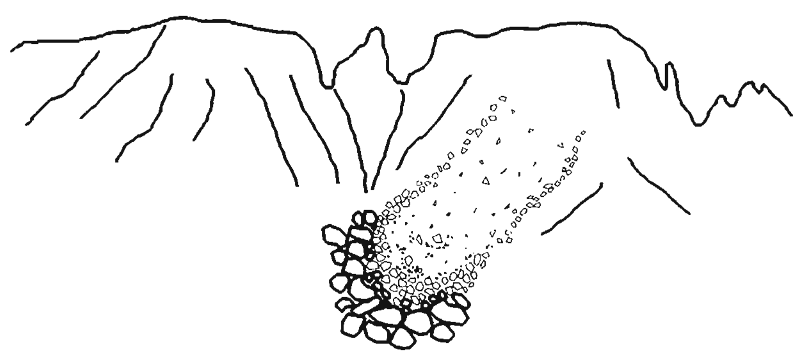
AN EVALUATION OF THE PERIGLACIAL MORPHOLOGY IN THE HIGH DRAKENSBERG AND ASSOCIATED ENVIRONMENTAL IMPLICATIONS



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PREFACE

The research described in this dissertation was carried out in the Department of Geography, University of Natal, Pietermaritzburg; Department of Geography and Environmental Studies, University of the Witwatersrand, Johannesburg and in the high Drakensberg, South Africa and Lesotho, from July 1992 to November 1997, under the supervision of Professor Kevin Hall.

These studies represent original work by the author and have not otherwise been submitted in any form for any degree or diploma to any other university. Where use has been made of the work of others, it has been duly acknowledged in the text.

Stefan W. Grab
University of Natal, Pietermaritzburg
November, 1997

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ABSTRACT

Although periglacial research in the high Drakensberg and Lesotho mountains has received growing interest amongst southern African geomorphologists, little detailed, quantitative information was available prior to this study. In an attempt to help overcome this deficit, a quantitative assessment on cryogenic landforms and processes operative in the high Drakensberg was undertaken.

Morphological and sedimentological assessments of sorted patterned ground, non-sorted steps, thufur, blockstreams, stone-banked lobes, debris deposits and turf exfoliation landforms were undertaken. In addition, geomorphic process assessments in the field included the measurement of turf retreat at turf exfoliation sites, the determination of frost-heave mechanisms within wetlands and sediment mobilization along the Mashai Stream. Ground temperatures were recorded for thufur from 1993 to 1996. The environmental implications of some of the findings are discussed.

Seasonal frost-induced sorted patterned ground emerges annually within a few weeks, demonstrating the effect of regular, diurnal freeze-thaw cycles during the winter months. It is found that the present climate is not conducive to maintaining or preserving miniature periglacial landforms below 3200 m a.s.l. during the summer months. Large relict sorted circles, stone-banked lobes and blockstreams are the most conspicuous periglacial landforms in the high Drakensberg and are products of at least seasonally-frozen ground. It is suggested that debris deposits found within high Drakensberg cutbacks are possible indicators for marginal niche and cirque glaciation during the Late Pleistocene. It is demonstrated that in climatically marginal periglacial regions, the microtopographically controlled freezing processes may be of paramount importance in maintaining and modifying the cryogenic landforms that occur. Pronounced temperature differentials are found during the winter months, when thufur are frozen for several weeks and depressions remain predominantly unfrozen. It is suggested that such contemporary temperature differentials induce thermodynamic forces and ultimately ground heave at sites in the high Drakensberg. The pronounced seasonal weather patterns in the high Drakensberg have promoted a cycle of geomorphic process events that operate synergistically and initiate particular erosion landforms. However, cryogenic activity during the colder period is overwhelmed by water induced erosion processes during the summer months in the high Drakensberg. It is concluded that the high Drakensberg is currently a marginal periglacial region, but that periglacial conditions prevailed during both the Pleistocene and some Late Holocene Neoglacial events.

CONTENTS

Page

CHAPTER FIVE : NON-SORTED PATTERNED GROUND

CHAPTER SIX : MASS-WASTING FORMS AND DEBRIS DEPOSITS

Page

CHAPTER EIGHT : DISCUSSION AND CONCLUSIONS

LIST OF FIGURES

Figure *Page*

CHAPTER 1

CHAPTER 2

Figure *Page*

CHAPTER 3

CHAPTER 4

Figure *Page*

CHAPTER 6

CHAPTER 7

CHAPTER 8

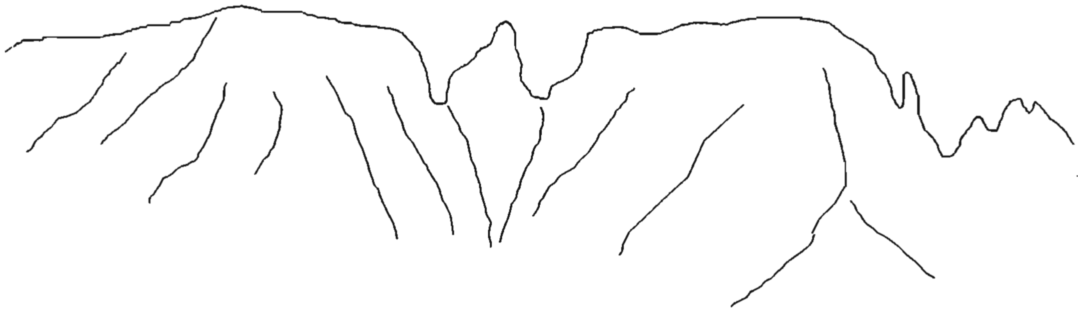
LIST OF TABLES

Table *Page*

CHAPTER 8

CHAPTER ONE

INTRODUCTION



CHAPTER 1

INTRODUCTION

1.1 PERIGLACIAL GEOMORPHOLOGY

The term "periglacial" was introduced by Polish geologist Walery von Łoziński in 1909 to describe climatic and geomorphic conditions peripheral to Pleistocene ice sheets and glaciers (Łoziński, 1909). Since the term was first introduced, there has been rigorous debate and criticism surrounding its definition (e.g. Linton, 1969; Washburn, 1973; Karte and Liedtke, 1981; Thorn, 1992; Warburton, 1992; Barsch, 1993). "Periglacial" is, at present, being used as a "blanket term which spans all cold climate studies" (Warburton, 1992; p35), but is predominately used by cold region geomorphologists. Although Warburton (1992) has recommended that the term "periglacial" should only describe a zone "adjacent to existing, or former glaciers or ice sheets" (p35), others (e.g. French, 1976; Harris, 1988) favour environmental rather than geographical definitions. For instance, Harris (1988) argues that permafrost is the most valuable definable characteristic of a periglacial environment. It would appear, however, that many researchers are prepared to accept more broad-based, less restricted definitions such as that provided by Thorn (1992, p1):

> "Periglacial geomorphology is that part of geomorphology which has as its primary object physically based explanations of the past, present, and future impacts of diurnal, seasonal, and perennial ground ice on landform and landscape initiation and development. Additional components of the subdiscipline include similar investigations of the geomorphic roles of snowpacks (but not glaciers) and fluvial, lacustrine, and marine ice."

Periglacial geomorphology established itself officially as an academic discipline within the International Geographical Union (IGU) in 1949. A *Commission on Periglacial Geomorphology* was approved in 1956 with Prof. J. Dylik elected as Chairman. In 1972, the commission was disbanded and replaced by the IGU *Co-ordinating Committee for Periglacial Research* and in 1989 renamed as the *Commission on Frost Action Environments*. During the eight year term from 1989 to 1996, the Commission on Frost Action Environments held annual scientific meetings under the leadership of Jean-Pierre Lautridou. A new IGU Commission on *Climatic Change and Periglacial Environments* was approved during the IGU congress in the Hague in August 1996. So as to meet the growing interest in permafrost studies, the International Permafrost Association (IPA) was founded in 1983, although meetings of the non-formal body had taken place for a number of years prior. In July 1989, the IPA became affiliated to the International Union of Geological Sciences. A primary objective of the IPA is the convening of International Permafrost Conferences of which six had been held by 1993.

Recently, there has been a considerable focus of interest in the Natural Sciences as it is hoped that they may offer new perspectives on global climatic and environmental change (Dixon and Abrahams, 1992). Recognizing that the cryosphere is a sensitive index of climatic change and a storehouse of information on past climates (Marcus *et al.*, 1992), there has been a resurgence of interest in cold climate research (c.f. Dixon and Abrahams, 1992). Focus of much recent discussion has been the many periglacial research problems facing the discipline (e.g. Washburn, 1985; French, 1987; Thorn, 1992). For instance, Hall (1991) and Thorn (1992) have cautioned the equating of form to process, such as the assumption of angular clasts being the product of freeze-thaw weathering. Freeze-thaw has often been considered the sole weathering process in many cold regions, however, as pointed out by Hall and Lautridou (1991), several weathering processes are now being recognized as significant in periglacial environments. Equally, it has for a long time been assumed that chemical weathering is of little importance in cold environments, but Dixon *et al.* (1984) have indicated that chemical weathering may be of great importance and Balke *et al.* (1991) have even shown it can be an active denudational component on the Antarctic continent.

Until recently, the focus of periglacial work had concentrated to a large extent on describing "form". More recently, however, there has been increasing emphasis on process studies, with the purpose of interpreting the previously described forms (Harris, 1988). Barsch (1993) has reiterated that present trends should follow scientifically sound physically-based models of periglacial processes as these may enable us to interpret relict landforms and predict geomorphic responses to environmental change.

1.2 PERIGLACIAL GEOMORPHOLOGY IN SOUTHERN AFRICA

A brief historical overview of southern African periglacial research is presented and a list of publications and researchers given in Table 1.1. Periglacial research in the Drakensberg appears to have begun with Troll's work at Mont-aux-Sources in 1944. Following a hiatus during the 1950's, a growing interest in cold region geomorphology during the 1960's resulted in 17 publications from 1960 to 1974 (Table 1.1; Figure 1.1). Foreign-based researchers who showed great interest in southern African periglacial studies at the time, contributed 50% of these publications. However, it was only Sparrow and Hastenrath who established themselves as southern African periglacial "specialists" during this time, as other workers only produced a single publication each. It would appear that increasing political pressure on South Africa impeded the further influx of foreign-based researchers over the next 10 years. This may, at least in part, be the reason why only nine publications were produced between 1974 and 1988 (Figure 1.1). However, as mentioned, there has been a global resurgence of interest in cold region geomorphology (Dixon and Abrahams, 1992), which has also had an impact in southern Africa. A new group of periglacial geomorphologists has emerged, producing no fewer than 40 scientific papers on southern African geocryological studies from 1988 to 1996 (Figure 1.1), of which 13 appear in international journals. This clearly shows the increasing desire of researchers to bring greater international interest to this little known region.

Of the 67 papers published by December 1996 that focus on southern African geocryological studies, 30 cover broad perspectives of the periglacial/glacial environment, as exemplified by such as Lewis (1988a; 1988b; 1996a; 1996b), Boelhouwers (1991a; 1991b) and Hanvey and Marker (1992) (Table 1.1). Although 27 papers have focused on specific cryogeomorphic forms, not one study has examined frost action processes *per se*. The absence of data and the extensive use of qualitative judgements has initiated critical debate that has resulted in several analytical papers being published in short succession (e.g. Le Roux and Marker, 1990; Hall, *et al.*, 1991; Hall, 1992; Boelhouwers, 1995a; Sumner, 1995; Hall, 1995; Grab and Hall, 1996). Clearly, there is a need for more empirical data from this region.

GEOCRYOLOGICAL STUDIES IN SOUTHERN AFRICA				
Theme	Reported By	Relict or Active	Altitude (m)	Location
Needle-Ice	Troll, 1944	Active	3000 - 3300	Drak. (Mont-Aux-Sources)
General	Ellenberger, 1960	Relict	Plateau	Lesotho
General	Alexandre, 1962	Relict/Active	Plateau	Lesotho
Nivation	Sparrow, 1964	Relict	----	Drakensberg/Lesotho
Slope Formation	Sparrow, 1965	Relict	----	Drakensberg
Cirque	Sparrow, 1967a	Relict	----	Drakensberg/Lesotho
General	Sparrow, 1967b	Relict	----	Drakensberg/Lesotho
General	Harper, 1969	Relict/Active	Plateau	Drakensberg/Lesotho
General	Linton, 1969	Relict	----	Southern Africa
General	Sparrow, 1971	Relict	>1500	Southern Africa
Nivation Cirque	Marker & Whittington, 1971	Relict	>2900	Drakensberg (Sani-Pass Area)
General	Hastenrath, 1972	Relict/Active	Plateau	Southern Africa
Discussion	Sparrow, 1973	Relict	----	Southern Africa
General	Hastenrath & Wilkinson, 1973	Relict/Active	Plateau	Lesotho
General	Butzer, 1973	Relict/Active	Plateau	Drakensberg/Lesotho
Hollow/Head deposit	Nicol, 1973	Relict	1630 - 2840	Golden Gate N.P.
Solifluction/Cirque	Sparrow, 1974	Relict	>1700	South Eastern Lesotho
General	Van Zinderen Bakker & Werger, 1974	Active	Plateau	Lesotho
Periglacial soils	Fitzpatrick, 1978	Relict	>2280	Drakensberg (Sani Pass)
Nivation/Glacial Hollows	Dyer & Marker, 1979	Relict	>3000	Lesotho
Perigl. & Glacial	Borchert & Sänger, 1981	Relict	----	Western Cape Mountains
General	Hagedorn, 1984	Relict	----	Western Cape Mountains
Ice-wedge casts	Lewis & Dardis, 1985	Relict	1980	E.Cape Drak. (Barkley Pass)
Patterned ground	Dardis & Granger, 1986	Active	3000 - 3200	Drak. (Champagne Castle)
Periglacial slope deposits	Hanvey et al, 1986	Relict	2225	E.Cape Drak. (Carlisle's Hoek)
Scree Tongues	Marker, 1987	Relict	>1800	E. Cape Drakensberg
Pleistocene Glaciation	Sänger, 1987	Relict	----	Western Cape Mountains
Valley Asymmetry	Boelhouwers, 1988b	Relict/Active	2000 - 3000	Drakensberg (Giant's Castle)
Discussion	Lewis, 1988a	Relict/Active	----	Southern Africa
Discussion	Lewis, 1988b	Relict/Active	----	Southern Africa
Pleistocene Glaciation	Sänger, 1988a	Relict	----	Western Cape Mountains
Glaciation/Periglacial	Sänger, 1988b	Relict/Active	----	Western Cape Mountains
Debris Slope Deposits	Lewis & Hanvey, 1988	Relict	----	E. Cape Drak. (Rhodes)
General	Marker, 1989	Relict/Active	1800 - 2837	Golden Gate H.N.P.
Nivation	Marker, 1990a	Relict/Active	2100 - 2300	Golden Gate H.N.P.
General	Marker, 1990b	Relict/Active	----	Golden Gate H.N.P.
Correspondence	Le Roux & Marker, 1990	Relict	----	Golden Gate H.N.P.
General	Boelhouwers & Hall, 1990	Relict/Active	>2900	Sani Pass Area & Lesotho
General	Boelhouwers, 1991a	Active	>2300	Drakensberg/Lesotho
General	Boelhouwers, 1991b	Relict/Active	>1800	Western Cape Mountains
Freeze - Thaw (Discussion)	Hall, 1991	----	----	Southern Africa
Cirque glaciation	Marker, 1991	Relict	>3000	Eastern Lesotho
Glaciation	Gellert, 1991	Relict	>1700	W. Cape Mountains
Solifluction	Lewis & Hanvey, 1991	Relict	±1900 - 2000	E.Cape Drak. (Glen Orchy)
Solifluction & cryo-nival	Hanvey & Lewis, 1991	Relict	±2000	E.Cape Drak. (Sonskyn)
Correspondence	Hall, Marker & Le Roux, 1991	Relict/Active	----	Southern Africa
Review	Hall, 1992	Relict/Active	----	Southern Africa
General	Marker, 1992	Relict	>1700	Drakensberg/Lesotho
General	Hanvey & Marker, 1992	Active	±3000 - 3275	Lesotho (Tlaeng Pass)
Valley Asymmetry	Meiklejohn, 1992	Relict	Plateau	Drakensberg/Lesotho
Rock Glaciers	Lewis & Hanvey, 1993	Relict	>1820	E.Cape Drakensberg
Quaternary/General	Lewis, 1994	Relict	>2000	E.Cape Drakensberg
Protalus Ramparts	Lewis, 1994	Relict	>2000	E.Cape Drakensberg
Thufur	Grab, 1994	Relict/Active	2860-3270	Drak./Lesotho (Mohlesi Valley)
Valley Asymmetry	Meiklejohn, 1994	Relict	Plateau	Antarctica/Drak./Lesotho
Rock Cutbacks	Hall, 1994	Relict	2500 - 3000	Giant's Castle (Drak.)
Periglacial Landforms	Boelhouwers, 1994	Relict/Active	3140 - 3300	Giant's Castle (Drak.)
Correspondence	Sumner & Hall, 1995	Relict	2500 - 3000	Giant's Castle (Drak.)
Review	Boelhouwers, 1995a	Relict/Active	----	Southern Africa
General	Boelhouwers, 1995b	Active	±1850 - 2000	Hexriver Mountains
Periglacial Gradients	Marker, 1995a	Relict	----	Southern Africa
Periglacial Landforms	Lewis, 1996a	Relict/Active	1550 - 2900	E. Cape Drakensberg
Glacial Landforms	Lewis, 1996b	Relict	> 1800	E. Cape Drakensberg
Correspondence	Grab & Hall, 1996	Relict	Plateau	Drakensberg/Lesotho
Niche Glaciation	Grab, 1996a	Relict	> 2900	Drakensberg
Sorted stripes	Grab, 1996b	Relict/Active	3380 - 3410	Drakensberg/Lesotho

Table 1.1 Geocryological studies in southern Africa.

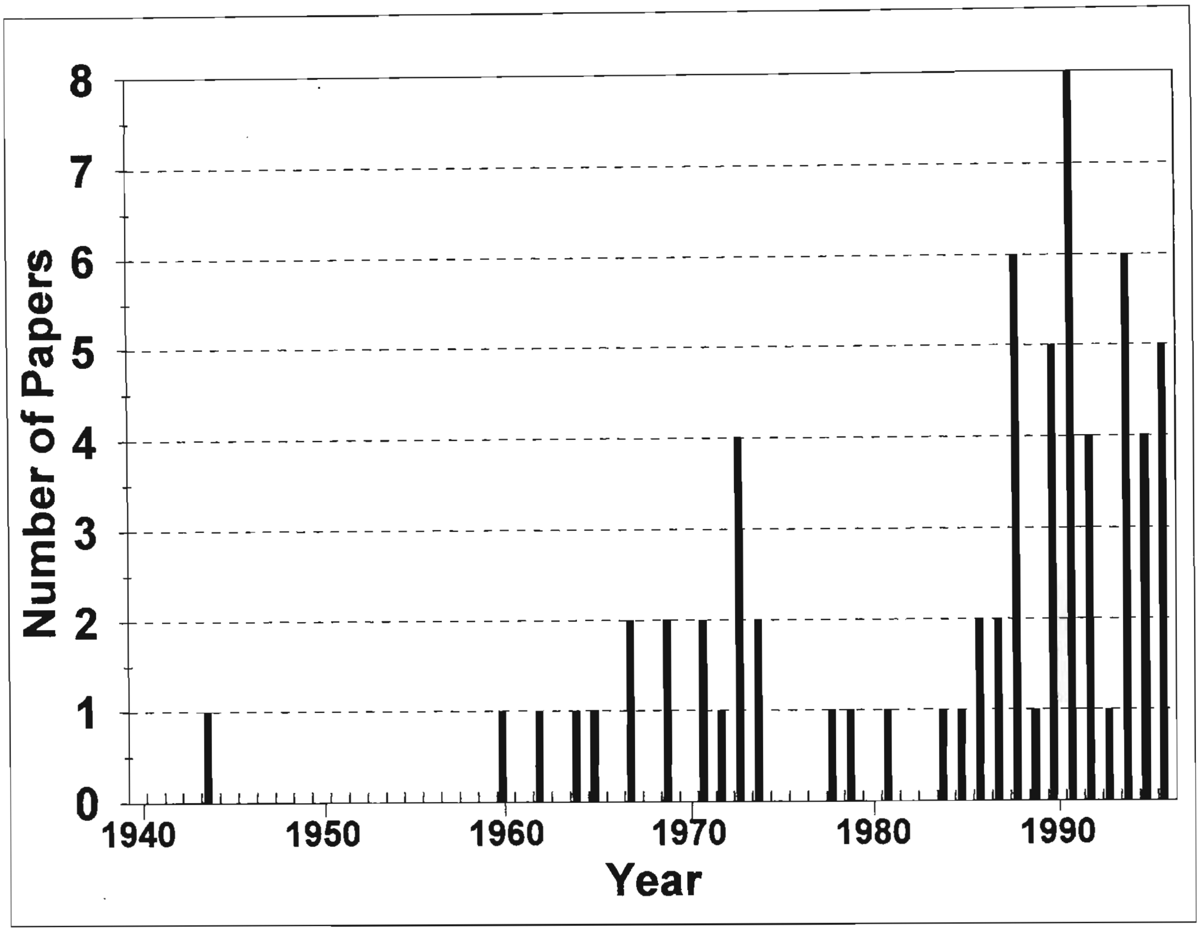


Figure 1.1 The distribution of publications on southern African geocryological studies.

1.2.1 Research Shortcomings from the Drakensberg Region

Although periglacial research is receiving growing interest in southern Africa, few papers of any substance have been published. Hall (1992) and Boelhouwers (1995a) have already addressed some of the periglacial research problems and show that rigour is lacking with respect to both terminology and interpretation. With the rapidly increasing focus on southern African periglacial topics, it seems appropriate to critically examine several of the perceived shortcomings regarding periglacial research in the Drakensberg region, as this dissertation should be viewed within such a framework.

1.2.1.1 Limited and Site Specific Research

A map showing the distribution of active periglacial features in the Drakensberg and Lesotho mountains (Figure 1.2a) has been presented in several papers (Lewis, 1988a, 1988b, 1994; Hanvey, 1990), yet its precision is to be questioned. The map shows large areas apparently devoid of periglacial features, but a review of the literature indicates that fieldwork has been limited to but a few specific sites (Figure 1.2b). Consequently, the map represents a generalization for the region as a whole, but extrapolated from only a few sites. Although it is difficult to overcome site-specific research, particularly in such an extensive and often inaccessible area, more caution is required when implying the results are applicable to the region as a whole.

1.2.1.2 Assumptions Regarding Cryogenic Processes

Although several comments have been made with respect to past cryogenic processes and landscape initiation in the Drakensberg, these are based primarily on assumptions deduced from inadequate temporal and spatial field observations and so result in contradictory findings. Researchers have not undertaken extensive field measurements and the gathering of field data in the Drakensberg has been extremely limited. For instance, Harper (1969) believed that basalt steps found in the Drakensberg are a product of frost wedging. Frost wedging is a process that Harper felt to be most prominent where the lithological structures are favourable. It was further assumed that frost wedging is intense at nivation sites. Essential data (e.g. rock temperature and rock moisture content) to support the ideas of frost wedging were, however, not produced by Harper. Equally, Marker (1990a) assumes that because the air temperature at Golden Gate Highlands National Park did not rise above 0°C for at least a week during the time of investigation, frost action would have been considerable where moisture was available. However, to determine whether frost action is "considerable," it is the rock/soil temperature and moisture content that must be known (Hall, 1991), rather than (the here recorded) air temperature.

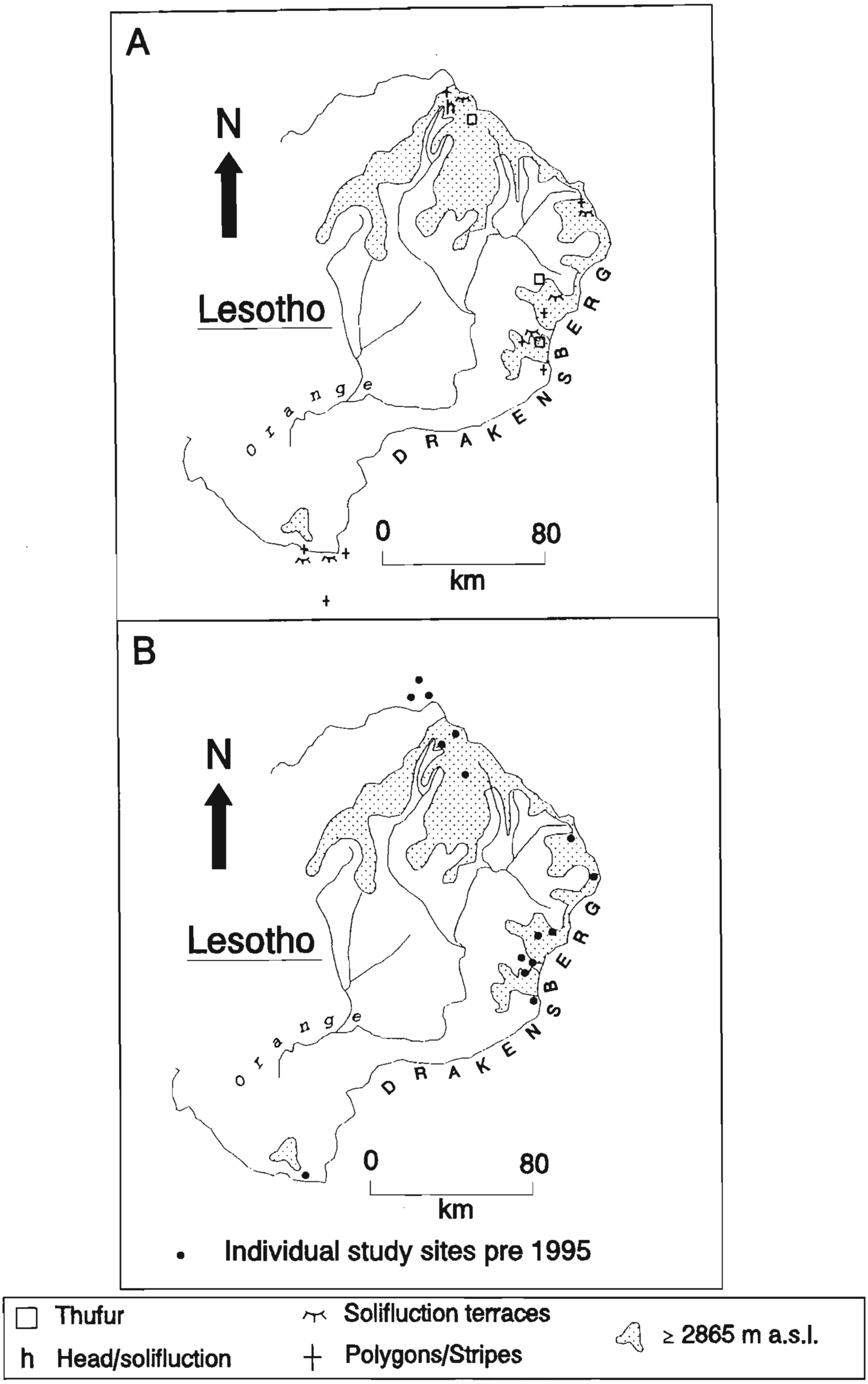


Figure 1.2 a. The distribution of active periglacial features in the high Drakensberg (according to Lewis, 1988a, 1988b). **b.** Individual periglacial fieldwork sites in the high Drakensberg.

Discussions regarding the occurrence of freeze-thaw epitomise the problem of inadequate data. In a discussion on fieldwork potential at Golden Gate Highlands National Park, Marker (1989) suggests that angular screes are produced when blocks break off cliffs due to frost shattering, while Sparrow (1967b), like many other researchers, believes that angular boulders indicate frost shattering. Hall and Lautridou (1991) stress that it is very difficult to deduce what processes take place in the field when rock temperature, rock moisture content and rock moisture chemistry are unknown, and Thorn (1992) notes that, for rock samples found in the field, there are no known criteria diagnostic of frost action. The assumption that frost shattering has caused angular screes cannot be accepted unless the essential field data are produced (Hall, 1991), particularly as several weathering processes may produce angular rock material (McGreevy and Whalley, 1982). What many researchers have not acknowledged is that several weathering processes, such as wetting and drying, salt weathering, thermal fatigue, chemical and biological weathering, may be significant within periglacial environments (Hall, 1991; Hall and Lautridou, 1991). Even if frost shattering did occur, it is most likely that it operates synergistically with other weathering processes.

It has been suggested that nivation and/or weak cirque glaciation occurred on the high plateau at least once during the Pleistocene (Sparrow, 1964; Marker and Whittington, 1971; Nicol, 1973; Dyer and Marker, 1979; Marker, 1991; Grab, 1996a). Harper (1969), Sparrow (1971) and Marker (1989) all support the idea that late-lying snowpatches enhance freeze-thaw weathering, with the underlying rock being broken-up. However, it has been shown by Hall (1980) that the geomorphic effectiveness of freeze-thaw cycles at nivation sites is limited due to the snow cover which reduces air temperature fluctuations from reaching the underlying rock surface. A more recent study examining geomorphic processes at snow patch sites in northern Sweden could find no proof of frost weathering associated with these sites (Nyberg, 1991). In fact, Ballantyne *et al.* (1989) found evidence of enhanced chemical, rather than mechanical, weathering at snowpatch sites. It would seem that southern African researchers have failed to recognise that nivation is, in many ways, an imprecise concept (Thorn and Hall, 1980; Thorn, 1988) and that much confusion surrounds the understanding of the process-association intrinsic to nivation (Thorn, 1988). A further problem is that the arguments presented to justify nivation in southern Africa are largely speculative and based on morphometric, rather than process, data.

1.2.1.3 Terminological and Definitional Problems

Terminological problems that have been the concern of periglacial geomorphology in southern Africa are largely the result of a lack of consistency or agreement with respect to the terms used by different authors. Another problem has been the lack of understanding regarding processes or their controlling factors.

It appears that some authors find it difficult to discern the difference between the "nivation hollow" and the "glacial cirque", not only morphologically but also with respect to their processes or controlling factors. Several authors (e.g. Harper, 1969; Sparrow, 1971; Marker and Whittington 1971) believe that the "cirque" is formed because of nivation being sufficiently prolonged to substantially enlarge the original "niche". From the range of terms used by southern African authors it is often difficult to determine if they are dealing with glacial cirques or nivation hollows; there being a potential important palaeoenvironmental distinction between the two. Similar landforms have been referred to as "cirque-shaped hollows", "cirques" (Sparrow, 1967a), "nivation cirques", "nivation hollows" (Marker and Whittington, 1971) and "cirque like hollows" (Dyer and Marker, 1979). This may all appear semantic argument and yet it is worth noting that at the first meeting of the BGRG Small Research Group on Geomorphometry it was agreed that the nivation hollow "lacks some element of the cirque form" (Evans and Cox, 1974; p151).

Since Muller (1947) defined the term "permafrost", it has always been understood as a ground temperature condition in which earth materials remain below 0°C for two or more years. Marker (1989) incorrectly categorizes ground that is only frozen during the winter months as a form of permafrost. From this, it appears that some authors are unfamiliar with the terms they use, or are uncertain about the features they describe. In short, much greater terminological precision is required in southern African cryogenic studies (Hall, 1992).

International definitions cannot be ignored and greater recourse to such glossaries as that of Harris *et al.* (1988) should be considered.

1.2.1.4 Altitudinal Zonation

Periglacial features in the Drakensberg have frequently been grouped into fixed altitudinal zones without adequate analysis of their actual distribution. While more data on altitudinal limits are required, there has also been a failure to establish climate-process links for the Drakensberg.

According to Hastenrath and Wilkinson (1973), needle ice has been found to break up vegetation cover at altitudes above 3000 m, but Boelhouwers (1991a) points out that needle ice growth may cause localized vegetation disruption as low as 2300 m a.s.l. Indeed, from the occurrence of needle ice, it has been concluded that diurnal freezing is geomorphically significant above 2300 m a.s.l. (Boelhouwers, 1991a). Personal observations show that localized needle ice may occur from the highest Drakensberg summits down to elevations of about 1000 m a.s.l. in the KwaZulu-Natal midlands, and may play a significant geomorphological role down to at least 1500 m a.s.l., where needle ice (5 cm in length) has been seen to grow and move sediments within undercut stream banks.

With respect to thufur, an altitudinal range from 2900 to about 3100 m has been given (Hastenrath, 1972) and supported (e.g. Lewis, 1988a, 1988b; Boelhouwers, 1991a). However, Grab (1994) has recently shown that thufur occur from at least 2860 m a.s.l. through to 3270 m a.s.l. in the Mohlesi Valley. Equally, the lower limit for polygonal patterned ground has been given at 2550 m a.s.l. (Lewis, 1988a), yet polygons in a perennial pond have been observed at an altitude of 2250 m (Grab, 1992b). According to Lewis (1988a), polygons that do not occupy such preferential positions have their lower limit at 2800 m a.s.l. Because it is moisture during the drier, colder months that is the limiting factor for the development of patterned ground, such patterns are most commonly found in preferential positions where moisture is collected and stored. Nelson (1989) questions the physical significance of the concept of "lower limit" because most small periglacial landforms are notably dependent on moisture for their development and continued activity. In reality, observations in the Drakensberg have been far too few and localized to enable the validation of altitudinal boundaries for thufur and patterned ground.

Marker (1995) suggests a Pleistocene Periglacial gradient in southern Africa, largely on the basis of two types of erosional hollows. Earlier, Dyer and Marker (1979) and Marker (1991) 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. However, Grab and Hall (1996) have shown that it is unlikely that snowpatches should survive longer on the warmer north-facing slopes than on the cooler south-facing slopes. The idea of a Pleistocene periglacial gradient for southern Africa, based predominantly on the hollows, therefore becomes problematical. It has not been acknowledged that placing Pleistocene periglacial gradients against latitude is rather complex, as factors such as regional climate, vegetation and lithology must be taken into account.

Although strict altitudinal limits are unlikely to be found, it may be possible to examine the altitudinal frequency distribution (Nelson, 1989) of certain processes and landforms within relatively homogeneous areas. To establish any frequency distribution, it would, however, be necessary to establish climate-process links for the Drakensberg region.

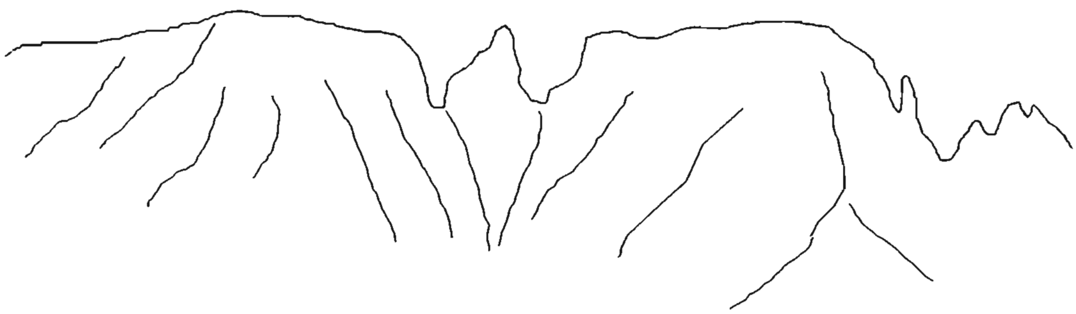
1.3 AIMS AND OBJECTIVES

As a result of the shortcomings outlined above, southern African periglacial geomorphology has achieved little with respect to broadening the understanding of the present-day Drakensberg Afro-alpine environment. Equally, such studies have been unable to provide substantive information on palaeoenvironments and environmental change. Here, an attempt is hence made to provide empirical data that may ultimately help improve the current status of southern African periglacial research and contribute towards the establishment of more applied work; thereby benefiting such as botanists and rural development projects.

The types of active and relict cryogenic landforms occurring in several parts of the Drakensberg are determined and their morphology described. An attempt is made to establish some of the environmental parameters required to initiate the formation of such landforms, and subsequently, to assess whether they are currently active or relict features. A further objective is to determine the past and present environmental implications of such landforms and processes. For instance, inactive thufur which are presently inhabited by rodents may be an indication that the present wetlands are retreating or becoming increasingly degraded. A spatial and temporal assessment of active processes and landforms is undertaken and some process studies (e.g. sediment movement by needle ice) examined. The study also evaluates the extent to which periglacial processes are overwhelmed by other geomorphic processes and/or anthropogenic influences in this area.

CHAPTER TWO

ENVIRONMENTAL SETTING



CHAPTER 2

ENVIRONMENTAL SETTING

2.1 THE STUDY REGION

The highest and most extensive mountain range in southern Africa is that of Quathlamba (Zulu name for "a mass of spears"), more commonly known as the Drakensberg (Dutch name for "Mountains of the Dragon"). This mountain range extends about 962 km from Pietersberg in the Northern Transvaal to Elliot in the Eastern Cape (Figure 2.1). The highest section of the Drakensberg forms a natural border between eastern Lesotho and KwaZulu-Natal, with the only access route for vehicles being at Sani Pass (Figure 2.1). The KwaZulu-Natal section of the Drakensberg extends about 280 km, attaining an average escarpment altitude of almost 3000 m and a maximum altitude of 3482 m at Thabana-Ntlenyana. The Natal Parks Board has divided the KwaZulu-Natal Drakensberg into 3 geographic sections: the northern, central and southern Drakensberg. For the purpose of this thesis, the term "high Drakensberg" refers to the higher altitude zones of the Drakensberg, generally above 2850 m a.s.l., including the Lesotho plateau and the higher reaches of the escarpment. The term "Little Berg" refers to the lower Drakensberg altitudes (ie between 2850 and 1600 m a.s.l.). The study region is located in the high Drakensberg (Lesotho plateau) (Figure 2.1), and maps have been produced to show the individual study areas (Figures 2.2, 2.3 and 2.4) and the distribution of selected geomorphic landforms (Figures 2.5, 2.6 and 2.7).

1. Sanqebethu Valley - Mafadi Summit region (central Drakensberg)

Latitude: 29° 11' 52" S to 29° 15' 00" E, Longitude: 29° 19' 50" E to 29° 25' 45" E

Altitudinal range: 2960 to 3450 m (Figure 2.2)

2. KwaNtuba - Mohlesi Valley region (southern Drakensberg)

Latitude: 29° 27' 15" S to 29° 31' 20" S, Longitude: 29° 16' 15" E to 29° 19' 20" E

Altitudinal range: 2840 to 3482 m (Figure 2.3)

3. Wilson's Peak - Mashai Valley region (southern Drakensberg)

Latitude: 29° 41' 40" S to 29° 45' 10" S, Longitude: 29° 7' 45" E to 29° 9' 22" E

Altitudinal range: 2900 to 3431 m (Figure 2.4)

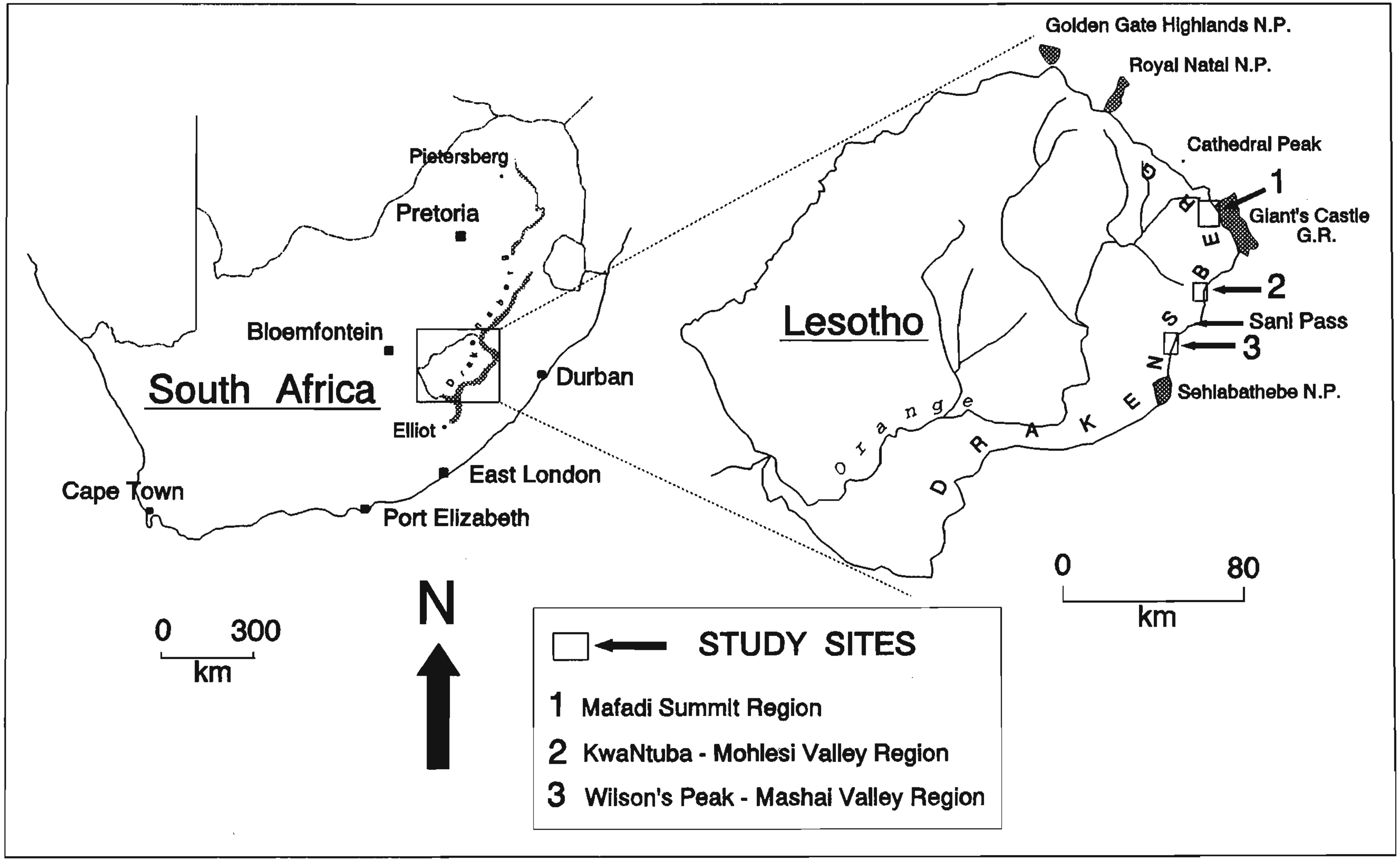


Figure 2.1 The study region and individual study sites.

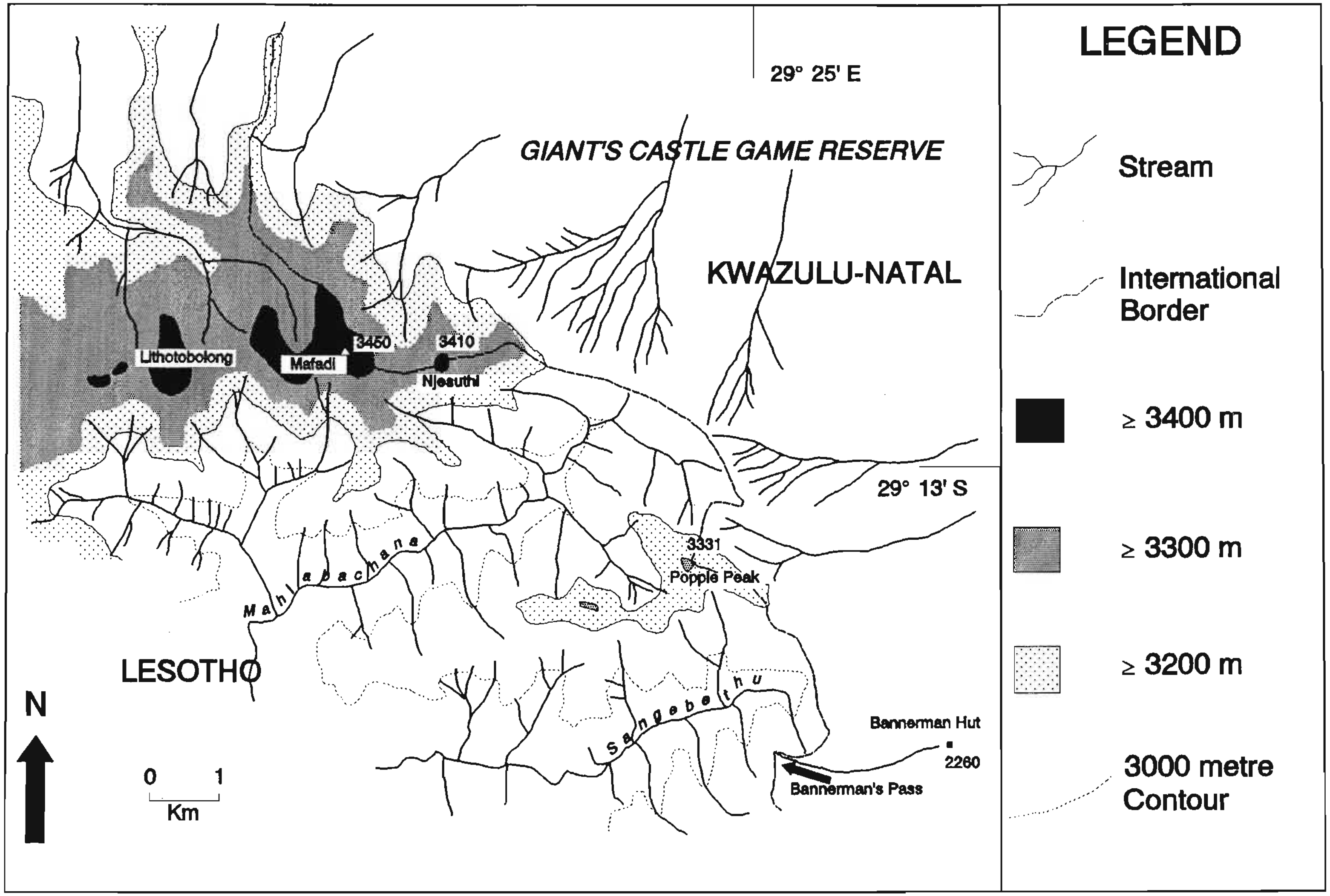


Figure 2.2 The Sanqebethu Valley - Mafadi Summit study region.

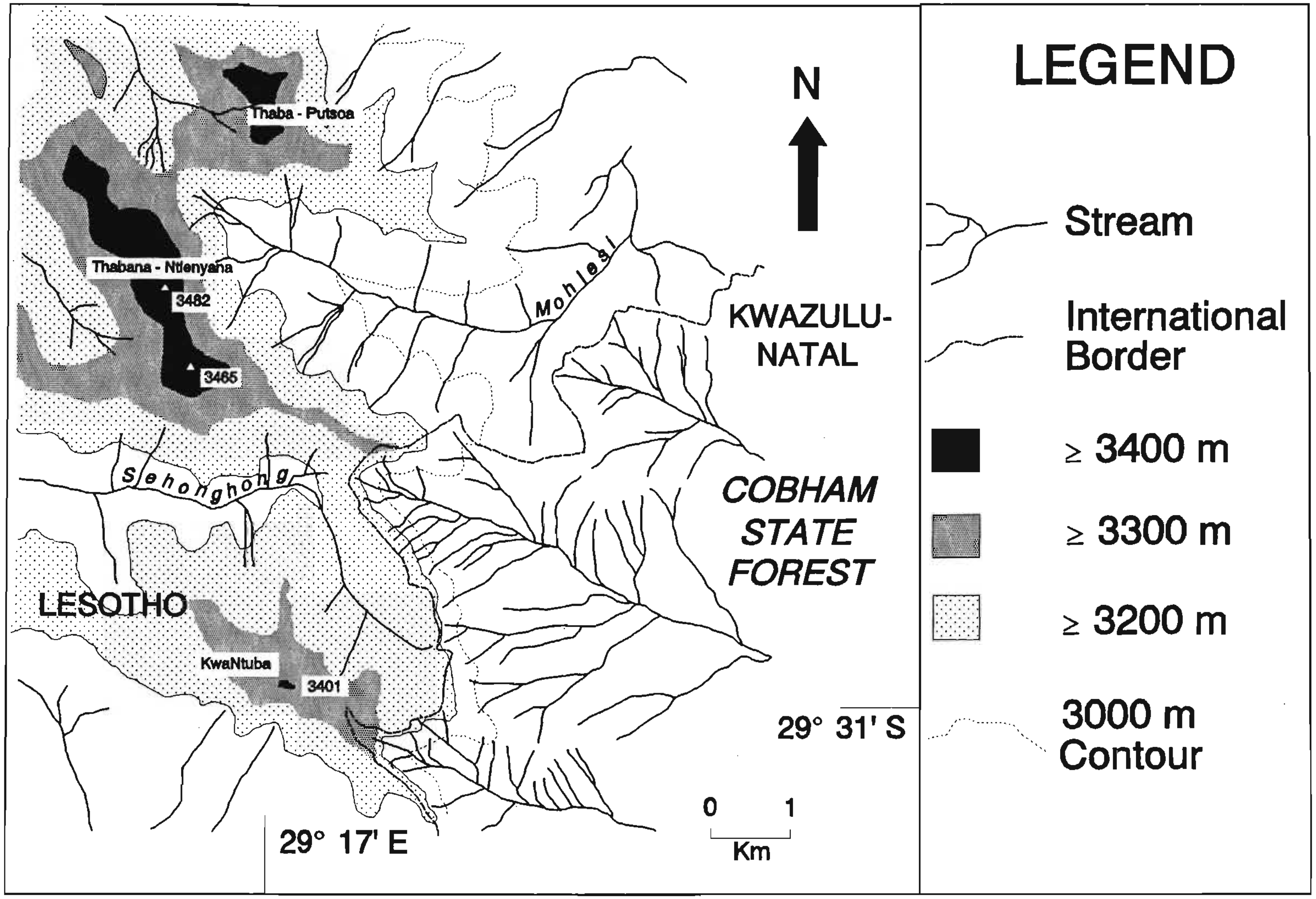


Figure 2.3 The KwaNtuba - Mohlesi Valley study region.

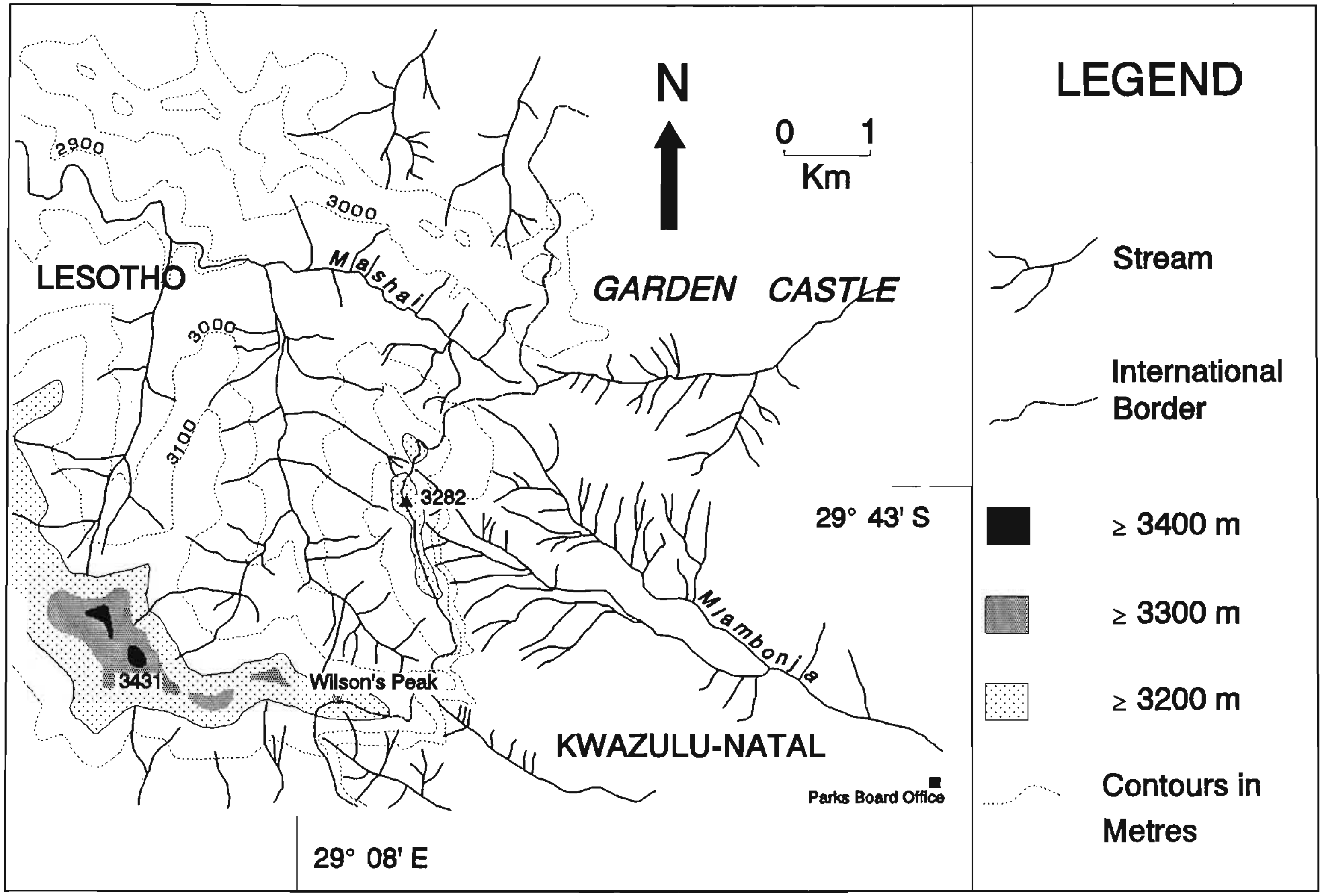


Figure 2.4 The Wilson's Peak - Mashai Valley study region.

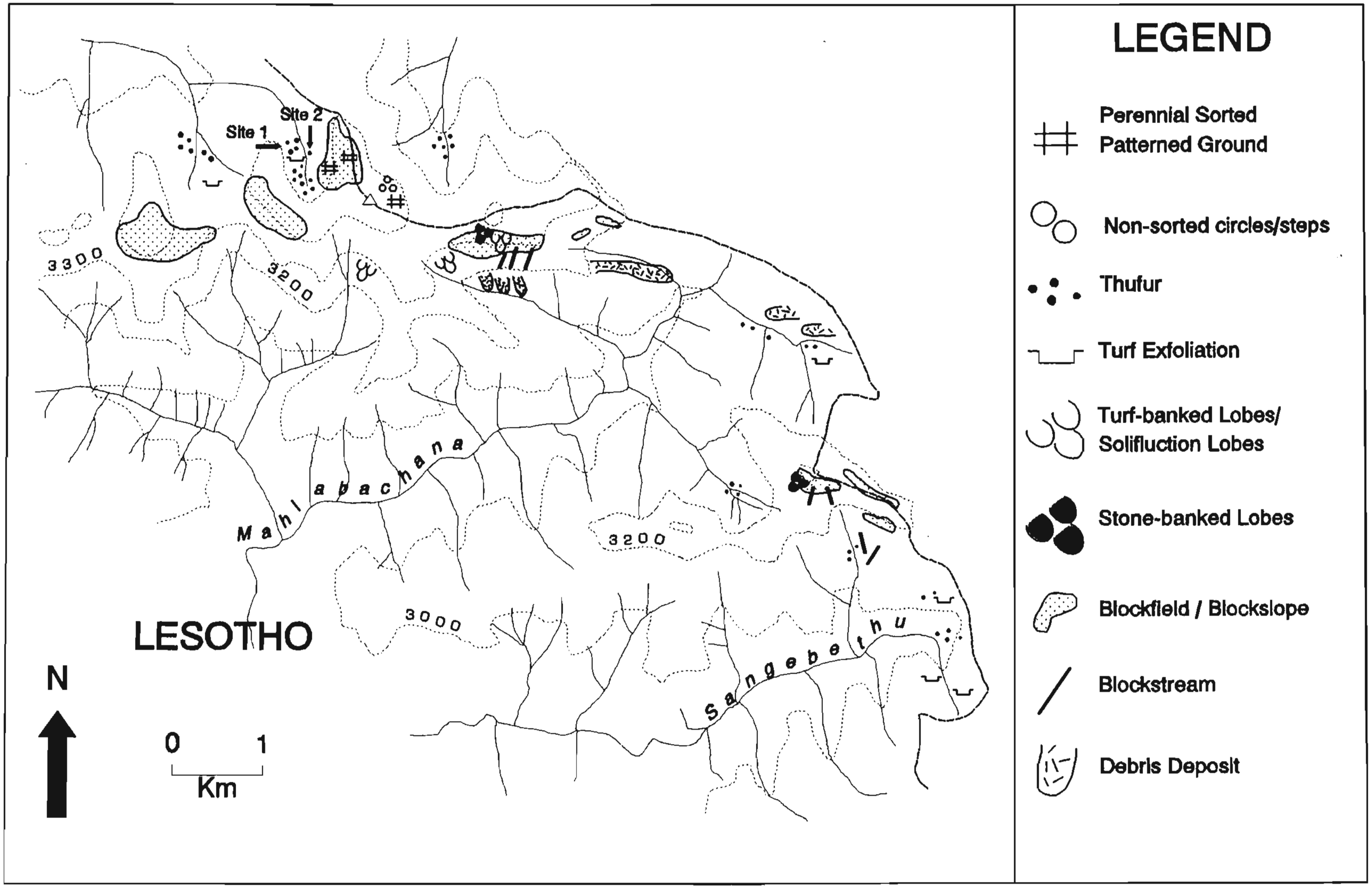


Figure 2.5 The distribution of selected geomorphic landforms in the Sanqebethu Valley - Mafadi Summit study region.

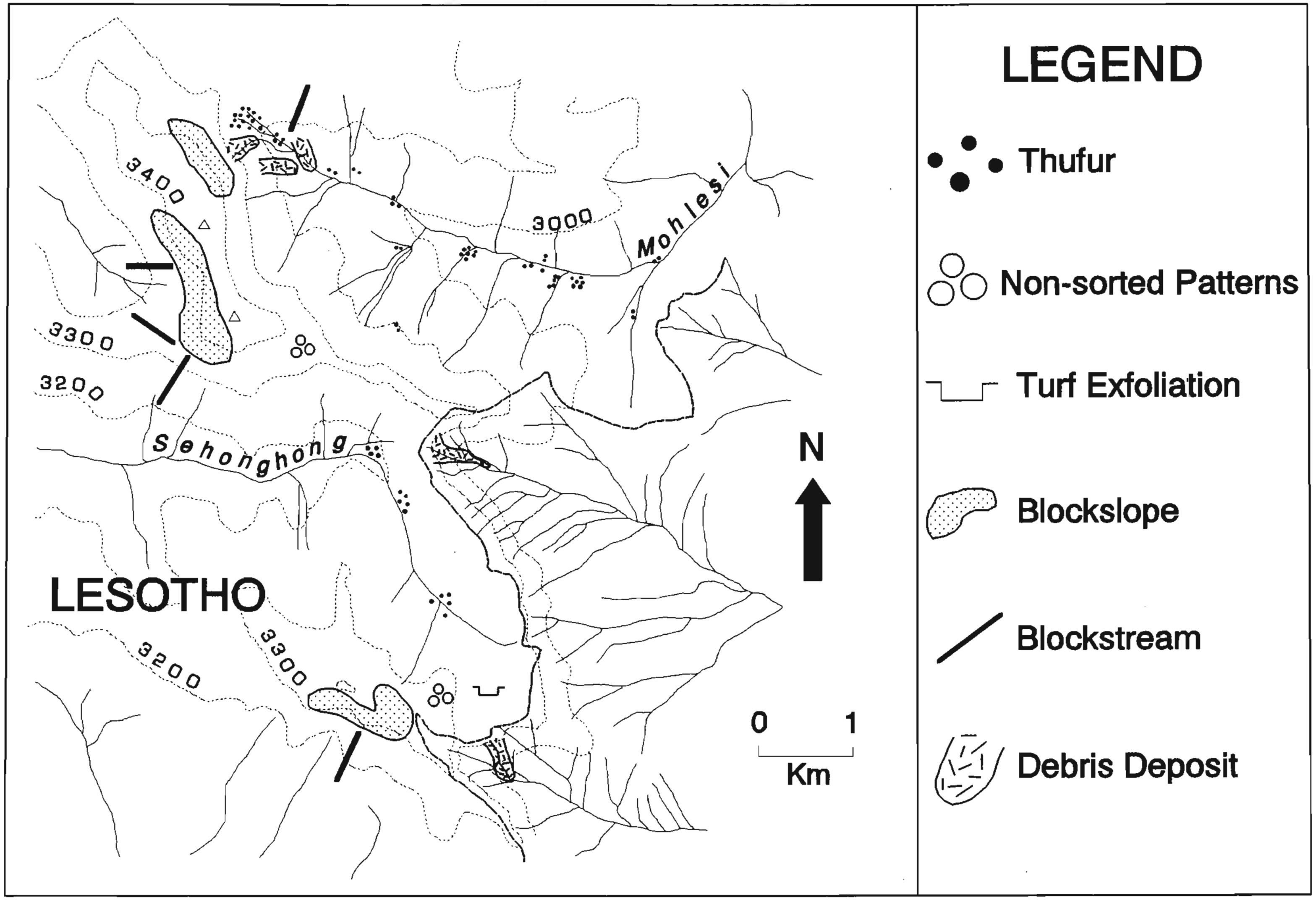


Figure 2.6 The distribution of selected geomorphic landforms in the KwaNtuba - Mohlesi Valley study region.

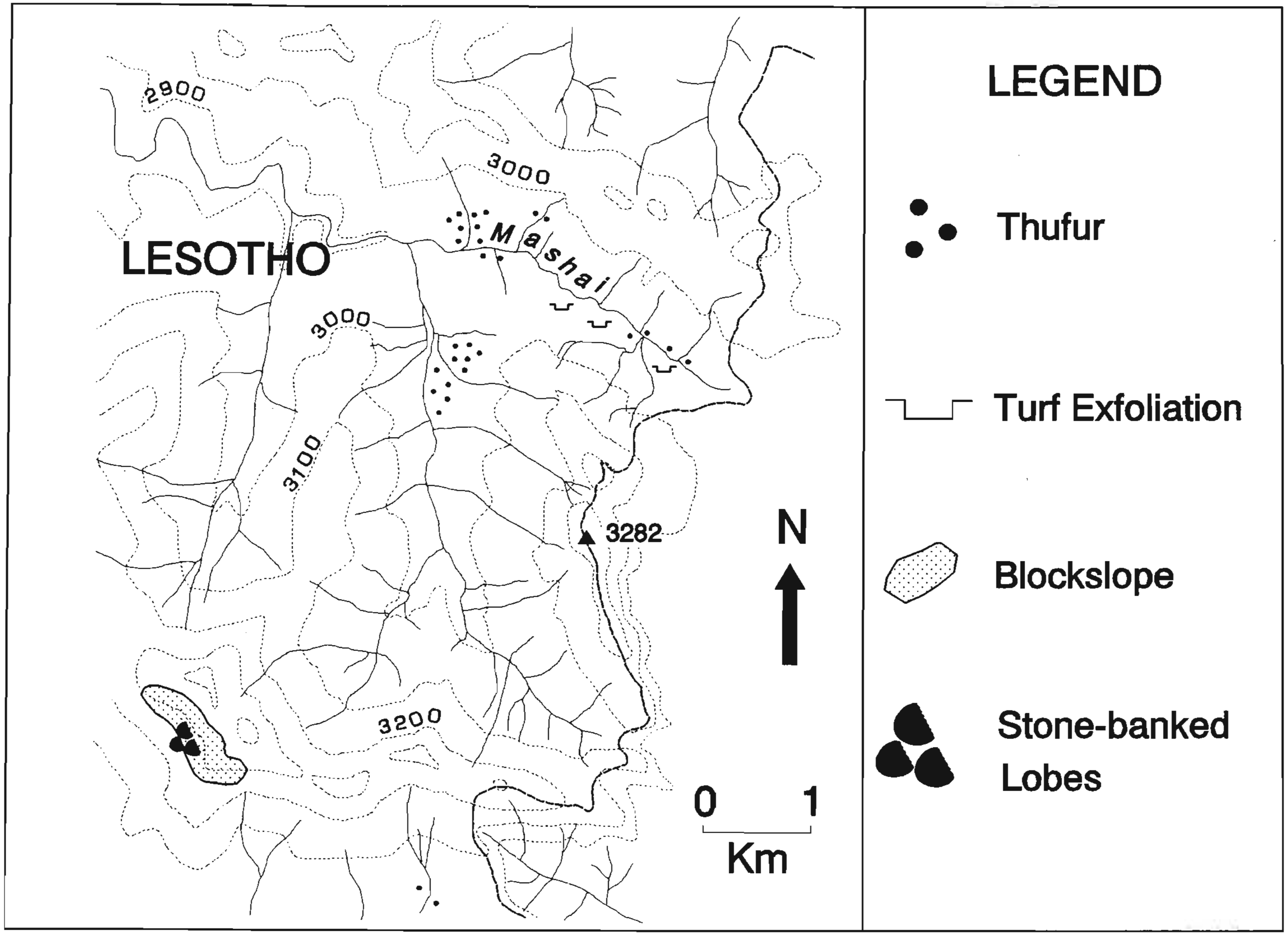


Figure 2.7 The distribution of selected geomorphic landforms in the Wilson's Peak - Mashai Valley study region.

2.2 GEOLOGY

Strata of Triassic to early Jurassic age are represented in two separate geochemical provinces in southern Africa (Cox *et al.*, 1967), of which the largest and youngest, known as the Karoo Basin, is centred on and around Lesotho (Dingle *et al.*, 1983). The Dwyka, Ecca, Beaufort, Stormberg and Drakensberg Groups form the lithostratigraphic sequence of the Karoo Supergroup (Table 2.1, Figure 2.8). The lower elevations of the Drakensberg (the "Little Berg") fall within the Stormberg Group while the high Drakensberg and Lesotho mountains are represented by the Drakensberg Group volcanic series. The Drakensberg plateau, now an erosional remnant, is estimated at 140 000 km^2 in extent (Tankard *et al.*, 1982), and still attains a thickness of 1350 m in the Mont-Aux-Sources region (King, 1982). The Stormberg Group comprises three sedimentary formations : the Molteno, Elliot and Clarens (Figure 2.9). The Clarens Formation adjoins lavas and various pyroclastics of the overlying Drakensberg Group (Dingle *et al.*, 1983) (Figure 2.9). The current study areas are concentrated primarily within the Drakensberg Group basalts. However, a number of subsidiary investigations have also been conducted in the Clarens Formation sandstones.

2.2.1 Drakensberg Group

The Drakensberg volcanic group, which forms the highest parts of the country, is the product of a phase of discrete volcanic centres which ejected material, followed by a phase of basaltic lava extrusions associated with dyke intrusions (Dingle *et al.*, 1983). The first basalts were formed during Upper Triassic times when shield volcanoes, diatremes and subsidence calderas formed the Moshesh's Ford Formation (Dingle *et al.*, 1983; Van Rooy and Van Schalkwyk, 1993). A radiometric date of about 187 million years has been given for the earliest Drakensberg/Lesotho basalts (Fitch and Miller, 1971). Botha and Theron (1967) have suggested that some of these early eruptions took place while sedimentation of the Elliot Formation was still in progress. Very mobile lavas later welled up from fissures fed by dykes, cutting through the Stormberg sediments and Moshesh's Ford Formation to produce basalts of the Kraai River and Lesotho Formations (Dingle *et al.*, 1983; Van Rooy and Van Schalkwyk, 1993). The Kraai River Formation, which typically displays columnar jointing and

	Group	Formation & subgroups	Lithostratigraphy	Vertebrate zonation		System
		Western area: Karoo Basin & Lesotho				
	Drakensberg Group	Lesotho Fm.	main flood basalts: thick tholeiitic flows & occasional interbedded pyroclastics			Jurassic
		Kraai River Fm.	fine grained thoheiitic basalts & pillow lavas			
		Moshesh's Ford Fm.	lavas, pyroclastics vent fill & interbedded sediments			
Karoo Super-group		Clarens Fm.	yellow & white massive & cross bedded feldspathic sandstones and minor grey-green shales & silts	no formal zonation	occasional light-limbed dinosaurs	to
	Stormberg Group	Elliot Fm.	greyish red & purple mudstones with minor yellow & grey sandstone		crocodiles & occ. light-limbed dinosaurs	
					light-limbed Anchisaurid dinosaurs	
					heavy- limbed Melanosaurid dinosaurs	Triassic
		Molteno Fm.	glittering sandstones, grits & conglomerates with grey & black shale, mudstones & coals		no tetrapods	
	Beaufort Group					

Table 2.1 Summary of the lithostratigraphic and biostratigraphic subdivision of the Jurassic and Triassic strata of the main Karoo outcrop (after Dingle *et al.*, 1983, p12 [*sic*]).

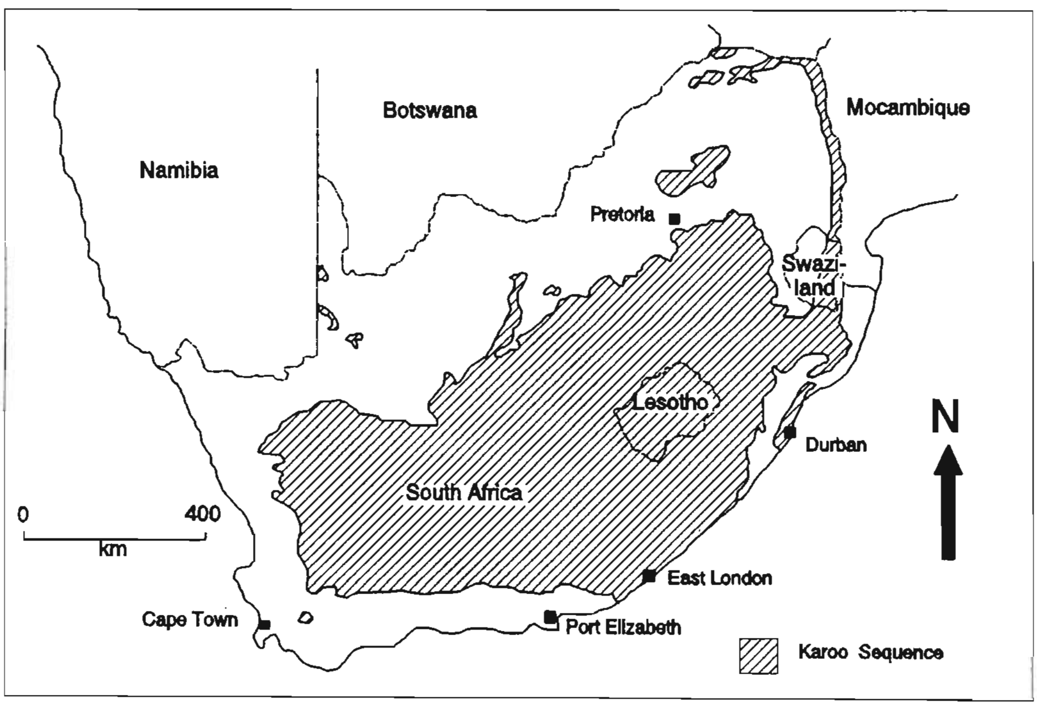


Figure 2.8 Distribution of the Karoo Sequence in South Africa, Lesotho & Swaziland (after Kent, 1980, p560).

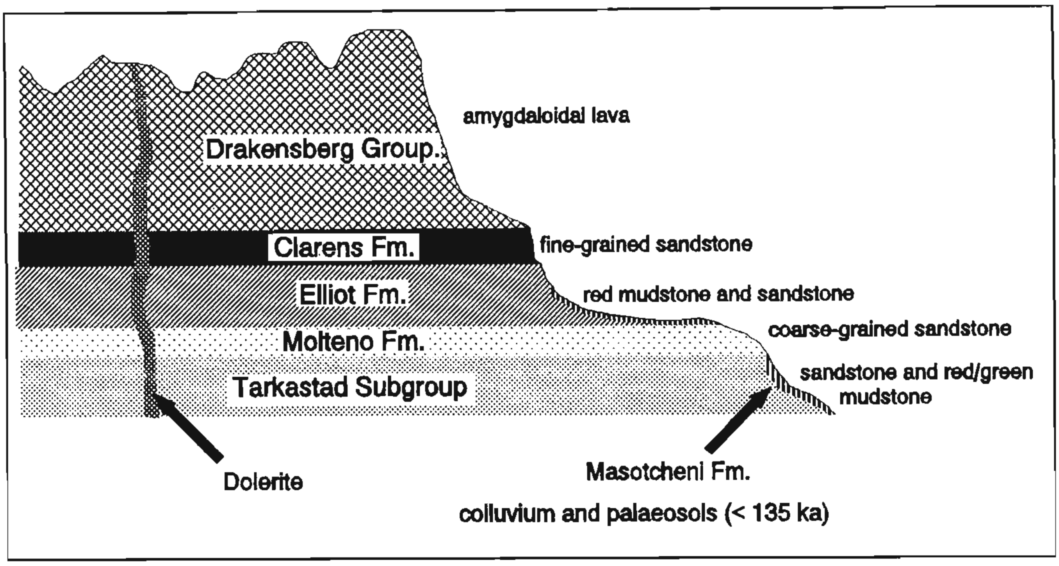


Figure 2.9 The typical topographical expression of each unit in the geological succession in the Drakensberg (after Pickles, 1985, p5).

pillow lavas, has only been recognized in the Eastern Cape regions (Dingle *et al.*, 1983), while pillow lavas have been reported from the Indedema Valley in the Cathedral Peak region (McCarthy, 1970). The basalts of the Lesotho Formation mark the final and main stage which consisted of thoheiitic flows with pipe amygdales forming at their bases (Lock *et al.*, 1974). The pipe amygdales were produced by the movement of gas bubbles through the viscous, cooling material (Haughton, 1969). At some places the Clarens sandstones merge with the lower lava beds. It has been suggested that this may be due to continued deposition and reworking of aeolian sands during the volcanic interruptions (Schmitz and Rooyani, 1987). Mineral composition of the Drakensberg basalts includes labradorite, augite and smaller quantities of magnetite and apatite (Visser, 1989). The texture of the basalt has been described as varying from compact to vesicular and amygdaloidal (Klug *et al.*, 1989).

2.2.2 Clarens Formation (Stormberg Group)

The sandstones of the Clarens Formation have often been referred to as "cave sandstone", owing to the cave-like rock overhangs found here. Many of these rock overhangs contain San rock art (commonly known as "Bushman rock art"), which is particularly conspicuous in the central (Giant's Castle Game Reserve) and southern (Bushman's Nek region) KwaZulu-Natal Drakensberg. The Clarens Formation forms a narrow outcrop around Lesotho and has been studied in detail along the KwaZulu-Natal Drakensberg by Eriksson (1979a, 1979b, 1983). Recently, Meiklejohn (1995) has examined rock weathering processes and San rock art deterioration within the Clarens Formation.

It is thought that the Clarens Formation was probably first laid down by an alluvial fan and subsequently by aeolian dune depositional processes (Eriksson, 1983). Haughton (1969) suggests that the lenses of shales within the rocks are indicative of limited amounts of water having been available during the depositional sequence. Fossil features identified within these sandstones include polygonal sand-filled mudstone cracks, planolites, burrows, fossil wood fragments and dinosaur footprints (Erikson,1983).

Although the thickness of the Clarens Formation can vary between 100 and 150 m (Eriksson, 1983), commonly, it is relatively constant throughout most of the KwaZulu-Natal Drakensberg. Boelhouwers (1988b) reports that the Clarens Formation occurs at an altitude of between 1720 and 1880 m in the Giant's Castle Game Reserve. However, according to Du Toit (1954), the Clarens Formation reaches a thickness of up to 300 m in the northern parts of Lesotho and adjacent parts of the Orange Free State.

Although the Clarens Formation is represented by the finest textured sandstones within the Stormberg stratigraphic sequence, a variety of sedimentary rock types ranging from mudstones to coarse conglomeratic sandstones have been identified within this formation (Eriksson, 1983). Eriksson (1983) found that 87% of the Clarens Formation comprises very fine sandstones and silts, 11% fine to medium-grained sandstones and 2% medium to coarse conglomeratic sandstones. The Clarens Formation is typified by quartz-rich grains (Eriksson,

1983) and interstitial material includes tourmaline, garnet, zircon and white mica (Du Toit, 1954). The colour of these sandstones is mainly yellowish or white (Dingle *et al.*, 1983) due to the chemical weathering of the feldspar (Eriksson, 1983).

Although sills and dykes of fine-grained dolerite have intruded various outcrops of the Karoo Supergroup, these become increasingly scarce when moving up the stratigraphic sequence. According to Boelhouwers (1988b), the dykes have widths of between 3 and 6 m, and lengths varying from tens of metres to several kilometres.

2.3 GEOMORPHIC EVOLUTION

Typical topographical features of Lesotho and the KwaZulu-Natal interior are the relatively level areas of land ("erosion surfaces") separating the steeper slopes. These so called "erosion surfaces" have frequently been related to erosion cycles (e.g. Dixey, 1938; King, 1944; Stockley, 1947). Although both Suess (1904) and Penck (1908) considered the high Drakensberg to have an erosional origin, these workers did not consider tectonic activity to be part of the evolution of such features. King (1944), on the other hand, considered the erosion surfaces at different elevations as of the same age due to tectonic uplift. This view was abandoned in 1959, when the main Drakensberg scarp was envisaged as a divide between the early Tertiary planation and the Gondwana/post-Gondwana landsurfaces (King and King, 1959). King and King (1959) believed that the regularity of erosion surfaces suggests a cyclic process within the Quaternary; this included uplift of land and extensive warping accompanied by strong seaward tilting, thereby raising the early Tertiary landscape and initiating the incision of river courses. King and King (1959) further considered the Lesotho highland summits to be of Jurassic age, a product of the post-Gondwana cycle rather than the African cycle. Major upwarping in the late Miocene is believed to have contributed to considerable gully development and gorge incision along the Great Escarpment (King and King 1959; King 1974). Some of the headwaters to the west of the escarpment, such as those near Sani Pass, have broad gentle valleys, which were considered to be the second erosion surface or "Kretacic cycle" (King 1972, 1974, 1976). The lowlands of Lesotho were ascribed to the "African" or late Cretaceous and early Tertiary surface (King, 1974).

A more recent viewpoint presented by Partridge and Maud (1987) argues against the preserved evidence of an erosion surface dating to before the breakup of Gondwanaland. Rather, Partridge and Maud (1987) believe that some 300 m may have been removed from the original surface and that the benches are of structural, rather than cyclic origin.

2.4 SOILS

Few studies have been undertaken to examine the alpine soils (Hughes, J. 1993, pers. comm.), possibly because soil surveys have focussed most attention on the agriculturally important lower areas (Fitzpatrick, 1978). Because the weathered material at higher altitudes appears not to be subjected to extensive transportation, mixing and redeposition, the soil properties can to some extent be related to parent material. Chemical weathering processes may, however, contribute to some mineralogical changes during the breakdown process. Although a detailed analysis of soil properties has never been undertaken above 3000 m a.s.l, Schmitz and Rooyani (1987) use climatic parameters in classifying these soils and believe that they have a "udic" moisture regime and "cryic" temperature regime.

The shallow soils of the high Drakensberg slopes are mostly residual and colluvial in origin. Small amounts of alluvial material have contributed to the somewhat deeper soils found in the valley flats. The sloping summits and basalt benches are characterized by shallow soils, about 0.15 m in depth, and have commonly been referred to as "lithosols" or "litholic soils" (Klug *et al.*, 1989). The valley floors and heads are represented by mollisols (soil classification after Miller and Donahue, 1990), which constitute a deeper and darker soil than that found on the slopes above (Schmitz and Rooyani, 1987). Peatlands and wetlands are commonly found at the valley heads (Backéus, 1988) where the soils attain thicknesses of over 0.18 m (Klug *et al.*, 1989) and have a base saturation of over 50% (Schmitz and Rooyani, 1987; Klug, *et al.*, 1989). The highly permeable and relatively stable histosols have, in the past, impeded erosion (Morris *et al.*, 1989), but the recent destruction of peatlands has resulted in organic soils being lost through oxidation (Backéus, 1988), thereby facilitating increased erosion rates. The occurrence of such humic-rich A1 horizons is also considered as an indication that current frost activity is somewhat less severe than in the past (Fitzpatrick, 1978).

Relatively high clay/silt contents of up to 68% have been measured for mollisols along valley flats (Grab, 1992a). The organic content for these soils is also high, ranging from 6 to 20% (Klug *et al.*, 1989; Grab, 1992a). 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; Grab, pers. obs.). According to Fitzpatrick (1978), frost processes play an important role in the formation of these soils. Most of the miniature cryogenic features such as sorted and non-sorted patterned ground have developed in mollisols because of the abundance of fine material and soil moisture.

There is some discrepancy in the literature pertaining to the comparative depth of soils on north- and south-facing slopes. Although Klug *et al.* (1989) point out that mineralogically and chemically, the soils on north- and south-facing slopes do not differ significantly, the soils on north-facing slopes are somewhat shallower than those on south-facing slopes. Contrary to this, Boelhouwers (1988a) has suggested that higher denudational rates on the south-facing slopes have resulted in shallower soils, whereas the drier, warmer north-facing slopes are represented by relatively deep and gradual weathering profiles.

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). Fitzpatrick (1978) has examined some of these soils and identifies properties that he thinks may indicate former permafrost, but other features, such as thermokarst or ice-wedge casts, indicative of the former presence of permafrost, have yet to be found. The present author believes that although the evidence presented by Fitzpatrick may indicate frost action, this evidence is inconclusive to suggest the former occurrence of permafrost, and, therefore, should be examined with caution.

2.5 VEGETATION

Vegetation characteristics of the high Drakensberg have been documented by several authors (e.g. Killick, 1963, 1978; Van Zinderen Bakker and Werger, 1974; Van Zinderen Bakker, 1981; Backéus, 1988; Schwabe *et al.*, 1989). The terms montane, subalpine and alpine have been used to describe the vegetation belts of the Drakensberg and Maluti mountains (Killick, 1963; Jacot-Guillarmod, 1971). Killick (1963) has characterized these belts according to three climax communities; *Podocarpus Latifolius* forest, between 1280 and 1829 m a.s.l. (montane), *Passerina-Philippia-Widringtonia* fynbos, between 1829 and 2865 m a.s.l. (subalpine), and *Erica-Helichrysum* heath, between 2865 and 3353 m a.s.l. (alpine) (Figure 2.10). The montane and subalpine belts are characterized by tussock grassland, scattered protea and woody communities (Killick, 1963) (Figure 2.10). Although the subalpine belt of the KwaZulu-Natal Drakensberg contains several fynbos species, these are rare or absent in the Lesotho alpine belt (Jacot-Guillarmod, 1971; Killick, 1978). The vegetation above 3000 m a.s.l. has been described by Van Zinderen Bakker (1981) as consisting predominantly of "miniature" plants and "cushion" plants. This alpine vegetation belt is characterized by climax heath communities, in particular woody species of *Erica* and *Helichrysum*, varying between 152 and 609 mm in height (Killick, 1963). The woody species are interspersed with alpine grassland dominated by species of *Festuca, Danthonia* and *Pentaschistis* (Killick, 1963). These plants of the alpine belt have a brief growing season (Morris *et al.*, 1993), possibly exercising control on plant size.

It appears that topographic position exercises considerable control on plant distribution and colonization (Granger and Schulze, 1977; Morris, *et al.*, 1993). Granger and Schulze (1977) conclude that topographically-induced variations in radiant energy will lead to corresponding variations in soil moisture status and thereby cause alterations in the plant environment. Similarly, Morris *et al.* (1993) examined the relationship between topographic positions and plant communities and found that topography, because it modifies solar radiation patterns, influences plant growth and may explain the distribution of vegetation communities. Klug *et al.* (1989) have noted that the drier inter-bench areas contain shrubs whereas bench areas support almost exclusively grass vegetation. Killick (1963, 1978) has found that heath

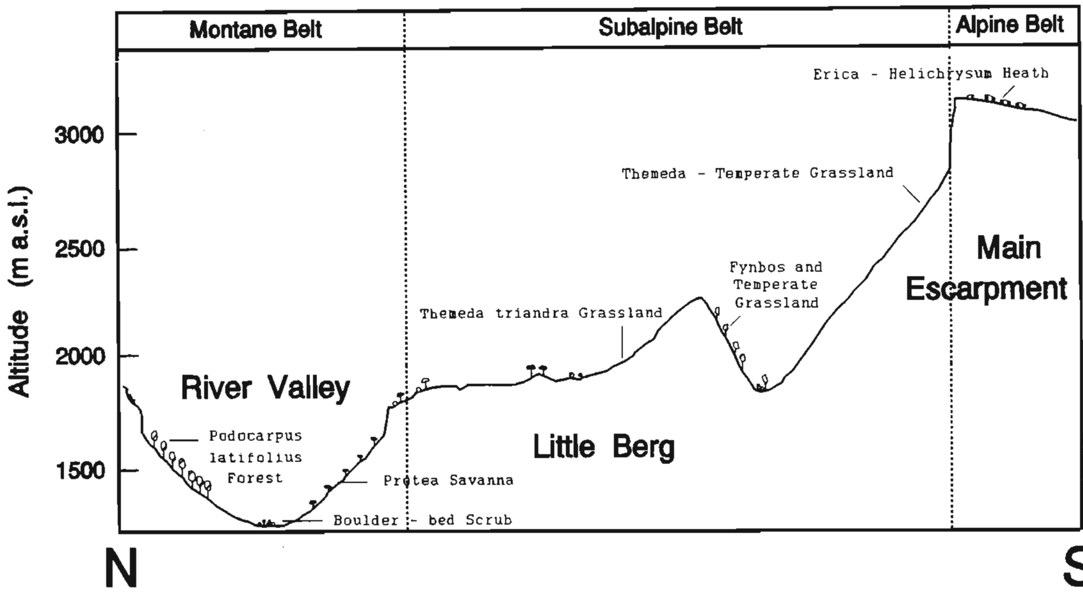


Figure 2.10 Profile through the Drakensberg region showing the vegetation belts with their primary plant communities (after Killick, 1963, p25-26).

vegetation frequently colonizes rocky or bouldery areas and grows generally taller than in the surrounding areas due to the boulders offering protection against fire and wind. Killick (1978) has, nevertheless, suggested that there are interrelationships between plant communities from different topographic positions (Figure 2.11).

