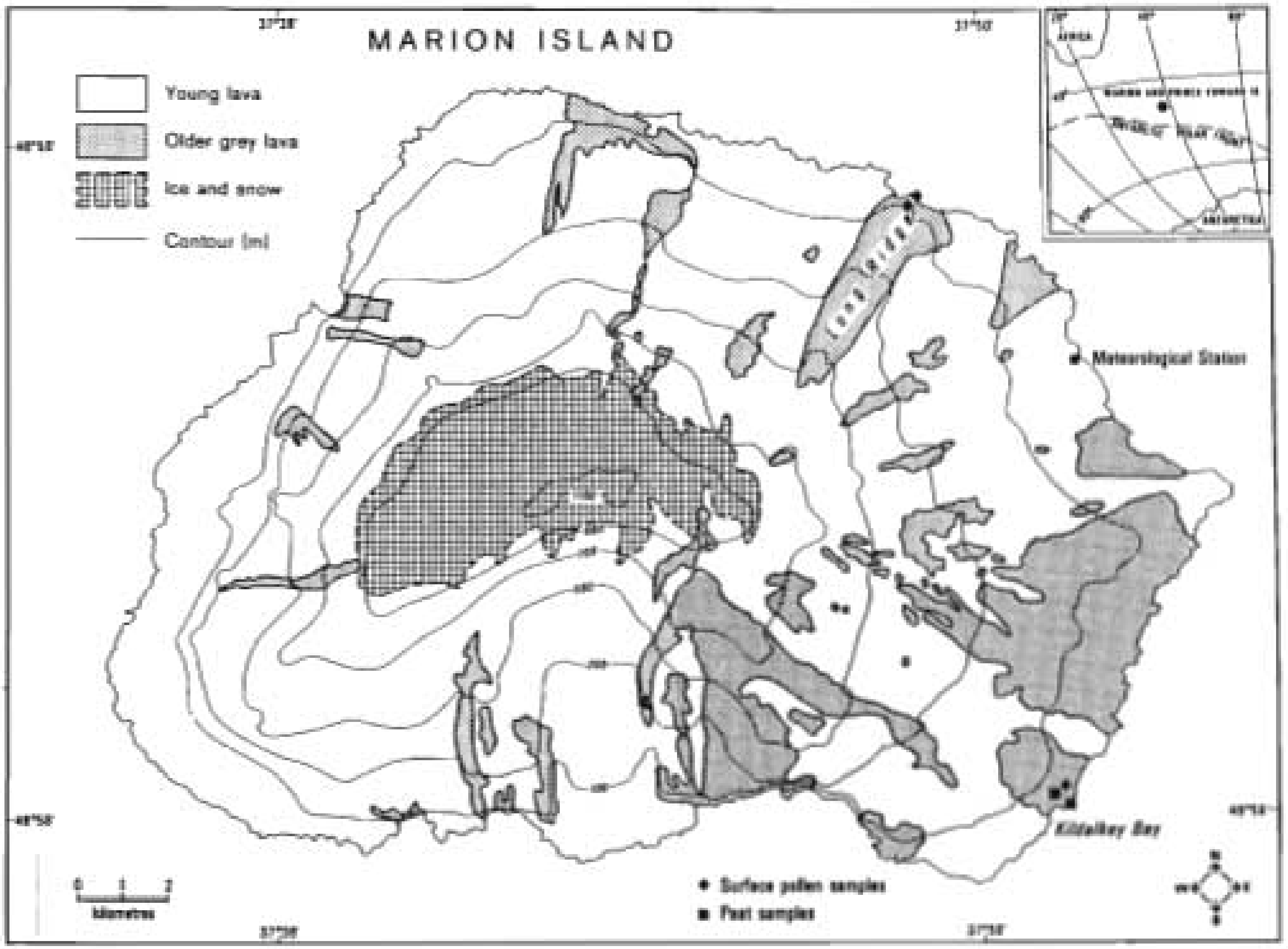

Fig.1. Locality map of Marion Island.

resembling that of the present time (Schalke and Van Zinderen Bakker, 1971). During recent investigations into the Quaternary glacial geology and palynology of the island, thin layers of peat and organic-rich deposits, considered stratigraphically to pre-date a series of glacial deposits (Hall, 1978), were found in a small bay (unofficially termed "Small Kildalkey Bay") adjacent to Kildalkey Bay (Figs.1 and 2). Pollen recorded in exploratory samples of these sediments provide evidence of the early vegetation on the island. The results can be interpreted in terms of pollen spectra of modern and Holocene vegetation. In addition to the peat, other interglacial deposits, such as rhythmites (varves), were collected from the Kildalkey Bay area plus from Ships Cove and Long Ridge (Fig.1), but these yielded no pollen.

STRATIGRAPHY

The age of the intercalated peat deposits in Small Kildalkey Bay cannot be determined precisely but, stratigraphically, they certainly predate at least two distinct glacial episodes (Hall, 1978). Small lenses of basal peat (ca. 0.4 m thick) occur on top of a grey basaltic lava and are overlain by fluvially deposited sand and gravel with peat inclusions (Fig.3). There then occurs a sequence of tills, the uppermost one or two metres of which exhibit a palaeosol with some accompanying fluvial deposits (rhythmites). Above the palaeosol are found a series of tills from the most recent glacial (Fig.3) upon which there is a post-glacial peat cover. Five radiocarbon dates from this post-glacial peat yielded dates ranging from 7120 ± 45 B.P. (GrN—8577) near the base to 900 ± 500 B.P. (GrN—8579) at 1 m below the top (Lindeboom, 1979).

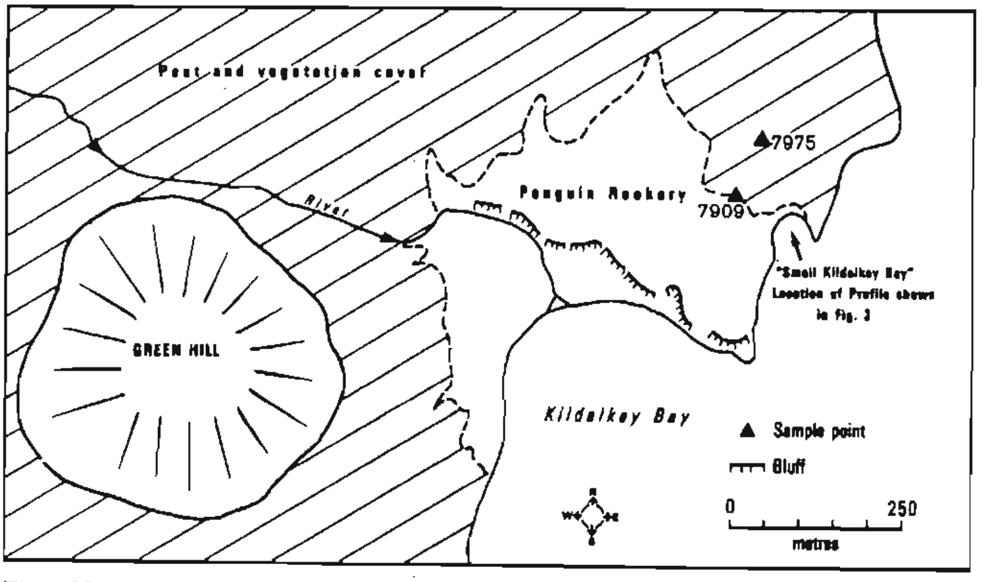

Fig.2. Map of Kildalkey Bay, showing the positions of the profile and two of the younger samples.

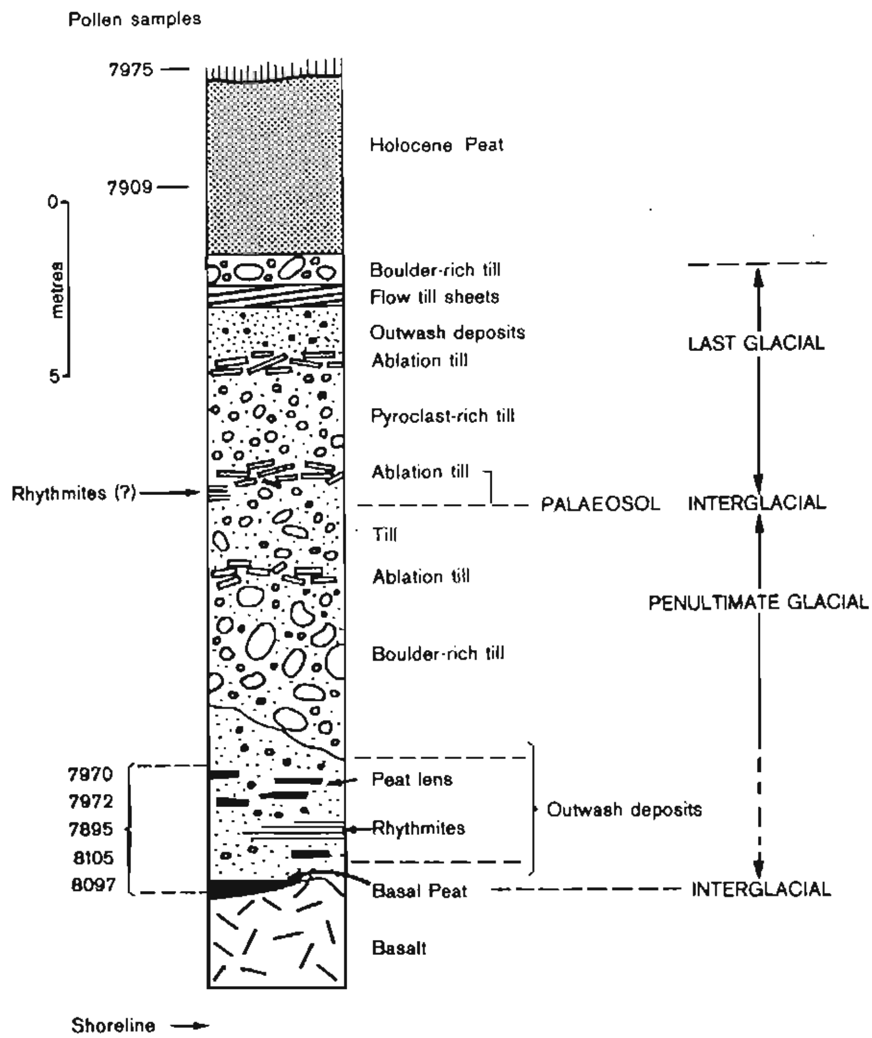

Fig.3. Profile of the glacial and interglacial deposits at Small Kildalkey.

The inference (Hall, 1978) is that the basal peat formed upon the basalt during an interglacial stage; unfortunately no date is available for the basalt. With the move towards glacial conditions, outwash dissected the peat cover and laid down the sands and gravels prior to the arrival of the ice and the deposition of till. The succeeding interglacial is marked by the palaeosol development and the most recent glacial is expressed by the sequence of tills which overlie the palaeosol. The dated surface peats are all of post-glacial age.

METHODS

Five samples of peat and fluvial material containing pollen were obtained from the lowest interglacial deposit (Fig.3). In order to aid palaeoenvironmental interpretation, a number of modern surface pollen samples were collected: one from Kildalkey Bay and two from Long Ridge. In addition, a sample was obtained from ±4 m below the top of the 6 m Holocene peat at Small Kildalkey Bay. The Holocene spectra described by Schalke and Van Zinderen Bakker (1971) were also utilised as examples of modern production.

The palynomorphs were extracted by means of acetolysis and heavy-liquid mineral separation, and the slides were mounted in glycerine-jelly. The loss of ignition and the organic carbon content, using the Walkley-Black method (Allison, 1965), were also determined and are given together with the resulting spectra in Fig.4.

POLLEN SPECTRA

During the pollen analyses the following types were recorded in the interglacial deposits: *Azorella*, *Acaena*, Gramineae, *Ranunculus*, *Montia*, *Lycopodium magellanicum*, Compositae (mainly *Cotula*) and small numbers of unidentified forms (Fig.4). In general, Gramineae pollen is the most abundant. Of the five samples (Fig.4), the lower three contain high numbers of *Cotula* whilst the upper two contain relatively high percentages of *Azorella* and *Acaena*.

It is interesting to note that the *Callitriche* and monolete Pteridophyte spores, which are more important in the Holocene and surface samples (Fig.4), are not recorded in the studied interglacial deposits.

DISCUSSION

The organic carbon content and the loss by ignition values (Fig.4) of the Holocene peat (average of 7 samples from different levels) are substantially higher than those found for the interglacial peat. The lowest (stratigraphically) interglacial sample (No. 8097 in Fig.4), however, produced a relatively high value and thus consists of a fairly pure peat, whilst the others are seen to contain large inorganic fractions.

In general, the similarity of the interglacial pollen assemblages with recent types (Fig.4) suggest that at the time of their formation the main components of the present vegetation system were already established and that the climatic regime must have been fairly similar to the present. Palynomorphs, which appear to be bryophyte spores, were observed in small numbers but these are not recorded due to uncertainty about their morphologies. Approximately the same concentration of these are recorded in the modern vegetation samples, of which bryophytes are an important constituent as is e.g. shown by Gremmen (1981). It can therefore be assumed that bryophytes may also have been established on the island at this stage.

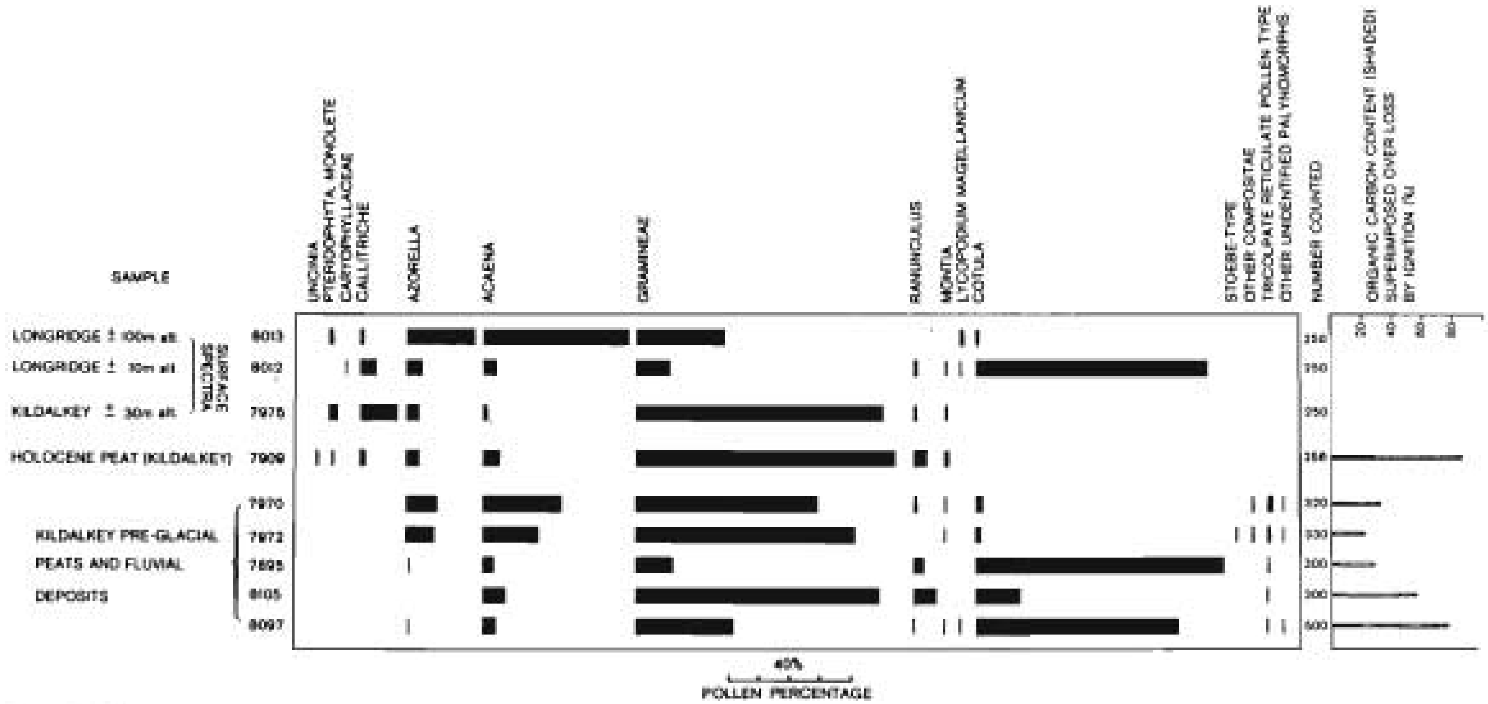

Fig.4. Pollen diagram of the peat deposits at Small Kildalkey compared with the results of some surface pollen spectra.

It is interesting that some forms were recorded which do not occur on the island at present. The most prominent are Compositae pollen, other than those of *Cotula*, and a tricolpate reticulate type (*Cardamine*?) (Fig.5). The regularity with which the latter is recorded tentatively suggests that it was produced on the island rather than transported there from distant areas, such as is probably the case with the others and the *Nothofagus* and *Podocarpus* pollen grains recorded by Schalke and Van Zinderen Bakker (1971). Current research on the recent pollen deposition and the foreign influx which will be applied to throw more light on the origin of such palynomorphs, shows that a number of types are regularly transported to the island but the mentioned tricolpate reticulate type does not appear to be one of them.

It is important to note that *Azorella* is very scarce or absent in the lower samples of the interglacial, less than 0.01% being recorded. This is in direct contrast with the much higher *Azorella* counts in the younger deposits and surface samples. In the late Pleistocene and Holocene spectra of Schalke and Van Zinderen Bakker (1971) the numbers of *Azorella* were also found to be fairly high, often reaching 20% or more, and rarely dropping below 5%. Its scarcity in the Kildalkey deposits suggests that perhaps *Azorella* did not grow on this part of the island. Whether it was more abundant elsewhere on Marion Island is uncertain.

The high percentages of *Cotula* in the lower interglacial samples suggest close proximity to the ocean, as indicated by the ±10 m altitude surface

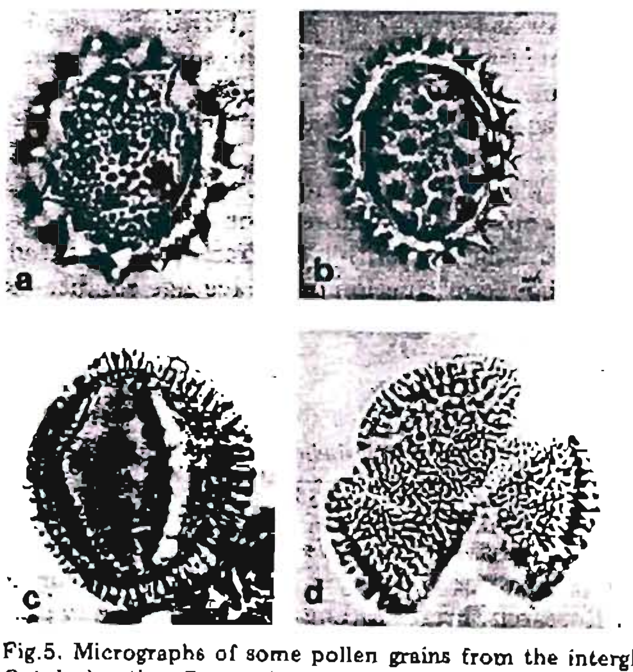

Fig.5. Micrographs of some pollen grains from the interglacial peats at Marion Island: a, *Cotula*; b, other Compositae; c and d, tricolpate reticulate type. All magnifications 1000×.

sample (No. 8012) which represents the present-day saltspray vegetation close to the shore (Fig.4). Thus the distance of the interglacial shore to sample site could have been comparable to the relationship found at the coastal saltspray or heavily manured coastal lowland vegetation types (described by Gremmen, 1981) where *Cotula* is prominent today. In contrast, the upper two interglacial samples do not suggest a strong littoral influence. These two samples compare with the ±100 m altitude present-day surface sample (No. 8013) from Long Ridge which represents a steep slope, with *Blechnum*, *Acaena*, *Poa cookii*, *Azorella* and *Cotula*. It is therefore possible that these later deposits formed along a well-drained drier slope, probably at a somewhat greater distance from the shore. Whether this represents the onset of cooler conditions and lower sea levels associated with the approach of the glacial stage or tectonic and/or drainage changes is uncertain, but evidence to date (Hall, 1978, 1982) suggests that the former is most likely.

In general, the assemblages from the interglacial deposits differ markedly from those of the younger Holocene peat, which correspond more closely with the modern surface spectrum of the *Poa cookii* grassland above it (sample 7975, Fig.4).

CONCLUSIONS

The pollen grains derived from interglacial deposits on Marion Island indicate that there existed a vegetation type and general climatic regime comparable to those of the present. Whilst the general vegetation assemblage is similar to that of today, certain forms such as *Callitriche*, Pteridophytes and *Azorella* were rare, or even absent, for at least part of the studied interval. The interglacial pollen spectra show a clear zonation which implies that the distance from the shore increased as a result of sea level lowering at the onset of the glacial.

In view of these findings it is clear that further detailed studies of this type on the old organic deposits could provide more information about the plant colonisation of the island. This data set within the framework of dated lavas would prove invaluable in the reconstruction of the Ice Age conditions in the Subantarctic.

ACKNOWLEDGEMENTS

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A MODEL FOR THE DEPOSITION OF THE PERMO-CARBONIFEROUS KRUITFONTEI BOULDER PAVEMENT AND ASSOCIATED BEDS, ELANDSVLEI, SOUTH AFRICA

by

J.N.J. VISSER and K.J. HALL

ABSTRACT

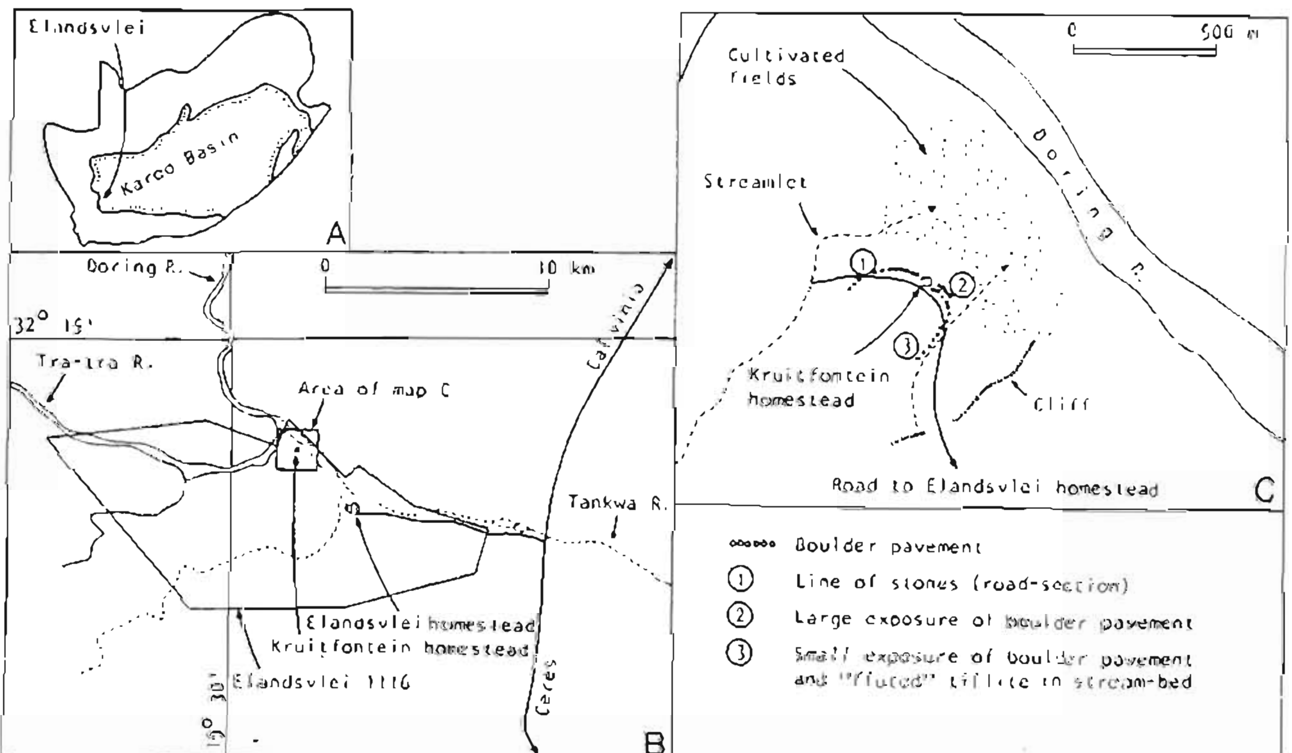
The Kruitfontein boulder pavement, with its veneer of grooved tillite, overlies a massive lodgement tillite. The pavement was formed by selective lodgement processes during palaeoice flow from the east. The grooved tillite was also deposited subglacially and the water-saturated sediment, largely derived from abrasion of the boulder bed, deformed plastically during ice movement. The laminated shaly diamictite, which conformably covers the grooved tillite, was deposited in either a subglacial or proglacial lake by suspension settling of clays, debris rain of coarse particles, and minor debris flows during stagnation or a temporary recession of the ice front. When the ice readvanced arenaceous tillite was deposited unconformably on top of the lake sediments.

CONTENTS

I. INTRODUCTION

The farm Elandsvlei 1116 lies along the western margin of the Karoo Basin (Fig. 1A) and is well known for its good exposures of the Permo-Carboniferous glacigene Dwyka Formation. The glacial beds, and in particular the boulder bed near the Kruitfontein homestead (Fig. 1B), drew attention of geologists as early as the turn of the century (Rogers and Du Toit, 1904). In recent years, interest ha again returned to the area as it is an important link i reconstructing the glaciation in the Karoo Basin.

Rogers and Du Toit (1904), Du Toit (1921), Stratter (1968), and Crowell and Frakes (1972), suggested palaeoice flow towards the east which implies centripetal ice flow into the Karoo Basin. Stratten (1968), and Crowell and Frake



(1972) based their conclusions on the bevelling of the inferred up-flow sides of clasts in the boulder pavement. Theron and Blignault (1975) rejected the evidence of bevelled clasts and suggested that ice had flowed in an opposite direction. In a later publication, Stratten (1977) attributed the conflicting directions of ice movement in the south-western Cape to glaciation of the area by different ice sheets.

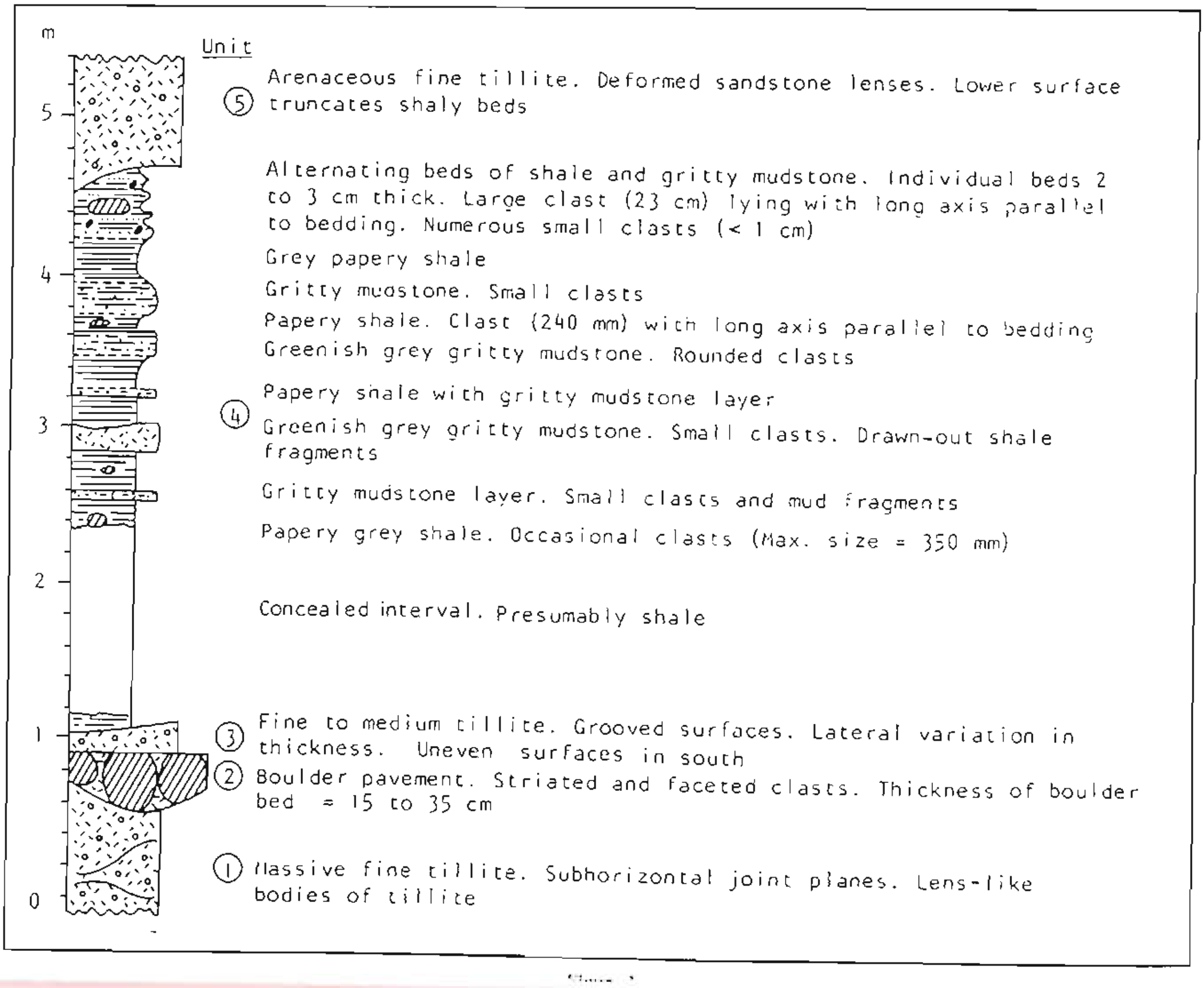
At Kruitfontein the boulder pavement is overlain by a soft-sediment pavement and the aim of the investigation was to establish a model for the deposition of the pavements and associated beds. For this purpose the lithology of individual beds, sedimentary structures, contact relationships, clast composition, clast shape, and directional data were studied at three sites near the Kruitfontein homestead (Fig. 1C). Measurements of clast sizes, shapes, and composition were obtained from 1 to 1,5 m^2 sample areas, of which two were located at the homestead and one at the road-section. For practical purposes 10 mm was taken as the lower limit for a clast and all material smaller was regarded as part of the matrix. Field studies were supplemented by laboratory investigation of tillite samples. In this paper the term "tillite" is applied to deposits laid down directly by ice, whereas the term "diamictite" is used in a non-genetic sense for all non-sorted sediments containing a wide range of particle sizes.

II. STRATIGRAPHY AND LITHOLOGY

The boulder pavement and associated beds, which occur about 17 m above the base of the Dwyka Formation, are 4 m thick at the Kruitfontein homestead. The sequence illustrated in Fig. 2 is subdivided into a number of lithological units for reference purposes.

The *basal tillite unit* (unit 1) consists of homogeneous tillite showing subhorizontal joint planes and containing sparse clasts. Clasts, mostly subrounded in shape, attain a maximum size of 2,5 m and consist predominantly of metamorphic rocks (mainly quartzite, gneiss, and schist). However, sandstone, siltstone, and shale clasts, derived from the underlying Cape Supergroup, are common towards the base of the unit and the tillite as a whole is also more arenaceous (Visser and Loock, 1982). The tillite matrix immediately below the boulder pavement consists volumetrically of about 50 per cent clay and silt, 40 per cent fine-grained sand (mainly quartz and feldspar) and 10 per cent granule-size fragments consisting of quartzite, schist, gneiss, lava, diabase (?), and dolomite. The granule and sand grains are, qualitatively, rounded to subrounded. In thin section deformed clay laminae are to be seen. Theron and Blignault (1975) reported a preferential imbrication of discoidal clasts in this unit.

The *boulder pavement* (unit 2) has a lateral extent of about 800 m, and good exposures are found to the south-east of the Kruitfontein homestead (Fig. 1C). The boulder bed consists largely of a single layer of boulders and pebbles and the bed thickness varies from 150 to 350 mm, depending on the size of the clasts and whether stacking of clasts occurred. In the latter instance the bottom clasts have been forced into the basal tillite by the overlying ones, possibly due to overriding ice. The lower contact of the bed is thus uneven, whereas the upper contact is a smoothed surface (Fig. 3).



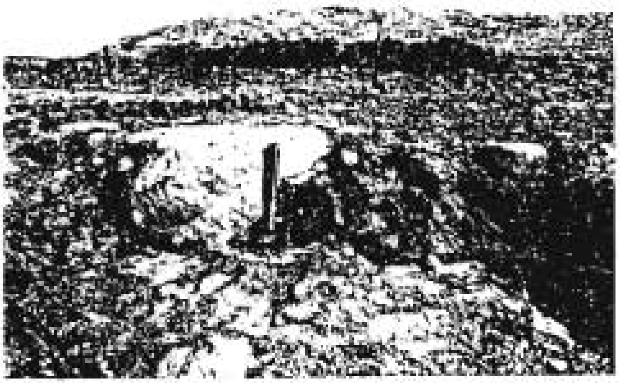


Figure 3
Uneven base and smoothed upper surface of boulder bed at Kruitfontein homestead. Basal tillite unit with sparse clasts in foreground.

Figure 4
Rounded clasts with good clast sorting in boulder bed

In three selected areas, the lithology was recorded of clasts measuring between 20 and 400 mm along the long axis as well as all other exposed clasts larger than 400 mm in size. The results (columns A and B, Table I) suggest that the clast composition is size dependent (Hall and Visser, in press) as large clasts of dolomite and schist are uncommon due to their susceptibility to mechanical breakdown, whereas the population of gneiss and quartzite clasts increases in the boulder fraction. The boulder bed is composed entirely of clasts derived from a distant source. Quartzite, which is the dominant clast type, includes white, grey, green, and purplish varieties. The green quartzites were derived from the Namaqualand Metamorphic Complex, whereas the purplish quartzite clasts came from the Matsap beds, implying a transport distance of between 300 and 550 km.

Several parameters were calculated to study the shape of the clasts. The Cailleux flatness index (= 184,5) shows that the clasts (n = 16) are not flat, whereas the maximum projection sphericity index (= 0,69) shows that they are not entirely disc shaped, but are approaching spheres. This conclusion is substantiated by the elongation index (= 0,78). The modified Wentworth roundness index (= 0,48), which refers to the roundness of corners, shows that the clasts are generally well-rounded and can best be defined as subrounded to rounded. The clast morphology of the boulder pavement is distinctly different from that of the basal tillite as well as of the overlying laminated diamictite in that the clasts tend to be more spherical and better rounded and a higher percentage are striated and faceted.

The matrix of the boulder pavement is very fine grained and clayey. Matrix deformation, often in association with shear structures, was observed where plastic flow of the soft

TABLE I
Clast Composition (%) of the Kruitfontein Boulder Pavement and Laminated Diamictite

Lithology	A Clasts > 400 mm n = 28	B Clasts 20–400 mm n = [illegible]	C Clasts [illegible] mm n = [illegible]
Quartzite sandstone	[illegible]	[illegible]	[illegible]
Conglomerate/grit	1	[illegible]	[illegible]
Dolomite	0	[illegible]	[illegible]
Chert/BIF/jasper	0	[illegible]	[illegible]
Granite/gneiss	18	[illegible]	[illegible]
Lava	[illegible]	[illegible]	[illegible]
Diabase	[illegible]	[illegible]	[illegible]
Schist	[illegible]	[illegible]	[illegible]

A Random sample from the boulder pavement
B Composite sample of three stations each measuring [illegible] m
C Random sample from the laminated diamictite
BIF = banded iron-formation

Whilst the maximum clast size recorded is 1,72 m, sizes are not evenly distributed in the boulder pavement and areas where clasts are moderately sorted (So = 0,55; formula of Folk and Ward, 1957) and poorly sorted (So = 1,04) are obvious (Fig. 4). The prolate clasts exhibit a west-north-west to east-south-east long axis orientation (Fig. 5A). On the eastern side of some large boulders, concentrations of small, unsorted, angular clasts were recorded (Fig. 6). Numerous bullet-shaped boulders, in which the ends represent stoss and lee sides (Boulton, 1978), were also found (Fig. 7).

Investigation of clasts (n = 16) taken from a 1 m² area at the road-section shows that 81 per cent exhibit striations and that 93 per cent have facets with a mean of 1,63 per clast and a standard deviation of 0,81. On some surfaces more than one set of striae are present. The clasts in the boulder

matrix occurred between clasts or where matrix was bulldozed in front of the embedded clasts. The latter phenomenon occurs on the lee side of bullet-shaped clasts and a faint cleavage due to shearing can be seen in the compressed matrix (Fig. 7).

The *grooved tillite* (unit 3) overlying the boulder pavement varies in thickness from 100 to 230 mm at the homestead. This thickness increases to about 500 mm in the stream-bed where the surface of the tillite is very uneven. The tillite has a fine-to-medium texture with very few clasts. Although Rogers and Du Toit (1904) stated that there is no marked difference between the tillite above and below the boulder pavement, the matrix of this upper tillite has a higher clay/silt content (60%) and a smaller percentage of sand-size grains than the basal unit. The sand-size grains in the upper tillite are also much more angular in thin section

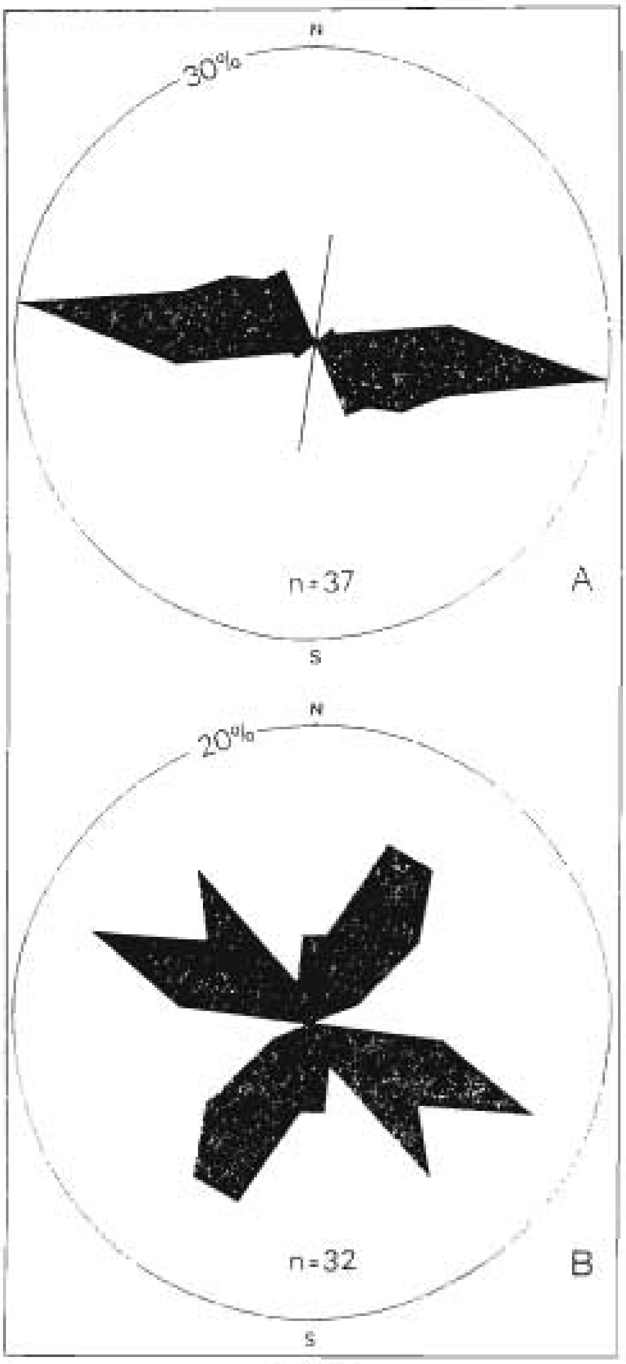


Figure 5
Mirror-image diagram of clast long-axis orientation in the boulder pavement (A) and laminated diamictite (B)

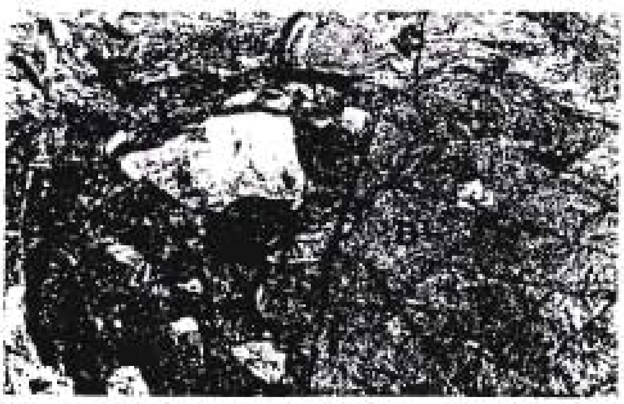


Figure 6
Small clasts concentrated on the side of a large boulder (?)

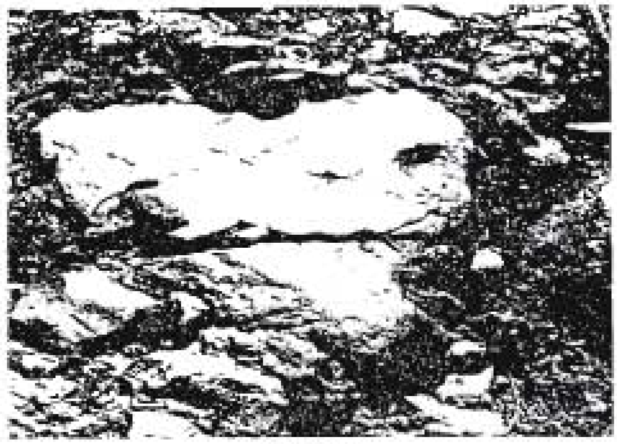


Figure 7
Large bullet-nosed striated boulder with stoss and lee sides (stoss side towards viewer). Ice flow in direction of the handle of the geological hammer. Note the compression of the tillite matrix on the down-flow side of the boulder (arrow) and the development of (leeward) cleavage.

The tillite at the homestead and in the stream-bed displays soft-sediment grooves (Fig. 8). In the south the grooved bed is overlain by tillite, the surface of which was shaped into ridges, low mounds and dome-shaped structures (Fig. 9). Evidence of soft-sediment deformation in these structures is also present.

The *laminated diamictite* (unit 4) consists of a grey papery shale and interbedded greenish-grey gritty mudstone containing scattered clasts up to 400 mm in diameter. The latter truncate the shale laminae as well as the thin gritty beds (Fig. 9). At first sight the clast

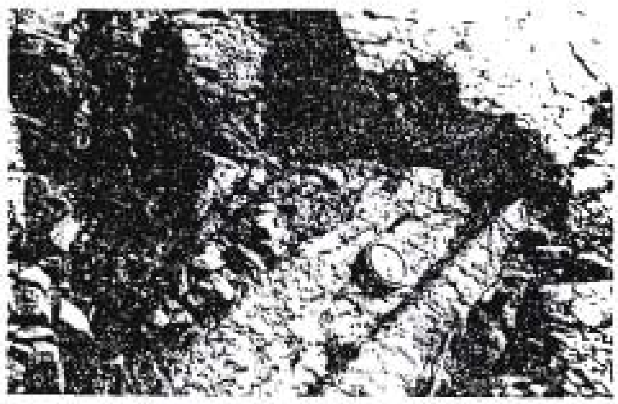


Figure 8
Intratillite pavement. The groove shows minor soft-sediment deformation.

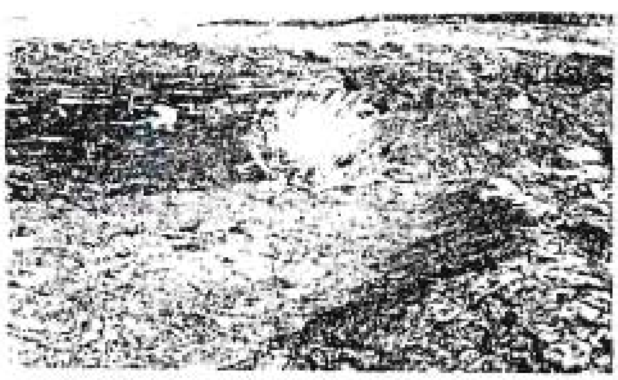


Figure 9
Tillite with uneven surface, conformably overlain by dropstone

composition appears to be different from that in the boulder pavement (Table I), but when consideration is taken of the trend of certain clast sizes that are lithology dependent, then the overall clast composition is comparable to that of the boulder pavement. The gritty beds contain mud fragments, some of which are drawn out, as well as small (<10 mm) extrabasinal clasts. The thickness of these beds varies from 20 to 200 mm and, although bed contacts are often sharp, gradational contacts were also found. The upper surfaces of some beds are undulating. The long axes of elongated clasts have a bimodal distribution in the AB plane; the one mode approximately parallel to the clast long-axis orientation in the boulder pavement and the other almost perpendicular to that direction (Fig. 5B). This unit blankets the uneven surface of the upper tillite thereby explaining the observed thickness variation.

An *arenaceous tillite* (unit 5) with large deformed sandstone lenses unconformably overlies the laminated diamictite without any visible deformation of the bedding on the contact (Fig. 10). The tillite is about 25 m thick and contains large clasts, up to 2 m in diameter, composed predominantly of gneiss with subordinate chert, lava, and quartzite.

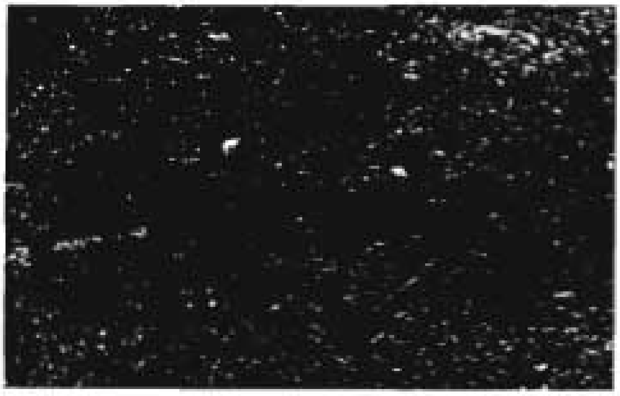


Figure 10
Arenaceous tillite (AT) unconformably overlying the laminated diamictite (LD). No deformation can be seen in the well-bedded unit.

III. GLACIAL DEPOSITION

Crowell and Frakes (1972) attributed the formation of the boulder pavement at Kruitfontein to erosion of frozen till during overriding by debris-laden ice. Theron and Blignault (1975) considered the Kruitfontein boulder bed as a lag deposit which could have formed by winnowing action along the buoyancy line and that minor advances of wet-based ice rotated and abraded the boulders forming the striated surfaces without the deposition of a significant thickness of till. Both above-mentioned theories, however, have shortcomings as the authors considered the deposition of this particular unit in isolation and not as part of the whole sequence.

A. Basal Tillite

The massive basal tillite, which is characterized by subhorizontal joint planes, a strong clast fabric, and material eroded from the immediately underlying bedrock, represents a lodgement till. This conclusion is, furthermore, supported by evidence that the tillite rests directly on striated pavements in the vicinity of Elandsvlei (Visser and Loock, 1982).

B. Boulder Bed

The boulder bed contains clasts of different sizes, shapes, and surface textures and shows varying grain to grain relationships as well as a smoothed upper surface, all of which are related to its mode of origin. Those clasts that have stoss and lee sides were formed at sites where abrasion and stream-lining of deeply embedded boulders by ice resulted in bullet-nosed ends. The lee sides of such boulders may represent remanié surfaces or surfaces reshaped by fracture at a point of contact with large overriding particles (Sharp, 1982). Some loose boulders from the pavement display evidence of acquiring the stoss and lee shape during lodgement before their upper surfaces had been finally abraded and smoothed in place by the overriding ice.

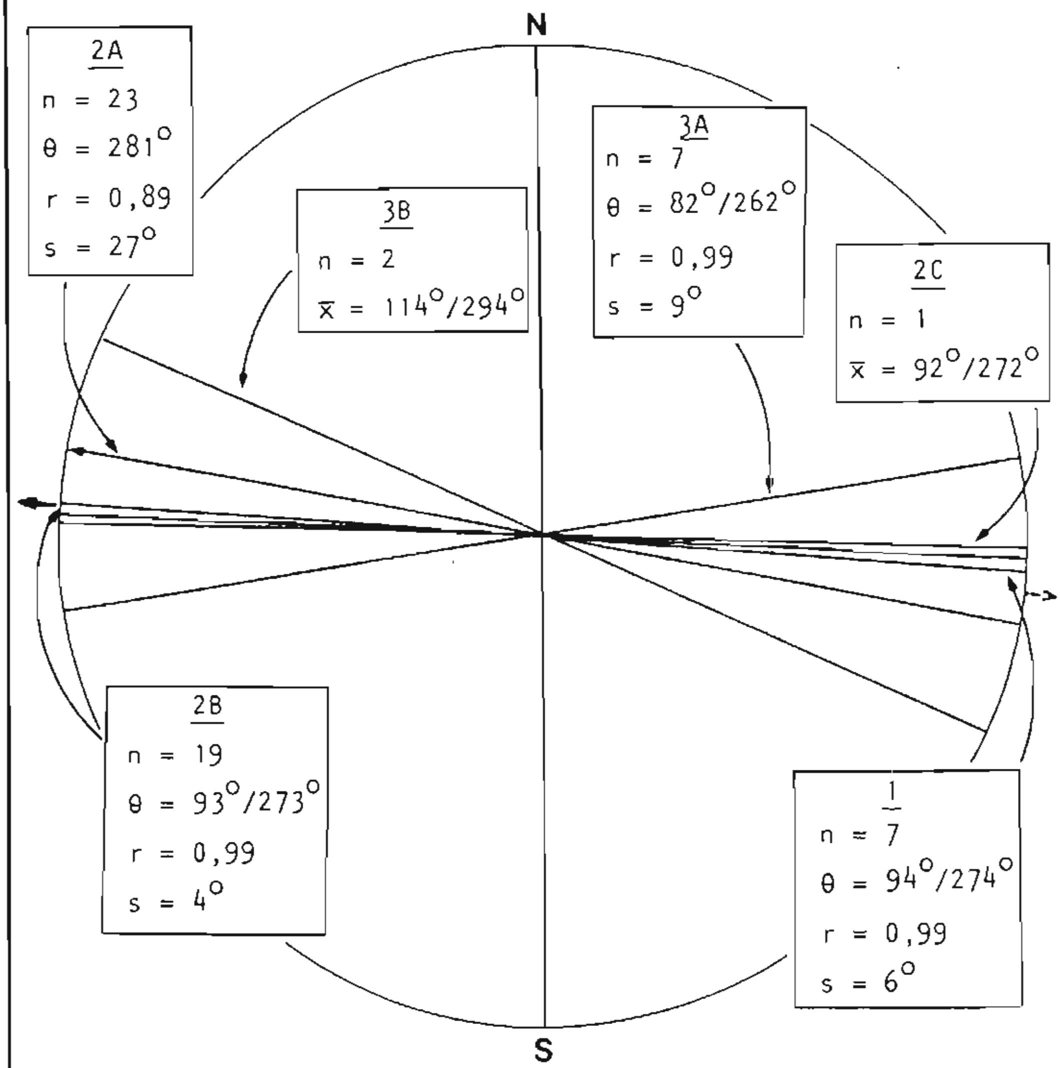
Dyson (1952) found that sediment-ridges accumulate subglacially on the lee side of large boulders and contain a high proportion of fine-grained material and rock flour. However, at Kruitfontein the abundance of small angular clasts implies an accumulation on the up-flow side of large boulders. These clast clusters probably formed where large embedded boulders, projecting above the depositional surface, acted as obstacles to further movement of debris at the ice-sediment interface.

The subrounded to rounded character of the clasts, especially those of pebble and cobble size, and the high percentage of striated and faceted clasts suggest transportation in the basal zone of the ice.

Thus, the presence of boulders with stoss and lee sides, clusters of small clasts on the up-flow side of boulders, bull-dozed and compressed matrix material in front of embedded boulders, and the strong long-axis fabric indicates subglacial deposition by wet-based ice (Shaw, 1983; Kruger, 1979). However, additional aspects that have to be borne in mind are the presence of well-rounded clasts and their good sorting locally. This could be explained by the incorporation of lag gravels (either from stream-beds or a beach) by regelation processes in the basal ice. When the basal ice conditions changed down-flow to warm melting, selective lodgement took place. The latter process can be visualized as a chain reaction whereby large boulders embedded in the basal till partly obstructed the flow of debris-laden ice so that more and more clasts became lodged on their up-flow sides. Continued movement of the debris-laden ice over the tops of the embedded clasts abraded and striated their upper surfaces.

The direction of the ice flow which deposited the boulders can be deduced by a study of the features related to the boulder pavement. Analysis of the directional data was done by means of the Von Mises distribution (Till, 1974), which gives a vector (θ), an estimate of the spread of angular values (r), and the circular normal equivalent of a standard deviation (s).

(i) Clasts with stoss and lee sides. Sharp (1982) found that the long-axis orientation of these clasts is a good indicator of former ice-flow direction. The orientation of 23 such clasts indicates that the ice flowed towards the west (281°) (2A, Fig. 11). Four clasts with an apparent bullet-nosed shape show a "reverse" orientation, but they can probably be attributed to the intersection of the latter abraded surface with the original shape of the clast. Some 80 to 90 per cent of the boulders have orientations within the statistical grouping.
(ii) Striae on embedded clasts: Twenty-three measurements were obtained from the three localities (1, 2B and 3A, Fig. 11). The orientation of the striae varies from 262 to 274° with a cluster value (r) at each station of 0.99 indicating that the values are consistent (the closer r is to unity the closer the points are clustered).
(iii) Sheared and compressed matrix with a faint cleavage (Fig. 7) occurs on the western sides of large boulders indicating compressional forces (i.e. ice flow) from the east.
(iv) Clusters of small, unsorted angular clasts occur on the eastern sides of large boulders suggesting ice flow from the east.



1.* Direction of striae on boulder pavement at road-section

2A. Ice-flow direction based on boulders with stoss/lee shape

2B. Direction of striae on boulders at homestead

2C. Direction of grooves on soft-sediment pavement at homestead

3A. Direction of striae on boulders in stream-bed

3B. Direction of grooves on soft-sediment pavement in stream-bed

n = number of readings. θ = vector. r = measure of concentration

s = standard deviation. $\bar{x}$ = lineation

---> Ice flow after Stratten (1968, pl. 25)

⟶ Ice flow after present study as well as Theron and Blignault (1975, p. 352)

*Numbers refer to localities on Fig. 1C

Figure 11
Palaeoice-flow directions at Kruitfontein.

(v) Certain clast lithologies can be successfully traced to their respective source areas, but it is the lithological association of clasts consisting of glassy purplish quartzite, dolomite, black chert, banded iron-formation, amygdaloidal lava, and jasper which is diagnostic of a source area (Ghaap Plateau) located about 500 km north-east of the Kruitfontein pavement.

The total of the directional measurements, substantiated by the clast lithology and other observations, suggest that the ice which formed the boulder pavement came from the east and north-east (Fig. 11). Visser (1982) found that the ice flowing down the Ghaap Plateau was initially directed towards the south and later spread out eastwards into the Karoo Basin. The above conclusion supports the interpretations of Theron and Blignault (1975) and Visser and Loock (1982) in respect of the direction of palaeoice flow during deposition of the basal tillite in the Elandsvlei area.

C. Soft-sediment Pavement

The intratillite pavement occurs close to the base of the argillaceous tillite, and in places tillite only a few millimetres thick separates it from the underlying boulder pavement. The slight difference in matrix composition between the basal tillite and this upper tillite could be attributed to the mixing of debris abraded from the boulder bed by overriding ice with debris derived from a distant source and carried in the basal ice. This suggestion is supported by the higher proportion of angular fragments in the upper tillite matrix.

The deposit represents a lodgement till of which the material was plastered on to the boulder bed as well as sculptured by the overriding ice. Plastic deformation of some grooves as well as the presence of small flow structures in the tillite covering the grooves indicate that the till was water saturated after deposition. The preservation of the soft-sediment grooves and the seepage of pore water from the till suggest that pressure on the soft sediment was instantly relieved locally, probably due to irregularities (crevasses, channels) in the glacier sole. Part of the water-saturated debris was then statically forced to flow up into such subglacial cavities during movement of the ice. This would explain the uneven surface of the tillite, the sporadic occurrence of the soft-sediment pavement, the thickness variation of the tillite unit, as well as the flow structures in the tillite. Von Brunn (1977) suggested the same origin for similar structures, although being much more regular, occurring on an intratillite pavement in northern Natal.

The soft-sediment grooves near the homestead and in the stream-bed strike 272 and 294°, respectively (2C and 3B, Fig. 11). No vector for palaeoice flow can be deduced, but the good correlation in strike between the grooves and the striae on the boulder pavement, and the proximity of the two pavements, could indicate ice flow also from the east.

D. Argillite with Dropstones

The thinly bedded to laminated clast-bearing sequence was deposited by suspension settling of clay and silt and debris rain from floating ice in a quiet water body. The thin gritty mudstone beds, with drawn-out mud clasts, probably represent minor debris flows. Absence of reworking of the top of the underlying tillite indicates that a large-scale transgression following an ice retreat could be considered as unlikely. It is, therefore, concluded that deposition of the argillaceous sediments took place in a glacial lake or meltwater pond.

The preferred bimodal long-axis orientation of the dropstones needs explanation. Engelbrecht (1973) also recorded a preferred clast orientation, although not as strong as in the associated massive tillite, in a dropstone argillite in the Barkly West area. He found that the long axes of the clasts are orientated parallel to the direction of ice flow and attributed the clast fabric to reorientation of the dropstones by active bottom hugging currents. However, at Kruitfontein, where clasts embedded themselves in the bottom muds, as is indicated by the truncated laminae, such an explanation is untenable.

At Kruitfontein one of the cross fabric modes also approximately parallels the direction of ice flow, which suggests that the fabric probably reflects the orientation of stones carried in the basal ice. Lindsay (1970) found fabrics in tillite with both parallel and transverse modes to the ice flow which he attributed to the presence of more than one shear domain near the ice margin where compression flow takes place. Thus where basal melting of debris-laden ice occurs, especially in water of very limited depth, it is suggested that clasts could settle on the bottom in such a way that their long axes reflect the clast fabric in the basal ice.

Although a proglacial lake is favoured for the deposition of the dropstone argillite, the presence of the clast fabric implies that the lake must have been subglacial or partly subglacial as ice floes on a proglacial lake would have yielded dropstones with a random clast long-axis orientation in the AB plane. However, other supporting evidence for deposition in a subglacial lake is lacking and, thus, the problem of the location of the lake relative to the ice front remains unresolved. Nevertheless, the blanketing effect of the lacustrine mud, which was deposited during a temporary stagnation and downwasting of the ice front, was essential in the preservation of the soft-sediment pavement and related glacial features.

During a later rejuvenation of the ice front minor erosion of the lake sediments occurred and arenaceous lodgement till, possibly containing esker sands (Theron and Blignault, 1975), was deposited unconformably upon the laminated diamictite.

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Boulder beds in the glaciogenic Permo–Carboniferous Dwyka Formation in South Africa

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ABSTRACT

Single-layer and massive boulder beds, which include boulder pavements, are sporadically distributed in the glaciogenic Permo–Carboniferous Dwyka Formation. These matrix-supported beds consist of moderately to poorly sorted, rounded boulders, cobbles and pebbles with a clast composition similar to those in the underlying or overlying diamictite. Alternatively, the clasts are composed of monolithic basement rock-types. The clasts show a long-axis orientation which, in the case of the boulder pavements, is parallel to the striae on the pavements.

The various types of boulder beds have a similar mode of deposition and their subglacial origin is evidenced by the clast orientation, clasts with stoss and lee sides, stacking of clasts, and the development of a cleavage in the matrix due to horizontal stresses exerted by the boulders in the subglacial sediment. Subglacial streams, kame mounds, subaqeously winnowed till, or boulder beaches supplied the coarse debris which was entrained in the basal ice by plastic flow and regelation. Selective lodgement of the transported boulders occurred down-glacier when the basal thermal conditions changed from cold-freezing to warm-melting. The formation of the different types of boulder beds is thought to depend primarily on the concentration of coarse debris in the basal ice.

BOULDER BEDS

Boulder beds with striated surfaces were first recognized in 1828 before the Glacial Theory was established (Flint, 1971). Boulder beds have been commonly referred to in descriptions of glacial sequences, but there are little published data pertaining to their texture, structure, composition, relationship to adjoining glacial beds, or their mode of deposition.

As no clear definition of what constitutes a boulder bed is available, references to such as a 'bouldery tillite' or 'conglomeratic layer' may in fact pertain to this form of deposit. A boulder bed is here defined as a tabular concentration of boulders, cobbles and pebbles, with either a close or a wide spacing in which the distance between clasts is less than the diameter of the clasts, forming single-layer or massive deposits within a diamictite (intratillite) or between diamictites (intertillite). Beds may be striated by overriding ice (the so-called 'boulder pavements').

Boulder beds are widely distributed in the Permo–Carboniferous Dwyka Formation which constitutes the basal unit of the approximately 15,000 m thick Karoo Sequence (Upper Carboniferous–Jurassic). The Dwyka Formation underlies an area of approximately 600,000 km^2 in the Karoo Basin (Fig. 1), attains a maximum thickness of about 800 m and consists predominantly of massive, bedded and laminated diamictites with subordinate mudstone, rhythmite shale, siltstone, sandstone and conglomerate. The sedimentary rocks are glacial and proglacial in origin and transgress upwards into argillaceous lacustrine and arenaceous fluvial deposits which constitute the major part of the Karoo Sequence.

Boulder beds and boulder pavements commonly occur along the periphery of the Karoo Basin (Fig. 1). Their apparent absence from the central and northern parts of the basin can be attributed to the obscuring cover of younger Karoo rocks. Although bore-hole cores are available it is impossible to recognize such

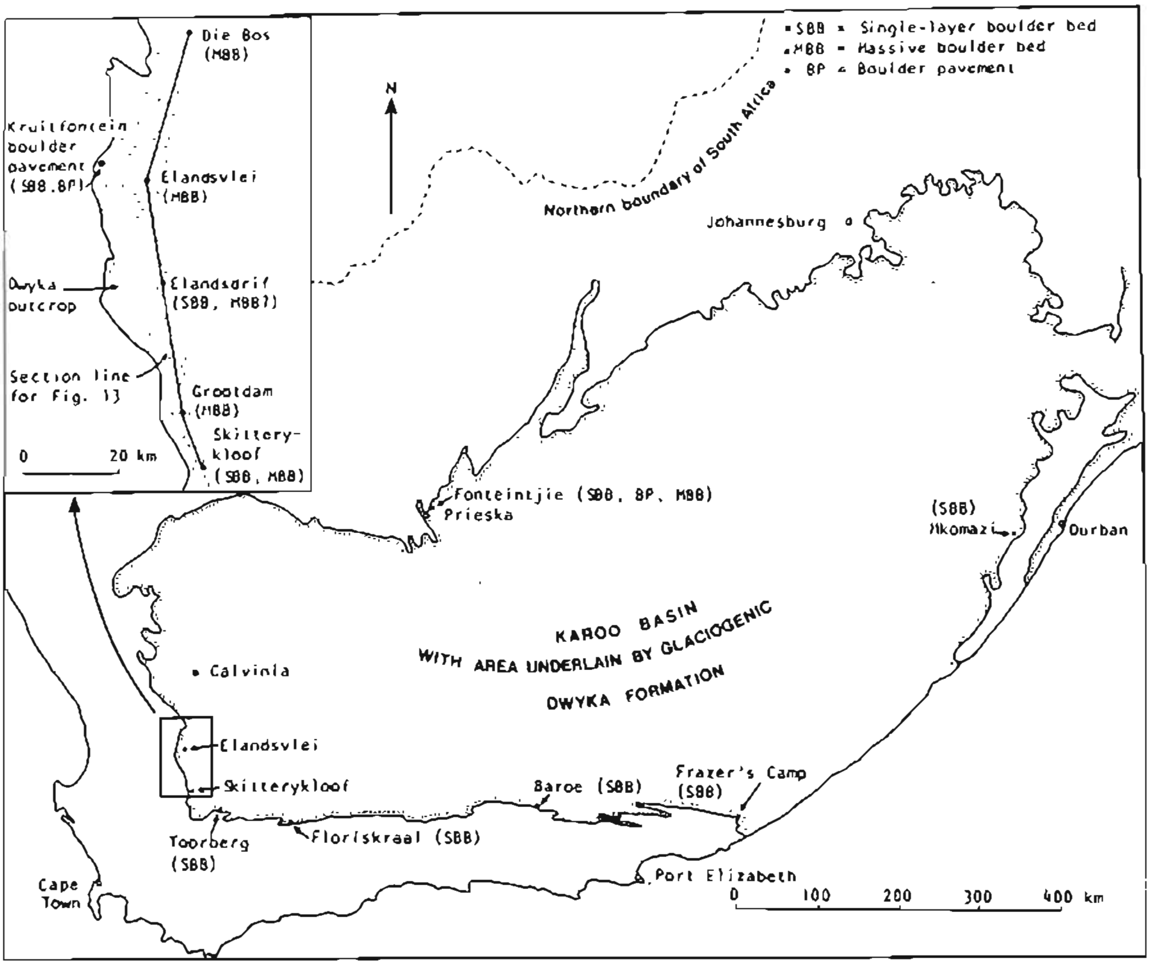


Fig. 1. Locality map showing the distribution of known boulder beds in the Dwyka Formation.

beds in small diameter cores and their presently known spatial distribution is thus primarily a function of the availability of good rock exposures.

This paper describes the boulder beds, their stratigraphic relationships and their mode of deposition. The term 'tillite' is only used for sedimentary rocks deposited directly by ice, whereas the term 'diamictite' is applied to any coarse-grained matrix-supported rock.

CLASSIFICATION AND DESCRIPTION

Distinction is commonly made between boulder or lag concentrates at the top of tillite beds and boulder pavements marking disconformable surfaces between tillites (e.g. Lindsay). As will be shown later all boulder beds are genetically related, and such a distinction thus becomes unnecessary. Theron & Blignault (1975) recognized two boulder bed forms in the Dwyka Formation: firstly, imbricated single layers of boulders which may or may not have striated surfaces, embedded in disrupted fashion in massive diamictite (a boulder pavement) and, secondly, single layers of boulders which apparently grade into boulder rudites with a disrupted framework and a diamictite matrix. Clast imbrication is often difficult to detect visually and we have therefore modified this classification of Theron & Blignault (1975) placing instead the emphasis on the layering. Boulder beds accordingly have been subdivided into single layers, with or without a glaciated surface (i.e. including boulder pavements), and massive beds consisting of a random clast distribution without noticeable internal layering. The

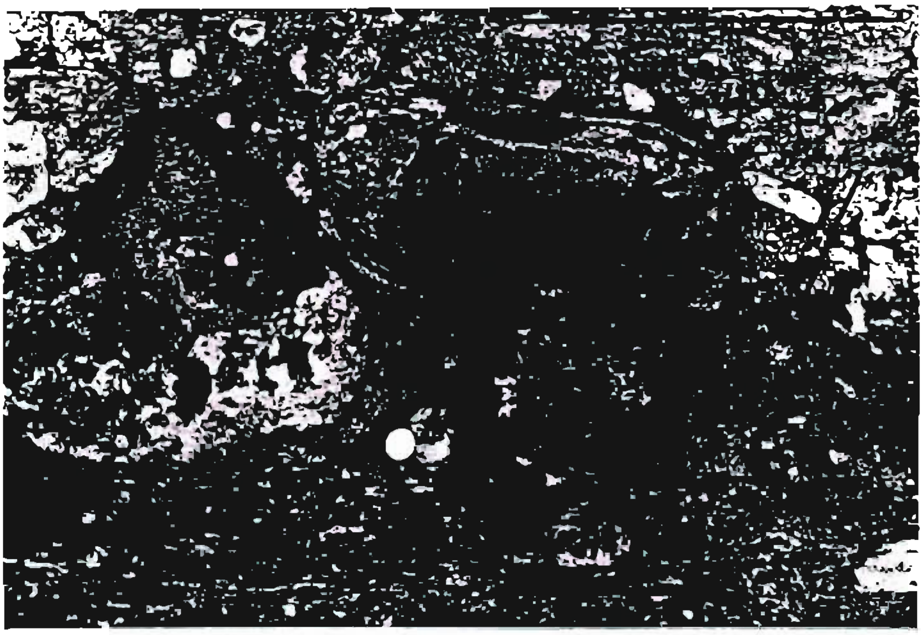


Fig. 2. Stacking of clasts (centre of photograph) and possible shearing in argillaceous matrix (arrow) next to a clast. Lower boulder bed at Fonteintjie, near Prieska.

field distinction between these two types is very easily made and no gradation between the two types was observed, although a minor vertical stacking of clasts (i.e. clast-supported framework) in single-layer boulder beds was observed (Fig. 2).

In the description of the boulder beds the term 'matrix' is used for all the finer material, including small stones or granules, in which the boulders and cobbles are embedded; it commonly contains structures which have a bearing on the origin of the beds.

(1) Single-layer boulder beds

(a) *Without a glaciated surface:* These beds are fairly widespread in the Dwyka Formation (Fig. 1) and maximum exposures of up to 100 m in length (e.g. Floriskraal and Skitterykloof) were recorded. The beds consist of closely to widely spaced rounded to well-rounded boulders and cobbles of up to 3 m diameter (Fig. 3), frequently exhibiting striations. Sorting varies from moderate to poor. Prolate clasts are mostly horizontally disposed, but locally pronounced up-glacier dips were recorded. A distinctive feature of some of these boulder beds is their almost monomictic composition. For example, in the southern part of the Karoo Basin gneiss clasts constitute 92% of the boulder bed on the farm Mt Stewart near Baroe, whilst the lower boulder bed at Fonteintjie near Prieska consists of 80% dolomite clasts.

The matrix between the boulders is very fine-grained and argillaceous with scattered angular grains and granules having the same composition as the clasts in the boulder bed or in the over- or underlying diamictite. The matrix is commonly massive, but flow structures and shearing were seen in places. In Fig. 2 faint bedding in the matrix is disrupted next to a clast forced into the matrix.

(b) *With a glaciated surface:* Boulder pavements with glaciated surfaces were recorded at Kruitfontein (Fig. 4) (Rogers & Du Toit, 1904) and Fonteintjie. The pavements were traced over distances of up to 800 m at Kruitfontein. Bed thickness varies from 0·15 to 0·6 m, depending on the size of the clasts and whether minor stacking of clasts occurred. The lower contact of the bed is thus eneven whereas the upper contact represents a smoothed surface.

Whilst the maximum clast size recorded in the boulder pavements is 1·72 m, the average clast size is much smaller (0·17 m at Kruitfontein; $n = 94$). Clast sizes are commonly unevenly distributed, but at Kruitfontein areas of moderately well-sorted clasts (Fig. 5) with a sorting coefficient of 0·53 occur, whereas a random clast sample of the pavement exhibits a

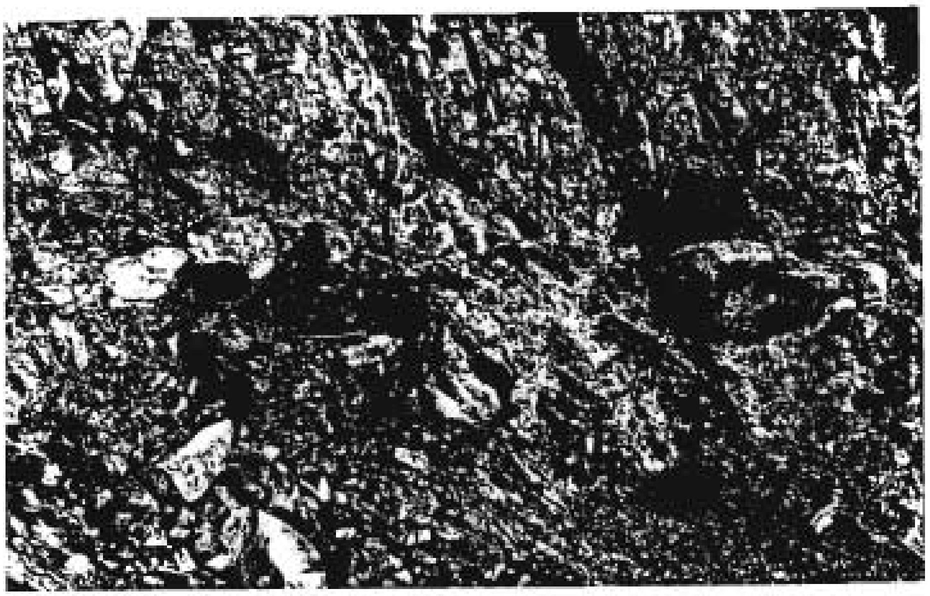


Fig. 3. Widely spaced boulders (arrows) in single-layer boulder bed at the base of the Dwyka Formation, near Toorberg. Bv = Bokkeveld beds.

poor coefficient of sorting ($S_0 = 1{\cdot}04$). Spacing of the clasts varies from close to fairly wide (Fig. 4).

The large prolate clasts exhibit a long-axis orientation (Fig. 6) which coincides closely with the striae direction on top of the boulders. At Kruitfontein the mean vector for the clasts is 281° and for the striae 273°. A crude imbrication appears to be due to the bottom clasts being forced into the underlying diamictite by the overlying ones.

Parameters for clast shape (maximum projection sphericity index, Cailleux flatness index and elongation index) indicate that the clasts in the Kruitfontein pavement are approaching spheres with discs slightly more abundant than rods (Table 1). The modified Wentworth roundness index which relates to the roundness of corners, shows that the clasts can best be qualitatively defined as subrounded to rounded (Fig. 5; Table 1). Numerous boulders were recorded with stoss and lee sides (Boulton, 1978) in which the upflow sides were abraded and stream-lined by the overriding ice, resulting in a bullet-nosed end. The lee side of such a boulder may represent a remanié surface or one reshaped by fracture at a point of contact with large overriding particles (Sharp, 1982). Of 16 clasts taken from a 1 m² area at Kruitfontein, 13 exhibit striations with often more than one striae set per clast and 15 have a mean of 1·63 facets per clast. The upper smoothed surface of the clasts truncates older bevelled surfaces and shows a well-developed set of striae (Fig. 4).

At Kruitfontein the boulder pavement is composed of clasts derived from the pre-Karoo basement. The population of each clast lithology varies with the clast size so that there is an increase in the number of gneiss and quartzite clasts in the boulder fraction, whereas large clasts of dolomite and schist are uncommon due to their proneness to mechanical breakdown (Table 1). At Fonteintjie 60% of the clast population consists of banded iron-formation derived from beneath the diamictite, and dolomite which outcrops in the near by source area (Fig. 7). Clasts of diabase (30%), lava and quartzite were distantly derived.

The matrix of the boulder pavements consists of an unsorted mixture of angular to rounded granules, grains and clay-size particles. The composition of the larger fragments more or less reflects the clast composition of the underlying diamictite. This is well illustrated at Fonteintjie (Fig. 8) where a massive yellowish-grey matrix (M1) was squeezed up between large clasts of banded iron-formation and is sharply overlain by reddish-brown faintly sheared matrix (M2). The reddish colouration of the upper matrix can be ascribed to the presence of numerous minute fragments of banded iron-formation. The matrix is commonly massive although in some thin sections faintly deformed laminae were observed. At Kruitfon-

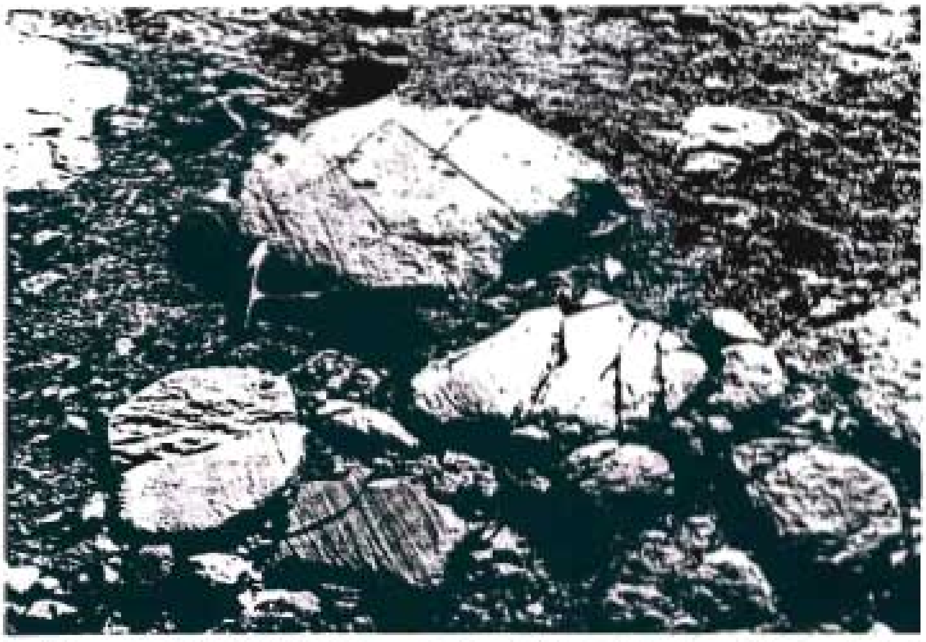


Fig. 4. Irregular clast spacing in Kruitfontein boulder pavement. Smoothed striated surfaces with ice flow from bottom to top.

Fig. 5. Moderately well-sorted and rounded clasts in Kruitfontein boulder pavement.

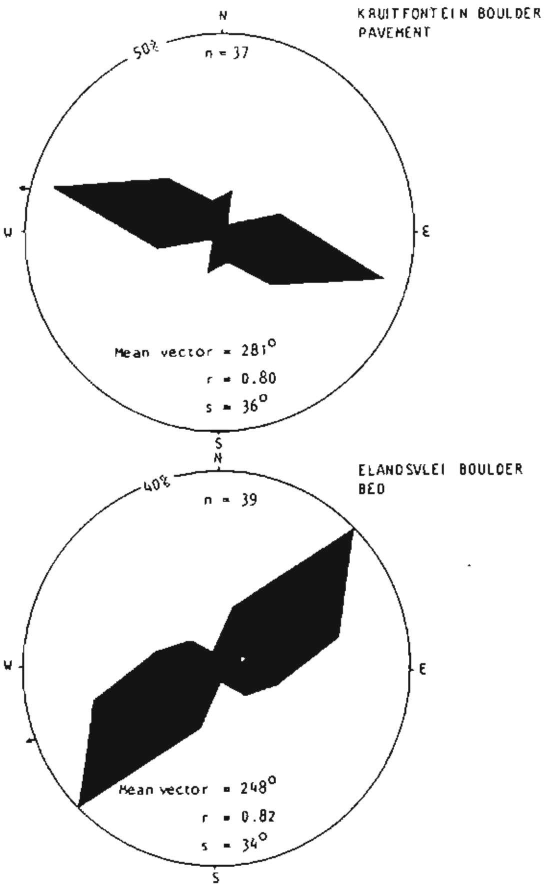


Fig. 6. Clast long-axis orientation in the Kruitfontein boulder pavement and at the top of the Elandsvlei boulder bed. The symbol r is an estimate of the spread of angular values and s is the circular normal equivalent of a standard deviation. The values imply that at Kruitfontein 80% of the orientations lie within an angle of $281° \pm 36°$ and at Elandsvlei 82% of the orientations lie within an angle of $248° \pm 34°$. Statistics based on the von Mises distribution (Till, 1974).

tein pavement the otherwise structureless matrix shows a near-vertical cleavage on the lee side of some clasts (Fig. 9). This can be attributed to horizontal stresses exerted in the matrix during lodgement of the large clasts.

(2) Massive boulder beds

The massive boulder bed at Elandsvlei was traced for about 32 km by Rogers & Du Toit (1904), but recent studies show that, except for a sand-covered interval, it extends for almost 100 km (Fig. 10). The bed has its maximum development, with a thickness of 12 m, in the north near Die Bos, where it consists of two units separated by massive diamictite. Southwards the boulder bed decreases in thickness to about 3 m at Skitterykloof. At Fonteintjie, near Prieska, a boulder bed with a thickness of 2·5 m was found. The upper contacts of the beds are fairly smooth, but the lower ones are very uneven so that at Elandsvlei vertical differences of up to 2 m over an horizontal distance of about 10 m were recorded. Lenses of bouldery conglomerate, up to 2 m long, also occur below the main bed at Elandsvlei.

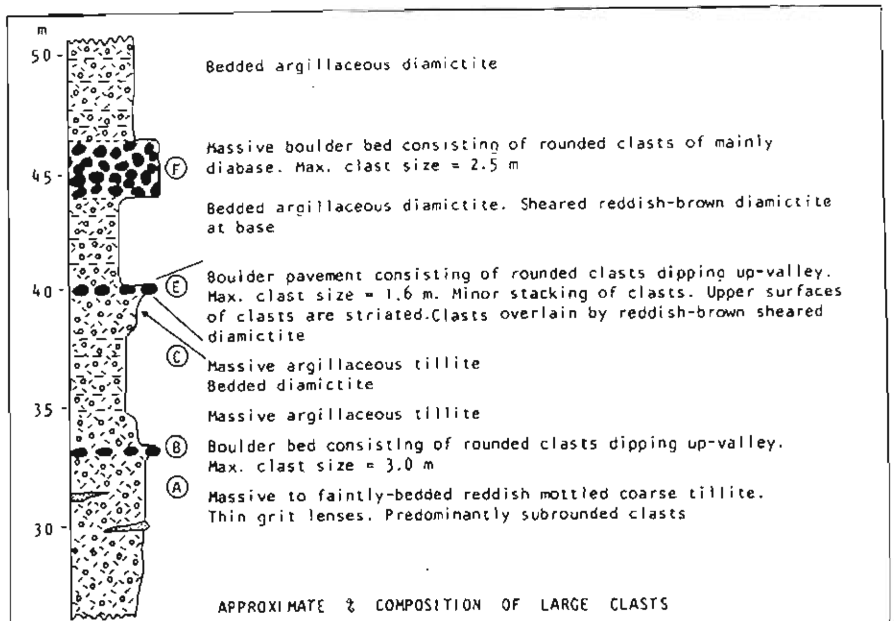
The boulder beds have a predominantly matrix-supported framework (Fig. 11) with clasts up to 2·3 m in diameter, but good examples of stacked clasts also occur (Fig. 12). The rounded to well-rounded clasts are moderately sorted and show a long-axis preferred orientation (Fig. 6). No imbrication was noticed, but a few clasts with a vertical long-axis orientation were recorded (Fig. 11). About 50% of the clasts at Elandsvlei are rod-shaped. Striations were seen on about 25% of the clasts. Rogers & Du Toit (1904) described well-striated boulders at Elandsvlei and in addition to these a few boulders with possible stoss and lee sides were found during the present study.

Distantly derived basement clasts predominate in both boulder beds. At Elandsvlei basic igneous rocks (lava and diabase) constitute about 56% of the clast population having a diameter of more than 0·2 m (Table 1), whereas in the same clast fraction quartzite comprises only 20%. Diabase clasts predominate at Fonteintjie (Fig. 7).

The matrix content of the boulder beds depends on the clast spacing, and measurements at two 1 m^2 areas at Elandsvlei indicate a matrix content of 53%. The matrix consists of small clasts and granules (average long-axis length at Elandsvlei = 4·5 cm; $n = 26$) in a gritty (coarse sand), clay-rich mixture. At Elandsvlei the coarse fragments are rounded to subrounded and consist predominantly of quartzite and basic igneous rocks (Table 1) whereas in the sand-size fraction quartz grains are dominant. Deformation of the matrix with the development of a faint cleavage can be seen below some large clasts. Carbonate cement occurs in the matrix and results in a nodular weathering of the diamictite at Elandsvlei.

STRATIGRAPHIC RELATIONSHIPS

Boulder beds, and in particular boulder pavements, have been related to the associated glacial beds in three different ways: Firstly, boulder beds interbedded



APPROXIMATE % COMPOSITION OF LARGE CLASTS

	A	B	C	D	E	F
Diabase/lava	30	5	15	40	30	45
Dolomite	10	80	45	5	20	5
Banded iron-formation	45	10	15	20	40	-

Fig. 7. Stratigraphic section and clast composition of boulder beds and adjoining diamictites at Fonteintjie, near Prieska.

Table 1. Clast characteristics of boulder beds and associated diamictites at Elandsvlei and Kruitfontein

	Elandsvlei			Kruitfontein	
	Clasts (>0·2 m) in boulder bed $n=87$	Matrix clasts $n=26$	Clasts (>0·2 m) 200 m below boulder bed $n=28$	Clasts (>0·2 m) in boulder pavement $n=28$	Clasts above pavement $n=30$
% Clast lithology					
Quartzite/conglomerate	20	42	31	58	30
Dolomite/chert	13	12	11	0	20
Banded iron-formation	1	0	0	0	7
Gneiss/granite	10	4	14	18	4
Diabase/norite	26	11	33	13	13
Lava	30	31	8	11	23
Schist	0	0	3	0	3
Clast shape					
% Blades	6	0	—	0	7
% Discs	44	46	—	56	20
% Rods	50	54	—	44	73
$\bar{x}$ roundness	0·44 ($s=0·15$)	0·34 ($s=0·18$)	—	0·48 ($s=0·20$)	0·26 ($s=0·15$)
$\bar{x}$ sphericity	0·73 ($s=0·08$)	0·68 ($s=0·08$)	—	0·69 ($s=0.09$)	0·67 ($s=0·10$)
% striated clasts	25	31	—	81	30

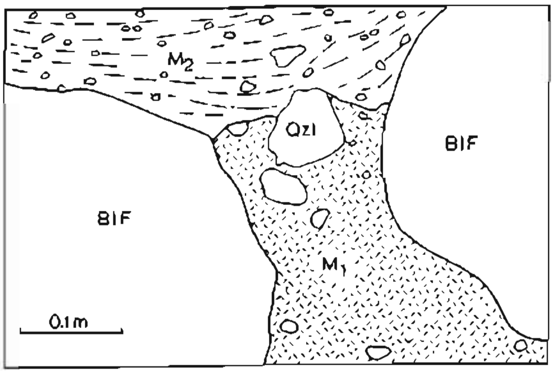


Fig. 8. Reddish-brown faintly sheared matrix (M_2) overlying massive yellowish-grey matrix (M_1) in the boulder pavement at Fonteintjie, near Prieska. Large clasts of banded iron-formation (BIF). Qzt = quartzite.

in a homogeneous diamictite sequence (Dreimanis, 1976); secondly, boulder beds that form an integral part of the diamictite overlying them (Dreimanis, 1976; Edwards & Føyn, 1981; Garnes, 1979; White, 1982; Moran *et al.*, 1980; Ojakangas & Matsch, 1981); and thirdly, boulder beds occurring at the top of diamictite sequences (Dreimanis, 1976; Lindsay, 1970; Ojakangas & Matsch, 1981; Crowell & Frakes, 1971, 1972; Elson, 1957; Dreimanis & Reavely, 1953; Theron & Blignault, 1975; Hansom, 1983).

In the Dwyka Formation the spatial distribution of boulder beds shows that both single-layer and massive boulder beds occur sporadically throughout the glacial basin fill (Figs 7 and 13). They are also unconstrained by the depositional facies in the Dwyka Formation and occur both in the valley (e.g. Fonteintjie) and platform (e.g. Floriskraal, Baroe, Toorberg and Frazer's Camp) facies associations (Visser, 1983).

Typical examples of the relationship of the boulder beds with the adjoining rock-types are illustrated in Figs 7, 10 and 13, whilst, in addition, Table 2 presents a summary of the rock-types in contact with the boulder beds. Table 2 shows that there is no obvious relationship regarding texture and structure between boulder beds and the adjoining rock-types. This is well illustrated in the Elandsvlei boulder bed where the under- and overlying diamictites change laterally in character from north to south without any corresponding change in the boulder bed (Fig. 10).

Comparison of the clasts as well as the overall composition of the boulder bed and the adjoining diamictite often indicates a stratigraphic relationship. At Fonteintjie near Prieska, the lower boulder bed (Fig. 7B) shows lithological affinities with the overlying diamictite (C) as both have a very high dolomite clast content, whereas the underlying diamictite (A) contains predominantly fragments of banded iron-formation. The boulder pavement (E) also shows a closer relationship with the overlying reddish-brown diamictite on account of its high banded iron-

Fig. 9. Cleavage developed in matrix next to clasts. This is indicative of horizontal stress exerted in the matrix during lodgement of the clasts. Kruitfontein boulder pavement.

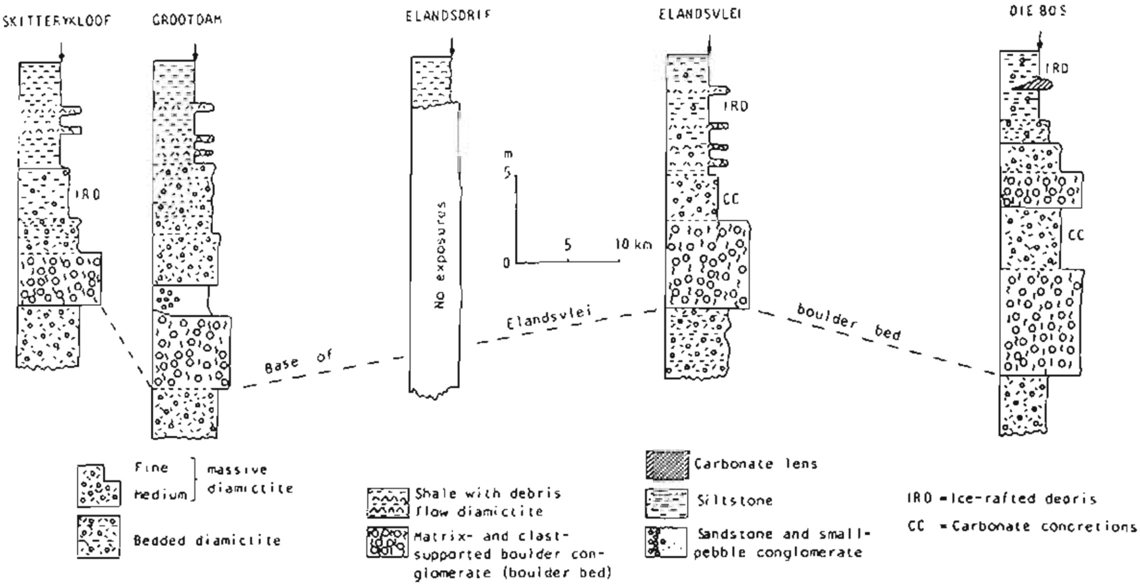


Fig. 10. Stratigraphic sections south of Calvinia illustrating the disposition of the massive Elandsvlei boulder bed (see Fig. 1 for section line).

formation clast content. In this case the actual contact between the two types of diamictite can be seen in the matrix (Fig. 8). A single-layer boulder bed at Skitterykloof which consists almost entirely of gneiss clasts, is probably related to the underlying diamictite in which gneiss clasts also predominate. There the overlying diamictite contains predominantly quartzite clasts and no gneiss. At Floriskraal (Fig. 13) the two single-layer boulder beds occur at the base of glacial units, whereas at Frazer's Camp, in the Eastern Cape, the boulder bed probably occurs at the top of a unit.

Commonly no conclusive relationship based on clast composition exists, even when applying statistical tests, between the massive boulder beds and the adjoining rock-types or matrix as the entire sequence shows an homogeneous clast composition. The only differences in clast composition can be attributed to the interdependence between clast composition and clast size. However, in some cases the single-layer boulder beds are in sharp contrast with the adjoining diamictites in that the clasts consist almost entirely of a single lithology derived from a distant source (e.g. Kruitfontein I in Fig. 13, boulder bed F in Fig. 7, and Fig. 3). Garnes (1979) found a similar relationship for boulder beds in Weichselian till of central South Norway.

Various parameters for clast shape were applied to find a correlation between the clasts in the boulder beds and those in the adjoining diamictite. At Kruitfontein there is a vast difference between the percentages of blade-, disc- and rod-shaped clasts in the boulder pavement and in the dropstone shale overlying it (Table 1). As clast shape is independent of clast size (Hall, 1983) this may indicate different clast populations, but the influence of lithology on the clast shape was not determined and the above conclusion may thus be invalid. The modified Wentworth roundness index for clasts from the boulder beds is higher ($>0{\cdot}4$) than for clasts from the matrix (0·34) or from the adjoining diamictite (0·26). This observation, i.e. that clasts in the boulder beds are fairly well rounded, is substantiated by the findings of Garnes (1979), Edwards & Føyn (1981) and Lindsay (1970) for boulder beds in Norway and Antarctica.

Clast fabric analyses were also applied as an aid to unravel the relationship between the boulder beds and the adjoining diamictites. Large prolate clasts in the reddish diamictite below the boulder bed at Fonteintjie (Fig. 7) have a long-axis orientation of 140°, and the boulders in the boulder bed have a near parallel long-axis orientation of 135°. These are in line with observations by Hälbich (1964) and Heath (1972) for Namibia where clasts in a diamictite several metres below a boulder bed also have a similar long-axis orientation to those in the boulder bed.

Thin section studies show that at Kruitfontein the

Fig. 11. Elandsvlei boulder bed illustrating the matrix-supported framework and the predominantly horizontal clast long-axis orientation.

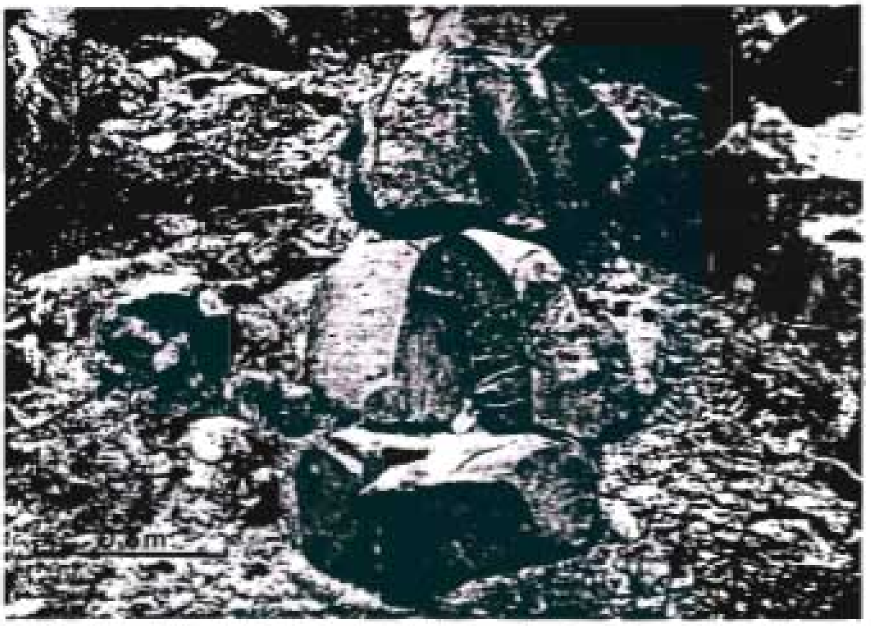


Fig. 12. Stacking of boulders at Elandsvlei boulder bed. Nodular weathering of matrix (arrows) due to the presence of carbonate cement.

Table 2. Relationship between boulder beds and adjoining rock-types

Adjoining rock-type	Boulder bed ($n=20$) Underlain by	Boulder bed ($n=20$) Overlain by	% of total contacts
Fine massive diamictite (average clast* size <25 mm)	7	5	30
Medium massive diamictite (average clast size = 25–100 mm)	8	6	35
Coarse massive diamictite (average clast size >100 mm)	2	4	15
Bedded diamictite	2	3	12·5
Local tillite	1	—	2·5
Conglomerate and grit	—	1	2·5
Siltstone and shale	—	1	2·5

*A clast was taken as being larger than 4 mm in diameter.

thin diamictite overlying the boulder pavement (Kruitfontein II in Fig. 13) differs mineralogically and texturally from the matrix of the boulder pavement. At Elandsvlei the overlying diamictite differs from the matrix of the massive boulder bed in that it contains a higher percentage of carbonate fragments and angular grains, whereas the mineralogical composition of the matrix in thin section almost perfectly reflects that of the clast composition of the massive boulder bed. However, in the single-layer boulder beds the matrix composition may differ considerably from the composition of the boulders in the bed (e.g. at Mt Stewart).

Evidence regarding the relationship of the boulder beds with the adjoining diamictites is thus by no means conclusive. In 65% of the observed occurrences where massive diamictite forms the contact rock with a boulder bed, the average clast size of the diamictite is less than 0·1 m, which is many times smaller than that of the boulder bed (Table 2), which implies that reworking of the diamictite could not have supplied the material for the boulder bed. At those boulder beds where conclusive evidence could be found, about 20% of the beds lithologically form an integral part of the overlying diamictite and about 10% are related to the underlying diamictite.

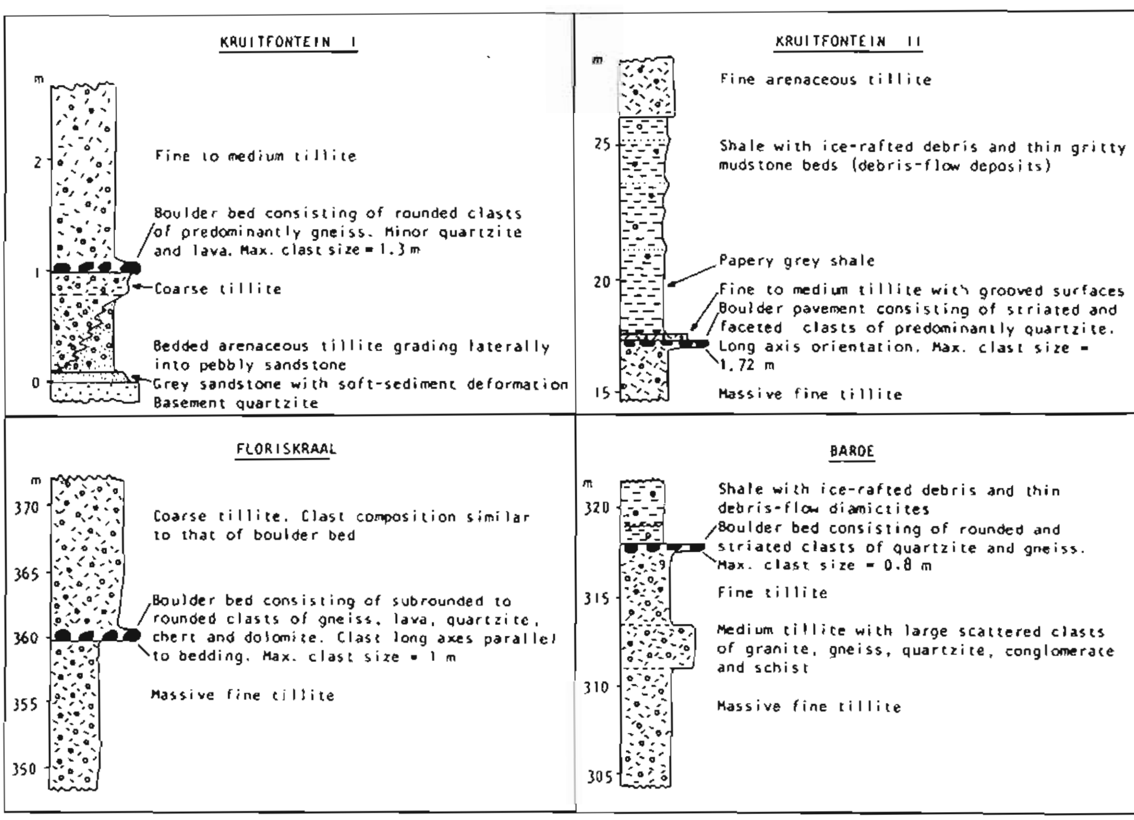


Fig. 13. Representative sections illustrating the occurrence of single-layer boulder beds (see Fig. 1 for locations).

diamictite as well as their long-axis preferred orientation, infer a common process of deposition and that the formation of the boulder beds was probably entirely a function of the availability of coarse material and the particular basal ice conditions. Possible reasons for selective lodgement instead of homogenization of the subglacial sediment (as is the case in ordinary tillite) are (i) that the basal load which consisted entirely of boulders, cobbles and pebbles greatly exceeded the competence of the basal ice due to a rapid change in basal thermal conditions, and (ii) due to bedrock topography a point was reached where the basal ice started to loose contact with the substratum. Such action would have released pressure on the basal ice, decreased the amount of basal melting, and prohibited the deposition of fine debris as well as the homogenization of the basal till. The latter process would also explain the limited extent and sporadic occurrence of some boulder beds.

In cases where shale overlies boulder beds (Baroe section in Fig. 13) little or no deposition of till occurred during or after formation of the boulder bed, a conclusion also reached by Theron & Blignault (1975). This phenomenon could be attributed to (i) a situation in which the ice lost contact with the substratum after deposition of the boulder bed whereby a release of pressure on the basal ice would have reverted the basal regime to cold freezing and prohibited the further deposition of till, and (ii) the basal ice probably became depleted in debris. Where the boulder bed forms a lithologically integral part of the overlying diamictite, continuous deposition of debris from the basal ice is implied except that the depositional process changed from selective lodgement to normal lodgement. Strong support for continuous deposition came from the lower boulder bed at Fonteintjie (Fig. 7) where the boulder bed B and diamictite C consist predominantly of dolomite derived from an up-valley source. There is, however, evidence for a break in deposition between the underlying diamictite (A), which consists of local material (banded iron-formation), and units B and C.

No suitable pebble and boulder concentrations which could have acted as a source for the boulder beds, have been found in the Dwyka Formation. In the Permo–Carboniferous Pagoda Formation in the Transantarctic Mountains, Lindsay (1970) was able to trace the boulders of a pavement to a small conglomerate-filled stream channel. Garnes (1979) also suggested that subglacial streams could have supplied the coarse debris for boulder beds in Norway. Lundqvist (1979) described and illustrated subglacial kame mounds in Småland, Sweden, which consist of well-rounded boulders and which would yield enough coarse debris for massive boulder beds. Another possible source is the erosion and winnowing of older tills as suggested by Dreimanis (1976), Flint (1971), Theron & Blignault (1975) and Ojakangas & Matsch (1981). This hypothesis has limitations regarding both the size of the coarse debris (underlying diamictites in the Dwyka Formation are commonly finer grained) and the presence of clasts belonging to a single rock-type. A possible suggestion is that the winnowed till could have been coarser up-glacier, but no plausible explanation for the concentration of certain clast lithologies in boulder beds can be given as such clasts (e.g. gneiss) would have hardly survived lengthy periods of reworking.

All evidence points to the transportation of predominantly rounded clasts, whether they were derived from a boulder lag formed subglacially (streams, kame mounds), subaqueously (winnowed till, streams) or subaerially (streams, beaches), to the point of lodgement down-glacier. During rejuvenation of a glacier, boulders, cobbles and pebbles could have become entrained in the basal ice by plastic flow (Sugden & Clapperton, 1981; Boulton, 1982). Smaller clasts could have been moved by regelation processes. The concentration of coarse debris in the basal ice would have determined whether a single-layer or a massive boulder bed would have formed. A boulder pavement would have resulted when debris-laden basal ice abraded and striated the top of the embedded boulders under steady state conditions.

It is difficult to visualize how some of the above processes can lead to the formation of boulder beds covering 'many tens of thousands of square miles' (White, 1982). Individual single-layer boulder beds in the Dwyka Formation probably do not exceed 100,000 m^2, whereas the thicker extensive massive boulder beds may exceed 100 km^2. It may be that the laterally extensive beds described in the literature are in fact discontinuous along a stratigraphic horizon.

The most likely hypothesis for the formation of boulder beds thus involves the transportation of rounded boulders and pebbles from an up-glacier source which might have been a subglacial or subaerial fluvial, beach or other type of lag deposit, and the selective lodgement of the large clasts where the basal thermal ice conditions changed to warm melting. The concentration of clasts in the basal ice and the topography of the ice–sediment interface might have influenced the type of boulder bed formed, its areal extent and its preservation.

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THE APPLICATION OF STONE COUNTS IN THE GLACIGENE PERMO-CARBONIFEROUS DWYKA FORMATION, SOUTH AFRICA

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ABSTRACT

Visser, J.N.J., Hall, K.J. and Loock, J.C., 1986. The application of stone counts in the glacigene Permo-Carboniferous Dwyka Formation, South Africa. Sediment. Geol., 46: 197–212.

Stone counts, based on the lithological identification of clasts in the +10 mm size fraction, in the glacigene Permo-Carboniferous Dwyka Formation proved useful in delineating the source areas for the basal diamictite unit. However, the method was found less suitable for higher stratigraphic units where erosion of older diamictons led to the mixing of debris from several source areas. By means of this technique, denudation by ice of the northeastern source area ("Ghaap" Plateau + Namaqua Mobile Belt) was able to be reconstructed.

The stone counts supplemented stratigraphic studies in assisting lithostratigraphic subdivision of massive diamictite sequences. Recognition of a valley and a platform lithofacies association, with the former having debris derived from subtillite material distributed throughout the sequence and the latter with it restricted to a thin basal zone, was also substantiated by stone counts. A severe limitation on the application of stone counts in pre-Pleistocene glacial sequences can be the lack of indicator rocks, so it is suggested instead that clasts forming diagnostic rock suites be used in defining the source areas. Palaeontological, isotopic and radiometric age data on the clasts may also assist in such studies.

INTRODUCTION

Since the previous century stone counts have played an important role in studies of the Quaternary geology of Europe and North America as they proved useful in establishing the source areas of glacial deposits, flow paths of the former ice cover and the relative age of the deposits. With the exception of Du Toit (1921) who suggested the use of indicator rocks along the northern margin of the Karoo Basin, the potential of stone counts in the glacigenic Dwyka Formation, which forms the basal unit of the 7500 m thick Karoo Sequence (late Carboniferous to early Jurassic), has not yet been fully exploited. It is in the southern part of the basin (area ca. 600,000 km^2) where, due to extensive folding, clast fabrics cannot be used to establish ice-flow patterns, and glacial pavements are absent, that there is a need to

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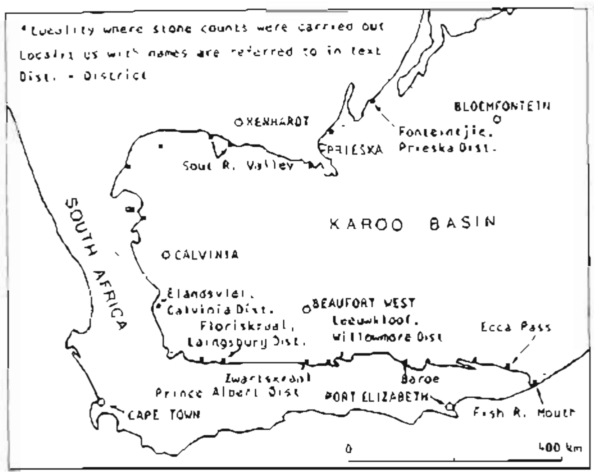

Fig. 1 Locality map of the western half of the Karoo Basin.

obtain information on the 800 m thick, massive diamictites of the Dwyka Formation.

In his study of the Dwyka Formation Stratten (1968) used stone counts based on the 50 largest clasts in a particular area, but unfortunately the counts were not carried out in a systematic way as stratigraphic control was lacking. He concluded (p. 100) that "from north to south the trend of pebble composition is inconsistent", but his results were undoubtedly negatively influenced by the methods employed.

The present study is part of a stratigraphic investigation of the Dwyka Formation in the western half of the Karoo Basin (Fig. 1) carried out over a period of five years. Stone counts were considered as an additional stratigraphic tool especially in areas where other information pertaining to the glaciation was lacking. It is not the purpose of this paper to give a resumé of our stone count results in the Karoo Basin, but rather to illustrate, by means of a few examples, the application of stone counts to the interpretation of pre-Pleistocene glacial sequences.

STONE COUNT METHODS APPLIED IN THE DWYKA

Although it is often tempting to compare the Permo-Carboniferous glaciers and ice sheets in the Karoo Basin with the Antarctic Ice Sheet, on a similar scale the entire Karoo Basin will fit into an ice-free Wilkes Basin in eastern Antarctica (Drewry, 1983). On the other hand, the characteristics of the diamictites in the southern part of the Karoo Basin and the ice-flow patterns compare favourably with the North American ice sheets of Pleistocene age (Mayewski et al., 1981). Stone count methods employed in the study of fairly recent glaciations can thus be adapted, with certain limitations, for pre-Pleistocene glacial studies.

Fresh exposures of the Dwyka Formation are sparse and the availability of

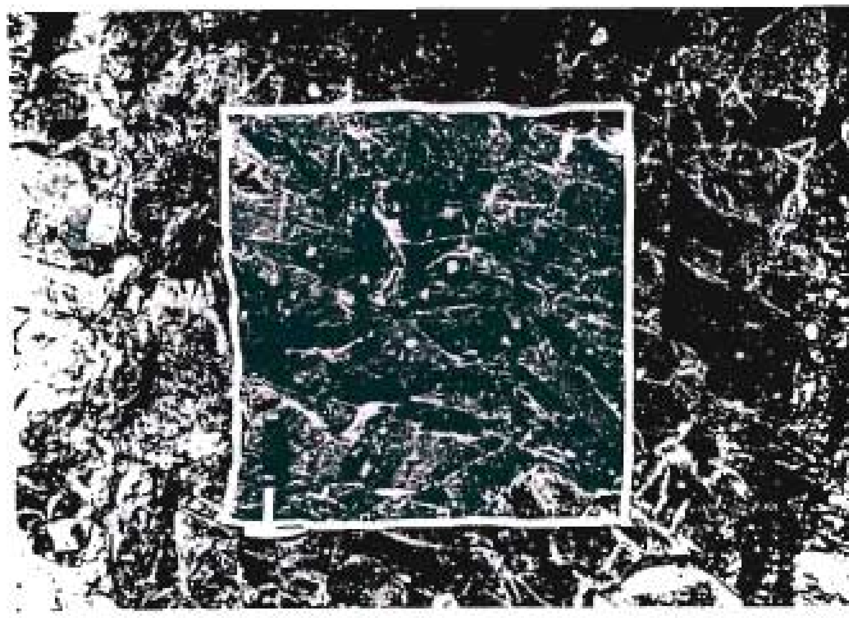

Fig. 2. Massive basal diamictite at Zwartskraal, Prince Albert District. A 1 m^2 area in which a stone count was carried out, is marked off. Fracture zones to the left and right of photograph limited the choice of the area.

artificial exposures (road cuttings and quarries), together with the clast packing density of the diamictites, commonly determined the method employed. At fresh exposures along road cuttings or in stream beds a 1 m^2 area was randomly selected and marked off by tape (Fig. 2). Within this area the composition of all clasts larger than 10 mm in diameter was recorded. Where outcrops were weathered the clasts were removed and their composition recorded at a later stage. In clast-poor diamictites (clast packing density = 2–12%) all the clasts in multiples of 1 m^2 areas were often recorded and for some dropstone argillites along the northern margin of the basin it was necessary to sample the entire quarry floor. No material was taken from the quarry walls so as to restrict the sample population to a specific stratigraphic level. This restriction often limits the number of clasts in a sample to less than 50 in the pebbly mudstones. However, it was found that counts of less than 50 stones at a sampling site gave comparable qualitative results despite such small sample numbers not giving reproduceable quantitative results.

As it was suspected that clast composition could be size dependent, the composition of all clasts with a diameter exceeding 200 mm was recorded at a few selected sections. These measurements were made in the area surrounding the 1 m^2 site where detailed counts were made.

All sample sites are stratigraphically controlled as detailed sections of the glacial sequence were measured prior to recording the clast lithologies. Counts were either carried out at evenly spaced vertical intervals (depending on the availability of outcrops) or restricted to recognisable diamictite units. Finally, in representing the data certain rock suites providing for (1) possible misinterpretations in the field (e.g. diabase or non-amygdaloidal andesite); (2) mutual relationships (e.g. chert and banded iron-formation); and (3) their uniqueness (e.g. reddish quartzite or diorite) were chosen for comparison of particular areas or stratigraphic units.

STONE COUNTS AS SOURCE AREA INDICATORS

In parts of the Karoo Basin stone counts were successfully employed to locate the direction of diamicton source areas. A basal diamictite unit with a maximum thickness of about 100 m was recognised in the southern part of the Karoo Basin (Visser and Loock, 1982). It consists of clast-rich to clast-poor diamictite which unconformably overlies quartzite and shale of the Ordovician to early Carboniferous Cape Supergroup. The diamictites are interpreted as lodgement tills (Visser and Loock, 1982), but data on the palaeoice-flow pattern are only available for Elandsvlei where glacial pavements occur and clast fabric studies could be carried out as the beds are unaffected by the Cape folding.

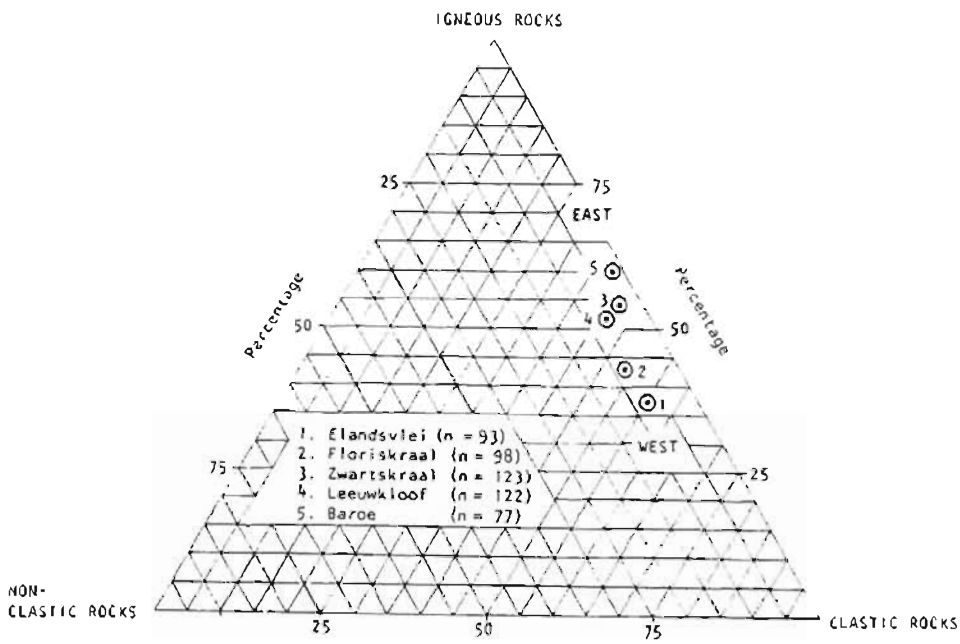

Fig. 3. Triangular plot of the total composition of clasts from the basal unit of the Dwyka Formation in the southern Karoo. Histograms of the same analyses are shown in Fig. 4. n = number of stones.

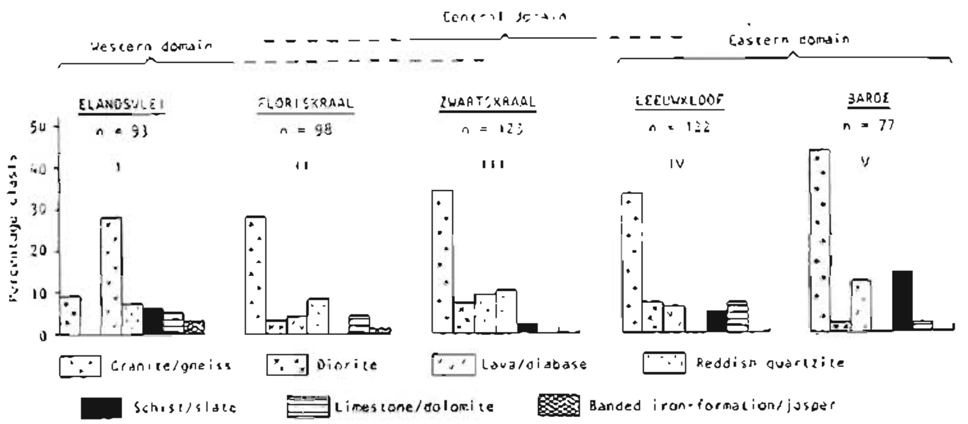

Fig. 4. Histograms of the partial composition of clasts from the basal unit of the Dwyka Formation in the southern Karoo. Seven rock suites which may be diagnostic, were chosen for comparison of the stone counts.

The results of five stone counts from within the lowermost 60 m of the unit show a linear trend on a triangular plot of the percentages of igneous, clastic and non-clastic rocks (Fig. 3). Quartzite clasts are predominant in the west, whereas granitic to gneissic clasts form the major portion of the clast population in the east. However, very little additional information on the source areas of the diamictites can be gained from such a plot.

Comparison of seven rock suites, depicted as histograms for the five sample sites (Fig. 4), shows the existence of three possible source domains. In the western domain (analysis I) lava/diabase clasts are abundant with carbonate rocks and banded iron-formation/jasper also present. A central domain (analysis III) is characterised by a high granite/gneiss clast content as well as up to 7% diorite clasts. The eastern domain (analyses IV and V) also has a high granite/gneiss clast content but with the additional attendance of fragments of diorite, carbonate rocks, schist and slate. Analysis II probably represents an overlap of the western and central domains, and the influence of the western domain may even extend to Zwartskraal (analysis III).

The reddish quartzite is only present in analyses I–III and, according to palaeoice-flow data at Elandsvlei, suggests a source to the north and northeast of Elandsvlei where, at a distance of 400–500 km, such quartzite outcrops presently occur. Extensive outcrops of diorite are unknown towards the north of the Karoo Basin and its presence in analyses II–V suggests a possible southern to southeastern source. The carbonate rocks present in analyses I and II consist mostly of stromatolitic dolomite (Fig. 5) which was derived from the "Ghaap" Plateau * located about

* The present name for the plateau is used in inverted commas to indicate that about 300 Ma ago the area included a much larger mountainous region.