Vegetation exerts considerable control on geomorphic processes and is, in return, affected by these processes. It is therefore important to recognise the role of vegetation when undertaking geomorphic studies. For example, Pérez (1987a) has shown that the vegetative growth of *E. semiglobulata* may be an adaptive response to needle ice induced disturbance, and further, it is believed that the destruction of vegetation cover in certain cold regions may cause a depression of the periglacial altitudinal belt (Hagedorn, 1980).

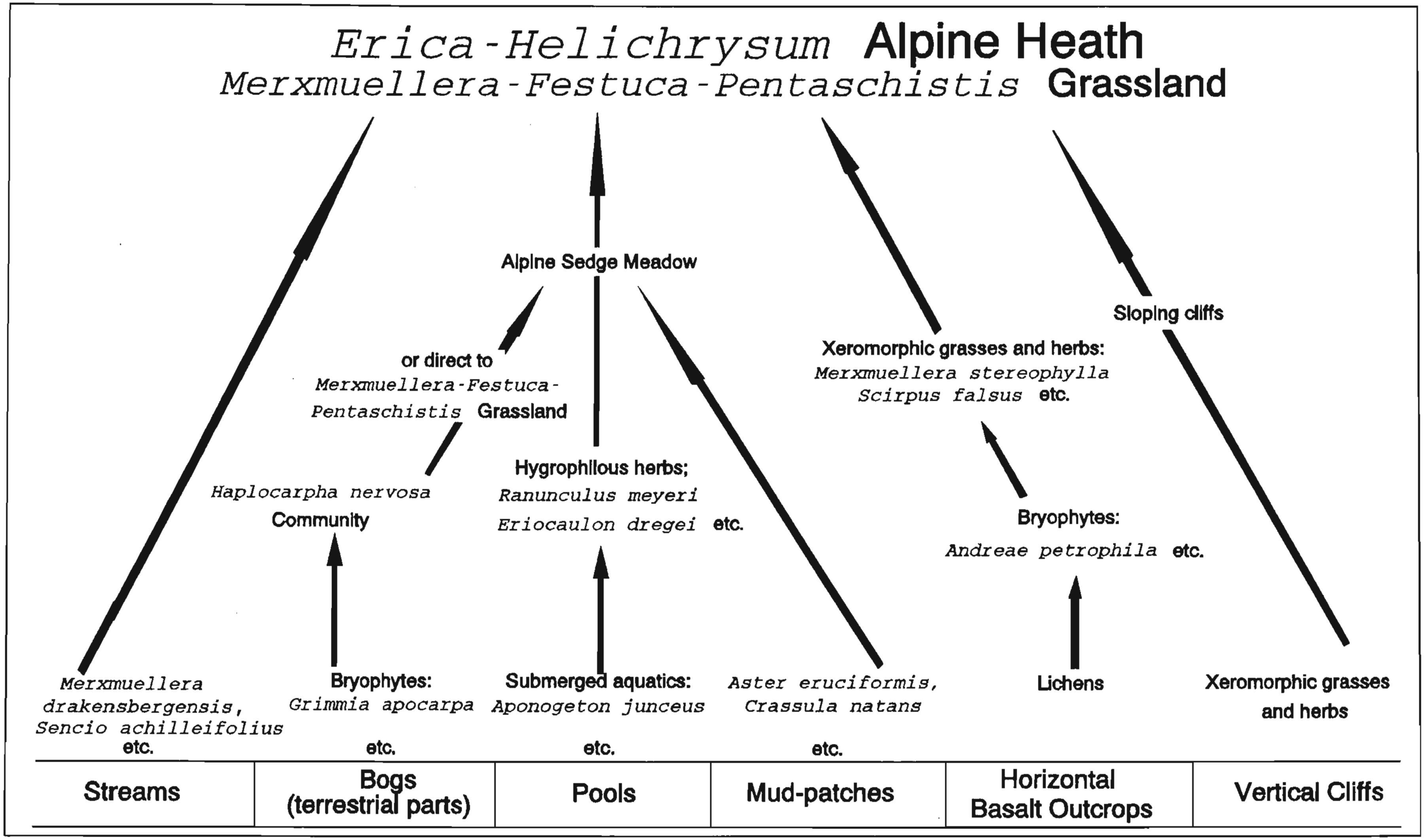


Figure 2.11 Suggested interrelationships of the plant communities in the alpine belt of the Drakensberg and Lesotho (after Killick, 1978, p549).

2.6 CLIMATE

Although a limited number of weather stations along the "Little Berg" have been operative for several decades, few climatic records exist for the high Drakensberg and Lesotho plateau. Further, data sets for the high altitude stations which have recorded some climatic data show frequent interruptions and/or are unreliable in their precision.

2.6.1 Precipitation

The climate over the interior of southern Africa is dominated by a high pressure cell during the winter months and low pressure cell during the summer months. Consequently, precipitation is distinctly seasonal; 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 and Maluti mountains are orographic thunderstorms. Convective instability is induced by slope convection along the escarpment, where cloud build-up may take place from late morning (Tyson *et al.*, 1976). Surface and upper air troughs (squall-line thunderstorms) moving in an easterly direction over the sub-continent are also an important source of precipitation during the summer months. Occasionally, anti-cyclones (cold fronts) may bring some precipitation to the Drakensberg and Maluti mountains, particularly during the autumn, winter and early spring months. The anti-cyclones are associated with the northward displacement of pressure belts during the autumn and early winter months.

The zone of maximum precipitation 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 west where a marked decrease in precipitation is encountered (Schulze 1979) (Figure 2.12). Killick (1963) illustrates this by comparing the mean annual precipitation of the Organ Pipes Pass summit area, receiving about 1609 mm per annum, with Mokhotlong (40 km to the west), which only receives about 562 mm per annum.

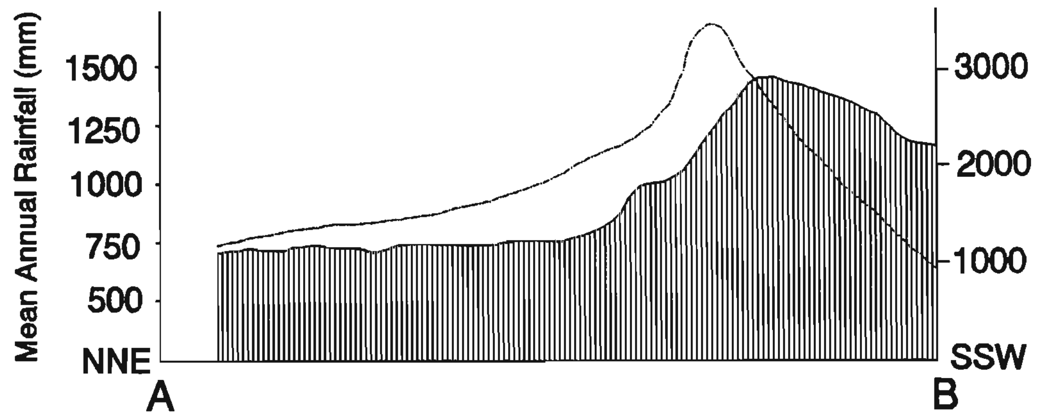


Figure 2.12 Rainfall : altitude relationship along **A** (Bergville) to **B** (Mothelsessane) (after Schulze, 1979, p48).

No detailed study has yet been undertaken to assess the importance and frequency of hail in the Drakensberg. The expected frequency of hail has consequently been contradictory and probably underestimated. Nänni (1956) suggests that hail occurs once every two years while Tyson *et al.* (1976) estimate that it may fall eight times per annum. From personal observations, it would appear that hail may occur far more frequently than has been suggested by Tyson *et al.* (1976). The present author has also observed that hail may lie for up to 24 hours on the higher slopes that are protected from insolation.

Light snowfalls may occur about eight times per annum (Harper, 1969; Tyson *et al.*, 1976) and lie for several days on south-facing slopes. Less frequent, heavier snowfalls do occur and may sometimes leave high plateau areas snow-bound for several weeks during the winter months. Snow may fall during any month of the year but is most frequent during autumn (April and May), winter (June to August) and spring (September and October).

2.6.2 Winds

East and south-east winds prevail between November and February, bringing moist air from the Indian Ocean and consequently increasing the likelihood for precipitation. The frequency of easterly winds decreases from March to May, when westerly winds gradually become more prevalent. During the winter months (June to August), westerly winds predominate, which are sometimes referred to as "bergwinds". The westerly winds become particularly frequent towards the end of winter and early spring (August and September), when they may blow with exceptionally great velocity and cause high levels of evaporation (Nänni, 1956).

2.6.3 Radiation

The valleys of eastern Lesotho are characterized by strong asymmetry, with steep south-facing and somewhat shallower north-facing slopes (Meiklejohn, 1992, 1994). This has, to a large extent, controlled the amount of radiation received on these slopes, with the north-facing slopes receiving as much as 470% more than south-facing slopes in mid-winter (Granger and Schulze, 1977). Although the most accentuated variation in insolation, induced by slope and aspect, are evident in July (Granger and Schulze, 1977), seasonal variations of noon time radiation levels in mid-summer may be significantly higher (± 77%) than in mid-winter (Tyson *et al.*, 1976).

2.6.4 Temperature

2.6.4.1 Introduction

Although temperature is considered the single most important component of mountain climates (Barry, 1981), air temperature data from higher altitudes in Lesotho are scarce and have consequently received little attention (e.g. Killick, 1978; Backéus 1988). Tyson *et al.* (1976) estimate that the absolute maximum and absolute minimum temperature for the "Little Berg" is +35.0°C and -12.5°C respectively. Air temperatures for the higher KwaZulu-Natal summits and the Lesotho plateau are considerably lower than those for the "Little Berg".

Schulze (1979) estimates that the higher peaks have a mean annual air temperature of about 7°C. According to Tyson *et al.* (1976), the first frosts occur in middle to late April and the last frosts in mid-October. Climatic variables such as cold air drainage frequently induce an irregular distribution of frost, which is believed to occur on up to 180 days per year in parts of Lesotho (Tyson *et al.*, 1976).

A further analysis of high altitude air temperature characteristics, including such as freezing indices and the number of frost days and freeze-thaw days per year, is of potential value when assessing the periglacial environment. The present author has therefore examined and analyzed screen air temperatures within three altitudinal zones of northern Lesotho (see Section **2.6.4.2**). It is hoped that this study may also make a small contribution to broadening the understanding of the high Drakensberg climate.

2.6.4.2 Study area and methodology

The investigated area is about 70 x 60 km and situated in northern Lesotho, within which four weather stations with long-term air temperature records were available (Figure 2.13). The three altitudinal zones discussed are those identified by Killick (1963) (Figure 2.10), who categorized them according to vegetation assemblages. For the purpose of this discussion, the subalpine zone has been altitudinally subdivided into "lower" (1829 to 2174 m a.s.l.), "mid" (2175 to 2519 m a.s.l.) and "upper" (2520 to 2865 m a.s.l.) zones. The temperature recording stations are Butha Buthe (1768 m a.s.l., montane zone), Mokhotlong (2375 m a.s.l., sub-alpine zone), Ox-Bow (2591 m a.s.l., sub-alpine zone) and Letseng-la-Draai (3050 m a.s.l., alpine zone) (Figure 2.13).

Several decades of reliable daily screen air temperatures for the stations were analysed and graphs produced to show mean statistical values (Figure 2.14). Terrestrial lapse rates based on long-term mean annual and mean monthly air temperatures have been calculated for the various altitudinal zones. Lapse rates calculated and compared include those of mean annual air temperature (M.A.A.T.), mean warm month temperature (M.W.M.T.) and mean cold month temperature (M.C.M.T.).

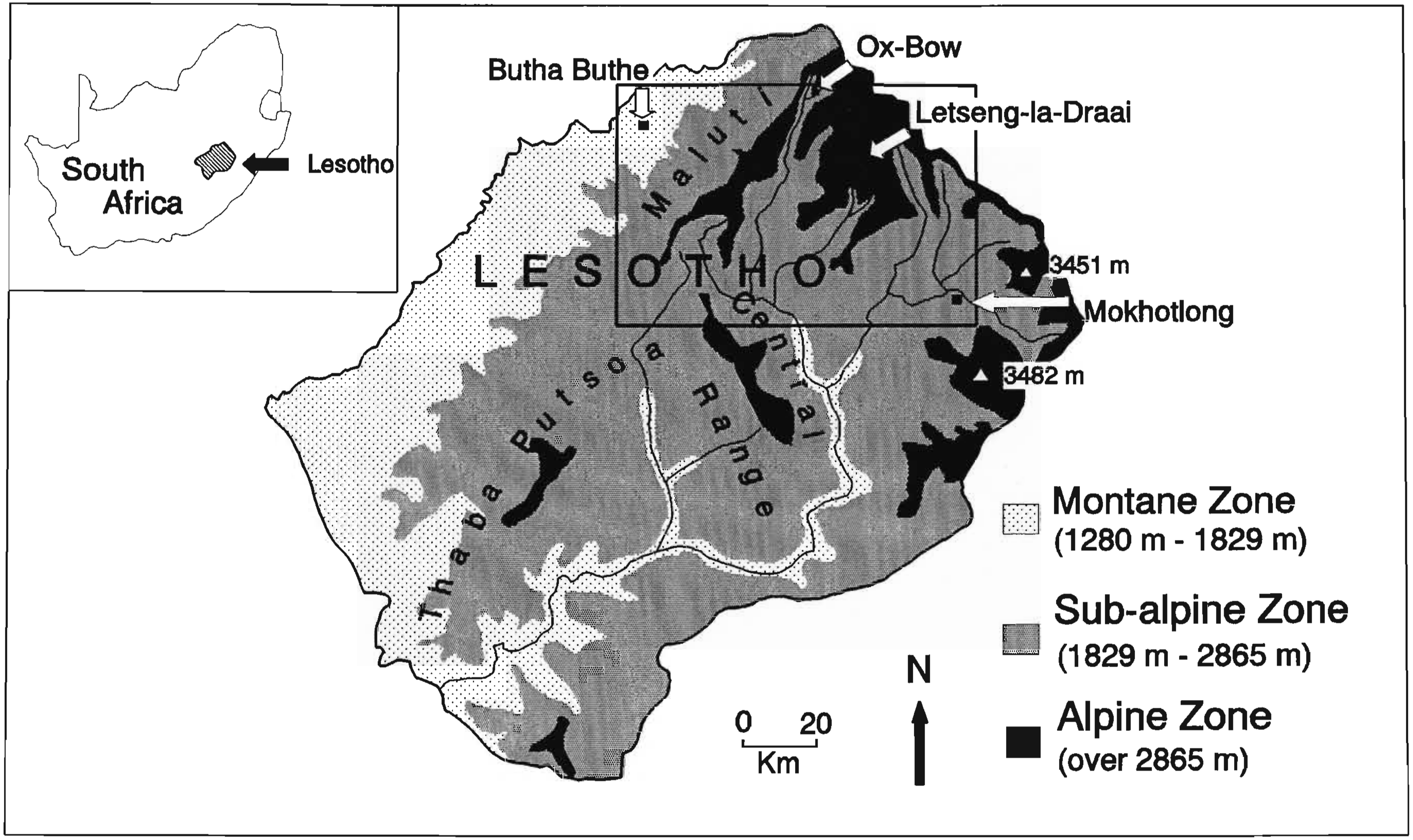


Figure 2.13 The location of primary weather stations in northern Lesotho.

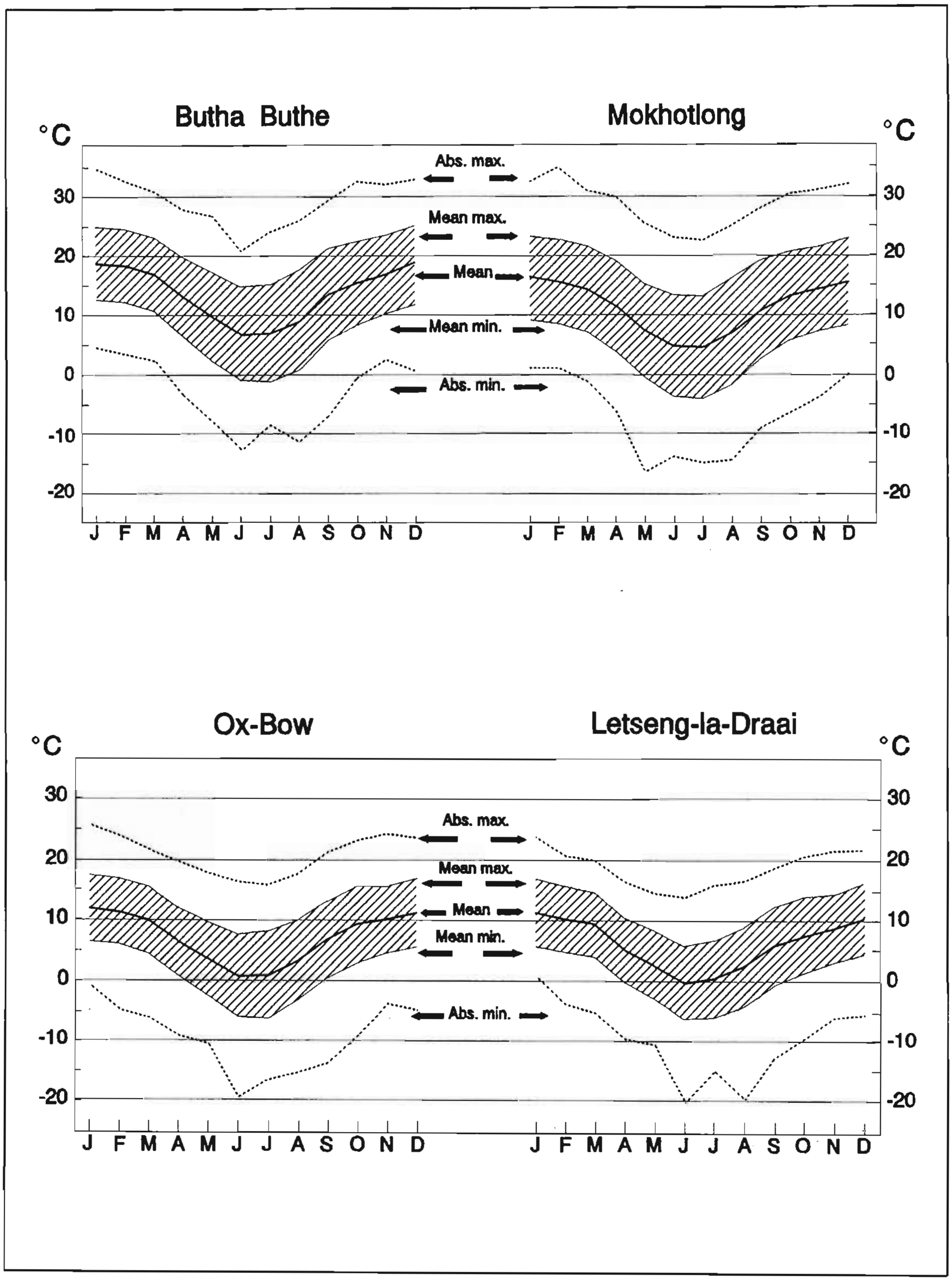


Figure 2.14 Monthly temperature graphs for Butha Buthe, Mokhotlong, Ox-Bow and Letseng-la-Draai. (after Grab, 1997).

The four locations for which air temperature data have been analysed share a number of common climatic controls, namely continentality, latitude, leeward position from the Great Escarpment and location near valley bottoms. It has been argued, however, that correlation of terrestrial lapse rates with climatic regimes is complicated in areas of high relief because considerable climatic differences may occur within short distances, owing to microclimatic variables (Hastenrath, 1968; Meyer, 1992). Although the data from Lesotho reflect local conditions, and may therefore not be genetically significant in a detailed analysis, they do, nevertheless, give a broader understanding of altitudinal air temperature variations in the region.

2.6.4.3 Lapse Rates

Ambrose (1976) has suggested a terrestrial lapse rate of 8°C/km for Lesotho but the present findings show a mean annual lapse rate of 6.2°C/km, which is closer to the "standard" terrestrial lapse rate of 5.5°C/km (Meyer, 1992). However, in reality, lapse rates in the various altitudinal belts of Lesotho differ considerably. The M.A.L.R. from the montane to the mid-subalpine zone is only 3.9°C/km, but reaches an exceptionally high value of 19.4°C/km in a belt between the mid- and upper-subalpine zones (Figure 2.15); the M.A.L.R. then drops to 3°C/km towards the alpine zone (Figure 2.15). Because the climate in the various zones is still inadequately understood, it is difficult to explain such large differences in the lapse rates. However, one possible explanation could be that winds become more frequent and considerably stronger towards the upper-subalpine zone, thereby enhancing the vertical fluxes of heat and moisture, so causing high rates of evapotranspiration at the higher elevations.

2.6.4.4 Frost days

A "frost day" occurs when the minimum screen air temperature is below 0°C during a 24 hour period (McIntosh, 1972). The difference between air and ground level freezing is discussed in some detail by Schunke and Stingl (1973) who caution that temperatures at ground level may be markedly different to those at screen level. Tyson *et al.* (1976) believe that "frost"

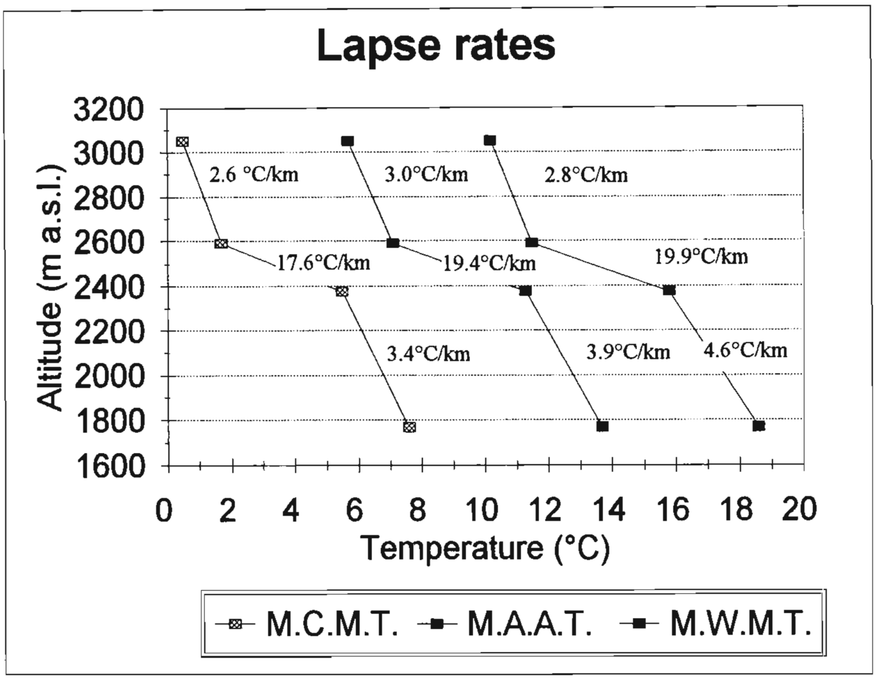


Figure 2.15 Lapse rates in northern Lesotho.

may occur on about 180 days in parts of Lesotho, but this figure probably represents ground level freezing. Although there are no quantitative recordings of ground level freezing, Van Zinderen Bakker (1981) claims to have observed such freezing every night for several weeks during mid-summer in northern Lesotho and so must be common in winter, thereby suggesting a potentially substantial number of occurrences each year.

Although less dramatic, the variable rate of additional number of frost days with altitude, follows the same trend as that for lapse rates (Figure 2.16). The greatest increase is from the mid- to upper-subalpine zone where there are an addition of 140.3 frost days/km. In the alpine zone, daily frost frequency is 85% from June to August. Because of the strongly seasonal cycle of sub-zero air temperatures, the number of additional frost days increases only marginally with altitude in the alpine zone, such that the highest summits possibly receive about 180 frost days per annum. Ground level freezing could occur on over 200 days per year at such altitudes.

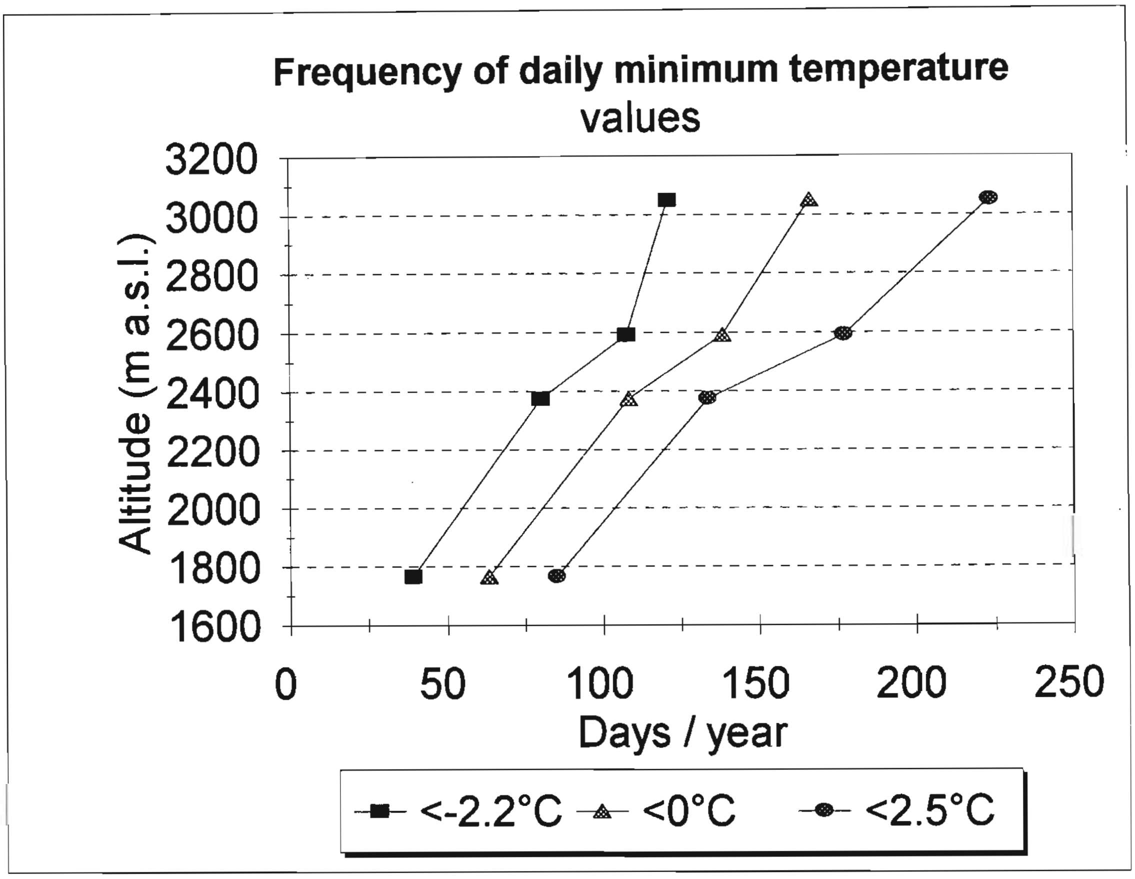


Figure 2.16 Frequency of daily minimum temperatures at various altitudes in northern Lesotho. (after Grab, 1997)

2.6.4.5 Freeze-thaw days

Schmidlin *et al.* (1987) have defined a "freeze-thaw day" as one where the maximum temperature is 0°C or above and the minimum temperature -2.2°C or below in an observation day. It becomes evident from Table 2.2 that air freeze-thaw cycles are most prolific in the upper-subalpine (mean = 107.9 per annum) and alpine (mean = 121.5 per annum) zones. The frequency of air freeze-thaw cycles rapidly decreases towards the montane zone where the mean is only 39.1 per annum.

	Butha Buthe	Mokhotlong	Ox-Bow	Letseng-la-Draai
m a.s.l.	1768	2375	2591	3050
January	-	-	-	-
February	-	-	0.1	0.2
March	-	-	0.5	0.5
April	0.1	2.1	6.4	6.5
May	2.4	8.8	17.5	20.4
June	13.9	22.9	24.1	27.5
July	15.4	25.2	25.2	28.1
August	6.4	17.8	20.0	22.7
September	0.9	3.1	10.0	10.2
October	-	0.3	3.9	4.2
November	-	-	0.2	1.0
December	-	-	0.1	0.2
Year	39.1	80.2	107.9	121.5

Table 2.2 The distribution of air freeze-thaw days at various altitudes in northern Lesotho : average values are based on ten years of data (data from Grab, 1997).

2.6.4.6 Freezing indices

Freezing indices were calculated by adding daily minimum air temperature values below 0°C within a year. This is similar to the frost index, obtained by multiplying mean air temperatures below 0°C by the time during which these temperatures occur, expressed as negative °C hours (Williams, 1961). The variable rate of increase of freezing indices with altitude, shows the same general trend as that for lapse rates and rates of additional frost days (Figure 2.17). The freezing index becomes most significant in the upper-subalpine and alpine zones where Letseng-la-Draai has a mean index value of -757 °C/year (Figure 2.17). This is primarily owing to seasonally prolonged frost action, greater frost frequency and high levels of radiative cooling, resulting in severe frost intensity at these higher altitudes (Tables 2.2, 2.3). In the alpine zone, mean minimum temperatures during June and July are less than -6°C and daily minimum temperatures may fall to below -10°C from May to September (Figure 2.14, Table 2.3). The coldest temperature yet recorded is -20.4°C at Letseng-la-Draai on 12 June 1967 (Figure 2.14, Table 2.3).

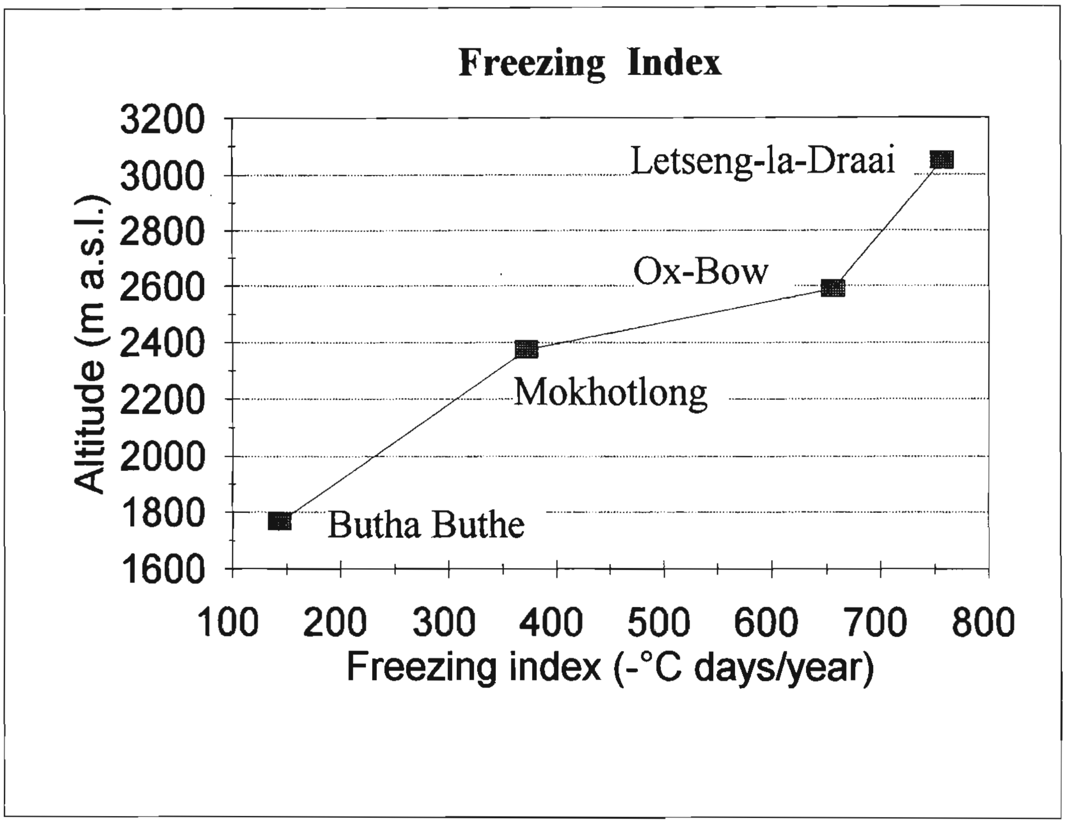


Figure 2.17 Freezing indices at various altitudes in northern Lesotho.

	J	F	M	A	M	J	J	A	S	O	N	D	Year
Mean Max.	16.6	15.3	14.3	10.1	7.9	5.4	6.3	8.3	11.8	13.5	13.9	15.6	11.6
Mean Min.	5.5	4.6	3.6	-0.3	-3.3	-6.4	-6.1	-4.3	-1.1	1.0	2.6	4.0	-0.2
Mean	11.1	10.0	9.0	4.9	2.3	-0.5	0.1	2.0	5.4	7.3	8.3	9.8	5.8
Absolute Max.	23.5	20.3	19.5	16.1	14.2	13.4	15.4	15.6	18.2	20.1	20.8	21.3	23.5
Absolute Min.	0.6	-4.0	-5.2	-9.9	-10.8	-20.4	-15.2	-19.9	-13.1	-9.9	-6.4	-5.8	-20.4
Mean Frost Days	0	0.9	1.3	14.8	28.1	29.4	30.3	28.3	17.3	10.7	3.8	1.7	166.6

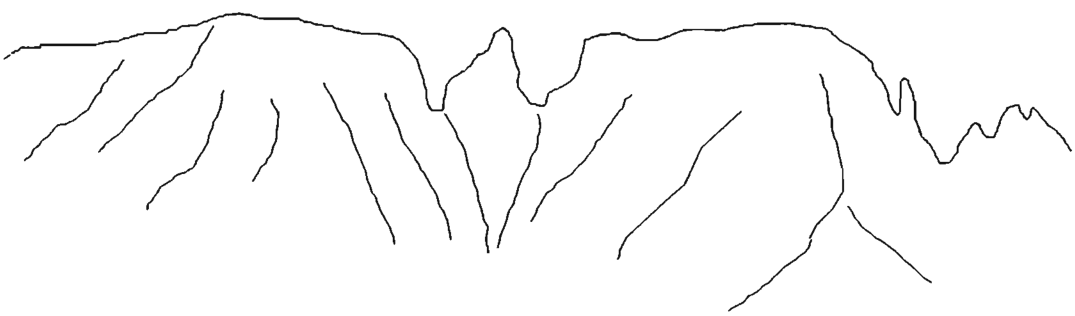
Table 2.3 Air temperature data for Letseng-la-Draai (3 050 m a.s.l.) : average values are based on ten years of data (after Grab, 1994, p114).

2.6.4.7 Summary

High altitude air temperature data from northern Lesotho show rapid intensification in cryogenic temperature characteristics from the mid- to upper-subalpine zones. Freezing cycles extend into autumn and spring months in the upper-subalpine and alpine zones while in the lower altitudinal zones, such cycles are restricted to a few weeks in winter. The lowering of air temperature with an increase in altitude will eventually result in changing geomorphic processes which may occur within particular zones. The zonal location and rate of geomorphic response depends largely on the rate of environmental change, such as changing air temperature.

CHAPTER THREE

METHODOLOGY



CHAPTER 3

METHODOLOGY

3.1 THESIS STRUCTURE

Because the aims and objectives of this study are somewhat heterogeneous, the dissertation has been structured accordingly. The data-reporting sections have been arranged in such a way that similar suites of periglacial phenomena are discussed within a chapter. Such chapters may not automatically connect with one another, as the landforms, processes and implications discussed, vary considerably. Therefore, separate discussions on the relevant issues are held in each of the data-reporting chapters.

3.2 SITE SELECTION

The main study sites, as outlined in Section **2.1**, include:

1. Sanqebethu Valley to Mafadi Summit region (central Drakensberg) (Figure 2.2).
2. KwaNtuba to Mohlesi Valley region (southern Drakensberg) (Figure 2.3).
3. Wilson's Peak to Mashai Valley region (southern Drakensberg) (Figure 2.4).

Initially, the entire latitudinal extent of the KwaZulu-Natal Drakensberg was surveyed for the selection of potential study sites. Several sites were then selected and assessed as to their logistic viability for research purposes. Potential sites had to be reasonably accessible from the KwaZulu-Natal side of the Drakensberg Escarpment as there is no logistically viable access from central and eastern Lesotho. Although the chosen sites are remote and require up to two days hiking to be reached, they have the least human disturbance and interference. For this reason (human impact), more accessible areas, such as that above Sani Pass, were not considered as cryogenic processes and landforms have been disturbed along such transport routes.

The selected sites cover a large altitudinal range and encompass the highest summits in the Drakensberg. Although the latitudinal extent covered is only 33' 18", personal observations during the initial survey indicated that the development of cryogenic landforms along the KwaZulu-Natal Drakensberg is not restricted so much by latitude, as by altitude. From this initial survey, it would appear that most, if not all, cryogenic landforms occurring in the high Drakensberg are found at one or more of the selected sites.

3.3 PHOTOGRAPHY AND MAPPING

A Minolta Dynax 8000i was used to photograph landforms, study sites and research equipment. The present author perceives photography as an important research tool for this dissertation as it supplies evidence for a variety of landforms not yet reported or photographed in the high Drakensberg. Photo scales used include :

Artline pen - length = 144 mm

Caliper rule - "closed" length = 210 mm

Camera lens cap - diameter = 55 mm

Geological hammer - length = 316 mm

Parker pen - length = 130 mm

Scale bar = 100 mm

Spade - length = 380 mm

Tape measure - diameter = 185 mm

Water bottle - length = 240 mm

A descriptive scale is used in some figure captions where an inadequate scale appears in the photograph. Some photographs taken during the winter months may appear over-exposed, owing largely to the often amber-coloured land surface, extremely high levels of radiation and low inclination of the sun. Although a Marumi C P L polar filter was occasionally used, such a filter also distorts natural colour and, therefore, was avoided where possible.

Once the landforms were identified, they were plotted onto 1:50 000 topographic maps and geomorphic field sketch maps produced. The cross-profiles and down-slope profiles obtained for larger features were digitized onto enlarged 1:50 000 base maps using the AUTOCAD program. Air photographs were used to help locate larger landforms, such as debris ridges, which were then plotted onto the base maps. Some 1:50 000 photographs were enlarged up to six times to enable the plotting of the large deposits on maps.

3.4 FIELD TECHNIQUES

3.4.1 Morphological and Sedimentological Assessment

Terrain characteristics of altitude, slope aspect and slope gradient were determined at each site and individual features measured. Equipment used included a Thommen altimeter, Brunton compass, abney level, tape measure and geological hammer. Heavier equipment (e.g. theodolite) could not be utilized as the sites are both too remote and too difficult to access with such items.

For features such as patterned ground, blockstreams, stone-banked lobes and debris deposits, the fabric, sediment granulometry and clast shape were determined. Weathering rind thicknesses were measured with calipers for clasts in sorted circles. Individual samples for granulometry were restricted to about 0.1 to 0.2 kg as these commonly had to be carried within a backpack for two or more days. A sample of between 20 and 30 clasts was measured at each site.

The following clast axis were measured:
a-axis = representing the longest axis,
b-axis = representing the intermediate orthogonal axis,
c-axis = representing the shortest orthogonal axis.

Unless otherwise indicated, the mean clast sizes are expressed as:

$$\frac{a+b+c}{3}$$

The dimensions of several landforms such as sorted and non-sorted patterns, stone-banked lobes and turf exfoliated depressions were measured using a tape. Where appropriate, both downslope and across-slope diameters of such features were measured. Measurement of stripe widths and depth of sorting were taken along lateral transects, using a cord. The centre height of some sorted patterns was determined by measuring the vertical distance between border centres and pattern centres. Sometimes a small trench was dug through thufur or sorted patterns to determine subsurface sedimentary and sorting characteristics. The sample size was commonly restricted by the available number of landforms at the sites. Further, the scarcity of sites in which particular cryogenic landforms (e.g. stone-banked lobes and sorted circles) have developed, has contributed to any conservative approach taken in their analysis.

Bulk density of thufur and turf-exfoliation scarps was measured using a miniature tubular soil corer for sample collection (Figure 3.1). Soil samples (usually 0.05 - 0.1 kg) were collected from most sites throughout the year and hermetically sealed for the later determination of moisture content, organic matter content and particle size distribution.

Figure 3.1 Miniature tubular soil corer used for measuring soil bulk density.

3.4.2 Geomorphic Process Assessment

Despite some criticism in the literature (e.g. Tallis, 1981; Birnie, 1993), stakes and pins are commonly used for the determination of erosion rates and terrace retreat. In this study, metal pins (0.40 and 0.50 m in length) were inserted vertically in front of turf exfoliation terraces as a guide to determining terrace retreat. Although the pins were randomly distributed around four turf exfoliated depressions and along a lower valley slope terrace, it was ensured that all sides (aspects) containing terraces were well represented. The number of pins used was restricted to 85, owing to weight constraints. The pins should have little effect on the terrace soil characteristics as they are located away from the eroding edge. The retreat of the terrace edges was measured throughout the year to determine any seasonal rates of retreat.

Wooden pegs (120 in total) were inserted in thufur apexes and in inter-thufur areas to determine possible heave within a 25 m^2 area of a wetland. Pegs were inserted to 5, 10 and 20 cm depth to determine any heave differentials with depth of insertion. Five thufur at the Mashai Valley site were examined for possible heave. Wooden pegs were inserted at the thufur apexes and inter-thufur depressions. The distance along the ground surface was measured between the inner-edge of the thufur-apex pegs and inter-thufur pegs, with measurements taken during warm and cold seasons to determine any seasonal thufur expansion or contraction.

Four pvc troughs (each 0.5 m long) were inserted along low gradient ($\leq 23°$) Mashai Stream banks (Figure 3.2). Above this, 0.5 m wide soil plots were constructed with rocks (Figure 3.3), so as to capture only those soil particles moving towards the stream channel within the plot. The sides of the troughs were anchored with boulders to limit their heaving. This precautionary measure appeared to be successful with three of the troughs. However, one of the troughs showed considerable evidence of heave and therefore its data were not included in the study. Trough sediment samples were collected approximately every two months to enable seasonal comparisons of soil movement rates and to aid interpretation of processes.

At two sites on the Mafadi Summit northwest-facing slope, clasts were painted along a transect, parallel to the slope contours (Figures 3.4 and 3.5). This was undertaken to determine clast movement within coarse and fine stripe patterns. Larger clasts were first brushed clean of dust before an "Aerolak, Lac-R-Spray" was used to paint the surface. Large rocks were marked and used to serve as benchmarks for the original transect position. Markers on adjacent rocks remained *in situ* and revealed that the rocks had not moved during the monitoring period. Although Caine (1981) reported that painted soil surfaces biases the measurements of movement, the results should, nevertheless, provide some estimates of

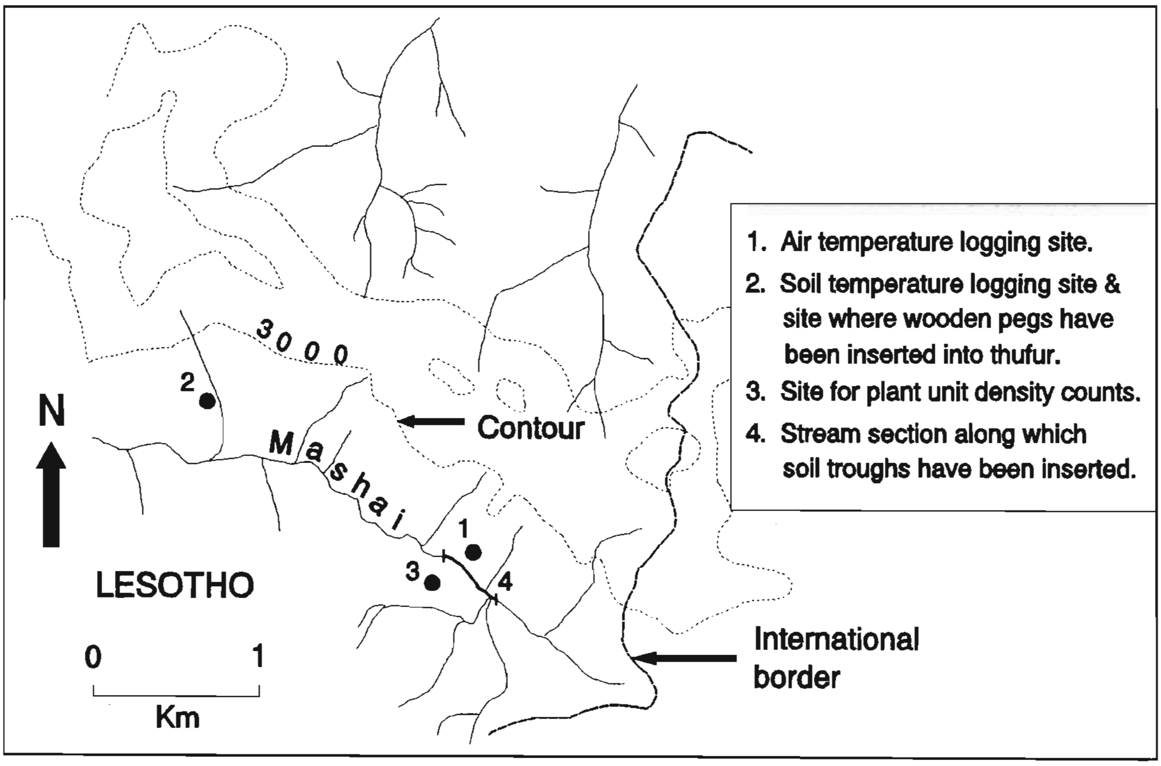


Figure 3.2 The Mashai Valley study site.

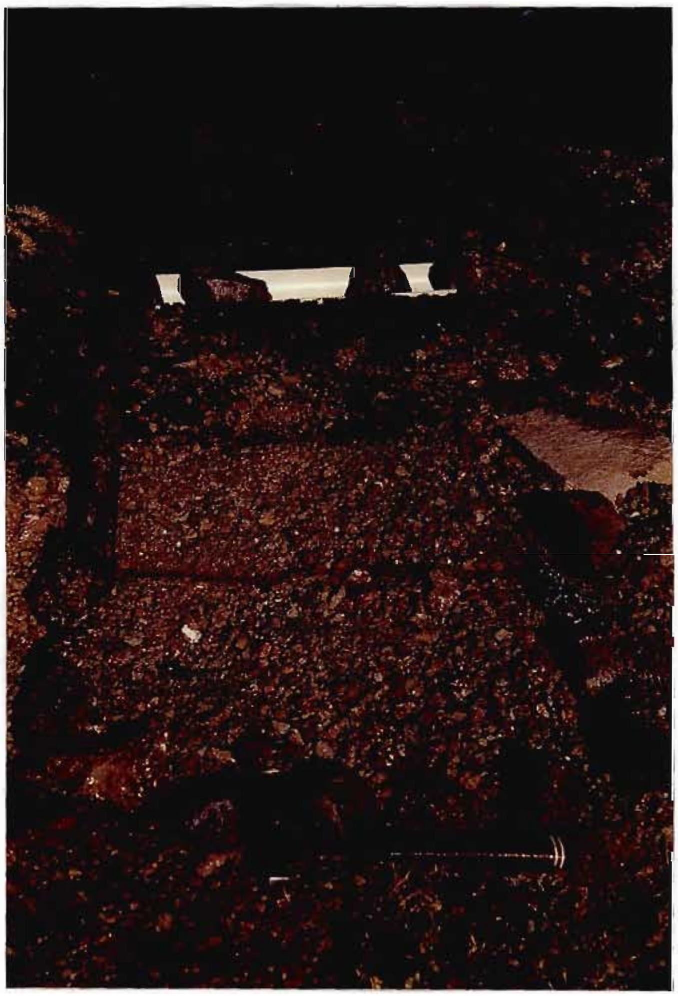


Figure 3.3 A soil plot and pvc trough used for the capture of moving soil particles.

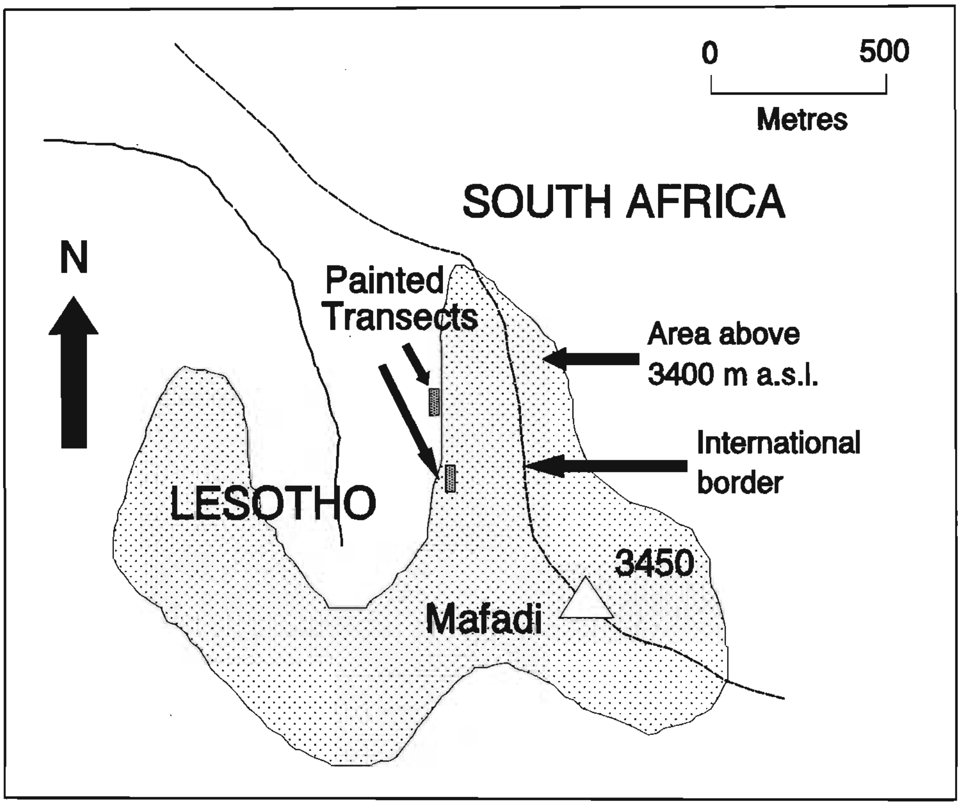


Figure 3.4 The location of painted transects near the Mafadi Summit.

Figure 3.5 Painted transects across sorted stripes near Mafadi Summit.

relative particle movement. Pérez (1993) has cautioned that measurement of stone movement should be taken over several years and that short term data should be avoided when evaluating the "long-term evolution of talus" (p200). As this is a short-term study and movement was measured only over one year (June 1993 to June 1994), no long-term inferences will be made. Owing to hazardous weather conditions in June 1994, data could only be collected from one transect. Although the area is grazed by sheep and goats, these were never seen on the stone fields, owing mainly to low forage availability. Although no signs of biotic disturbance was evident, data are, nevertheless, cautiously used as herdsmen and their horses were occasionally seen on the slopes.

Detailed short-term measurements of particle movement by needle ice were determined within experimental plots along the Mashai stream. Four transects of uniform slope gradient were used for the monitoring of a single mornings' movement. Twenty four painted gravel-size markers were spaced 5 cm apart. The markers were carefully selected to ensure similar size and shape. A further three transects, consisting of 50 markers spaced 2 cm apart, were designed to monitor particle movement over a nine week period. The location of the transect start positions were fixed by large immobile rocks or fixed rods. Caution was taken not to alter the soil surface conditions while placing particles along the transect. Movement of the marked particles was determined by measuring the distance along the fall line from the centre of the particle to the original transect. Some painted markers could not be recovered, and thus movement rates are probably conservative. It is likely that most markers lost, were due to burial. For instance, one particle was uncovered 15 mm below the surface.

Needle ice growth and ablation cycles were determined during eight nights in the Mashai Valley. Tinytalk temperature loggers (see Section **3.4.3**) were used to measure air and ground temperatures at the needle ice growth site. Air temperature was recorded at 2 cm and 150 cm above the ground surface while ground temperatures were recorded at 1 cm below the surface. Needle ice growth was measured at the same position within a cluster of needle ice throughout the night at hourly intervals using calipers. These data were then plotted against both air and ground temperatures.

3.4.3 Temperature Recording

Ten *Tinytalk*^{TM} miniature temperature loggers (Figure 3.6) (manufactured by Orion Components, U.K.), with a range of +46°C to -37°C, were used for the monitoring of air and ground temperatures in the Mashai Valley. The Tinytalk loggers are capable of storing up to 1800 measurements on an E-PROM. The launch setting permits the operator to select the duration of the logging program and consequently the measurement interval. The available logging duration varies from 15 minutes (0.5 sec. logging interval) to 360 days (4.8 hr. logging interval). All the loggers were calibrated in water and ice baths before installation in the field. The tinytalk temperature thermistors have a resolution of +/- 0.2°C between 0°C and 80°C, which may be important when considering the "zero curtain" effects in soils. During 1993/1994 air temperature was recorded every 1.6 hours (logging interval for 120 day duration) at 2960 m a.s.l. in the Mashai Valley, approximately 50 m from the Mashai Stream, towards the south-facing slope (Figure 3.2). The logger was placed beneath a rock scarp, so as to avoid direct sunlight. The recording height above the ground was 0.80 m, and the sensor attached to a rod 0.30 m from the rock-scarp surface (Figure 3.7). The passage of air was unimpeded at the site. In the Mashai Valley, it is logistically impossible, primarily due to vandalism, to continuously record temperature at the standard height of 1.50 m and provide technical screening against direct sunlight.

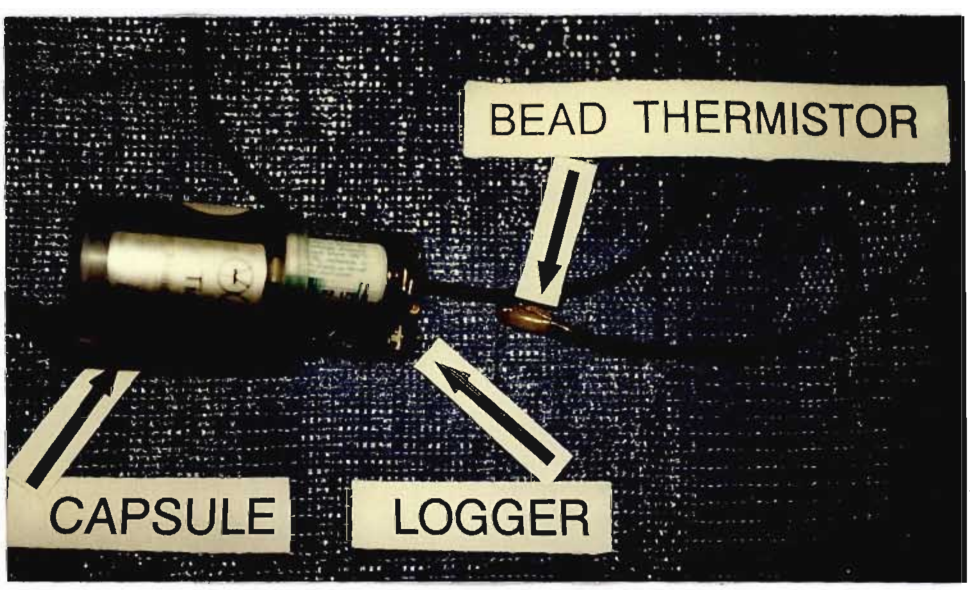


Figure 3.6 The *Tinytalk*^{TM} miniature temperature logger.

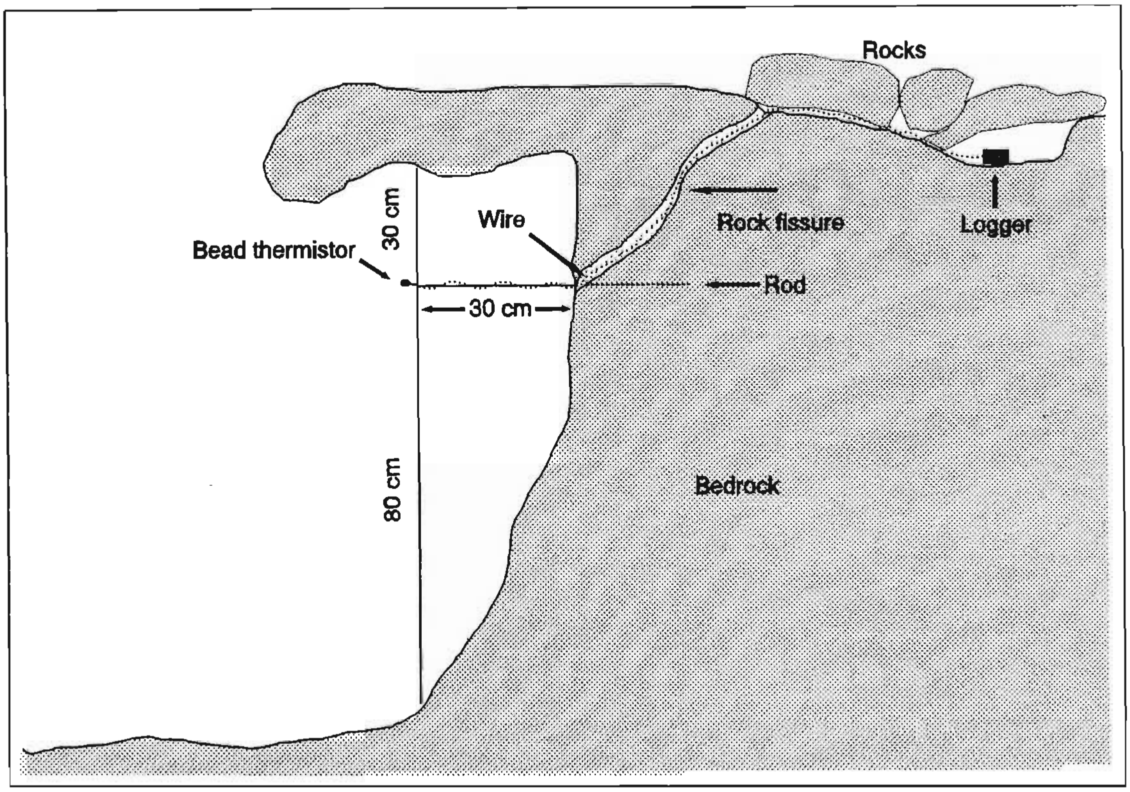


Figure 3.7 Diagrammatical sketch showing the location of the Tinytalk logger and bead thermistor beneath a rock-scarp in the Mashai Valley, which was used for the recording of air temperature.

1993/1994 Temperature Recording

One sensor was placed within a thufa at 12 cm depth and another within an inter-thufur depression at 12 cm depth. A further two sensors were placed at a depth of 5 and 15 cm within the wetland. Two sensors, also at 5 and 15 cm depth, were buried at the same altitude adjacent to the wetland (Figure 3.2). It was ensured that all sites had uniform vegetation cover. The logging interval for ground temperatures was 4.8 hours, deployed for a period of 360 days. Although precautionary measures were taken to hermetically seal the loggers, several of the loggers placed within the wetland had their protective canisters cracked, possibly owing to frost-induced pressure. Consequently, data were lost from these loggers.

1995/1996 Temperature Recording

A large thufa was selected for the 1995/1996 thufur thermal monitoring. The thufa had an across-slope diameter of 73 cm, an upslope diameter of 55 cm and a height of 32 cm. During 1995, thermistors were inserted into the sides of the thufa to a distance of 15 and 20 cm on the north, east, south and west aspects (Figure 3.8). In 1996, thermistors were placed into the sides of the thufa north and south aspects to a distance of 1 cm, 5 cm and 10 cm.

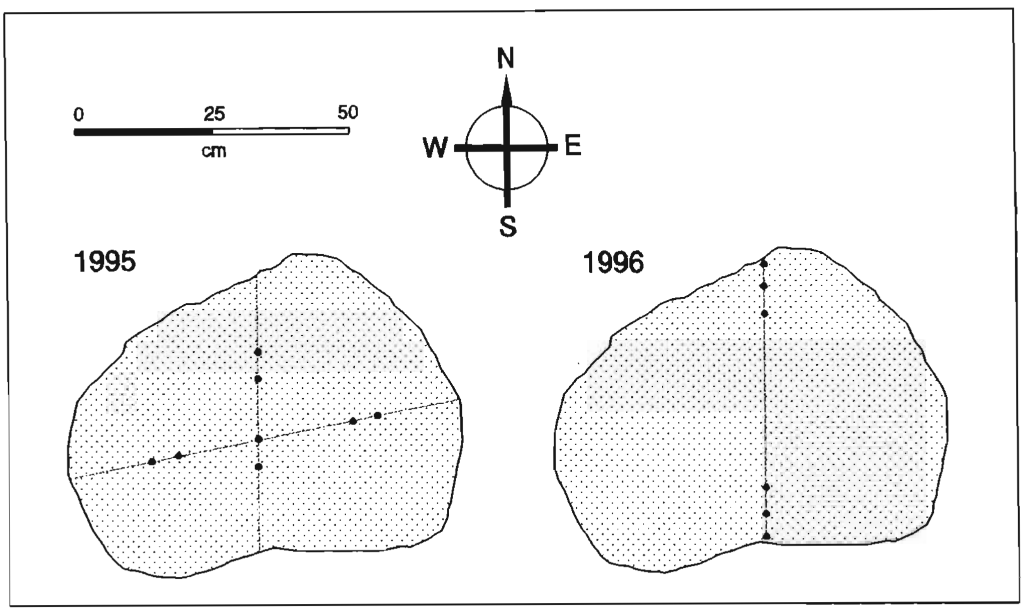


Figure 3.8 Plan view of thermistor positions within a thufa.

3.4.4 Vegetation Analysis

Plant specimens were collected and placed in a portable press. Plants were then identified with assistance from the University of Natal's Grassland Science and Botany Departments.

Plant unit density counts

Individual plant unit numbers in 4 cm^2 grids were counted within three turf exfoliated depressions during December of 1993 and March, May and June of 1994 in the Mashai Valley (Figure 3.2). This was undertaken to determine any seasonal cyclicity of vegetation growth within turf exfoliated depressions. The percentage grids with plant cover was also determined. Ten grids were used at a regular interval across each of the three depressions.

3.5 LABORATORY TECHNIQUES

Each soil sample was weighed, then oven dried at 105°C for 48 hours. Soil moisture content was determined and expressed as a percentage weight. Soil colour of some oven dried samples was described using the Munsell soil colour chart (1990 revised edition). Organic matter content was obtained through the loss of weight by incineration at 700°C. Particle size analyses (Folk, 1968) were performed by sieving soils through a mesh series ranging from 8000μm (-3.0 ϕ) to 63μm (4.0 ϕ). Each mesh was thoroughly cleaned before undertaking sieving for 15 minutes. Soils are described according to the Wentworth size class.

3.5.1 Fine Fraction Analysis

The fine fraction (< 63μm) of some soil samples was assessed using a SA-CP3 Shimadzu centrifugal particle size analyser (Figure 3.9). This photometric method is a desirable means of detection since no contact with the particles is required and the sedimenting system is undisturbed. The fine samples were reduced to less than 1 g by coning and quartering. A 20% glycerine solution was prepared and approximately 100 cm^3 put in a beaker. The sample obtained by the reduction procedure was added and stirred well with a spatula until none of the sample was left floating on the liquid surface (Figure 3.10). It was ensured that the sample solution exhibited an absorbency in the range of 80 and 120 (the absorbency is expressed as log[Io/I] x 100, where I is transmissivity in real time). The pure dispersant liquid was then poured into a cell and the zero percent set (Figure 3.10). The suspension was again stirred to eliminate possible segregation due to sedimentation and a small amount poured into the cell (Figure 3.10). Automatic sampling was then undertaken. The built-in microcomputer converts the change in particle concentration detected by the photometric system into the particle size distribution and prints out the results. To ensure statistical accuracy, several measurements (ie three or more) were taken with the reduced sample.

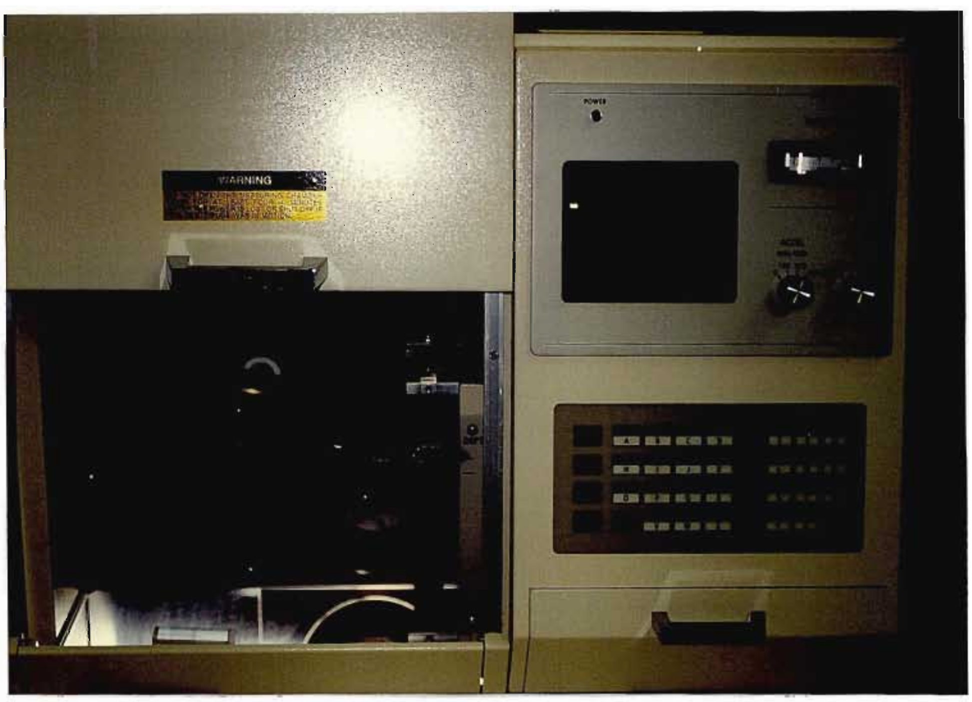


Figure 3.9 The SA - CP3 Shimadzu centrifugal particle size analyser, used for fine fraction analysis.

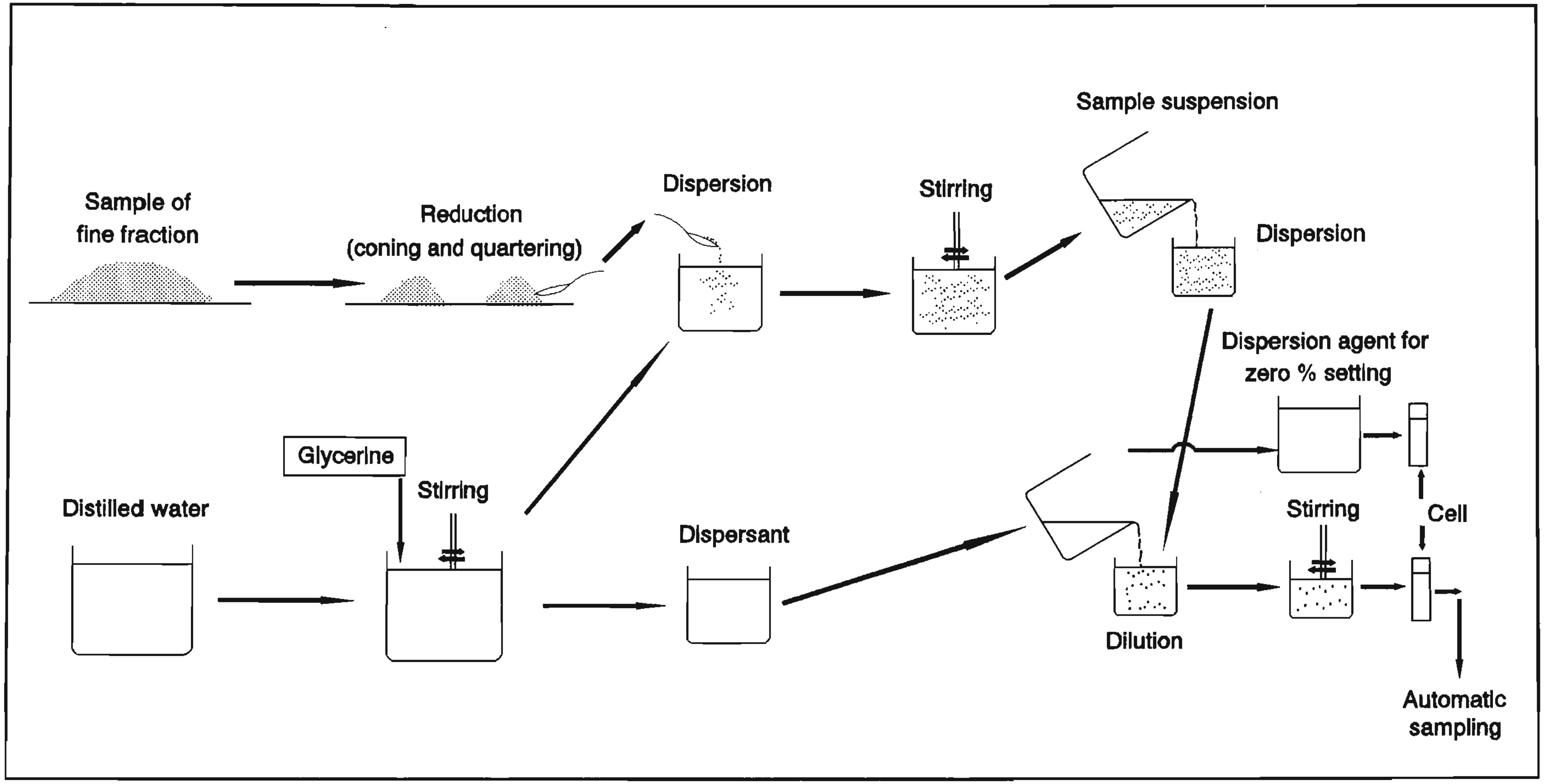


Figure 3.10 Procedure for preparing sample suspension.

3.6 STATISTICAL TECHNIQUES

3.6.1 Spearman's Rank Correlation Coefficient

The Spearman's rank correlation coefficient (r_s or RSQ) was frequently used to assess the degree of relationship between two variables:

$$r_s = 1 - \frac{6.\Sigma D^2}{n^3 - n}$$

Where: D^2 = the square of the difference between rank values for each pair of observations and n = the number of samples.

3.6.2 Fabric Analysis

Flatness and roundness indices were calculated according to Cailleux (1945) and Dobkins and Folk (1970) (as suggested by Barrett, 1980) respectively. Roundness values for larger clasts and boulders were determined using the visual technique of Krumbein (1941). Mean values and standard deviations (s) were calculated and some samples tested to determine whether they are of the same population, using the Mann-Whitney U test:

$$U = n_1 n_2 + \frac{n_1(n_1+1)}{2} - R_1$$

Where: n_1 = the sample size of sample 1

n_2 = the sample size of sample 2

R_1 = the sum of the ranks in sample 1

Because the samples were usually small (n = 10 to 20), the significance of the Mann-Whitney U-statistic was tested using the critical values for the test at the 0.05 level of rejection.

Where H_o is accepted, the two samples are from the same population.

Where H_o is rejected, the two samples are from different populations.

Rose diagrams were produced to show the orientation of clasts found within debris deposits and stone-banked lobes. Some of these data were tested for probability of random occurrences by means of the Chi-square test:

$$x^2 = \Sigma \frac{(O-E)^2}{E}$$

Where: Σ = the sum of

O = the observed frequency

E = the expected frequency

Where H_o is accepted, the fabric is random.

Where H_o is rejected, the fabric is orientated (the 0.05 significance level was used).

For some sorted patterns, the percentage weight of a given fraction in the pattern border was compared against the same fraction in the pattern centre. A lateral-sorting index (Sl) for each fraction of the pattern could then be calculated (modified after Ballantyne and Matthews, 1983), where:

$$Sl = \frac{Bo}{Ce}$$

Bo = percentage by weight of a given fraction in the border sample.

Ce = percentage by weight of the same fraction in the centre sample.

Where Sl = 0 there is "perfect sorting", with all material of a given size sorted into the centres.

Where Sl = 1.0 this "indicates no lateral sorting of a given fraction".

Where Sl > 1.0 this "indicates that a given fraction is deficient in the centre sample compared with the equivalent border sample".

This is the inverse of the sorting index introduced by Ballantyne and Matthews (1983, p344).

Similarly, the percentage weight of a given fraction at the pattern surface was compared against the same fraction at depth (sub-surface). For these a vertical-sorting index was calculated, where:

$$Sv=\frac{Su}{De}$$

Su = percentage by weight of a given fraction at the pattern surface.

De = percentage by weight of the same fraction at depth (subsurface).

Where Sv = 0 there is "perfect sorting", with all material of a given size sorted at depth (subsurface).

Where Sv = 1.0 this "indicates no vertical sorting of a given fraction".

Where Sv > 1.0 this indicates that a given fraction is deficient at depth (subsurface) compared with the equivalent surface sample.

Sorting indices were also calculated for paired striped pattern samples (modified after Pérez, 1992a) where:

$$Si=\frac{Cs}{Fs}$$

Cs = percentage by weight of a given fraction in the coarse stripe.

Fs = percentage by weight of a given fraction in the fine stripe.

Where Si = 0 there is "perfect" sorting with all material of a given size laterally sorted into the fine stripe.

Where Si = 1.0 there is no lateral sorting of a given fraction.

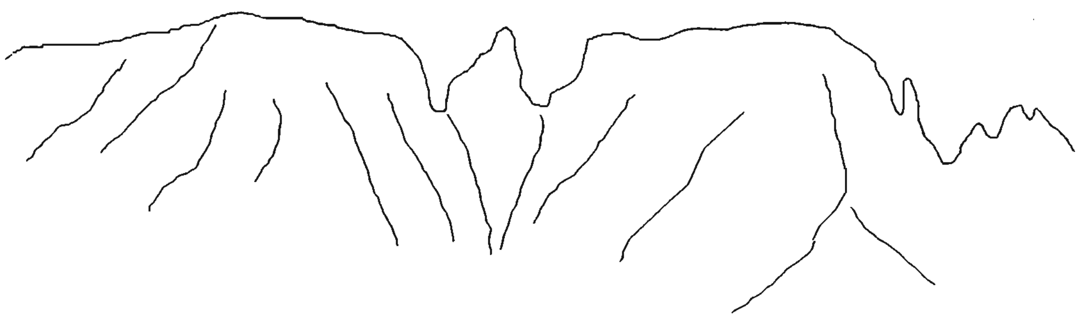
Where Si > 1.0 this indicates that a given fraction is deficient in the fine stripe (modified after Pérez, 1992a).

3.7 SUMMARY

The variety of field-based, laboratory and statistical methods used in this dissertation have been outlined. Time and logistical constraints, of which some are mentioned in this chapter, have wielded considerable control on the methods used. A somewhat broad approach is taken to ensure that most of the data needs outlined in Section **1.3** are met.

CHAPTER FOUR

SORTED PATTERNED GROUND



CHAPTER 4

SORTED PATTERNED GROUND

4.1 INTRODUCTION

According to Washburn (1956), patterned ground is a "group term for the more or less symmetrical forms, such as circles, polygons, nets, steps and stripes, that are characteristic of, but not necessarily confined to, mantle subject to intensive frost action" (p824). Although patterned ground commonly has a cold climate origin (Washburn, 1956), several patterns have been attributed to non frost-related processes (Kelletat, 1985; Hallet, 1990; Van Vliet-Lanoë, 1991; Ahnert, 1994). Dybeck (1957) has ascribed some soil polygon formation to stones sinking into a "liquid" mud, whereas Kelletat (1985) has observed rain-splash and rain-wash produce sorted patterns remarkably similar in morphology to those with a frost-related origin. Recently, Bennett (1993) has described stone stripes from North Wales, attributed primarily to sheep disturbing debris cones upslope of the stripes. Numerous authors have, however, produced evidence which indicates that frost-related processes play a significant role in the development of many sorted patterns (e.g. Goldthwait, 1976; Hall, 1979, 1983, 1994b; Washburn, 1979; Ballantyne and Matthews, 1982; Van Vliet-Lanoë, 1988, 1991; Hallet, 1990). Washburn (1956) reviewed as many as 19 proposed mechanisms for the formation of various types of patterned ground found in Arctic and alpine regions. However, the formative processes of many types of patterned ground are still poorly understood (French, 1976; Washburn, 1979; Hallet, 1990), in part because different patterns are controlled by different mechanisms (Hallet, 1990), and further, because many are said to have a polygenetic origin (Washburn, 1979).

The absence of precise information on formative processes taking place in well developed patterns has limited the value of patterned ground in palaeoclimatic reconstructions (Hallet and Prestrud, 1986). Despite the problems of identifying processes and being able to accurately date such patterns, they may, nevertheless, be of potential value in Quaternary research. For instance, some patterns have been used to suggest the former occurrence of permafrost (e.g.

Williams, 1975; Goldthwait, 1976; Wayne, 1983), even though their inter-relationship is still imperfectly understood (Washburn, 1979, 1980; Hallet and Prestrud, 1986).

Washburn (1950, 1956, 1973) has classified patterned ground, utilizing the pattern shape and the sorting of stones as the main criteria. The classification includes the sorted and non-sorted patterns: circles, nets, polygons, steps and stripes. Sorted patterns "are made prominent because of the segregation of stones and fines" (Gleason *et al.*, 1986, p216), while non-sorted patterns are distinguished by "ground cover or colour variations" (Krantz, 1990, p117). This chapter examines some sorted patterned ground types occurring in the Drakensberg, while Chapter 5 focuses on the non-sorted varieties. An ongoing problem has been the defining of a standard distinction between large sorted patterned ground forms and the miniature forms (Washburn, 1979). Parameters such as freezing conditions (Nicholson, 1976), the size of stones in the coarse border (Washburn, 1979) and the spacing between pattern centres and borders (Goldthwait, 1976) have all been used to differentiate between large and small forms. Patterns with a mesh diameter or stripe width between 10 and 25 cm (Troll, 1944), or not exceeding about 20 cm (Wilson and Clark, 1991), have respectively been regarded as "small" or "miniature". This dissertation will use the classification given by Wilson and Clark (1991) and refer to patterns with a mean mesh diameter or stripe width of under 20 cm as "miniature", and those equal to or exceeding 20 cm as "large".

Although patterned ground has been the focus of much research worldwide (Warburton, 1990), little work of any substance has been produced regarding these landforms in southern Africa. Many researchers (Troll, 1944; Van Zinderen Bakker, 1965; Harper, 1969; Hastenrath, 1972; Hastenrath and Wilkinson, 1973; Dardis and Granger, 1986; Lewis, 1988a, 1988b; Boelhouwers and Hall, 1990; Boelhouwers, 1991a; Hanvey and Marker, 1992; Grab, 1992b) have reported patterns such as sorted circles, polygons, stripes and non-sorted circles from the Drakensberg and Lesotho. Detailed assessments and quantitative data, however, are lacking. A precise analysis of a variety of patterned ground types occurring in the high Drakensberg is presented and provides the first account of miniature patterned ground from an alpine region in southern Africa.

4.2 MINIATURE SORTED CIRCLES (MASHAI VALLEY)

Sorted circles are defined by Washburn (1956, p827) as "patterned ground whose mesh is dominantly circular and has a sorted appearance commonly due to a border of stones surrounding finer material". Although several periglacial studies have focused on sorted circles (e.g. Ballantyne and Matthews, 1982; Rissing and Thorn, 1985; Hallet and Prestrud, 1986; Washburn, 1989), few have examined miniature varieties.

In the high Drakensberg, miniature sorted circles occur primarily near streams and wetlands, where adequate moisture is available throughout the year. The patterns sometimes form in river gravels where the microtopography is often irregular, but occur most frequently where the surface slope is between 0° and 3°. Occasionally, patterns develop on basalt steps near areas of ground seepage.

4.2.1 Characteristics

Two comparable varieties of miniature sorted circles were observed at the Mashai Valley study site. The first variety (A) consists of a somewhat raised, fine cone-shaped centre which is surrounded by coarse gravels and small pebbles (Figures 4.1 and 4.2). The second variety of circle pattern (B) is also characterized by a fine cone-shaped centre, but is enclosed by sub-rounded and "blocky" stones and/or boulders (Figures 4.1 and 4.3). These commonly occur amongst boulders and clasts which are underlain by finer sediments. Hastenrath (1973) observed similar patterns on Mount Kenya and referred to these as "fine earth mounds" (p174 & 175). The centres of variety (A) patterns are raised about 3.5 cm and average 6.3 cm in diameter (Table 4.1). Although most centres appear rounded, slight elongation sometimes occurs. Lateral sorting between centres and borders is evident, with the centres containing fewer pebbles and considerably finer particles than the borders (Figure 4.4). The clay/silt content for pattern surfaces is 0.3% for centres and 0% for borders (Figure 4.4). Sorting may occur to a depth varying between 10 and 15 mm.

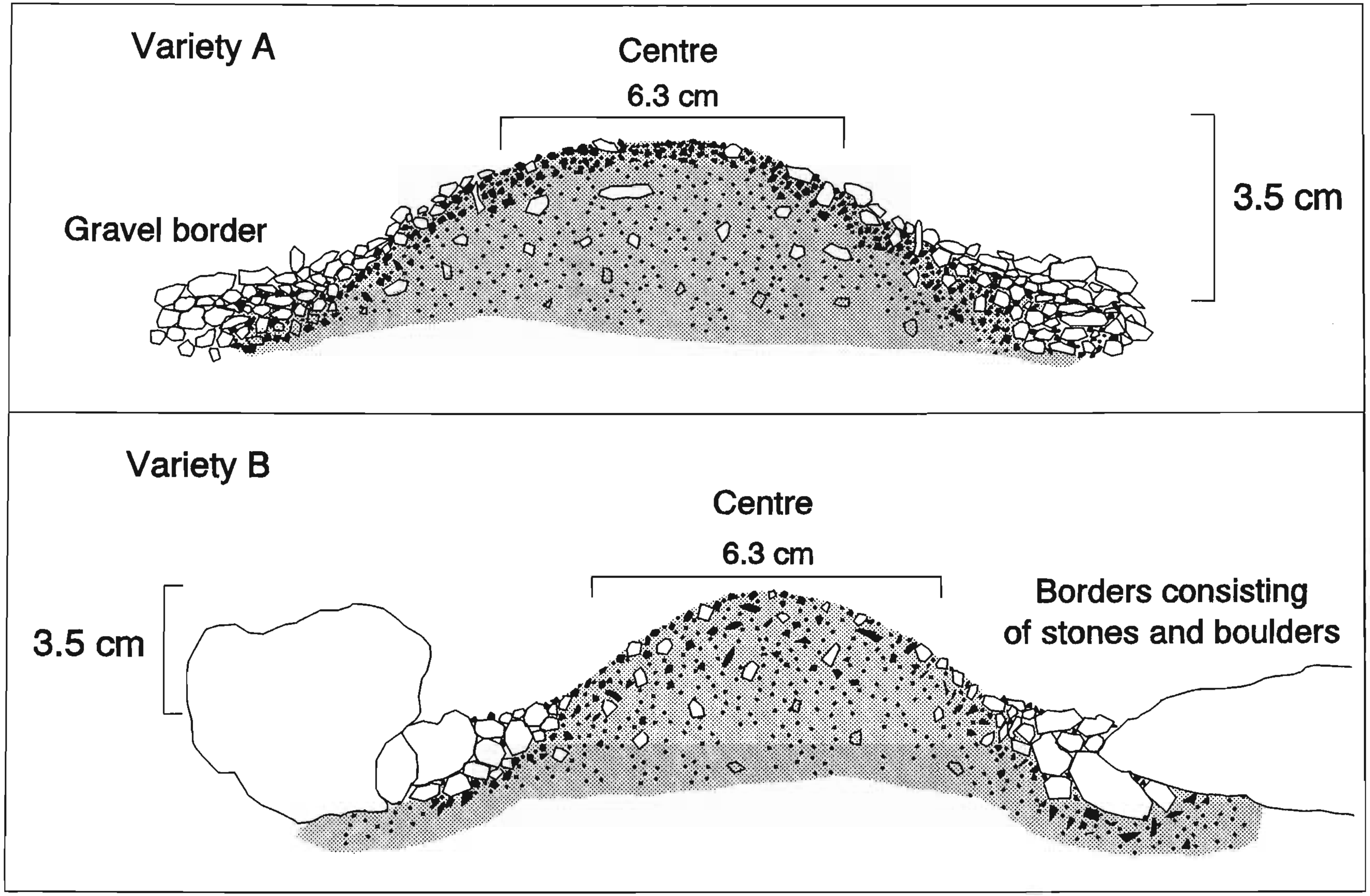


Figure 4.1 Sections through sorted circle varieties (A) and (B) from the Mashai Valley.

Figure 4.2 Variety (A) sorted circles from the Mashai Valley.

Figure 4.3 Variety (B) sorted circles from the Mashai Valley.

	Centre	Border
% moisture by weight	13.1%	5.2%
Diameter	6.3 cm	8.4 cm
Height above border	3.5 cm	-

Table 4.1 Mean characteristic values of sorted circles (Variety A) from the Mashai Valley, June 1993 (n = 10).

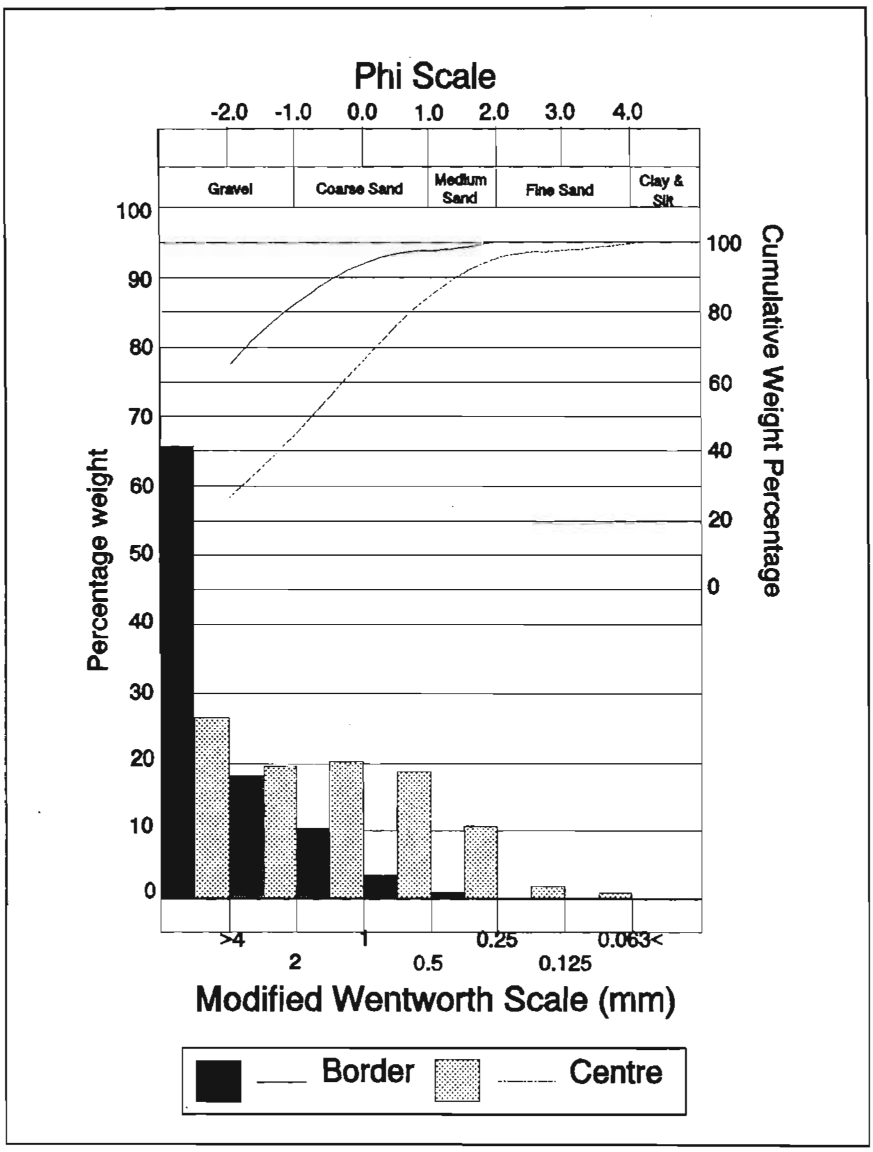


Figure 4.4 Particle size distribution for border and centre samples of sorted circles (Variety A) from the Mashai Valley (n=10).

4.2.2 Discussion

Granulometry and moisture are recognised as the important controls affecting the frost susceptibility of soils (Goldthwait, 1976; Thorn, 1976; Meentemeyer and Zippin, 1981; Van Vliet-Lanoë, 1991). Goldthwait (1976) has indicated that the host material must contain at least 10% fines for the development of frost-sorted patterns. Although the miniature sorted circle centres had a winter soil moisture content of 13.1% (June, 1993) by weight (Table 4.1), the frost susceptible soil component (i.e. clay and silt) was absent. The absence of clay and silt may be attributed to deflation and also to the entrainment and removal of such particles during periods of fluvial action. Thorn (1976) has indicated that sites containing less fines require more freeze-thaw cycles to initiate pattern development. It appears, therefore, that it is the abundance of soil freeze-thaw cycles during the winter months that permits pattern development at the Mashai Valley site. Several characteristics may indicate that the circle patterns found at the Mashai Valley site are frost-sorted. Particle size characteristics confirm strong lateral sorting; with medium and coarse gravels sorted into the borders, while finer material is more abundant in the circle centres (Figure 4.4). Other characteristics which may indicate frost-action at the site are:

1. The occurrence of most patterns in groups rather than as individuals (Goldthwait, 1976).
2. Elevated circle centres (Goldthwait, 1976; Washburn, 1979).
3. Ice-cored circle centres during winter.

Observations have shown that the accumulation of cobbles and gravel around centres of finer material is primarily the result of needle ice lifting material in the centres. Cobbles and gravel are then moved laterally under the influence of gravity. It has also been suggested (Ballantyne and Matthews, 1983; Washburn, 1989) that such sorting may occur because surface material moving away from pattern centres is generally coarser than that heaved to the surface in the middle of pattern centres.

4.3 MINIATURE SORTED POLYGONS AND NETS (MASHAI VALLEY AND MAFADI SUMMIT)

Sorted polygons are defined by Washburn (1956, p831) as "patterned ground whose mesh is dominantly polygonal and has a sorted appearance commonly due to a border of stones surrounding fine material". Sorted polygonal patterns appear to be common in periglacial environments (Washburn, 1980) but are sometimes referred to as "sorted nets" because of their irregularity and the absence of straight edges (Washburn, 1979; Ballantyne, 1987a, 1991). In fact, Washburn (1956) has suggested that patterned ground whose "mesh is intermediate between that of a sorted circle and a sorted polygon" should be referred to as a "sorted net" (p830). In the Drakensberg, sorted circles and sorted polygons frequently occur in close proximity, and ultimately merge to form sorted nets. The patterns referred to in this sub-section are somewhat more multi-angular than rounded in appearance, and shall therefore be referred to as "miniature sorted polygons".

Miniature sorted polygons have been reported predominantly from alpine regions such as the Rocky Mountains, U.S.A. (Butler and Malanson, 1989); Scottish Highlands (Ballantyne, 1991); Kuh-i-Jupar, Iran (Kuhle, 1974); Atlas Mountains, Morocco (Couvreur, 1973) and Mount Kenya and Kilimanjaro, East Africa (Hastenrath, 1973). It is possible that Arctic and sub-Arctic regions, owing to extensive permafrost and intense frost action, host larger polygonal patterns than are commonly found in alpine regions. Further, the larger polygonal patterns appear to have received greater international attention than miniature forms in such high latitude regions (e.g. Jonasson and Sköld, 1983; Hallet, 1990). Miniature sorted polygons from the Drakensberg have been described by Hastenrath and Wilkinson (1973); Lewis (1988a); Hanvey and Marker (1992) and Grab (1992b), who have presented some limited data on pattern dimensions and depths of sorting.

In the Drakensberg, polygonal patterns have been recorded on the basaltic plateau as well as at lower altitudes in the Clarens Formation sandstones. Lewis (1988a) and Grab (1992b) have reported patterns in sandstone tarns at altitudes of 2550 and 2420 m respectively. The best developed polygonal patterns were observed in May 1989 within a tarn near Sehlabathebe N.P.

(see Figure 2.1) at 2250 m a.s.l. The tarns are seasonally filled with water, drying out towards winter. Although such seasonal patterns are likely to be polygenetic in origin, they appear to be initiated by desiccation cracking, as has also been suggested for miniature patterns in the maritime Antarctic (e.g. Chambers, 1967). In the high Drakensberg, polygonal patterns are most commonly located within and adjacent to stream beds and ground seepage sites, but have also been reported from pan-shaped depressions (Hastenrath and Wilkinson, 1973).

4.3.1 Characteristics

Twenty miniature sorted polygons were examined at the Mashai Valley site and another ten patterns at the Mafadi Summit site. The patterns in the Mashai Valley were examined in river gravels near the air temperature logging site (Figure 3.2) and the Mafadi Summit patterns were found on non-sorted steps (Figure 2.5). Pattern dimensions and characteristics for the two sites are given in Table 4.2.

Mean centre dimensions range from 11.6 cm (Mafadi) to 16 cm (Mashai), with a maximum value of 23.4 cm at Mafadi Summit (Table 4.2). Boelhouwers (1991a) has reported polygonal patterns with diameters ranging from 5 to 20 cm for the high Drakensberg. Polygonal centres become somewhat elongated downslope when the surface gradient is over 1°. At Mashai Valley, where patterns developed on a 3° surface, the mean across-slope to downslope elongation is 1: 1 46, whereas at Mafadi there is no significant elongation (1: 0 94), possibly owing to the shallower 1° surface gradient (Table 4.2). It is worth noting that Wilson and Clark (1991) found an almost identical elongation of 1: 1 45 for miniature sorted nets in East Falkland where the surface gradient was 3° to 6°.

The mean distance between centres and the height of centres above their respective borders is similar between the two sites, despite a 460 m altitudinal difference (Table 4.2). Although the centre height above the borders is commonly 3 cm, some patterns show little difference between the border and centre height. This may result from rapid movement of gravels and coarse sand into the borders, accompanied by only slight heave of polygonal centres. The distance between centres (i.e. border diameter) is often considerable for such miniature

Site	MASHAI	Std. Dev.	MAFADI	Std. Dev.
Sample Size	20	-	10	-
Altitude (m)	2920	-	3380	-
Gradient	3° (irregular)	-	1° (irregular)	-
Aspect	40°	-	60°	-
Centre Diameter - Ds (cm)	18.9	2.97	11.4	6.38
Centre Diameter - As (cm)	13.2	2.62	11.9	3.66
Centre Diameter - Mean (cm)	16	2.46	11.65	4.68
Ratio of As / Ds	1: 1.46	0.26	1: 0.94	0.32
Distance between centres (cm)	7.87	1.7	7.51	1.11
Centre ht above border (cm)	2.66	0.57	2.81	0.96
Max. depth of sorting (cm)	6	-	7.3	-
Ds = Downslope As = Across-slope				

Table 4.2 Dimensions and characteristics for sorted polygons from the Mashai Valley and Mafadi Summit.

patterns, averaging 7 to 8 cm. As mentioned earlier, this is possibly due to the rapid migration of coarse particles towards pattern borders. As particles migrate towards borders, they bank-up against clasts already occupying the borders, consequently increasing the lateral dimensions of such borders. A positive correlation (r_s = 0.59; significant at the 0.01 level) between polygonal border and centre dimensions was found at Mashai Valley; however, despite a similar value, no correlation was found at Mafadi Summit (r_s = 0.54; not significant at the 0.01 level) (Figure 4.5). Both the Mashai and Mafadi patterns show a strong positive correlation (r_s = 0.80 and 0.89 respectively; significant at the 0.01 level) between centre diameters and centre heights (Figure 4.5).

4.3.1.1 Pattern Sorting

1. Mashai Valley

Although fine fractions are absent at the Mashai site, the patterns display sorting typical of such miniature patterned ground (Goldthwait, 1976). Figure 4.6 shows how various particle sizes are sorted laterally and vertically. Samples were also taken mid-way between pattern centres and border centres. A gradual increase in the percentage of sand and decrease in the percentage of gravels occurs from borders to centres (Figure 4.6). The lateral extent of sorting is more pronounced at pattern surfaces than at 4 cm depth (Figure 4.6), as is also indicated by

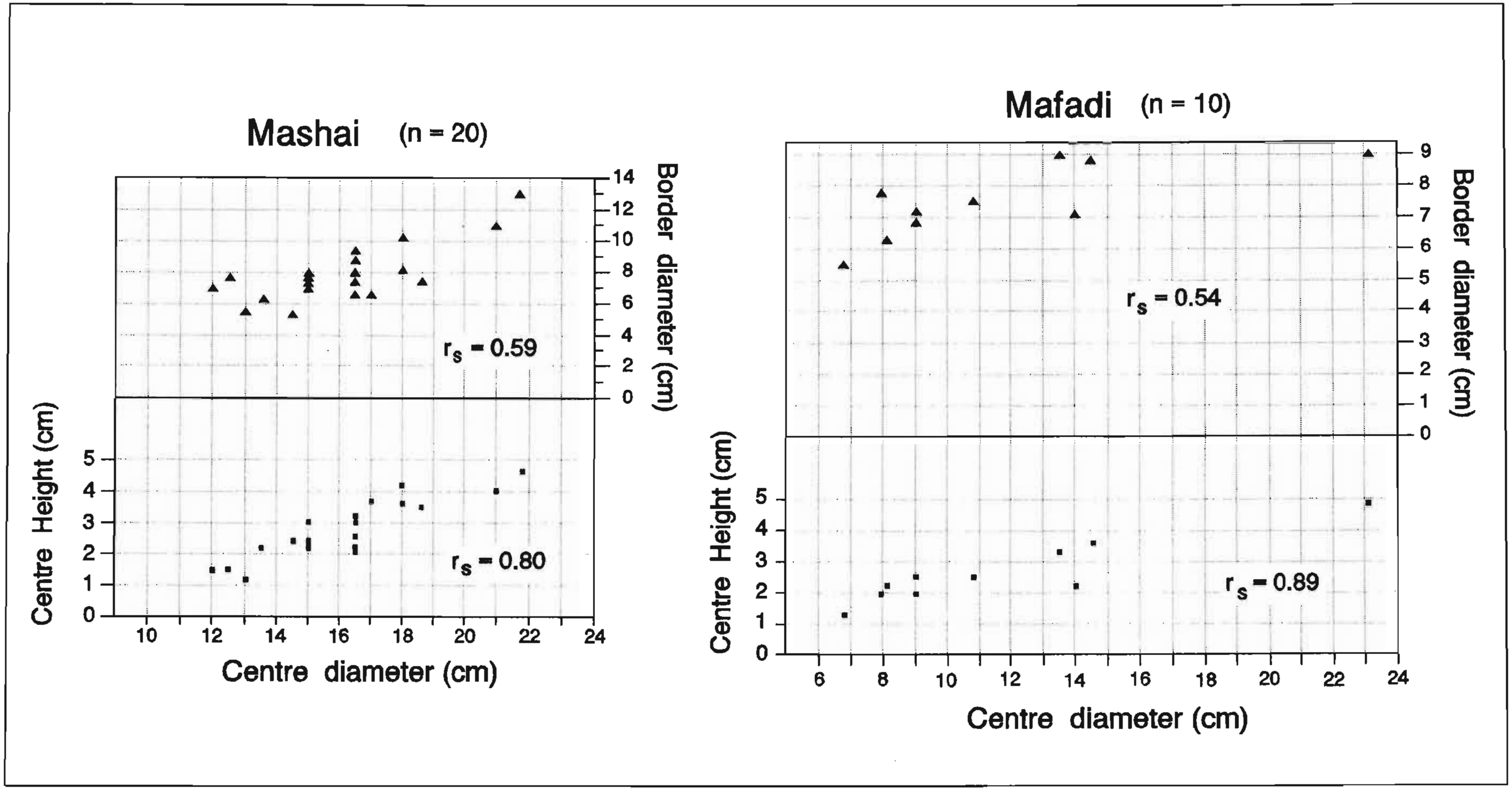


Figure 4.5 Correlation of polygonal centre diameters with centre heights and border diameters at Mashai Valley and Mafadi Summit.

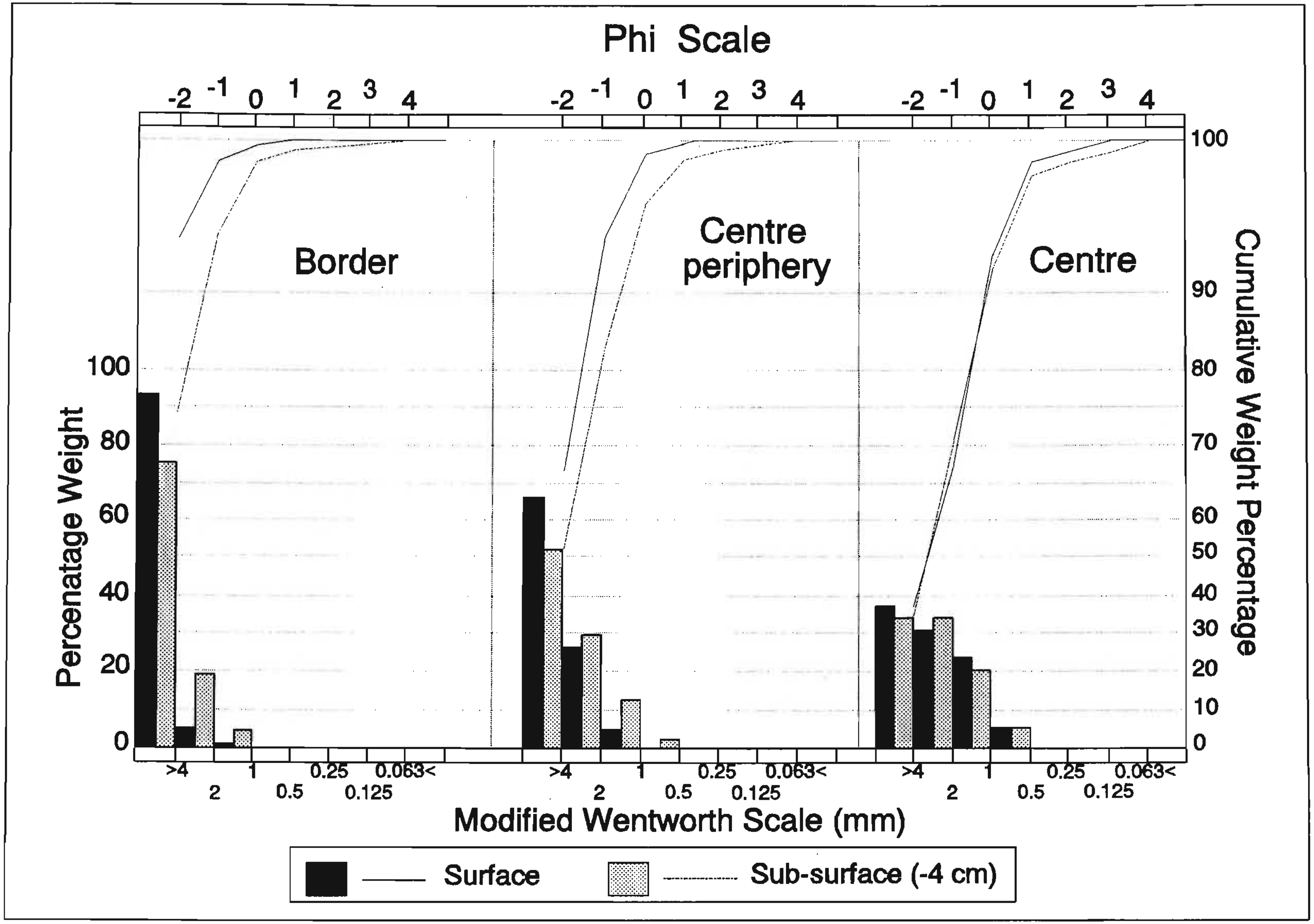


Figure 4.6 Particle size distribution for polygonal borders, centre peripheries and centres, Mashai Valley (n = 20).

SORTING INDICES FOR POLYGONAL PATTERNS (MASHAI SITE)								
Size Grade (Phi)	>-2	-1	0	1	2	3	4	4<
Sl surface	2.5	0.2	0.1	0	0	0	0	0
Sl sub-surface (-4 cm)	2.1	0.6	0.2	0.1	0.1	0.3	0.5	0.5
Sv centre	1.1	0.9	1.1	1	0.8	0.6	0	0
Sv inter-border-centre	1.2	0.9	0.4	0.2	0.2	0	0	0
Sv border	1.3	0.3	0.3	0.1	0	0	0	0

SORTING INDICES FOR POLYGONAL PATTERNS (MAFADI SITE)										
Size Grade (Phi)	>-3	-2.5	-2	-1	0	1	2	3	4	4<
Sl surface	25.5	6.6	0	0	0	0	0	0	0	0
Sl sub-surface (-4 cm)	-	-	1.3	0.9	0.9	0.8	0.7	0.7	0.5	0.4
Sv centre	-	-	1.1	1.2	1	0.8	0.6	0.7	0.6	0.5
Sv border	-	-	1.6	0	0	0	0.1	0.1	0.1	0

Table 4.3 Sorting indices for polygonal patterns at the Mashai Valley and Mafadi Summit sites.

the sorting indices (Table 4.3). For instance, coarse to medium gravels (< -2 Phi) have a lateral sorting index (Sl) of 2.5 at pattern surfaces and 2.1 at 4 cm depth, indicating that centres are more deficient of such particle sizes at their surfaces than at depth, relative to the adjacent borders (Table 4.3). Conversely, finer sand and clay/silt is all sorted into centres at pattern surfaces (Sl = 0), but is marginally less noticeable at depth (Sl = 0.1 to 0.5) (Table 4.3). The extent of vertical sorting increases progressively from centres to borders and attains a maximum depth of 6 cm at pattern borders. The depth of sorting at centres is seldom more than 1 cm. Sorting indices show that for most size fractions in centres, values are close to 1.0, indicating no vertical sorting for these size fractions (Table 4.3).

2. Mafadi Summit

Sorting values for the Mafadi patterns show a very similar trend to that of the Mashai patterns (Table 4.3 and Figure 4.7). The percentage clay/silt for the Mafadi patterns is very much higher than for the Mashai patterns, attaining a maximum of 17.2% at 4 cm depth for pattern centres (Figure 4.7). Lateral sorting within miniature polygonal centres was determined by taking samples at 2 cm intervals across the pattern centre (Figure 4.8). Gradual sorting across the centre is clearly evident (Figure 4.8). There is a general decrease in the percentage gravels and increase in the percentage fines towards the midpoint of pattern centres (Figure 4.8).

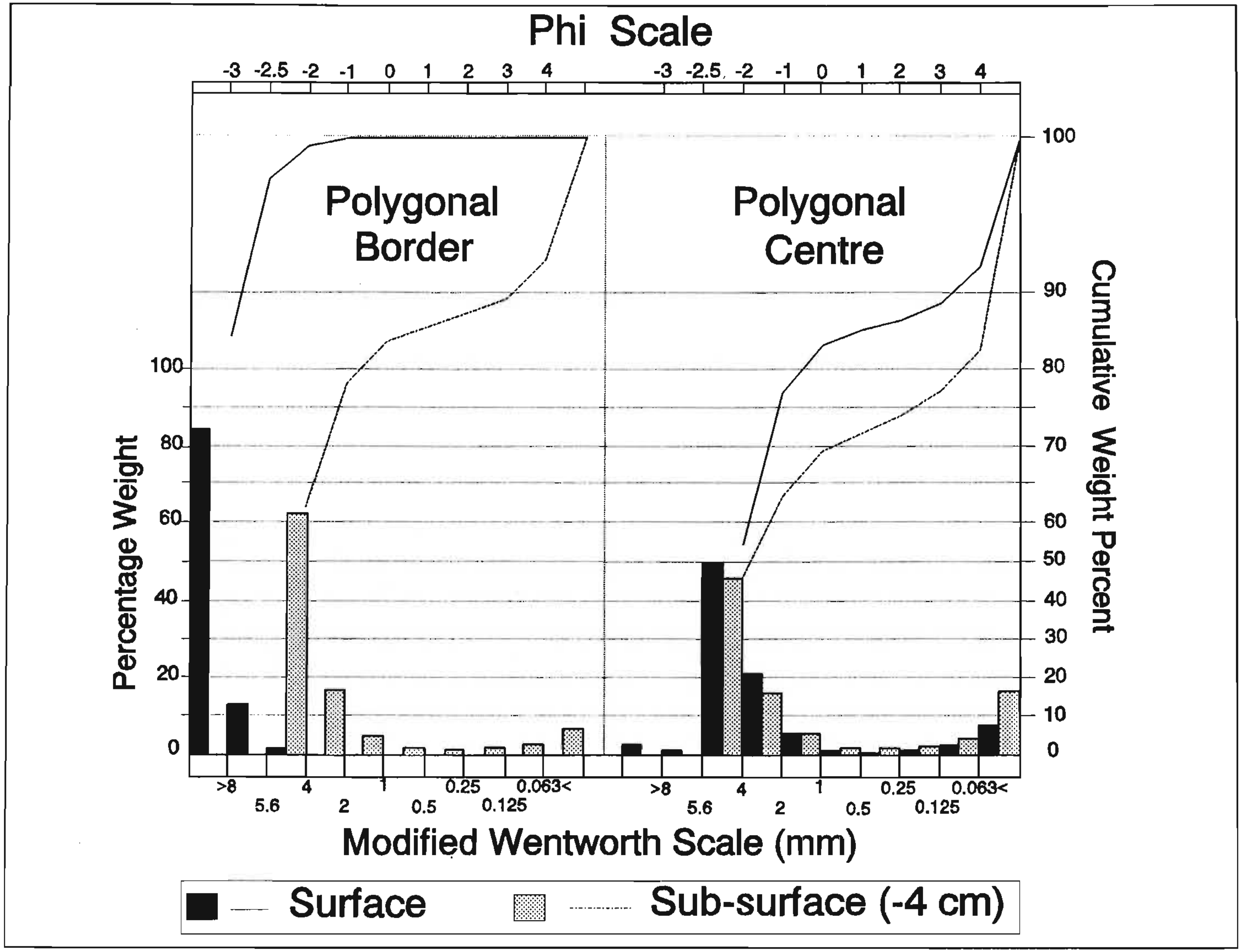
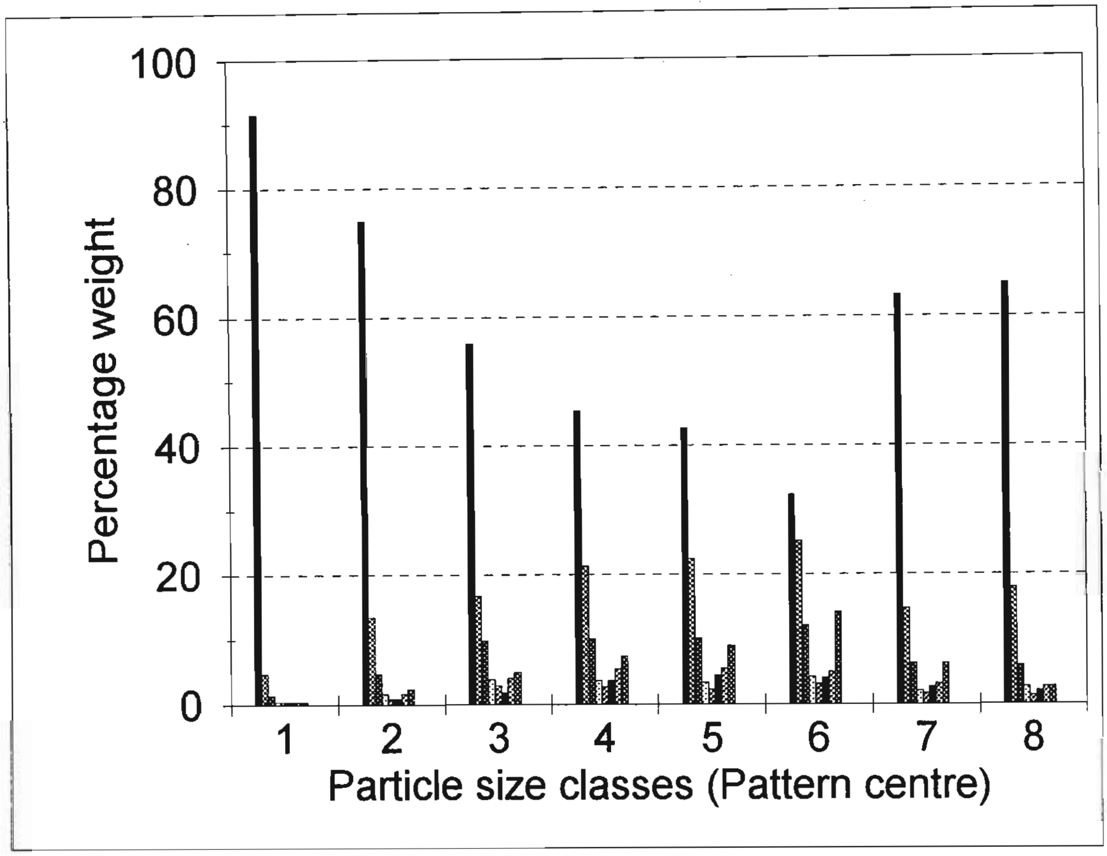


Figure 4.7 Particle size distribution for border and centre samples of sorted polygons from the Mafadi Summit (n = 10).



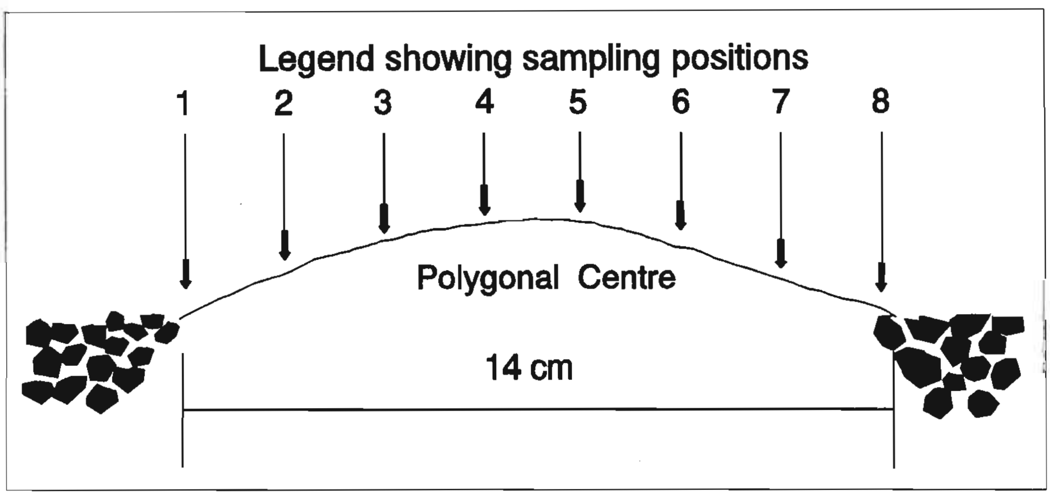


Figure 4.8 Sorting across a polygonal centre at the Mafadi Summit site. Samples were taken at 2 cm intervals across the pattern centre. Size classes represented at each sampling point are, from left to right: *>4μm, 4-2μm, 2-1μm, 1-0.5μm, 0.5-0.25μm, 0.25-0.125μm, 0.125-0.063μm, <0.063μm.*

4.3.2 Discussion

Miniature sorted polygons and nets occasionally appear at suitable sites in the high Drakensberg. However, the absence of moisture during winter prevents widespread pattern development. Sites containing moisture throughout the year are usually found adjacent to rivers and streams where the microtopography is often steep, rendering such sites more suitable for striped patterned ground. A further limiting factor for the development of polygons and nets is the absence of fines along such drainage zones. It appears that site-specific factors such as surface gradient and soil composition, depth and moisture are the primary controls for the development of polygons and nets in this region. Altitudinal differences of 460 m on the summit do not appear to have significantly influenced miniature pattern dimensions (Table 4.2).

Desiccation and thermal contraction cracking are frequently considered to be primary controls for the development of some polygonal sorted patterns (Büdel, 1982; Ballantyne and Matthews, 1983; Krüger, 1994). Polygonal patterns observed along the Mashai Stream begin to emerge in autumn when soils are still saturated, and are therefore unlikely to be initiated by desiccation cracking. It is more likely that desiccation cracking influences pattern development towards the end of winter when soils dry out rapidly due to cold, dry and extremely windy conditions. Desiccation and polygonal crack development are, however, common on drier slopes and may be of greater importance in the development of polygonal patterns such as those found on Mafadi Summit. Similarly, contraction cracking (Krüger, 1994) and/or convection cracking (Guangpan and Min, 1993) may help initiate pattern formation. However, the frost cracking model only describes polygonal patterns (Krantz, 1990) and therefore is unlikely to explain the origin of circular patterns at the Mashai site.

Stream levels in the study area subside rapidly from April, leaving saturated sediments exposed and subject to rapid cooling. This commonly promotes an uneven microtopography with heaved centres and depressions, possibly initiated by differential swelling and frost heaving (Van Vliet-Lanoë, 1988, 1991; Van Vliet-Lanoë *et al.*, 1990). Nicholson (1976) proposed that patterns may develop as a result of such uneven microtopography, with the dome-shaped areas likely to undergo greater heave than the adjacent depressions, which in turn encourages new areas of heave. Observations from the Mashai Valley would support such a formative process for polygonal patterns. During April/early May of 1993 and 1994 it was observed that small sand banks adjacent to and within stream beds developed uneven surfaces with raised areas and depressions, yet little sorting was visible (Figure 4.9). By late May/early June, the uneven surfaces flattened, yet still showed irregularity, but now with more visible particle sorting. Coarser gravels occupied the depressions, whereas the raised areas consisted predominantly of finer gravels and sand. By mid-winter (July), much of the irregularity had disappeared, to be replaced by well sorted polygonal patterns (Figures 4.10 and 4.11).

Moisture appears to be drawn towards the raised areas, where it freezes and ultimately increases the distance between soil particles (Guangpan and Min, 1993). These raised areas eventually become polygonal centres. Slight vertical sorting at centres is possibly the result of larger particles moving upward (by the frost-pull process) in the direction of heat flow, relative to finer material (Anderson, 1988). Ballantyne and Matthews (1983) have suggested that the decline of coarser particles at pattern centre surfaces is not very rapid, primarily because clasts are continuously being heaved up from below. Fines at centre surfaces may also be subjected to deflation, as a result of intense and frequent winds during the winter months. Such processes may help explain the general downward fining of particles within pattern centres.

Lateral sorting appears to be induced by mechanical heave at centres, with gravels moving under gravity. Gradwell (1957) also reported that the growth and collapse of needle ice may result in coarse material migrating towards "cracks" (borders). As needle ice was found growing from within polygonal centres on several occasions, it is possible that it may have contributed towards particle movement away from such centres. Gradwell (1957) and Krüger (1994) have further suggested that rain-splash and/or rain-wash may aid such lateral sorting. The general absence of rain during the winter months would, however, exclude such rain-

Figure 4.9 Gravel surface adjacent to the Mashai Stream, April 1994. Although uneven surfaces are present, there is little visible sorting.

Figure 4.10 Gravel surface adjacent to the Mashai Stream, June 1994. Well developed polygonal patterns are visible.

Figure 4.11 Polygonal patterns adjacent to the Mashai Stream, June 1994.

induced sorting mechanisms in the high Drakensberg. The movement of gravels towards borders eventually widens the borders, reduces the height difference between borders and centres and strengthens vertical sorting at borders.

These seasonal polygons and nets are ultimately destroyed by rain-splash, rain-wash and other fluvial processes (e.g. stream flow) towards October/November. The polygonal-sorted patterns are a product of frost-related processes. The tendency for clasts to occupy troughs surrounding centres of finer material, slightly elevated pattern centres, the presence of needle ice and the occurrence of patterns in groups, are a few properties which may support a frost-related origin (e.g. Goldthwait, 1976; Washburn, 1979; Ballantyne and Matthews, 1982; Butler and Malanson, 1989; Wilson and Clark, 1991).

4.4 MINIATURE SORTED STRIPES AND PARTICLE DISPLACEMENT ON MAFADI SUMMIT

According to Washburn (1956, p836), sorted stripes are "patterned ground with a striped pattern and a sorted appearance due to parallel lines of stones and intervening strips of dominating finer material orientated down the steepest available slope". Numerous studies have examined varieties of sorted stripes from alpine regions (e.g. Troll, 1958; Hastenrath, 1973, 1974, 1977; Heine, 1977; Wilson, 1992) and the sub-Antarctic islands (e.g. Hall, 1979, 1983, 1994b; Heilbronn and Walton, 1984) in particular. Detailed reviews of sorted stripes are given in Washburn (1979) and Pérez (1992a), but as mentioned by Pérez (1992a), little work has dealt with their morphological or pedological characteristics. Further, most studies 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. Beaty, 1974; Mackay and Mathews, 1974a; Pérez, 1984). Relatively few findings have attributed miniature sorted stripes found in periglacial regions to other factors such as ground freeze-thaw cycles (e.g. Hall, 1994b), rill development (e.g. Caine, 1963; Ballantyne, 1987a; Warburton, 1987), rainstorm erosion (Kelletat, 1985) or pre-existing corrugated topography (Muir, 1983).

For ease of differentiation, those stripes perceived to be entirely a product of needle ice action are referred to as "needle ice stripes" and those thought to have a more polygenetic origin as "sorted stripes". Although needle ice stripes have frequently been described from the Drakensberg and Lesotho mountains (e.g. Troll, 1958; Hastenrath and Wilkinson, 1973; Hanvey and Marker, 1992), most accounts refer to non-sorted or pebble-sorted (mean particle size < 64 mm) varieties. Cobble-sorted (mean particle size > 64 mm) stripes have not yet been reported from the high Drakensberg. Miniature sorted stripes of the cobble-sorted variety were located on an extensive stone field on the Mafadi northwest face between 3380 and 3410 m a.s.l.. Morphologically (size and sorting characteristics), the stripes differ from those typically attributed to needle ice action. Similar cobble-sorted stripes have not been found elsewhere but may occur on other high altitude summits such as Lithotobolong (3400 to 3440 m a.s.l.) and Thabana-Ntlenyana (3400 to 3482 m a.s.l.)(Figures 2.2 and 2.3).

Figure 4.12 The Mafadi Summit stone fields on which a variety of perennial sorted patterns are found.

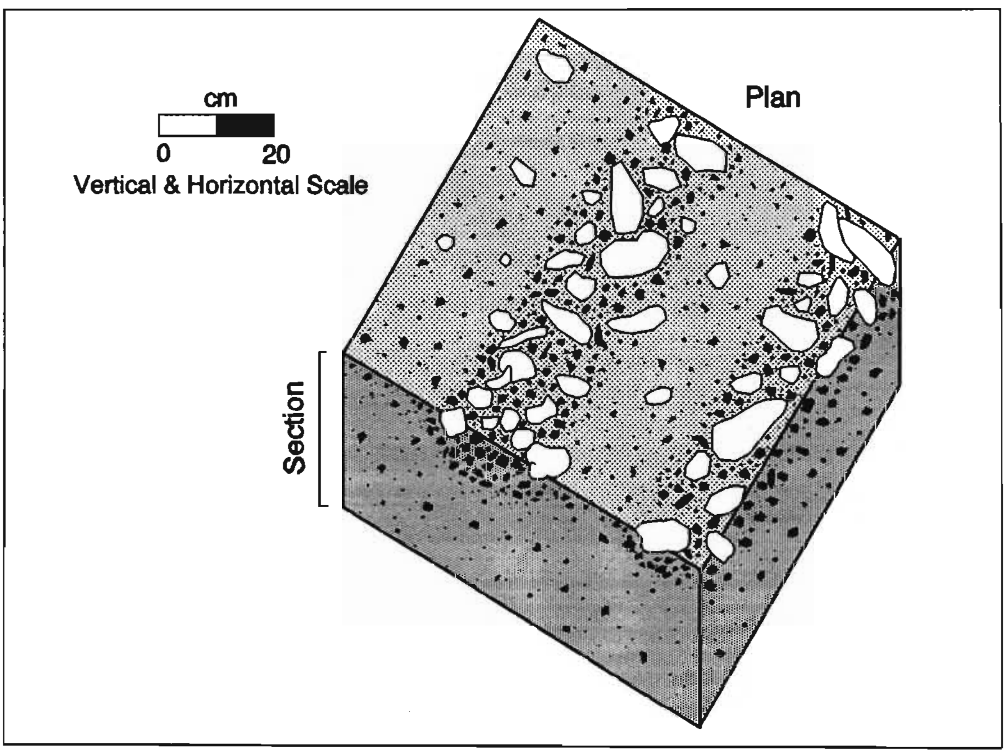


Figure 4.13 Sketch showing plan view and section through miniature cobble-sorted stripes at Mafadi Summit.

4.4.1 Characteristics

The observed sorted stripes occur sporadically on the steeper and lower sections of the Mafadi northwest stone field (Figure 4.12). While sorted stripes on Mafadi Summit are located where the slope gradient is commonly 15° (± 3°), in other areas they are said to occur on gradients between 3° and 30° (Pérez, 1992a). Mean stripe orientation on Mafadi Summit was found to be within 3° of slope direction (293°) (Table 4.4). Pérez (1992a) has described similar patterns from the Venezuelan Andes and found that stripe departure from slope direction was only 1.7°. It would therefore appear that such stripes are "orientated down the steepest available slope", as is defined in their original definition (Washburn, 1956, p836). The correspondence of slope direction with stripe orientation indicates that the patterns have been significantly affected by the downslope shift of particles (e.g. Pérez, 1992a).

The almost parallel lines of alternating coarse and fine clasts are usually only one to three metres in length and sometimes converge or rapidly diffuse lower down the slope. Coarse stripe widths average 12.6 cm and fine stripe widths 16.2 cm, giving a mean coarse to fine stripe ratio of 1: 1 3 (Table 4.4). Wilson (1992) found a similar value where the coarse (9.4 cm) to fine (15.6 cm) stripe ratio was 1: 1 7. Stripes at any one site on Mafadi Summit seldom exceed six or seven pairs. The clasts within coarse stripes usually occupy troughs while the finer stripes sometimes form ridges.

4.4.1.1 Particle Size and Sorting

The stripes have formed in gravely soils and even within fine stripes at 4 cm depth there is 49.6% gravel (Table 4.4 and Figure 4.14). Although coarse stripes have 98.9% gravel at their surfaces, there is appreciable silt and clay at 4 cm depth (23.3%) (Table 4.4 and Figure 4.14). Figures 4.13 and 4.14 and Tables 4.4 and 4.5 show vertical and lateral sorting characteristics for the Mafadi sorted stripes. It is noticeable that there is strong vertical sorting of the coarse stripes where values are close to 0 for most size fractions. As can be expected, vertical sorting in the fine stripes is less pronounced. Maximum sorting depth of the coarse stripes is 6.7 cm,

	% Moisture	Wentworth Grade (%)				
		Gravel	Coarse sand	Medium sand	Fine sand	Silt/clay
Fine Stripe - surface	2.1	78.5	7.9	2.3	2.7	8.6
Fine Stripe - sub-surface (-4cm)	13.7	49.6	8.1	5.4	6.6	30.3
Coarse Stripe - surface	1.5	98.9	0.5	0.2	0.1	0.3
Coarse Stripe - sub-surface (-4cm)	12.2	58	6.7	5.9	6.1	23.3

Mean coarse stripe clast size (n=40)
a - axis = 7.4 cm (std. dev. = 5.4)
b - axis = 5.5 cm (std. dev. = 4.0)
c - axis = 3.1 cm (std. dev. = 2.3)

Slope Gradient = 15°
Slope Aspect = 290°
Mean Stripe Direction = 293°

Mean coarse stripe width = 12.6 cm (std. dev. = 3.9)
Mean fine stripe width = 16.2 cm (std. dev. = 4.5)
Ratio of coarse : fine stripes = 1: 1·3
Maximum depth of sorting = 6.7 cm
Stripe length is commonly between 90 - 130 cm

Table 4.4 Dimensions and characteristics for miniature sorted stripes from the Mafadi Summit.

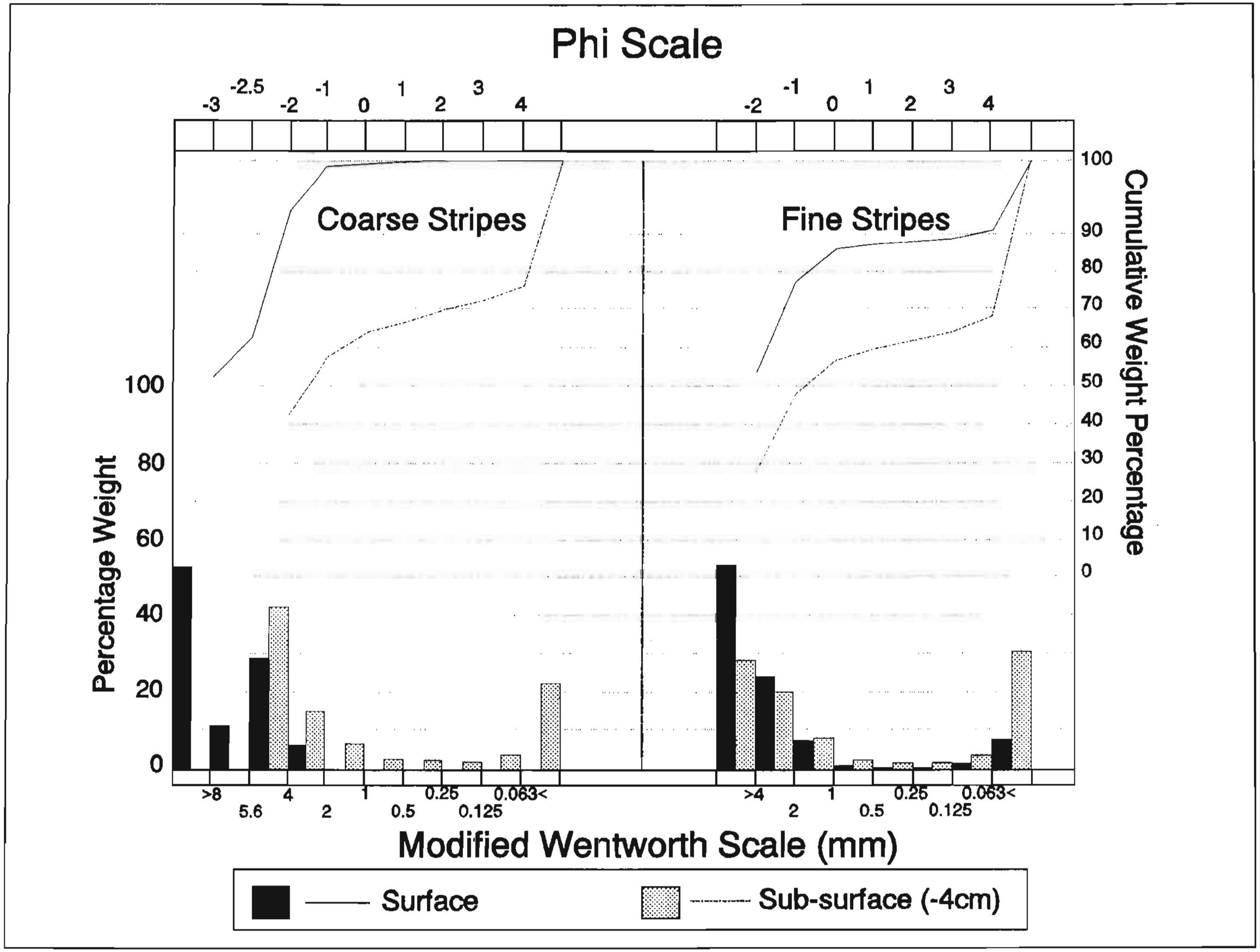


Figure 4.14 Particle size distribution for coarse and fine stripes from the Mafadi Summit (n = 5 pairs of stripes).

while that of the fine stripes is usually only 3 to 4 cm (Table 4.4). At the surface there is also strong lateral sorting, with particles less than 2 mm predominating in the fine stripes. However, at 4 cm depth lateral sorting becomes negligible (Table 4.5). While particle sizes larger than 2 mm are concentrated in the coarse stripes, other particle sizes show sorting indices approaching 1.0, thus indicating little or no sorting at 4 cm depth (Table 4.5).

Sorting Indices for sorted stripes (Mafadi summit)								
Particle size (Phi)	> -2	-1	0	1	2	3	4	4<
Sv coarse stripe	2.2	0.4	0.1	0.1	0	0	0	0
Sv fine stripe	1.9	1.2	1	0.5	0.4	0.3	0.4	0.3
Si surface	1.7	0.2	0.1	0.1	0	0	0	0
Si sub-surface (-4cm)	1.5	0.7	0.8	1.1	1.1	0.9	0.9	0.8

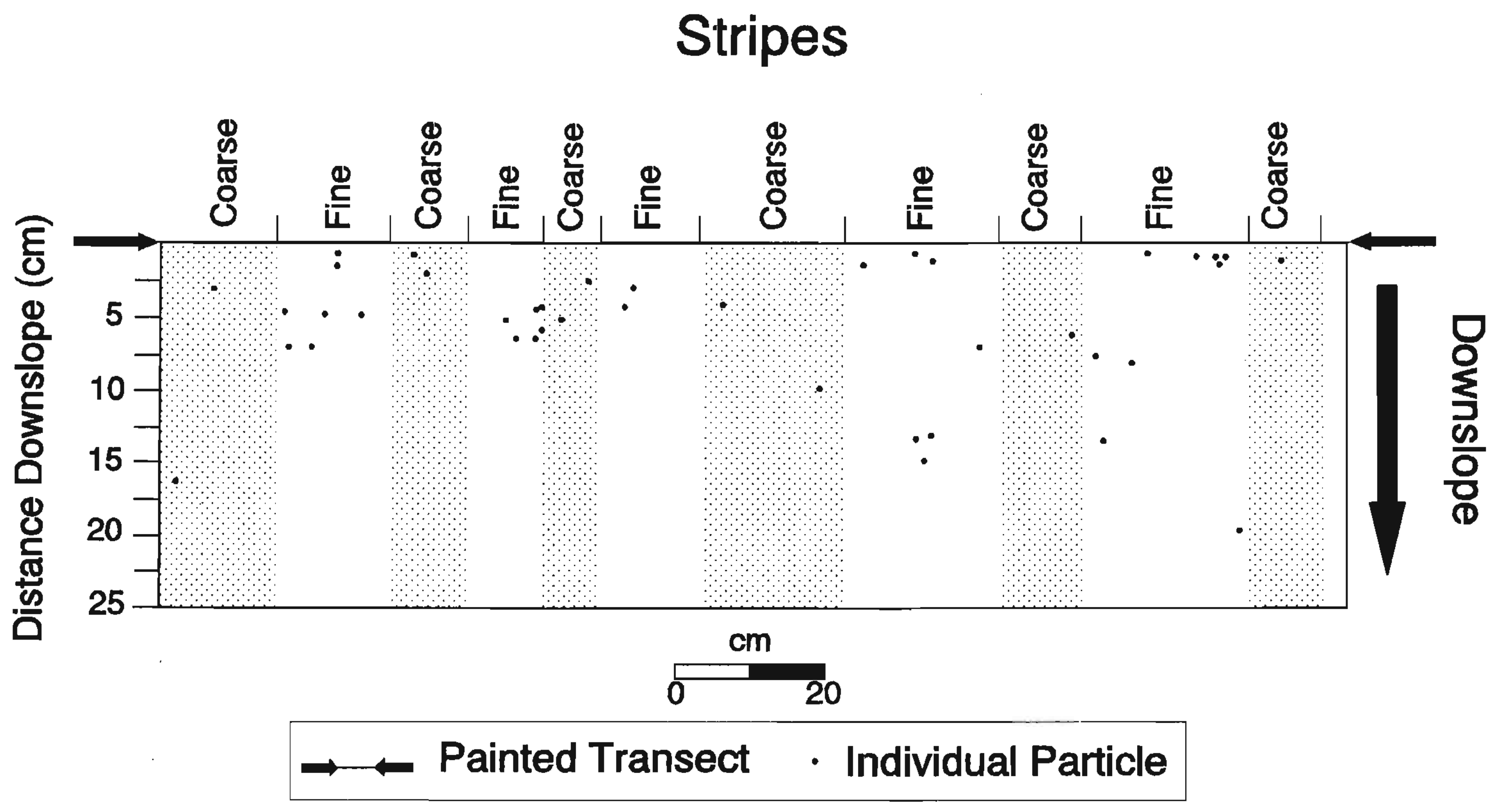
Table 4.5 Sorting indices for cobble-sorted stripes at Mafadi Summit.

4.4.1.2 Particle Movement

Particle movement on talus slopes and within striped patterns, apparently similar to those at Mafadi Summit, has been measured by such as Mackay and Mathews (1974b), Gardner (1979), Pissart *et al.* (1981) and Pérez (1985, 1987b, 1988, 1993). Such an analysis has not previously been undertaken in the high Drakensberg.

A line was painted across 11 sorted stripes (5 fine and 6 coarse stripes) to compare particle movement along coarse and fine stripes (Figures 3.4, 3.5 and 4.15). As only one colour (red) was used, the evaluation is unable to account for across-slope particle displacement. Results are therefore not considered unequivocal, but are used to demonstrate a general trend in particle movement over a limited period of one year.

Observations made over one year found that many particles along the transect had remained stationary, but a total of 41 individual particles ranging from medium sand to cobbles were



	n	Transect Distance	Particles moved	Mean rate of displaced particle movement (cm. yr^{-1})
Fine Stripes	5	80.0 cm	31 (1 per 2.6 cm)	4.75 (max. = 19.5)
Coarse Stripes	6	71.2 cm	10 (1 per 7.1 cm)	4.27 (max. = 16.1)

Figure 4.15 Individual downslope particle movement along sorted stripes at Mafadi Summit (from June 1993 to June 1994).

detected to have moved (within the top 10 mm of soil) downslope from the 150.2 cm long transect. Of the particles that had moved, it is apparent that a greater number had moved within the fine stripes (31) than within the coarse stripes (10) (Figure 4.15). The average rate of displaced particle movement within the fine stripes was 4.75 cm (from June 1993 to June 1994)(max. = 19.5 cm) while that for the coarse stripes during the same period was 4.27 cm (max. = 16.1 cm) (Figure 4.15). Observations and measurements from Barnett (1966), Mackay and Mathews (1974b) and Pérez (1988) have shown that rates of particle movement are also dependent on slope gradients, with more rapid movement occurring on steeper gradients. As the present study examined particle movement on only one slope for a limited period (12 months), no comparative or longer term pattern assessments could be made.

Boulders have been found to obstruct downslope particle movement, whereby material is dammed upslope and deflected around the boulder edges (Pérez, 1993). Immediately downslope of such boulders are areas relatively free of coarser particles, sometimes referred to as "fine earth flags" (e.g. Hastenrath, 1973; Pérez, 1987b, 1993). Pérez (1993) examined particle movement across such a fine earth flag and found little or no movement. However, particles adjacent to the fine earth flag had moved a considerable distance. Although such fine earth flags have not been reported from southern Africa, a few miniature fine earth flags were found below larger rocks on the Mafadi stone field (Figure 4.16a and 4.16b). These fine earth flags could provide further evidence for downslope particle movement on the Mafadi northwest face, or alternatively, for deflation processes at the site.

Figure 4.16a Fine earth flags are visible on the downslope sides of larger rocks to the left and right of the geological hammer (Mafadi northwest face, 3415 m a.s.l.).

Figure 4.16b A large rock with clasts banked against its upslope side (left) and a fine earth flag on its downslope side (right).

4.4.2 Possible mechanisms

A number of processes such as surface-wash (Mackay and Mathews, 1974b; Francou, 1984, 1988; Kelletat, 1985; Selby, 1985), runoff creep (De Ploey and Moeyersons, 1975; Pérez, 1988), dry debris slides (Pérez, 1987b, 1988) and snowcreep (Mackay and Mathews, 1974b) have been suggested as possible contributing mechanisms to particle movement on slopes in periglacial regions. Both surface and/or subsurface rills are sometimes considered as an initial cause for sorted stripes (e.g. Hay, 1943; Ballantyne, 1987a; Warburton, 1987; Wilson, 1992). Rills develop on the Mafadi slopes after heavy or prolonged rainfall events, especially towards late summer (pers. obs.). During such events, finer particles are washed downslope, leaving gravels and cobbles to occupy the rills. However, the often regular spacing, and occurrence of several pairs of stripes, would suggest that other mechanisms must have been operative in the stripe development.

The extremely strong winds, especially during the dry season, are likely to entrain smaller particles on the Mafadi slopes. Wind cannot, however, be considered as a primary mechanism for the formation of sorted stripes, as particles could equally well be blown or moved in any direction, depending on the wind bearing. Small dry debris slides observed near some of the stripes are likely to be caused by Basuto herdsmen and their horses travelling across the slopes, but, like the wind, are not the cause of the stripes.

On visiting the Mafadi site in April 1994 (mid-autumn), needle ice was widespread on the lower slopes below the stone field, but absent at the site where miniature sorted stripes occur. It appears that the soils dry out rapidly on these slopes, owing to their exposure to wind and the high radiation receipt, thereby limiting needle ice development during the cold period. Further, needle ice activity essentially only causes sorting at the surface and is therefore unlikely to have produced the Mafadi stripes where sorting reaches a depth of almost 7 cm. In June 1994, segregation ice was found beneath the fine stripes, which had a diurnal frozen layer of 4 to 5 cm and closely corresponds to the depth of sorting. From such observations, it would appear that needle ice was not a significant initiator of sorted stripes at the Mafadi site. Although the processes could not be determined quantitatively at the site, it is possible that diurnal freeze-thaw cycles within the upper 4 cm of surficial material may have induced frost heave and thereby contributed towards particle movement and sorting. Thermal creep, caused by frequent alternating volumetric expansion and contraction following significant daily temperature fluctuations, is another possible process. For instance, during June 1994, the surface temperature fluctuated from + 25°C to - 16°C within a 24 hour period at the site.

4.5 LARGE SORTED CIRCLES ON MAFADI SUMMIT

Despite some references and photographic evidence of miniature sorted patterned ground in the high Drakensberg (Troll, 1944; Hastenrath, 1972; Hastenrath and Wilkinson, 1973; Hanvey and Marker, 1992), very little information pertaining to larger sorted patterns is available. In fact, the absence of photographic evidence and qualitative or quantitative data has cast much doubt on whether larger sorted patterns actually occur in southern Africa.

Although Hanvey and Marker (1992) report polygonal patterns from the vicinity of Tlaeeng Pass (3275 m a.s.l.) which have a maximum diameter of "1 metre or less" (p357), these are said to occur on a "thin soil cover on bedrock" (p357) and are considered as "small scale" (p357). Harper (1969) reported "stone rings and polygons" (p89) up to 6 m in diameter from the Thabana-Ntlenyana Massif above 3354 m a.s.l. Harper further suggests that they have formed during "recent times" (p89). The Thabana-Ntlenyana Massif was surveyed but no evidence of large sorted patterns, let alone patterns 6 m in diameter, which would be an exceptionally sized feature in any environment, was found. Reports of patterns up to 6 m in diameter without quantitative data (e.g. sorting characteristics) or photographic evidence should be met with scepticism until better descriptions and stronger supporting data are presented.

Dardis and Granger (1986) reported 40 to 50 "sorted stone circles" from the Champagne Castle region (3000 to 3200 m a.s.l.) and they believe these to be active. The patterns are said to have diameters varying from 0.5 to 1.3 m. The photographs supplied by Dardis and Granger (1986) show stones lying on top of the soil surface. From the photographs, it appears that the patterns are random associations that give the appearance of sorted forms. Similar features were identified in the Sani Pass area, and it can be seen in Figure 4.17 that the stones have been displaced to form a circular pattern shape, yet these patterns show no vertical sorting.

The only large sorted circular patterns found in the study area were on the Mafadi Summit and are, as yet, the only ones known to the present author. Some empirical data were obtained for the Mafadi circles and are presented in the following section (**4.5.1**). A detailed analysis is presented for one group of patterns, namely those referred to as "Site 1" in section **4.5.1**. The patterns at Site 2 were only discovered in July 1994 and so insufficient time was available for a more detailed study. The author attempted to reach the site during three subsequent winters but during each of these, the Mafadi Summit region was snowbound.

Figure 4.17 Displaced stones forming a random patterning.

4.5.1 Site 1

4.5.1.1 Location

The large sorted circles are situated on a broad, gentle (2°), north-facing (350°) interfluve at an altitude of 3420 m, near the Mafadi Summit (3450 m a.s.l.) (Figures 2.2, 2.5 and 4.12). The broad summit area consists mostly of a stony mantle with occasional bedrock outcrops (Figure 4.12). At one site, perpendicular to the interfluve, is a linear block accumulation about 60 m long and 8 to 10 m wide. It is within this block accumulation that the sorted circles have developed. The occurrence of a linear concentration of large rocks (a-axis varies between 17 and 64 cm) across an interfluve implies that the block accumulation is structurally controlled. It is possible that this localized concentration of large rocks is as a result of the high bedrock joint concentration, consequently facilitating enhanced weathering. The relatively deep (30 to 35 cm) unconsolidated clastic surface is underlain by a poorly sorted diamiction of unknown depth. Wilson and Sellier (1995) have also reported sorted patterned ground within blockfields underlain by fine material, which is said to be associated with well-jointed rocks that are susceptible to macrogelivation. However, the area adjacent to the Mafadi block accumulation has a weathered surface profile of only 5 to 10 cm. Similar concentrations of large rocks (blockfields) were only found on the Mafadi southwest face. Because the surface gradients of most blockfields, including those on Lithotobolong and Njesuthi, are relatively steep (over 5°), such sites are more conducive to stone-banked lobe and blockstream development (See Chapter **6**).

4.5.1.2 Characteristics

Ten well-defined sorted circles and several less-defined patterns occur in the block accumulation near the Mafadi Summit. Such sorted circles, found amid blocks or boulders, have also been referred to as debris islands (Washburn, 1979). Pattern occurrence is restricted by the small block accumulation surface area and are best developed in the central, widest area of the block mass. Owing to the scarcity of well-defined patterns, all of the ten surface patterns were measured and two were sectioned.

Most of the patterns are slightly oval in appearance and measurements show that all ten patterns are elongated downslope (Table 4.6). The mean downslope elongation is 1: 1 31 on the 2° slope gradient (Table 4.6). Washburn (1989, p942) categorizes patterns that are irregular in plan and only roughly approximate a circular outline as "sorted circles". Although the Mafadi patterns are somewhat elongate, they do nevertheless appear roughly circular and therefore are referred to as "sorted circles". Pattern centre diameters range from 31 to 143.5 cm and average approximately 74 cm (Table 4.6). Ballantyne and Matthews (1982) examined sorted patterns which are morphologically similar to those at Mafadi Summit and found that 41.4% have diameters between 20 and 60 cm. Fifty percent of the Mafadi Summit patterns have centre diameters between 20 and 60 cm, 20% between 60 and 80 cm and 30% exceeding 120 cm. A positive correlation (r_s = 0.82; significant at the 0.01 level) is found between the pattern centre diameters and the pattern centre height above the adjacent borders (Table 4.6 and Figure 4.18). The pattern centre height above the adjacent borders varies from 11.7 cm (centre diam. = 31 cm) to 41.8 cm (centre diam. = 143.5 cm), but averages 23.5 cm (Table 4.6 and Figure 4.18). These pattern centre elevations are apparently not extreme, as Ballantyne and Matthews (1982) found a mean elevation of 40 cm for patterns similar in diameter to those at the Mafadi Summit.

Pattern	Diameter (cm)			As/Ds	Centre height above border (cm)	Mean Clast Size (cm)			Vegetation in pattern centre
	As	Ds	Mean			Centre	Intermediate border	Primary border	
1	27	35	31	1: 1·30	11.7	2.8	8.4	19.7	*lichen*
2	37	42	39.5	1: 1·13	20.2	2.9	7.0	27.1	*lichen*
3	105	141	123	1: 1·34	32.5	5.2		23.7	*lichen*
4	33	51	42	1: 1·54	18.5	3.5		23.5	*lichen*
5	70	78	74	1: 1·11	16.2	3.8	10.0	30.2	*lichen / heath*
6	66	88	77	1: 1·33	30.2	3.9	10.8	29.3	*lichen*
7	114	131	122.5	1: 1·15	30.0	5.0		29.3	*lichen / grass*
8	108	179	143.5	1: 1·66	41.8	6.9		30.8	*lichen*
9	33	40	36.5	1: 1·21	17.0	2.9		27.8	*lichen*
10	42	55	48.5	1: 1·31	17.0	3.6		26.8	*lichen*
Mean	63.5	84	73.8	1: 1·31	23.5	4.1	9.1	26.82	
Std. Dev.	32.65	47.5	39.64	0.17	9.03	1.23	1.46	3.35	

As = Across-slope
Ds = Downslope

Table 4.6 Dimensions and characteristics for large sorted circles from the Mafadi Summit.

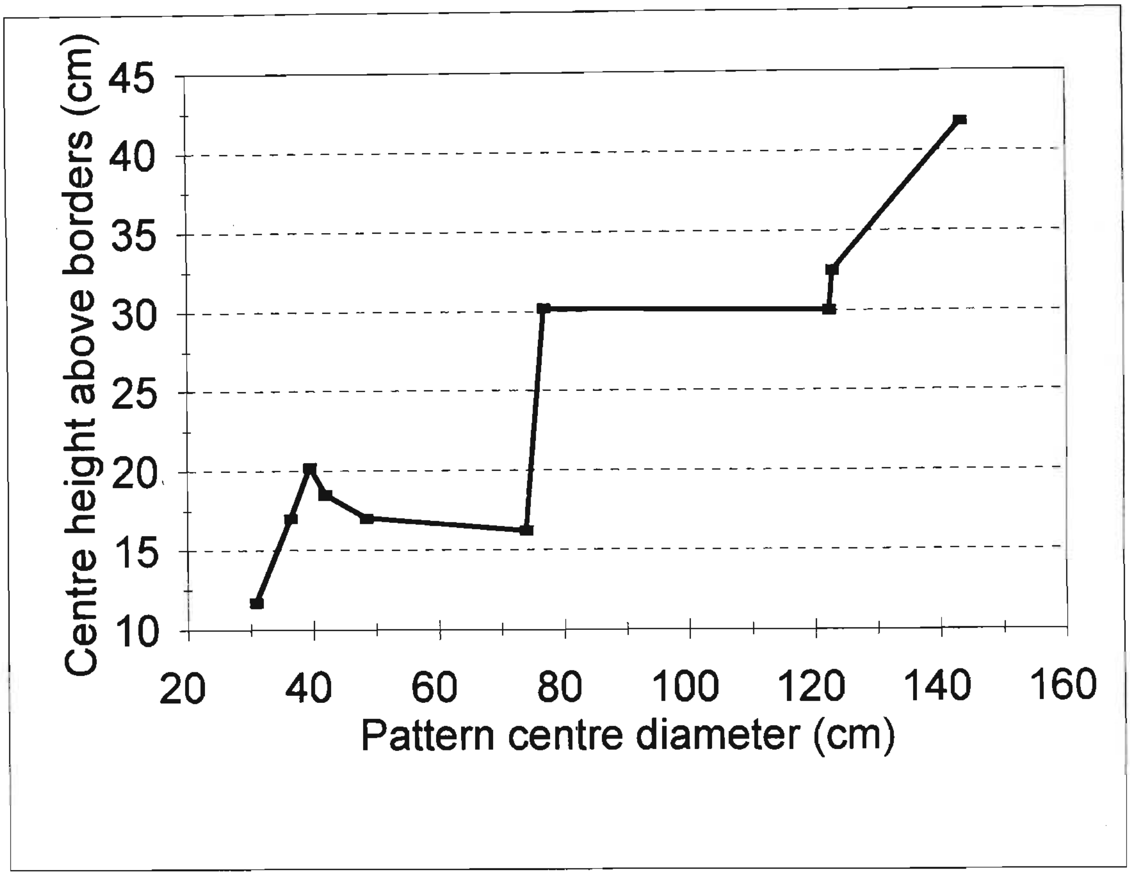


Figure 4.18 Graph showing the relationship between the diameter of sorted circle pattern centres with the height of centres above borders. RSQ = 0.82, indicating a positive correlation, significant at the 0.01 level.

The pattern centre surfaces are occupied by sub-rounded clasts averaging 4 cm in diameter (a-axis). A positive correlation (r_s = 0.91; significant at the 0.01 level) is found between the mean clast size (at the pattern centre) and centre diameter. Mean clast size varies from 2.8 cm (centre diam. = 31 cm) to 6.9 cm (centre diam. = 143.5 cm) (Table 4.6 and Figure 4.19). The centres are encircled by primary and, in some cases, secondary borders which are absent of fines at the ground surface (Figure 4.20). Although there is a considerable increase in mean clast size between centres and secondary borders (from 4 to 9 cm), a much more pronounced increase occurs between secondary and primary borders (from 9 to 26.8 cm) (Table 4.6 and Figure 4.19). The primary borders consist of large rocks with Krumbein Roundness (1941) values of between 3 and 4, and average approximately 17 kg (max. = about 58 kg) in mass. Most of the larger rocks at pattern borders dip towards the centres (Figures 4.21 and 4.22),

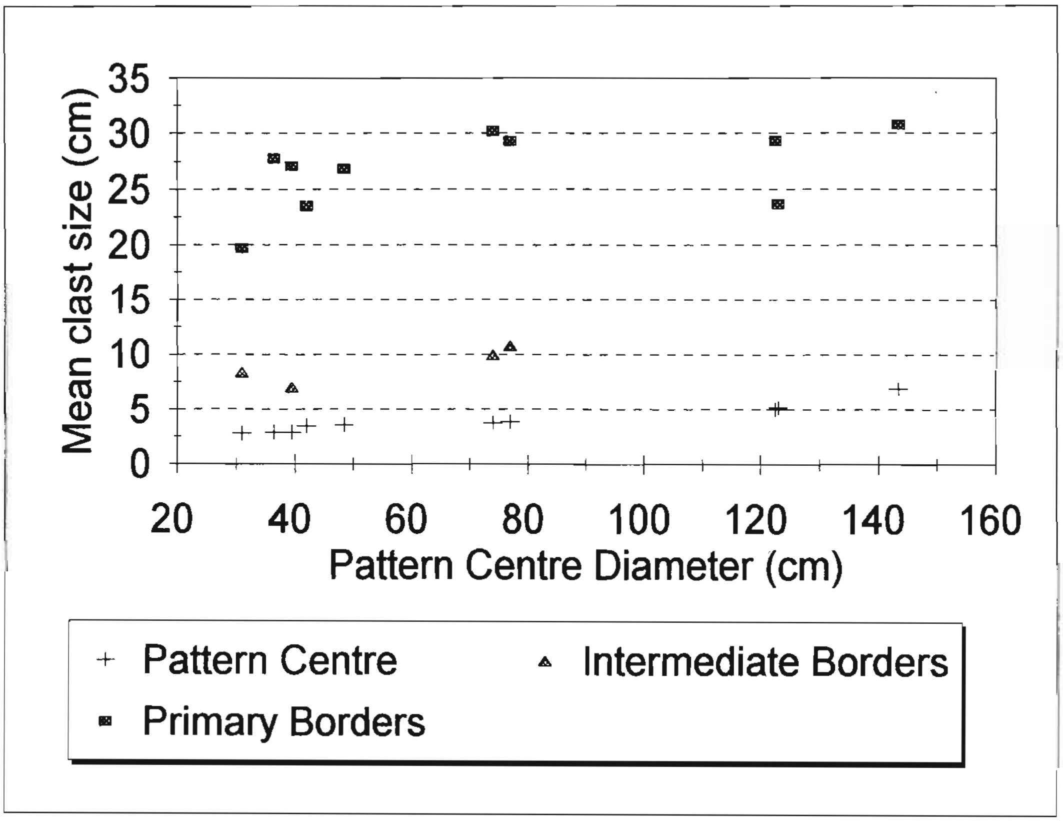


Figure 4.19 Plot of mean clast size at circle centres, intermediate borders and primary borders against pattern diameters. For the pattern centre, RSQ = 0.911, indicating a positive correlation, significant at the 0.01 level. For pattern primary borders, RSQ = 0.217, indicating no correlation.

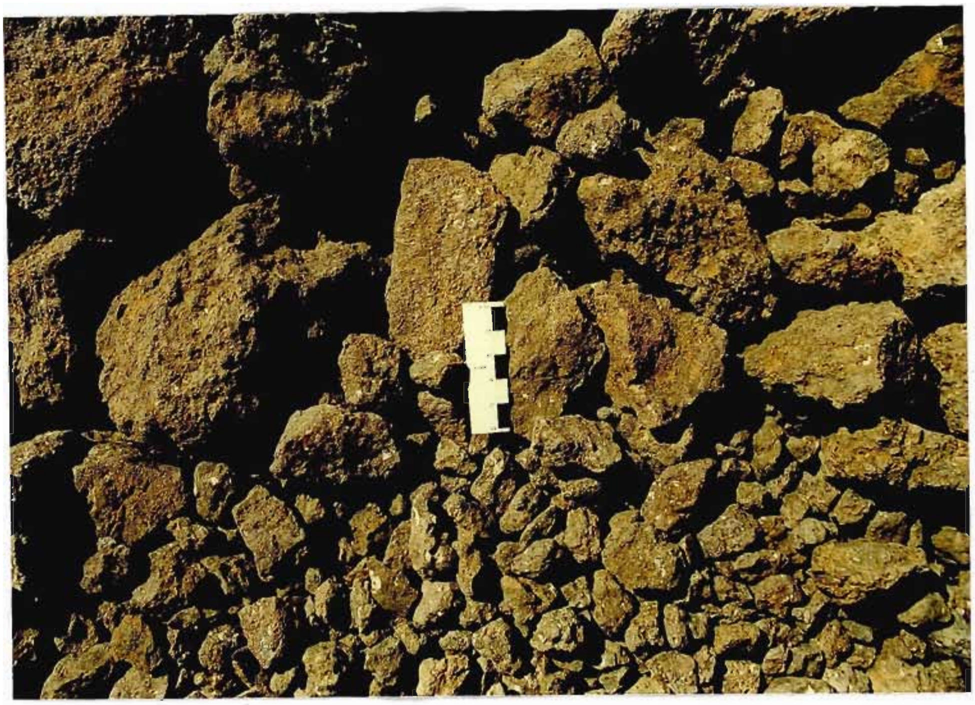


Figure 4.20 Section through a sorted circle centre and its secondary border. The bar-scale is 10 cm in length.

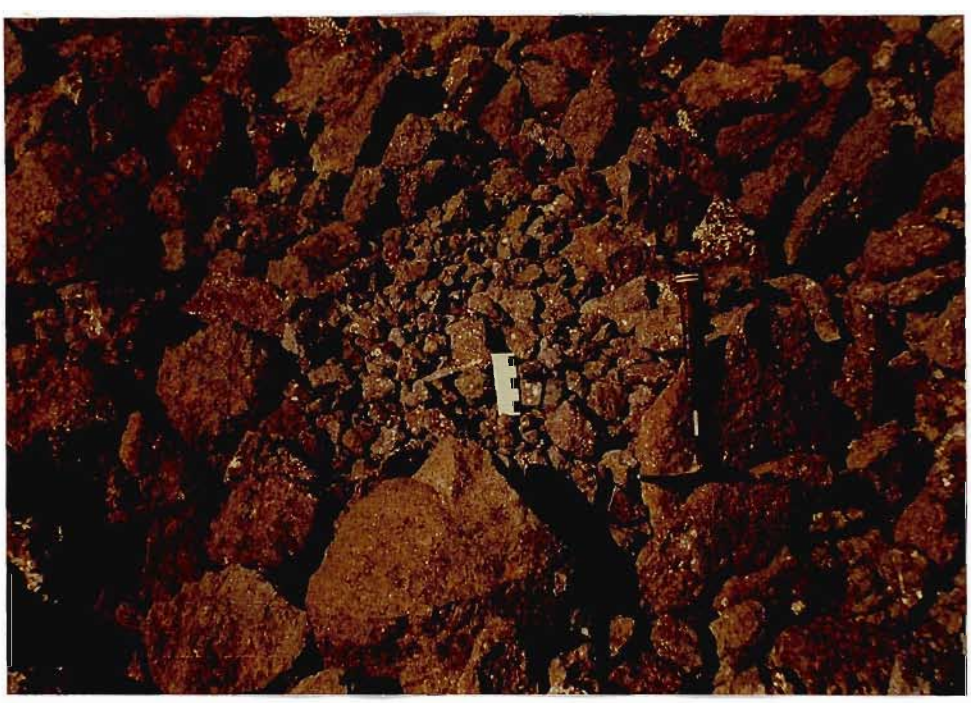


Figure 4.21 A large sorted circle, showing many rocks at the pattern border dipping towards centres.

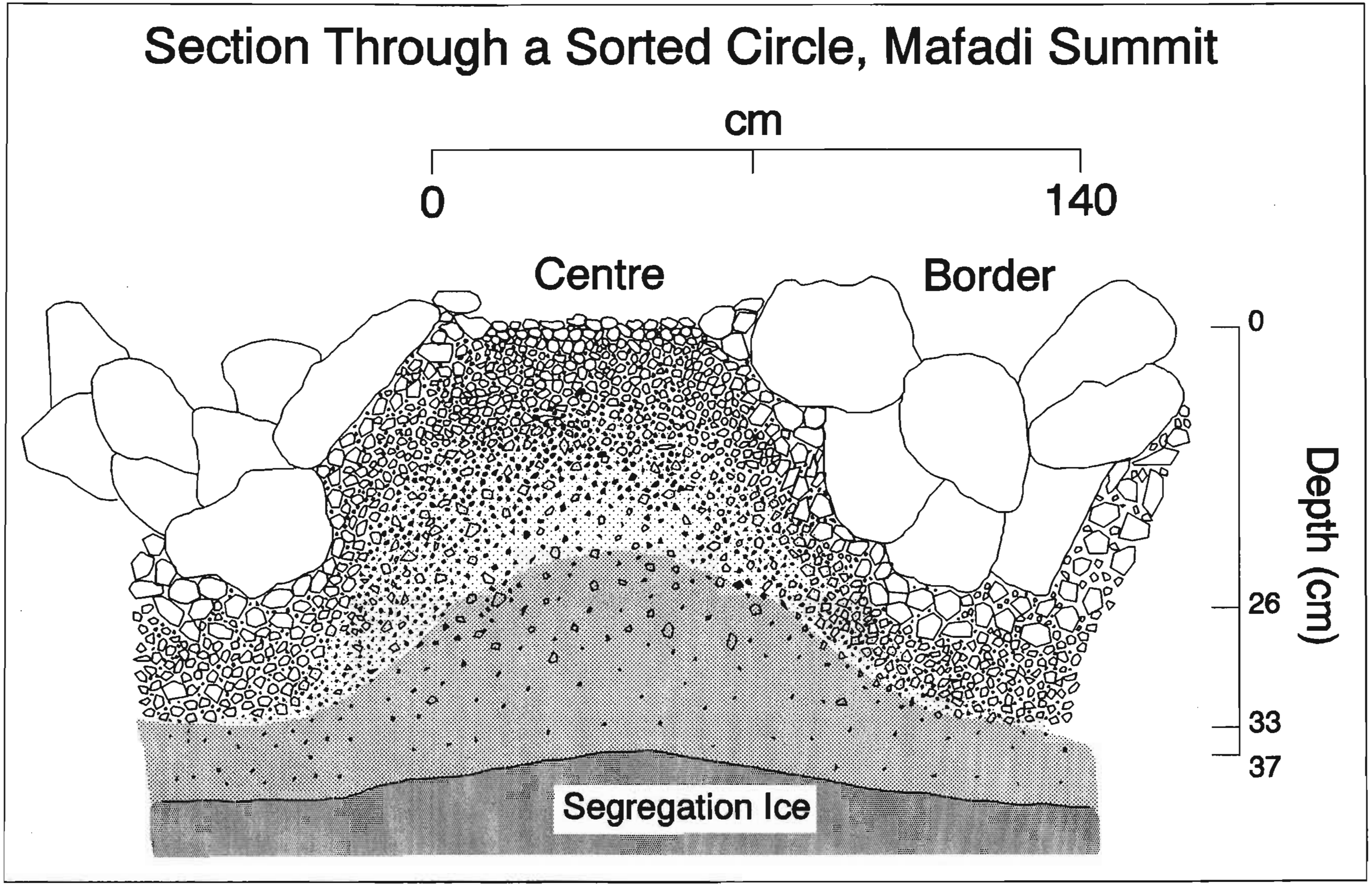


Figure 4.22 Vertical section through a large circle from the Mafadi Summit.

as has also been found for circular patterns in Nova Scotia, Canada (Mooers and Glaser, 1989). Although the primary border clast sizes do vary somewhat between patterns, there is no correlation (r_s = 0.22) with pattern centre diameter (Figure 4.19).

The size and shape of clasts from a pattern centre were compared against those of the surrounding stone field. Figure 4.23 shows that clasts from the pattern centre are significantly more rounded and less flat than clasts from the surrounding stone field. Although clasts were found to be similar in size (Figure 4.23) and of the same population (the computed value of U [83] was greater than the critical value [72] at the 0.05 significance level), analysis of their roundness indicates separate populations (the computed value of U [21] was less than the critical value [72] at the 0.05 significance level).

Sparse lichen growth is found on many clasts in the circle centres and borders. While all ten patterns show the presence of lichen on clasts, only two patterns show other types of vegetation, namely heath and grass (Table 4.6). The heath and grass favour the central perimeters and *Helichrysum* was observed growing from within clastic borders (Figure 4.24).

Particle size and sorting characteristics

Sections through the circles reveal that clastic borders occupy trough-like depressions to a depth of 26 cm, below which are cobbles and coarse gravels to a depth of 33 cm. From 33 cm is an unsorted diamiction, that was frozen (June 1994) below 40 cm depth at borders but somewhat heaved towards pattern centres (Figure 4.22). Observations show that finer material at depth appears to have been squeezed or heaved towards pattern centres (Figure 4.22). Particle size distribution was determined with depth below a pattern centre (Figures 4.25 and 4.26). Gravels and cobbles predominate to approximately 28 cm depth, but gradually incorporate very small quantities of sand and silt/clay from 8 cm depth. It is only below 30 cm that frost susceptible material occurs, and the percentage silt/clay rapidly increases to 36.9% at 37 cm depth (Figure 4.25).

Vertical variations in the coarse fraction size of material were determined to assess subsurface sorting of gravels through a pattern centre (Figure 4.26). Gravels at the pattern centre surface

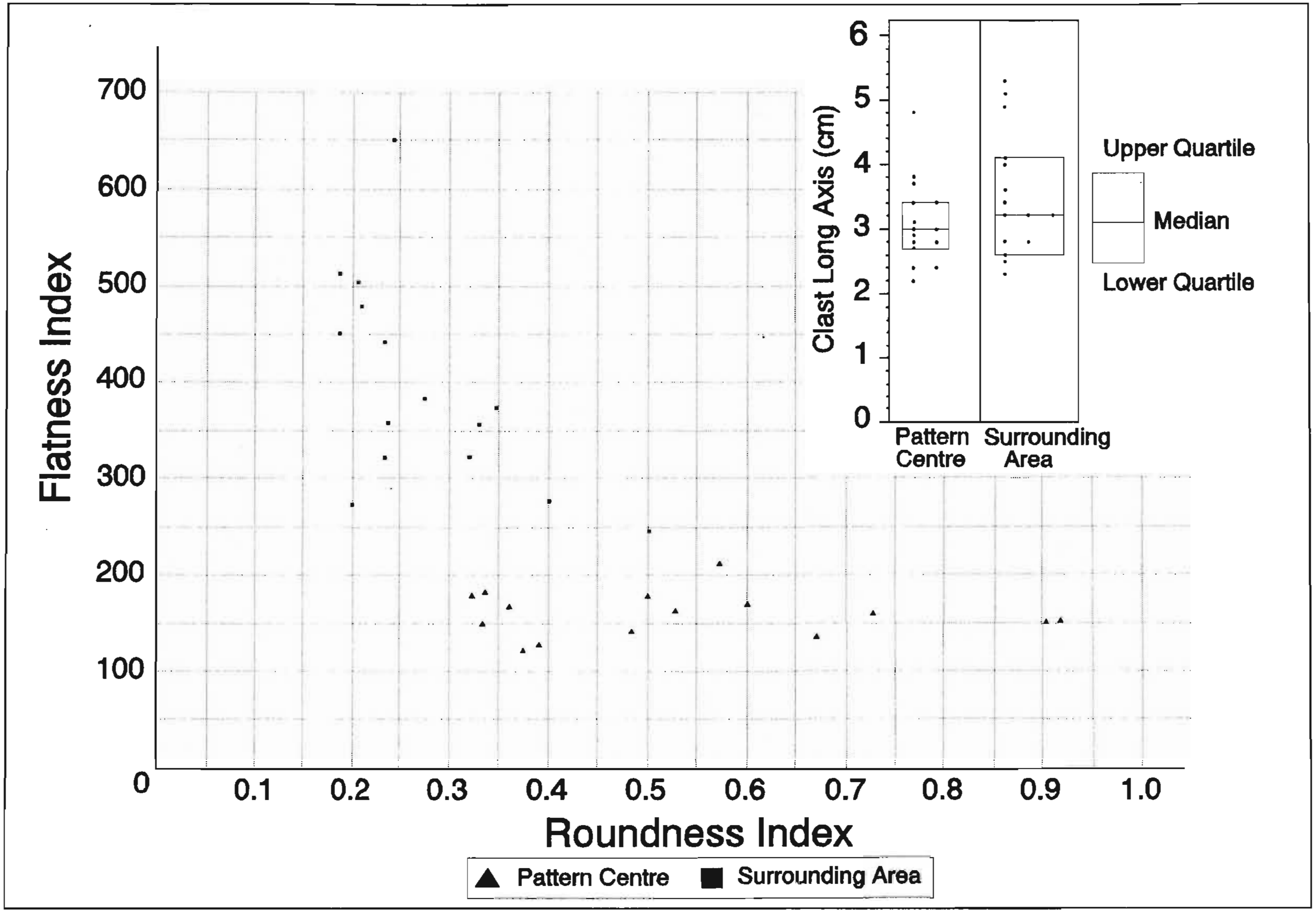


Figure 4.23 Shape and size of clasts within a large sorted circle centre and on the surrounding stone field at Mafadi Summit (n = 15 per sample).

Figure 4.24 Vegetation growing on large sorted circles (as indicated by arrows) includes grass at the central peripheries and *Helichrysum* within the clastic borders. The bar-scale is 10 cm in length.