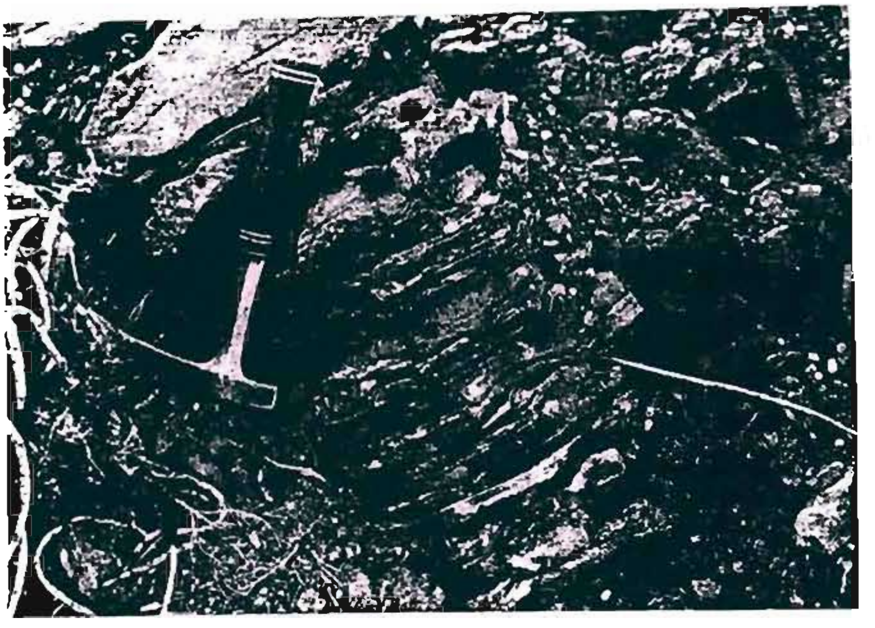

Fig. 5. Clast of stromatolitic dolomite with black chert lenses (to the right of the hammer) in the basal diamictite (*dmt*) at Floriskraal. Clast was derived from the "Ghaap" Plateau located to the north-north-east of Floriskraal.

400–500 km to the northeast of Elandsvlei. The carbonate rocks in analyses IV and V on the other hand consist of light grey limestone (Fig. 6A) of which some stones contain archaeocyathids (Fig. 6B and C). The source of these clasts was probably to the south and southeast in present-day Antarctica (Oosthuizen, 1981; Buggisch and Webers, 1982). The increase in non-durable rock-types, like schist and slate, in analyses III–V suggests a nearby source in the southeast and east as such source areas are unknown to the north of Leeuwkloof and Baroe.

It can be concluded that the basal diamictites at Elandsvlei were largely derived from the northeast ("Ghaap" Plateau source) whereas the sediments at Zwartskraal probably had a southern source. Interference of palaeoice flow from the northeast and south probably occurred from just west of Floriskraal up to Zwartskraal. The basal diamictites at Leeuwkloof and Baroe were probably derived from the southeast and east with less interference by other ice entering the basin as was the case towards the west.

An approach of this kind involves complications including the mixing of debris from several sources during transport and deposition, the complexity of ice-flow paths, and the lithological character of the debris. Mixing of debris can be illustrated by the example of some stone counts from the Eastern Cape Province where the uppermost unit of the Dwyka Formation consists of a clast-poor speckled diamictite with small deformed sandstone bodies. The speckled character of the diamictite is attributed to the presence of coarse sand-sized grains of white quartzite, milky quartz and feldspar. Four analyses from this unit show a decrease in the number of granite/gneiss clasts from west to east and a corresponding increase in the total

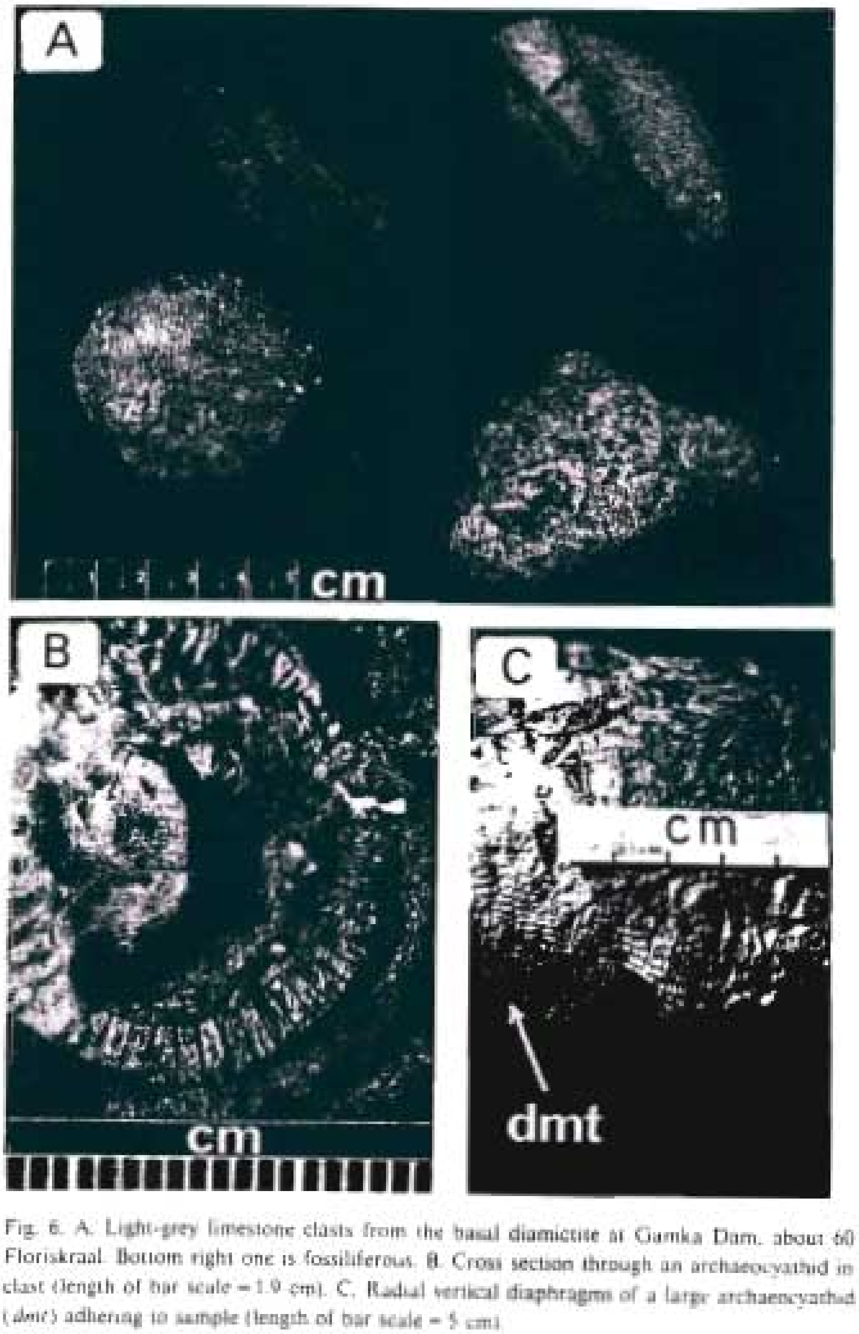

Fig. 6. A. Light-grey limestone clasts from the basal diamictite at Gamka Dam, about 60 km east of Floriskraal. Bottom right one is fossiliferous. B. Cross section through an archaeocyathid in a limestone clast (length of bar scale = 1.9 cm). C. Radial vertical diaphragms of a large archaeocyathid. Diamictite (*dmt*) adhering to sample (length of bar scale = 5 cm).

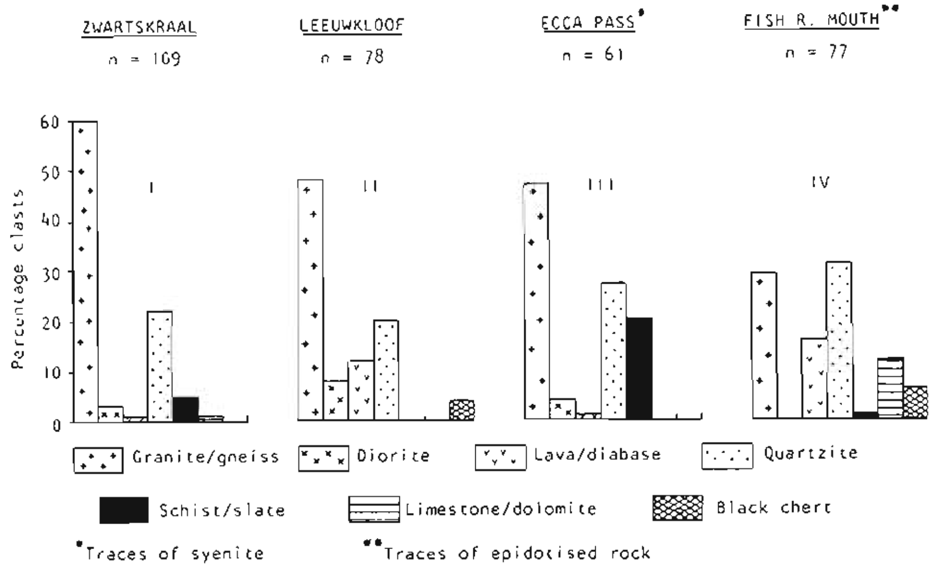

Fig. 7. Histograms of the partial composition of clasts from the speckled diamictite unit at the top of the Dwyka Formation in the southern Karoo.

quartzite clast content (Fig. 7). Although two possible source domains (a southern to southeastern one based on analyses I–III and a northeastern to eastern one based on analysis IV can be vaguely recognised, the influence of the source areas was not as well defined as in the basal diamictite. This can be attributed to the mixing of debris from different source areas as numerous fragments of diamictite and black shale are present in the speckled diamictite which suggests erosion and reworking of older diamictons.

Commonly the straight-line direction of palaeoice flow can be broadly defined, but in practice the flow paths of the ice were much more complex as the local ice-flow pattern for a region did not necessarily represent the direction to the source of the clast type. This can be attributed to the influence of the topography and the ice budget on the ice-flow direction as well as multiple glaciation of the area. This is illustrated by an example from Elandsvlei where the main sources for the Dwyka Formation (585 m thick) were the mid-Proterozoic basement rocks of the Namaqualand Metamorphic Complex and the early Proterozoic sedimentary rocks and volcanics of the "Ghaap" Plateau region. An intratillite pavement in, and clast fabric studies carried out on, the basal diamictite suggest palaeoice flow from the east (towards 281°), whereas clast fabrics from near the top of the formation suggest a palaeoice flow from the east-northeast (towards 248°). Debris derived from the "Ghaap" Plateau is restricted to the basal diamictite towards the east of Elandsvlei (at Floriskraal and in a borehole near Beaufort West), whereas north of Calvinia such debris is largely confined to higher stratigraphic units with gneissic basement

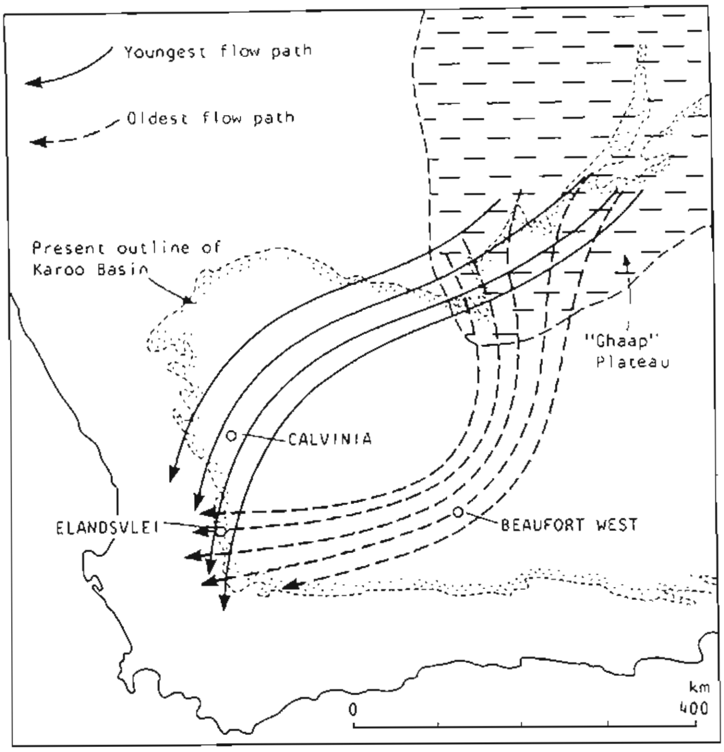

Fig. 8. Different ice-flow paths for debris from the "Ghaap" Plateau to Elandsvlei. Based on clast composition, clast fabrics, glacial pavements and basement morphology.

clasts predominating in the basal diamictite. The skewed distribution of "Ghaap" Plateau debris in the Dwyka glacials can be explained by postulating an older and a younger ice-flow path towards Elandsvlei (Fig. 8). These flow directions were dependent on the influence of different ice lobes which again was a function of the ice budget in the region.

In evaluating the source area of a diamictite based on the clast composition, it is assumed that clasts of all lithologies behaved the same during glacial transport. Flint (1971) suggested that the proportion of a given lithology in a diamictite depends on the area of outcrop of the source rock up-glacier, erodibility, durability in transport, and distance of transport. The effect of some of these parameters is illustrated by the sympathetic relationship between clast composition and size in the Dwyka diamictites (Table 1; Hall and Visser, 1984). Less durable rock types like dolomite, limestone and schist would account for lower percentages in the +200 mm size fraction than for example quartz porphyry, granite, gneiss or quartzite if the same transport distance is assumed. However, it was also found that clast size is affected by the bed thickness of the source rocks (Hall and Visser, 1984) which would explain the abundance of chert in the −200 mm size fraction (Table 1). Therefore in some

TABLE I

The relationship between clast composition and clast size in the basal diamictite at Elandsvlei [a]

Clast lithology	Relative abundance of clasts [b]		% change [c]
	< 200 mm (n = 98)	> 200 mm (n = 48)	
Quartzite	141	210	+ 49
Granite/gneiss	28	60	+ 114
Chert	21	10	− 110
Dolomite	24	0	− 2400
Schist	24	0	− 2400

[a] All clasts were distantly derived (100–500 km).

[b] Lava/diabase was taken as a unit (= 100), because it was equally abundant in both size fractions.

[c] (+) = more abundant in + 200 mm fraction; (−) = more abundant in − 200 mm fraction.

counts where a reasonable percentage of chert was found, the presence of dolomite or limestone in the source areas cannot be ruled out.

STONE COUNTS AS AN INDICATOR OF GLACIAL EROSION IN THE SOURCE AREAS

The evolution of the source area during a glaciation can be deduced if the regional ice-flow pattern remained more or less constant and a thick sequence of glacial debris accumulated. This postulation can be illustrated by means of the 585 m thick sequence at Elandsvlei which had the "Ghaap" Plateau towards the northeast as one of its major sediment sources. The glaciers, on their way from the "Ghaap" Plateau to the depository, traversed the Namaqualand basement rocks which led to the mixing of debris from these two sources in their tractional load. Presently the Ghaap Plateau consists mainly of a basal non-clastic zone (dolomite and banded iron-formation), overlain by volcanic rocks and then a clastic sequence (Matsap quartzite) dipping towards the west, but during the late Carboniferous the quartzites probably capped the less durable rock-types to build the higher ground.

In the Elandsvlei section three coarse clast-rich diamictites, interpreted as lodgement tills, were sampled and the composition of the + 200 mm clast fraction recorded. The clast percentages of four rock suites for which the straight-line distance of travel was known, were plotted against the stratigraphic thickness (Fig. 9). It was found that the percentages of granite/gneiss and quartzite clasts decrease with an increase in the stratigraphic thickness of the formation, whereas the percentages of lava/diabase and dolomite/chert/BIF clasts show the opposite trend.

This suggests, firstly, the presence of more proximally derived clasts in the lower part of the sequence and, secondly, that quartzites covered a larger area during the onset of the glaciation. Theron and Blignault (1975) found that along the western

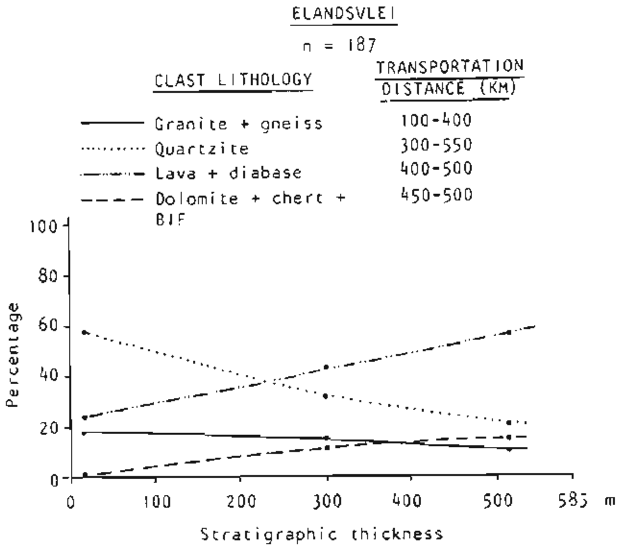

Fig. 9. Vertical variation in clast percentages of four rock suites from the Dwyka Formation at Elandsvlei. Only clasts > 200 mm in diameter were counted.

margin of the Karoo Basin upper units of the Dwyka Formation transgress the lower ones in a northward direction. This would mean that more of the Namaqualand basement became covered by diamicton as the glaciation proceeded and that the relative importance of the Namaqualand Metamorphic Complex as a source for the younger glacial units at Elandsvlei dwindled. A stage would have been reached where the reworking of older diamictons by successive ice advances and the derivation of material from nunataks were the only sources of Namaqualand basement clasts for the younger glacials. Denudation of the "Ghaap" Plateau exposed large areas of non-clastic and volcanic rocks which led to the clast increase of these rock-types in the younger glacial units at Elandsvlei. The steep trend in the graphs for quartzite and lava/diabase clasts (Fig. 9) suggests that denudation of the source area was the major contributing factor in the changing clast composition at Elandsvlei.

STONE COUNTS AS A STRATIGRAPHIC TOOL

Two applications of stone counts in glacial stratigraphy will be discussed.

(1) In the Sout River Valley, Kenhardt District, and at Leeuwkloof, Willowmore District, stone counts were carried out at regular intervals in the Dwyka Formation (Fig. 10). Afterwards, the sections were described in detail and several lithological units based on lithology, texture, colour and sedimentary structures were recognised.

An analysis of the stone counts in the Sout River valley indicated five possible groupings (Table 2). Two of the groupings (I and II) correspond with stratigraphic

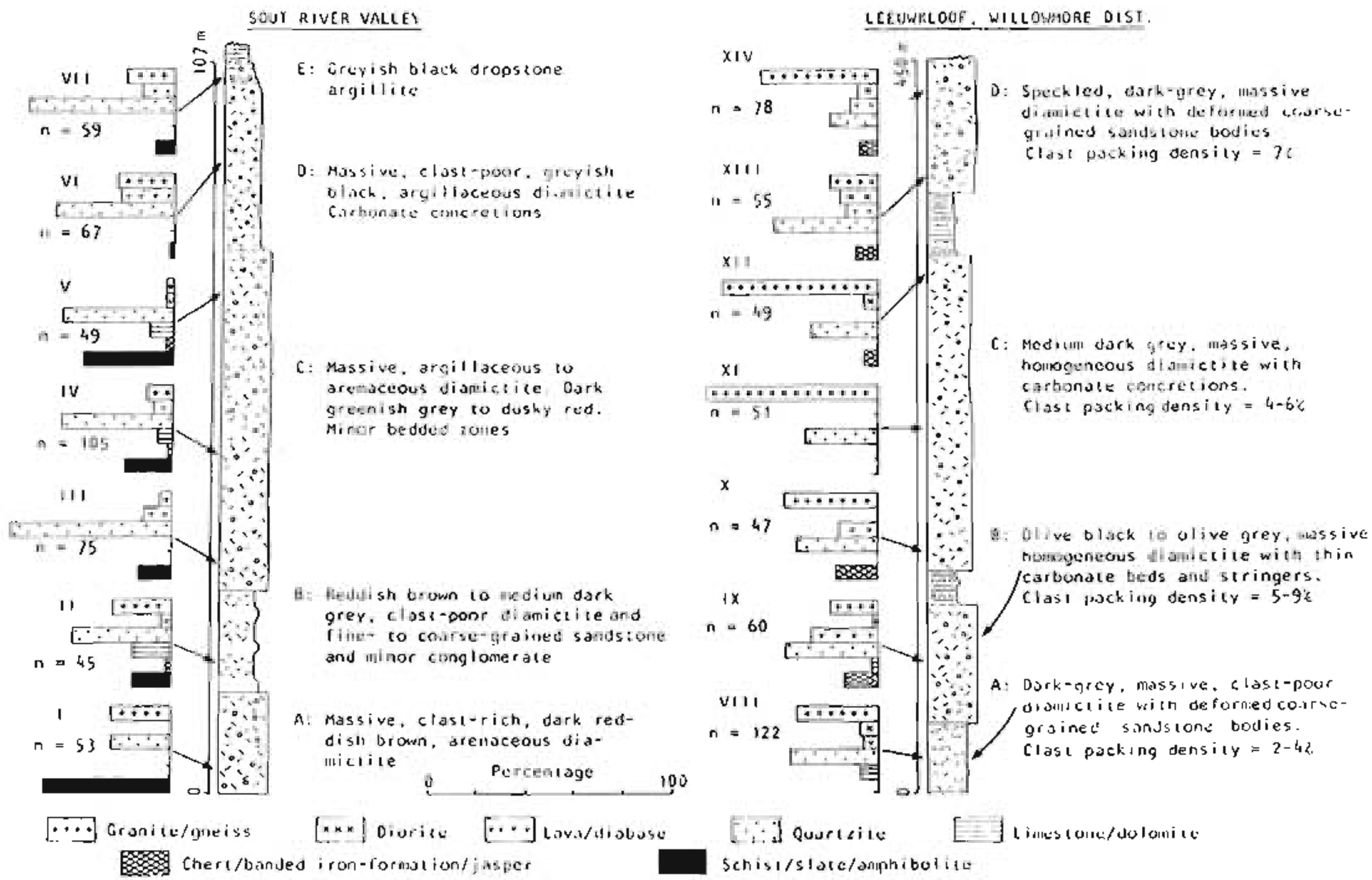

Fig. 10 Stone counts from two Dwyka sections in the western part of the Karoo Basin. The section in the Sout River valley represents a complete sequence of the Dwyka Formation, but the section at Leeuwkloof represents only the lower two thirds of the sequence. The stratigraphic units are based on the lithology, clast packing density, colour and sedimentary structures. Arrows indicate sample positions.

TABLE 2

Correlation between lithological units and percentage clast composition (Fig. 10)

Sout River Valley		Leeuwkloof	
Stratigraphic unit	Stone count grouping	Stratigraphic unit [a]	Stone count grouping
E, D	VI/VII	D	XIII/XIV
C	IV/V, III	C	XI/XII
B	II	B	IX/X
A	I	A	VIII

[a] Interbedded shale and mudstone not considered as part of a unit.

units A and B, respectively, but unit C (massive argillaceous to arenaceous diamictite; Fig. 10) comprises two stone count groupings. However, further field studies of this unit indicated the presence of stacked diamictites which would explain the variation in clast composition. The uppermost stone count grouping (VI/VII) comprises lithological units D and E. This is, however, not unexpected as the thin dropstone argillite (unit E) at the top of the formation represents the disintegration of the ice lobe that deposited the underlying massive, clast-poor argillaceous diamictite (unit D).

At Leeuwkloof stratigraphic units A and D correspond well with stone count groupings VIII and XIII/XIV, respectively (Table 2). However, stone count grouping IX/X apparently corresponds with lithological unit B and the lower part of unit C which suggests that the interbedded mudstone with thin debris-flow diamictites represents only a minor oscillation of the ice front and not a major interglacial. Field studies show similarity between the two diamictite units (B and C), except for a colour change and a slightly higher clast content for unit B (Fig. 10). The upper half of stratigraphic unit C corresponds with stone count grouping XI/XII which suggests a stratigraphic break somewhere near the middle of unit C. This was, however, not recognised in the field.

The vertical variation in clast composition can thus be of use as a supplementary tool in recognising stratigraphic units in a glacial sequence. At Leeuwkloof this distinction is less obvious as far-travelled lithologies probably became more uniformly distributed, due to multiple erosion of older diamictons, within a given ice lobe. The interaction of source areas by the interference of ice lobes may, however, reach such proportions that stone counts are unreliable stratigraphic indicators as was found by Shetsen (1984) for tills in Alberta, Canada.

(2) Visser (1983), in a study of the lithofacies distribution in the Dwyka Formation, recognised a valley and a platform facies association. Each of these facies associations consists of a number of related lithofacies representing glacial deposition in a particular environment (e.g. marine platform). In a plot of the distribution

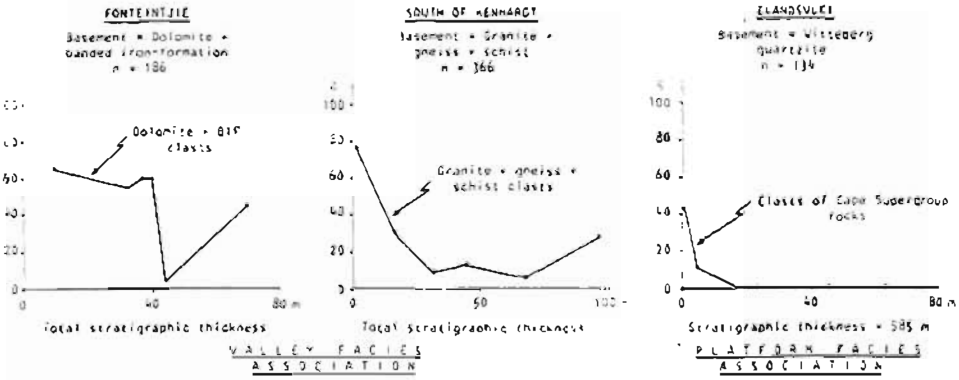

Fig. 11. Percentage distribution of subtillite rock-types present as clasts in diamictite sequences from a valley and a platform lithofacies association in the Dwyka Formation.

of subtillite rocks as clasts in the stratigraphic sequence (Fig. 11). it can be seen that. despite some fluctuations in the percentages, subtillite rocks are present as clasts throughout the full sequence along the northern margin of the Karoo Basin (Fonteintjie and south of Kenhardt) where the valley facies association crops out. However, at about 17 m above the base of the Dwyka Formation at Elandsvlei which forms part of the platform facies association. clasts derived from the underlying Cape rocks are completely replaced by distantly derived material in the diamictite. This latter pattern is applicable to all the outcrops of the Dwyka Formation along the southern margin of the basin.

The relationship between the clast composition and facies association can be attributed to the palaeotopography. During valley glaciation debris was also derived from the sides of the valley and this became eventually mixed with the basal transport load during deposition. whereas on the platform. with its subdued topography. incorporation of local bedrock material was restricted only to the tractional load which thus resulted in a thin zone of locally derived debris at the base of the glacial sequence.

CONCLUSIONS

(1) Although the coarse gravel fraction (6–63 mm) is preferred for stone counts in Europe (Lüttig. 1958; Meyer. 1983) the entire +10 mm clast population proves to be suitable for a systematic study of clast composition in the highly indurated Dwyka Formation. The size limit of 10 mm is considered the minimum for the positive identification of the lithology in the field. A study of the +10 mm fraction furthermore lessens the possibility of mixing of debris from various sources during glacial transport as well as the selective removal of certain components during weathering.

(2) Stone counts can be successfully employed, but not with the same confidence limits as in Quaternary deposits, in delineating the source areas for some of the Permo-Carboniferous glacial deposits in the Karoo Basin. Ice-flow paths are, however, commonly complex and the local ice-flow pattern did not necessarily reflect the direction to the source of the clast type. In a stratigraphic study of Palaeozoic and older glacial sequences, especially where massive diamictites are involved, stone counts are an essential tool. However, stone counts cannot be used in isolation in an effort to solve glacial problems, but must be supplemented by other lithological methods (Lüttig, 1958).

(3) Stone count studies in pre-Pleistocene glacial sequence, however, have numerous limitations. As the source areas of the glacial sediments are no longer recognisable, the application of indicator rocks which is a common practice in Europe (Lüttig, 1958), is not possible. The use of diagnostic rock suites can, to some extent, replace this shortcoming. As the postulated source areas were probably denuded during, and subsequent to, glaciation their original elevation, extent and composition are largely unknown. Without this knowledge the degree of mixing of debris derived from several sources cannot be fully interpreted.

(4) In the Permo-Carboniferous Dwyka Formation more elaborate studies on the clasts have to be done. The palaeontology of the non-clastic rocks as well as their isotopic composition need further study, whereas radiometric age determinations on the wide variety of gneisses, granites, diorites and lavas present as clasts in the diamictite would indicate whether they came from the Archaean craton, late Proterozoic basement or an early Palaeozoic orogeny. All these lines of additional evidence should make for more accurate predictions on the distribution of the possible source areas.

ACKNOWLEDGEMENTS

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Chapter 9

Other Papers

"Who are you going to believe, me or your own eyes?"

Groucho Marx

Material presented in this section comprises papers, reports, reviews and other such documents that do not readily fit to the preceding eight chapters. In one way or another, though, they do fit to the broader pattern of information presented in this thesis. The first paper is simply a report on the Binghampton meeting on periglacial geomorphology whilst the second relates to thoughts on the journal *Polar Geography* after 20 years. The following paper details evidence regarding Quaternary changes to sea levels on sub-Antarctic Marion Island as a result of glacial, isostatic and eustatic variation. This paper includes the use of fossil seal wallows as landforms to help identify former sea levels (as per discussions in Chapter 6 - e.g. Hall and Williams, 1981). Some of the level changes are also a result of the glacio-isostatic responses discussed in Chapter 8 (e.g. Hall, 1982). There then follows a paper dealing with conservation issues in the sub-Antarctic relating also to Marion Island. The other 1983 paper questions the status of geomorphology at universities in South Africa where, despite a rich history in this field, recent undertakings had been minimal and qualitative. Finally, there is an invited rapporteur presentation, relating to terrestrial environments, from the "Arctic and Antarctic: A Study in Contrasts" conference held in Ottawa. As much as the two are polar regions, the Arctic and the Antarctic differ enormously in terms of their environments and their future. The document by Loken and Hall (2000) is a follow-up to the paper by Hall (199b) and the volume in which it appears. While Hall (1999b) is a rapporteur discussion from the conference ("Poles Apart: Arctic and Antarctic. A Study in Contrasts"), Loken and Hall (200) is a resulting working document used as a basis for strategy and funding within Canada. The reference Hall, *et al.,* (In Press f) is to a bibliography on permafrost and periglacial landforms and processes of the southern hemisphere that was undertaken as part of the mandate of the Southern hemisphere Working Group of the International Permafrost Association. Although the work, as presented with this thesis, is actual "work in progress' (as it will continue to expand until the 2003 final deadline for production as a CD-ROM) it already has some 1000 references and constitutes the first such undertaking for periglacial attributes of the Southern hemisphere. It comprises a substantial body of material in a host of different languages and, although 'only' a bibliography, is nevertheless cited here as it is a "first" and comprises a deceptively extensive and difficult piece of research.

Material presented here includes:

- Finkler, H., Fondahl, G., Hall,K. And Polezer, G. 1996. *Polar Geography* at 20 Years. *Polar Geography*, 20, 1-2.
- Hall, K.J. 1977. Some observations on the former sea levels of Marion Island. *South African Journal of Antarctic Research*, 17, 19-22.
- Hall, K.1999b. Terrestrial Environments, *In* A. Lewkowicz (ed.): *Poles Apart: Arctic and Antarctic. A Study in Contrasts.* Ottawa, University of Ottawa Press, 213-216.
- Loken, O and Hall, K. (eds) 2000. *Antarctic and Bipolar Science.* Canadian Committee for Antarctic research, Canadian Polar Commission,

38pp.

- Hall, K., Boelhouwers, J. and Lamont, N. In Press f. Bibliography on Southern hemisphere permafrost and periglacial landforms and processes. *International Permafrost Association, Southern hemisphere Working Group*, 86pp. [Presented as a separate volume owing to its size]

The above is a fairly eclectic mixture of papers but they do relate, in some manner, to the overall presentation of this thesis and, in so doing, often show the degree of involvement by this author.

POLAR GEOGRAPHY AT 20 YEARS

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This volume celebrates the 20th anniversary of *Polar Geography* (known earlier as *Polar Geography and Geology* and, much earlier, by the present name, *Polar Geography*). It also initiates the second year of publication under a new editorial management with a revised set of goals. For two decades this journal has been most recognized for its contribution in facilitating Western scholars' access to Russian scientific articles on the polar regions of the globe. Given the political divide that hindered Western field research in a significant sector of the Circumpolar North, access to such material became indispensable for furthering our understanding of similar processes across cold regions.

The dramatic changes in the former Soviet Union and opening of the Russian North has allowed increased work by Westerners, often in collaboration with Russian colleagues. It also allows Russians to publish freely in Western journals. To this end, while *Polar Geography* will still provide translations of key articles appearing in Russian scholarly journals, it will increasingly publish original research by Russians and by Westerners working in the Russian North (see the Vilchek et al., Pavlov, and Cooper et al. articles in this issue). The journal will continue to include articles on the other parts of the Circumpolar North as well as the Russian North (as it has since its launching), but will no longer focus on the southern polar region.

The decision to constrain the geographical focus of the journal complements our intent to not only pursue a focus on the Russian North in particular, but to offer a multidisciplinary spectrum of articles that embrace more completely the human as well as physical dimensions of Arctic and Subarctic environments. While publishing articles in the field of human geography occasionally, the journal's emphasis has been on physical geography and its cognate disciplines (e.g., resource geology). To emphasize our intent to broaden the topical coverage, we reverted to its earlier name. This certainly does not signal a retreat from

Polar Geography, 1996, **20**, 1, pp. 1-2.

resource geology and related aspects of physical geography, but rather underscores the inclusion of the social-sciences component, which will provide for a more complex and integrated understanding of the Circumpolar North.

To accomplish our aim of offering the basis for such an understanding, *Polar Geography* has identified three broad themes of special significance to the Circumpolar North at the end of this millennium and beginning of the next:

- climate change, and especially its impact on permafrost
- environmental impacts of anthropogenic activities
- northern and aboriginal policy.

We will devote special attention to these themes, while publishing articles of excellent scientific merit in resource geology and other fields as well.

As we approach the dawn of the new millennium and face increasing challenges of climatic change, resource development, and social transformation, the Circumpolar North assumes increased importance for scientists and policymakers alike. We look forward to providing a dynamic and engaging forum of leading scholarly research that addresses these challenges.

Reprinted from S. Afr. J. Antarct. Res., Vol. 7, 1977

Some observations on the former sea levels of Marion Island

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Investigation of raised beaches together with mathematical extrapolation from a river long profile suggest two definite former sea levels with the possibility of a third. True raised beaches were found at c. *+3 m and* c. *+6 m with extrapolated levels of +5,9 m and +10,8 m. Stone roundness and flatness indices were used as additional aids to differentiation of various levels. Comparison is made with former sea levels found on other sub-Antarctic islands.*

'n Ondersoek van strandterrasse met behulp van wiskundige ekstrapolering van 'n rivierlangse profiel af dui op twee beslste vroeëre seevlakke, met die moontlikheid van 'n derde. Egte strandterrasse is by c. *+3 m en* c. *+6 m gevind met geëkstrapoleerde vlakke van +5,9 m en +10,8 m. Klippe se rondheids- en platheidsindekse is as verdere hulpmiddels by die onderskeiding van die verskillende vlakke aangewend. 'n Vergelyking word getrek met vroeëre seevlakke wat op ander sub-Antarktiese eilande gevind is.*

Introduction

As part of the study of the Quaternary history of Marion Island (46°54'S, 38°45'E) an investigation of former sea levels was undertaken. Observations by Verwoerd (1971) had shown no evidence for former levels and he suggested that this may be due to the "differential tectonic settling of the volcano . . . so that they were drowned". The volcanic nature of Marion is such that many of the recent black lava flows, which line the coast, could be morphologically mistaken for a former marine level owing to their bench-like form. However, despite the problems of possible tectonic disturbance and the extensive lava flows a number of distinct raised beaches could be recognised at several locations around the island.

Two main methods were used to determine the former marine levels, namely direct observation (and subsequent survey) and mathematical extrapolation from the long profile of a river. As an aid to direct observation use was made of stone roundness and flatness indices to help differentiate the

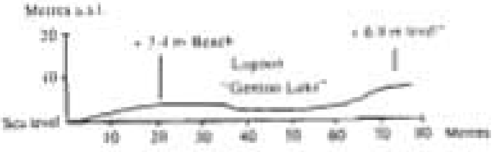

Fig. 1. Surveyed profile of the raised beach at Transvaal Cove.

various levels. One river was surveyed and used in the calculations, the results of which are in good accord with those found for the actual beaches.

Results

True raised beaches

True raised beaches were observed at a number of locations around the island. At all sites wave-smoothed rocks, wave-rounded pebbles and boulders, and wave-cut cliffs at the back of the beach were observed. At Transvaal Cove (Fig. 1) and Trypot Beach raised beaches were surveyed and heights of +3,4 m and +2,9 m, respectively, were found. A further level of +6,9 m was indicated at Transvaal Cove but the lack of rounded pebbles and boulders suggests that this may represent a lava flow rather than a true marine-cut level. Just to the north of Cabbage Point a classic raised beach sequence (Fig. 2) from +3,3 m to +6,1 m was found and surveyed (Fig. 3). The lower beach started above a wave-cut cliff and rose, across a pebble-bould beach, to a cliff-backed level at +6,1 m.

Fig. 2. A view across the raised beach sequence to the north of Cabbage Point.

At this location stone roundness (Pi), using Cailleux's equation $Pi = (2r/a) \times 1000$ and flatness (Fi), using $Fi = [(a + b)/2c] \times 100$ indices were determined for both levels. The following results were obtained.

+3,3 m level	+6,1 m level
Pi = 422,9 (s = 125,6)	Pi = 381,2 (s = 121,3)
Fi = 448,1 (s = 61,0)	Fi = 492,0 (s = 117,6)
n = 100	n = 100

The above figures clearly indicate differentiation between the two levels, with the greater roundness and lower flatness indices at the +3,3 m level. In addition, Pi indices given below were also calculated for the present beach and the +2,9 m level at Trypot Beach:

Present beach	+2,9 m level
Pi = 357,5 (s = 120,3)	Pi = 417,1 (s = 119,5)

Again a clear-cut distinction is seen between the two levels but it is interesting to note that the mean Pi index is similar for the +2,9 m level and the +3,3 m level (417,1 and 422,9 respectively). However, owing to the variations in the structures of the lavas at the two locations the resulting beach deposits can only be assumed with 80 per cent certainty to come from the same population (Mann Whitney U- test with level of significance of U calculated using z-score). Thus there appears to be evidence to indicate two levels, *c.* +3 m and *c.* +6 m, above the present-day beach.

These two levels were observed at several locations around the island (Fig. 4). Between Transvaal Cove and Trypot Beach a small beach was found at +6,1 m cut in the black lava and backed by a 3-m cliff. The whole coastline from Cabbage Point up to Skua Ridge shows a *c.* +3 m level with occurrences of the +6 m level. The two levels were also observed at the following locations: Watertunnel Stream, Goodhope Bay, Kildalkey Bay, Bullard Beach, Ships Cove, Fur Seal Bay, and around the edges of Cape Davis. Both levels are thought to occur at Kaalkoppie and Swartkop Point but observations here were limited.

An interesting observation of what may be 'fossil' seal wallows, which are now well above the present sea level, was made. Four large (up to 800 m²) abandoned wallow complexes were found which are no longer used by the seals and in which plant regeneration has reached an advanced stage. The locations of these wallows suggest that they were in active use at the time of higher sea levels when the present wallow areas would have been under water. Above Transvaal Cove, where at the present time there is a large concentration of seals on the +3 m beach and up to +6 m, there is an abandoned wallow area recognisable at +10 m a.s.l. A similar type of area occurs at the back of Trypot Beach at approximately +7 m a.s.l., whilst immediately to the back of the raised beach sequence shown in Fig. 2 there is a 'fossil' wallow which even shows an old 'crawlway' from the beach top into the wallow. Logically the seals must have used wallows at higher elevations at times of higher sea levels and the situations described above appear to be good evidence.

River long profile

The long profile of the Van den Boogaard River, close to

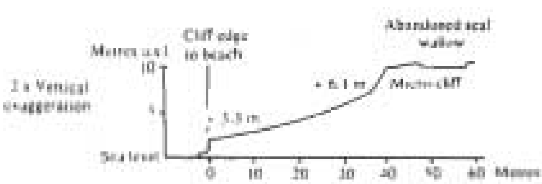

Fig. 3. Surveyed profile of the raised beach sequence for the unnamed beach to the north of Cabbage Point.

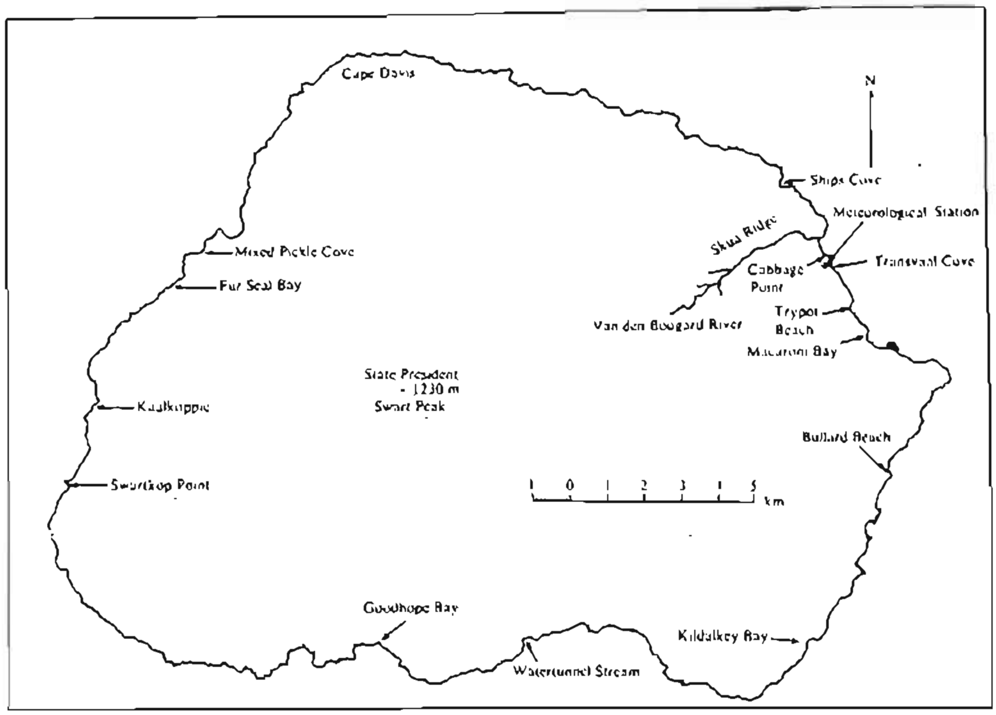

Fig. 4. Map of Marion Island to show the named locations where raised beaches have been found.

the Meteorological Station, was measured by means of a Kern level. From the long profile obtained a number of distinct nick-points were discernable (Fig. 5). Whilst an exponential curve (Table 1), with a coefficient of determination (r^2) of 0,83, best fitted the whole profile a linear equation was found to fit each segment best.

A line generated for the seaward segment (1st segment) suggests a former level of +5,9 m with an r^2 of 0,98 (Table 2). The inland (2nd) segment, when projected, gives an extrapolated level of +10,9 m with an r^2 of 0,95 (Table 3). Thus two levels, +5,9 m and +10,9 m, are suggested. The former is of great interest as it is so very close to the values found for the true raised beaches, namely c. +6 m. With the close correspondence of this extrapolated level to those actually found it is tempting to assume that the +10,9 m value may also be indicative of a former level. In fact a wave-cut platform at Macaroni Bay does reach to +10 m and may be related to this extrapolated level.

Conclusions

It would appear that there are two distinct former sea levels on Marion Island with the possibility of a third; namely at

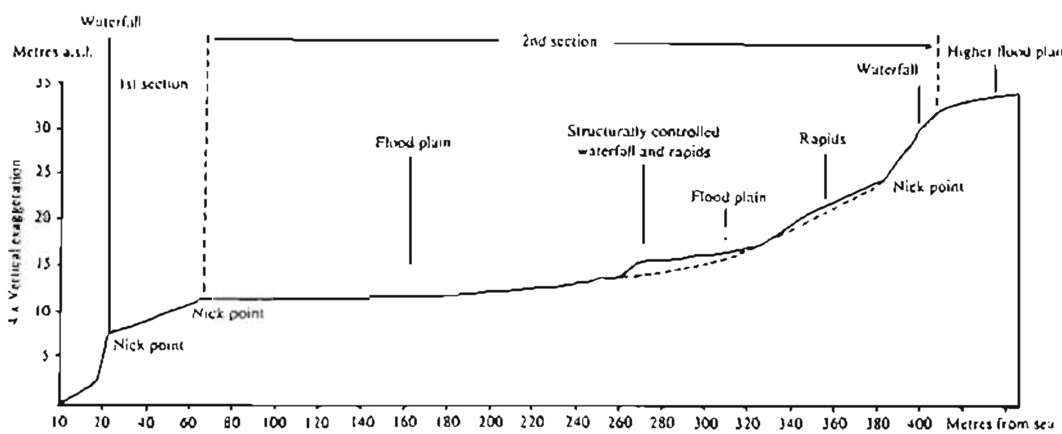

Fig. 5. Levelled profile of the Van den Boogaard River from which the former sea levels were extrapolated.

Table 1

Exponential curve generated for the whole of the measured profile of the Van den Boogaard River

x	y
70	11,2
100	11,3
140	11,4
180	11,6
200	11,8
280	13,8
320	16,5
340	18,1
380	24,0
400	30,0

x = distance inland from present sea level (metres)
y = height above present sea level (metres)

constants: $a = 7{,}853$
$b = 0{,}00269$
$r^2 = 0{,}8228$
$y = 7{,}853 \exp(0{,}00269\,x)$

Table 2

Linear regression of seaward section of long profile of river

x	y
70	11,2
60	10,0
50	9,3
40	8,8
30	8,0
20	7,5

x = distance inland from present sea level (metres)
y = height above present sea level (metres)

$a_0 = 5{,}92$
$a_1 = 0{,}07$
$r^2 = 0{,}98$
$y = 0{,}07\,x + 5{,}92$

Extrapolation of former marine level:

x	$\hat{y}$
10	6,63
5	6,28
0	5,92

Table 3

Linear regression of inland section of long profile of river

x	y
70	11,2
100	11,3
140	11,4
180	11,6
200	11,8

x = distance inland from present sea level (metres)
y = height above present level (metres)

$a_0 = 10{,}86$
$a_1 = 0{,}0043$
$r^2 = 0{,}95$
$y = 0{,}0043\,x + 10{,}86$

x	$\hat{y}$
60	11,12
50	11,08
30	10,99
10	10,91
0	10,86

c. +3 m, *c.* +6 m and *c.* +10,9 m. A +3 m level has been noted on Kerguelen and Crozet where it is considered to be of eustatic origin (Nougier, 1970). It has also been suggested (Bellair, 1969) that this +3 m level relates to the Climatic Optimum (~ 5500 B.P.). Nougier (pers. comm.) has noted a +6 m level on Kerguelen, Crozet and St. Paul-Amsterdam and suggests that the +11 m level can be recognised on Crozet. Sugden and John (1973) found both *c.* +3 m and *c.* +6 m levels in the South Shetlands but obtained dates for these levels of *c.* 300 B.P. (A.D. 1650) and 750 to 500 B.P. (A.D. 1200 to 1450) respectively. Araya and Hervé (1966) noted +2,7 m and +5,4 m levels on Elephant Island in the South Shetlands. Finally, Everett (1971) describes levels of +6,1 m and +10,6 m for Livingston Island.

Unfortunately no specific dates are available for the Marion levels so it is difficult to correlate them with those found on other islands or to suggest whether they are related to the Climatic Optimum or the more recent dates of Sugden and John (1973). However, despite the doubts of Verwoerd (1971) it would appear that raised beaches can be recognised around the island and that their accordance in height suggests that Marion has been tectonically stable during and since their formation. A detailed investigation of the former sea levels of Marion Island, especially if dates can be produced, would be most profitable for a better understanding of the post-glacial sea level changes of the southern hemisphere.

Acknowledgements

I thank Professor van Zinderen Bakker, Director of the Institute for Environmental Sciences, who directed this work and read drafts of the paper. The work and help of the Department of Transport is gratefully acknowledged.Thanks go also to Wolf Gruellich (Marion 32), who helped with all the surveying.

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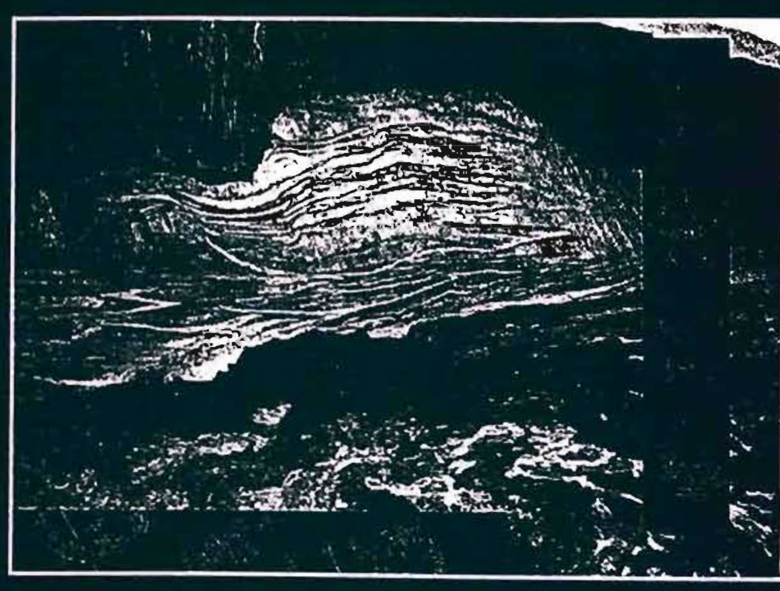
POLES APART

A Study in Contrasts

Edited by
Antoni G. Lewkowicz



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Theme 3: Polar Ecology and Resources

Theme 4: Aspects of Governance in Polar Regions

Theme 5: Protection of the Terrestrial Environment

Kevin Hall

COMMENTARY: PROTECTION OF THE TERRESTRIAL ENVIRONMENT

As Wally Herbert (1978, p. 6), the cartographer/explorer of both polar regions, suggests, the only real parallel between the Arctic and the Antarctic is that they are both polar. If these two regions are so dissimilar, then, in terms of the protection of their terrestrial environments, there might be stronger similarities between the Arctic and Bikini Atoll (despite the obvious climatic differences) and the Antarctic and Mars. The former involved the role of nations, particularly of one nation impacting on another both directly and indirectly (i.e., use of one person's land by an extraneous national group and the results of that use impacting on groups/nations beyond the immediate area) to the detriment of the environment. The second comparison has all territorial claims held in abeyance and so there are, at least from the purist viewpoint, no direct national problems nor impacts of one upon another. Rather, as Chester (1991, p. 118) states "The Antarctic Treaty System has been, and continues to be, a very important world diplomatic forum that has helped manage the only continent on Earth that is 'owned' by no one." In that sense, the Antarctic has been, and continues to be, a "continent for science" as too, at this time, is the planet Mars. Neither has yet been impacted by nuclear weapons or military activity and there are strenuous endeavours to protect the environment; so much so that Van der Lugt (1997) states "The Antarctic Treaty System can today be described as an international environmental regime." The same cannot be said for either the Arctic, or those Pacific atolls, where indigenous peoples have been severely impacted by military and nuclear undertakings in which they had no say. Thus, despite the

Poles Apart: A Study in Contrasts Edited by A.G. Lewkowicz

seeming non sequitur of my statement, the common assumption of similarities between the poles may be more apparent than real.

In terms of governance, the distinction is between the laws of individual countries versus that of a singular controlling body (the Antarctic Treaty) such that, at least under the Treaty System, the Antarctic is not subjected to the demands of nationalism or economics. It is, perhaps, colonialism of the "purest" kind. How much longer this can continue is moot but the reaffirming of the Antarctic Treaty does bode well for the near future at least. This is very different from the dynamism in the Arctic where the break-up of the former Soviet Union, coupled with the increasing demands of indigenous peoples, and the exploitation of natural resources, has produced a range of "stakeholders," all vying for the same environment. The presence of indigenous peoples and mining in the Arctic also provide a major contrast with the Antarctic. The South has no indigenous peoples. "In a place where there has never been a native people, nor permanent inhabitants, everyone strictly speaking, is a tourist" (Chester, 1991, p. 121). The true, fee-paying, tourist to the Antarctic is perhaps comparable to the exploiters of the Arctic: they exert demands regarding access and facilities. However, as with some arctic resource extractors, they can also bring positive attributes—jobs in the North, and recognition of, and hence the desire to preserve, the unique environment of the South. So far, the relatively exclusive/elitist nature of antarctic tourism has worked in favour of environmental preservation. Although the Antarctic Treaty has no fixed way of regulating tourism, the majority of companies go out of their way to follow any and all proposed environmental protection guidelines. Further, that very exclusiveness (based largely on time and costs) has meant that many important upper management and government decision-makers from around the world have been "educated" with respect to the pristine nature of the Antarctic and the need to preserve it as a World Heritage. That same treaty has also negated, so far, mineral exploitation and thus the economic demands associated with this have not impacted on the environment. This stands in stark contrast to the extensive and varied exploitation undertaken in the Arctic, often to the detriment of the terrestrial environment. Another interesting contrast is that whilst the demands by the indigenous peoples of the North in regard to their environment are made through different regulating bodies within different countries, as a "continent for science," it is the scientists in the Antarctic who (in a sense) constitute the "indigenous peoples" and who make their "demands" through a single regulatory, non-political, non-national body, the Antarctic Treaty.

A key issue here, with respect to the environment, is that terrestrial life in Antarctica is unlike that in all other major regions of the world.

Here discussion is with respect to continental Antarctica and the issue of the sub-antarctic and maritime antarctic regions are not considered, although they constitute a significant area and a major environmental concern. Antarctic vegetation is composed almost exclusively of low-lying mosses and lichens and there are no vertebrates that do not rely on the marine environment for food. Terrestrial life, excluding species of snow-dwelling bacteria and algae, is restricted to the small area (approximately 2–3 percent of the continent) that is snow-free in summer. There are only two flowering plants at a few scattered localities and terrestrial life consists only of springtails (about 2 mm in length) and mites! It is a system so fragile and close to the margins of survivability that, in many cases, it cannot recover from human impact; even the introduction of external bacteria from humans may impact and alter the terrestrial system. Climatic change could have a major effect on plant life whereby the available propagates from non-indigenous species are more able to compete than are the indigenous ones (Hall and Walton, 1993). This is very different to much of the Arctic, even the High Arctic, where the organisms are less "fragile," more complex and diverse and, in some localities, already greatly affected by extraneous species. Even though plant species diversity has always been less in the Antarctic than the Arctic, perhaps the current antarctic environment gives a snapshot of the Arctic at the end of the last glacial.

In spite of arguments to the contrary, I think there is still a good case to be made for the Antarctic providing an analogue for some attributes of the Arctic at the end of the last glacial and that is one reason why the Antarctic is so important for the Arctic. Certainly antarctic scientists do not, and in fact cannot, undertake their work without recourse to arctic information. The corollary is not so: few arctic scientists refer to or even know of their antarctic counterparts. In fairness this is partly due to the disseminated nature of antarctic publishing and that rarely, unlike this meeting, do the two ever function in the same forum. However, for many terrestrial environmental situations in the Arctic, the present-day conditions in part of the Antarctic (and the Antarctic encompasses a range of environments from dry, hyper arid through to cool, extremely wet) provide a good analogue for the environment as the arctic glaciers retreated. Periglacial conditions, particularly the cold, dry kind that transformed to cool and wet, can be found along the climatic gradient of the Antarctic—from sand wedge polygons through to mud flows. Certain landforms (e.g., cryoplanation terraces) that are argued, by some, not to be operative in the Arctic are certainly active in the Antarctic (see Hall, 1997) and so they do provide a modern analogue for European conditions at the end of the last glacial. Clearly this does not and cannot apply to all aspects, but the very severity of antarctic conditions must be akin to those that no longer exist in the North but for which there is evidence of their former presence. Here between poles, co-operation brings the Arctic and the Antarctic, through the continuum of time, closer together.

So perhaps the similarities are fewer than we first assume, and thus the very title of this Symposium—A Study in Contrasts—is perfect. However, by recognizing these differences and utilizing complementary experiences from both polar regions, we may be able to better preserve and, at the same time, use the poles for the future of all. Encouraging polar scientists, peoples, and politicians to recognise both these contrasts and mutual benefits will, with a move away from the parochialism that the groups at both poles seem to have, help us to better understand our planet and to maintain the polar environments against future pressures.

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Canadian Committee for Antarctic Research
Comité canadien de recherches antarctiques

ANTARCTIC AND BIPOLAR SCIENCE

Report of a Workshop
held at the
Arctic Institute of North America
Calgary, Alberta
October 16, 1999

Edited by Olav H. Loken and Kevin Hall

CANADIAN POLAR COMMISSION
COMMISSION CANADIENNE DES AFFAIRES POLAIRES

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Foreword

In July 1998, Canada became a full member of the Scientific Committee on Antarctic Research (SCAR), with the Canadian Polar Commission (CPC) as the adhering body. At the same time, the Canadian Committee for Antarctic Research (CCAR) was established as Canada's National Antarctic Committee for SCAR, and began developing a strategic plan for Canada's participation in antarctic and bipolar science. A discussion paper, *Antarctic and Bipolar Science: A Strategic Plan for Canada*, was distributed in the late summer of 1999 as a means of stimulating discussion and obtaining input from a wide group of stakeholders; much valuable feedback was received.

For more detailed discussions, CCAR arranged a workshop at the Arctic Institute of North America, University of Calgary, in October 1999. A number of Canadians with firsthand experience in antarctic and arctic science were invited. Roughly half the 19 participants were from the academic sector, representing eight Canadian universities. Other participants were scientists from four federal government departments and agencies, and representatives of two Canadian tour companies operating in Antarctica. The following is an account of the workshop.

Introduction

The workshop opened with an overview of the Canadian Polar Commission's perspective on polar science, presented by Commission Chairperson Michael Robinson. Professor Warwick Vincent, Chair of CCAR, provided additional background on the development of the discussion paper. As a prelude to future planning workshop participants were then invited to outline their current and proposed activities with respect to antarctic science; summaries of these are reprinted in Section 2 of the report. (Several groups active in antarctic research were not represented at the workshop; to the extent that information is available, activities at seven universities and two private companies are mentioned briefly.) In addition to research, scientific issues related to the Antarctic Treaty System were also presented.

Canadian antarctic researchers participate in several international research projects and are involved in a wide range of activities within several scientific disciplines (e.g., studies of UV effects on biological systems; genetic characteristics of seal populations; offshore sediment sampling in areas until recently covered by large ice shelves; studies related to Lake Vostok; permafrost investigations; and weathering studies.) Lakehead University is planning a second field trip to Antarctica, and a private company plans to arrange educational tours for high school students starting next austral summer. A private company has announced plans to start an antarctic krill fishery in late 2000.

In the second half of the workshop participants addressed issues related to the discussion papers and the means by which Canada can benefit through future participation in antarctic science. After considering several options, the group concluded that, in view of the important roles the two polar regions—Arctic and Antarctic—play in the global ecosystem, Canadian scientists should seek increased participation in Antarctic and bipolar science. Canada's commercial interests and treaty obligations in Antarctica lend further support to this argument. It was felt, as well, that Canadian participation should be based on Canada's expertise in arctic science. An essential prerequisite for increased participation in antarctic science would be the strengthening of Canada's arctic science programs which have been seriously eroded by recent budgetary cutbacks. Increased awareness of antarctic issues among the public, bureaucrats, politicians, and scientists is also required to overcome the perception that funding provided for antarctic research automatically reduces allocations to arctic studies. This discussion is summarized in Section 3 of the report.

The report includes five appendices: *Workshop Agenda; Workshop Participants*; *Canadian Contributions to Antarctic and Bipolar Science,* which lists 78 publications written since an earlier bibliography was compiled in mid-1997; *Canadian Scientists in Antarctica During the 1999–2000 Austral Summer*, which lists 18 individuals; and *List of Acronyms*.

Opening Statements

1.1 Welcome

Mike Robinson
Chairperson, Canadian Polar Commission
Executive Director, Arctic Institute of North America

Mr. Robinson welcomed everyone to AINA's facilities, located on the campus of the University of Calgary, and provided a brief overview of the Institute. AINA was created by Act of Parliament in 1945 as a non-profit, tax-exempt research and educational organization. Originally based at McGill University in Montreal, the Institute moved to the University of Calgary in 1976. In 1979 the Institute became part of the University of Calgary as a university research institute.

The Institute's mandate is to advance the study of Canada's North through the natural and social sciences, and the arts and humanities, and to acquire, preserve, and disseminate information on physical, environmental, and social activities in the North.

Kluane Lake Research Station (KLRS) is one of two research facilities operated by AINA. Established in 1991, near the Alaska Highway, 220 km north-west of Whitehorse, Yukon, on the south shore of Kluane Lake, the station has fostered research projects in a number of disciplines—glaciology, geomorphology, geology, biology, botany, zoology, hydrology, limnology, climatology, high-altitude physiology, anthropology, and archaeology. Kluane Lake Research Station is supported by an infrastructure grant from the Natural Sciences and Engineering Research Council.

Devon Island Research Station (DIRS) is located on the north coast of Devon Island, Nunavut, approximately 315 km north-east of Resolute. Since its establishment in 1960, the station has provided a base for numerous research programs on all aspects of the Truelove Lowland environment and on the interrelationships between the glacier ice of Devon Island, the marine environment of Jones Sound, and the ambient atmosphere.

Mr. Robinson noted that Canada has invested heavily in the intellectual capital and technical expertise of its polar scientific community. However, research and development in the Arctic is hampered by an aging support infrastructure, a situation which demands that Canada adopt a strategy to bolster the state of Arctic research and support the intellectual growth and regeneration of the scientific community in the North. Canada's strengths in arctic science should serve as a support to important research initiatives in Antarctica. This will require organizational development, fiscal planning, and a coherent strategy for raising public awareness. Canada must look beyond the traditional parameters of research and funding strategies to find new initiatives. Bipolar science is, in fact, global science, and Canada can and should be a leader in this area.

As Chairperson of the CPC, Mr. Robinson thanked the members of CCAR for the important work they have undertaken with respect to Antarctic research, and for keeping the Commission apprised of Canadian research activities in the region. He indicated that the Commission is impressed with Committee's work and is looking forward to carrying on a productive and rewarding partnership with CCAR as the national advisory body to the Commission on Antarctic issues.

Mr. Robinson outlined the Polar Commission's mandate for monitoring the state of Canadian polar knowledge and reporting regularly and publicly. He noted that the CPC is a "multi-tasking" body, advising the Minister of Indian and Northern Affairs on polar matters, promoting the development and dissemination of polar knowledge, providing information on polar research to Canadians, enhancing Canada's profile as a circumpolar nation, and promoting co-operative initiatives among organizations involved in polar research. As such, the Commission engages a range of intellectual resources, from traditional ecological knowledge and wisdom, to scientific and professional expertise, to governmental and policy-making capacities.

In February 1999, a new board of directors was appointed to the Commission: Josie Sias, a Crow Clan elder from Yukon; Richard Binder, a traditional Inuvialuit resource person; Julie Cruikshank, an anthropologist from the University of British Columbia; Jean Dupuis, a community science and local government specialist from northern Quebec; Peter Johnson, President of the Association of Canadian Universities for Northern Studies, from Ottawa; and Wayne Adams, a former provincial environment minister from Halifax.

In concluding, Mr. Robinson wished CCAR members much success with their meeting, and expressed his desire to learn more about Canada's research activities in the Antarctic.

1.2. Background and Rationale for the Strategy Development

Prof. Warwick Vincent
Chair, Canadian Committee for Antarctic Research

In 1839–43, British explorer James Clark Ross led an expedition to Antarctica jointly sponsored by the Royal Navy and the Royal Society of London. It was a voyage that yielded a series of remarkable discoveries, including a sea route to latitude 78°S, the 900-km-long ice cliffs at the edge of the Ross Ice Shelf, and unusual marine life-forms at a depth of several kilometres in the Southern Ocean. These accomplishments, in the region now known as the Ross Sea Sector of Antarctica, were the result of Ross's consummate skill in polar exploration and science, acquired in the course of his five expeditions to the Canadian Arctic earlier in the century.

Over the last 100 years, Canadians have continued the Ross tradition of North–South research and exploration, and many place names in Antarctica bear testament to the links with Canada. For example, the Canada Glacier and Canada Stream in the Taylor Valley of southern Victoria Land were named as a result of the 1911 expedition to the region by Griffith Taylor who went on to become Canada's first professor of geography. These days, Canada is most conspicuous in Antarctica because of its many commercial interests in the south polar region: ecotourism, clothing, radar imagery, aviation services, snow and ice vehicles, marine harvesting (beginning in 2000), housing, camp facilities, and other goods and services based on Canada's broad expertise in cold regions science and technology. In addition, a small number of Canadian scientists conduct research in Antarctica each year, generally within programs led and supported by other nations. There is increasing awareness, however, that Canada also has much to gain by developing its research interests in Antarctica, and by placing its northern research in a bipolar, global perspective.

The opportunities for Canadian scientists to conduct bipolar and Antarctic research have recently been augmented by Canada's acceptance in July 1998 as a full member of the Scientific Committee on Antarctic Research (SCAR), the umbrella organization which oversees all work in the Antarctic region. As a required step in this process, the Canadian Committee for Antarctic Research (CCAR) was established as the National Antarctic Committee for Canada. One of CCAR's immediate tasks was to formulate a science strategy for Canada in Antarctica. As a first step in this process, CCAR prepared a discussion paper, *Antarctic and Bipolar Science: A Strategic Plan for Canada*, which was distributed in August–September 1999 to interested parties throughout Canada, including polar scientists, Canadian businesses active in Antarctica, science policy groups, and government agencies concerned with fulfilling Canada's Antarctic Treaty obligations.

Early on in its discussions, CCAR sought to define Canadian antarctic and bipolar science. The strategy document takes a broad view of science that includes the natural sciences, medical and other life sciences, the social sciences (including political science), and engineering and technology development. The term "bipolar" is taken to mean any research in which a comparison is made between the Arctic and Antarctica, or in which information from both polar regions is relevant. This refers to research programs that involve *in situ* data collection in both polar regions, but also includes research conducted in one polar region with comparisons to data published or otherwise available from the other polar region. For some Canadian researchers, the bipolar component or comparison is simply part of a much larger global analysis.

Antarctic research similarly encompasses a broad range of approaches: *in situ* data collection (e.g., the Canadian Museum of Nature's collaboration with National Science Foundation [NSF] researchers on the marine biology of McMurdo Sound; McGill University's collaboration with the NSF McMurdo Long-Term Ecological Research [LTER] program on permafrost, and ground ice research and work by the University of Northern British Columbia with the British Antarctic Survey [BAS] in the Antarctic Peninsula region); remote sensing (e.g., the use RADARSAT imagery of the Antarctic ice sheet by the Canada Centre for Remote Sensing); analysis of material from Antarctica (e.g., analysis of microbial life forms in the deep ice from Lake Vostok, East Antarctica, at Université du Québec à Montréal); experiments using organisms from Antarctica (e.g., the University of Western Ontario's research on photosynthesis by micro-algae from the McMurdo Dry Valley lakes; work at the University of British Columbia on Southern Ocean viruses; Laval University's work on Antarctic cyanobacteria); and theoretical and modelling studies based on the analysis of existing data (e.g., work at the Institute of Ocean Sciences on a model for circulation in Lake Vostok). By "Canadian" we mean the participation by Canadian institutions in this Antarctic or bipolar research, either as principal investigators or as collaborators.

Current Canadian Activities

2. Antarctic and Bipolar Science

2.1. Universities

University of Alberta—Curtis Strobeck

Prof. Curtis Strobeck, PhD student Corey Davis, and Dr. Ian Stirling of the Canadian Wildlife Service continue to study the genetic characteristics of Antarctic seals, based on laboratory analyses of specimens collected by scientists from several countries. Davis and Stirling will spend six weeks in the Antarctic during the austral summer of 1999-00 collecting additional specimens. At the same university; Prof. Foght studies hydrocarbon degrading bacteria at fuel-contaminated sites in Antarctica in co-operation with a New Zealand scientist. She expects to continue her studies during a sabbatical in New Zealand next year with emphasis on bipolar comparisons of active bacteria.

University of British Columbia—Peter Suedfeld

Prof. Suedfeld is continuing his studies of human adaptation to living in isolated and environmentally challenging situations. Prof. Curtis Suttle and students at the same university study the effects of UV radiation on viruses in the Southern Ocean, and extract DNA samples to determine the genetic diversity of natural viral communities. They also have ice cores from Ellesmere Island for studies of viral content and seek to obtain samples from the Vostok core for comparative studies.

University of Northern British Columbia—Kevin Hall

Prof. Kevin Hall, Department of Geography, studies weathering and periglacial processes in the Mars Oasis, Alexander Island, and will do fieldwork in the area in November-December 1999, as part of the Arctic-Antarctic Exchange Program. He co-operates with the British Antarctic Survey (BAS) and with a French scientist.

Profs. Hall and J. Arocena, (also of UNBC) analyse weathering rinds from the Cape Roberts drill core as a means of reconstructing past terrestrial environmental conditions in co-operation with J. Smellie, BAS.

University of Calgary, Arctic Institute of North America—Gerald Holdsworth

A conceptual synoptic-scale polar front cyclone model, incorporating the fractionation scheme for the stable isotopes of water (snow), has been developed for the Pacific Northwest region and applied to explain the vertical variations of the stable isotopes of snow in the St Elias Mountains up to 5920 m. One important result of this model is that for high-altitude ice core sites (> about 3200 m), changes in the strength of cyclones will result in changes in thickness of the warm front zone of the cyclone. Any ice core site located in this zone will receive an isotopic signal related to changes in wind speed as well as from temperature. These isotopic changes can be abrupt and may swamp the temperature signal. An example of this is the isotopic shift seen in ice cores at the Greenland summit (72°N; 3220 m) traversing the Last Glacial Maximum to Holocene (LGM-Holocene) transition. The cyclone model is applicable here and can be used to show that a wind-induced shift on top of a temperature shift could have caused most of the disagreement between the isotopic thermometer (ca 9°C) and the borehole thermometer (ca 18°C) results.

The model can be applied to the ice core data from Vostok, Antarctica (79°S; 3500 m) which is predominantly supplied by cyclonic precipitation. Instead of the roughly 8–9°C (LGM-Holocene) temperature shift derived from the "isotopic thermometer", it can be shown realistically how the shift may have been twice this value, making it more in accordance with the Greenland borehole thermometry result.

***Lakehead University, School of Outdoor Recreation, Parks and Tourism**—Margaret Johnston*

M. Johnson has studied Antarctic tourism, specifically the regulation of behaviour. Her primary interest is in the strategies used to control the activities of tourists and the tourism industry and how this is reflected in tourist use of sites in the Antarctic. Johnson has explored the general framework of tourist regulation and in December 2000 hopes to be part of a research project led by Dr. Bernard Stonehouse of the Scott Polar Research Institute to examine tourism planning at the Polish Arctowski base.

Related to this research is an Antarctic field trip offered for senior-year students at Lakehead (12 students in 1998 and 12 in 1999). This course focuses on the historical development of tourism, patterns, issues, and regulation. It includes a field trip aboard a cruise ship during which students gain firsthand experience of Antarctic tourism and are able to set that experience into the context provided by the course material.

The Antarctic tourism course and a course on the Geography of Polar Regions are part of the Minor in Northern Studies. The Northern Studies Committee provided some support to the students who participated last year in the Antarctic field trip, and will contribute again this year.

University of Toronto—Marianne Douglas

A number of University of Toronto researchers are currently involved in bipolar and, more specifically, Antarctic research.

Physicist Kent Moore and his students study air-sea interactions with regard to open ocean polynyas with support from NSERC, the Office of Naval Research, and colleagues at the British Antarctic Survey (BAS). Real-time monitoring of the ice pack off Antarctica allows for the early detection of developing polynyas and increased understanding of the formation of Antarctic Bottom Water. They are also studying the physics of an anomalous cloud type resulting from strong fall and winter inversions.

Physics Professor R. Peltier and associates are developing models to track the Holocene stability of the great polar ice-sheets, determine the extent of continental Antarctic ice-mass loss between 21,000 and 6,000 Yr BP, and place Earth rotational constraints on the global sea-level rise. Supported by NSERC and NERC funds, Prof. Peltier collaborates with colleagues in Britain and Denmark.

Prof. B. Netterfield (Astronomy and Physics departments) was involved in the BOOMERANG Telescope experiment and participated in the field phase during the austral summer 1998–1999. This circumpolar balloon flight, roughly along latitude 78°S in early 1999, examined cosmic microwave background. The work was conducted in collaboration with NASA, NSF, the Italian Space Agency, and NSERC.

Geologist Prof. N. Eyles participated in the Ocean Drilling Program (ODP) Leg 178 (Spring 1998) when cores were collected off the west coast of the Antarctic Peninsula. His PhD student, N. Januszczak, will participate in Leg 188 (Spring 2000) when target locations in the Prydz Bay area, East Antarctica, will be drilled. Core data from offshore drilling is used to establish sedimentation models to track the growth history of the continental margins. The project is funded by NSERC and ODP.

Prof. M. Douglas (Geology Department) spent January 1999 in the Dry Valleys of Antarctica, as part of the Canadian Arctic-Antarctic Exchange program, to conduct a bipolar comparison of freshwater algae and paleolimnology. This project was partially sponsored by the Polar Continental Shelf Project (PCSP), the Northern Scientific Training Program (NSTP), and NSF.

University of Ottawa—Hugh French

H. French participated in Project 2a.1.3 of the Programma Nazionale di Ricerche in Antartide (PNRA) during the 1998–1999 austral summer. The PNRA maintains a research station at Terra Nova Bay, Northern Foothills, Northern Victoria Land, latitude 74/50' S. A wide range of scientific research activities is supported by way of excellent logistics support and a station capable of supporting 120 persons which is open from October to February each year. The station was established in 1984–85. Principal investigator for Project 2a.1.3 is Professor

Francesco Dramis, Department of Geology, University of Rome. Working with Dr. Mauro Guglielmin, a geologist attached to the Lombardy Regional Survey, Prof. French's responsibilities were for permafrost distribution and periglacial/ snow-ice phenomena. Assisted by technician Luigi Bonnetti, also of the Lombardy Regional Survey, H. French set up two permafrost monitoring stations in the vicinity of Terra Nova station during the period October 25–December 5, 1999.

H. French has had earlier involvement with PNRA: February 1994 and November 1996. Invited lecturer at universities in Rome, Camerino, Naples, Florence, and Milan.

A draft three-year agreement on Arctic–Antarctic co-operation in permafrost and earth science research between the University of Ottawa and the University of Rome is currently being circulated among the interested parties. PNRA and PCSP have been proposed as the respective logistics sponsors. The University of Ottawa has agreed in principle and is awaiting finalization from the Italian side.

In the 1999-2000 austral summer, A. G. Lewkowicz (Department of Geography) will be at Terra Nova Bay October 25–December 6. In the 2000–2001 austral summer, H. French will again visit Terra Nova Bay and possibly Cape Roberts or the Dome C drilling sites.

Université Laval—Warwick Vincent

More than 80 graduate students, their professors and many undergraduates work in the North each year associated with three Université Laval research institutes: Centre d'études nordiques (CEN); Groupe interuniversitaire de recherches en océanographie du Québec (GIROQ); and Groupe d'études inuit et circumpolaires (GETIC). U Laval maintains a permanently staffed research station on the shores of Hudson Bay (CEN, Kuujjuarapik) similar to some winter-over stations in Antarctica; a CEN field camp on Bylot Island analogous to semi-permanent Antarctic field stations; and a network of automated weather stations (AWS) mostly throughout subarctic Québec and linked by satellite back to CEN (similar to the American AWS network in Antarctica). U. Laval is also the centre of operations for the 30 million CAN$ Northern Open Water program (NOW) in Arctic Canada (details at the GIROQ website: www.fsg.ulaval.ca/giroq/now).

Ongoing links with Antarctic research and researchers have been a small but active part of this focus on high latitude science. U. Laval has a Memorandum of Understanding with the Alfred Wegener Institute, Germany, for collaborative work in the Antarctic as well as Arctic. Many of the overseas participants in NOW conduct similar work in Antarctica, and Prof. Louis Legendre (GIROQ) is co-organizer of an ongoing series of Gordon Conferences on polar oceans. For example, he chaired the 1999 meeting in California that involved 120 invited researchers active in Antarctic and/or Arctic oceanography. Prof. Warwick Vincent (CEN) has been conducting research in the Ross Sea sector of Antarctica from 1979 onwards. He and his students work primarily on global change processes in northern lakes, but they also continue to be involved in a variety of Antarctic initiatives such as the " Long Term Ecological Research " program in the

McMurdo Dry Valleys (National Science Foundation, USA) and the "Polar Microbial Consortia" program led by New Zealand. The laboratory maintains a culture collection of high latitude micro-organisms, including Antarctic cyanobacteria. Dr. Vincent has an ongoing interest in environmental management in the polar regions and helped lead the development of an Environmental Code of Conduct for the McMurdo Dry Valleys.

***Université du Québec à Rimouski, Institut des Sciences de la mer de Rimouski (ISMER)**—Serge Demers*

The Institut des Sciences de la mer de Rimouski has two important research programs in Antarctica. The first deals with the biodegradation of oil in Antarctica. This program is conducted in collaboration with the Institut Français pour la recherche et la technologie polaires, with the aim of studying the long-term effect of contamination by oil in three particular environments of the Austral Ocean: sea ice, coastal soil, and subantarctic intertidal sediment.

The second considers the responses of marine ecosystems to enhanced Ultraviolet-B (UV-B) radiation at different latitudes. Adverse effects on organisms have been observed at several different trophic levels. Moreover, organisms from various ecosystems (i.e. tropical, temperate, and polar) have been shown to be affected. Most investigations have been reductionist in approach, considering single species or limited assemblages of organisms. Such studies are difficult to extrapolate to predict community responses, and harder still to assess with respect to latitudinal differences. In comparison with marine ecosystems at high latitudes, those at low latitudes experience an environment that is more stable with respect to a number of environmental variables and, although acclimatized to higher levels of irradiance, are hypothesized to be less able to accommodate anticipated increases in UV-B radiation. Our project examines the responses of planktonic ecosystems at various latitudes to realistic future levels of UV radiation from North to South poles. Results will be of value in elucidating the sensitivity of the base of the food web to UV radiation. This program will be carried out in collaboration with researchers from Argentina, Brazil and United States.

***McGill University**—Wayne Pollard*

Three departments have current or recent research interests relating to Antarctica, including: Geography (W. Pollard, D. Andersen, and D. Mueller), Biology (Eleanor Bell), and Oceanic and Atmospheric Studies (L. Mysak).

Research interests at McGill include:

Landscape and Paleoclimatic Evolution—W. Pollard in collaboration with Peter Doran and Robert Wharton, United States Antarctic Program (USAP)–McMurdo Long-Term Ecological Research program (LTER), Chris Mackay, NASA Ames-USAP, and Warren Dickenson, Victoria University, New Zealand. This research was the first to utilize PCSP's Canadian Arctic–Antarctic Exchange Program and resulted in a paper presented to the 8th International Symposia on Antarctic Earth Science; a second paper has been submitted for publication.

Extreme Environment Ecology—Dale Andersen and Derek Mueller both focus on aspects of microbial ecology. Andersen, who has extensive Antarctic experience with NASA Ames, is currently studying saline springs on Axel Heiberg Island. Mueller is doing a bipolar comparison of microbial communities in cryoconite holes (Canada Glacier, White Glacier). Dr. Eleanor Bell, a biologist, has studied plankton in a stratified saline lake in the Vestfold Hills, East Antarctica, work which is linked to that of Chris Mackay (NASA Ames), Peter Doran and Chris Fritzen (McMurdo LTER), Marianne Douglas (University of Toronto), and Warwick Vincent (Université Laval). D. Mueller also went south as part of the Canadian Arctic–Antarctic Exchange Program.

Ocean Circulation and Sea Ice— Work in this area has been conducted by Lawrence Mysak, Oceanic and Atmospheric Studies.

McGill also maintains a working relationship with the McMurdo LTER and NASA where their work at Expedition Fiord is being used for comparative analysis with research in Antarctica. Other activities include W. Pollard's participation on CCAR and the SCAR Geology Working Group.

University of New Brunswick*—Jack Terhune*

Prof. Jack Terhune continues collaboration with Australian scientists in the study of vocalization dialects among Antarctic seals. As part of this study, a graduate student, will spend the austral winter of 2000 at the Australian Mawson station.

University of Western Ontario*—Norman Huner*

Prof. Huner, post-doctoral fellow A. Ivanov, and PhD student R. Morgan study photo-chemical processes in a green alga collected from Lake Bonney, McMurdo Dry Valleys.

Trent University*—John Marsh*

Prof. Marsh and his students continue research related to Antarctic and Arctic tourism, and Prof. March also continues to assist the World Conservation Union's World Commission on Protected Areas in addressing protected area issues in Antarctica.

Queen's University*—Robert Gilbert*

Prof. Gilbert and graduate student Åsa Chong have spent two summers along the Antarctic Peninsula studying fiord sediments, building on experience from similar studies in arctic Canada and Greenland. Gilbert will return in May 2000 to collect sediment cores in areas that have become accessible for the first time due to the recent extensive break-up of the northern part of the Larsen Ice Shelf.

Université du Québec à Montréal—David Bird

Prof. Bird and students study the microbial ecology of bacteria in arctic and antarctic environments, including microbial activities in the Vostok ice core.

2.2 Federal Departments and Agencies

Canadian Museum of Nature—Kathy Conlan

Three scientists are currently conducting research on antarctic biology.

Dr. Michel Poulin worked on the benthic and epizoic diatoms from the Maritime Antarctic from 1994–1997. The materials were supplied from the polar studies of the Alfred Wegener Institute for Polar and Marine Research, Germany. From 1998–1999, research shifted to sea-ice diatoms and work with C. Riaux-Gobin in the French sector. Dr. Poulin will be visiting Antarctica with the French team in October–December 1999.

Dr. Kathleen Conlan has had seven field trips to Antarctica since 1991, and has been active in antarctic research and public outreach for nine years. Her research in the Antarctic deals with the ecology of sea floor disturbance, reproductive synchrony to the annual plankton bloom, and biodiversity and evolution. In addition to her research publications, Dr. Conlan has written popular articles on Antarctica and has appeared on television and radio, and in magazines and newspapers. She has made special efforts to educate children about the Arctic and Antarctic by giving talks in schools and linking with groups by e-mail while in Antarctica. She plans to write an autobiography for children of her Antarctic experiences, continue conducting bipolar research, and expand polar teaching. Through her representation of Canadian biologists in SCAR, Dr. Conlan hopes to enhance opportunities for Canadian scientists to work in Antarctica.

Environment Canada, Atmospheric and Climate Science Directorate/AES*—Barry Goodison*

Observation and modelling are essential in our weather, climate, and air quality initiatives. By nature, the processes involved are commonly global, even when the particular focus is on a region. Weather and climate models are global and require global observations, including both polar regions. The Climate Research Branch (CRB) led the international WMO Solid Precipitation Measurement Intercomparison, and several enquiries have been received on precipitation measurement and the associated errors for Antarctica; this is one area in which Canadian researchers can partner with others to address a common observational challenge (contact: Barry Goodison). For climate modelling, atmospheric and ocean processes are basically similar in both polar regions, but the process of sea ice formation is different and is currently not well represented in models. The CRB Climate modelling group, CCCma, is planning a specific experiment to study the effect of the recession of the Antarctic ice shelves on regional climate and ocean water mass properties; this is a global model experiment, but focussed on the Antarctic problem. Greg Flato is the contact person with the Climate Research Branch.

The Air Quality Research Branch (AQRB) has strong polar interests. The extent to which ozone depletion, as observed in Antarctica, might be replicated in the Arctic has been a particular research focus. A special observatory was built in 1992 at Eureka, N.W.T., as a centre for Canadian and international Arctic ozone studies. There are still considerable uncertainties in the science of Arctic ozone depletion, more than with the Antarctic case. Canadian scientists run models that simulate both polar regions, but their measurements are almost exclusively done in the Arctic. The increasing UV-B radiation in the Arctic is also being monitored. EC scientists participate in NASA ER-2 projects and these are sometimes focused on Antarctic ozone, the last being the Arctic Southern Hemisphere Experiment (ASHOE) during the spring of 1995 (David Wardle, AQRB). A proposed Branch project with Germany to determine if the mercury depletion phenomenon first observed in spring 1995 at Alert also occurs during the spring in Antarctica, remains unfunded. This information is used in the formulation of hemispherical and/or global mercury models currently being developed (Bill Schroeder, AQRB).

CRYSYS (Cryospheric System Study) focuses on the variability and change in the cryospheric system in Canada. Led by CRB (Principal Investigator, Barry Goodison) the study currently involves 15 Canadian universities and four federal departments. The cryosphere is a global phenomenon, and CRYSYS will provide a Canadian focus for the proposed new World Climate Research Program project CLIC (Climate and Cryosphere), which will have a strong Antarctic focus. CLIC could offer opportunities for Canadian scientists with cryosphere/climate interests to link with foreign investigations in Antarctica.

Fisheries and Oceans Canada--Eddy Carmack

High-latitude ocean climate research, by its very nature, requires a bipolar perspective. For example, the global thermohaline circulation, sometimes called "the great conveyor belt", is driven by convective forcing in both hemispheres, and comparative studies of ocean convection are essential if we are to adequately understand and model the ocean's role in climate. Thermohaline (T/S) structures in both Arctic and Antarctic seas show remarkable similarity to one another, again lending argument to comparative study. Indeed, such bipolar considerations form the basis for high-latitude ocean researches planned under the World Climate Research Program's CLIVAR (Climate Variability and Prediction) and CLIC (Climate and Cryosphere) programmes. DFO scientists have participated in the planning phases of these programmes (e.g. A Clarke, Bedford Institute of Oceanography; E. Carmack, Institute of Ocean Sciences). However, participation in fieldwork in the southern Ocean has not been possible due to funding constraints and other priorities.

Much ongoing international work in the Southern Ocean is presently co-ordinated through the International Antarctic Zone programme (IAnZone) and this body forms a natural basis for modest Canadian participation in future international research. Again, similarities to Arctic programs and Canadian strengths are obvious: IAnZone is currently focusing on shelf-basin interaction, processes which have been the focus of Canadian work in the Beaufort Sea over the past two decades.

Efforts to bring oceanographic modelling approaches to Antarctic freshwater bodies are also underway. One example is recent work to predict circulation in Lake Vostok, a freshwater lake roughly the size of Lake Ontario that is covered by almost 4000 m of glacial ice. (see Wuest and Carmack, 2000). Lake Vostok is also considered as an analogue to planetary exploration; in turn, proposals are being developed to test new technologies.

Natural Resources Canada, Geological Survey of Canada—Roy Koerner

Presently the Geological Survey of Canada (GSC) Arctic program falls into two basic parts: Glacier/ice-cap mass balance; and ice-core/climate studies.

The GSC continues to measure the mass balance of: the northern catchment of the Agassiz Ice Cap (started 1977); the Meighen Ice Cap (1960); the Melville Ice Cap (1964); and the Devon Ice Cap (1961). These measurements show no overall trend in either winter snow accumulation or summer melt over the 30–40 year period, indicating the eastern Arctic is not presently taking part in continuing global warming. The ice core/pollution program, begun in 1964 with the Polar Continental Shelf Project, has drilled four surface-to-bed cores on the Devon Ice Cap, five on the Agassiz Ice Cap, and one on the Meighen Ice Cap. These have provided climatic/snow chemistry records covering the last glacial period. The program includes annual snow sampling on the four mass balance ice caps and wherever ice drilling is in progress.

There is no active glaciology field program in Antarctica. The work presently undertaken is with the use of satellite imagery, modelling, and the use of data generated by various Antarctic field programs of other nations.

Tom James, a post-glacial rebound modeller, is working on Antarctica "both in terms of contributions to sea-level since the last Glacial Maximum, and in terms of present-day contributions to sea level change. This is effected by changing surface load caused by ice sheet fluctuations on the solid Earth." Most of the work is done in close collaboration with E. Ivins (Jet Propulsion Lab). James and Ivins are also refining the ice sheet history of the Antarctic Peninsula. James is also working on Roosevelt Island data to calculate the uplift rate that could be measured on the nearest bedrock exposure.

Laurence Gray (Canada Centre for Remote Sensing) continues to work on the RADARSAT AMM interferometric data for ice motion with his CCRS colleagues and with Joughin (Jet Propulsion Laboratory) and Bindschadler (NASA) on the tributary system feeding the West Antarctic Ice Streams and the ice flux draining through the Filchner Ice Shelf.

Drs. James and Gray have published several papers in international journals on this Antarctic work.

2.3 Commercial Activities

Expedition Logistics—*Geoff Green*

Geoff Green represented the Canadian Antarctic tour operator, Marine Expeditions Inc., and his own company, Expedition Logistics. Mr. Green was a representative at this year's meeting of the International Association of Antarctica Tour Operators (IAATO) in Hamburg, Germany. A Canadian expedition leader, Mr. Green has led 42 expeditions to Antarctica, as well as many to the Arctic, and contracts his leading and consulting services to several organizations around the world, including the World Wildlife Fund, the Smithsonian Institution, the Discovery Channel, and the National Audubon Society. He also conducts an educational lecture series on the polar regions to student groups and corporations across Canada and the U.S..

Mr. Green described his "Students to Antarctica" expedition, which is a project to take 100 high school students on an educational journey to Antarctica in December 2000. This is a ship-based expedition leaving from Ushuaia, Argentina, and concentrating on the Antarctic Peninsula. A tailored educational program will be an integral part of each student's experience, and live Internet links will allow students from across Canada and the world to share in the journey. It was explained that the "Students to Antarctica" expedition is a flagship for annual student expeditions to Antarctica and the Canadian Arctic, as well as a catalyst in developing an Antarctic–Arctic educational curriculum for Canadian schools. Mr. Green proposed that CCAR consider becoming involved with the student expedition as an academic sponsor. This could involve CCAR assistance in developing an Antarctic educational curriculum and/or CCAR members implementing or participating in shipboard research projects during the expedition. Benefits to CCAR/CPC would include: exposure to schools across Canada; the raising of public awareness of CCAR/CPC and Canadian polar science; facilitation of polar research projects; and the potential for funding spin-offs to the CCAR community.

Icefield Instruments Inc.

Icefields Instruments Inc., Whitehorse, Yukon, has developed ice-coring equipment and improved methods for analysis of core samples and has worked on glaciers in many countries. The company is currently working for the U.S. component of the International Trans-Antarctic Scientific Expeditions (ITASE).

2.4 Related Activities—*Olav H. Loken*

As it was not possible to gather all Canadian scientists involved in Antarctic and bipolar science some activities have not been reported. The following paragraphs address some of these gaps, but CCAR is also seeking information on other initiatives or activities which may be relevant. From west to east in Canada, CCAR noted the following activities:

Alan Saunders, Canadian Polar Commission, is National Antarctic Data Co-ordinator and Canadian representative to the Joint Committee on Antarctic Data Management (JCADM), a joint committee of SCAR and COMNAP. In June 1999, the Commission hosted the first joint meeting of the JCADM and the International Arctic Environmental Data Directory (ADD) Council. *Arctic and Antarctic Data Management: The Bipolar Context*, a report based on the proceedings of the meeting, is now being prepared in conjunction with L. Belbin, Australian Antarctic Division, M. Thorley, British Antarctic Survey, D. Henry, GRID-Arendal, and T. Northcutt, Global Change Master Directory. Representatives of polar research groups from close to 40 countries participated in the four-day meeting.

Polar Continental Shelf Project (PCSP), Natural Resources Canada, works in collaboration with the Canadian Committee on Antarctic Research in managing the Canadian Arctic–Antarctic Exchange Program. Established in 1996, the Program is designed to encourage scientific collaborations among Canadian Arctic research scientists and their Antarctic colleagues. In addition, the Canadian Polar Commission has designated the PCSP to represent Canada on the Council of Managers of National Antarctic Programs (COMNAP), which serves as a co-ordinating body among national Antarctic operators, and on the affiliated logistics co-ordinating body for Antarctica, Standing Committee on Antarctic Logistics and Operations (SCALOP). Efforts have advanced to establish a similar co-ordination mechanism for circumpolar Arctic countries, the Forum of Arctic Research Operators (FARO) which Canada (PCSP) currently chairs.

Several recent and ongoing developments affect the policy setting in which Canadian bi-polar and antarctic research is carried out and supported. Within Canada, international and antarctic relations will now be focused on the newly created Aboriginal and Circumpolar Affairs Division of the Department of Foreign Affairs and International Trade (DFAIT). Antarctic policy matters are the responsibility of the Deputy Director, Elaine Koren. Canada's Ambassador for Circumpolar Affairs, Mary Simon, has also been appointed Ambassador to Denmark and Greenland, while keeping her circumpolar portfolio. She has stated her intention to use her double responsibilities and enhanced connections with Greenland to pursue increased circumpolar scientific co-operation, especially in the social sciences.

Dr. Fred Roots is scientific adviser to both the Office of the Circumpolar Ambassador and the Aboriginal and Circumpolar Affairs Division. Since 1991, Dr. Roots has been the Canadian delegate to the Antarctic Treaty Consultative Meetings, and attended the XXIII ATCM in Lima, Peru in 1999. During that meeting, on behalf of Canada, Dr. Roots: prepared and presented the report from Canada; prepared and presented the report on circumpolar scientific developments in the Arctic that are relevant to the Antarctic; prepared a Working Paper on the relevance of the Antarctic Treaty to the World Conference on Science, which was sponsored jointly by Canada and Ecuador; attended all sessions of the ATCM Committee on Environmental Protection, and on Science and Operations; participated in the special international workshop on criteria and delineation of protected areas in Antarctica; provided the Canadian contact and address for a web site on information exchange on environmental issues in Antarctica; and agreed on behalf of Canada to adoption of guidelines for Environmental Impact Assessment on Antarctica. At the

United Nations (UNESCO/ICSU) World Conference on Science held in Budapest, Hungary June–July 1999, Dr. Roots was instrumental in ensuring that research in the polar regions was noted specifically in the *Science Agenda for the 21st Century* that was endorsed by 154 countries. It may be noted that the Canadian presentation to the plenary assembly was the only national statement that specifically referred to "Antarctic and polar research" by name.

Other science-related, policy-connected activities include:

- promotion and liaison of research and international co-ordination of technical specifications for ship navigation in Arctic and Antarctic waters, under the International Maritime Organization (working group led by Canada)

- assessment of proposals for research on climate change in polar regions, in response to Canadian commitments under the Kyoto Protocol.

3. Canadian Strategy for Antarctic and Bipolar Science

Rapporteur's Report
Kevin Hall

The discussion centred around six potential strategies as outlined by the workshop chair:

1) withdraw from the Antarctic Treaty and SCAR;
2) be a "great polar nation" via substantial dollar input;
3) maintain the status quo in Antarctic/bipolar research;
4) encourage a moderate scaling up of "*in situ* research" in Antarctica;
5) increase awareness in Canada regarding the importance of antarctic/bipolar science; and
6) strengthen Canada's northern research.

Points 4 and 5 were the key elements of the discussion paper produced by CCAR and circulated to all participants prior to the workshop.(*Antarctic and Bipolar Science: A Strategic Plan for Canada*)

Extensive debate enabled all participants to make their views known. In general, it was felt that Canada's accession to the Antarctic Treaty, as a non-consultative party, was done without a broad commitment to fulfilling Treaty obligations, and many governmental bodies now affected by it have been—and probably still are—unaware of all the implications. "Education" was therefore identified as a major consideration. It became clear that there is a great need to improve awareness, from kindergarten through to the ministerial level, of the CPC and CCAR, of Canada's history in the Antarctic, and of the significance of Antarctic studies to Canada's scientific undertakings and economic initiatives (e.g., relevance of antarctic glacial/oceanic/climatic elements to the northern hemisphere; income derived by Canadian companies as a result of Antarctic tourism, goods, or services.) This increased awareness also needs to be communicated to the private sector, as Canada is "potentially poised to be central to bipolar commercial undertakings". This educational component is critical and needs to be packaged in a professional way.

A central issue underpinning the entire debate, and one of particular sensitivity within the Canadian context, is the current need for more Arctic research and the inadequacy of the current logistical infrastructure that supports it. Current budgets are woefully small, older research stations are no longer viable while others are used to capacity, and in many instances Canadian research can only be done by "piggy-backing" on foreign undertakings without which very little could be achieved. The parochialism with respect to the Arctic also engenders serious antagonism in some quarters with respect to increased support for a Canadian Antarctic program. The situation is often portrayed as a zero-sum game in which any funding for the Antarctic is perceived as a loss for Arctic research, which is itself grossly underfunded. The existing situation is viewed as conflictual rather than mutually supportive, and one which scientific peers and colleagues have frequently exacerbated—rightly or wrongly—by open opposition or lack of support at the granting (e.g., NSERC/SSHRC) level. In some respects, the problems facing Arctic research in Canada have further complicated an already untenable situation in which Canada finds itself party to an agreement without strong and broadly based commitment or an adequate knowledge of the historical context.

Any strategy for the future must be viewed as credible, as the issue is one of national importance. At present, much of the research work in the Arctic is undertaken on a "shoestring" budget; at the same time, there is little support for Antarctic research. (Following the workshop, in response to the CCAR Discussion Paper, some government groups have expressed strong optimism for future support for bipolar and Antarctic research.) At the national level, this should be recognized in comparison with other countries. For example countries, such as Peru, Lithuania, Venezuela, and Ukraine, are prepared to fund Antarctic research, including an Antarctic base and logistical support; other "non-polar" nations, such as Holland, France, Germany, the United Kingdom, and Italy have active programs in both the Arctic and Antarctic. Canada is paying lip-service to its polar role by underfinancing polar research, even in its own "backyard".

With these comments, it was suggested that withdrawal from international Antarctic Treaty obligations (point 1) was impractical and would be counter-productive. A significant increase in funding (point 2) was viewed as unrealistic; however, government should be prepared to answer the question, "Why not?", particularly when other countries can achieve this despite their own economic problems. The *status quo*, while it would allow for some work to continue, was not seen as a satisfactory way forward. In discussion around all three points, it was noted that Canada has lacked an organization and voice to tackle these issues adequately and in a timely fashion.

The consensus among participants was that a moderate scaling up of Canadian Antarctic research (point 4) is the most appropriate strategy, but it was noted that this can only be achieved in conjunction with increased awareness of the importance of antarctic/bipolar science (point 5) and a strengthening of Canada's northern research (point 6). Strengthening Arctic research would create a critical mass of Canadian polar scientists whose expertise would be valuable in the

Antarctic, and an organizational infrastructure in which that Antarctic experience would, in turn, be utilized in Canada. With this as the way forward (points 4, 5, and 6) the creation of a "Canadian Foundation for Antarctic Research" should be strongly considered, with input from the private sector. Such a body would provide a small measure of financial support to help maintain and encourage Canada's present (and not inconsequential) Antarctic involvement. The steps by which to implement this and obtain financial support will be examined.

There is strength in Canadian Antarctic research, but future development depends on the health of Canada's Arctic research programs. Canada has placed itself in a contentious situation under the Antarctic Treaty, and this must be addressed, in part through the establishment of a better system of communications. Overwhelmingly, the consensus was that Canada clearly has the potential to be a real player of significance and that it is a duty that must be undertaken for future generations if Canada is to play any meaningful role in the Arctic, the Antarctic or science in general.

4. Closing Remarks—*Warwick F. Vincent, Chair CCAR*

A number of important themes and conclusions have emerged from this workshop, and we must now follow up on these. The first is the remarkable breadth of antarctic and bipolar science activities undertaken by Canadians over the last few years. This is underscored not only by the diversity of scientific interests among the workshop participants, but also by the number of Canadian scientists in Antarctica each year (Appendix 4) and the large output of publications by Canadians in the international scientific literature (Appendix 3). This significant level of contribution is to be expected given the longstanding expertise of Canadians in cold regions research. CCAR will continue to identify these contributions.

Second, our discussions reaffirmed the strong commercial presence of Canadians in the Antarctic region. Given the vital importance of cold regions science and technology to everyday life in much of Canada throughout winter, this Canadian presence in Antarctica again seems logical and likely to expand in the future.

Third, there is limited awareness of Canada's obligations and commitment to the Antarctic Treaty. The Treaty, specifically the Protocol on Environmental Protection to the Antarctic Treaty, devotes this region to "Peace and Science" (Article 2), the latter necessarily focused on cold regions science and technology. Such a focus makes this agreement by 44 adherent nations a remarkably "Canadian" international treaty, but we need to look more closely at defining Canada's role as a signatory.

Fourth, any discussion about Antarctica in Canada is necessarily linked to discussions about the Arctic, and Canadian arctic research. CCAR has a role to play in helping place Canadian arctic research in a more global, bipolar context (thereby also contributing toward our Antarctic Treaty commitments) and in helping the CPC and others strengthen northern research in Canada.

Fifth, given Canada's other priorities, it is not appropriate to consider a major scaling up of Canadian research activities in Antarctica. There was little enthusiasm for a Canadian Antarctic base, although cost-sharing of existing facilities may be of interest. Reciprocal arrangements such as the Canada Arctic–Antarctic Exchange Program, run by the Polar Continental Shelf Project in collaboration with CCAR, would seem a preferable approach, to the benefit of all countries. Some support does seem necessary for Antarctic research undertaken by Canadians. NSERC has affirmed that antarctic research is eligible for consideration under all of its programs. Of course, these programs are already stretched to the limit and in the past have had difficulty supporting northern research to the extent that it deserves. An alternative option that was endorsed by this workshop is the establishment of a Canadian Foundation for Antarctic Research that could work in partnership with Canadian industry and others toward providing a small amount of support to Canadian researchers (e.g., travel costs).

CCAR will now work with these ideas and with additional input from the written responses to the Strategy discussion paper, and will, in concert with the CPC, prepare a Canadian antarctic/bipolar science plan that builds on the discussion paper and reflects these ideas.

Finally I would like to thank all of you who contributed toward making this workshop a success. Our workshop was hosted by the Arctic Institute of North America (AINA), and on behalf of CCAR I thank the Executive Director of AINA, Michael Robinson, and his staff for their gracious hospitality in such an appropriate setting. We are also grateful to Professor Kevin Hall for his superb work as rapporteur during the workshop and for his co-editing and contributions to this report. I thank all the participants at the Calgary meeting for their stimulating presentations and feedback (and look forward to it continuing); and the Canadian Polar Commission for their support of CCAR activities. Finally, I extend a special thanks to Dr. Olav Loken, secretary to CCAR, for his excellent work in editing the discussion paper and the current report, and for ensuring the smooth organization and running of this productive meeting. If you have additional comments on the strategy paper or on issues raised in this report, we would be very pleased to hear from you.

APPENDIX A

Agenda—CCAR Workshop
October 16, 1999, 0900-1700hrs

Arctic Institute of North America, MacKimmie Library Tower,
11th floor, 2500 University Drive, University of Calgary.

1. Introduction and Welcome (Warwick Vincent, Chair, CCAR/CCRA)

2. Welcome to Calgary/A CPC Perspective (Michael Robinson, Chairperson, Canadian Polar Commission.)

3. Overview of Discussion Paper. How did it come about? What do we seek to achieve? Where are we at? (Olav Loken)

4. The Context. Who in Canada is doing/planning what in Antarctic related actvities? Information exchange. A series of short presentations (5-10 min. each)

    a) In the universities

    b) In government departments/agencies

    c) In the private sector

LUNCH

Afternoon session
Rapporteur : Prof. Kevin Hall (University of Northern British Columbia)

5. Comments on the discussion paper (all).

6. Discussion of key points to arrive at consensus about the next steps.

7. Canadian Foundation for Antarctic Research (CFAR). What is the best way to go about it?

8. Summary: Action to Be Taken

9. Closing Remarks

APPENDIX B

ATTENDANCE AT CCAR WORKSHOP, CALGARY, AB, OCT. 16/1999

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APPENDIX C

Recent Canadian Contributions to Antarctic and Bipolar Science

This list contains papers published since the *Bibliography of Canadian Contributions to Antarctic and Bipolar Science* was compiled in mid-1997, and is thus an extension of that document. It also contains a few older publications we were not aware of when the previous document was compiled. Efforts have been made to include all publications, but we suspect some are missing. We would welcome your help in identifying omission so that the list can be as complete as possible.

Names of Canadian co-authors are <u>underlined</u>, except when all are Canadians. As before the publications are listed according to SCAR subsidiary groups in the following sequence:

Biology
- *UV effects*
- *Marine*
- *Terrestrial-Freshwater*

Hydrology

Geodesy and Geographical Information

Geology
- *Bedrock*
- *Geomorphology*

Glaciology

Human Biology and Medicine

Solid-Earth Geophysics

Environmental Affairs

Global Change and the Antarctic
- *Paleoclimates*

Others

Biology

UV Effects

Belzile, C., S. Demers, D.R.S. Lean, B. Mostajir, S. Roy, S. de Mora, M. Gosselin, D. Bird, J.-P. Chanut, and M. Levasseur, 1998: "An experimental tool for the study of the effects of ultraviolet radiation of planktonic communities: a mesocosm approach". *Environ. Technol.* 19 : 667–682

Belzile, C., S. Johannessen, M. Gosselin, S. Demers, and W.L. Miller. "Transmission of ultraviolet radiation through first-year sea ice in the North Water Polynya (76–79°N) in late spring. *Limnol. Oceanogr.*

Chatila, K., S. Demers, B. Mostajir, J.-P. Chanut, and P. Monfort (In press). "Bacterivory of a natural heterotrophic protozoan population exposed to different intensities of ultraviolet radiation". *Aquat. Micro. Ecol.*

de Mora, S., S. Demers, and M. Vernet (In press). *The effects of UV Radiation in the marine environment.* Cambridge University Press. 320 pp.

Fauchot, J., M.. Gosselin, M.. Levasseur, B. Mostajir, C. Belzile, S. Demers, and S. Roy: "Influence of UV-B radiation on nitrogen utilization by a natural assemblage of phytoplankton". *J. Phycol.*

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Mostajir, B., S. Demers, S. de Mora, C. Belzile, J.-P. Chanut, M. Gosselin, S. Roy, P. Z. Villegas, J. Fauchot, J. Bouchard, D. Bird, P. Monfort and M.. Levasseur 1999: "Experimental test of the effect of ultraviolet-B radiation in a planktonic community". *Limnol. Oceanogr.* 44: 586–596.

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Mousseau, L., M. Gosselin, M. Levasseur, S. Demers, B. Mostajir , P. Villegas, and S. Roy . "Long-term effects of ultraviolet-B radiation on community structure and on simultaneous transport of carbon and nitrogen by estuarine phytoplankton during a mesocosm study". *Mar. Ecol. Progr. Ser.*

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Quesada, A. and W.F. Vincent, 1997: "Strategies of adaptation by Antarctic cyanobacteria to ultraviolet radiation". *Eur. J. Phycol.*. 32: 335–342

Rae, R., Howard-Williams, C., Hawes, I., and W.F. Vincent. (In press) "Temperature-dependence of photosynthetic recovery from solar damage in Antarctic phytoplankton". SCAR, Christchurch.

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Robinson, D.H., K.R. Arrigo, Z. Kolber, M. Gosselin, and C.W. Sullivan 1998: "Phytophysiological evidence of nutrient limitation of platelet ice algae in McMurdo Sound, Antarctica". *Journal of Phycology* 34: 788-797.

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Geodesy and Geographical Information

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Geology

Bedrock

Eyles, N., J. Daniels, L. Osterman, and N. Januszczak. 2000: "Ocean Drilling program Leg 178 (Antarctic Peninsula): sedimentology of glacially-influenced continental shelf 'topsets' and 'foresets'". *Marine Geology.*

Rees, M.N., B.R. Pratt, and A.J. Rowell, 1989: "Early Cambrian reefs and associated lithofacies of the Shackleton Limestone, Transantarctic Mountains". *Sedimentology,* v. 36, pp. 341–361.

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Geomorphology

Bockheim, J.G. and K. Hall. 1999: "Permafrost, active layer dynamics and periglacial environments of continental Antarctica", In: J. Lee-Thorp and H. Clift (eds.): *Book of Abstracts, XV International Congress, International Union for Quaternary Research, 26.*

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Glaciology

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Human Biology and Medicine

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Solid-Earth Geophysics

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Environmental Affairs

Aislabie, J., J. Foght, and D. Saul, 2000: "Aromatic hydrocarbon-degrading bacteria isolated from soil near Scott Base, Antarctica". *Polar Biology*.

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Global Change and the Antarctic

Paleoclimates

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Others

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Lewkowicz, A.G. (Ed.) 1999: *Poles Apart: A Study in Contrasts*. Proceedings of an International Symposium on Arctic and Antarctic Issues, University of Ottawa, September 25–27, 1997. University of Ottawa Press, 237 pp.

Yannick Huot, a M.Sc. student at Dalhousie University, was a participant in a National Science Foundation post-graduate course on Antarctic ecology and biochemistry, McMurdo Station, in January 2000.

N. Januszczak, PhD student, University of Toronto, will be involved in offshore drilling in the Prydz Bay area, East Antarctica, as part of the Ocean Drilling Program in Spring 2000.

Prof. Margaret Johnston, Lakehead University, visited the Antarctic Peninsula area with a class of 12 students in December 1999.

Dr. Marian Kuc, of Ottawa, conducted botanical studies on the Antarctic Peninsula in January 2000, in co-operation with the Chilean Antarctic Program.

Aron Lawton, of Trent University is conducting a tourist survey on the Antarctic Peninsula.

Prof. A.G. Lewkowicz, University of Ottawa, conducted periglacial studies at Terra Nova Base, North Victoria Land, as part of the Italian Antarctic Program, November–December 1999.

Stewart J. Moorehead, a University of Waterloo graduate, and now a PhD student at Carnegie Mellon University in Pennsylvania, is undertaking a rock and meteorite search at Elephant Moraine, Victoria Land, from January to March 2000, as part of the United States Antarctic Program.

Dr. Michel Poulin, Canadian Museum of Nature, is conducted under-ice algae studies at Dumont d'Urville station, East Antarctica, as part of the French Antarctic Program, January 2000.

Kristi Skebo, previously with the Churchill Northern Studies Centre, worked as a field assistant for a New Zealand ecologist at Cape Bird, Ross Island, as part of the New Zealand Antarctic Program, October–November 1999.

Dr. Ian Stirling, Canadian Wildlife Service, Edmonton, studied the ecology of Antarctic seals, onboard the *US-RV Nathaniel B Palmer* in the Ross Sea and McMurdo Sound, January–February 2000.

Valérie Villeneuve, an M.Sc. student at Université Laval undertook microbiological studies on the McMurdo Ice Shelf with the New Zealand Antarctic Program as part of the Canada Arctic–Antarctic Exchange Program, January 2000.

APPENDIX D

Canadian Scientists in Antarctica during the Austral Summer 1999–2000

The following Canada-based scientists have indicated plans to undertake science related activities in Antarctica during the austral summer 1999–2000. Every effort has been made to make the following list as comprehensive as possible; however, some names may have been omitted. CCAR would appreciate information on any other research activities which should be included.[1]

Patrick Abgrall, PhD student, University of New Brunswick, will spend the 2000 austral winter at Mawson Station as part of the Australian Antarctic Program, studying vocalization among Antarctic seals.

Peter Adams, MP, formerly of Trent University, visited the Scott Base/McMurdo area in late January 2000, as a guest of Antarctica New Zealand, to discuss exchange protocol and co-operation in polar science.

Corey Davis, PhD student, University of Alberta, was on board the *US-RV Nathaniel B Palmer*, in the Ross Sea and McMurdo Sound in January–February 2000, to study the ecology of Antarctic seals.

Mike Gerasimof, Icefield Instruments Inc., Whitehorse,YT, conducted shallow ice-coring in West Antarctica as part of United States ITASE program, November–December1999.

Prof. Robert Gilbert, Queen's University, will conduct offshore sediment sampling in May 2000 in the Larsen Ice Shelf area onboard the *US-RV Nathaniel B Palmer*, as part of United States Antarctic Program.

Prof. Kevin Hall, University of Northern British Columbia, conducted weathering studies in Mars Oasis, Alexander Island, November–December 1999, as part of the Canada Arctic–Antarctic Exchange Program with the British Antarctic Survey.

Phil Holme, previously with the Geological Survey of Canada, and now a PhD student at Victoria University in Wellington, New Zealand, studied sedimentary geology, January–February 2000, in the McMurdo Sound area as part of New Zealand Antarctic Program.

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[1]The emphasis here is on Canada-based scientists. For example, government and university scientists from Canada have been included, but Canadian scientists on the staff of foreign universities have not. Likewise, participants in a field trip arranged as part of a formal university course have been included, but Canadians engaged as guides or lecturers on Antarctic cruise vessels are not listed.

APPENDIX E

List of Acronyms

AINA—Arctic Institute of North America
AMD—Antarctic Master Directory
ATCM—Antarctic Treaty Consultative Meeting
AWS—Automated weather stations

BAS—British Antarctic Survey

CCAMLR—Convention for the Conservation of Antarctic Marine Living Resources
CCAR—Canadian Committee for Antarctic Research
CCCma—Canadian Centre for Climate modelling and analysis
CEN—Centre d'études nordiques
CLIC—Climate and Cryosphere program
CLIVAR—Climate Variability and Prediction Research program
COMNAP—Council of Managers National Antarctic Programmes
CRB—Climate Research Branch (DOE)
CPC—Canadian Polar Commission
CRYSYS—Cryosheric System Study (in Canada)

DIRS—Devon Island Research Station

FARO—Forum of Arctic Research Operators

GETIC—Groupe d'études inuit et cirqcumpolaires
GIROQ—Groupe interuniversitaire de recherches en océanographie du Québec

IAATO—International Association of Antarctic Tour Operators
ICSU—International Council of Scientific Unions
ITASE—International Trans-Antarctic Scientific Expeditions

JCADM—Joint Committee on Antarctic Data Management

KLRS—Kluane Lake Research Station

LGM-Holocene—Last Glacial Maximum to Holocene transition
LTER—Long Term Ecological Research program

NASA ER-2—National Aeronautic and Space Administration—ER-2 aircraft flights
NERC—Natural Environment Research Council
NOW—Northern Open Water program
NSERC—Natural Sciences and Engineering Research Council of Canada

NSF—National Science Foundation
NSTP—Northern Scientific Training Program

PCSP—Polar Continental Shelf Project
PNRA—Programma Nazionale di Ricerche in Antartide

SCALOP—Standing Committee on Antarctic Logistics and Operations
SCAR—Scientific Committee for Antarctic Research
SSHRC—Social Sciences and Humanities Research Council of Canada

TNB—Terra Nova Base

UNESCO—United Nations Educational, Scientific and Cultural Organization
USAP—US Antarctic Program

WMO—World Meteorological Organization

Chapter 10

Conclusions

"The important thing in science is not so much to obtain new facts as to discover new ways of thinking about them."

Sir William Bragg

As stated in the Introduction, the main aim of this undertaking was to reconsider the nature of mechanical weathering in cold regions, with a particular emphasis on the Antarctic. The concept of cold region mechanical weathering has, as was shown in the Introduction and commented on in many chapters, been dominated by the assumption of freeze-thaw as the predominant, if not the only, process. Thus, the work reported on here was initiated with the aim of collecting as much relevant field data as possible, coupled with laboratory simulations based upon those field data, to facilitate a meaningful evaluation of the nature and rate of the weathering taking place. Integral with this study of weathering processes was the evaluation of their impact upon landforms, particularly taffoni, nivation and cryoplanation - all cold region landforms that are considered to owe a significant component of their development to mechanical weathering. The papers pertinent to these topics were presented in Chapters 2 to 5.

In many ways it would seem that the underlying problem with the concept of weathering in cold environments, a concept that has been with us for over a century, is largely an artefact of our data. With the early presumption of the role of cold researchers fixated upon the measurement of 'cold', they used 'cold' as the underlying driving force for processes, and viewed cold climates in a zonal manner based on that same perception of the role of 'cold'. With technical and logistical landforms.

The material presented in this thesis comprises an attempt to reconsider the basic concepts of weathering in cold environments, not from any pre-ordained perspective but rather from a data-rich approach. Hopefully the outcome is that it may have helped change our perceptions of the weathering processes in cold regions it. Ironically, the high-frequency data actually proved the operation, within rock, of freeze-thaw in a field situation by monitoring of an exotherm (the latent heat released as water changed to ice). Thus, freeze-thaw is not "dead", but rather "alive and well" and in need of better investigation! Overall, the essence is that weathering in cold regions is just not that simple: the simplistic picture created in many textbooks or assumed in many papers does not acknowledge the complexity of the actual processes taking place. Rather weathering is highly complex, both in space and time. Synergistic relationships between a whole host of mechanical, chemical *and* biological processes may be operating and this synergy changes in space and time. Here time has to be considered from that of diurnal variability, through that of the seasons, to the longer-term (millenia); the complications of future climatic change provide further challenges and urgency to our need for understanding. The weathering synergy at any one place is also the outcome of rock properties and how these change through time.

A significant outcome of the study is that thermal conditions are only part of the picture. While it is clear that the thermal attributes have dominated the literature, the role of water is equally important - but not so recognised. The contradiction is that the presence of water is as necessary as the thermal conditions for freezing to allow for the operation of freeze-thaw weathering. Once that notion is accepted a whole range of possibilities, not the least the inadequacy of moisture-related data, for weathering processes changes dramatically. With water the potential for chemical processes increases dramatically, particularly when *rock*, rather than air, temperatures are viewed. Indeed, a whole range of recent studies has shown that chemical weathering, although spatially and/or temporally constrained by moisture availability, is more prevalent in cold regions than has hitherto been considered. With these moisture-related thoughts in mind, our perceptions of the applicability of laboratory experimentation becomes all the more fragile. If not based on realistic thermal and moisture conditions, can laboratory experiments have any real meaning to the field situation? The laboratory findings in this thesis, based as they are on actual field conditions, show very different outcomes to those which use hypothetical conditions; conditions conducive to effective breakdown by freeze-thaw (i.e. large amounts of water and adequate freezing conditions).