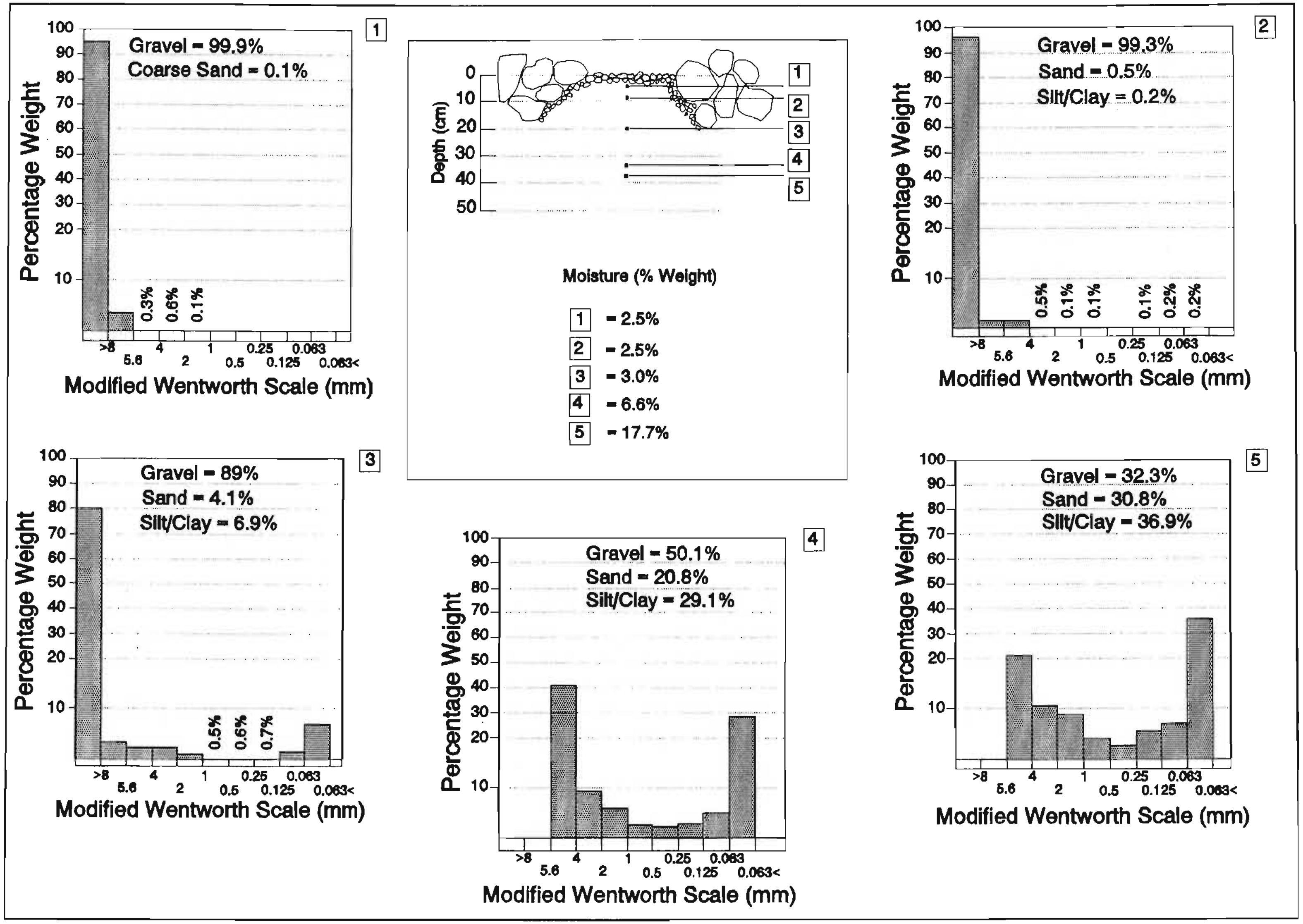


Figure 4.25 Particle size and moisture changes with depth below a large sorted circle centre at Mafadi Summit (measurements were taken in July 1994).

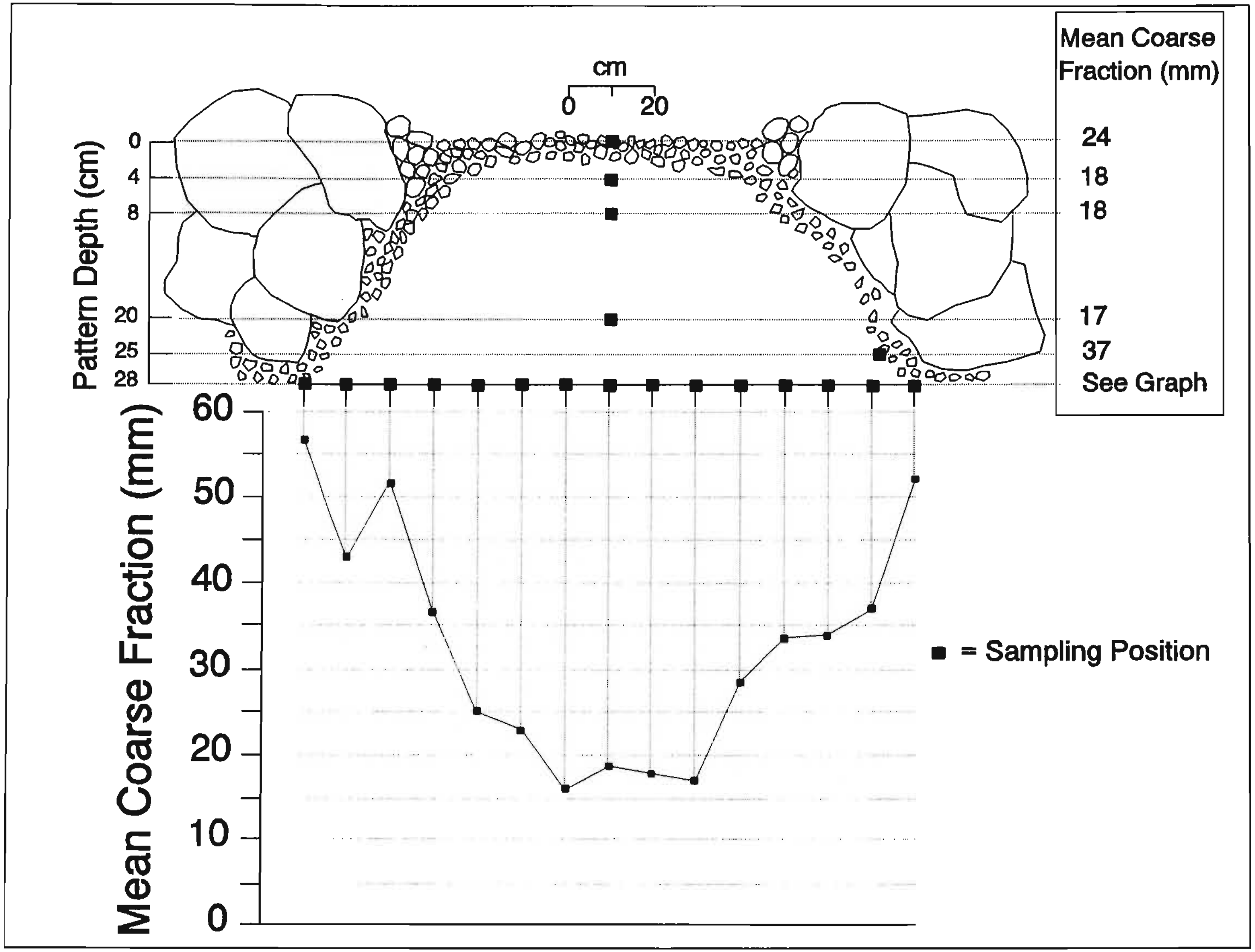


Figure 4.26 Mean course fraction size variations within a large sorted circle at Mafadi Summit (n = 10 per sampling position).

are considerably larger (mean = 24 mm) than those measured at 4 cm depth (mean = 18 mm) (Figure 4.26). The lateral variation in mean coarse fraction size was determined at 28 cm below the pattern centre (Figure 4.26). At this depth, coarse fractions directly below the pattern midpoint average between 15 and 20 mm, but increase laterally towards borders where they average between 50 and 60 mm. From such measurements and observations it is apparent that fine material has been moved towards centres to approximately 20 to 25 cm below the surface.

Moisture

Measurements to determine the percentage moisture content were taken in July 1994, during the dry season, and indicated that most of the pattern material had already desiccated. The percentage moisture by weight for the gravels at 4 and 8 cm depth was 2.5%, and only increased to 3% at 20 cm depth (Figure 4.25). The increase in the percentage fines below 30 cm coincides with a rapid increase in moisture. For instance, the percentage moisture increases from 6.6% at 33 cm depth to 17.7% at 37 cm depth (Figure 4.25). Although segregation ice was encountered from 37 cm below surface, the true thickness of the frozen layer could not be determined.

4.5.2 Site 2

Three large sorted circles are located on the Mafadi east-northeast face (77°) at an altitude of 3410 m. The circles are situated on a gently-sloping (3°) rock bench, approximately 10 to 15 m wide. The pattern distribution and numbers appear to be restricted by the width of the rock bench. The slope gradient on either side of the rock bench is ≥7° and has shallow soils, thereby restricting the development of large patterns. The mean pattern centre diameters are 74.5, 94 and 134 cm, all showing slight elongation down the 3° terrace (Table 4.7). The wide borders which average 78.3, 82.5 and 123.5 cm, give the patterns a particularly large dimension (Table 4.7 and Figure 4.27). The circles are occupied by secondary and somewhat wider primary borders (Table 4.7 and Figure 4.28). The largest circle has a maximum downslope diameter (centre + border) of 4.0 m and across-slope diameter of 3.62 m.

Terrace Gradient = 3°		Depth of Sorting = ~ 50 cm							
Slope Aspect = 77°									
Altitude = 3410 m a.s.l.									
Pattern	**Centre Diameter (cm)**			**Border Diameter (cm)**					Secondary border :
	Across-Terrace	Down-Terrace	Mean	Left	Right	Upslope	Down-slope	Mean	Primary Border
1		104	94	91	64	96	62	78.3	1: 2
2	71	78	74.5	88	74	107	61	82.5	1: 3
3	107	161	134	84	144	122	144	123.5	1: 1·4

Table 4.7 Dimensions and characteristics for large sorted circles at Mafadi Summit, Site 2.

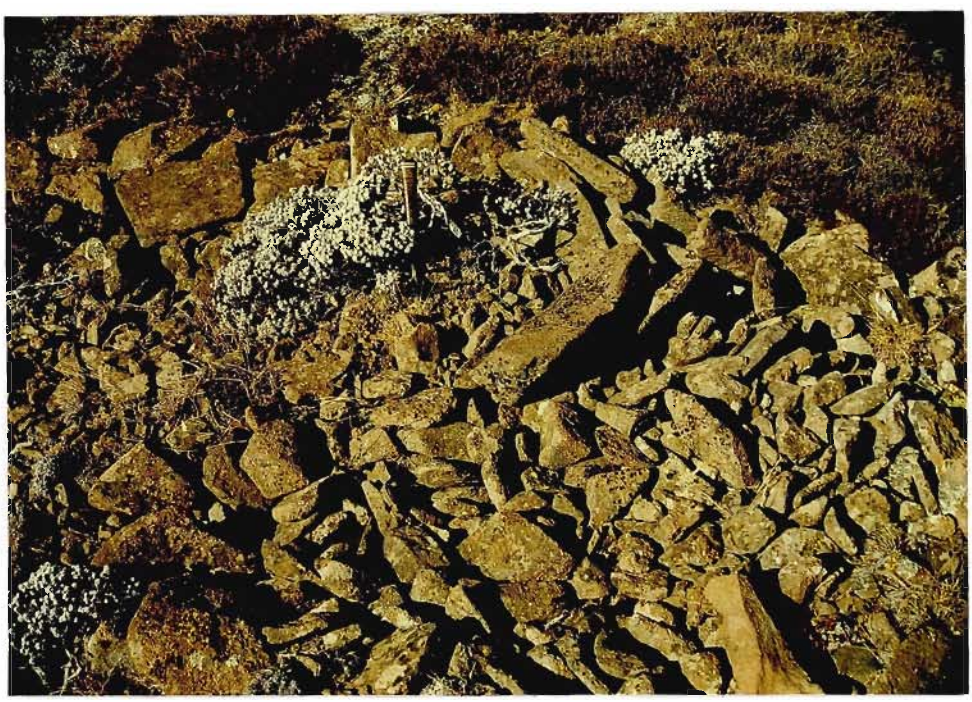


Figure 4.27 A large sorted circle at Mafadi Summit, Site 2. The wide borders give the patterns a particularly large dimension.

Figure 4.28 A large sorted circle at Mafadi Summit, Site 2. The circle is occupied by a secondary and somewhat wider primary border.

The clast size values for a pattern centre and its border are given in Figure 4.29. The central cell contains soil with small pebbles (mean pebble size = 2.3 cm), while the surrounding secondary border has a mean clast size of 7.5 cm and the primary border 22.0 cm (Figure 4.29). Many of the plate-like blocks at the primary borders appear to have their a-axis orientated directly towards pattern centres (Figure 4.28). A section through the pattern gave a visual indication that sorting occurs both laterally and vertically to a depth of about 0.5 m.

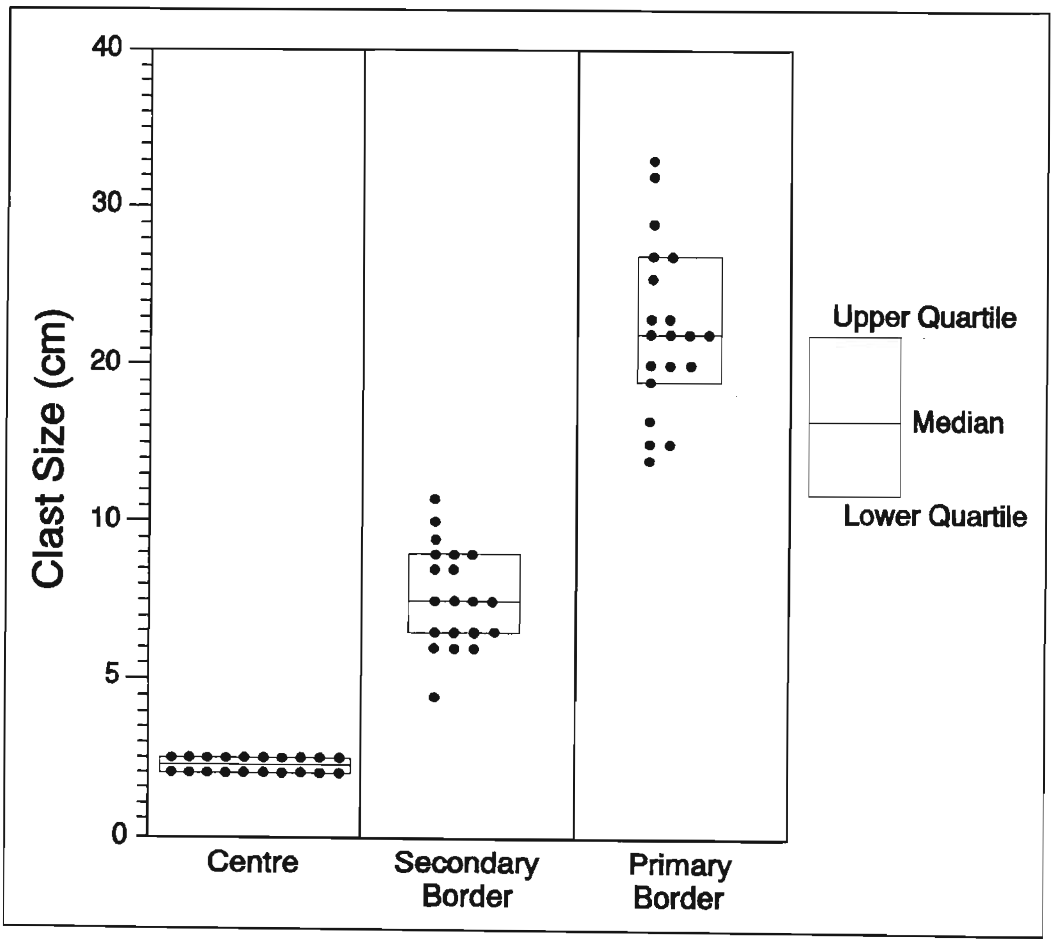


Figure 4.29 Clast size variations for large sorted circle centres, secondary borders and primary borders, Mafadi Summit, Site 2.

4.5.3 Discussion

As no rigorous analysis has been undertaken for the sorted circles at Site 2, only general comments can be made on these patterns. The circles at Site 2 show vertical and lateral sorting with raised cell areas and large troughs filled with plate-like blocks. Although the circles have independent borders, they are in close proximity (3 metres) to each other. These pattern characteristics could suggest that the circles have a periglacial origin (Goldthwait, 1976). The circle centres have a dense vegetation cover (Figure 4.28) and the clasts at borders have lichen growing on them. Plants such as *Helichrysum* and moss are found growing within the borders. Given such vigorous vegetation growth within the circles, it would appear that these large sorted patterns are inactive under present environmental conditions. Further discussion in this section shall pertain to the large sorted circles at Site 1.

4.5.3.1 Pattern Origin

Several pattern characteristics indicate a periglacial origin for the sorted circles. The vertical and lateral particle size sorting, occurrence of patterns in groups, elevated centres and trough-shaped border depressions are observed attributes which have been ascribed to a frost origin (e.g. Goldthwait, 1976; Thorn, 1976; Ballantyne and Matthews, 1982; Wilson and Clark, 1991). As mentioned in Section **4.3.2**, the predominantly circular patterns cannot be explained by cracking (Krantz, 1990), therefore, alternative processes should be considered. Equally, the patterns are unlikely to be a product of present-day diurnal frost cycles, even though it has been argued that a high frequency of freeze-thaw cycles may rapidly develop frost related features (Black, 1976; Gleason *et al.*, 1986). The large size of the patterns, the large mass of some sorted clasts and the absence of moisture and frost susceptible material near the ground surface would not favour a diurnal freeze-thaw origin. Seasonal ice at 37 cm depth may be an indication that seasonal freezing and thawing is a more likely process than diurnal freeze-thaw.

Van Vliet-Lanoë (1988) reports that in situations where dry unconsolidated gravels, at or near the surface, are underlain by frost susceptible material, silts may easily protrude through gravels due to differential frost heave. Segregation ice (ice lensing) is common in many periglacial environments, causing such deformations or cryoturbations (Van Vliet-Lanoë, 1988). Segregation ice was found at depths below 36 cm and must have contributed towards limited heave, thereby forcing finer materials into the less saturated gravels to about 20 cm below surface.

Soil convection hypotheses have been widely favoured to explain the origin of sorted circles and polygons (e.g. Ray *et al.*, 1983; Hallet and Prestrud, 1986; Gleason *et al.*, 1986, 1988; Hallet, 1987, Hallet *et al.*, 1988; Anderson, 1988; Krantz, 1990). Convection, initiated by moisture-induced density gradients, may be created during heaving (Hallet and Prestrud, 1986; Hallet, 1987; Hallet *et al.*, 1988), particularly in a situation where coarse, dry materials overlie saturated, frost susceptible material (French, 1976). Should the process of moisture-induced free convection be well developed, then it would be expected that finer sediments are moved upwards into the drier, porous overlying material, while larger rocks collapse (Thorn, 1976) or move downward (French, 1976) and concentrate over ice troughs (Gleason *et al.*, 1986). The pattern morphology of injected fines below centres, primary-border clasts dipping towards centres, and depressions ("gutters") within borders (Figures 4.22 and 4.30), could have originated from a moisture-induced free convection process during the time of pattern initiation (Figure 4.31). Present moisture characteristics have shown desiccated gravels underlain by frozen, frost susceptible material. The frozen soil has also formed slight corrugations at depth, with crests below pattern centres and troughs below borders (Figure 4.22). These characteristics may indicate some limited convective activity still present at depth.

The Rayleigh free convection model is similarly used to describe pattern formation but requires cooler buoyant water near the surface to descend and become exchanged with rising warmer, less dense water (Gleason *et al.*, 1986, 1988; Krantz, 1990). This may be possible during the thawing of frozen ground because, contrary to most liquids, water has a density maximum 4K above its freezing point, owing to the formation of hydrogen bonds which store the molecules in a more tightly-packed configuration (Krantz, 1990). As moisture is absent near pattern surfaces during winter and during the early spring thaw, Rayleigh free convection would be

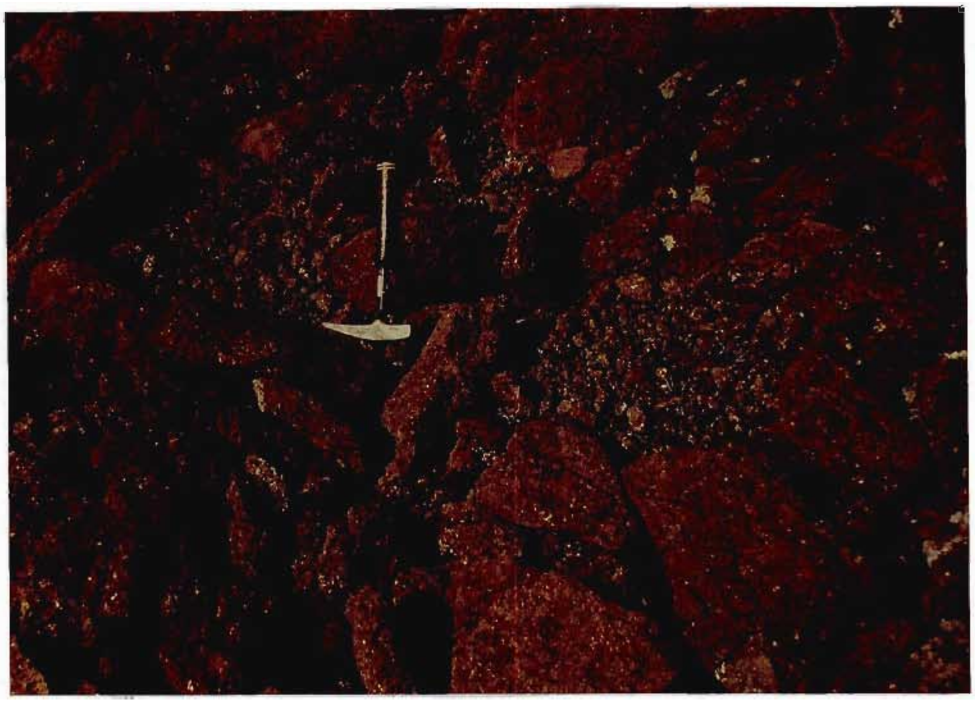


Figure 4.30 Large sorted circles at Mafadi Summit, which are characterized by : (a) depressions ("gutters") within the rock-dominated borders and (b) somewhat raised centres.

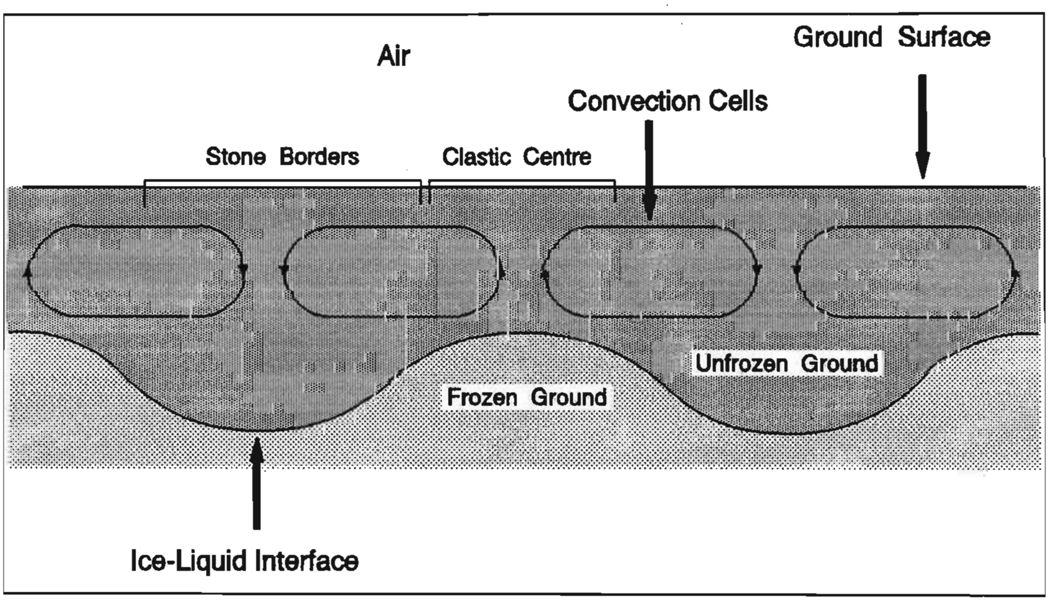


Figure 4.31 One possible hypothesis for the development of large sorted circles at Mafadi Summit is the convection cell model (after Krantz, 1990, p122).

unable to develop at the Mafadi site under present environmental conditions. The precise mechanism(s) that initiated these sorted circles is/are difficult to determine but it would appear that some soil convection and/or cryoturbation and/or diapiric processes may have been operative during a wetter and/or cooler period.

4.5.3.2 Pattern activity : present and past

The presence of seasonal freezing and fine material injected into pattern centres at depth indicates that some cryogenic activity is still operative today. Air and, more pertinently, ground temperature conditions, are sufficiently low during winter months to enable seasonal freezing to depths of over 37 cm on Mafadi Summit. Although the present material at such depth is frost susceptible, it is the absence of moisture during the freezing period which appears to prevent more intense frost action. The 17% soil moisture content at 37 cm depth was measured during the exceptionally wet winter of 1994. It is possible that in the normally dry winter years, the percentage moisture would be less than that measured in 1994, and consequently the potential for cryogenic activity would be lower.

A number of pattern characteristics suggest relative stability under present environmental conditions. According to Ballantyne and Matthews (1982), the presence of lichen and plant growth within patterns indicates relatively stable conditions. Lichen growth is found on many surface clasts of the Mafadi sorted circles, although the lichen are usually small in radius. Rapid desiccation owing to insolation and wind are likely to prevent more vigorous lichen colonization and growth. The presence of lichen is, however, not unequivocal evidence for stable conditions as lichen have been reported growing on an active medial moraine at Storbreen (Matthews, 1973; Griffey, 1978). *Helichrysum* and grass were found growing around the periphery of two circle centres and from within several pattern borders (Figure 4.24). Ballantyne and Matthews (1982) also observed vascular plants growing around the outer parts of pattern centres and attribute this to less cryoturbation at pattern centre peripheries, than within borders. The overwhelming predominance of gravels and cobbles within centres to a depth of 20 cm is unsuitable for vegetation establishment at the Mafadi Summit patterns. Vegetation establishment is more likely around centre peripheries or within borders where there is proximity to finer, wetter soils at depth. Larger clasts at pattern borders also offer greater protection against insolation, rapid desiccation and wind action, and may therefore contribute towards favourable micro-environments for vegetation establishment and survival. The presence of vegetation does, nevertheless, indicate relatively stable conditions.

Weathering rind thicknesses for two large border clasts were examined. Although measurements are limited and therefore not conclusive, they do show that the exposed clast surfaces have thicker weathering rinds (5 to 12 mm) than the submerged surfaces (1 to 3 mm). The exposed surfaces of such clasts are also discoloured and so it would appear that larger clasts have undergone little movement for a considerable time.

The general absence of sand and fines to at least 20 cm depth is some indication that the injection of fine material into centres is presently impeded. Further, any fine material that may have accumulated near the surface during a more active period, would have been removed by deflation and rain wash processes over the years. Such sorted circle centres often show slight updoming when active (Nicholson, 1976), yet the Mafadi pattern centres showed no indication of even slight heaving, even during mid-winter when frost processes are most operative at sites where moisture is available in the high Drakensberg.

As mentioned, most clasts at the pattern centre surfaces are sub-rounded. A similar finding was made by Ballantyne and Matthews (1982) who attributed such shapes to an origin by traction at the glacier sole. As there is no evidence for recent or past glaciation at the Mafadi Summit, such an argument cannot be applied here. The sub-rounding of the surficial clasts appears to be a function of weathering over a prolonged period, possibly a product of microgelivation as described by Ballantyne (1987a). These gravels and cobbles are not subjected to transportation and redeposition, owing to the surrounding borders functioning as an obstruction. Similarly, the clasts are too large to be washed into the gravels below. The clasts, therefore, appear to be held in a closed system. The pattern centres are subject to possible ponding during heavy rainfall events and are exposed to insolation for much of the day, conceivably enhancing chemical weathering processes at the site. Repeated rain-splash and rain-wash processes within the closed pattern centre systems may also contribute to particle rounding over a prolonged period. Wind abrasion could also contribute to the particle rounding at the site, particularly as extremely strong and frequent winds blow for several months from May to September (personal observations). Bedrock outcrops at Mafadi Summit (3450 m a.s.l.) are smooth and well rounded, and are possibly a product of wind abrasion. In contrast, the more angular clasts on the surrounding stone field are subjected to erosion, transportation and redeposition, and so are not left exposed at one site for as prolonged a period as those within the sorted circle centres. The particles at pattern centre surfaces are significantly larger (mean = 24 mm) than those at 4 cm depth (mean = 18 mm) (Figure 4.26). Therefore, the presence of such large sub-rounded particles within pattern centres may indicate that the clasts have remained *in situ* for a prolonged period, which could serve as further evidence for current pattern stability.

Thorn (1976) and Walters (1993) have suggested that large sorted circles may go through a cycle of growth, stillstand and eventual decay. Such a cycle of sorted circle evolution and decay may be considered for the Mafadi circles. It is envisaged that the large sorted circles formed during a cooler, wetter period, as moisture is considered to be an important control in the development of large sorted patterned ground (Goldthwait, 1976; Hallet and Prestrud, 1986; Krantz *et al.*, 1988; Guangpan and Min, 1993). During this period of circle growth there would have been horizontal and vertical pattern extension, with material at depth being moved up towards the surface within pattern centres. Possible climatic warming and/or increased seasonal desiccation may have caused the movement of material into centres to become less active, allowing for the accumulation of gravels below centres and thereby offering greater insulation, and consequently, restricting convection or cryoturbation (Thorn, 1976). Over time, finer material near the pattern centre surfaces may have been removed by deflation and rainwash. With the removal of fines from centres, larger voids are created and so desiccation is enhanced. Therefore, any weak intrusion at depth would generate some pressure upon the air held within the capillary soil spaces, thereby restricting pressure on the gravels above (French, 1976). Further, as the depth to ground water increases as a response to climatic/seasonal/local drying, the magnitude of frost heaving in soils would be expected to decrease (Hansen and Guoqing, 1993). Walters (1993) has suggested that sorted circles become smaller as stones occupy cells and fines are removed, eventually progressing to an end point where no fines exist on the surface. From detailed observations and measurements it appears that the large sorted circles at Mafadi are approaching/have approached such an end point. The presence of seasonal freezing and some limited sorting at depth may help preserve the present circle morphology. However, should these processes become inoperative over a long time period, it is likely that the sorted circle morphology will eventually decay.

4.6 FURTHER IMPLICATIONS OF SORTED PATTERNED GROUND IN THE HIGH DRAKENSBERG

4.6.1 Present Environment

The absence of moisture during winter prevents widespread pattern development. Sites containing moisture throughout the year are usually found adjacent to rivers and streams where the microtopography is often steep, rendering such sites more suitable for striped patterned ground. A further limiting factor for the development of miniature sorted patterns is the absence of fines along such drainage zones. Site-specific factors such as surface gradient and soil composition, depth and moisture are the primary controls for the development of miniature sorted patterns in this region. Such a finding may suggest that the near-surface soil freezing and thawing characteristics in the high Drakensberg alpine belt do not undergo significant change with altitude during the winter months. This could be attributed to a low lapse rate of only 3°C per kilometer in the alpine belt (Grab, in press). The size of miniature sorted patterns in the high Drakensberg is more likely to be determined by the localized soil moisture and particle size characteristics and the duration of pattern development and/or preservation. Marques *et al.* (1990) have described some frost-related features such as "ice cementation, pipkrake and high laminar porosity and gaps around stones" (p163), which emerged within four months after a fire in a Mediterranean zone. Similarly, miniature sorted patterned ground, thought to have emerged over a short time, has been described by Wilson and Clark (1991). No published accounts on recurrently forming, seasonal, frost-related patterned ground from other regions, however, have been found. The ability of frost-induced sorted patterns to develop within five to six weeks demonstrates the effect of regular, diurnal freeze-thaw cycles during the high Drakensberg winter months. Perennial miniature sorted patterns have only been observed at altitudes over 3200 m in the high Drakensberg. From the present distribution of seasonal and perennial miniature sorted patterns, it would appear that at altitudes below 3200 m, the present climate is less conducive to maintaining or preserving miniature cryogenic landforms during the summer months. Most miniature sorted patterns in the high Drakensberg develop near wetlands, streams and ground seepage zones, and are consequently destroyed during the wetter months.

The presence of primarily seasonal miniature sorted patterns is possibly the reason why most of these patterns are poorly developed. Clearly, fluvial processes during the warmer months dominate over cryogenic processes during the colder period, in surface modification. Although the absence of moisture during the cold period restricts the potential for frost heave on the drier slopes, it appears that the duration and intensity of freeze is still geomorphically significant at particular sites, such as at Mafadi, where perennial sorted patterned ground varieties occur. It would seem that conditions favourable for pattern development have become less favourable during recent years and are consequently only periodically active. For instance, the extensive cover of lichen on larger clasts and the establishment of plants on coarse stripes could be indicative of a degradational process. The coarse stripes offer natural gutters for the runoff during high precipitation events and concentrated flow assists in the destruction of the stripes at their downslope margins (personal observations). The ever increasing number of livestock and people crossing the higher summit slopes has undoubtedly contributed to the degradational process of sorted stripes.

It would appear from earlier discussions that the miniature sorted patterned ground varieties are a product of diurnal frost cycles while the larger sorted patterns are restricted to altitudes of over 3400 m and are a product of seasonal freezing. As seasonal freezing is restricted to the higher summits and the less-insolated south-facing slopes, this has limited the distribution of larger sorted patterns. The steep slope gradients, shallow soils and absence of adequate material supply, required for the formation of larger patterns, has further restricted their distribution. Unlike in many periglacial regions, no glacial sediments have yet been found in the high Drakensberg. It has been said that up to 60% of the world's patterned ground found in periglacial regions occurs in till (Goldthwait, 1976). The restricted distribution and occurrence of larger sorted patterned ground may also be as a result of rather short periods of cold, wet conditions during the Holocene, or more particularly, during recent decades. It is also possible that larger sorted patterned ground varieties were more widespread in the past, but have subsequently been destroyed owing to climatic and/or regional environmental change during the Holocene. It appears that only those patterns found on the higher summits have been given the time to develop and/or been preserved, suggesting intensified cryogenic activity at higher altitudes.

4.6.2 Large sorted patterns: Possible age and Palaeoenvironmental inferences

It has not been possible to date the large sorted patterns, owing to an absence of datable material. Therefore, several deductions made in this section are somewhat speculative.

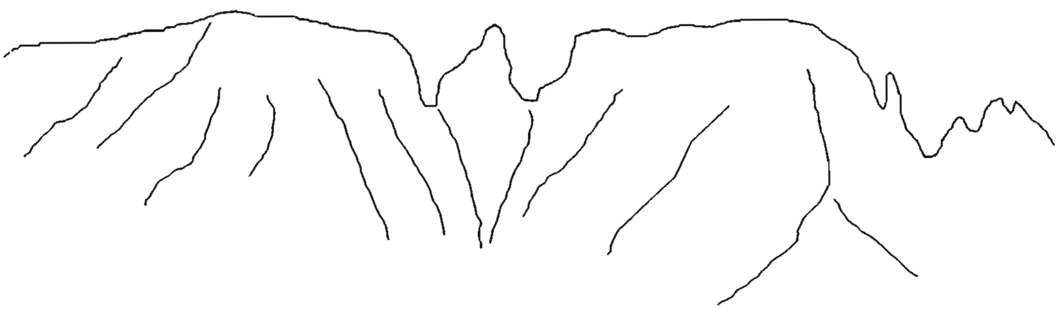
It has been suggested that large sorted patterned ground over one metre in diameter and with large boulders is indicative of at least marginal permafrost (Williams, 1975; Goldthwait, 1976; Washburn, 1980). Ballantyne and Matthews (1982) and Hall (1994b) have suggested caution with this as large sorted patterns may occur in response to local environmental conditions rather than regional climate. Black (1969) has also suggested that an impermeable horizon may help promote intense frost action near the surface when permafrost is absent. The large sorted patterns at Mafadi Summit may not necessarily imply former permafrost but certainly suggest seasonal ice which may have been periodically prolonged at times. Such large sorted patterns are also believed to have developed when the MAAT is between 0 and -4°C (Karte, 1983; Wayne, 1983). Washburn (1980) has suggested a MAAT of between 0 and -2°C for larger (more than 1 m) sorted patterns. Patterns very similar in dimension to the Mafadi sorted circles (Site 1) occur at Slettmarkbreen where the MAAT is about -1.6°C (Ballantyne and Matthews, 1982), about 5°C cooler than at Mafadi Summit. On the basis of such findings, it would appear that the patterns of the dimension found at Mafadi Summit are likely to form where the MAAT is approximately 0 to -2°C. The present MAAT on Mafadi Summit is estimated at between 3 and 4°C (extrapolated from Letseng-la-Draai air temperature data at 3050 m a.s.l., Table 2.3). A drop in MAAT or seasonal air temperature (winter) of about 3 to 4°C, accompanied by relatively wet summer and autumn conditions may be sufficient to initiate such pattern formation where frost susceptible sediments occur. S.A. Harris (1981) has cautioned many palaeoclimatic studies that assume a constant lapse rate, showing that lapse rates on Plateau Mountain (U.S.A.) are most pronounced at higher altitudes and increase very substantially in winter. Should Holocene palaeo-lapse rates in the higher Drakensberg have been significant and/or increased during colder months/periods, then these patterns may have been active at various intervals during the Late Holocene. The sorted circles (Site 1) appear relatively "fresh" (given the well preserved morphology) and are unlikely to have survived since the Pleistocene, given that such patterns may decay rapidly with changing regional/local environmental conditions (Ballantyne and Matthews, 1982). It is doubtful whether such patterns could have survived what is believed to have been a warm phase in the region, 5000 to 8000 yrs BP (Scott, 1982; Mitchell, 1994). It is perhaps more likely that the patterns formed during the Late Holocene (approx. 1000 to 3000 yrs BP) when conditions are thought to have been slightly cooler and wetter than at present (Scott, 1982, 1984, 1990; Partridge *et al.*, 1990; Mitchell, 1994). The patterns may have been in a state of dynamic equilibrium since then, with marginal activity during cooler oscillations (e.g. during the Little Ice Age from 1300 to 1850 AD) and decay during warmer oscillations. If this were the case, then it may correspond with pulses of human occupation found in eastern Lesotho (Mitchell, 1994). The absence of organic horizons (Hanvey and Marker, 1994) and the presence of these patterns for possibly 1000 to 3000 years may also indicate that the summit areas have had restricted and/or sparse vegetation cover over the past few 100 to 1000 years. The scarcity of Late Holocene archaeological findings from eastern and highland Lesotho, dating between 1000 to 4000 yrs BP (Mitchell, 1994), may support some of the above inferences.

4.7 SUMMARY

This chapter has presented a detailed account on a variety of miniature and large sorted patterned ground types occuring in the high Drakensberg. While seasonal patterns occur at any altitude in the high Drakensberg, perennial patterns appear to be restricted to altitudes above 3200 m. Large relict sorted circles are possibly the most conspicuous periglacial landforms yet identified in this Afro-alpine region and are indicative of a past colder and possibly wetter period. However, it is not only sorted patterns, but also non-sorted patterns, that have required a detailed assessment in the high Drakensberg. The following chapter (Chapter 5) will, therefore, specifically examine the morphology and possible processes associated with non-sorted patterns in the high Drakensberg.

CHAPTER FIVE

NON-SORTED PATTERNED GROUND



CHAPTER 5

NON-SORTED PATTERNED GROUND

5.1 INTRODUCTION

An introduction to patterned ground and its classification into sorted and non-sorted varieties was given in Section **4.1**, and therefore, will not be repeated here. Non-sorted patterns such as mudboils (Shilts, 1978; Zoltai and Tarnocai, 1981), circles (Mooers and Glaser, 1989), stripes (Soons and Price, 1990: Mark, 1994), steps (Washburn, 1956) and hummocks (Grab, 1994) have been described in the literature. While several papers have focussed on sorted patterned ground from seasonally frozen areas such as the Comeragh Mountains, south-east Ireland (Wilson, 1992), the East African Rift mountains (Hastenrath, 1973) and the sub-Antarctic islands (Hall, 1979, 1983; Wilson and Clark, 1991), there appears to have been less attention devoted to non-sorted patterned varieties. Larger varieties of patterned ground are sometimes associated with the presence of permafrost, yet the relationship between patterned ground formation and permafrost has remained imperfectly understood. Caution should be exercised when distinguishing between permafrost and non-permafrost related patterned ground. For instance, large non-sorted patterns located at sea level in Nova Scotia, Canada, were found to be the product of frequent freeze-thaw cycles within saturated soils overlying impermeable bedrock (Mooers and Glaser, 1989) where permafrost was absent. Although the occurrence of thufur (Grab, 1994), circular plant growths (Hastenrath and Wilkinson, 1973) and non-sorted circles with bare centres (Boelhouwers, 1991a) is known in the high Drakensberg, empirical data are scarce. This chapter focuses primarily on thufur, which are the most common non-sorted patterns found in the high Drakensberg. However, towards the end of the field-survey, the author did locate non-sorted steps on Mafadi Summit, and so these are also briefly described.

5.2 THUFUR

5.2.1 Introduction

Over the last decade periglacial geomorphologists have focused attention on frost-induced mounds (e.g. Bobov, 1972; Washburn, 1983; Pollard and French, 1984; Nelson *et al.*, 1985; Lagerbäck and Rodhe, 1986; Seppälä, 1986, 1988a, 1988b; Hinkel, 1988; Pollard, 1991; Harris, 1993; Sone and Takahashi, 1993). One such form of frost-induced mound is the "thufa" (plural "thufur") (Thorarinsson, 1951), a term used most frequently for features in Iceland and Greenland (e.g. Schunke, 1977). Other terms used in the literature to describe similar forms of frost-related hummocks include "earth hummocks" (Lundqvist, 1969; Costin and Wimbush, 1973; Pettapiece,1974; Zoltai and Pettapiece, 1974; Tarnocai and Zoltai, 1978; Zoltai and Tarnocai, 1981; Scotter and Zoltai, 1982; Ellis, 1983; Gerrard, 1992; Kojima, 1994), "turf hummocks" (Raup, 1963, 1965), "earth grass mounds" (Dionne, 1966), "mud hummocks" (Mackay and Mackay, 1976), "soil hummocks" (Williams and Smith, 1989) and "frost hummocks" (Williams, 1961). Although some authors regard several of these terms as synonymous, there has been dispute concerning the usage of the above-mentioned terminology. For instance, Schunke and Zoltai (1988) and Gerrard (1992) regard the term "earth hummock" and "thufur" as being synonymous, while Harris (1988) makes a distinction between the two. It would seem that while earth hummocks are commonly associated with intense heave in permafrost environments (Zoltai and Pettapiece, 1974; Zoltai *et al.*, 1978; Harris, 1988), thufur are produced by the local displacement of surface soil material due to seasonal frost penetration (Schunke, 1977; Schunke and Zoltai, 1988) and hence may not require permafrost (Harris, 1988). Schunke and Zoltai (1988) have suggested that those hummocks occurring in permafrost environments should be referred to as "earth hummocks" and those in non-permafrost environments as "thufur". The following discussion uses the latter recommendation and will thus refer to the non-permafrost related Lesotho features as "thufur".

Thufur have a circumpolar occurrence (Tarnocai and Zoltai, 1978) where they have been reported from Iceland (Thoroddsen, 1913; Schunke, 1977; Gerrard, 1992), Greenland (Raup, 1963, 1965), and Arctic and sub-Arctic Canada (Zoltai and Pettapiece, 1974; Tarnocai and

Zoltai, 1978). In addition, thufur have been recorded in alpine areas of North America (Billings and Mooney, 1959; Scotter and Zoltai, 1982), the Snowy Mountains, Australia (Costin and Wimbush, 1973), the Daqingshan Mountains, Inner Mongolia (Cui Zhijiu and Song Changqing, 1992), northern Japan (Koaze *et al.*, 1974), Sweden (Lundqvist, 1969) and in the northern Pennines, U.K. (Tufnell, 1975).

In Africa, thufur are said to occur in the Ruwenzori Range (Heinzelin, 1952) and the Lesotho Mountains (Van Zinderen Bakker, 1965; Hastenrath and Wilkinson, 1973; Van Zinderen Bakker and Werger, 1974; Boelhouwers, 1991a; Hanvey and Marker, 1992; Grab, 1992a, 1994), but empirical data are scarce. Although some of the current work on thufur has been published (e.g. Grab, 1994; Grab, in press), this chapter aims to provide further detailed empirical information on thufur occurring at several of the study sites.

5.2.2 Thufur or zoogeomorphic hummocks?

Lynch and Watson (1992) have suggested that the thufur are produced by the burrowing action of rodent-moles (*Cryptomys hottentotus*), and subsequently by the enlarging and cleaning activities of "ice rats" (*Otomys sloggetti robertsi*). Although Lynch and Watson (1992) regard the origin of these mounds as somewhat "controversial" (p147), they believe that vegetation colonization and repeated top-dressing may eventually form the hummock landscape. Elsewhere, in Cumbria, United Kingdom, Pemberton (1980) examined the possibility that hummocky microrelief may result from abandoned and grassed-over molefields. However, it was concluded that this hypothesis did not explain the origin of the hummocks. More recently, Cox *et al.* (1987) have attributed thufur-like mounds in the Western Cape Province (South Africa) to termites (*Microhodotermes viator*) and/or bathyergid mole rats (*Cryptomys hottentotus*). However, such a zoogeomorphic origin seems unlikely at the present study sites, as is outlined below.

Detailed observations at several of the present study sites were undertaken to ascertain whether or not zoogeomorphic hummocks occur there. It was observed that molehills frequently occur amongst the thufur (which are vegetated), yet, not one molehill was found with vegetation established on it. Several reasons may account for the absence of vegetation on the molehills. The wetland soils are fine-textured with high percentages of fine sand and silt/clay (averaging 54%) (Grab, 1994). The molehills are not compact and therefore soils are easily removed from the sites. Overland flow of the wetlands occurs frequently as the water table rises above the surface, consequently destroying the molehills. Rain-splash and rain-wash also contribute towards molehill disintegration (personal observations). In fact, molehills are usually surrounded by bare patches indicating the former sites of other molehills. During the winter months, intense needle ice action occurs on the molehills, often disrupting the mounds completely. Desiccation and wind action towards the end of winter further contribute towards the erosion of such molehills. Clearly, vegetation is unable to establish itself on the molehills at any given time due to the on-going disturbances. Further, there is no indication whatsoever, that *Otomys sloggetti* contribute towards the enlarging of such molehills. Six sectioned hummocks were unable to reveal infilled or collapsed mole tunnels. Earthworms, termites and other fauna are mostly absent from the hummocks. Thus, there is no evidence to suggest the occurrence of zoogeomorphically induced vegetated hummocks in the high Drakensberg. It has, however, been shown that the vegetated hummocks found within alpine wetlands in the high Drakensberg are indeed frost-induced hummocks (thufur) (Grab, 1994).

5.2.3 General Characteristics of Thufur

5.2.3.1 Distribution

It has already been shown (Section **1.2.2.4**) that thufur have for a long time been considered to occur over a restricted altitudinal range from 2900 to about 3100 m (Hastenrath, 1972; Lewis, 1988a, 1988b; Boelhouwers 1991a) in the Lesotho highlands. However, from the present study and initial survey, it would appear that thufur may occur from at least 2750 m a.s.l. (Tlanyaku Valley, Cathedral Peak area) to 3360 m a.s.l. (Mafadi Summit region) (Table 5.1). It is doubtful whether thufur occur below 2700 m a.s.l. in Lesotho, as they have never been reported from such low altitudes. Thufur remnants between 2750 and 2900 m a.s.l. are frequently disrupted from livestock overgrazing and trampling.

Site			Sample number (n)	Slope aspect	Altitude (m a.s.l.)	Slope gradient (degrees)	Diameter Dd (cm)	Diameter Ad (cm)	Diameter Mean (cm)	Ratio of Ad to Dd	Height (cm)	Inter-thufur spacing (cm)	% of thufur broken up
Mohlesi													
	1	mean	10	N	2880	1	43.2	42.2	42.7	1 : 1.02	18.0	16.8	0%
	2	mean	30	N	2900	6	52.5	47.2	49.8	1 : 1.11	16.9	16.5	20%
	3	mean	10	N	2910	9	45.8	41.7	43.7	1 : 1.10	14.8	-	-
	4	mean	10	N	2980	5	50.0	47.7	48.8	1 : 1.05	17.0	-	-
	5	mean	10	NE	3110	6	42.2	35.3	38.7	1 : 1.19	15.6	-	-
	6	mean	10	E	3220	7	52.4	45.7	49.0	1 : 1.15	16.8	37.2	-
	7	mean	10	SE	2910	8	63.3	54.4	58.8	1 : 1.16	14.7	-	-
Mashai													
	1	mean	40	W	2950	3	83.8	76.4	80.1	1 : 1.09	20.2	-	-
		std. dev.	-		-	-	48.6	48.5	-	-	11.1	-	-
		range	-		-	-	-	-	38 - 249	-	6 - 59	-	-
	2	mean	457	SW	2960	4	45.5	44.8	45.2	1 : 1.02	20.3	33.27	19.2%
		std. dev.	-		-	-	16.1	16.7	-	-	6.4	25.11	-
		range	-		-	1 - 6	-	-	15 - 133	-	6 - 40	-	-
Popple Peak													
	1	mean	22	N	3170	2	48.6	38.5	43.5	1 : 1.26	19.8	-	63.6%
		std. dev.	-		-	-	18.9	16.5	-	-	4.8	-	-
		range	-		-	-	-	-	21 - 75	-	16 - 32	-	-
Mafadi													
	1	mean	25	N	3360	2	61.8	47.5	54.6	1 : 1.30	21.8	15.2	26.2%
		std. dev.	-		-	-	28.5	12.9	-	-	3.8	8.7	-
		range	-		-	-	-	-	30 - 112	-	14 - 32	7 - 44	-
	2	mean	15	N	3360	1	86.5	82.1	84.3	1 : 1.05	8.4	76.7	0%
		std. dev.	-		-	-	23.8	23.9	-	-	5.6	37.9	-
		range	-		-	-	-	-	33 - 138	-	3 - 22	22 - 122	-

Dd = Downslope diameter / Ad = Across-slope diameter

Table 5.1 Thufur dimensions and characteristics at several of the study sites.

According to Schunke (1977), the upper altitudinal boundary for thufur is where the amount of fine material covered by grasses diminishes significantly, such that the development of thufur is no longer possible. However, several of the identified vegetation types growing on and adjacent to thufur at the study sites have no apparent altitudinal limit within this region. It is therefore not plant cover which determines the upper limit for thufur in the high Drakensberg, but rather the availability of sufficient soils and abundant near-surface moisture, which are necessary prerequisites for thufur development to occur (Schunke, 1977). Because large parts of the high Drakensberg are semi-arid, thufur have been restricted to wetland sites and thus have a sporadic distribution. The wetlands form broad margins up side-channels and spread headward into the basins, often producing triangular riverheads. Many of these wetlands are considered to be bogs because they produce peat and manifest an abundance of raised peat-based landforms such as elongate ridges and hummocks (Van Zinderen Bakker, 1955, 1965, 1981; Jacot-Guillarmod, 1962, 1963; Grobbelaar and Stegmann, 1987; Schwabe,1989). Thufur are restricted to the wetlands and appear best developed in the bogs owing to the deep, fine, organic horizons that facilitate moisture retention during colder months. The distinct valley asymmetry found throughout the high Drakensberg (Meiklejohn, 1992, 1994) has controlled the spatial distribution of wetlands and consequently thufur swarms (Grab, 1994). The relatively steep south-facing slopes have resulted in fewer wetlands developing on such slopes (Grab, 1994) (Table 5.1). It is only along the broader, gentler valleys, such as the Sani and Mashai, that thufur are well developed on south-facing slopes.

5.2.3.2 Vegetation

Although it has been reported (Schunke and Zoltai, 1988) that in some high Arctic regions hummocks may have bare apexes, thufur are usually overgrown with vegetation (Zoltai and Pettapiece, 1974; Schunke and Zoltai, 1988). The vegetation growing on and around thufur in North America consists mostly of tundra, meadow or sub-Arctic woodland (Schunke and Zoltai, 1988), and even trees have occasionally been found growing on the hummocks (Zoltai and Pettapiece, 1974; Zoltai, 1975). Vegetation growing on thufur in most alpine regions is predominantly meadow grass and sedges, which includes *Koeleria capensis*, *Lobelia galpinii* and *Sebaea marlothii* in the high Drakensberg (Grab, 1994). Variations in vegetation cover and vegetation types have been reported to occur between thufur apexes and depressions (Schunke, 1977; Williams and Smith, 1989). Most thufur sites in the high Drakensberg show few vegetational variations between apexes and depressions. Table 5.2 lists 14 vegetation types identified at a thufur site in the Mashai Valley and results show that 10 of these vegetation types grow on and adjacent to the thufur. *Athrixia fontana*, *Helichrysum flanaganii* and *Helichrysum trilineatum* do not favour the waterlogged depressions and appear, therefore, to be restricted to the apexes (Table 5.2). Conversely, the more aquatic *Polygala ohlendorfiana* was not found growing on the thufur apexes, but rather on the lower thufur sides and depressions (Table 5.2). Occasionally, vegetation cover is somewhat thicker on the thufur apexes than in the depressions, as has also been suggested by Van Zinderen Bakker and Werger (1974).

Vegetation	Apex	Depression
Lobelia galpinii	X	X
Koeleria capensis	X	X
Sebaea marlothii	X	X
Alchemilla	X	X
Athrixia fontana	X	-
Cotula hispida	X	X
Cotulaleariloba	X	X
Crassula natalensis	X	X
Felicia	X	X
Helichrysum flanaganii	X	-
Helichrysum trilineatum	X	-
Othonna	X	X
Polygala ohlendorfiana	-	X
Wahlenbergia	X	X
X present / - absent		

Table 5.2 Comparison of thufur apex and depression vegetation characteristics at the Mashai Valley site.

5.2.3.3 External morphology

Schunke (1977) has classified various thufur forms according to their external appearance, which include: hump-shaped, plateau-shaped, shield-shaped and ridge-shaped forms. Thufur found on gentle gradients such as at Mafadi, are often circular with flat tops (plateau shaped), while at steeper sites thufur are commonly elongate giving them a more ridge- or hump-shape.

The thufur sides are also variable in appearance. At flat sites where thufur are closely spaced, the sides tend to be very steep and reach almost 90° at times. On steeper slopes the thufur sides tend to be low angled (20° to 30°) and more convex towards their apexes.

Although the appearance of thufur in the high Drakensberg does not differ much between sites, a somewhat morphologically different variety of non-sorted circle was located about 60 metres from a thufur swarm at the Mafadi study site (Figures 5.1 and 5.2). While the thufur comprise distinctly raised hummocks with little or no vegetational changes, the more recently discovered non-sorted circles are only slightly raised (mean height = 8.4 cm) but have a more vigorous vegetation growth (Figures 5.1 and 5.2) than the surrounding areas. It has not yet been determined whether these raised vegetation rings, which include *Helichrysum flanaganii*, are a formative stage for thufur development, which have otherwise been referred to as "embryonic earth hummocks" (Schunke and Zoltai, 1988, p240).

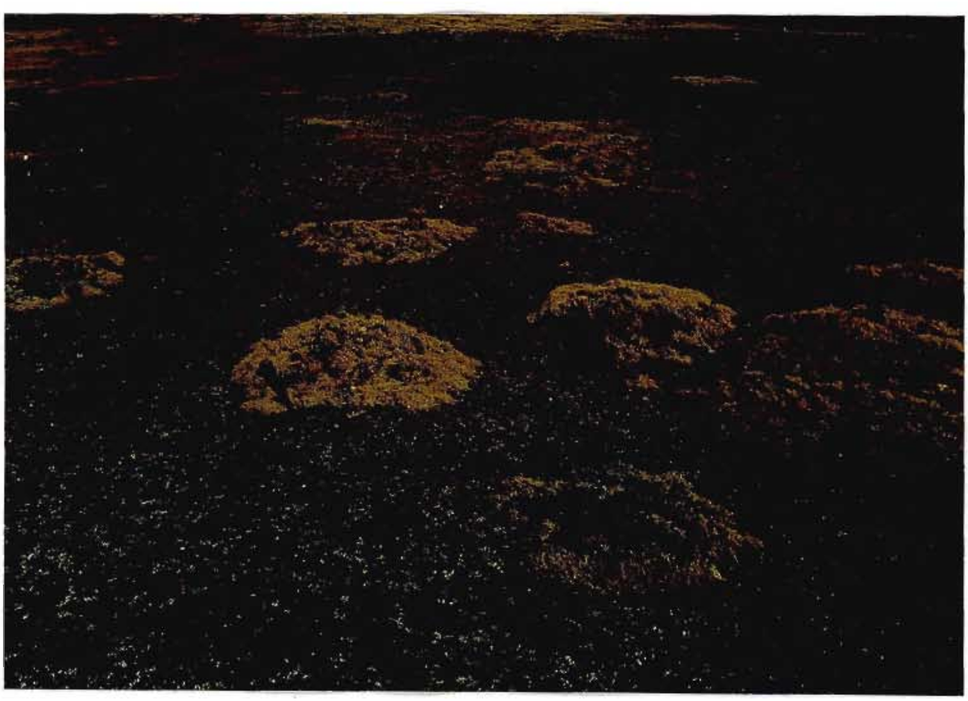


Figure 5.1 Non-sorted circles with vigorous vegetation growth, Mafadi, Site 2.

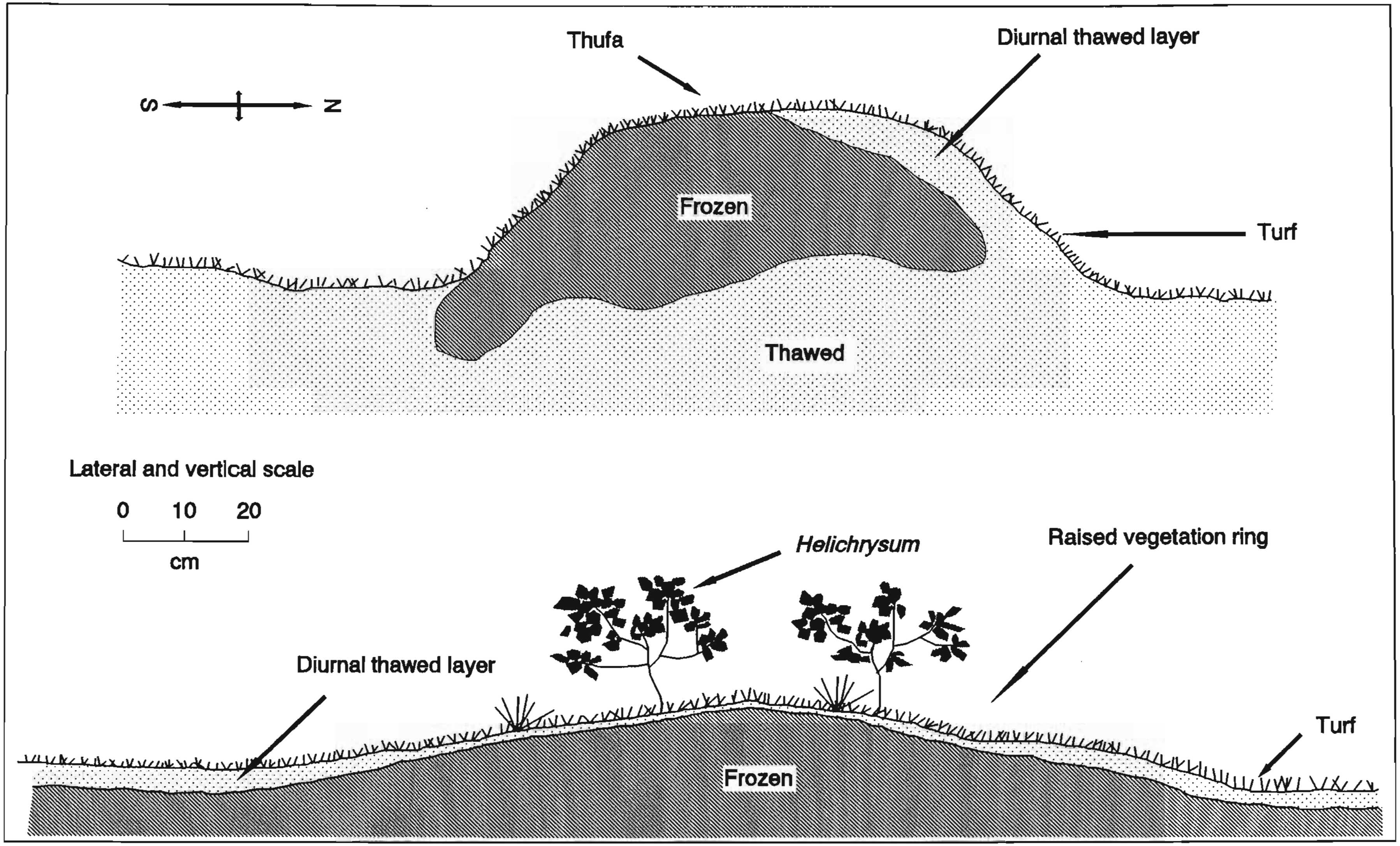


Figure 5.2 Comparing thufur and raised vegetation rings at Mafadi Summit. Ground freezing was assessed during July 1994.

It would appear that thufur in alpine areas are considerably smaller than those reported from Arctic and sub-Arctic regions (Grab, 1994). For instance, Gerrard (1992) reports that the Icelandic thufur may be up to 75 cm high and average 100 to 200 cm in diameter. Equally, Canadian sub-Arctic thufur are said to average 50 cm in height and 160 cm in diameter (Zoltai and Pettapiece, 1974), while in the western sub-Arctic they may average as much as 320 cm in diameter (Tarnocai and Zoltai, 1978). Yet, in the Canadian Rocky Mountains, K.Hall (1995, pers. comm.) observed thufur averaging less than 15 cm in height. External dimensions for 649 thufur measured at 12 different locations in the high Drakensberg (Table 5.1) indicate an overall average height of 19.5 cm and mean diameter of 48.9 cm. These thufur are considerably larger than those reported from the Snowy Mountains, Australia, where the mean height is 15 cm and mean diameter 30 cm (Costin and Wimbush, 1973). Grab (1994) has attributed the somewhat restricted thufur growth in most alpine areas to a brief freezing period and shallow freeze penetration. Although Gerrard (1992) found little dimensional variations between thufur at his Icelandic sites, Tarnocai and Zoltai (1978) measured hummocks that ranged between 14 and 78 cm in height at their Mackenzie Valley site. Considerable variations in thufur size may occur at individual sites and between sites in the high Drakensberg (Table 5.1). For instance, thufur height varied from 3 cm (Mafadi) to 40 cm (Mashai) and their diameters from 15 cm (Mashai) to 249 cm (Mashai) (Table 5.1). It has been argued that variations in ground conditions, such as available moisture, soil characteristics , vegetation cover and snow depth, may have a significant influence on the development of thufur (Schunke and Zoltai, 1988; Gerrard, 1992), and therefore, it should not be surprising to expect variations in thufur size at particular sites where surface conditions may vary. Observations in the high Drakensberg have shown that the wetter, more protected wetland centres are occupied by the largest thufur, while the wetland periphery areas have poorly developed thufur. Elsewhere, thufur have been reported to attain their greatest vertical development in the central parts of depressions (Scotter and Zoltai, 1982). Available measurements show that thufur size in the high Drakensberg appears not to be controlled by altitudinal changes (Table 5.1).

Grab (1994) has shown that thufur on gentler slopes (1° to 5°) have greater vertical development than those on steeper slopes (8° and 9°), indicating that steeper slopes in alpine regions may impede vertical thufur growth. Results from the present study indicate that the vertical development of thufur is usually greatest where slope gradients are 3° or less (Table 5.1). The steeper slopes in the Mohlesi Valley (6° to 9°) have mean thufur heights not exceeding 17 cm (Table 5.1). Although thufur have been reported from slopes of 9° and 10° in the Sani Pass region, Lesotho (Boelhouwers and Hall, 1990), and up to 15° in parts of Iceland (Schunke, 1977), Tarnocai and Zoltai (1978) have confirmed that vertical development of thufur rarely takes place where the slope gradient is over 7°, but rather, become increasingly elongated downslope on steeper gradients.

Although no distinct trend of elongation with slope gradient is found between the study sites (Table 5.1), mean values indicate that elongation tends to be in the downslope direction at all sites. Tarnocai and Zoltai (1978) found that steeper slopes show increasing ratios between across-slope to downslope diameters, yet in the high Drakensberg, the lowest ratio of 1 : 1.02 occurs on a 3° slope and the highest ratio of 1 : 1.3 is found on a 2° slope (Table 5.1). Clearly, elongation is not a function of slope gradient alone; factors such as site-specific surface conditions, thufur spacing and slope aspect are also likely to influence thufur elongation. Elongation regularly occurs diagonally across-slope or parallel to the slope contour on slope gradients of up to 3°.

According to Schunke (1977), the inter-thufur spacing in Iceland is usually not more than 10 to 15 cm, while Tarnocai and Zoltai (1978) have indicated that the distance between hummocks at their North American sites is usually less than the hummock diameter. At Mafadi, mean inter-thufur spacing varies from 15.2 cm (Site 1) to 76.7 cm (Site 2) and commonly averages between 16 and 40 cm at the other sites (Table 5.1). The mean inter-thufur distance is less than the average thufur diameter at all of the present study sites (Table 5.1). Although factors such as slope aspect, slope gradient and altitude do not appear to influence thufur spacing, it is apparent that site-specific surface conditions have exerted a role here. Tarnocai and Zoltai (1978) have explained that hummock spacing increases where sites are poorly drained and where the groundwater table is near the surface, yet, in the high Drakensberg, the opposite occurs. The drier wetland periphery areas have widely spaced thufur while the wetter centres have more densely populated thufur swarms. At sites where thufur are closely spaced, snow and/or ice may remain in the inter-thufur spaces for several weeks because of the micro-topographically induced shading and protection from desiccating winds (Figure 5.3). Where such snow is thick and compact, it is also likely to act as an insulator, whereas the thufur apexes are more exposed to freezing.

Where thufur are closely spaced, they sometimes merge to form elongate thufur (Figures 5.4 and 5.5). In some instances three or four thufur may coalesce to form a morphologically complex thufa (Figures 5.4 and 5.5). Often, the original thufur may be discerned as they are slightly elevated above the "joined areas" (Figures 5.4 and 5.5). Although Mark (1994) has explained that non-sorted stripes and hummocks may merge to form "beaded hummocks" (p271), no further references discussing the coalescence of thufur were found.

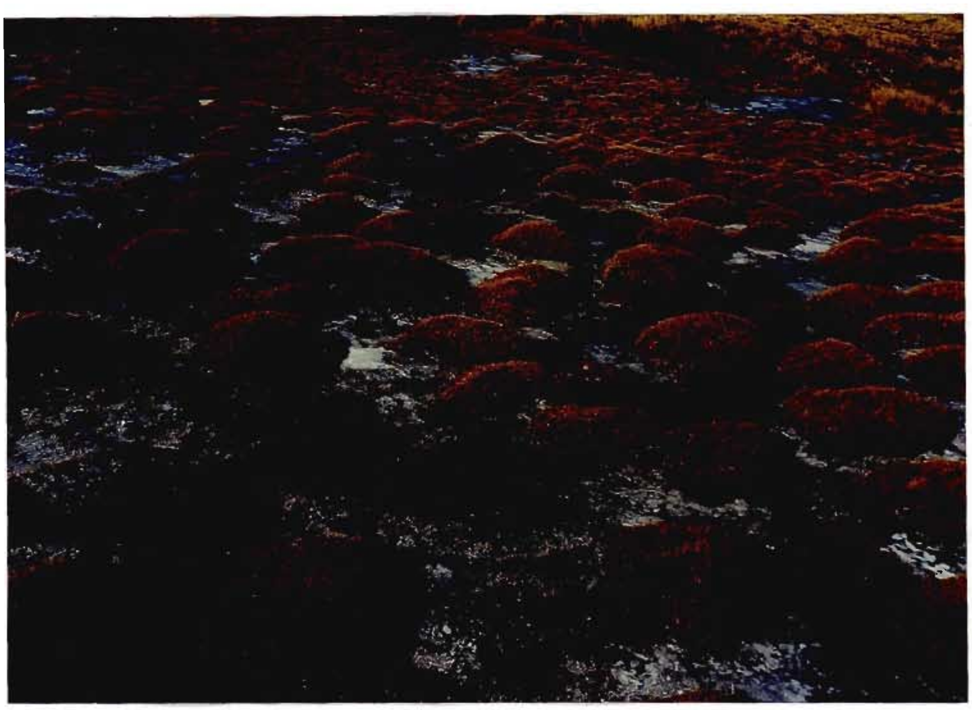


Figure 5.3 Snow and/or ice may remain in the inter-thufur spaces for several weeks during winter. The thufur average about 60 cm in diameter.

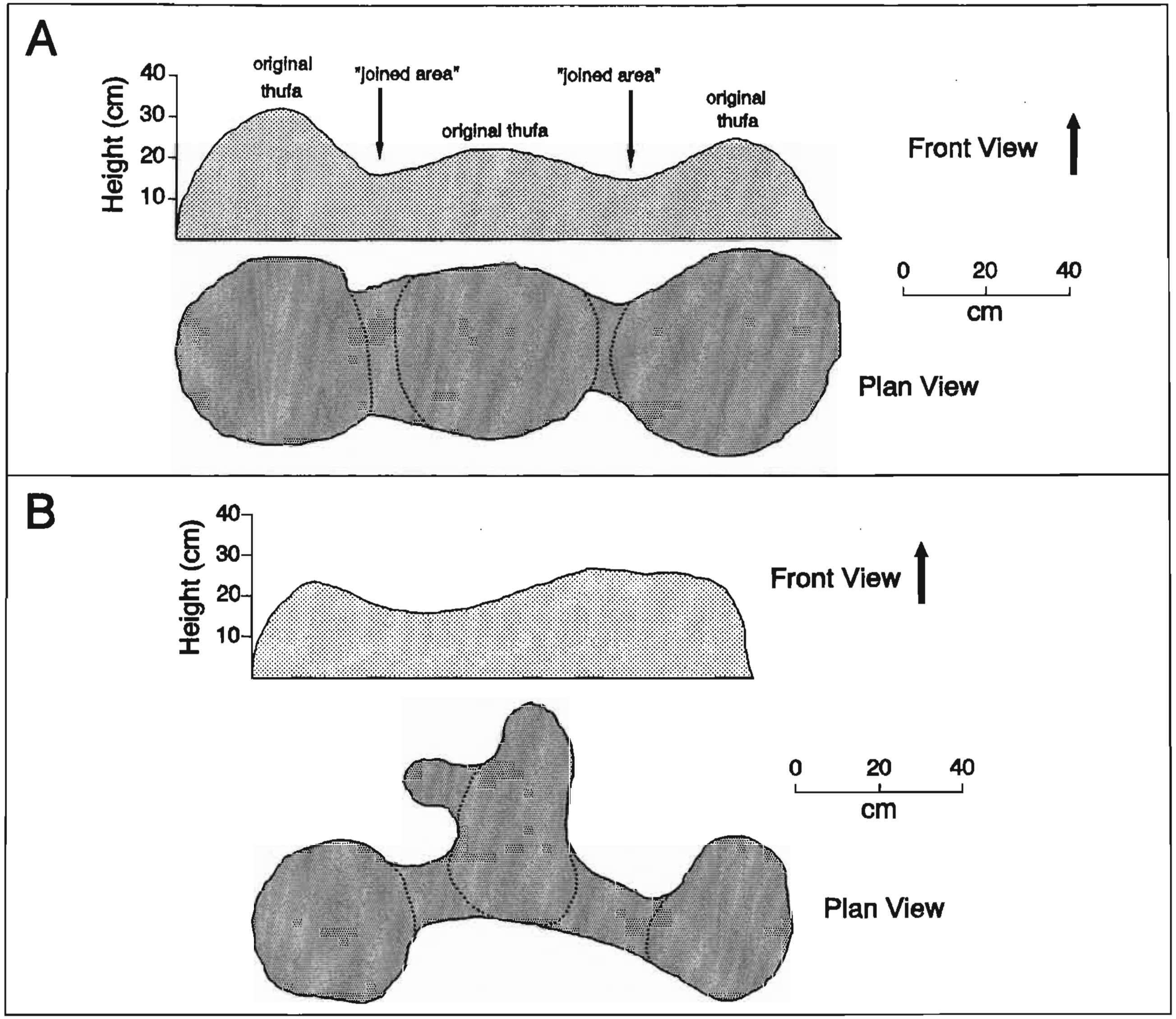


Figure 5.4 Closely spaced thufur sometimes merge to form (A) elongate thufur or (B) coalesce to form a morphologically complex thufa.

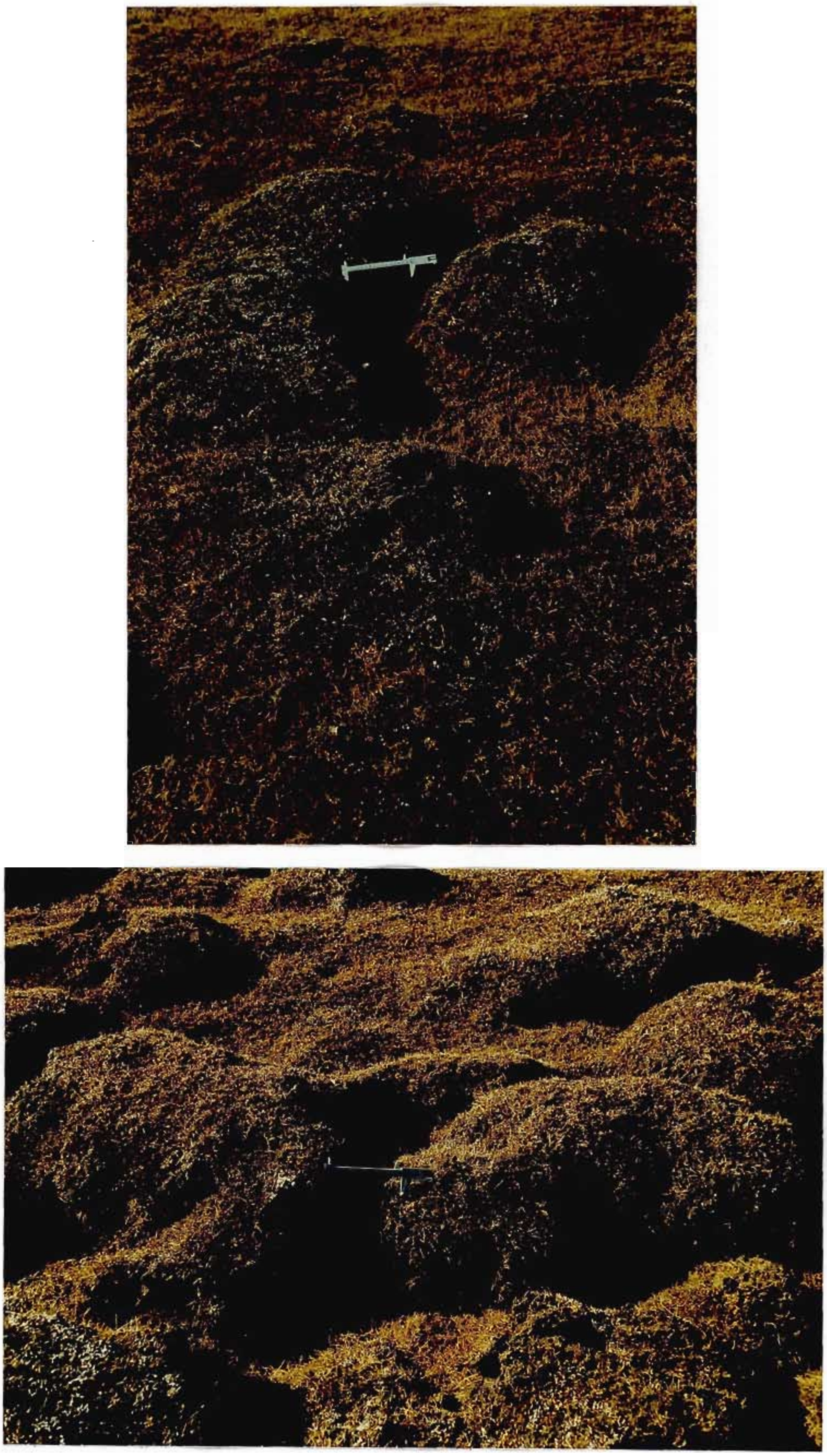


Figure 5.5 Photographs showing elongate and morphologically complex thufur at Mafadi, Site 1.

5.2.3.4 Internal characteristics

Prior to the present study, the internal characteristics of thufur found in southern Africa had not been examined in any great detail (Grab, 1994). Sectioned thufur indicate that a thick rooting zone extends down to 1.8 cm on the thufur apexes and 3 cm on their lower sides (Grab, 1994), however, root strands may extend a further 15 to 20 cm into the organic soils. The thufur soils have been described as "dark, heavy clay-like" (Hastenrath and Wilkinson, 1973, p160), "organic peaty" (Hanvey and Marker, 1992, p357) and "homogeneous dark grey" (Grab, 1994, p114). Further observations have shown that thufur soil characteristics may vary slightly between sites in the high Drakensberg. While the soil colour is usually dark grey at most sites (Grab, 1994), at other sites such as at the Mashai Valley, soil colour may at times be black (Table 5.3). At one sampling site in the Mashai Valley, a light yellowish brown sandy layer was encountered at 50 cm depth below the depression (Table 5.3). A similar finding was made in an earlier study (Grab, 1992a) in the Mohlesi Valley, where a lighter coloured soil of coarser texture was found at 25 cm depth. Since convoluted upper horizons are usually absent or very weakly defined, most of these soils may be described as homogeneous.

Organic matter was found at all depths within a sectioned thufa profile (Mohlesi Valley) but the soils nearest the apex and sides contain the highest percentage organic matter (14%) (Table 5.4). The percentage of organic matter decreases progressively below the apex to 5.6% at 40 cm depth (Table 5.4). Comparisons between three thufur and their adjacent depressions in the Mashai Valley show that at 10 cm depth the thufur have 30.7% organics (mean dry sample weight) and the depressions 17.3% (Table 5.3). These values are somewhat higher than the mean value of 15.6% found for Icelandic thufur (Schunke, 1977). At several sites in the high Drakensberg, thufur have formed entirely within thick peat deposits.

Thufur commonly contain fine-textured soils with high percentages of silt and clay sized particles. The percentage of silt/clay may, however, vary considerably from region to region. For instance, a study in Canada found mean silt/clay contents of 58 to 99% (Tarnocai and Zoltai, 1978), while those in Iceland may average from 12% (Gerrard, 1992) to 60% (Schunke

		% weight Apex	Depression		
February	moisture	39.8	62.8		
	mineral soil & organics	60.2	37.2		
May	moisture	35.3	42.5		
	mineral soil & organics	64.7	57.5		
July	moisture	31.5	21.5		
	mineral soil & organics	68.5	78.5		
September	moisture	34.5	37.3		
	mineral soil & organics	65.5	62.7		
Soil colour	(dry) 10 cm depth	5Y 2.5/1 Black	5Y 2.5/1 Black	(dry) 50 cm depth below the depression	10YR 6/4 Yellowish Brown
Soil texture	gravel	0.0	24.7		
	sand	71.7	66.0		
	silt & clay	28.3	9.3		
Organics	dry sample	30.7	17.3		

Table 5.3 Mean soil characteristics for thufur apexes and depressions in the Mashai Valley. Three thufur apexes and three depression samples were collected at 10 cm depth during each sampling month of 1994.

Depth (cm)	Wet sample (% weight) Moisture	Organics	Soil	Dry sample (% weight) Organics	Soil	% silt/clay	Bulk density (g cm)
Apex							
5	61.3	14.0	24.7	37.9	62.1	68.1	0.88
15	64.0	8.7	27.3	24.3	75.7	68.4	1.27
20	66.9	6.4	26.7	18.8	81.2	55.4	1.28
30	65.5	5.7	28.8	15.4	84.6	48.0	1.30
40	66.4	5.6	28.0	14.5	85.5	29.0	1.48
Depression							
15 cm	68.6	9.1	22.3	21.1	79.0		1.31
Mean	65.5	8.3	26.3	22.0	78.0	53.8	1.30
Std. dev.	2.5	3.2	2.4	8.5	8.5	16.3	1.30

Table 5.4 Mean soil characteristics for a thufa apex and its adjoining depression. Samples were collected in the Mohlesi Valley during September 1992 (after Grab, 1994).

and Zoltai, 1988). The mean silt/clay content for thufur in the high Drakensberg is found to vary from 28.3% (Mashai Valley) (Table 5.3) to 54% (Mohlesi Valley) (Table 5.4). The silt/clay content for a Mohlesi Valley thufa diminishes rapidly with depth below the hummock, reaching a maximum of 68.4% at 15 cm depth and a minimum of 29% at 40 cm depth (Table 5.4). Soil texture characteristics for the Mashai Valley thufur indicate that finer material (silt/clay) is displaced into the apexes while coarser material (gravel) remains in the depressions (Table 5.3 and Figure 5.6). For instance, the apexes had 28.3% fines while the depressions only had 9.3% fines. Conversely, the apexes were absent of gravel while the depressions contained 24.7% gravel (Table 5.3 and Figure 5.6). Such particle size differences between thufur apexes and depressions may result from sorting associated with the differential freeze-thaw activity between these sites.

Although Tarnocai and Zoltai (1978) found that the soil moisture content for hummocks at their study sites increases with depth, no great variation in the distribution of moisture with depth was found within a sectioned thufa in the Mohlesi Valley (Table 5.3) (Grab, 1994). However, the study has found that the moisture content is lowest nearest the thufur surfaces (61.3%) (Table 5.4), as was also found by Tarnocai and Zoltai (1978). It would appear that desiccation, particularly during winter months, affects near-surface soils considerably. From Tables 5.3 and 5.4 and Figure 5.7, it can be deduced that soil moisture content in thufur apexes and adjoining depressions vary both spatially and temporally. The moisture content may be influenced by site-specific wetland conditions (i.e. drainage conditions) or seasonal precipitation and/or snow/ice thaw events. At several sites, the depressions are submerged and only the thufur apexes protrude above the surface water during the wet summer months. Seasonal fluctuations in soil moisture content were assessed for three thufur apexes and their adjoining depressions during 1994 (Table 5.3 and Figure 5.7). As shown in Figure 5.7, the seasonal soil moisture content for the apexes undergoes less significant change than in the depressions. While the soil moisture content for the depressions is considerably higher (62.8%) than that for the apexes (39.8%) during the summer months, the reverse occurs towards mid-winter (dry period). It is likely that prolonged drought and desiccation rapidly reduces the soil moisture content during the drier months. Because the freezing front normally penetrates the apexes more rapidly than the depressions (personal observations and

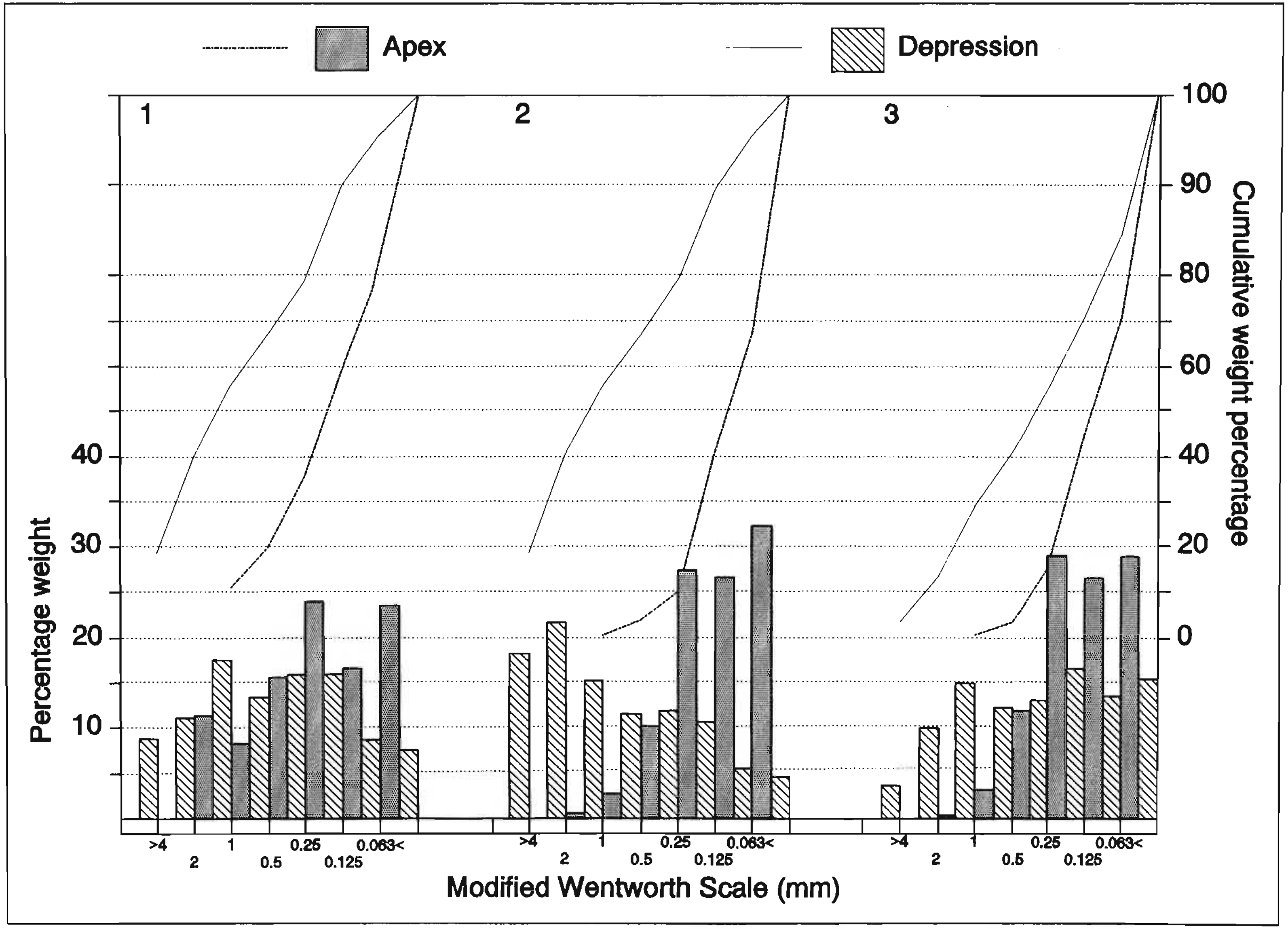


Figure 5.6 Particle size distribution for three thufur apexes and their adjoining depressions in the Mashai Valley (samples taken at 10 cm depth).

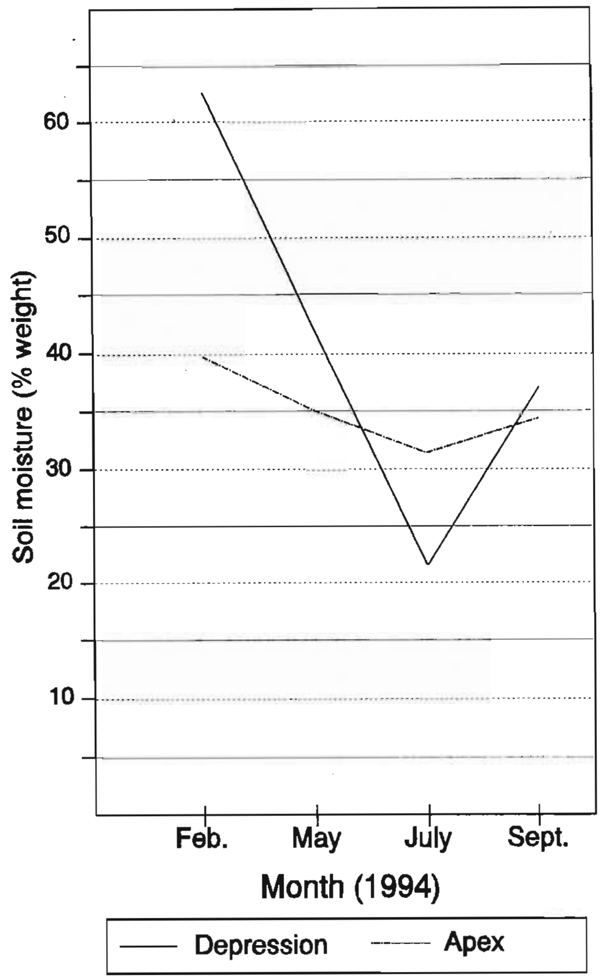


Figure 5.7 Comparison of mean thufur apex and depression soil moisture content in the Mashai Valley. Three apex and three depression samples were collected at 10 cm depth during each sampling month of 1994.

measurements), it is likely that moisture from the depressions is drawn towards the apexes, hence the accelerated loss of moisture in the depressions (Table 5.3 and Figure 5.7). "Hollow spaces" were not found within the sectioned thufur at the study sites, which agrees with findings made in the Northern Hemisphere (Schunke and Zoltai, 1988). Bulk density values for a thufa in the Mohlesi Valley vary from 0.88 g cm^{-3} near the apex surface (5 cm depth) to 1.48 g cm^{-3} at a depth of 40 cm (Table 5.4). It appears that bulk density values are influenced by soil organic matter content, hence the progressive increase in bulk density values with depth below the thufa apex (Table 5.4). It has also been suggested that a wide range of bulk density values could indicate cryoturbation (Tarnocai and Zoltai, 1978).

5.2.3.5 Thufur breakup

The breakup of a thufa involves the disruption of one or more of its sides, with consequent removal of vegetation and soil from the disrupted surface and subsequent disintegration of the thufa (Grab, 1994). Although the rupturing of larger landforms such as frost mounds and pingos has received some detailed attention (e.g. Åkerman and Malmström, 1986; Hinkel, 1988; Pollard, 1991), thufur disruption or degradation has only briefly been mentioned (e.g. Costin and Wimbush, 1973; Hastenrath and Wilkinson, 1973; Van Zinderen Bakker and Werger, 1974; Crampton, 1977; Scotter and Zoltai, 1982; Schunke and Zoltai, 1988; Boelhouwers and Hall, 1990) and empirical data are absent.

Most thufur swarms at the present study sites are disturbed and it appears that in recent years the rate of thufur breakup has accelerated. The percentage of disrupted thufur is commonly between 20 and 30% but may be as high as 63.6% as at the Popple Peak site (Table 5.1). It appears that thufur breakup has a zonal distribution, affected primarily by wetland conditions and consequent grazing pressures.

Approximately 50% of thufur at most sites show slight cracking and splitting during the winter months. This could be the result of frost cracking, and/or internal pressure (Crampton, 1977), and/or desiccation cracking. On some thufur where polygonal cracking has occurred, the soil moisture content near the surface has remained high (over 60% by weight) throughout the dry winter months, and therefore, desiccation cracking is unlikely to play a role. However, at some of the drier sites desiccation may play a significant role in breaking up the thufur. At one site in the Mashai Valley, deep open fissures on thufur apexes have dissected the thufur surfaces. The fissures are 5 to 10 cm deep and correspond to the depth of a desiccated peaty surface. Once this has occurred, polygonally-shaped peaty blocks with sparsely vegetated surfaces may be dislocated from the underlying thufur soils.

At Site Two in the Mohlesi Valley, where approximately 20% of the thufur examined are disrupted (Table 5.1), 62% of those affected are on upslope (south-facing) sides, while only 8% of those affected are on downslope sides (Figure 5.8). Boelhouwers (1991a) considers

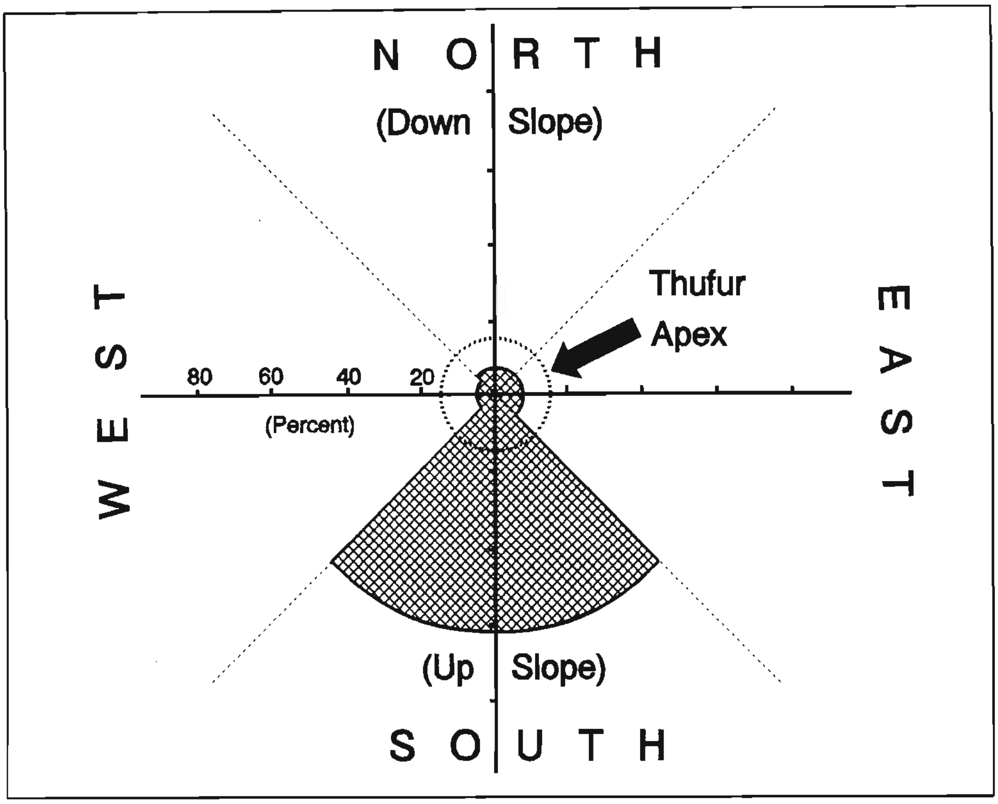


Figure 5.8 Percentage thufur disrupted on various hummock aspects and hummock apexes at Site 2 in the Mohlesi Valley (n = 47) (after Grab, 1994).

creep and solifluction as possible causes for the upslope side of thufur disruption. In most cases, solifluction becomes less pronounced as slope gradients decrease, which therefore limits the breakup on upslope sides of thufur. Those thufur occurring on slope gradients of over 2° show considerable breakup on upslope sides, whereas those found on a 1° slope (Site One - Mohlesi Valley) show no such indication, in spite of abundant surface moisture being available. Findings from Site Two in the Mohlesi Valley also show that 76% of thufur broken on upslope sides occur where surface water and/or ice is present throughout the year. Surface water and ice are believed to contribute significantly towards turf breakup on the upslope and across-slope sides of thufur. Surface ice may become attached to the thufur on upslope and across-slope sides, whereas on downslope sides a small breach occurs between the turf and the surface ice. Several thufur broken on their upslope sides have had soil removed from within the mound, the centre of which is frequently filled with water or ice (Figure 5.9). This may cause downslope pressure on the turf-covered across-slope and downslope sides owing to

Figure 5.9 Several thufur broken on their upslope sides are frequently filled with ice during the winter months.

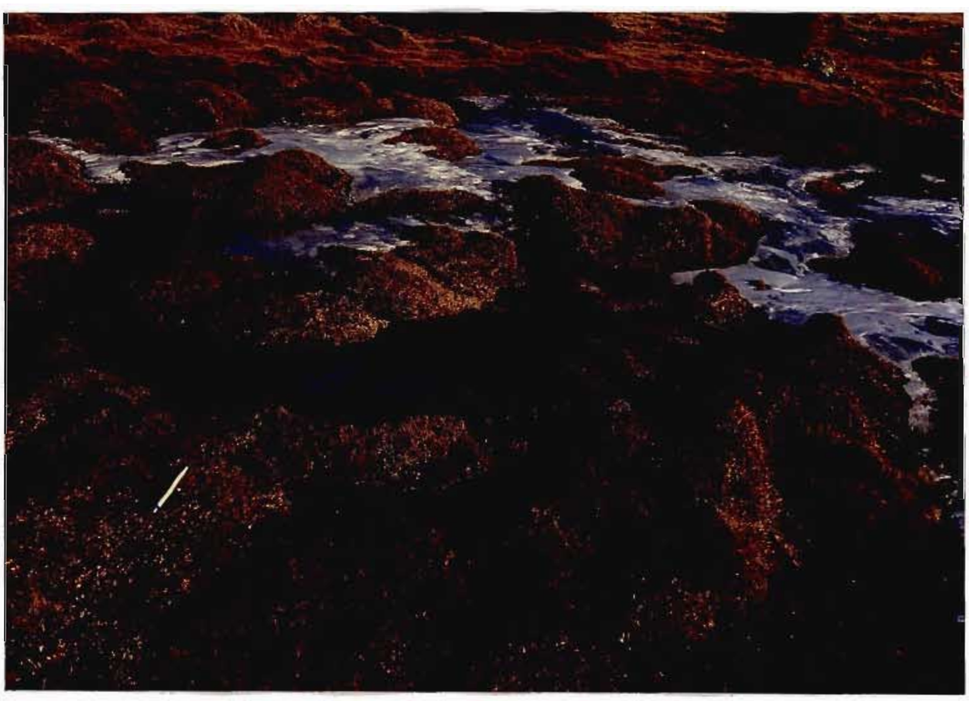


Figure 5.10 Turf is sometimes forced downslope to form small turf-banked steps.

hydraulic forces; in some places the turf is seen to be tilted downslope, thereby contributing towards the formation of turf-banked steps (Figure 5.10), which may be up to 600 cm in length, 107 cm wide and 42 cm high. Needle ice may also play an important role in the breakup of thufur, since it disrupts turf in the inter-thufur depressions and in many places disconnects it from the underlying soil. Van Zinderen Bakker and Werger (1974) suggest that needle ice may be a factor in the development of "crater-like thufur". The disruption of turf in the inter-thufur areas has, in many instances, continued onto the lower parts of adjoining thufur, particularly the south-facing sides where soils remain cold and wet throughout most of the winter. Consequently, the soils are exposed and no longer held together by turf, making them more susceptible to erosion. Lehrsch *et al.* (1991) have shown that aggregate stability may be affected by freezing and therefore it must also affect the soil's response to erosive forces. Van Zinderen Bakker and Werger (1974) also suggest that the westward sides of thufur are commonly broken, and believe this is probably due to wind action, however, observations at Site Two in the Mohlesi Valley show that only 6% of disrupted thufur are broken on the westward sides and 8% on the eastward sides (Figure 5.8). Thus, although wind action is considered to be an important erosional agent in the region, it is not believed to be the primary cause of thufur breakup.

Animal and human interference may also contribute to the disruption of thufur. For instance, Schunke and Zoltai (1988) have shown that entire thufur swarms have been destroyed by the levelling of agricultural lands in Iceland. Scotter and Zoltai (1982) attribute some of the thufur vegetation destruction in the Rocky Mountains to the burrowing by ground squirrels (*Spermophilus columbianus*). Rodents have built burrows and nests amongst thufur at several of the present study sites. Lynch and Watson (1992) mention that "ice rats" (*Otomys sloggetti robertsi*) inhabit "hummocks" in the Lesotho highlands, particularly at altitudes above 2500 m. The ice rats usually inhabit the drier wetland fringes where the desiccated hummocky peatlands provide insulation during the cold winter months (Morris and Grab, 1997). Further, grazing animals favour the grasses/sedges associated with the wetlands and so consequently trample the hummocks (Grab, 1994). Cattle, sheep and horses were observed grazing amongst thufur at several of the study sites, particularly from November to May. The recent drought (1993 - 1995) in the high Drakensberg has resulted in lower water tables and less saturated

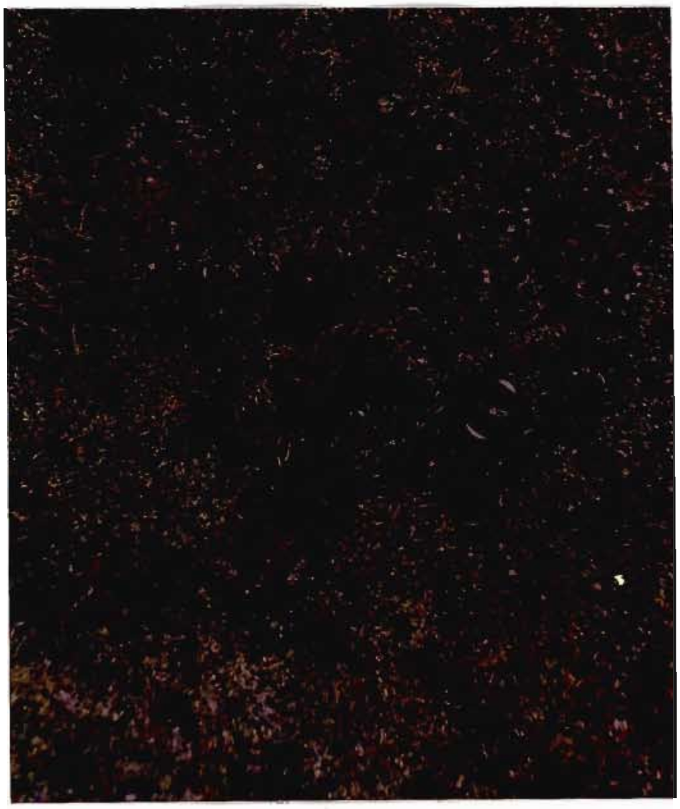


Figure 5.11 A slip scar on a thufa side, owing to livestock trampling within the wetland.

less saturated conditions even in the wetland centres, and consequently, grazing animals have trampled thufur at such sites. The animals try to avoid trampling on the thufur but, where thufur spacing is close, slip scars are frequently seen on thufur sides as a result of trampling (Figure 5.11). Hoofprints, some of them up to 17 cm in depth, were found on several of the thufur. In fact, 14% of the thufur broken open at Site One in the Mohlesi Valley, are broken on their apexes (Figure 5.8), and this is attributed to animal trampling. In some cases water has collected in these hoofprints, while elsewhere needle ice growth has contributed to the breaking up of turf around particular hoofprints and slip scars. At some of the wetter sites, however, renewed vegetation growth covering the original "scars" is evident.

5.2.3.6 Stone-cored thufur

Boelhouwers and Hall (1990) and Hanvey and Marker (1992) have reported stone-cored thufur from sites in the high Drakensberg. Isolated groups of stone-cored thufur were also located at a few of the present study sites. Dionne (1966) examined such thufur varieties and suggests that the thufur develop as a result of boulders migrating to the surface by frost push and pull processes. As the boulders rise towards the surface, the ground heaves to form "grass-mound fields" (Dionne, 1966, p100). Eventually, the grass and organic cover is destroyed, leaving only the exposed boulders (Dionne, 1966).

Several exposed boulders amongst thufur in the Mashai Valley are well covered with lichen, possibly indicating that they have remained exposed at the surface for a considerable time. Such boulders sometimes have a turf-ring at their base (Figure 5.12), as is also reported by Dionne (1966). The turf-rings are possible remnants of the original thufur from an earlier stage when the boulder was still submerged. Examples of the various stages of stone-cored thufur development and subsequent disintegration are evident at the Mafadi Summit site and at sites in the Mashai Valley. An example of an exposed split boulder with turf protruding through the fracture was also found in the Mashai Valley (Figure 5.13).

Figure 5.12 A large boulder with a pronounced turf-ring around its base.

Figure 5.13 An exposed split boulder with turf protruding through the fracture.

5.2.4 Thermal characteristics - 1993/1994

Few observations have been published on the thermal characteristics of frost-induced mounds such as thufur. Although Tarnocai and Zoltai (1978) and Grab (1992a) recorded temperatures from within thufur, these were only taken at one point in time and are therefore unable to show long-term trends. The most detailed study on thufur temperature characteristics is that by Mark (1994) from New Zealand, which compared soil temperatures within a thufa apex and its adjoining depression at 5 and 20 cm depth over an extended period (approximately 12 months). Although Mark's study provides much valuable information, further thufur temperature data are required before unequivocal conclusions can be made regarding their thermal regimes. The objective of the 1993/1994 study was to provide both comparative and new data on the thermal characteristics of thufur apexes and their adjoining depressions in the high Drakensberg. Soil temperatures have never before been recorded continuously over a prolonged period (i.e. more than a few days) in the high Drakensberg. The thufur temperature data presented here, therefore also provide the first thermal assessment of southern African alpine soils.

5.2.4.1 Overview of annual air and soil temperatures in the Mashai Valley

Mean monthly soil temperatures for a thufa apex (@ 12 cm depth) and its adjoining depression (@ 12 cm depth) both follow the mean monthly air temperature cycle (Tables 5.5 and 5.6 and Figures 5.14, 5.15 and 5.16). The mean monthly air temperature reached its maximum value in January (10.2°C) and minimum value in July (-2.9°C) (Table 5.5 and Figure 5.14). In response, ground temperatures reached their mean monthly maximum values in January (13.1°C in the thufa apex; 13.9°C in the depression) and mean monthly minimum values in July (-1.9°C in the thufa apex; 0.3°C in the depression) (Table 5.6 and Figures 5.15 and 5.16). An absolute maximum air temperature of 24.5°C was recorded in October 1993, and an absolute minimum air temperature of -16.9°C was recorded in June 1994 (Table 5.5). The absolute maximum thufa apex and depression ground temperatures, both recorded in January 1994, were 19.1 and 16.5°C respectively (Table 5.6). The absolute minimum ground temperatures for the thufa apex and depression, reached in July 1994, were -5.5 and -0.1°C respectively (Table 5.6). The mean annual air temperature (September 1993 to August 1994) was 5.0°C (Table 5.5) while the mean annual thufa apex and depression ground temperatures were 6.2 and 7.5°C respectively (Table 5.6). Air temperatures drop most significantly from March to June (Table 5.5 and Figure 5.14), which is also reflected by a pronounced drop in ground temperatures (Table 5.6 and Figures 5.15 and 5.16). For instance, the mean monthly air temperature from March (7.8°C) to June (-1.9°C) dropped by 9.7°C and is accompanied by a rapidly increasing number of frost days (Table 5.5). During the same period, the mean monthly thufa apex and depression ground temperatures dropped by 11.3 and 11.2 °C respectively (Table 5.6).

5.2.4.2 Seasonal ground temperature variations for a thufa apex and its adjoining depression

Spring 1993 (September to November)

The mean monthly air temperature increase during the spring months of 1993 was irregular. In fact, early spring (September) was exceptionally warm and late spring (November) unusually cool, such that Novembers' mean air temperature was 0.4°C cooler than that in October (Table 5.5). In spite of this, both the thufa apex and depression ground temperatures

	1993				1994								
	Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	March	April	May	June	July	Aug.	Year
Mean max.	14.4	15.5	14.1	16.2	17.3	15.7	14.8	11.4	9.5	4.8	5.4	7.0	12.2
Mean min.	0.6	3.1	2.7	5.5	5.8	4.3	2.6	-0.3	-3.1	-6.5	-8.5	-5.2	0.1
Mean	6.2	8.0	7.6	9.9	10.2	9.3	7.8	4.6	2.1	-1.9	-2.9	-0.4	5.0
Absolute max.	18.5	24.5	21.5	22.6	21.9	19.8	18.1	15.6	12.9	9.4	12.3	14.6	24.5
Absolute min.	-5.3	-0.1	-1.8	0.9	3.3	0.9	-3.3	-5.5	-13.0	-16.9	-15.3	-16.6	-16.9
Frost days	14	1	6	-	-	-	4	13	27	29	31	27	152

Table 5.5 Summary of the Mashai Valley air temperature data.

		1993				1994								
		Sept.	Oct.	Nov.	Dec.	Jan.	Feb.	March	April	May	June	July	Aug.	Year
Mean max.	Apex	8.01	11.56	13.75	14.98	16.39	15.18	13.00	8.43	3.01	-0.60	-1.07	-0.58	8.50
	Depression	5.18	8.89	11.43	13.09	14.88	14.71	13.28	9.66	5.05	1.14	0.33	0.37	8.16
Mean min.	Apex	2.30	5.64	6.84	8.89	10.02	9.25	7.42	3.54	0.87	-1.91	-2.93	-1.31	4.05
	Depression	3.43	7.19	9.33	11.30	12.93	12.57	11.20	7.62	3.80	1.06	0.20	0.36	6.75
Mean	Apex	4.86	8.42	10.19	11.97	13.14	12.10	10.11	6.00	1.91	-1.21	-1.94	-1.00	6.21
	Depression	4.28	8.00	10.33	12.17	13.88	13.67	12.31	8.76	4.49	1.10	0.30	0.36	7.47
Absolute max.	Apex	12.60	14.40	18.50	17.80	19.10	17.80	15.90	13.10	6.70	-0.10	-0.60	-0.30	19.10
	Depression	7.50	11.40	14.40	14.60	16.50	16.50	15.00	13.50	7.50	2.50	0.40	0.40	16.50
Absolute min.	Apex	0.20	3.30	3.00	5.90	7.80	7.20	4.80	1.20	-1.00	-4.30	-5.50	-3.80	-5.50
	Depression	1.20	4.80	5.90	9.40	12.00	11.10	10.30	4.80	2.00	0.70	-0.10	0.20	-0.10
Mean daily range	Apex	5.71	5.87	6.88	6.09	6.37	5.84	5.58	4.89	2.20	1.31	1.89	0.73	4.45
	Depression	1.74	1.72	2.08	1.75	1.95	2.13	2.08	2.04	1.25	0.07	0.12	0.01	1.41
Sub-zero days	Apex	-	-	-	-	-	-	-	-	8	30	31	18+	~87
	Depression	-	-	-	-	-	-	-	-	-	-	6	-	6
Mean daily max. rate of change (°C/4.8hrs)	Apex	5.16	4.00	4.67	4.10	4.72	4.96	5.18	4.51	1.56	1.05	1.43	0.69	3.50
	Depression	1.36	1.35	1.88	1.50	1.63	1.65	1.65	1.28	0.90	0.09	0.15	0.01	1.12
Absolute max. rate of change (°C/4.8hrs)	Apex	8.00	5.70	7.60	7.00	8.30	8.10	7.90	7.50	3.70	2.50	3.30	2.20	8.30
	Depression	2.10	2.20	3.10	2.70	3.00	2.70	2.80	2.60	1.60	0.50	0.30	0.20	3.10

Table 5.6 Summary of the thufa apex and depression soil temperature data at 12 cm depth, Mashai Valley.

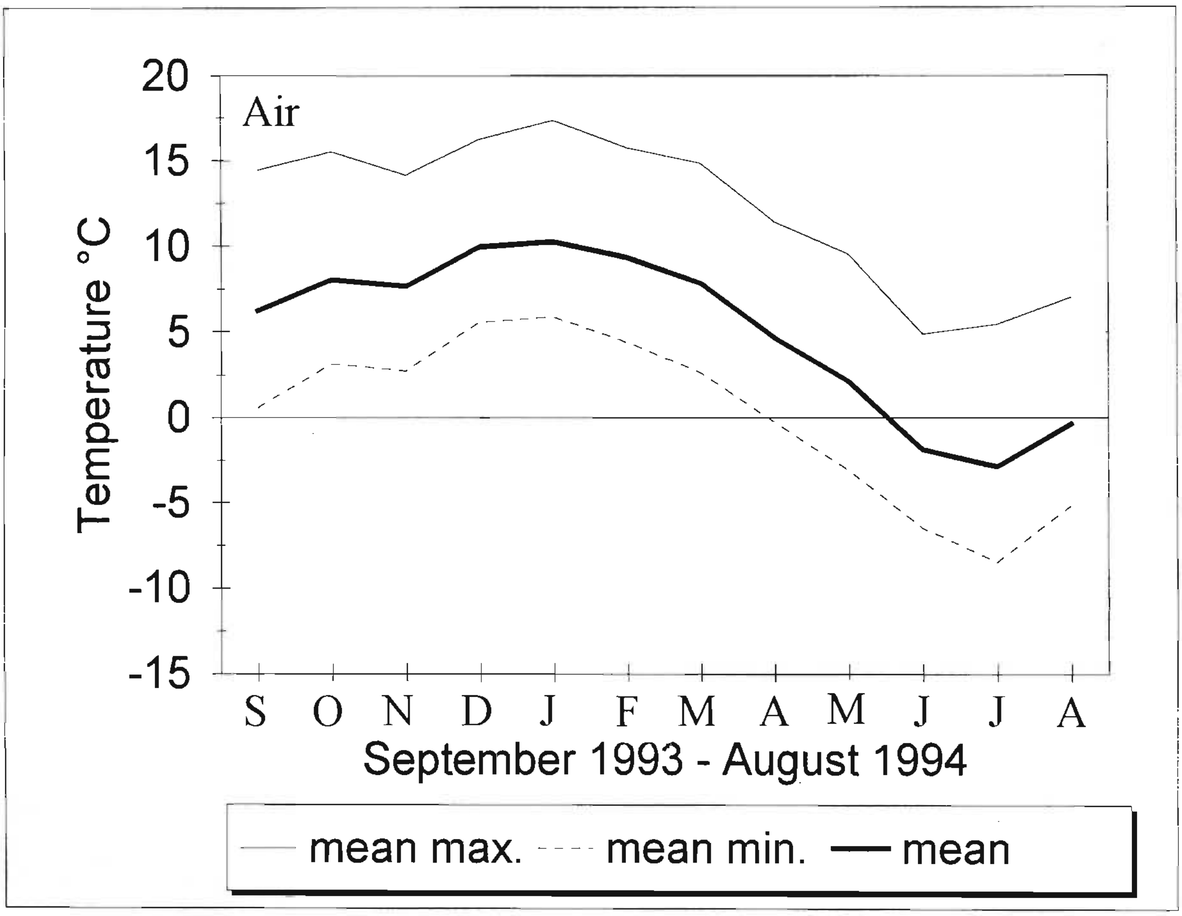


Figure 5.14 Mashai Valley air temperature trends from September 1993 to August 1994.

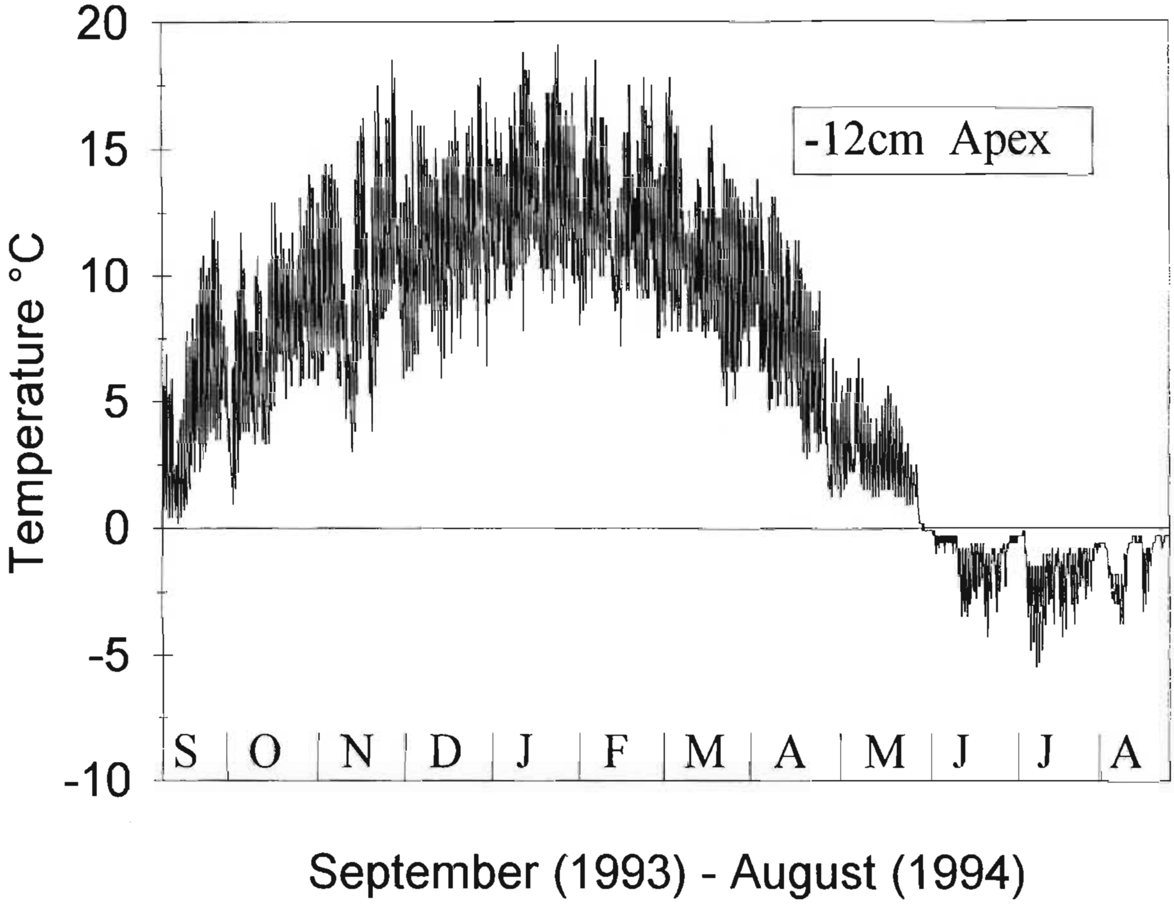


Figure 5.15 Thufa apex temperature trends from September 1993 to August 1994.

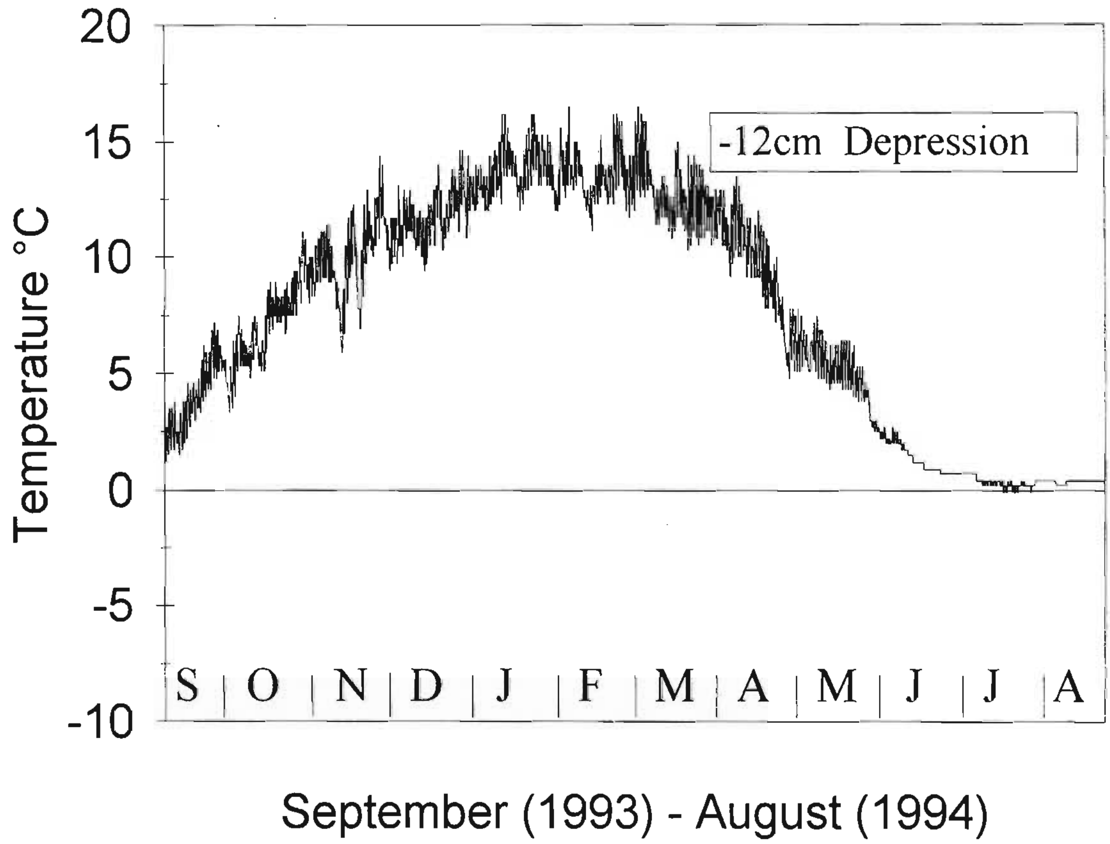


Figure 5.16 Thufa depression soil temperature trends from September 1993 to August 1994.

increased progressively during the spring months. The most pronounced ground temperature increase occurred during September and October when the mean monthly ground temperatures increased by up to 3.7°C per month (Table 5.6); possibly a function of increased radiation. In response to the slightly cooler conditions in November, ground temperatures rose less rapidly than during the two previous months (Table 5.6).

Although the mean ground temperatures in the depression during early and mid-spring (Sept. and Oct.) was slightly cooler than the mean thufa ground temperature (Table 5.6), by November the mean depression temperature (10.3°C) had fractionally exceeded the mean apex temperature (10.2°C). This trend can also be observed in Figure 5.17, which compares the apex and depression ground temperatures during 4 days in September (early spring) with 4 days in November (late spring). Here it can be seen that during September, ground temperatures in the depression were considerably lower than in the apex. Although the daily minimum apex ground temperatures dropped slightly lower than depression ground temperatures (by 1 to 1.5°C), the daily maximum temperatures for the apex were about 5°C higher than the maximum depression temperatures (Figure 5.17). By November, depression temperatures had risen sharply and, although the daily apex temperatures still varied more than depression temperatures, the mean daily temperatures were similar for the apex and depression (Figure 5.17).

September and October were the only months when mean depression temperatures were lower than the mean apex temperatures during the 12 months recording period. It would appear that the depression ground temperatures respond more slowly to diurnal and seasonal air temperature changes than do thufur apex ground temperatures. Therefore, the more rapid warming of thufur apex temperatures during early spring resulted in warmer apex soils than depression soils during this period.

Summer 1993/4 (December to February)

Ground temperatures continued to rise until January, when they reach their peak (Table 5.6 and Figures 5.15 and 5.16). The cooler air temperatures during February (Table 5.5 and Figure 5.14) account for an immediate and considerable response in the mean monthly thufa

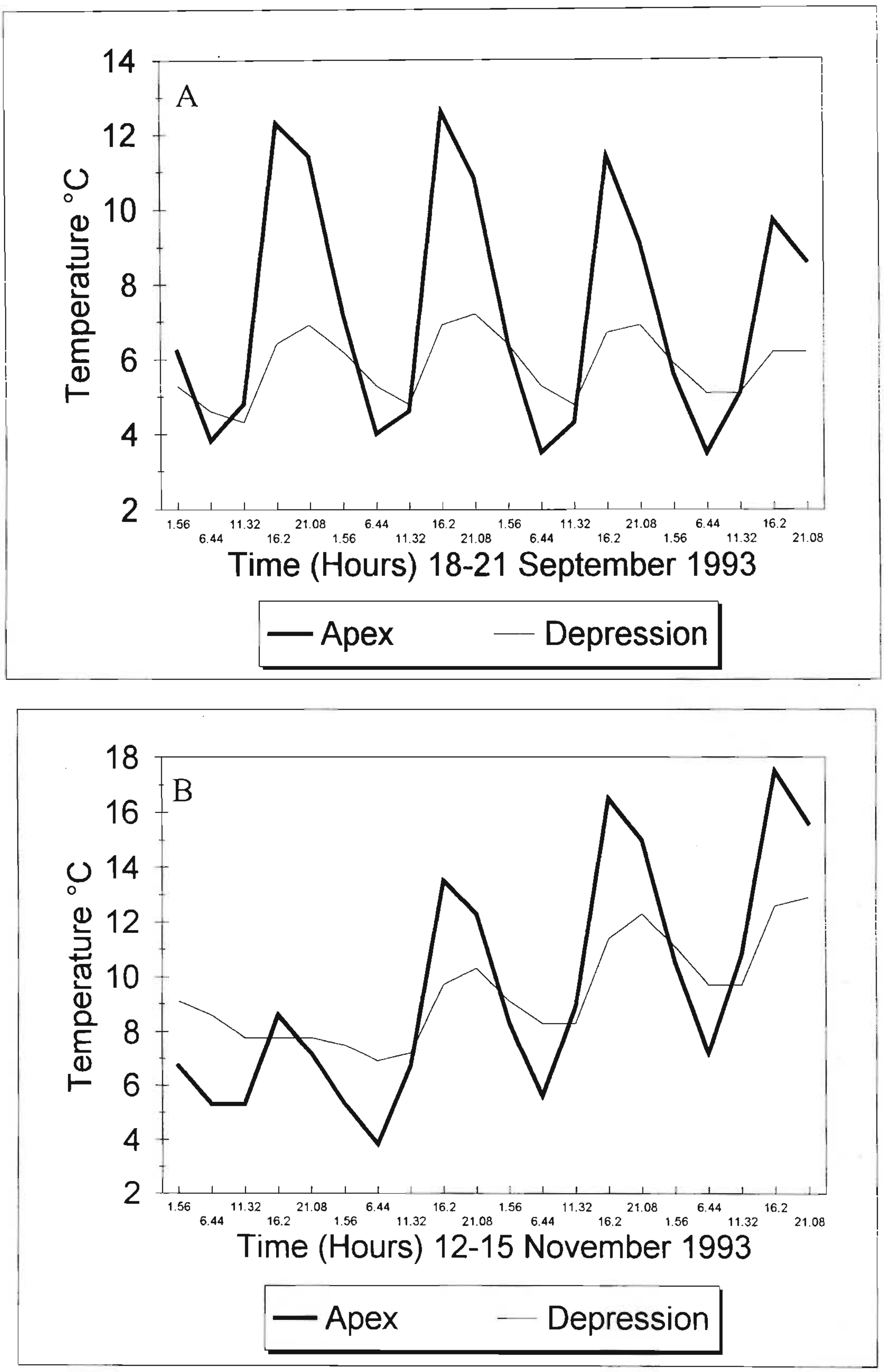


Figure 5.17 Comparing the thufa apex and depression temperatures during (A) early spring and (B) late spring.

apex ground temperatures, which dropped by 1.0°C from January to February (Table 5.6). The depression ground temperatures, however, show a delayed and gradual response to the early seasonal cooling and dropped by only 0.2°C from January to February. This resulted in continuously higher mean depression temperatures than apex temperatures during this period (Table 5.6 and Figure 5.18). Figure 5.18 shows that towards the end of summer, depression temperatures had mean daily maximum temperatures only slightly (less than 1°C) lower than those for the thufa apex, while the mean daily minimum temperatures were significantly (4 to 5°C) higher than those for the apex. The amplitudinal trend of daily maximum and minimum temperature values for the apex and depression had therefore reversed since early spring.

Autumn 1994 (March to May)

Rapid and progressive seasonal cooling of soils occurs from March to May, with the most significant drops in temperatures being recorded in mid-April and again at the end of May. The thufa apex ground temperatures continued to respond more rapidly to the seasonal cooling than did those in the depression (Table 5.6). This is also shown in Figure 5.19, which compares ground temperatures over 4 days in early autumn (March) with those over 4 days in late autumn (May). It is evident that by mid-May, the apex ground temperatures had dropped so significantly, that they were continuously lower than the depression temperatures.

Towards the end of May, mean depression temperatures were still between 2 and 3°C while the apex temperatures dropped sharply to -0.2°C (Figures 5.20 and 5.21). It would appear that, as the thufa apex ground temperatures remained at -0.2°C for several days, this represents a phase change of soil moisture from liquid to frozen state (Figure 5.20). Williams and Smith (1989) have indicated that soil freezing usually commences at just below 0°C (at about -0.1°C). C. Harris (1981) and Williams and Smith (1989) have also explained that soil temperatures may remain constant ("zero curtain") until pore water is frozen, after which soil temperatures may continue to fall. This evidently appears to have taken place in thufur apex soils towards the end of May, when the insulating effect of snow was still absent (Figure 5.20).

Winter 1994 (June to August)

The recorded soil temperatures for the thufa apex during the three winter months remained below 0°C and reached a minimum of -5.5°C (Table 5.6 and Figures 5.20, 5.22 and 5.24). It would appear, however, that towards mid-August the "zero curtain" was again reached at times with possible successive freeze and thaw events as air temperatures began to slowly rise (Figure 5.24.). A total of 87 sub-zero days were recorded during the logging period of 350 days (at 12 cm depth within the apex) (Table 5.6).

The rate of cooling in the thufa depression is significantly slower than in the apex, and the "zero curtain" appears to be considerably prolonged (20 days) (Figures 5.21 and 5.23). It would appear that the depression temperatures remained around the "zero curtain" for several days in July (Figure 5.23). The depression temperatures were almost constant at 0.4°C during August, possibly owing to an insulating snow cover or "zero curtain" effect (Figure 5.25). Only six sub-zero days were recorded for the depression soils at 12 cm depth (Table 5.6).

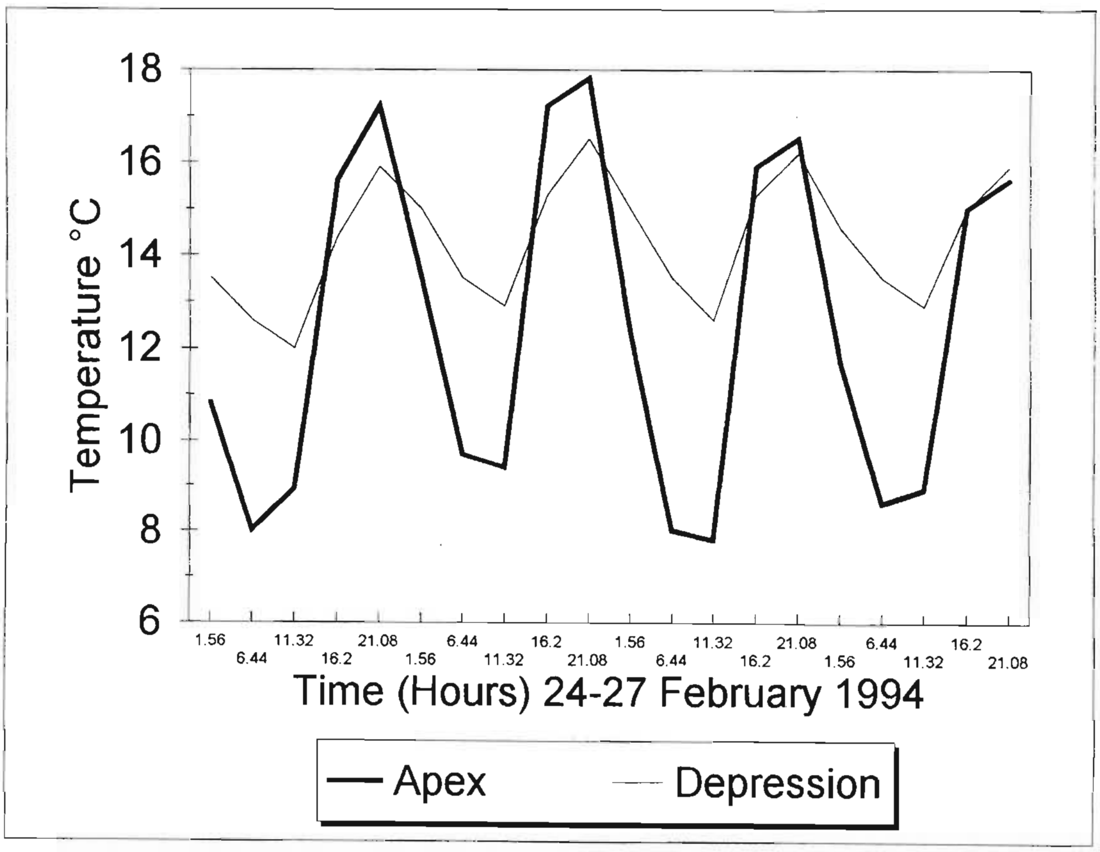
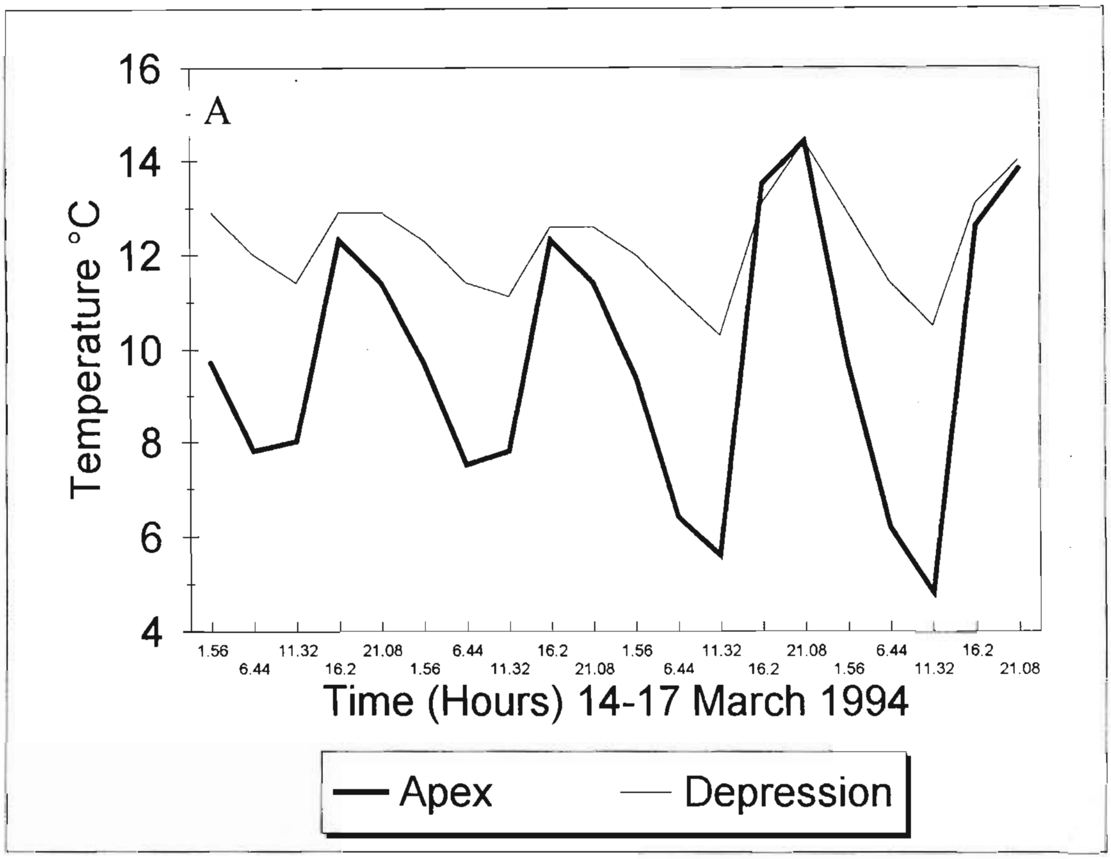


Figure 5.18 Comparing the thufa apex and depression temperatures towards the end of summer (February 1994).



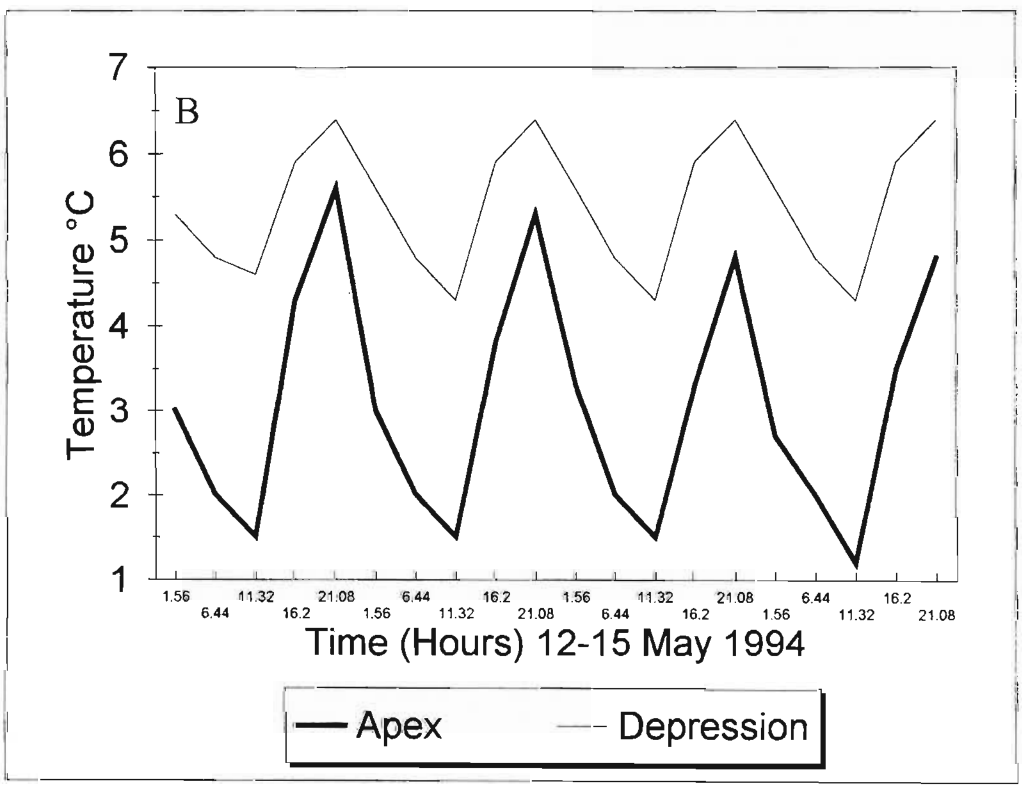


Figure 5.19 Comparing the thufa apex and depression temperatures during (A) early autumn and (B) late autumn.

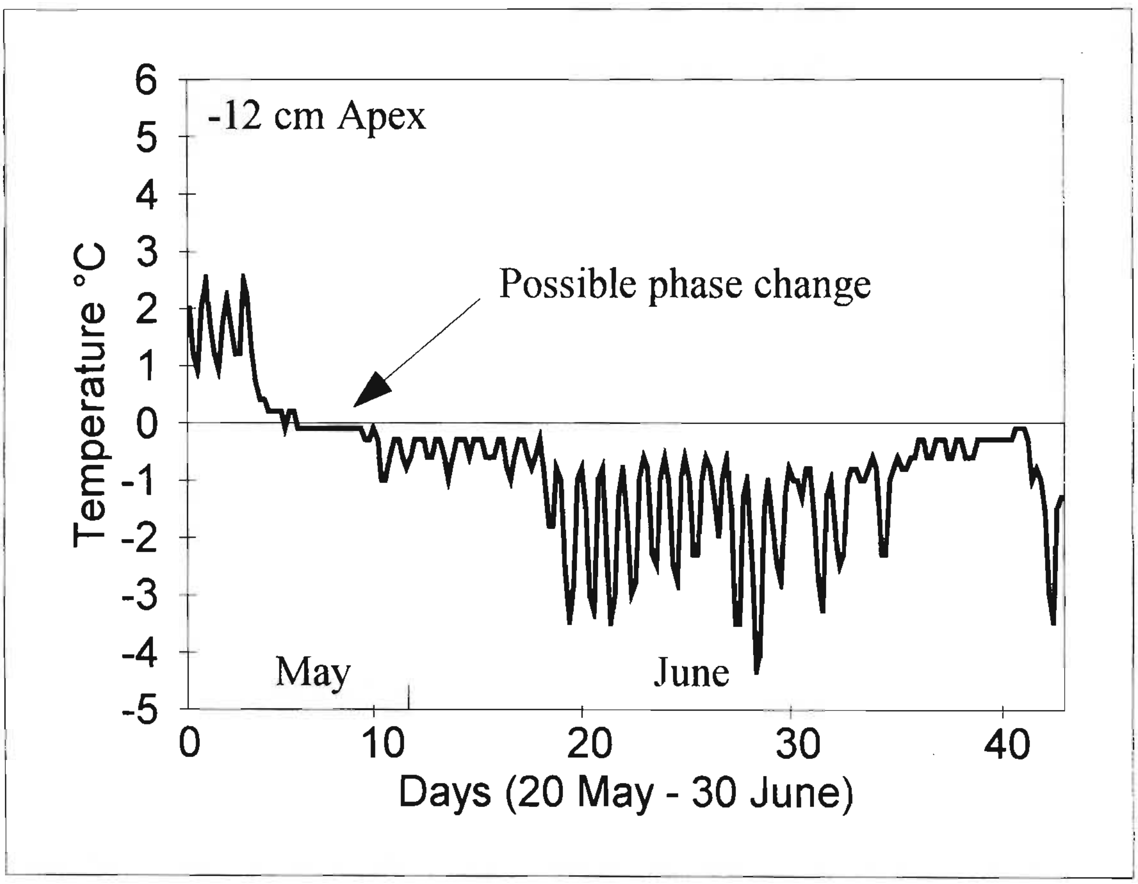


Figure 5.20 Thufa apex temperature trends from 20 May to 30 June, 1994.

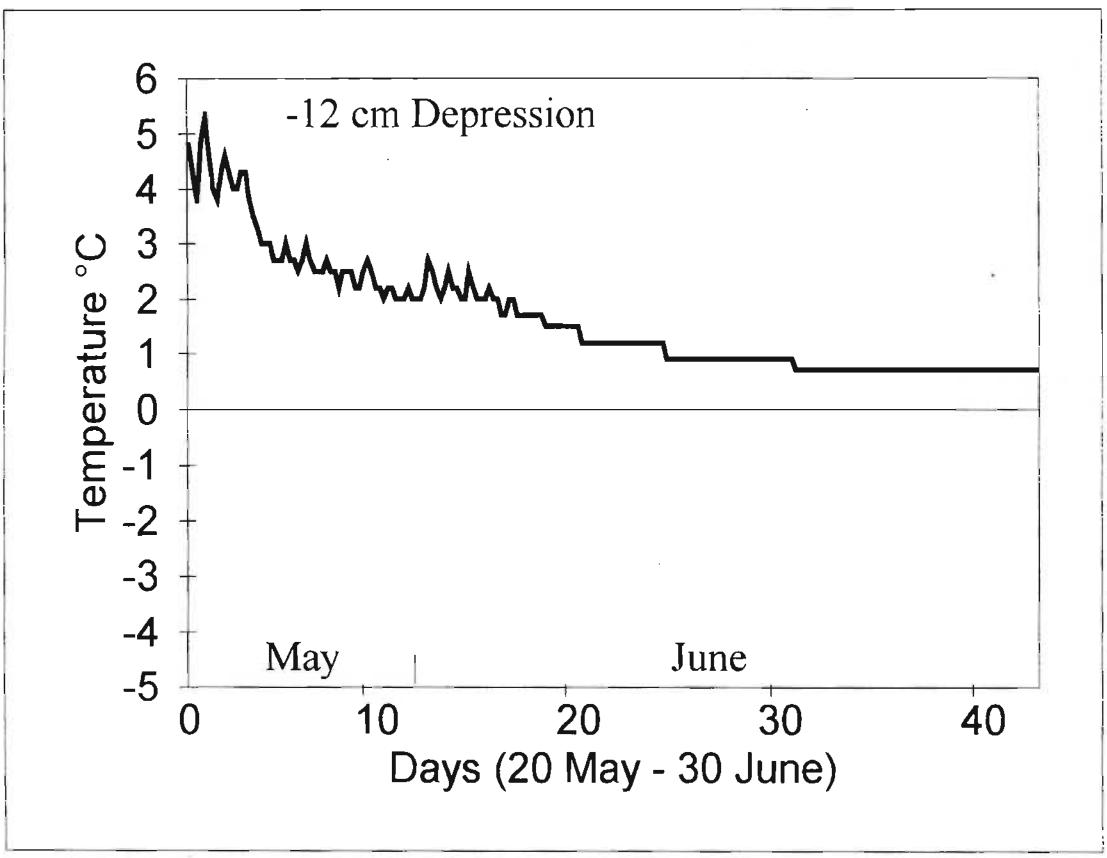


Figure 5.21 Thufa depression temperature trends from 20 May to 30 June, 1994.

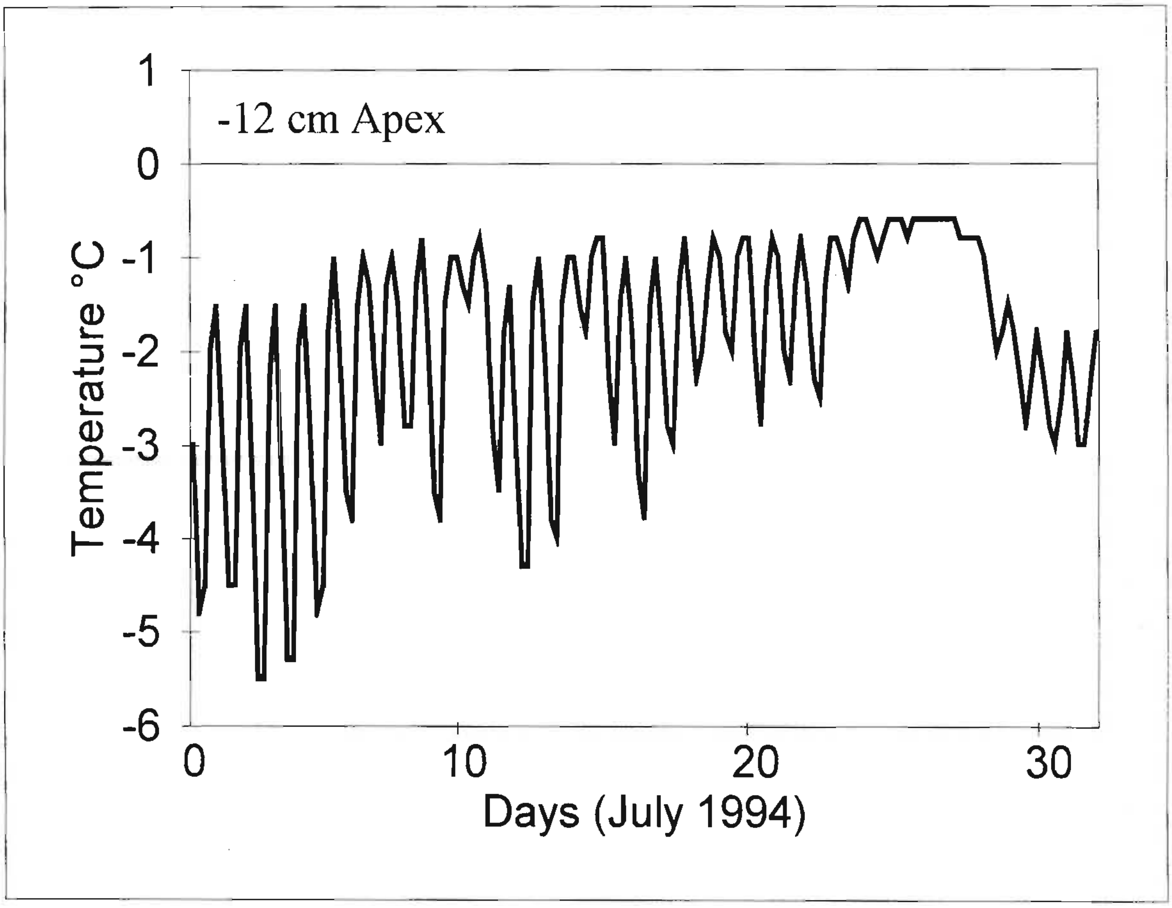


Figure 5.22 Thufa apex temperature trends during July, 1994.

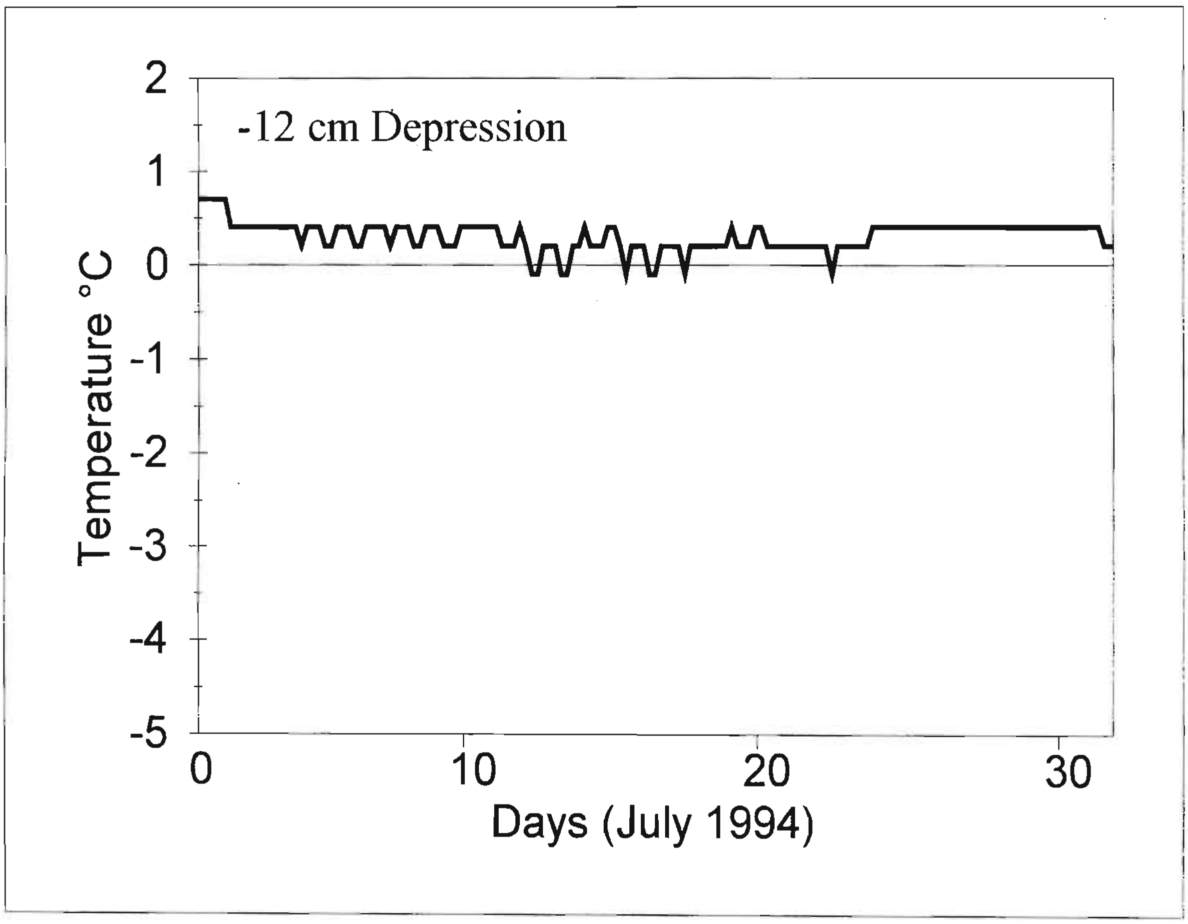


Figure 5.23 Thufa depression temperature trends during July, 1994.

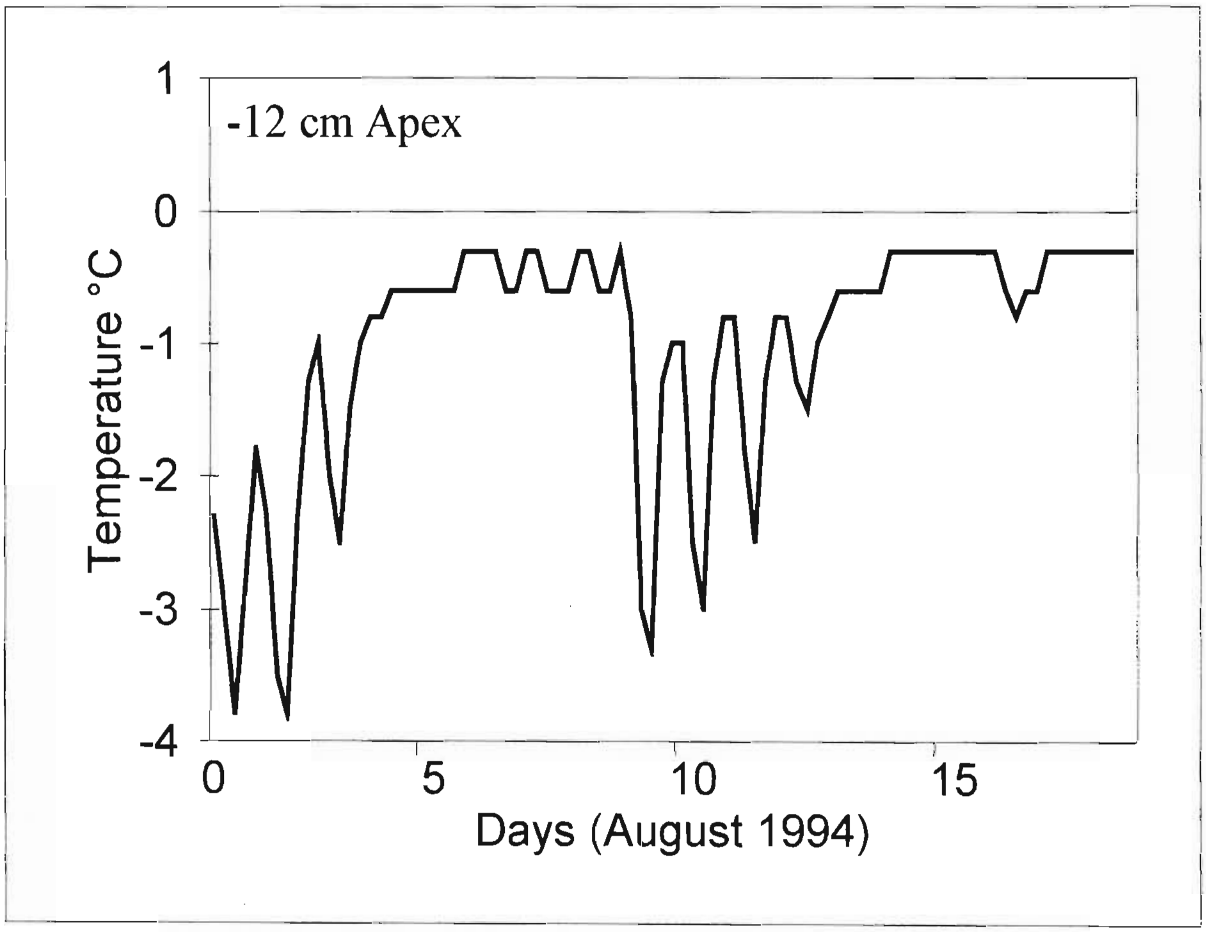


Figure 5.24 Thufa apex temperature trends during early August, 1994.

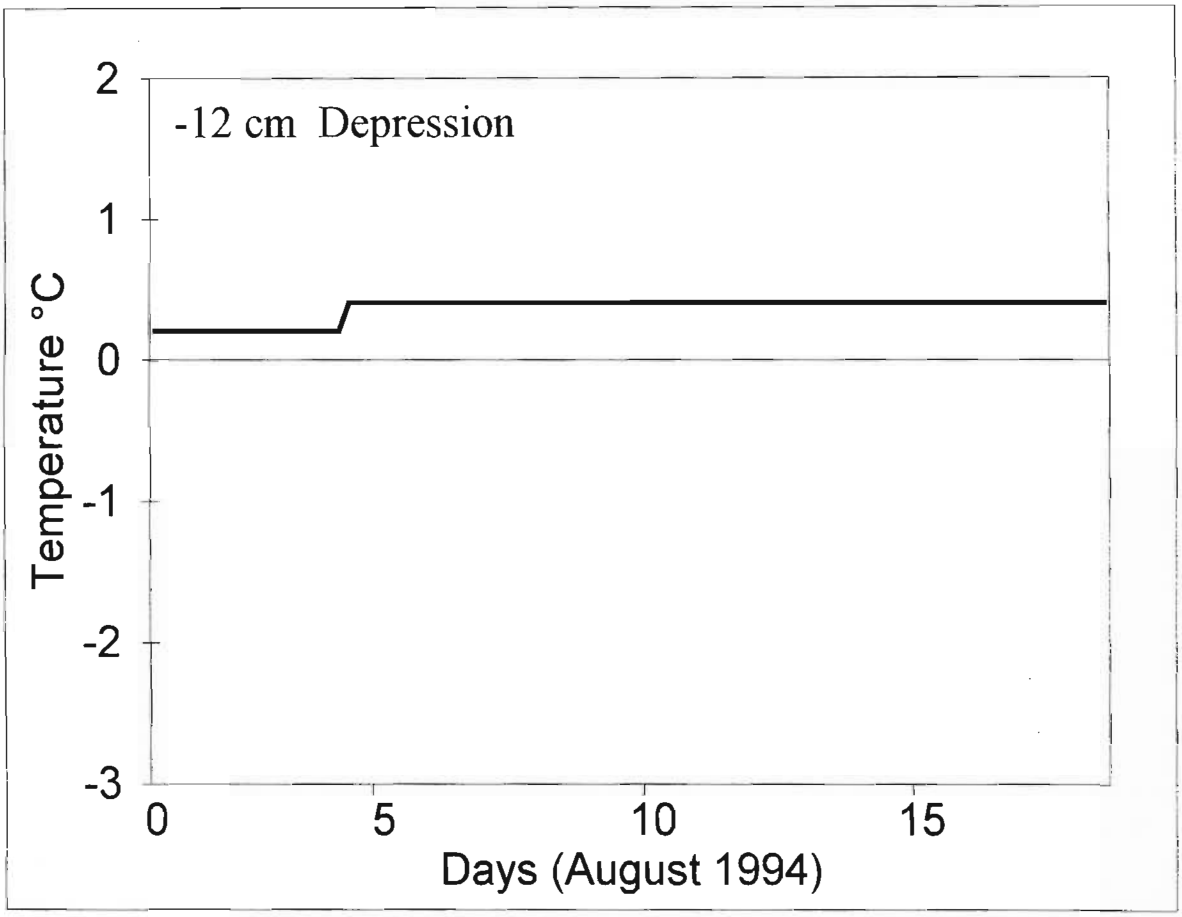


Figure 5.25 Thufa depression temperature trends during early August, 1994.

5.2.4.3 General trends in ground temperature

Mean daily temperature range

The mean daily range in soil temperature is more pronounced in the thufa apex (mean = 4.5°C) than in the depression (mean = 1.4°C) throughout the year (Table 5.6 and Figure 5.26). The highest mean daily range in soil temperature (6.9°C) occurred in the apex during November, while the lowest (0.0°C) occurred in the depression during August (Table 5.6 and Figure 5.26). It is noticeable from Figure 5.26, that as the mean daily range in air temperature is diminished during late autumn and summer months, the mean daily range in the apex soil temperature increases (Figure 5.26). Conversely, when the mean daily range in air temperature increases during some autumn and winter months, the mean daily range in soil temperature for the thufa apex and depression is markedly reduced (Figure 5.26). In part, this may be as a result of the seasonal variations in soil moisture content (Figure 5.7). It is possible that the higher soil moisture content during the summer months increases the thermal conductivity, thereby contributing towards a greater daily range in soil temperatures than during the drier autumn and winter months (Figures 5.7 and 5.26.).

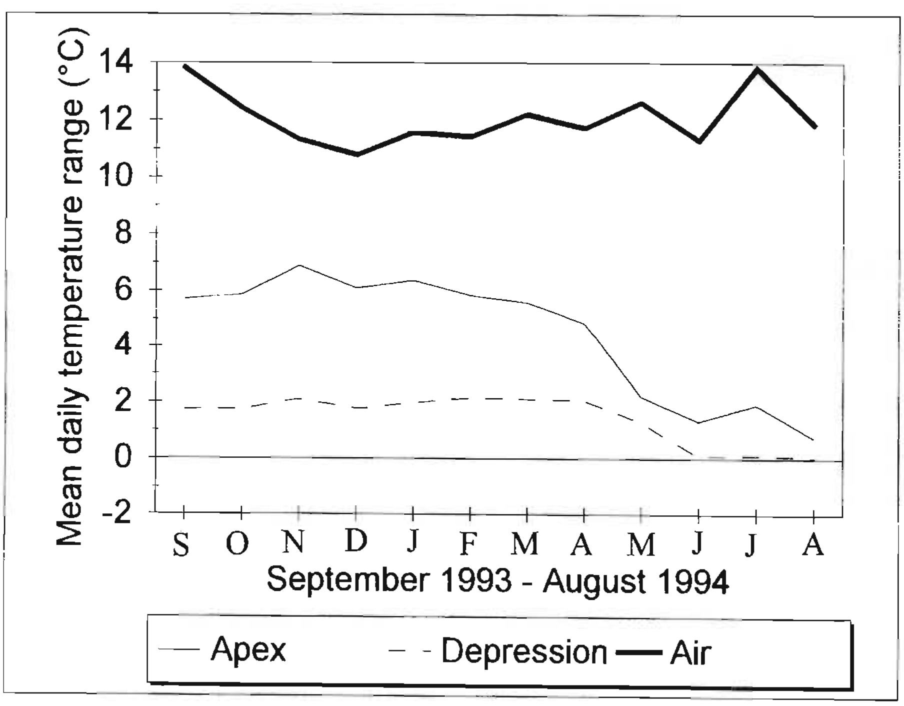


Figure 5.26 Mean daily range in air, thufa apex and depression temperatures from September 1993 to August 1994.

Ground temperature response time

The mean daily maximum rate of temperature change is greater in the thufa apex than in the depression throughout the year (Table 5.6 and Figure 5.27). The absolute maximum rate of temperature change recorded for the apex (@ 12 cm depth) was 1.7°C/hr, while that for the depression (@ 12 cm depth) was 0.6°C/hr. The overall increase in rates of temperature change during the summer months (Table 5.6 and Figure 5.27) may be attributed to the greater thermal conductivity during this period.

The response of apex and depression ground temperatures to changing air temperatures over a period of eight days was assessed in detail during the last week in July 1994 (Figure 5.28). The air temperature graph shows progressive daily cooling from the 24th to 27th July and a subsequent gradual rise in air temperature from the 28th to 31st July (Figure 5.28). Ground temperatures at 12 cm depth in the apex show a delayed response to the air temperature fluctuations. In response to mean daily air temperature cooling, which began on the 24th July, the mean daily apex temperatures started dropping on the 28th July (Figure 5.28). Daily mean apex temperatures then continued to drop until the 30th July, during which time the mean daily air temperature was increasing (Figure 5.28). The depression temperature, possibly around the "zero curtain", remained unaltered during this period.

Figure 5.29 shows the number of days during each month against the time of recorded minimum and maximum temperatures at 12 cm depth for a thufa apex and its adjoining depression. The time of recorded minimum apex temperatures during the warmer months (October to February) was usually at 06h44, while during the autumn months an increasing number of days logged minimum recorded temperatures at 11h32 (Figure 5.29). In part, this is probably as a result of the higher rates of ground temperature change during the warmer, wetter months. This may also be attributed to the prolonged cooling during autumn mornings when the time of first insolation received on south-facing slopes is considerably delayed. The time of recorded minimum temperatures in the depression is usually at 11h32 throughout the year. Only from October to December are minimum temperatures commonly recorded at 06h44 in the depression (Figure 5.29). This again demonstrates the very much slower temperature response time and rate of temperature change in the depression, compared with that in the apex. Such trends are reiterated when examining the times of recorded maximum temperatures (Figure 5.29). The time of recorded maximum apex temperatures is usually at 16h20, except during April and May, when it is more frequently recorded at 21h08 (Figure 5.29). Conversely, the time of recorded maximum depression temperatures is usually at 21h08, except during November and December, when it is commonly recorded at 16h20 (Figure 5.29).

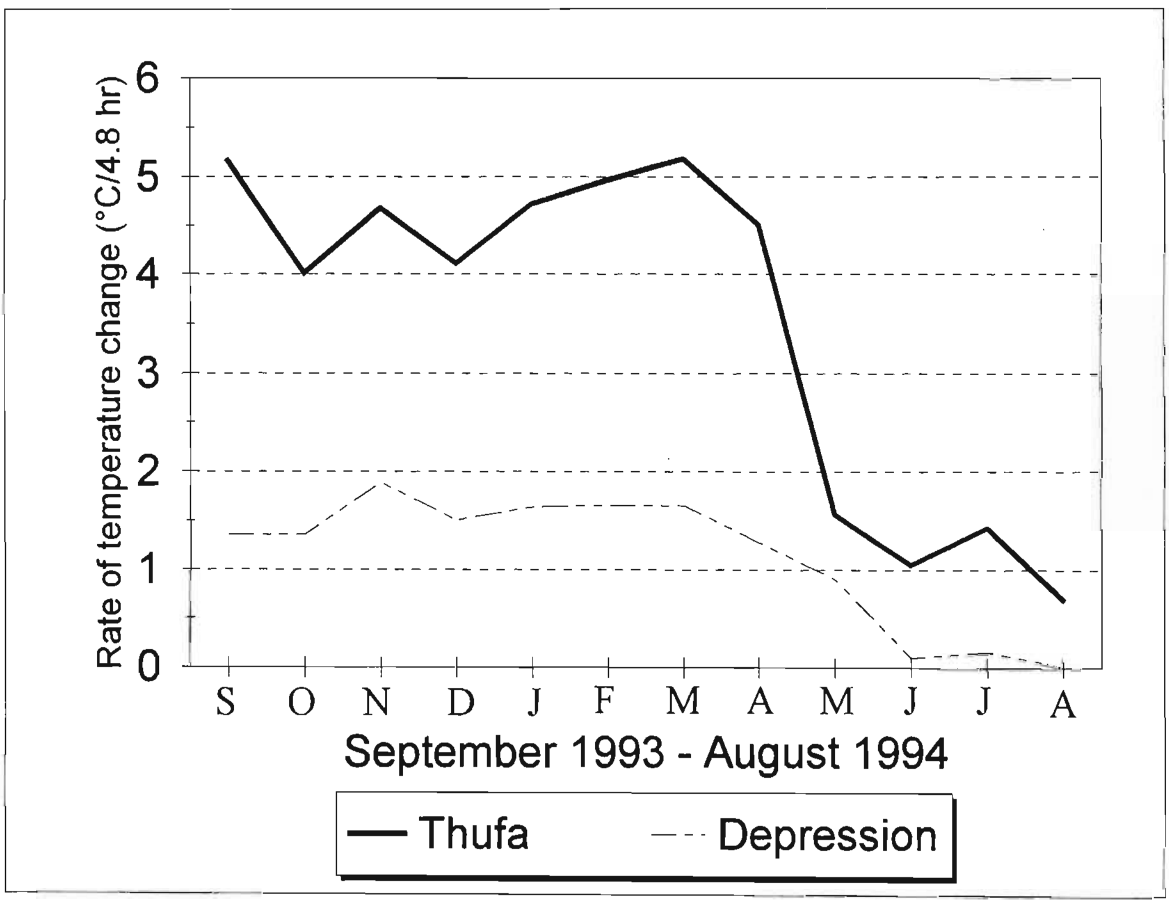


Figure 5.27 Mean daily maximum rate of temperature change for a thufa apex and depression.

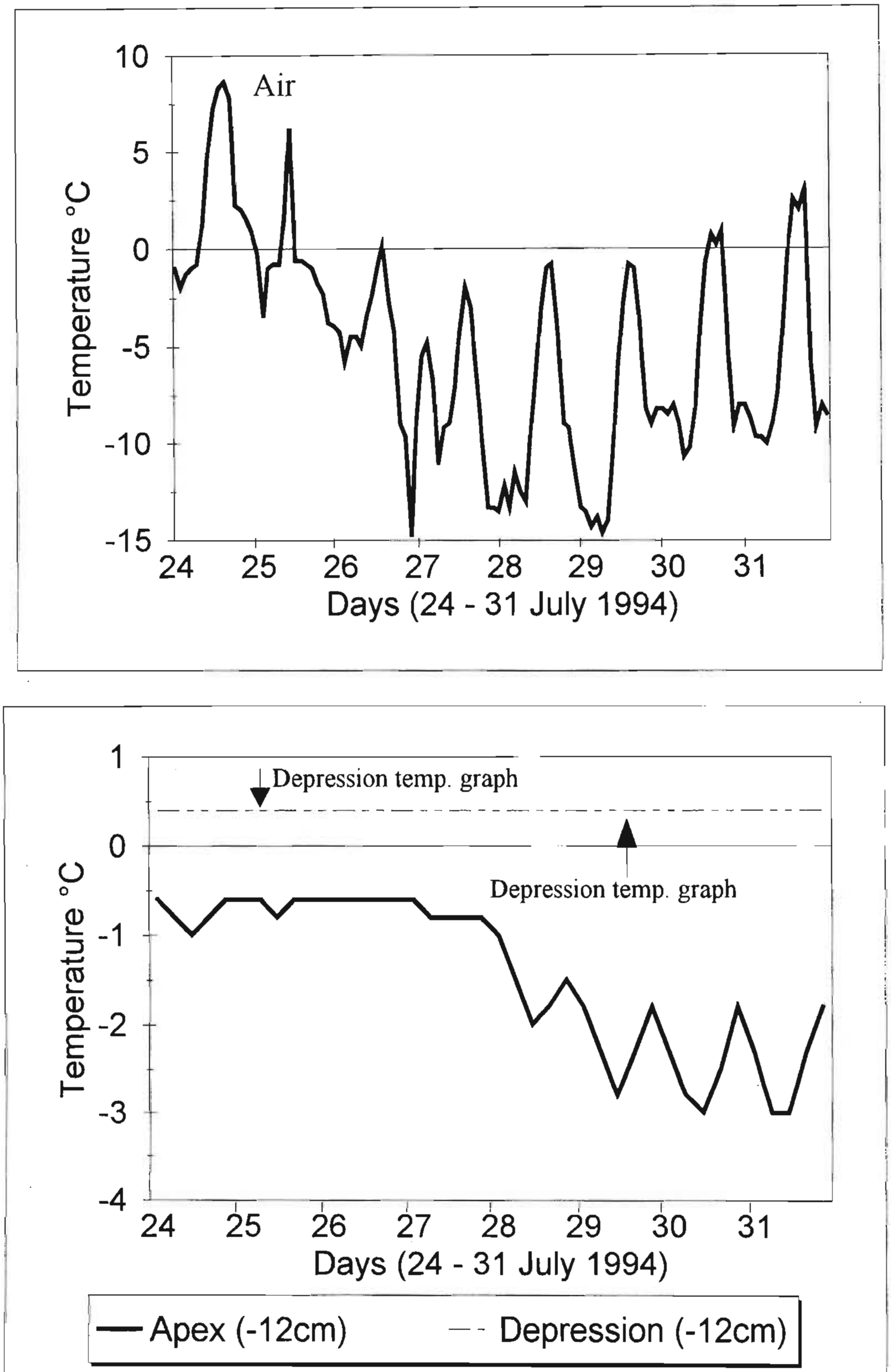


Figure 5.28 The response of the thufa apex and depression soil temperatures to changing air temperatures over a period of eight days during July 1994.