If anything, the findings here point towards an exciting future where so little is really known about weathering in cold regions, and what a diversity of regions there are. Thus, there is great potential for better, more rigorous and longer-term studies than those given here. Hopefully ,perhaps, the material within this thesis opens the door to thinking of weathering in cold regions, in a new way such that the science can go forward, not by admiring our models "...or by proclaiming how well they seem to fit our observations..." but rather by seeking conflict "...between our models and the real world" (McCarroll, 1997, p.1).

"The great tragedy of Science - the slaying of a beautiful hypothesis by an ugly fact"

T. H. Huxley

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CORE LOG, PHOTOGRAPHS AND GRAIN SIZE ANALYSES FROM THE CIROS-1 DRILLHOLE WESTERN McMURDO SOUND, ANTARCTICA.

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MAY 1987

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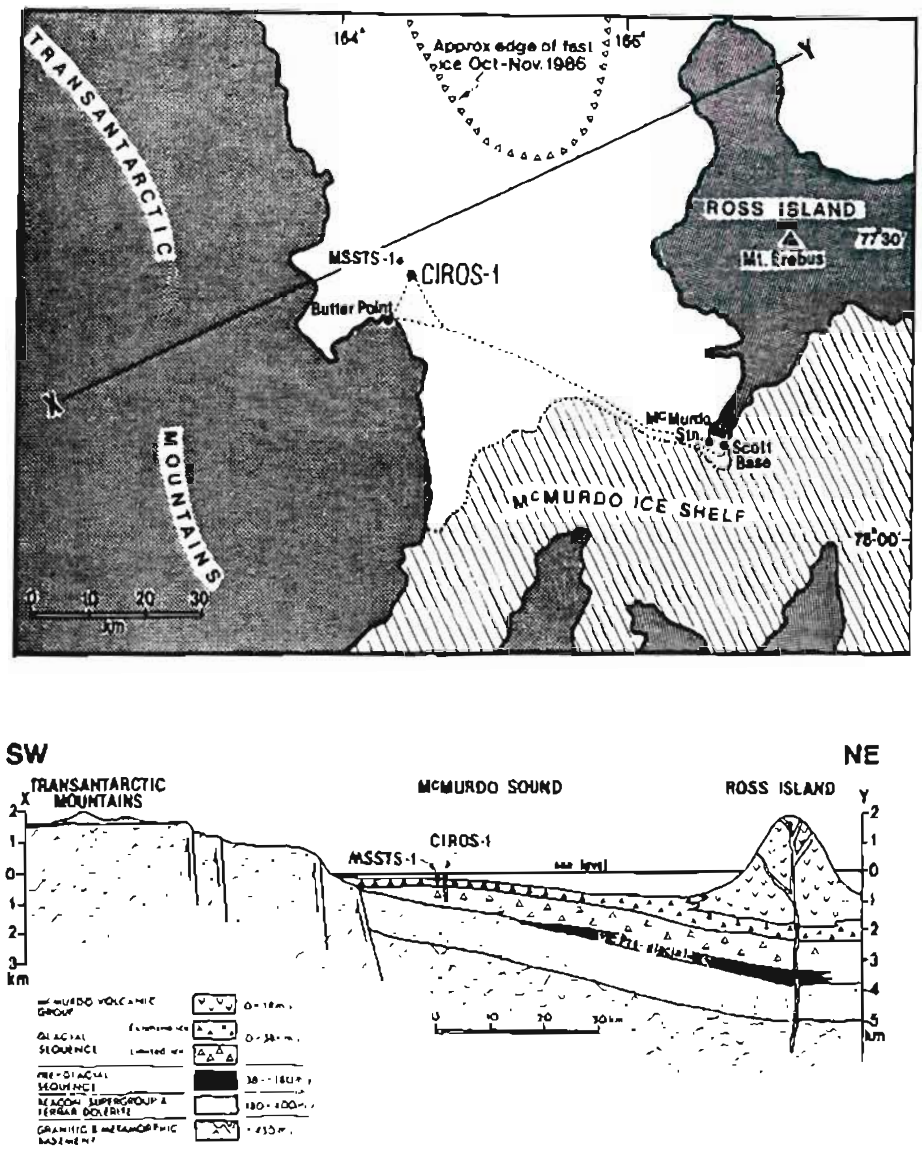


Figure 1. Map of McMurdo Sound, western Ross Sea, and geological cross-section, showing the location of MSSTS-1 and CIROS-1.

Table 1:

Site data for CIROS-1.

Position:	12 km northeast of Butter Point, McMurdo Sound. (77°34' 54.7397"S; 164°29' 55.8950"E)
Water depth:	197.5 m. Sea ice thickness: 2 m.
First core taken:	15 October 1986. Last core taken 14 November 1986.
Depth drilled:	702.14 m sub seafloor.
Depth cored:	26.69 to 702.14 m (98% recovered).
Lithology:	Diamictite, mudstone, sandstone and mixed conglomerate.
Age of oldest sediment:	Early Oligocene (foraminifera and diatoms)

DRILL SITE OPERATIONS

The core was processed as drilling proceeded in a 24 hour operation. A.R. Pyne supervised operations with J.N. Ashby, E.F. Hardy, C. Mills and B.D. Morris processing and readying the core for lithological description.

Core Depth

All core depths are recorded as metres beneath the sea floor after correction for a water depth of 197.5 m and tide which ranged up to 0.8 m.

The assigning of depths for the core within a core run were determined from a variety of features, such as percent recovery, continuity with the previous core run, core spin and catcher marks and known no recovery zones. Drilled depths were determined from the tensioned sea casing anchored to the seafloor and are therefore independent of tidal movement, which affected the sea ice and drilling platform.

Core Splitting

The CIROS-1 core was split lengthwise in 1 m lengths using either a specially constructed diamond saw splitting table for hard core, or a knife for those few intervals of very soft core. Core splits were boxed separately. One split was cleaned ready for core photography, description, sampling and eventual shipment to New Zealand. The other split was packaged for shipment to the United States.

Core storage facilities:

New Zealand Geological Survey
P O Box 30-368
Lower Hutt
New Zealand.

Sample requests to: Dr. P.J. Barrett, Victoria University of Wellington

Antarctic Research Facility
Department of Geology
Florida State University
Tallahasse, Florida
United States of America

Sample requests to: Mr. Dennis S. Cassidy

Core Photography

The New Zealand boxed core split was photographed on-site with Ilford 220 black and white and Ektachrome 120 colour films. Bounced colour corrected flash light was used to illuminate core boxes and for closeup core photography. Black and white photographs of the core boxes are presented in Appendix 1.

PROCEDURE AND NOMENCLATURE USED IN CORE LOGGING

Detailed core logging was performed on-site by P.J. Barrett, K.J. Hall, M.J. Hambrey and P.H. Robinson. The log format is a standard graphical and descriptive presentation using nomenclature and symbols (Fig. 3) modified for core from Andrews (1982) and Shell Standard Legend (1976).

Core Features

The CORE column is a sketch depicting as near as possible the nature of the cut core face. Features described during core processing, such as core breaks, fractures, joints and large clasts were combined with more detailed and additional features recognised in core logging including veins, faults, strength and style of bedding, and their attitude (in degrees) relative to horizontal.

Core breaks are represented as a bold line; fault, fracture, vein and clast outlines are represented as fine lines, and bedding is depicted as dashes of varying length depending upon continuity.

Grain Size Estimation

Average grain size (Wentworth scale) has been visually estimated through the entire core and is depicted as a vertical line in the MEDIAN GRAIN SIZE column. Those lithologies with wide grain size distribution were given an average grain size for the matrix only (that is, for sediment 4 mm and finer). The coarser fraction (gravel) was visually estimated from the cut core face using comparison cards, and plotted as GRAVEL %. The percentage of intraclasts is shown in the gravel column in brackets. Grain size estimates have not been adjusted to conform with the results of the grain size analyses.

Lithology and Sedimentary Features

The LITHOLOGY column is divided in two. The left side graphically illustrates gross lithology, modified where appropriate for lithologic qualifiers (muddy, sandy, pebbly, etc). Also shown are the degree of induration and cementation, the presence of assessory minerals (pyrite, mica, etc), and the various clast types and shapes. The nature, thickness and strength of bedding, bedding features, fossil types and miscellaneous sediment features are graphically portrayed using standard sedimentary symbols plotted in the right column.

Clast Composition and Shape

Clast composition, size roundness, surface texture (striae) and percentage were determined for most units and subunits. These data will be presented elsewhere (Hall, in Barrett, in prep), but some data on clast composition and striae are recorded in the LITHOLOGY column, as well as in the description of each subunit, which summarises clast size and roundness data.

Lithological Description

The core was divided into subunits at the time of logging, each "subunit" representing a distinct lithofacies or facies association. Major unit boundaries were also designated on-site where marked changes in lithofacies occurred.

Each subunit description is set to a standard format:

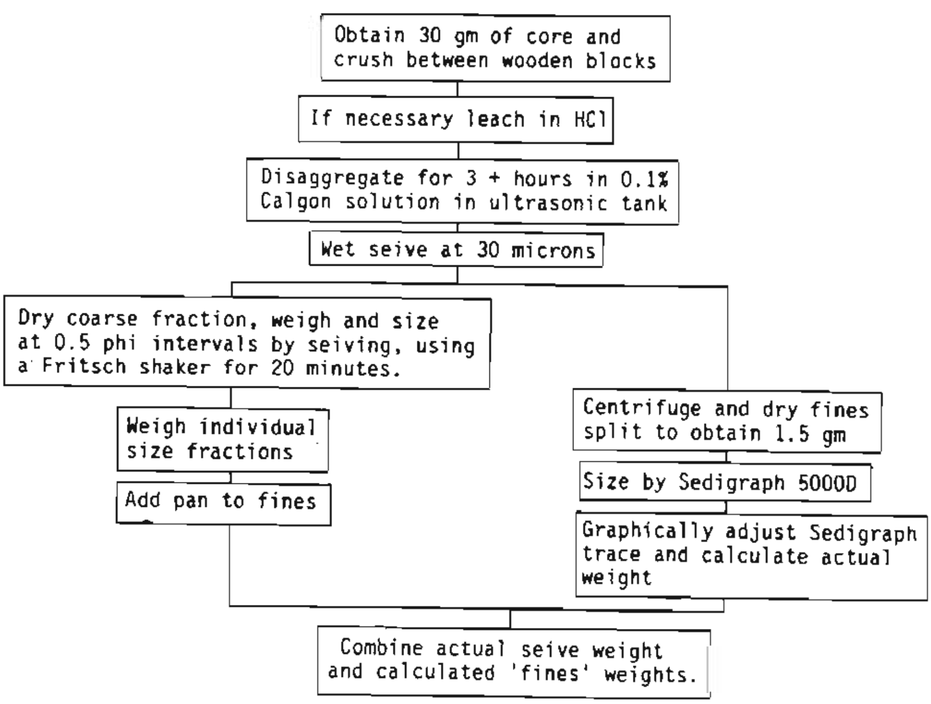
Nature of upper contact
Unit and subunit number
Subunit interval (depth)
Gross lithology(ies)
General description and colour
Stratification
Clasts
Other features

The subunit description begins with the degree of induration, cementation and stratification, a full textural name following the modified Folk textural classification (Andrews 1982), and colour (based on the Munsell Colour Chart). Other information presented for fresh wet core includes the nature of stratification, clast percentage, size, roundness and composition, sedimentary structures, (including soft sediment deformation features), biogenic material and tectonic features.

GRAIN SIZE ANALYSIS

Fifty-one samples were selected both to represent the range of facies and to cover the entire core. About 20 g was crushed between wooden blocks to pass through a 2 mm sieve. Calcareous samples were leached in hydrochloric acid. Others were stirred in 0.1% Calgon solution in an ultrasonic tank for 3 hours to aid disaggregation. Samples were checked for aggregates and some required up to 12 hours ultrasonic treatment for full disaggregation. The remainder of the analytical procedure is outlined in figure 3. The coarse (coarser than 5 phi) fraction was sized at ½ phi intervals by sieves and a Fritsch automatic shaker for 20 minutes on "intermittent". The fine fraction from the pan was added to the fines from wet sieving, all of which was split to provide a 1.2 - 1.6 g sample for settling analysis using the Sedigraph 5000D (Stein 1985). Macpherson (1986) discusses both the sieve calibration and correction of Sedigraph curves. The results of the grain size analyses are summarized in figure 4, and presented in detail in Appendix 2.

Figure 3. Flow chart for grain size analysis of CIROS-1.



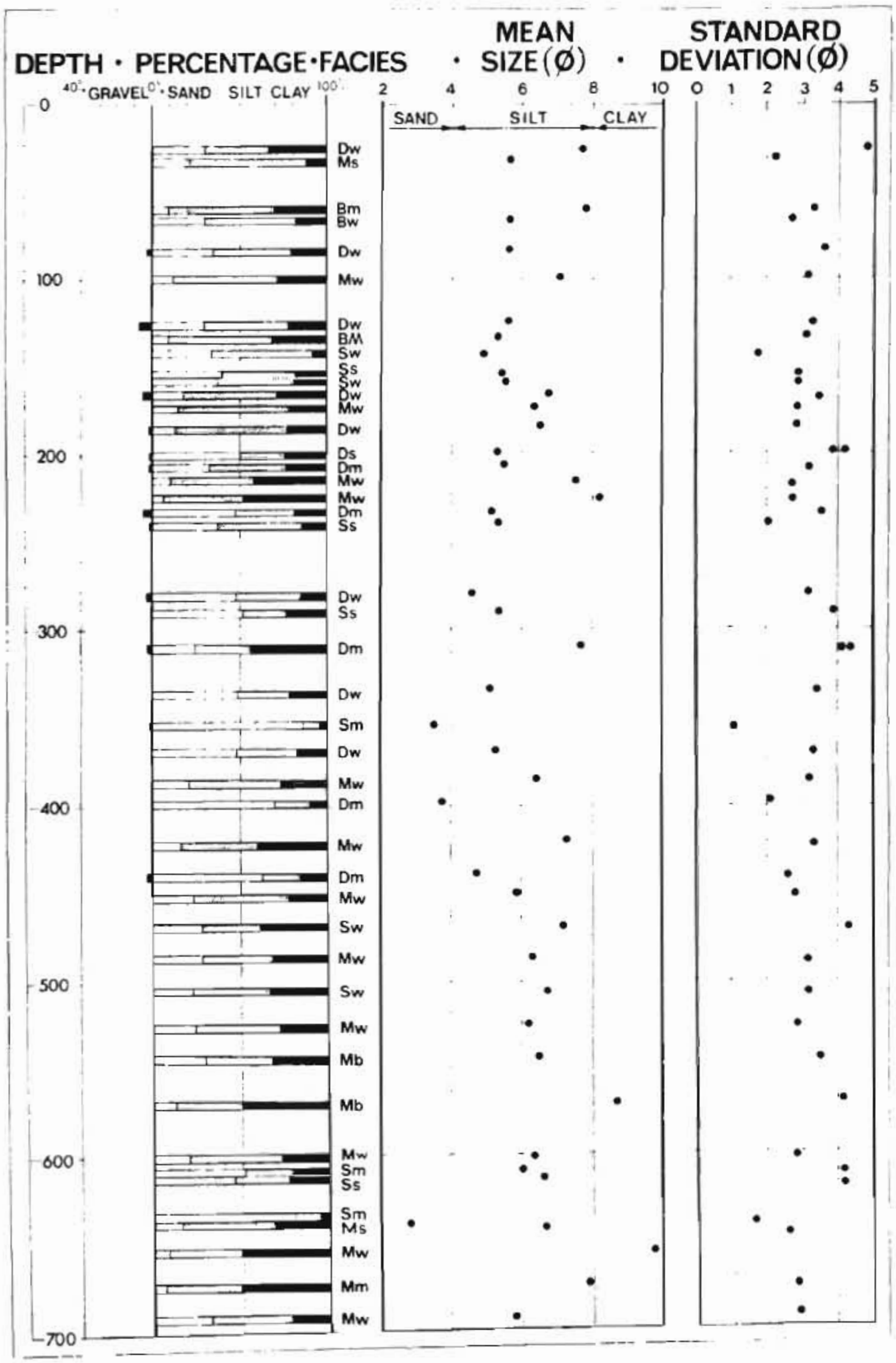


Figure 4. Grain size data for CIROS-1. Percent gravel was taken from the core log; percent sand, silt and clay were determined by analysis (see text for method and Appendix 2 for complete analyses).

Lithofacies coding: B - breccia, D - diamictite, M - mudstone and S - sandstone, with suffixes, m - massive, w - weakly stratified, s - moderately to well stratified and b - bioturbated.

ACKNOWLEDGEMENTS

The CIROS project was a joint venture between Victoria University of Wellington, who coordinated the science programme, Geophysics Division DSIR who supervised the drilling, and Antarctic Division DSIR, who provided the logistic support. The drilling rig was provided by the US National Science Foundation. Special thanks to Peter Gallagher and Terry Best (NZ Geological Survey) for their considerable effort in drafting the core logs. Figures 1 - 4 were drafted by Ted Hardy (VUW). Frank Williams (VUW) carried out the grain size analyses. John Casey (VUW photographic facility) helped develop the photographic system for the core and processed the core photographs.

Finally we thank the drillers, planners and those in the support system for the project. Without their efforts such an abundance and quality of core would not have been possible.

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Kevin Hall

LEGEND

LITHOLOGY

Diamictite

Mudstone

Rhythmite

Sandstone

Conglomerate

LITHOLOGY QUALIFIERS

muddy

sandy

pebbly

carbonaceous

calcareous

LITHOLOGY ACCESSORIES

Py Pyrite

C Chlorite

M Mica

FOSSIL TYPES

Fossils in general

(broken)

Mollusc (articulated)

Scaphopod

Mollusc (single valve)

Gastropod

Echinoid

Sponge spicule

Leaf

Wood fragments

Indeterminate plant debris

Foraminifera in general

CLAST TYPES

Basement Complex

Ferrar Dolerite

Beacon Supergroup

McMurdo Volcanic

Indurated sediment

Quartz

Limestone

Mudstone clast

Sandstone clast

Diamictite

CONTACTS

Sharp

Gradational

Erosional (irregular)

SEDIMENTARY STRUCTURES

Weakly bedded

Well bedded

Inclined bedding

Graded bedding (normal)

Wavy bedding

Flaser bedding

Asymmetrical ripples

Bioturbation structures

Mottling

Contorted bedding

Load structures

Flame structures

Bedded

Horizontal bedding

Cross bedding

Graded bedding (reverse)

Discontinuous bedding

Lenticular bedding

Symmetrical ripples

Burrows

Convolute bedding

Contorted and disrupted bedding

Sand dykes

Water escape structures

MISCELLANEOUS SYMBOLS

-o- Lonestones

Concretions/Nodules

frac Fractures

Slickensides

Ca Calcite vein

No core recovery

Dropstones

△ Mouldic and vugular porosity

F Faults

T Indurated

Cl Clay vein

Striations

Clast rind / Weathering rind

jts Joints

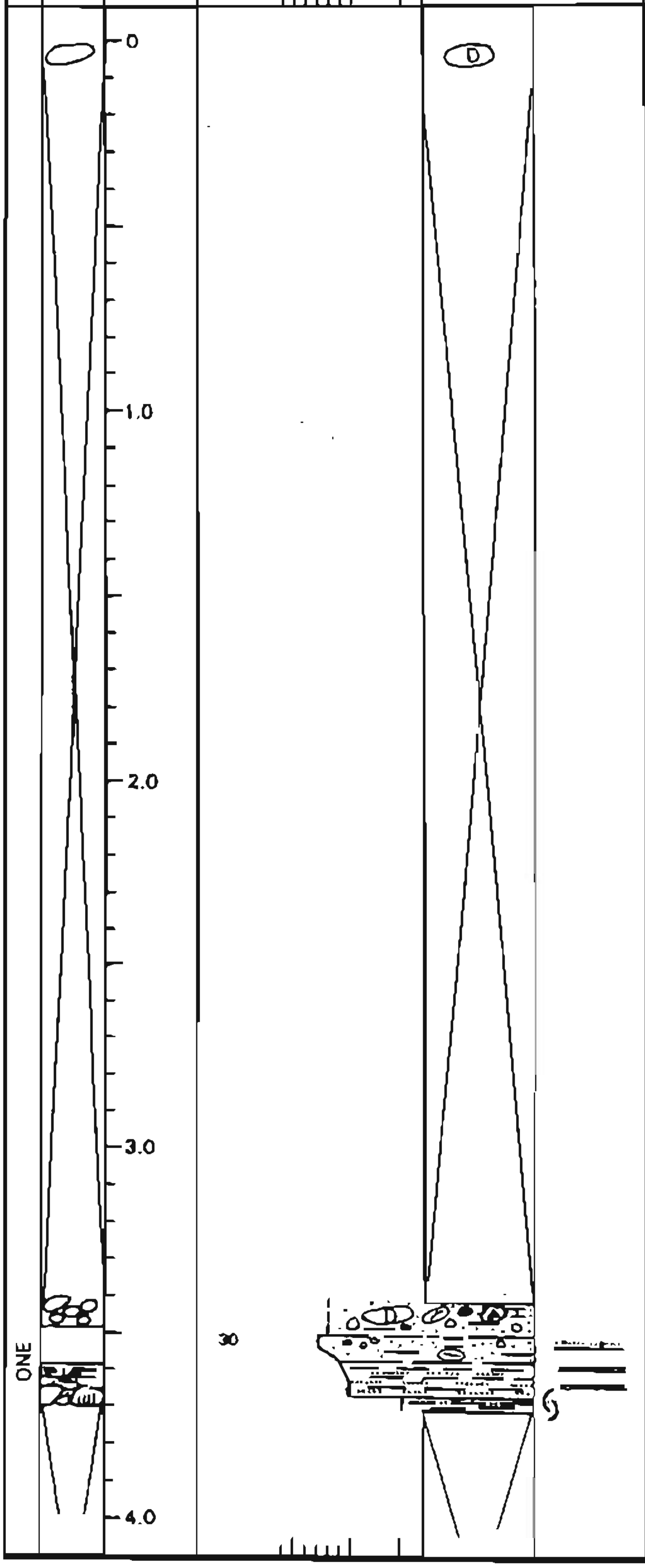
cmt Cemented

PROJECT : CIROS -1

SHEET NO. : 1

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 1.1

3.41 - 3.70 m

MUD AND SAND TO DIAMICTON

Firm, well stratified (mm-bedded) mud (grey 10Y 5/1) interbedded with soft, moderately well sorted, fine sand (grey 10Y 4/1) (rhythmites) overlain by soft, slightly calcareous, unstratified poorly sorted, **slightly gravelly muddy sand (diamicton)**(greyish olive 7.5Y 5/2), which coarsens upwards. The sand throughout consists of largely basaltic grains.

Clasts in the diamicton are mainly 15 to 60 mm long and include dolerite, basement and basic volcanics. There is also a 68 mm long pebble of carbonate–cemented diamictite and a dolerite >80 mm. Clasts range from angular to subrounded, and are on average subangular, with several striated

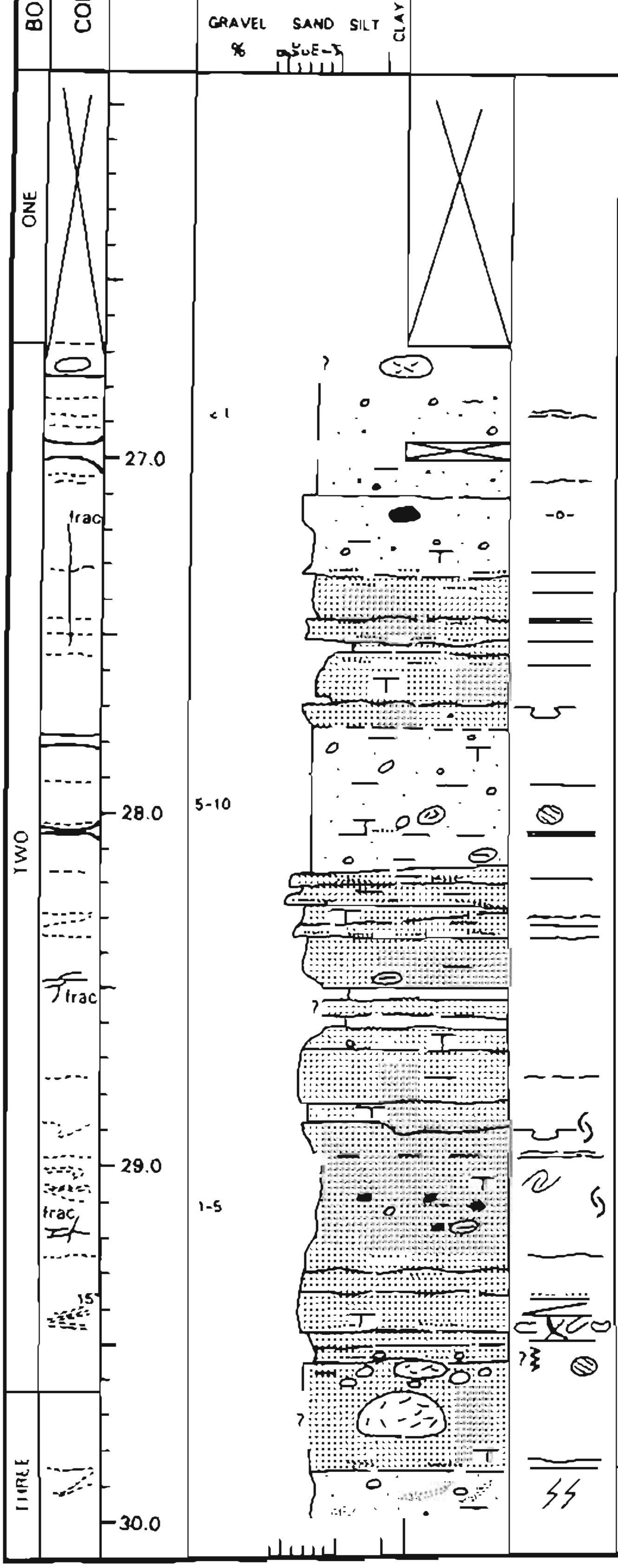
Unstratified thin bed of **silty clay** (olive grey 10Y 4/2) between 3.66 and 3.70 m

PROJECT : CIROS -1

SHEET NO. : 2

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



— contact in non-cored interval -

Sub-unit 2.1

26.69 - 29.85 m

SANDSTONE AND DIAMICTITE

Hard, moderately well stratified (cm- and mm-bedded), moderately well-sorted to poorly sorted, **coarse to medium sandstone** (grey 10Y 4 1) with minor interbeds of weakly stratified (mm-bedded), moderately well sorted, **muddy fine sandstone** and **sandy mudstone** (greenish grey 10G 5 1).
Clasts are on average subrounded and two are striated Clasts form 1 to 10% of the core. They range in size mainly from 5 to 30 mm, but 3 are >50 mm Lithologies include basement, dolerite and basic volcanic rocks.

Beds of hard, weakly stratified (mm- and cm-bedded), slightly **conglomeratic muddy medium to fine sandstone (diamictite)** (dark greenish grey 7 5GY 4 1) from 26 79 to 27.32 and from 27 75 to 28 15 m

Between 29.05 and 29.30 m bedding is contorted and disrupted with deformation on a scale of 1 to 3 cm

?Zeolite-lined cavities and surfaces with the form of burrows and bivalve shells in coarse sandstone bed from 29 45 to 29.48 m. Horizontal slickenside-like lineations in mudstone above clast at 29 57 m

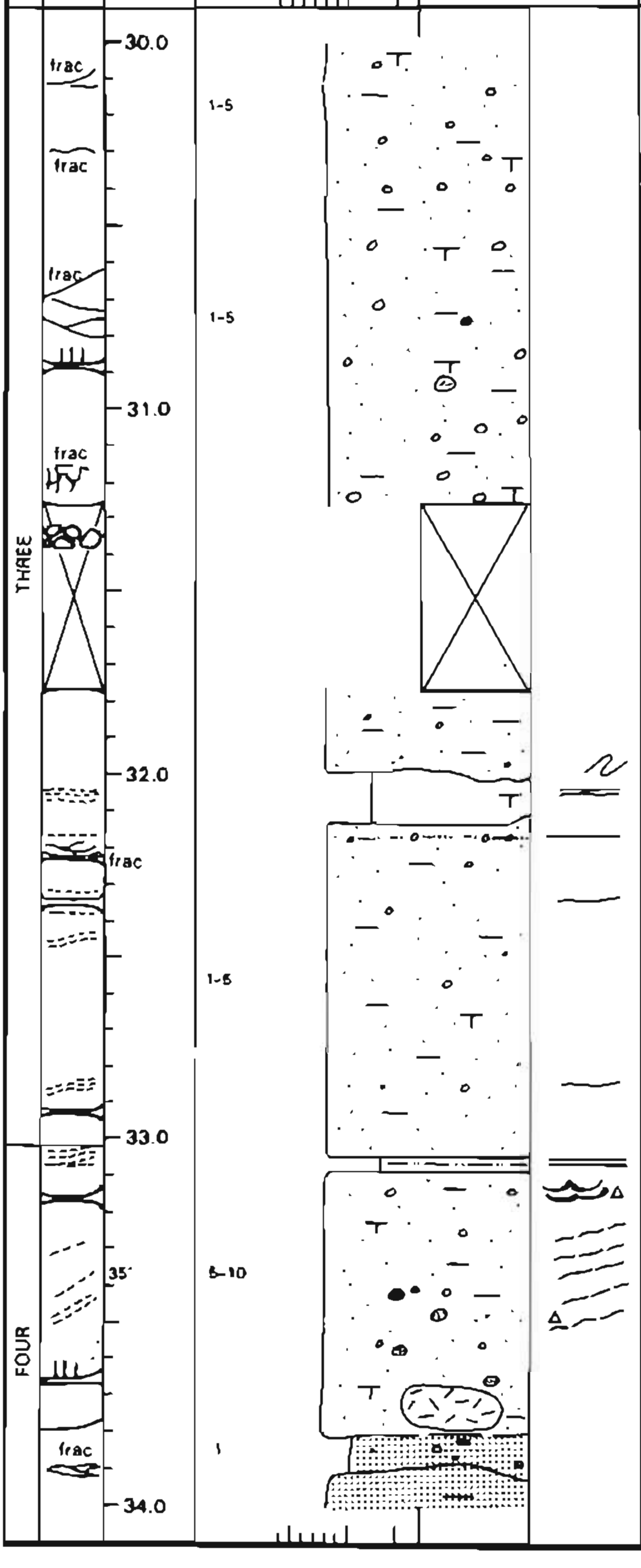
— contact gradational -

PROJECT : CIROS -1

SHEET NO. : 3

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 2.2

29.85 - 33.06 m

DIAMICTITE

Hard, **slightly conglomeratic muddy medium to fine sandstone (diamictite)** (dark greenish grey 10GY 4/1).
Unstratified above 31.98 m, weakly and diffusely stratified below.
Clasts up to 15 mm form 1 to 5% of rock. Clasts range from angular to rounded but on average are subangular. Lithologies include basement and basic volcanic rocks.
Clast-free, stratified (cm-bedded), moderately well sorted, **sandy mudstone** (dark greenish grey 10G 4/1) with gradational contacts from 31.98 to 32.33 m.
Stringer of coarse sandstone to fine conglomerate at 32.33 m.

— contact sharp -

Sub-unit 2.3

33.06 - 33.80 m

DIAMICTITE

Hard, weakly stratified (mm- to cm-bedded), **conglomeratic muddy medium sandstone (diamictite)** (dark greenish grey 10GY 4/1).
Bedding dips steeply (up to 35°) and is emphasised by dissolution.
Clasts average 2 to 8% of core and are mostly between 5 and 18 mm long, though one large granite clast is > 140 mm. Bedrock clasts are largely basic volcanic. Intraclasts of sandstone and grit are also common. All are on average subrounded.

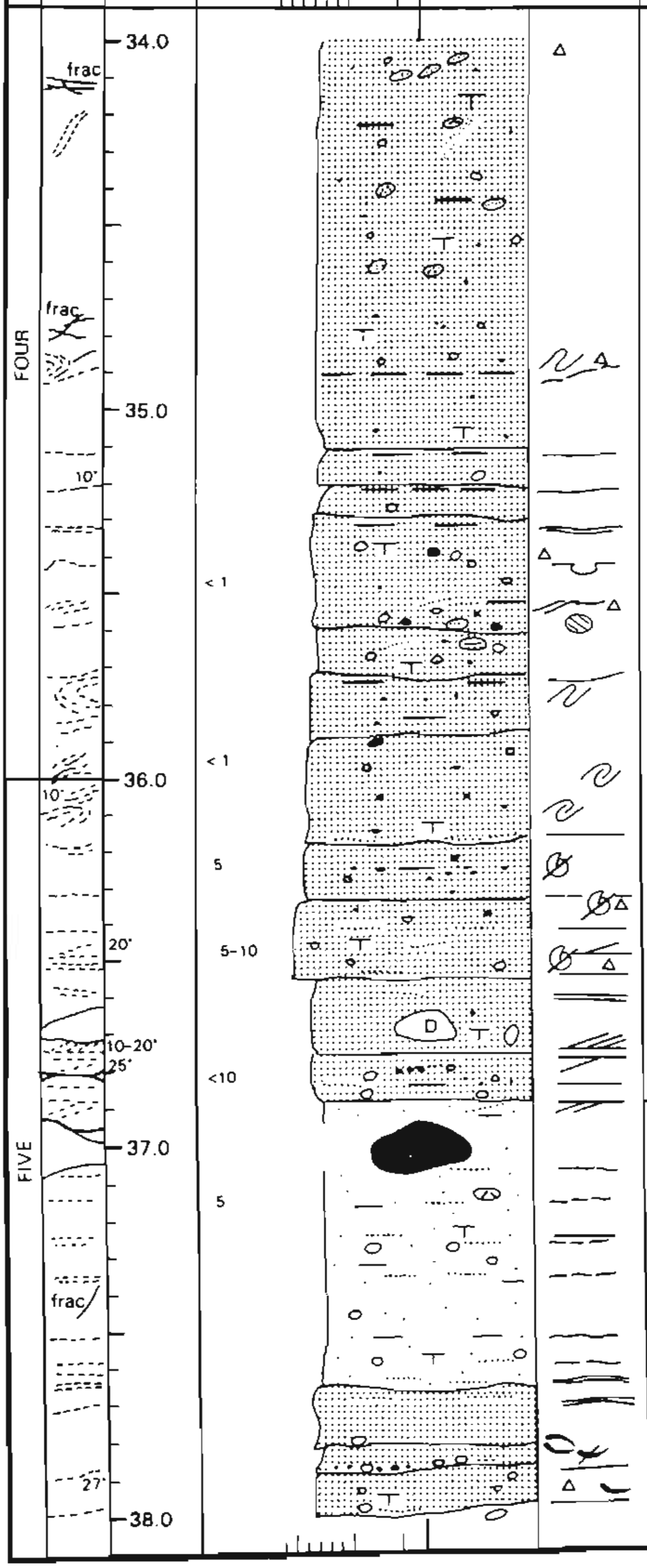
— contact at base of large clast -

PROJECT : CIROS -1

SHEET NO. : 4

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 2.4

33.80 – 36.88 m

SANDSTONE

Interbeds of hard, unstratified, moderately well sorted, **very fine sandstone** (dark greenish grey 10GY 3 1) and weakly stratified (cm-bedded), very poorly sorted, **coarse sandstone** (dark greyish yellow 2.5Y 4 2).
Stratification is irregular, discontinuous and cross-bedded with dips up to 25 . Common contorted and disrupted bedding.
Clasts form <1% in upper part of sub-unit to around 10% near base. Bedrock clasts are subangular to subrounded and from 5 to 10 mm long. One is striated and several are faceted. Lithologies include basic volcanics, basement and dolerite. Slightly larger (up to 22 mm) subrounded coarse sandstone and mudstone intraclasts also occur.
Vugular and shell mouldic porosity widespread.

– contact sharp -

Sub-unit 2.5

36.88 – 38.82 m

DIAMICTITE AND SANDSTONE

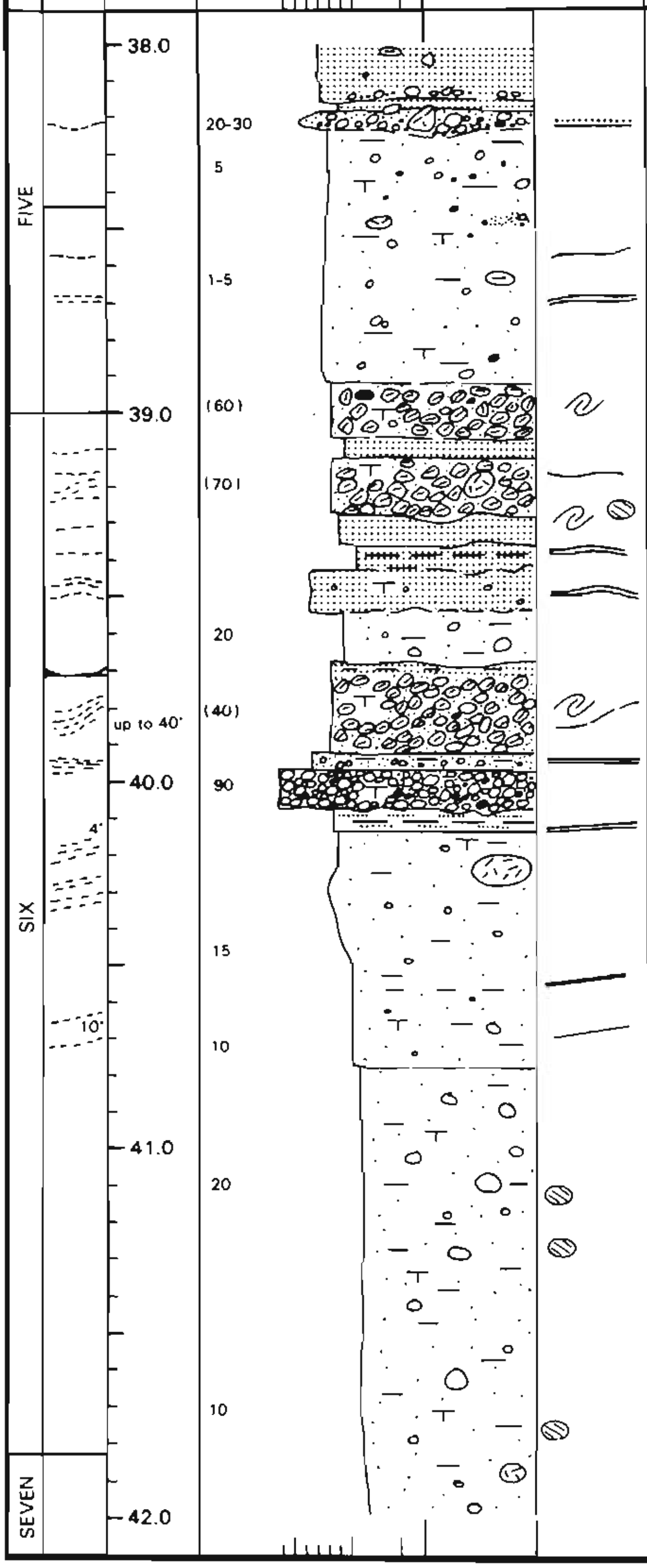
Hard, weakly stratified (cm-bedded), **conglomeratic muddy medium to fine sandstone (diamictite)** (dark greenish grey 10Y 4 1), separated by hard, weakly stratified to well stratified (cm-bedded), moderately poorly sorted, **coarse to fine sandstone** (dark greenish grey 7.5GY 4 1 to 10GY 3 1).
Discontinuous and irregular stratification marked by coarse sandstone partings.
Clasts in the diamictite form 5% of the core. They range in size mainly from 5 to 25 mm, but several are >50 mm and a volcanic clast at 37.00 m is >160 mm. Clasts on average are subangular and several are faceted. Lithologies include basement(40%), basic volcanics(30%), dolerite(10%) and chert.
Cavities in coarse sandstone from 37.82 to 37.95 m have the form of bivalve shells (one articulated)

PROJECT : CIROS -1

SHEET NO. : 5

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



- contact sharp -

Sub-unit 2.6

38.82 – 40.08 m

MUDSTONE BRECCIA AND DIAMICTITE

Three beds of moderately hard, poorly sorted **brecciated mudstone** (dark bluish grey 5BG 4/1) (38.8 to 39.07 m, 39.14 to 39.28 m, 39.68-39.92 m).

Bedding on mm- to cm-scale, moderately well developed but contorted with dips up to 40°.

From 40 to 80% of the rock is formed of clasts up to 20 mm long. Between 70 and 90% are intraformational mudstone, the remainder being of basement and a few basic volcanic (mainly subangular to subrounded).

Bed of soft, unstratified, **conglomeratic muddy fine to very fine sandstone (diamictite)** (greenish grey 10Y 6/1) from 39.55 to 39.68 m

Clasts form 20% of core, and include basement and minor volcanics as well as intraformational mudstone. They are mainly subangular to subrounded. There is one striated clast.

Basal bed of unstratified, moderately well sorted, **pebble conglomerate** with 10% matrix. Pebbles range up to 10 mm and are rounded to subangular. Lithologies include basement (80%) and basic volcanic (20%).

- contact erosional -

Sub-unit 2.7

40.08 - 40.78 m

DIAMICTITE

Hard, weakly to well stratified (mm- to cm-bedded), **conglomeratic muddy fine to very fine sandstone (diamictite)** (greenish grey 10G 5/1).

Stratification is subhorizontal to inclined (4–10°); weakly bedded at the base to well bedded in the upper 6 cm which is interlaminated sandstone and mudstone (rhythmite).

Clasts form 10% of core and range in size from 5 to 10 mm, though one basement clast is >60 mm. Basement is the predominant clast type.

Possible shell fragments. Thin bed of fine volcanic sandstone and quartz, feldspar, brown amphibole and minor glass at 40.35 m.

- contact sharp -

Sub-unit 2.8

40.78 - 42.84 m

DIAMICTITE

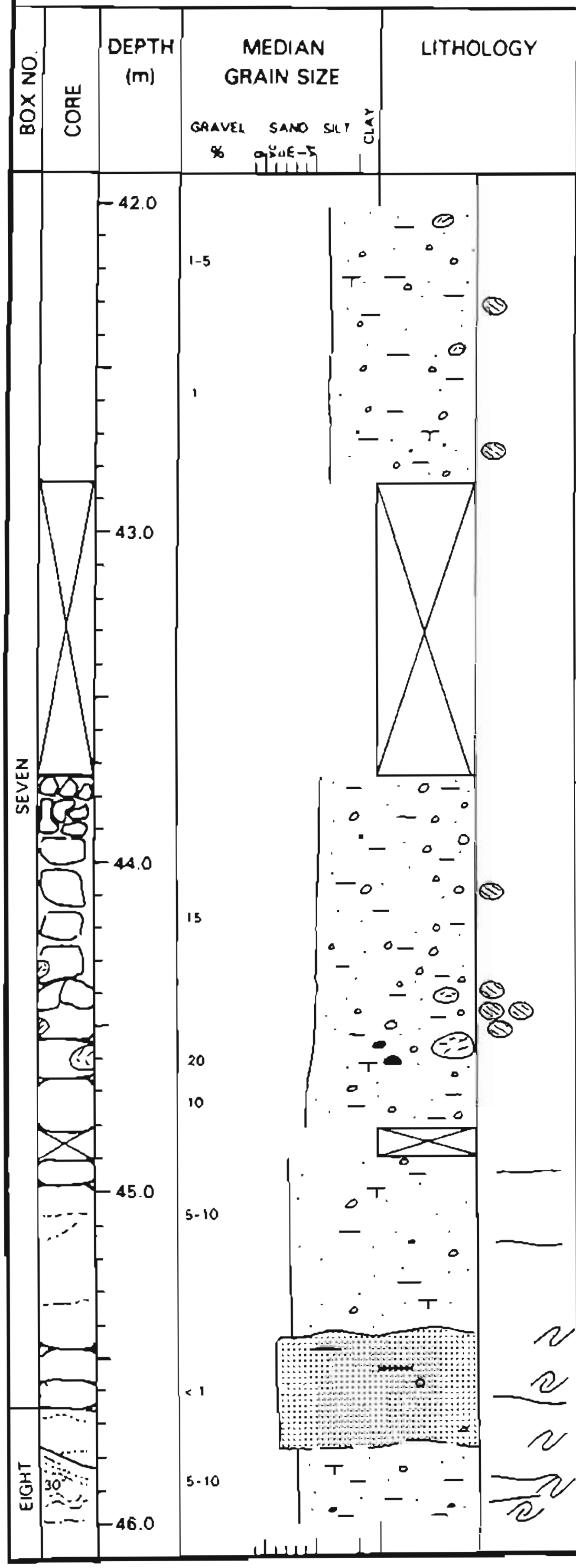
Variably hard to soft, unstratified, conglomeratic sandy mudstone (diamictite) (greenish grey 10G 5/1)

Clasts form 20% of core in upper part of sub-unit but proportion declines to 1% near base. They range from 5 to more than 30 mm long; they are angular to subrounded and on average are subrounded. Almost half are striated. About 70% are of basement and 30% are basic volcanic lithology.

PROJECT : CIROS -1

SHEET NO. : 6

SCALE : 1:20



DESCRIPTION

— contact in interval of lost core -

Sub-unit 2.9

43.72 - 44.81 m

DIAMICTITE

Soft, unstratified, **conglomeratic sandy mudstone** (diamictite)(dark greenish grey 10GY 4 1).
Clasts form around 15% of core and range from 10 to 20 mm, but one gneiss is >90 mm. They range from angular to subrounded and on average are subrounded. About half are striated and faceted, and about 60% are basic volcanic, 35% are basement and 5% mudstone intraclasts
Possible shell fragments

— contact in interval of lost core -

Sub-unit 2.10

44.90 - 45 20 m

DIAMICTITE

Hard, weakly stratified **conglomeratic sandy mudstone** (diamictite) (dark greenish grey 10Y 4 1)
Original weak stratification almost completely destroyed by bioturbation
Clasts form 8% of core and range from 5 to 20 mm long. Most are angular to subrounded. About 60% are basic volcanic, 40% are basement. Minor mudstone intraclasts and a few vein quartz
A few shell fragments. Veins of calcite < 1 mm thick pervade sub-unit

— contact sharp to diffuse (loaded) -

Sub-unit 2 11

45.42 - 45.76 m

SAND

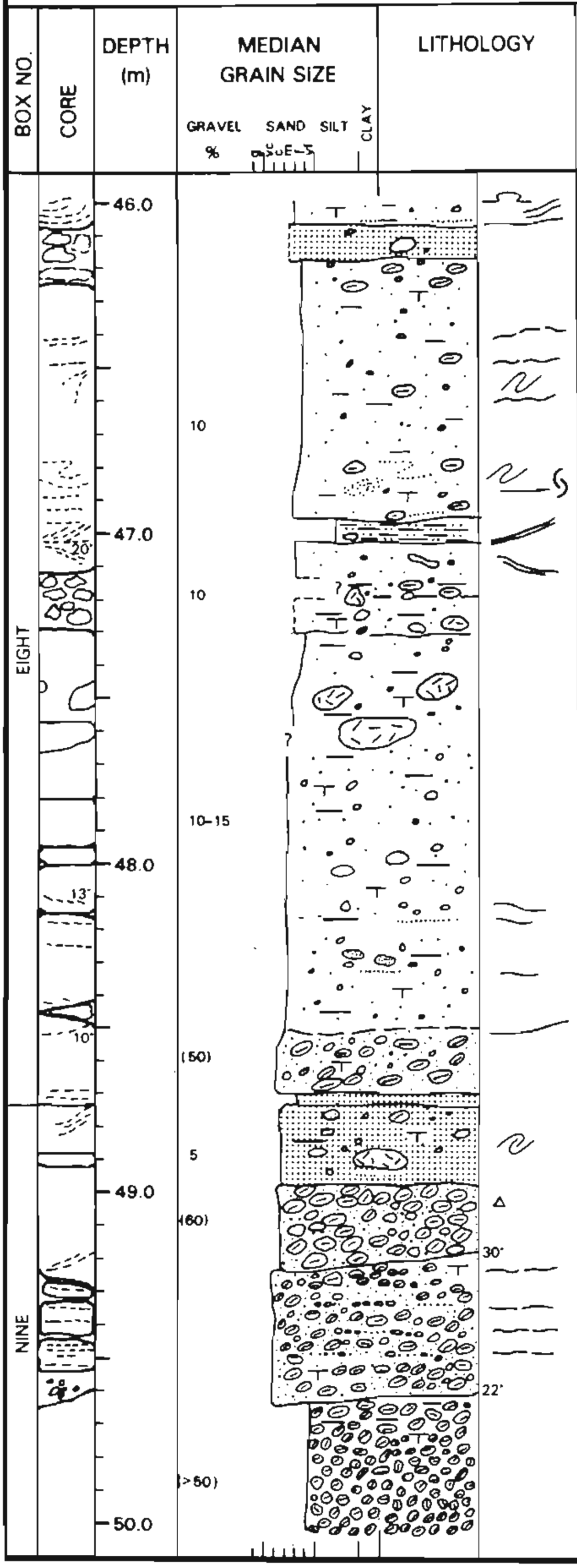
Soft, weakly stratified, poorly sorted **medium to fine sand**(olive black 10Y 3 1)
Stratification uneven and partly disrupted
Some angular mudstone intraclasts. High volcanic content

— contact gradational -

PROJECT : CIROS -1

SHEET NO. : 7

SCALE : 1:20



DESCRIPTION

Sub-unit 2.12

45.76 - 48.50 m

DIAMICTITE

Hard, weakly stratified (cm-bedded), **conglomeratic muddy fine sandstone (diamictite)** (greenish grey 5BG 5/1).
Original stratification partly disrupted, producing diamictite clast breccias, and in places contorted, dipping up to 30°.
Bedrock clasts form 5 to 15% of core and range from 4 to 20 mm long. About 60% are volcanic (including red porphyritic felsite) and 40% are basement.
Thin beds of unstratified, moderately well-sorted, **medium sandstone** (greenish grey 10GY 6/1) from 46.08 to 46.18 m and of weakly stratified, **mudstone** (dark bluish grey 5BG 4/1) from 46.96 to 47.04 m.

— contact sharp and loaded -

Sub-unit 3.1

48.50 - 62.13 m

MUDSTONE BRECCIA

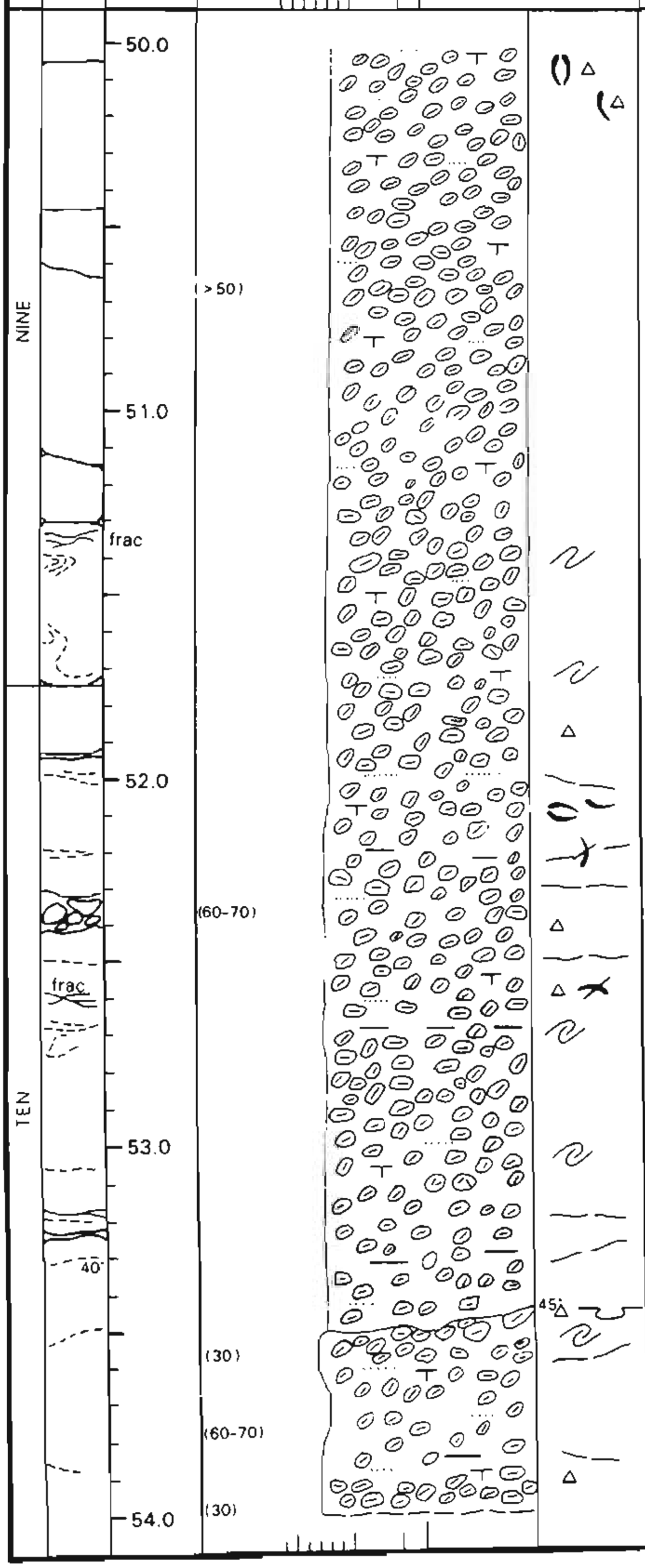
Hard, unstratified to weakly stratified, very poorly sorted, breccia of intraformational **mudstone** clasts in a **sandy mudstone** matrix (greenish grey 5BG 5/1 to dark greenish grey 7.5GY 4/1).
Stratification is locally contorted and disrupted, showing dips up to 30 and at 51.40 m the hinge of an isoclinal fold.
Matrix- and clast-supported with 80 to 90% mudstone clasts and the remainder bedrock derived. Mudstone clasts are on average 10 to 15 mm long and range up to 25 mm.
Thin bed of well-stratified, pebbly coarse to fine sandstone (dark greenish grey 7.5GY 4/1) from 48.73 to 48.98 m.

PROJECT : CIROS -1

SHEET NO. : 8

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 3.1

48.50 - 62.13 m

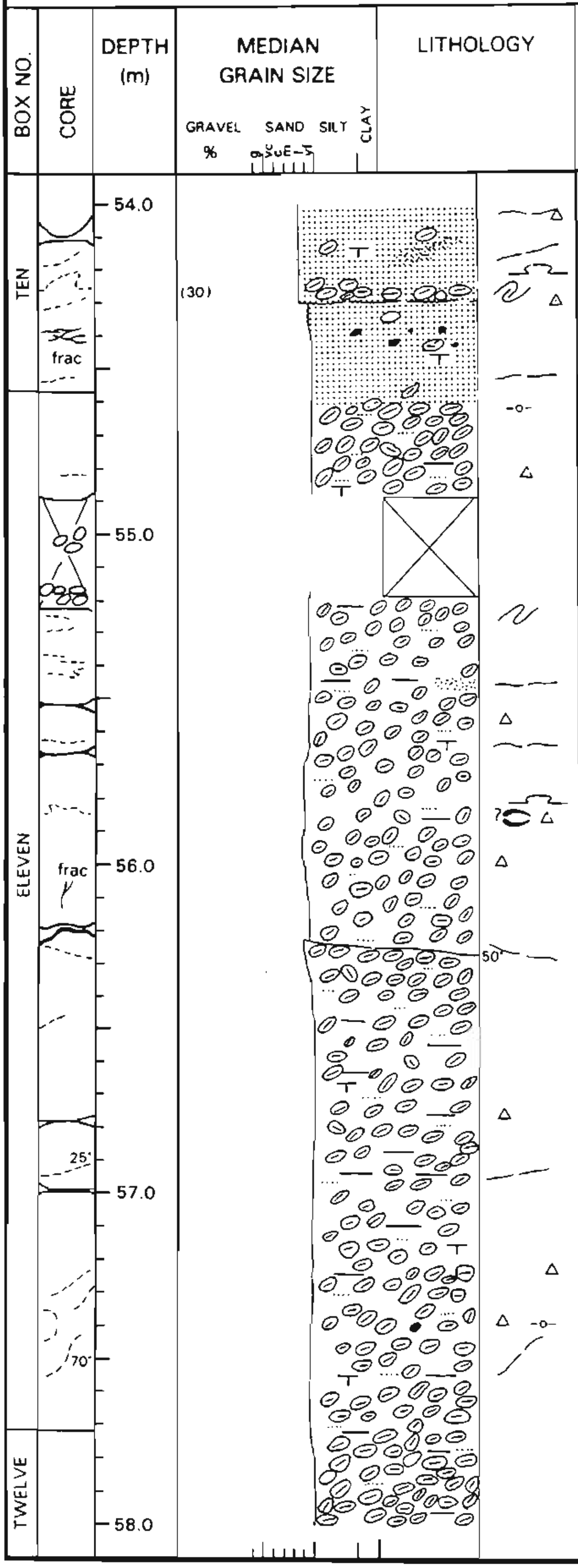
MUDSTONE BRECCIA

Hard, unstratified to weakly stratified, very poorly sorted, breccia of intraformational **mudstone** clasts in a **sandy mudstone matrix** (greenish grey 5BG 5/1 to dark greenish grey 7.5GY 4/1). Stratification is locally contorted and disrupted, showing dips up to 30° and at 51.40 m the hinge of an isoclinal fold. Matrix-and clast-supported with 80 to 90% mudstone clasts and the remainder bedrock derived. Mudstone clasts are on average 10 to 15 mm long and range up to 25 mm. Thin bed of well-stratified, pebbly coarse to fine sandstone (dark greenish grey 7.5GY 4/1) from 48.73 to 48.98 m

PROJECT : CIROS-1

SHEET NO. : 9

SCALE : 1:20



DESCRIPTION

Sub-unit 3.1

48.50 - 62.13 m

MUDSTONE BRECCIA

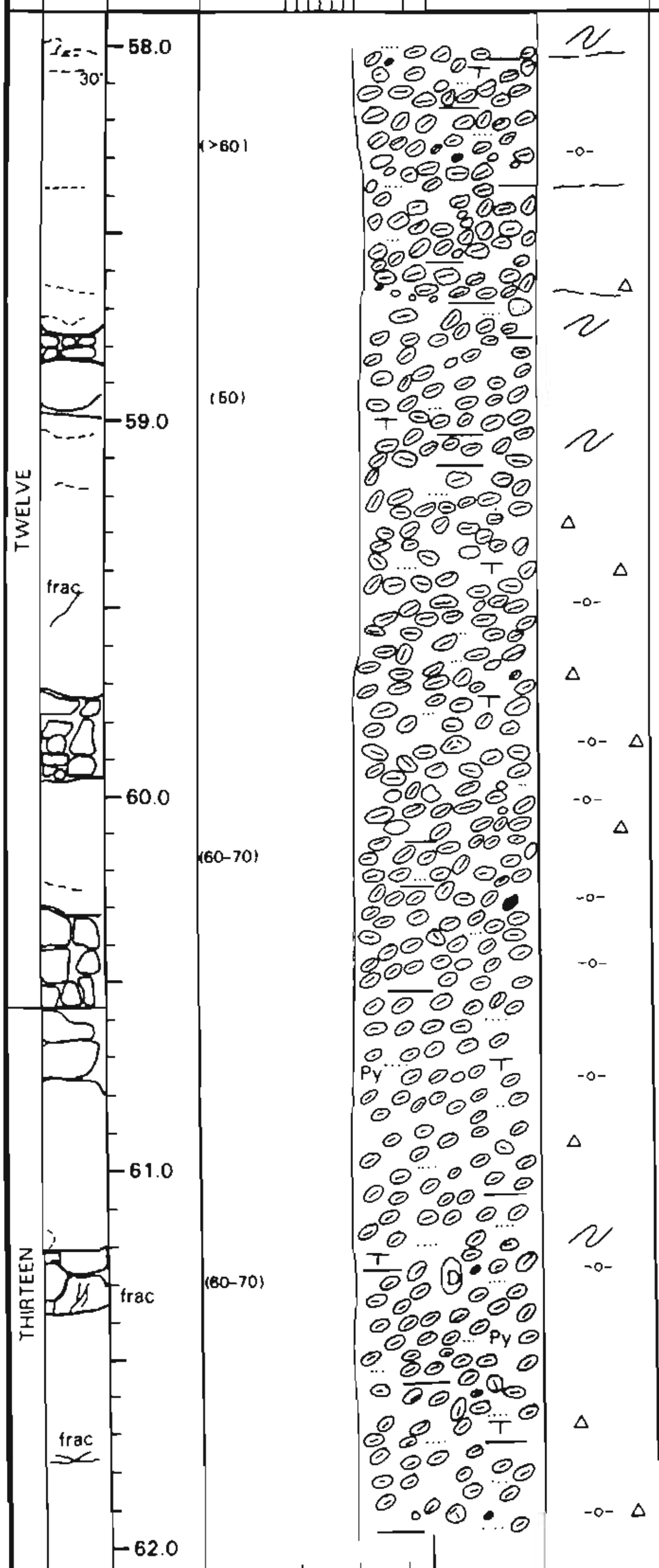
Hard, unstratified to weakly stratified, very poorly sorted, breccia of intraformational **mudstone clasts** in a **sandy mudstone matrix** (greenish grey 5BG 5/1 to dark greenish grey 7.5GY 4/1). Stratification is locally contorted and disrupted, showing dips up to 30° and at 51.40 m the hinge of an isoclinal fold. Matrix-and clast-supported with 80 to 90% mudstone clasts and the remainder bedrock derived. Mudstone clasts are on average 10 to 15 mm long and range up to 25 mm. Thin bed of well-stratified, pebbly coarse to fine sandstone (dark greenish grey 7.5GY 4/1) from 48.73 to 48.98 m.

PROJECT : CIROS -1

SHEET NO. : 10

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 3.1

48.50 - 62.13 m

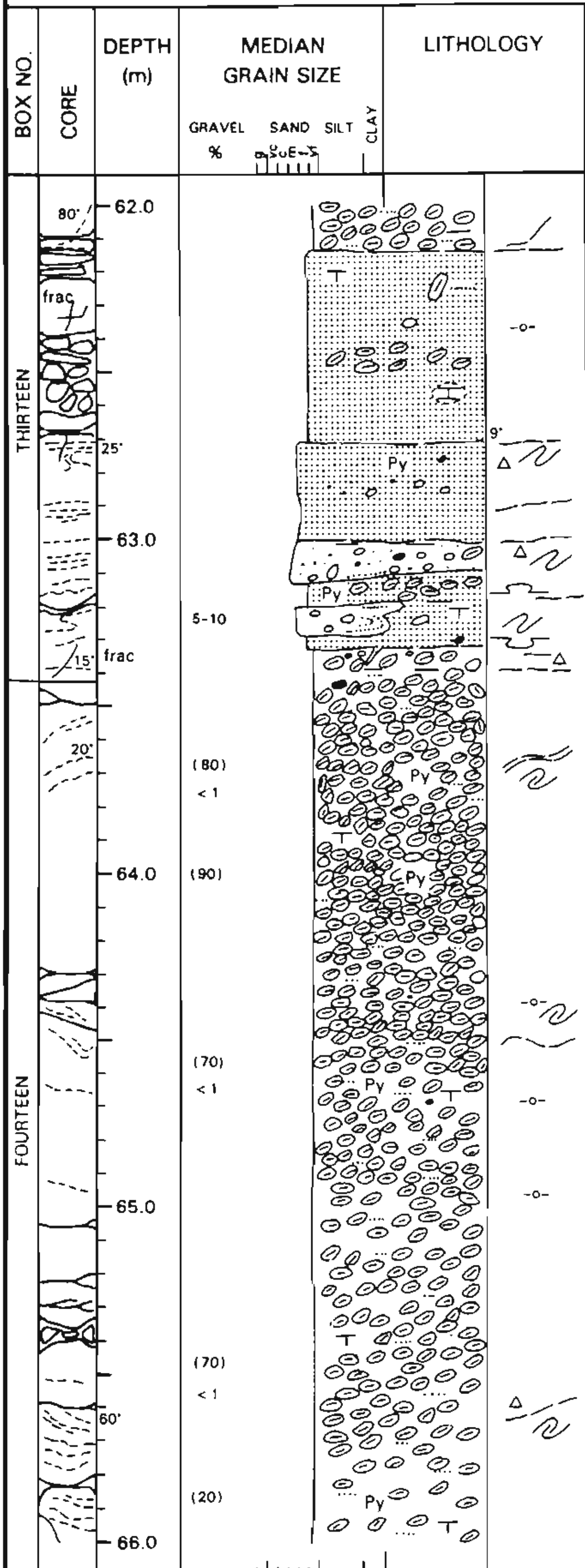
MUDSTONE BRECCIA

Hard, unstratified to weakly stratified, very poorly sorted, breccia of intraformational **mudstone** clasts in a **sandy mudstone matrix** (greenish grey 5BG 5/1 to dark greenish grey 7.5GY 4/1). Stratification is locally contorted and disrupted, showing dips up to 30° and at 51.40 m the hinge of an isoclinal fold. Matrix-and clast-supported with 80 to 90% mudstone clasts and the remainder bedrock derived. Mudstone clasts are on average 10 to 15 mm long and range up to 25 mm. Thin bed of well-stratified, pebbly coarse to fine sandstone (dark greenish grey 7.5GY 4/1) from 48.73 to 48.98 m.

PROJECT : CIROS -1

SHEET NO. : 11

SCALE : 1:20



DESCRIPTION

— contact gradational -

Sub-unit 3.2

62.13 - 63.32 m

SANDSTONE AND DIAMICTITE

Hard to firm, unstratified to weakly and diffusely stratified (mm- to cm-bedded), **muddy fine to very fine sandstone** (grey 10Y 4 ′1) with partial intraformational brecciation.
Contorted bed (63.00 to 63.12 m) of weakly stratified **diamictite**(grey 10Y 4 ′1). Clasts comprise 5 to 10% of core and range up to 24 mm in length. Most are subangular to subrounded. About 70% are basement and 30% basic volcanic lithology.
Small crystals of pyrite disseminated through the core.

— contact sharp, possibly slumped -

Sub-unit 3.3

MUDSTONE BRECCIA

63.32 - 73.10 m

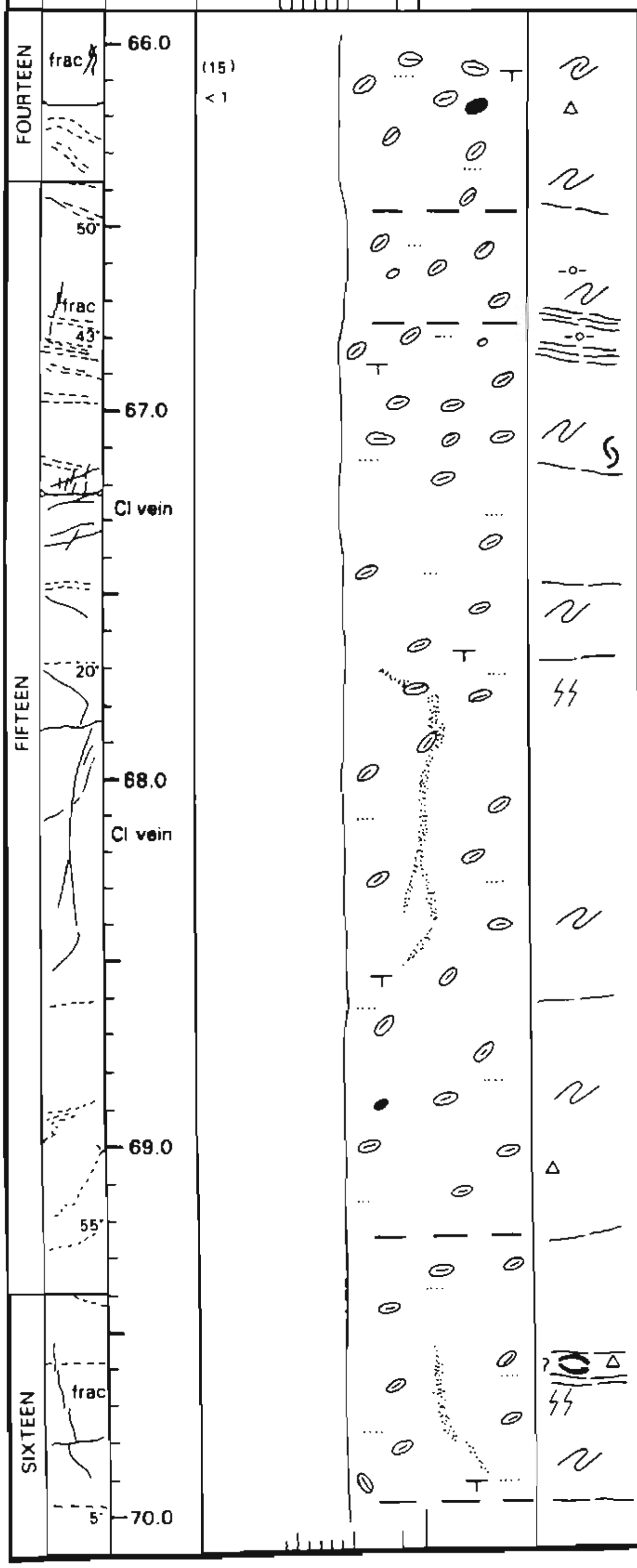
Hard, unstratified to weakly stratified (mm-bedded), very poorly sorted, **sandy mudstone breccia** (intraformational) (dark olive grey 5GY 4 ′1).
Stratification is locally distorted and disrupted into mudstone clasts up to 20 mm long that form 10 to 90% of the core (less towards base of sub-unit). Bedding dips as much as 55 .
Bedrock clasts, mostly granule size but a few more than 20 mm long, form < 1% of the core. Lithologies include dolerite, basement and basic volcanic rocks.
Veins of clay, some light grey and others black, occur below 65 m. Some run parallel with tectonic fractures, which are mostly steeply inclined. Thin vertical synsedimentary dykes of sandstone from 67.70 to 68.53 m and 69.55 to 69.90 m. Fault breccia at 72.10 m.

PROJECT : CIROS -1

SHEET NO. : 12

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 3.3

63.32 - 73.10 m

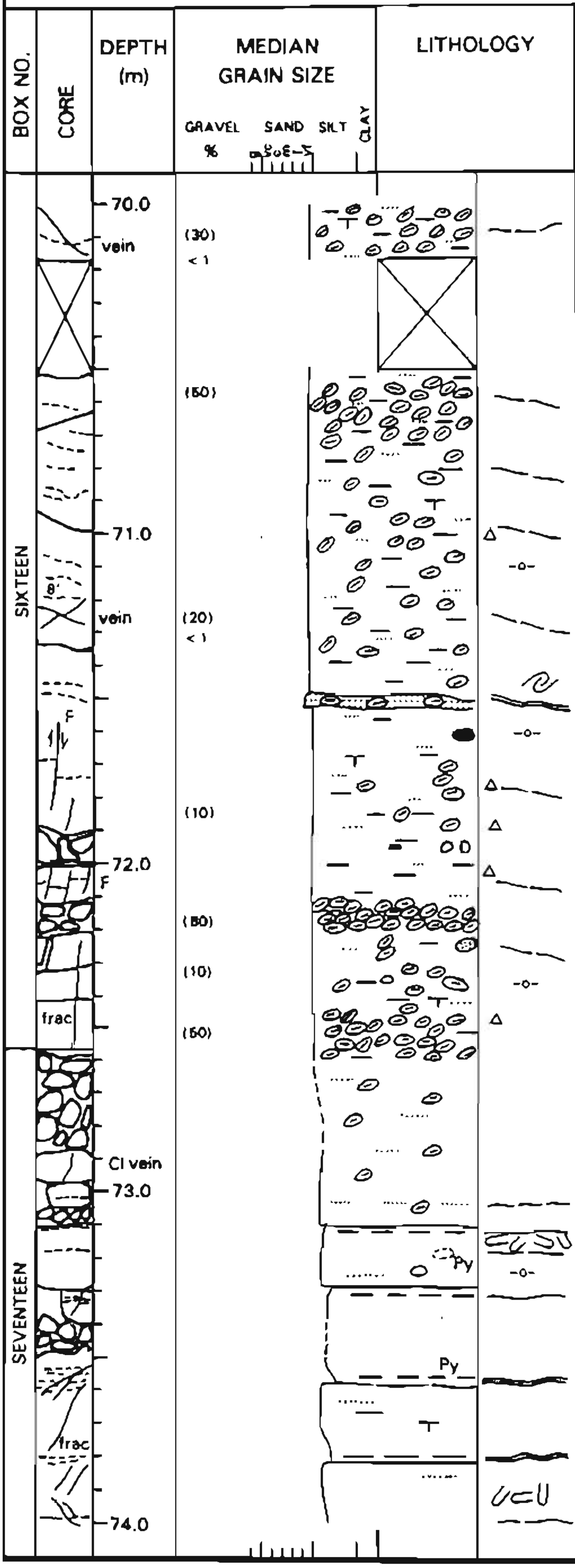
MUDSTONE BRECCIA

Hard, unstratified to weakly stratified (mm-bedded), very poorly sorted, **sandy mudstone breccia** (intraformational) (dark olive grey 5GY 4/1). Stratification is locally distorted and disrupted into mudstone clasts up to 20 mm long that form 10 to 90% of the core (less towards base of sub-unit). Bedding dips as much as 55°. Bedrock clasts, mostly granule size but a few more than 20 mm long, form < 1% of the core. Lithologies include dolerite, basement and basic volcanic rocks. Veins of clay, some light grey and others black, occur below 65 m. Some run parallel with tectonic fractures, which are mostly steeply inclined. Thin vertical synsedimentary dykes of sandstone from 67.70 to 68.53 m and 69.55 to 69.90 m. Fault breccia at 72.10 m.

PROJECT : CIROS-1

SHEET NO. : 13

SCALE : 1:20



DESCRIPTION

Sub-unit 3.3

63.32 - 73.10 m

MUDSTONE BRECCIA

Hard, unstratified to weakly stratified (mm-bedded), very poorly sorted, **sandy mudstone breccia** (intraformational) (dark olive grey 5GY 4/1). Stratification is locally distorted and disrupted into mudstone clasts up to 20 mm long that form 10 to 90% of the core (less towards base of sub-unit). Bedding dips as much as 55°. Bedrock clasts, mostly granule size but a few more than 20 mm long, form < 1% of the core. Lithologies include dolerite, basement and basic volcanic rocks. Veins of clay, some light grey and others black, occur below 65 m. Some run parallel with tectonic fractures, which are mostly steeply inclined. Thin vertical synsedimentary dykes of sandstone from 67.70 to 68.53 m and 69.55 to 69.90 m. Fault breccia at 72.10 m.

— contact sharp -

Sub-unit 3.4

73.10 - 75.60 m

MUDSTONE

Moderately hard to hard, well stratified (mm- to cm-bedded), **sandy mudstone** (dark greenish grey 10GY 4/1).

Stratification is near horizontal, discontinuous and wavy.

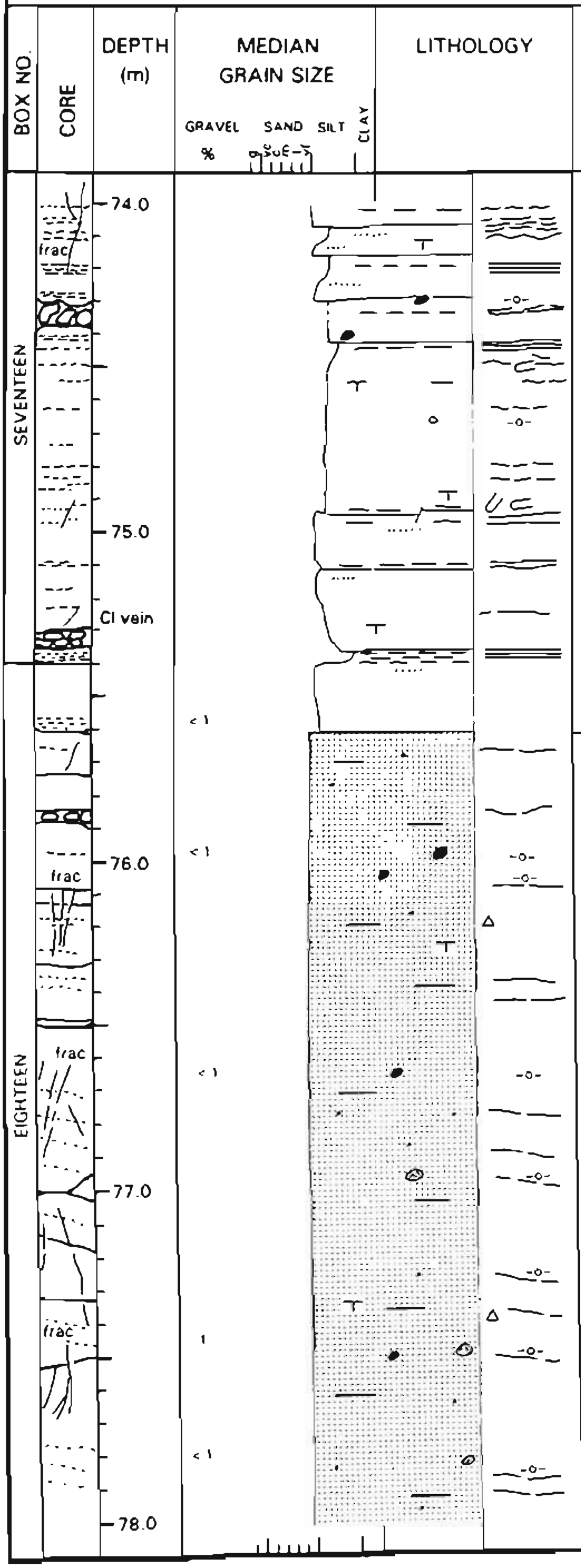
Rare lonestones up to 8 mm long of basement and basic volcanic composition.

Burrows 1 to 10 mm wide, both inclined and horizontal, associated with well stratified clayey beds (dark grey N3). Nodules of disseminated pyrite from 73.26 to 73.60 m.

PROJECT : CIROS -1

SHEET NO. : 14

SCALE : 1:20



DESCRIPTION

Sub-unit 3.4

73.10 - 75.60 m

MUDSTONE

Moderately hard to hard, well stratified (mm- to cm-bedded), **sandy mudstone** (dark greenish grey 10GY 4/1) Stratification is near horizontal, discontinuous and wavy Rare lonestones up to 8 mm long of basement and basic volcanic composition Burrows 1 to 10 mm wide, both inclined and horizontal, associated with well stratified clayey beds (dark grey N3). Nodules of disseminated pyrite from 73 26 to 73 60 m

— contact sharp -

Sub-unit 3 5

75 60 - 79 14 m

SANDSTONE

Hard, unstratified to weakly stratified (cm-bedded) poorly sorted **muddy very fine sandstone** (dark olive grey 2.5GY 4 1)

Occassional slight disruption of bedding resulting in lenticular pseudobedding

Stratification dips < 10

Bedrock clasts up to 8 mm form < 1% of core but are slightly more common towards the base Most are subangular to subrounded They include basement and basic volcanic lithologies

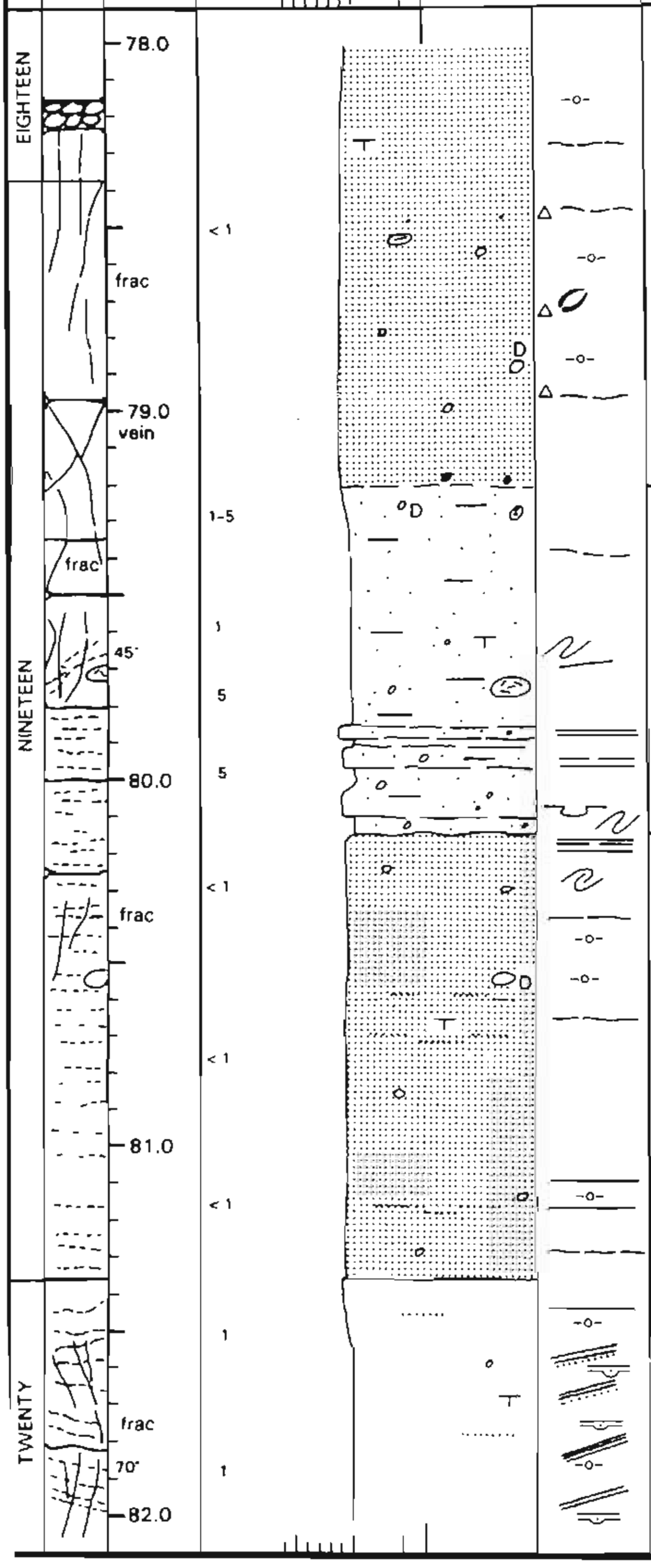
The sub-unit has numerous subvertical fractures and associated mouldic porosity

PROJECT : CIROS -1

SHEET NO. : 15

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



contact gradational over 20 cm -

Sub-unit 3.6

79.14 - 80.13 m

DIAMICTITE

Hard, weakly to well stratified, **slightly conglomeratic muddy very fine sandstone (diamictite)** (dark olive grey 2.5GY 4 1).
Bedding inclined (upto 45), diffuse, weak and on cm scale in upper part, well stratifed with matrix alternating between muddy and sandy and on mm scale in lower part.
Bedrock clasts form between 2 and 5% of core, and range up to 30 mm long. Most are subangular to subrounded. They include basement (50%), dolerite (30%) and basic volcanic (20%).
Numerous synsedimentary fractures, occasionally parallel to stratification but mainly steeply dipping. Contorted fine sandstone lamina 3 mm thick at 79.70 m.

contact sharp -

Sub-unit 3.7

80.13 to 81.46 m

SANDSTONE

Hard, mainly broken, weakly stratified (mm- to cm-bedded), poorly sorted, **muddy fine sandstone** (dark olive grey 2.5GY 4 1).
Stratification loaded and slightly disrupted near top
Clasts mainly rare scattered granules, but include a 30 mm long dolerite.
Dark **gritty sandstone** layer 15 mm thick at 81.15 m. Occasional thin, less muddy, pale grey brown sandy laminae towards base Intact core shows irregular subvertical synsedimentary fractures.

contact sharp -

Sub-unit 3.8

81.46 - 82.36 m

SANDSTONE TO MUDSTONE

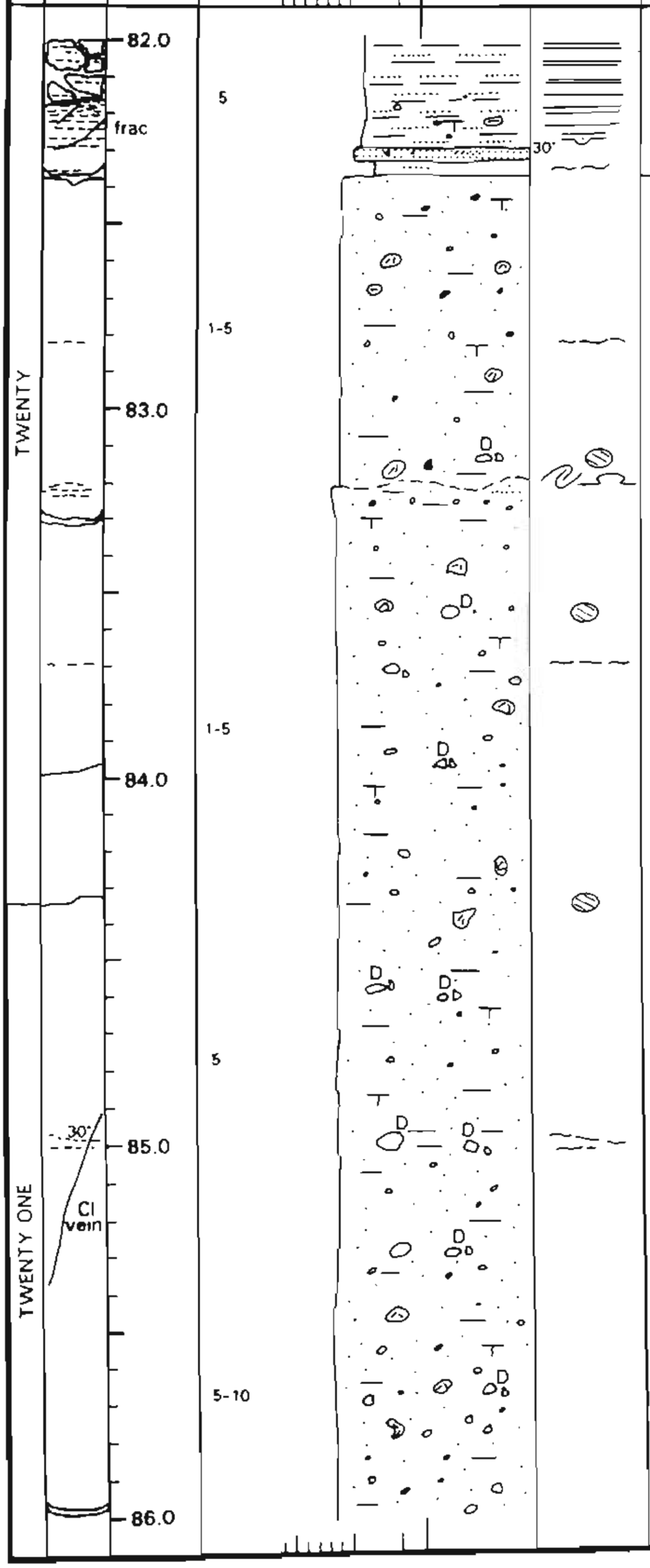
Hard, well-stratifed (mm- to cm-bedded), poorly sorted, **muddy very fine sandstone to sandy mudstone** (dark greenish grey 10GY 4 1).
Stratification mainly regular but disrupted by synsedimentary faulting. Dips vary from 0 to 12 .
Small lonestones (up to 5 mm) and several dropstones form up to 10% of the core.
About 20% of the interval has rhythmic bedding, with graded alternations of **medium to fine sandstone** and **muddy fine sandstone.** Occasional black laminae. At 81.15 m a low angle fault displaces rhythmites downwards in direction of bedding dip

PROJECT : CIROS -1

SHEET NO. : 16

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



— contact sharp -

Sub-unit 4.1

82.36 - 88.91 m

DIAMICTITE

Moderatly hard, unstratified to weakly stratified (mm- to cm-bedded), **slightly conglomeratic muddy fine to very fine sandstone (diamictite)** (dark olive grey 2.5GY 4 1)

Stratification is wispy in places, inclined (up to 30), discontinuous and occasionally contorted and disrupted

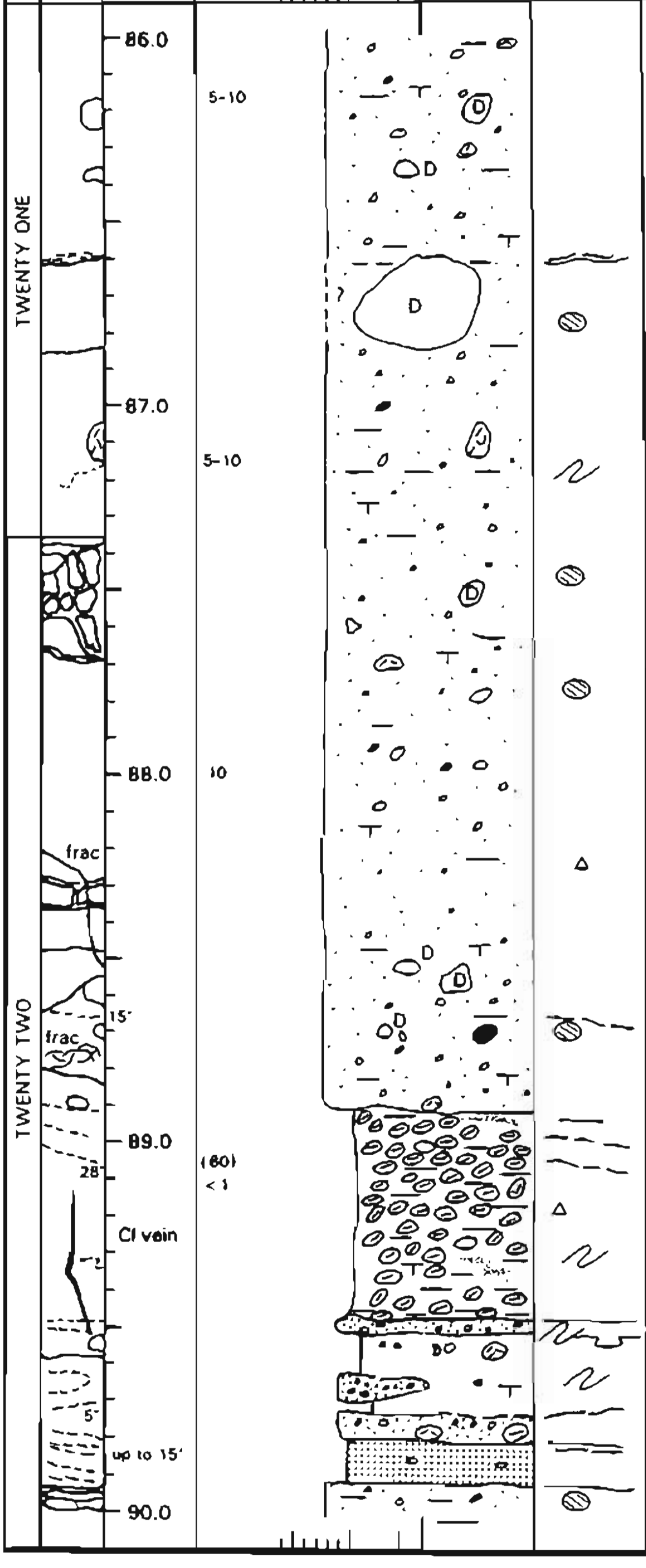
Clasts form between 1 and 10% of the core (increasing towards base of sub-unit). They range from subangular to rounded and average subrounded. About a quarter are striated and a similar proportion faceted. There are equal proportions of basement and dolerite and <2% volcanic rocks.

PROJECT : CIROS -1

SHEET NO. : 17

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



— contact sharp -

Sub-unit 4.2

88.91 - 89.47 m

MUDSTONE BRECCIA

Hard, weakly stratified, poorly sorted breccia of intraformational mudstone clasts in a **sandy mudstone matrix** (dark greenish grey 10GY 4 1)
Stratification is discontinuous and contorted, with dips up to 28
Breccia is clast-supported with 60% mudstone clasts on average 6 to 8 mm, but up to 12 mm long, and subangular to subrounded. Bedrock clasts up to 5 mm form < 1% of the core.
Near vertical clay-filled vein runs from 89.15 to 89.51 m.

— contact sharp, deformed -

Sub-unit 4.3

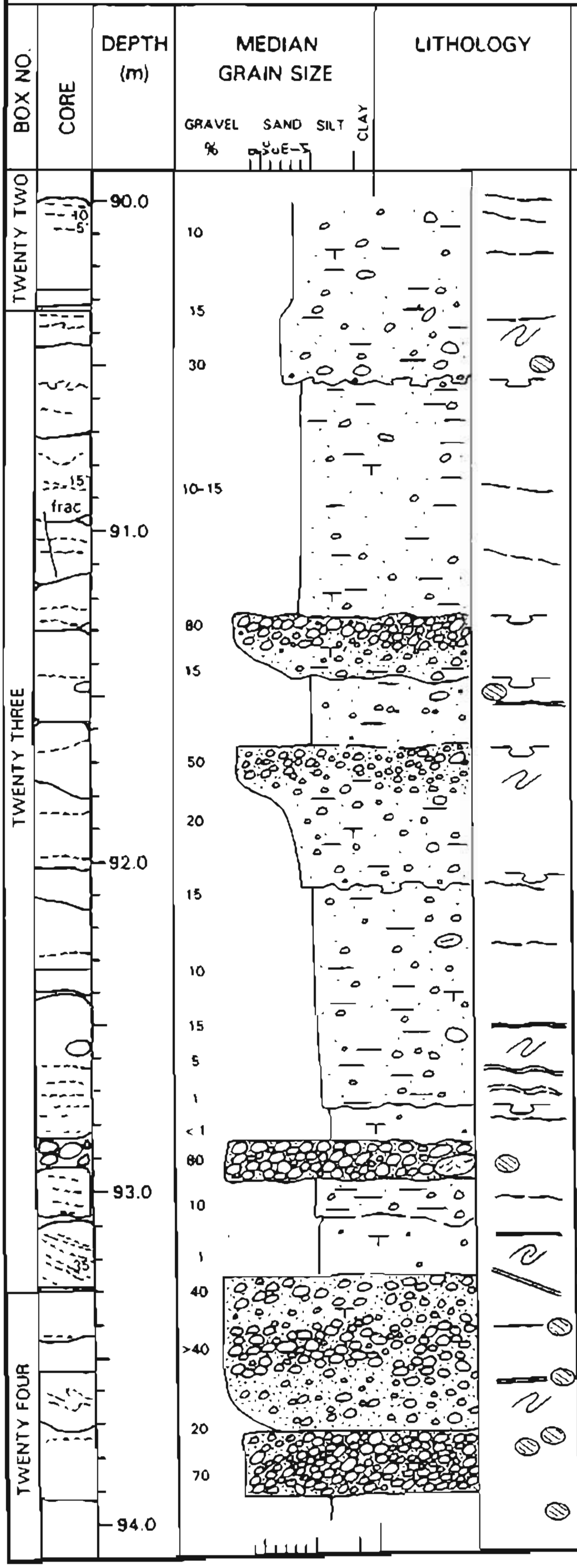
89.47 - 93.26 m

DIAMICTITE, CONGLOMERATE AND MUDSTONE

PROJECT : CIROS -1

SHEET NO. : 18

SCALE : 1:20



DESCRIPTION

– contact rapid transition –

Sub-unit 4.4

93.26-93.73 m

CONGLOMERATE

Firm, poorly sorted, weakly stratified, **pebble conglomerate**(matrix – dark olive grey 2 5GY 4 1)

Clasts form 40 to 80% of core but grade down to 20% at base Most are 4 to 20 mm, but range up to 45 mm long. They range from angular to rounded, averaging subrounded and over half are striated About 70% are basement, 25% dolerite and 5% basic volcanic.

Synsedimentary flow fold with horizontal fold axis near base

– contact sharp –

Sub-unit 4.5

93.73 - 93.91 m

CONGLOMERATE

Hard, unstratified, moderately well sorted, **sandy pebble conglomerate**.

Clasts form 70% of the core and range from 4 to 20 mm long They are subangular to subrounded Some are striated Most are basement and dolerite.

Matrix is **medium to fine sandstone** (dark olive grey 2 5GY 4 1)

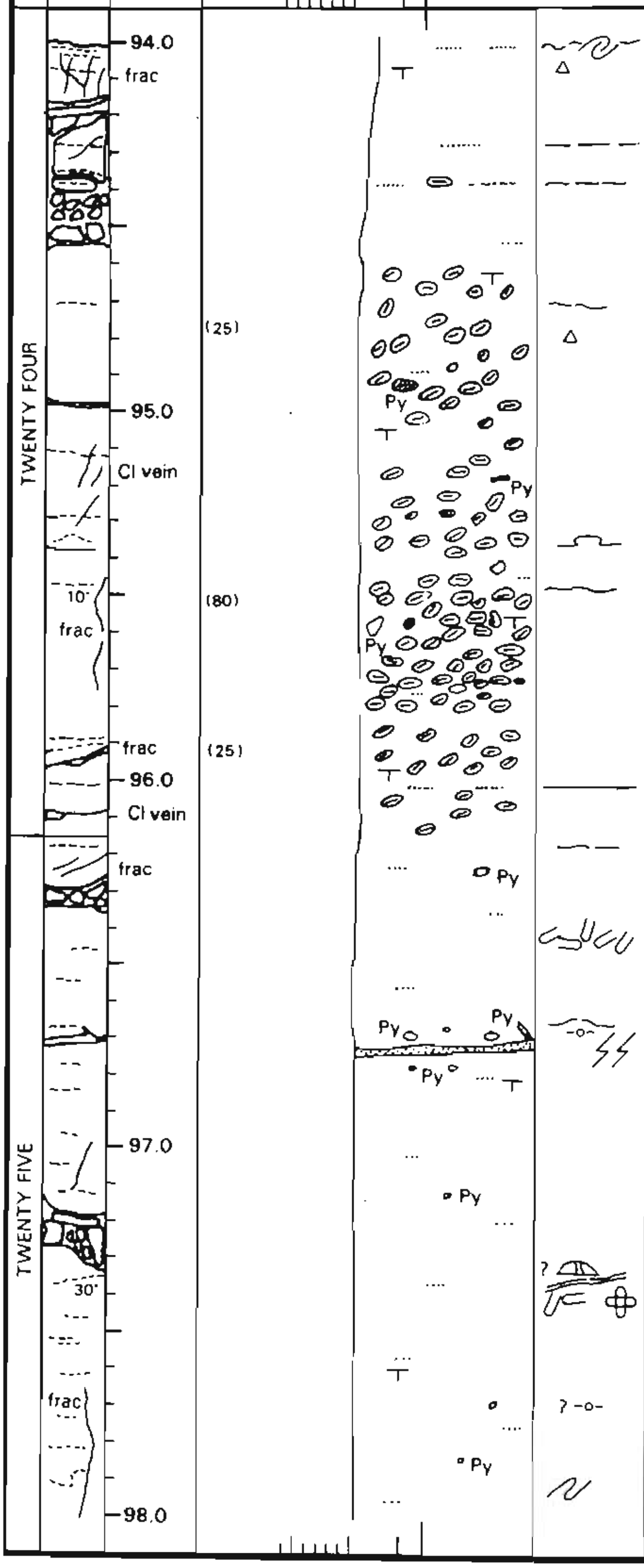
– contact sharp –

PROJECT : CIROS -1

SHEET NO. : 19

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 5.1

93.91 – 121.16 m

MUDSTONE

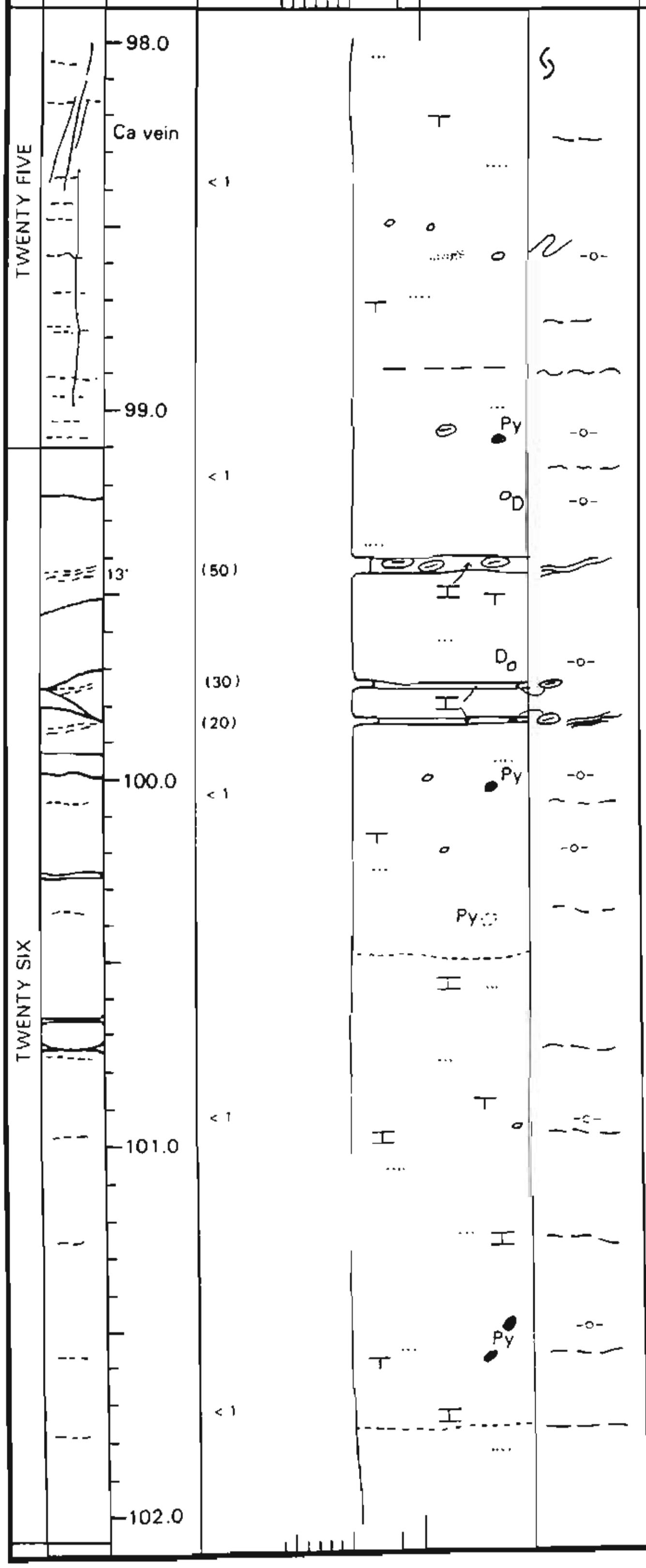
Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, **sandy mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1).
Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40°.
A few scattered lonestones mostly 5 to 10 mm long, but one granite at 105.50 m is > 160 mm. Lithologies are basement, dolerite and volcanic.
The sub-unit is brecciated with sandy mudstone matrix between 94.60 and 96.10 m. Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7.5Y 4/1). No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation. Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide.
Three pale grey flake breccia beds of strongly calcareous, **mudstone** at 99.40 to 99.44, 99.73 to 99.74 and 99.84 to 99.85 m. Dykes of poorly sorted, **fine sandstone** at 96.10, 96.72, 109.10 and 111.60 m. Possible echinoids at 97.35, 111.25 and 111.53 m. Scattered nodular and disseminated pyrite crystals.

PROJECT : CIROS-1

SHEET NO. : 20

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 5.1

93.91 - 121.16 m

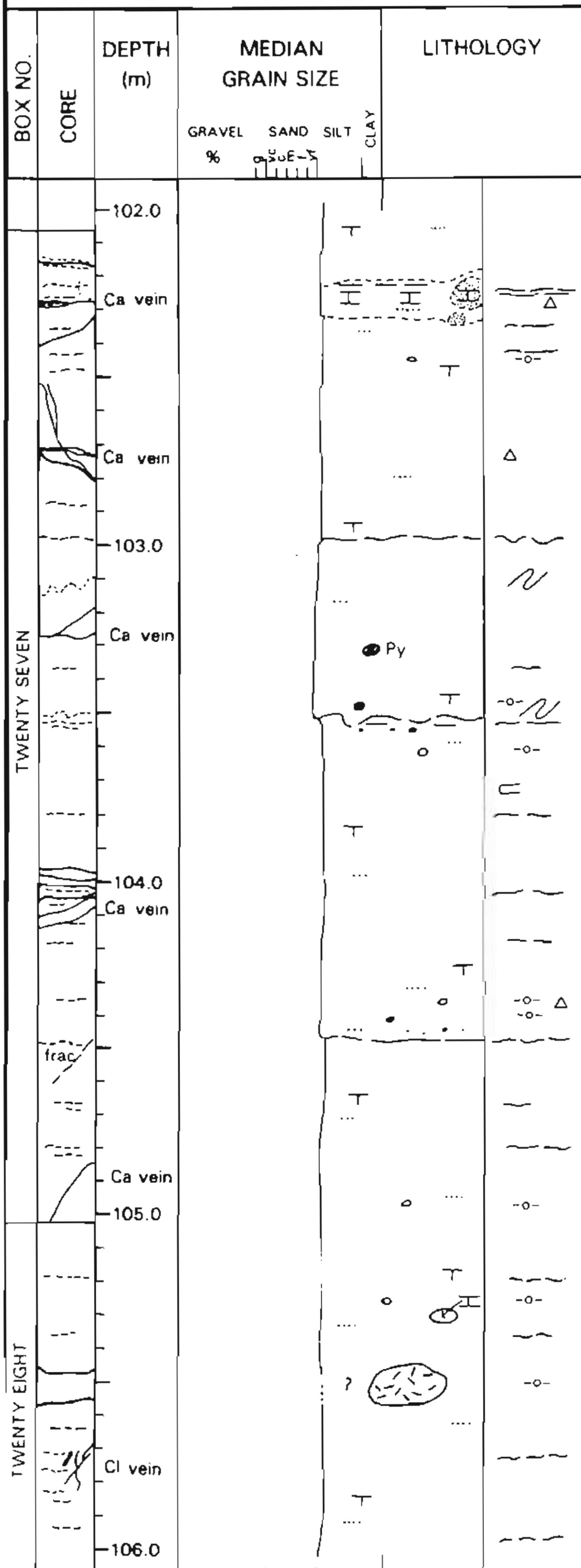
MUDSTONE

Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, **sandy mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1) Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40°. A few scattered lonestones mostly 5 to 10 mm long, but one granite at 105.50 m is > 160 mm. Lithologies are basement, dolerite and volcanic. The sub-unit is brecciated with sandy mudstone matrix between 94 60 and 96 10 m. Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7.5Y 4.1) No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide. Three pale grey flake breccia beds of strongly calcareous**mudstone** at 99.40 to 99.44, 99.73 to 99.74 and 99.84 to 99 85 m Dykes of poorly sorted **fine sandstone** at 96.10, 96.72, 109 10 and 111.60 m. Possible echinoids at 97.35, 111.25 and 111.53 m. Scattered nodular and disseminated pyrite crystals

PROJECT : CIROS -1

SHEET NO. : 21

SCALE : 1:20



DESCRIPTION

Sub-unit 5.1

93.91 - 121.16 m

MUDSTONE

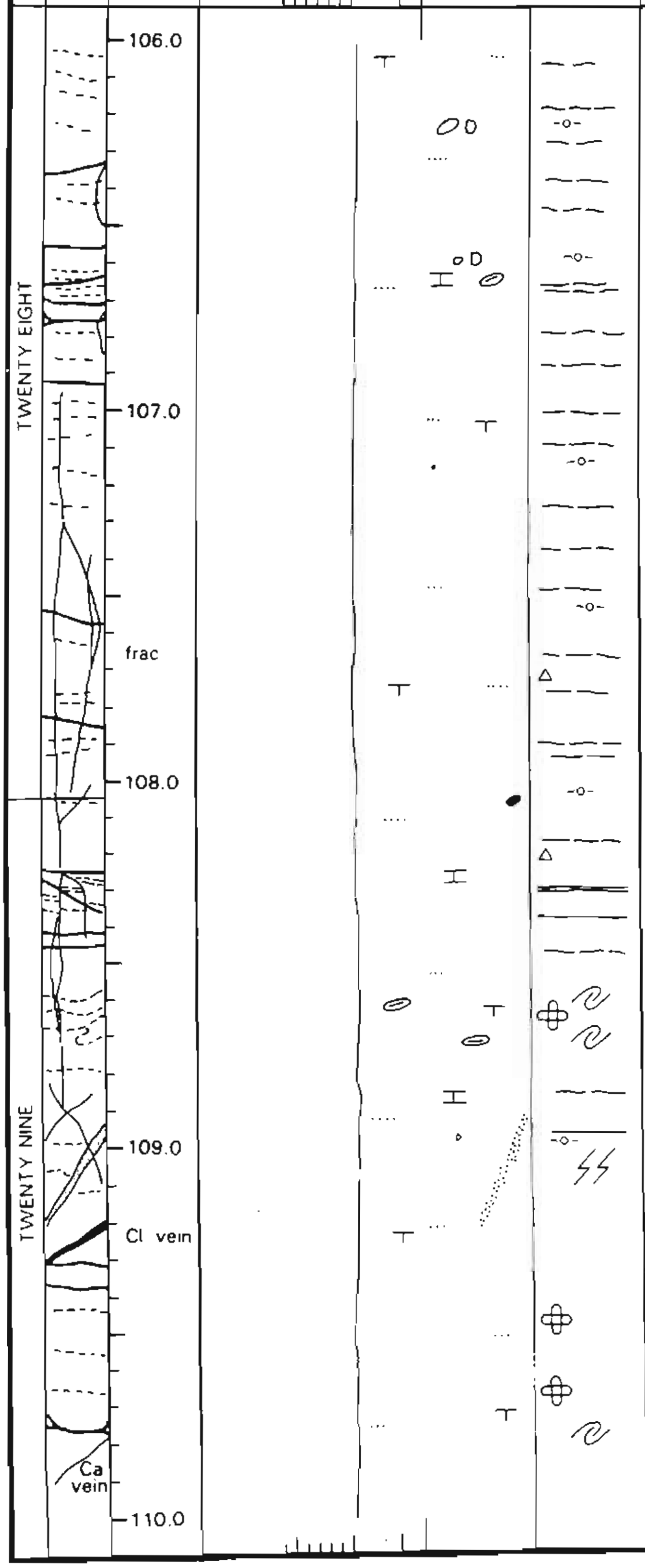
Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, **sandy mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1). Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40°. A few scattered lonestones mostly 5 to 10 mm long, but one granite at 105.50 m is > 160 mm. Lithologies are basement, dolerite and volcanic. The sub-unit is brecciated with sandy mudstone matrix between 94.60 and 96.10 m. Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7.5Y 4.1). No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation. Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide. Three pale grey flake breccia beds of strongly calcareous**mudstone** at 99.40 to 99.44, 99.73 to 99.74 and 99.84 to 99.85 m. Dykes of poorly sorted **fine sandstone** at 96.10, 96.72, 109.10 and 111.60 m. Possible echinoids at 97.35, 111.25 and 111.53 m. Scattered nodular and disseminated pyrite crystals.

PROJECT : CIROS -1

SHEET NO. : 22

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 5.1

93.91 - 121.16 m

MUDSTONE

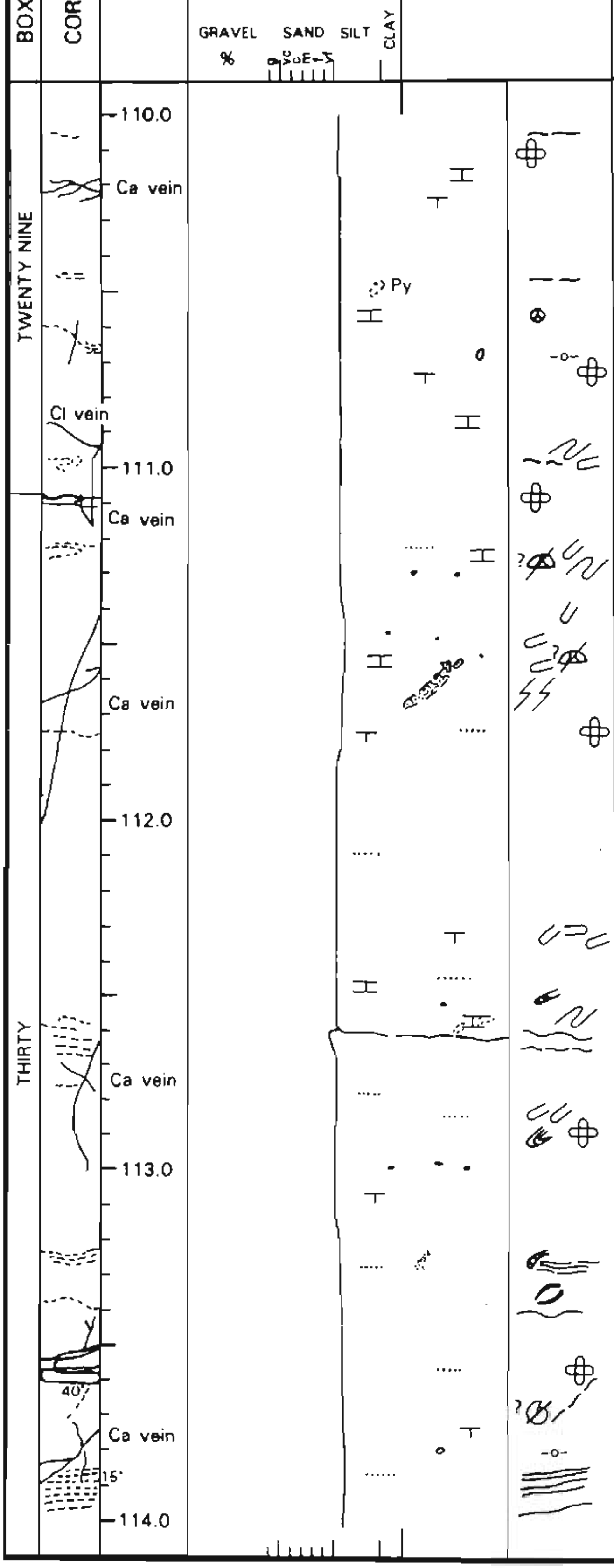
Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, sandy **mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1). Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40°. A few scattered lonestones mostly 5 to 10 mm long, but one granite at 105.50 m is > 160 mm. Lithologies are basement, dolerite and volcanic. The sub-unit is brecciated with sandy mudstone matrix between 94.60 and 96.10 m. Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7.5Y 4.1). No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation. Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide. Three pale grey flake breccia beds of strongly calcareous**mudstone** at 99.40 to 99.44, 99.73 to 99.74 and 99.84 to 99.85 m. Dykes of poorly sorted fine **sandstone** at 96.10, 96.72, 109.10 and 111.60 m. Possible echinoids at 97.35, 111.25 and 111.53 m. Scattered nodular and disseminated pyrite crystals.

PROJECT : CIROS -1

SHEET NO. : 23

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 5.1

93.91 - 121.16 m

MUDSTONE

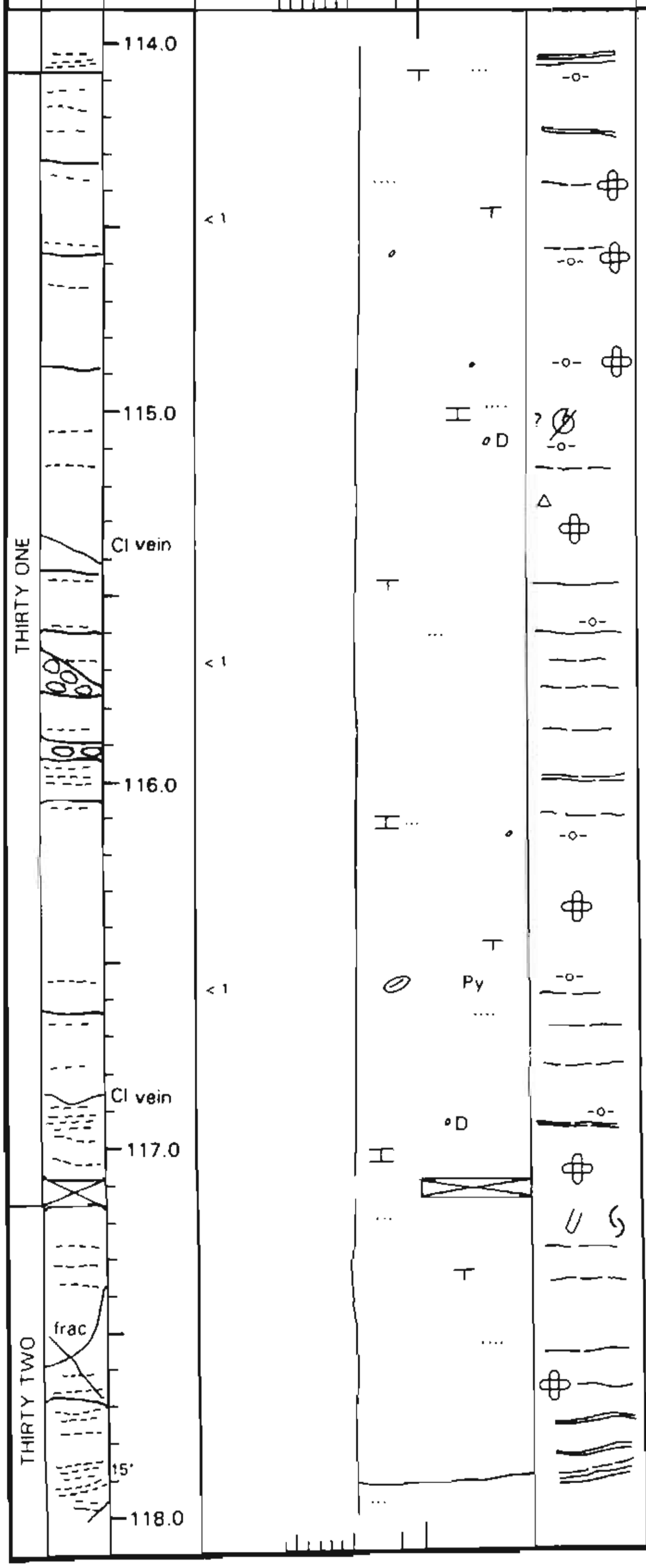
Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, **sandy mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1). Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40°. A few scattered lonestones mostly 5 to 10 mm long, but one granite at 105.50 m is > 160 mm. Lithologies are basement, dolerite and volcanic. The sub-unit is brecciated with sandy mudstone matrix between 94.60 and 96.10 m. Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7.5Y 4 1) No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation. Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide. Three pale grey flake breccia beds of strongly calcareous**mudstone** at 99.40 to 99.44, 99.73 to 99.74 and 99.84 to 99.85 m. Dykes of poorly sorted **fine sandstone** at 96.10, 96.72, 109.10 and 111.60 m. Possible echinoids at 97.35, 111.25 and 111.53 m. Scattered nodular and disseminated pyrite crystals.