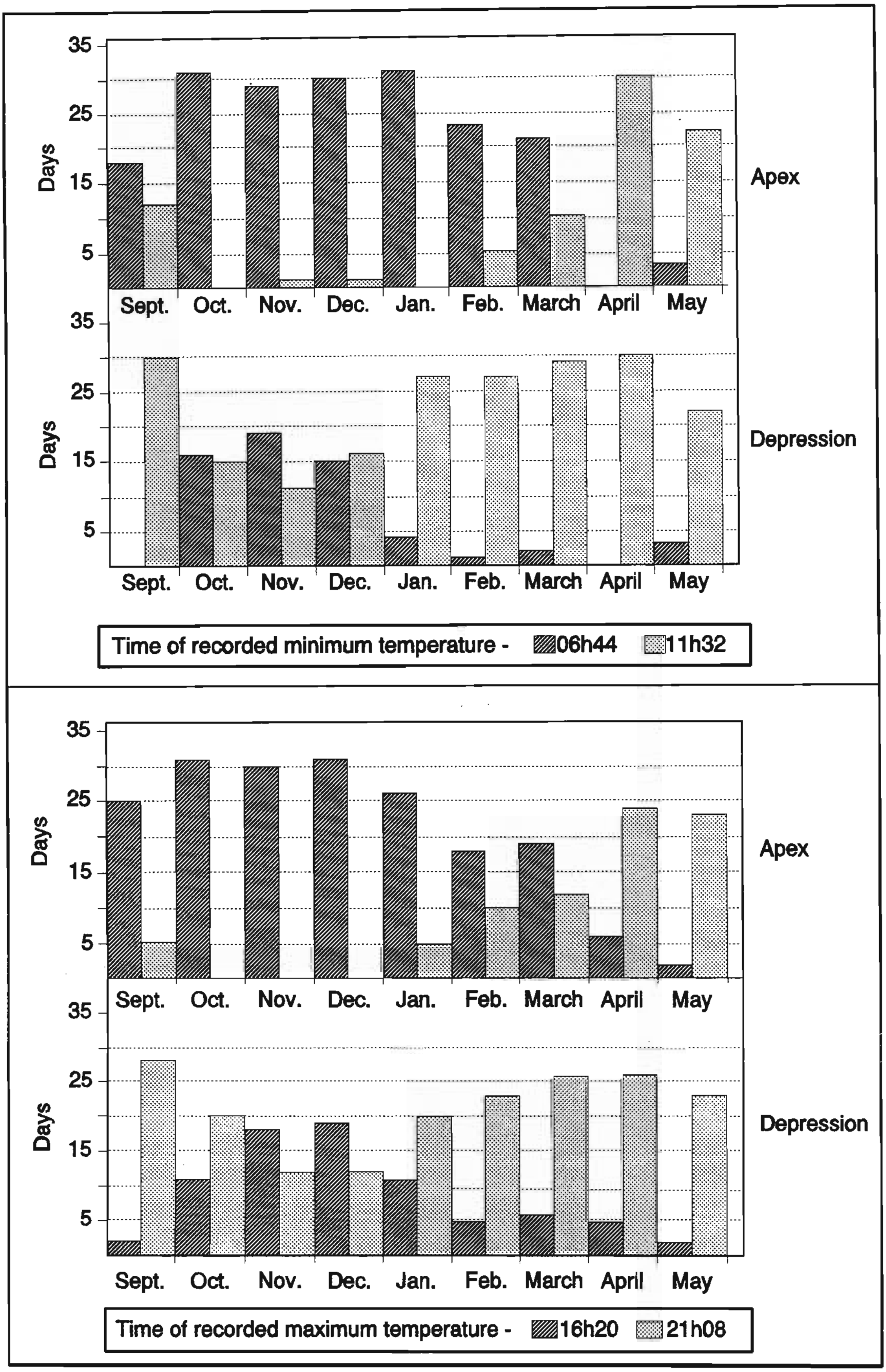


Figure 5.29 The number of days during each month against the time of recorded minimum and maximum temperatures at 12 cm depth for a thufa apex and its adjoining depression.

5.2.5 Thermal characteristics - 1995/1996

A substantial amount of recent literature has focussed on the ground thermal regime of periglacial environments and cryogenic landforms as a means to better describe landform genesis and examine various associated environmental implications (e.g. Josefsson, 1990; Harris, 1995; Wang and Allard, 1995, Ballantyne, 1996). It appears, however, that much of this research has focussed primarily on ground temperature variations within permafrost regions (e.g. Smith, 1975; Sone and Takahashi, 1993; Pavlov, 1994; Romanovsky and Osterkamp, 1995; Beltrami, 1996; Thórhallsdóttir, 1996; Osterkamp and Romanovsky, 1996). Significantly, such as Brown (1978) and Williams and Smith (1989), have acknowledged that a considerable variation in ground thermal conditions may occur within a small area, owing largely to varying environmental factors such as vegetation, soil moisture, snow cover and microclimate. However, it appears that information pertaining to ground thermal variations within areas of a few square metres or centimetres is relatively scarce. Nelson *et al.* (1985) have briefly examined the aspect-controlled temperature variations within a palsa and Seppälä (1994) monitored the temperature differentials between the snowcovered and non-snowcovered aspects of a palsa. Other than the single record of northern and southern aspects of a thufa taken by Kojima (1994), such detailed studies have yet to be undertaken on smaller scale mounds such as thufur. It was considered that the microtopography of such frost-induced mounds might promote thermal changes over a few centimetres within the ground.

Thufur thermal regime and energy balance studies are required on a detailed spatial scale to help improve our understanding of thufur development. In addition, such a study could indicate the indirect control of microtopography in promoting ground thermal changes within a single hummock. The objective of the 1995/1996 study was, therefore, to assess the temperature variations between different aspects and depths within a thufa, in order to broaden the understanding of the freeze and thaw processes within thufur in alpine environments.

5.2.5.1 General trends in ground temperature and freezing: aspect and depth controls

Data from 1995

During the 1995 winter, temperatures were logged at 15 and 20 cm depth on the northern, southern, eastern and western aspects of a thufa. Pronounced temperature differentials occur on varying aspects at different depths within the thufa (Figures 5.30 and 5.31 and Table 5.7). Soil freezing commenced on the 23 May (Figure 5.30). The higher insolation receipts on the thufa northern and eastern exposures permits only shallow diurnal freeze during the onset of winter. Gradual freeze penetration then occurred through the thufa from the southern aspect towards the western and northern aspects (Figures 5.30 and 5.31). Owing to the high insolation receipt and mean daily maximum temperature of 2°C at 15 cm depth on the north-facing aspect, indications are that freeze does not penetrate the thufa from the surface on this aspect, but rather from the southern and western aspects. It took 13 days for the freeze to penetrate from 15 cm depth to 20 cm depth on the southern aspect, indicating slow freeze fringe migration during the onset of winter (end of May) (Table 5.7). Freeze penetration became more rapid during the first half of June, such that the freeze fringe only took four days to progress from 15 cm depth to 20 cm depth on the western aspect (Table 5.7). The ground temperature on the sun-exposed eastern aspect of the thufa remained unfrozen at both 15 and 20 cm depth for the entire winter period.

The lowest recorded thufa temperature of -2.3°C was measured in July at 15 cm depth on the southern aspect (Figure 5.30 and Table 5.7). The lowest temperature at 20 cm depth was -1.3°C, recorded on the western aspect during July (Figure 5.31 and Table 5.7). The ground at 15 cm depth on the southern aspect remained frozen for the entire logging duration, which terminated on the 15 September. This continuous frozen state is also reflected by a high "freeze day" (days during which the maximum temperature does not exceed 0°C) index of $\geq$107 days at 15 cm depth on the thufa southern aspect (Table 5.7). The "freeze day" and "freeze index" values (Table 5.7) give an indication of the duration and intensity of freeze at various depths on the different aspects. Clearly, the duration and intensity of freeze is most pronounced at shallower depths on the southern and western aspects and decrease towards the

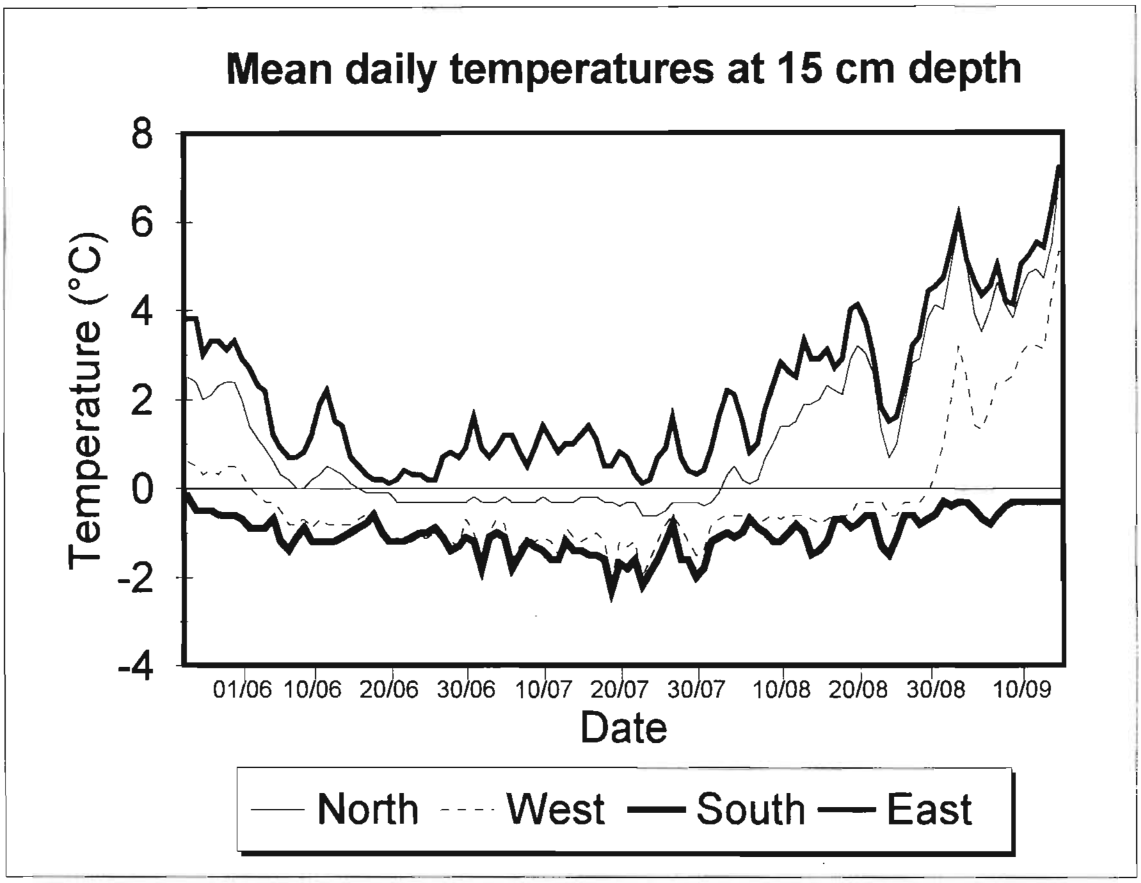


Figure 5.30 Mean daily temperatures at 15 cm depth for a thufa north, west, south and east facing aspects (May - September 1995).

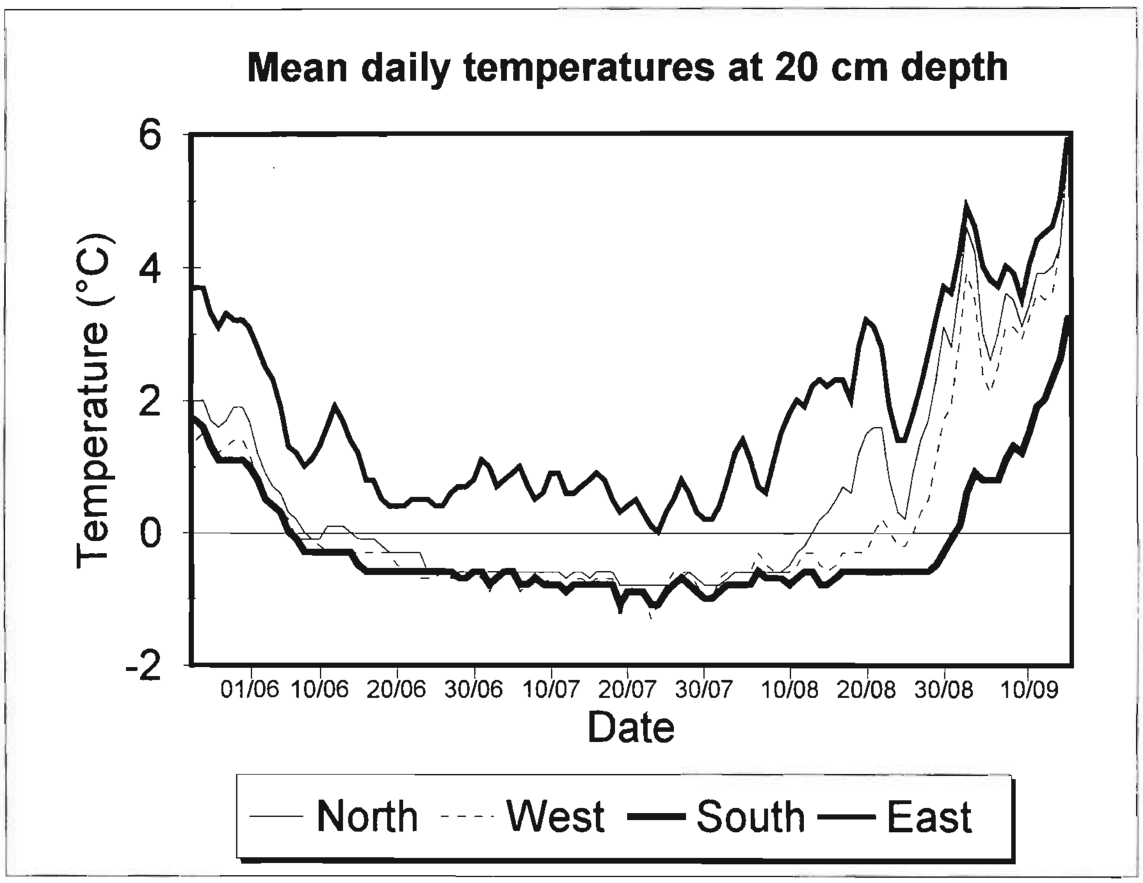


Figure 5.31 Mean daily temperatures at 20 cm depth for a thufa north, west, south and east facing aspects (May - September 1995).

	Temperature (°C)								Freeze days		Freeze index	
	Mean		Maximum		Minimum		Mean daily range				(°C days/month)	
JUNE	15 cm	20 cm	15 cm	20 cm	15 cm	20 cm	15 cm	20 cm	15 cm	20 cm	15 cm	20 cm
north	0.2	0.0	2.0	1.7	-0.3	-0.6	0.16	0.16	21	22	-3.8	-6.5
west	-0.8	-0.2	0.3	1.2	-1.3	-0.7	0.54	0.14	28	24	-23.2	-10.4
south	-1	-0.3	-0.6	1.0	-1.4	-0.7	0.56	0.07	30	24	-31	-12
east	1.0	1.2	2.9	3.1	-0.7	0.4	1.44	0.57	0	0	-	-
JULY												
north	-0.3	-0.7	-0.2	-0.6	-0.6	-0.8	0.12	0.05	27	31	-10	-21
west	-1.2	-0.8	-0.6	-0.6	-2.1	-1.3	1.06	0.38	31	31	-36.9	-25
south	-1.5	-0.8	-0.8	-0.6	-2.3	-1.1	0.91	0.18	31	31	-46.8	-26.2
east	0.8	0.6	1.6	1.1	0.1	0.0	1.5	0.66	0	0	-	-
AUGUST												
north	1.7	0.5	4.1	3.1	-0.3	-0.8	2.57	0.94	0	12	-	-6
west	-0.5	-0.1	1.0	1.9	-0.8	-0.7	0.25	0.39	29	20	-15.7	-9.9
south	-0.9	-0.6	-0.3	-0.1	-1.5	-0.9	0.59	0.09	31	31	-28.6	-20
east	2.6	2.0	4.7	3.7	0.8	0.4	3.35	1.4	0	0	-	-
SEPT. (1-15)												
north	4.7	3.7	6.9	5.6	3.5	2.6	4.25	2.55	0	0	-	-
west	2.8	3.3	5.3	5.6	1.3	2.1	3.77	2.62	0	0	-	-
south	-0.4	1.4	-0.3	3.2	-0.8	0.1	0.14	0.65	15	0	-	-
east	5.2	4.3	7.2	5.9	4.1	3.5	4.28	1.69	0	0	-	-
							TOTAL					
							North		48	65	13.8	33.5
							West		88	75	75.8	45.3
							South		107	86	106.4	58.2

Duration of freeze penetration (days) (from 15 cm to 20 cm depth)		**Duration of thaw (days) (from 20 cm to 15 cm depth)**	
north	-1	north	-16
west	4	west	9
south	13	south	16+
east	-	east	-

Table 5.7 Temperature and freeze characteristics for various thufa aspects (1995).

northern and eastern aspects. Of interest is the considerably longer duration of freeze at 20 cm depth (65 days) than at 15 cm depth (48 days) on the northern aspect (Figures 5.30 and 5.31 and Table 5.7). This difference is due to an early thaw that commenced from the thufa northern aspect. From the available data (see Figures 5.30 and 5.31 and Table 5.7) it is apparent that thawing occurred from the northern and eastern aspects and gradually progressed towards the western and southern aspects. The thaw process was considerably delayed on the colder southern aspect where it took over 16 days (logging period terminated during this time) to thaw from 20 cm depth to 15 cm depth, whereas on the western aspect it took 9 days (Figures 5.30 and 5.31 and Table 5.7). The data also show that thawing on the southern and western aspects took place from within the thufa, rather than from the surface inwards.

Mean daily temperature range

During winter, the shallower depths of the thufa eastern aspect experience the largest temperature range as temperatures did not go sub-zero. The lowest temperature recorded at 15 cm depth on the eastern aspect was 0.1°C (Table 5.7). The temperature range at 20 cm depth on the eastern aspect is considerably smaller than at 15 cm depth because of the cooling effect from within the thufa (Figures 5.32 and 5.33 and Table 5.7). The smallest temperature ranges occurred during mid-winter on the thufa northern, southern and eastern aspects at 20 cm depth. This is possibly due to two reasons; firstly, the "zero curtain" effect and, secondly, the effect of depth which reduces the temperature range. The "zero curtain" effect is evident, especially when comparing the daily temperature range with the temperature trend displayed in Figures 5.30 and 5.31. For instance, at 20 cm depth on the northern aspect, temperatures drop to below 0°C on the 14 June but remain between 0°C and -0.5°C until the 11 July (Figure 5.31). During this zero curtain time, there was no daily temperature variability (Figure 5.33). However, when temperatures dropped to below -0.5°C during the following two weeks, there were small recorded daily temperature fluctuations (Figures 5.31 and 5.33). This trend is most evident when viewing the daily temperature range for the thufa southern aspect at 20 cm depth (Figure 5.33), when two distinct periods of "zero curtain" occur. The temperature range at the shallower 15 cm depth shows distinct peaks during mid-winter on the southern and western aspects (Figure 5.32). These higher temperature ranges are associated

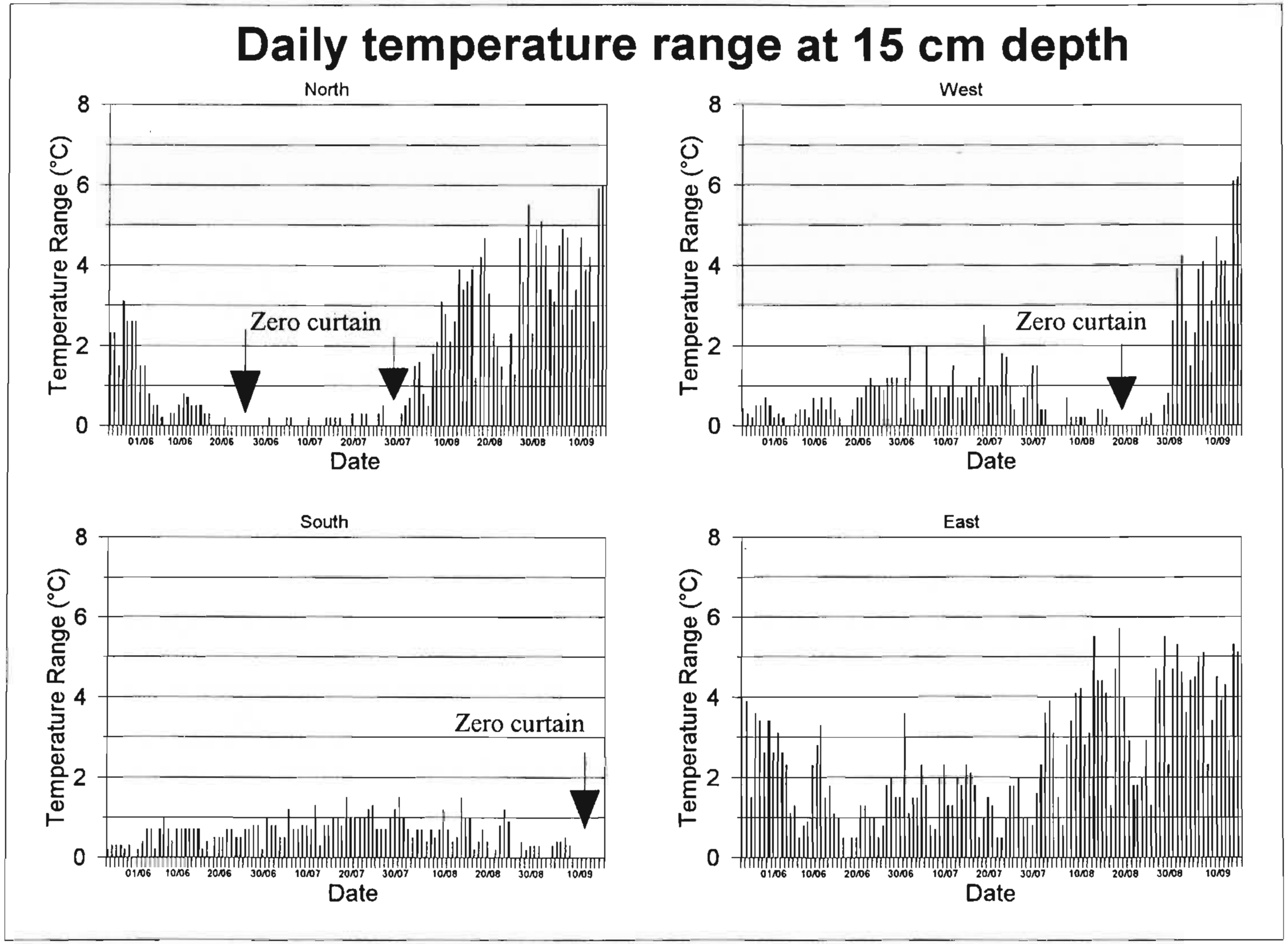


Figure 5.32 Daily temperature range for various thufa aspects at 15 cm depth (May - September 1995).

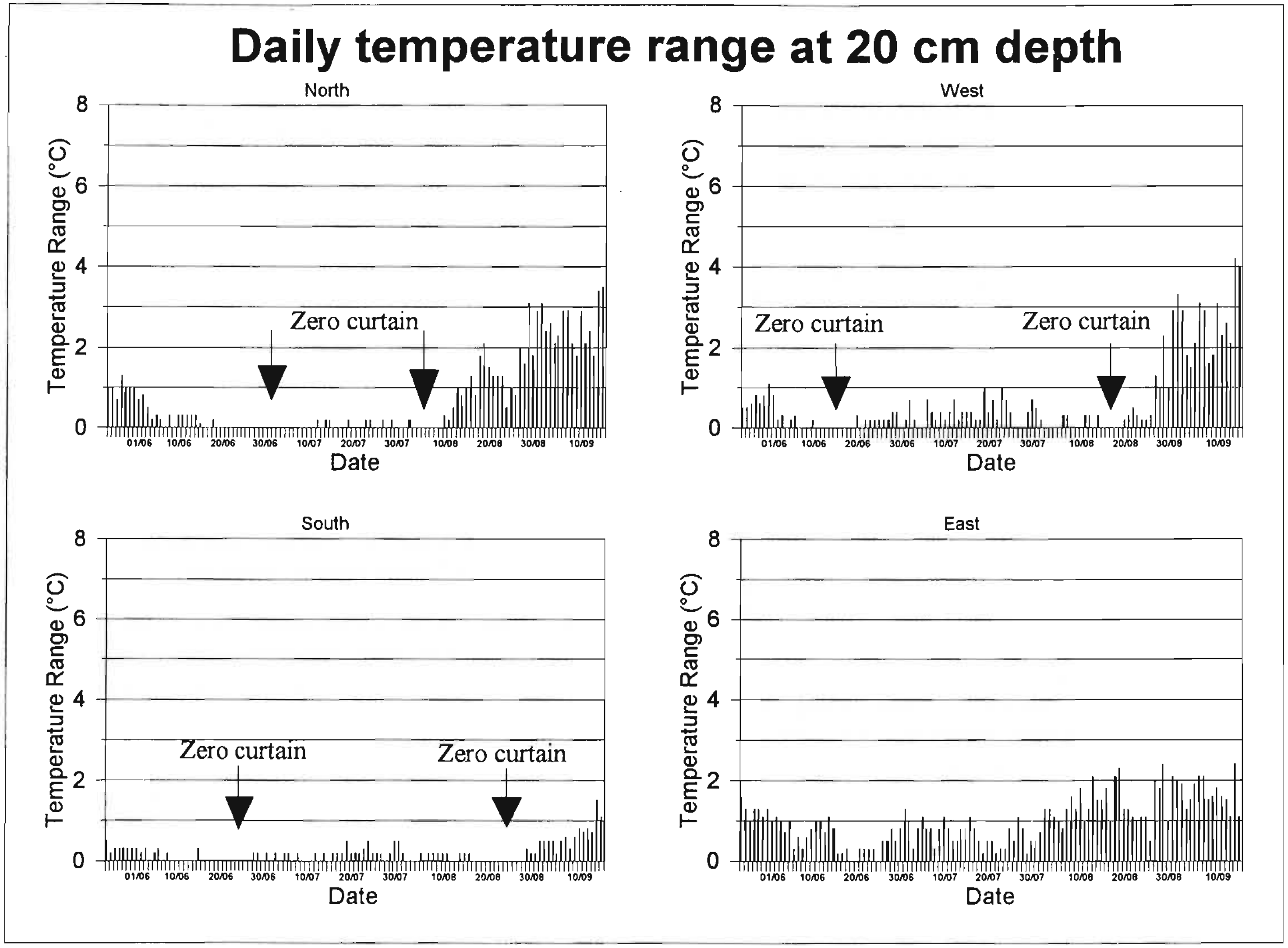


Figure 5.33 Daily temperature range for various thufa aspects at 20 cm depth (May - September 1995).

with somewhat lower temperatures (-2.3°C) and consequent increased conductivity when the "zero curtain" effect is overcome. In contrast, the temperature at 15 cm depth on the northern aspect remains between -0.6°C and 0°C and consequently the "zero curtain" effect is not overcome, thereby reducing daily temperature fluctuations. It is also evident from the graphs that once the "zero curtain" effect is overcome during the thawing process at the end of winter, the daily temperature range increases rapidly (Figures 5.32 and 5.33), as do mean daily temperatures (Figures 5.30 and 5.31).

Daily temperature fluctuations

Between the 5th and 8th of June the freeze fringe had penetrated to 15 cm depth on the western aspect and was slowly approaching the northern aspect of the thufa (Figure 5.34). Of interest is the almost "mirror image" temperature curves between the eastern and southern/western thufa aspects. This is also visible between the 5th and 8th of July when the freeze fringe was situated at 15 cm depth on the thufa northern aspect (Figure 5.35). Ground temperatures at 15 cm depth on the thufa southern and western aspects are lowered to such an extent that maximum daily air temperatures are unable to supply enough heat energy to cause pronounced daily temperature peaks.

Towards the end of winter (25 August - 1 September), thufa temperatures at 15 cm depth on the eastern and northern aspects show a continuous gradual rise which is almost parallel on the two aspects (Figure 5.36). Temperatures on the southern and western aspects remain within the "zero curtain" for much of this time and therefore do not show the same daily amplitude in range as do those on the northern and eastern aspects. However, once the "zero curtain" effect is overcome on the western aspect, the ground temperature rapidly rises and permits a greater daily temperature range (Figure 5.36). Within a few days (5 - 8 September) the daily ground temperature fluctuations on the western aspect parallel those on the northern and eastern aspects, albeit somewhat depressed (Figure 5.37). During the same time period the ground temperature on the southern aspect did not undergo a similar daily amplitude temperature range.

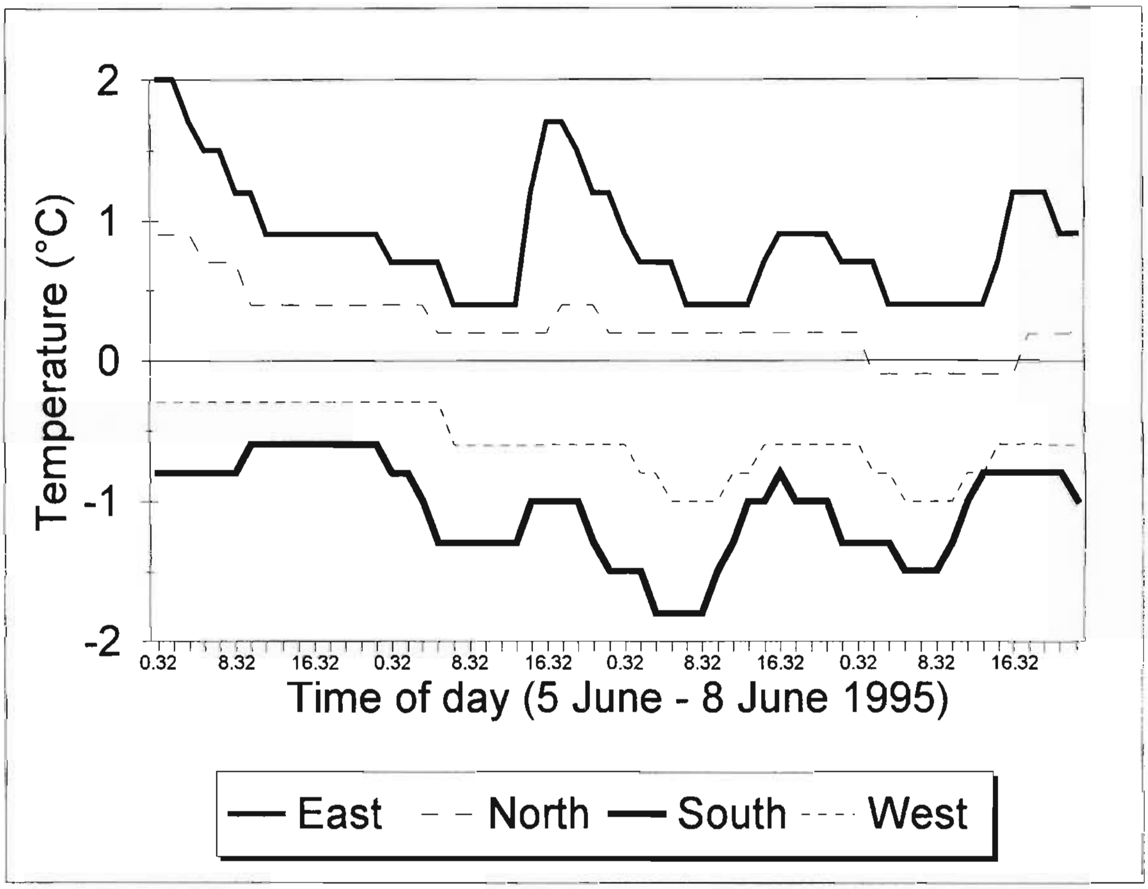


Figure 5.34 Temperature trends for various thufa aspects at 15 cm depth from the 5th June to 8th June 1995.

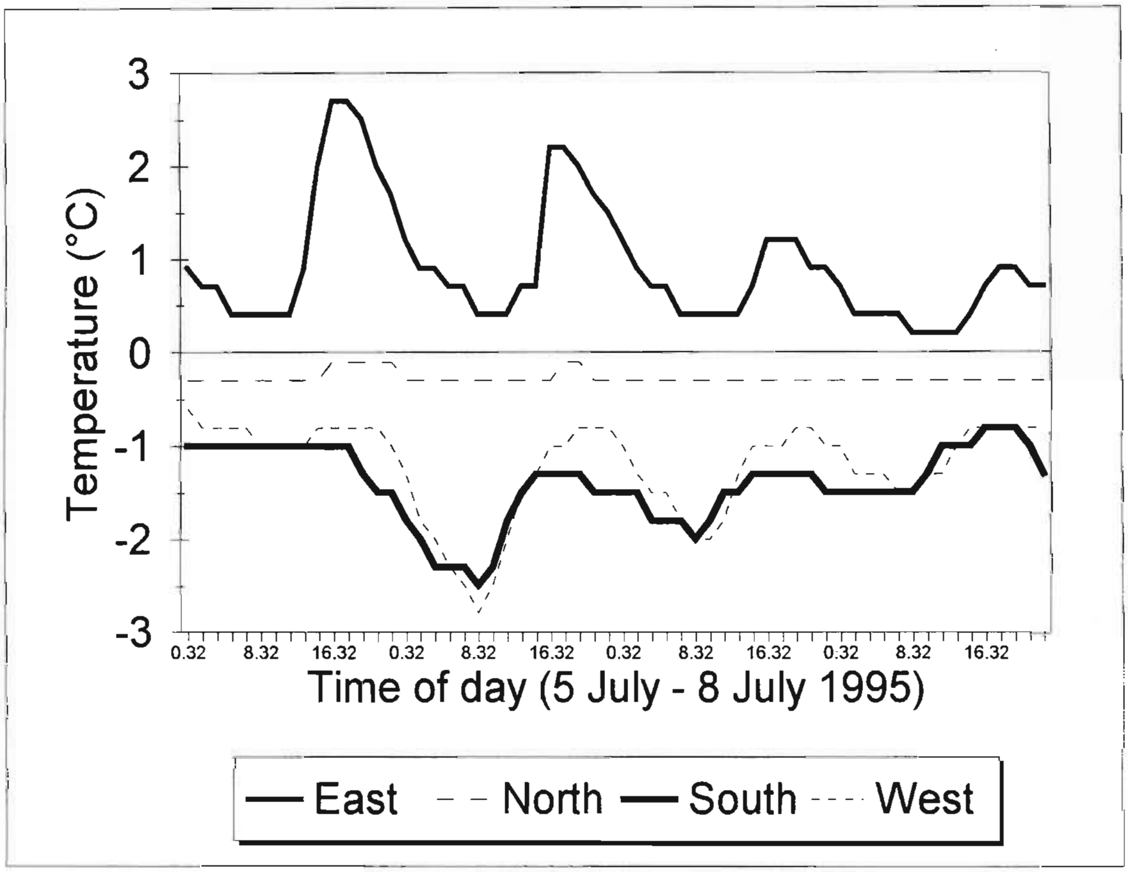


Figure 5.35 Temperature trends for various thufa aspects at 15 cm depth from the 5th July to 8th July 1995.

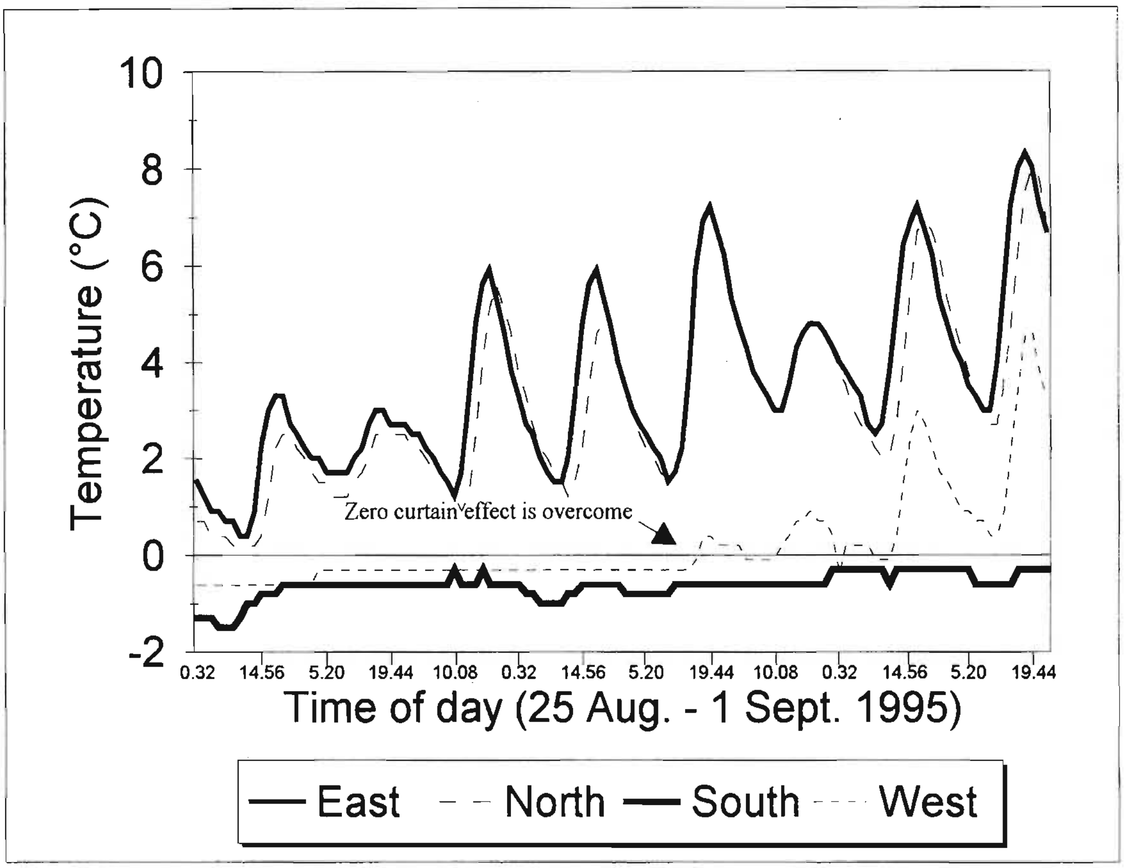


Figure 5.36 Temperature trends for various thufa aspects at 15 cm depth from the 25th August to 1st September 1995.

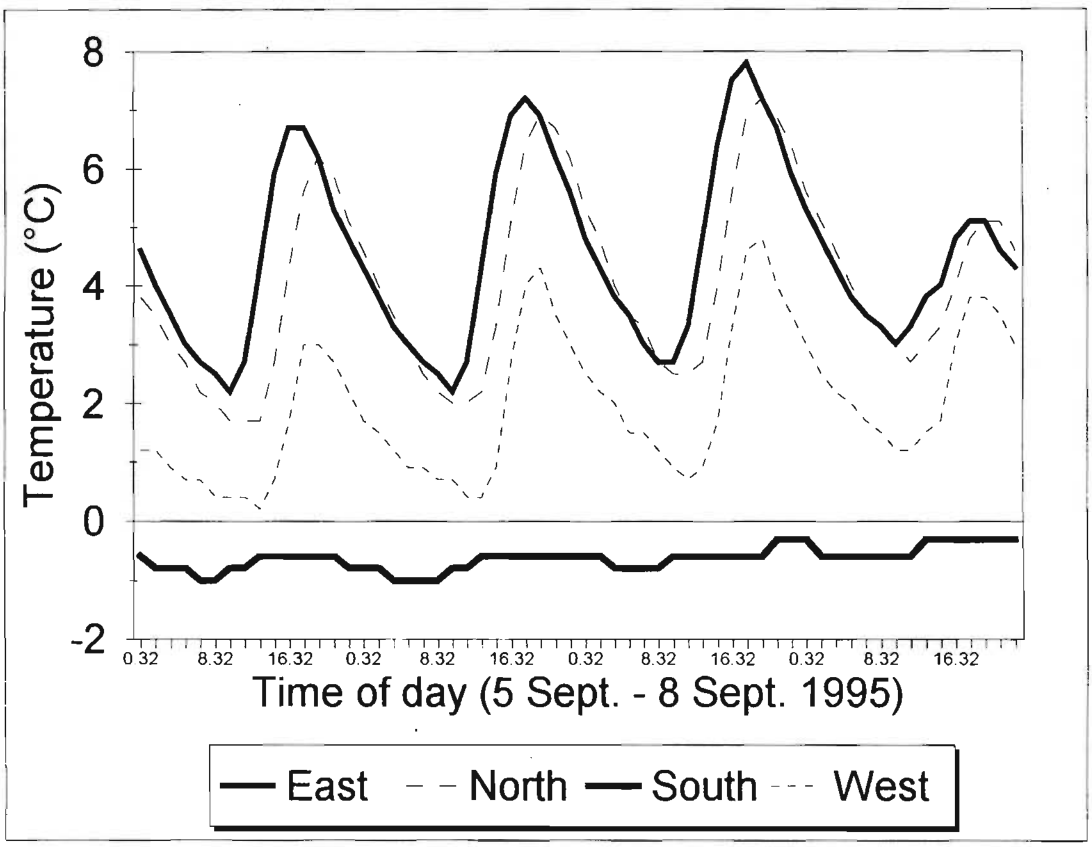


Figure 5.37 Temperature trends for various thufa aspects at 15 cm depth from the 5th September to 8th September 1995.

Data from 1996

During the 1996 winter, temperatures were recorded at 1 cm, 5 cm and 10 cm depth on the northern and southern aspects of a thufa. By the 1st of June, the ground had frozen to at least 10 cm depth on the thufa southern aspect (Figures 5.38, 5.39 and 5.40). In contrast, the north-facing aspect remained unfrozen until the 19th of June. Owing to the high insolation receipt, mean ground surface temperatures (@ 1 cm depth) continued to fluctuate between 0°C and 2.7°C throughout June on the northern aspect (Figure 5.38 and Table 5.8).

A heavy snowfall was recorded between the 6th and 8th of July when snow accumulated to over 1 metre on the plateau (Rescue Services radio report). The effect of the snowfall on thufur thermal properties is clearly evident from the graphs (Figures 5.38, 5.39 and 5.40). Mean temperatures differed by 2 to 5°C between the thufa northern and southern aspects at 1 cm depth during the snow-free month of June. However, once snow had accumulated at the site and remained for several weeks, these aspect controlled temperature differentials were reduced to about 1°C. Once the snow had thawed on the northern aspect, there was again a marked difference in mean ground temperatures between northern and southern aspects. The insulating effect of over 1 metre of snow almost nullifies the temperature differentials between the thufa northern and southern aspects at 5 and 10 cm depth during July and August (Figures 5.39 and 5.40).

Mean temperatures remained around the "zero curtain" level until the 30th of August on the thufa northern aspect, thereafter mean temperatures rose rapidly for several days (Figures 5.38, 5.39 and 5.40). The rapid temperature rise may indicate that the thinning snowcover disappeared, and hence the insulating capacity to protect ground temperatures from insolation was lost. Snow is protected by shading on the thufa southern aspect, consequently surviving longer here. It is also evident that the snow-pack thickness is more gradually reduced on the southern aspect than on the northern aspect. This results in a more progressive and gradual rise in ground temperatures on the thufur southern aspects than on the northern aspects (Figures 5.38, 5.39 and 5.40).

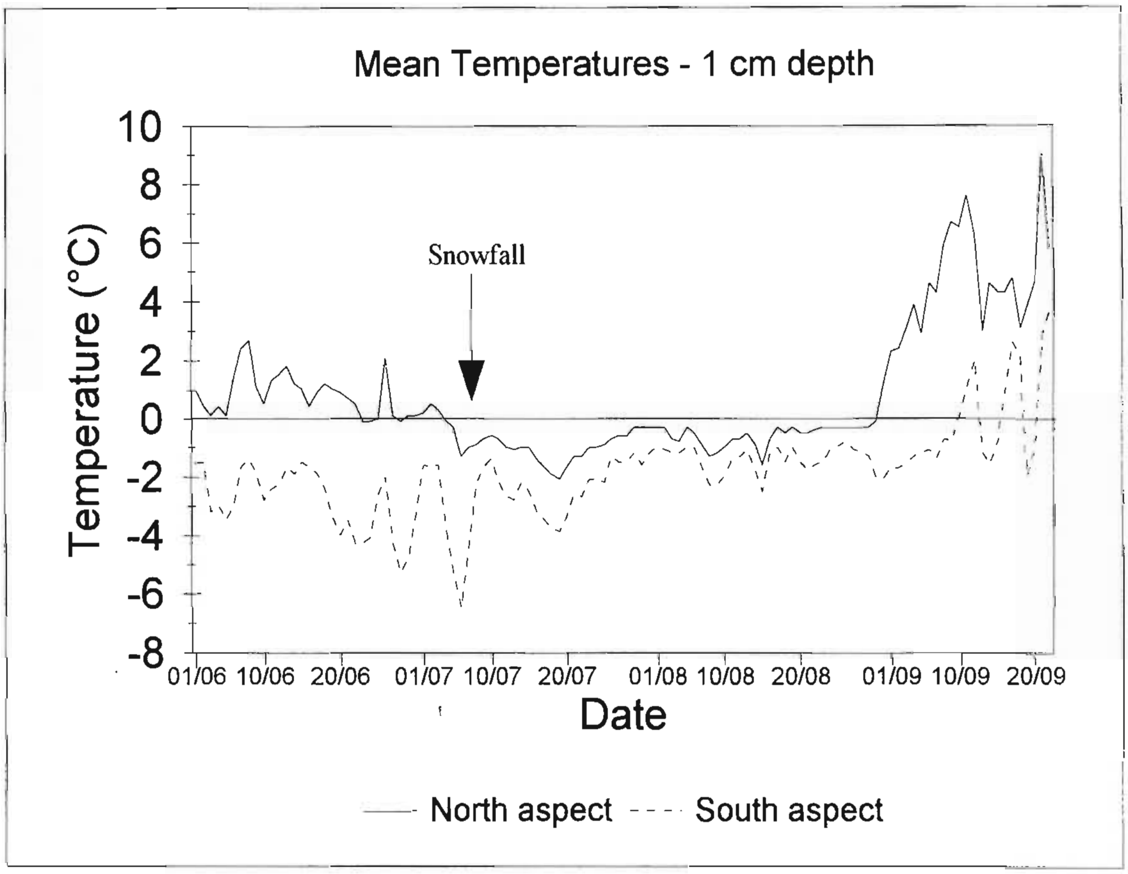


Figure 5.38 Mean daily temperatures at 1 cm depth for a thufa north and south aspect (June - September 1996).

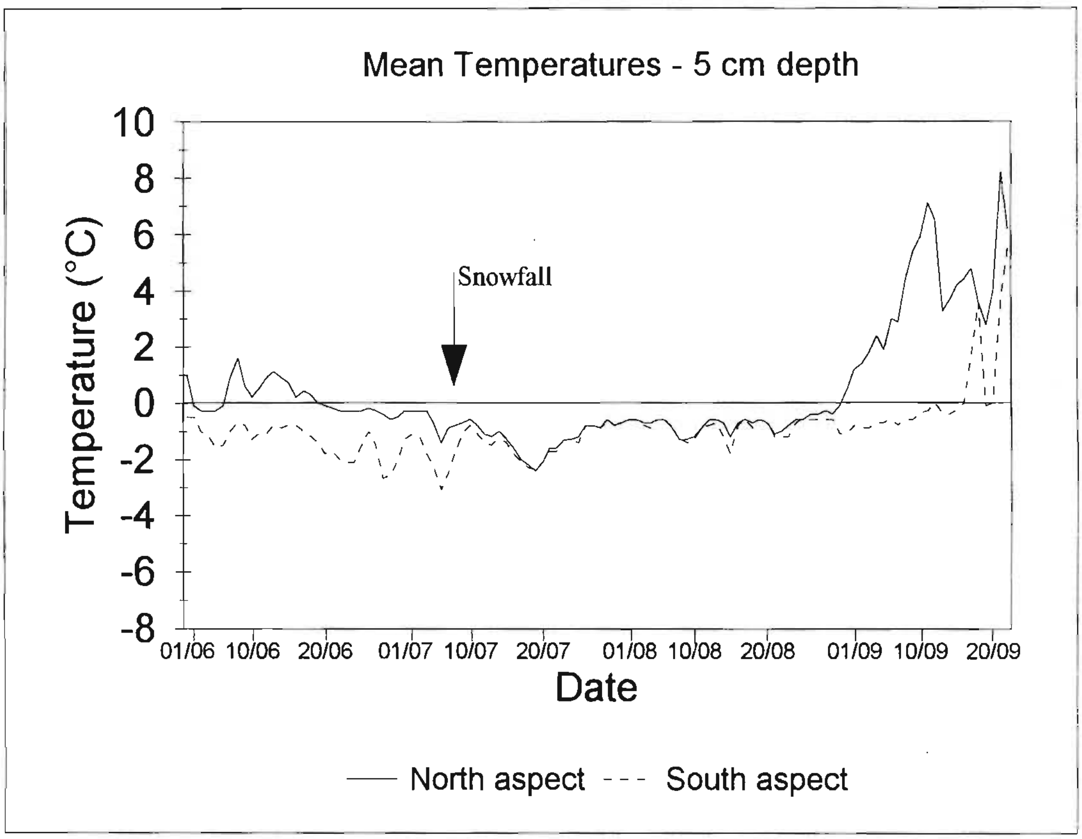


Figure 5.39 Mean daily temperatures at 5 cm depth for a thufa north and south aspect (June - September 1996).

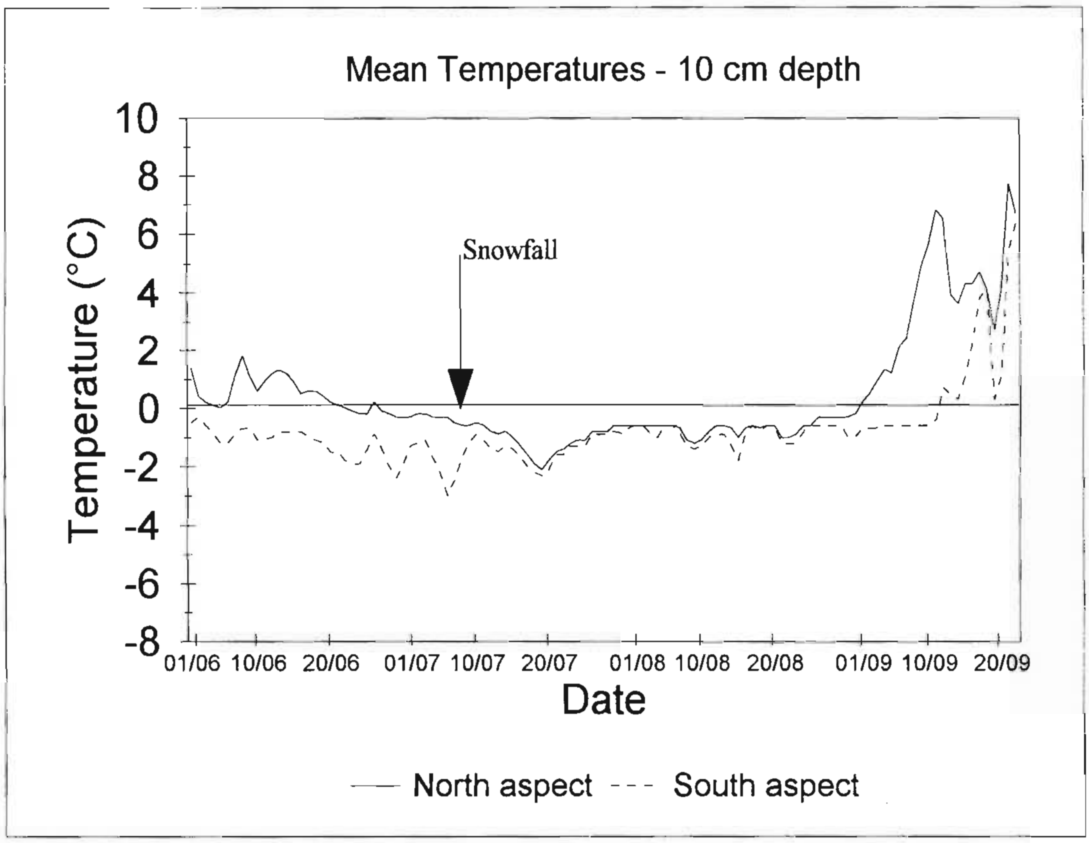


Figure 5.40 Mean daily temperatures at 10 cm depth for a thufa north and south aspect (June to September 1996).

	Depth	Mean (°C)	Max. (°C)	Min. (°C)	Std. Dev.	Mean Daily Range (°C)	Freeze days	Freeze Index °C days/month	Thaw Index °C days/month
JUNE									
North aspect	1 cm	0.8	5.8	-0.6	3.15	6.3	0	-17.1	172.9
	5 cm	0.2	1.4	-0.3	1.27	1.6	9	-8.2	43.4
	10 cm	0.4	1.2	0	0.94	1.2	6	-2.8	38.3
South aspect	1 cm	-2.8	-0.6	-4.5	2.00	3.9	23	-136.6	5.3
	5 cm	-1.3	-0.8	-1.9	0.75	1.2	30	-58.9	0
	10 cm	-1.2	-0.7	-1.7	0.66	0.9	30	-50	0
JULY									
North aspect	1 cm	-0.8	-0.3	-1.1	1.02	0.8	26	-34.6	12.7
	5 cm	-1.1	-0.9	-1.3	0.60	0.4	31	-40.2	0
	10 cm	-0.9	-0.7	-1	0.53	0.2	31	-30	0
South aspect	1 cm	-2.6	-1.6	-3.5	1.75	1.9	28	-108.9	0
	5 cm	-1.5	-1.2	-1.8	0.67	0.6	31	-55.8	0
	10 cm	-1.5	-1.2	-1.8	0.62	0.5	31	-54.8	0
AUGUST									
North aspect	1 cm	-0.5	0	-0.8	1.11	0.7	29	-23.9	12.5
	5 cm	-0.7	-0.4	-0.9	0.65	0.5	29	-27.9	5.5
	10 cm	-0.6	-0.6	-0.8	0.29	0.2	31	-24.4	0
South aspect	1 cm	-1.4	-1.1	-2	0.69	0.9	31	-62	0
	5 cm	-0.9	-0.7	-1.2	0.47	0.5	31	-37.2	0
	10 cm	-0.9	-0.7	-1.1	0.43	0.2	31	-35.8	0
SEPT. (1 - 20)									
North aspect	1 cm	4.7	15.0	0	6.72	14.9	0	-5.5	330.4
	5 cm	4.0	9.9	0.7	4.27	9.2	0	-2.4	218.1
	10 cm	3.7	7.5	1.1	3.45	6.4	0	-1.9	164.3
South aspect	1 cm	-0.1	2.6	-2.4	3.11	5.0	6	-52.3	60.7
	5 cm	0.3	1.1	-0.5	2.10	0.7	14	-14.8	29.3
	10 cm	0.9	1.7	0	2.36	1.7	11	-9	44.5

Table 5.8 Temperature characteristics and freeze / thaw indices for various thufa aspects (1996)..

Mean daily temperature range

It is clearly evident that the mean daily temperature range during June is greater on the northern aspect than on the southern aspect at all depths of measurement (Figures 5.41 and 5.42). As freezing intensified on the southern aspect during June, the temperature range progressively increased at all depths (Figure 5.42). In contrast, the daily temperature range decreased during this time on the thufa north-facing aspect, possibly owing to the "zero curtain" effect (Figure 5.41). The pronounced effect that snow cover has on reducing daily temperature fluctuations is clearly seen in Figures 5.41 and 5.42; it being most evident on the thufa north-facing aspect where the mean daily temperature ranges recorded at 1 cm and 10 cm depth were 0.8°C and 0.2°C respectively (Table 5.8). The insulating effect of the snow maintains temperatures within the "zero curtain" level on the thufa northern aspect for several weeks, hence the low daily temperature range. On the thufa southern aspect, mean daily temperatures at 1 cm depth are depressed to well beyond the "zero curtain" effect, therefore permitting a greater daily amplitude in temperature variations than on the northern aspect. Further support for the above observation is obtained when examining the daily temperature ranges for the thufa southern aspect at 5 and 10 cm depths. At these depths, temperatures remain closer to the "zero curtain" threshold than at 1 cm depth and consequently mean daily temperature ranges are lower than those at 1 cm depth. On the northern aspect there is very little or no variation in daily temperature ranges between 5 and 10 cm depth as the insulating effect of snow maintains temperatures within the "zero curtain". The pronounced increase in the mean daily temperature range on the 30th August is most likely associated with the rapid disappearance of an insulating snowcover on the thufa northern aspect (Figure 5.41).

Freeze and thaw indices

Weekly and monthly freeze (Figure 5.43) and thaw (Figure 5.44) indices were calculated for the northern and southern aspects for the various thufa depths (see also Table 5.8). On the southern aspect the weekly freeze index reaches its highest level during the last week in June owing to continued mid-winter cooling (Figure 5.43). The freeze indices on the southern aspect are substantially higher than on the northern aspect. For instance, during the fourth

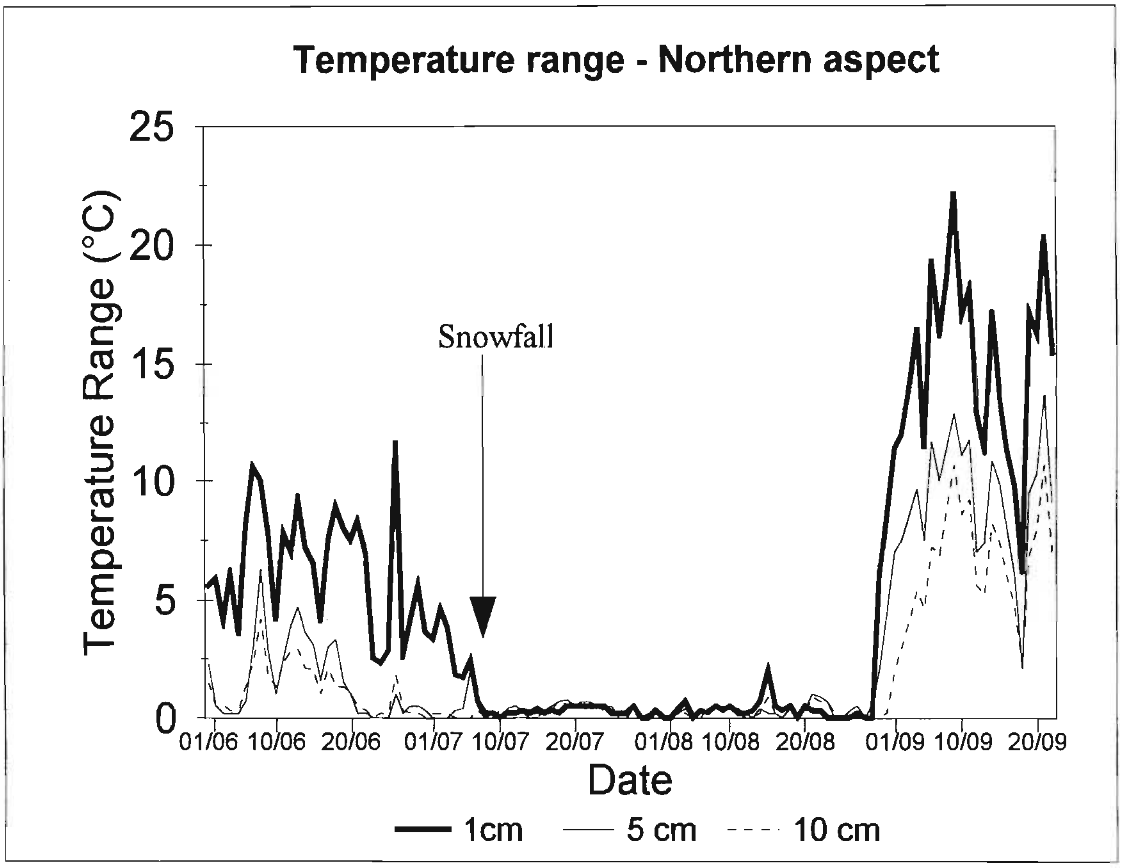


Figure 5.41 The daily temperature range for a thufa northern aspect at 1 cm, 5 cm and 10 cm depth (June - September 1996).

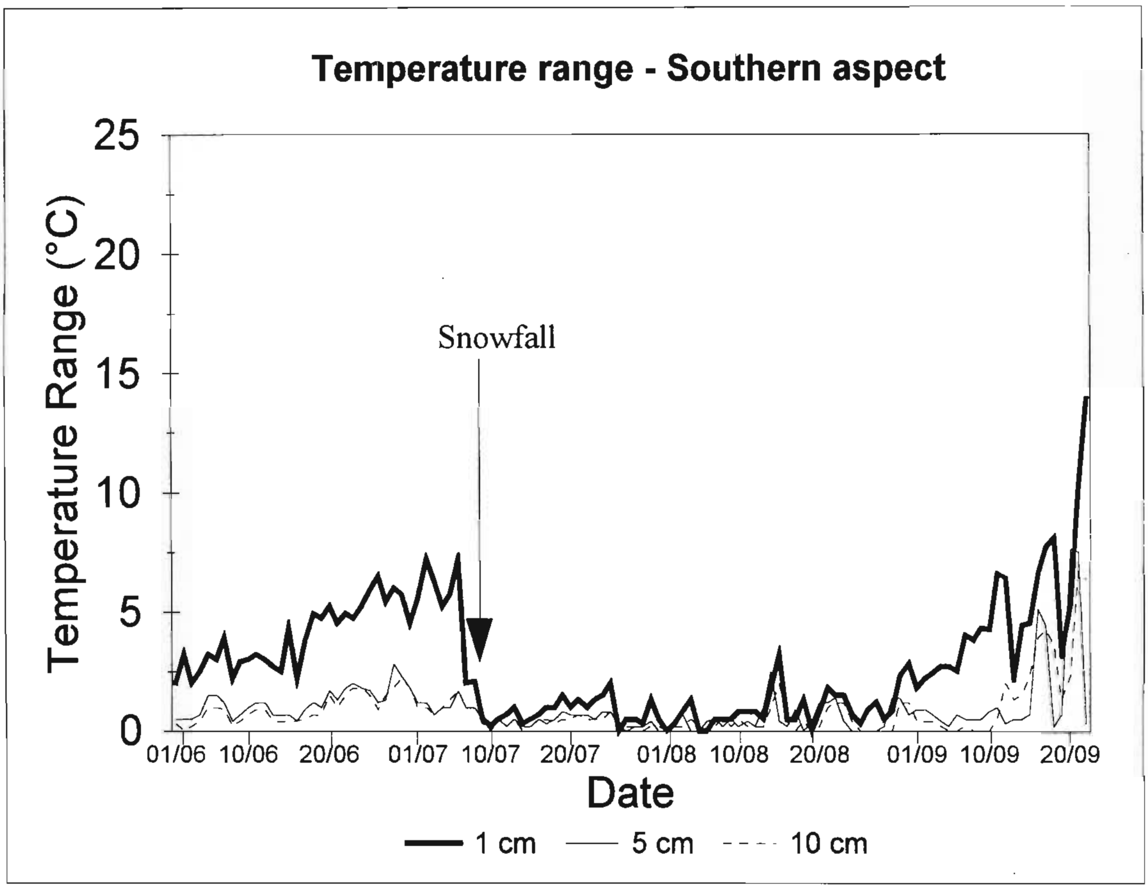


Figure 5.42 The daily temperature range for a thufa southern aspect at 1 cm, 5 cm and 10 cm depth (June - September 1996).

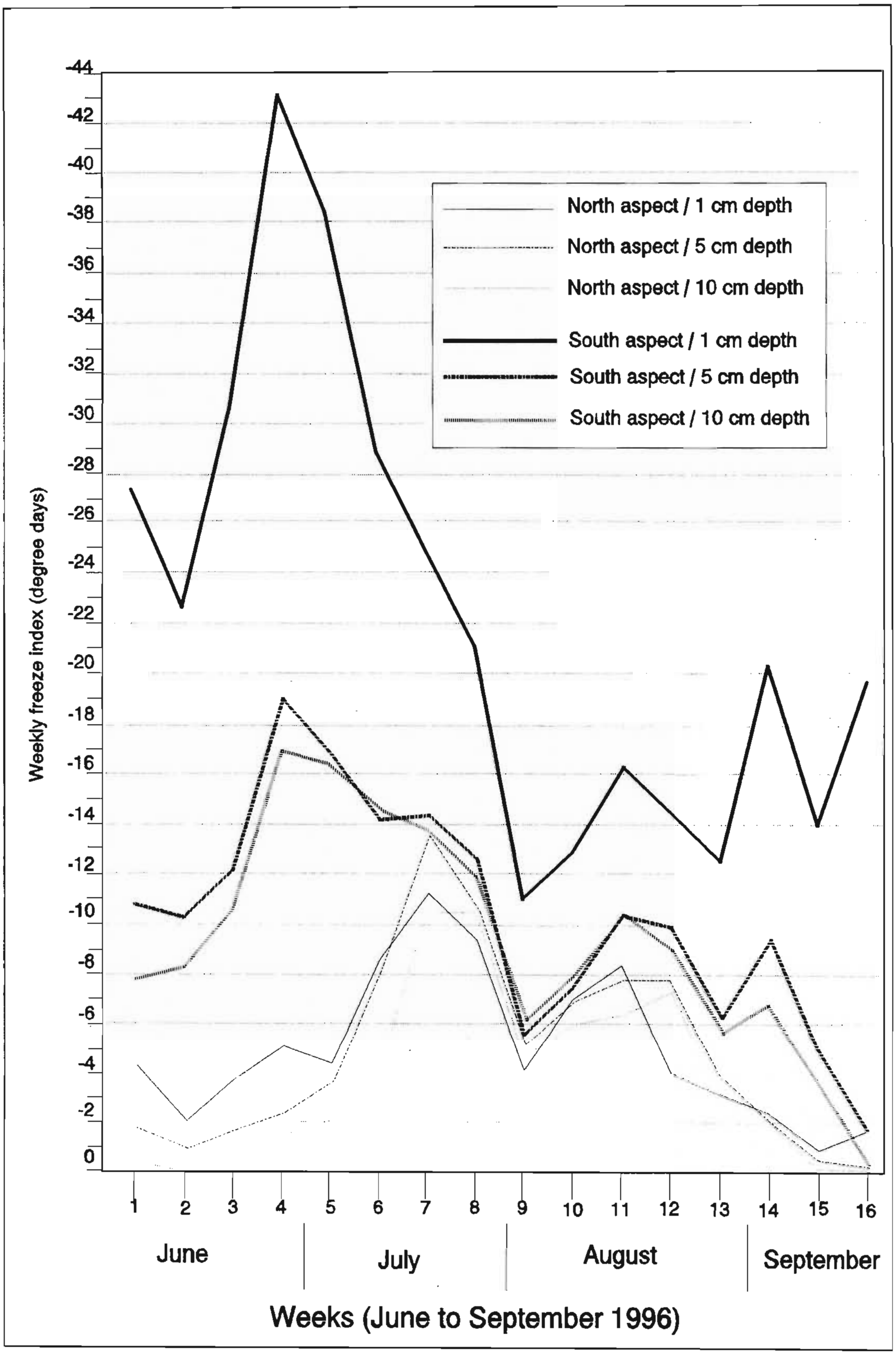


Figure 5.43 Weekly freeze indices for various depths on a thufa northern and southern aspect (June - September 1996).

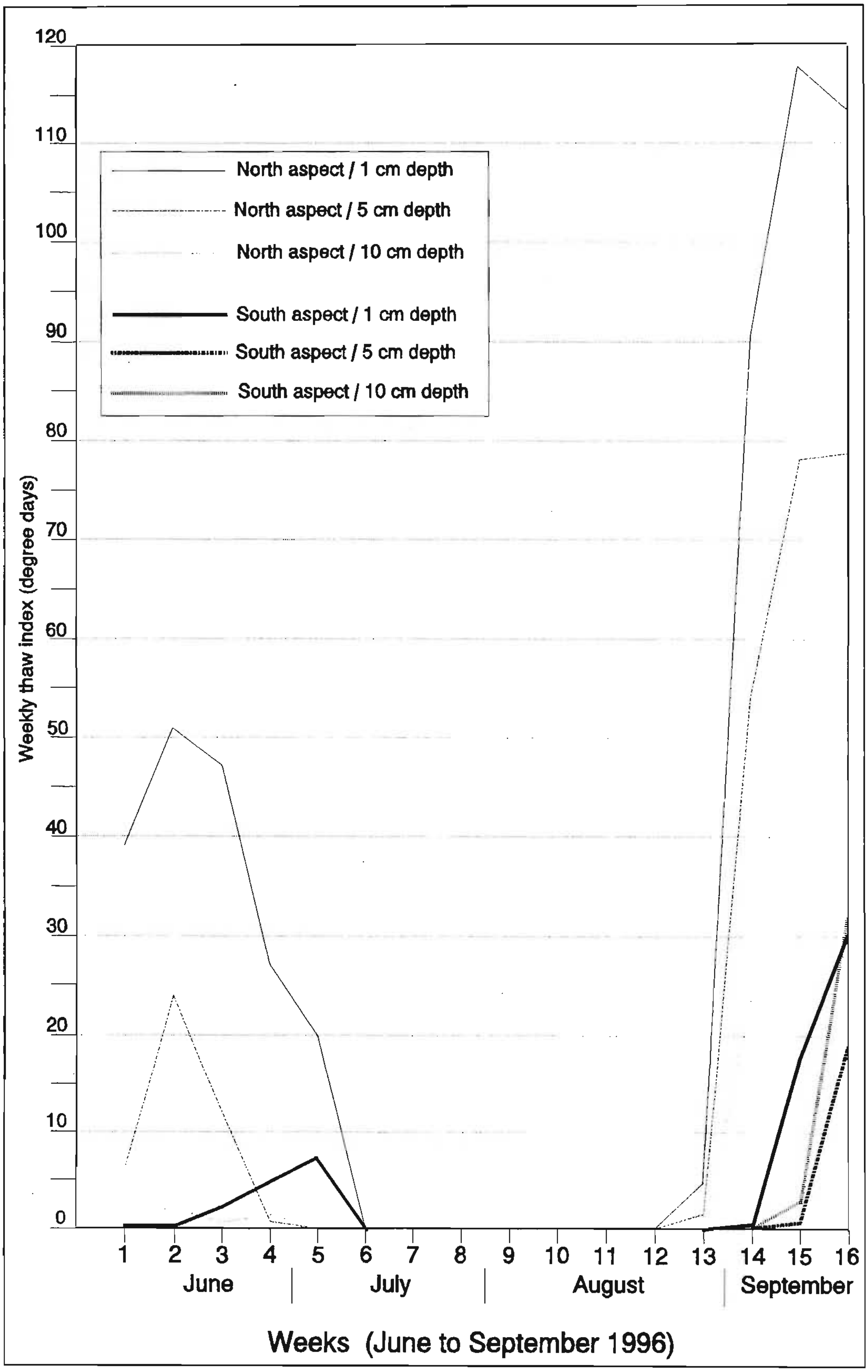


Figure 5.44 Weekly thaw indices for various depths on a thufa northern and southern aspect (June - September 1996).

week in June, a freeze index of -43 was recorded at 1 cm depth on the southern aspect while at the same depth on the northern aspect the index value was only -5.1. No thaw index was recorded at 5 and 10 cm depth on the southern aspect from June to August, indicating continuously frozen conditions. During the period of snowcover in July, thaw indices reached zero at all recording positions within the thufa, indicating that insolation receipt at the thufa surface was insufficient to induce positive daily ground temperatures. The freeze indices on the northern aspect continued to rise until mid-winter indicating continued cooling owing to the protective snow cover and very low daily air temperatures. As the snowcover diminished and daily air temperatures recovered somewhat towards the end of July, freeze indices stabilized and later dropped (Figure 5.43). The higher freeze indices towards mid-August may be attributed to a reduced snowcover offering less insulation against nocturnal radiative cooling. The drop in the freeze index value on the southern aspect at 1 cm depth is substantial; dropping from -43 (end of June) to -11 (early August) (Figure 5.43). This drop is clearly the result of a deep snowcover insulating the ground.

Freeze days

The number of freeze days gives an indication of the duration of continuous freeze at a position within the thufur. The southern aspect at 5 cm depth was continuously frozen for the longest period, at 106 days (Table 5.8). In comparison, the northern aspect at 5 cm depth was frozen for 69 days while at 10 cm depth it was frozen for 68 days. From the data (Table 5.8), it is evident that the freeze fringe penetrates on the southern aspect and progresses towards the northern aspect. The thaw front during late August/early September moves through the thufa towards the southern aspect. These findings from 1996 thus support those made during 1995.

5.2.6 Heave characteristics from the Mashai Valley

5.2.6.1 Peg Heave

Of the 120 wooden pegs inserted at a thufur site to determine possible heave, 104 were retrieved for measurement after a 3 month period (June to August). The recovery success for pegs inserted into the thufur apexes was 95%, while that for the depressions was 78%. Results displayed in Figure 5.45 show that the greatest peg heave occurred where pegs were inserted 20 cm deep into the depression (mean heave = 21.7 mm). Pegs inserted to 5, 10 and 20 cm depth all show greater average heave in the depressions than on the apexes (Figure 5.45). A general increase in mean peg heave is recorded with an increased depth of peg insertion (Figure 5.45). For instance, the mean apex peg heave at 5 cm depth of insertion is 6.4 mm, while at 20 cm depth of insertion it is 10.8 mm (Figure 5.45). Similarly, the mean depression peg heave at 5 cm depth of insertion is 6.5 mm, while at 20 cm depth of insertion it is 21.7 mm (Figure 5.45).

Williams and Smith (1989) have described two types of heaving, namely "primary" and "secondary" heaving. Primary heaving is associated with ice segregation developing mainly near the frost line, while secondary heaving occurs more slowly within frozen layers. Williams and Smith (1989) have indicated that secondary heave, although slower than primary heave, has the potential for larger heaving pressure than primary heave. It would appear that pegs inserted to 5 and 10 cm depth were within a frozen layer for much of the time, while those inserted to 20 cm depth, had their lower sections within an unfrozen soil. Heaving may occur when pressure differentials are created between different phases of soil moisture. Soil moisture will migrate from the unfrozen areas towards the freezing plane and even to some point within the frozen layer (Williams and Smith, 1989). As the pegs inserted to 5 and 10 cm depth were within a frozen layer for much of the time, secondary heave would have been more likely than primary heave. The pegs inserted to 20 cm depth would have experienced primary heave for a much longer period because of unfrozen soils at their lower sections. Because the temperature gradients within a frozen mass are expected to be less pronounced than those between a frozen and an unfrozen mass, it would be expected that those pegs inserted to a greater depth would heave more substantially than those inserted to a shallower depth. The pegs inserted to 20 cm depth in the depression are likely to have had their lower areas within an unfrozen soil for much of the period. The pegs inserted to 20 cm depth in the thufur apexes would, however, have had a greater extent of their lower areas within a frozen mass as the freezing plane progressed through the apexes. This may account for the pegs inserted to 20 cm depth in the apexes, having heaved less than those inserted to a

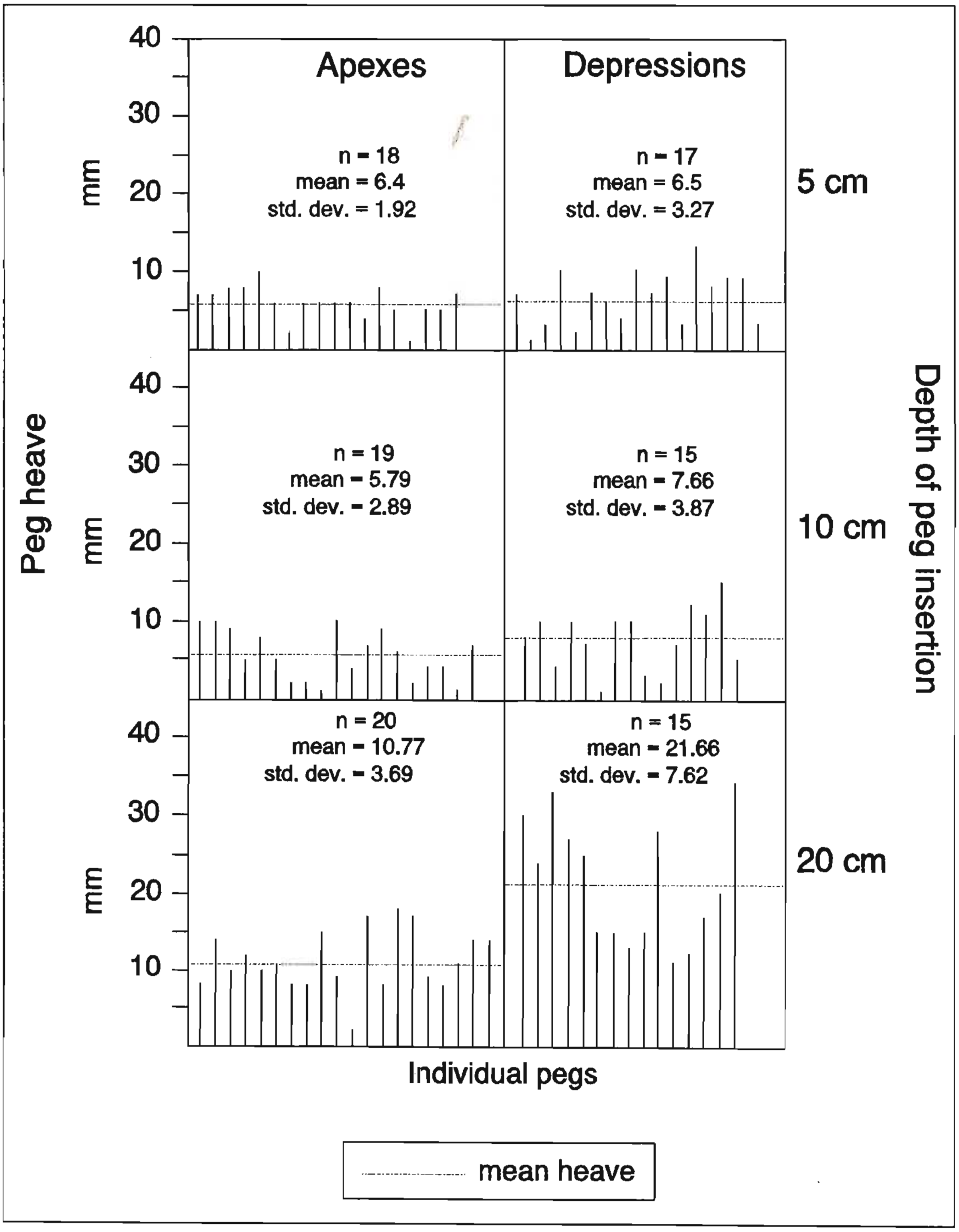


Figure 5.45 Comparison of peg heave at different depths within thufur apexes and depressions (June to August 1994).

corresponding depth in the depressions. This argument is supported by the 1993/94 ground temperature data at 12 cm depth within a thufa apex and its adjoining depression (See Section **5.2.4**). The peg heave results may not be conclusive enough to explain the heave differentials between thufur apexes and their adjoining depressions, however, they do indicate a soil heaving process at thufur sites.

5.2.6.2 Assessment of thufur apex heave, relative to the adjoining depressions

Results were obtained for four of the five thufur examined for potential heave during the winter of 1994 (Table 5.7). Results show that the surface distance between the thufur apexes and their adjoining depressions increased for all four thufur from March to July, and then decreased from July to September (Table 5.7). The maximum potential vertical apex heave, relative to the adjoining depressions, varied from 0.18 to 0.86 cm (Table 5.7). From this, it would appear that some thufur in the Mashai Valley annually expand or heave very slightly during the winter months and then slowly subside during the thaw period.

Thufur	Approximate slope of measured thufa surface (ϕ)	Surface distance :- apex to depression (cm) March	May	July	Sept.	Max. deviation of surface distance (cm) (d)	Max. potential vertical thufa heave, relative to adjoining depression (cm) ($d \sin \phi$)
1	27°	30.8	31.3	32.7	31.0	1.9	0.86
2	29°	30.8	31.1	31.6	30.8	0.8	0.39
3	25°	31.0	peg lost				
4	21°	32.7	32.8	33.2	32.8	0.5	0.18
5	18°	25.7	26.5	26.8	26.0	1.1	0.34

Table 5.9 Thufur heave characteristics during 1994, Mashai Valley.

5.2.7 Discussion

5.2.7.1 Thermal regime of thufur

The available thufa apex and depression soil temperature data from the Mashai Valley provide information which should help broaden the understanding of the thufur thermal regime. This discussion therefore attempts to explain the possible factors contributing towards the thermal characteristics outlined in Sections **5.2.4.** and **5.2.5.**

Thufur apex - depression temperature differentials

Tarnocai and Zoltai (1978) concluded that the inter-thufur depressions are the coldest parts of thufur, however, this is based upon a single recording time. The present study has now been able to show that the mean depression ground temperatures are considerably warmer than the mean apex ground temperatures at 12 cm depth. This study has also found that only during early spring months are mean depression temperatures lower than mean apex temperatures. In New Zealand, Mark (1994) found that depression temperatures were higher than apex temperatures in the summer months. However, both this and Mark's (1994) study have now indicated strong temperature differentials during the winter months when thufur apexes were frozen for several weeks and depressions remained predominantly unfrozen. It has been said that when such temperature differentials between frozen thufur apexes and generally unfrozen depressions are significant, this may provide a basis for maintenance of the existing microtopography (Mark, 1994). Further, it has been suggested that the amount of ground expansion would be a function of the duration and extent of such apex-depression temperature differentials, as well as the moisture conditions and rates of ground freezing (Rieger, 1983; Williams and Smith, 1989).

It has already been shown (Section **5.2.4.2**) that apex soil temperatures were rapidly depressed to below freezing over a few days in May and that the duration of the freeze was for 87 days. This contrasts significantly with the slow cooling and relatively unfrozen depression soils. A similar observation has been reported by Van Vliet-Lanoë (1991), who recorded a hummock temperature (upper 5 cm) of -0.1°C while the depression temperature was +1.2°C. It may therefore be argued that the temperature differentials between the thufur apexes and depressions were significant; possibly enough so to induce thermodynamic forces and ultimate ground heave. The soil moisture characteristics discussed in Section **5.2.3.4**, indicate that the rate of moisture lost to the depressions towards winter was considerably higher than that lost to the apexes; to the extent that depressions were left drier than apexes in mid-winter (Figure 5.7). From this, it would appear that soil moisture migrated from the depressions towards the freezing plane in the thufur apexes. Cryosuction is likely to have induced a volumetric expansion and heaving process here (c.f. Razbegin, 1988; Williams and Smith, 1989). The rise in depression soil moisture content during early spring is attributed to rapidly ablating snow or possible rainfall (Figure 5.7).

The thermal gradients within soils may be generated by several factors including such as air temperature, solar radiation, terrestrial radiation and evapotranspiration (C. Harris, 1981). The uneven microtopography of thufur apexes and depressions induces such thermal gradients and consequently helps maintain its morphology. However, this has not been examined in any detail for thufur, as long-term ground temperatures have, until recently, been unavailable. Here, an attempt is hence made to explain the possible contributing factors to the apex - depression temperature differentials that may ultimately induce volumetric expansion and heave.

The heat capacity of the apexes and depressions is likely to be different as the individual thufa apex represents a smaller volume than the depression areas. The smaller volume of the apex is, therefore, more likely to undergo rapid temperature changes and experience greater diurnal and seasonal temperature extremes. Further, the apexes usually have part of their surface continuously exposed to direct insolation during cloudless sunshine hours, while at the same time the depression surfaces may be shaded. This may also account for the higher mean daily range and mean daily maximum rate of change of temperature in the thufur apexes, than that encountered in the depressions. During spring warming, the smaller volume of the apexes absorbs heat more rapidly than the depressions, thereby accelerating the warming of apex soils. The greater heat capacity of depression soils results in a more gradual seasonal rise in ground temperatures there.

The soil moisture content of the depressions during summer is over 20% greater per unit weight than in the apexes (Figure 5.7). Such high moisture differentials may also influence a soils thermal regime. Although the thermal conductivity is likely to increase with an increase in moisture content (C. Harris, 1981), the thermal diffusivity is lowered as a result of greater heat capacity. Thus, the greater percentage moisture in the depressions during the summer months is likely to reduce the thermal diffusivity there, and consequently, contribute towards the lower mean daily range and mean daily maximum rates of change of temperature in the depressions. The lower thermal diffusivity and greater heat capacity in the depressions may together result in generally higher temperatures in the depressions than in the apexes during summer and autumn months. In addition, thufur apexes in the Mashai Valley were found to have on average over 13% more organics (dry weight) than the depressions (Table 5.3). The higher percentage of organics aids greater evaporative cooling during the summer months (Williams and Smith, 1989; Josefsson, 1990) and thereby contributes to keep apex temperatures somewhat cooler than the depression temperatures. These factors also help to explain the lag time of recorded maximum and minimum temperatures in depressions for much of the year.

Because of their relatively low heat capacity, apex ground temperatures are lowered relatively quickly during autumn months. The depressions, however, are able to store the heat absorbed during summer months for much longer periods and therefore drop less rapidly than apex temperatures. Several snowfalls during the winter recording period (June to August) left depression areas under an almost continuous blanket of snow while thufur apexes were left exposed (Figure 5.3). The irregular thufur topography permits snow depth to vary inversely with the topography (Mackay, 1993), resulting in the depression areas being occupied with the deepest snowcover. Because of the low thermal conductivity of snow, it has an insulating effect on the ground which prevents heat at the surface from escaping into the atmosphere (Smith, 1975; Williams and Smith, 1989). Consequently, during the winter months, the thufur apexes were subjected to intense nocturnal radiative cooling while the depressions were protected for extended periods owing to the insulating effect of snow. Such factors have almost certainly contributed towards the steep thermal gradients between thufur apexes and depressions during the winter months of 1994.