PROJECT : CIROS -1

SHEET NO. : 24

SCALE : 1:20

BOX NO. | CORE | DEPTH (m) | MEDIAN GRAIN SIZE (GRAVEL % · SAND · SILT · CLAY) | LITHOLOGY | DESCRIPTION



Sub-unit 5.1

93.91 - 121.16 m

MUDSTONE

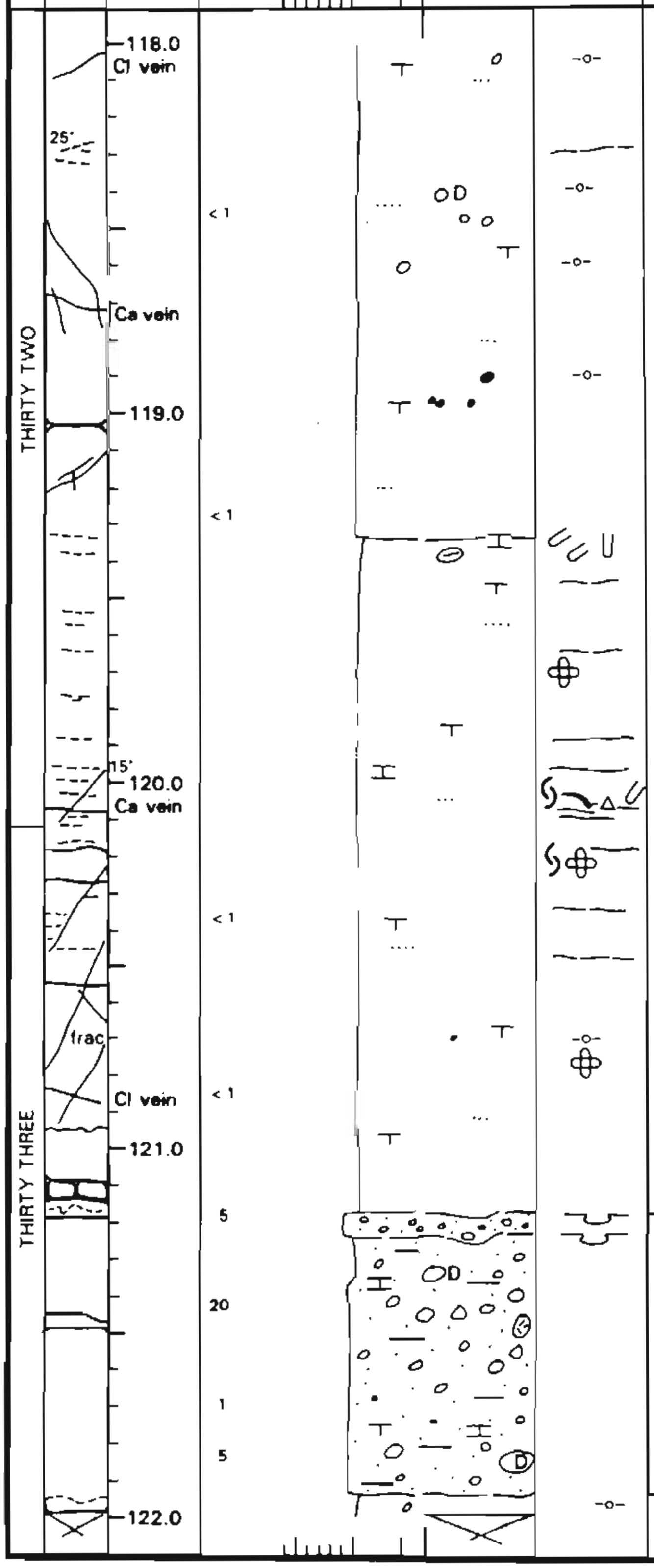
Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, **sandy mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1) Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40° A few scattered lonestones mostly 5 to 10 mm long, but one granite at 105 50 m is > 160 mm Lithologies are basement, dolerite and volcanic The sub-unit is brecciated with sandy mudstone matrix between 94.60 and 96 10 m Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7 5Y 4 1) No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation. Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide Three pale grey flake breccia beds of strongly calcareous**mudstone** at 99 40 to 99.44, 99.73 to 99 74 and 99.84 to 99 85 m. Dykes of poorly sorted **fine sandstone** at 96.10, 96.72, 109 10 and 111.60 m. Possible echinoids at 97.35, 111 25 and 111.53 m Scattered nodular and disseminated pyrite crystals

PROJECT : CIROS-1

SHEET NO. : 25

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 5.1

93.91 - 121.16 m

MUDSTONE

Hard, weakly stratified (mm- and cm-bedded), moderately well sorted, **sandy mudstone** (greenish grey to olive grey 5BG 4/1 to 5GY 4/1). Stratification is wispy mainly horizontal, but is locally contorted and dips up to 40°. A **few scattered** lonestones mostly 5 to 10 mm long, but one granite at 105.50 m is > 160 mm. Lithologies are basement, dolerite and volcanic. The sub-unit is brecciated with sandy mudstone matrix between 94.60 and 96.10 m. Below this the sub-unit becomes slightly calcareous and changes colour to dark olive grey to grey (5GY 4/1 to 7.5Y 4.1). No brecciation can be seen but the pervasive, wispy stratification and mottling, along with definite burrowing locally, indicate moderately strong bioturbation. Most burrows are 1 to 2 mm wide and range from horizontal to vertical. A few burrows are 10 mm wide. Three pale grey flake breccia beds of strongly **calcareousmudstone** at 99.40 to 99.44, 99.73 to 99.74 and 99.84 to 99.85 m. Dykes of poorly sorted **fine sandstone** at 96.10, 96.72, 109.10 and 111.60 m. Possible echinoids at 97.35, 111.25 and 111.53 m. Scattered nodular and disseminated pyrite crystals.

— contact gradational over a few mm and slightly deformed -

Sub-unit 6.1

121.16 - 121.93 m

DIAMICTITE

Hard, slightly calcareous, weakly stratified, **conglomeratic muddy fine sandstone (diamictite)** (grey 7.5Y 4/1).

Stratification is on a cm scale, and is diffuse and irregular. Bounding contacts show soft sediment deformation.

Clast content is variable, forming 1 to 20% of the core. They range up to 30 mm long. Most are subangular to subrounded. About 60% are basement, 30% are dolerite and <10% are volcanic.

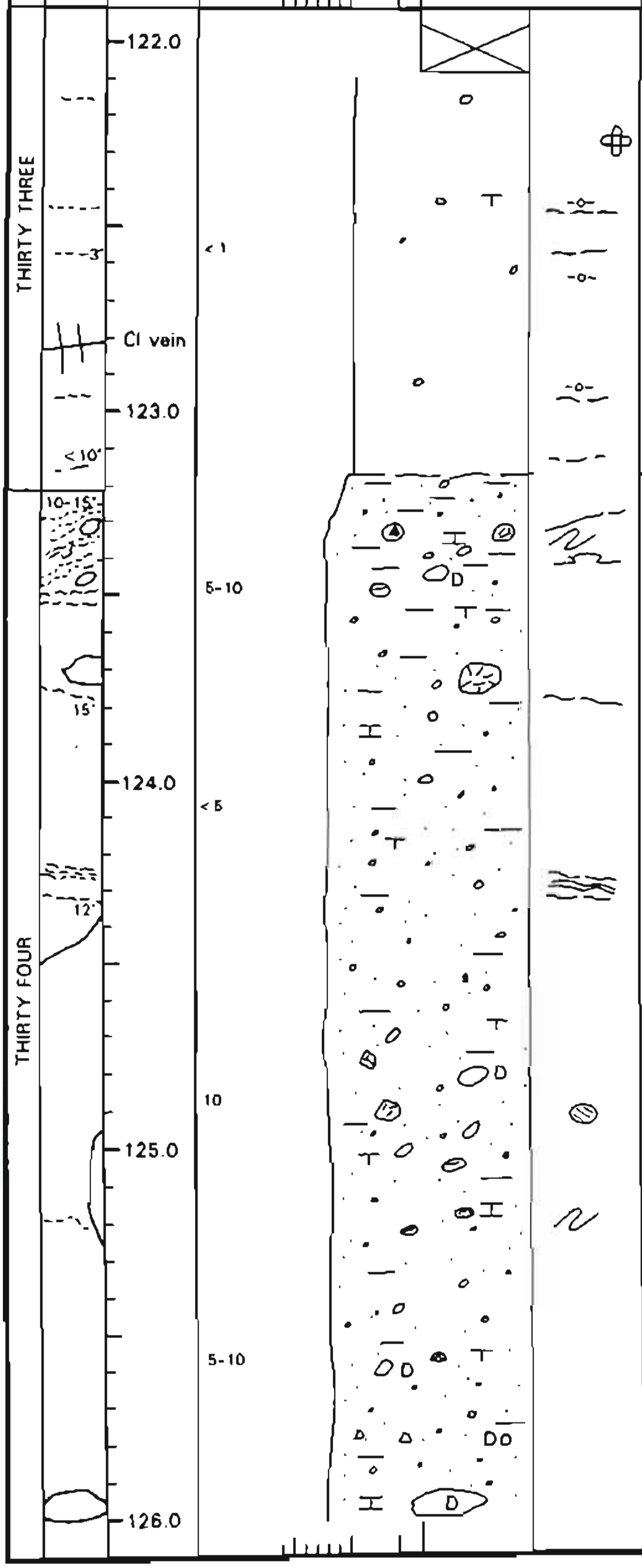
— contact gradational over a few mm and slightly deformed -

PROJECT : CIROS -1

SHEET NO. : 26

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 6.2

121.93 - 123.15 m

MUDSTONE

Hard, weakly stratified (mm-bedded), moderately well sorted to poorly sorted, **sandy mudstone** (grey 7.5Y 4 1).
Stratification is wispy and irregular, suggestive of bioturbation. It is mainly horizontal but locally dips at up to 10.
A few subangular bedrock clasts from 5 to 11 mm long.

— contact gradational over 2 to 5 cm -

Sub-unit 6.3

123.15 - 126.36 m

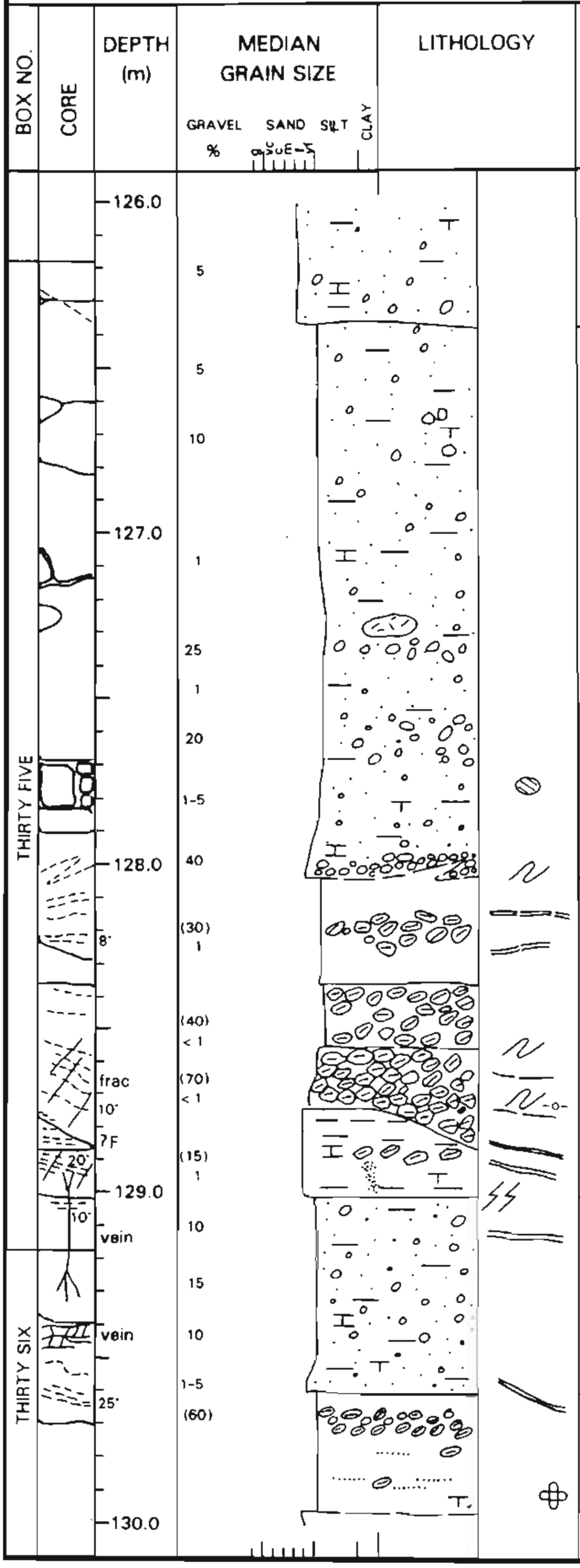
DIAMICTITE

Hard, slightly calcareous, unstratified to weakly stratified (mm- to cm-bedded), **conglomeratic muddy fine sandstone** (grey 10Y 4 1).
Stratification includes horizontal and inclined (10–15) bedding, contorted in places.
Clasts form 5 to 10% of the core, averaging 10 to 15 mm, and reaching 60 mm in length. Clasts are mostly subrounded, one is striated. They consist of equal proportions of basement and dolerite. Occasional mudstone and sandstone intraclasts.

PROJECT : CIROS -1

SHEET NO. : 27

SCALE : 1:20



DESCRIPTION

— contact sharp and inclined -

Sub-unit 6.4

126.36 - 128.03 m

DIAMICTITE

Hard, calcareous, unstratified, **conglomeratic sandy mudstone (diamictite)** (grey 10Y 4 1).
Clasts in core vary from 1 to 40% with diffuse local concentrations especially beneath a large gneiss clast (127.21 to 127.30 m) and at base. One clast is striated. Most are 10 to 20 mm long and subangular to subrounded. About 65% are basement, 30% dolerite and 5% volcanic.

— contact sharp and contorted -

Sub-unit 6.5

128.03 - 129.01 m

MUDSTONE BRECCIA

Hard, slightly calcareous, well stratified, poorly sorted, **sandy mudstone** and **muddy sandstone** (grey 10Y 4 1).
Partially brecciated with up to 70% angular mudstone clasts.
Rare (<1%) granules and pebbles.
Synsedimentary faulting, thrusting, inclined and contorted bedding, and muddy very fine sandstone dyke at base.

— contact sharp -

Sub-unit 6.6

129.01 - 129.60 m

DIAMICTITE

Hard, calcareous, unstratified to well stratified, **conglomeratic sandy mudstone (diamictite)** (grey 10Y 4 1)
Well stratified at top, unstratified in middle and weakly stratified at base.
Clasts form 2 to 15% of the core, averaging 10 mm, but ranging up to 20 mm long. Most are angular to subangular. About 70% are dolerite, 25% basement and 5% volcanic. Bed includes a few mudstone intraclasts up to 30 mm.
Numerous veins associated with partial brecciation at 129.45 m.

— contact sharp and dipping at 25 -

Sub-unit 6.7

129.60 - 129.96 m

MUDSTONE AND CONGLOMERATE

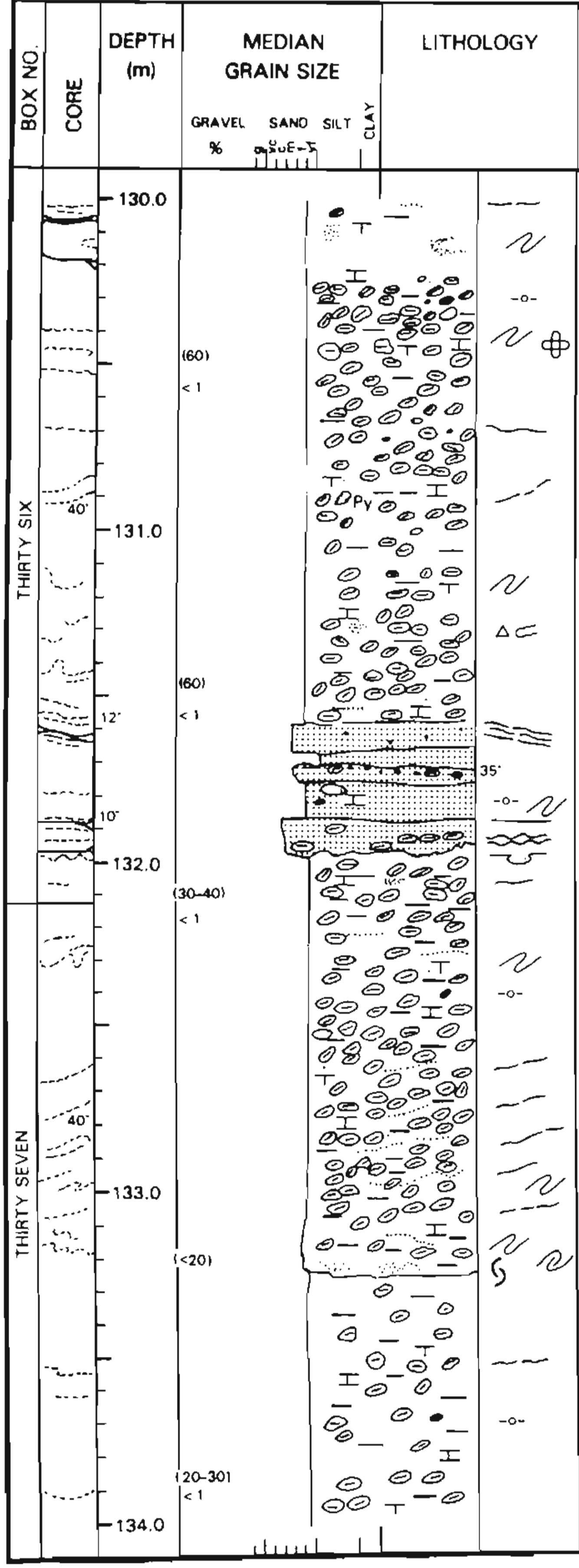
Hard, weakly stratified, very poorly sorted, **sandy mudstone** and **muddy very fine sandstone** (grey 7.5Y 4 1).
Mudstone beds partially brecciated and surrounded by sandstone
Conglomerate from 129.62 to 129.70 m has about 60% mudstone intraclasts, 20% basement and 20% volcanic.
Bed contacts are gradational. Possible bioturbation structures at base

— contact gradational -

PROJECT : CIROS -1

SHEET NO. : 28

SCALE : 1:20



DESCRIPTION

Sub-unit 7.1

129.96 - 136.01 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratified, poorly sorted, intraformational **mudstone clast breccia** with a sandy mudstone matrix (grey 10Y 4/1).

Stratification is generally discontinuous and variable in dip up to 40°.

Mudstone clasts average 8 to 12 mm long. Bedrock clasts form <1% of the core, and are either basement or dolerite.

The sub-unit includes thin beds of **fine to very fine sandstone** (olive black N 3/1) that are contorted and occasionally disrupted. Sandstone dykes occur irregularly throughout the sub-unit.

Two beds of **muddy medium to very fine sandstone** (dark greenish grey 7.5GY 4/1 to 3/1) from 129.96 to 130.25 m and 131.58 to 131.97 m with irregular and contorted bedding.

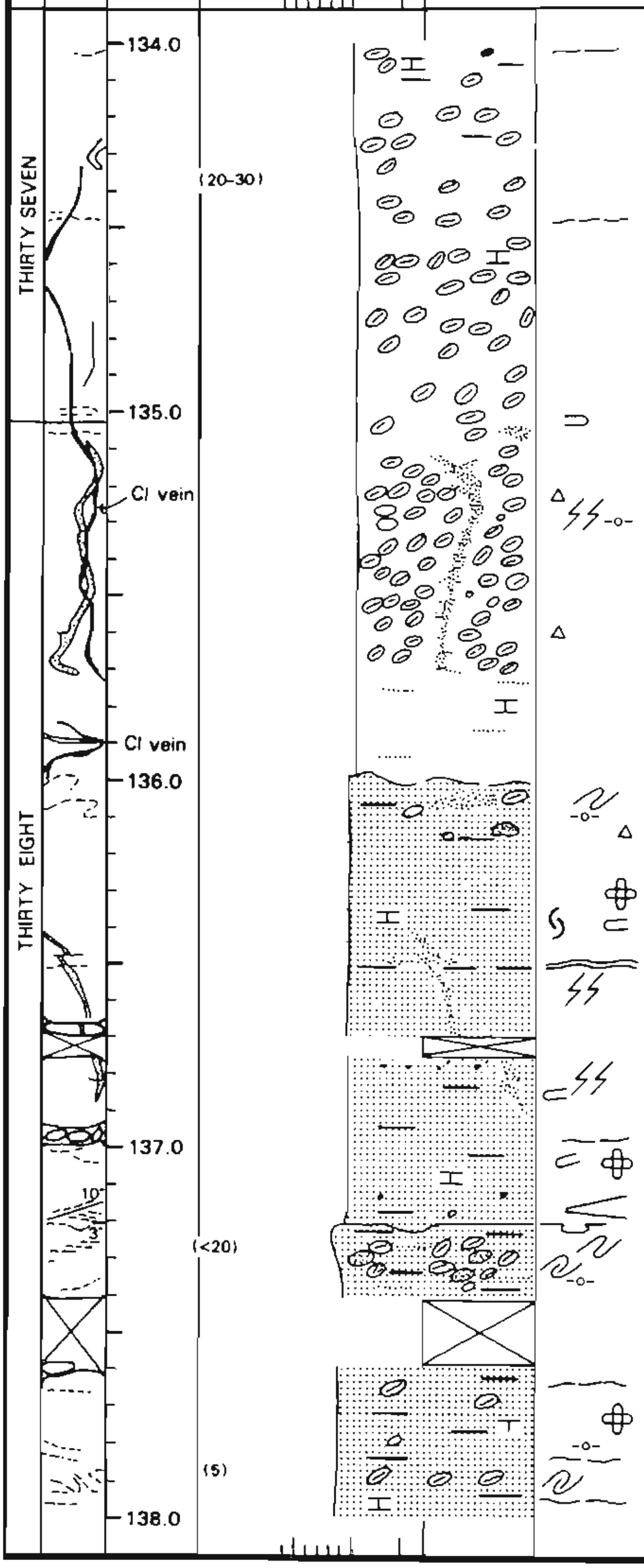
A few burrows and bioturbation structures.

PROJECT : CIROS -1

SHEET NO. : 29

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 7.1

129.96 - 136.01 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratified, poorly sorted, intraformational **mudstone clast breccia** with a sandy mudstone matrix (grey 10Y 4/1). Stratification is generally discontinuous and variable in dip up to 40°. Mudstone clasts average 8 to 12 mm long. Bedrock clasts form < 1% of the core, and are either basement or dolerite. The sub-unit includes thin beds of **fine to very fine sandstone** (olive black N 3/1) that are contorted and occasionally disrupted. Sandstone dykes occur irregularly throughout the sub-unit. Two beds of **muddy medium to very fine sandstone** (dark greenish grey 7.5GY 4/1 to 3/1) from 129.96 to 130.25 m and 131.58 to 131.97 m with irregular and contorted bedding. A few burrows and bioturbation structures.

— contact gradational over 20 cm -

Sub-unit 8.1

136.01 - 152.94 m

SANDSTONE

Hard to moderately hard (but soft in a few places), calcareous, unstratified to weakly stratified and well stratified (mm-bedded), moderately well sorted, **fine to very fine sandstone** (dark olive grey 5GY 3/1 to 4/1).

Muddy near top of sub-unit. Patches of greater carbonate cementation are slightly lighter (olive grey 2.5GY 5/1).

Burrowing commonly disrupts stratification, and in many places destroys it completely. Load features, dish structures and contorted stratification (with recumbent isoclinal folds) at several levels.

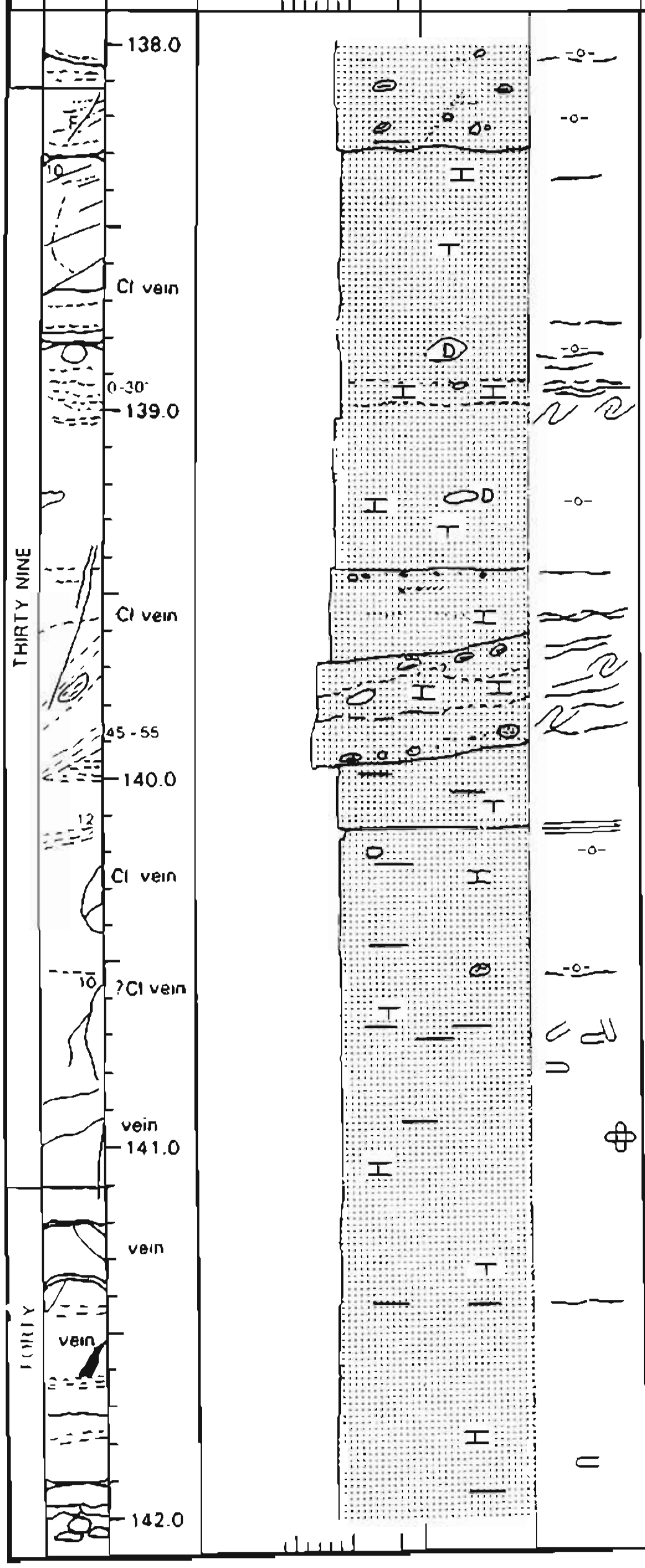
Clasts are dispersed throughout but they form much less than 1% of the core and most are <5 mm long. The larger ones range from 10 to 55 mm and are angular to rounded. Dominant lithologies are dolerite and basement. Other features include synsedimentary faults and sandstone dykes.

PROJECT : CIROS -1

SHEET NO. : 30

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 8.1

136.01 - 152.94 m

SANDSTONE

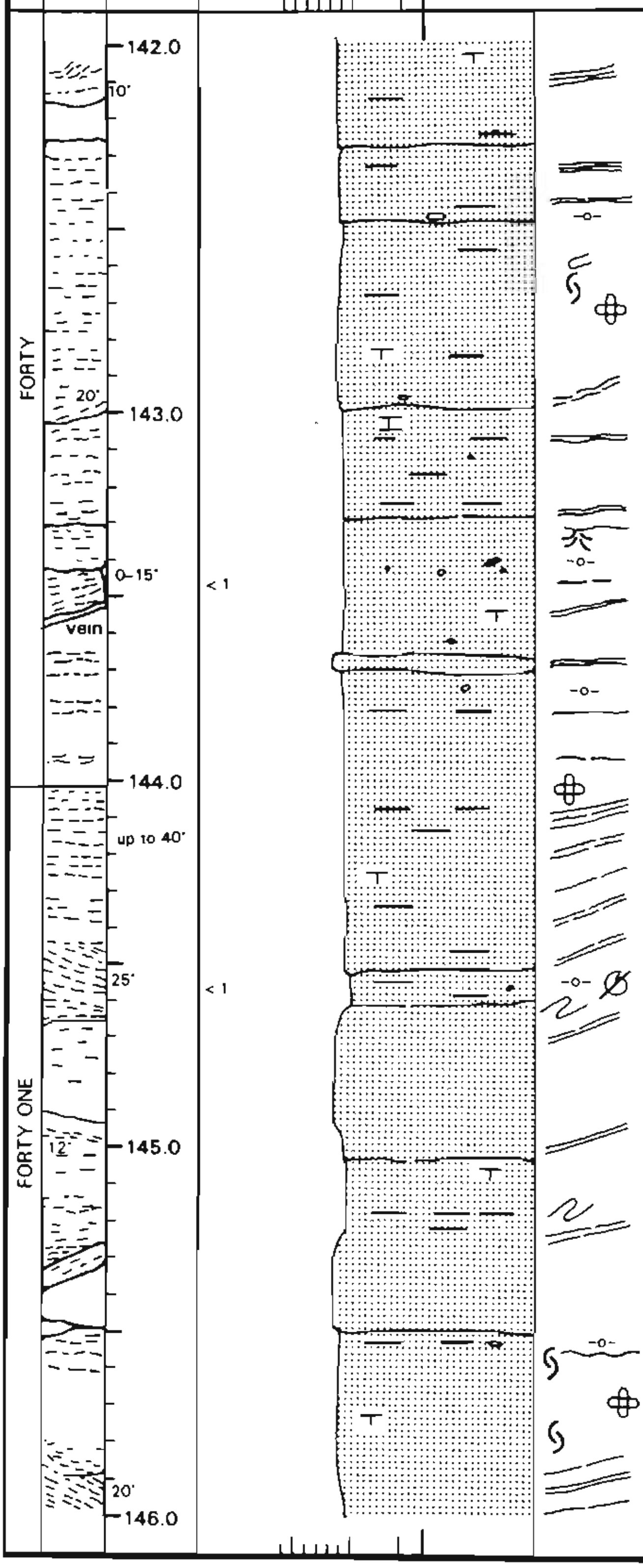
Hard to moderately hard (but soft in a few places), calcareous, unstratified to weakly stratified and well stratified (mm-bedded), moderately well sorted, **fine to very fine sandstone** (dark olive grey 5GY 3/1 to 4/1) Muddy near top of sub-unit. Patches of greater carbonate cementation are slightly lighter (olive grey 2 5GY 5/1) Burrowing commonly disrupts stratification, and in many places destroys it completely Load features, dish structures and contorted stratification (with recumbent isoclinal folds) at several levels. Clasts are dispersed throughout but they form much less than 1% of the core and most are < 5 mm long The larger ones range from 10 to 55 mm and are angular to rounded Dominant lithologies are dolerite and basement Other features include synsedimentary faults and sandstone dykes

PROJECT : CIROS -1

SHEET NO. : 31

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub–unit 8.1

136.01 - 152.94 m

SANDSTONE

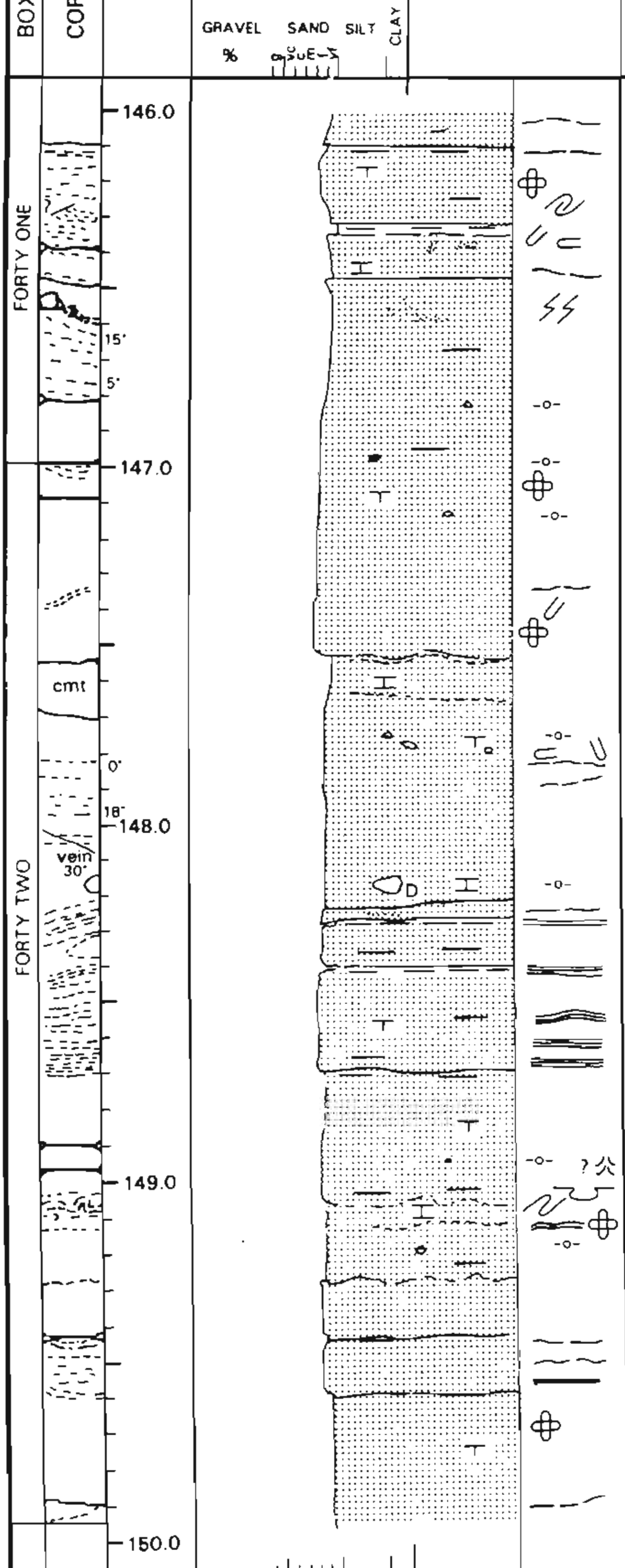
Hard to moderately hard (but soft in a few places), calcareous, unstratified to weakly stratified and well stratified (mm–bedded), moderately well sorted, **fine to very fine sandstone** (dark olive grey 5GY 3/1 to 4/1). Muddy near top of sub–unit. Patches of greater carbonate cementation are slightly lighter (olive grey 2.5GY 5/1). Burrowing commonly disrupts stratification, and in many places destroys it completely. Load features, dish structures and contorted stratification (with recumbent isoclinal folds) at several levels. Clasts are dispersed throughout but they form much less than 1% of the core and most are < 5 mm long. The larger ones range from 10 to 55 mm and are angular to rounded. Dominant lithologies are dolerite and basement. Other features include synsedimentary faults and sandstone dykes.

PROJECT : CIROS -1

SHEET NO. : 32

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 8.1

136.01 - 152.94 m

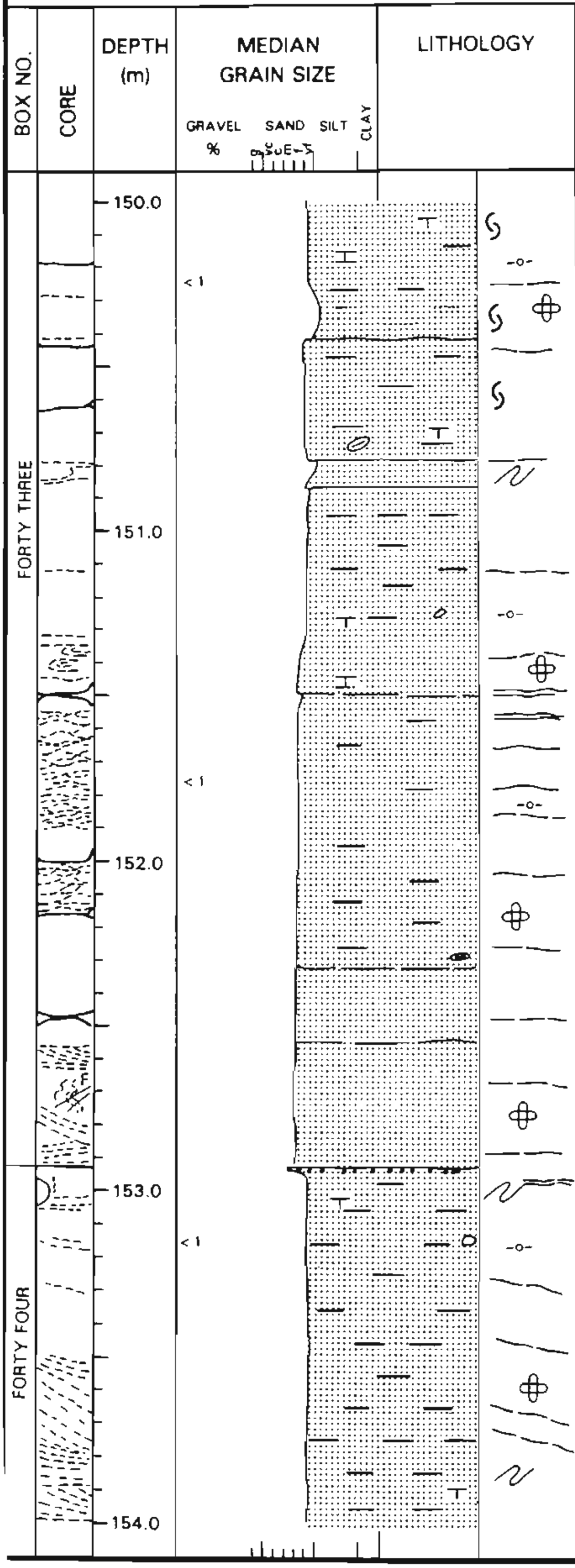
SANDSTONE

Hard to moderately hard (but soft in a few places), calcareous, unstratified to weakly stratified and well stratified (mm–bedded), moderately well sorted, **fine to very fine sandstone** (dark olive grey 5GY 3/1 to 4/1). Muddy near top of sub-unit. Patches of greater carbonate cementation are slightly lighter (olive grey 2.5GY 5/1). Burrowing commonly disrupts stratification, and in many places destroys it completely. Load features, dish structures and contorted stratification (with recumbent isoclinal folds) at several levels. Clasts are dispersed throughout but they form much less than 1% of the core and most are < 5 mm long. The larger ones range from 10 to 55 mm and are angular to rounded. Dominant lithologies are dolerite and basement. Other features include synsedimentary faults and sandstone dykes.

PROJECT : CIROS -1

SHEET NO. : 33

SCALE : 1:20



DESCRIPTION

Sub-unit 8.1

136.01 - 152.94 m

SANDSTONE

Hard to moderately hard (but soft in a few places), calcareous, unstratified to weakly stratified and well stratified (mm-bedded), moderately well sorted, **fine to very fine sandstone** (dark olive grey 5GY 3/1 to 4/1). Muddy near top of sub-unit. Patches of greater carbonate cementation are slightly lighter (olive grey 2.5GY 5/1). Burrowing commonly disrupts stratification, and in many places destroys it completely. Load features, dish structures and contorted stratification (with recumbent isoclinal folds) at several levels. Clasts are dispersed throughout but they form much less than 1% of the core and most are < 5 mm long. The larger ones range from 10 to 55 mm and are angular to rounded. Dominant lithologies are dolerite and basement. Other features include synsedimentary faults and sandstone dykes.

— contact sharp with concentration of granules -

Sub-unit 8.2

152.94 - 156.31 m

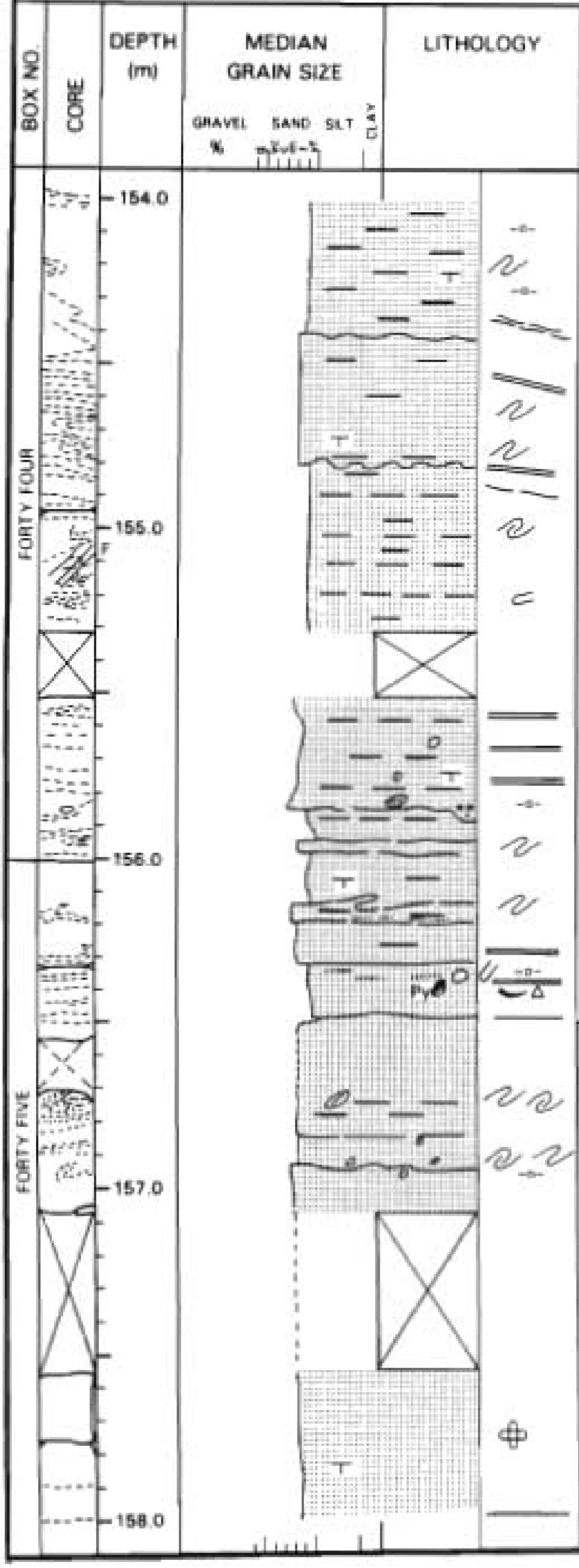
SANDSTONE

Mainly hard, well stratified (mm- and cm-bedded), moderately sorted, **muddy fine sandstone** (grey 7.5GY 4 1) and minor sandy mudstone.
Stratification largely destroyed by bioturbation between 153.20 and 153.80 m. Elsewhere it varies from subhorizontal to overturned in isoclinal slump folds (horizontal fold axes) defined by sandy and muddy beds on a scale of 1 to 10 cm. Faulting followed folding. Bioturbated interval could be large foreset.
A few scattered clasts up to 8 mm long form < 1% of the core.

PROJECT : CIROS -1

SHEET NO. : 34

SCALE : 1:20



DESCRIPTION

Sub-unit 8.2

152.94 - 156.31 m

SANDSTONE

Mainly hard, well stratified (mm- and cm-bedded), moderately sorted, **muddy fine sandstone** (grey 7.5GY 4/1) and minor sandy mudstone. Stratification largely destroyed by bioturbation between 153.20 and 153.80 m. Elsewhere it varies from subhorizontal to overturned in isoclinal slump folds (horizontal fold axes) defined by sandy and muddy beds on a scale of 1 to 10 cm. Faulting followed folding. Bioturbated interval could be large foreset. A few scattered clasts up to 8 mm long form < 1% of the core.

— contact sharp –

Sub-unit 9.1

156.31 - 158.57 m

SANDSTONE

Hard (in parts soft), weakly stratified (mm- and cm-bedded), moderately well sorted, **fine to very fine sandstone** (greenish grey to dark greenish grey 7.5GY 5/1 to 4/1)

Olive grey (2.5GY 6/1) patches of calcareous cement

Stratification ranges from subhorizontal and discontinuous to disrupted and contorted. Normally graded bedding from 158.12 to 158.41 m

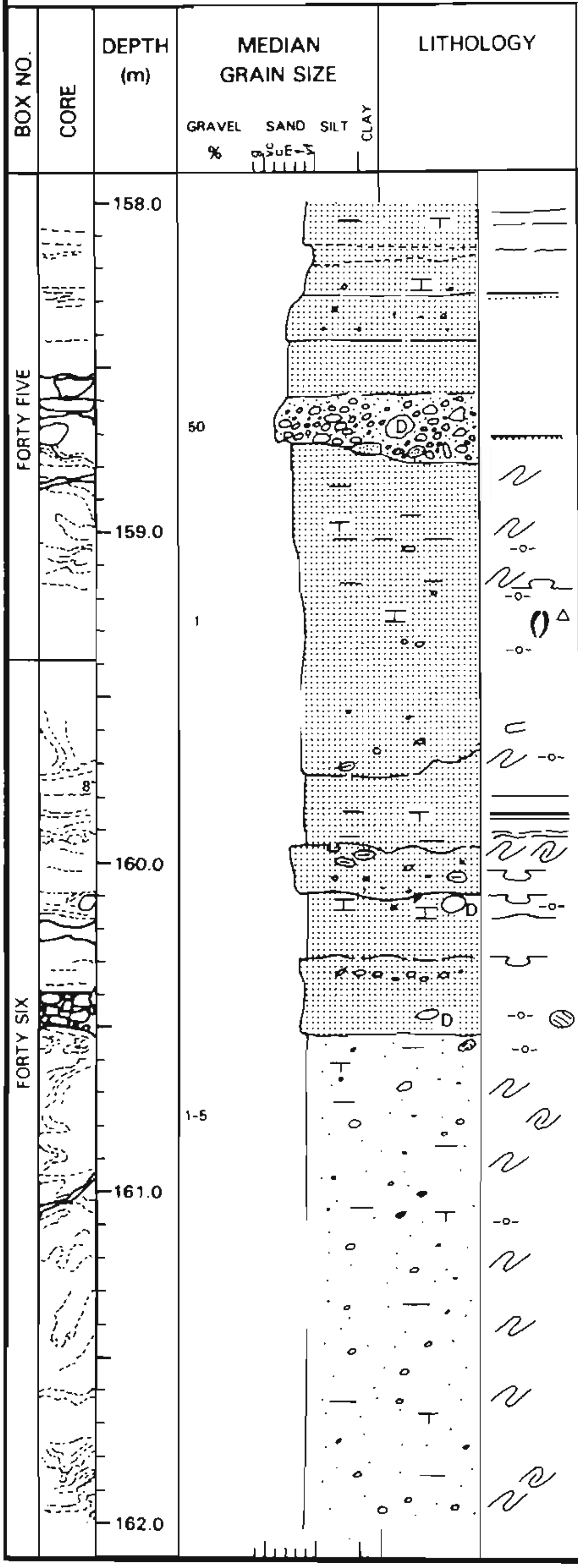
Rare bedrock lonestones and mudstone intraclasts

Some burrows, shell moulds and nodular pyrite

PROJECT : CIROS -1

SHEET NO. : 35

SCALE : 1:20



DESCRIPTION

– gradational contact –

Sub-unit 9.2

158.57 - 158.77 m

CONGLOMERATE

Moderately hard, unstratified, very poorly sorted, **sandy pebble conglomerate.**

Normally graded from clast- to matrix-supported, with matrix colour dark greenish grey (7.5GY 5/1).

Clasts form 50% of the core. They are subangular to rounded, with an average size of 6 to 8 mm and the largest at 58 mm. Basement and dolerite clasts are in equal proportions and 10% are sandstone intraclasts

– contact sharp (erosional) –

Sub-unit 9.3

158.77 - 160.51 m

SANDSTONE

Hard, weakly stratified, **conglomeratic muddy fine sandstone**(dark olive grey 2.5GY 4/1).

Stratification defined by disrupted stringers of muddy sandstone and sandy mudstone, in places highly contorted.

Bedrock clasts form between 1 and 2% of the core. Most are 4 to 10 mm, but they range up to 33 mm. Most are subangular to subrounded. About 60% are basement, 30% dolerite and 10% volcanic. Dolerite clast at base is striated and faceted.

Bed from 159.94 to 160.08 m is a moderately poorly sorted, medium sandstone with numerous sandstone rip-up clasts in the upper 6 cm.

– contact sharp –

Sub-unit 9.4

160.51 - 165.10 m

DIAMICTITE

Hard, weakly stratified, **conglomeratic muddy very fine sandstone (diamictite)** (greenish grey 10GY 5/1).

Minor conglomeratic sandy mudstone.

Stratification highly disrupted and contorted, with flame structures and brecciation.

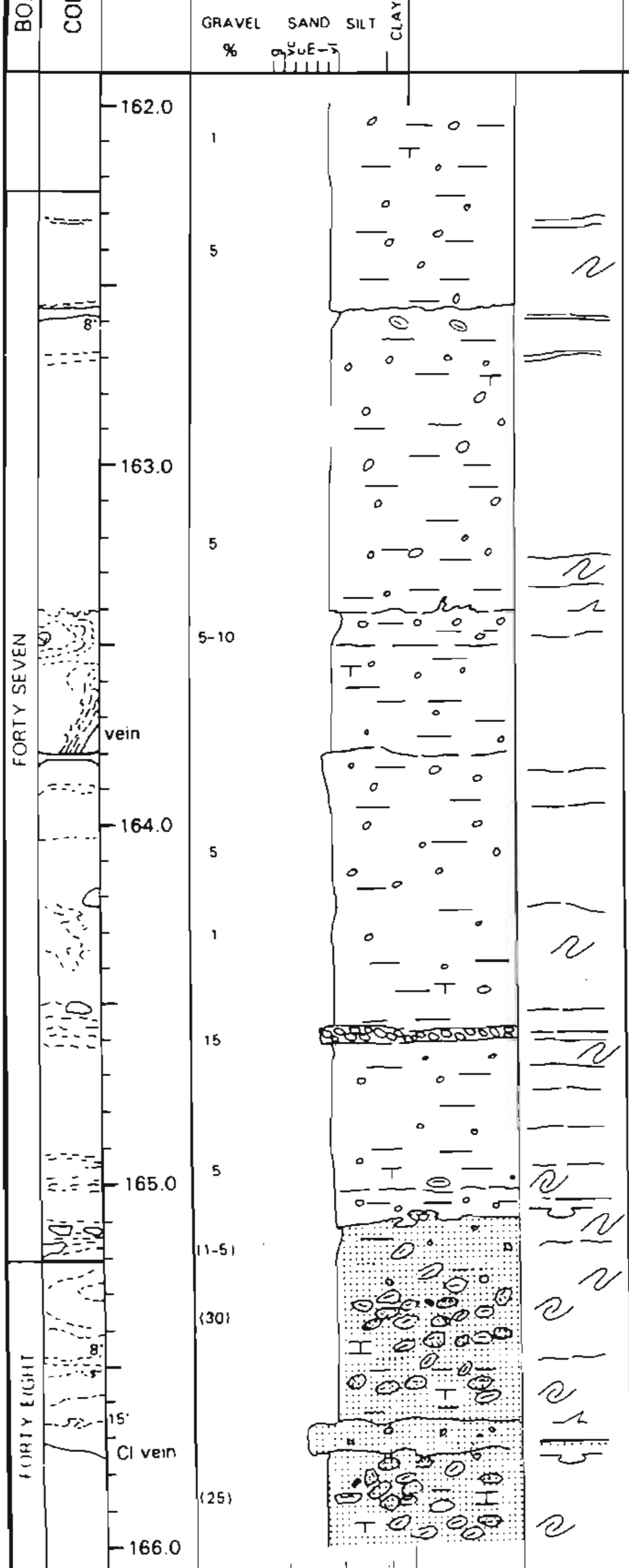
Clasts form 1 to 5% of the core, except for a layer at 164.60 m where they form 15%. Most are 5 to 15 mm, but a few are up to 30 m and are subangular to subrounded. About 60% are basement, 30% are dolerite and 10% are volcanic.

PROJECT : CIROS -1

SHEET NO. : 36

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 9.4

160.51 - 165.10 m

DIAMICTITE

Hard, weakly stratified, **conglomeratic muddy very fine sandstone (diamictite)** (greenish grey 10GY 5/1). Minor **conglomeratic sandy mudstone.** Stratification highly disrupted and contorted, with flame structures and brecciation. Clasts form 1 to 5% of the core, except for a layer at 164.60 m where they form 15%. Most are 5 to 15 mm, but a few are up to 30 m and are subangular to subrounded. About 60% are basement, 30% are dolerite and 10% are volcanic.

– contact sharp and erosional -

Sub-unit 9.5

165.10 - 168.16 m

SANDSTONE, DIAMICTITE AND CONGLOMERATE

Hard, weakly stratified (mm- and cm-bedded), moderately sorted, **muddy fine sandstone** (dark greenish grey 10GY 4/1) interbedded with poorly sorted, **coarse to medium sandstone** and **conglomerate**.

Stratification in the sandstone is disrupted and contorted and beds contain sandstone and mudstone intraclasts.

The conglomerates are clast-supported and contain 50% sandstone intraclasts, 30% dolerite and 20% basement and others. Clasts are subangular to rounded and range from 4 to 50 mm. One clast is striated.

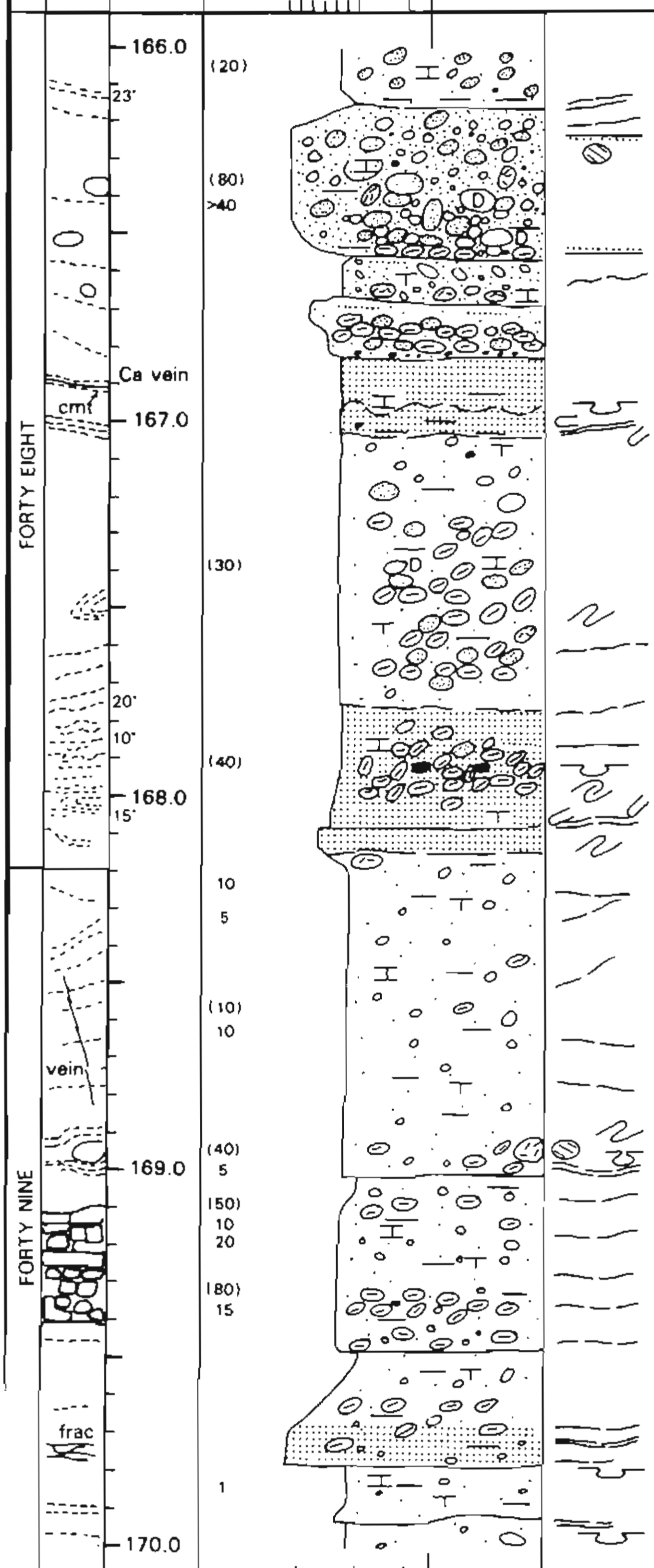
Beds of weakly stratified **diamictite** (greenish grey 10GY 5/1) from 160.00 to 160.14 m and 167.20 to 167.75 m with 20 and 60% subrounded sandstone and mudstone clasts (4 to 30 mm) respectively.

PROJECT : CIROS -1

SHEET NO. : 37

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 9.5

165.10 - 168.16 m

SANDSTONE, DIAMICTITE AND CONGLOMERATE

Hard, weakly stratified (mm- and cm-bedded), moderately sorted, **muddy fine sandstone** (dark greenish grey 10GY 4/1) interbedded with poorly sorted, **coarse to medium sandstone** and **conglomerate.** Stratification in the sandstone is disrupted and contorted and beds contain sandstone and mudstone intraclasts. The conglomerates are clast-supported and contain 50% sandstone intraclasts, 30% dolerite and 20% basement and others. Clasts are subangular to rounded and range from 4 to 50 mm. One clast is striated. Beds of weakly stratified diamictite (greenish grey 10GY 5/1) from 160.00 to 160.14 m and 167.20 to 167.75 m with 20 and 60% subrounded sandstone and mudstone clasts (4 to 30 mm) respectively.

— contact gradational over 3 cm –

Sub-unit 9.6

168.16 - 170.09 m

DIAMICTITE

Hard, weakly stratified, **conglomeratic muddy fine sandstone (diamictite)** (dark olive grey 5GY 4 1).

Stratification partially destroyed by bioturbation and slight to partial synsedimentary brecciation (in places mudstone fragments form up to 60% of core) with muddier clasts surrounded by veins of sandy material. Bedding contorted in one or two places.

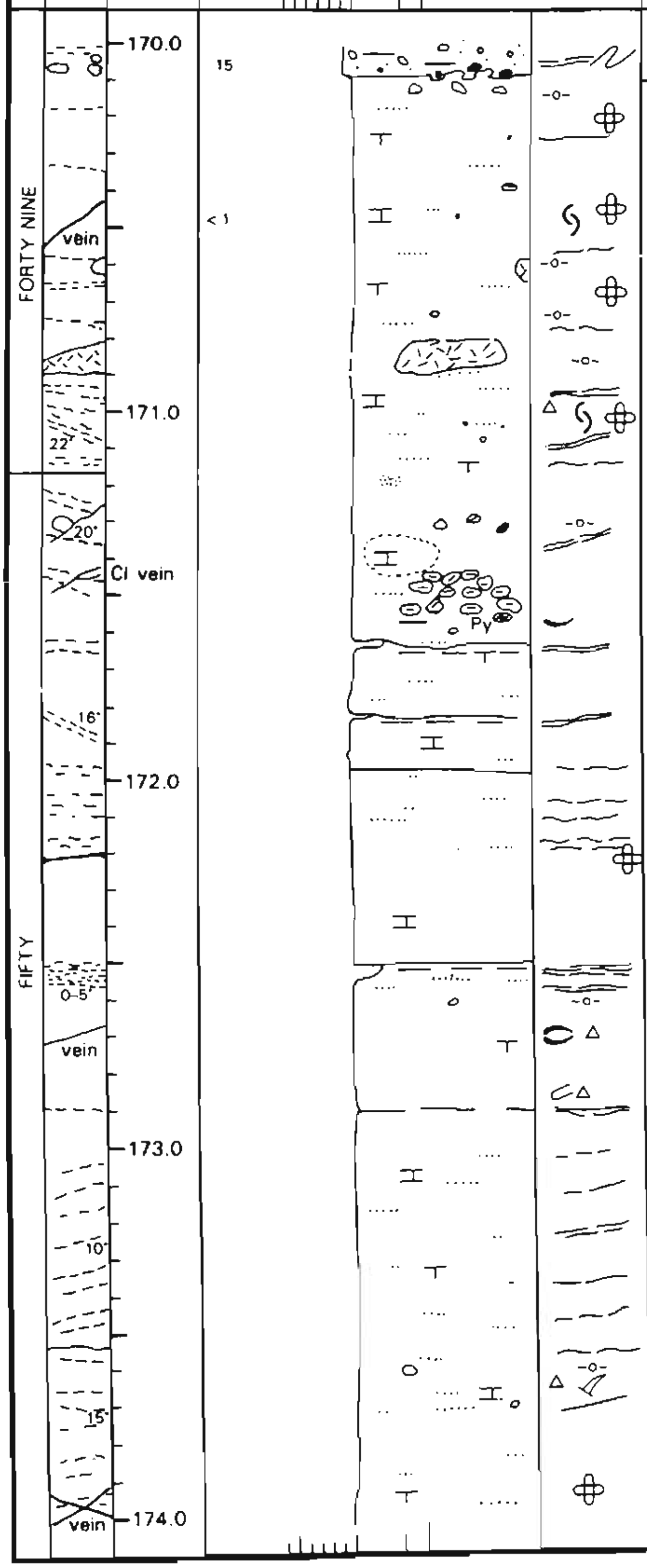
Bedrock clasts form between 5 and 15% of the core. Most are 4 to 20 mm, but several are more than 50 mm. They are subangular to subrounded. About 70% are basement and 30% dolerite. One gneiss clast is striated. Soft sediment boudinage at 169.00 m

PROJECT : CIROS -1

SHEET NO. : 38

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



—contact erosional -

Sub–unit 9.7

170.09 - 176.59 m

MUDSTONE

Hard, calcareous, weakly stratified, poorly sorted, **sandy mudstone to muddy very fine sandstone** (dark olive grey 2.5GY 4 1).

Irregular and discontinuous stratification with inclined (up to 22) and contorted bedding

Dispersed pebble size clasts (and one granite >80 mm) form <1% of the core . A dolerite clast is striated.

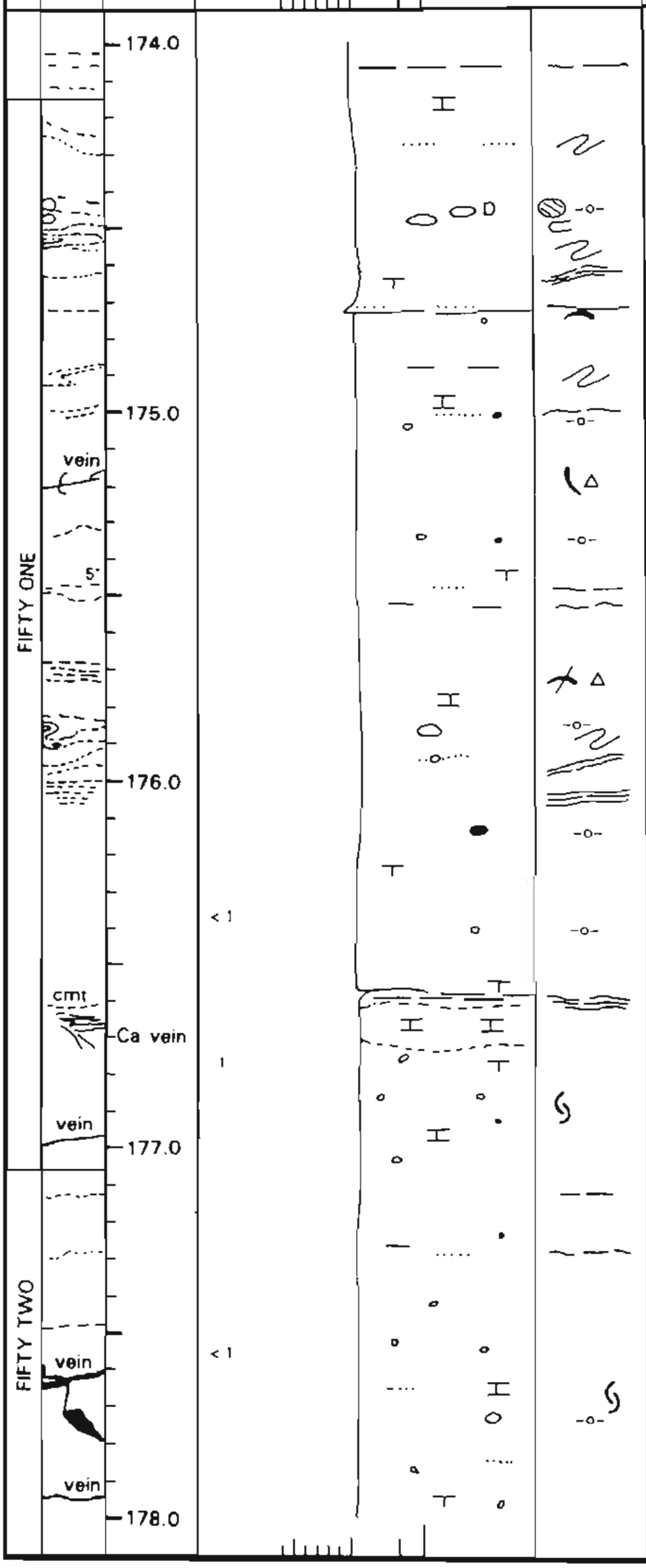
Scattered moulds and fragments of bivalves, and burrows and bioturbation structures.

PROJECT : CIROS-1

SHEET NO. : 39

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 9.7

170.09 - 176.59 m

MUDSTONE

Hard, calcareous, weakly stratified, poorly sorted, **sandy mudstone to muddy very fine sandstone** (dark olive grey 2.5GY 4/1). Irregular and discontinuous stratification with inclined (up to 22°) and contorted bedding. Dispersed pebble size clasts (and one granite > 80 mm) form < 1% of the core. A dolerite clast is striated. Scattered moulds and fragments of bivalves, and burrows and bioturbation structures.

— contact sharp but relation between sub-units gradational -

Sub-unit 9.8

176.59 - 180.30 m

MUDSTONE AND DIAMICTITE

Hard, calcareous, weakly stratified, **slightly conglomeratic sandy mudstone (diamictite** in more conglomeratic parts with gradational boundaries to mudstone beds) (dark olive grey 2.5GY 4 1).

Stratification is discontinuous, irregular and occasionally contorted. Some contorted sandy laminae. Locally mottled appearance is probably due to bioturbation.

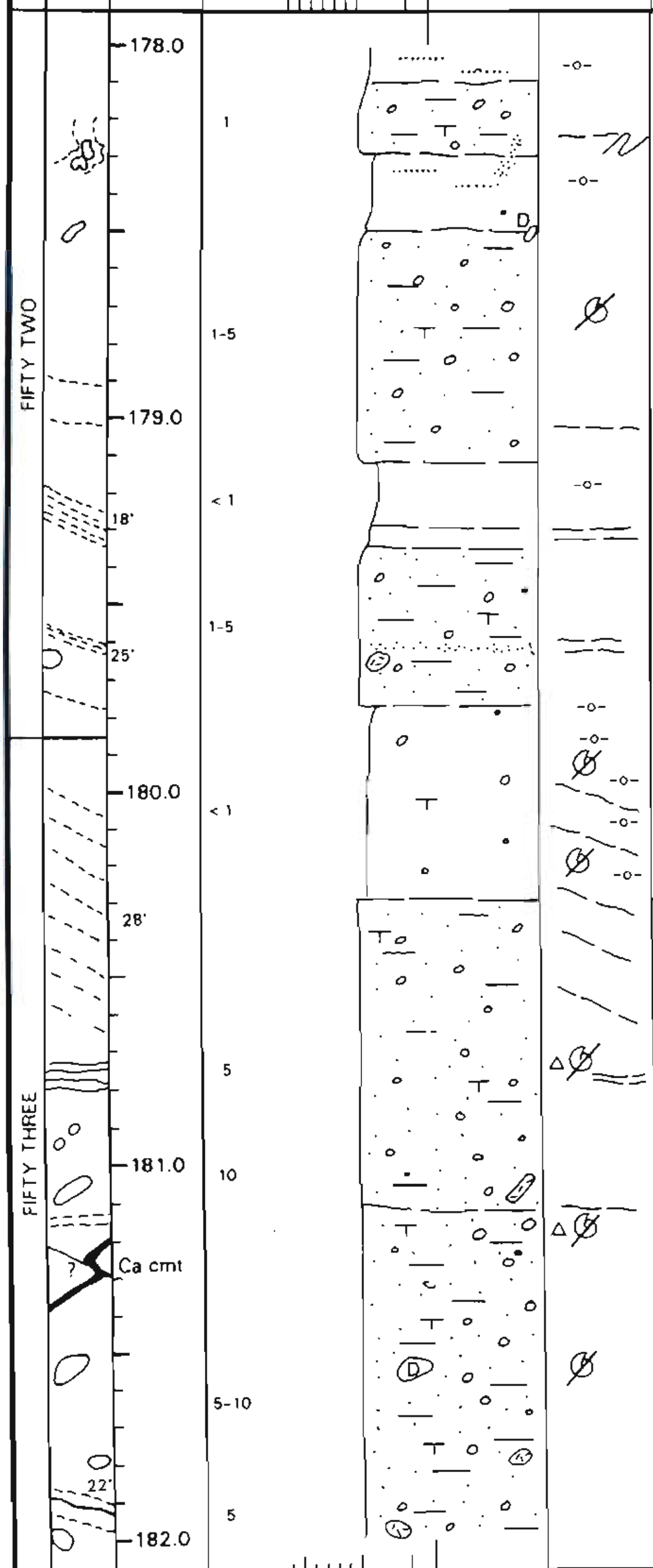
Clasts (1 to 2% of core) are up to 25 mm long and mainly subangular to subrounded. Lithologies include basement, dolerite amd volcanic rocks.

PROJECT : CIROS -1

SHEET NO. : 40

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 9.8

176.59 - 180.30 m

MUDSTONE AND DIAMICTITE

Hard, calcareous, weakly stratified, **slightly conglomeratic sandy mudstone** (**diamictite** in more conglomeratic parts with gradational boundaries to mudstone beds) (dark olive grey 2.5GY 4/1) Stratification is discontinuous, irregular and occasionally contorted. Some contorted sandy laminae. Locally mottled appearance is probably due to bioturbation Clasts (1 to 2% of core) are up to 25 mm long and mainly subangular to subrounded Lithologies include basement, dolerite amd volcanic rocks.

— contact transitional over 20 cm -

Sub-unit 9.9

180.30 - 182.79 m

DIAMICTITE

Hard, calcareous, weakly stratified, **conglomeratic sandy mudstone (diamictite)** (dark olive grey 2.5GY 4 1)

Sub-unit very homogeneous

Stratification is wispy and inclined (upto 50).

Clasts form 5 to 10% of the core. Most are 10 to 20 mm, but range up to 52 mm. One dolerite clast is striated. About 50% are basement, 30% are dolerite, 15% Beacon sandstone and 5% basalt

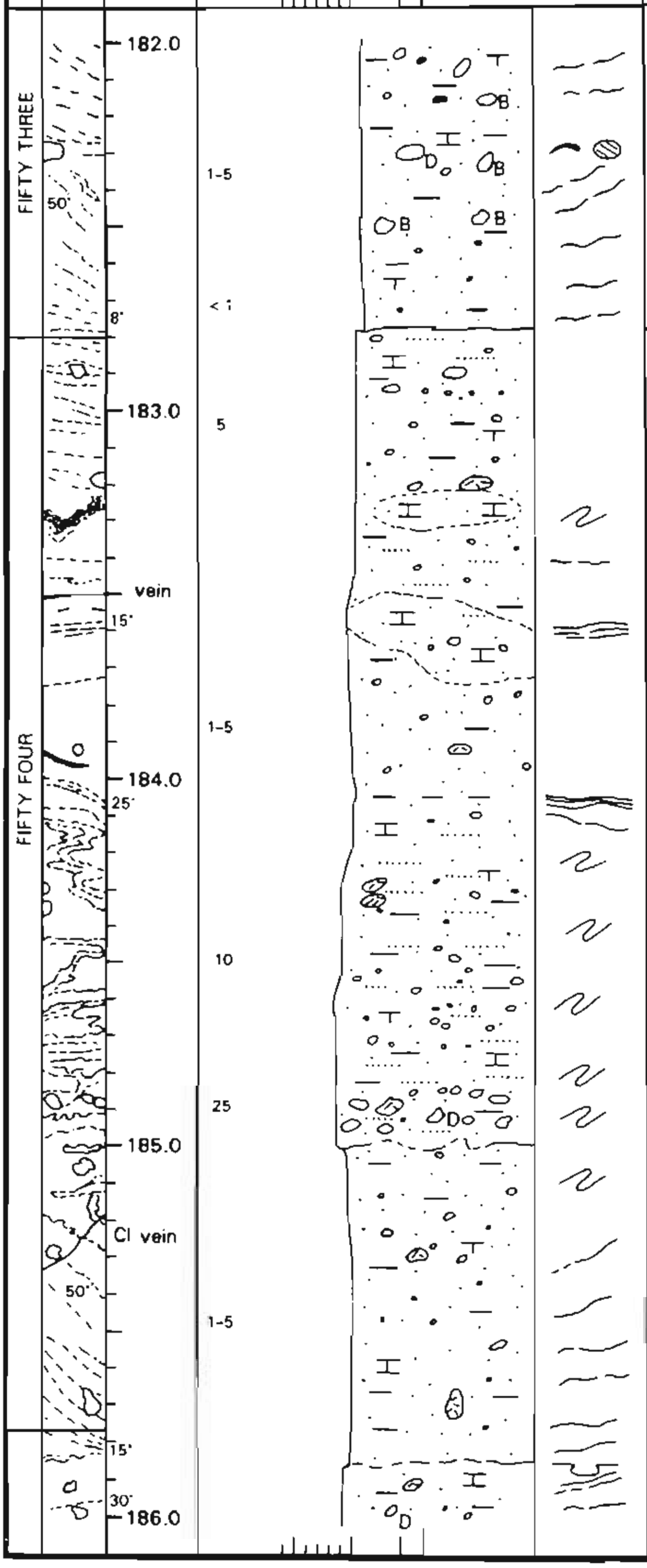
Shell fragments evident from moulds Faulted synsedimentary dyke from 181.20 to 181.35 m

PROJECT : CIROS -1

SHEET NO. : 41

SCALE : 1:20

BOX NO. | CORE | DEPTH (m) | MEDIAN GRAIN SIZE (GRAVEL % SAND SILT CLAY) | LITHOLOGY | DESCRIPTION



— contact sharp -

Sub–unit 9.10

182.79 - 200.73 m

DIAMICTITE

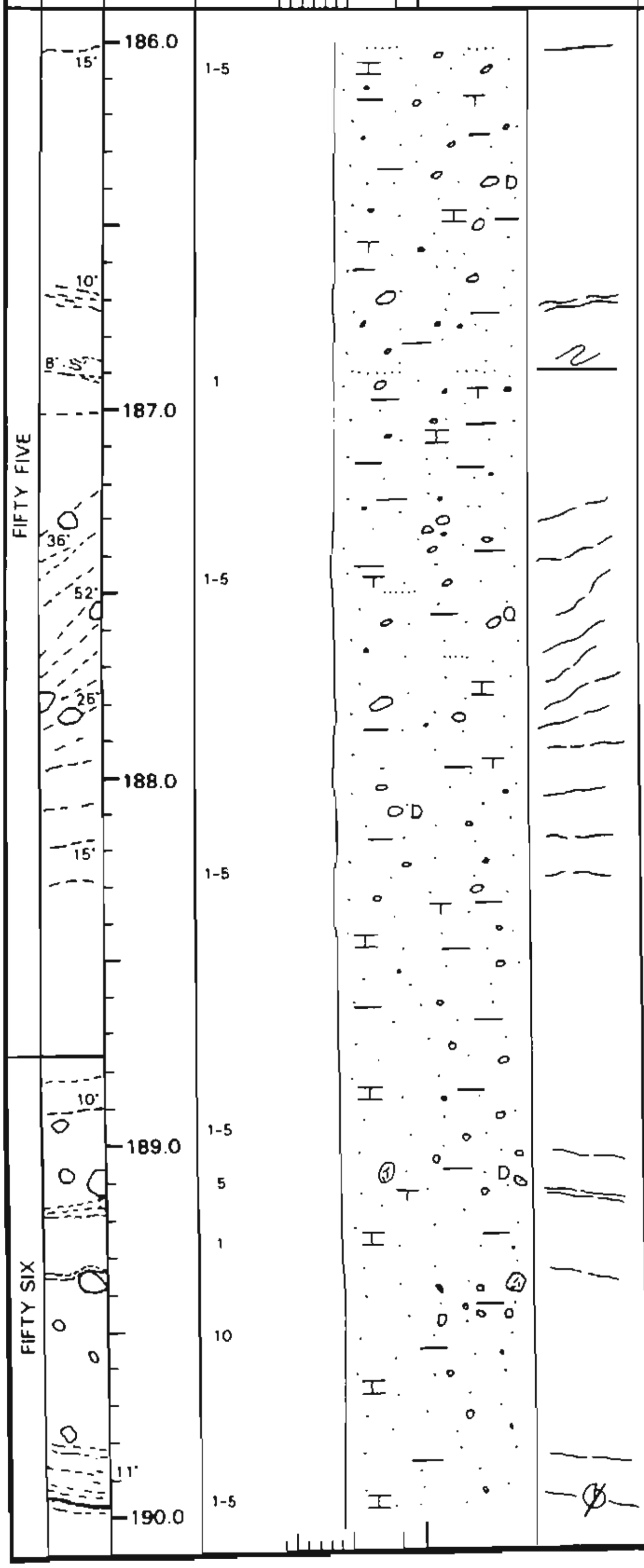
Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy fine to very fine sandstone (diamictite)** (dark olive grey 5GY 4/1).
Stratification discontinuous, irregular, inclined (up to 52°) and contorted.
Occasional thin unstratified intervals.
Clasts form between 1 and 5% of the core, except from 184.90 to 185.00 m where they form 25%. Most are 6 to 12 mm, but range up to 53 mm. long and they range from angular to rounded. Several are striated and faceted. Lithologies include basement, dolerite and the occasional Beacon sandstone.

PROJECT : CIROS -1

SHEET NO. : 42

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 9.10

182.79 - 200.73 m

DIAMICTITE

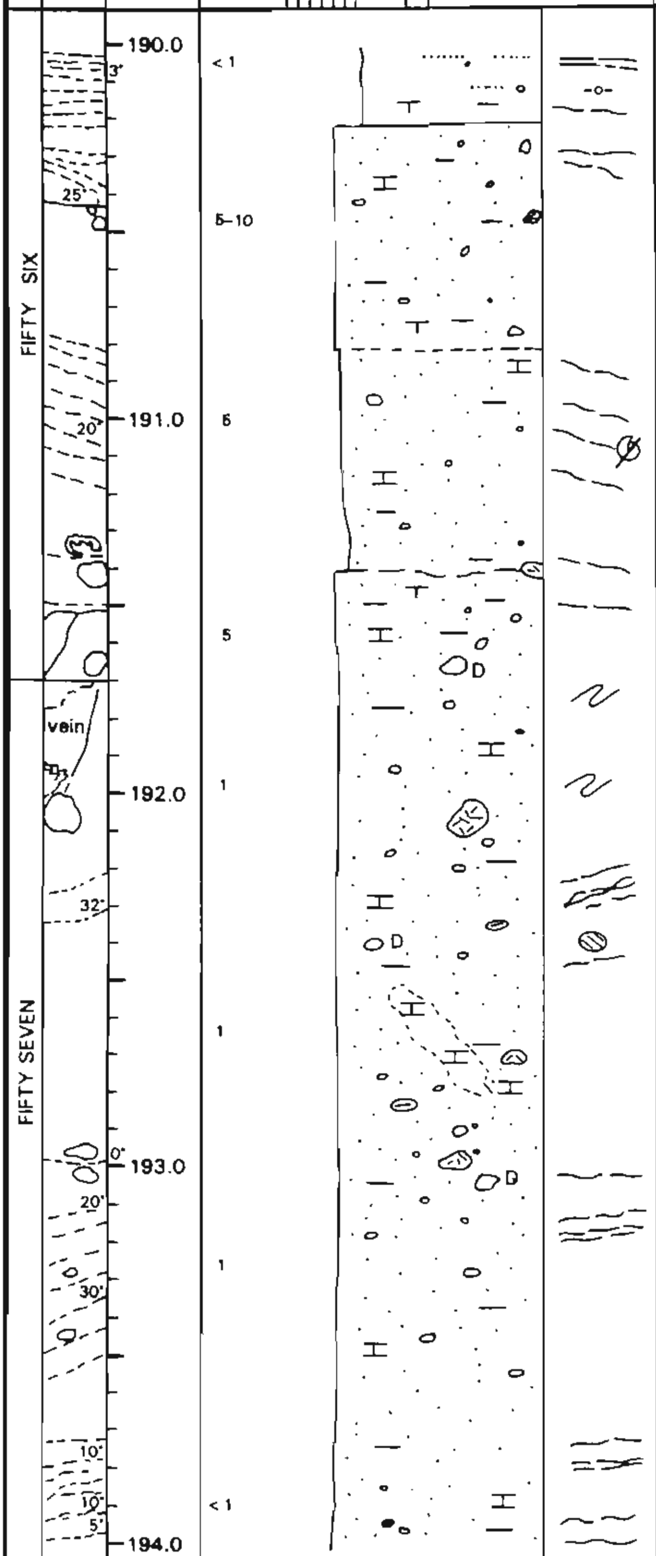
Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy fine to very fine sandstone (diamictite)** (dark olive grey 5GY 4/1). Stratification discontinuous, irregular, inclined (up to 52°) and contorted. Occasional thin unstratified intervals. Clasts form between 1 and 5% of the core, except from 184.90 to 185.00 m where they form 25%. Most are 6 to 12 mm, but range up to 53 mm, long and they range from angular to rounded. Several are striated and faceted. Lithologies include basement, dolerite and the occasional Beacon sandstone.

PROJECT : CIROS-1

SHEET NO. : 43

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 9.10

182.79 - 200.73 m

DIAMICTITE

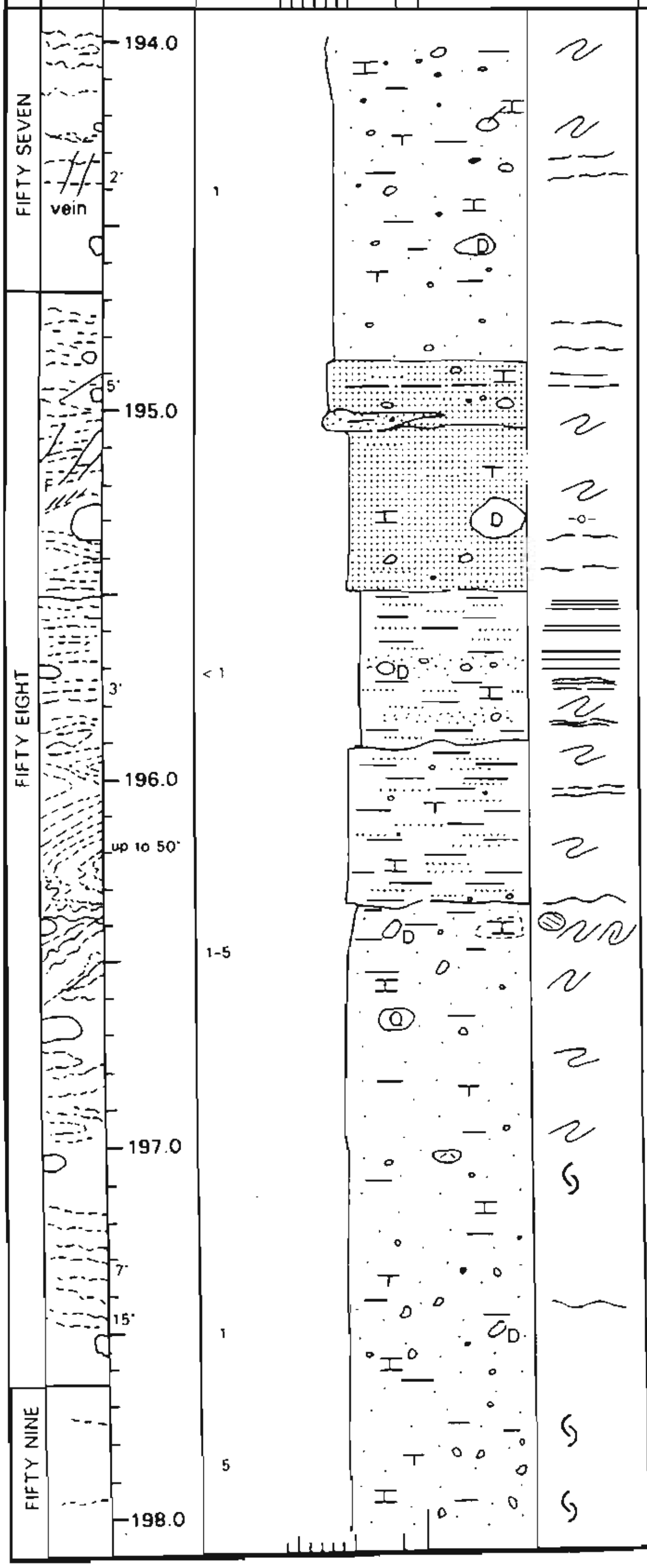
Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy fine to very fine sandstone (diamictite)** (dark olive grey 5GY 4/1). Stratification discontinuous, irregular, inclined (up to 52°) and contorted. Occasional thin unstratified intervals. Clasts form between 1 and 5% of the core, except from 184.90 to 185.00 m where they form 25%. Most are 6 to 12 mm, but range up to 53 mm, long and they range from angular to rounded. Several are striated and faceted. Lithologies include basement, dolerite and the occasional Beacon sandstone.

PROJECT : CIROS -1

SHEET NO. : 44

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 9.10

182.79 - 200.73 m

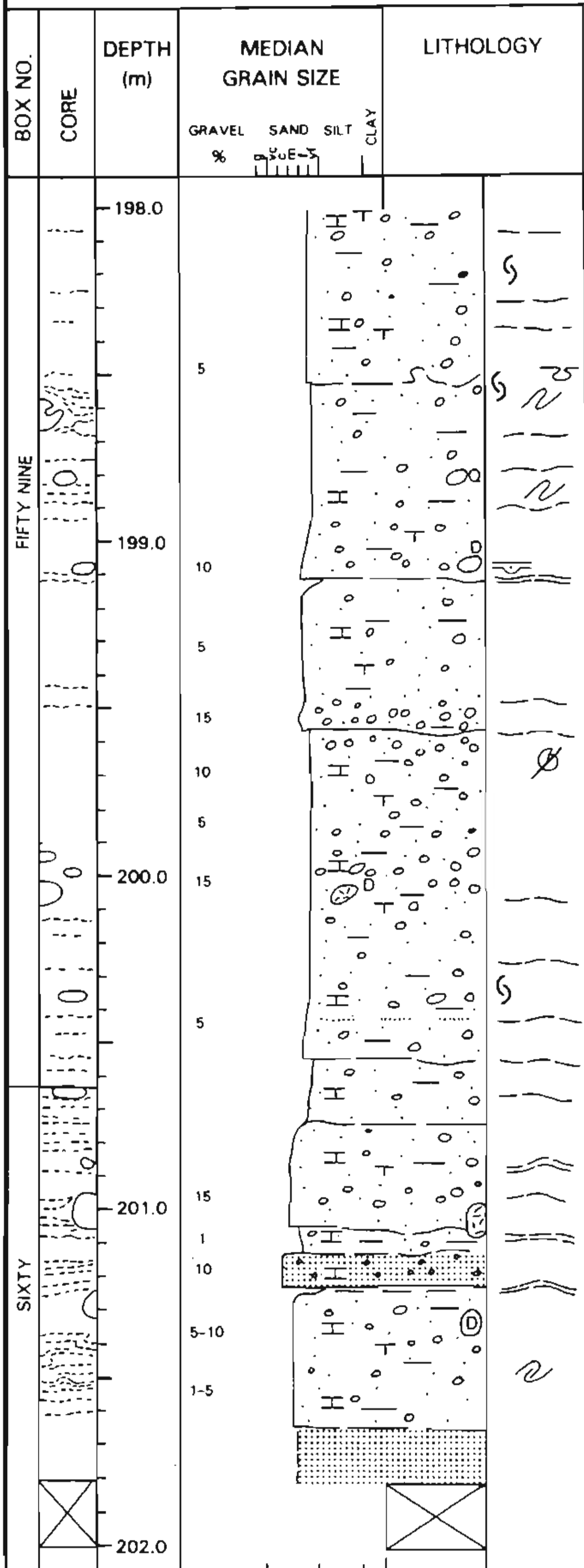
DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy fine to very fine sandstone (diamictite)** (dark olive grey 5GY 4/1). Stratification discontinuous, irregular, inclined (up to 52°) and contorted. Occasional thin unstratified intervals. Clasts form between 1 and 5% of the core, except from 184.90 to 185.00 m where they form 25%. Most are 6 to 12 mm, but range up to 53 mm, long and they range from angular to rounded. Several are striated and faceted. Lithologies include basement, dolerite and the occasional Beacon sandstone.

PROJECT : CIROS -1

SHEET NO. : 45

SCALE : 1:20



DESCRIPTION

Sub-unit 9.10

182.79 - 200.73 m

DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy fine to very fine sandstone (diamictite)** (dark olive grey 5GY 4/1). Stratification discontinuous, irregular, inclined (up to 52°) and contorted. Occasional thin unstratified intervals. Clasts form between 1 and 5% of the core, except from 184.90 to 185.00 m where they form 25%. Most are 6 to 12 mm, but range up to 53 mm, long and they range from angular to rounded. Several are striated and faceted. Lithologies include basement, dolerite and the occasional Beacon sandstone.

— contact gradational over 2 cm -

Sub-unit 10.1

200.73 - 201.65 m

DIAMICTITE

Moderately hard, well stratified, **conglomeratic muddy medium sandstone (diamictite)** (olive grey 2.5GY 6 1).

Stratification irregular, disrupted and in places contorted. Some wispy mudstone laminae and intraclasts.

Bedrock clasts form from 4 to 15% of the core Most are from 4 to 20 mm but they range up to 75 mm. Most are basement and dolerite

Bed from 201.13 to 201.23 m of well stratified, **slightly conglomeratic coarse sandstone** with pebbles up to 20 mm.

— contact gradational over 1 cm -

Sub-unit 10.2

201.65 - 202.01 m

SAND

Soft, weakly stratified, moderately well sorted, **medium to fine sand** (olive grey 2.5GY 6 1).

Grains rounded and mostly quartz. No clasts.

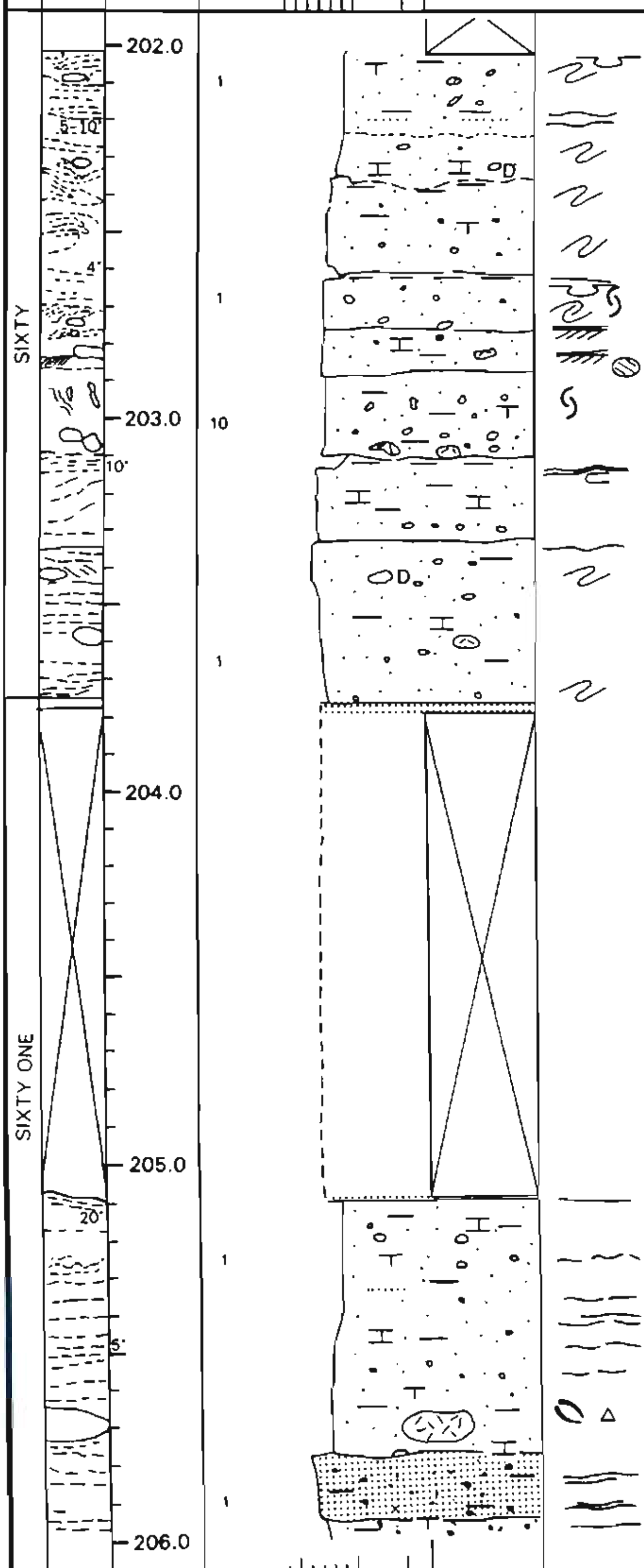
Lost core (201.82 to 202.01 m) probably of this lithology and included in this sub-unit.

PROJECT : CIROS -1

SHEET NO. : 46

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



— contact unknown due to core loss -

Sub-unit 10.3

202.01 - 203.75 m

DIAMICTITE

Hard, calcareous, well stratified (mm- and cm-bedded), **conglomeratic muddy medium to fine sandstone (diamictite)** (greenish grey 7.5GY 5 1).

Stratification irregular, discontinuous, wispy and contorted. Occasional load features and cross-bedding.

Basement and dolerite clasts generally form about 1% of the core (but up to 10% from 203.00 to 203.10 m). They range in size from 4 to 32 mm and average about 10 mm. Roundness ranges from angular to rounded and on average is subrounded. One basement clast is striated.

— contact sharp -

Sub-unit 10.4

203.75 - 205.09 m

SAND

Soft, unstratified, well sorted, **medium sand** (olive grey 2.5GY 6 1).
Most of the sub-unit (203.78 to 205.08 m) not recovered.

— contact sharp -

Sub-unit 10.5

205.09 - 210.72 m

DIAMICTITE

Hard, calcareous, unstratified to well stratified (mm- and cm-bedded), **slightly conglomeratic muddy fine sandstone (diamictite)**(greenish grey 7.5GY 5 1).

Stratification ranges from wispy discontinuous irregular bedding with average dips of 5-10 to disrupted and contorted bedding.

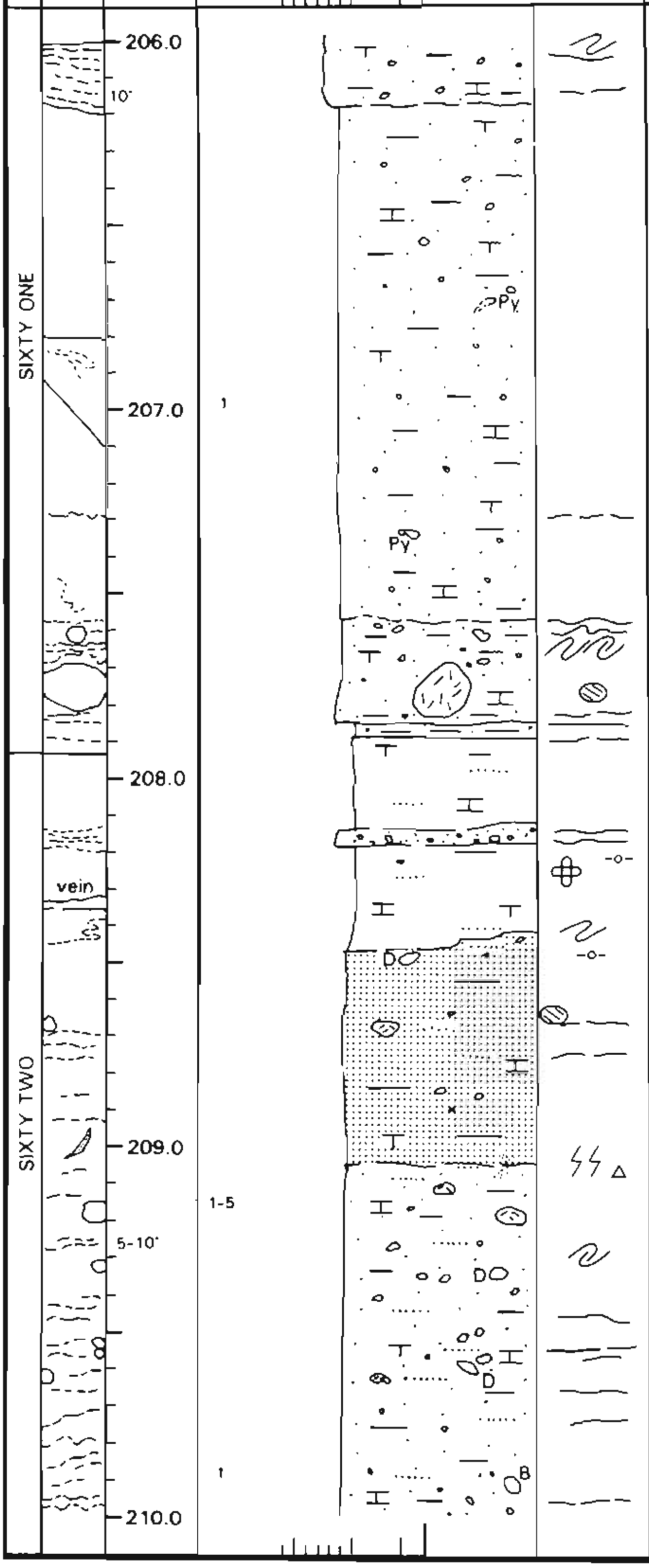
Clasts form 1-2% of the core but locally reach 5%. Average size is 10 mm, but one diorite is 105 mm. Two clasts are striated. They range from subangular to rounded. Lithologies include dolerite, basement and Beacon sandstone.

PROJECT : CIROS -1

SHEET NO. : 47

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



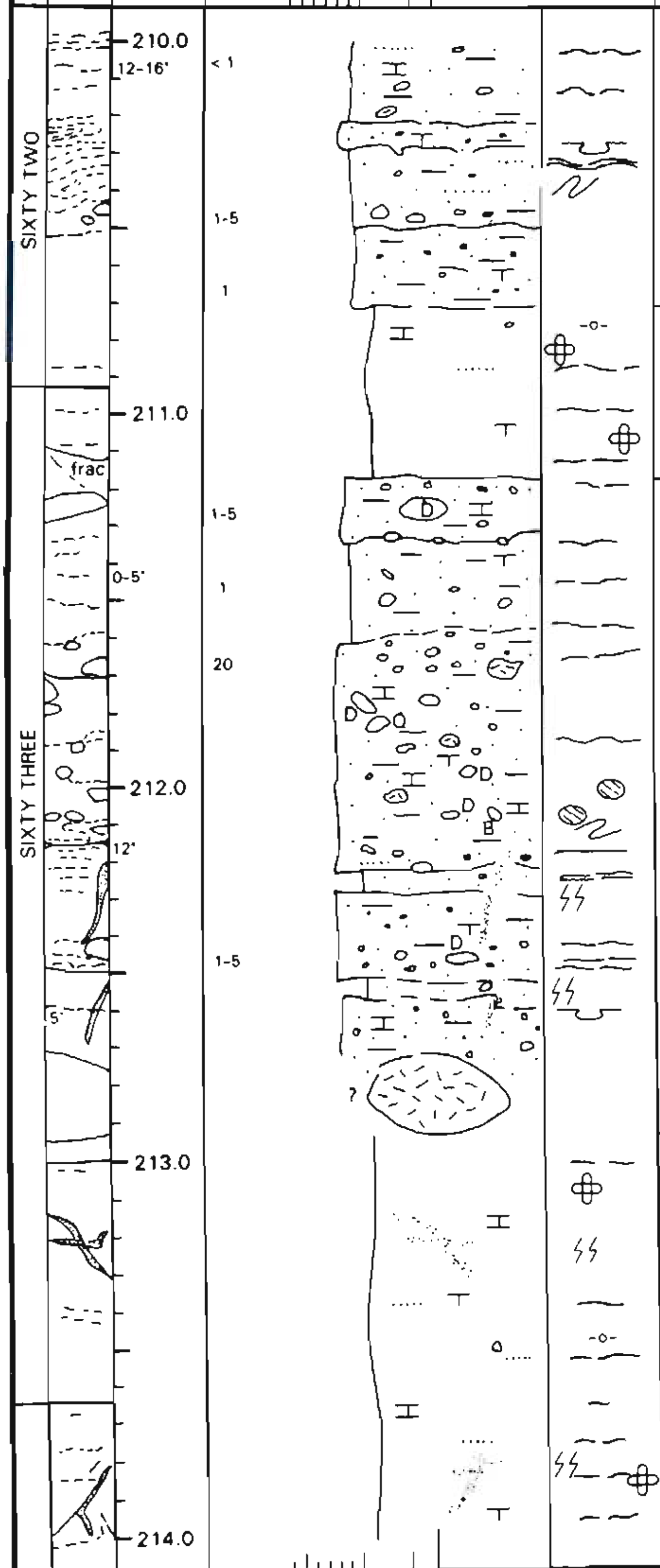
Interbeds of unstratified **sandy mudstone** (greenish grey 7.5GY 5 1) and weakly stratified **muddy very fine sandstone** (olive grey 2.5GY 6/1) at 207.84 to 208.45 m and 208.45 to 209.05 m respectively.

PROJECT : CIROS -1

SHEET NO. : 48

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



— contact gradational over 3 cm -

Sub-unit 10.6

210.72 - 211.16 m

MUDSTONE

Hard, calcareous, weakly stratified, moderately well sorted, **sandy mudstone** (dark olive grey 2.5GY 4 1).
Stratification is subhorizontal, wispy and partially destroyed by bioturbation

— contact gradational -

Sub-unit 10.7

211.16 - 212.93 m

DIAMICTITE

Hard, calcareous, weakly stratified (in places well stratified), **conglomeratic muddy fine sandstone (diamictite)** (greenish grey 10GY 6 1)
Stratification is discontinuous and irregular, with occasional load structures and contorted bedding.
Clasts in upper and lower part of the sub-unit form 1 to 5% of the core but in central part form up to 20%. Average clast size is 8 to 12 mm with one gneiss clast >240 mm. Some clasts are striated, and most are subangular to subrounded. About 60% are dolerite, 30% are basement and 10% Beacon sandstone

— contact at base of gneiss boulder -

Sub-unit 11.1

212.93 - 225.46 m

MUDSTONE

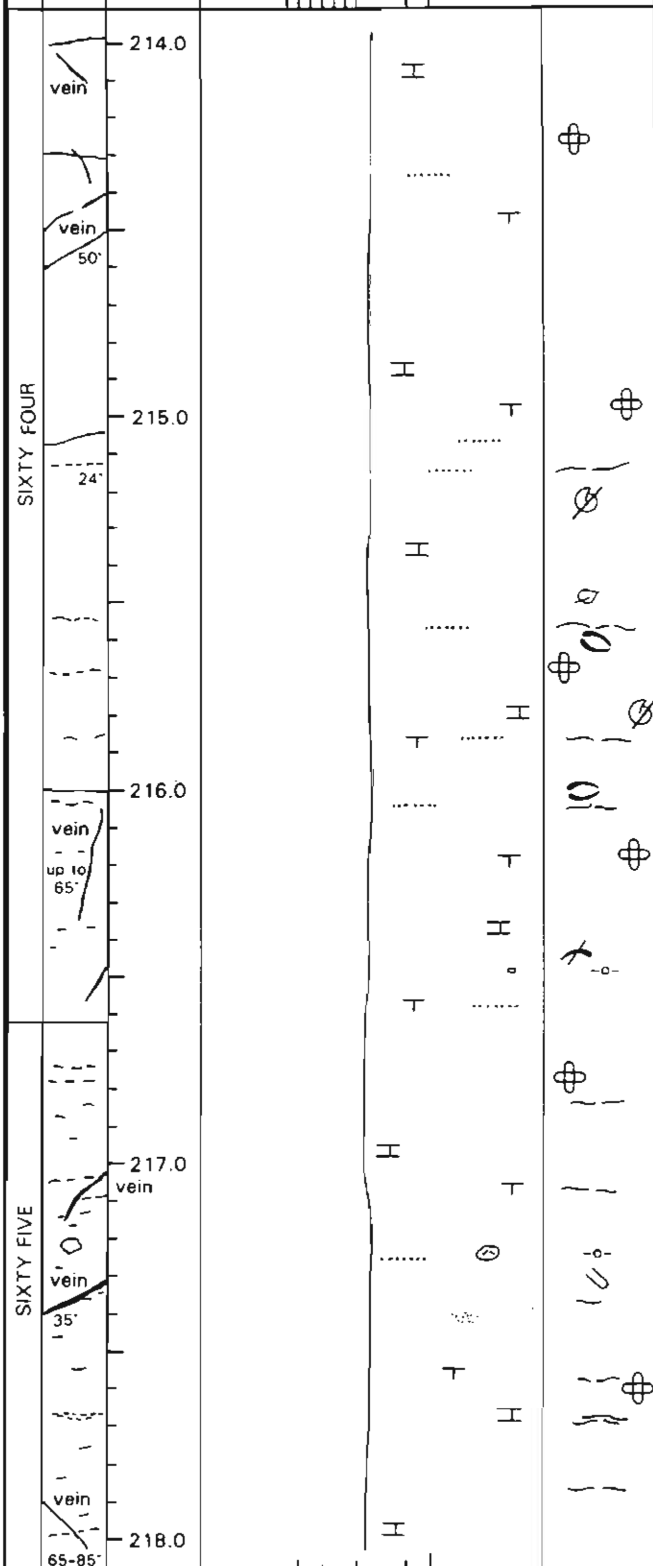
Hard, calcareous, unstratified to weakly stratified, **sandy mudstone** (dark greenish grey 5G 4 1)
Strongly bioturbated.
Leaf impression and whole articulated pectinid at 215.47 and 215.58 m
Possible carbonaceous fragments below 225.00 m

PROJECT : CIROS -1

SHEET NO. : 49

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % SAND SILT CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 11.1

212.93 - 225.46 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Strongly bioturbated. Leaf impression and whole articulated pectinid at 215.47 and 215.58 m. Possible carbonaceous fragments below 225.00 m.

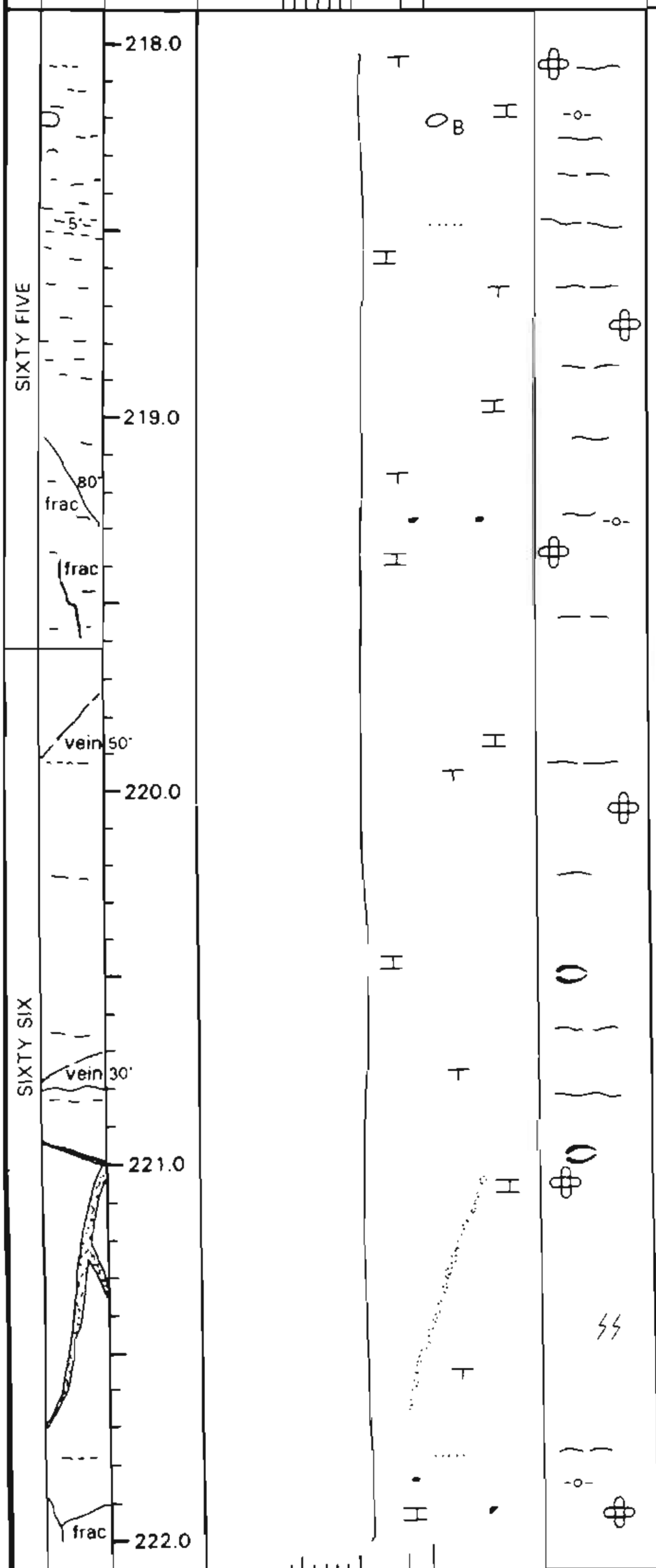
PROJECT : CIROS -1

SHEET NO. : 50

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION

GRAVEL % SAND SILT CLAY



Sub-unit 11.1

212.93 - 225.46 m

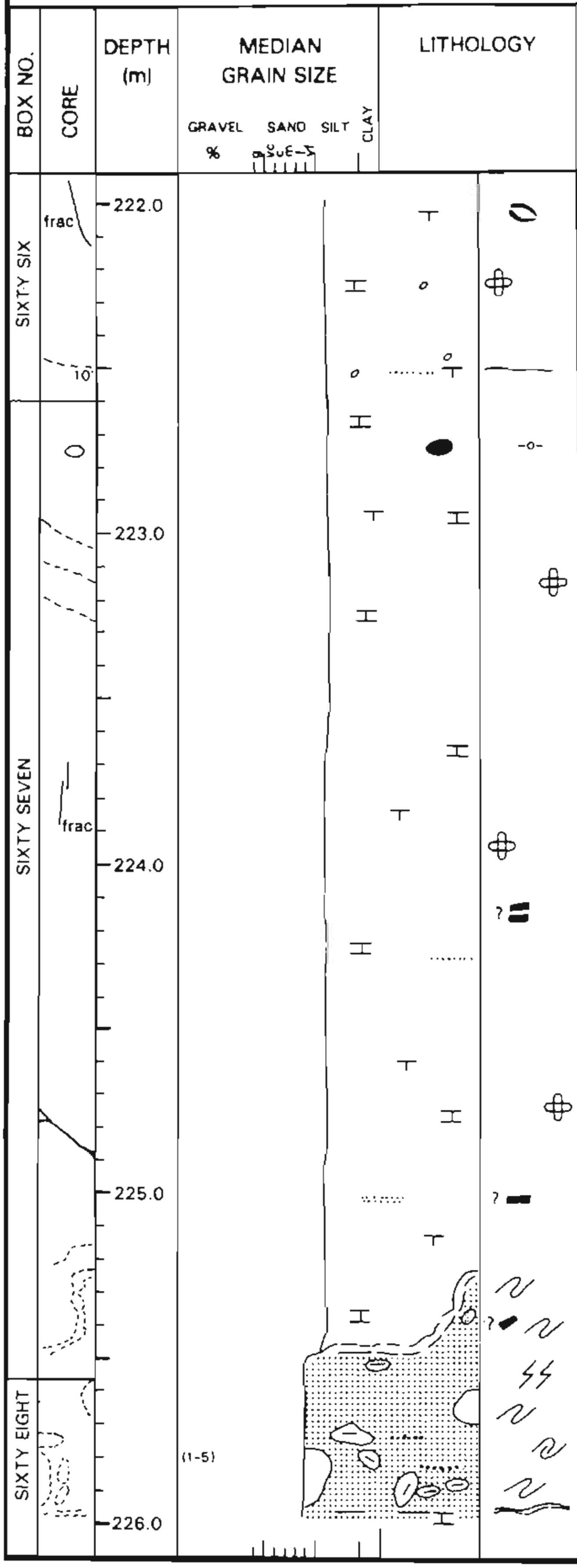
MUDSTONE

Hard, calcareous, unstratified to weakly stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Strongly bioturbated. Leaf impression and whole articulated pectinid at 215.47 and 215.58 m. Possible carbonaceous fragments below 225.00 m

PROJECT : CIROS -1

SHEET NO. : 51

SCALE : 1:20



DESCRIPTION

Sub-unit 11.1

212.93 - 225.46 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Strongly bioturbated. Leaf impression and whole articulated pectinid at 215.47 and 215.58 m. Possible carbonaceous fragments below 225.00 m.

— contact sharp. Mudstone injected with sandstone from beneath -

Sub-unit 12.1

225.46 - 226.07 m

SANDSTONE DYKE IN MUDSTONE

Hard, calcareous, weakly stratified and intensely deformed, poorly sorted, **coarse to fine sandstone** dyke (olive grey 2.5GY 5 1) intruding **sandy mudstone** (grey 10Y 5 1 to 4 1).

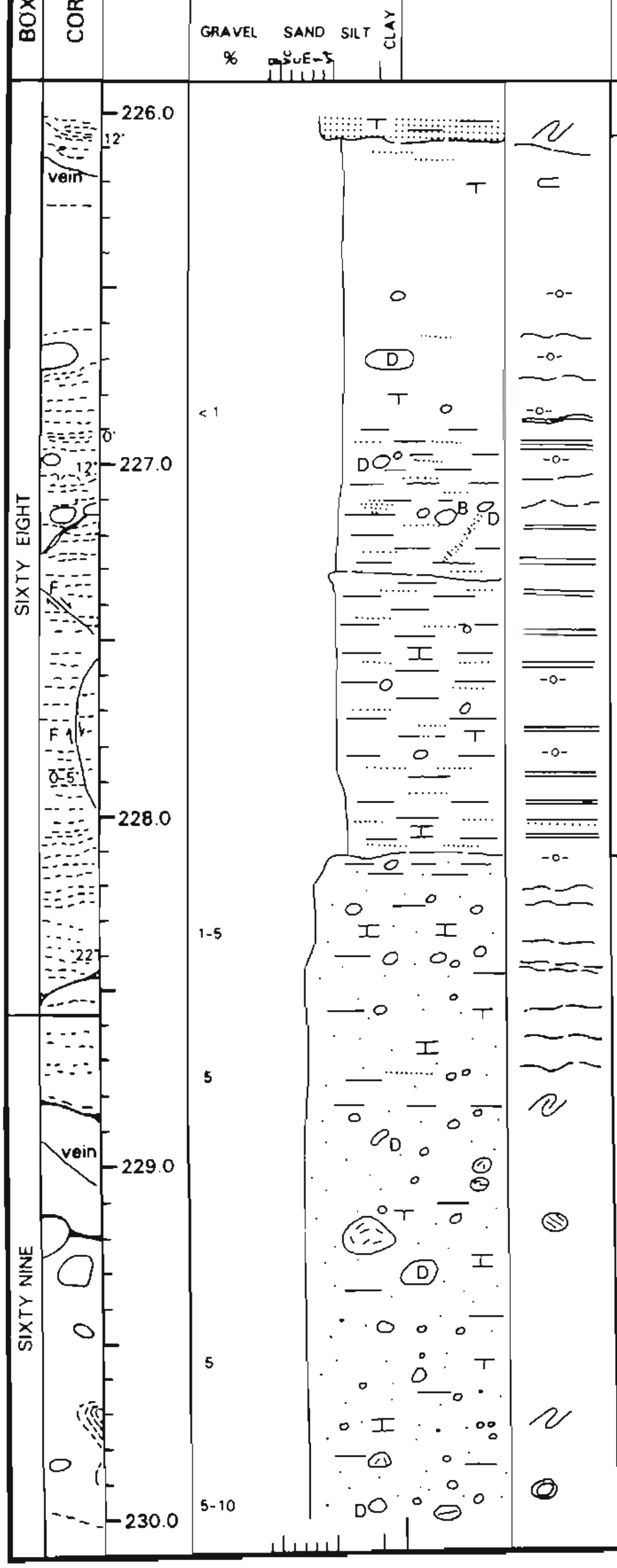
Large mudstone clasts within the sandstone. Possible carbonaceous fragments at 225 40 m.

PROJECT : CIROS -1

SHEET NO. : 52

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



— contact gradational over 5 cm -

Sub-unit 12.2

226.07 - 228.12 m

MUDSTONE

Hard, calcareous, weakly to well stratified (mm- and occasionally cm-bedded), fine **sandy mudstone** (grey 10Y 5 1 to 4 1).
Stratification in places rhythmic and weakly graded.
Clasts form < 1% of the core and range from subangular to well rounded. Most are 4 to 20 mm but one is > 53 mm. Lithologies are of basement, dolerite and Beacon sandstone.

— contact interfingering over 10 cm -

Sub-unit 12.3

228.12 - 233.28 m

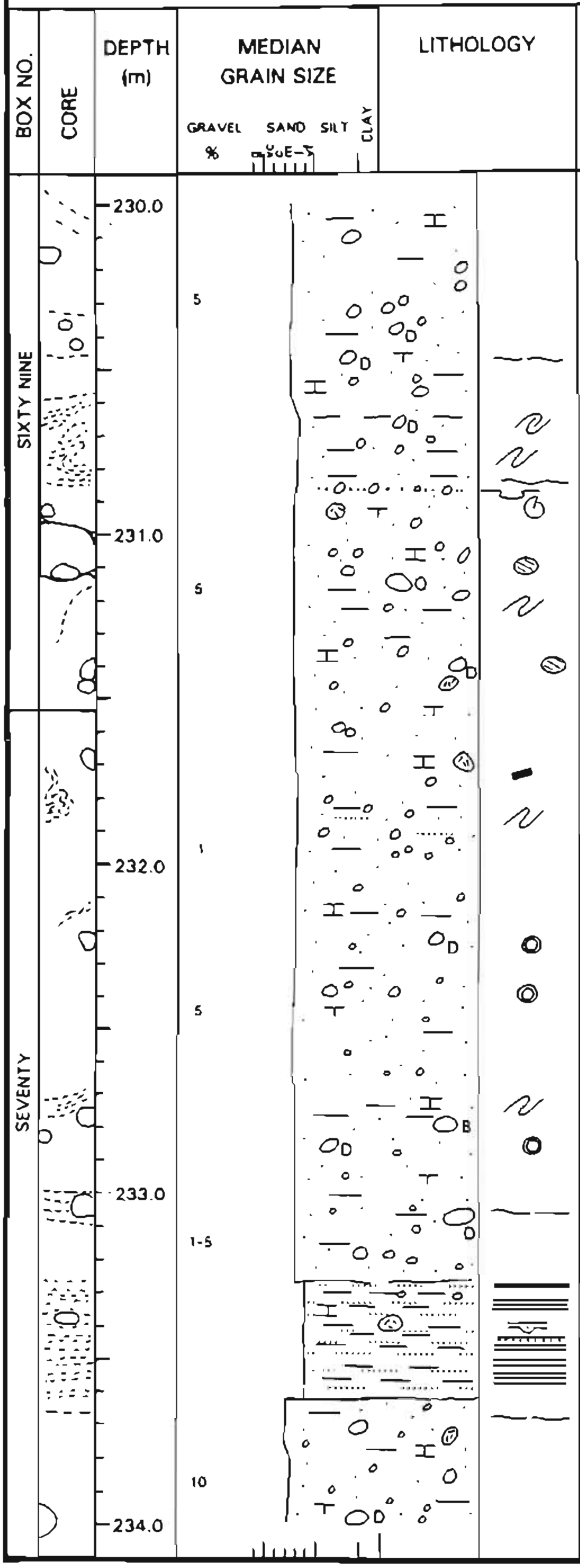
DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy medium sandstone** (dark olive grey to greenish grey 2.5GY 4 1 to 10GY 6 1-4 1).
Stratification varies from wispy and contorted to nonexistent.
Clasts form on average 5% of the core, and are fairly evenly distributed except for a clast-poor stratified zone from 230 62 to 230 90 m. Most range from 5 to 50 mm, but some greater than 80 mm. They are mainly subangular to subrounded, and several are striated. Most are dolerite but there are also basement and Beacon sandstone clasts and a carbonaceous fragment at 231 72 m

PROJECT : CIROS -1

SHEET NO. : 53

SCALE : 1:20



DESCRIPTION

Sub-unit 12.3

228.12 - 233.28 m

DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, conglomeratic muddy medium sandstone (dark olive grey to greenish grey 2.5GY 4/1 to 10GY 6/1–4/1). Stratification varies from wispy and contorted to nonexistent. Clasts form on average 5% of the core, and are fairly evenly distributed except for a clast-poor stratified zone from 230.62 to 230.90 m. Most range from 5 to 50 mm, but some greater than 80 mm. They are mainly subangular to subrounded, and several are striated. Most are dolerite, but there are also basement and Beacon sandstone clasts and a carbonaceous fragment at 231.72 m.

— contact gradational over a few mm -

Sub-unit 12.4

233.28 - 233.63 m

SANDSTONE AND MUDSTONE

Interbeds of hard, calcareous, well stratified (mm-bedded), fine sandstone grading upwards to sandy mudstone (rhythmites) with a few dispersed clasts including a granite dropstone.

— contact sharp -

Sub-unit 12.5

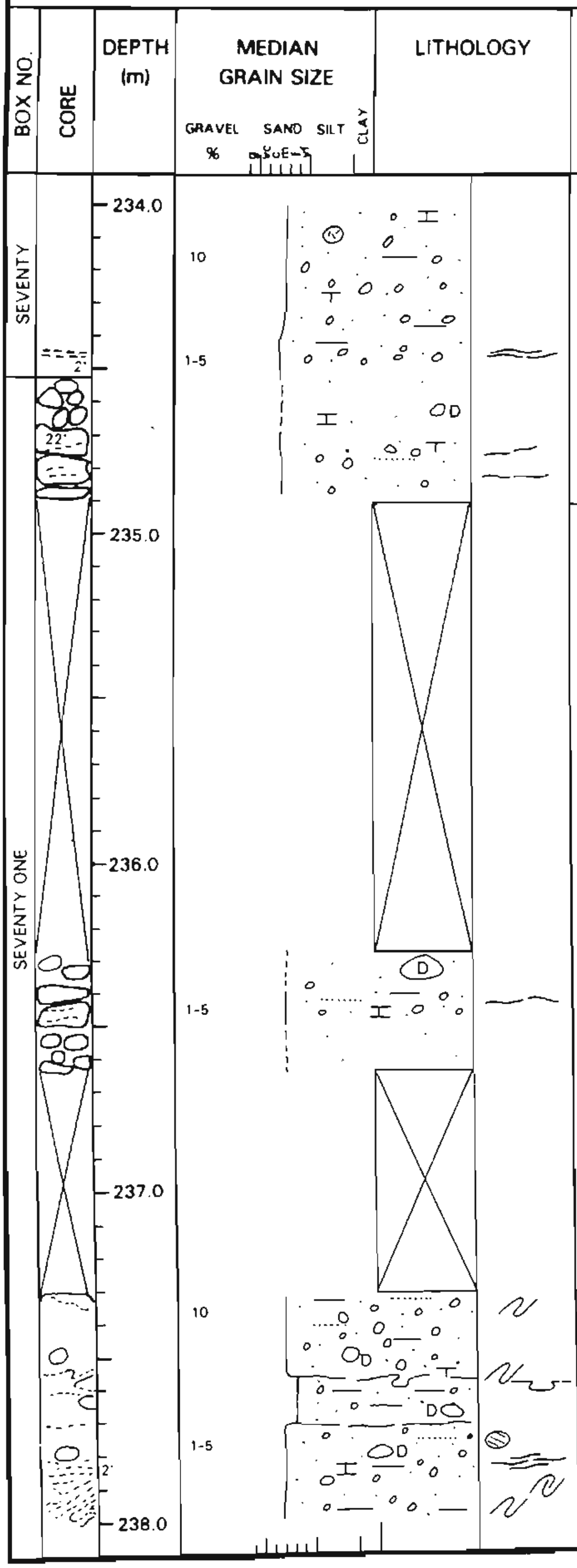
233.63 - ?234.91 m

DIAMICTITE

PROJECT : CIROS -1

SHEET NO. : 54

SCALE : 1:20



DESCRIPTION

Sub-unit 12.5

233.63 - ?234.91 m

DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy medium sandstone (diamictite)** (dark olive grey 2.5GY 4/1). Occasional wisps of sandstone laminae. Clasts mainly form 5 to 10% of the rock, and range in size up to 32 mm. Most are subangular to subrounded. About 70% are dolerite, 25% basement and about 5% are sedimentary and other

- contact unknown due to core loss -

Sub-unit 12.6

?234.91 - ?237.32 m

?SAND AND DIAMICTITE

Poor core recovery suggests soft **sand** from 234.91 to 236.29 m and from 236.65 to 237.32 m. Core recovered between these intervals is firm but broken fragments of weakly stratified **conglomeratic muddy medium sandstone (diamictite)** (dark olive grey 2.5GY 4 1)

- contact unknown due to core loss -

Sub-unit 12.7

?237.32 - 239.14 m

DIAMICTITE

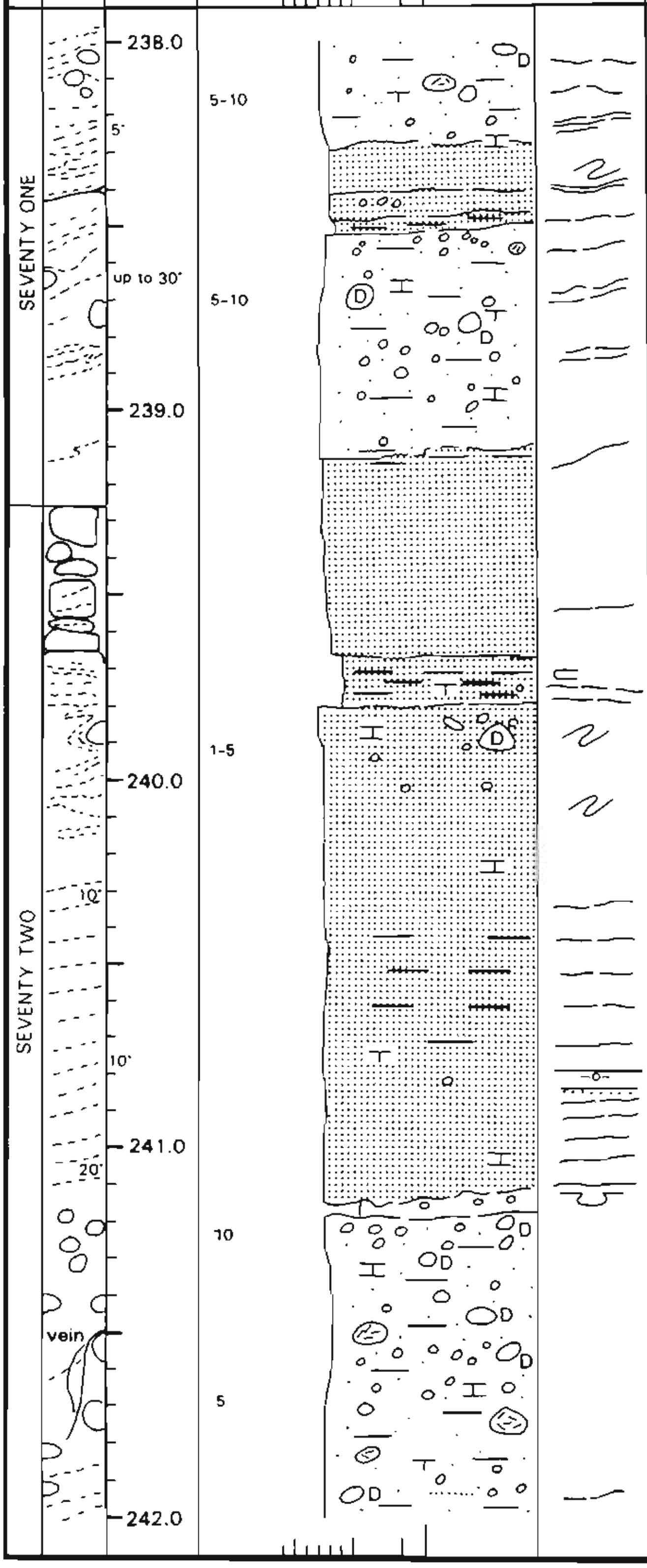
Hard, calcareous, weakly and well stratified, **conglomeratic muddy medium sandstone (diamictite)** (dark olive grey to dark greenish grey 2.5GY 4 1 to 7.5GY 4 1).
Stratification is irregular and discontinuous with disrupted and contorted bedding towards the top, and inclined bedding dipping from 5 to 55 towards the base.

PROJECT : CIROS -1

SHEET NO. : 55

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Clasts form 1-10% of the core, averaging 12 mm, but up to 37 mm, long. They are subangular to well rounded and are mainly dolerite (70%), basement (25%) and Beacon sandstone (5%). One clast is striated.

— contact sharp -

Sub-unit 12.8

239.14 - 241.19 m

SANDSTONE

Variably hard and soft, locally calcareous, unstratified to well stratified, moderately well sorted, **medium sandstone**(dark greenish grey 10GY 4 1).

Beds of **muddy fine sandstone** and **sandy mudstone** from 239.69 to 239.81 m and from 241.15 to 241.19 m respectively.
Stratification is discontinuous and locally contorted; some intervals are subhorizontal and others inclined up to 20 . A few beds are graded.
Clasts form <1% except for the upper 20 cm where they form 1-2% of the core. They are mostly about 10 mm but one dolerite is >55 mm. They range from subangular to subrounded.

— contact sharp -

Sub-unit 12.9

241.19 - 248.71 m

DIAMICTITE

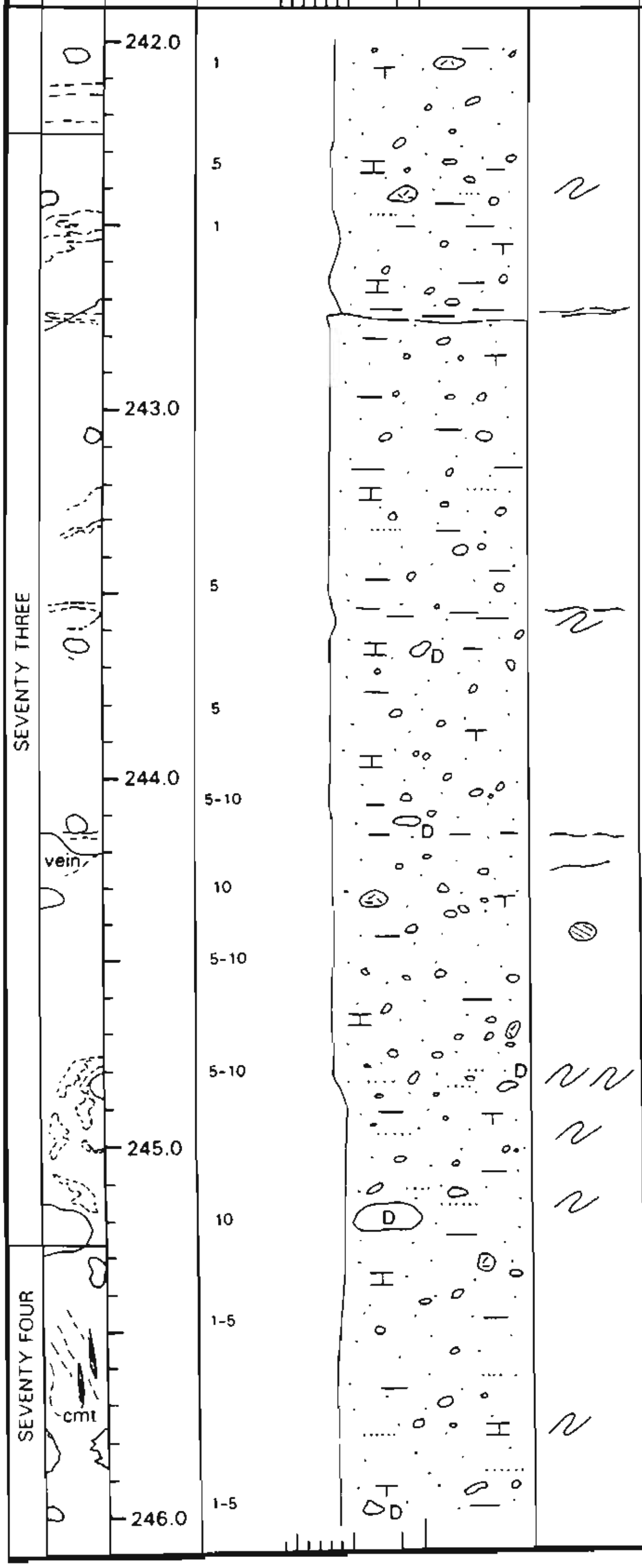
Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy medium to very fine sandstone (diamictite)** (dark greenish grey to greyish olive 10GY 4 1 to 7.5Y 6 2 and 2.5GY 5 1).
Stratification is wispy and discontinuous, occasionally disrupted and commonly contorted, dipping up to 23
Clasts form mainly between 5 and 10% of the core, averaging about 10 mm but reaching >90 mm long. Most are subangular to subrounded but a few are well-rounded. One is striated. About 70% are dolerite, 25%

PROJECT : CIROS -1

SHEET NO. : 56

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % · SAND · SILT · CLAY)	LITHOLOGY	DESCRIPTION



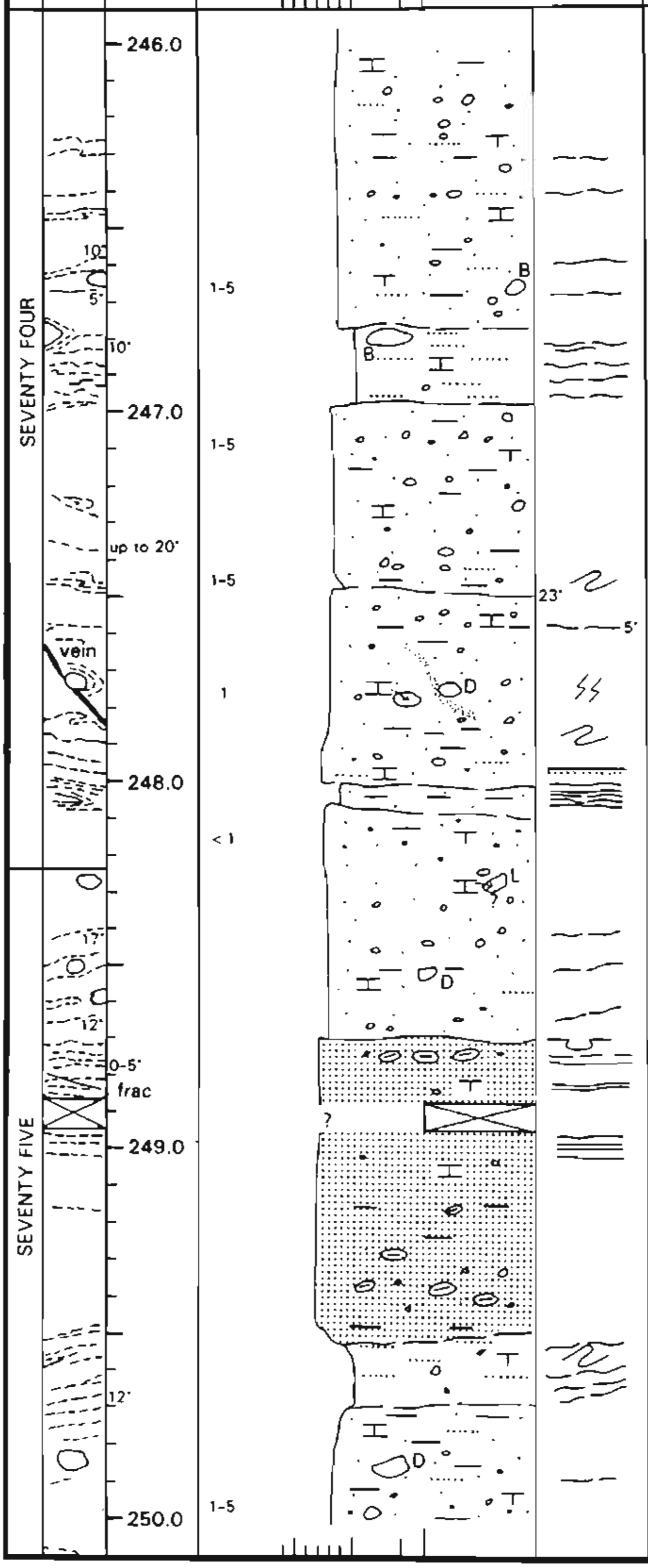
basement and Beacon sandstone and others less than 5%.
Thin beds of sandy mudstone and muddy very fine sandstone from 246.79 to 246.98 m and from 248.02 to 248.09 m respectively.

PROJECT : CIROS-1

SHEET NO. : 57

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 12.9

241.19 - 248.71 m

DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy medium to very fine sandstone (diamictite)** (dark greenish grey to greyish olive 10GY 4/1 to 7.5Y 6/2 and 2.5GY 5/1). Stratification is wispy and discontinuous, occasionally disrupted and commonly contorted, dipping up to 23°. Clasts form mainly between 5 and 10% of the core, averaging about 10 mm but reaching > 90 mm long. Most are subangular to subrounded but a few are well rounded. One is striated. About 70% are dolerite, 25% basement and Beacon sandstone and others less than 5%. Thin beds of sandy mudstone and muddy very fine sandstone from 246.79 to 246.98 m and from 248.02 to 248.09 m respectively.

— contact sharp -

Sub-unit 12.10

248.71 - 249.52 m

SANDSTONE

Hard, calcareous, weakly stratified (mm-bedded), poorly sorted, **muddy coarse to medium sandstone** (grey 7.5Y 5/1).
Stratification irregular, subhorizontal and inclined (5° maximum dip).
Scattered subrounded mudstone clasts near top and base.

— contact gradational -

Sub-unit 12.11

249.52 - 287.71 m

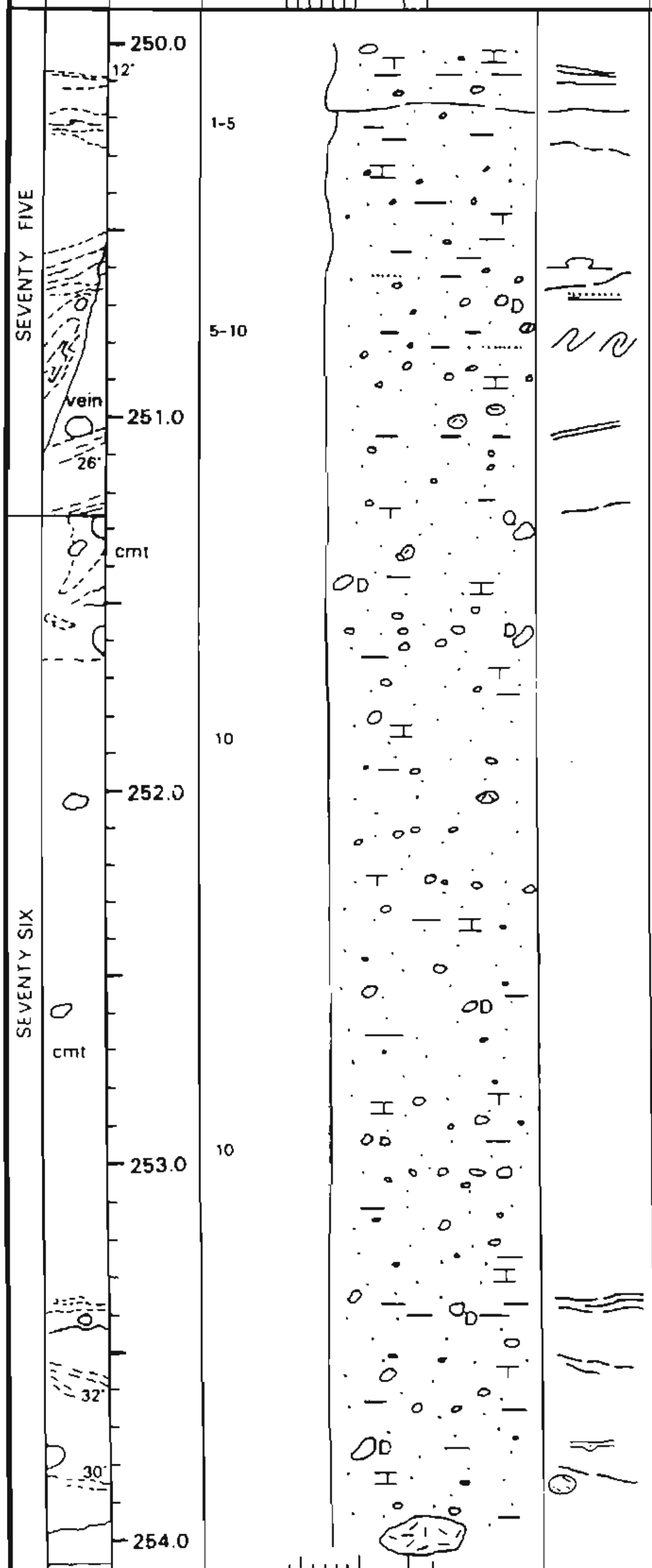
DIAMICTITE AND SANDSTONE

PROJECT : CIROS -1

SHEET NO. : 58

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub–unit 12.11

249.52 – 287.71 m

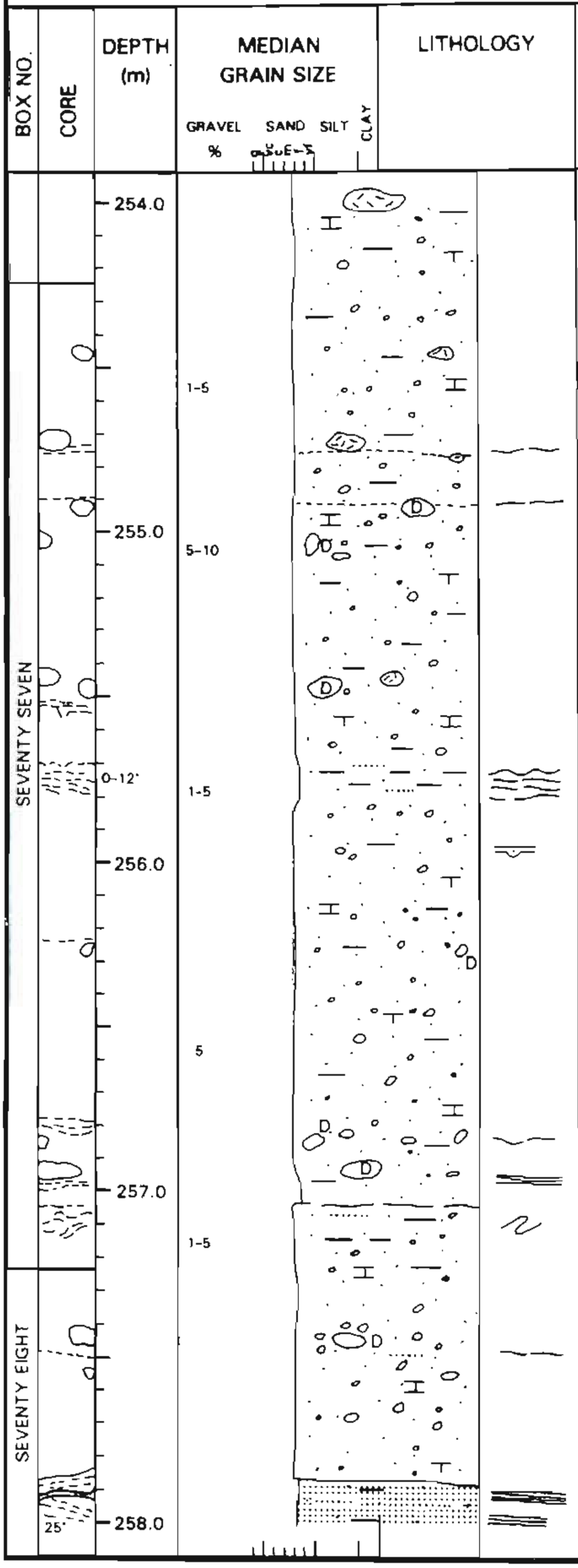
DIAMICTITE AND SANDSTONE

Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well–stratified (mm–bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic.** and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1) Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280 70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266 41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7 5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 59

SCALE : 1:20



DESCRIPTION

Sub-unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

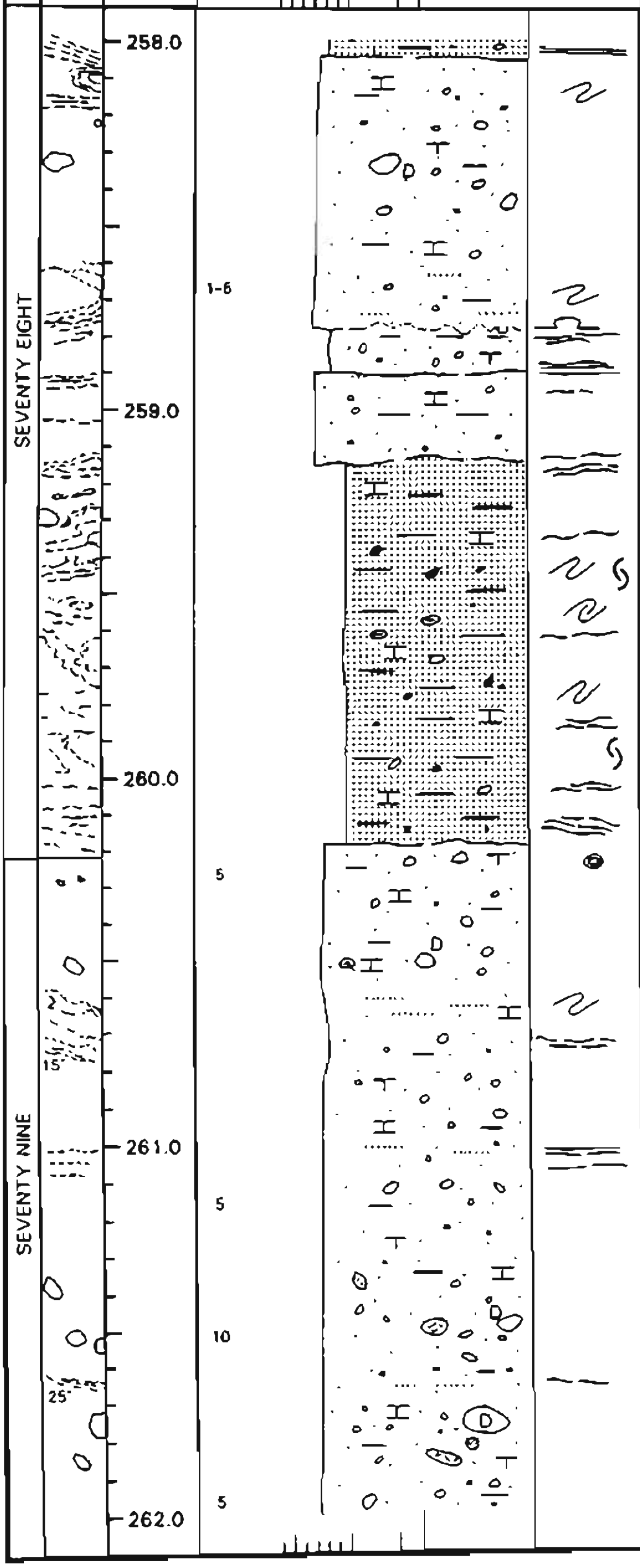
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic,** and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 60

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

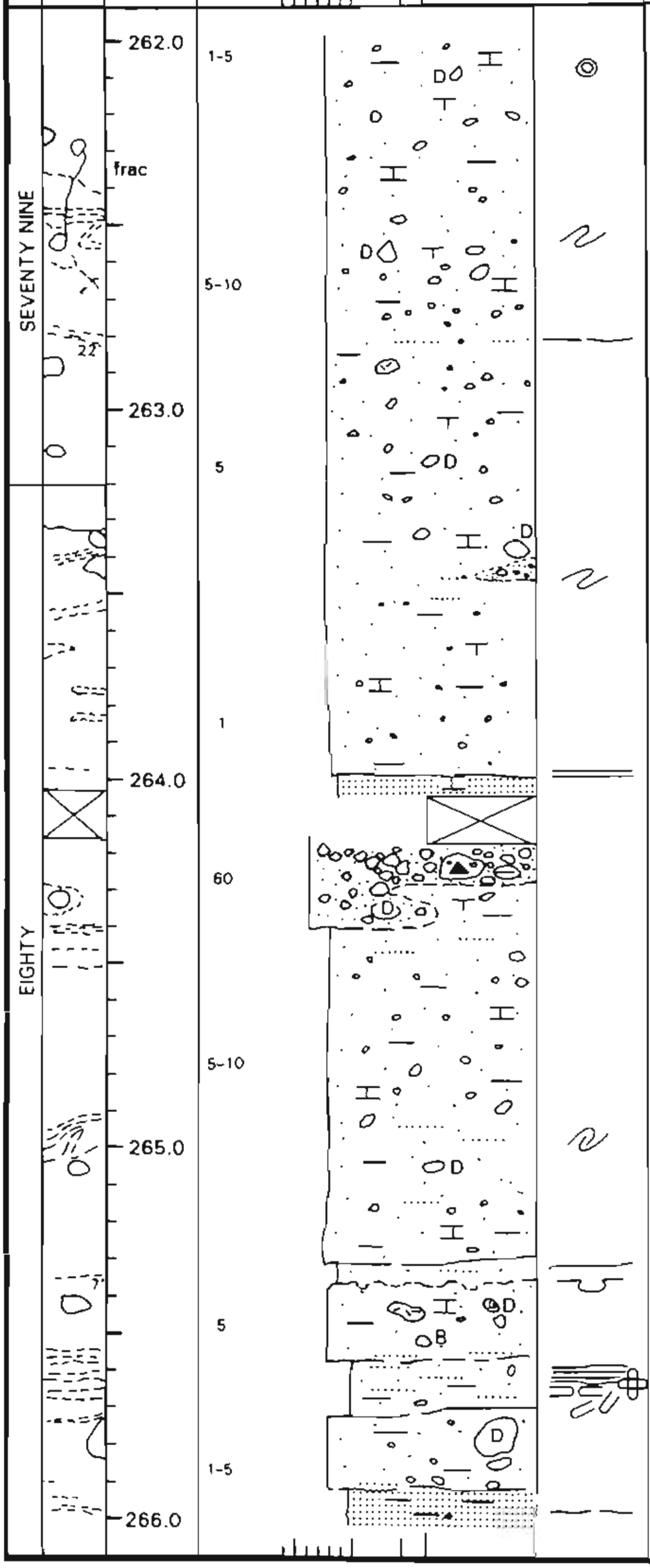
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone** (diamictite) (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic**, and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form rhythmites at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 61

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

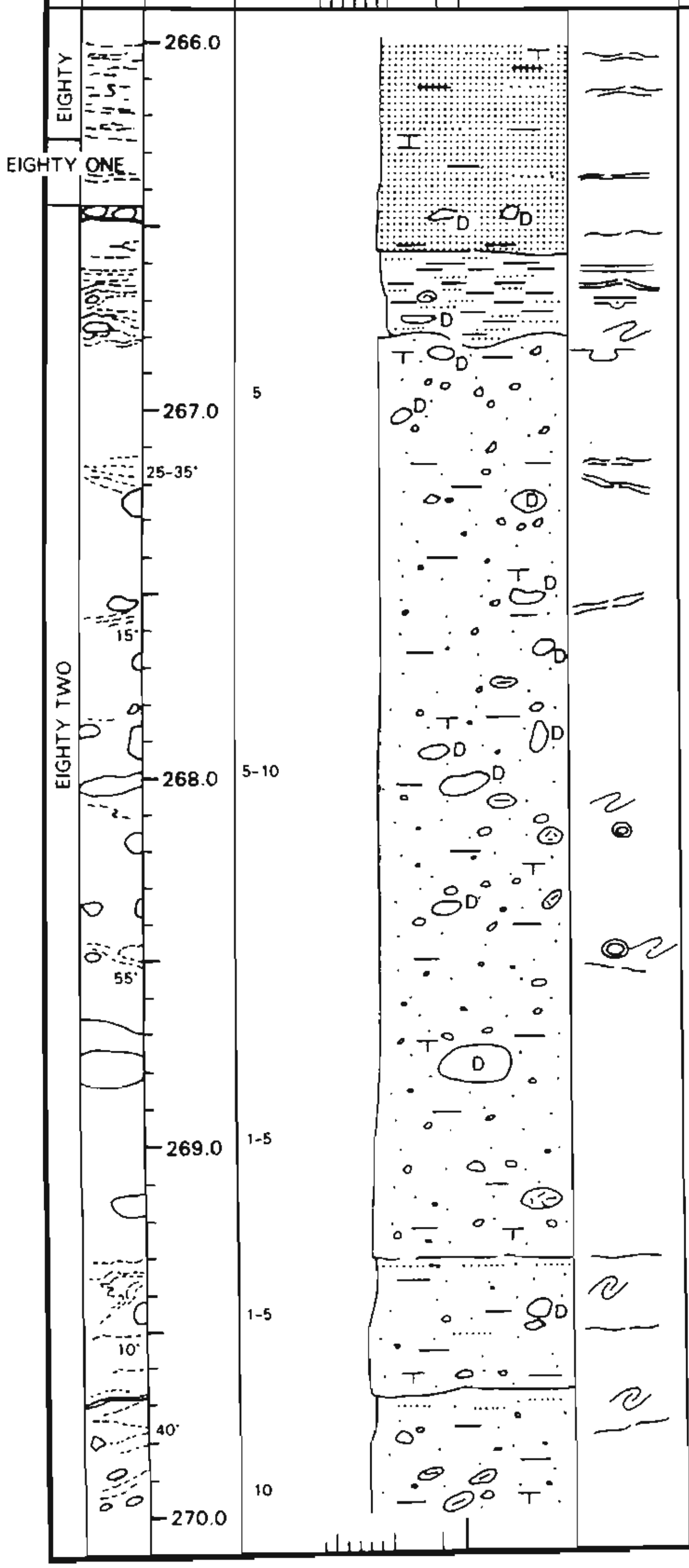
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic**, and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 62

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

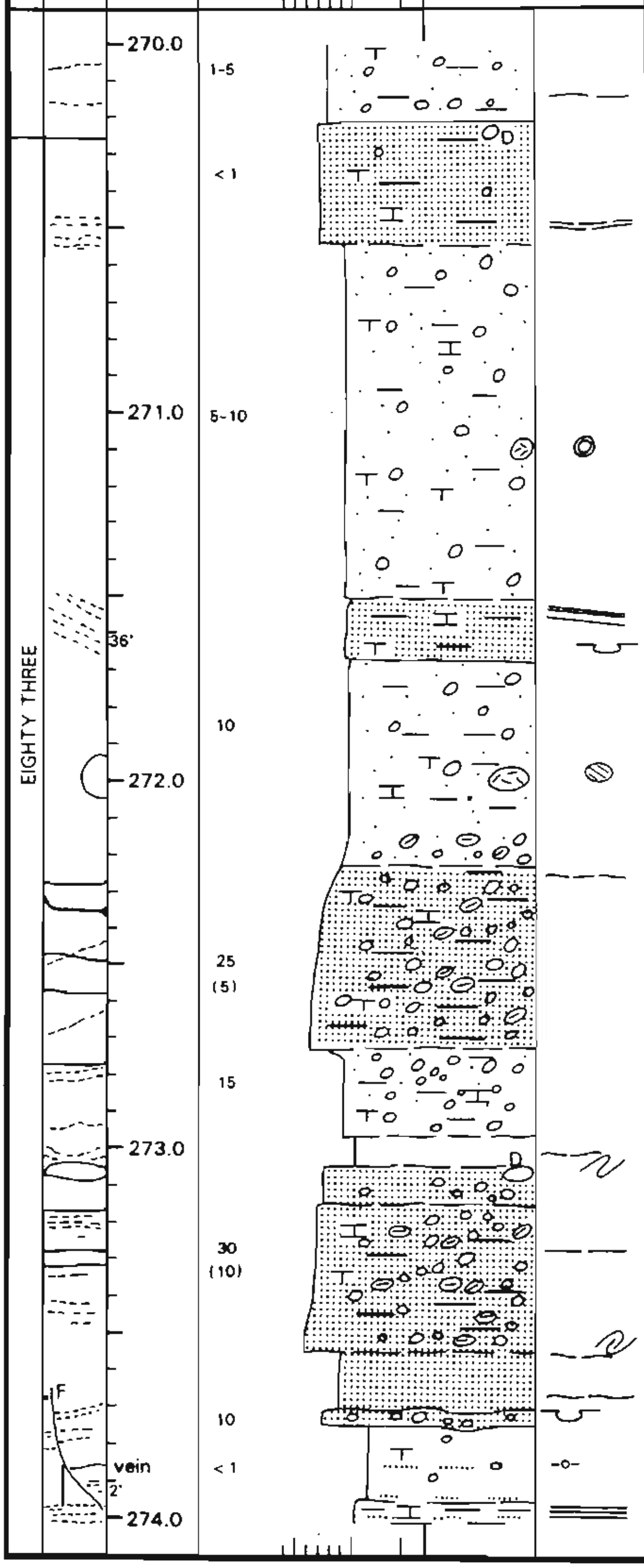
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic,** and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 63

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

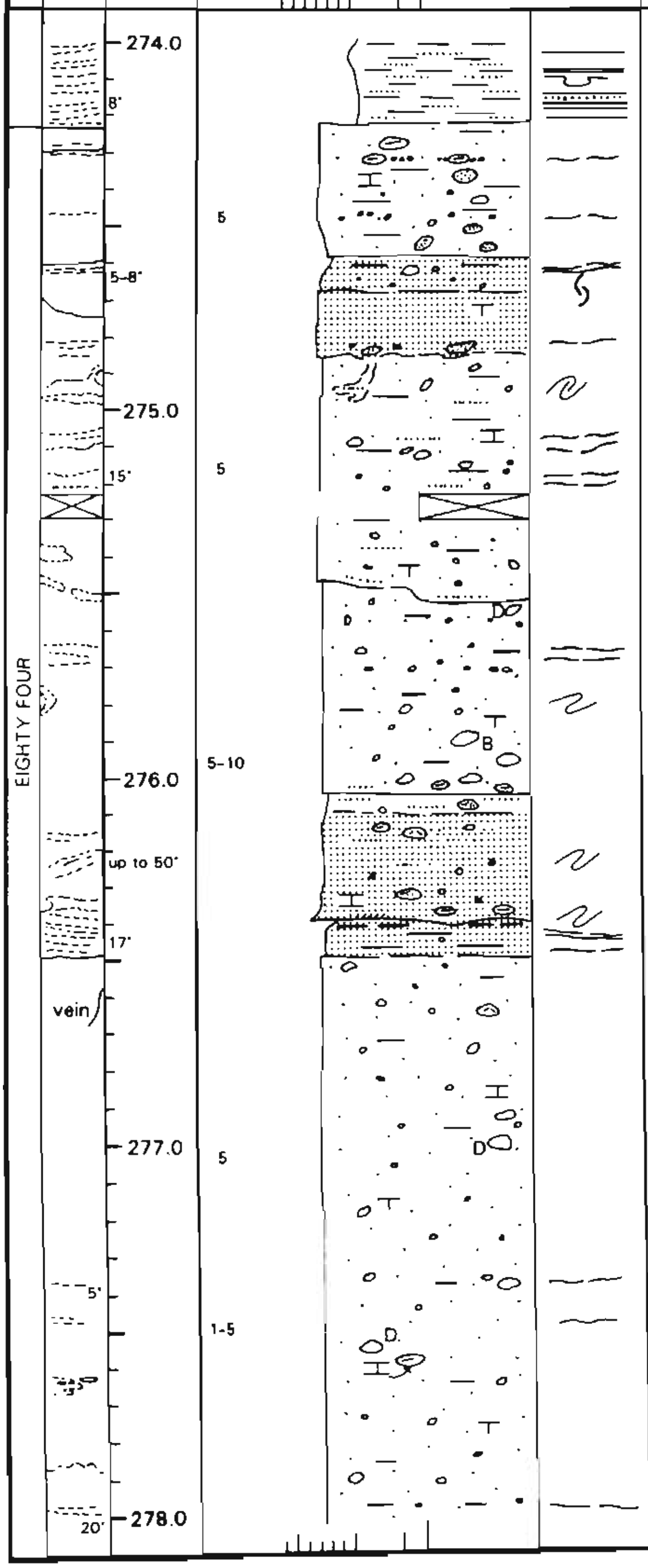
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic**, and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 64

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub–unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

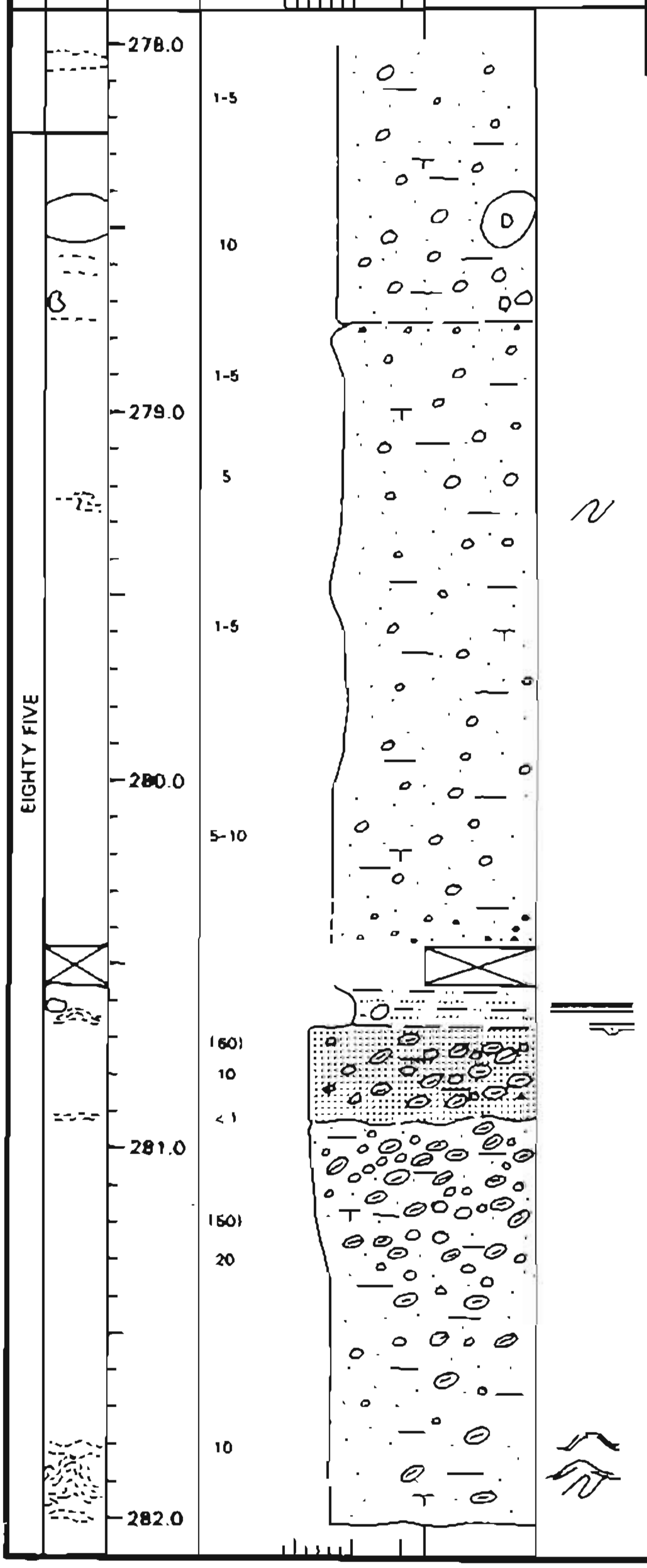
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm–bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic**, and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 65

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 12.11

249.52 - 287.71 m

DIAMICTITE AND SANDSTONE

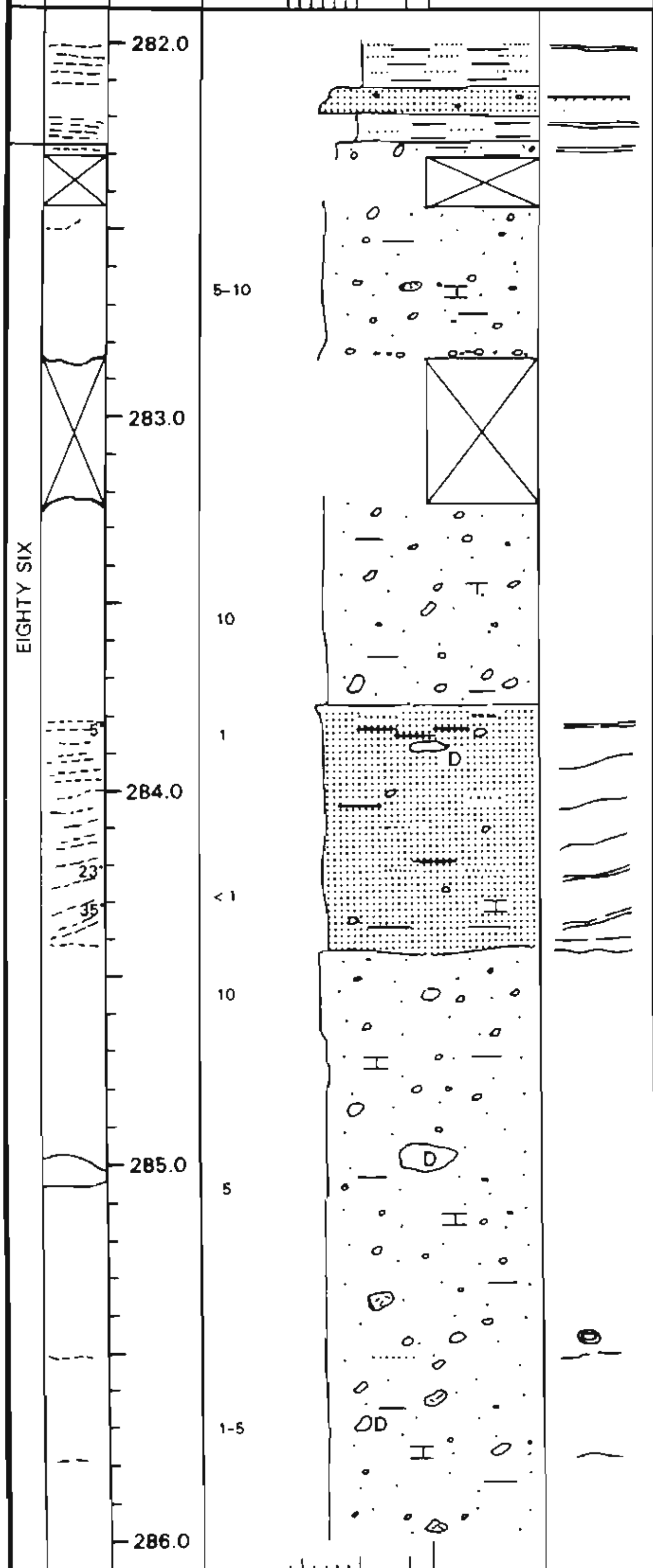
Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic**, and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 66

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 12.11

249.52 - 287.71 m

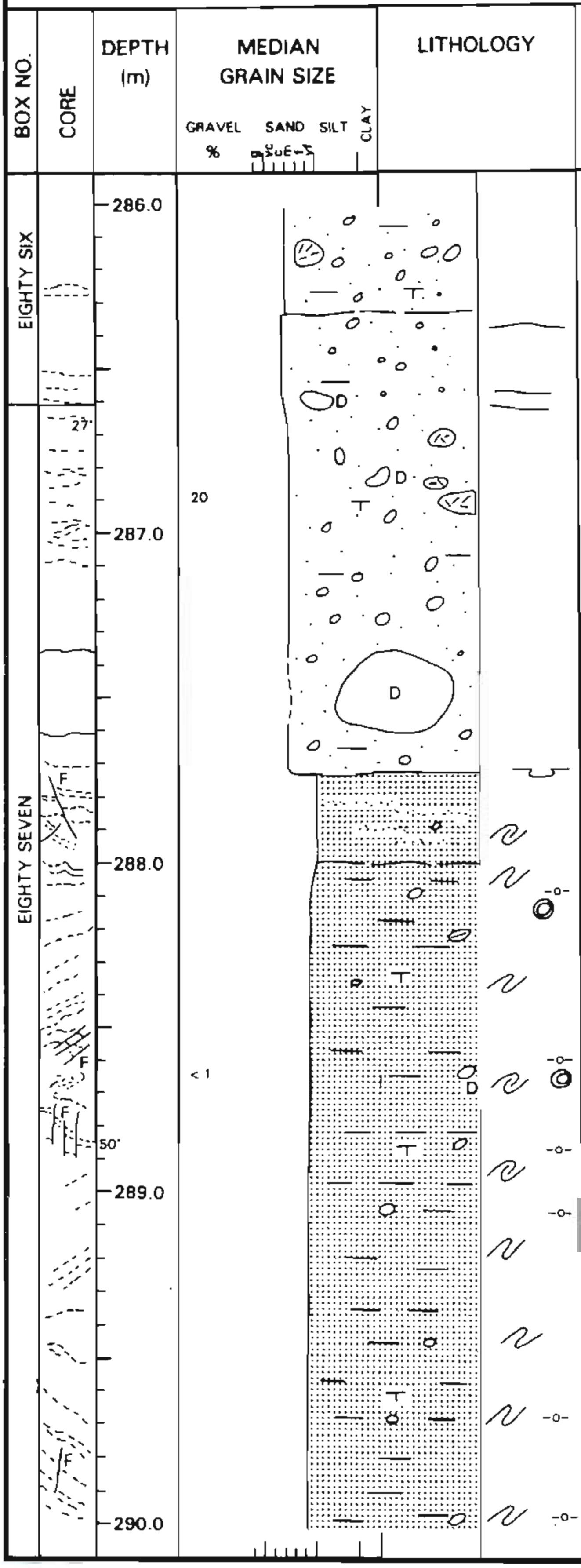
DIAMICTITE AND SANDSTONE

Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic**, and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

PROJECT : CIROS -1

SHEET NO. : 67

SCALE : 1:20



DESCRIPTION

DIAMICTITE AND SANDSTONE

Alternating diamictite and sandstone. Hard, calcareous, unstratified to well stratified, **conglomeratic muddy medium to fine sandstone (diamictite)** (grey 10Y 5/1 to 4/1 and dark greenish grey to greenish grey 7.5GY 4/1 to 5/1). Stratification varies through subhorizontal, wavy, inclined (up to 55°), disrupted, convoluted and contorted. Occasional graded bedding and load casts. Clasts form between 2 and 10% of the core, most ranging from 5 to 12 mm, but with one dolerite > 110 mm. They range from subangular to rounded, averaging subrounded. Several are striated and faceted. Some weakly stratified diamictites contain sandstone and mudstone intraclasts, locally comprising 30% of the rock. Most are dolerite (60%) and basement (35 %) with some Beacon sandstone (< 5%). **Sandstone** ranges from weak to well-stratified (mm-bedded), **very fine, fine and medium, mostly muddy** and in places **slightly conglomeratic,** and mottled (greenish grey and grey 10GY 6/1 and 10Y 6/1). Stratification shows a similar range of features to that in the diamictites. Weakly stratified, **muddy very fine sandstone** (mottled greenish grey 10GY 6/1 and grey 10Y 6/1) from 259.17 to 260.20 m. Two **conglomeratic sandstone** beds between 272.30 and 273.50 m have up to 30% gravel, of which 60% of the clasts are intraformational and the remainder bedrock. There is also a sandy conglomerate at 280.70 m of which 90% of the clasts are intraformational. The sandstone is interstratified on a mm scale with mudstone to form **rhythmites** at 266.41 m (35 cm thick), 273.95 m (29 cm thick), 280.56 m (11 cm thick), 282.00 m (14 cm thick) and 282.20 m (12 cm thick). The rhythmites are greyish olive (7.5Y 6/2) and contain dropstones.

— contact sharp and loaded -

Sub-unit 13.1

287.71 - 290.75 m

SANDSTONE

Hard, calcareous, weakly to well stratified, **slightly conglomeratic muddy fine sandstone** (dark greenish grey 10Y 4/1).

Stratification defined by sandstone and mudstone laminae.

Dispersed dolerite, basement and volcanic clasts up to 25 mm long. They range from angular to subrounded and some are weathered.

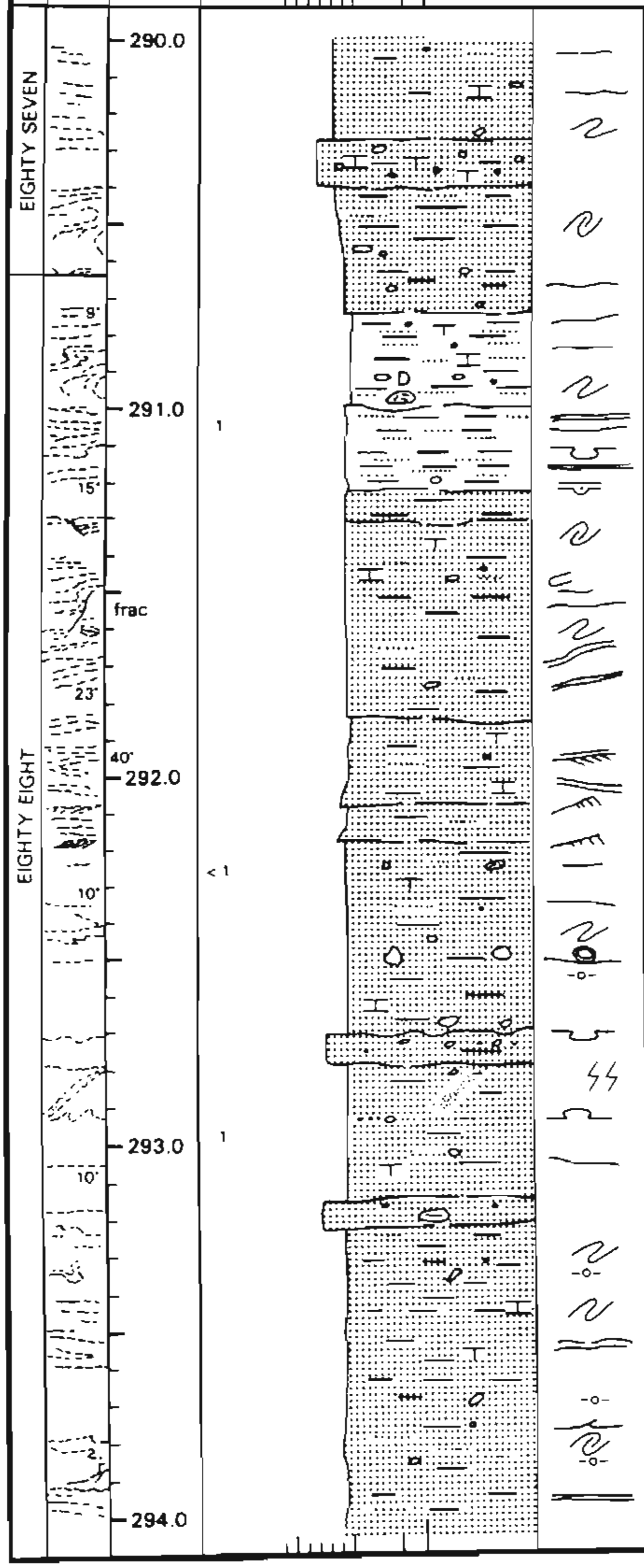
Much synsedimentary folding and faulting.

PROJECT : CIROS -1

SHEET NO. : 68

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 13.1

287.71 - 290.75 m

SANDSTONE

Hard, calcareous, weakly to well stratified, slightly **conglomeratic muddy fine sandstone** (dark greenish grey 10Y 4/1). Stratification defined by sandstone and mudstone laminae. Dispersed dolerite, basement and volcanic clasts up to 25 mm long. They range from angular to subrounded and some are weathered. Much synsedimentary folding and faulting.

—contact gradational -

Sub-unit 13.2

290.75 - 294.80 m

SANDSTONE AND MUDSTONE

Hard (but locally soft), calcareous, well stratified, **muddy fine sandstone** (dark olive grey 2.5GY 4/1), interbedded on mm- to cm-scale with **sandy mudstone** (dark greenish grey 7.5GY 4/1)

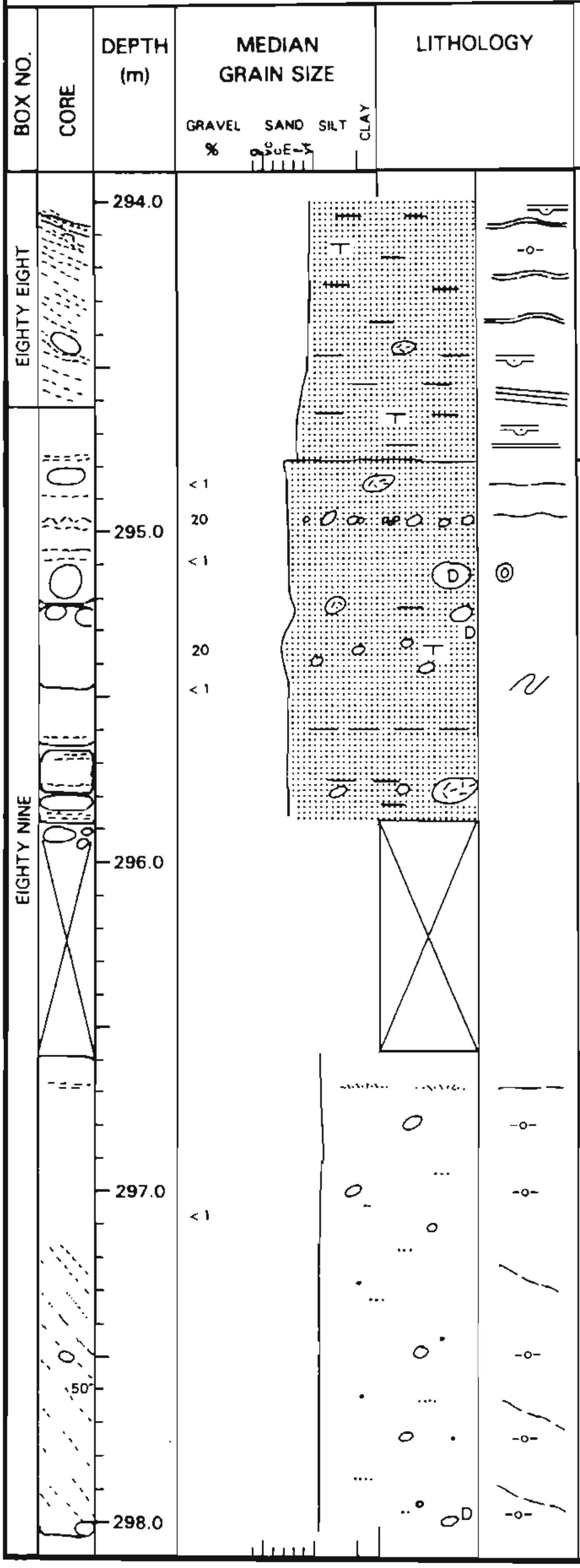
Laminae are rhythmic to lenticular with ripple lamination in thinly bedded interval (290.75 to 292.20 m). Synsedimentary deformation includes load structures and variable degrees of contorted strata. Synsedimentary sandstone dyke at 292.90 m. The middle of the sub-unit is less well stratified and sandier, with beds up to several cm thick

Dispersed clasts, including some dropstones, form <1% of the core. Most are 10 to 20 mm but they range to >50 mm long. They are subangular to subrounded and a few are weathered.

PROJECT : CIROS -1

SHEET NO. : 69

SCALE : 1:20



DESCRIPTION

— contact sharp -

Sub-unit 13.3

294.80 - ?295.89 m

SANDSTONE

Moderately hard to soft, weakly stratified, poorly sorted, **conglomeratic muddy medium sandstone** (dark olive grey 2.5GY 4/1).
Stratification is very irregular and contorted.
Clasts represent up to 20% of the core. Most are subrounded and average 20 mm. Large clasts (> 100 mm) occur in diffuse concentrations of gravel of varied shape. Predominant lithologies are dolerite and basement.
Well sorted **sand** at 295.66 m. Similar material may have been washed out from interval of no recovery (295.89 to 296.60 m).

— contact in interval of no recovery -

Sub-unit 14.1

?296.60 - 300.30 m

MUDSTONE

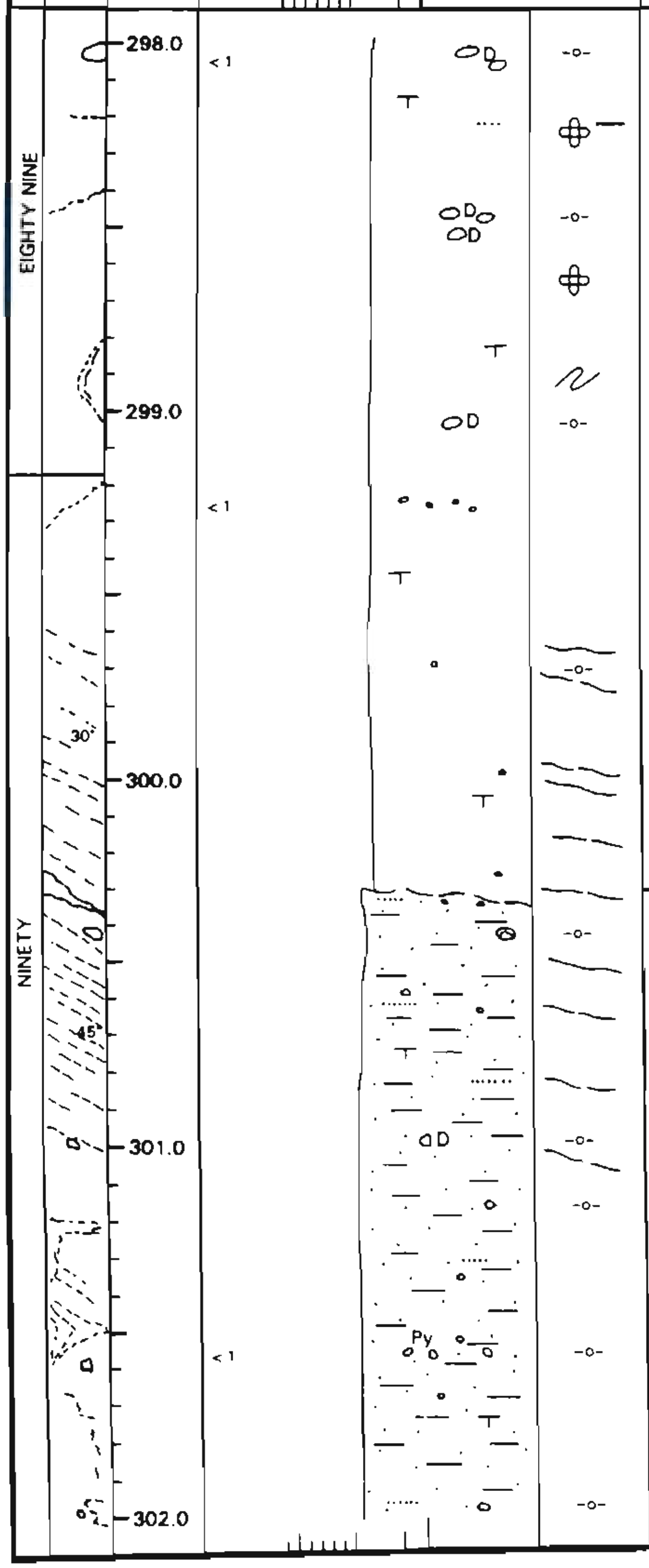
Hard, calcareous, unstratified to weakly stratifed, **slightly sandy mudstone** (dark greenish grey 7.5GY 4/1).
Stratification is wispy and irregular due to strong bioturbation.
Occasional lonestones mostly from 4 to 10 mm but up to 22 mm long.

PROJECT : CIROS -1

SHEET NO. : 70

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 14.1

?296.60 - 300.30 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratifed, slightly **sandy mudstone** (dark greenish grey 7.5GY 4/1). Stratification is wispy and irregular due to strong bioturbation. Occasional lonestones mostly from 4 to 10 mm but up to 22 mm long.

— contact gradational over 20 cm -

Sub-unit 14.2

300.30 - 309.75 mm

DIAMICTITE

Hard, slightly calcareous, unstratified and weakly stratified, slightly **conglomeratic sandy mudstone (diamictite)** (grey 10Y 4 1).
Stratification is irregular, discontinuous and subhorizontal to inclined (up to 45) from top of sub-unit to 305.50 m; below this the sub-unit is unstratified

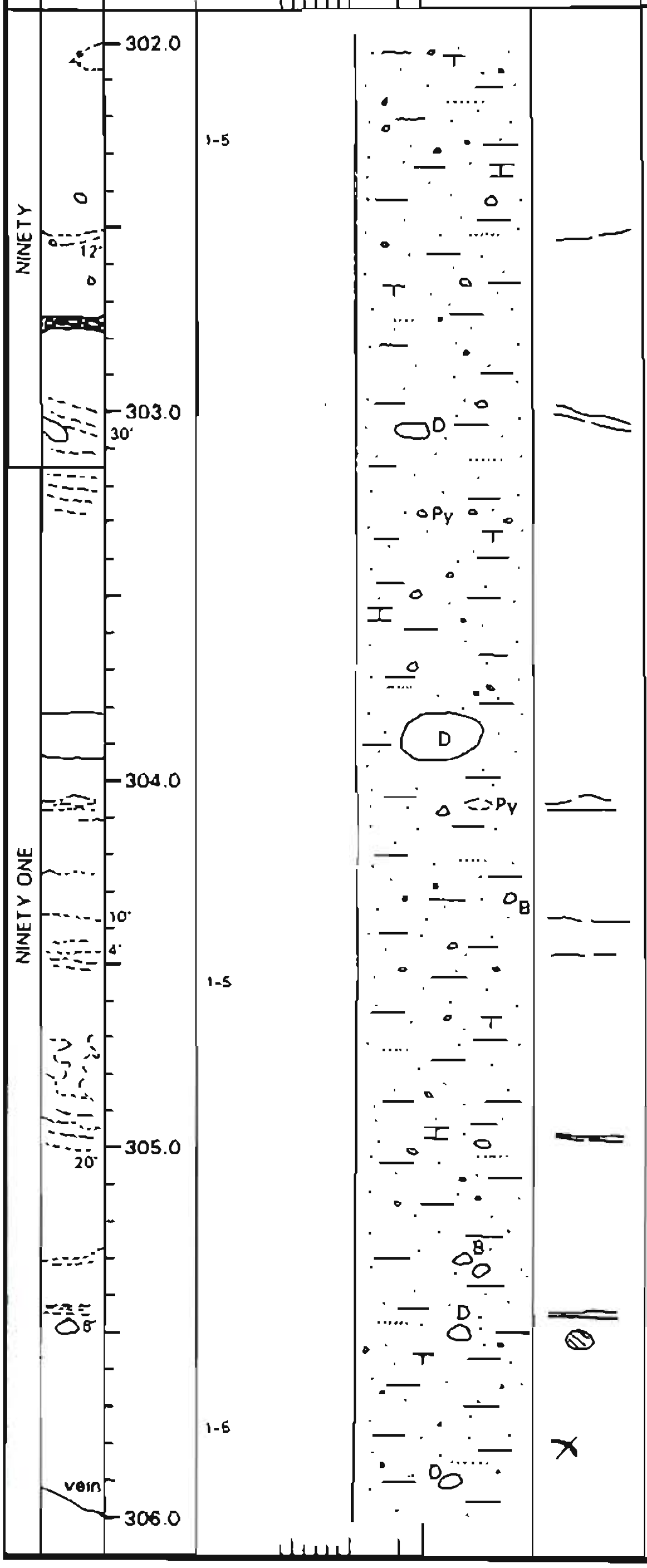
Clasts form 1 to 3% of the core, and are mostly between 4 and 10 mm across, though one dolerite clast is > 124 mm. Clasts range from subangular to rounded, and on average are subrounded. One is striated. Most are dolerite (65%), with some basement (30%) and Beacon sandstone (< 5%).

PROJECT : CIROS -1

SHEET NO. : 71

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 14.2

300.30 - 309.75 m

DIAMICTITE

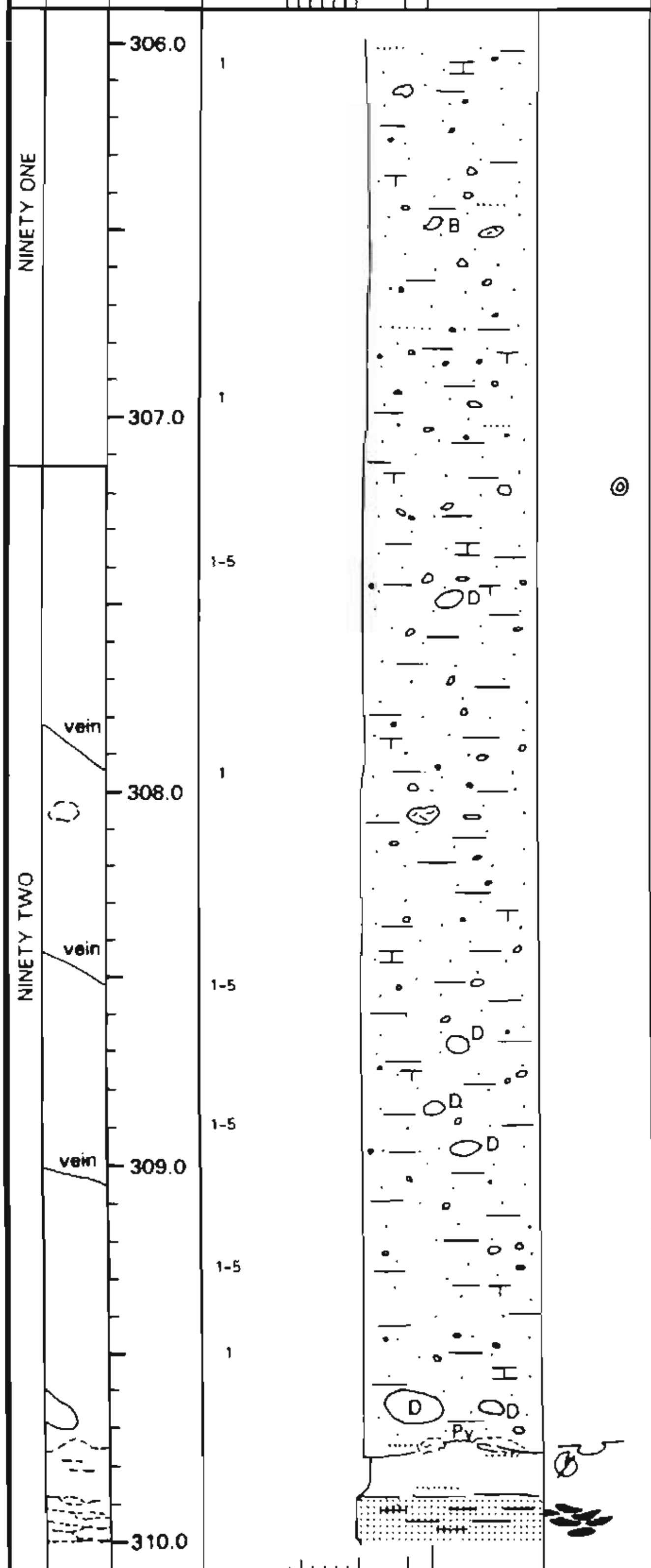
Hard, slightly calcareous, unstratified and weakly stratified, slightly **conglomeratic sandy mudstone (diamictite)** (grey 10Y 4/1). Stratification is irregular, discontinuous and subhorizontal to inclined (up to 45°) from top of sub-unit to 305.50 m, below this the sub-unit is unstratified. Clasts form 1 to 3% of the core, and are mostly between 4 and 10 mm across, though one dolerite clast is > 124 mm. Clasts range from subangular to rounded, and on average are subrounded. One is striated. Most are dolerite (65%), with some basement (30%) and Beacon sandstone (< 5%).

PROJECT : CIROS -1

SHEET NO. : 72

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 14.2

300.30 - 309.75 m

DIAMICTITE

Hard, slightly calcareous, unstratified and weakly stratified, **slightly conglomeratic sandy mudstone (diamictite)** (grey 10Y 4/1). Stratification is irregular, discontinuous and subhorizontal to inclined (up to 45°) from top of sub-unit to 305.50 m; below this the sub-unit is unstratified. Clasts form 1 to 3% of the core, and are mostly between 4 and 10 mm across, though one dolerite clast is > 124 mm. Clasts range from subangular to rounded, and on average are subrounded. One is striated. Most are dolerite (65%), with some basement (30%) and Beacon sandstone (< 5%)

—contact sharp, wavy and loaded -

Sub-unit 15.1

309.75 - 310.72 m

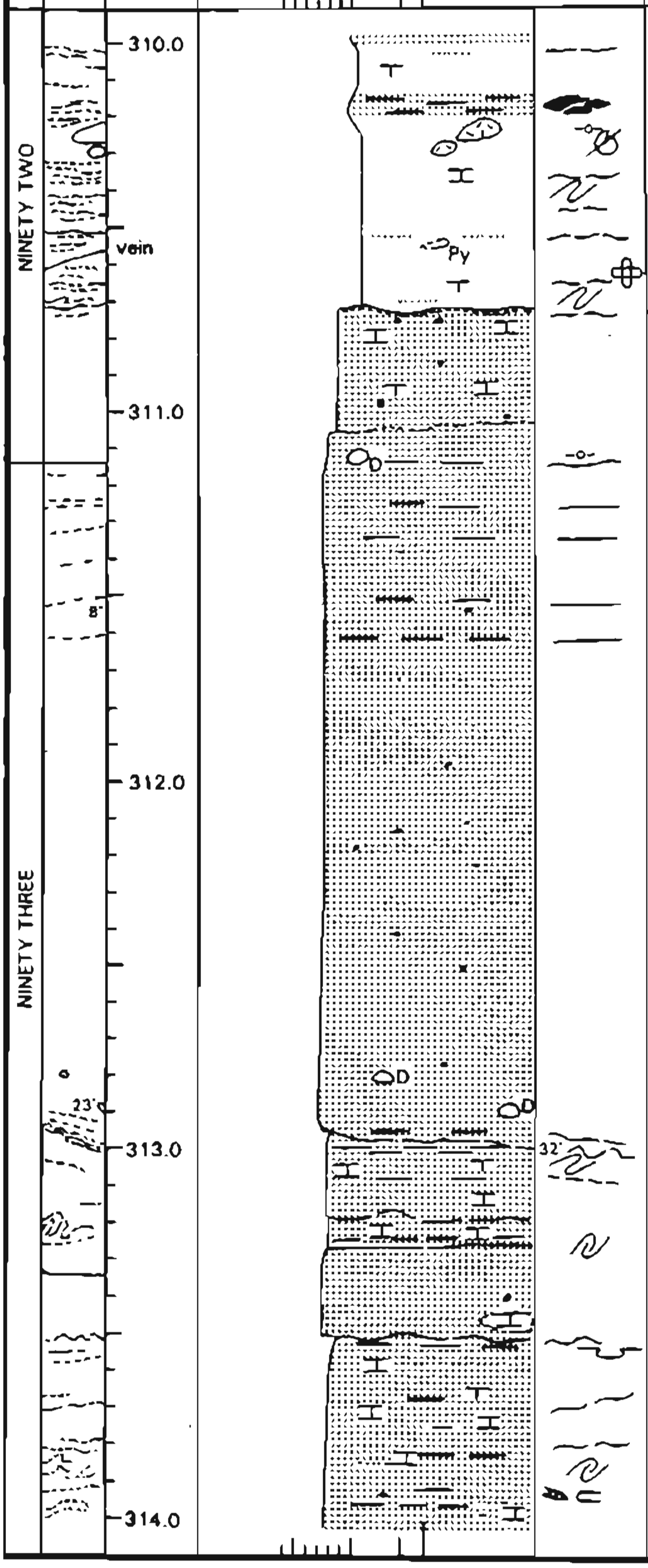
MUDSTONE

PROJECT : CIROS-1

SHEET NO. : 73

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Hard, calcareous, weakly to well stratified (mm- and cm-bedded), **sandy mudstone** (dark greenish grey 10GY 4/1).
Stratification irregular, wispy and wavy, locally contorted and probably disrupted by bioturbation.
Only two basement lonestones, >29 and >47 mm long.
Beds from 309.89 to 309.98 m and 310.18 to 310.23 m of **muddy very fine sandstone** with irregular **mudstone** lenses. Shell fragment at 310.82 m.

—sharp contact -

Sub-unit 15.2

310.72 - 325.97 m

SANDSTONE

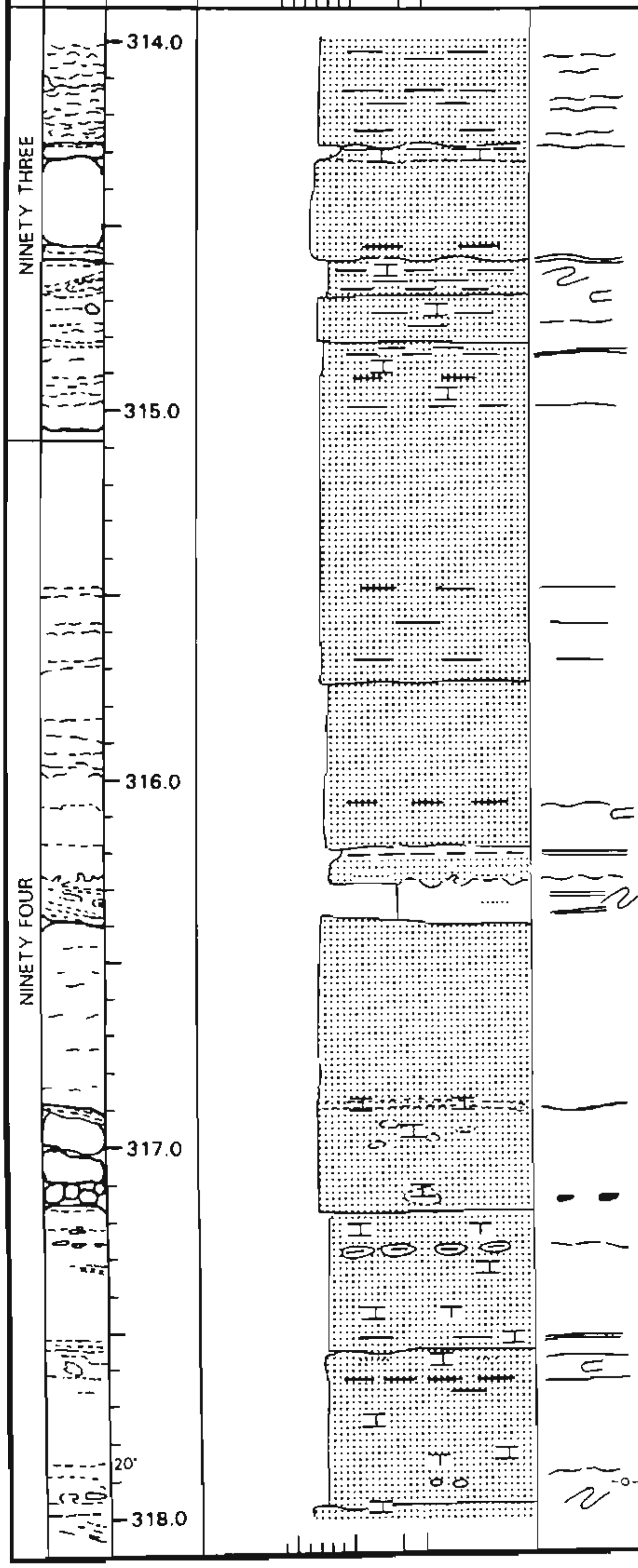
Hard, calcareous, well stratified (mm- and cm-bedded), **muddy fine to very fine sandstone** (grey 7.5GY 6/1 to N 6/0 where cemented).
Interbeds of soft, unstratified to weakly stratified (cm-bedded), moderately well sorted, **medium sandstone** (greyish olive 7.5Y 6/2).
Stratification in hard beds is irregular, wavy and discontinuous. Most is subhorizontal with a little ripple lamination, but some intervals show bedding inclined up to 45° (possibly slumped), and others show isoclinal folding and shearing. Bioturbation structures are common below 321.0 m.

PROJECT : CIROS -1

SHEET NO. : 74

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 15.2

310.72 - 325.97 m

SANDSTONE

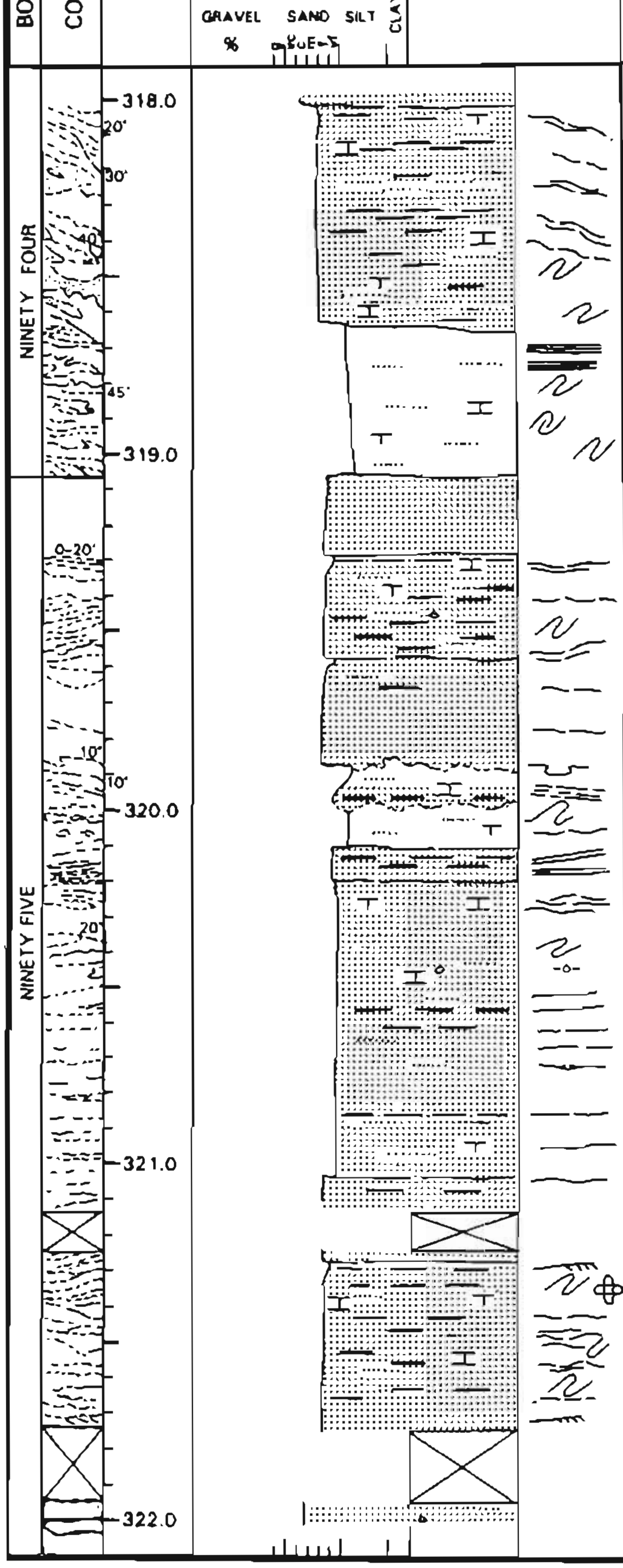
Hard, calcareous, well stratified (mm- and cm-bedded), **muddy fine to very fine sandstone** (grey 7.5GY 6/1 to N 6/0 where cemented). Interbeds of soft, unstratified to weakly stratified (cm-bedded), moderately well sorted **medium sandstone** (greyish olive 7.5Y 6/2). Stratification in hard beds is irregular, wavy and discontinuous. Most is subhorizontal with a little ripple lamination, but some intervals show bedding inclined up to 45° (possibly slumped), and others show isoclinal folding and shearing. Bioturbation structures are common below 321.0 m.

PROJECT : CIROS -1

SHEET NO. : 75

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 15.2

310.72 - 325.97 m

SANDSTONE

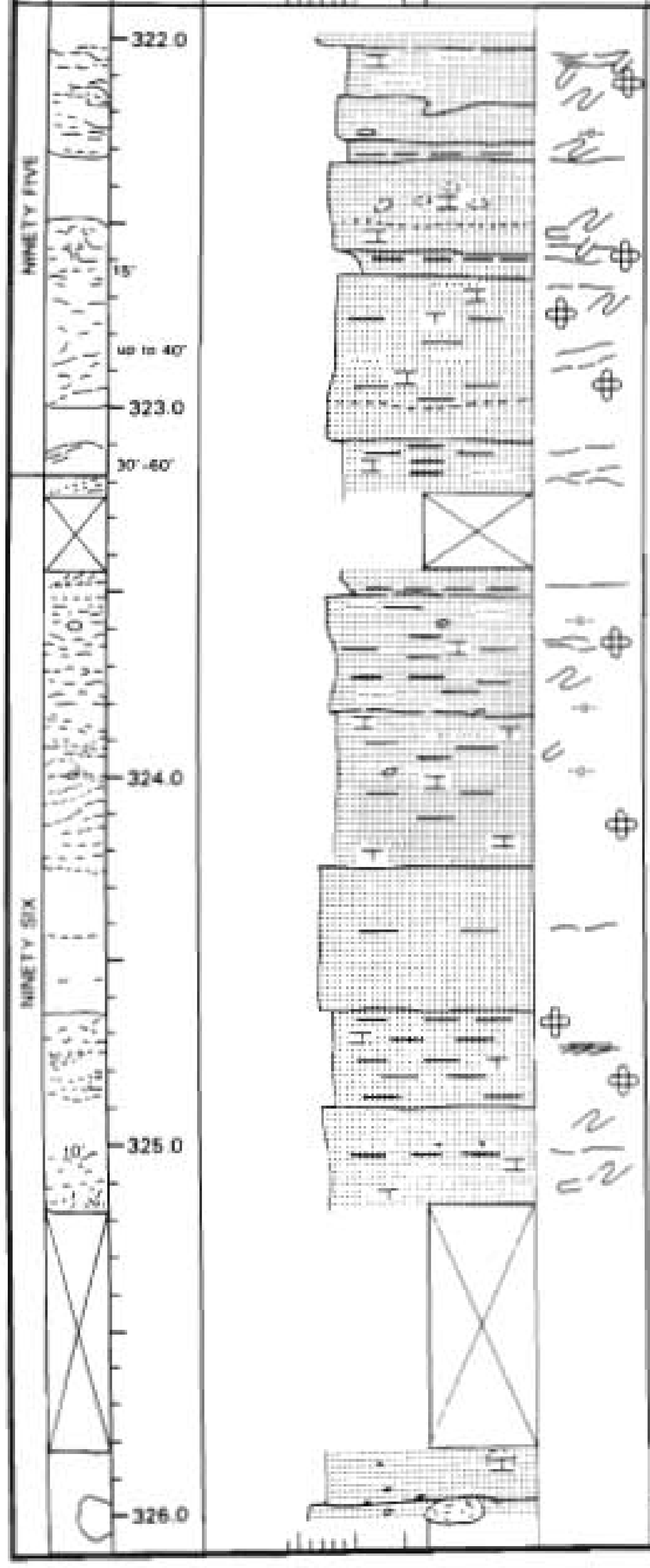
Hard, calcareous, well stratified (mm- and cm-bedded), muddy fine to very fine sandstone (grey 7.5GY 6/1 to N 6/0 where cemented). Interbeds of soft, unstratified to weakly stratified (cm-bedded), moderately well sorted medium sandstone (greyish olive 7.5Y 6/2). Stratification in hard beds is irregular, wavy and discontinuous. Most is subhorizontal with a little ripple lamination, but some intervals show bedding inclined up to 45° (possibly slumped), and others show isoclinal folding and shearing. Bioturbation structures are common below 321.0 m.

PROJECT : CIROS -1

SHEET NO. : 76

SCALE : 1:20

BOX NO	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 15.2

310.72 - 325.97 m

SANDSTONE

Hard, calcareous, well stratified (mm- and cm-bedded), **muddy fine to very fine sandstone** (grey 7.5GY 6/1 to N 6/0 where cemented). Interbeds of soft, unstratified to weakly stratified (cm-bedded), moderately well sorted **medium sandstone** (greyish olive 7.5Y 6/2). Stratification in hard beds is irregular, wavy and discontinuous. Most is subhorizontal with a little ripple lamination, but some intervals show bedding inclined up to 45° (possibly slumped), and others show isoclinal folding and shearing. Bioturbation structures are common below 321.0 m.

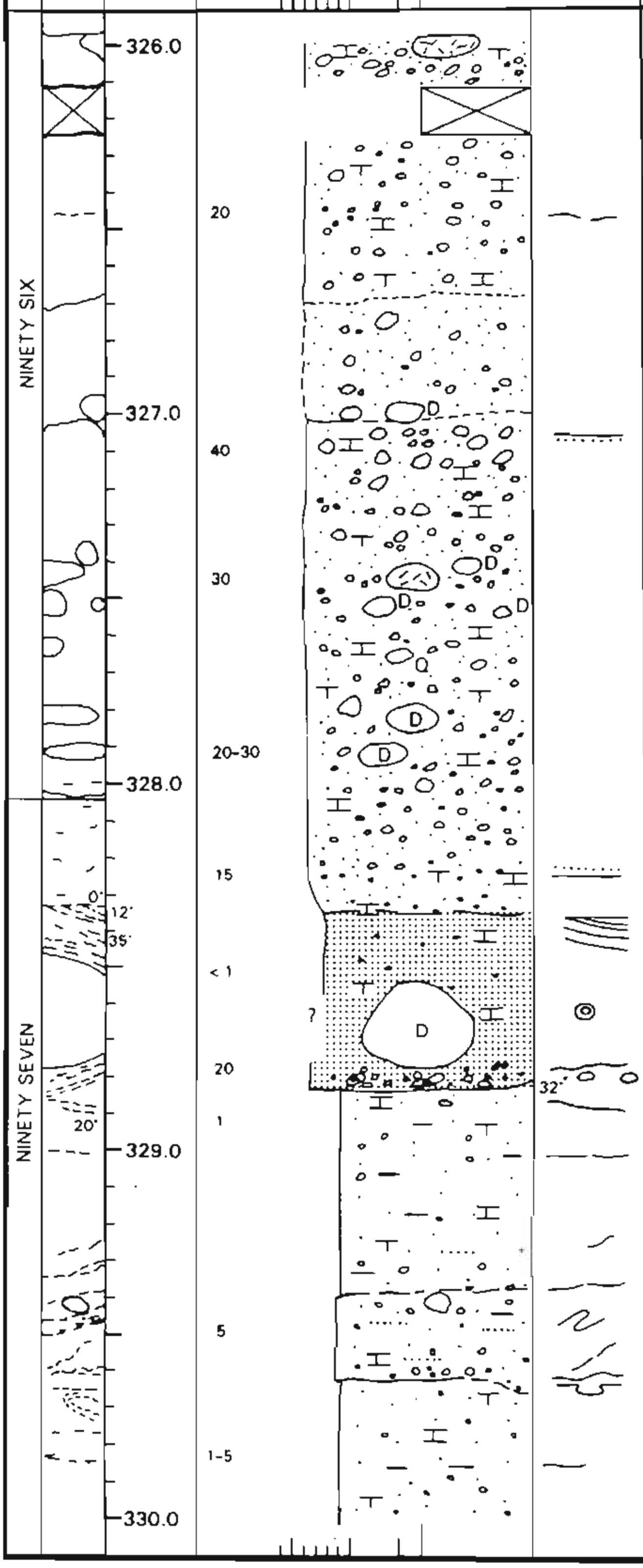
contact at top of gneiss clast -

PROJECT : CIROS -1

SHEET NO. : 77

SCALE : 1:20

BOX NO. | CORE | DEPTH (m) | MEDIAN GRAIN SIZE (GRAVEL % — SAND — SILT — CLAY) | LITHOLOGY | DESCRIPTION



Sub-unit 16.1

325.97 - 328.83 m

CONGLOMERATE

Hard (in places soft), calcareous, unstratified, very poorly sorted, **sandy pebble conglomerate.**

Matrix is poorly sorted, **very coarse to medium sandstone** (grey 7.5Y 6/1), with grains of similar composition to the clasts. Reverse grading occurs from 328.25 m up and normal grading from 327.40 m up.

Clasts are matrix-supported and range in size from small pebbles (4 mm) to boulders (>280 mm from 328.50 to 328.78 m) and are generally subrounded to rounded. Most are basement (60%), with 40% dolerite.

Dolerite boulder at 328.50 m has weathering rind on top surface but has fresh lower surface. Beneath it lies a gritty pebble conglomerate with weak clast long axis alignment subparallel to the base of the sub-unit.

— contact sharp and inclined at 32° -

Sub-unit 16.2

328.83 - 342.27 m

DIAMICTITE

Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy very fine sandstone (diamictite)** (greenish grey to grey 5G 5/1 to 10Y 4/1).

Stratification irregular and discontinuous and mainly subhorizontal, but locally inclined (up to 45°) and contorted. Load structure at 329.62 m.

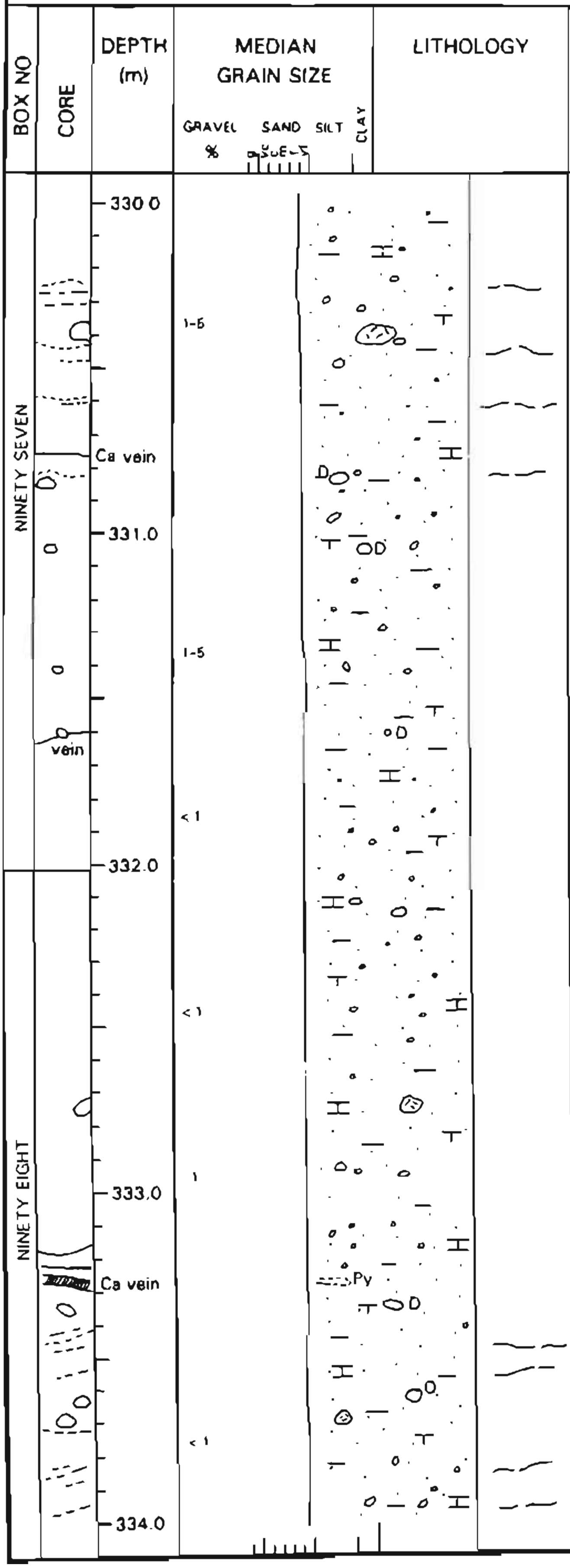
Clasts form on average about 1%, but locally up to 5%, of the core. Most are <6 mm, but one is >45 mm, long. They are generally subrounded, and range from angular to rounded. Basement and dolerite clasts are present in about equal proportions. A few have weathering rinds 1 to 2 mm thick. A clast at 335.74 m is a dropstone.

A few carbonaceous (peaty) and bioclastic fragments up to 5 mm across. Also some disseminated and nodular pyrite.

PROJECT : CIROS -1

SHEET NO. : 78

SCALE : 1:20



Sub-unit 16.2

328.83 - 342.27 m

DIAMICTITE

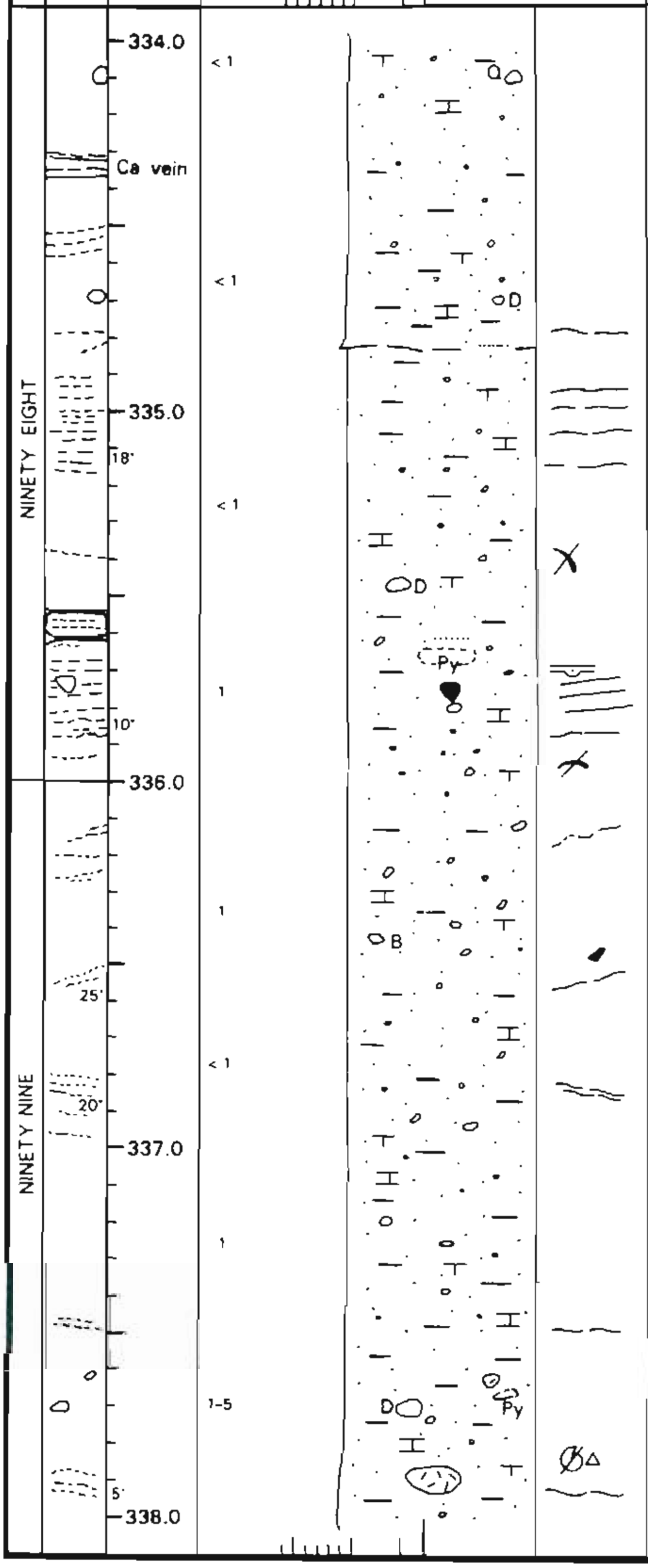
Hard, calcareous, unstratified to weakly stratified, conglomeratic muddy very fine sandstone (diamictite) (greenish grey to grey 5G 5/1 to 10Y 4/1). Stratification irregular and discontinuous and mainly subhorizontal, but locally inclined (up to 45°) and contorted. Load structure at 329.62 m. Clasts form on average about 1%, but locally up to 5%, of the core. Most are < 6 mm, but one is > 45 mm, long. They are generally subrounded, and range from angular to rounded. Basement and dolerite clasts are present in about equal proportions. A few have weathering rinds 1 to 2 mm thick. A clast at 335.74 m is a dropstone. A few carbonaceous (peaty) and bioclastic fragments up to 5 mm across. Also some disseminated and nodular pyrite.

PROJECT : CIROS -1

SHEET NO. : 79

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 16.2

328.83 - 342.27 m

DIAMICTITE

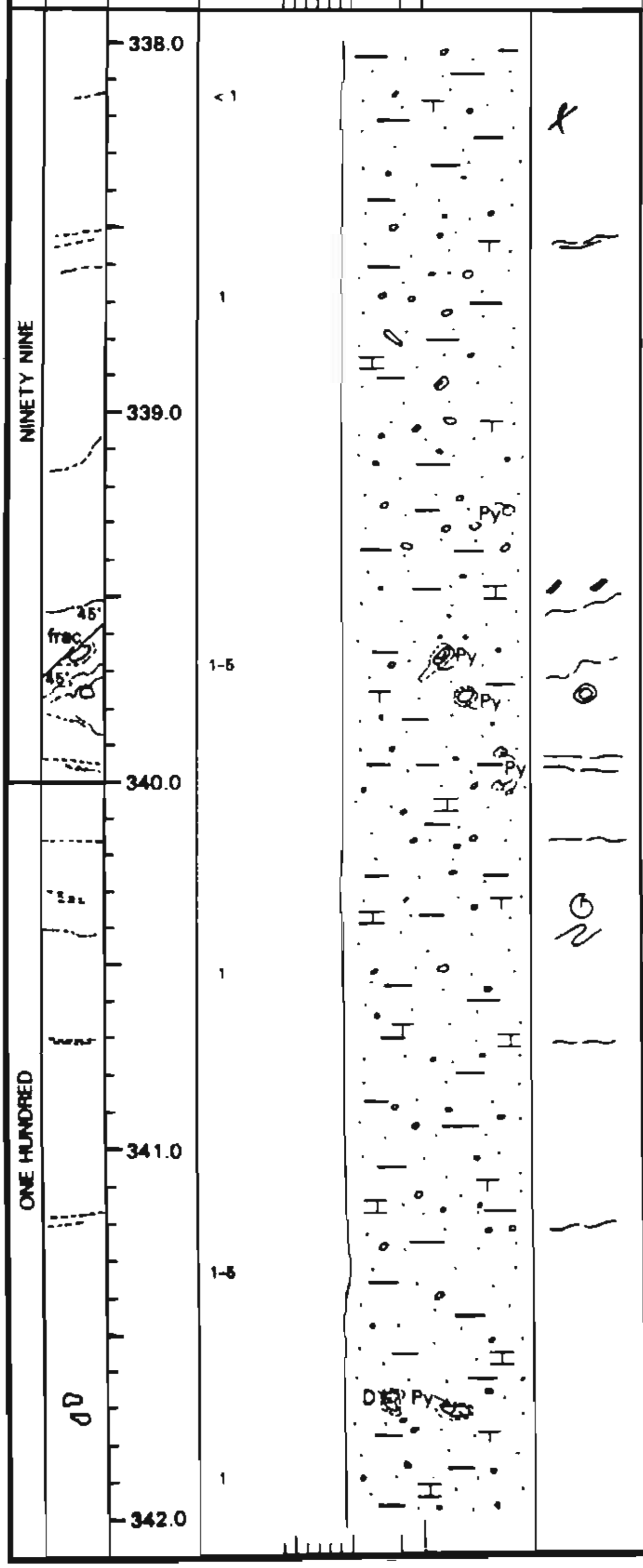
Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy very fine sandstone (diamictite)** (greenish grey to grey 5G 5/1 to 10Y 4/1). Stratification irregular and discontinuous and mainly subhorizontal, but locally inclined (up to 45°) and contorted. Load structure at 329.62 m. Clasts form on average about 1%, but locally up to 5%, of the core. Most are < 6 mm, but one is > 45 mm, long. They are generally subrounded, and range from angular to rounded. Basement and dolerite clasts are present in about equal proportions. A few have weathering rinds 1 to 2 mm thick. A clast at 335.74 m is a dropstone. A few carbonaceous (peaty) and bioclastic fragments up to 5 mm across. Also some disseminated and nodular pyrite.

PROJECT : CIROS -1

SHEET NO. : 80

SCALE : 1:20

BOX NO. | CORE | DEPTH (m) | MEDIAN GRAIN SIZE | LITHOLOGY | DESCRIPTION



Sub–unit 16.2

328.83 - 342.27 m

DIAMICTITE

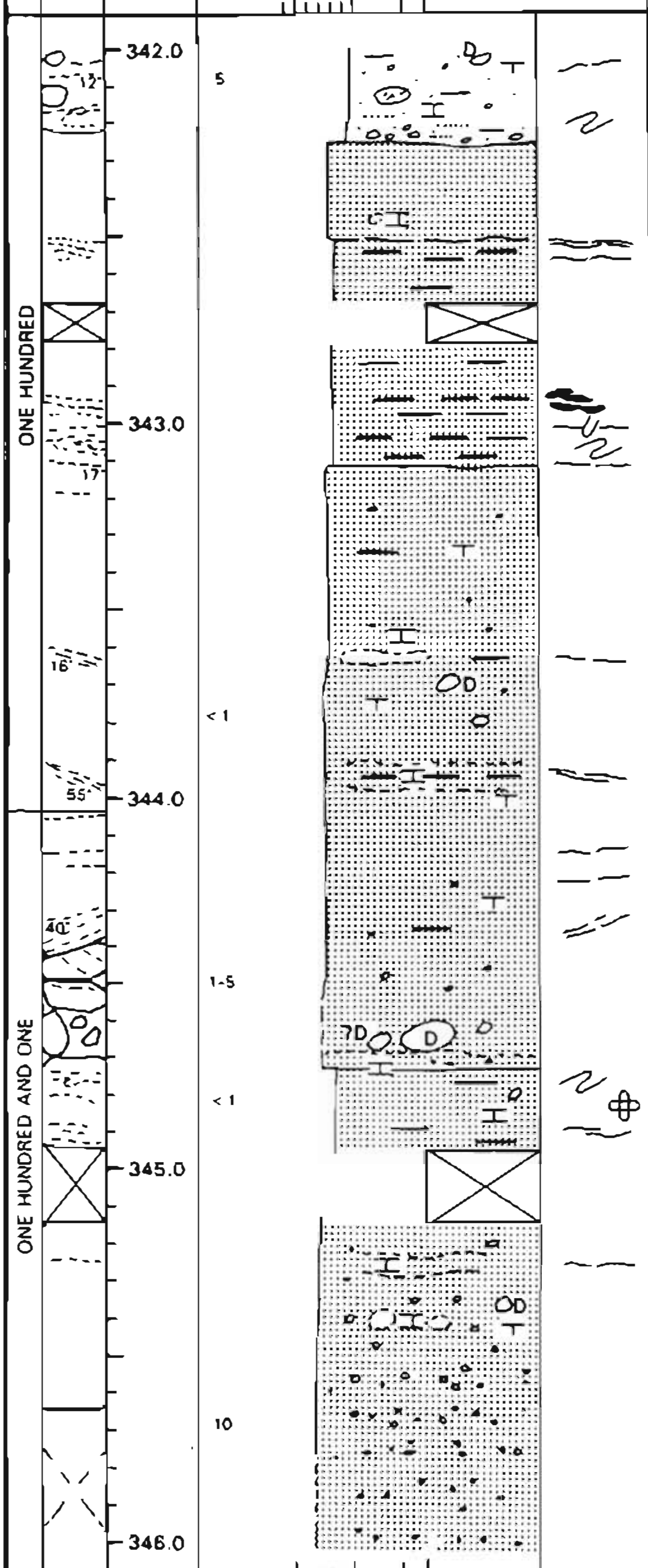
Hard, calcareous, unstratified to weakly stratified, **conglomeratic muddy very fine sandstone (diamictite)** (greenish grey to grey 5G 5/1 to 10Y 4/1). Stratification irregular and discontinuous and mainly subhorizontal, but locally inclined (up to 45°) and contorted. Load structure at 329.62 m. Clasts form on average about 1%, but locally up to 5%, of the core. Most are < 6 mm, but one is > 45 mm, long. They are generally subrounded, and range from angular to rounded. Basement and dolerite clasts are present in about equal proportions. A few have weathering rinds 1 to 2 mm thick. A clast at 335.74 m is a dropstone. A few carbonaceous (peaty) and bioclastic fragments up to 5 mm across. Also some disseminated and nodular pyrite.

PROJECT : CIROS -1

SHEET NO. : 81

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



— sharp contact -

Sub-unit 17.1

342.27 - 347.39 m

SANDSTONE

Moderately hard (in places soft), variably calcareous, unstratified to weakly stratified, moderately to poorly sorted, **coarse to fine sandstone** (greyish olive 7.5Y 5/2).

Stratification is generally irregular and discontinuous with dips up to 55 and occasional contorted beds.

Clasts usually form <1% of the rock (except from 345.5 to 346.08 m, where small pebbles form 10% of the rock). One clast is >110 m long

Basement and dolerite form about equal proportions

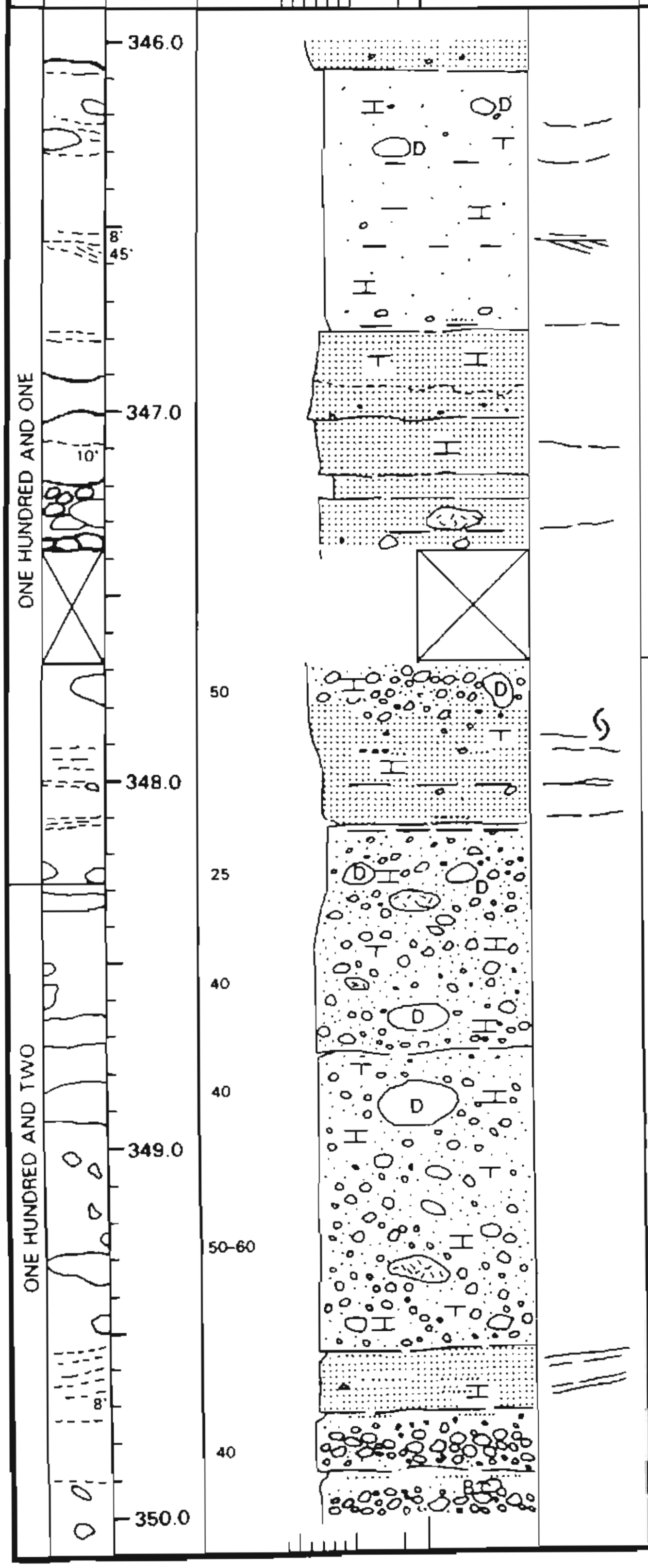
Burrows and contorted bedding from 342.91 to 343.09 m

PROJECT : CIROS -1

SHEET NO. : 82

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 17.1

342.27 - 347.39 m

SANDSTONE

Moderately hard (in places soft), variably calcareous, unstratified to weakly stratified, moderately to poorly sorted, **coarse to fine sandstone** (greyish olive 7.5Y 5/2). Stratification is generally irregular and discontinuous with dips up to 55° and occasional contorted beds. Clasts usually form < 1% of the rock (except from 345.5 to 346.08 m, where small pebbles form 10% of the rock). One clast is > 110 m long. Basement and dolerite form about equal proportions. Burrows and contorted bedding from 342 91 to 343.09 m.

— contact unknown due to core loss –

Sub-unit 17.2

347.69 - 351.34 m

CONGLOMERATE

Hard, very calcareous, weakly stratified very poorly sorted, **sandy pebble conglomerate.** Dispersed framework in matrix of coarse to medium sandstone (grey 7.5Y 6 1) and grit.

Stratification is irregular, discontinuous and subhorizontal (up to 8)

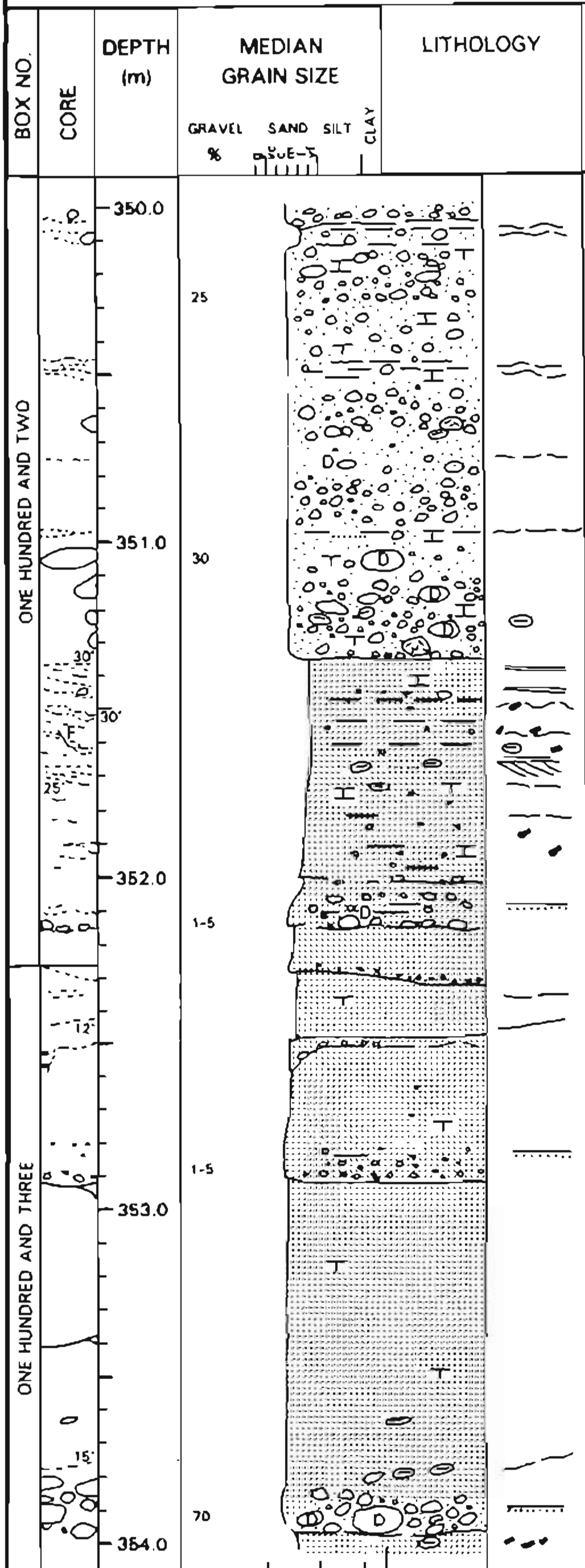
Clasts generally form about half of the core. Most are 12 to 18 mm, but range up to 114 mm long. On average they are subangular. About 40% are basement, 55% dolerite with 5% diamictite and mudstone intraclasts.