Surface cover may also insulate ground temperatures, but, as discussed in Section **5.2.3.2** , there are few vegetational differences between thufur apexes and depressions in the high Drakensberg. Further, the vegetation growing at most thufur sites is not sufficient to induce much of an insulating effect. Only at Site Two, near Mafadi Summit, is vegetation thought to influence the thermal regime of non-sorted raised vegetation rings (Figure 5.1). Van Zinderen Bakker and Werger (1974) have suggested that areas of sparse vegetation cover become frozen before areas of more prolific vegetation cover. Although this may be true in many instances, this is not always the case. The thicker vegetation cover on the raised vegetation rings (Mafadi Summit) (Figures 5.1 and 5.2) is such that it provides only marginal insulation during terrestrial radiation, but provides protective shading against solar radiation. This has resulted in a very thin (0.5 cm) "diurnal thawed layer" on the raised vegetation ring, while the surrounding areas have a considerably deeper "diurnal thawed layer" (4 to 5 cm) (Figure 5.1). Others (Lundqvist, 1969; Schunke, 1977) have suggested that the insulating effects of plants may cause depressions to become repeatedly frozen while the thufur apexes are left unfrozen. The present study has now shown that under most circumstances, this is not the case (see also: Van Vliet-Lanoë, 1991; Mark, 1994).

Freeze - thaw processes within thufur

It is well known that factors such as soil moisture, aspect, vegetation and snowcover will influence the aggradation or degradation of frozen ground (Åhman, 1977; Seppälä, 1986). However, the role of microtopography in influencing the aggradation and degradation of frozen ground has not been considered in detail. The present study on the thufur thermal regime offers some new information on micro-topographically controlled ground freezing processes.

The soil freezing process within thufur in Lesotho begins on the southern aspects and gradually progresses towards the western aspects owing to topographic shading and much lower levels of insolation than what would be expected on northern and eastern aspects (Figure 5.46). During snow-free conditions this may induce pronounced temperature differentials of between 2 and 5°C for different thufur aspects. Thufur size would undoubtedly have an affect on these differences as well as on the freeze-thaw processes within the hummock. For instance, where thufur are very small, warming on the northern and eastern aspects will counteract freeze penetration from the southern aspect, thereby limiting the depth of freeze and possibly thufur expansion. Hence, it is possible that a size threshold needs to be overcome before a thufa can expand or grow rapidly, particularly in marginal periglacial environments. However, the control of thufur size on micro-topographically controlled freezing processes still needs to be

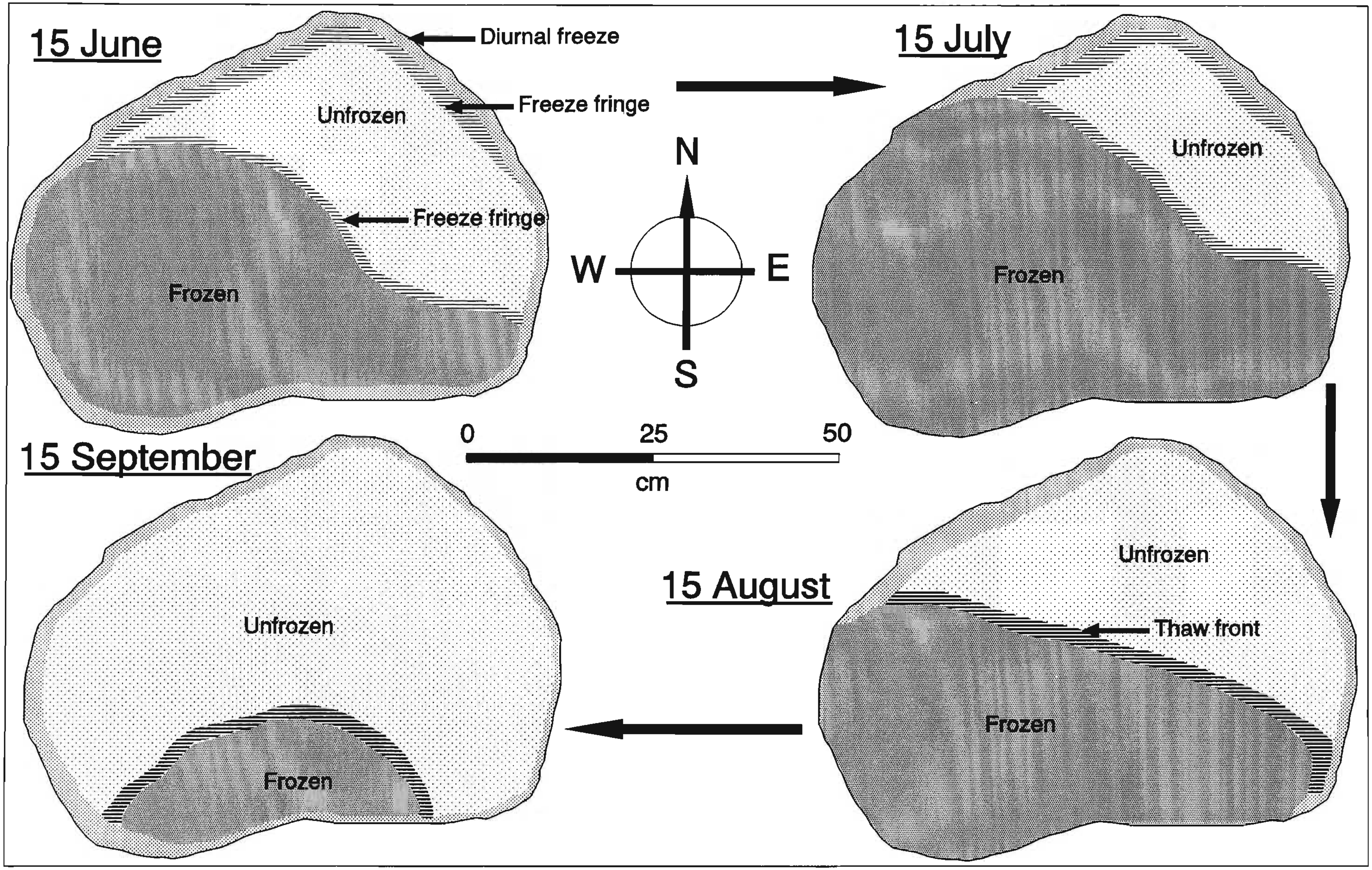


Figure 5.46 Ground freezing within a thufa (June - September 1995).

during the onset of winter. Towards mid-winter the freeze process intensifies on the southern aspect and the level of insolation received on the northern and eastern aspects decreases, permitting more rapid freeze penetration towards the thufur northern aspect, and to a much lesser extent the eastern aspect (Figure 5.46 and 5.47).

From the 1996 data it is evident that diurnal freeze occurs only at a shallow depth on the thufur northern aspects. Towards mid-winter a freeze fringe develops at a shallow depth (possibly 3 to 15 cm) on the northern aspects while diurnal freeze-thaw cycles occur nearer the surface (1 cm depth). Therefore, under snow-free conditions two freeze fronts may develop: a primary front on the thufur southern and western aspects that migrates north and east, and a secondary front that develops on the thufur northern aspect but usually does not penetrate far into the thufur owing to heat flow at the surface and considerable heat storage on the thufur northern and eastern aspects (Figure 5.47). Depending on external factors (air temperature and the insulating effect of snow) and thufur size factors, the primary and secondary fronts may meet towards mid-winter (usually early July), causing almost completely frozen thufur. However, the thufur northern aspects remain unfrozen for a considerable time into the winter season and is also the first aspect to be subjected to thaw (the eastern aspect usually remains unfrozen throughout winter, except for diurnal freeze at the surface). It was thus found that thufur southern aspects may be continuously frozen for 107 days or more while the northern aspect may be frozen for only 65 days.

There is a significant difference in the freeze intensity and duration on the opposing thufur aspects, which would also exercise some control on the heaving pressures. Marginal freeze on the thufur northern aspect may cause some segregation ice to develop near the frost line causing primary heave (Figure 5.47). The more intense freeze on the southern and western aspects of thufur causes segregation ice formation (personal observations from sectioned thufur). Cryosuction causing water-flow towards the penetrating freeze fringe enables progressive growth of lenses towards the thufur centres and northern aspect. These lenses may then initiate secondary frost heave (Krantz and Adams, 1996) on the western and southern sides of thufur (Figure 5.47). Clearly, the micro-topographically controlled freezing process in marginal periglacial environments (or at high latitudes where aspect may exercise an

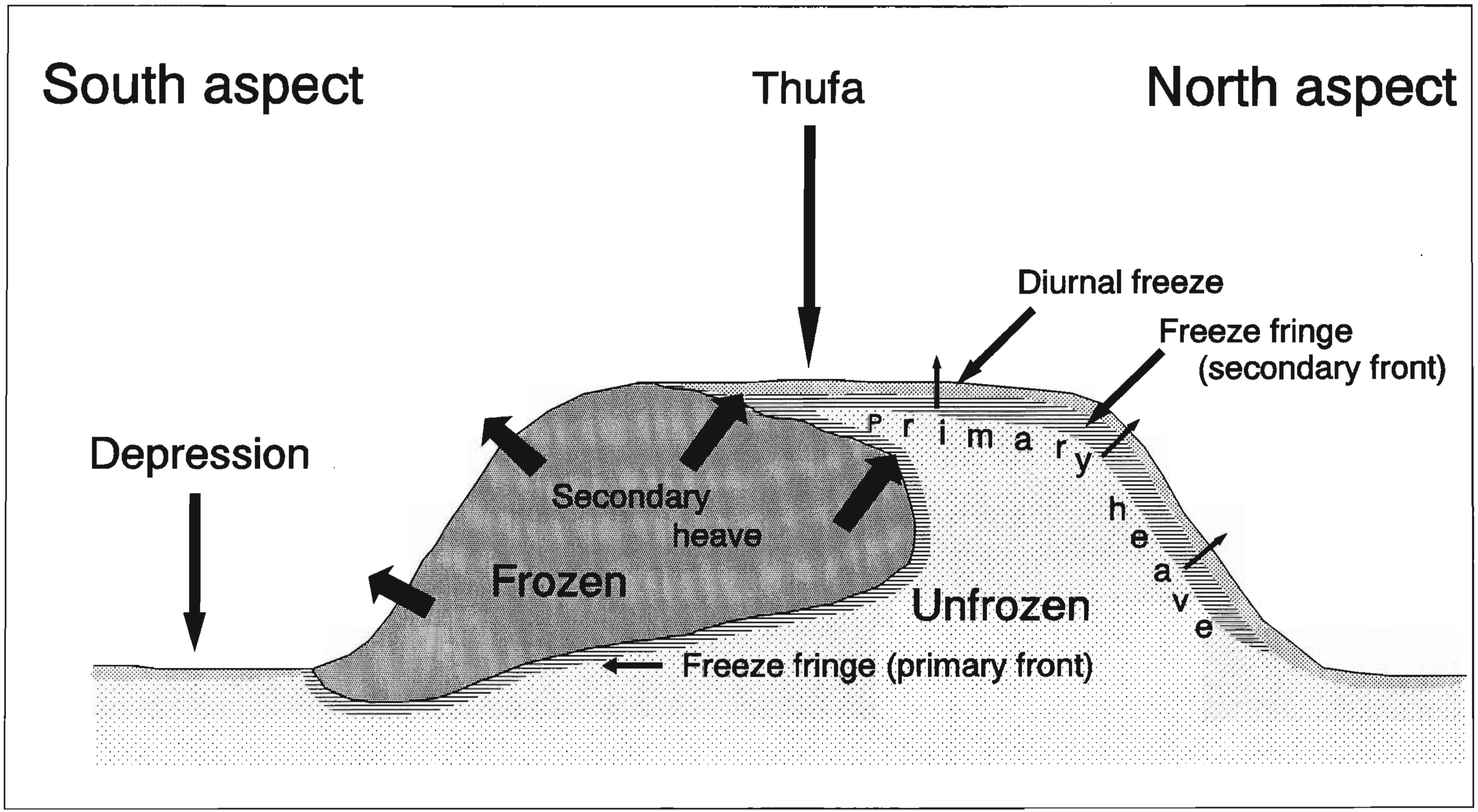


Figure 5.47 Side-sectional view of soil freezing processes within thufur.

important control) is more complicated than is commonly recognized. Past research (e.g. Mackay, 1980; Schunke, 1981; Van Vliet-Lanoë, 1991) has assumed that freezing occurs from the apexes down into the thufur and the application of the circulation cell model is largely based on the assumption of such a freezing process. What these models have failed to recognize is that the freezing process within thufur may be topographically restricted and that resultant heave processes may be more complicated. The present findings suggest that limited freeze may cause localized heave within thufur. It is imperative, however, to recognize that microclimatic, microtopographic, vegetational and other factors can influence the nature of thufur freezing processes. Thus, there may be a variety of thufur developmental processes in different environments, causing a diversity of different thufur morphologies.

The effect of snow on freeze-thaw processes within thufur

According to Matsuoka (1996), snow tends to reduce daily temperature ranges and prevents diurnal freeze-thaw penetration into the ground. It is commonly accepted that snowcover in cold environments acts as a barrier against heat loss (Williams and Smith, 1989) and that warmer ground temperatures during winter are associated with higher minimum temperatures due to the insulating effect of snow (Smith, 1975). However, the geomorphic effect of snow can be more complicated. For instance, Seppälä (1994) found that snowdrifts along the edge of a palsa reduced seasonal frost penetration on the palsa sides and consequently limited the lateral extension of the frozen core. In other areas it has been observed that snow drifts protect vegetated earth hummocks from winter blizzards, thereby permitting vegetation to become well established at such sites (Kojima, 1994).

The July 1996 snowfall reduced ground temperature differentials between northern and southern thufur aspects from 5°C to 1°C at 5 and 10 cm depth. However, it emerged from the study that pre-snowfall ground thermal conditions on opposing thufur aspects considerably influences the thermal regime during the period of snowcover. The variations in snowdepth and the duration of snowcover on northern and southern aspects also have a significant effect on freeze-thaw activity within thufur. Thus, on the southern aspect where the ground had already frozen, the snow acted as a barrier to heat loss and prevented nocturnal cooling, thereby increasing mean ground temperatures. On the northern aspect, however, where temperatures remained slightly above 0°C at 1 cm depth, the snow substantially reduced insolation receipt at the surface, thereby permitting ground cooling to below 0°C. The deep snowcover therefore reduces the rate of freeze penetration into the thufur from the southern and western aspects but permits shallow ground freeze to develop on the thufur northern and eastern aspects. Ground temperatures on the northern aspect were lowered during the time of snowcover, as opposed to the snow-free period during June.

The snowcover substantially reduces ground temperature ranges at all depths within thufur as it induces a lower thermal diffusivity between the air and the ground (Williams and Smith, 1989). Thorn (1979) suggested that the reduction in daily temperature ranges is largely owing to a drop in minimum, rather than rise in maximum, temperatures. However, the extent to which snowcover affects daily ground temperature ranges depends on the existing ground temperatures. Where temperatures have already been lowered below the zero curtain effect, temperature ranges were between 1 and 2°C during the period of snowcover. On the northern aspect where temperatures were maintained within the zero curtain effect, temperature ranges were reduced to less than 1°C. It is therefore suggested that greater precaution be taken when predicting reductions of temperature ranges with increasing snow depth (e.g. Jumikis, 1977).

Differential snowmelt occurs on the various thufur aspects. On the thufur northern and eastern aspects, snow thaws more rapidly than on the southern and western aspects where there is prolonged and gradual melt. Consequently, ground temperatures on the northern aspect rise very rapidly as the snow cover disappears and this permits ground thaw commencement to considerable depth (to over 15 cm) within about 2 days of thaw beginning at the thufur surface. The slow reduction of the snowpack depth on the southern and western aspects induces a gradually increasing intensity of nocturnal cooling, yet keeps daily temperatures at the surface below 0°C. This consequently increases the diurnal temperature range and freeze indices on thufur southern and western aspects.

The soil freeze process during June of 1996 was similar to that recorded in 1995 (Figure 5.48). The snowcover during July and August of 1996, however, facilitated freeze on the northern and eastern aspects, with only a small core of unfrozen ground remaining at depth on the northeastern aspect (Figure 5.48). Temperatures at the thufa surface on the northern and eastern aspects remained within the zero curtain throughout the period of snowcover. However, once the snow had disappeared, ground surface temperatures on the northern and eastern aspects were again subjected to diurnal freeze-thaw cycles.

The present study has demonstrated that initial ground temperatures need to be taken into consideration when determining the effect of snow on ground thermal properties. Even though thufur are relatively small cryogenic landforms, variations in snowpack depth, longevity and rate of melt, occur on the different thufur aspects, which contribute towards the aspect-controlled ground temperature differentials and freeze-thaw processes found within thufur.

5.2.7.2 A model for thufur elongation and coalescence

Although it is recognised that non-sorted patterns may become elongated and even coalesce (Van Vliet-Lanoë, 1991; Lewkowicz and Gudjonsson, 1992; Mark, 1994), there appears to have been little attempt to model such pattern alteration. The model, depicted in Figure 5.49, examines thufur elongation and coalescence, and is based upon detailed field observations and available ground temperature data over a period of three years in the high Drakensberg.

It was observed that thufur freezing is micro-topographically controlled, with a freeze fringe developing on the thufur south- and west-facing sides (Figure 5.49). Although near-surface soil at the thufur north- and east-facing sides may experience nocturnal freeze, this usually thaws during the day as north- and east-facing thufur sides receive considerable insolation. The freeze fringe then gradually progresses into the thufur from the south- and west-facing sides throughout much of winter (Figure 5.49). It is likely that the probable direction of maximum heave would thus be concentrated on the south- and west-facing thufur sides. Over a long period of possibly several decades, this may contribute towards thufur elongation in a southerly and/or westerly direction.

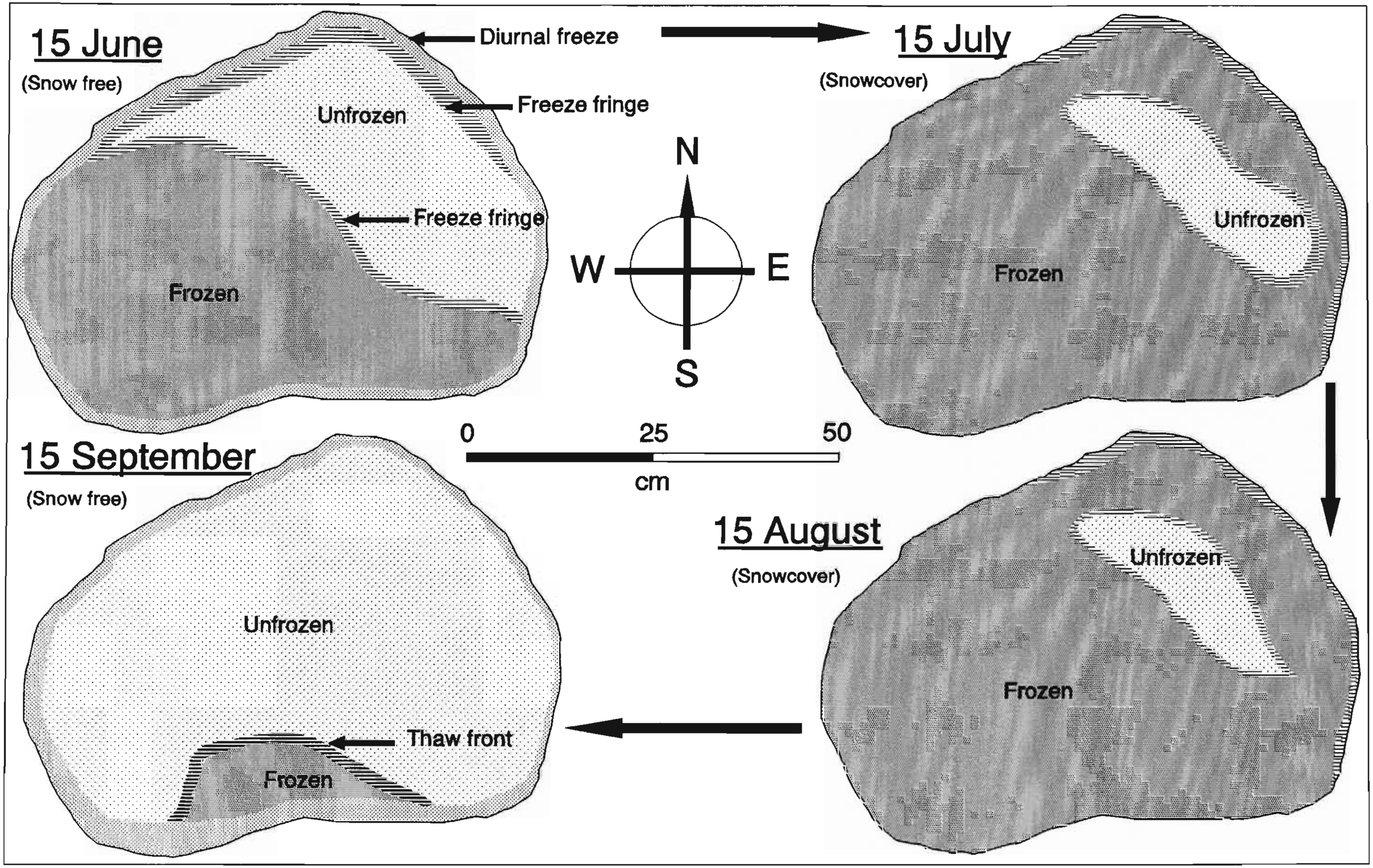


Figure 5.48 Ground freezing within a thufa (June - September 1996).

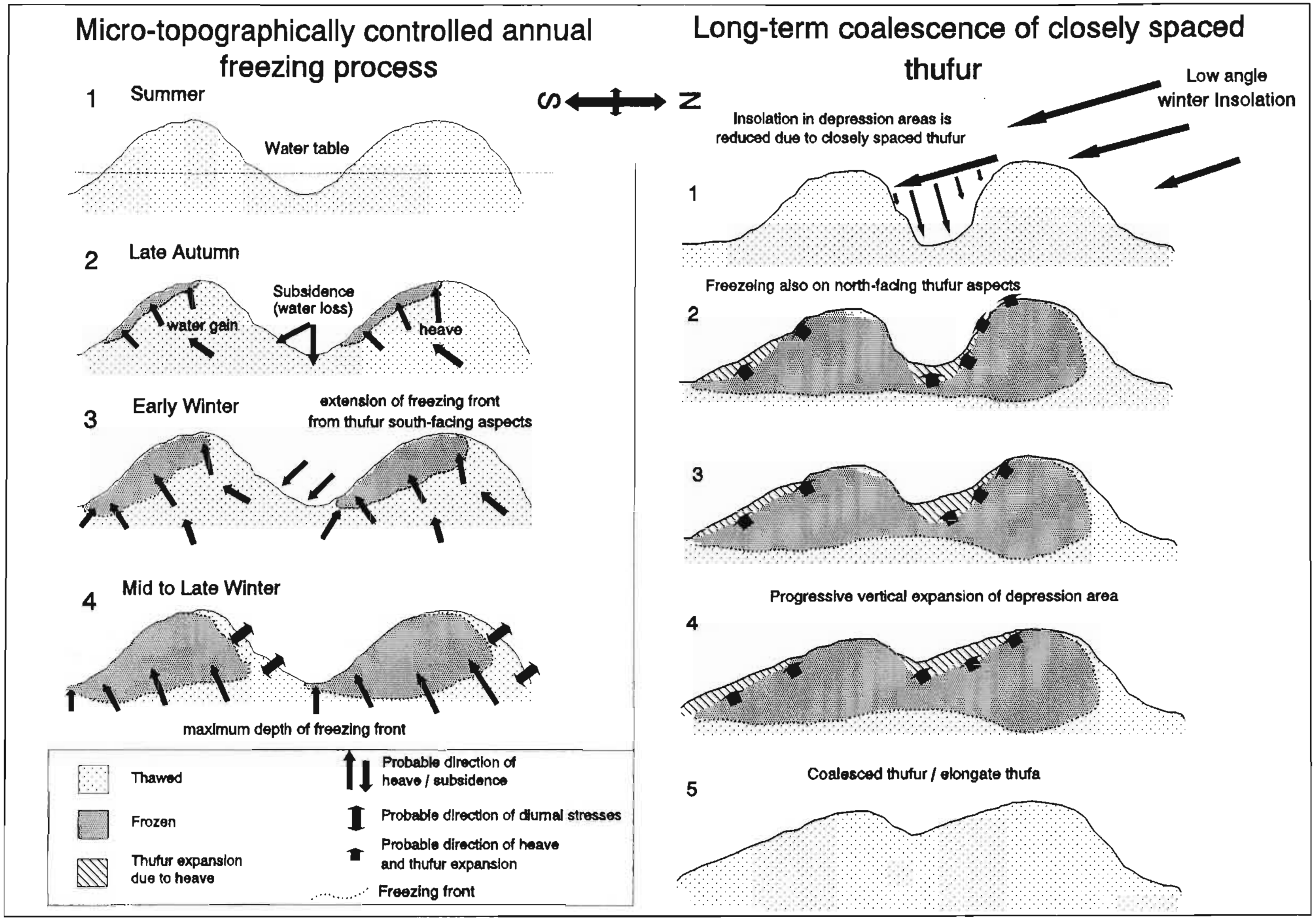


Figure 5.49 A model for thufur elongation and coalescence.

Where thufur are closely spaced, the north- and east-facing sides may become more protected against insolation owing to increased shading (Figure 5.49). Here it was observed that a freezing front penetrates the thufur from all sides (Figure 5.49). Over a period of possibly several decades, the continued thufur expansion (elongation), primarily from the south- and west-facing sides, may ultimately induce coalescence with an adjoining thufa (Figure 5.49).

It would appear that microtopography, such as thufur, significantly influences the soil freezing process in the high Drakensberg. In a climatically marginal periglacial region such as the high Drakensberg, the micro-topographically-controlled freezing processes may be of paramount importance in maintaining and modifying the cryogenic landforms that occur there. It is possible that in other cold environments the causes for thufur elongation and coalescence are somewhat different from those described here, and therefore, the model may not be applicable to all thufur studies (c.f. Lewkowicz and Gudjonsson, 1992).

5.2.7.3 Thufur development and activity

Although several hypotheses have been presented to explain the possible formation of thufur (e.g. Schunke, 1977, 1981; Mackay and Mackay, 1976; Tarnocai and Zoltai, 1978; Mackay, 1980), the precise mode of their formation has remained unclear (Zoltai and Pettapiece, 1974; Schunke and Zoltai, 1988; Gerrard, 1992). Van Vliet-Lanoë (1991) has suggested that differential frost heaving is probably the most consolidatory process that explains such pattern formation. This hypothesis, however, probably better explains processes occurring in pre-existing patterns rather than the actual initiation of the landforms.

It is commonly argued that fine-grained soils with a high percentage of silts are most susceptible to heave and consequent thufur development (Chamberlain, 1981; Zoltai and Tarnocai, 1981; Van Vliet-Lanoë, 1991). According to Van Vliet-Lanoë (1991), silts are frequently injected into the hummocks, sometimes creating a clearly visible contact with sandy soils. As already discussed in Section **5.2.3.4**, fine-textured soils are displaced into the thufur apexes while depressions contain considerably coarser material. Thus, although convoluted horizons may not be visible, the soil characteristics suggest recent cryogenic activity. This contention is further supported by the thermal characteristics of the thufur, as well as heave and moisture characteristics during the winter months. Zoltai and Pettapiece (1974) and Zoltai (1975) have also found that in dry years earth hummocks may show limited or no cryoturbation. However, in years where more moisture is available during the frost periods, there would be an increase in earth hummock development. Similarly, thufur development in the high Drakensberg and Lesotho mountains may be restricted during mild or "snow rich" winters but become active again during exceptionally cold periods.

Several discussions have examined parameters such as MAAT and the freezing indices which control the activity of certain frozen ground phenomena (Kersten, 1959; Williams, 1961; Lundqvist, 1969; Karte, 1983; Wayne, 1983). Although Karte (1983) suggests a MAAT of 3°C or lower for thufur activity, Lundqvist (1969) believes that thufur may form where the MAAT is at least as high as 5°C, while Koaze *et al.* (1974) have located active thufur where the MAAT is about 6°C. Even though the MAAT at altitudes above 3000 m in the high Drakensberg is probably between 4 and 6°C, the air temperature data are characterized by both seasonal and diurnal frost cycles, such that intense frost action is restricted to the winter months, which had a mean temperature of -1.7°C during 1994 in the Mashai Valley (Table 5.5). The general absence of cloud cover and snow, and a mean minimum air temperature of less than -6°C during June and July, permits ground freezing to depths of 30 cm or more, which may persist for two or three winter months. These findings support current thufur activity at favourable sites in the high Drakensberg under present climatic conditions, but it is restricted to the four coldest months, namely May to August. Dating of thufur in the high Drakensberg has never been undertaken and therefore it is difficult to determine a relative age for the landforms. However, the thufur fields have mostly developed in deep, fine-textured organic wetland soils and so it would appear that their maximum age is restricted to the age of the wetlands. Most wetlands are thought to be between 2000 and 5000 years old (Van Zinderen Bakker, 1955; Hanvey and Marker, 1994).

5.2.7.4 Thufur as Palaeoclimatic indicators

Recently, Nyberg and Lindh (1990) and Cui and Song (1992) have shown the value of using periglacial landforms such as permafrost mounds and earth hummocks as climatic indicators. Nyberg and Lindh (1990) also mention that periglacial landforms are useful in the reconstruction of short-term climatic fluctuations, particularly as periglacial landforms adjust to new prevailing environments (Priesnitz, 1988). It is possible that the geomorphic systems in the high Drakensberg are adjusting to a changing climate, but quite how it is changing is still not known. Although thufur may provide palaeoclimatic information, Tarakanov (1991) and Gerrard (1992) warn that thufur cannot be used as the sole criterion for an interpretation of past and present physico-geographical conditions. It is therefore suggested that the thufur, together with other periglacial landforms in the high Drakensberg, could be used as climatic and environmental indicators.

5.3 NON-SORTED STEPS

Non-sorted steps are defined as "patterned ground with a steplike form and a non-sorted appearance due to a downslope border of vegetation embanking an area of relatively bare ground upslope" (Washburn, 1956, p834). Patterns with a semicircular or oval appearance and corresponding with Washburn's (1956) definition, were located on a 3° terrace slope on the Mafadi northeast face, at approximately 3400 m a.s.l. (Figure 2.5).

5.3.1 Characteristics

The three non-sorted patterns measured have bare clastic surfaces which are surrounded by grasses (Figures 5.50 and 5.51). The vegetated fronts on the pattern downslope sides appear to have been pushed forwards to form turf-banks of up to 42 cm high (Figures 5.50 and 5.51). The across-slope pattern diameters are somewhat larger (mean = 3.32 m) than the downslope diameters (mean = 2.37 m) (Table 5.10) and so it would appear that the patterns have developed across the slope to form non-sorted steps. A study of non-sorted steps in the Kosciusko area, Australia, also found greater across-slope widths than downslope lengths

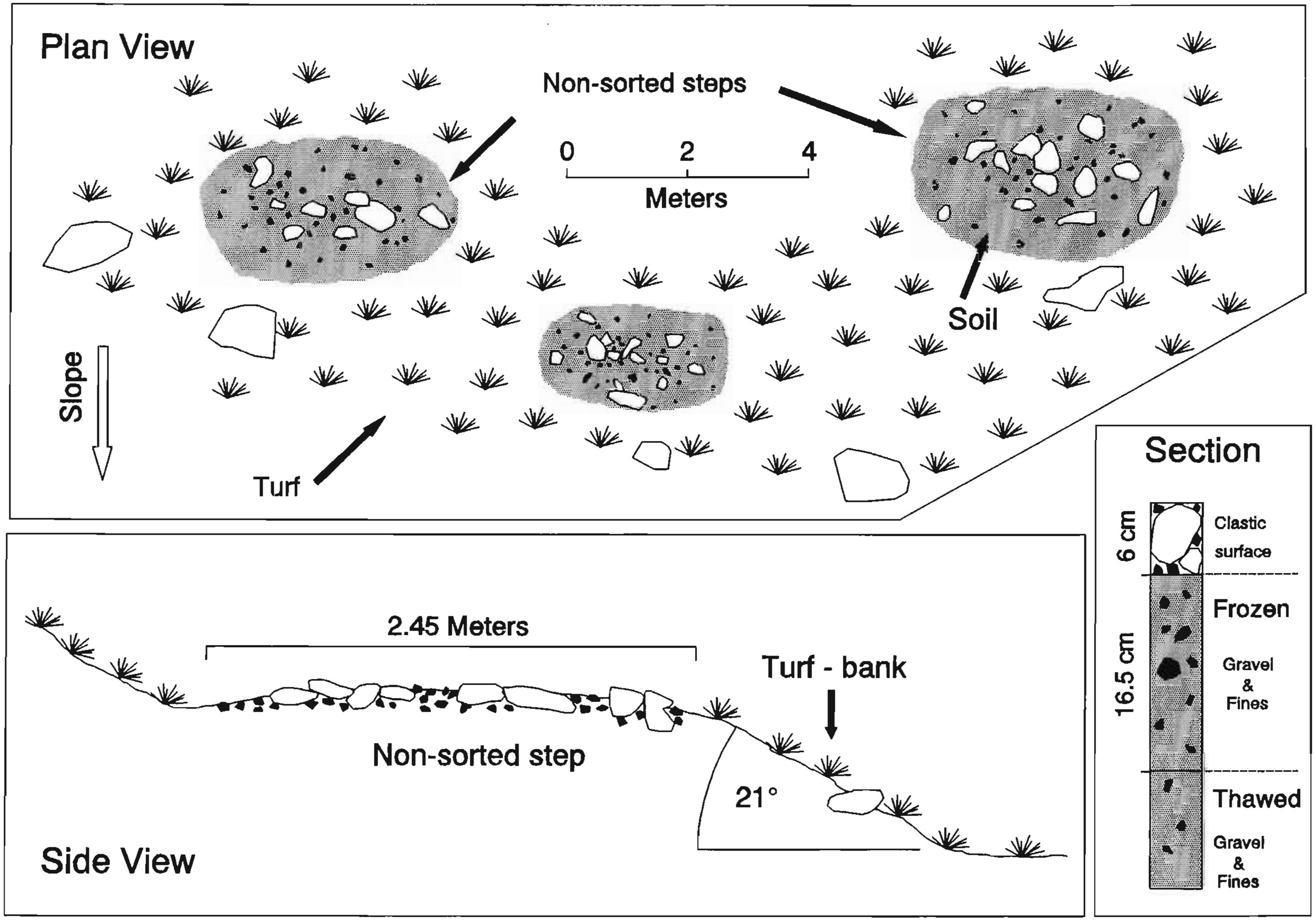


Figure 5.50 Sections through non-sorted steps at Mafadi Summit (July 1994).

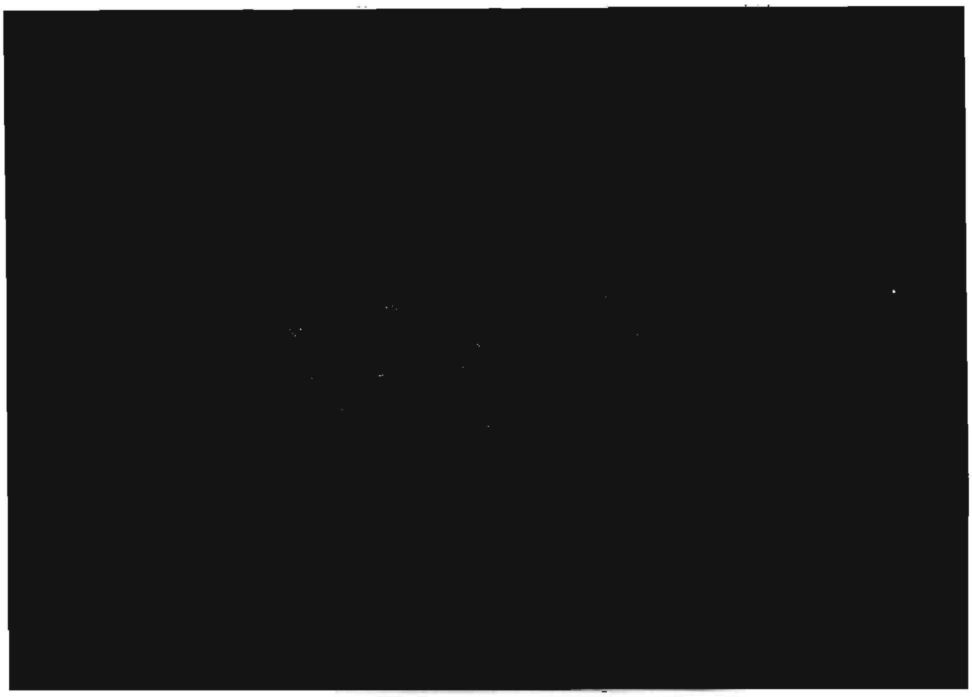


Figure 5.51 Non-sorted steps on Mafadi Summit.

Pattern		Diameter		Ratio of	Heave above surrounding	Tread
	Dd	Ad	Mean	Ad to Dd	ground vegetation	height
		(cm)			(cm)	(cm)
1	2.45	3.48	2.96	1 : 0.7	6.4	41.0
2	1.77	2.82	2.29	1 : 0.63	6.8	42.0
3	2.89	3.67	3.28	1 : 0.79	7.0	32.0
Dd = Downslope diameter / Ad = Across-slope diameter						

Table 5.10 Characteristics of non-sorted steps at Mafadi Summit.

(Costin *et al.*, 1967). During July 1994, the pattern centres were heaved between 6.4 and 7.0 cm above the surrounding vegetated area (Table 5.10) and were frozen to a depth of 22.5 cm. Miniature sorted circles and polygons developed on the gravely surface, indicating secondary sorting within the non-sorted steps. The surface consists primarily of flat, angular clasts. Downward fining from the surface is evident, as is displayed in Figures 5.50 and 5.52. The percentage gravel drops from 96.4% at the surface to 62.2% at 6 cm depth, while the percentage silt/clay rises from 0.8% at the surface to 13.4% at 6 cm depth (Figure 5.52).

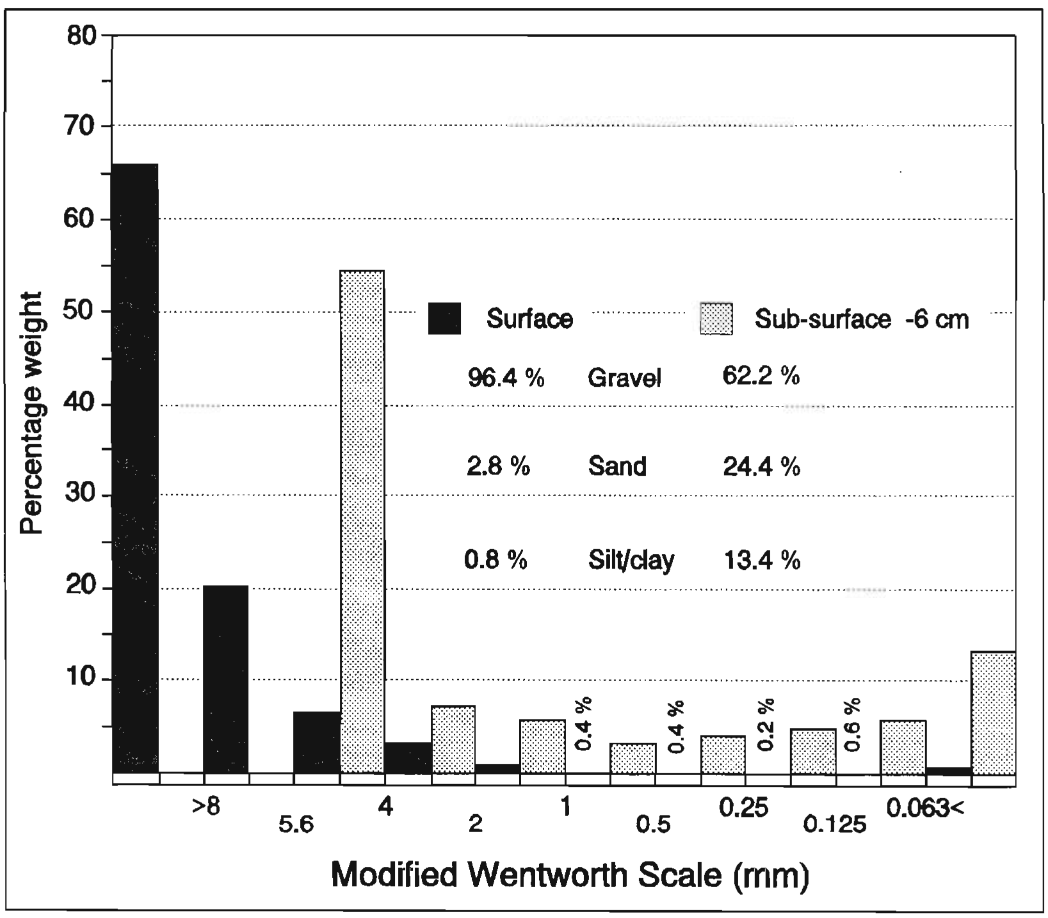


Figure 5.52 Particle size distribution for non-sorted steps at Mafadi Summit.

5.3.2 Discussion

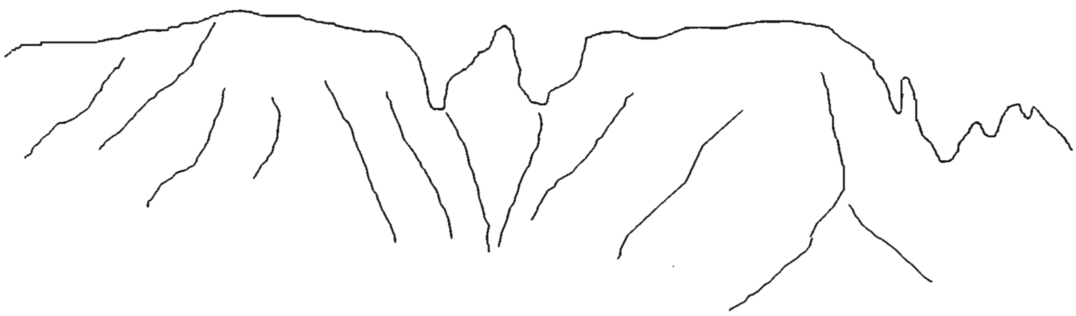
Non-sorted steps such as those described here (Section **5.3.1**), have so far only been found on Mafadi Summit, suggesting that these are uncommon in the high Drakensberg. The steps have formed on a terraced slope and appear to be lithologically constrained. The slope terraces have probably been subjected to considerable weathering owing to the observed high joint densities. At such sites, the weathered profile (regolith) is sufficiently deep (> 50 cm), and contains adequate fines (13.4%), to enable frost heave mechanisms to occur. The patterns appear to be seasonally active or may be in a state of quasi equilibrium. The absence of vegetation on the steps and the slight heaving of pattern centres during the winter months provides some evidence for contemporary cryogenic activity. The formative processes could, however, not be determined and further research is required to ascertain whether such non-sorted steps are still actively forming today.

5.4. SUMMARY

This chapter has described thufur and non-sorted steps from the high Drakensberg. In addition, the first detailed assessment of thufur thermal characteristics has been presented and discussed. Monitoring thufur activity in the high Drakensberg may also provide valuable information on the changing wetland conditions. It is therefore suggested that future research be aimed at determining the value of thufur as local environmental indicators. In addition to sorted and non-sorted patterned ground, other conspicuous high Drakensberg cryogenic landforms include various mass-wasting forms and debris deposits; these are to be discussed in the following chapter (Chapter **6**).

CHAPTER SIX

MASS-WASTING FORMS AND DEBRIS DEPOSITS



CHAPTER 6

MASS-WASTING FORMS AND DEBRIS DEPOSITS

6.1 INTRODUCTION

Washburn (1979, p192) describes mass-wasting as "the movement of regolith downslope by gravity without the aid of a stream, a glacier, or wind". During the last few decades, periglacial researchers have focused considerable attention on mass movement features and slope processes in cold environments (e.g. Flint and Skinner, 1977; C. Harris, 1981; Lewkowicz, 1988; Hirakawa, 1989; Bennett and French, 1991).

Solifluction is considered to be one of the most widespread processes of soil movement in periglacial regions (Egginton and French, 1985) and is defined as "the slow flowing from higher to lower ground of masses of waste saturated with water" (Andersson, 1906, p95). C. Harris (1981) has shown that the understanding of mass-wasting processes and their terminology has encountered some problems. For instance, C. Harris (1981) discusses the conflicting usage of the term "solifluction" and recommends that it be re-defined. Because solifluction may occur in non-periglacial environments, terms such as "congelifluction" (Dylik, 1951) and "gelifluction" (Baulig, 1956, 1957) have been introduced to refer to solifluction processes specific to periglacial environments. Washburn (1967, 1973, 1979) has accepted the term gelifluction to describe soil flow associated with seasonally frozen ground and permafrost. However, the term solifluction is still preferred where uncertainty of the origin of mass-wasted topographic features prevails. Terminological confusion and uncertainty regarding mass movement features occurring in the Drakensberg and Cape mountains has not escaped southern African researchers. For instance, Linton (1969) uses the terms "gelifluvial movement" and "gelifluction sludges", while Lewis (1988a) refers to "congeli-solifluction" to describe apparent solifluction processes and landforms.

Although several publications (Hanvey *et al.*, 1986; Hanvey and Lewis, 1991; Lewis and Hanvey, 1991) have reported and described slope deposits from the East Cape Drakensberg, little has been said about deposits occurring in the high Drakensberg. Lewis (1994a) and Lewis and Hanvey (1993) have respectively argued for apparent protalus ramparts and rockglaciers in the East Cape Drakensberg, but after having completed a three year survey, the present writer has yet to find conclusive evidence for such remnant landforms in the high Drakensberg. Nonetheless, mass-wasting forms and deposits are common on south-facing slopes in the high Drakensberg, but have yet to be examined in detail. Although Boelhouwers (1994) has given an account of mass-wasting forms that occur at Giant's Castle (3140 to 3300 m a.s.l.), detailed studies of individual landforms are still required. It has been argued that mass movement phenomena are important in the evolution of the paraglacial / periglacial landscape (Johnson, 1984). An understanding of mass-wasting forms and processes in the high Drakensberg may help to explain landscape evolutionary processes and features such as asymmetric valleys.

Troll (1944) grouped solifluction into a "macro" (non- needle ice induced) and "micro" (needle ice induced) component. This chapter examines only "macro-solifluction" forms and excludes "micro-solifluction" forms produced by needle ice action. Several landforms, such as ploughing blocks and protalus ramparts, have as yet not been found in the high Drakensberg and will therefore not be discussed in this dissertation.

6.2 DETRITAL ACCUMULATIONS

A great variety of detrital accumulations such as blockfields, blockstreams and stone-banked terraces have been described in the literature (e.g. Dahl, 1961; Dahl, 1966; Benedict, 1976; White, 1976; Caine, 1983; Tyurin, 1983; Nesje *et al.*, 1988; Nesje and Dahl, 1990). In southern Africa, such large detrital accumulations have been reported from the Cape mountains (Borchert and Sänger, 1981; Hagedorn, 1984; Sänger, 1987) and high Drakensberg (Sparrow, 1971; Hastenrath and Wilkinson, 1973; Boelhouwers, 1994). Given the widespread occurrence of such landforms in the high Drakensberg, it is important that they are discussed here. Several block-material accumulations observed at the study sites will, therefore, be discussed in the context of the available literature.

6.2.1 Blockfields and Blockslopes

A well recognized and widely used definition for blockfields appears not to have been forthcoming. However, according to Nesje *et al.* (1988, p149-150) and Nesje and Dahl (1990, p226), after Fairbridge (1968), autochthonous blockfields "normally consist of *in situ* angular boulders and stones up to several metres thick and formed through mechanical and chemical weathering of local bedrock". Washburn (1969, 1979) explains that blockslopes are similar to blockfields, but occur on slopes rather than gentle gradients. Synonymous terms used to describe blockfields include "blockmeere" (e.g. Meier, 1987), "felsenmeere" (e.g. Sugden and Watts, 1977) and "boulder fields" (e.g. Kleman and Borgström, 1990). Others (King and Hirst, 1964; Embleton and King, 1975; Washburn, 1979), however, have separated boulder fields from blockfields, where the former is said to consist of more rounded boulders and is sometimes associated with a marine origin, while the latter is more angular and a product of mechanical and chemical weathering.

Washburn (1979, p219) has defined blockfields as "considerable areas, broad and usually level or of only gentle gradient (less than 10°), covered with moderate-sized or large angular blocks of rock". However, Washburn's definition firstly requires blockfields to cover *considerable areas*, yet the minimum extent of the area is uncertain and not specified, and secondly, the rocks are required to be of moderate to large size, which would possibly exclude some of the so called blockfields described by Nesje *et al.* (1988), where only stones and smaller rocks are found. This would also apply to some sites in the high Drakensberg, such as on Mafadi Summit, which is strewn with stones and small rocks. These sites would, therefore, require a separate term, even though the detrital accumulation may have a similar origin to that of a blockfield. Other definitions are more general and easier to adopt. For instance, Potter and Moss (1968, p255) define blockfields as "..one of several different types of rubbly mountain deposit..", and, according to Embleton and King (1975, p167), blockfields are "accumulations of coarse detritus on level or gently sloping mountain-top surfaces..". The present author favours the definition used by Fairbridge (1968), Nesje *et al.* (1988) and Nesje and Dahl

(1990), as this definition eliminates two problems mentioned earlier, namely that of the local extent and material size of blockfields, yet is considerably more specific than those used by Potter and Moss (1968) and Embleton and King (1975).

6.2.1.1 Blockfields and blockslopes in the high Drakensberg

The presence of block-accumulations is well known in the high Drakensberg (Sparrow, 1971; Boelhouwers, 1994). However, as some of these are found on slopes of 16° or more (Boelhouwers, 1994), they are best referred to as blockslopes (following Washburn, 1979). The Drakensberg blockfields are usually restricted to the higher mountain summits and high altitude north-facing slopes, where the slope gradients are shallow (less than 10°). Blockslopes are found on the steeper south-facing slopes but may sometimes merge downslope into blockfields where there are somewhat broader basalt terraces. Sparrow (1971) has commented that blockfields occur above 1600 m a.s.l. in Lesotho, where they are said to sometimes cover extensive areas. Yet, the blockfields/slopes at the study sites are restricted to slopes above 3190 m a.s.l. and have a strong preferential occurrence on south- and southwest-facing slopes (Figure 6.1). In other regions, such as Scandinavia, blockfields also predominate at higher altitudes (Dahl, 1966) and it has been suggested that this is connected with altitudinally defined weathering boundaries (Nesje *et al.*, 1988).

The basalt steps described by Harper are explained by "frost wedging" attacking the scarp base (Harper, 1969, p82-83). Such basalt steps are common throughout Lesotho, and at higher altitudes these are covered by blockfields/slopes. Topographical and lithological conditions are said to play an important role in blockfield/slope development (Dahl, 1966; Potter and Moss, 1968; Tyurin, 1983; Ballantyne, 1984). It would appear that the basalt steps are zones of higher joint concentration, which are taken advantage of by mechanical and chemical weathering processes Boelhouwers (1988b). The presence of seasonal ice within rock fissures and at scarp bases supports the contention that frost-related mechanical weathering processes could be operative, as has similarly been suggested by Tyurin (1983) for other regions. Ground seepage along zones of weakness may also enhance other mechanical and chemical weathering processes. Several such scarps at higher altitudes (over 3100 m) have been

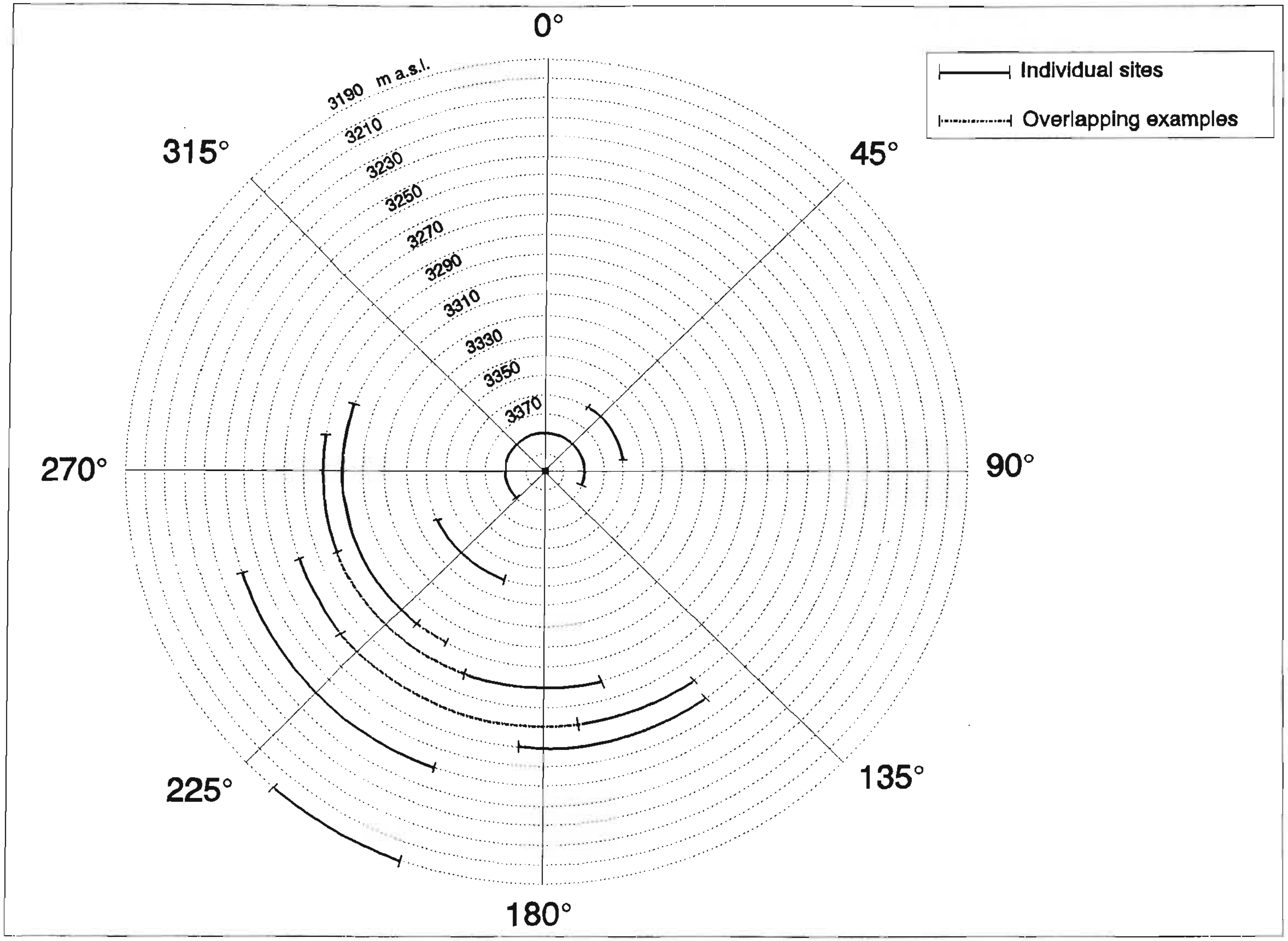


Figure 6.1 Altitudinal and aspect controlled locality of various blockfields/blockslopes in the study area.

weathered / eroded back substantially so as to be left with isolated "tor-like" scarp remnants with blockslopes at their bases. At lower altitudes, slopes below rock scarps are strewn with fewer rocks (less than 50% cover) and thus it would appear that weathering processes at higher altitudes have been more conducive to block-material accumulation than at lower altitudes.

Most of the block accumulations at high altitudes are not scree as rock scarps are usually less than four metres in height or absent at such sites, and so, free-fall and rolling of debris is minimal. Weathering appears to be the primary origin for most blockslope accumulations at the study sites. It is also possible that thermal creep, snow creep, gelifluction and other mass movement processes have displaced material downslope to form para-autochthonous or allochthonous blockslopes. It is difficult, however, to determine whether the accumulations are autochthonous or allochthonous, as this depends on the degree of block-material resorting (Ohlson, 1964). The presence of lobate and linear block accumulations upslope and downslope from some blockslopes, could suggest that material is feeding the blockslope from above and leaving it from below. In this way, block-material is moved from one blockslope-unit (basalt step) to that immediately below it (lower basalt step).

6.2.2 Lobate detrital accumulations

Lobate detrital accumulations include stone-banked lobes, sheets and terraces. The stone-banked lobes have been examined in greater detail than other detrital forms, and are thus discussed separately in this chapter (see Section **6.3**). C. Harris (1981) has indicated that solifluction terrace forms and lobate forms have no real distinction, such that stone-banked terraces may have lobate fronts, but are somewhat larger than stone-banked lobes. Stone-banked terraces are defined as "terrace- or garland-like accumulations of stones and boulders overlying a relatively stone-free moving sub soil" (Benedict, 1970, p174). Stone-banked terraces are usually the product of solifluction, with lengths of up to 400 m having been measured in Ireland (Coxon, 1988). Stone-banked terraces are reported to occur at the downslope margins of blockfields and where stone stripes spread out laterally or coalesce (Benedict, 1976). Although stone-banked sheets (Boelhouwers, 1994) and stone-banked terraces occur in the high Drakensberg, they are uncommon. At Njesuthi Summit, some stone-banked lobes have coalesced to form stone-banked terraces. Stone-banked terraces have also been observed at the downslope margins of a few blockslopes. It appears that these lobate detrital accumulations are solifluction forms, comprising debris transported from higher to lower slopes, thereby contributing to the overall slope evolution.

6.2.3 Linear block accumulations

Terms used to describe various forms of linear block accumulations include "sorted stripes" (Boelhouwers, 1994), "stone stripes" (Price, 1974; Coxon, 1988), "boulder streams" (Whittecar and Ryter, 1992), "rubble streams" (Richmond, 1962), "blockstreams" (Andersson, 1906; Jennings, 1956), "rock-streams" (Small *et al.*, 1970), "rock-rivers" (Costin, 1955) and "rock stream stripes" (Tyurin, 1983). Several of the above mentioned terms are synonymous or describe similar landforms. Some terminological confusion has arisen though, possibly because many linear block-forms have been inadequately studied, and as a consequence, definitions are vague. For instance, Embleton and King (1975) regard blockslopes and blockstreams as synonymous, while, according to Washburn (1979, p219), blockstreams are similar to blockslopes <u>but</u> "confined to valleys or forming narrow linear deposits extending down the steepest available slope". Tyurin (1983) uses the term "rock stream" to describe felsenmeer (blockfields), which is somewhat misleading, as this descriptive term is not intended to describe only linear block-forms. Other linear, downslope block features described by Tyurin (1983, p1285) are referred to as "rock stream stripes".

The contextual usage of the term "stone stripes" (Price, 1974) should also be questioned, as the landforms described by Price and others are not always "true" sorted or non-sorted stripes, as defined by Washburn (1956). The stripes described by Price (1974) are separated by considerably wider areas and are also sometimes orientated diagonally across the slope. According to Washburn (1956, 1979), stripes require *parallel lines* of some sort, and therefore, if the linear block accumulations do not form a parallel linear network, then they should not be referred to as stripes. It appears that the so called stone stripes are common on periglacial slopes and are a product of mass movement from the plateau edge (Coxon, 1988). The term "blockstream" has not been used for the above-mentioned landforms. This is possibly because the term "blockstream" has been reserved for larger linear block accumulations, rather than the relatively short and narrow ones found on higher mountain slopes. There is perhaps no need to differentiate between short, narrow block accumulations with somewhat longer, wider accumulations, particularly as these may have a common developmental origin.

6.2.3.1 Blockstreams in the high Drakensberg

Blockstreams occurring in the high Drakensberg have only briefly been mentioned by Hastenrath and Wilkinson (1973). A blockstream up to 1.2 km in length has been reported from a southeast-facing slope in the Thabana-Ntlenyana region, at an altitude between 2960 and 3180 m (Hastenrath and Wilkinson, 1973). Blockstreams are found on similar aspects and share a common altitudinal distribution to that of blockfields/slopes discussed in Section **6.2.1.1**. The blockstreams are particularly common on the southern and western aspects of Popple Peak, but are also found on other high summits such as Njesuthi and Thabana-Ntlenyana (Figures 2.5 and 2.6)

A blockstream was examined on the southwestern aspect of Popple Peak, at about 3325 m a.s.l. The summit of Popple Peak consists of a broad, flat area of rock debris. This debris appears to be moving over the shallow scarp face at various sites along the southern and western summit areas, often accumulating as lobate block deposits (Figures 6.2 and 6.3). A blockstream extends from below such a lobate block deposit and appears to be fed by the block mass above it (Figure 6.3). The blockstream extends 14.8 m downslope and attains a fairly constant width of about 3 m (Figures 6.3 and 6.4). A closer examination of the blockstream revealed that its elevation above the surrounding area is most pronounced near its upslope margin where it is being fed by the lobate block deposit, and again near its front where movement appears to have halted (Figures 6.3 and 6.5). A vertical section of the lobate block deposit indicates an open block network approximately 52.5 cm thick, underlain by what appear to be horizontal rock slabs which have been displaced from the rock scarp; bedrock is encountered at a depth of 119.5 cm (Figure 6.3). The open block network from the blockstream varied in thickness from 42 cm (blockstream front) to 74 cm (upslope margin)

Figure 6.2 An example of where debris appears to be moving over a shallow scarp face to produce a lobate block deposit at its base.

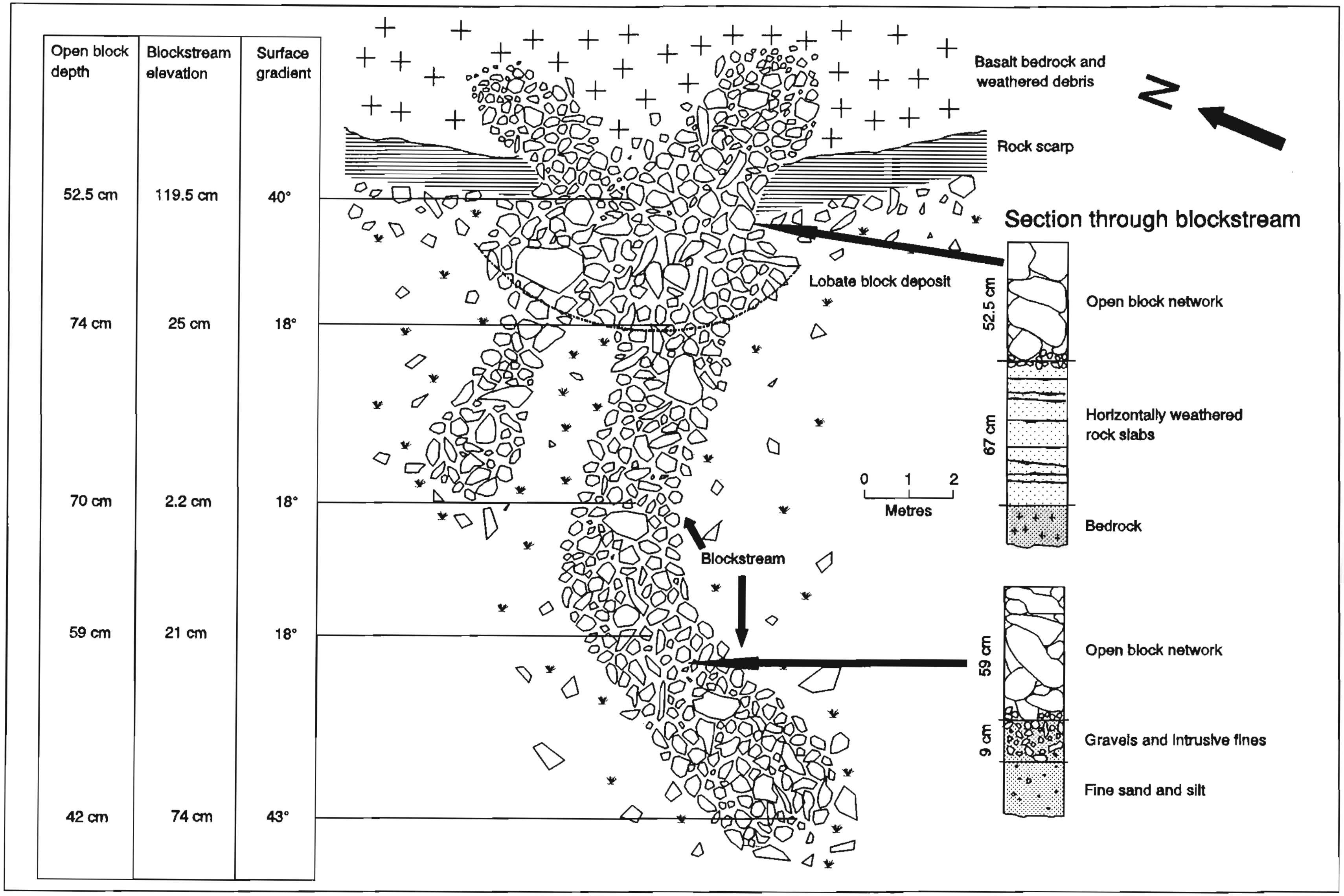


Figure 6.3 Sketch of the Popple Peak lobate block deposit and blockstream (3325 m a.s.l.).

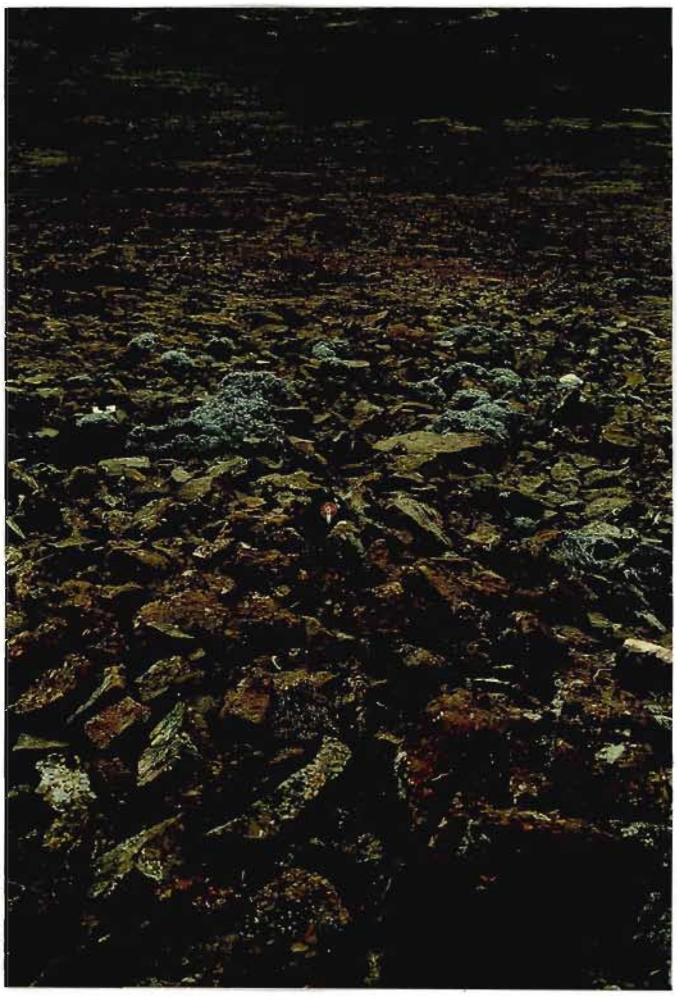


Figure 6.4 The Popple Peak blockstream which attains a length of 14.8 m and a width of 3 m.

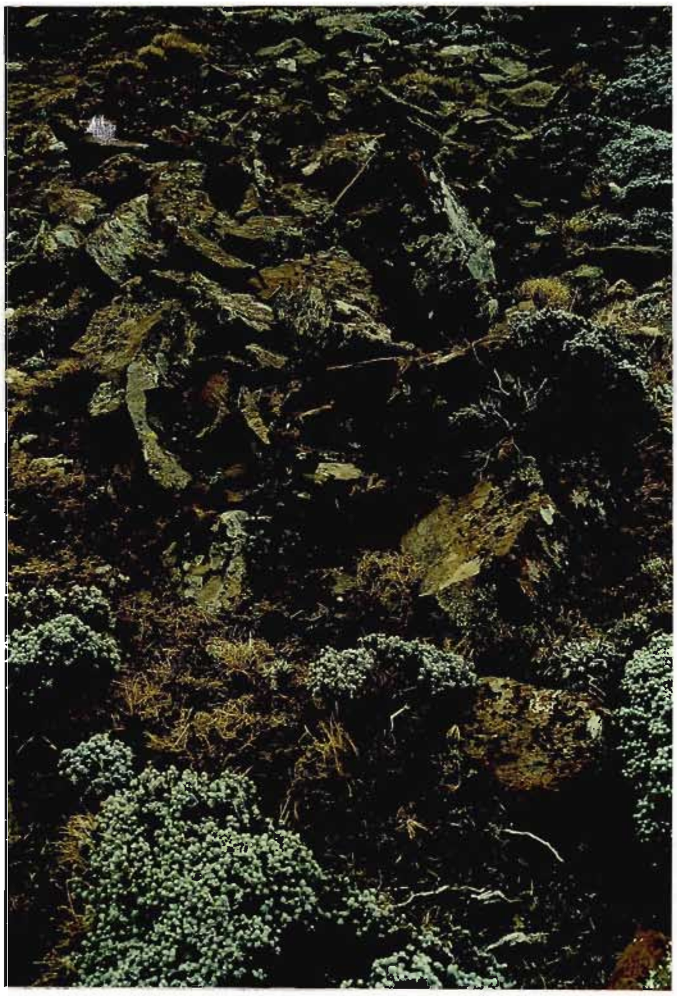


Figure 6.5 The frontal lobe of the Popple Peak blockstream which is slightly raised above the surrounding area.

(Figure 6.3). A section through the blockstream revealed an imbricated open block network of 59 cm thick, underlain by a 9 cm gravel unit containing few intrusive fines. Below the gravel unit is an organic silty sand with minimal gravel (Figures 6.3 and 6.6).

Figure 6.6 To show vertical sorting through a blockstream. An open block network at the surface overlies a gravel unit, below which are organic rich sediments.

It is possible that the blockstreams are structurally controlled and are a product of solifluction. Areas of greater joint density along the summit rock scarps would be most prone to weathering. Movement of debris from the flat summits over the rock scarps at zones of weakness could help remove weathered material, thereby resulting in the observed lobate block accumulations (Figures 6.2 and 6.3). The presence of broken horizontal slabs, disconnected from the underlying bedrock at the rock scarp, helps support this contention. The blockstreams are a downslope expression of these block lobes, particularly at gradients of over 15°. Tyurin (1983, p1285) has similarly discussed the structural control on rock stream development, where zones of weakness may produce "rock stream stripe facies" or "rock stream rampart-flow facies".

The soil structure of the blockstream is similar to that of stone-banked lobes and sheets at Giant's Castle, which is considered favourable for segregation ice development (Boelhouwers, 1994). This, together with the imbrication which suggests a former flow process, may indicate that macro-solifluction has been operative here. The presence of such blockstreams that are restricted to the cooler southern and western slopes of only the higher summits, would also support a periglacial mass-wasting hypothesis. Blockstreams emanating from below blockfields/slopes/lobes may be a downslope extension of such block accumulations. The extensive growth of moss and lichen covered rocks, together with well vegetated blockstream fronts, may be an indication that these landforms are presently inactive.

Larger blockstreams (100 to 200 m in length) were found between 3100 and 3200 m a.s.l. on the southern side of Popple Peak (Figure 6.7). These blockstreams are about 20 to 30 m wide and have an open block network of unknown depth. The rock blocks have an average a-axis of well over 0.5 m. These blockstreams require a more detailed examination than was possible here, before an attempt at explaining their origin can be made. Many of the basalt blocks have

Figure 6.7 A large blockstream (100 to 200 m in length) on the south side of Popple Peak (between 3100 and 3200 m a.s.l.).

weathered into a crumbly mass, and it would appear that they have remained *in situ* for a considerable time. Given the relatively low altitude of some blockstreams, and the highly weathered nature of many surface blocks, it is suggested that the larger blockstreams are of Late Pleistocene age.

6.3 STONE-BANKED LOBES

It appears that little periglacial work has concentrated on stone-banked lobes, resulting in few available data (Benedict, 1976) and their formation being incompletely understood (Embleton and King, 1975). Several authors (Sharp, 1942; Galloway, 1961; Benedict, 1966, 1970; R.B. King, 1972; Quinn, 1975, 1987; Hall, 1983; Harrison and Macklin, 1991; Van Steijn *et al.*, 1995) have presented some characteristics on stone-banked lobes (Table 6.1), however, data on particle movement, size, sorting, orientation and dip are scarce. In Africa, stone-banked lobes have been reported from the high Simien in Ethiopia (Hurni, 1982) and from the high Drakensberg (Dardis and Granger, 1986; Boelhouwers, 1994).

Benedict (1970, p176) has defined stone-banked lobes as "lobate masses of rocky debris underlain by relatively stone-free, fine-textured, moving soil", while Embleton and King (1975, p112) describe these landforms as "gelifluction and other deposits confined by crescent-shaped stony embankments". The definition provided by Embleton and King (1975) fails somewhat because it restricts the term to *crescent-shaped* forms, yet lobate stony embankments need not necessarily be crescent shaped, as both "V" and "U" shaped forms have also been found (pers. obs.). There is undoubtedly confusion surrounding the definition, terminology and characteristics of stone-banked lobes. For instance, Quinn (1987) differentiates between "stone-lobes" and "stone-banked lobes", where the former is absent of fines and the latter has fines present. Relict stone-banked lobes may have had fines washed out, yet, the lobate morphology may still be that of a stone-banked lobe, and therefore, should be referred to as such. "Tongue-shaped steps" have been described from the Mt. Kosciusko region, Australia (Costin *et al.*, 1967, p983), yet appear to be stone-banked lobes. Clearly, there is a need to standardise terminology and present a more rigorous definition for these features.

Author	Location	Slope gradient	Tread length (m)	Riser Height (m)	Lobe width (m)
Sharp (1942)	St. Elias Range, Canada	5° - 15°	up to 7.6	up to 0.6	1.2 - 2.5
Galloway (1961)	Scottish Highlands	8° - 20°	-	0.6 - 3	3.6 - 30.5
Lewis & Lass (1965)	Faroe Islands	>6°	-	0.6	-
Benedict (1970)	Colorado Front Range, U.S.A.	12° - 24°	-	<1	-
King (1972)	Cairngorms, Scottland	20° - 35°	9 - 180	-	-
Embleton & King (1975)	-	10° - 25°	-	-	-
Quinn (1975, 1987)	Knocknadobar, Ireland	-	Mean = 22.31	Mean = 5.63	-
Harris (1976)	Okstindan, Norway	2° - 24°	-	0.5 - 1.5	-
Hall (1981)	Marion Island	Mean = 13°	Mean = 0.93	Mean = 0.32	-
Harrison & Macklin (1991)	Okstindan, North Norway	7°	30 & 30.4	1.8 & 1.3	-
Van Steijn et al (1995)	Andes & French Alps (review)	4° - 35°	several metres	-	2 - 3
High Drakensberg					
Dardis & Granger (1986)	Champagne Castle	-	10 - 60	0.2 - 0.5	2 - 6
Boelhouwers (1991a)	Giant's Castle	-	0.9 - 4.5	0.2 - 0.7	0.5 - 2
Boelhouwers (1994)	Giant's Castle	14° - 18°	8 - 30	0.3 - 3	2 - 7

Table 6.1 Characteristics of stone-banked lobes as reported by various authors.

Lewis (1988b) has commented that stone-banked lobes have only been convincingly reported from Marion Island, however, more recently Boelhouwers (1991a, 1994) has produced some data on stone-banked lobe dimensions from the high Drakensberg (Table 6.1). Although Boelhouwers (1994) gives a more comprehensive account on stone-banked lobes found at Giant's Castle, data are still extremely limited and a further assessment is imperative if their environmental implications are to be understood. The stone-banked lobes reported by Dardis and Granger (1986) occur at altitudes as low as 3000 m and are also found on the warmer northwest- and northeast-facing flanks of a river valley. These lobes are said to be "actively eroding rock outcrops" (Dardis and Granger, 1986, p91), yet no explanation as to how lobes erode rock outcrops is given. Many stone-banked lobes on the cooler south-facing slopes at altitudes above 3300 m are presently inactive, so it is difficult to envisage the larger (up to 60 m in length) lobes at the warmer sites reported by Dardis and Granger (1986), as being active.

6.3.1 Characteristics

A variety of stone-banked lobe forms, varying somewhat in size and morphology, have been found on the higher summits at the study sites (Figures 2.5 and 2.7). Boelhouwers (1991a, 1994) has recorded stone-banked lobes at altitudes between 3150 and 3240 m at his Giant's Castle study site. At the present study sites, stone-banked lobes were found between 3300 and 3330 m a.s.l. at Popple Peak, between 3340 and 3400 m a.s.l. at Njesuthi Summit and at about 3300 m a.s.l. near Wilson's Peak (Figures 2.5 and 2.7). The smaller varieties are common at Popple Peak, while larger forms were only found on the Njesuthi Massif. The stone-banked lobes occur on south-, southwesterly- and westerly-facing slopes, primarily at aspects between 250° and 270°. The present distribution of stone-banked lobes indicates that they are restricted to the colder, rock-strewn slopes where vegetation is or has been sparse. The stone-banked lobes are commonly found near blockslopes and debris mantled summits. At Popple Peak, stone-banked lobes may be seen overriding low summit scarps, as debris is moved from the summit onto lower slope terraces (Figures 6.8 and 6.9).

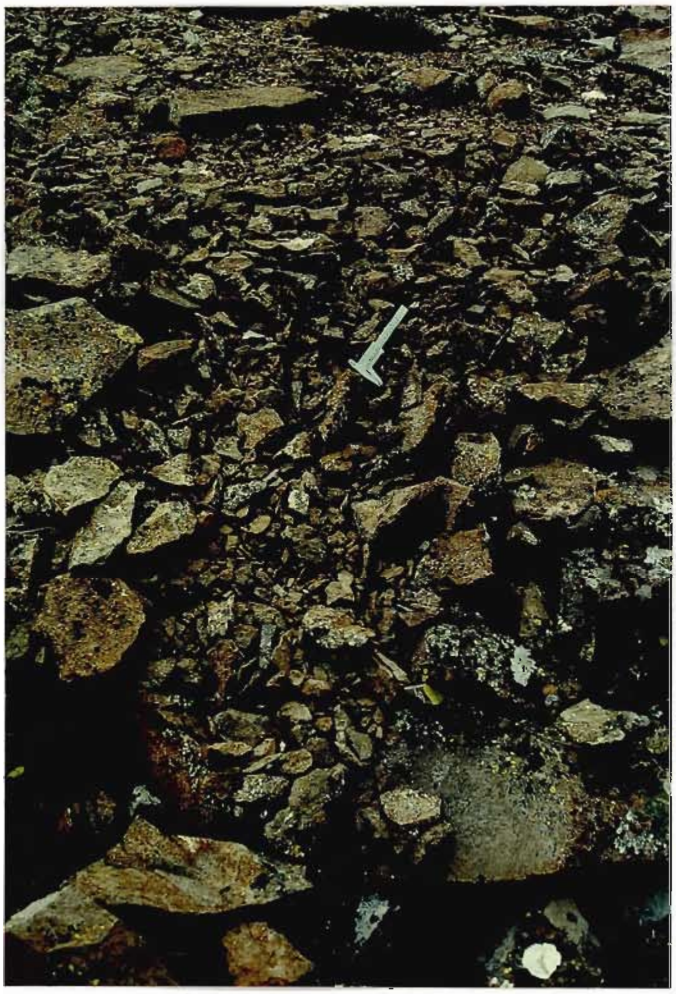


Figure 6.8 A stone-banked lobe moving over a shallow rock scarp at Popple Peak. Note the "fine-edged front" of the lobe.

Figure 6.9 A large stone-banked lobe moving over a shallow rock scarp at Popple Peak. Note the "rounded front" of the lobe.

Researchers have reported stone-banked lobes on slope gradients varying from 5° to 35° (Sharp, 1942; R.B. King, 1972). It appears, however, that they occur most frequently on slopes of 10° to 25°, as is suggested by Embleton and King (1975) (Table 6.1). In the high Drakensberg, stone-banked lobes have been observed on gradients of 14° to 18° at Giant's Castle (Boelhouwers, 1994) (Table 6.1), about 19° at Popple Peak and 15° at Njesuthi Summit.

6.3.1.1 Morphological varieties

A variety of stone-banked lobe morphologies are found in the high Drakensberg, and for ease of discussion, these are described separately.

Variety a (Refer to Figures 6.8 and 6.9)

These lobes are common on the southern and southwestern aspects of Popple Peak (± 3325 m a.s.l.). The stone-banked lobes vary from 1.4 to 4.0 m in length and comprise platy clasts averaging 7 cm (a-axis) in size at the surface. The highly weathered rock mantle on the near-level Popple Peak has had material moved along preferred but confined zones towards and over shallow rock scarps (Figures 6.8 and 6.9). At such sites, it appears that larger clasts have moved more rapidly and have been sorted into peripheral areas of the moving debris masses, thereby forming stone-banked lobes. Lobe shape is possibly controlled by bedrock morphology and large rock obstructions, resulting in a variety of stone-banked lobe shapes. For instance, Figure 6.8 shows a "fine-edged" lobe front, while the lobe depicted in Figure 6.9 has a "rounded front".

Variety b (Refer to Figures 6.10 and 6.11)
Variety *b* lobes are found on the boulder strewn slopes of Popple Peak, but are best developed on the Njesuthi Summit slopes. These lobes have a crescent-shaped stony embankment, similar to those described by Embleton and King (1975) and Hall (1981). Boulders and larger cobbles form the frontal bank, behind which are found smaller cobbles and gravels (Figures 6.10 and 6.11). These lobes sometimes contour, rather than occupy, the steepest available slope, thereby forming a stepped micro-relief. Similarly, Benedict (1970) has reported stone-banked terraces from the Colorado Front Range, which are believed to have resulted from the coalition of closely spaced lobes. Stone sorting phenomena on the lobe treads are common during the winter months, as has also been reported for lobes on Marion Island (Hall, 1981).

Variety c (Refer to Figures 6.12 and 6.13)
Variety *c* lobes have only been found on the Njesuthi Massif and contain more blocky material and have greater vertical and lateral dimensions than variety *a* lobes. Variety *c* lobes have surface rocks averaging about 36 cm (a-axis) in diameter. These lobes show less distinct sorting than variety *b* lobes, as the treads themselves are comprised of rock blocks. Nevertheless, a gradual decrease in rock size is found upslope from the frontal bank. Variety *c* lobes sometimes have a raised frontal bank, such that they are slightly elevated above rocks immediately upslope (Figure 6.13).

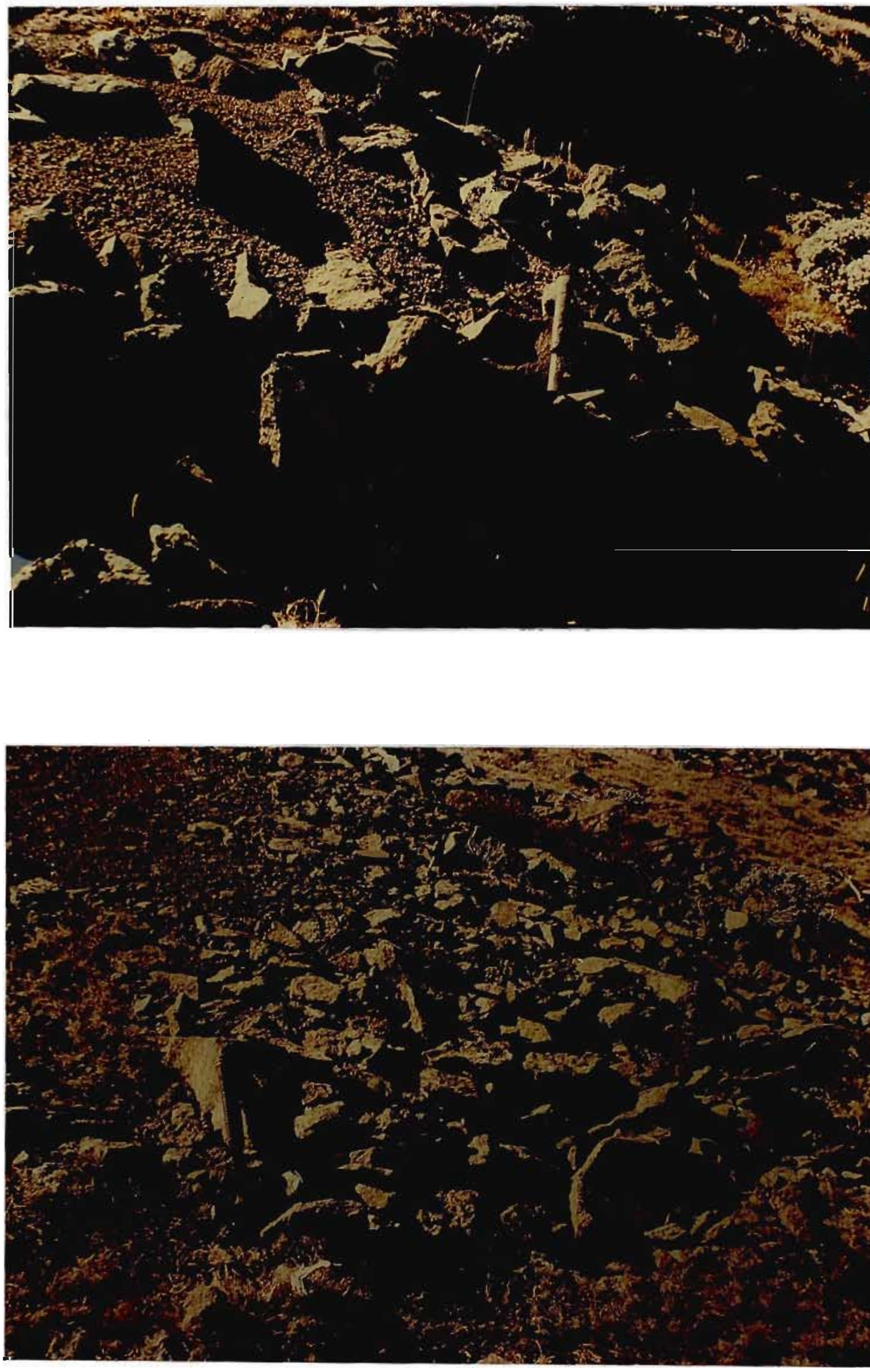


Figure 6.10 Examples of crescent-shaped stone-banked lobes on Njesuthi Summit.