Beds of hard, calcareous, weakly stratified, **gritty medium sandstone** from 347 88 to 348.15 m and 349.55 to 349.72 m.

PROJECT : CIROS -1

SHEET NO. : 83

SCALE : 1:20



DESCRIPTION

Sub-unit 17.2

347.69 - 351.34 m

CONGLOMERATE

Hard, very calcareous, weakly stratifed, very poorly soned, **sandy pebble conglomerate**. Dispersed framework in matrix of coarse to medium **sandstone** (grey 7 5Y 6/1) and grit. Stratification is irregular, discontinuous and subhorizontal (up to 8°). Clasts generally form about half of the core Most are 12 to 18 mm, but range up to 114 mm long. On average they are subangular About 40% are basement, 55% dolerite with 5% diamictite and mudstone intraclasts. Beds of hard, calcareous, weakly stratified, gritty medium **sandstone** from 347 88 to 348 15 m and 349.55 to 349.72 m

— contact sharp and possibly erosional -

Sub-unit 17.3

351.34 - 365.85 m

SANDSTONE

Hard and soft (varying with degrees of carbonate cementation), unstratified to weakly stratified, moderately sorted to poorly sorted, **coarse to very fine sandstone** (grey to greyish olive 10Y 5/1 to 5Y 5/2).

Stratification is irregular and discontinuous, and ranges from subhorizontal to inclinations of 30 .

Thin interbeds of weakly stratifed, **sandy mudstone** (dark greenish grey 5G 4 1) at 360.15, 363.52 and 363.88 m Hard and soft matrix-supported **sandy pebble conglomerate** occurs in irregular beds through the sub-unit.

Some conglomerates fine upward into sandstone.

Clasts mostly range from 8 to 12 mm, but a few are >60 mm and one dolerite is >230 mm Most are subrounded About 50% are basement, 40% dolerite and the remainder are mudstone intraclasts

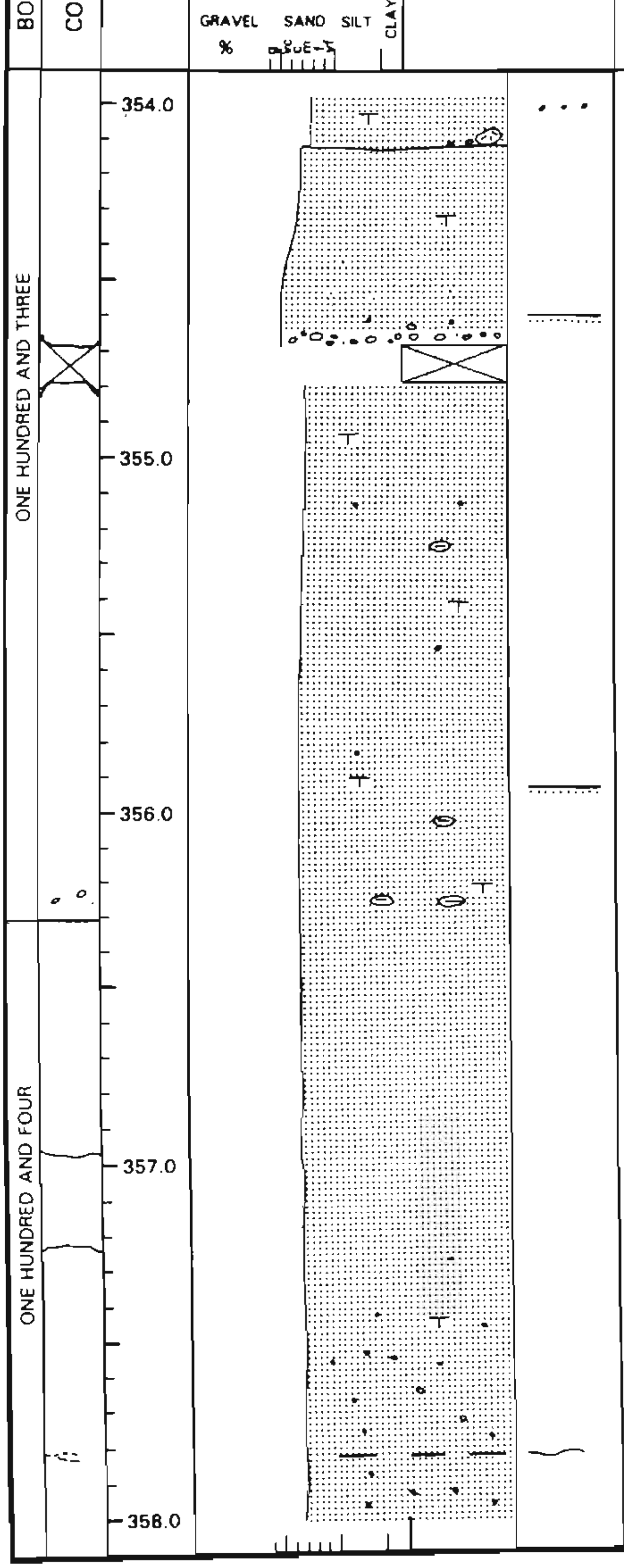
Carbonaceous fragments were found between 351 49 and 351.95 and at 353 98 m The interval from 363 31 to 365 85 m is strongly calcareous and bluish grey (5PB 5/1)

PROJECT : CIROS -1

SHEET NO. : 84

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub–unit 17.3

351.34 - 365.85 m

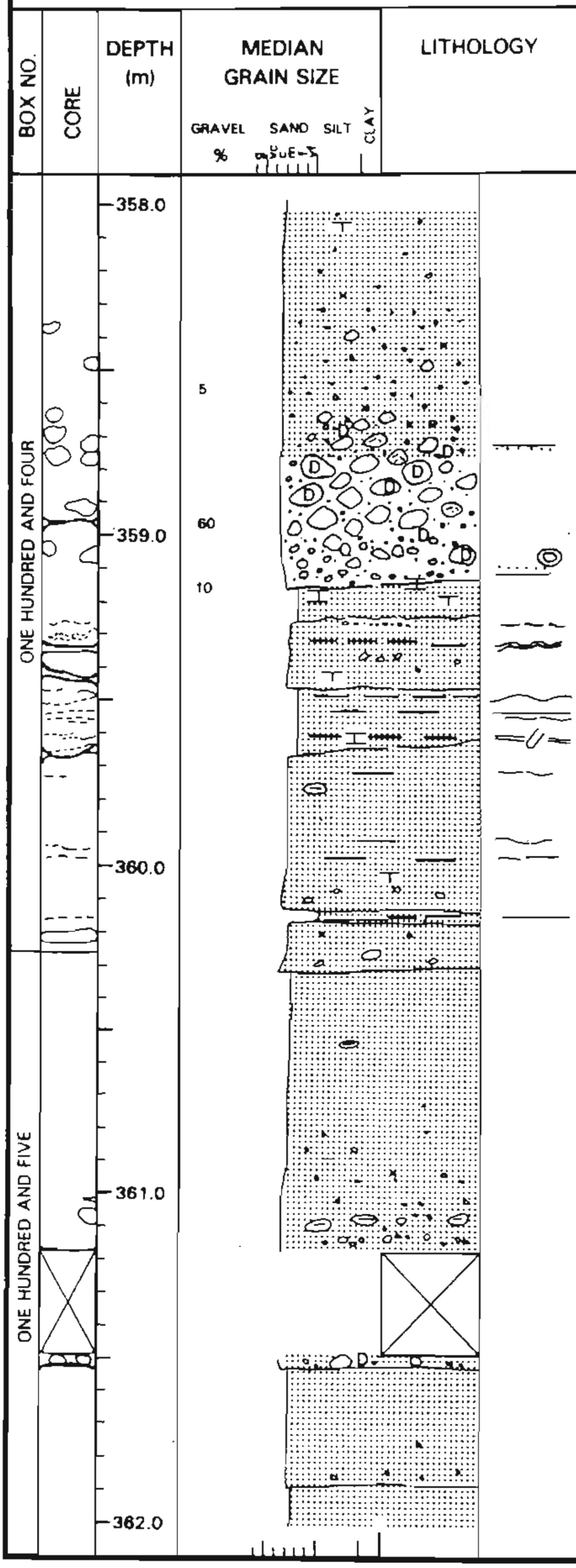
SANDSTONE

Hard and soft (varying with degrees of carbonate cementation), unstratified to **weakly** stratified, moderately sorted to poorly sorted, **coarse to very fine sandstone** (grey to greyish olive 10Y 5/1 to 5Y 5/2). Stratification is irregular and discontinuous, and ranges from subhorizontal to inclinations of 30°. Thin interbeds of weakly stratifed, **sandy mudstone** (dark greenish grey 5G 4/1) at 360.15, 363.52 and 363.88 m. Hard and soft matrix-supported **sandy pebble conglomerate** occurs in irregular beds through the sub–unit. Some conglomerates fine upward into sandstone. Clasts mostly range from 8 to 12 mm, but a few are > 60 mm and one dolerite is > 230 mm. Most are subrounded. About 50% are basement, 40% dolerite and the remainder are mudstone intraclasts. Carbonaceous fragments were found between 351.49 and 351.95 and at 353.98 m The interval from 363.31 to 365.85 m is strongly calcareous and bluish grey (5PB 5/1).

PROJECT : CIROS -1

SHEET NO. : 85

SCALE : 1:20



DESCRIPTION

Sub-unit 17.3

351.34 - 365.85 m

SANDSTONE

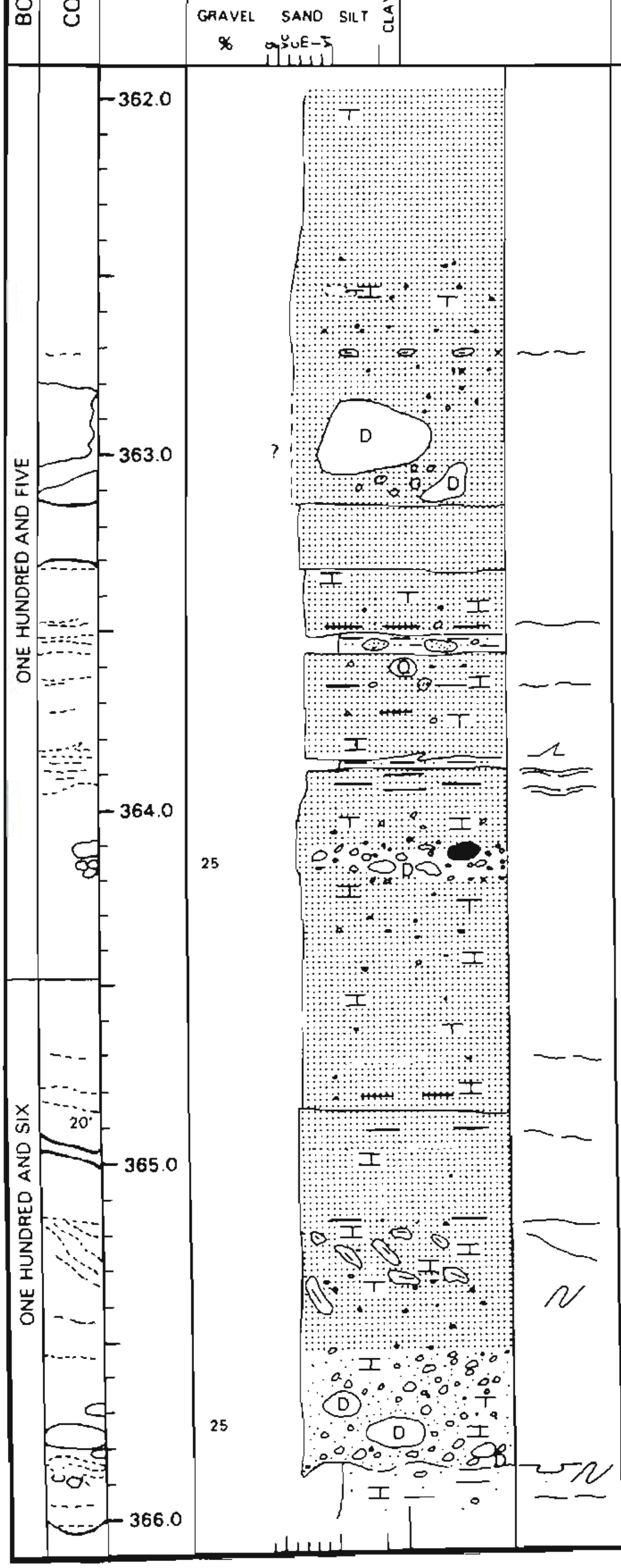
Hard and soft (varying with degrees of carbonate cementation), unstratified to weakly stratified, moderately sorted to poorly sorted, **coarse to very fine sandstone** (grey to greyish olive 10Y 5/1 to 5Y 5/2). Stratification is irregular and discontinuous, and ranges from subhorizontal to inclinations of 30°. Thin interbeds of weakly stratifed, **sandy mudstone** (dark greenish grey 5G 4/1) at 360.15, 363.52 and 363.88 m. Hard and soft matrix-supported **sandy pebble conglomerate** occurs in irregular beds through the sub-unit. Some conglomerates fine upward into sandstone. Clasts mostly range from 8 to 12 mm, but a few are > 60 mm and one dolerite is > 230 mm. Most are subrounded. About 50% are basement, 40% dolerite and the remainder are mudstone intraclasts. Carbonaceous fragments were found between 351.49 and 351.95 and at 353.98 m. The interval from 363.31 to 365.85 m is strongly calcareous and bluish grey (5PB 5/1).

PROJECT : CIROS -1

SHEET NO. : 86

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 17.3

351.34 - 365.85 m

SANDSTONE

Hard and soft (varying with degrees of carbonate cementation), unstratified to weakly stratified, moderately sorted to poorly sorted, **coarse to very fine sandstone** (grey to greyish olive 10Y 5/1 to 5Y 5/2). Stratification is irregular and discontinuous, and ranges from subhorizontal to inclinations of 30°. Thin interbeds of weakly stratified, **sandy mudstone** (dark greenish grey 5G 4/1) at 360.15, 363.52 and 363.88 m. Hard and soft matrix-supported **sandy pebble conglomerate** occurs in irregular beds through the sub-unit. Some conglomerates fine upward into sandstone. Clasts mostly range from 8 to 12 mm, but a few are > 60 mm and one dolerite is > 230 mm. Most are subrounded. About 50% are basement, 40% dolerite and the remainder are mudstone intraclasts. Carbonaceous fragments were found between 351.49 and 351.95 and at 353.98 m. The interval from 363.31 to 365.85 m is strongly calcareous and bluish grey (5PB 5/1)

—contact sharp and injected -

Sub-unit 18.1

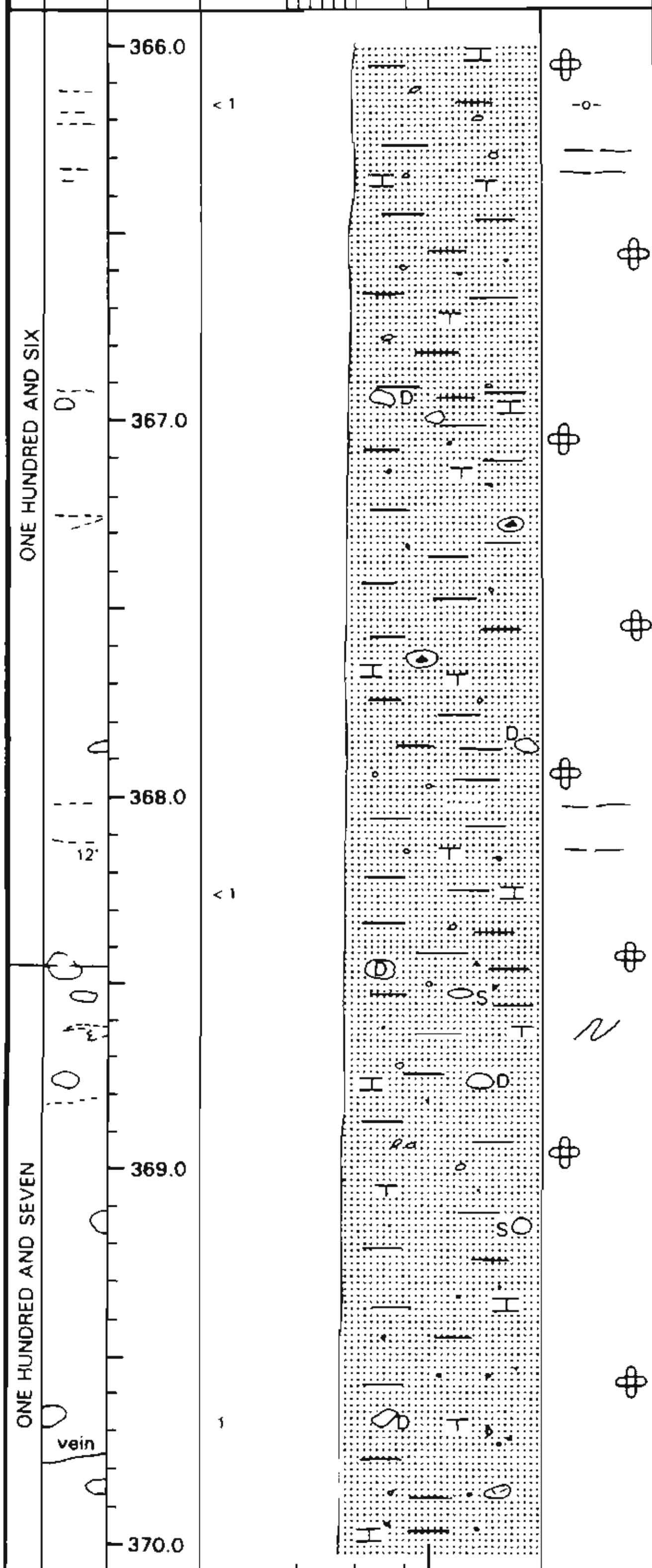
365.85 - 388.28 m

PROJECT : CIROS -1

SHEET NO. : 87

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.1

365.85 - 388.28 m

SANDSTONE (?DIAMICTITE)

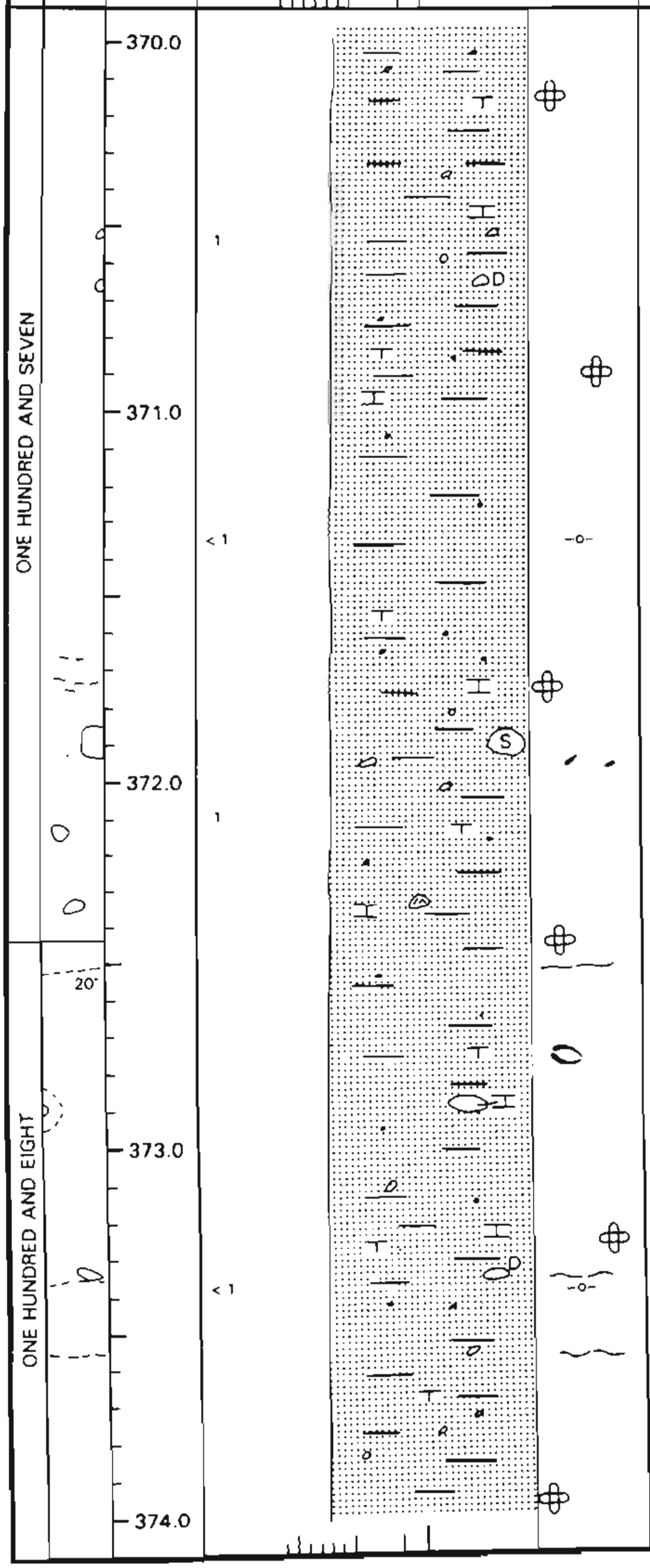
Moderately hard to hard, calcareous, unstratified to weakly stratified, poorly sorted, **slightly conglomeratic muddy fine sandstone (possibly sandy diamictite)** (greenish grey 7.5GY 5/1 to 10GY 5/1). The sub-unit fines downward to **very fine sandstone.** Strong bioturbation has disrupted bedding such that the sub-unit is essentially unstratified with only wispy remnants of stratification. Clasts form at the most 1% of the core and below 380 m form < 1%. Average clast size is about 8 mm and the largest is an indurated sedimentary clast > 45 mm at 371.90 m. Clasts are subangular to subrounded, and about 25% are striated. Lithologies include dolerite (65%), basement (30%) and indurated sedimentary clasts (5%). Articulated and fragmented bivalves occur below 376.25 m. Carbonaceous fragments 2 or 3 mm across are scattered through several parts of the sub-unit.

PROJECT : CIROS -1

SHEET NO. : 88

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.1

365.85 - 388.28 m

SANDSTONE (?DIAMICTITE)

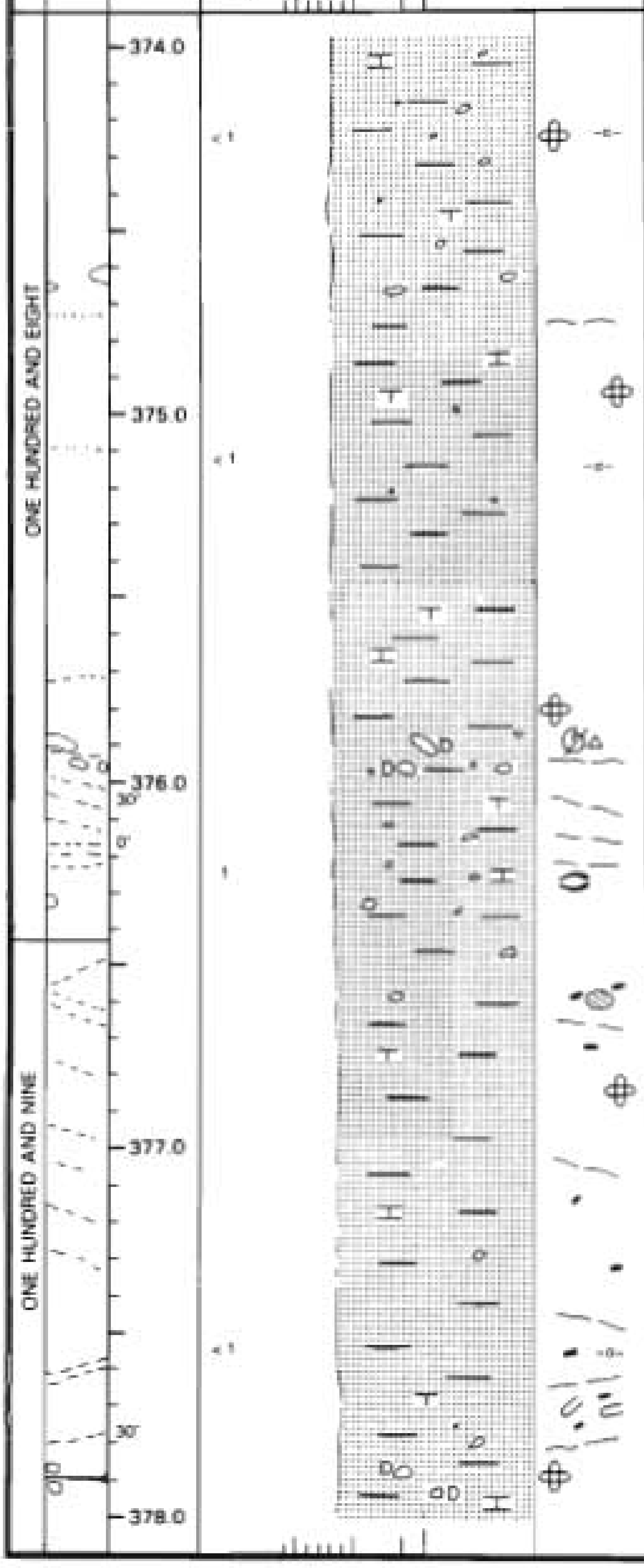
Moderately hard to hard, calcareous, unstratified to weakly stratified, poorly sorted, **slightly conglomeratic muddy fine sandstone (possibly sandy diamictite)** (greenish grey 7.5GY 5/1 to 10GY 5/1). The sub-unit fines downward to **very fine sandstone.** Strong bioturbation has disrupted bedding such that the sub-unit is essentially unstratified with only wispy remnants of stratification. Clasts form at the most 1% of the core and below 380 m form < 1%. Average clast size is about 8 mm and the largest is an indurated sedimentary clast > 45 mm at 371.90 m. Clasts are subangular to subrounded, and about 25% are striated. Lithologies include dolerite (65%), basement (30%) and indurated sedimentary clasts (5%). Articulated and fragmented bivalves occur below 376.25 m. Carbonaceous fragments 2 or 3 mm across are scattered through several parts of the sub-unit.

PROJECT : CIROS -1

SHEET NO. : 89

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.1

365.85 - 388.28 m

SANDSTONE (?DIAMICTITE)

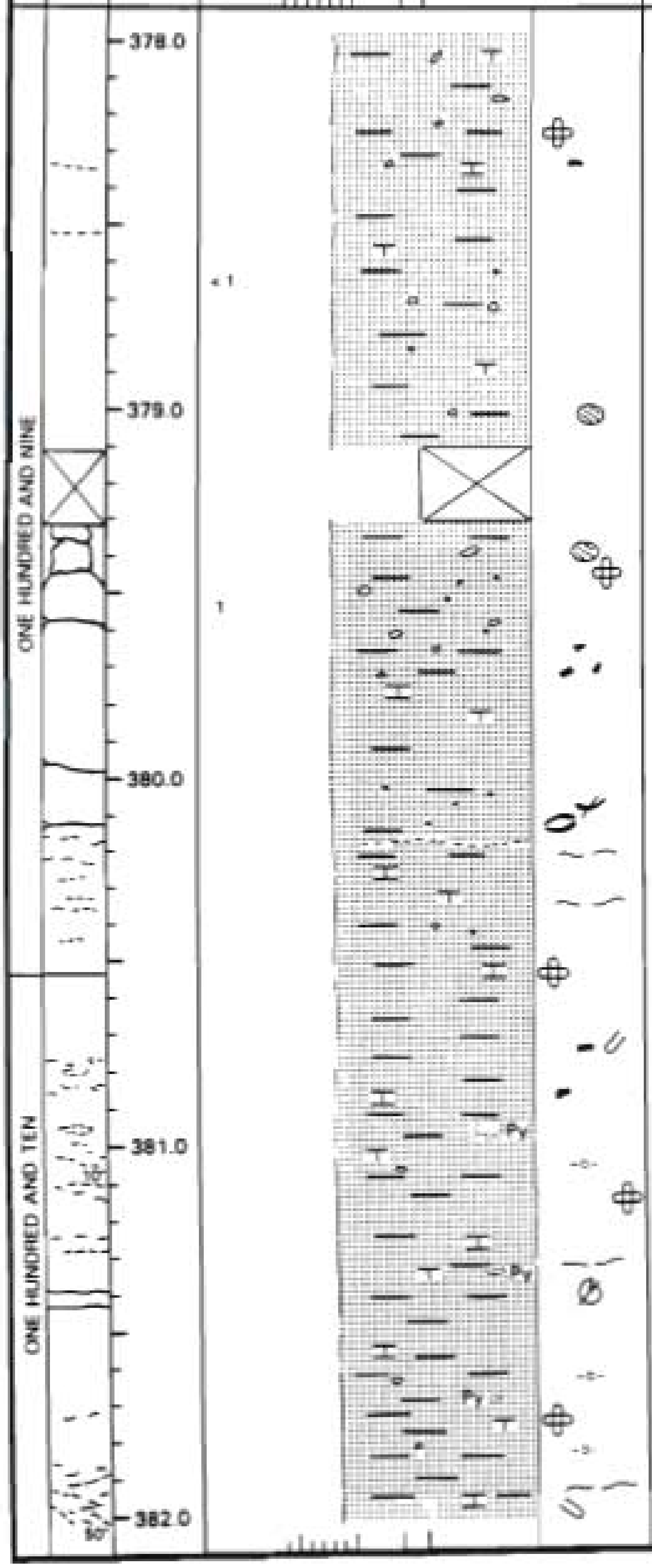
Moderately hard to hard, calcareous, unstratified to weakly stratified, poorly sorted, **slightly conglomeratic muddy fine sandstone (possibly sandy diamictite)** (greenish grey 7.5GY 5/1 to 10GY 5/1). The sub-unit fines downward to **very fine sandstone**. Strong bioturbation has disrupted bedding such that the sub-unit is essentially unstratified with only wispy remnants of stratification. Clasts form at the most 1% of the core and below 380 m form < 1%. Average clast size is about 8 mm and the largest is an indurated sedimentary clast > 45 mm at 371.90 m. Clasts are subangular to subrounded, and about 25% are striated. Lithologies include dolerite (65%), basement (30%) and indurated sedimentary clasts (5%). Articulated and fragmented bivalves occur below 376.25 m. Carbonaceous fragments 2 or 3 mm across are scattered through several parts of the sub-unit.

PROJECT : CIROS -1

SHEET NO. : 90

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.1

366.85 - 388.28 m

SANDSTONE (?DIAMICTITE)

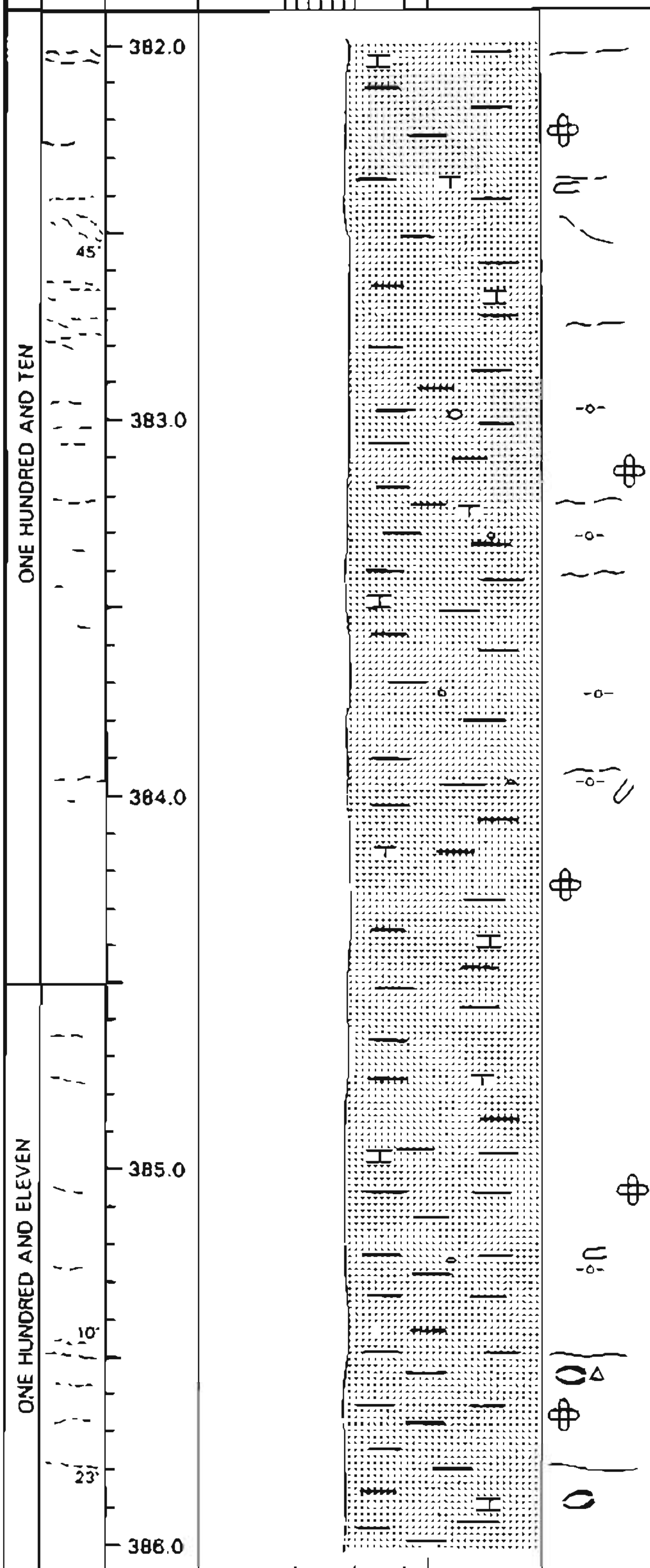
Moderately hard to hard, calcareous, unstratified to weakly stratified, poorly sorted, **slightly conglomeratic muddy fine sandstone (possibly sandy diamictite)** (greenish grey 7.5GY 5/1 to 10GY 5/1). The sub-unit fines downward to **very fine sandstone.** Strong bioturbation has disrupted bedding such that the sub-unit is essentially unstratified with only wispy remnants of stratification. Clasts form at the most 1% of the core and below 380 m form < 1%. Average clast size is about 8 mm and the largest is an indurated sedimentary clast > 45 mm at 371.90 m. Clasts are subangular to subrounded, and about 25% are striated. Lithologies include dolerite (65%), basement (30%) and indurated sedimentary clasts (5%). Articulated and fragmented bivalves occur below 376.25 m. Carbonaceous fragments 2 or 3 mm across are scattered through several parts of the sub-unit

PROJECT : CIROS -1

SHEET NO. : 91

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.1

365.85 - 388.28 m

SANDSTONE (?DIAMICTITE)

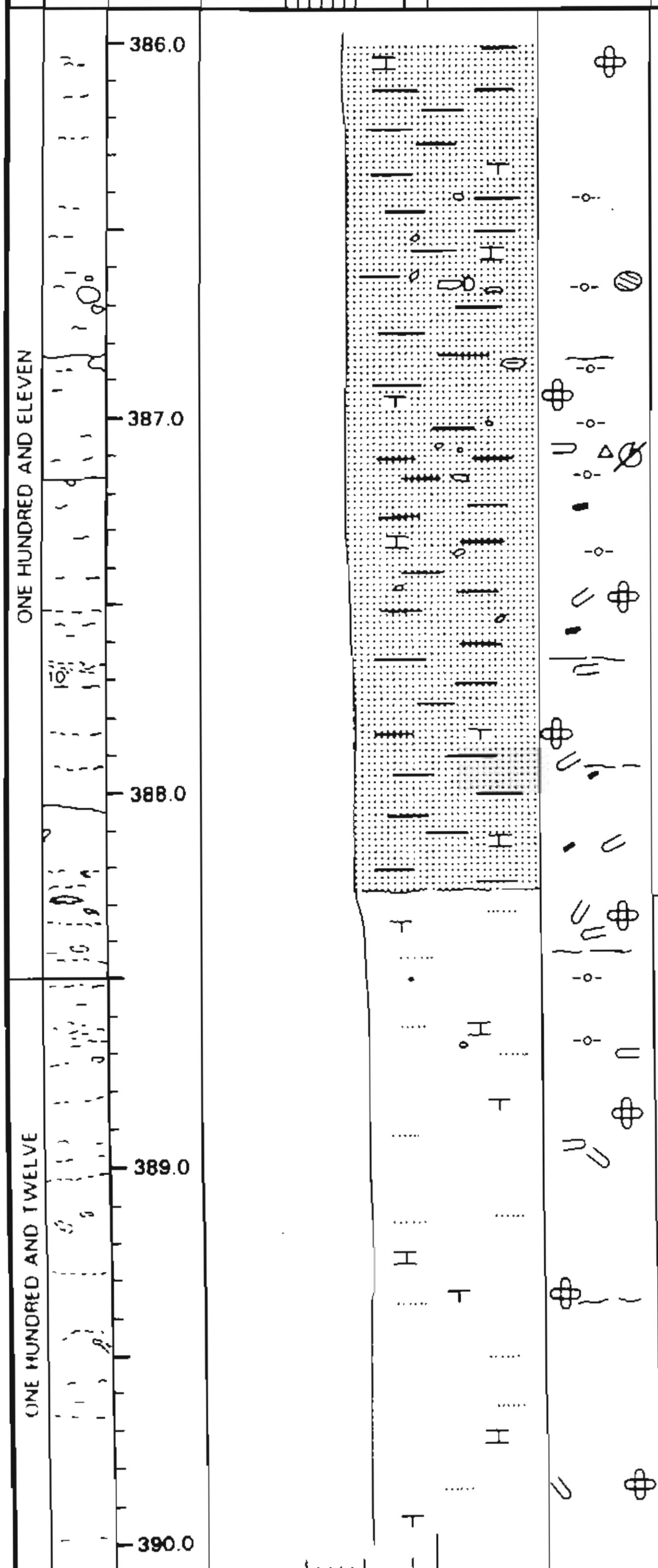
Moderately hard to hard, calcareous, unstratified to weakly stratified, poorly sorted, **slightly conglomeratic muddy fine sandstone (possibly sandy diamictite)** (greenish grey 7.5GY 5/1 to 10GY 5/1). The sub-unit fines downward to **very fine sandstone**. Strong bioturbation has disrupted bedding such that the sub-unit is essentially unstratified with only wispy remnants of stratification. Clasts form at the most 1% of the core and below 380 m form < 1%. Average clast size is about 8 mm and the largest is an indurated sedimentary clast > 45 mm at 371.90 m. Clasts are subangular to subrounded, and about 25% are striated. Lithologies include dolerite (65%), basement (30%) and indurated sedimentary clasts (5%). Articulated and fragmented bivalves occur below 376.25 m. Carbonaceous fragments 2 or 3 mm across are scattered through several parts of the sub-unit.

PROJECT : CIROS -1

SHEET NO. : 92

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.1

365.85 - 388.28 m

SANDSTONE (?DIAMICTITE)

Moderately hard to hard, calcareous, unstratified to weakly stratified, poorly sorted, **slightly conglomeratic muddy fine sandstone (possibly sandy diamictite)** (greenish grey 7.5GY 5/1 to 10GY 5/1). The sub-unit fines downward to **very fine sandstone**. Strong bioturbation has disrupted bedding such that the sub-unit is essentially unstratified with only wispy remnants of stratification. Clasts form at the most 1% of the core and below 380 m form < 1% Average clast size is about 8 mm and the largest is an indurated sedimentary clast > 45 mm at 371.90 m. Clasts are subangular to subrounded, and about 25% are striated. Lithologies include dolerite (65%), basement (30%) and indurated sedimentary clasts (5%) Articulated and fragmented bivalves occur below 376.25 m. Carbonaceous fragments 2 or 3 mm across are scattered through several parts of the sub-unit

—contact gradational over 5 cm –

Sub-unit 18.2

388.28 - 397.09 m

MUDSTONE

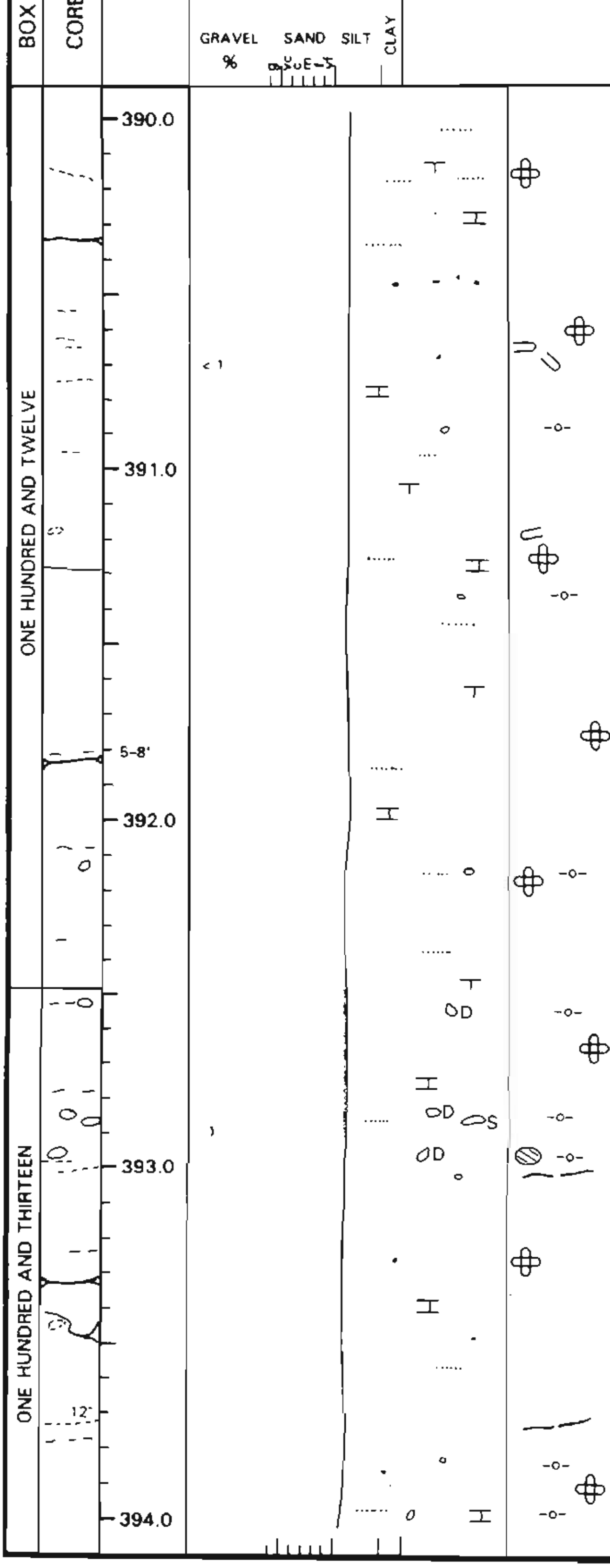
Hard, calcareous, weakly stratified, **sandy mudstone** (grey N 5 1). Stratification is wispy and irregular because of strong bioturbation. Rare scattered clasts averaging 6 mm, but reaching 18 mm long. One clast is striated Lithologies include dolerite, basement and indurated sediment.

PROJECT : CIROS -1

SHEET NO. : 93

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.2

388.28 - 397.09 m

MUDSTONE

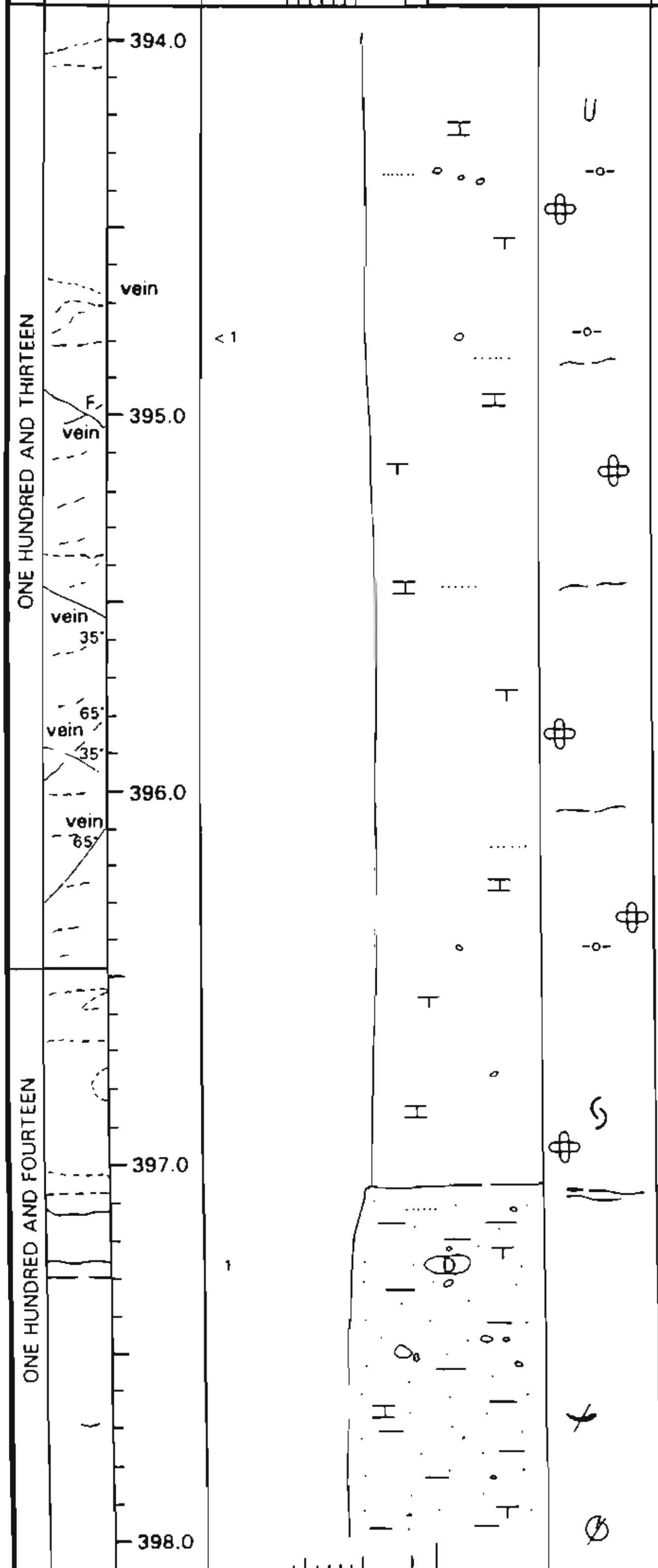
Hard, calcareous, weakly stratified, **sandy mudstone** (grey N 5/1). Stratification is wispy and irregular because of strong bioturbation. Rare scattered clasts averaging 6 mm, but reaching 18 mm long. One clast is striated. Lithologies include dolerite, basement and indurated sediment.

PROJECT : CIROS -1

SHEET NO. : 94

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.2

388.28 - 397.09 m

MUDSTONE

Hard, calcareous, weakly stratified, **sandy mudstone** (grey N 5/1). Stratification is wispy and irregular because of strong bioturbation. Rare scattered clasts averaging 6 mm, but reaching 18 mm long. One clast is striated. Lithologies include dolerite, basement and indurated sediment.

— contact gradational over 3 cm -

Sub-unit 18.3

397.09 - 399.84 m

DIAMICTITE

Hard, calcareous, unstratified, **slightly conglomeratic muddy fine to very fine sandstone (diamictite)** (grey 7.5Y 5 1)

Stratification is unrecognisable due to strong bioturbation.

Clasts form between 1 and 3% of the rock. Maximum clast size >43 mm - a basalt at 397.27 m Several clasts are striated. Lithologies include basement, dolerite and indurated sediment

Scattered articulated bivalves and shell fragments.

The sub-unit contacts are transitional with increased clast percent and poor sorting differentiating the diamictite from mudstone above and sandstone below.

PROJECT : CIROS -1

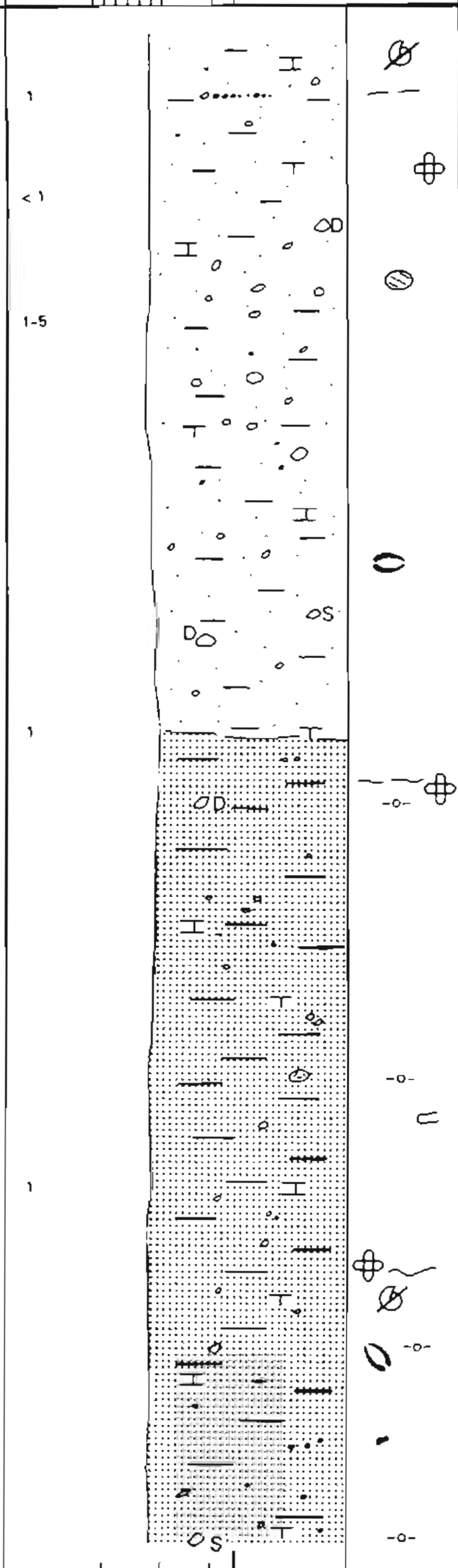
SHEET NO. : 95

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION

BOX NO.: ONE HUNDRED AND FOURTEEN; ONE HUNDRED AND FIFTEEN

DEPTH (m): 398.0, 399.0, 400.0, 401.0, 402.0



Sub-unit 18.3

397.09 - 399.84 m

DIAMICTITE

Hard, calcareous, unstratified, **slightly conglomeratic muddy fine to very fine sandstone (diamictite)** (grey 7.5Y 5/1). Stratification is unrecognisable due to strong bioturbation. Clasts form between 1 and 3% of the rock. Maximum clast size > 43 mm - a basalt at 397.27 m. Several clasts are striated. Lithologies include basement, dolerite and indurated sediment. Scattered articulated bivalves and shell fragments. The sub-unit contacts are transitional with increased clast percent and poor sorting differentiating the diamictite from mudstone above and sandstone below

— contact gradational over 10 cm -

Sub-unit 18.4

399.84 - 406.22 m

SANDSTONE (?DIAMICTITE)

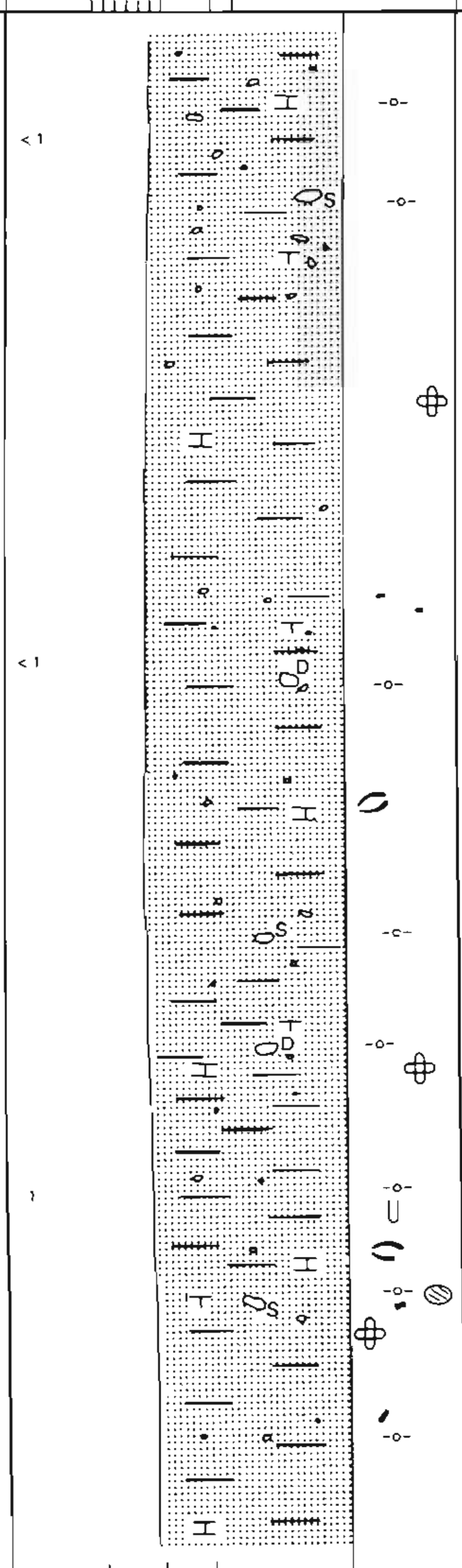
Hard, calcareous, unstratified, **slightly conglomeratic muddy fine to very fine sandstone (?diamictite)** (grey 10Y 5 1).
Stratification has been virtually obliterated by strong bioturbation.
Clasts are rare (<1% of the core). They are mostly subrounded lonestones up to 19 mm long and composed of dolerite, indurated sediment or rare basement lithologies.
Articulated bivalves, shell debris and carbonaceous fragments also occur in various parts of the sub-unit

PROJECT : CIROS -1

SHEET NO. : 96

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION
ONE HUNDRED AND FIFTEEN		402.0	< 1		Sub-unit 18.4
		403.0	< 1		399.84 - 406.22 m
		404.0			SANDSTONE (?DIAMICTITE)
ONE HUNDRED AND SIXTEEN		405.0	1		
		406.0			



Sub-unit 18.4

399.84 - 406.22 m

SANDSTONE (?DIAMICTITE)

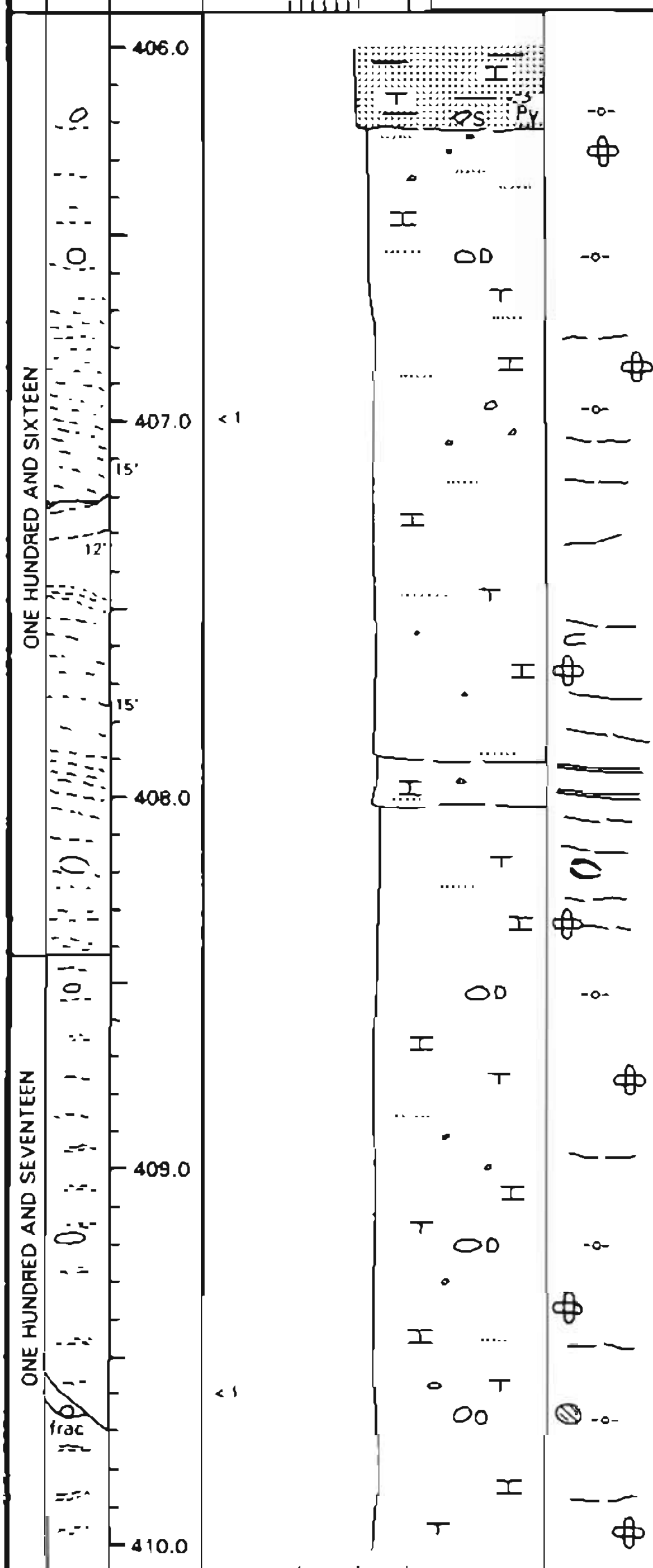
Hard, calcareous, unstratified, **slightly conglomeratic muddy fine to very fine sandstone (?diamictite)** (grey 10Y 5/1). Stratification has been virtually obliterated by strong bioturbation. Clasts are rare (< 1% of the core). They are mostly subrounded lonestones up to 19 mm long and composed of dolerite, indurated sediment or rare basement lithologies. Articulated bivalves, shell debris and carbonaceous fragments also occur in various parts of the sub-unit.

PROJECT : CIROS -1

SHEET NO. : 97

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



—contact gradational over 10 cm –

Sub–unit 18.5

406.22 - 425.70 m

MUDSTONE

Hard, calcareous, unstratified to well-stratified, **sandy mudstone** (dark greenish grey 5G 4/1).

Stratification varies from unstratified and strongly bioturbated through moderately bioturbated, wavy, discontinuous bedding to subhorizontal lamination, in places dipping up to 15°.

The bioturbation is mostly evident from colour mottling but a number of individual burrows can also be seen in the core face. The burrows are 4 to 8 mm across and are either horizontal or inclined at angles of up to 40°.

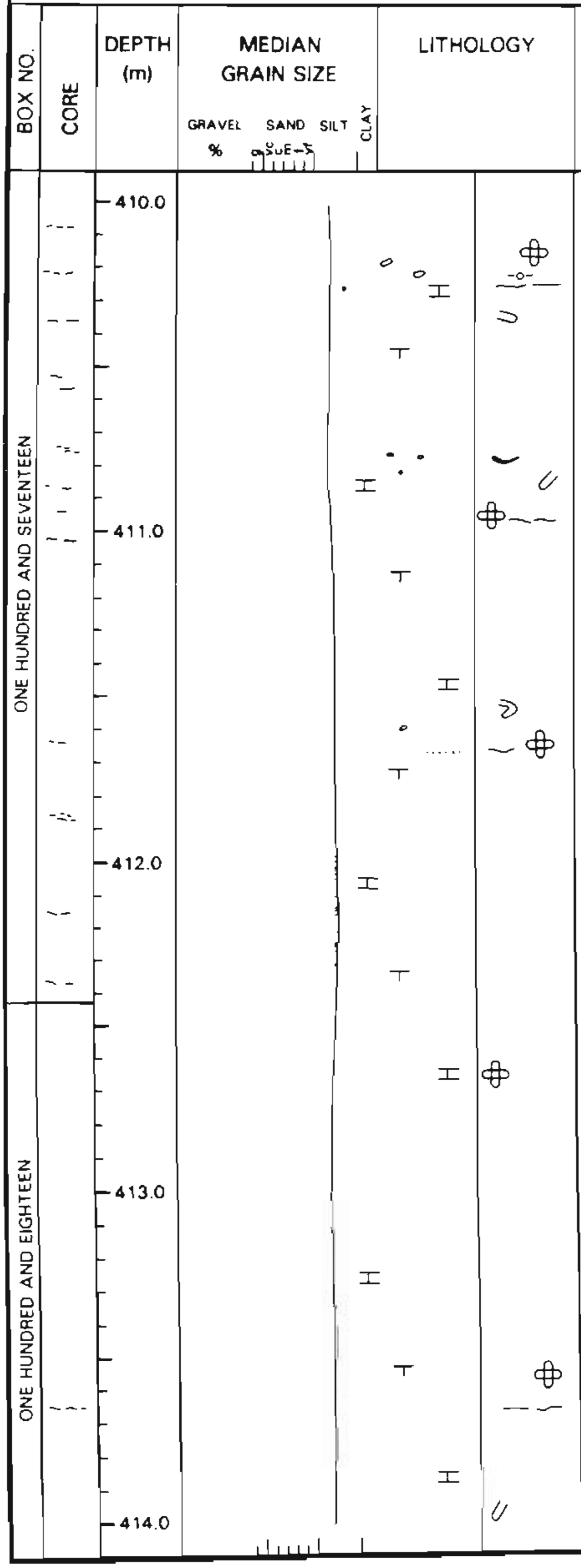
A few lonestones up to 19 mm across occur in the upper 2 m but below this clasts are extremely uncommon and rarely exceed granule size. One clast is striated.

The core also contains rare articulated and fragmented bivalves.

PROJECT : CIROS -1

SHEET NO. : 98

SCALE : 1:20



DESCRIPTION

Sub-unit 18.5

406.22 - 425.70 m

MUDSTONE

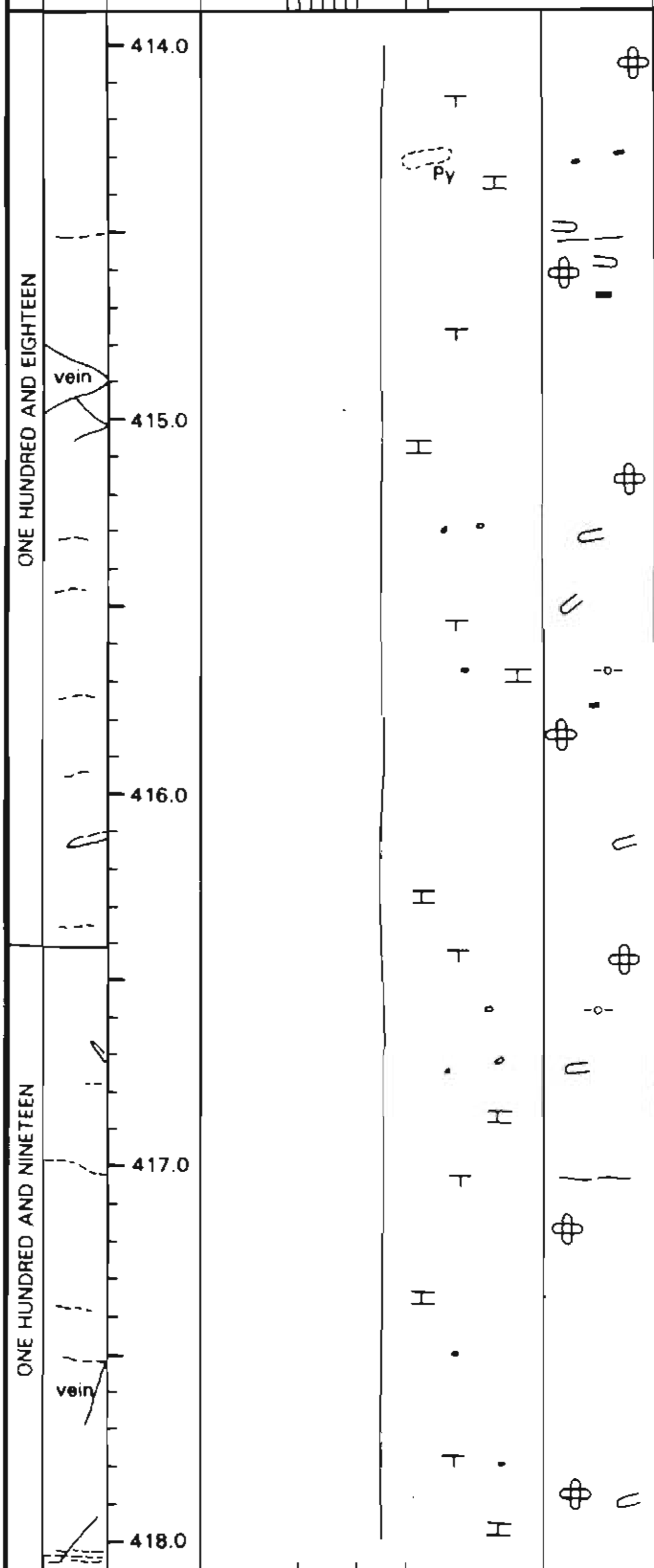
Hard, calcareous, unstratified to well-stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Stratification varies from unstratified and strongly bioturbated through moderately bioturbated, wavy, discontinuous bedding to subhorizontal lamination, in places dipping up to 15°. The bioturbation is mostly evident from colour mottling but a number of individual burrows can also be seen in the core face. The burrows are 4 to 8 mm across and are either horizontal or inclined at angles of up to 40°. A few lonestones up to 19 mm across occur in the upper 2 m but below this clasts are extremely uncommon and rarely exceed granule size. One clast is striated. The core also contains rare articulated and fragmented bivalves.

PROJECT : CIROS -1

SHEET NO. : 99

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % SAND SILT CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.5

406.22 - 425.70 m

MUDSTONE

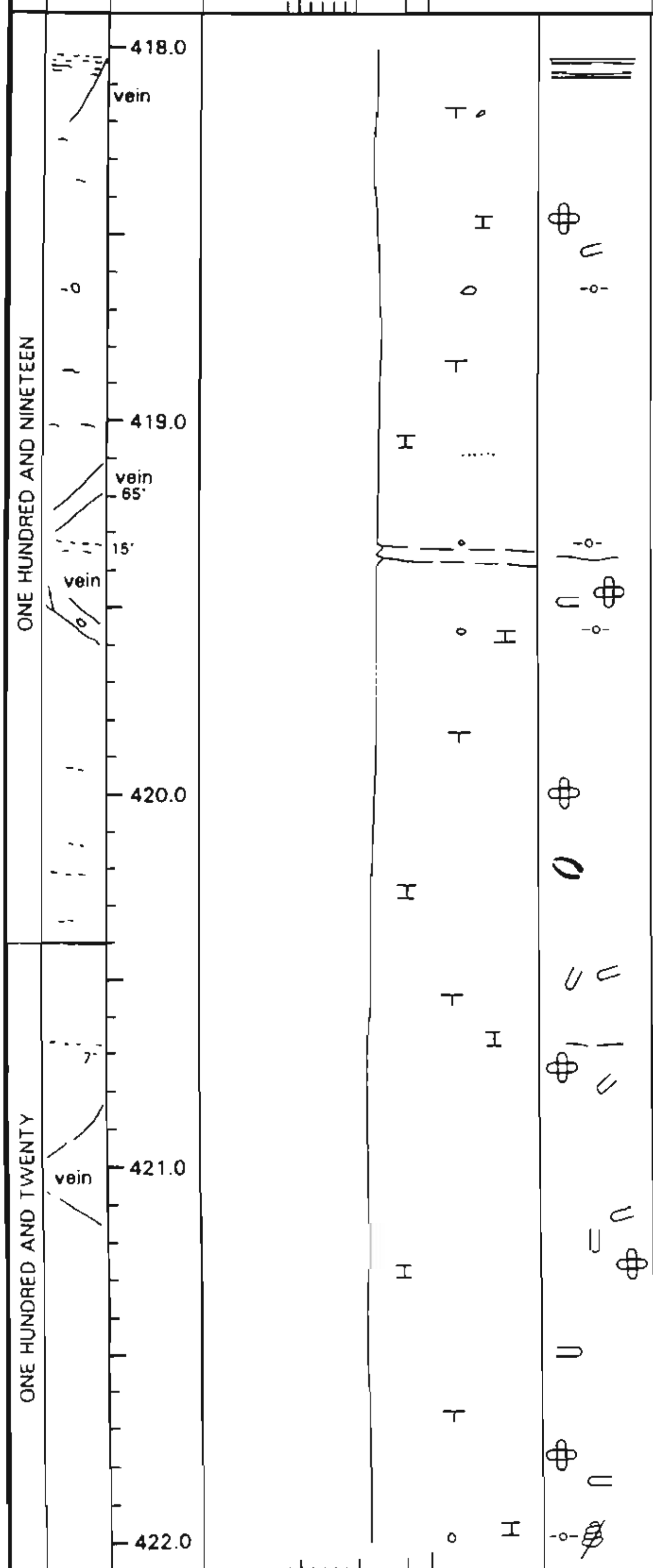
Hard, calcareous, unstratified to well-stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Stratification varies from unstratified and strongly bioturbated through moderately bioturbated, wavy, discontinuous bedding to subhorizontal lamination, in places dipping up to 15°. The bioturbation is mostly evident from colour mottling but a number of individual burrows can also be seen in the core face. The burrows are 4 to 8 mm across and are either horizontal or inclined at angles of up to 40°. A few lonestones up to 19 mm across occur in the upper 2 m but below this clasts are extremely uncommon and rarely exceed granule size. One clast is striated. The core also contains rare articulated and fragmented bivalves.

PROJECT : CIROS -1

SHEET NO. : 100

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.5

406.22 - 425.70 m

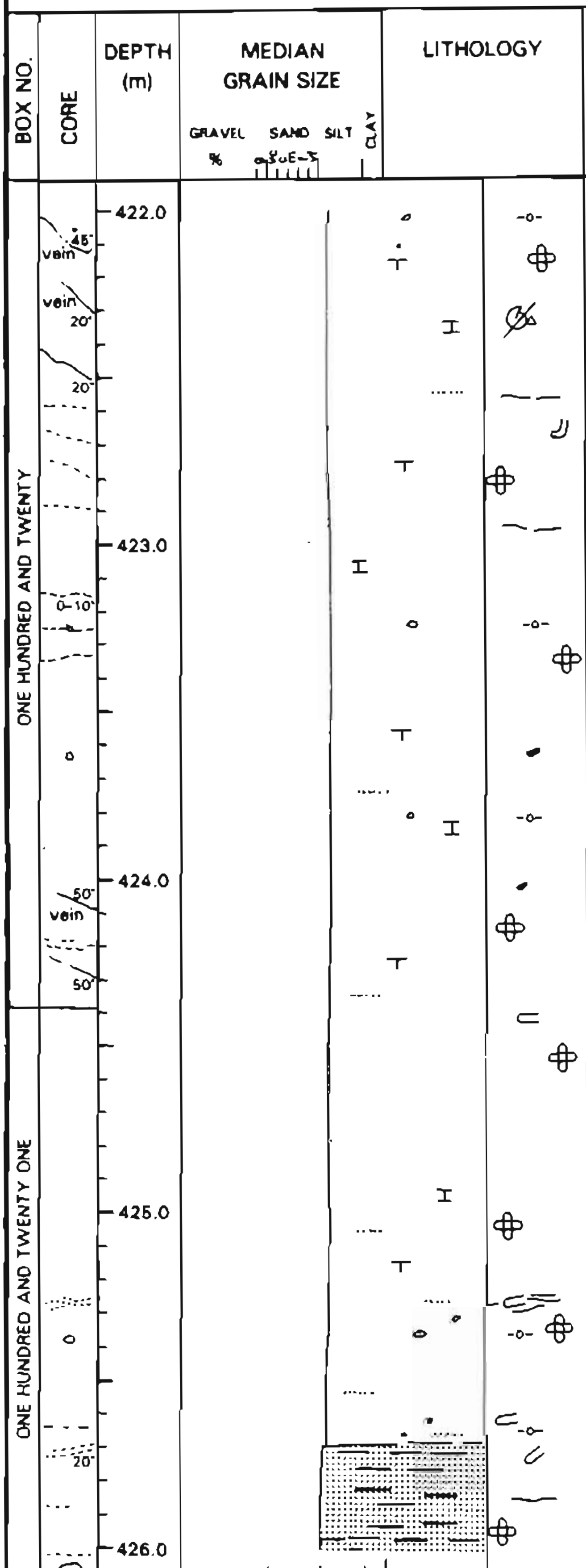
MUDSTONE

Hard, calcareous, unstratified to well-stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Stratification varies from unstratified and strongly bioturbated through moderately bioturbated, wavy, discontinuous bedding to subhorizontal lamination, in places dipping up to 15°. The bioturbation is mostly evident from colour mottling but a number of individual burrows can also be seen in the core face. The burrows are 4 to 8 mm across and are either horizontal or inclined at angles of up to 40°. A few lonestones up to 19 mm across occur in the upper 2 m but below this clasts are extremely uncommon and rarely exceed granule size. One clast is striated. The core also contains rare articulated and fragmented bivalves.

PROJECT : CIROS -1

SHEET NO. : 101

SCALE : 1:20



DESCRIPTION

Sub-unit 18.5

406.22 - 425.70 m

MUDSTONE

Hard, calcareous, unstratified to well-stratified, **sandy mudstone** (dark greenish grey 5G 4/1). Stratification varies from unstratified and strongly bioturbated through moderately bioturbated, wavy, discontinuous bedding to subhorizontal lamination, in places dipping up to 15°. The bioturbation is mostly evident from colour mottling but a number of individual burrows can also be seen in the core face. The burrows are 4 to 8 mm across and are either horizontal or inclined at angles of up to 40°. A few lonestones up to 19 mm across occur in the upper 2 m but below this clasts are extremely uncommon and rarely exceed granule size. One clast is striated. The core also contains rare articulated and fragmented bivalves.

Sub-unit 18.6

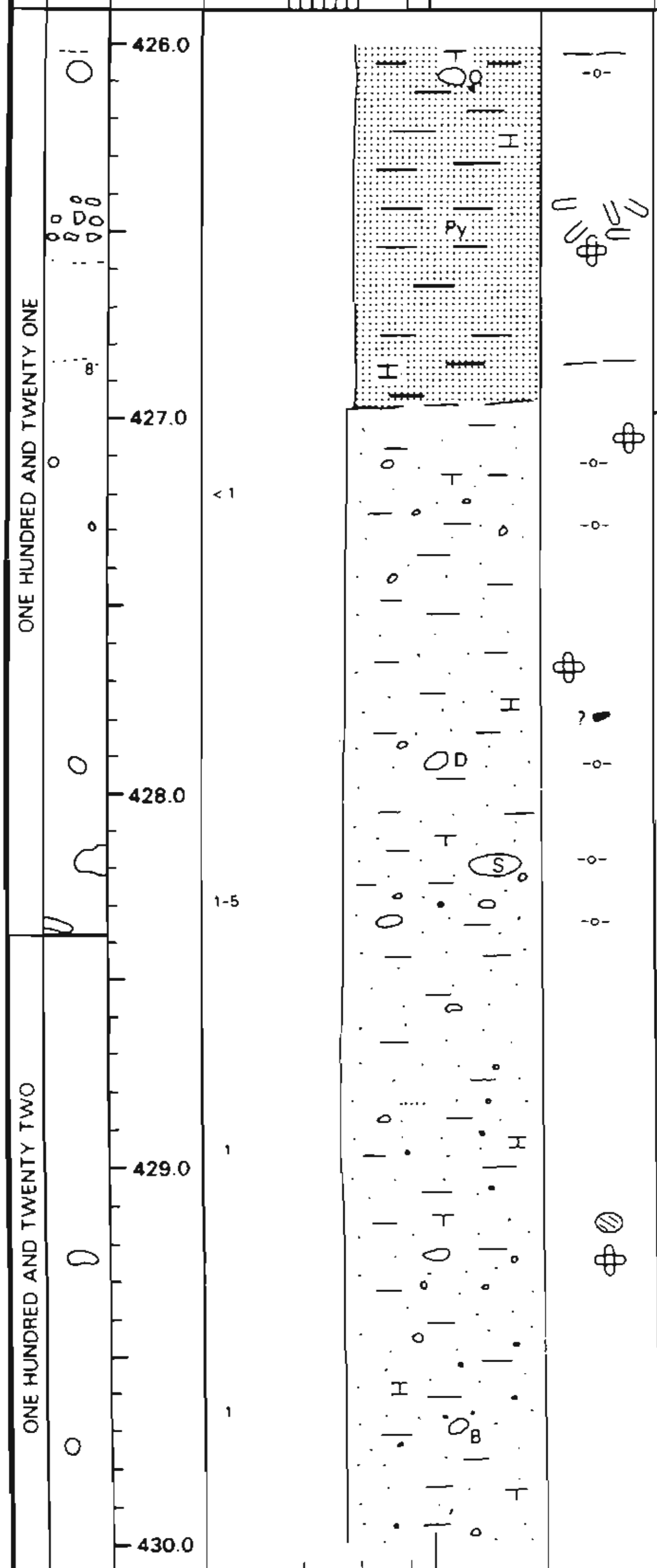
425.70 - 426.98 m

PROJECT : CIROS -1

SHEET NO. : 102

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.6

425.70 - 426.98 m

SANDSTONE

Hard, calcareous, weakly stratified, **muddy very fine sandstone** (dark greenish grey 5G 4/1). Stratification is wispy and irregular and disrupted by bioturbation. One subangular to subrounded clast of vein quartz 33 mm long at 426.07 m.

— contact gradational -

Sub-unit 18.7

426.98 - 430.99 m

DIAMICTITE

Hard, slightly calcareous, unstratified, **conglomeratic muddy fine sandstone (diamictite)** (grey 10Y 5 1).

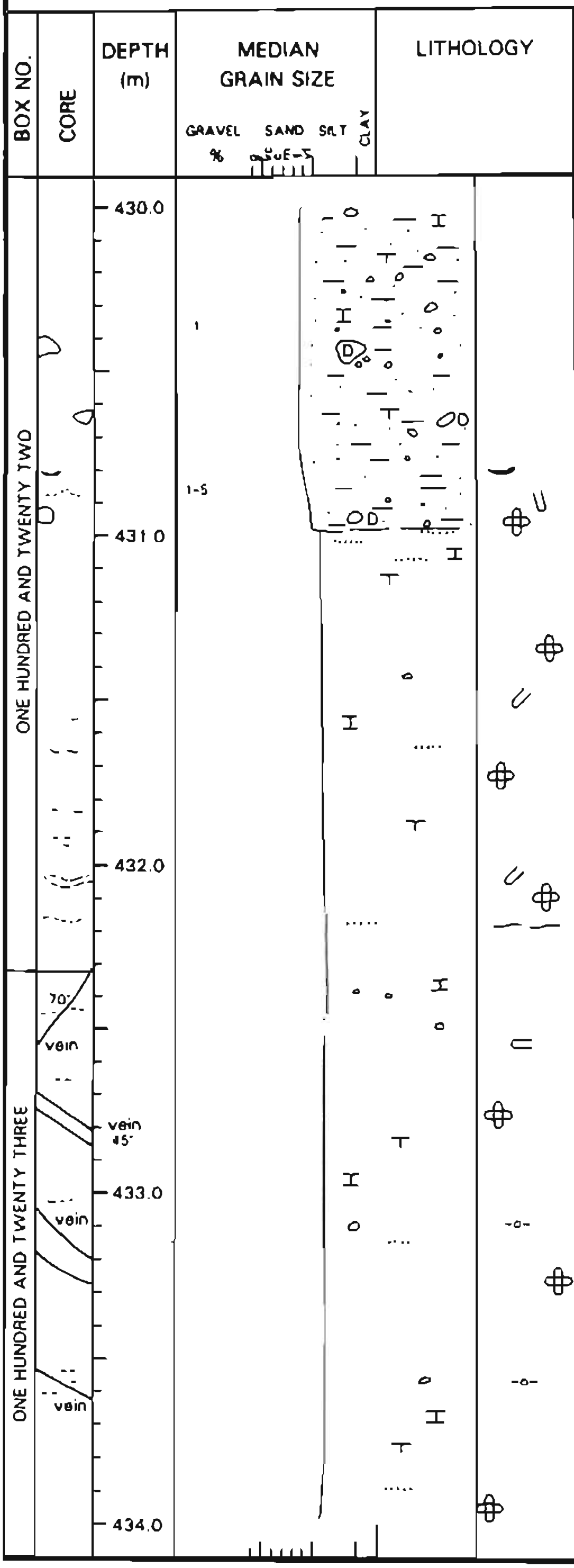
Clasts form less than 2% of the core, and average 4 to 10 mm long, though two are >50 mm. They range in shape from subangular to subrounded and are mainly dolerite and indurated sediment. One clast is striated

Sparse shell and carbonaceous fragments.

PROJECT : CIROS -1

SHEET NO. : 103

SCALE : 1:20



DESCRIPTION

Sub-unit 18.7

426.98 - 430.99 m

DIAMICTITE

Hard, slightly calcareous, unstratified, conglomeratic muddy fine sandstone (diamictite) (grey 10Y 5/1). Clasts form less than 2% of the core, and average 4 to 10 mm long, though two are > 50 mm. They range in shape from subangular to subrounded and are mainly dolerite and indurated sediment. One clast is striated. Sparse shell and carbonaceous fragments

— contact gradational over 3 cm -

Sub-unit 18.8

430.99 - 435.52 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratified, sandy mudstone (grey N 5 1).

Wispy, irregular and discontinuous bedding indicates strong bioturbation. Some individual horizontal and inclined burrows seen.

Lonestones rare and no longer than 5 mm.

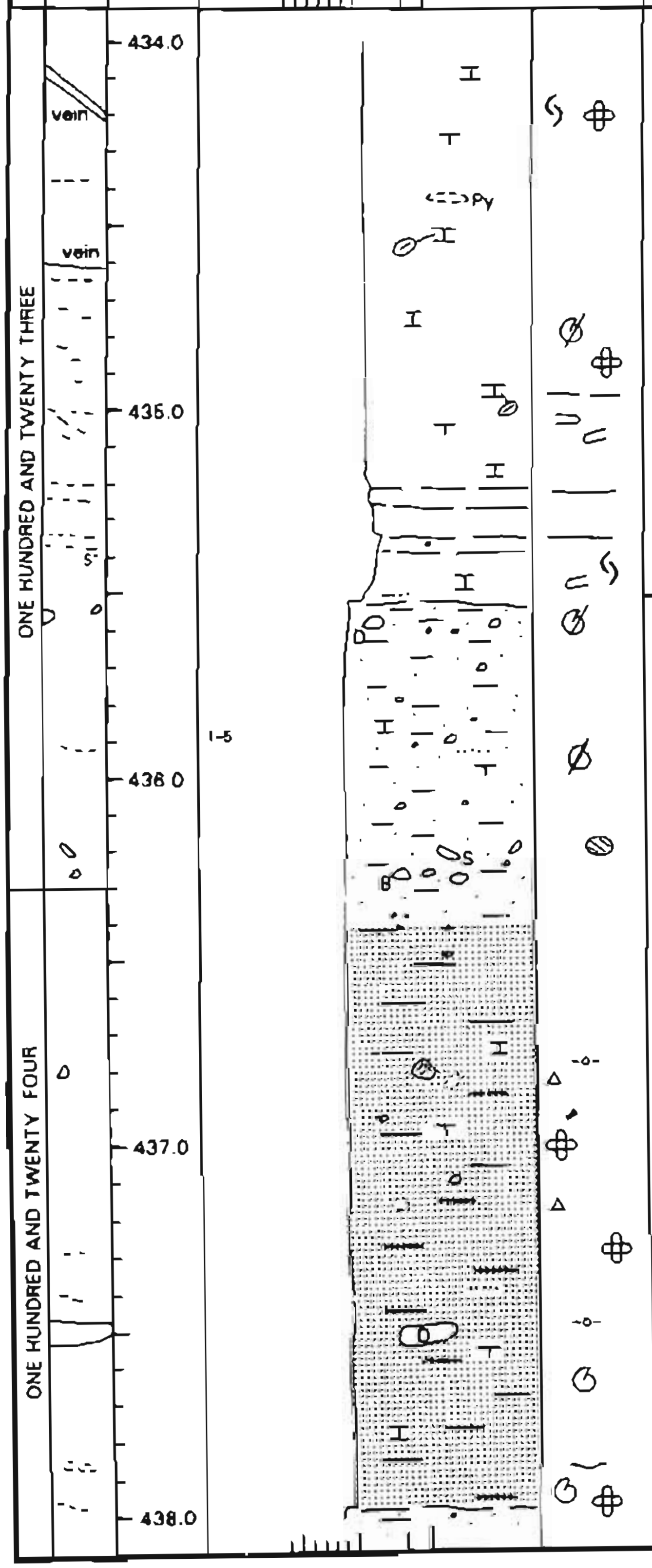
Indeterminate bioclastic debris at 434.80 m

PROJECT : CIROS -1

SHEET NO. : 104

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.8

430.99 - 435.52 m

MUDSTONE

Hard, calcareous, unstratified to weakly stratified, sandy mudstone (grey N 5/1). Wispy, irregular and discontinuous bedding indicates strong bioturbation. Some individual horizontal and inclined burrows seen. Lonestones rare and no longer than 5 mm. Indeterminate bioclastic debris at 434.80 m.

– contact gradational over 3 cm -

Sub-unit 18.9

435.52- 437.97 m

DIAMICTITE to SANDSTONE

Hard, slightly calcareous, unstratified, slightly conglomeratic muddy very fine sandstone (diamictite) (dark bluish grey 5B 4/1) from 435.52 to about 436.42 m grading down to muddy very fine sandstone of the same colour.

Bioturbation and broken shell fragments are common in the sandstone. Clasts are subrounded and form around 2% of the diamictite and < 1% of the sandstone. They range up to 43 mm long. One clast is striated. Lithologies include dolerite, indurated sediment, basement and Beacon sandstone.

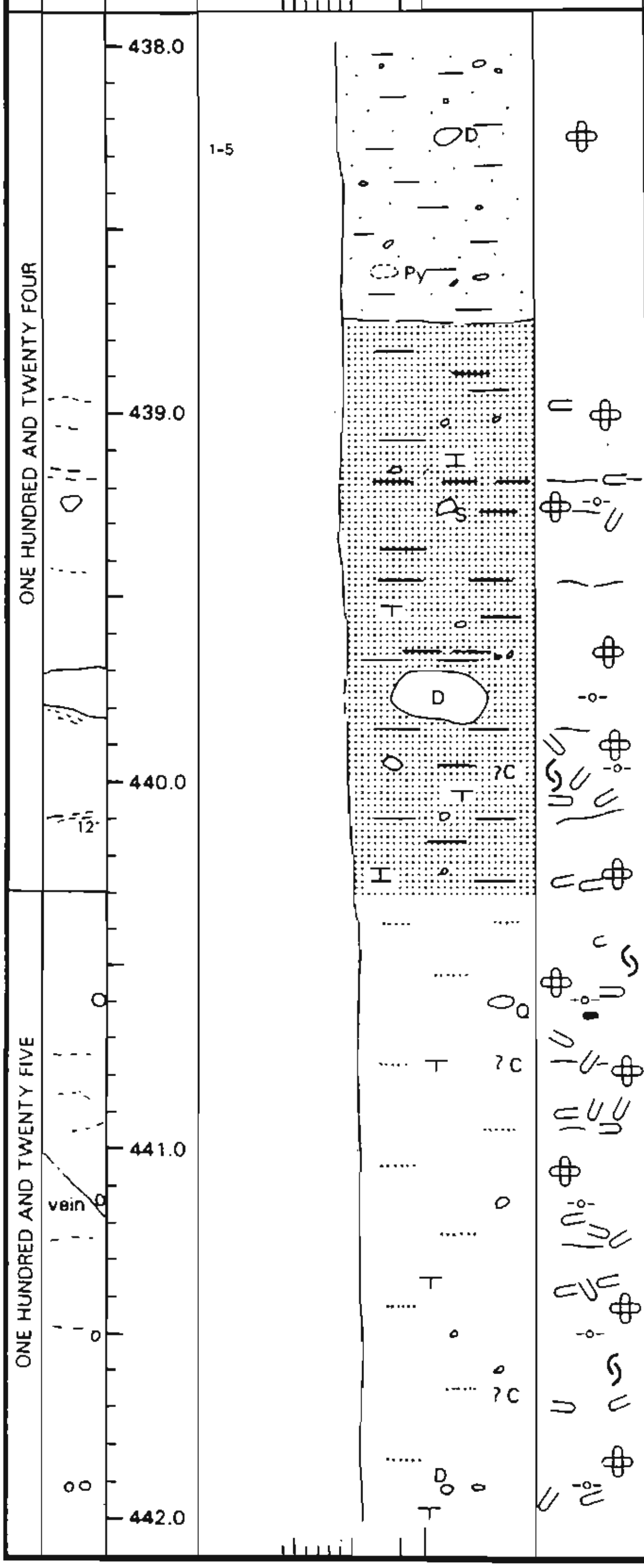
– contact gradational over 3 cm -

PROJECT : CIROS-1

SHEET NO. : 105

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.10

437.97 –438.76 m

DIAMICTITE

Hard, slightly calcareous, unstratified, **slightly conglomeratic muddy fine to very fine sandstone (diamictite)** (dark bluish grey 5B 4/1).
Bioturbated in places.
Clasts form 1-2% of the core, and range up to 24 mm long. On average they are subrounded and dolerite.

— contact gradational over 3 cm –

Sub-unit 18.11

438.76 - 440.30 m

SANDSTONE

Hard, slightly calcareous, unstratified to weakly stratified, poorly sorted, **muddy fine to very fine sandstone** (greenish grey to dark greenish grey 10GY 5/1 to 4/1).
Stratification is discontinuous and wispy, disrupted by strong bioturbation.
Rare (< 1%) lonestones on average 8 to 10 mm long, but one is > 147 mm.
Some dark greenish grey mottling, possibly chlorite.

— contact gradational over 20 cm –

Sub-unit 18.12

440.30 – 445.88 m

MUDSTONE

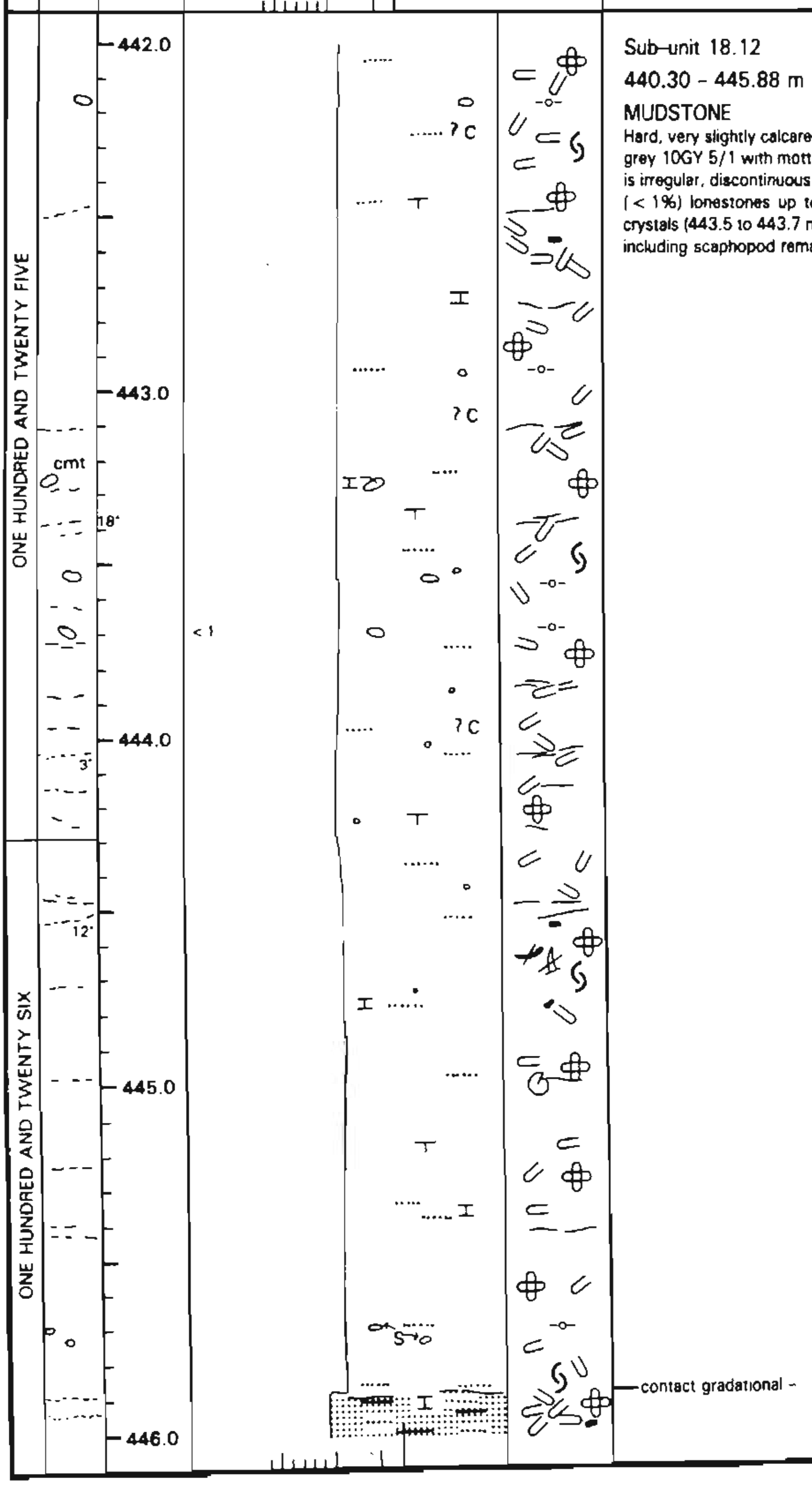
Hard, very slightly calcareous, weakly stratified, **sandy mudstone** (greenish grey 10GY 5/1 with mottles of dark greenish grey 10GY 4/1).
Stratification is irregular, discontinuous and wispy as a result of strong bioturbation.
Rare (< 1%) lonestones up to 17 mm long. A few have weathered feldspar crystals (443.5 to 443.7 m)
Some carbonaceous fragments and shell debris, including scaphopod remains.

PROJECT : CIROS -1

SHEET NO. : 106

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub–unit 18.12

440.30 - 445.88 m

MUDSTONE

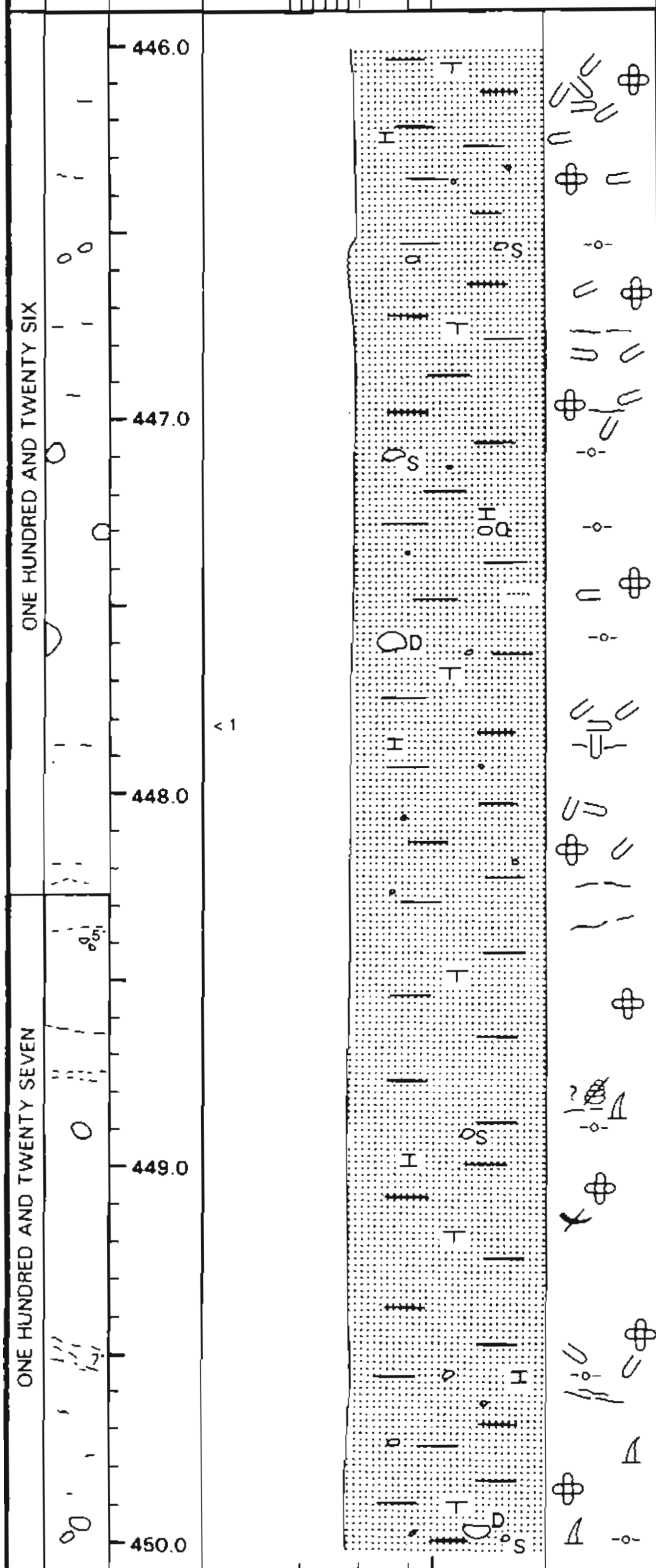
Hard, very slightly calcareous, weakly stratified, **sandy mudstone** (greenish grey 10GY 5/1 with mottles of dark greenish grey 10GY 4/1). Stratification is irregular, discontinuous and wispy as a result of strong bioturbation. Rare (< 1%) lonestones up to 17 mm long. A few have weathered feldspar crystals (443.5 to 443.7 m). Some carbonaceous fragments and shell debris, including scaphopod remains.

PROJECT : CIROS-1

SHEET NO. : 107

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.13

445.88 – 450.58 m

SANDSTONE

Hard, slightly calcareous, weakly stratified, poorly sorted, **muddy fine to very fine sandstone** (dark greenish grey 10GY 4/1).

Stratification sparse, discontinuous and wispy, and shows extensive burrowing.

Rare lonestones of subangular to subrounded indurated sediment, dolerite and quartz up to 30 mm long.

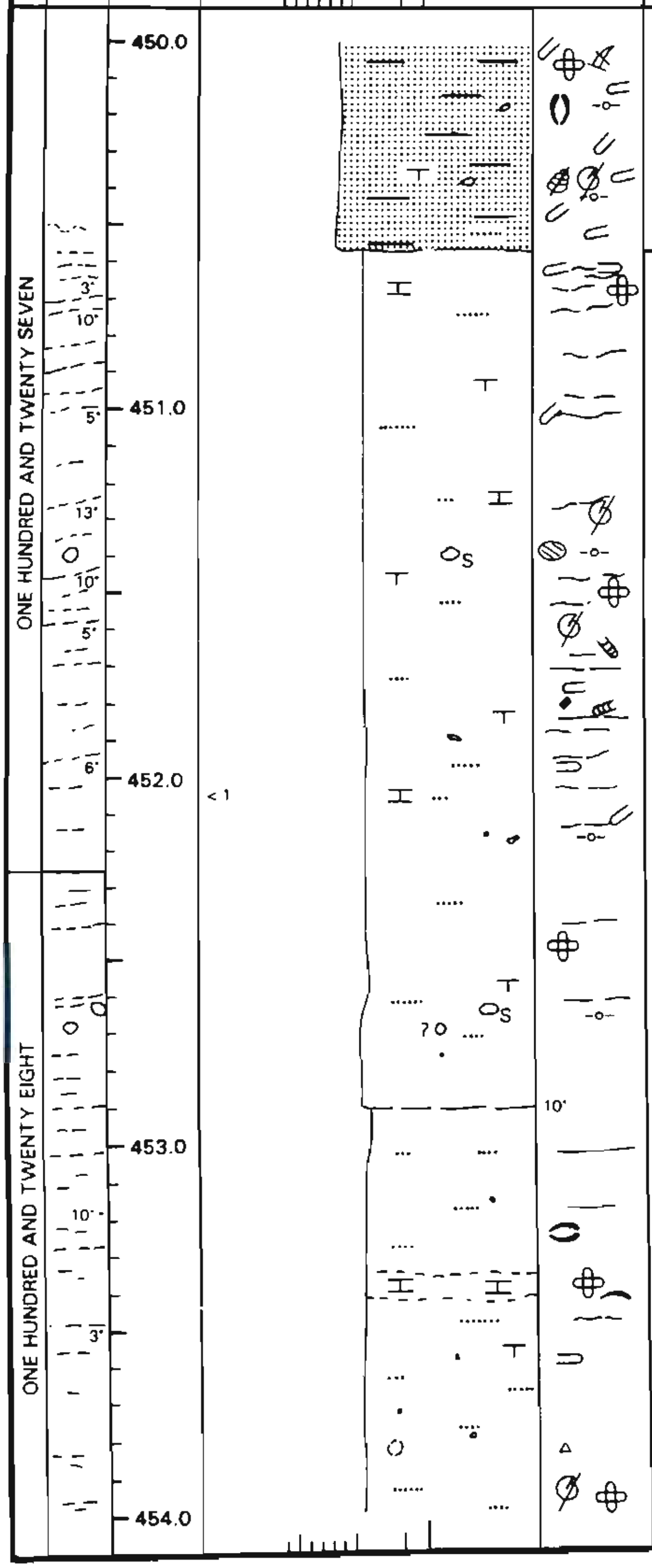
Broken skeletal debris from bivalves, scaphopods and ?gastropods.

PROJECT : CIROS -1

SHEET NO. : 108

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



— contact gradational over 3 cm -

Sub-unit 18.14

450.58 – 461.75 m

MUDSTONE

Hard, calcareous, weakly stratified, **sandy mudstone** (grey to olive black 7.5Y 4/1 to 3/1).

Stratification is discontinuous, wispy and inclined (3 to 22°), and shows extensive burrowing, some with spreiten.

Rare lonestones of angular to subrounded indurated sediment and limestone (deformed during or slightly after deposition). One striated clast.

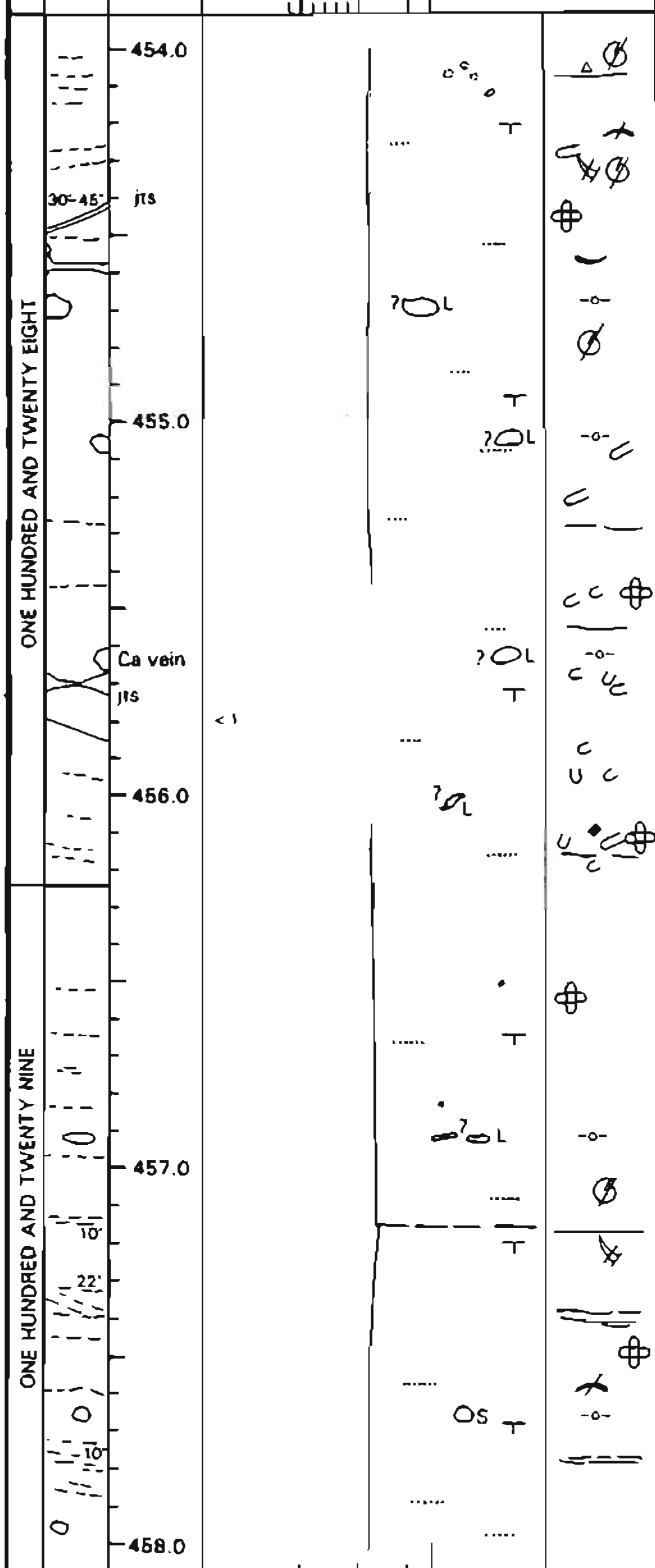
Scattered single and articulated bivalve shells and broken bivalve and scaphopod debris. Some carbonaceous fragments.

PROJECT : CIROS -1

SHEET NO. : 109

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.14

450.58 - 461.75 m

MUDSTONE

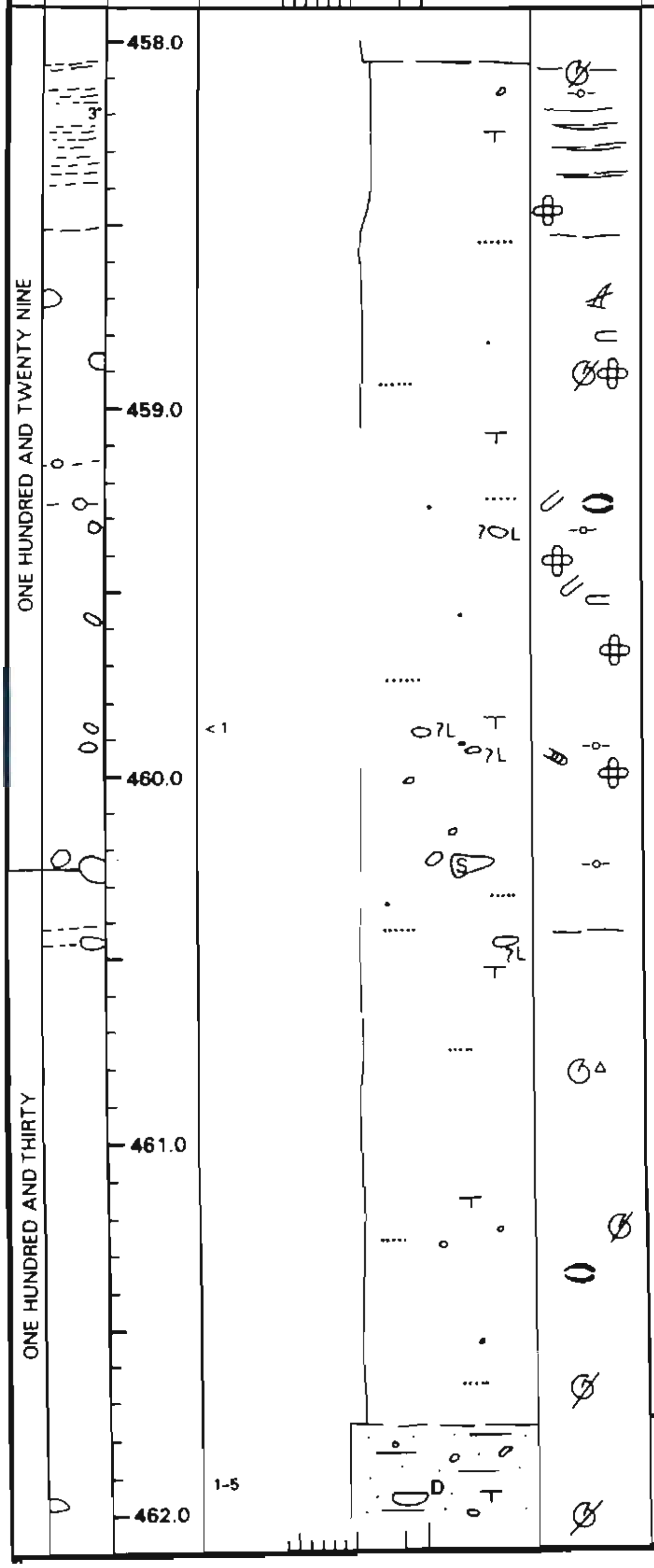
Hard, calcareous, weakly stratified, sandy mudstone (grey to olive black 7.5Y 4/1 to 3/1). Stratification is discontinuous, wispy and inclined (3 to 22°), and shows extensive burrowing, some with spreiten. Rare lonestones of angular to subrounded indurated sediment and limestone (deformed during or slightly after deposition). One striated clast. Scattered single and articulated bivalve shells and broken bivalve and scaphopod debris. Some carbonaceous fragments.

PROJECT : CIROS -1

SHEET NO. : 110

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.14

450.58 - 461.75 m

MUDSTONE

Hard, calcareous, weakly stratified, **sandy mudstone** (grey to olive black 7.5Y 4/1 to 3/1). Stratification is discontinuous, wispy and inclined (3 to 22°), and shows extensive burrowing, some with spreiten. Rare lonestones of angular to subrounded indurated sediment and limestone (deformed during or slightly after deposition). One striated clast. Scattered single and articulated bivalve shells and broken bivalve and scaphopod debris. Some carbonaceous fragments

Sub-unit 18.15

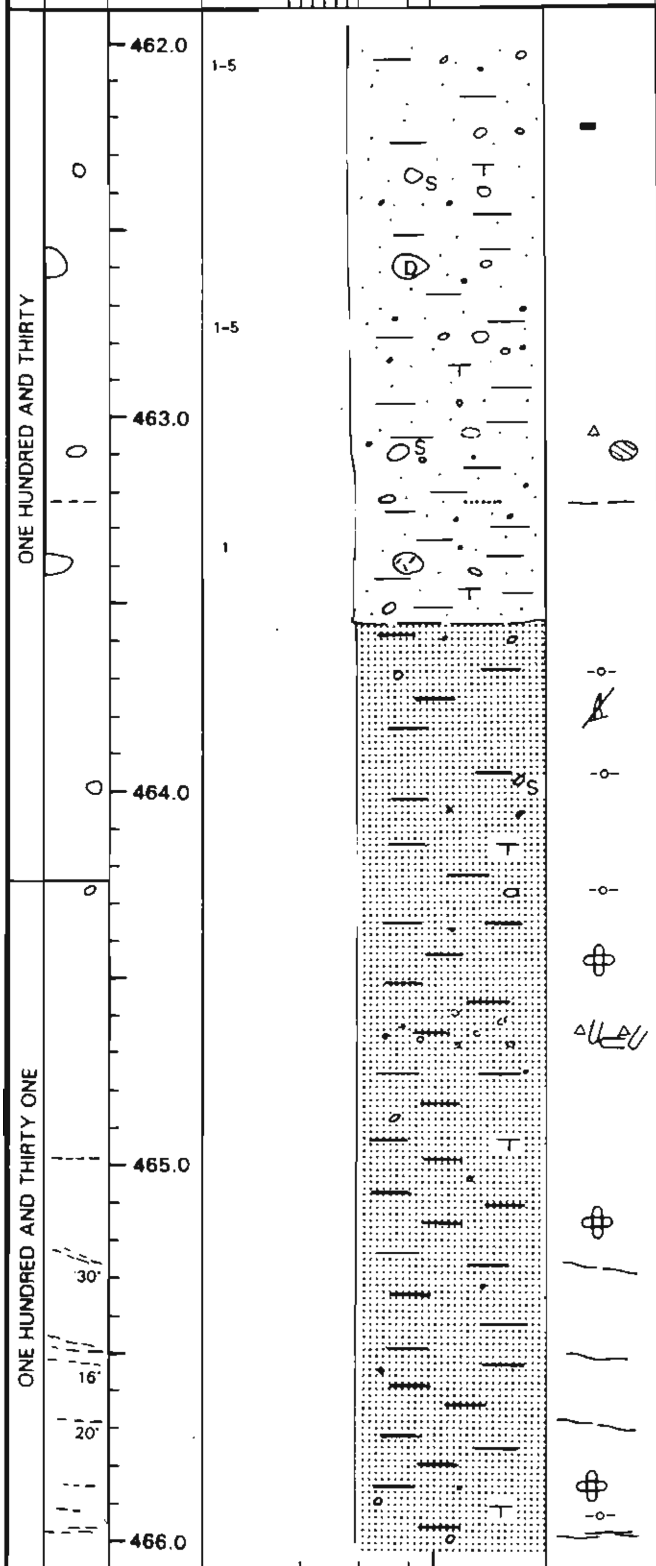
461.75 - 463.57 m

PROJECT : CIROS-1

SHEET NO. : 111

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



DIAMICTITE

Hard, unstratified to weakly stratified in places, **slightly conglomeratic muddy fine to very fine sandstone (diamictite)** (grey 7.5Y 4/1 to dark bluish grey 5B 4/1).

Clasts form 1 to 3% of the core, range up to >32 mm long and are angular to well rounded, but on average subrounded. Lithologies include dolerite, indurated sediment and basement.

Minor bioclastic debris.

—contact gradational over 3 cm –

Sub-unit 18.16

463.57 - 466.69 m

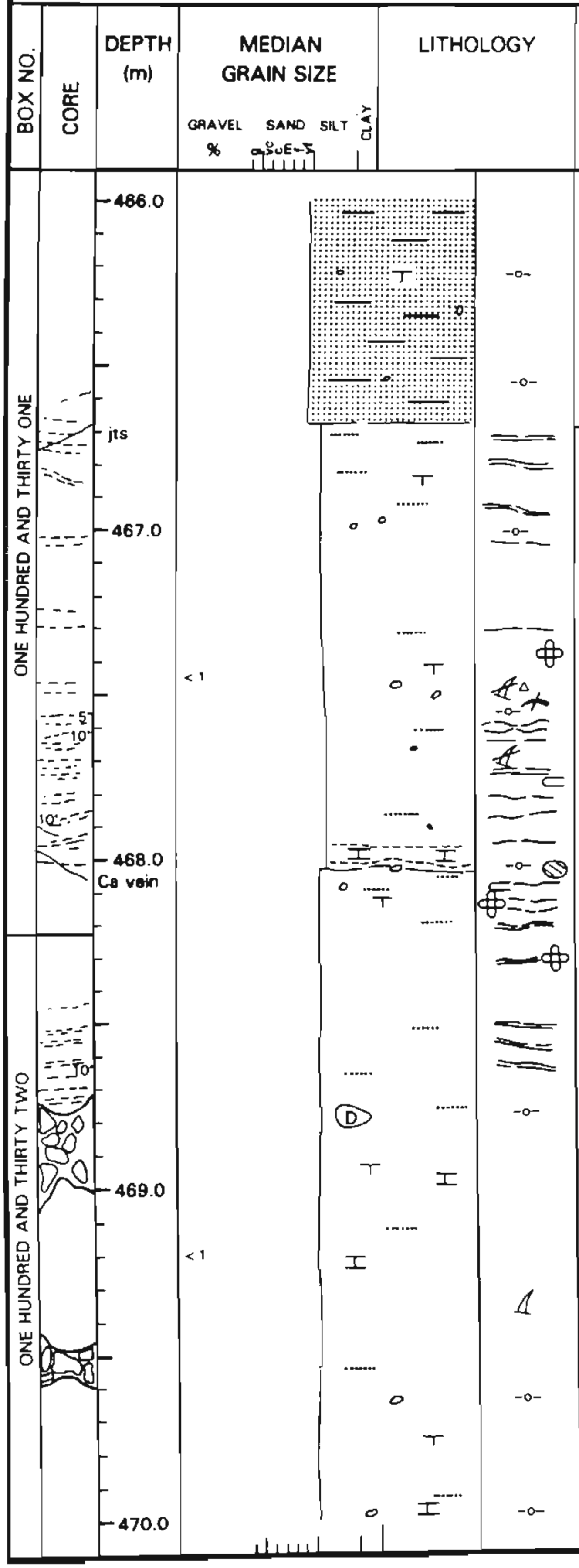
SANDSTONE

Hard, weakly stratified, poorly sorted, **muddy very fine sandstone** (grey 7.5Y 4/1). Stratification is wispy and discontinuous, and both subhorizontal and inclined (up to 30°). Much of the disruption is from burrowing. Rare (< 1%) lonestones up to 5 mm long. One striated clast A few broken scaphopods (463.80 m) and burrow moulds (464.59 to 464.70 m).

PROJECT : CIROS -1

SHEET NO. : 112

SCALE : 1:20



DESCRIPTION

— contact gradational –

Sub–unit 18.17

466.69 - 502.49 m

MUDSTONE

Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1).

Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m).

Stratification variable, including wispy, irregular, near–horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation Also larger scale (10–50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m.

Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated.

Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods.

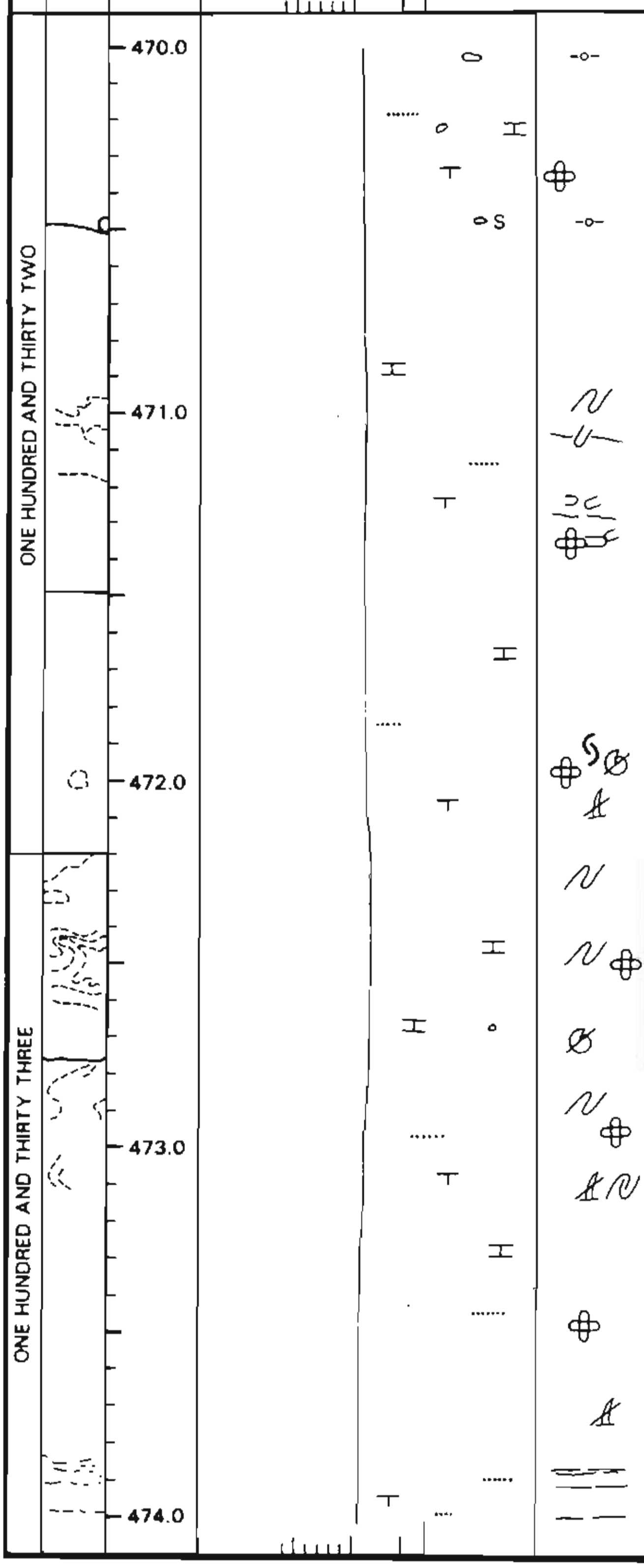
Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS -1

SHEET NO. : 113

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.17

466.69 - 502.49 m

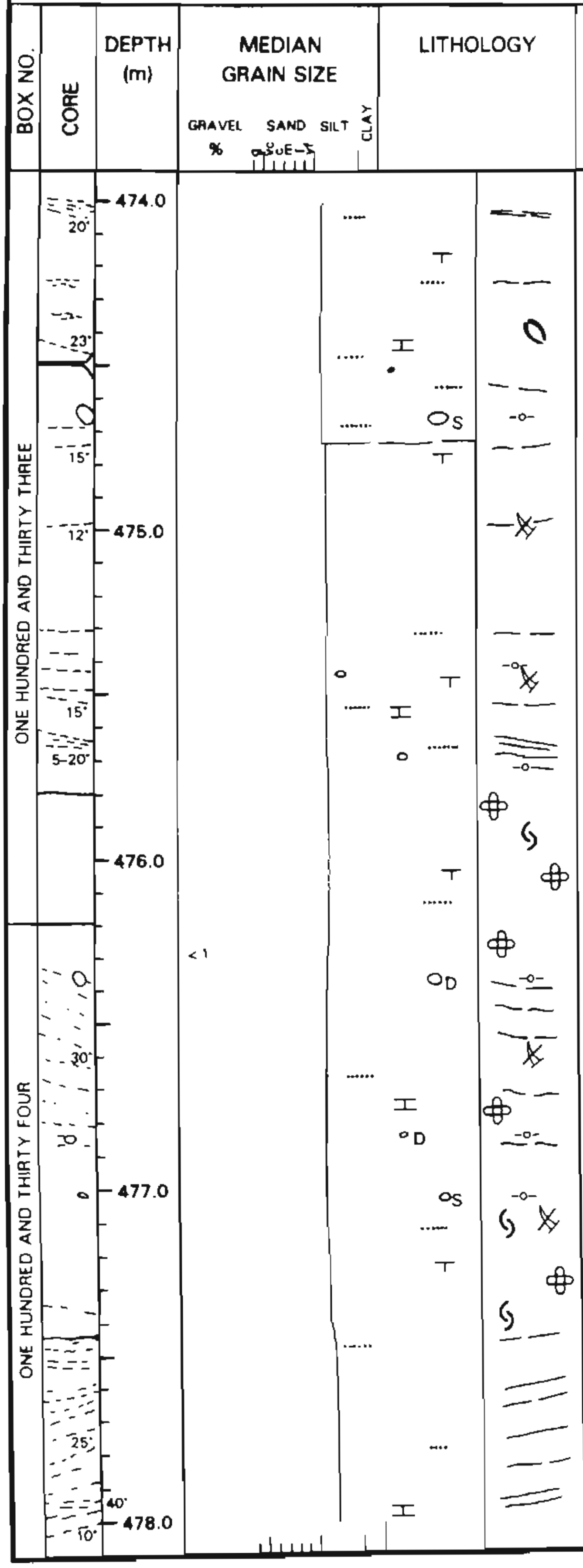
MUDSTONE

Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m) Stratification variable, including wispy, irregular, near-horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10–50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS -1

SHEET NO. : 114

SCALE : 1:20



DESCRIPTION

Sub-unit 18.17

466.69 - 502.49 m

MUDSTONE

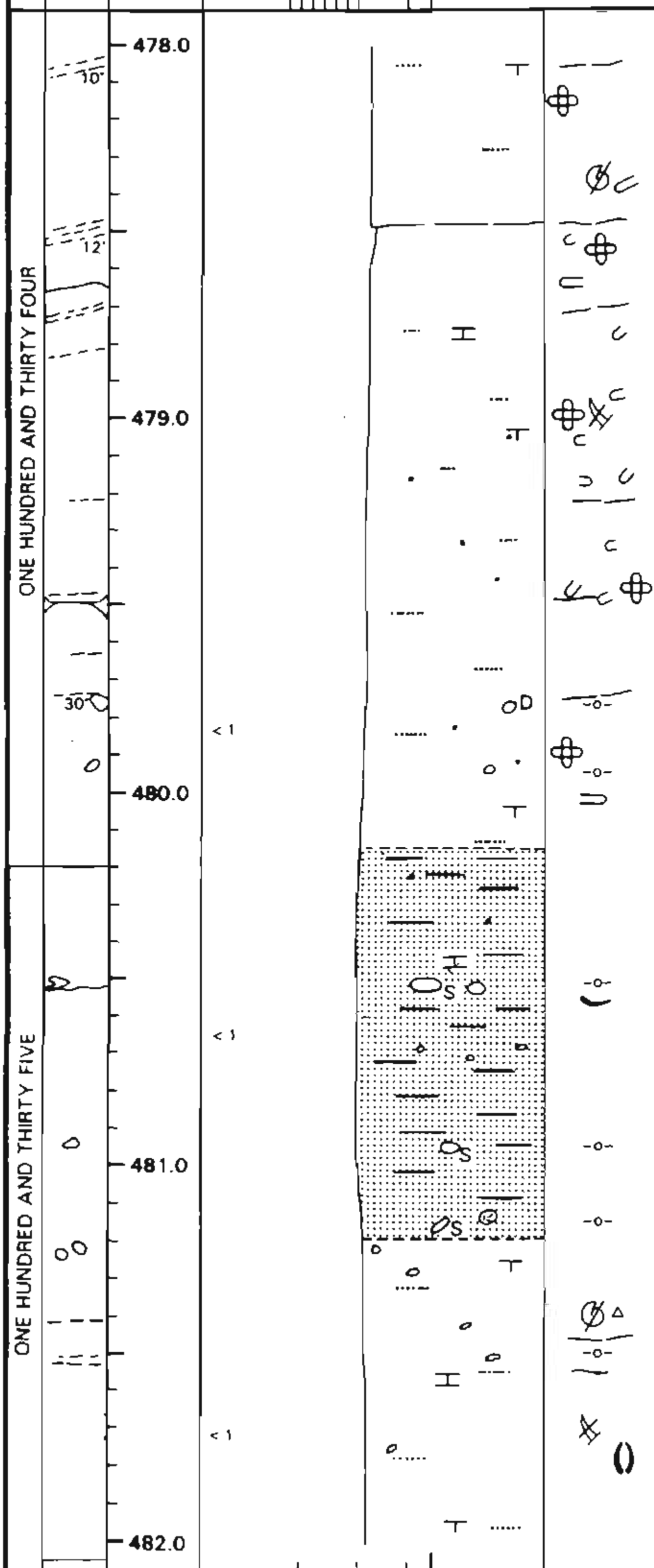
Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m). Stratification variable, including wispy, irregular, near-horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10-50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS-1

SHEET NO. : 115

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.17

466.69 - 502.49 m

MUDSTONE

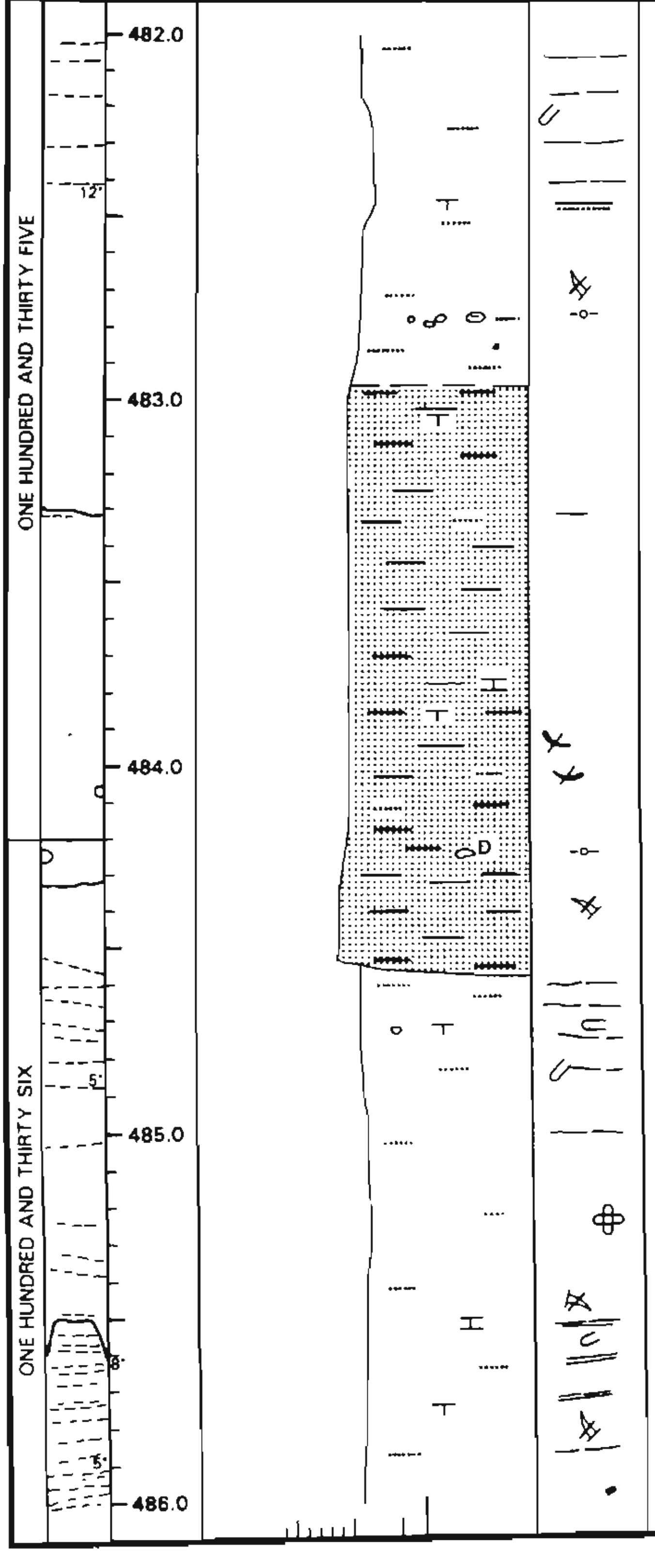
Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m). Stratification variable, including wispy, irregular, near-horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10–50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS -1

SHEET NO. : 116

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.17

466.69 - 502.49 m

MUDSTONE

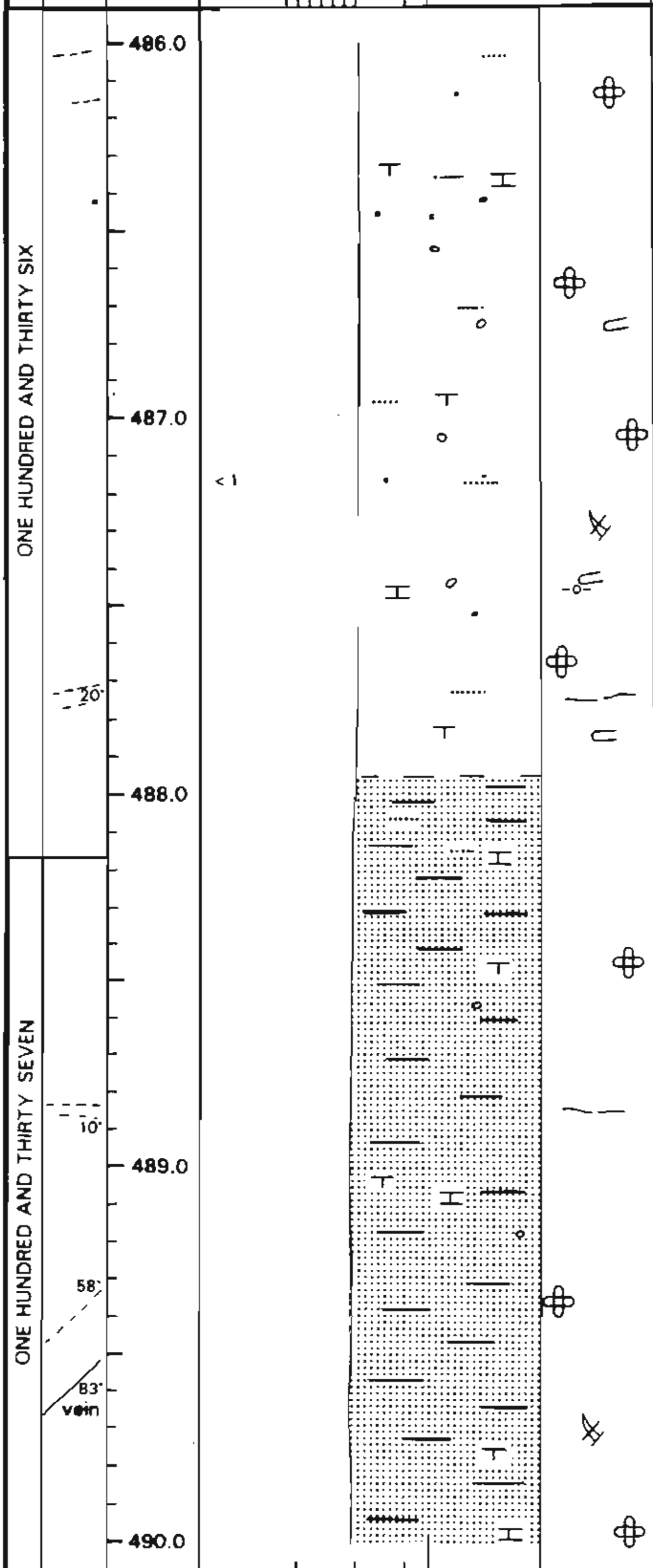
Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m). Stratification variable, including wispy, irregular, near-horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10-50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basemem. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS -1

SHEET NO. : 117

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub–unit 18.17

466.69 - 502.49 m

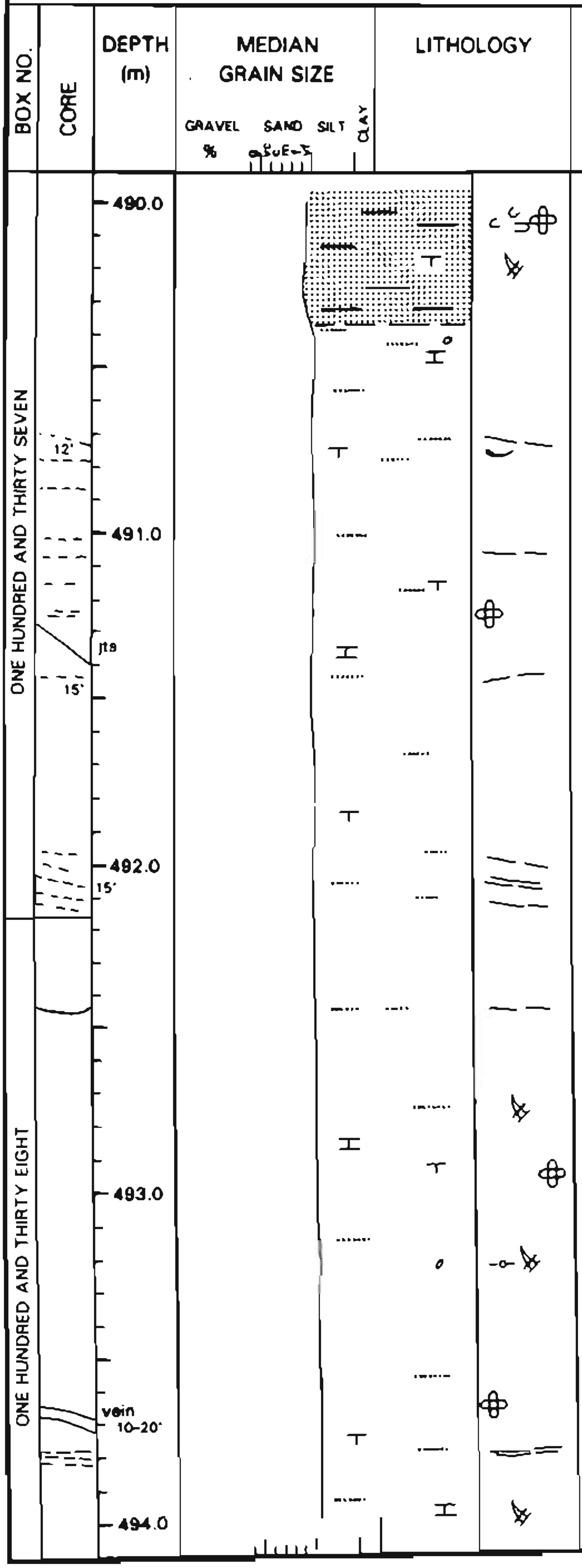
MUDSTONE

Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m). Stratification variable, including wispy, irregular, near–horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10–50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS -1

SHEET NO. : 118

SCALE : 1:20



DESCRIPTION

Sub-unit 18.17

466.69 - 502.49 m

MUDSTONE

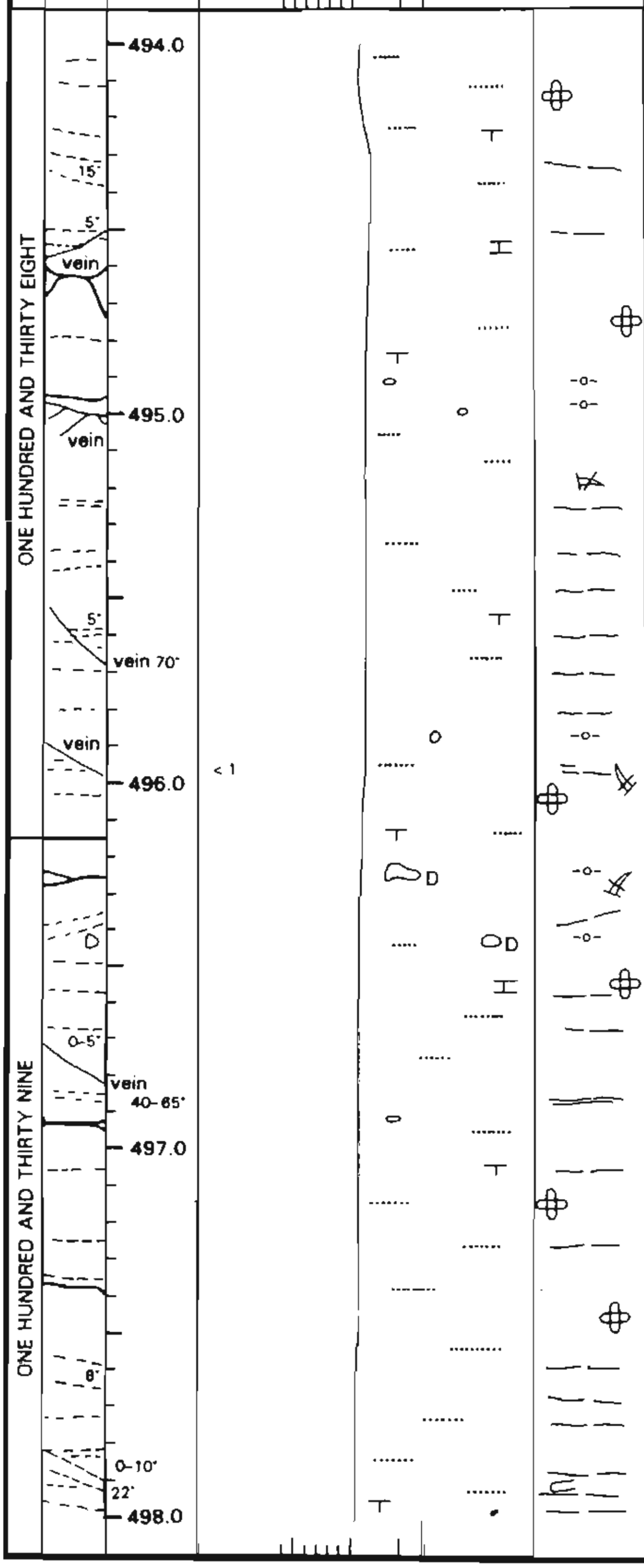
Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2 5GY 4/1) Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m). Stratification variable, including wispy, irregular, near-horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10-50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS -1

SHEET NO. : 119

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.17

466.69 - 502.49 m

MUDSTONE

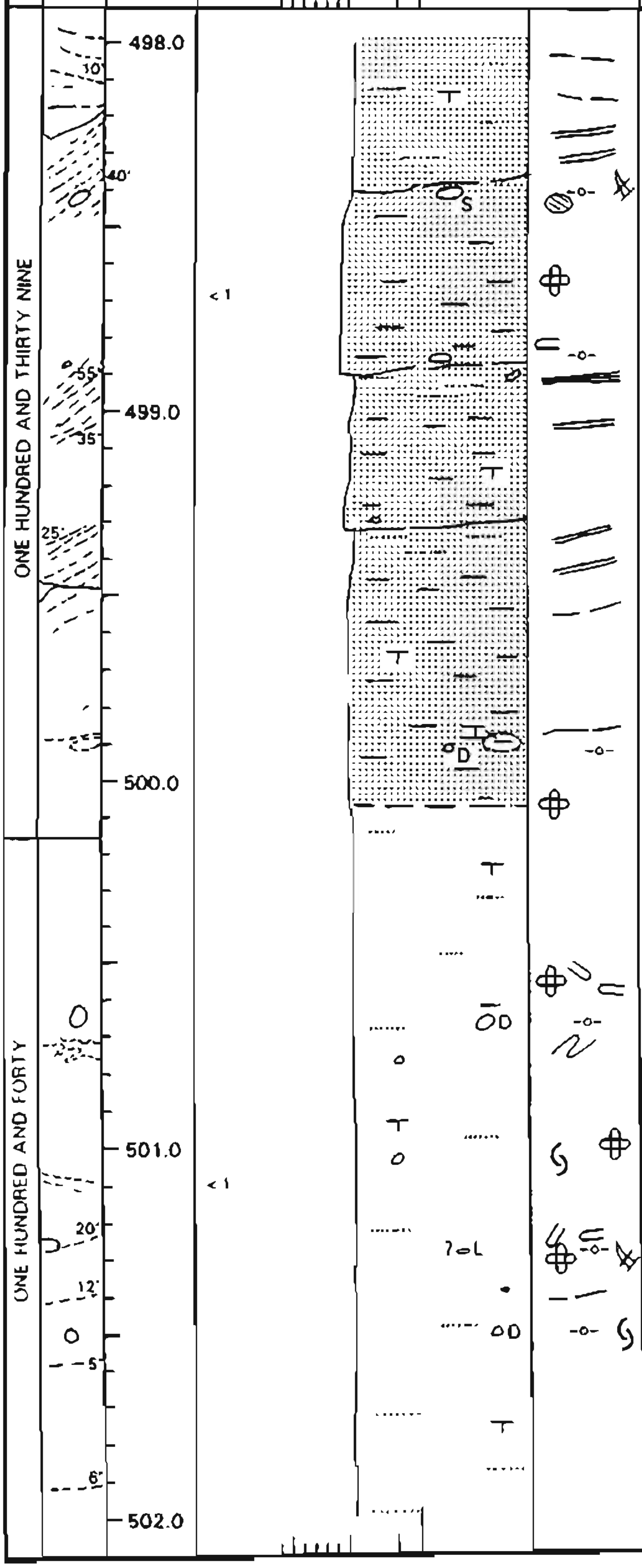
Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone** (grey 10Y 4/1) (480.15 to 481.29 m, 482.97 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m). Stratification variable, including wispy, irregular, near-horizontal and inclined (up to 35°) beds. Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10–50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS-1

SHEET NO. : 120

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.17

466.69 - 502.49 m

MUDSTONE

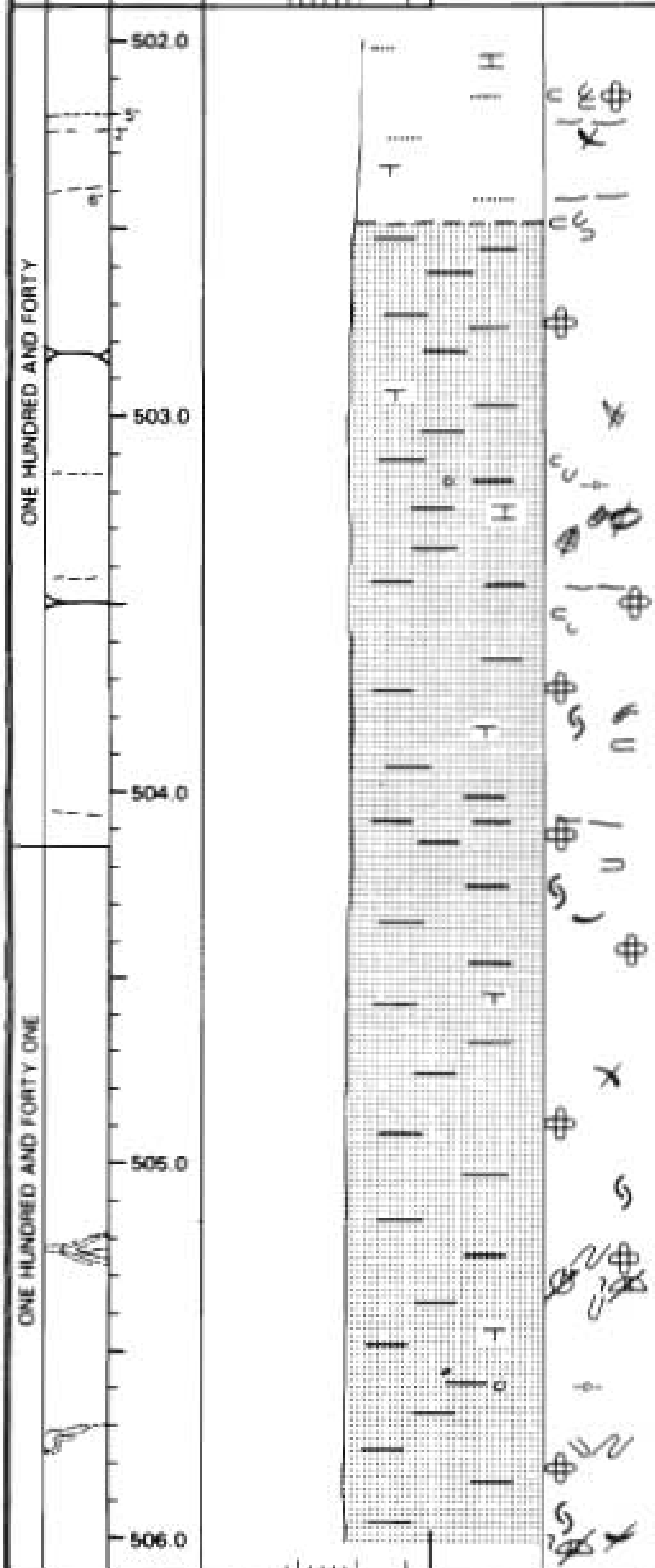
Firm to hard, slightly calcareous, unstratified to weakly stratified, **sandy mudstone** (dark olive grey 5GY 4/1 to 2.5GY 4/1). Includes intervals 1 to 2 m thick of hard, slightly calcareous, weakly to well stratified, **muddy fine to very fine sandstone**(grey 10Y 4/1) (480.15 to 481.29 m, 482.57 to 484.55 m, 487.96 to 490.38 m and 498.40 to 500.08 m) Stratification variable, including wispy, irregular, near-horizonta and inclined (up tc 35°) beds Much of the bedding has been disrupted by strong bioturbation. Also larger scale (10-50 cm) bedding contortion from 472.21 to 473.13 m and at 500.70 m. Rare (< 1%) lonestones average 5 to 10 mm (one > 57 mm) and are mostly subangular to subrounded. Lithologies are predominantly dolerite and indurated sediment with minor basement. Two clasts are striated. Widespread bioturbation is evident not only from colour mottling, but also occasional well defined burrows. Some scattered fossil shell material including articulated and broken bivalves and broken scaphopods. Thin, wispy, inclined and subhorizontal shear planes in several places.

PROJECT : CIROS-1

SHEET NO. : 121

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



— contact gradational -

Sub-unit 18.18

502.49 - 507.12 m

SANDSTONE

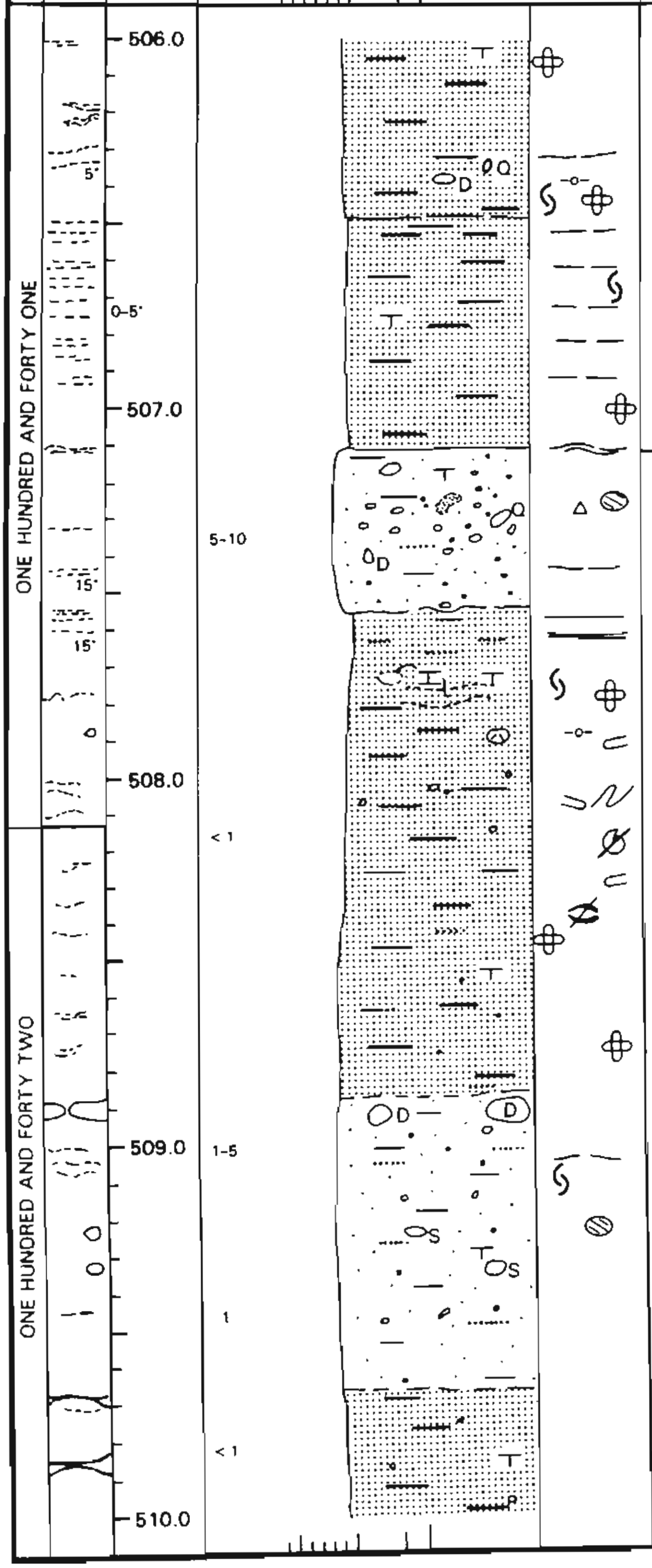
Firm to hard, unstratified to weakly stratified in places, moderately to poorly sorted, **muddy very fine sandstone** (grey 10Y 4/1). Stratification is irregular, wispy, near horizontal and inclined (up to 20°). Much bedding disruption due to bioturbation. A few small (3 to 6 mm) lonestones of dolerite and basement. Fossil material includes articulated and broken bivalves, scaphopods and echinoids.

PROJECT : CIROS -1

SHEET NO. : 122

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 18.18

502.49 - 507.12 m

SANDSTONE

Firm to hard, unstratified to weakly stratified in places, moderately to poorly sorted, **muddy very fine sandstone** (grey 10Y 4/1). Stratification is irregular, wispy, near horizontal and inclined (up to 20°). Much bedding disruption due to bioturbation. A few small (3 to 6 mm) lonestones of dolerite and basement. Fossil material includes articulated and broken bivalves, scaphopods and echinoids.

— sharp contact -

Sub-unit 18.19

507.12 - 515.08 m

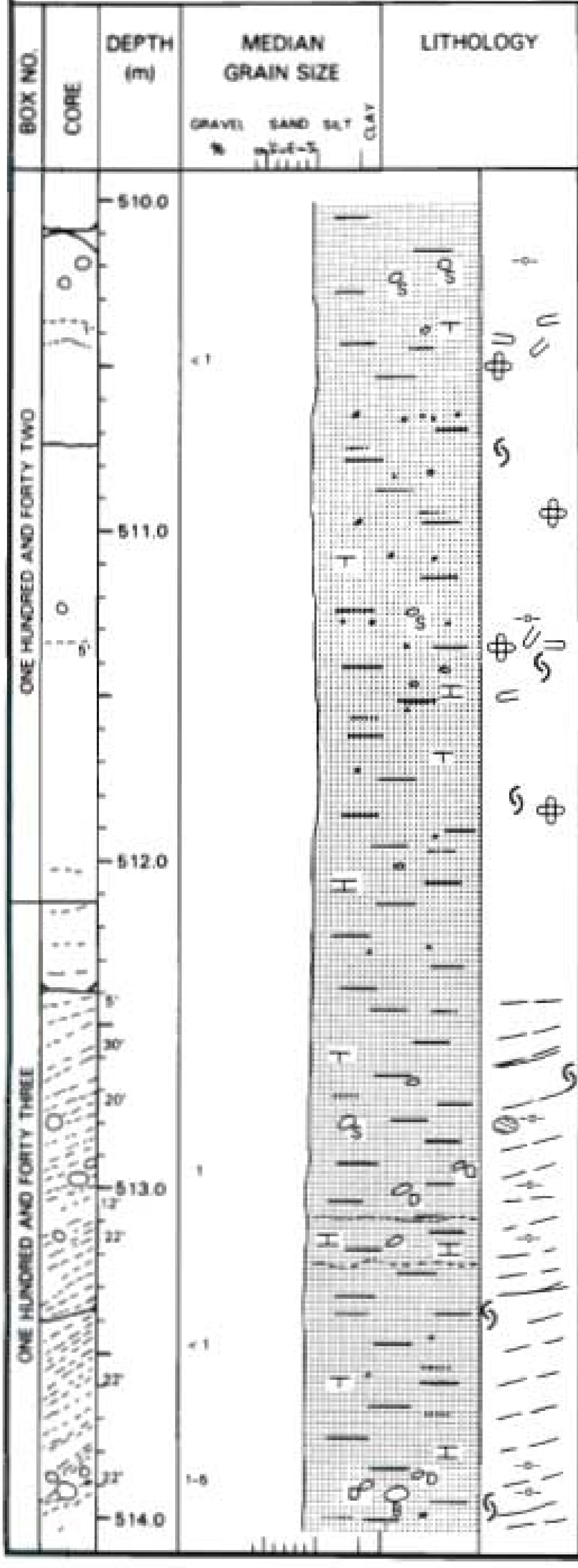
SANDSTONE AND ?DIAMICTITE

Hard, slightly calcareous (in patches), unstratified to well stratified, poorly sorted, **muddy very fine to fine sandstone** (dark olive grey 5GY 4/1 to 3/1). Stratification is generally discontinuous and wispy due to strong bioturbation. Strata from 512.42 to 515.08 m are mm- and cm-bedded, near horizontal, inclined (up to 30°) and contorted (recumbent fold at 514.90 m). Clasts occur as infrequent, subangular to subrounded lonestones. Some are striated. Most are less than 8 mm, but one Beacon sandstone clast is 25 mm. Other lithologies include dolerite, indurated sediment and basement. Shell debris, mottling and other indications of strong bioturbation are common. Two beds of hard, unstratified, **slightly conglomeratic muddy fine sandstone (diamictite)** (colour same as rest of sub-unit) from 507.12 to 507.56 m and from 508.88 to 509.67 m. Clasts form 5-10% of the core in the upper bed and 1-2% in the lower bed. Clasts are subangular to subrounded and are mainly dolerite and indurated sediment.

PROJECT : CIROS -1

SHEET NO. : 123

SCALE : 1:20



DESCRIPTION

Sub-unit 18.19

507 12 - 515.08 m

SANDSTONE AND ?DIAMICTITE

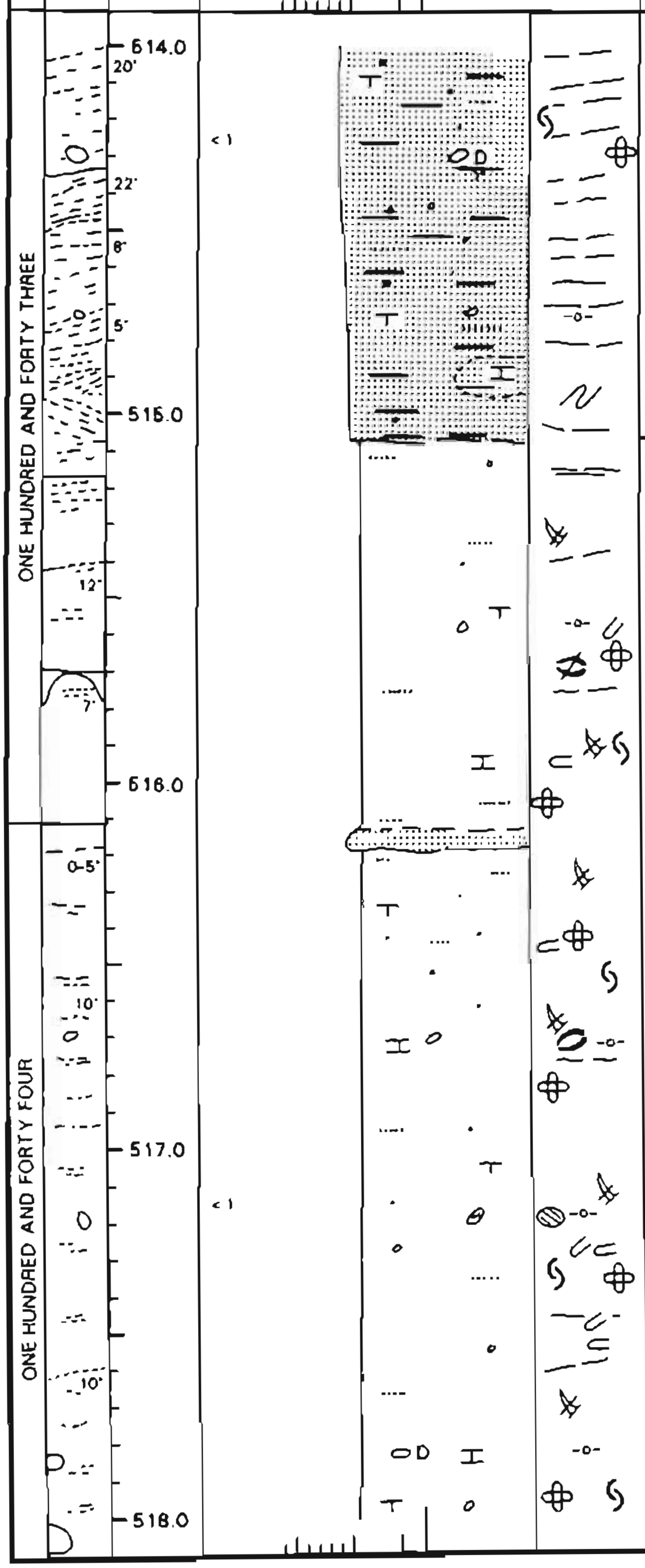
Hard, slightly calcareous (in patches), unstratified to well stratified, poorly sorted, **muddy very fine to fine sandstone** (dark olive grey 5GY 4/1 to 3/1). Stratification is generally discontinuous and wispy due to strong bioturbation. Strata from 512.42 to 515.08 m are mm- and cm-bedded, near horizontal, inclined (up to 30°) and contorted (recumbent fold at 514.90 m). Clasts occur as infrequent, subangular to subrounded lonestones. Some are striated. Most are less than 8 mm, but one Beacon sandstone clast is 25 mm. Other lithologies include dolerite, indurated sediment and basement. Shell debris, mottling and other indications of strong bioturbation are common. Two beds of hard, unstratified, **slightly conglomeratic muddy fine sandstone (diamictite)** (colour same as rest of sub-unit) from 507.12 to 507.56 m and from 508.88 to 509.67 m. Clasts form 5-10% of the core in the upper bed and 1-2% in the lower bed. Clasts are subangular to subrounded and are mainly dolerite and indurated sediment.

PROJECT : CIROS -1

SHEET NO. : 124

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % · SAND · SILT · CLAY)	LITHOLOGY	DESCRIPTION



— contact gradational over 2 cm -

Sub–unit 18.20

515.08 - 525.71 m

MUDSTONE

Hard, slightly calcareous, weakly and well stratified, **sandy mudstone** (grey 10Y 4/1 with mottling of light grey 7.5Y 7/1).

The mudstone grades into poorly sorted, **muddy very fine sandstone** from 519.81 to 520.15 m and 522.14 to 522.48 m

Stratification is generally wispy, irregular and discontinuous due to strong bioturbation. It is variably subhorizontal and inclined (up to 20) from 521 03 to 522.14 m, with dislocated and contorted bedding between 522 48 and 523 29 m

Clasts are rare (< 1%), but range from 5 to > 42 mm long, and are composed of indurated sediment, dolerite and basement One clast is striated. A basalt clast at 522.58 m has weathered feldspars. A limestone intraclast at 523 30 m has a diffuse margin.

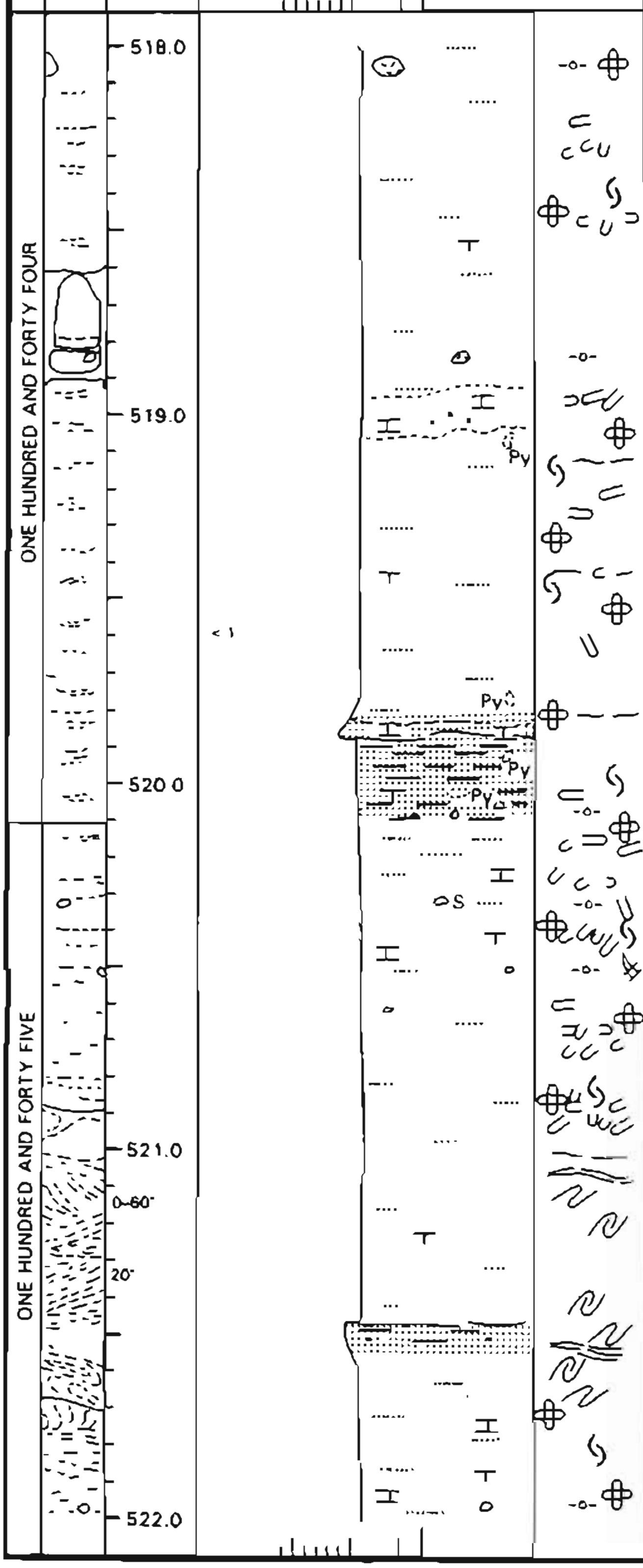
Bioturbation is strong with individual horizontal and inclined burrows recognized. Common sizes are 1 mm, 2 to 5 mm and around 15 mm Skeletal material includes articulated whole and broken bivalves, scaphopods and a lot of smaller indeterminate debris A few pyrite nodules

PROJECT : CIROS -1

SHEET NO. : 125

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 18.20

515.08 - 525 71 m

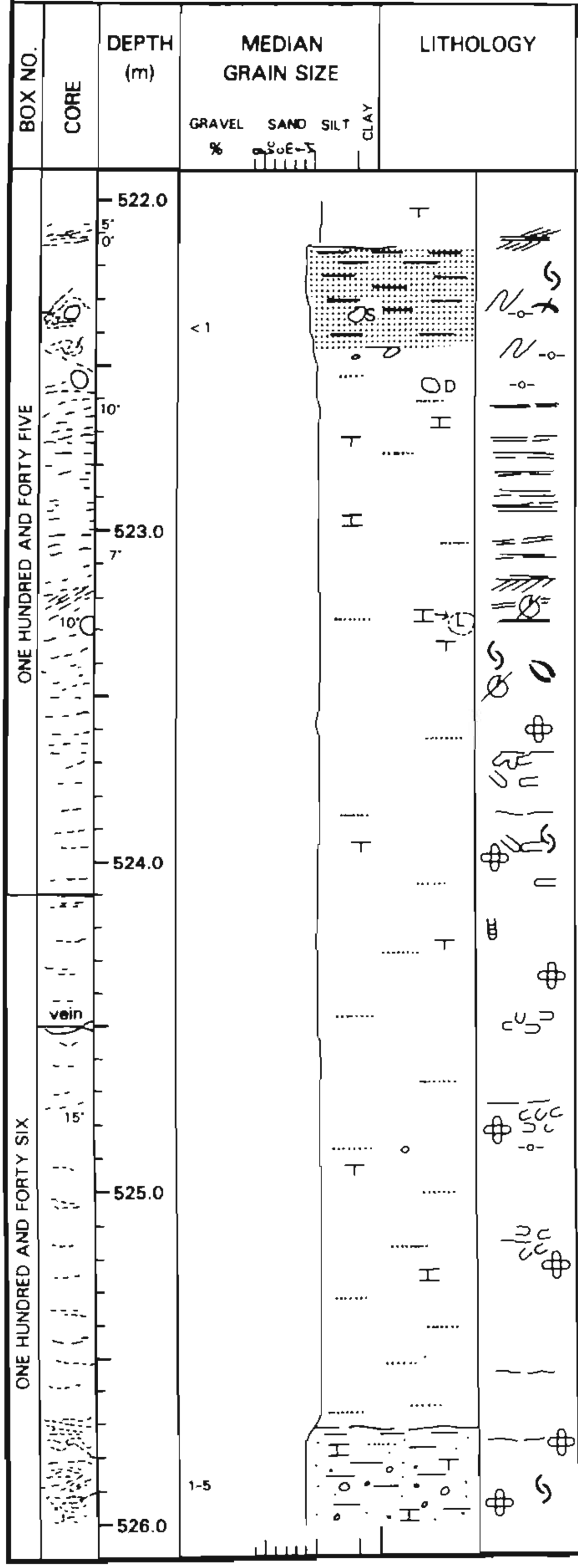
MUDSTONE

Hard, slightly calcareous, weakly and well stratified, **sandy mudstone** (grey 10Y 4/1 with mottling of light grey 7.5Y 7/1) The mudstone grades into poorly sorted, **muddy very fine sandstone** from 519.81 to 520.15 m and 522 14 to 522.48 m. Stratification is generally wispy, irregular and discontinuous due to strong bioturbation. It is variably subhorizontal and inclined (up to 20°) from 521.03 to 522 14 m, with dislocated and contorted bedding between 522.48 and 523.29 m Clasts are rare (< 1%), but range from 5 to > 42 mm long, and are composed of indurated sediment, dolerite and basement One clast is striated A basalt clast at 522 56 m has weathered feldspars A limestone intraclast at 523.30 m has a diffuse margin. Bioturbation is strong with individual horizontal and inclined burrows recognized. Common sizes are 1 mm, 2 to 5 mm and around 15 mm. Skeletal material includes articulated whole and broken bivalves, scaphopods and a lot of smaller indeterminate debris. A few pyrite nodules.

PROJECT : CIROS -1

SHEET NO. : 126

SCALE : 1:20



DESCRIPTION

Sub-unit 18.20

515.08 - 525.71 m

MUDSTONE

Hard, slightly calcareous, weakly and well stratified, **sandy mudstone** (grey 10Y 4/1 with mottling of light grey 7.5Y 7/1). The mudstone grades into poorly sorted, **muddy very fine sandstone** from 519.81 to 520.15 m and 522.14 to 522.48 m. Stratification is generally wispy, irregular and discontinuous due to strong bioturbation. It is variably subhorizontal and inclined (up to 20°) from 521.03 to 522.14 m, with dislocated and contorted bedding between 522.48 and 523.29 m. Clasts are rare (< 1%), but range from 5 to > 42 mm long, and are composed of indurated sediment, dolerite and basement. One clast is striated. A basalt clast at 522.56 m has weathered feldspars. A limestone intraclast at 523.30 m has a diffuse margin. Bioturbation is strong with individual horizontal and inclined burrows recognized. Common sizes are 1 mm, 2 to 5 mm and around 15 mm. Skeletal material includes articulated whole and broken bivalves, scaphopods and a lot of smaller indeterminate debris. A few pyrite nodules.

— contact gradational over 5 cm —

Sub-unit 18.21

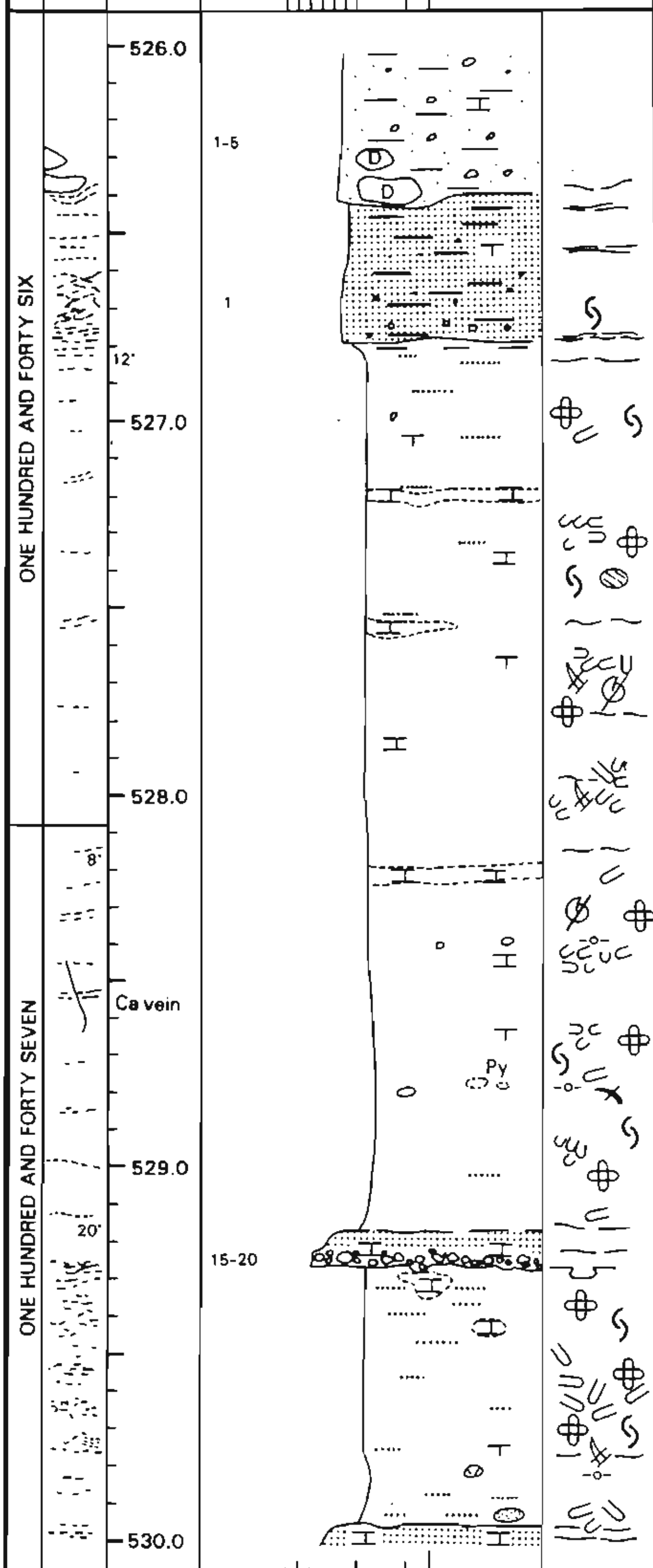
525.71 - 526.79 m

PROJECT : CIROS -1

SHEET NO. : 127

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



DIAMICTITE AND SANDSTONE

Hard, calcareous, unstratified, **conglomeratic muddy fine to very fine sandstone** (diamictite)(olive black 7.5Y 3/1 with grey 7.5Y 6/1 mottle).

Between 526.41 and 526.89 m is a hard, weakly stratified, **slightly conglomeratic muddy fine to very fine sandstone** (colour same as rest of sub-unit).

The sandstone has wispy, irregular, subhorizontal bedding. Stratification has been destroyed by bioturbation.

Clasts in the diamictite represent <5% of the core and are composed of dolerite and basement. Two basalt clasts (>32 mm) mark the diamictite - sandstone boundary. The sandstone has scattered (<1%) small pebbles (<8 mm long).

- sharp contact -

Sub-unit 18.22

526.79 - 529.14 m

MUDSTONE

Hard, calcareous, weakly stratified, **mudstone** (olive black 7.5Y 3/1)

Sandy in places.

Stratification is discontinuous, diffuse and very irregular. Bioturbation structures are common, along with individual burrows (1 mm and 2 to 5 mm across).

One striated lonestone at 527.42 m.

Scaphopod and other shell debris present.

- contact gradational over 2 cm -

Sub-unit 19.1

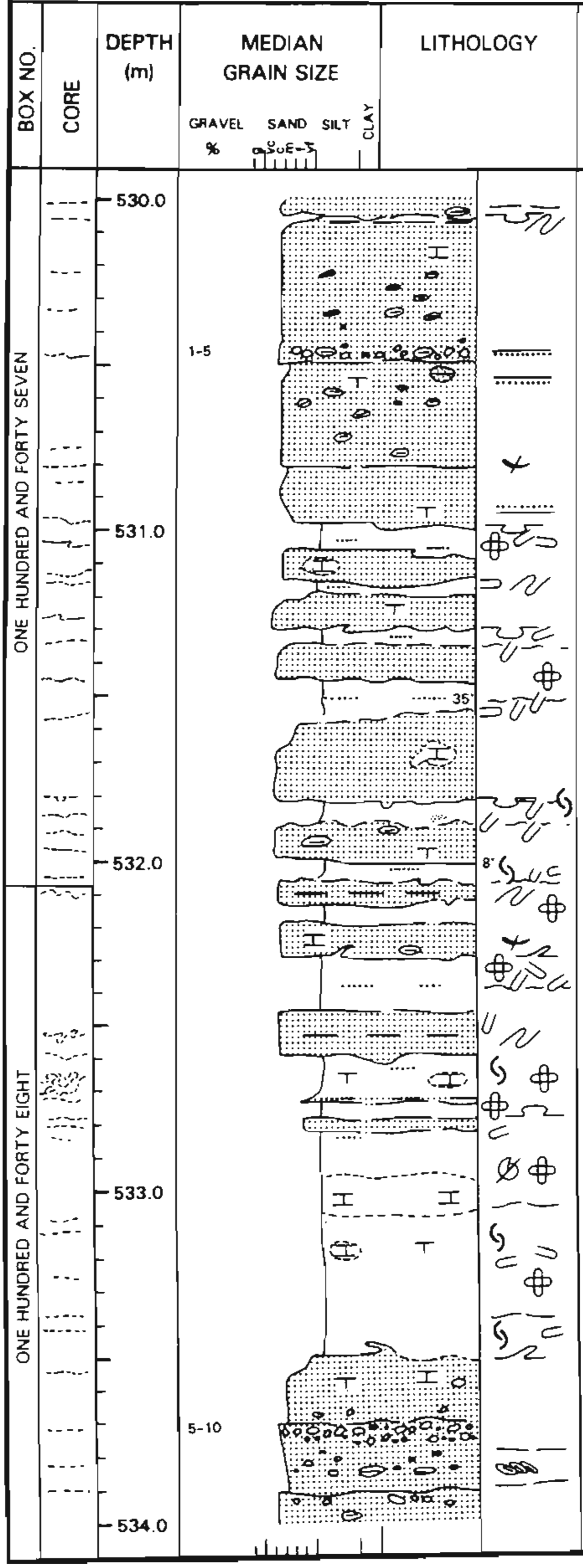
529.14 - 539.28 m

MUDSTONE AND SANDSTONE

PROJECT : CIROS -1

SHEET NO. : 128

SCALE : 1:20



DESCRIPTION

Sub-unit 19.1

529.14 - 539.28 m

MUDSTONE AND SANDSTONE

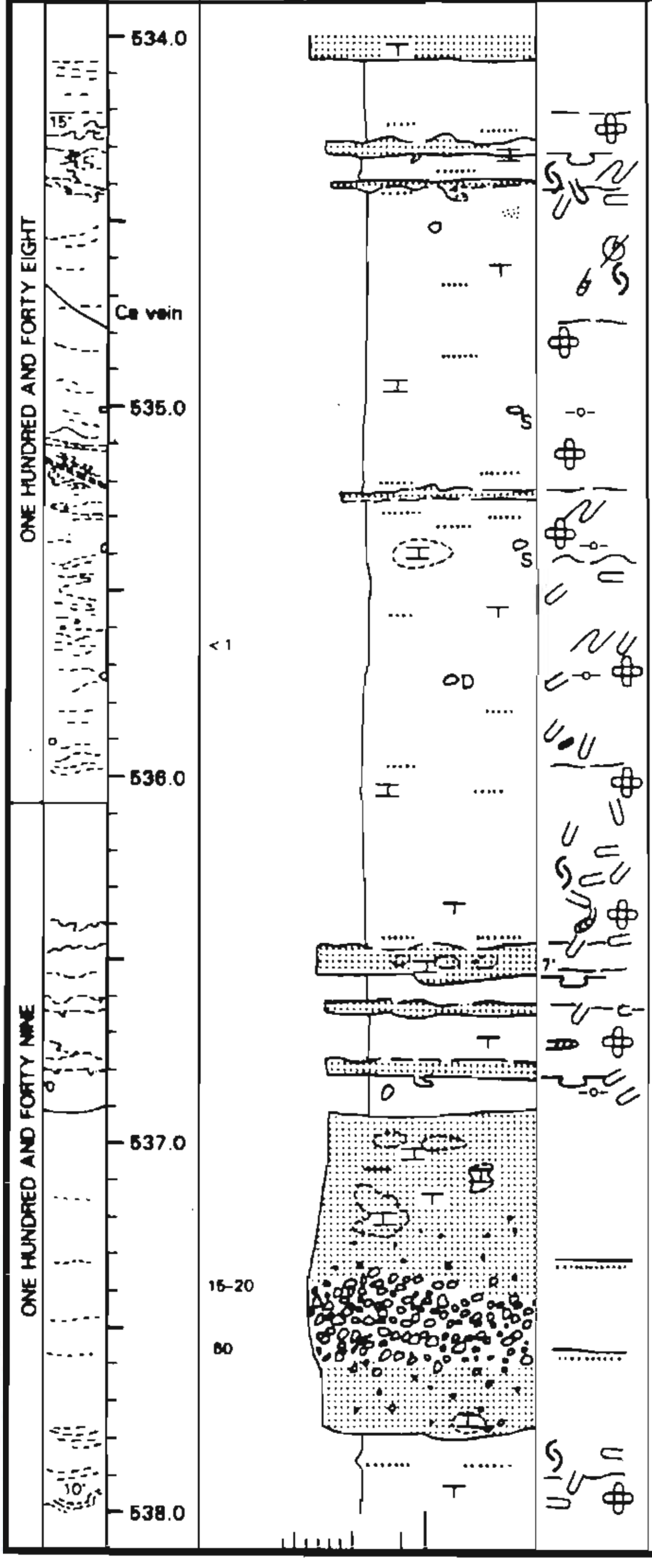
Beds from 0.05 to 1.60 m thick of moderately hard, slightly calcareous, **sandy mudstone** (olive black 7.5Y 3/1 with mottles of light grey 10Y 7/1) interbedded with beds from 0.07 to 1.00 m thick of unstratified and weakly stratified, moderately to poorly sorted, **coarse to medium sandstone** (grey N 6/1). **Pebble** (4–16 mm) **conglomerates** at 529.23, 533.71 and 537.37 to 537.60 m. Stratification is wispy, discontinuous, subhorizontal and inclined (up to 10°), with some contorted and graded bedding. Load and flame structures occur at the base and top of sandstone beds. Bedding is much disrupted by bioturbation. Some sandstones contain thin pebble beds with subangular to subrounded clasts up to 17 mm. Lithologies are basement (60%), indurated sediment (30%) and others (10%) including mudstone and sandstone intraclasts. Weak mudstone clast imbrication at 533.85 m. A few lonestones are scattered throughout the rest of the sub-unit. Most are less than 10 mm long and consist of indurated sediment. Biogenic features include spreiten-filled burrows, skeletal debris and carbonaceous fragments.

PROJECT : CIROS -1

SHEET NO. : 129

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.1

529.14 - 539.28 m

MUDSTONE AND SANDSTONE

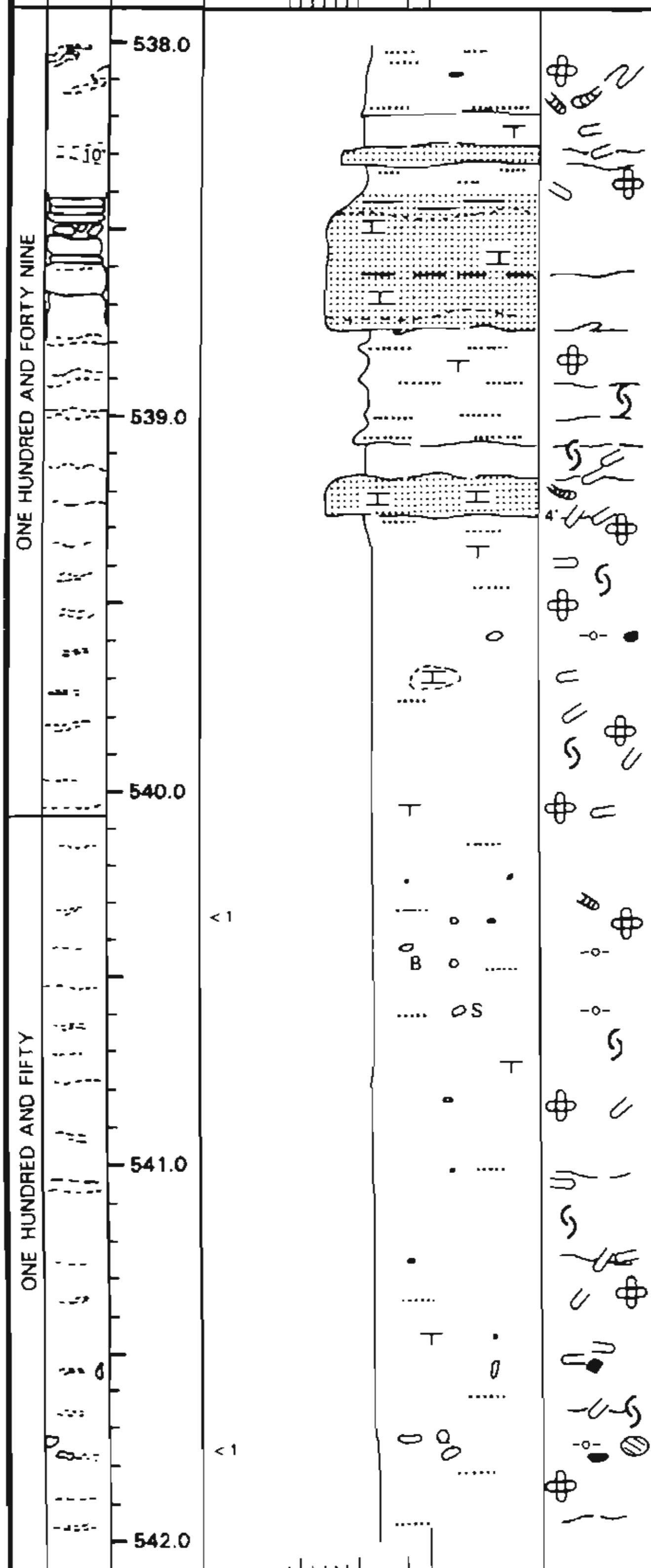
Beds from 0.05 to 1.60 m thick of moderately hard, slightly calcareous, **sandy mudstone** (olive black 7.5Y 3/1 with mottles of light grey 10Y 7/1) interbedded with beds from 0.07 to 1.00 m thick of unstratified and weakly stratified, moderately to poorly sorted, **coarse to medium sandstone** (grey N 6/1). **Pebble** (4-16 mm) **conglomerates** at 529.23, 533.71 and 537.37 to 537.60 m. Stratification is wispy, discontinuous, subhorizontal and inclined (up to 10°), with some contorted and graded bedding. Load and flame structures occur at the base and top of sandstone beds. Bedding is much disrupted by bioturbation. Some sandstones contain thin pebble beds with subangular to subrounded clasts up to 17 mm. Lithologies are basement (60%), indurated sediment (30%) and others (10%) including mudstone and sandstone intraclasts. Weak mudstone clast imbrication at 533.85 m. A few lonestones are scattered throughout the rest of the sub-unit. Most are less than 10 mm long and consist of indurated sediment. Biogenic features include spreiten-filled burrows, skeletal debris and carbonaceous fragments.

PROJECT : CIROS -1

SHEET NO. : 130

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



— contact sharp –

Sub-unit 19.2

539.28 - 544.93 m

MUDSTONE

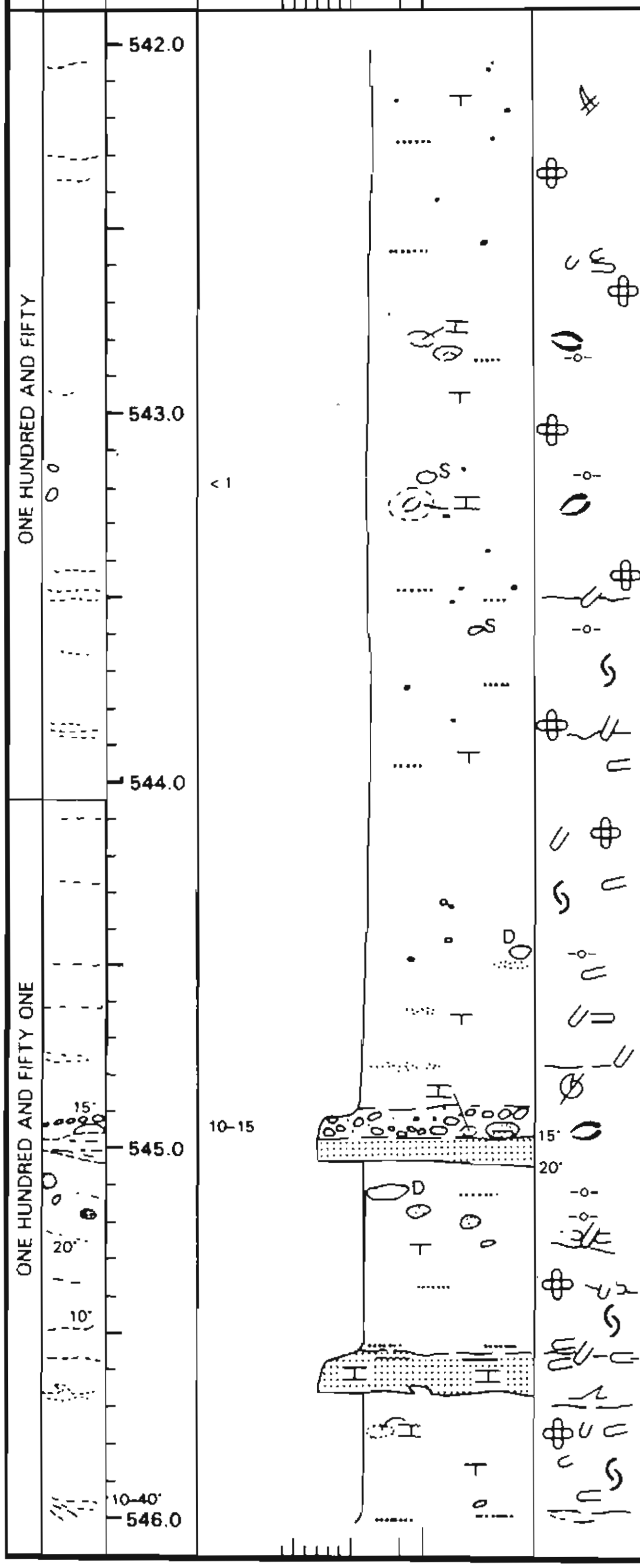
Hard, weakly stratified **sandy mudstone** (dark greenish grey 5G 4/1).
Stratification is wispy and discontinuous and the core strongly mottled due to bioturbation.
Lonestones are rare and mostly less than 6mm long. One is striated.
Lithologies are mainly indurated sediment, but there are also a number of carbonaceous mudstone clasts.
Biogenic features include horizontal and inclined burrows, and broken scaphopod and bivalve shells.

PROJECT : CIROS-1

SHEET NO. : 131

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.2

539.28 - 544.93 m

MUDSTONE

Hard, weakly stratified **sandy mudstone** (dark greenish grey 5G 4/1). Stratification is wispy and discontinuous and the core strongly mottled due to bioturbation. Lonestones are rare and mostly less than 6mm long. One is striated. Lithologies are mainly indurated sediment, but there are also a number of carbonaceous mudstone clasts. Biogenic features include horizontal and inclined burrows, and broken scaphopod and bivalve shells.

— contact sharp -

Sub-unit 19.3

544.93 - 551.98 m

MUDSTONE AND SANDSTONE

Beds from 0.14 to 1.20 m thick of hard, weakly stratified and strongly bioturbated, **sandy mudstone** (dark bluish grey 5B 4/1 to grey 5Y 6/1 where calcareous), interbedded with beds from 0.10 to 0.40 m thick of soft to firm, unstratified and weakly stratified, moderately to poorly sorted, **coarse to medium sandstone** (grey 5Y 5/1).

Stratification in both lithologies is wispy, discontinuous and irregular. Many beds have a consistent 5 to 10° dip. Bedding in the mudstone is disrupted by bioturbation which is reflected in the mottling and evident from the vertical and inclined burrows, one of which has spreiten.

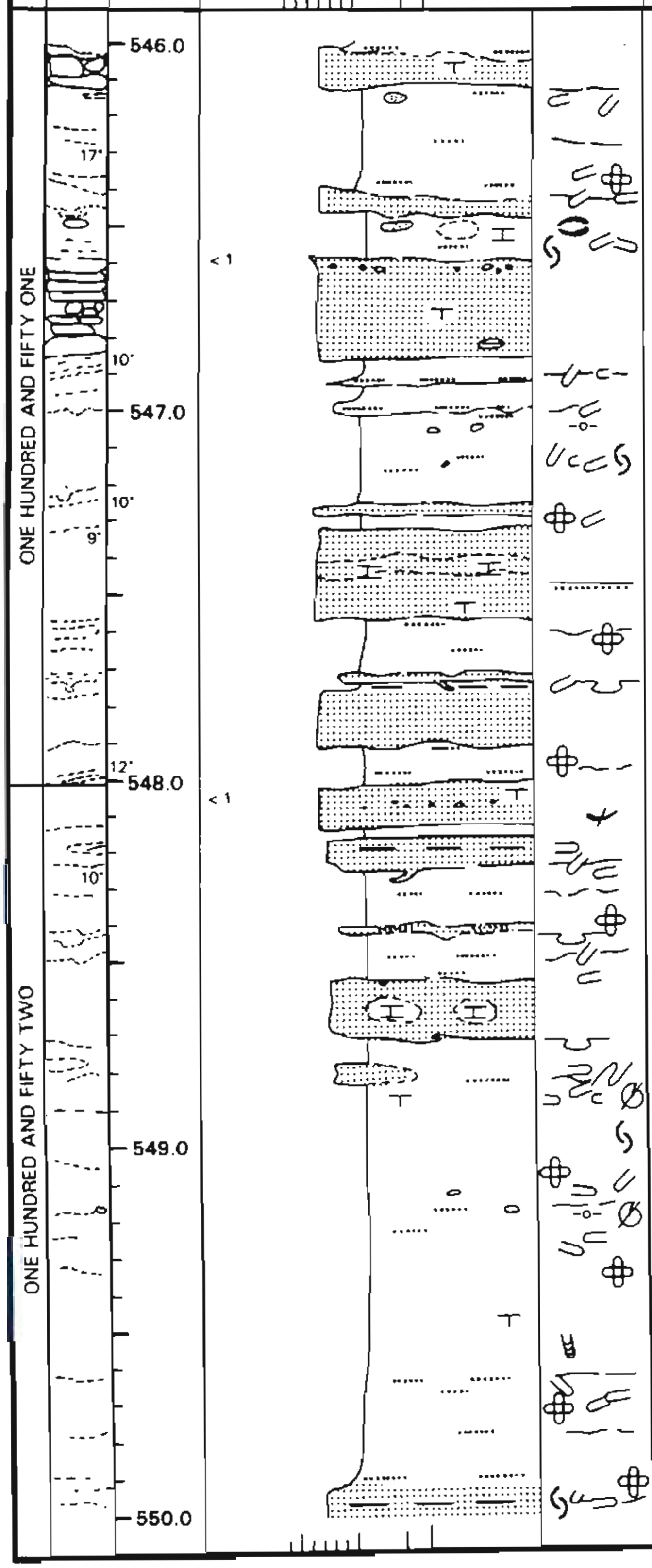
The sandstone beds mostly have sharp bases and irregular, gradational, bioturbated tops. Many of the bases show load structures. The bed from

PROJECT : CIROS -1

SHEET NO. : 132

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



550.67 to 551.08 m coarsens upwards.
Lonestones are rare (<1%) and mostly less than 8 mm though there are one or two dolerite and sandstone intraclasts between 12 and >34 mm.

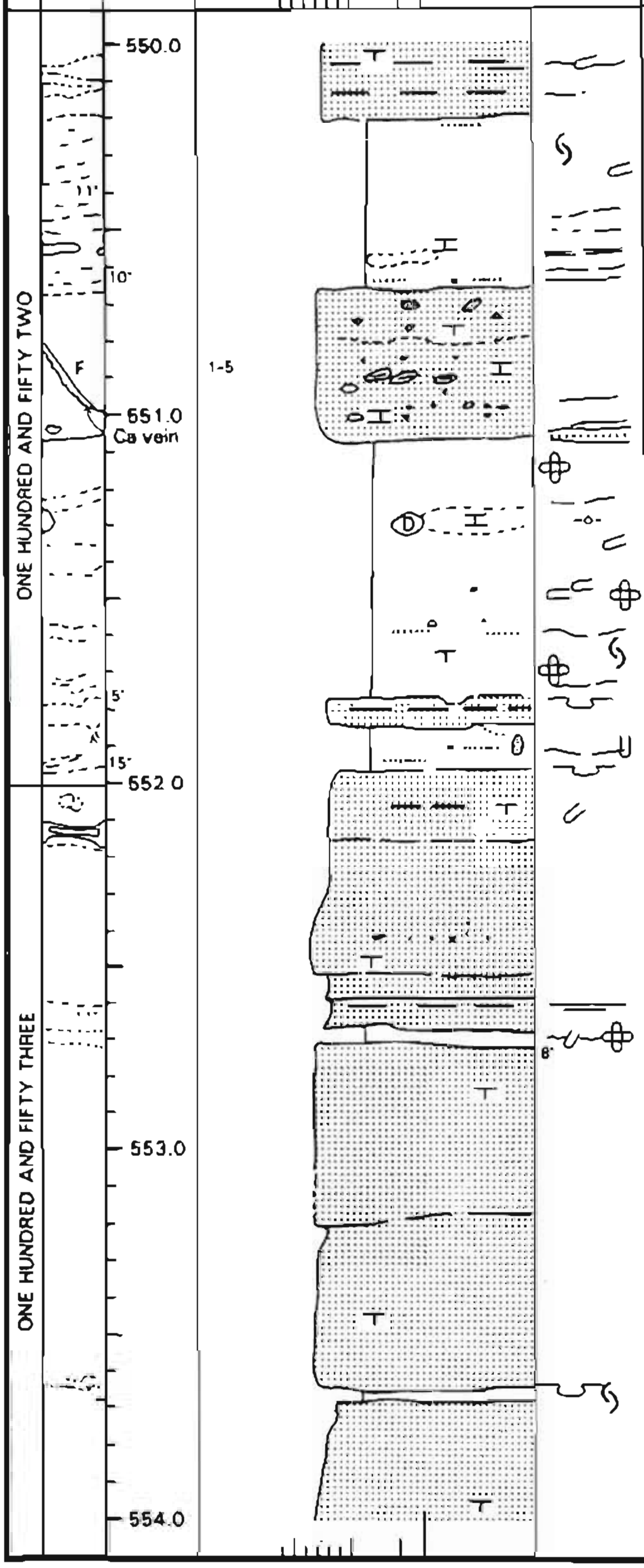
Some articulated and broken bivalve fragments.

PROJECT : CIROS-1

SHEET NO. : 133

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.3

544.93 - 551.98 m

MUDSTONE AND SANDSTONE

Beds from 0 14 to 1 20 m thick of hard, weakly stratified and strongly bioturbated, **sandy mudstone** (dark bluish grey 5B 4/1 to grey 5Y 6/1 where calcareous), interbedded with beds from 0.10 to 0.40 m thick of soft to firm, unstratified and weakly stratified, moderately to poorly sorted, **coarse to medium sandstone** (grey 5Y 5/1). Stratification in both lithologies is wispy, discontinuous and irregular. Many beds have a consistent 5 to 10° dip. Bedding in the mudstone is disrupted by bioturbation which is reflected in the mottling and evident from the vertical and inclined burrows, one of which has spreiten. The sandstone beds mostly have sharp bases and irregular, gradational, bioturbated tops. Many of the bases show load structures. The bed from 550.67 to 551.08 m coarsens upwards. Lonestones are rare (< 1%) and mostly less than 8 mm though there are one or two dolerite and sandstone intraclasts between 12 and > 34 mm. Some articulated and broken bivalve fragments.

— contact sharp -

Sub-unit 19.4

551.98 - 555.70 m

SANDSTONE

Firm to hard, unstratified to weakly stratified in places, poorly sorted, (occasionally) slightly **conglomeratic coarse to fine sandstone** (grey 5Y 5 1 to light grey 5Y 7 1 and dull yellow brown 10YR 5 3 from 553 00 to 553 40 m)

Several mottled and bioturbated mudstone beds 2 to 5 cm thick

Stratification dips consistently between 5 and 15 . Some beds first coarsen and then fine upwards. Base of sandstones occasionally show load structures

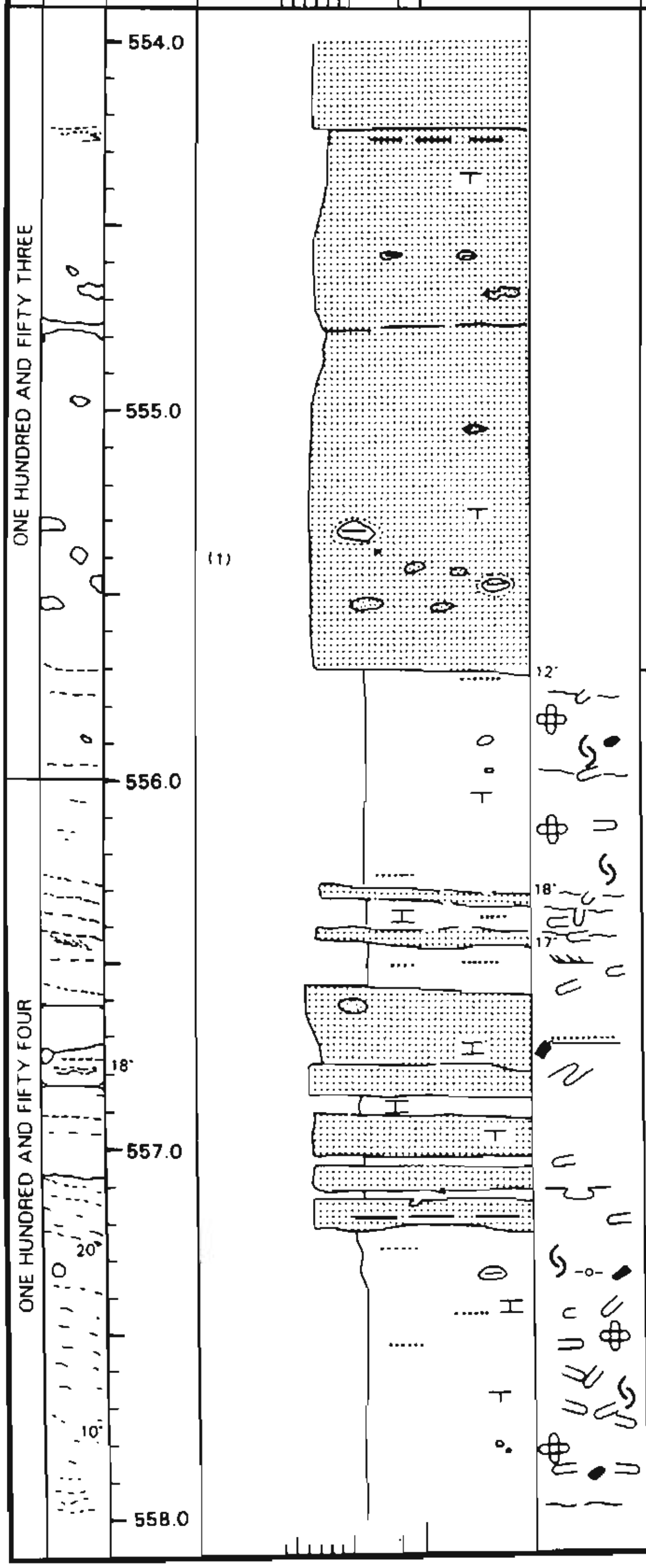
Mudstone and sandstone intraclasts below 554 50 m

PROJECT : CIROS -1

SHEET NO. : 134

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.4

551.98 - 555.70 m

SANDSTONE

Firm to hard, unstratified to weakly stratified in places, poorly sorted, (occasionally) **slightly conglomeratic coarse to fine sandstone** (grey 5Y 5/1 to light grey 5Y 7/1 and dull yellow brown 10YR 5/3 from 553.00 to 553.40 m). Several mottled and bioturbated mudstone beds 2 to 5 cm thick. Stratification dips consistently between 5 and 15°. Some beds first coarsen and then fine upwards. Base of sandstones occasionally show load structures. Mudstone and sandstone intraclasts below 554.50 m.

— contact sharp -

Sub-unit 19.5

555.70 - 558.27 m

MUDSTONE AND SANDSTONE

Hard, weakly stratified, **sandy mudstone** (grey 10Y 5/1 to 4/1) in beds 0.03 to 0.90 m thick alternate with several thin (0.05 to 0.25 m) beds of hard, slightly calcareous, unstratified, poorly sorted, **coarse to medium sandstone** (greyish olive 7.5Y 5/2).

Stratification in the mudstone is wispy, irregular and has dips varying up to 20°. Stratification is disrupted by strong bioturbation. Sandstone bases show load features, and one bed is reverse graded from medium to very coarse sandstone from 556.56 to 556.78 m.

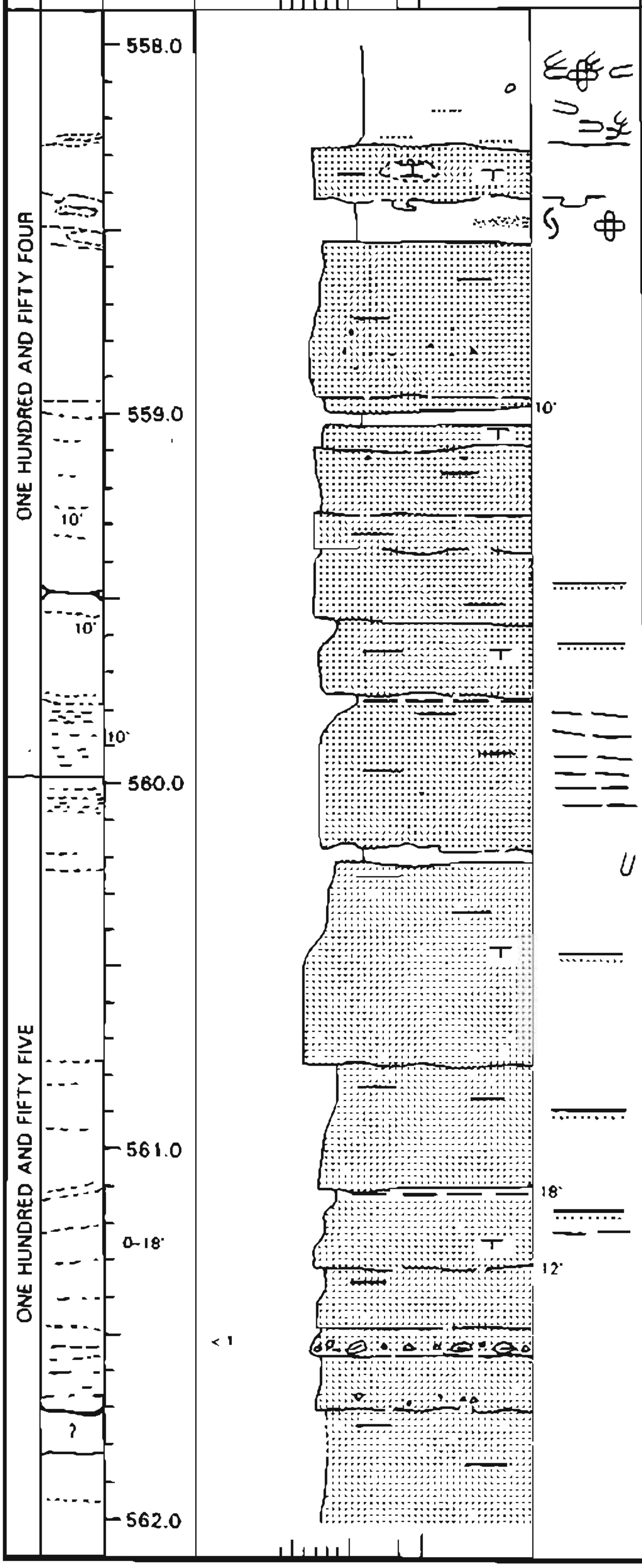
A few small carbonaceous mudstone and coal fragments up to 7 mm long.

PROJECT : CIROS -1

SHEET NO. : 135

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



— sharp contact –

Sub-unit 19.6

558.27 - 566.76 m

SANDSTONE AND MUDSTONE

Hard, slightly calcareous, weakly to well stratified, moderately to moderately well sorted **coarse to fine sandstone** (greyish olive 7.5Y 5 2 to 4 2 with greyish white to grey N 7 0 to N 6 0 from 562.55 to 565.57 m) with several interbeds (0.03 to 0.50 m thick) of weakly stratified, **sandy mudstone** (grey 7.5Y 4 1)

Stratification in the sandstone is generally wispy with sharp lower boundaries, commonly with load structures, and diffuse (bioturbated) upper boundaries. Some contorted bedding at 558.50 m. The beds are graded from coarse or medium sandstone up to fine sandstone. The mudstone also has irregular and discontinuous bedding that is disrupted by burrowing activity. The burrows occur in various sizes (1-2 mm, 5-8 mm and one is 24 mm wide) and are inclined or horizontal.

Slightly conglomeratic medium sandstone with elongate mudstone intraclasts up to 14 mm long and basement granules and small pebbles up to 6 mm at 561.5 m.

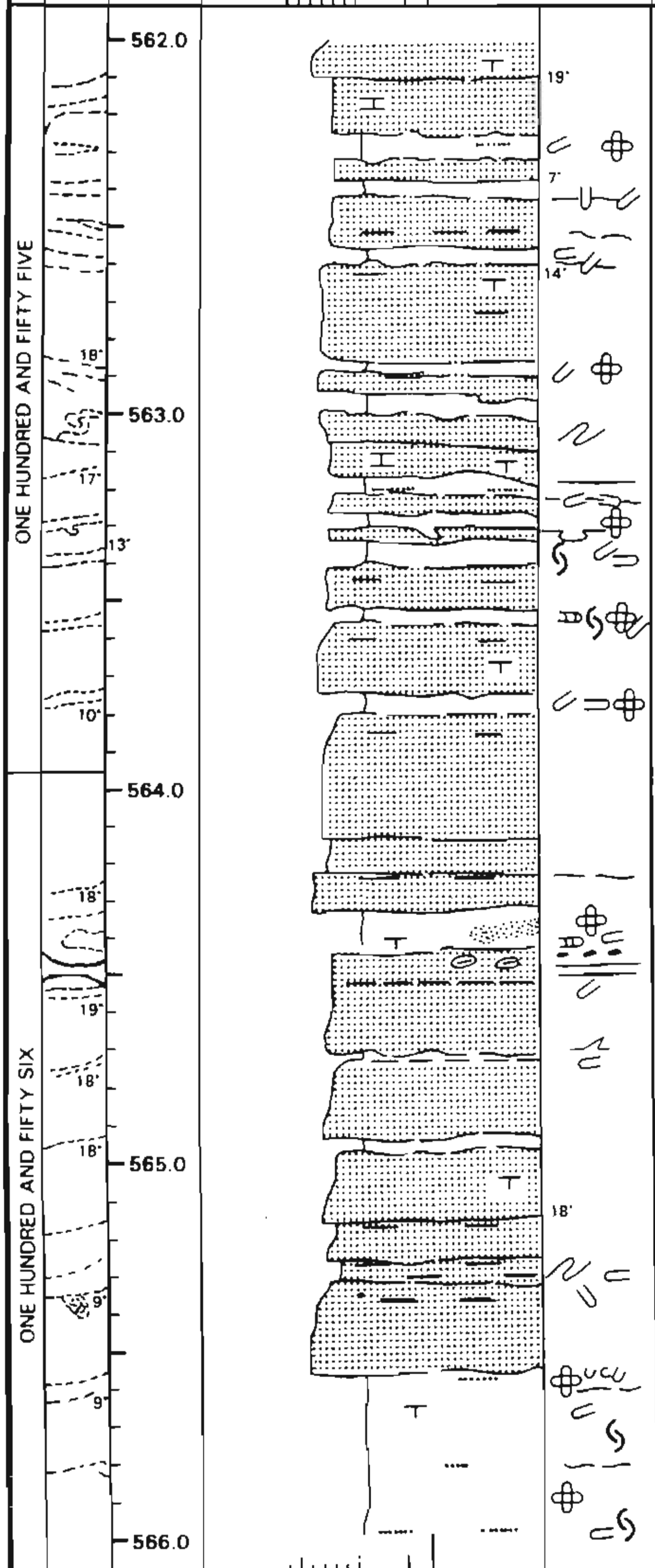
Granule size carbonaceous fragments at 564.47 m

PROJECT : CIROS - 1

SHEET NO. : 136

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.6

558.27 - 566.76 m

SANDSTONE AND MUDSTONE

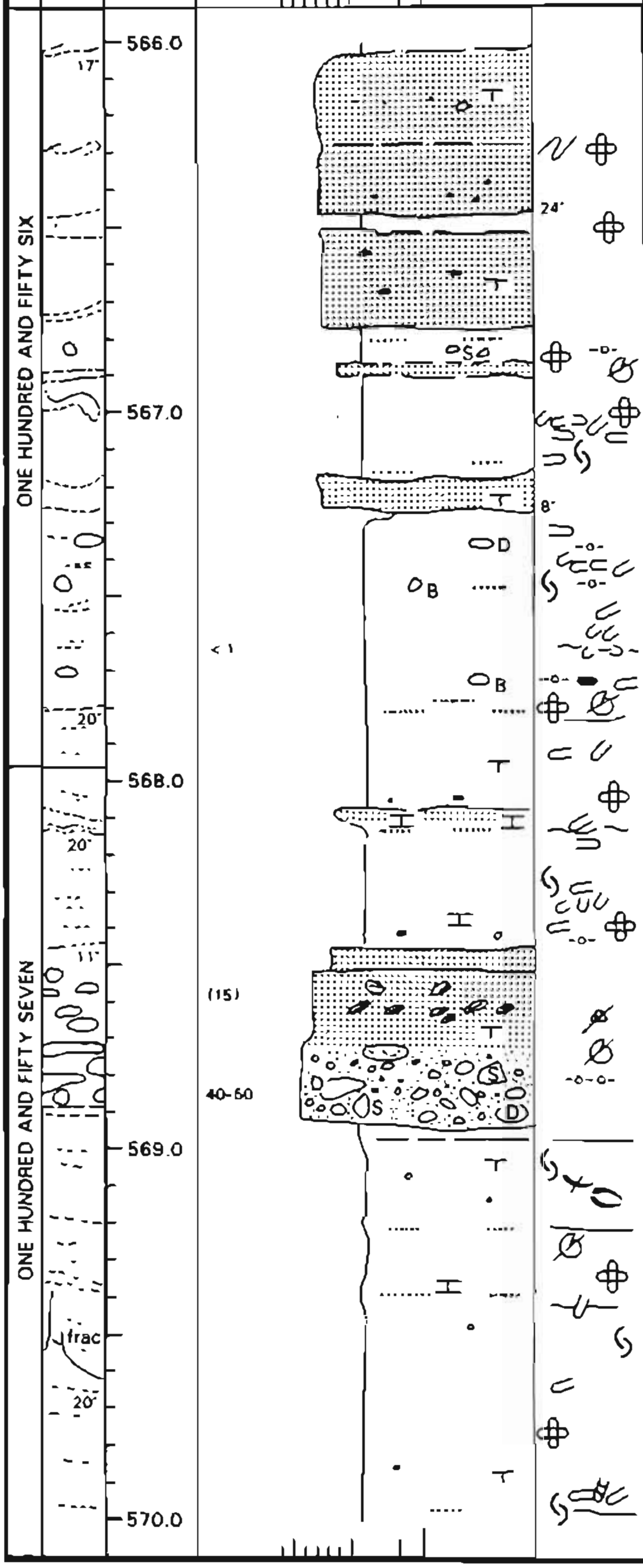
Hard, slightly calcareous, weakly to well stratified, moderately to moderately well sorted, **coarse to fine sandstone** (greyish olive 7.5Y 5/2 to 4/2 with greyish white to grey N 7/0 to N 6/0 from 562.55 to 565.57 m) with several interbeds (0.03 to 0.50 m thick) of weakly stratified, **sandy mudstone** (grey 7.5Y 4/1). Stratification in the sandstone is generally wispy with sharp lower boundaries, commonly with load structures, and diffuse (bioturbated) upper boundaries. Some contorted bedding at 558.50 m. The beds are graded from coarse or medium sandstone up to fine sandstone. The mudstone also has irregular and discontinuous bedding that is disrupted by burrowing activity. The burrows occur in various sizes (1-2 mm, 5-8 mm and one is 24 mm wide) and are inclined or horizontal. **Slightly conglomeratic medium sandstone** with elongate mudstone intraclasts up to 14 mm long and basement granules and small pebbles up to 6 mm at 561.5 m. Granule size carbonaceous fragments at 564.47 m.

PROJECT : CIROS -1

SHEET NO. : 137

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



— contact sharp -

Sub-unit 19.7

566.76 - 568.45 m

MUDSTONE

Hard, non-calcareous to slightly calcareous (568.09 to 568.46 m), **sandy mudstone** (dark olive grey 2.5GY 4/1), with interbeds (< 10 cm thick) of unstratified, **coarse to medium sandstone**.

The mudstones are virtually unstratified due to bioturbation which is recognised as mottling and burrowing. Burrows are 1-2 mm and 6-10 mm wide.

Lonestones (10 to > 24 mm long) of indurated sediment (including coal fragments), Beacon sandstone and dolerite are scattered through the sub-unit.

— sharp contact -

Sub-unit 19.8

568.45 - 568.92 m

SANDSTONE AND CONGLOMERATE

Hard, unstratified, **coarse sandstone** (greyish olive 7.5Y 5/2) with scattered elongate mudstone intraclasts up to 20 mm long and forming 15% of the core overlying a **pebble conglomerate** with a **very coarse sandstone** matrix. Lithologies include basement (50%), dolerite (30%) and indurated sediments (10%). Clasts have a weak long axis orientation subparallel to bedding. A chambered calcite test 3 mm across at 368.69 m.

— contact sharp -

Sub-unit 19.9

568.92 - 582.57 m

MUDSTONE

Hard, weakly calcareous (in places strongly calcareous), **sandy mudstone** (dark greenish grey 7.5GY 4/1 mottled light olive grey 2.5GY 7/1). Stratification is irregular and wispy, due to strong bioturbation.

Lonestones are present but rare. They are mostly subrounded and composed of dolerite, indurated sediment, coal and basement. One dolerite clast is > 110 m at 571.33 m. Two clasts are striated.

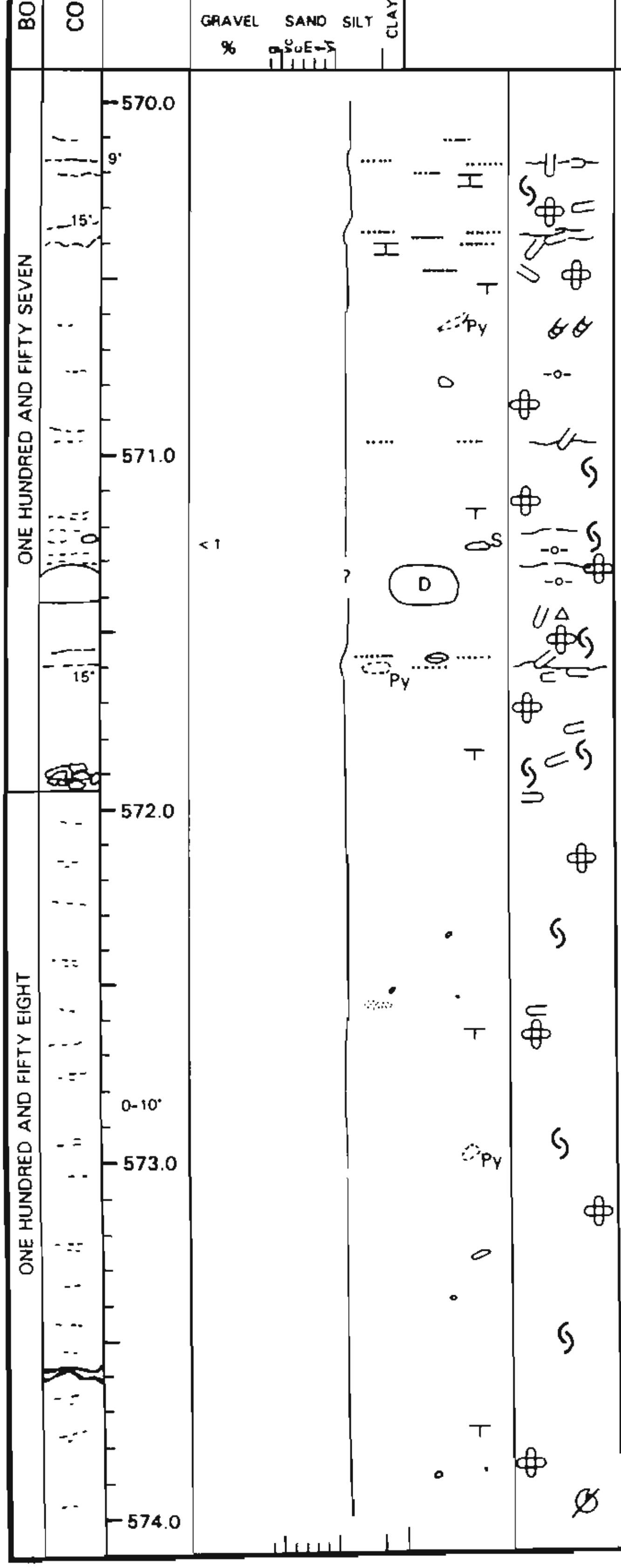
Biogenic features include both inclined and horizontal burrows, some with spreiten and either 1-2 mm or 6-10 mm in size. Broken and articulated bivalve and other shell debris are scattered through the core. Some pyrite nodules.

PROJECT : CIROS -1

SHEET NO. : 138

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 19.9

568.92 - 582.57 m

MUDSTONE

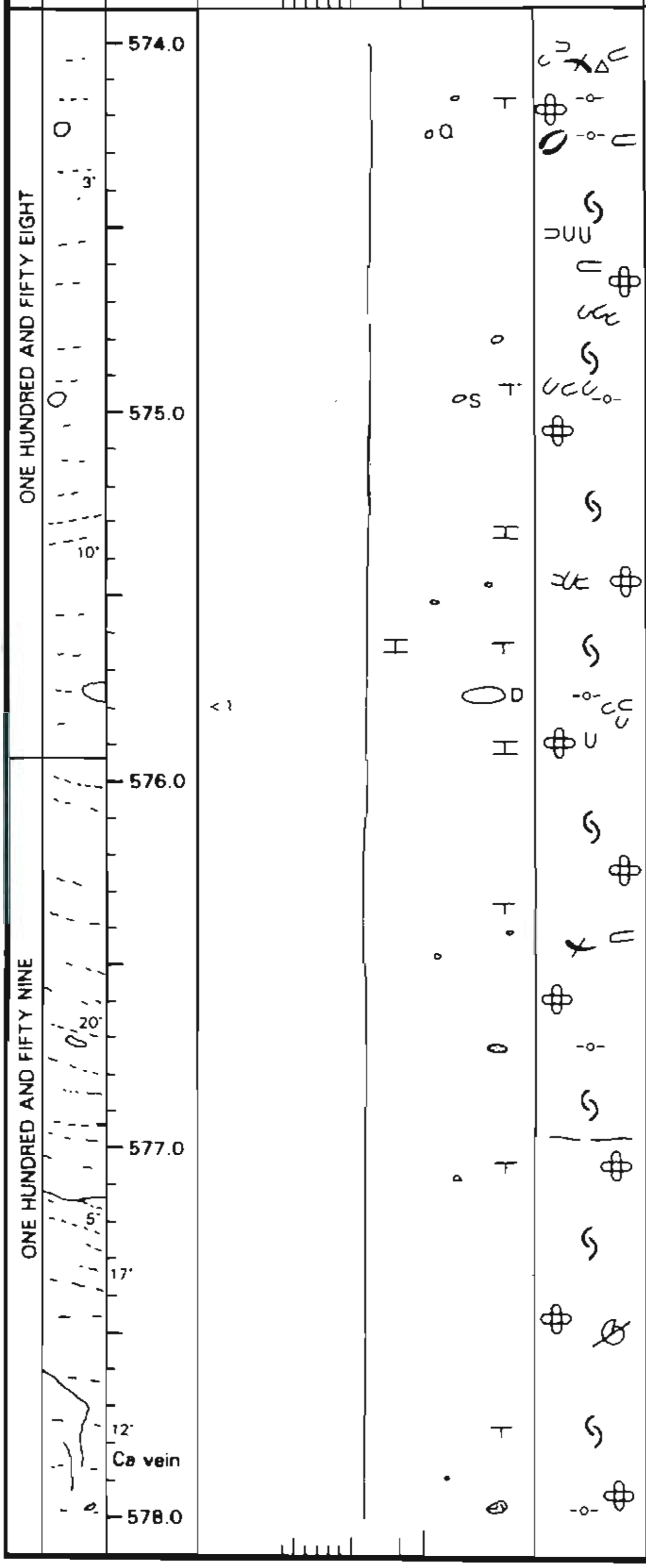
Hard, weakly calcareous (in places strongly calcareous), **sandy mudstone** (dark greenish grey 7.5GY 4/1 mottled light olive grey 2.5GY 7/1). Stratification is irregular and wispy, due to strong bioturbation. Lonestones are present but rare. They are mostly subrounded and composed of dolerite, indurated sediment, coal and basement. One dolerite clast is > 110 m at 571.33 m. Two clasts are striated. Biogenic features include both inclined and horizontal burrows, some with spreiten and either 1-2 mm or 6-10 mm in size. Broken and articulated bivalve and other shell debris are scattered through the core. Some pyrite nodules.

PROJECT : CIROS -1

SHEET NO. : 139

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 19.9

568.92 - 582.57 m

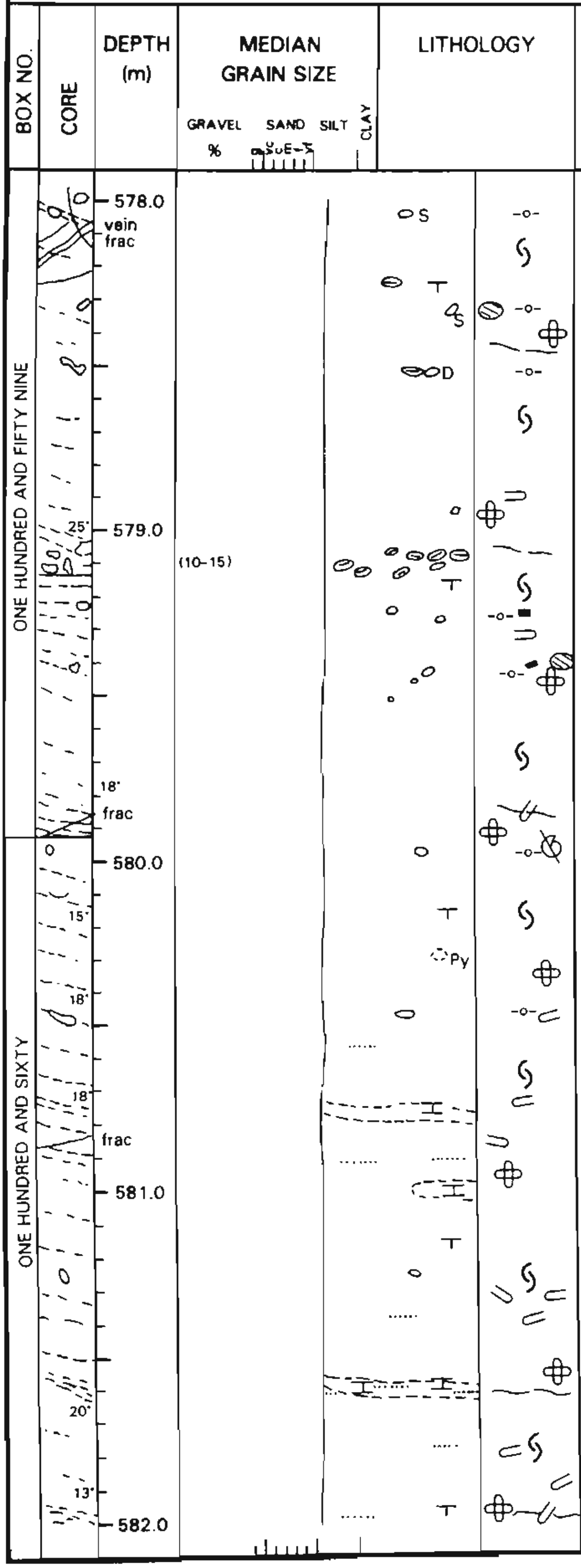
MUDSTONE

Hard, weakly calcareous (in places strongly calcareous), **sandy mudstone** (dark greenish grey 7.5GY 4/1 mottled light olive grey 2.5GY 7/1). Stratification is irregular and wispy, due to strong bioturbation. Lonestones are present but rare. They are mostly subrounded and composed of dolerite, indurated sediment, coal and basement. One dolerite clast is > 110 m at 571.33 m. Two clasts are striated. Biogenic features include both inclined and horizontal burrows, some with spreiten and either 1-2 mm or 6-10 mm in size. Broken and articulated bivalve and other shell debris are scattered through the core. Some pyrite nodules.

PROJECT : CIROS -1

SHEET NO. : 140

SCALE : 1:20



DESCRIPTION

Sub-unit 19.9

568.92 - 582.57 m

MUDSTONE

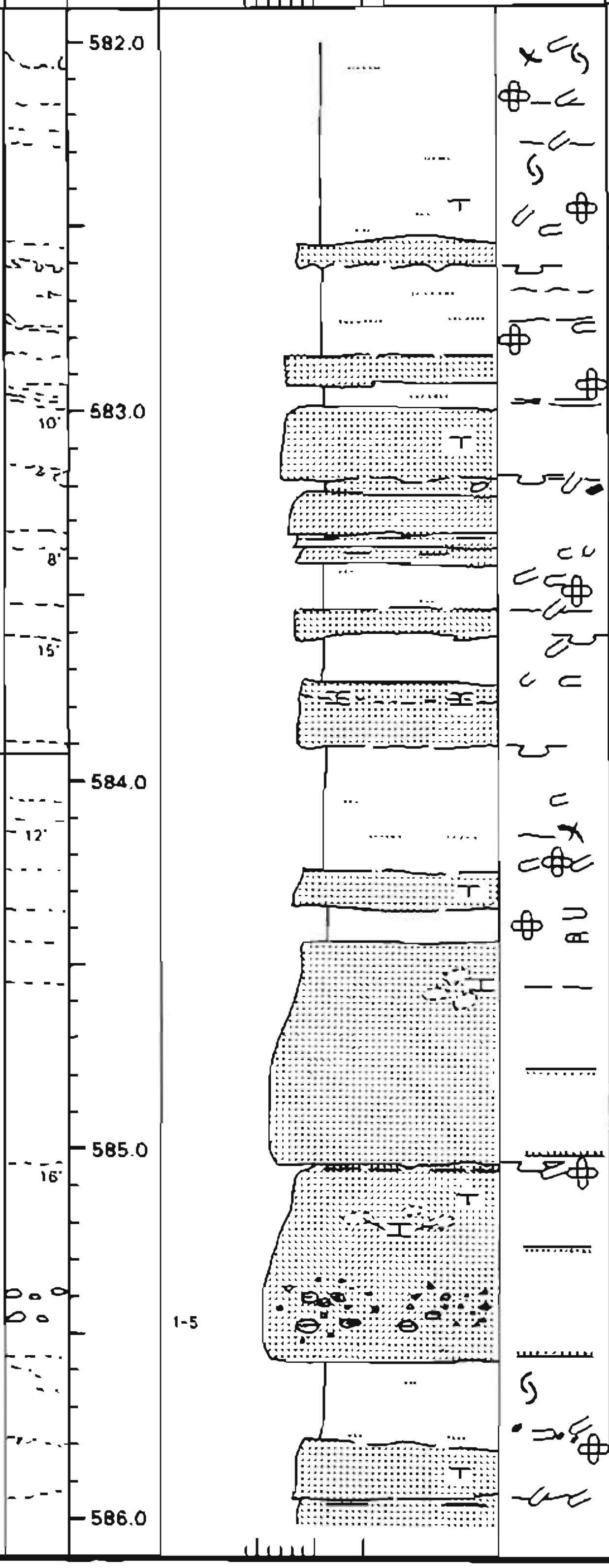
Hard, weakly calcareous (in places strongly calcareous), **sandy mudstone** (dark greenish grey 7.5GY 4/1 mottled light olive grey 2.5GY 7/1). Stratification is irregular and wispy, due to strong bioturbation. Lonestones are present but rare. They are mostly subrounded and composed of dolerite, indurated sediment, coal and basement. One dolerite clast is > 110 m at 571.33 m. Two clasts are striated. Biogenic features include both inclined and horizontal burrows, some with spreiten and either 1-2 mm or 6-10 mm in size. Broken and articulated bivalve and other shell debris are scattered through the core. Some pyrite nodules.

PROJECT : CIROS -1

SHEET NO. : 141

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



ONE HUNDRED AND SIXTY

ONE HUNDRED AND SIXTY ONE

582.0

583.0

584.0

585.0

586.0

1-5

— contact sharp -

Sub-unit 19.10

582.57 - 591.31 m

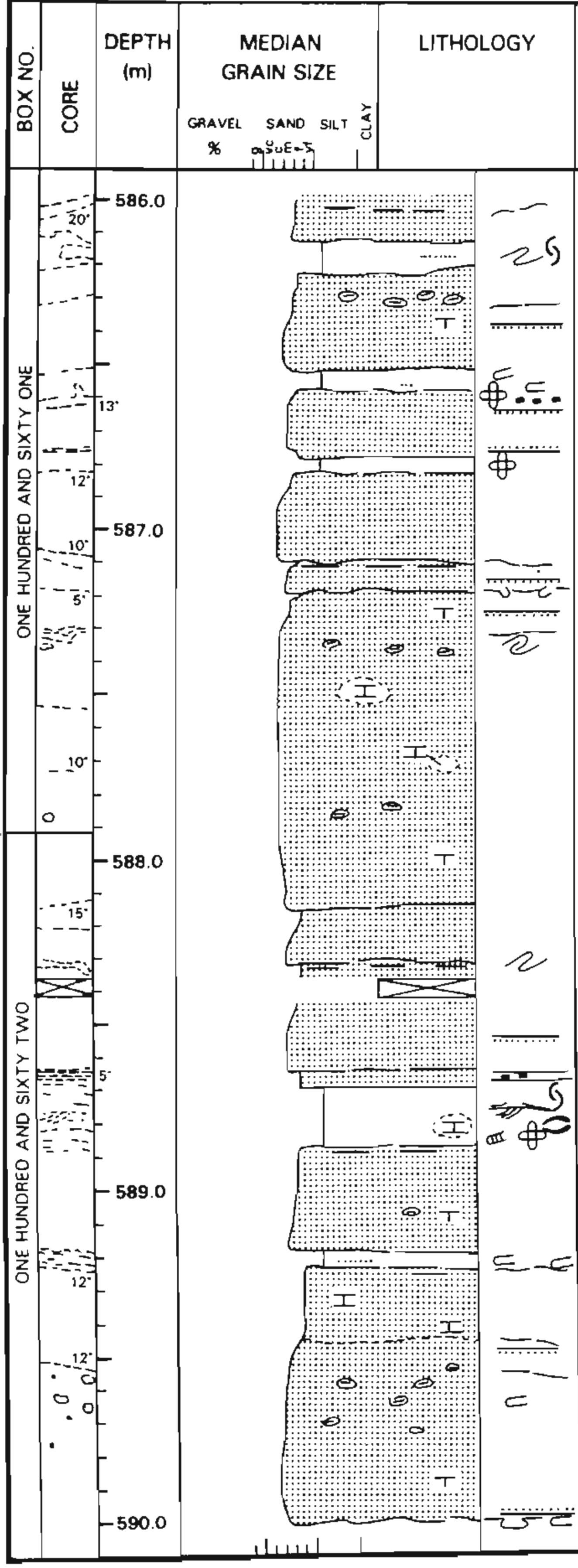
SANDSTONE AND MUDSTONE

Hard, unstratified to weakly stratified, moderately well sorted, medium to fine sandstone (grey 7.5Y 5/1) with interbeds of bioturbated sandy mudstone (dark greenish grey 5G 4/1 to 3/1).
The sandstone is unstratified and normally graded from medium to fine sandstone, though three thick beds coarsen and then fine upwards (585.04 to 585.58 m, 586.58 to 586.79 m and 589.22 to 590.00 m). The mudstone has load structures, small scale cross-bedding and disrupted and contorted bedding, but most is destroyed by bioturbation. There are also inclined and horizontal burrows up to 12 mm wide, a few with spreiten.
Rounded, elongate mudstone clasts up to 16 mm long occur in some sandstone beds.

PROJECT : CIROS -1

SHEET NO. : 142

SCALE : 1:20



DESCRIPTION

Sub-unit 19.10

582.57 - 591.31 m

SANDSTONE AND MUDSTONE

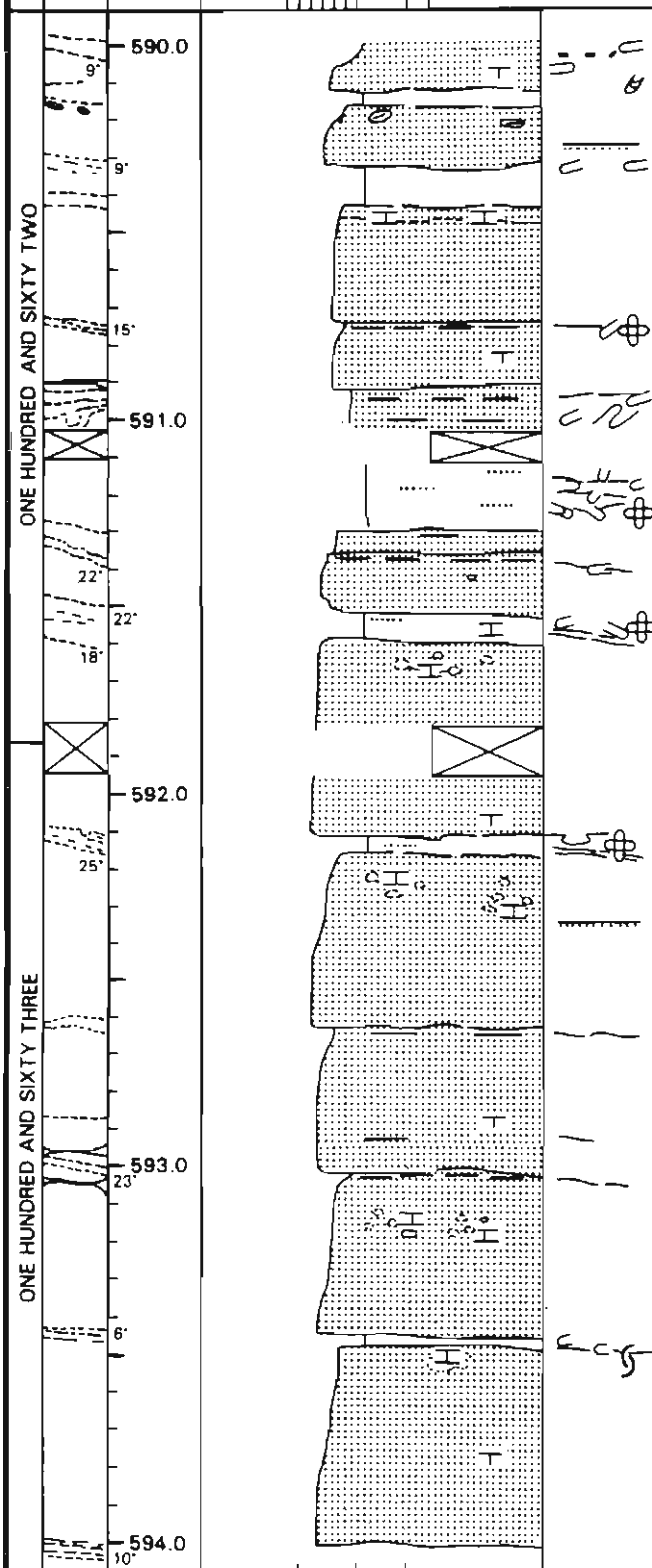
Hard, unstratified to weakly stratified, moderately well sorted, **medium to fine sandstone** (grey 7.5Y 5/1) with interbeds of bioturbated **sandy mudstone** (dark greenish grey 5G 4/1 to 3/1). The sandstone is unstratified and normally graded from medium to fine sandstone, though three thick beds coarsen and then fine upwards (585.04 to 585.58 m, 586.58 to 586.79 m and 589.22 to 590.00 m). The mudstone has load structures, small scale cross-bedding and disrupted and contorted bedding, but most is destroyed by bioturbation. There are also inclined and horizontal burrows up to 12 mm wide, a few with spreiten. Rounded, elongate mudstone clasts up to 16 mm long occur in some sandstone beds.

PROJECT : CIROS -1

SHEET NO. : 143

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.10

582.57 - 591.31 m

SANDSTONE AND MUDSTONE

Hard, unstratified to weakly stratified, moderately well sorted, **medium to fine sandstone** (grey 7.5Y 5/1) with interbeds of bioturbated **sandy mudstone** (dark greenish grey 5G 4/1 to 3/1). The sandstone is unstratified and normally graded from medium to fine sandstone, though three thick beds coarsen and then fine upwards (585.04 to 585.58 m, 586.58 to 586.79 m and 589.22 to 590.00 m). The mudstone has load structures, small scale cross-bedding and disrupted and contorted bedding, but most is destroyed by bioturbation. There are also inclined and horizontal burrows up to 12 mm wide, a few with spreiten. Rounded, elongate mudstone clasts up to 16 mm long occur in some sandstone beds.

— contact sharp -

Sub-unit 19.11

591.31 - 597.49 m

SANDSTONE

Predominantly hard, unstratified, normally graded, poorly sorted, **very coarse to fine sandstone** (grey 7.5Y 5/1) interbedded with hard, weakly stratified, **sandy mudstone** (dark greenish grey 5G 4/1 to 3/1)

The sandstone beds range in thickness from 0.14 to 0.60 m. They are usually normally graded (from very coarse to fine sandstone), and lower contacts show load and flame structures.

Some beds have bedrock clasts up to 7 mm long and which form up to 3% of the core. They also include mudstone intraclasts up to 20 mm long that lie subparallel to bedding.

The sandstones have patches and layers of carbonate cement

The mudstone beds are thin and lack much stratification because of the strong bioturbation.

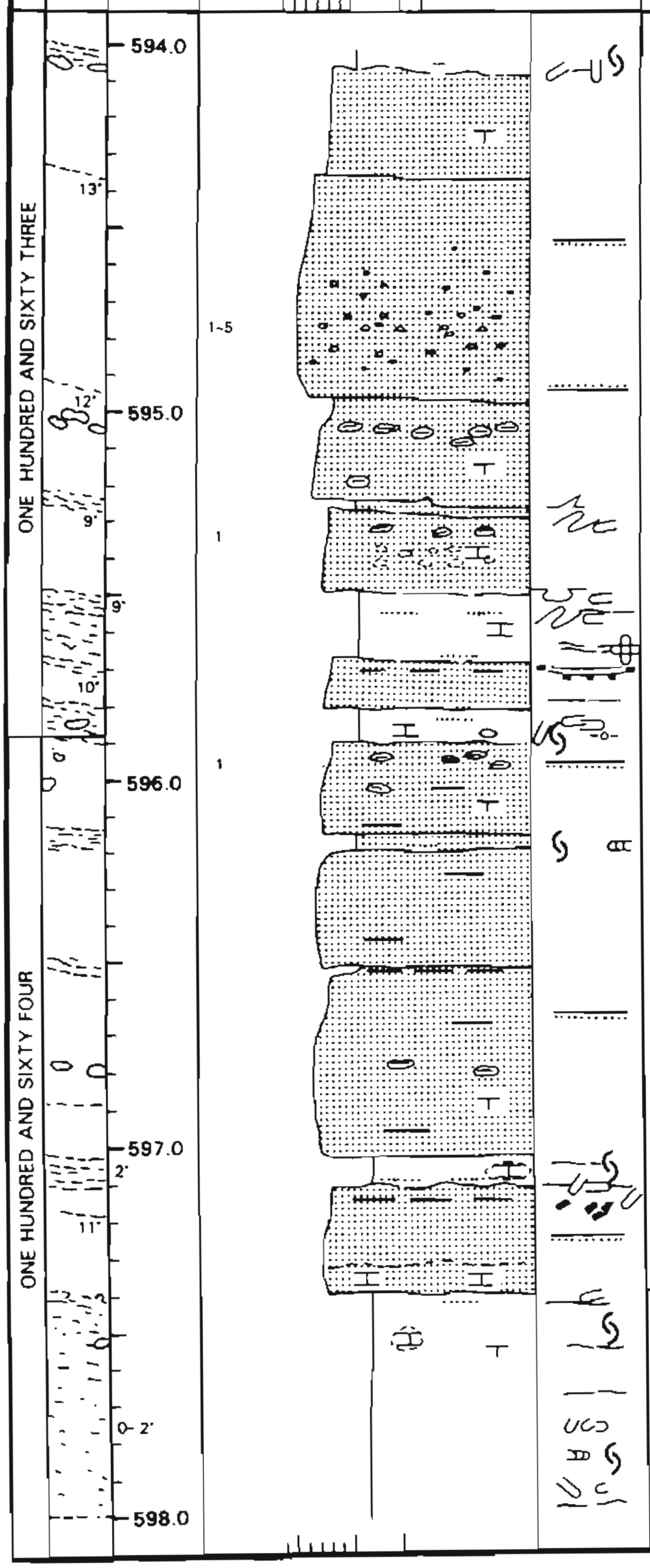
Flecks of carbonaceous material at 595.70 and 597.15 m.

PROJECT : CIROS -1

SHEET NO. : 144

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 19.11

591.31 - 597.49 m

SANDSTONE

Predominantly hard, unstratified, normally graded, poorly sorted, **very coarse to fine sandstone** (grey 7.5Y 5/1) interbedded with hard, weakly stratified, **sandy mudstone**(dark greenish grey 5G 4/1 to 3/1)
The sandstone beds range in thickness from 0.14 to 0.60 m. They are usually normally graded (from very coarse to fine sandstone), and lower contacts show load and flame structures.
Some beds have bedrock clasts up to 7 mm long and which form up to 3% of the core. They also include mudstone intraclasts up to 20 mm long that lie subparallel to bedding
The sandstones have patches and layers of carbonate cement.
The mudstone beds are thin and lack much stratification because of the strong bioturbation.
Flecks of carbonaceous material at 595.70 and 597 15 m.

—contact sharp -

Sub-unit 19.12

597.49 - 603.62 m

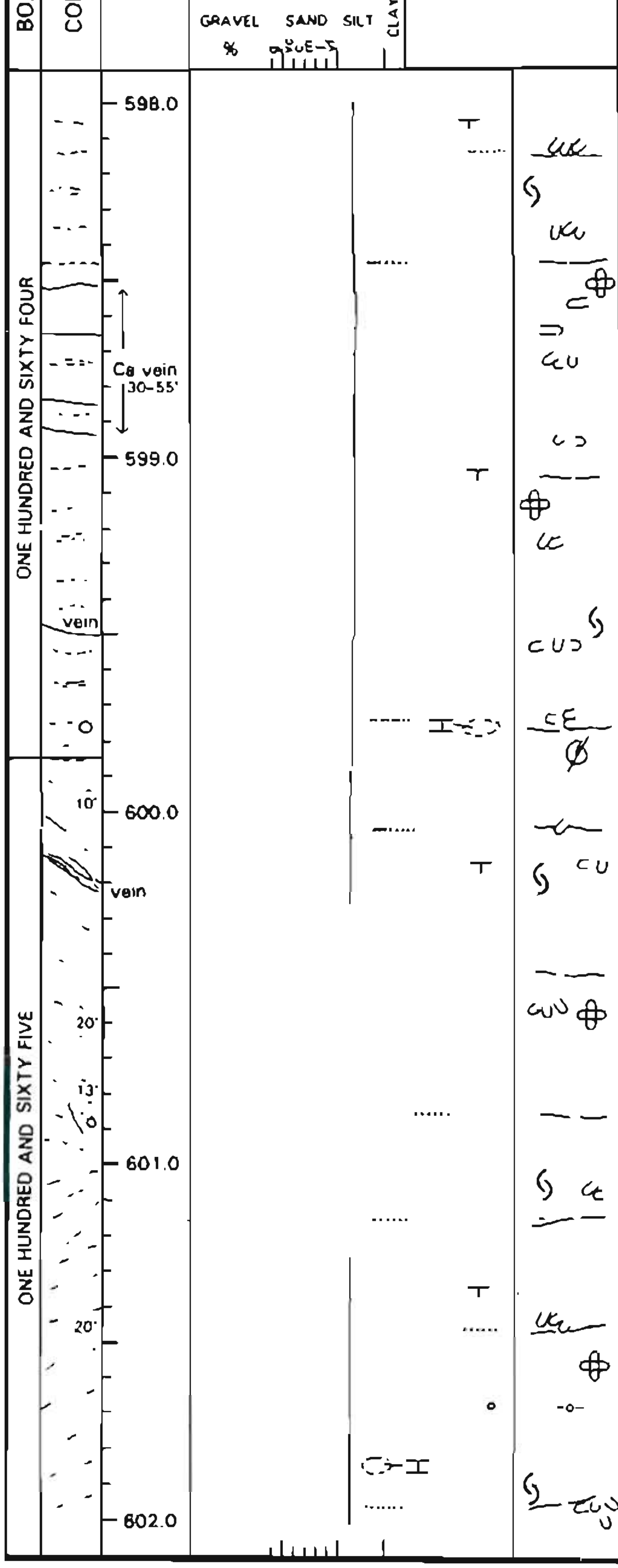
MUDSTONE

PROJECT : CIROS -1

SHEET NO. : 145

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 19.12

597.49 - 603.62 m

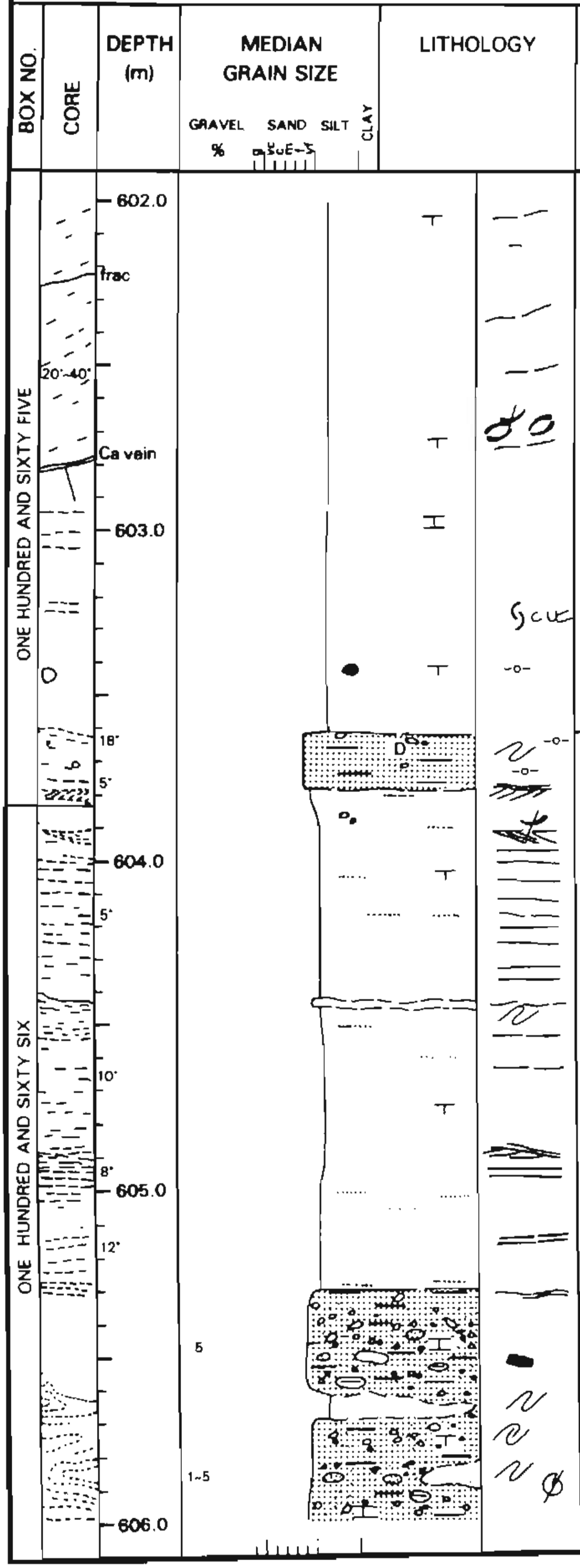
MUDSTONE

Hard, weakly stratified **mudstone** (grey to olive black 5Y 4/1 to 3/1), sandy in places. Stratification is wispy and irregular, dipping up to 20° and disrupted by strong bioturbation. Burrows are inclined or subhorizontal and are either 1-2 mm or 8-10 mm wide. Thick-walled bivalves (articulated and broken) occur at 602.70 m. Scattered calcareous nodules, and one scoriaceous clast at 603.43 m.

PROJECT : CIROS -1

SHEET NO. : 146

SCALE : 1:20



DESCRIPTION

Sub-unit 19.12

597.49 - 603.62 m

MUDSTONE

Hard, weakly stratified **mudstone** (grey to olive black 5Y 4/1 to 3/1), sandy in places. Stratification is wispy and irregular, dipping up to 20° and disrupted by strong bioturbation. Burrows are inclined or subhorizontal and are either 1-2 mm or 8-10 mm wide. Thick-walled bivalves (articulated and broken) occur at 602.70 m. Scattered calcareous nodules, and one scoriaceous clast at 603.43 m.

— contact sharp ~

Sub-unit 19.13

603.62 - 613.39 m

SANDSTONE AND MUDSTONE

Hard, in places slightly calcareous, unstratified to well stratified, **muddy very fine sandstone** (olive black 7.5Y 3/1), with intervals of **conglomeratic muddy very fine sandstone** (605.31 to 606.02 m, 606.44 to 606.52 m and 607.48 to 607.60 m) and a **sandy pebble conglomerate** (611.28 to 611.52 m).

Hard, weakly to well stratified, **sandy mudstone** beds (also olive black) occur throughout the sub-unit. Some minor mottling.

Stratification in both sandstone and mudstone is quite variable and includes horizontal, inclined, disrupted and contorted bedding. Isoclinal folding, sigmoidal bedding (up to 25 cm amplitude) and cross-bedding are present.

The conglomeratic sandstone and the conglomerate contain clasts of basement, dolerite, indurated sediment, coal as well as sandstone and mudstone intraclasts. Most are 4 to 6 mm, but range up to 45 mm long. Similar lithologies occur as lonestones in the mudstone beds. One striated basement lonestone at 607.50 m.

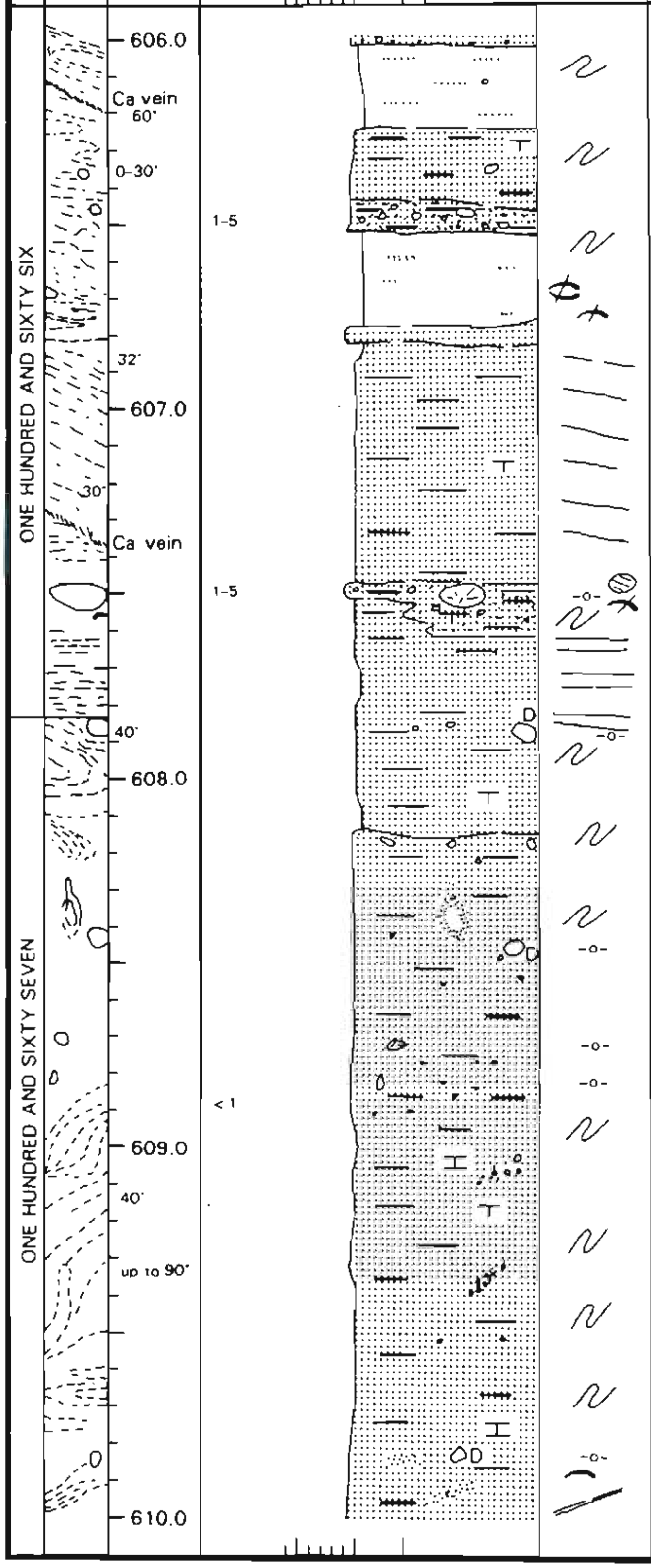
Other features include broken shell debris and broken and articulated bivalves.

PROJECT : CIROS -1

SHEET NO. : 147

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % SAND SILT CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.13

603.62 - 613.39 m

SANDSTONE AND MUDSTONE

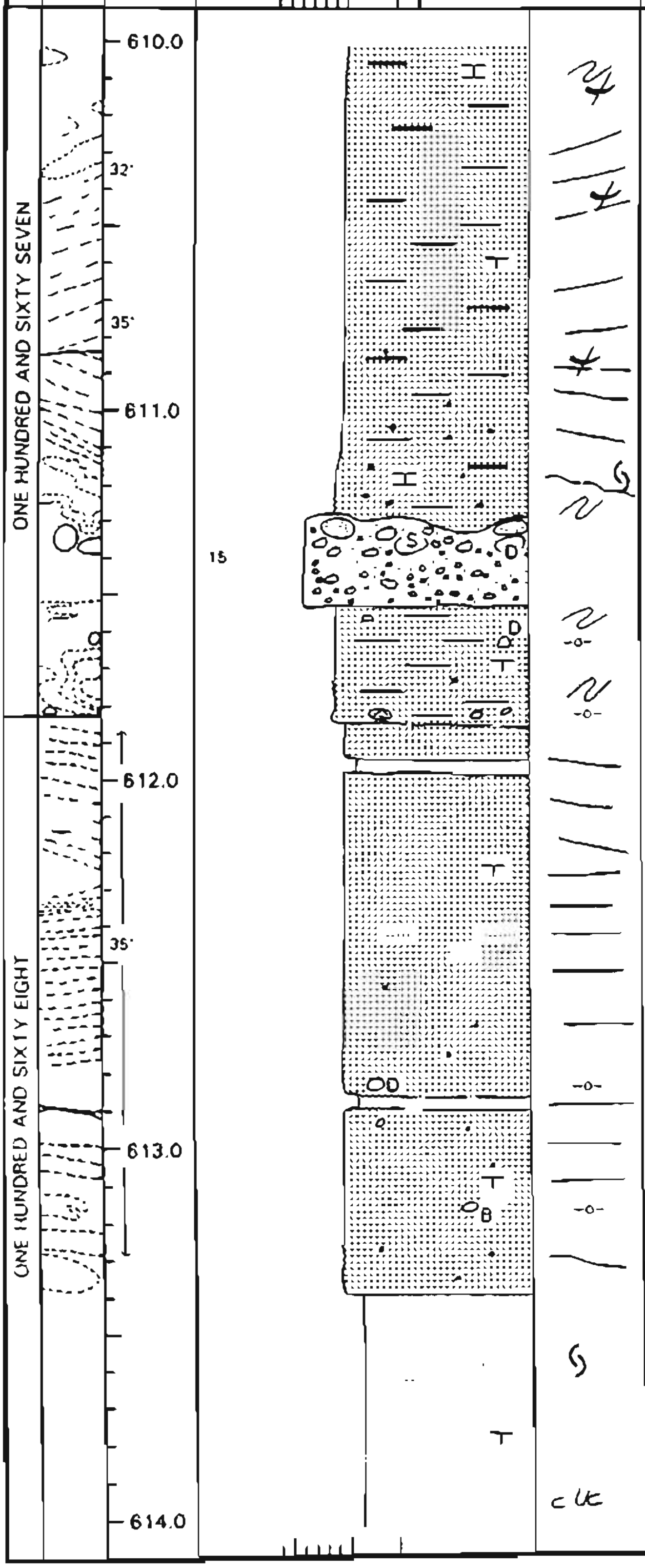
Hard, in places slightly calcareous, unstratified to well stratified, **muddy very fine sandstone** (olive black 7.5Y 3/1), with intervals of **conglomeratic muddy very fine sandstone**(605.31 to 606.02 m, 606.44 to 606.52 m and 607.48 to 607.60 m) and a **sandy pebble conglomerate** (611.28 to 611.52 m). Hard, weakly to well stratified, **sandy mudstone** beds (also olive black) occur throughout the sub-unit. Some minor mottling. Stratification in both sandstone and mudstone is quite variable and includes horizontal, inclined, disrupted and contorted bedding. Isoclinal folding, sigmoidal bedding (up to 25 cm amplitude) and cross-bedding are present. The conglomeratic sandstone and the conglomerate contain clasts of basement, dolerite, indurated sediment, coal as well as sandstone and mudstone intraclasts. Most are 4 to 6 mm, but range up to 45 mm long. Similar lithologies occur as lonestones in the mudstone beds. One striated basement lonestone at 607.50 m. Other features include broken shell debris and broken and articulated bivalves

PROJECT : CIROS -1

SHEET NO. : 148

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.13

603.62 - 613.39 m

SANDSTONE AND MUDSTONE

Hard, in places slightly calcareous, unstratified to well stratified, **muddy very fine sandstone** (olive black 7.5Y 3/1), with intervals of **conglomeratic muddy very fine sandstone**(605.31 to 606.02 m, 606.44 to 606.52 m and 607.48 to 607.60 m) and a **sandy pebble conglomerate** (611.28 to 611.52 m). Hard, weakly to well stratified, **sandy mudstone** beds (also olive black) occur throughout the sub-unit. Some minor mottling. Stratification in both sandstone and mudstone is quite variable and includes horizontal, inclined, disrupted and contorted bedding. Isoclinal folding, sigmoidal bedding (up to 25 cm amplitude) and cross-bedding are present. The conglomeratic sandstone and the conglomerate contain clasts of basement, dolerite, indurated sediment, coal as well as sandstone and mudstone intraclasts. Most are 4 to 6 mm, but range up to 45 mm long. Similar lithologies occur as lonestones in the mudstone beds. One striated basement lonestone at 607.50 m. Other features include broken shell debris and broken and articulated bivalves.

- sharp contact -

Sub-unit 19.14

613.39 - 621.78 m

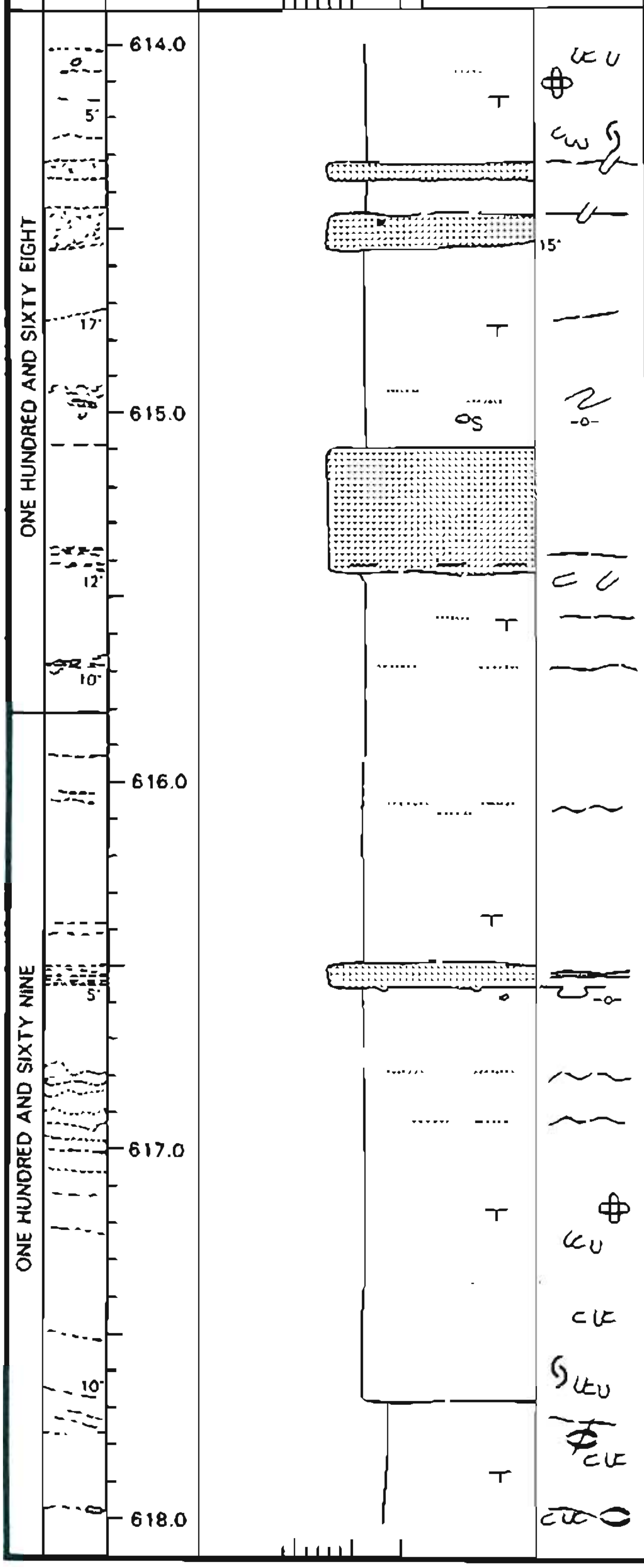
MUDSTONE

PROJECT : CIROS -1

SHEET NO. : 149

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 19.14

613.39 - 621.78 m

MUDSTONE

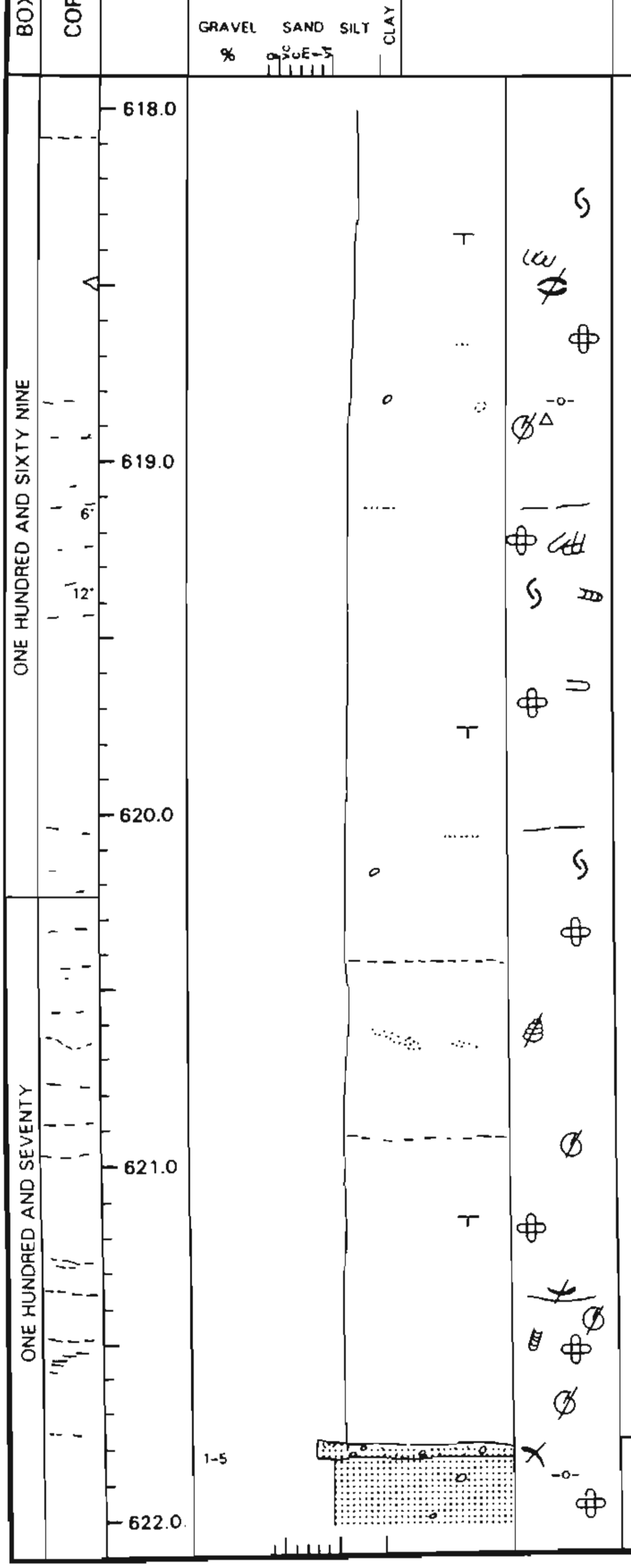
Hard, unstratified and weakly stratified, **sandy mudstone**(olive black 7.5Y 3/1 with two darker intervals of 5Y 3/1 at 617.67 and 620.42 m). Several beds 5 to 32 cm thick of hard, unstratified and weakly stratified, poorly sorted, **muddy medium sandstone** (greyish olive 7.5Y 5/2 to 6/2). The mudstones have irregular wispy bedding, best shown by thin (< 2 cm) sandstone stringers, much disrupted by strong bioturbation. The sandstones are either unstratified or contain irregular, subhorizontal (< 5°) bedding. Lonestones are very rare and no larger than 7 mm. Both small (1-2 mm) and large (up to 6 mm), inclined and horizontal burrows are present in most sandstone beds. Broken shell debris and some whole and articulated bivalves are found below 617.68 m

PROJECT : CIROS -1

SHEET NO. : 150

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 19.14

613.39 - 621.78 m

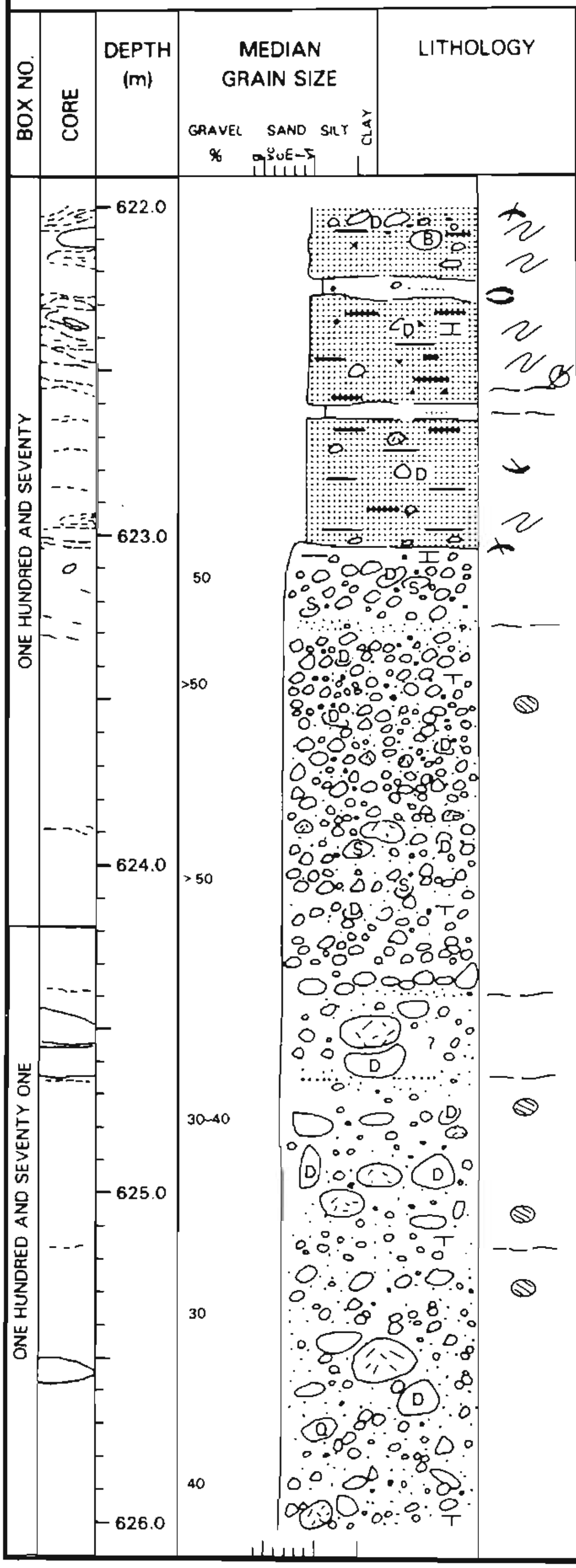
MUDSTONE

Hard, unstratified and weakly stratified, **sandy mudstone**(olive black 7.5Y 3/1 with two darker intervals of 5Y 3/1 at 617.67 and 620.42 m). Several beds 5 to 32 cm thick of hard, unstratified and weakly stratified, poorly sorted, **muddy medium sandstone** (greyish olive 7.5Y 5/2 to 6/2). The mudstones have irregular wispy bedding, best shown by thin (< 2 cm) sandstone stringers, much disrupted by strong bioturbation. The sandstones are either unstratified or contain irregular, subhorizontal (< 5°) bedding. Lonestones are very rare and no larger than 7 mm. Both small (1-2 mm) and large (up to 6 mm), inclined and horizontal burrows are present in most sandstone beds. Broken shell debris and some whole and articulated bivalves are found below 617.68 m.

PROJECT : CIROS -1

SHEET NO. : 151

SCALE : 1:20



DESCRIPTION

Sub-unit 19.15

621.78 - 623.04 m

SANDSTONE

Hard, weakly stratified, poorly sorted, **muddy very fine sandstone** (greyish olive 7.5Y 6 2) with thin **sandy mudstone** beds (622.21 to 622.27 m and 622.60 to 622.64 m).

Stratification is subhorizontal to inclined and in places contorted with isoclinal folding.

Lonestones (<1% of the core) are composed of dolerite, basement and Beacon sandstone. Most are 8 to 12 mm and subrounded. One clast is >30 mm.

— sharp contact —

Sub-unit 20.1

623.04 - 630.12 m

CONGLOMERATE

Hard, non-calcareous, weakly stratified, **sandy pebble conglomerate**, both matrix- and clast-supported.

The matrix is poorly sorted, **very coarse to medium sandstone** which is brownish black (10YR 3 1 to 3 2) except from 623.04 to 623.12 m (greyish brown 5Y 4 2) and from 623.12 to 623.50 m (dark reddish grey to reddish black 2.5YR 3 1 to 2 1).

Thin irregular sandstone layers form the only stratification in this interval. Contorted bedding at 626.56 m.

Clasts form 30 to 70% of the rock. They range from 4 to >95 mm and are angular to well rounded. Several are striated. They are composed of basement (45%), dolerite (40%) and indurated sediment (5%).

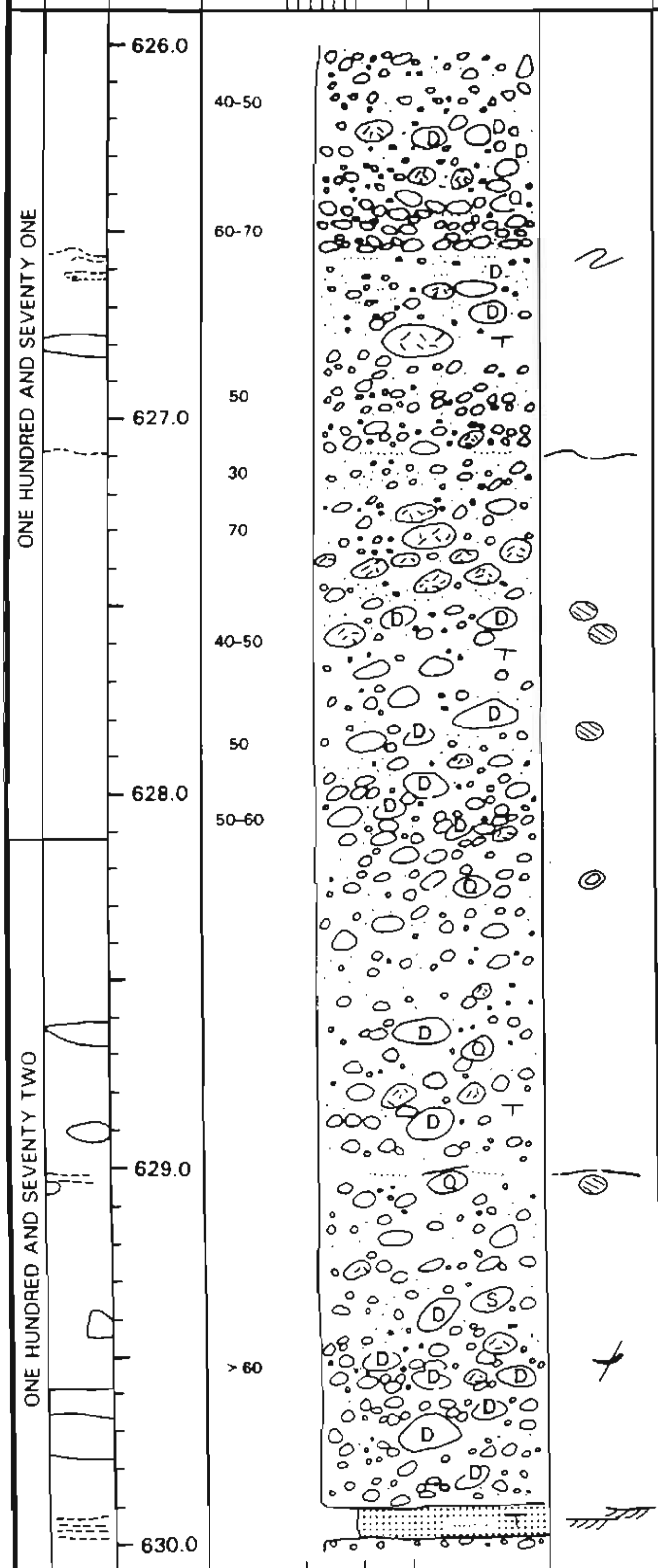
Well-stratified, cross-bedded, **fine sandstone** between 629.89 and 629.98 m. Broken bivalves at 623.05 and 629.42 m.

PROJECT : CIROS -1

SHEET NO. : 152

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 20.1

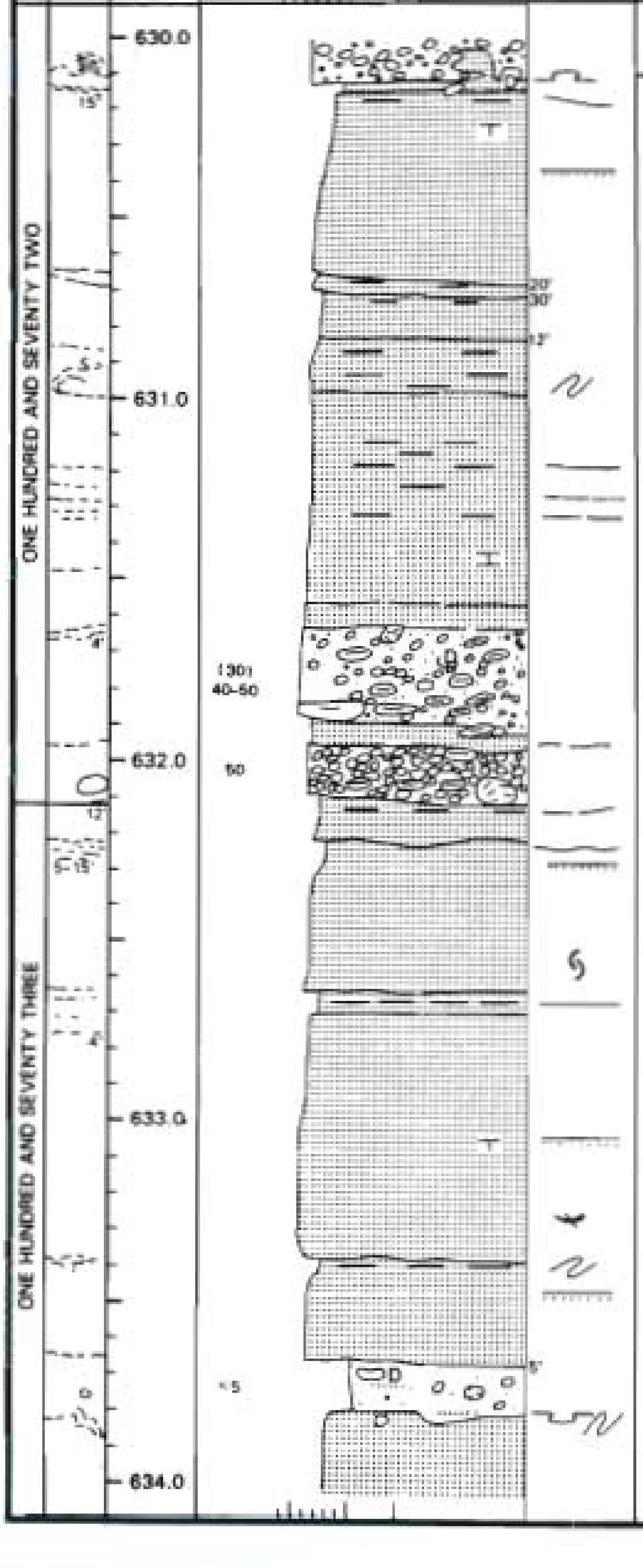
623.04 - 630.12 m

CONGLOMERATE

Hard, non-calcareous, weakly stratified, **sandy pebble conglomerate**, both matrix- and clast-supported. The matrix is poorly sorted, **very coarse to medium sandstone** which is brownish black (10YR 3/1 to 3/2) except from 623.04 to 623.12 m (greyish brown 5Y 4/2) and from 623.12 to 623.50 m (dark reddish grey to reddish black 2.5YR 3/1 to 2/1). Thin irregular sandstone layers form the only stratification in this interval. Contorted bedding at 626.56 m. Clasts form 30 to 70% of the rock. They range from 4 to > 95 mm and are angular to well rounded. Several are striated. They are composed of basement (45%), dolerite (40%) and indurated sediment < 5%). Well-stratified, cross-bedded, fine sandstone between 629.89 and 629.98 m. Broken bivalves at 623.05 and 629.42 m.

PROJECT : CIROS-1	SHEET NO. : 153 SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE GRAVEL % · SAND · SILT · CLAY	LITHOLOGY	DESCRIPTION



— contact sharp and loaded -

Sub-unit 20.2

630.12 - 634.34 m

SANDSTONE AND CONGLOMERATE

Hard, weakly stratified, moderately sorted, **muddy very coarse to fine sandstone** (dark olive grey 2.5GY 4/1 and brownish black 10YR 2/2 from about 632.50 to 634.34 m).

The colour results from brown interstitial material that forms small globules on drying and is presumed to be a waxy hydrocarbon residue. The material does not smell, burn or fluoresce with ultraviolet light.

Two **conglomeratic layers** between 631.64 and 632.09 m.

Stratification is subhorizontal to inclined (up to 30°) with beds fining upwards from very coarse sandstone on loaded bases to muddy very fine sandstone at the top.

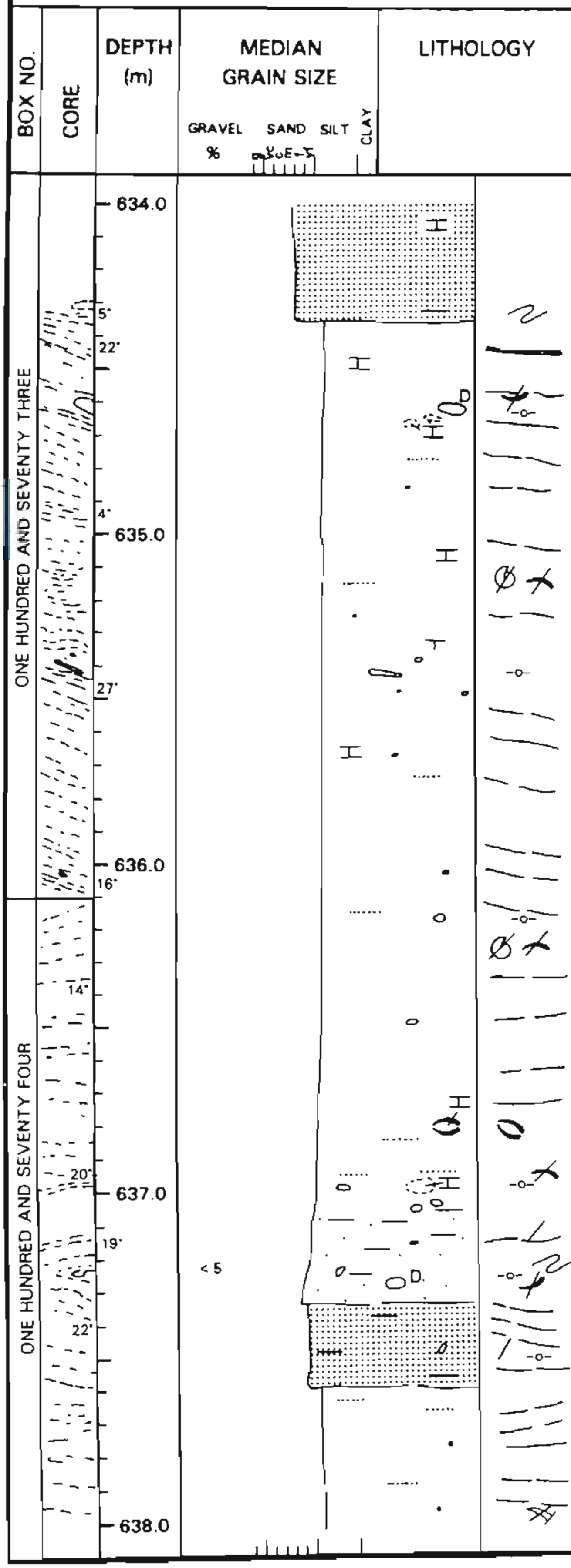
Clasts in the conglomerate form 30 to 50% of the core. They are subangular to subrounded and range from 6 to >35 mm long. The conglomerate is composed of mudstone intraclasts (60%), basement (20%) and dolerite (20%).

One bivalve fragment at 633.35 m.

PROJECT : CIROS-1

SHEET NO. : 154

SCALE : 1:20



DESCRIPTION

— contact sharp and slumped -

Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**.

The mudstone has wispy irregular bedding inclined up to 40 , and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding

The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m.

The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m.

Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (<5%). They occur as scattered lonestones in the mudstone and sandstone and form <5% of the diamictite. Sizes are variable (1 to >220 mm) with an average of 8-12 mm. A few are striated

Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten.

Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris.

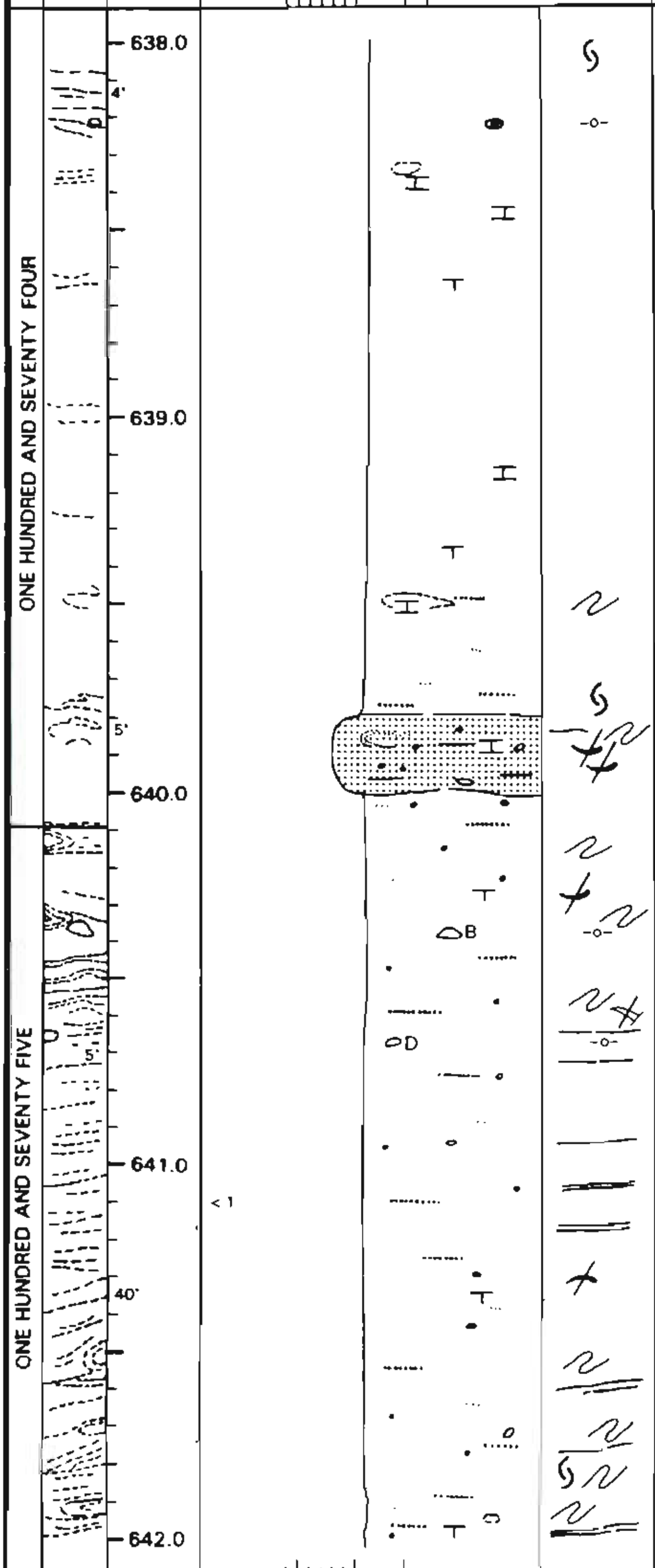
Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS -1

SHEET NO. : 155

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

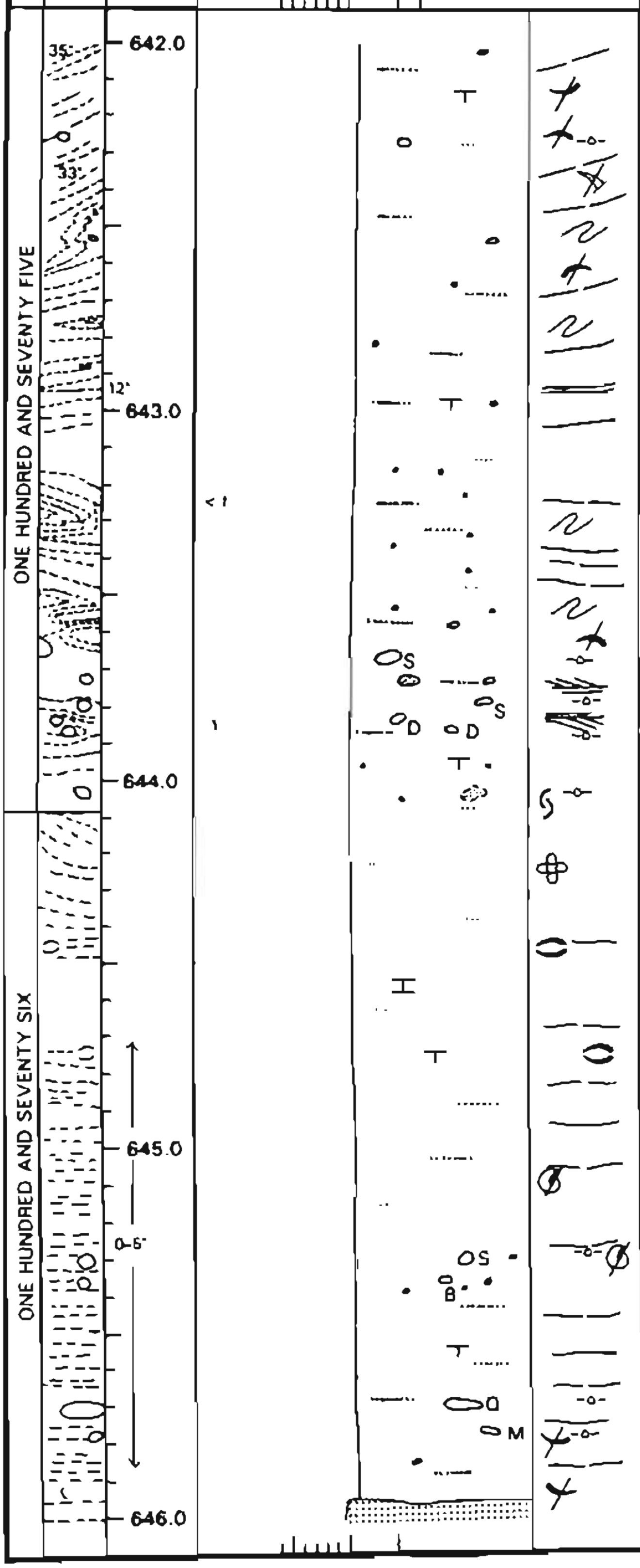
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS-1

SHEET NO. : 156

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub–unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

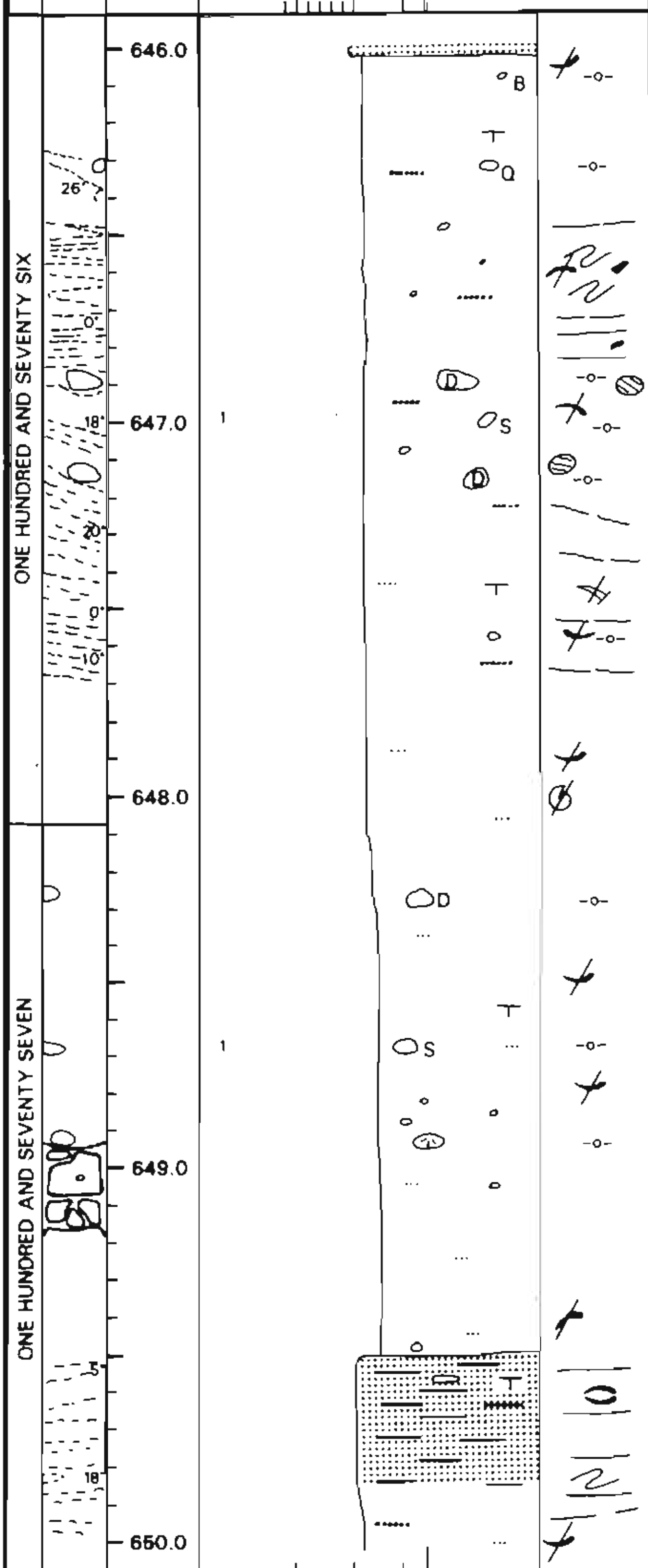
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphopods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS -1

SHEET NO. : 157

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

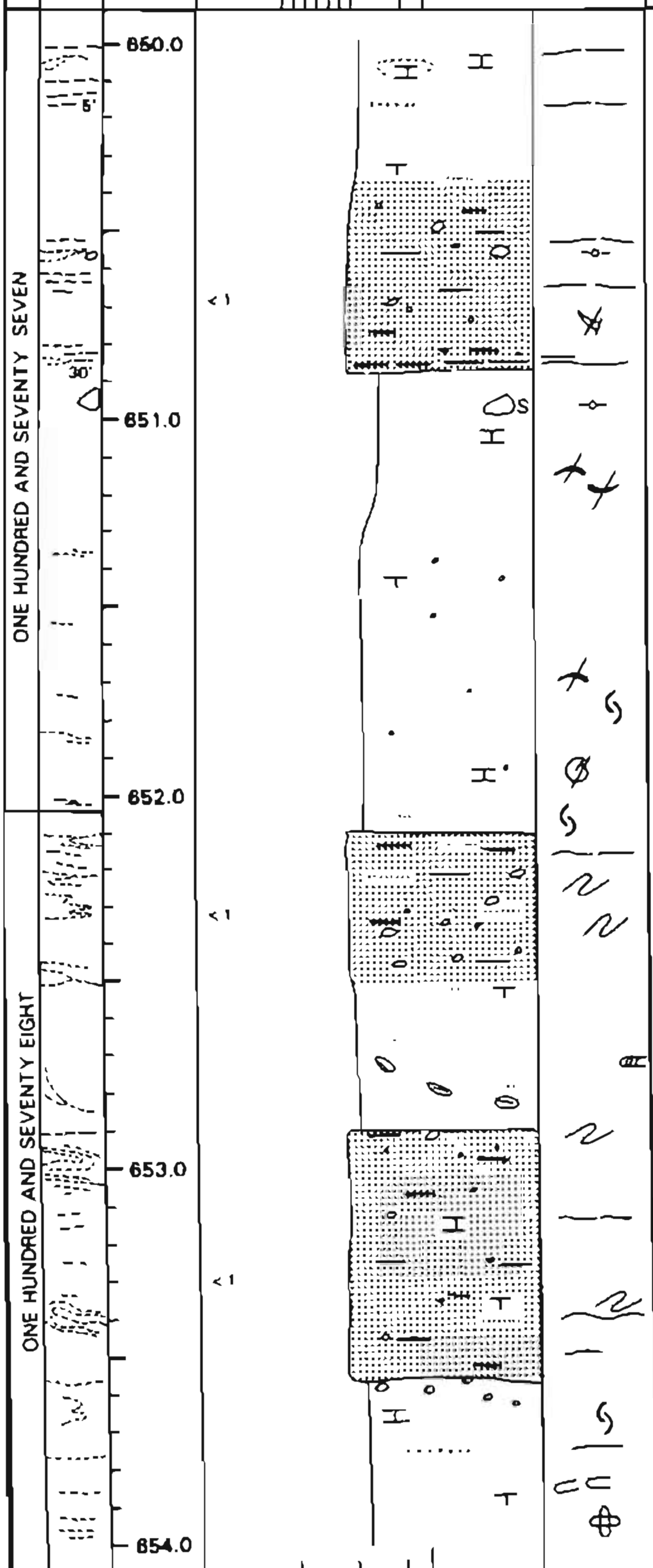
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1–2 mm) and large (6–8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS -1

SHEET NO. : 158

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

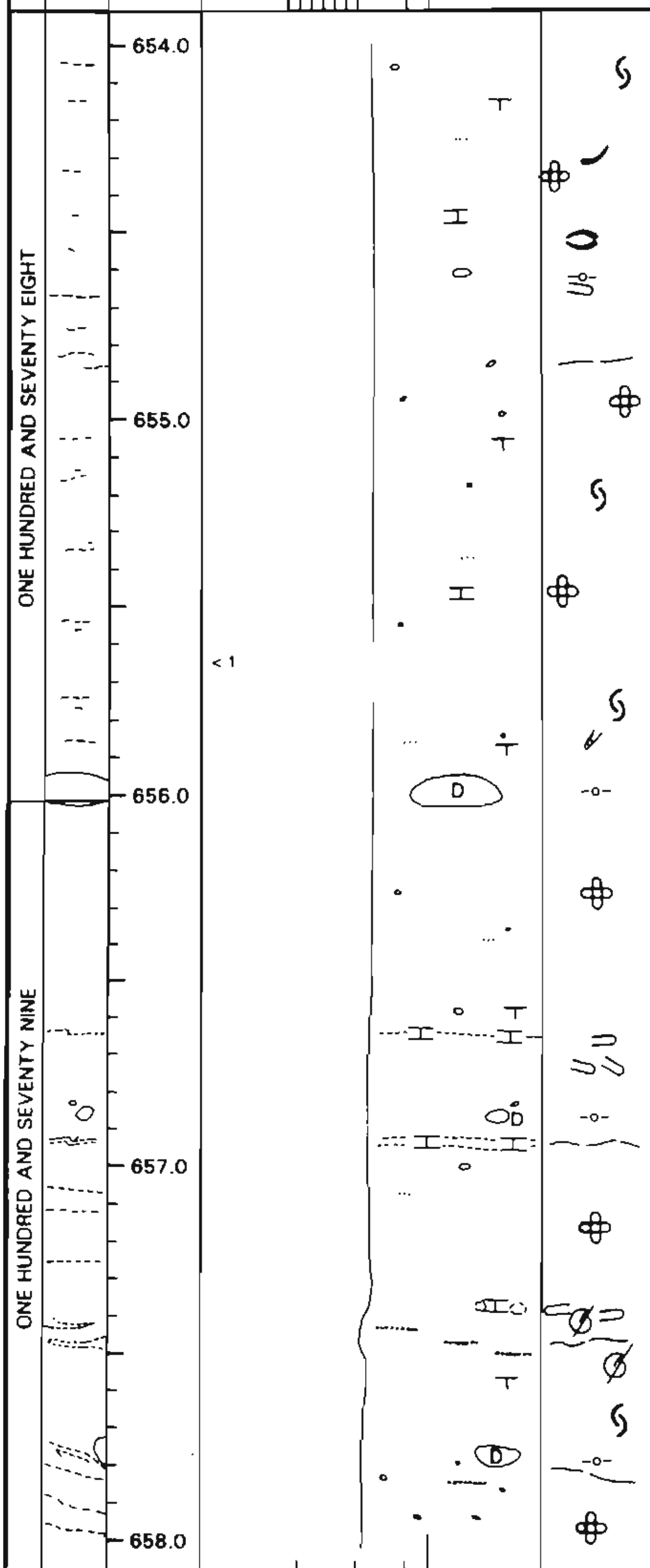
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphopods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS -1

SHEET NO. : 159

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.1

634.34 – 670.48 m

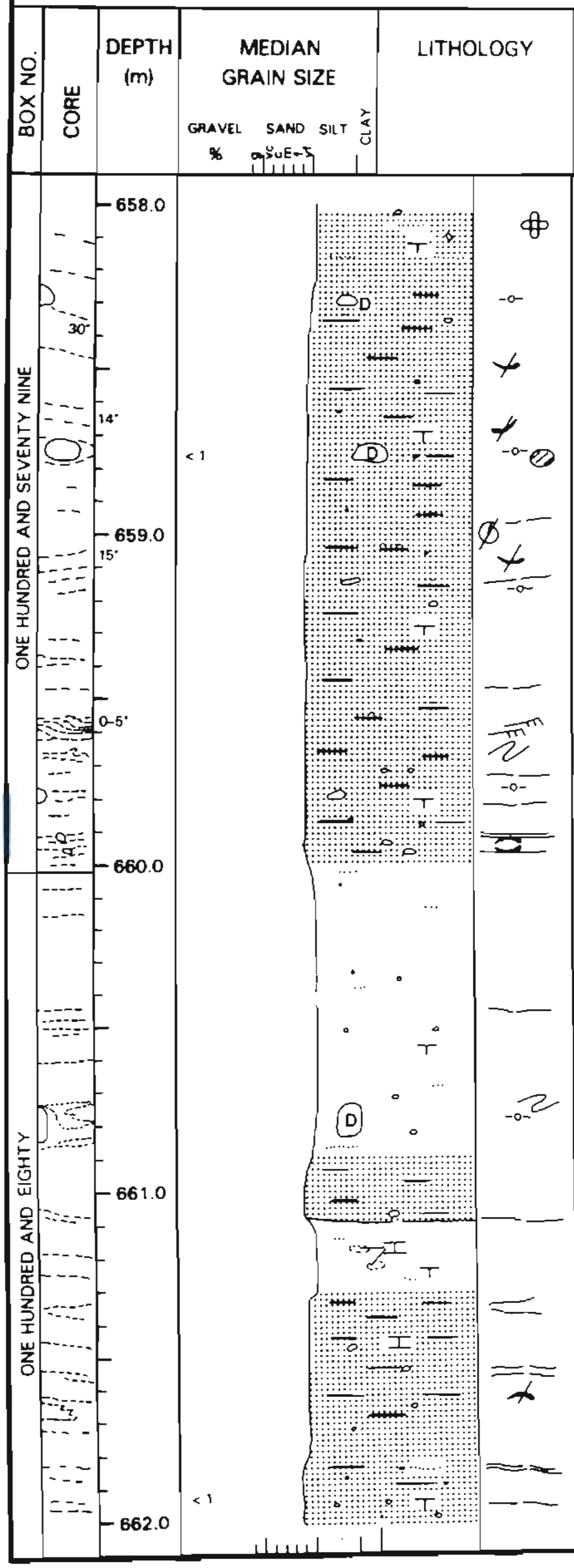
MUDSTONE, SANDSTONE AND DIAMICTITE

Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8–12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1–2 mm) and large (6–8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS -1

SHEET NO. : 160

SCALE : 1:20



DESCRIPTION

Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

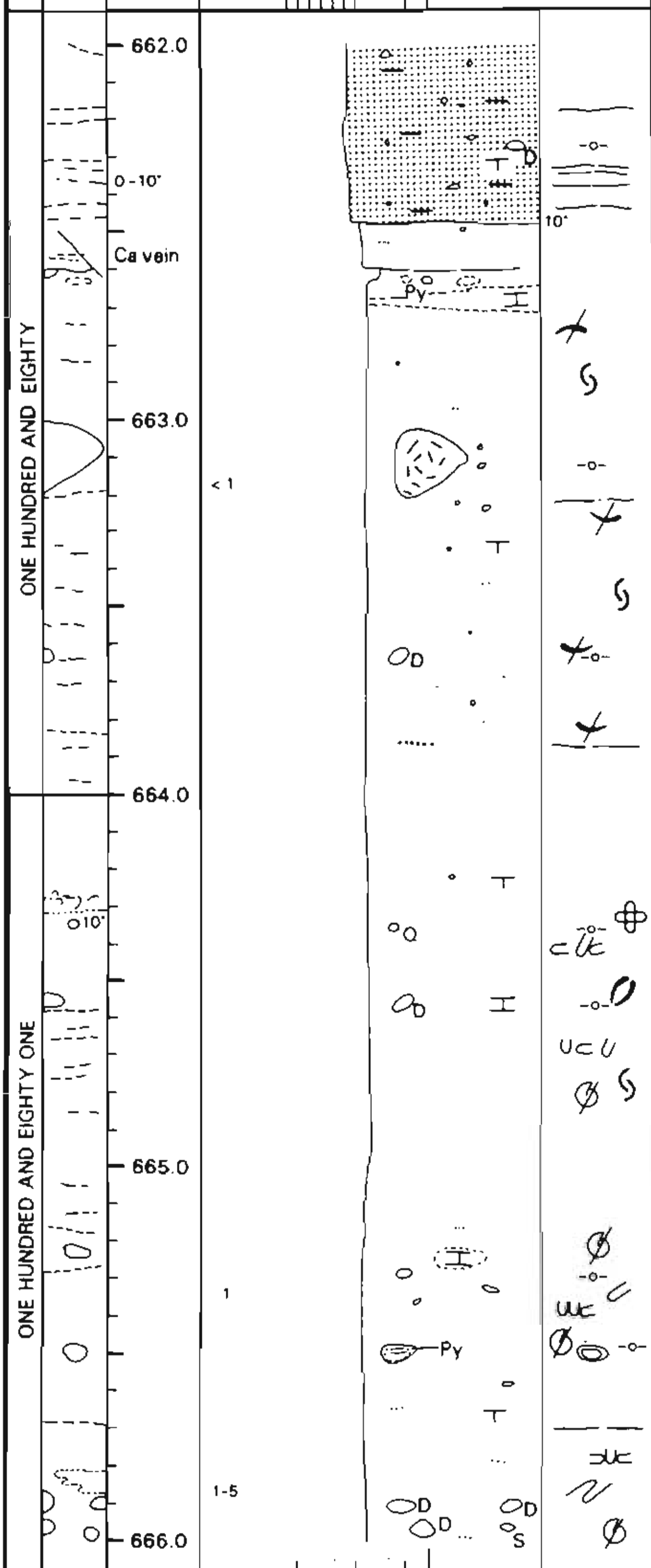
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently

PROJECT : CIROS -1

SHEET NO. : 161

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

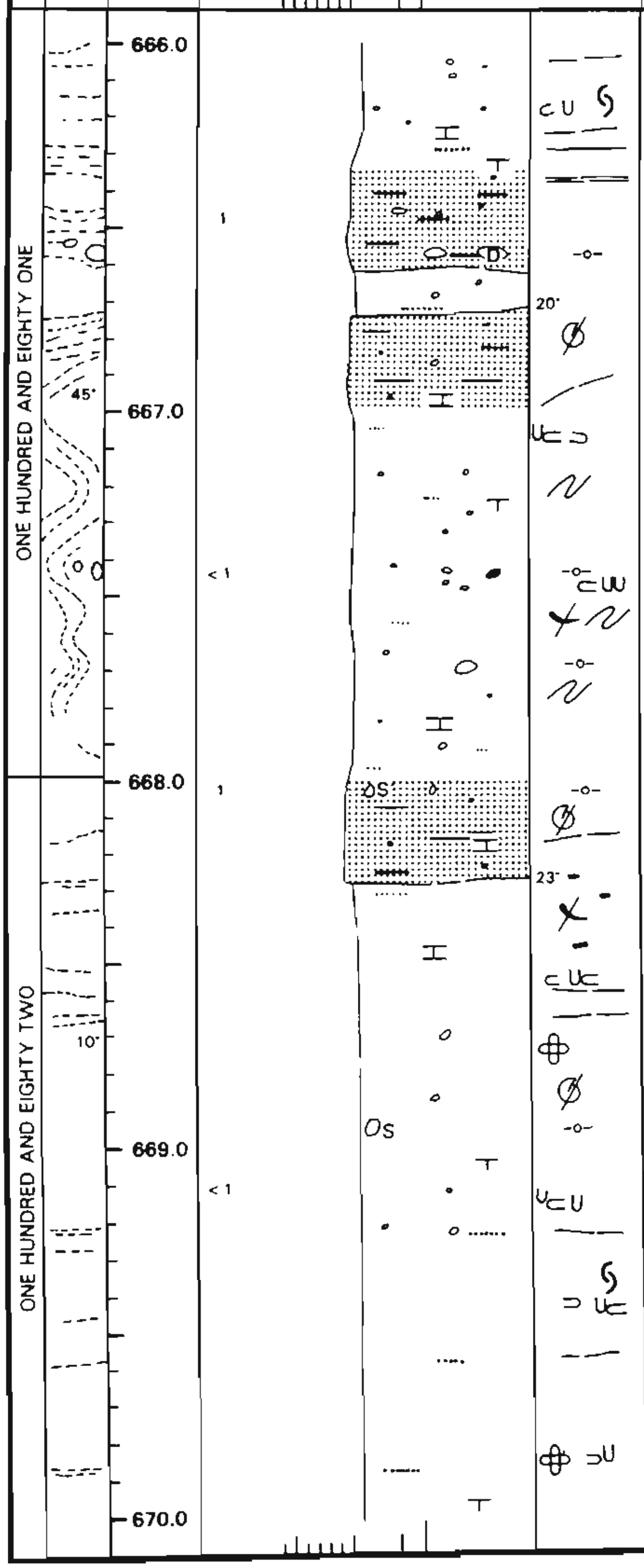
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS -1

SHEET NO. : 162

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE	LITHOLOGY	DESCRIPTION



Sub-unit 21.1

634.34 - 670.48 m

MUDSTONE, SANDSTONE AND DIAMICTITE

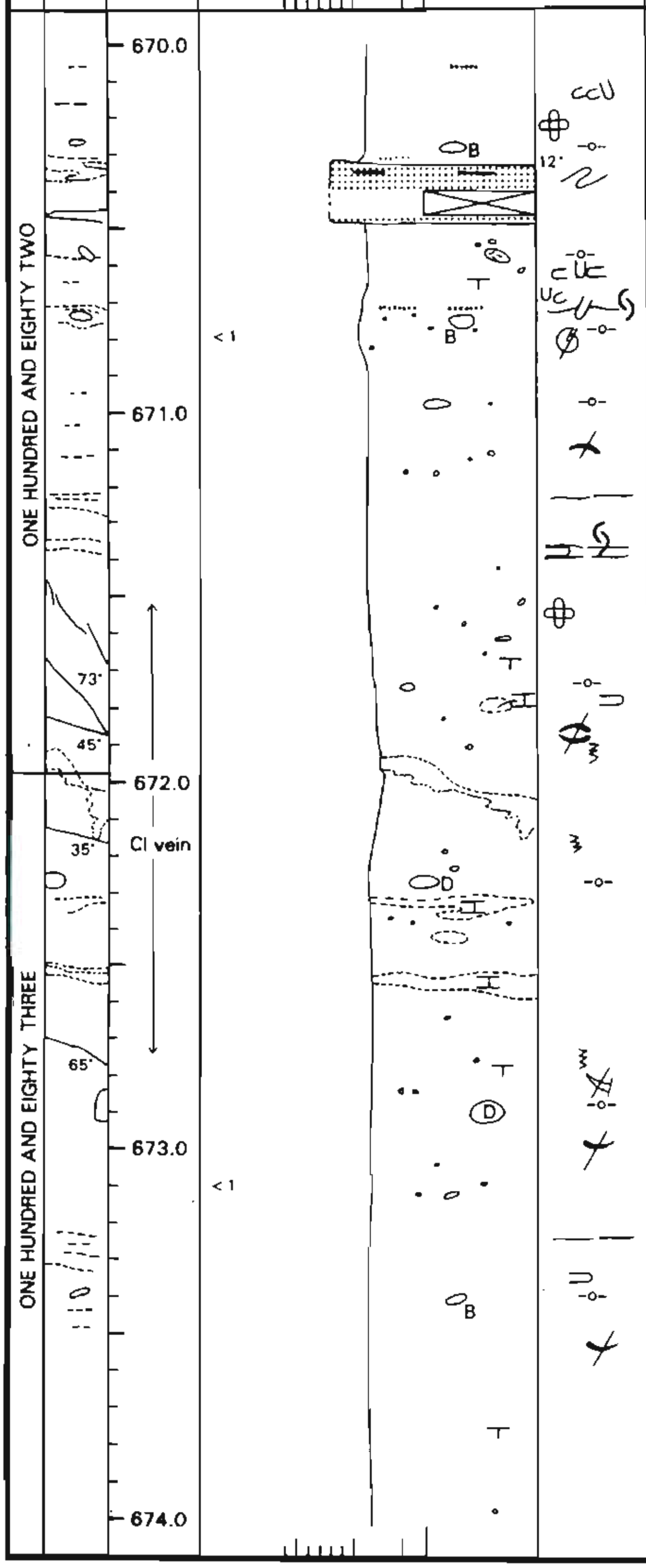
Hard, slightly calcareous, unstratified to well stratified, **sandy mudstone** (dark greenish grey 7.5GY 3/1 and 5G 3/1 to olive black 7.5Y 3/1) with interbeds of weakly stratified, **very fine sandstone** and occasional **conglomeratic muddy very fine sandstone** and **sandy mudstone (?diamictite)**. The mudstone has wispy irregular bedding inclined up to 40°, and strong isoclinal and recumbent folding from 639.50 to 644.0 m. Cross-bedding at 643.70 m. Below 647.69 m the mudstone is either unstratified with mottling or obviously bioturbated with disrupted bedding. The sandstone, more common below 649.49 m, has wispy subhorizontal bedding with occasional contorted and inclined (up to 45°) bedding. Asymmetric climbing ripples at 659.57 m. The ?diamictite (637.05 to 637.32 m and 665.27 to 668.29 m) has wispy bedding with isoclinal folding and a 1 m interval of steeply dipping (20), contorted bedding between 666.87 and 667.94 m. Clasts are predominantly dolerite (40%), basement (25%), indurated sediment (20%) and Beacon sandstone (< 5%). They occur as scattered lonestones in the mudstone and sandstone and form < 5% of the diamictite. Sizes are variable (1 to > 220 mm) with an average of 8-12 mm. A few are striated. Bioturbation structures with associated mottling are recognised in the lower sub-unit; also individual small (1-2 mm) and large (6-8 mm) inclined and sub-horizontal burrows, some with spreiten. Fossil shell fragments occur throughout. They include articulated and broken bivalves, scaphapods and general bioclastic debris. Carbonaceous fragments, and calcareous nodules and layers occur infrequently.

PROJECT : CIROS-1

SHEET NO. : 163

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



— sharp contact -

Sub-unit 21.2

670.48 – 685.82 m

MUDSTONE

Hard, slightly calcareous, unstratified to weakly stratified (slightly) **sandy mudstone** (grey 7.5Y 4/1, dark bluish grey 5BG 3/1 to dark olive grey 2.5GY 4/1).

Stratification is wispy and irregular with intervals of strong bioturbation where bedding is destroyed.

Lonestones are predominantly dolerite with minor basement, indurated sediment and Beacon sandstone. Some of the clasts are weathered. An indurated sediment lonestone is striated.

Two thin beds (674.46 to 674.52 and 675.50 to 675.53 m) are composed of **carbonaceous mudstone** (black N 2/1) and carbonaceous fragments. The interval is bioturbated with recognisable small (1–2mm) and large (6–10mm) burrows. Broken shell debris, including bivalves and scaphapods are common. Pyrite and carbonate nodules occur infrequently.

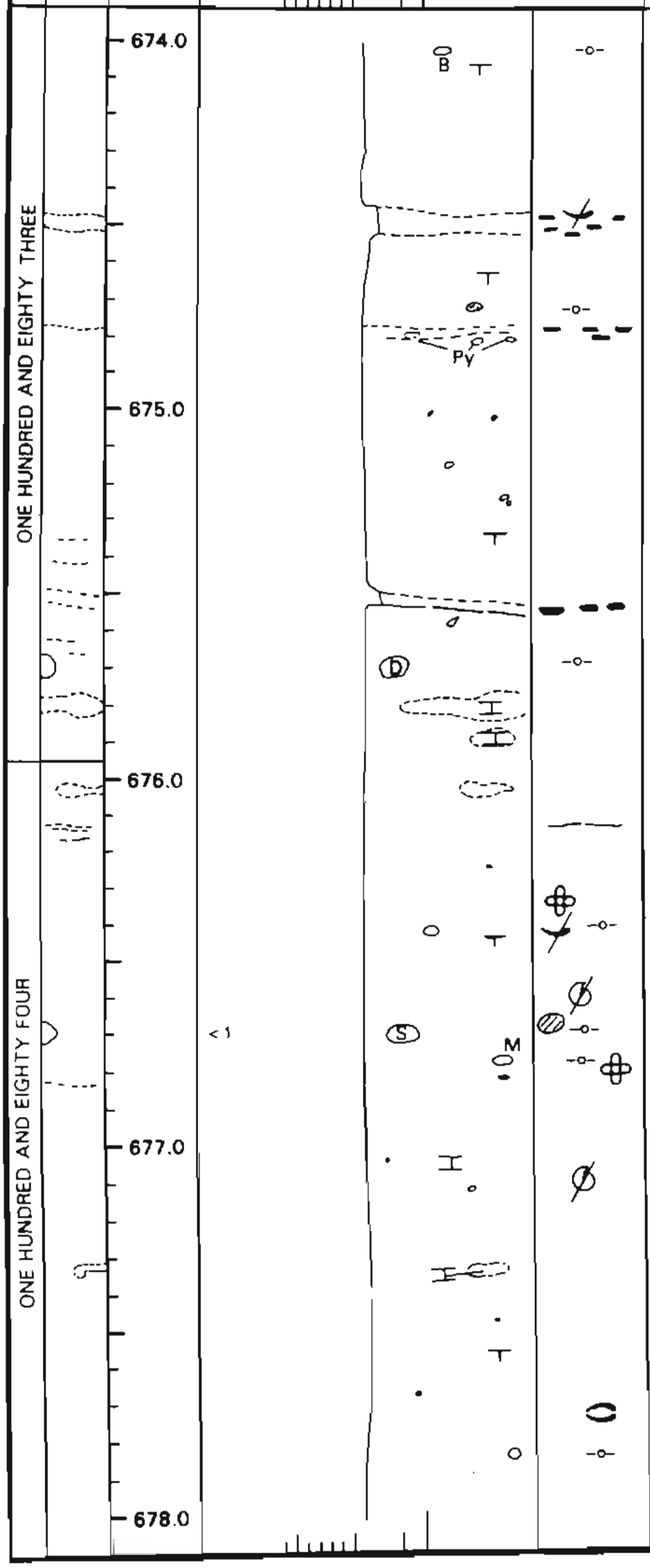
Slickensided surfaces dip between 35° and 65° at 671.83, 672.12, 672.70 and 680.17 m.

PROJECT : CIROS-1

SHEET NO. : 164

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.2

670.48 - 685.82 m

MUDSTONE

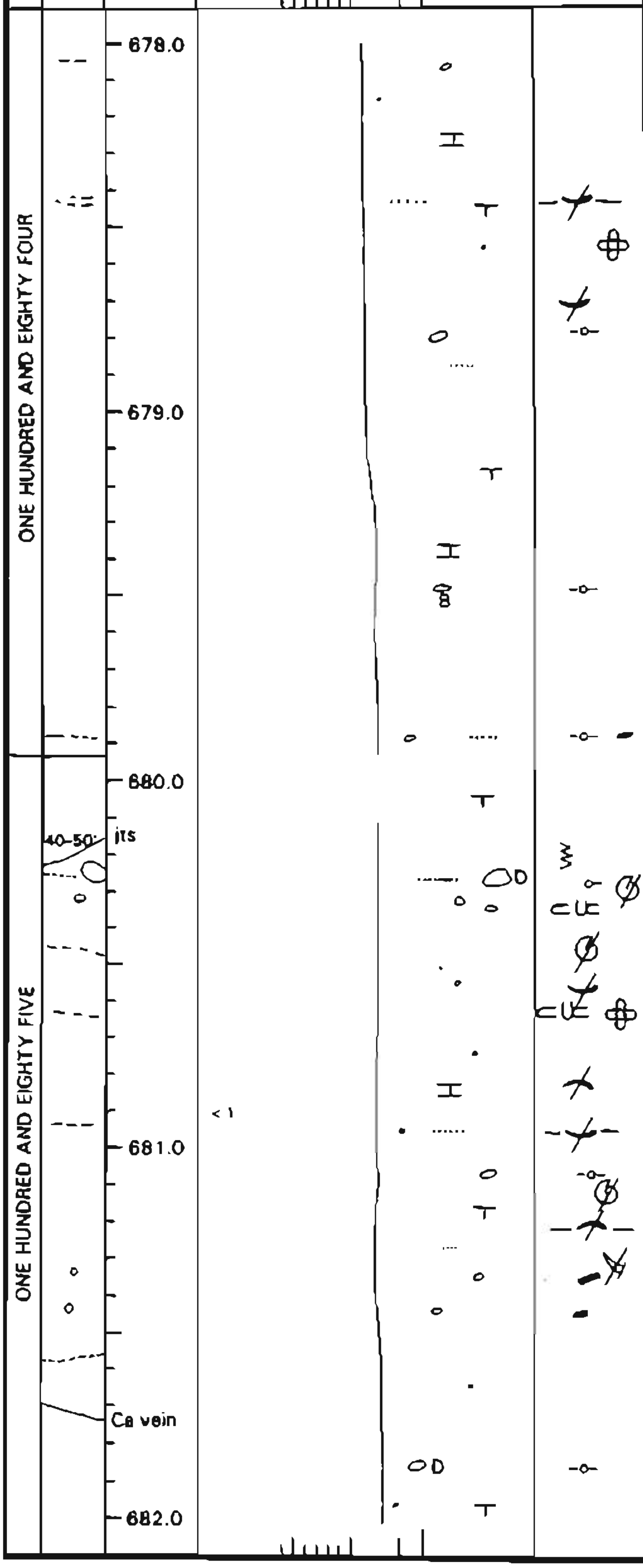
Hard, slightly calcareous, unstratified to weakly stratified (slightly) **sandy mudstone** (grey 7.5Y 4/1, dark bluish grey 5BG 3/1 to dark olive grey 2.5GY 4/1). Stratification is wispy and irregular with intervals of strong bioturbation where bedding is destroyed. Lonestones are predominantly dolerite with minor basement, indurated sediment and Beacon sandstone. Some of the clasts are weathered. An indurated sediment lonestone is striated. Two thin beds (674.46 to 674.52 and 675.50 to 675.53 m) are composed of carbonaceous **mudstone** (black N 2/1) and carbonaceous fragments. The interval is bioturbated with recognisable small (1-2mm) and large (6-10mm) burrows. Broken shell debris, including bivalves and scaphapods are common. Pyrite and carbonate nodules occur infrequently. Slickensided surfaces dip between 35° and 65° at 671.83, 672.12, 672.70 and 680.17 m.

PROJECT : CIROS-1

SHEET NO. : 165

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % SAND SILT CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 21.2

670.48 - 685.82 m

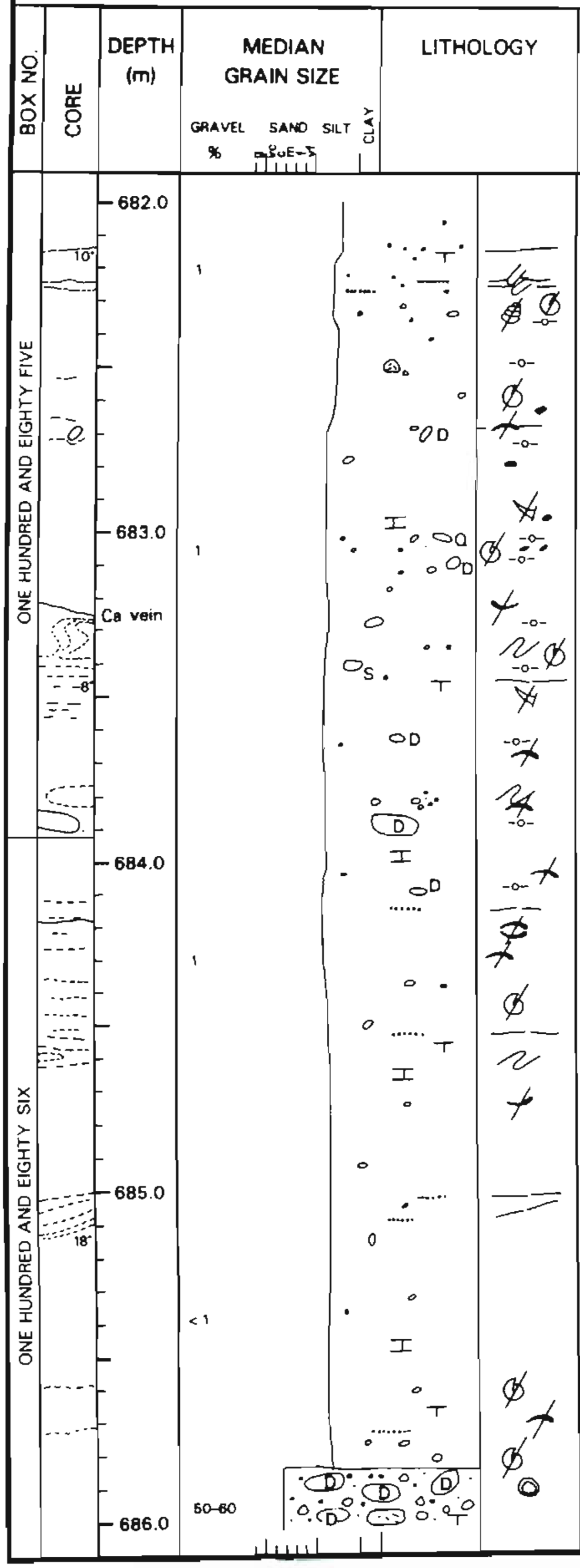
MUDSTONE

Hard, slightly calcareous, unstratified to weakly stratified (slightly) sandy mudstone (grey 7.5Y 4/1, dark bluish grey 5BG 3/1 to dark olive grey 2.5GY 4/1). Stratification is wispy and irregular with intervals of strong bioturbation where bedding is destroyed. Lonestones are predominantly dolerite with minor basement, indurated sediment and Beacon sandstone. Some of the clasts are weathered. An indurated sediment lonestone is striated. Two thin beds (674.46 to 674.52 and 675.50 to 675.53 m) are composed of carbonaceous mudstone (black N 2/1) and carbonaceous fragments. The interval is bioturbated with recognisable small (1-2mm) and large (6-10mm) burrows. Broken shell debris, including bivalves and scaphapods are common. Pyrite and carbonate nodules occur infrequently. Slickensided surfaces dip between 35° and 65° at 671.83, 672.12, 672.70 and 680.17 m.

PROJECT : CIROS-1

SHEET NO. : 166

SCALE : 1:20



DESCRIPTION

Sub–unit 21.2

670.48 - 685.82 m

MUDSTONE

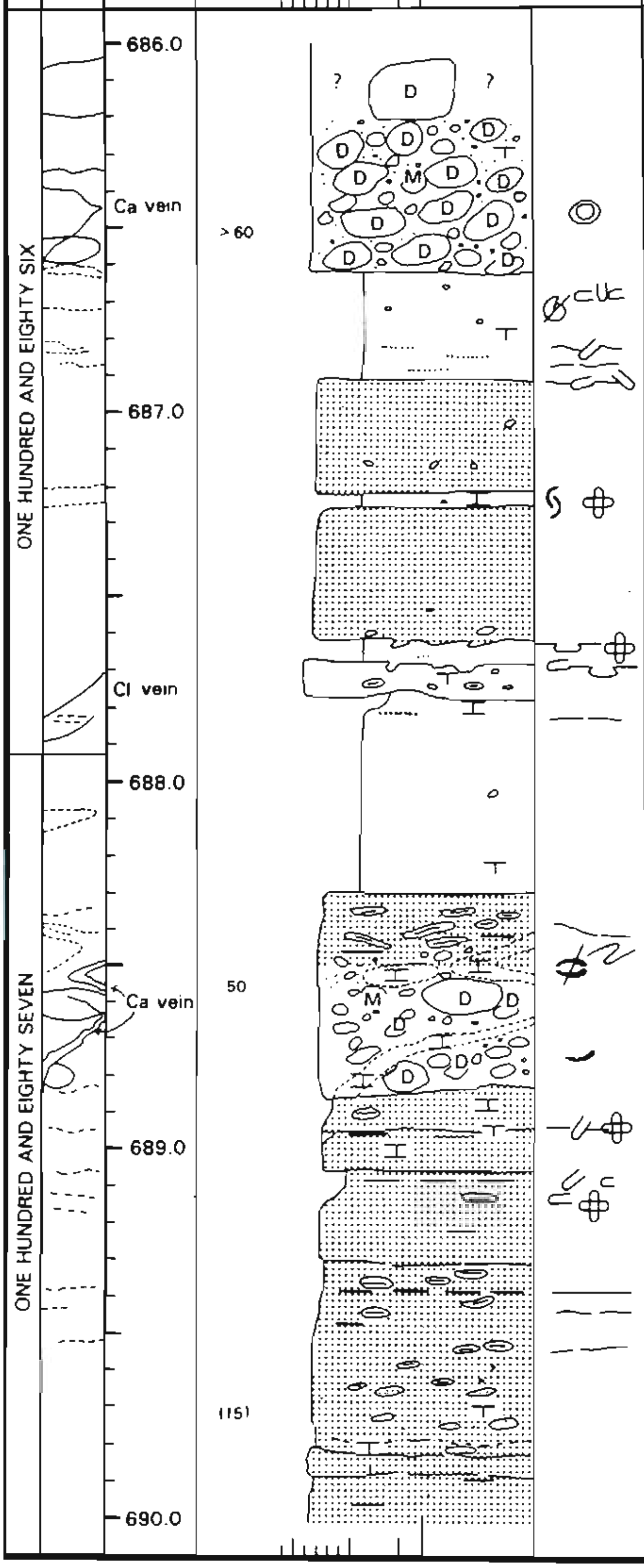
Hard, slightly calcareous, unstratified to weakly stratified (slightly) **sandy mudstone** (grey 7.5Y 4/1, dark bluish grey 5BG 3/1 to dark olive grey 2.5GY 4/1). Stratification is wispy and irregular with intervals of strong bioturbation where bedding is destroyed. Lonestones are predominantly dolerite with minor basement, indurated sediment and Beacon sandstone. Some of the clasts are weathered. An indurated sediment lonestone is striated. Two thin beds (674.46 to 674.52 and 675.50 to 675.53 m) are composed of carbonaceous **mudstone** (black N 2/1) and carbonaceous fragments. The interval is bioturbated with recognisable small (1–2mm) and large (6–10mm) burrows. Broken shell debris, including bivalves and scaphapods are common. Pyrite and carbonate nodules occur infrequently. Slickensided surfaces dip between 35° and 65° at 671.83, 672.12, 672.70 and 680.17 m.

PROJECT : CIROS-1

SHEET NO. : 167

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL %, SAND, SILT, CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 22.1

685.82 - 698.88 m

CONGLOMERATE, SANDSTONE AND MUDSTONE

Hard, slightly calcareous, unstratified, **sandy pebble conglomerate** which is clast- and matrix-supported. The matrix is poorly sorted, **coarse sandstone** (grey 7.5Y 5/1).

The interbeds are unstratified to well stratified, poorly sorted, **coarse sandstone** (olive grey 2.5GY 4/1 to 5GY 5/1 and carbonate-cemented greyish white N 7/1), and weakly stratified **sandy mudstone** (grey 7.5Y 4/1 to olive black 7.5Y 3/1).

Stratification in the sandstone and mudstone is very wispy and irregular, with minor contorted bedding, load structures and cross-bedding.

The conglomerates between 685.82 and 692.41 m contain predominantly subrounded clasts of dolerite (>90%) with minor basement (marble), indurated sediment and intraclasts of sandstone and mudstone. Clasts represent 15 to 70% of the rock, with sizes from 4 to 150 mm, averaging 20 mm. Below 692.91 m the conglomerates are matrix-supported and composed of mudstone and sandstone intraclasts (about 90%), dolerite (<10%) and basement (1%). The intraclasts represent from 15 to 50% of the rock; they are elongate and sub-horizontal, subrounded to subangular, and range from about 5 to >45 mm (average 18 mm).

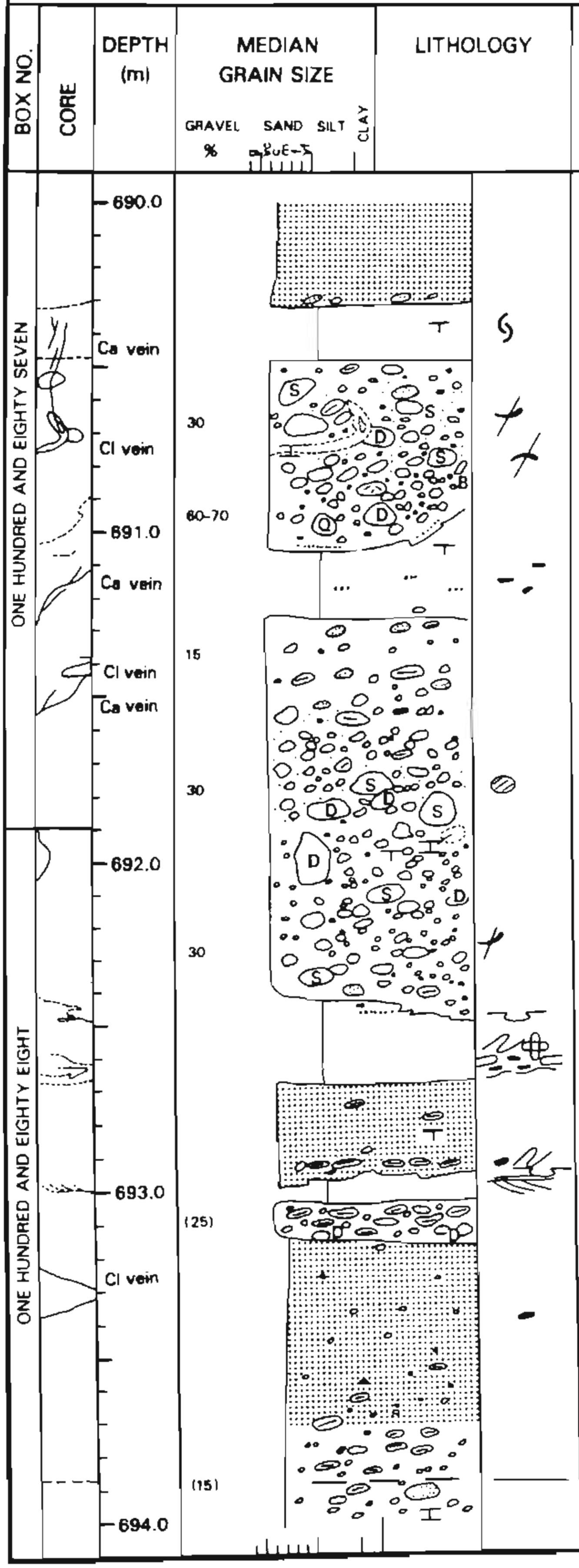
Carbonaceous fragments and flecks occur below 693.30 m. Bioturbation structures occur in mudstone beds. Bivalve fragments are scattered through all lithologies.

A weakly stratified, **slightly conglomeratic muddy coarse to medium sandstone** between 697.60 and 698.00 m may be considered as a **sandy diamictite**.

PROJECT : CIROS-1

SHEET NO. : 168

SCALE : 1:20



DESCRIPTION

Sub-unit 22.1

685.82 - 698.88 m

CONGLOMERATE, SANDSTONE AND MUDSTONE

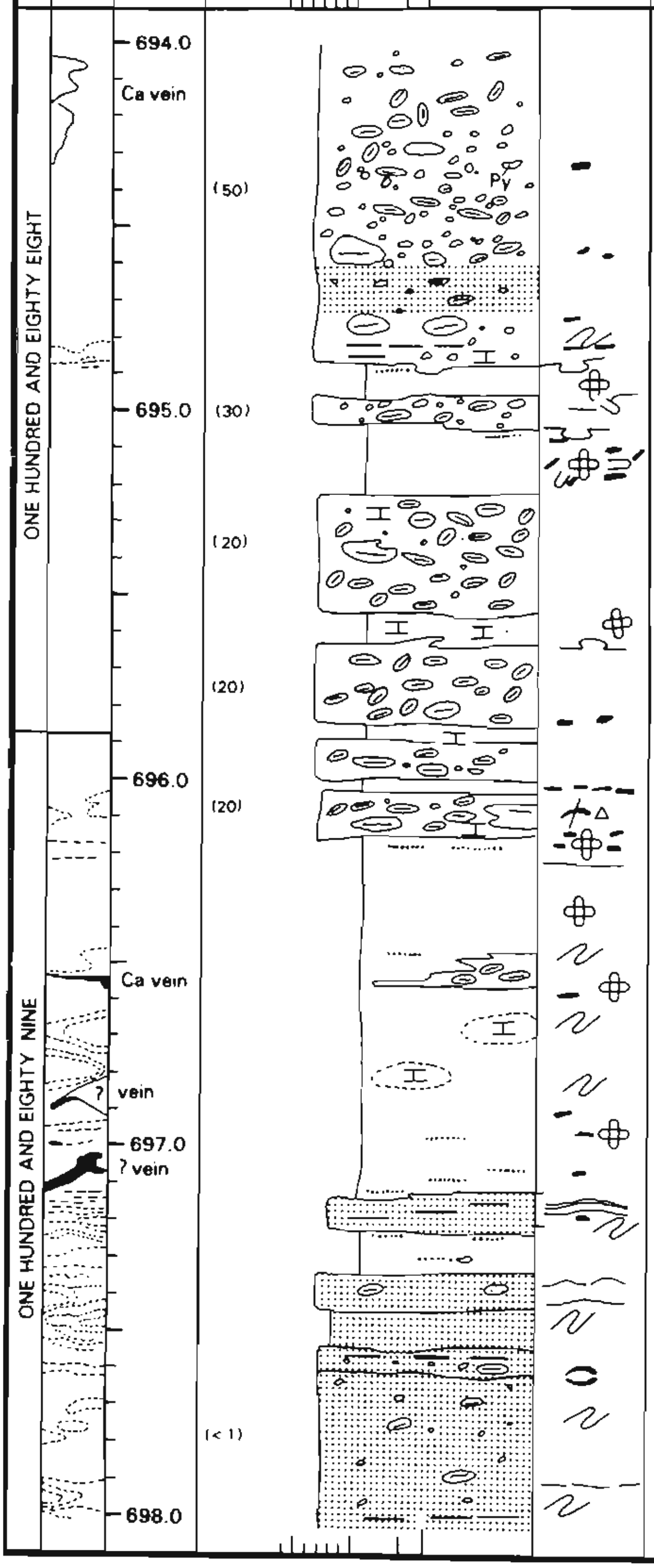
Hard, slightly calcareous, unstratified **sandy pebble conglomerate** which is clast- and matrix-supported. The matrix is poorly sorted, **coarse sandstone** (grey 7.5Y 5/1). The interbeds are unstratified to well stratified, poorly sorted, **coarse sandstone** (olive grey 2.5GY 4/1 to 5GY 5/1 and carbonate-cemented greyish white N 7/1), and weakly stratified **sandy mudstone** (grey 7.5Y 4/1 to olive black 7.5Y 3/1). Stratification in the sandstone and mudstone is very wispy and irregular, with minor contorted bedding, load structures and cross-bedding. The conglomerates between 685.82 and 692.41 m contain predominantly subrounded clasts of dolerite (> 90%) with minor basement (marble), indurated sediment and intraclasts of sandstone and mudstone. Clasts represent 15 to 70% of the rock, with sizes from 4 to 150 mm, averaging 20 mm. Below 692.91 m the conglomerates are matrix-supported and composed of mudstone and sandstone intraclasts (about 90%), dolerite (< 10%) and basement (1%). The intraclasts represent from 15 to 50% of the rock; they are elongate and sub-horizontal, subrounded to subangular, and range from about 5 to > 45 mm (average 18 mm). Carbonaceous fragments and flecks occur below 693.30 m. Bioturbation structures occur in mudstone beds. Bivalve fragments are scattered through all lithologies. A weakly stratified, **slightly conglomeratic muddy coarse to medium sandstone** between 697.60 and 698.00 m may be considered as a **sandy diamictite**.

PROJECT : CIROS-1

SHEET NO. : 169

SCALE : 1:20

BOX NO.	CORE	DEPTH (m)	MEDIAN GRAIN SIZE (GRAVEL % / SAND / SILT / CLAY)	LITHOLOGY	DESCRIPTION



Sub-unit 22.1

685.82 - 698.88 m

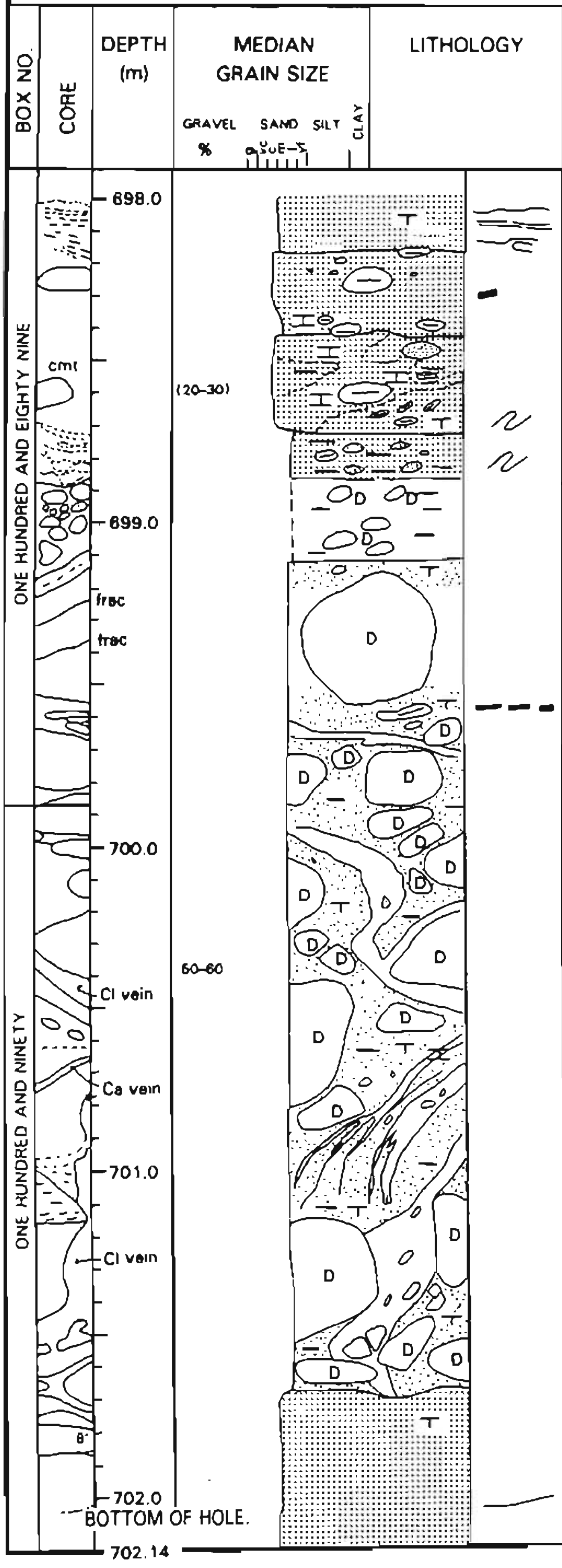
CONGLOMERATE, SANDSTONE AND MUDSTONE

Hard, slightly calcareous, unstratified **sandy pebble conglomerate** which is clast- and matrix-supported. The matrix is poorly sorted, **coarse sandstone** (grey 7.5Y 5/1). The interbeds are unstratified to well stratified, poorly sorted, **coarse sandstone** (olive grey 2.5GY 4/1 to 5GY 5/1 and carbonate-cemented greyish white N 7/1), and weakly stratified **sandy mudstone** (grey 7.5Y 4/1 to olive black 7.5Y 3/1). Stratification in the sandstone and mudstone is very wispy and irregular, with minor contorted bedding, load structures and cross-bedding. The conglomerates between 685.82 and 692.41 m contain predominantly subrounded clasts of dolerite (> 90%) with minor basement (marble), indurated sediment and intraclasts of sandstone and mudstone. Clasts represent 15 to 70% of the rock, with sizes from 4 to 150 mm, averaging 20 mm. Below 692.91 m the conglomerates are matrix-supported and composed of mudstone and sandstone intraclasts (about 90%), dolerite (< 10%) and basement (1%). The intraclasts represent from 15 to 50% of the rock; they are elongate and sub-horizontal, subrounded to subangular, and range from about 5 to > 45 mm (average 18 mm). Carbonaceous fragments and flecks occur below 693.30 m. Bioturbation structures occur in mudstone beds. Bivalve fragments are scattered through all lithologies. A weakly stratified, **slightly conglomeratic muddy coarse to medium sandstone** between 697.60 and 698.00 m may be considered as a **sandy diamictite**.

PROJECT : CIROS-1

SHEET NO. : 170

SCALE : 1:20



DESCRIPTION

— sharp contact -

Sub-unit 22.2

698.88 - 701.67 m

FAULT BRECCIA IN BOULDER CONGLOMERATE

Hard, poorly sorted, **sandy boulder conglomerate** with no obvious stratification.

Clast composition is almost entirely dolerite with minor mudstone intraclasts (one with carbonaceous flecks at 699.59 m). Clasts range from 70 mm up to 470 mm. They appear angular to subangular, and are both matrix- and clast-supported.

The matrix is coarse to medium sandstone (dark olive grey 2.5GY 3 1) which in places is very coarse sandstone and grit.

Hard, **claystone** (bluish grey 10BG 5/1) fills large (up to 30mm wide) joints and shear planes. Calcite veins occur from 700.67 to 700.98m.

Veins, joints and shear planes separate large clasts, but in situ brecciation is apparent.

— sharp contact -

Sub-unit 22.3

701.67 - 702.14 m

SANDSTONE

Hard, unstratified, poorly sorted, **coarse to medium sandstone** (dark olive grey 2 5GY 3 1)

Table 2. Downhole depth for the core interval in each core box.

Box #	Core Top	Core Bottom	Box #	Core Top	Core Bottom
1	0.0	3.70	5	36.00	38.91
2	26.79	29.63	6	38.91	41.83
3	29.63	33.01	7	41.83	45.64
4	33.01	36.00	8	45.64	48.73
9	48.73	51.73	13	60.52	63.42
10	51.73	54.57	14	63.42	66.38
11	54.57	57.72	15	66.38	69.40
12	57.72	60.52	16	69.40	72.58
17	72.58	75.39	21	84.33	87.36
18	75.39	78.36	22	87.36	90.32
19	78.36	81.36	23	90.32	93.28
20	81.36	84.33	24	93.28	96.16
25	96.16	99.10	29	108.05	111.07
26	99.10	102.06	30	111.07	114.09
27	102.06	105.03	31	114.09	117.14
28	105.03	108.05	32	117.14	120.13
33	120.13	123.21	37	131.11	135.03
34	123.21	126.19	38	135.03	138.11
35	126.19	129.18	39	138.11	141.10
36	129.18	132.11	40	141.10	144.02
41	144.02	146.97	45	156.00	159.37
42	146.97	149.94	46	159.37	162.24
43	149.94	152.93	47	162.24	165.21
44	152.93	156.00	48	165.21	168.19
49	168.19	171.16	53	179.85	182.81
50	171.16	174.15	54	182.81	185.77
51	174.15	177.06	55	185.77	188.75
52	177.06	179.85	56	188.75	191.69
57	191.69	194.67	61	203.74	207.93
58	194.67	197.64	62	207.93	210.92
59	197.64	200.63	63	210.92	213.64
60	200.63	203.74	64	213.64	216.63
65	216.63	219.62	69	228.58	231.56
66	219.62	222.60	70	231.56	234.54
67	222.60	225.58	71	234.54	239.28
68	225.58	228.58	72	239.28	242.27
73	242.28	245.27	77	254.25	257.24
74	245.27	248.24	78	257.24	260.23
75	248.24	251.26	79	260.23	263.22
76	251.26	254.25	80	263.22	266.27
81	266.27	266.45	85	278.25	282.28
82	266.45	270.26	86	282.28	286.63
83	270.26	274.24	87	286.63	290.65
84	274.24	278.25	88	290.65	294.62
89	294.62	299.18	93	311.14	315.08
90	299.18	303.16	94	315.08	319.07
91	303.16	307.15	95	319.07	323.19
92	307.15	311.14	96	323.19	328.06
97	328.06	332.03	101	344.04	348.29
98	332.03	336.01	102	348.29	352.27
99	336.01	340.00	103	352.27	356.32
100	340.00	344.44	104	356.32	360.26
105	360.26	364.48	109	376.43	380.54
106	364.48	368.47	110	380.54	384.51
107	368.47	372.45	111	384.51	388.50
108	372.45	376.43	112	388.50	392.49

Table 2. Downhole depth for the core interval in each core box (cont'd).

Box #	Core Top	Core Bottom	Box #	Core Top	Core Bottom
113	392.49	396.48	117	408.44	412.43
114	396.48	400.46	118	412.43	416.41
115	400.46	404.45	119	416.41	420.40
116	404.45	408.44	120	420.20	424.39
121	424.39	428.38	125	440.30	444.29
122	428.38	432.32	126	444.29	448.27
123	432.32	436.31	127	448.27	452.26
124	436.31	440.30	128	452.26	456.24
129	456.24	460.24	133	472.21	476.20
130	460.24	464.23	134	476.20	480.20
131	464.23	468.22	135	480.20	484.20
132	468.22	472.21	136	484.20	488.18
137	488.18	492.17	141	504.14	508.13
138	492.17	496.16	142	508.13	512.13
139	496.16	500.15	143	512.13	516.12
140	500.15	504.14	144	516.12	520.11
145	520.11	524.10	149	536.07	540.05
146	524.10	528.09	150	540.05	544.04
147	528.09	532.08	151	544.04	548.02
148	532.08	536.07	152	548.02	552.01
153	552.01	556.00	157	567.97	571.96
154	556.00	559.99	158	571.96	575.95
155	559.99	563.98	159	575.95	579.94
156	563.98	567.97	160	579.94	583.93
161	583.93	587.93	165	599.86	603.85
162	587.93	591.89	166	603.85	607.84
163	591.89	595.88	167	607.84	611.83
164	595.88	599.86	168	611.83	615.82
169	615.82	620.23	173	632.12	636.11
170	620.23	624.18	174	636.11	640.10
171	624.18	628.13	175	640.10	644.09
172	628.13	632.12	176	644.09	648.07
177	648.07	652.05	181	664.01	667.99
178	652.05	656.03	182	667.99	671.98
179	656.03	660.02	183	671.98	675.96
180	660.02	664.01	184	675.96	679.94
185	679.94	683.92	189	695.88	669.87
186	683.92	687.91	190	699.87	702.14 TD
187	687.91	691.90			
188	691.90	695.88			