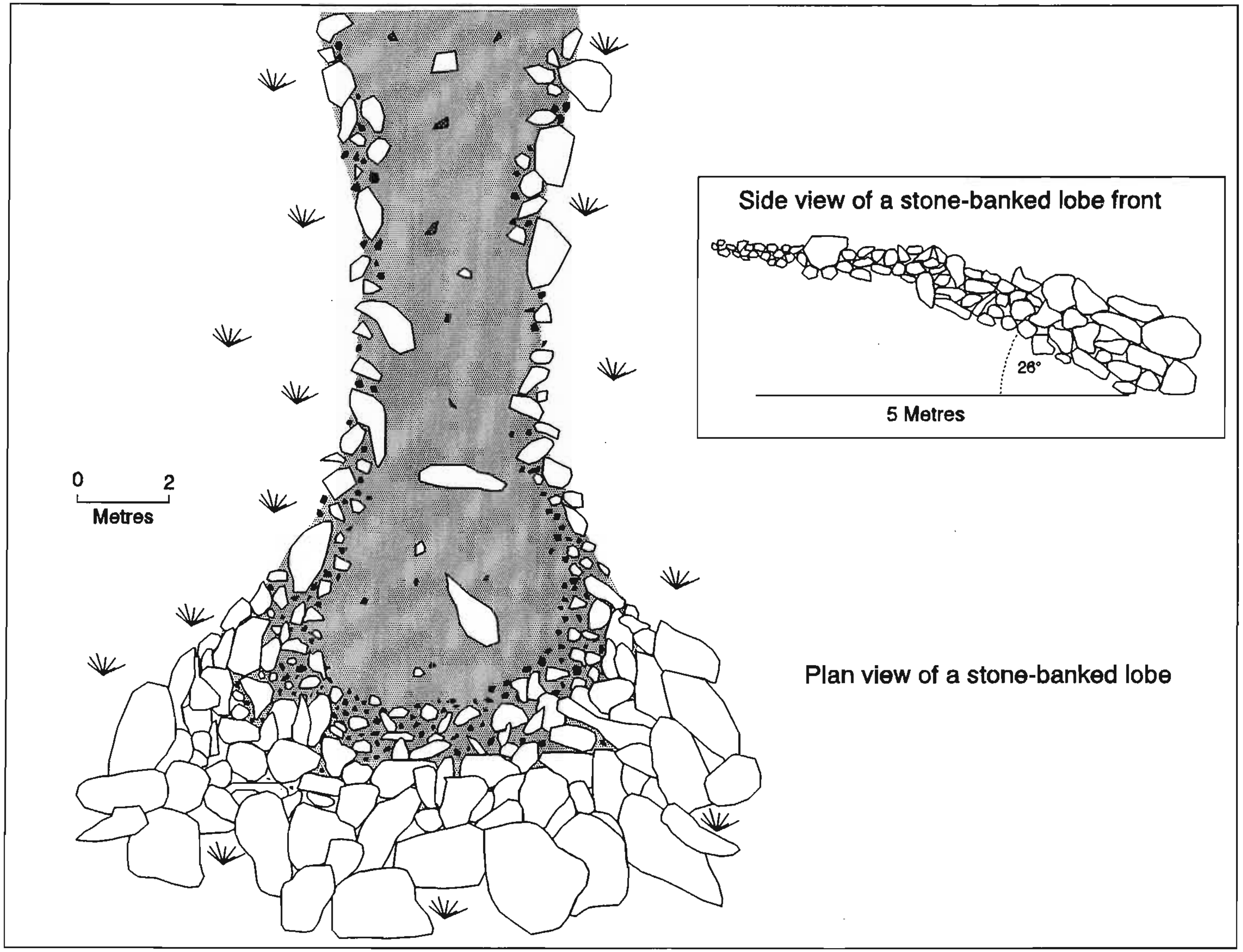


Figure 6.11 Sketch of a crescent-shaped stone-banked lobe found at Njesuthi Summit.

Figure 6.12 The front of a large stone-banked lobe containing much blocky material, Njesuthi Summit.

Figure 6.13 A variety *c* stone-banked lobe with a somewhat raised frontal bank.

6.3.1.2 Dimensions

Eleven stone-banked lobes (varieties *b* and *c*) were examined at Njesuthi Summit for which various dimensional data and their correlation values are given in Table 6.2. The stone-banked lobes have a mean tread length of 8.81 m and a mean frontal bank length of 2.37 m. The smallest lobe measured has a tread length of 3.42 m and a frontal bank length of 0.9 m, while the largest lobe has a tread length of 18.1 m and a frontal bank length of 5.01 m (Table 6.2). Total lobe lengths vary from 4.32 to 23.11 m. From Table 6.2, it can be seen that stone-banked lobes have mean tread lengths that vary considerably from one region to another. The lobes on Marion Island have a mean tread length of only 0.93 m (Hall, 1981), while those measured by R.B. King (1972) in the Cairngorms are said to be between 9 and 180 m (Table 6.1). The stone-banked lobes measured at Giant's Castle are 0.9 to 30 m in length (Boelhouwers, 1991a, 1994) (Table 6.1), and therefore, compare favourably with those found at Njesuthi. A positive correlation, significant at the 0.05 level ($r_s = 0.71$, $t = 3.01$), is found between the tread length and frontal bank length for the Njesuthi Summit lobes (Table 6.2).

Several stone-banked lobes on Njesuthi Summit have turf-banks extending downslope from the frontal stone-banks (Figure 6.14). The frontal stone-banks merge into the turf-banks, suggesting that the flow mechanisms associated with the stone-banked lobes may have pushed up the turf to produce turf-banks. Turf has also sometimes been pushed up against the frontal stone-bank sides (Figure 6.14), possibly owing to the advance of the lobe. The frontal turf-banks measured, extend 3.26 to 7.0 m downslope from the stone-banked fronts (Table 6.2).

Frontal bank widths (across-slope dimension of the frontal bank) at Njesuthi Summit vary from 2.31 to 10.06 m (mean = 4.35 m) (Table 6.2) and compare favourably with those measured by Boelhouwers (1994) at Giant's Castle, where they are 2 to 7 m wide (Table 6.1). A positive correlation, significant at the 0.05 level, is found between frontal bank width and tread length ($r_s = 0.88$, $t = 5.5$), and similarly between frontal bank width and frontal bank length (downslope dimension of the frontal bank) ($r_s = 0.83$, $t = 4.47$) (Table 6.2).

Lobe	Tread	Frontal bank					
	Length (m)	Width (m)	Length (m)	Incl. turf - banked front	Gradient	Height (m)	Incl. turf - banked front
1	5.35	2.32	1.60		30°	0.80	
2	9.58	5.3	2.93		31°	1.51	
3	3.42	2.9	0.90	3.26	23°	0.35	1.27
4	12.5	5.2	2.09	5.95	22°	0.78	2.29
5	11.7	4.82	2.09	7.00	21°	0.75	2.51
6	7.34	2.85	1.73		30°	0.86	
7	5.36	2.31	1.70		32°	0.90	
8	4.82	2.7	2.33		24°	0.95	
9	14.47	6.55	3.72		18°	1.15	
10	4.28	2.82	2.00		26°	0.88	
11	18.1	10.06	5.01		26°	2.20	
Mean	**8.81**	**4.35**	**2.37**		**25.7°**	**1.01**	
Std. Dev.	**4.59**	**2.27**	**1.09**		**4.37**	**0.46**	

Correlation values for stone-banked lobe measurements

Correlation	Correlation value	t - test value	Acceptance at 0.05 significance level	
Tread length / Frontal bank length	0.71	3.01	X	X = accepted
Tread length / Frontal bank height	0.49	1.68	-	- = rejected
Tread length / Frontal bank width	0.88	5.50	X	
Frontal bank width / Frontal bank length	0.83	4.47	X	
Frontal bank width / Frontal bank height	0.66	2.62	X	

Table 6.2 Characteristics of stone-banked lobes at Njesuthi Summit.

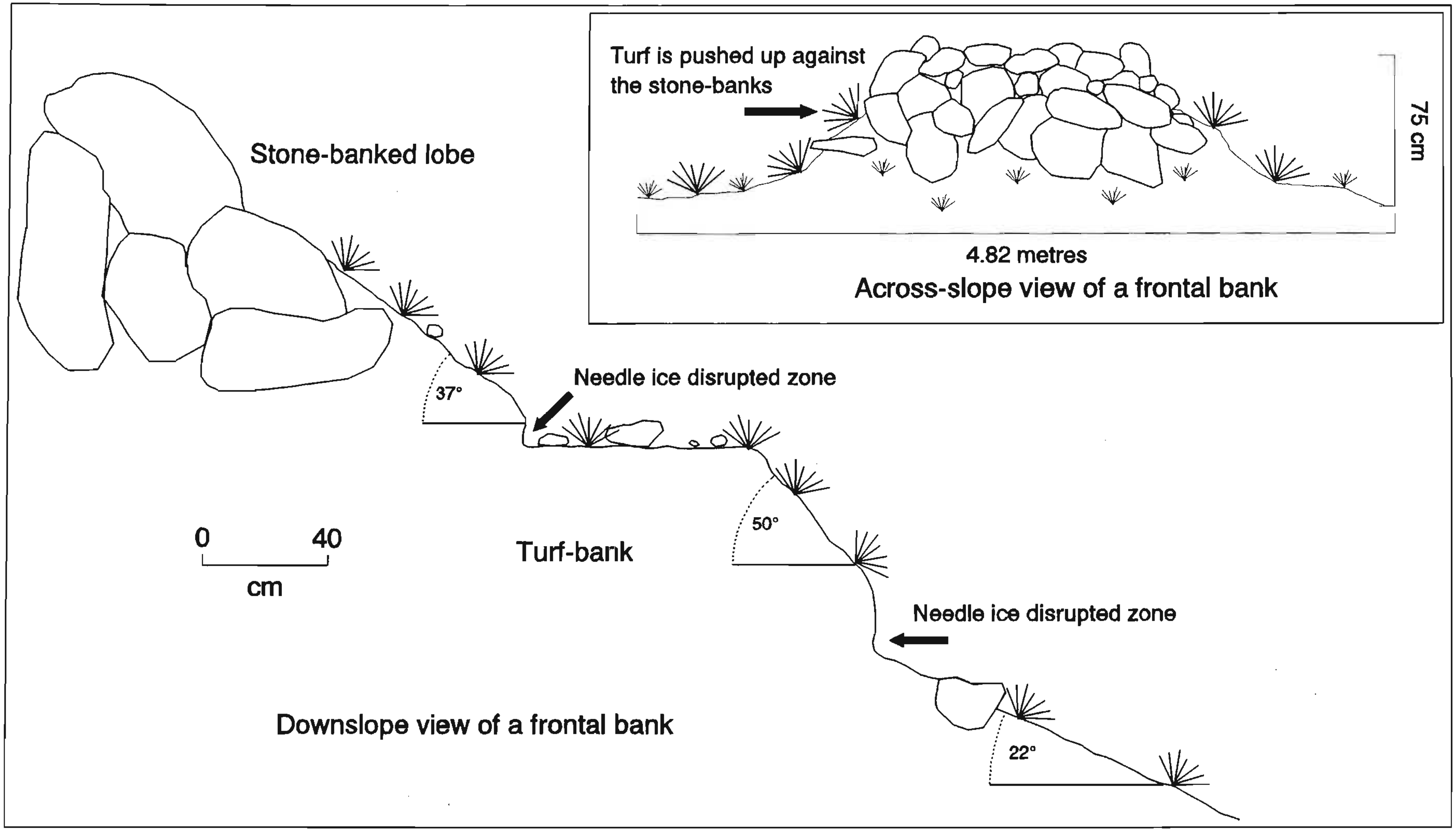


Figure 6.14 Sketch of frontal stone-banks merging into turf-banks

A lower, less positive correlation is found between frontal bank width and frontal bank height ($r_s = 0.66$, $t = 2.62$) (Table 6.2). The frontal bank heights vary from 0.35 to 2.2 m (mean = 1.01 m), and again, compare favourably with the Giant's Castle lobes which are 0.3 to 3 m high (Boelhouwers, 1994) (Table 6.2). Hall (1981) found a positive correlation between the tread length and frontal bank height for stone-banked lobes on Marion Island. Hall (1981) explains that this may be as a result of the tread slope increasing at a different rate to that of the field slope. However, results from Njesuthi Summit show no correlation between frontal bank height and tread length ($r_s = 0.49$, $t = 1.68$) (Table 6.2). As the Njesuthi lobes were measured on a uniform slope of about 15°, the argument presented by Hall (1981) cannot be applied here. Comparisons of lateral lobe dimensions show strong positive correlations, significant at the 0.05 level, whereas comparisons with vertical dimensions show little or no correlation (Table 6.2). This may be owing to the varied build-up of larger clasts at the lobe fronts (accumulation zones), which ultimately determines frontal bank height. The frontal bank heights of stone-banked lobes at Njesuthi Summit appear to be a function of the rapidity and quantity of debris accumulation, rather than tread dimensions. The occurrence of frontal banks raised above the adjoining treads (Figure 6.13) may attest to this.

6.3.1.3 Fabrics and sorting

Few workers have examined the fabrics and sorting characteristics for stone-banked lobes. Although Boelhouwers (1994) has investigated some fabric and sorting structures for stone-banked lobes at Giant's Castle, much more data are required to improve the understanding of stone-banked lobes; some such data were collected in this study.

Figure 6.15 shows clast size variations across two stone-banked lobes at Njesuthi Summit. Both lobes show larger clasts occupying the sides while centres have somewhat finer material. Although this is clearly evident in lobe **II**, lobe **I** shows only slight clast size variation across it (Figure 6.15). It is possible that after a considerable time of lobe inactivity, the clast-size variations become less pronounced owing to *in situ* rock break-up and removal/displacement of material. The high standard deviations of clast size in lobe **II** may be as a result of the combined effects of differential sorting and slope wash processes. Figure 6.16 indicates a

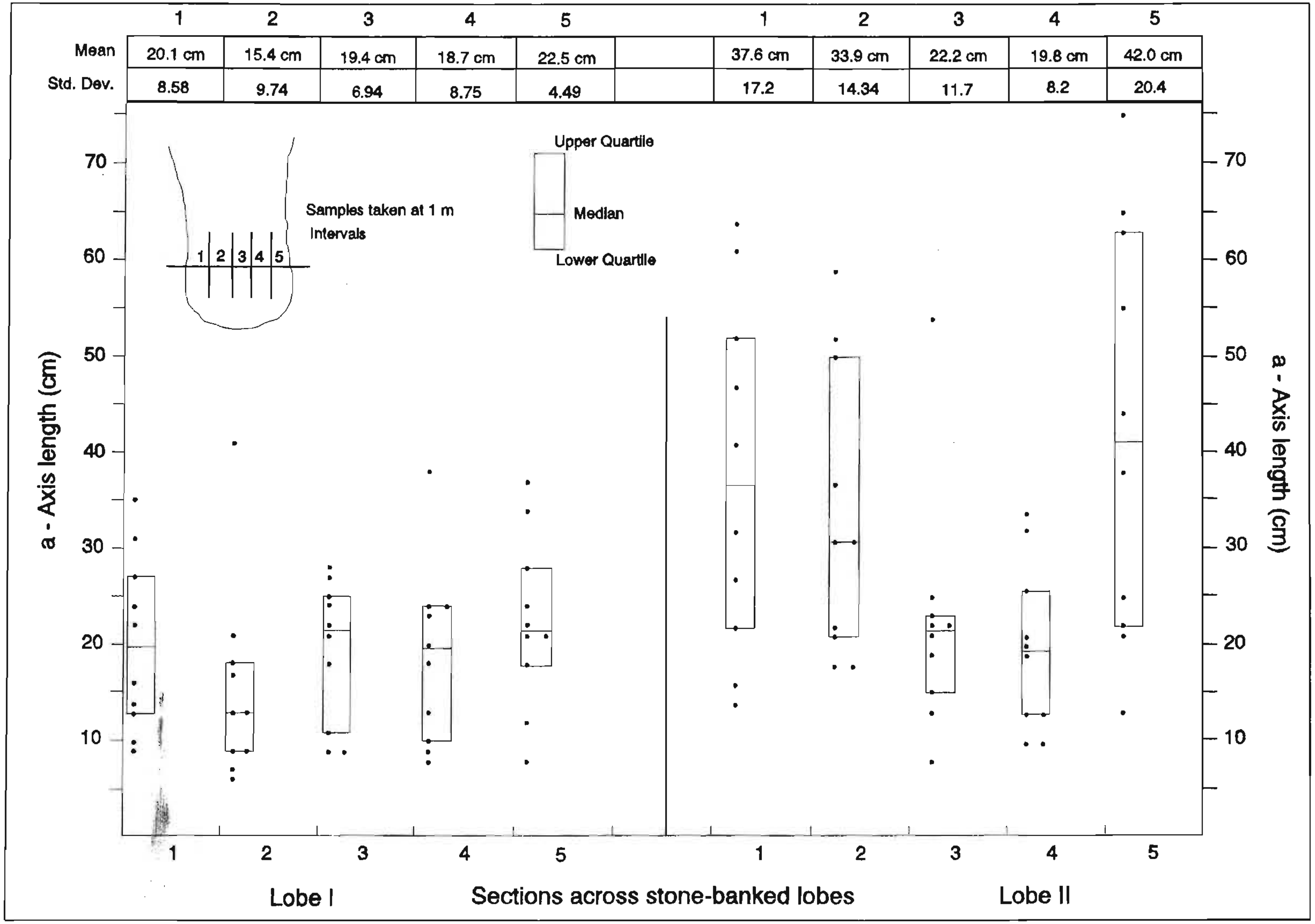


Figure 6.15 Clast size variations across two stone-banked lobes at Njesuthi Summit.

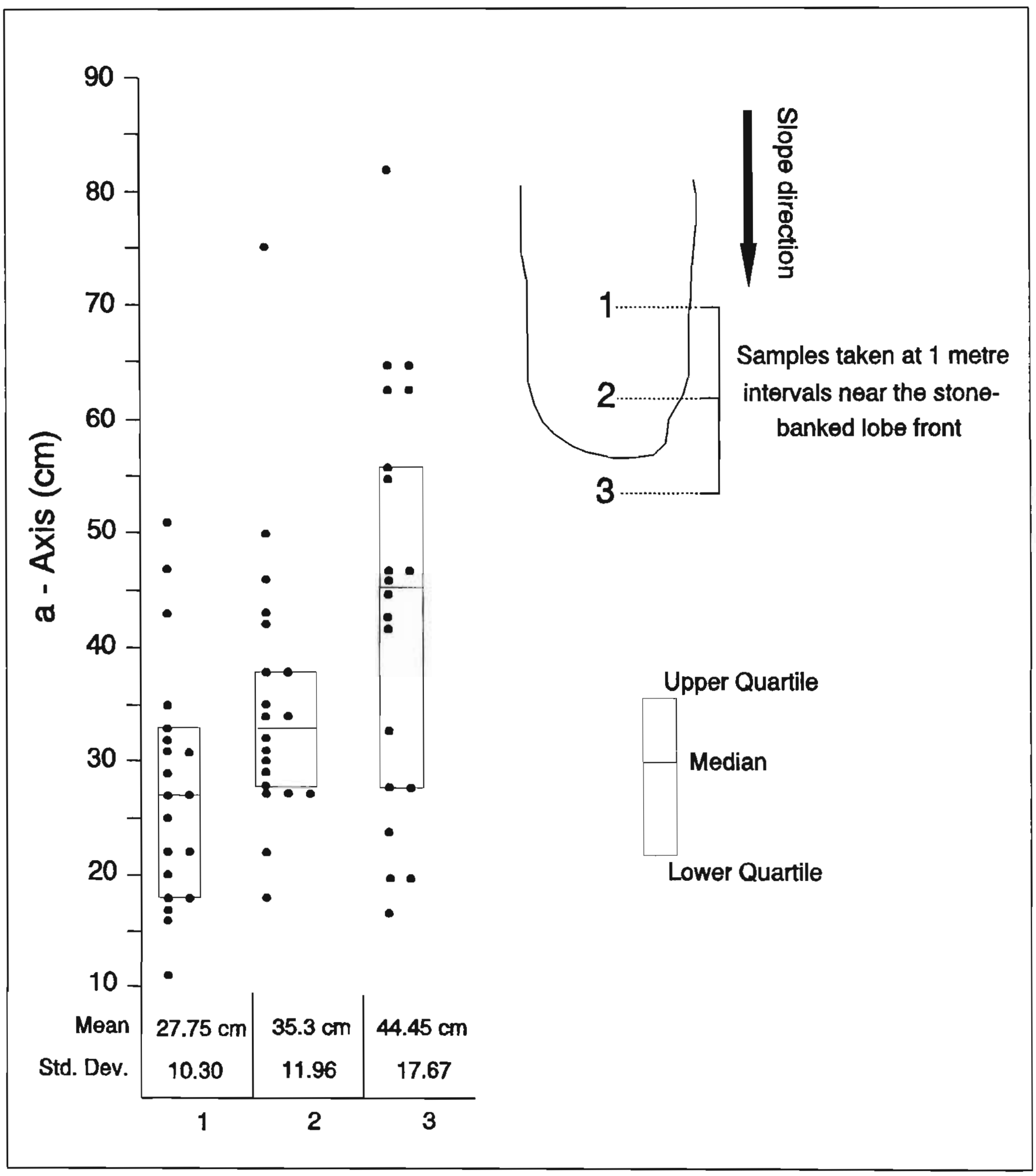


Figure 6.16 Downslope clast size variations through a stone-banked lobe front, Njesuthi Summit.

coarsening towards the frontal banks, an observation also made by Boelhouwers (1994). The clast sizes may rapidly diminish towards the tread where only gravel is found in variety *b* lobes. The lobe treads also show a zone of coarser material at their peripheries, while in the centre are mainly gravels and isolated clasts (Figures 6.10 and 6.11).

A vertical section through the front of a stone-banked lobe at Njesuthi Summit clearly shows a downward fining of material (Figure 6.17). The material is sorted into distinct zones, as has also been shown by Boelhouwers (1994). The first zone is an open block network, usually 20 to 30 cm thick, which ends abruptly to be replaced by angular, platy cobbles and gravel. The cobble/gravel zone consists of two units, the first (about 11 cm deep) contains mostly cobbles and coarse gravel with little matrix material, while the second (about 16 cm deep) comprises a greater percentage of fine gravel. The cobble/gravel zone then rapidly becomes more matrix supported, to be replaced by a compacted loam (Figure 6.17). This finding is similar to that made by Boelhouwers (1994), and would suggest consistent vertical sorting characteristics for such lobes in this high Drakensberg region. Vertical sorting for a small stone-banked lobe at Popple Peak shows a shallow clastic surface underlain by a gravely diamict (Figure 6.18). The surface is covered by angular, platy clasts which appear to be moving over the finer material below.

Clast orientation at the open block surface is primarily in the direction of flow (i.e. downslope), as is shown in Figure 6.19. Chi-squared tests and Figure 6.19 show a preferred orientation (significant at the 0.05 level) in the flow direction for open blocks at the tread. At the frontal banks, however, some clasts become re-orientated transverse to the lobe (χ^2 = 15.1). Similarly, Lundqvist (1949) found clasts in a stone-banked lobe front with alignment both transverse and parallel to the lobe front. At Njesuthi Summit, most clasts at the periphery of frontal banks are strongly aligned across-slope and only weakly downslope, so as to be orientated towards the frontal bank centre. Figure 6.19 also shows the increased imbrication from the lobe tread towards the frontal bank. Clasts in the tread usually dip parallel to the slope or in the upslope direction. Towards the frontal bank, however, clasts are seldom found dipping parallel to the slope, but rather show strong imbrication with a dip into the slope (Figure 6.19). The fabrics for smaller clasts at 34 cm depth show no orientation at the lobe

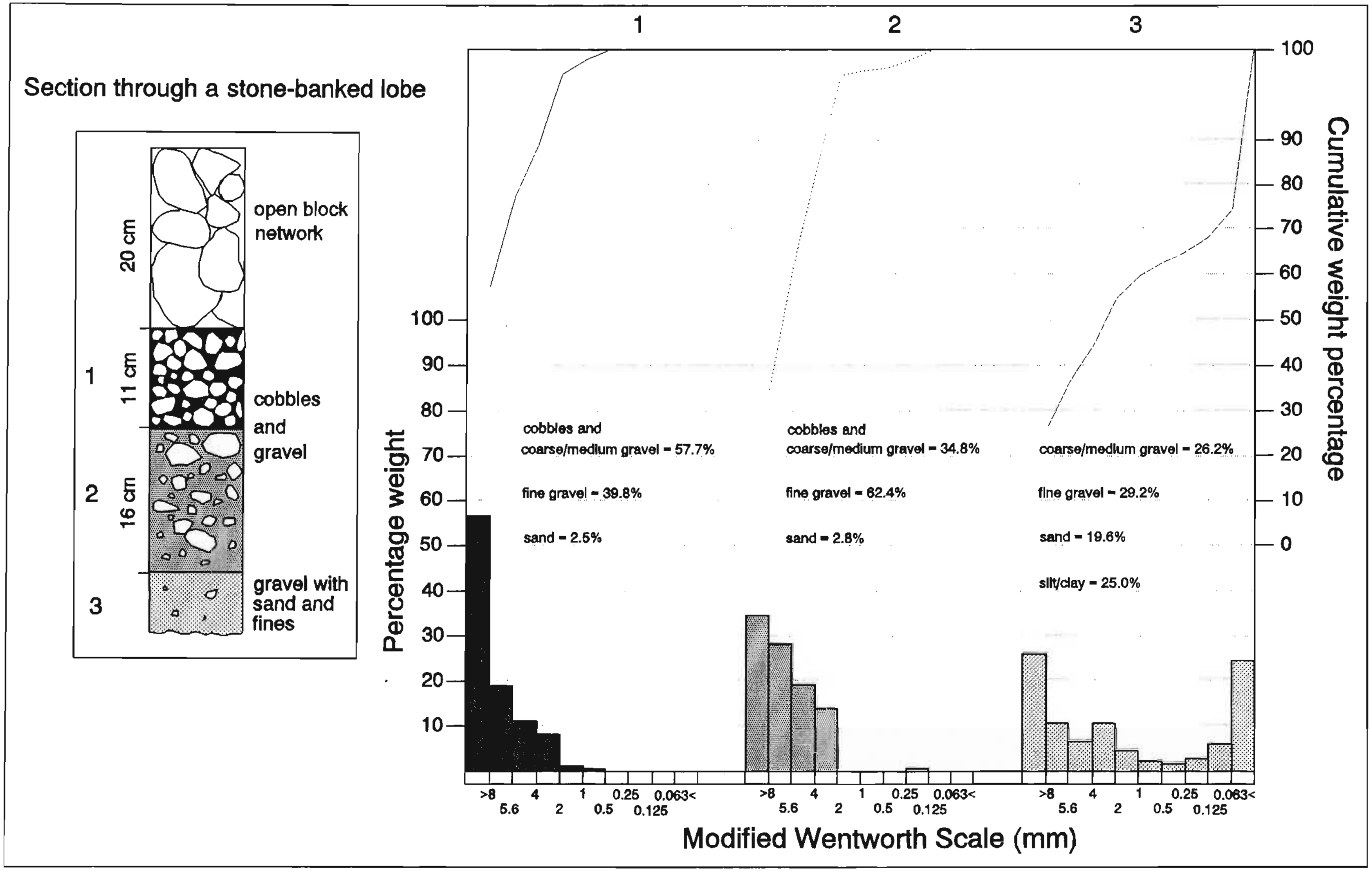


Figure 6.17 Vertical sorting and particle size distribution through a stone-banked lobe at Njesuthi Summit.

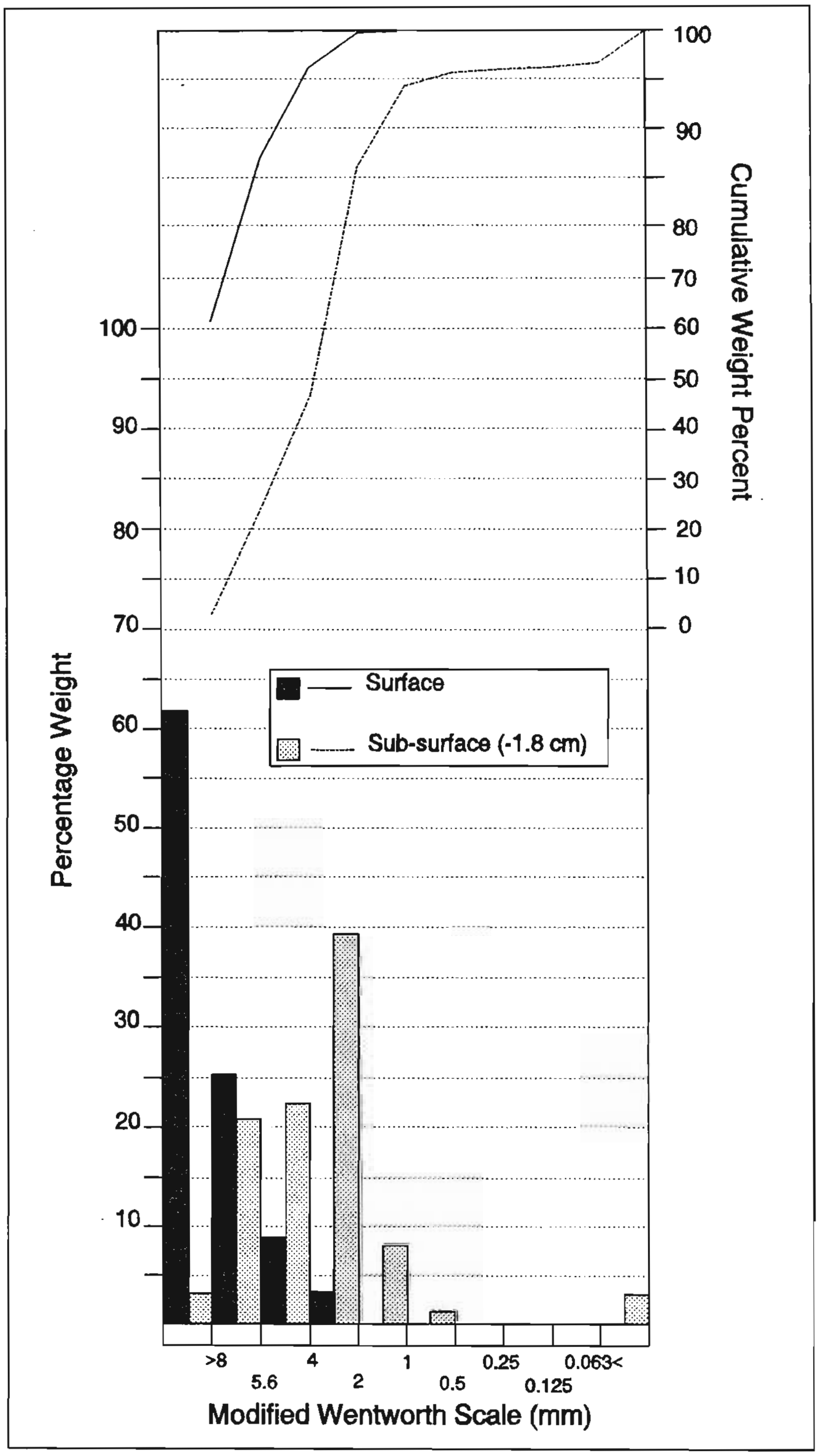


Figure 6.18 Particle size distribution through a stone-banked lobe at Popple Peak.

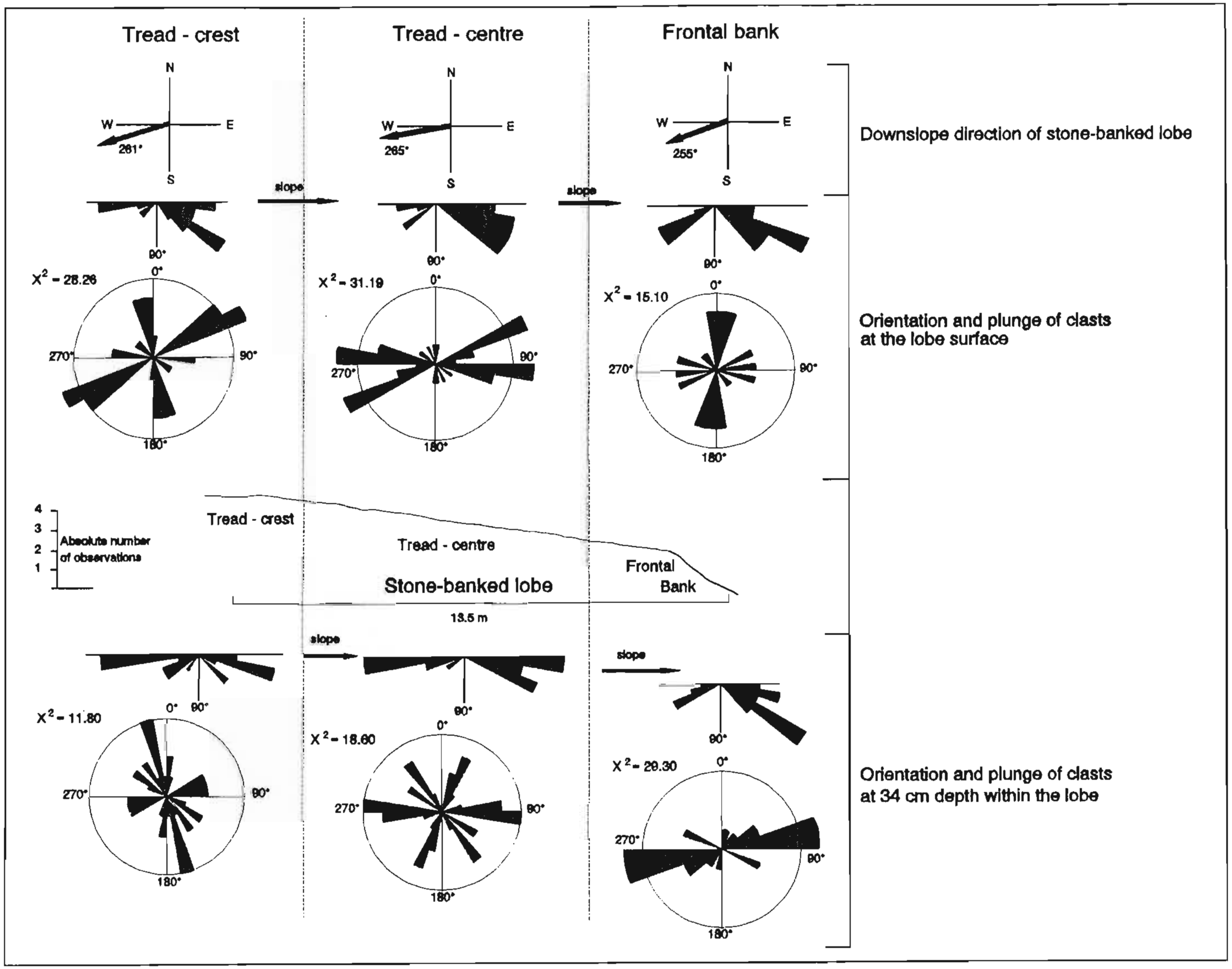


Figure 6.19 Orientation and plunge of clasts at various positions within a stone-banked lobe.

tread (χ^2 = 11.8 for the tread crest and 18.6 for the tread centre), but a preferred downslope orientation at the frontal bank (Figure 6.19). Subsurface clasts at the tread dip primarily parallel to the slope, with no visible imbrication. At the frontal bank, however, clasts dip strongly into the slope, thus showing some imbrication.

6.3.2 Discussion

6.3.2.1 Possible Processes and origin

Both gelifluction and frost creep are thought to be the primary mechanisms responsible for stone-banked lobe development (Van Steijn *et al.*, 1995), particularly where a debris mass becomes saturated owing to drainage being impeded by permafrost or seasonally frozen ground at depth (C. Harris, 1981). Several workers (Sharp, 1942; Jahn, 1958; Washburn, 1979; C. Harris, 1981) believe that the larger stones move downslope more rapidly than fine material, which gives rise to a growing accumulation of coarse material, ultimately forming the stone-banks. Such a process could be envisaged for the stone-banked lobes near Njesuthi Summit. The Njesuthi lobes show characteristics such as lobate fronts and imbricated blocks orientated in the flow direction. Clasts at the tread surface show stronger orientation in the flow direction and greater imbrication, than clasts at depth. These characteristics could support a solifluction origin, where movement rates are said to decrease with depth (C. Harris, 1981; Van Steijn *et al.*, 1995). The absence of large clasts in the middle of lobes may be indicative of rapid movement near the lobe centres (Price, 1974), while the accumulation of stones at the periphery of lobe treads could be the result of slower movement there. Lobes with raised frontal banks may be an indication that deceleration was very abrupt, which consequently limited lateral movement to form the typical crescent shape. It is postulated that the lobes formed during a period of extended seasonal ground freezing, possibly associated with greater winter and/or early spring precipitation than today. Rapid snowmelt may also be a source of moisture, causing saturation of such debris (C. Harris, 1981).

Several stone-banked lobes on Popple Peak may have developed under somewhat different controlling mechanisms and environmental conditions. Because the debris mantle is shallow here, bedrock may act as an impermeable layer, therefore permafrost or seasonal ground freezing may not be required to cause saturation. Owing to the small clast size and steep gradients, rain-wash cannot be ruled out as a possible mechanism for stone movements at this site. However, frost creep and/or solifluction are almost certain processes, as clasts are sorted, with larger stones concentrated at lobe fronts and peripheries (Figures 6.10 and 6.11). It is also likely that gravity is a primary control for such lobe development, possibly much more so than those found on the gentler Mafadi slopes.

The downslope advance of stone-banked lobes has given rise to turf-banks and turf-banked terraces immediately in front, and to a lesser extent, on the sides of frontal banks (Figure 6.14). The turf-banks merge into the stone-banked lobes and are raised above the surrounding vegetated areas. The vegetated turf-banks may be up to 7 m long, consisting mostly of soil, small stones and occasional large blocks which protrude through the turf banks. At greater depth, more blocks belonging to the frontal stone-banks may be found. It is suggested that some stone-banked lobes have had a "ploughing" effect, similar to that of ploughing blocks, whereby cryostatic and/or hydrostatic pressure at the lobe fronts may cause turf to be pushed up against the sides of the stone-banks. Ploughing blocks are commonly known to cause similar mound development on the downslope and across-slope sides of moving blocks (Tufnell, 1972; Chattopadhyay, 1983; Reid and Nesje, 1988; Gorbunov, 1991; Wilson, 1993a).

6.3.2.2 Possible implications for south-facing, high altitude slope processes

According to Benedict (1976), stone-banked lobes occur less frequently than turf-banked lobes, owing to their specialised requirements. The presence of stone-banked lobes and stone-banked terraces on some higher south-facing Drakensberg summits would suggest that several of these specialised requirements have been met. As there is no scarp or slope above the lobes, debris was produced *in situ*. It is suggested that the lithology at such sites is conducive to rock breakup, and was exploited by extensive weathering processes. The occurence of stone-banked lobes rather than extensive solifluction sheets, suggests selective or limited movement rather than an overall downslope flow of regolith (Sugden, 1971). It is possible that lithology and/or vegetation cover and/or snow cover may have exercised control on the selective distribution of solifluction (and gelifluction) on such summit slopes. It appears, however, that vegetation cover has been sparse on such slopes and possibly underwent rapid change and/or destruction during the onset of more severe frost periods. It may, therefore, be postulated that such slopes have been subject to considerable frost action, rock breakup, debris accumulation and solifluction/gelifluction processes. These processes have consequently resulted in a sparsely vegetated south-facing summit zone with a distinct convexo-concave slope profile. C. Harris, (1981) has supported the argument that such summit convexo-concave slope profiles in periglacial environments are associated with some of the processes outlined above.

6.3.2.3 Present and past activity

The smaller lobes (variety *a*) on Popple Peak appear fresh and active. Many clasts, particularly those in the centre of stone-banked lobes, show imbrication and sorting, and further, do not show the weathered discolouration displayed by bedrock and material surrounding the lobes (Figures 6.8 and 6.9). In addition, most of the clasts forming the stone-banked lobes have few or no lichen growing on them, whilst the rocks immediately adjacent to the stone-banked lobes are extensively lichen and moss covered.

Although the considerably larger lobes (varieties *b* and *c*) appear "fresh" in morphology (i.e. are well defined and not undergoing degradation), their current activity is thought to be sporadic. Some smaller variety *b* lobes may be marginally active where both lichen and vegetation cover is sparse, however, the larger lobes all appear to be inactive. Most of the lobes show rocks at frontal banks firmly imbedded in the turf-banks or vegetated fronts. Both rhizomatous and woody plants grow within the stone-banked lobes and frontal turf-lobes. Although woody plants may tolerate slight surficial disturbance, they are said to be prone to strong frost action (Jonasson, 1986). Both this, and the absence of fines on many stone-banked lobe treads, could be indicative of current lobe inactivity.

The exact age of such lobes, as with most other relict landforms discussed in this dissertation, is difficult to determine. The data and discussion presented would, nevertheless, indicate that a variety of stone-banked lobes of variable size have been active during Holocene times. Under the present climatic regime, only smaller stone-banked lobes at higher altitudes (possibly above 3200 m) appear active. The relative freshness, but apparent inactivity, of larger lobes could suggest that they may have been active during Late Holocene cold periods. This would be in accordance with findings from Marion Island (Hall, 1981) and the European Alps (Veit and Höfner, 1993). Veit and Höfner (1993) have suggested increased phases of gelifluction activity during the Late Holocene around 3000 yrs BP and 1250 yrs BP in the European Alps. If such a phase also occurred in the Southern Hemisphere during this time, it is likely that the larger stone-banked lobes at higher elevations in the Drakensberg were active then. The stone-banked lobes are in close proximity to relict sorted circles (refer to Section **4.5**) and these two periglacial landform types may have been operative during the same time and shown comparative phases of activity, although this is difficult to prove and so is somewhat speculative.

6.4 DEBRIS DEPOSITS ON THE HIGH DRAKENSBERG PLATEAU: THE NJESUTHI DEBRIS DEPOSIT

Debris deposits are frequently encountered at the base of south-facing slopes, primarily at altitudes above 3100 m. Many of the deposits may be primary or secondary products of periglacial slope processes, particularly as they occur at higher elevations and on south-facing aspects. Such south-facing slopes are considerably steeper than north-facing slopes (Meiklejohn, 1992, 1994), and thus are more conducive to slope instability and mass-wasting processes than north-facing slopes. Further, such high altitude slopes are frequently littered with poorly-consolidated weathered detritus (blockslopes and blockstreams), which is the debris source for the deposits found at slope basal zones. Despite their potential value for evaluating landscape evolution in the region, such debris deposits have, as yet, not been reported from the high Drakensberg plateau.

Regular solifluction and high magnitude/low frequency slope processes are considered to be important to the total mass movement occurring in periglacial environments (Egginton and French, 1985). It is envisaged that such processes have been significant in the high

Drakensberg during the past. The varying location, size and morphology of the deposits are possibly the result of different formative processes. It is difficult, however, to evaluate most of these relict deposits as they are now usually completely vegetation covered.

6.4.1 The Njesuthi debris deposit

6.4.1.1 Characteristics

The Njesuthi debris deposit is located at the base of the Njesuthi southwest face (200° to 220°), between 3090 and 3160 m a.s.l. Although the slope gradient averages about 15° at the depositional site, this increases rapidly to 27°, upslope of the deposit. The slope above the deposit reaches an altitude of 3410 m and is extensively covered by weathered material. The upper reaches of the slope have stone-banked and turf-banked lobes, below which are blockslopes and blockstreams extending down to 3190 m a.s.l. Between 3190 and 3250 m a.s.l. is a pronounced zone of blockstreams and boulder-strewn slopes with gradients between 25° and 30°. The blockstreams are between 8.6 and 24.07 m wide and up to 120 m in length. Between the blockstreams are smooth bedrock outcrops 10.85 to 35.31 m wide. Below are several pairs of flow levees consisting of large angular blocks, but these are partly covered by grasses. Extending downslope from the levees are three large lobate debris deposits (Figure 6.20). The fronts of these deposits have been incised by fluvial action.

One of the debris lobes was surveyed and a sketch map and profiles produced (Figures 6.21 and 6.22). The lobe is about 230 m long and 135 m wide near its front. The lobe has a mean surface gradient of about 16°, but this increases to 27° at its front where it has been incised (Figures 6.21 and 6.22). The lobe has three ridges extending from the upper reach towards the lobate front (Figures 6.21 and 6.22). Near the front, the debris has spread out laterally, producing a relatively flat surface microtopography; here only two ridges are visible on the down-valley side of the deposit (Figures 6.21 and 6.22). The deposit rises up to 12 m above the surrounding topography.

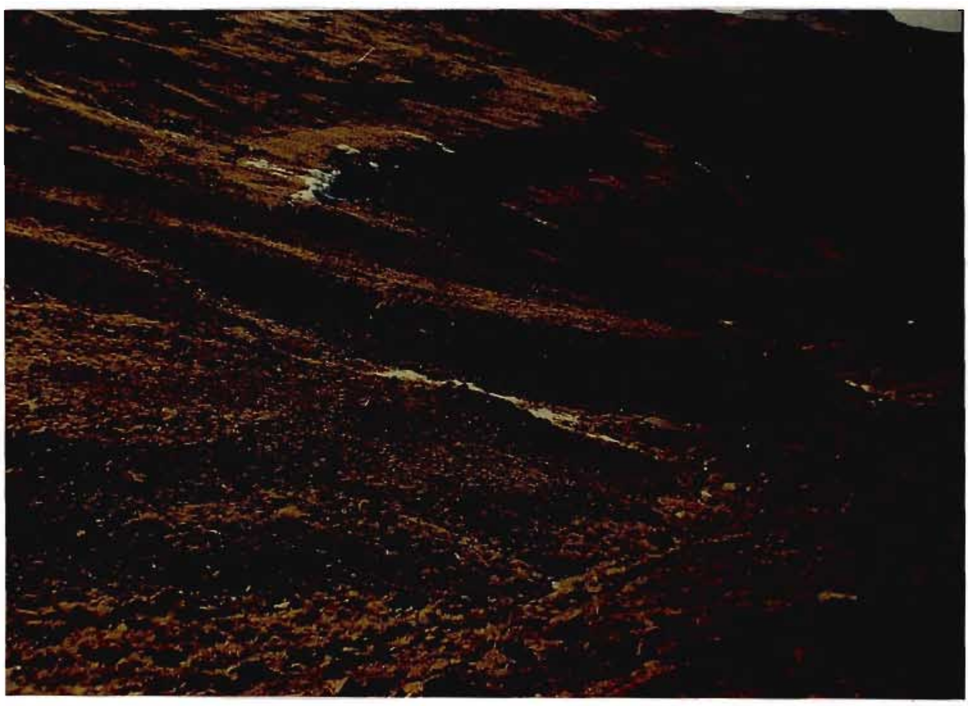


Figure 6.20 Three large debris deposits on the Njesuthi south-facing slopes.

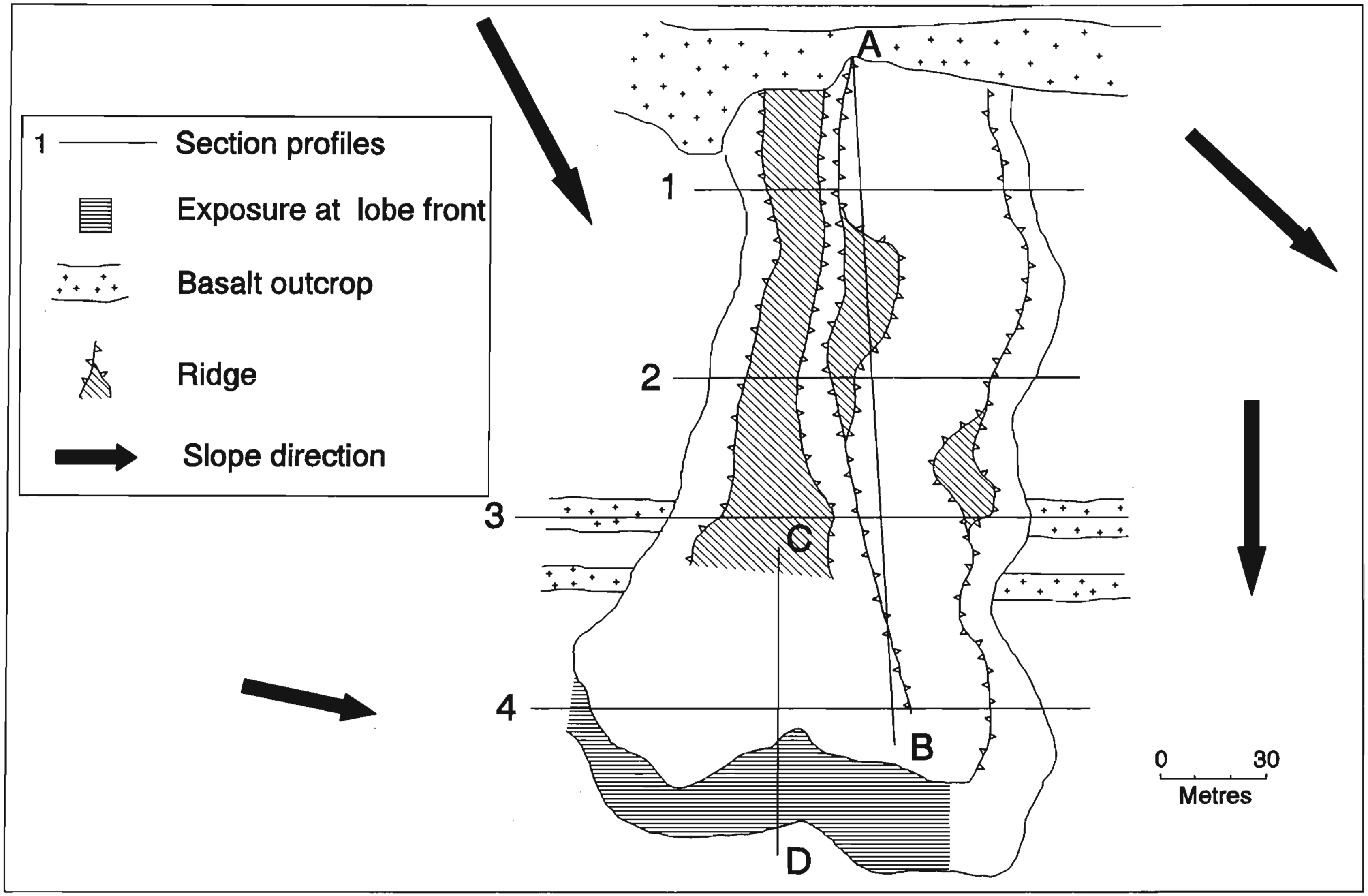
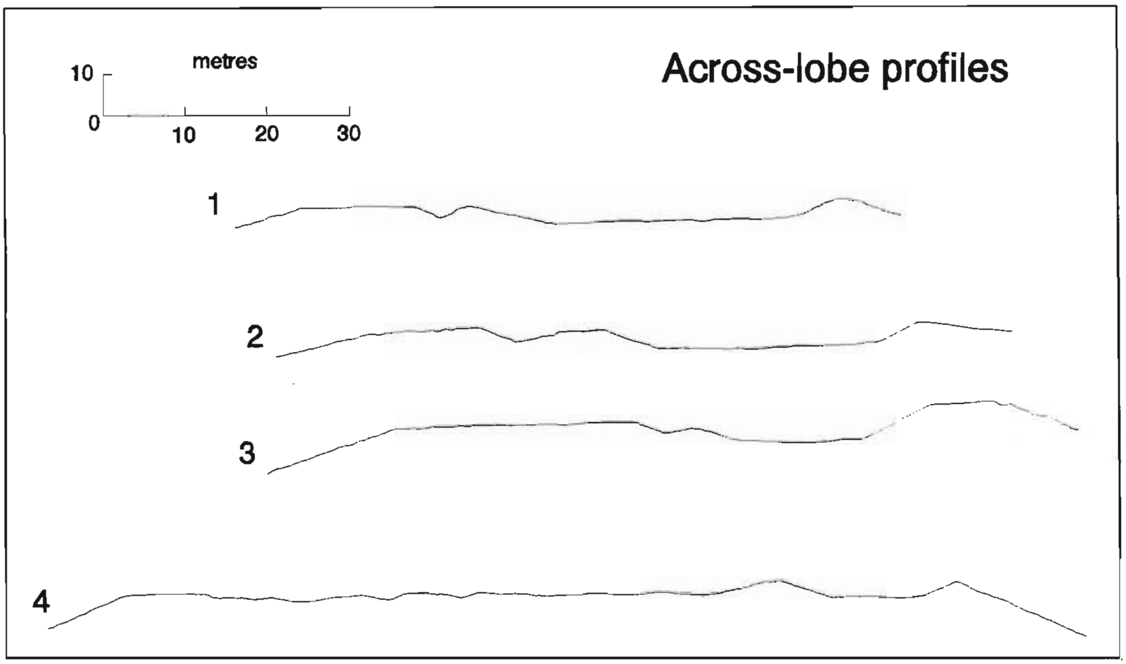


Figure 6.21 Sketch map of a debris lobe on the Njesuthi south face. Letters and numbers refer to the profile sections shown in Figure 6.22.



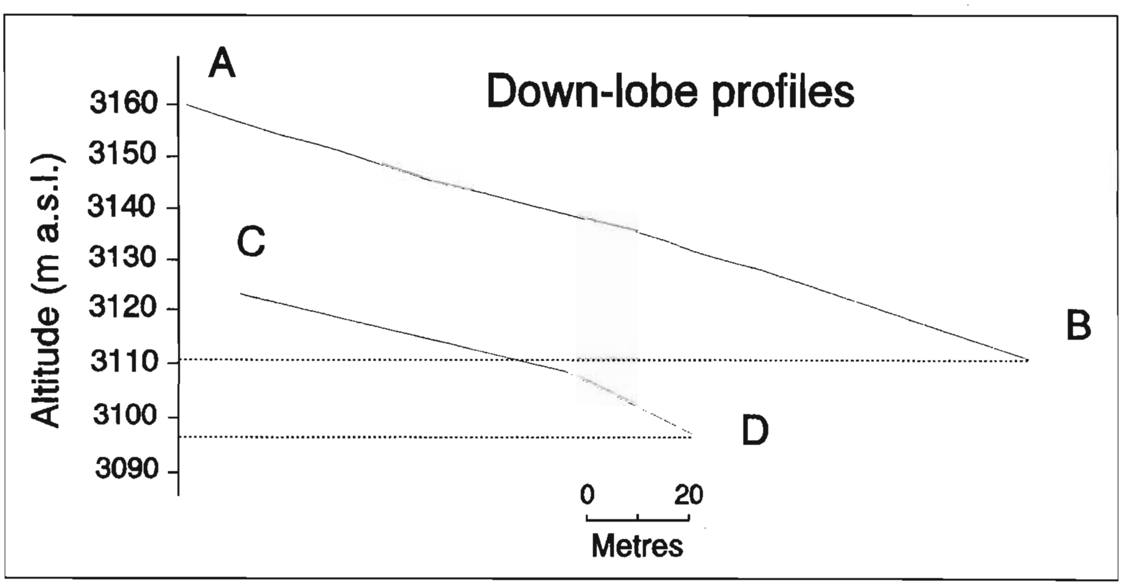


Figure 6.22 Across-lobe and down-lobe profiles for the Njesuthi debris lobe (please also refer to Figure 6.21).

Sedimentological characteristics are difficult to assess because of the extensive vegetation cover on the surface. However, the debris deposit has been undercut at its front, exposing some sedimentological features. The deposit consists of rocks with interstitial sand and fine sediments. The upper 2 m are primarily boulder dominated, but with increasing depth, fewer boulders and more fine material is encountered. The clasts are very angular (mean roundness index = 0.13) and relatively flat (mean flatness index = 199.58) (Figure 6.23). The mean clast size for the debris deposit was compared against that for a blockstream found on the slope above it (Figure 6.23). The blockstream shows distinct lateral sorting with a mean a-axis clast size of 32.6 cm in the central part and 69.0 cm on the periphery of the block mass (Figure 6.23). The mean a-axis clast size for the blockstream is 50.8 cm and compares very favourably with that for the debris deposit below, where the mean a-axis clast size is 51.3 cm (Figure 6.23).

Although clast a-axis orientation and dip were assessed, the results are specific to only one site, and therefore conclusions cannot be made for the deposit as a whole. The result shows no orientation (χ^2 = 15.44, H_o is accepted, significant at the 0.05 level) (Figure 6.24). There is considerable scatter in clast dip values, but most dip between 20° and 60°, thus exceeding the local slope angle of the debris terminus (Figure 6.24).

6.4.1.2 Possible origin of debris accumulation

The possible reasons for instability and debris deposition are difficult to assess, but several hypotheses are able to account for the debris accumulation.

1. *Solifluction*

According to an earlier discussion (see Section **6.3.2.1**), solifluction has been operative on the higher slopes above the deposits. Although Harris (1977) has argued for landsliding rather than solifluction on slopes steeper than 15° or 16°, Rapp (1960) has reported solifluction lobes on slopes of 15° to 25° in northern Sweden. It may thus be possible that solifluction processes have operated to lower elevations on the Njesuthi south-facing slopes, despite slope gradients of up to 27°. The blockstreams display lateral sorting with larger blocks on the sides, which

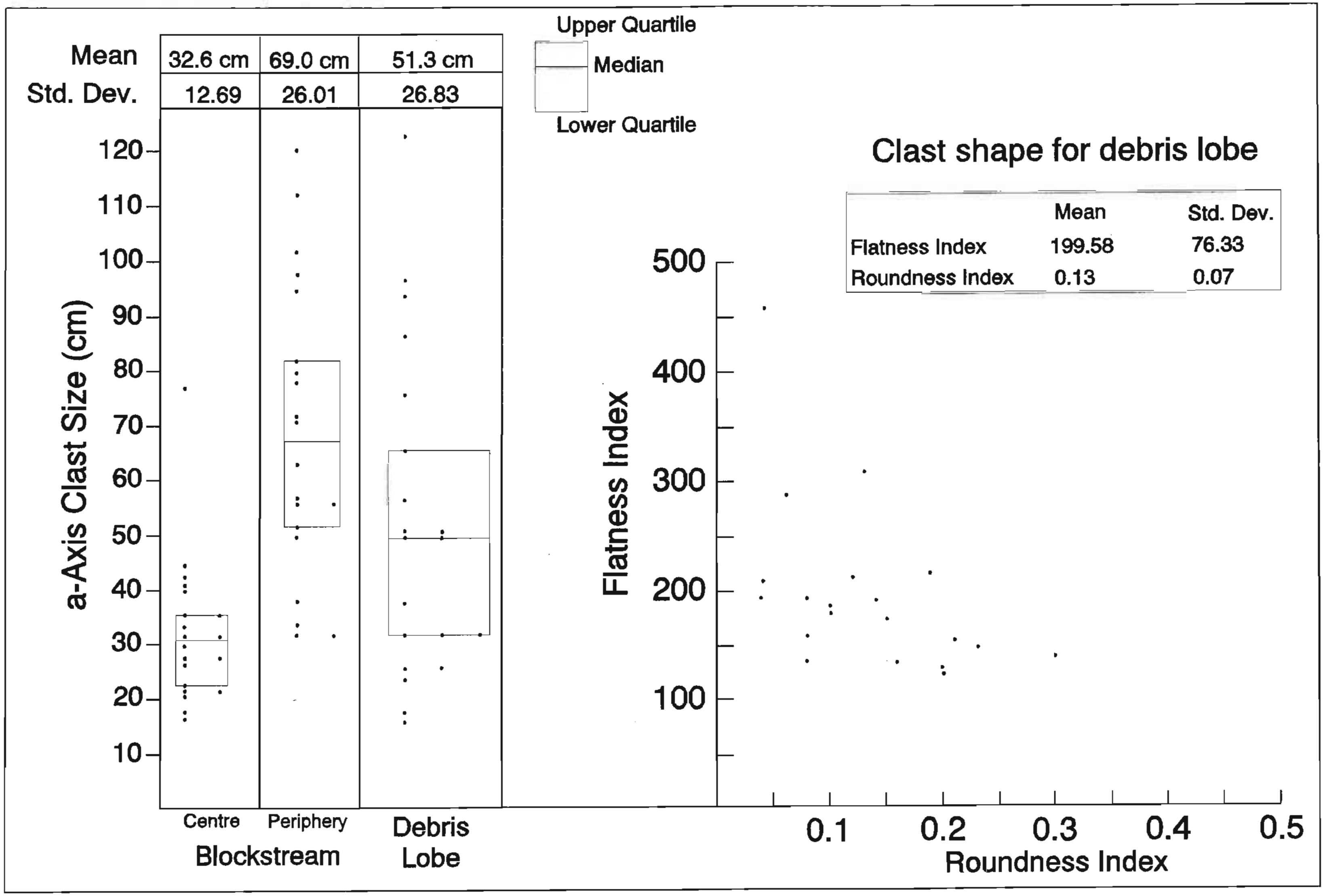


Figure 6.23 Clast shapes for the Njesuthi debris lobe and comparison of clast size between the debris lobe and a blockstream above the lobe.

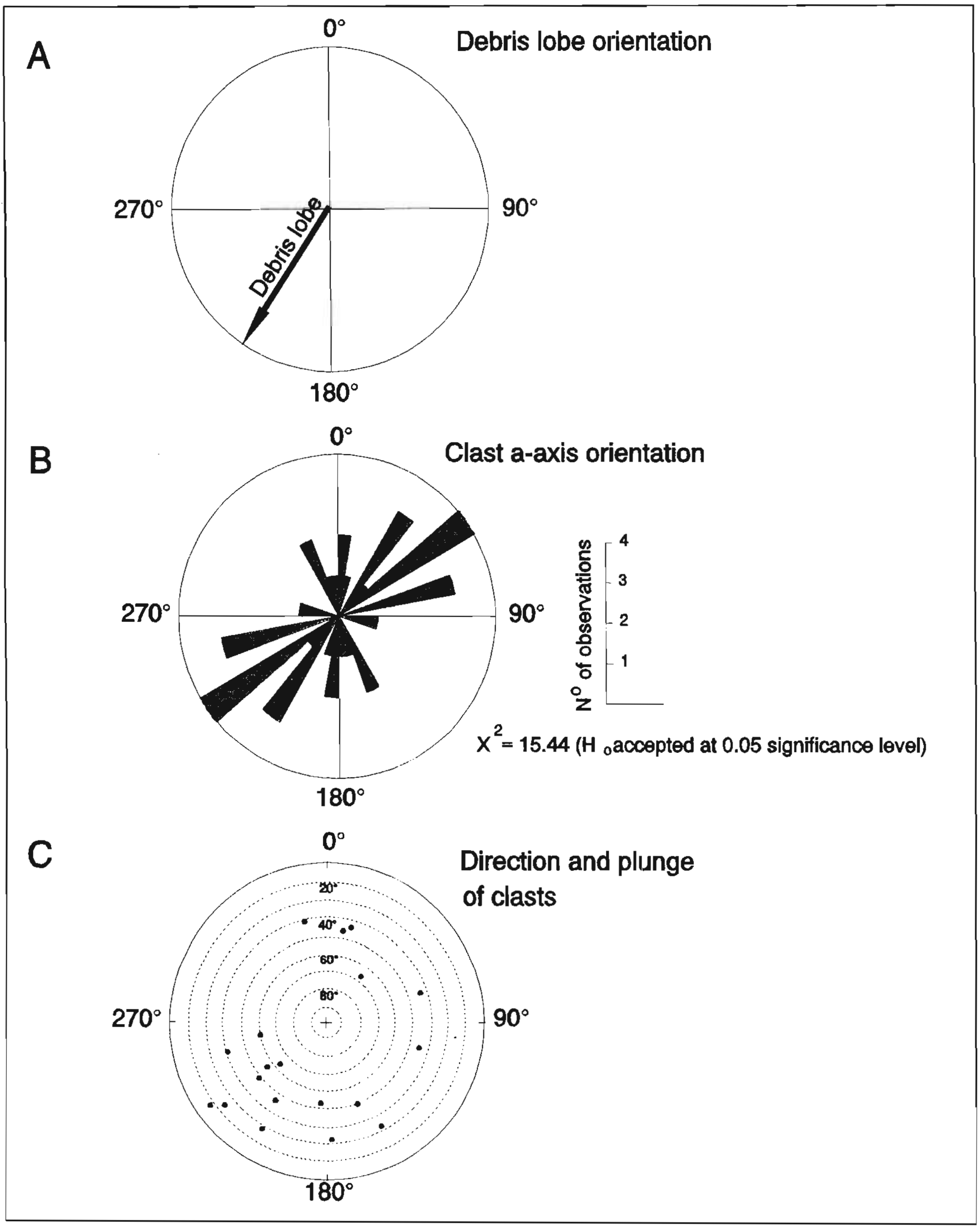


Figure 6.24 Clast orientation and dip at the Njesuthi debris lobe front.

may be ascribed to solifluction processes and dispersive pressure (Nieuwenhuijzen and Van Steijn, 1990), where movement on the sides is slower than in the central part of the debris mass. As the regolith is shallow on this slope, sediments may become saturated during heavy and prolonged rainfall events and/or rapid snowmelt and/or seasonal thaw events, thereby increasing the possibility of solifluction.

The exposed section at the Njesuthi debris lobe does not show characteristics such as shear planes, inverted stratigraphy or buried soil/organic horizons, which are typical of many solifluction lobes (Worsley and Harris, 1974; Matthews *et al.*, 1986; Smith, 1987a; Hirakawa, 1989). While solifluction appears to have been operative on the slopes above the deposit, there is no morphological or sedimentological evidence to suggest that the deposit is a solifluction lobe.

2. *Protalus rampart(s)*

The typical morphology of a protalus rampart is that of a single ridge or ramp crest (Harris, 1986; Ballantyne and Kirkbridge, 1986; Ballantyne, 1987b; Wilson, 1993b) with a shallow backing depression, indicating the former presence of a snowpatch (Wilson, 1993b). Other attributes may include symmetrical ridges (Harris, 1986) and arcuate ridge forms (Ballantyne and Kirkbridge, 1986). As the morphological characteristics of the Njesuthi debris deposits show no such attributes, they are not protalus ramparts.

3. *Moraines*

Moraines have yet to be found on the high Drakensberg plateau. Other glacial evidence, such as glacial striations, bullet-nosed clasts and roches moutonnées, are also absent from the region. The morphological attributes of the Njesuthi deposit and the absence of glacial evidence, does not support a glacial origin for such debris masses.

4. *Rockglacier development*

Rockglaciers are "lobate or tongue shaped bodies of frozen debris with interstitial ice and ice lenses, which move downslope or downvalley by deformation of the ice contained within it" (Barsch, 1988, p72), and thus are features of cohesive flow (Barsch, 1992). The Njesuthi deposit displays several characteristics which are typical of a rockglacier. These include ridges, lobate fronts, surfaces dominated by blocks and a downward fining of material with depth (Barsch, 1977, 1988). Although transverse ridges and depressions are ascribed to the flow structures associated with deformation of interstitial ice (Haeberli, 1985; Barsch, 1988), such surface relief is not necessarily displayed by all rockglaciers (Barsch, 1988). There is sufficient debris supply material on the slopes above, which is a requirement for rockglacier development (Whalley and Martin, 1992). Although the present author is unable to disprove rockglacier development at the site, more conclusive data are required before the presence of relict rockglaciers can be ascertained with certainty. On any account, the debris accumulation at the site cannot be explained in terms of rockglacier development, and thus an alternative hypothesis should be considered.

5. *Debris flows and avalanches*

A debris flow is "rapid mass movement of blocky, mixed debris of rock and soil by flow of wet, lobate mass" (Rapp and Nyberg, 1981, p183). The erosional part of such debris flows usually show a slide scar or gully at slope gradients of 25° to 40° (Rapp and Nyberg, 1981). The Njesuthi debris deposit shows several characteristics typical of a debris flow. According to Johnson (1984), debris flows are locally controlled and particular sites may be prone to repeated debris flow activity. This may explain the occurrence of three separate deposits alongside each other on the Njesuthi south face. The deposit also features : (1) a lobate mass, (2) a large accumulation of angular boulders near the surface and, (3) an increasing volume of fines at depth. These characteristics are all typical for debris flow deposits (Suwa and Okuda, 1983; Johnson, 1984). Further, the ridges and levees immediately upslope of such deposits are said to be indicative of a flow process (Lewin and Warburton, 1994). The levees are frequently found on steeper slopes and the lobes on somewhat gentler slopes below (André, 1990; Nieuwenhuijzen and Van Steijn, 1990). As all these attributes are present at the study site, there is strong indication that these deposits are the product of debris flows.

The former occurrence of debris avalanches and slides on the slopes above is also possible, given the steep gradients and buildup of talus at various sites along the slope profile. The coincidence in size of clasts in both the blockstream and the debris deposit below, may indicate that the steep blockstream site is a potential debris source. It is conceivable that avalanches were initiated on such steep slopes to subsequently become debris flows on the shallower slopes below. The debris deposits are not thought to be products of mudflows as these would require more fine-grained debris than is present in the debris flow (Rapp and Nyberg, 1981).

6.4.2 Discussion : a model for high-altitude, south-facing slope evolution

Although the modelling of slope evolution in periglacial regions appears to have received little attention (Lewkowicz, 1988), it has been argued that mass movement phenomena are important components in the evolution of such landscapes (Johnson, 1984). The Njesuthi south-facing slope profile contains a variety of mass movement features within a range of altitudinal zones which have also been identified on several other high altitude south-facing slopes. The observed findings may be able to offer some insight to the past slope processes that were operative on such slopes. Based on the findings already made, a model showing slope-evolutionary processes for some high-altitude, south-facing slopes in the Drakensberg has been produced (Figures 6.25 and 6.26). The model is intended to explain processes that have been operative in the recent past (Late Pleistocene and Holocene).

The summit zone predominantly comprises a blockslope resulting from intense weathering. The mass movement processes near the summit appear to have been dominated by solifluction and creep, producing stone-banked and turf-banked lobes. These relatively slow mass-wasting processes near the summit are within a convexo-concave slope profile, and have possibly helped maintain such a profile (Figures 6.25 and 6.26). The combination of weathering and slow mass movement enables considerable blockslope development down to 3300 m a.s.l. where the slope rapidly becomes steeper and somewhat more linear. The steeper gradient offers increased gravity effects and thus the blockslopes are replaced by blockstreams. Talus appears to have built up here and was possibly held locally by seasonal ice and a snow pack, which acted as constraining agents (c.f. Johnson, 1984). Because the slopes down to about 3200 m a.s.l. commonly have gradients between 20° and 30°, critical threshold values to initiate rapid, and often catastrophic mass movement, could have been reached on several occasions during the recent past. A variety of factors, such as snow melt, rapid thawing of

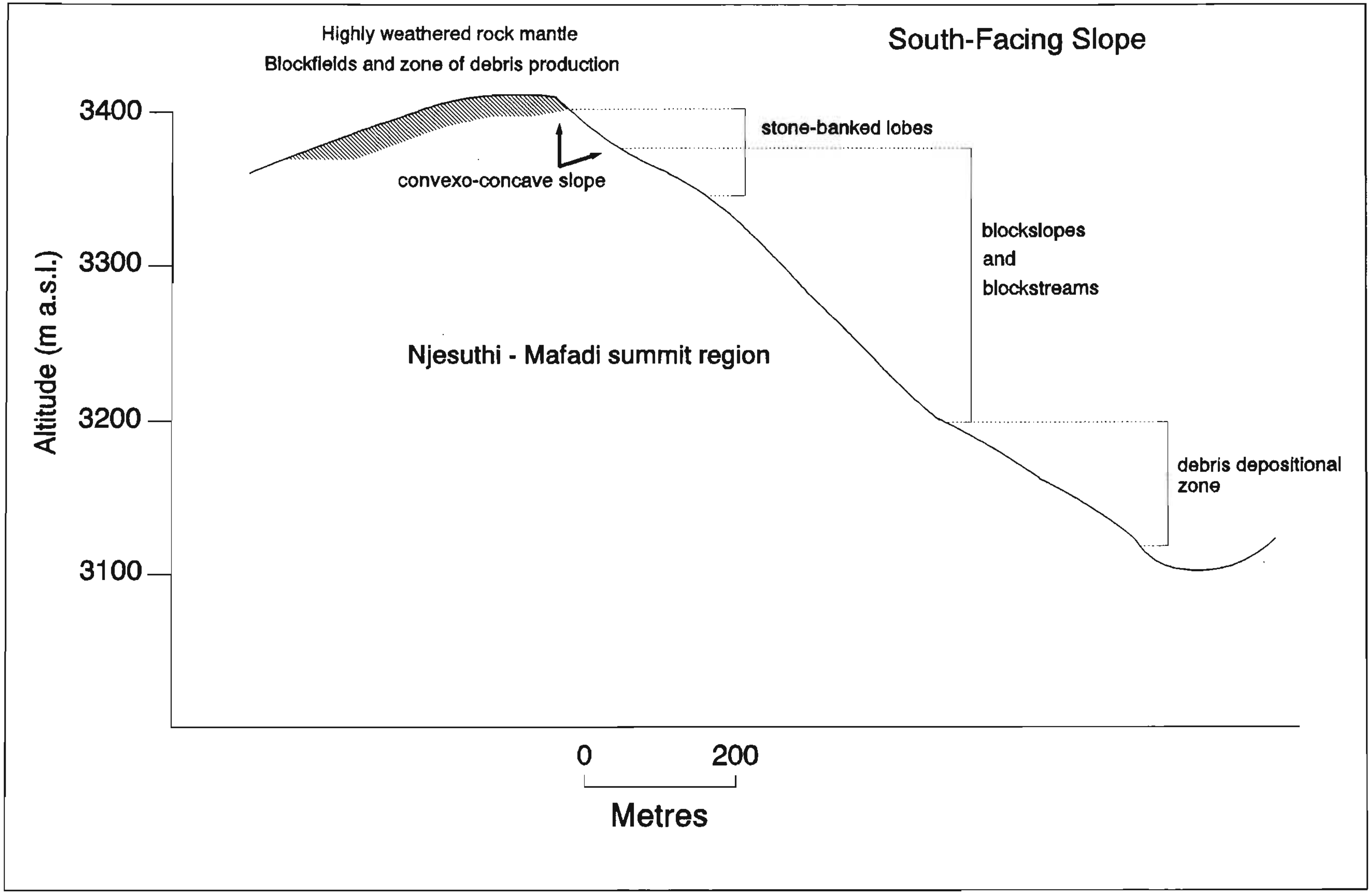


Figure 6.25 A profile through the Njesuthi south-facing slope which shows zones in which particular geomorphic landforms predominate.

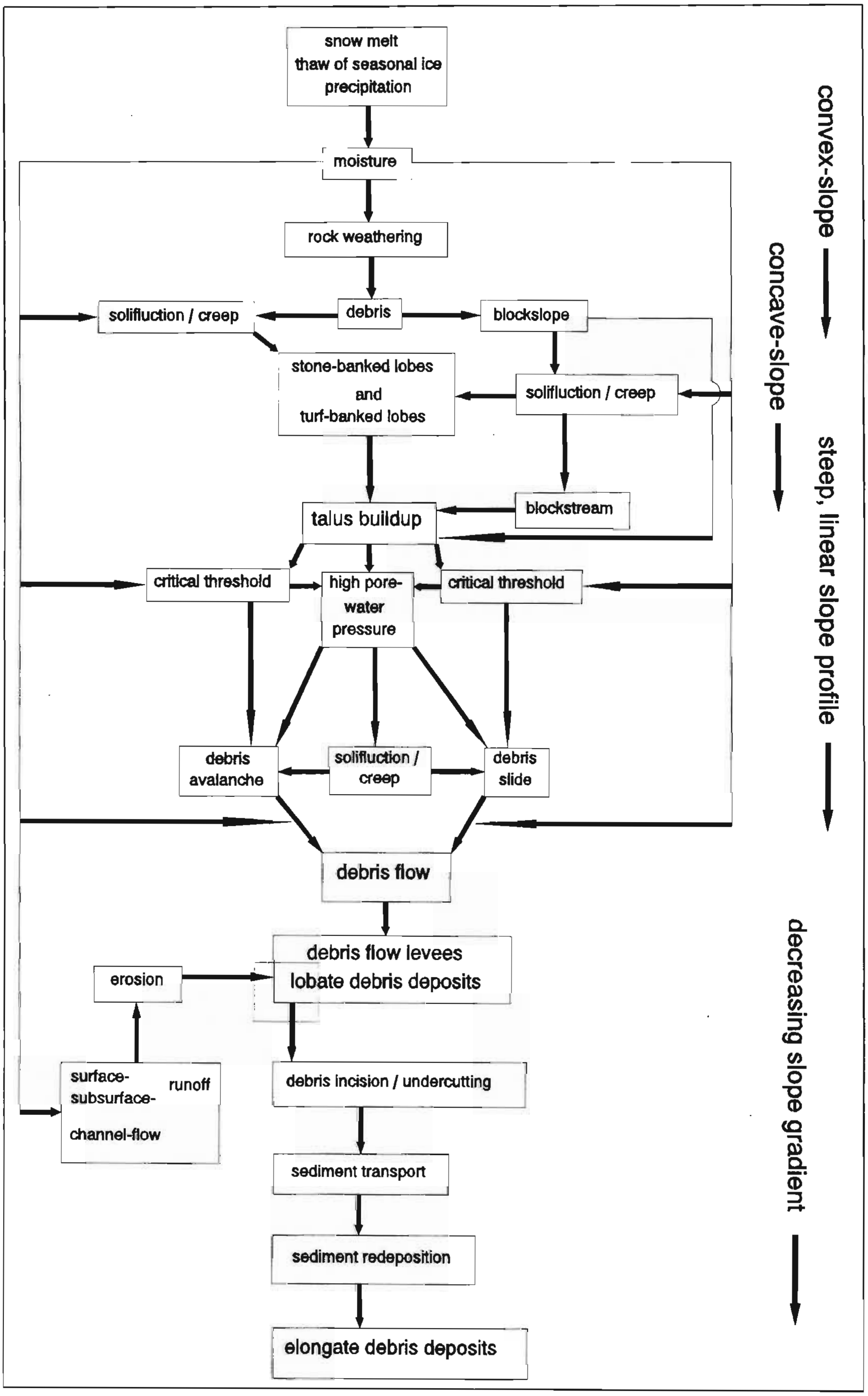


Figure 6.26 A model for high Drakensberg south-facing slope evolutionary processes.

frozen ground and intense and/or prolonged precipitation may have been trigger mechanisms for debris avalanching and debris slides. It is possible that during rapid climatic/seasonal change, several such factors operated synergistically. Many rock outcrops along this slope profile display very smooth surfaces, usually not found elsewhere. These smooth rock surfaces may offer further evidence for the occurrence of past debris avalanches and/or slides. Below 3200 m a.s.l. the slope gradients become increasingly shallow and are characterised by a depositional zone of flow levees and lobate debris deposits. Debris slides and avalanches on higher, steeper slopes may have transformed to debris flows on the lower gentler slopes, particularly where the debris mass became increasingly saturated. The lobate debris deposits have subsequently been incised and undercut by run-off and channel-flow. It is possible that during periods of rapid snow/ice melt and/or high rainfall events, debris incision was enhanced and the sediment redeposited downstream over a 1.5 km stretch, forming elongate debris deposits (Figure 6.26).

The objective here is to offer an empirical evaluation of the likely processes that have operated on the higher Drakensberg south-facing slopes during the recent past. From the landforms (e.g. blockslopes, blockstreams, stone-banked lobes) occurring on such slopes, and from the suggested processes (solifluction, creep, thawing of frozen ground, snowmelt) having operated there, it becomes apparent that periglacial mass-wasting has been a considerable component to the overall slope-evolution in the high Drakensberg.

6.5 DEBRIS DEPOSITS WITHIN CUTBACKS ALONG THE HIGH DRAKENSBERG ESCARPMENT : THE NHLANGENI DEBRIS DEPOSIT

Recently, Hall (1994a) has presented an hypothesis for the origin of cutbacks found along the Drakensberg escarpment. This hypothesis is based on the occurrence of large boulders and levees in the lower cutback zone and suggests that meltwater from a rapidly ablating ice body may be responsible. There has been no published account on the occurrence of macro-scale debris deposits near the summit zone of cutbacks. However, Grab (1996a) has so far identified three cutbacks which contain such deposits at or near their summits. Much of the discussion in this section is extracted from Grab (1996a).

Eleven cutbacks, all of which face an easterly direction, have been examined for possible debris deposits (Figure 6.27 and Table 6.3). Although several small debris deposits occur in cutbacks where the pass altitude is above 3000 m, more extensive deposits are restricted to those with an altitude of 3200 m or more (Table 6.3). The cutbacks containing deposits are bounded by some of the highest summits in the Drakensberg, and further, have large rock niches/hollows on their south-facing aspects. The slopes to the west of some cutbacks (e.g. Bannerman and Nhlangeni) contain blockstreams and "roche moutonnée" - shaped tors. The largest debris deposit found, occurs in the Nhlangeni cutback (29° 17' 48" E ; 29° 29' 15" S) (Figures 2.6 and 6.27 and Table 6.3), where a detailed field investigation was undertaken and upon which much of the hypothesis presented here is based.

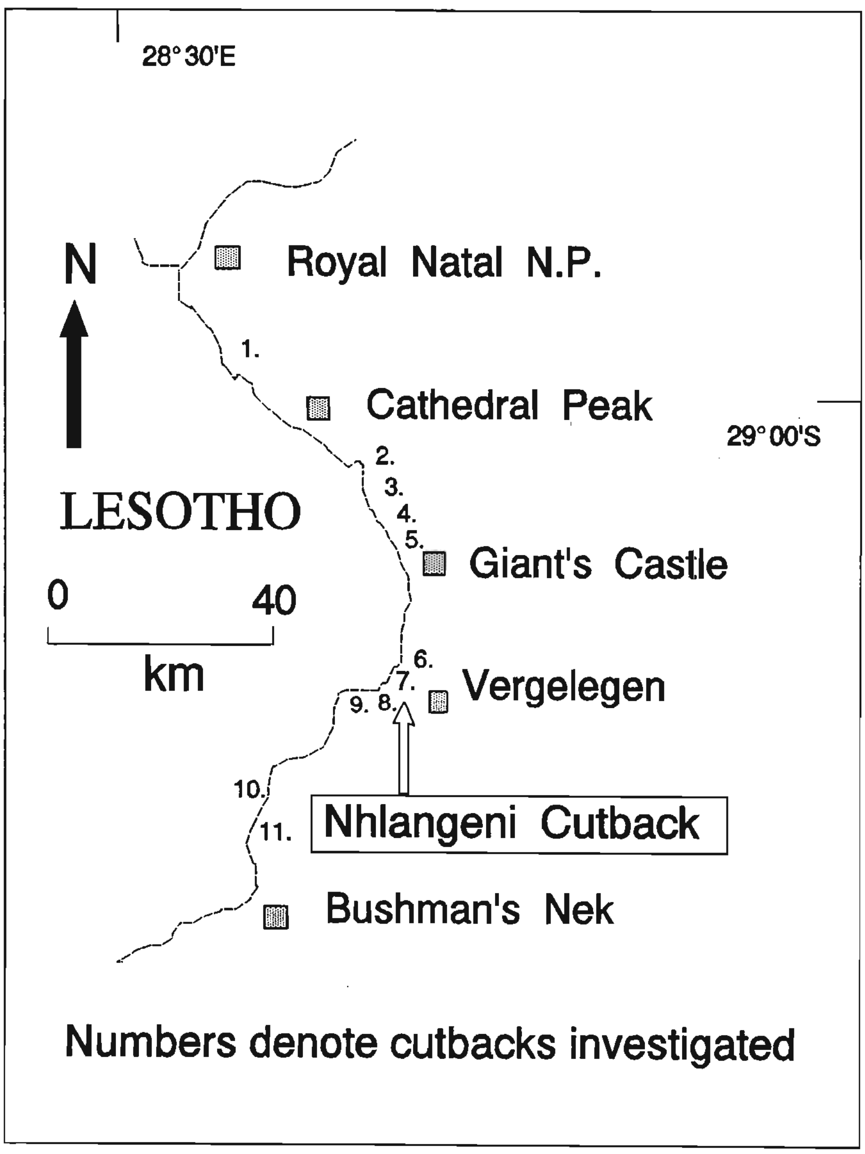


Figure 6.27 The high Drakensberg cutbacks investigated for possible debris deposits.

CUTBACK/PASS	LATITUDE	PASS ALTITUDE (m a.s.l.)	MAX. ALTITUDE OF CUTBACK SIDEWALLS (m a.s.l.)	BLOCKSTREAMS ON LEEWARD SIDE	EXTENSIVE DEBRIS RIDGES/DEPOSITS	LARGE ROCK NICHE/HOLLOW WITHIN CUTBACK SIDEWALLS
1. Mbundini	28 50' 48" S	3100	3220	no	no	no
2. Ship's Prow	29 05' 45" S	3300	3377	?	yes	yes (south facing)
3. Judge	29 13' 20" S	3140	3191	no	no	no
4. Bannerman	29 15' 10" S	3050	3295	yes	no	no
5. Ka-Langalibalele	29 16' 50" S	2950	2985	yes	no	no
6. Mkomazi	29 28' 30" S	2850	3068	no	no	no
7. Nhlangeni	29 29' 15" S	3200	3301	yes	yes	yes (south facing)
8. Kwa Ntuba	29 30' 00" S	3250	3355	yes	yes	yes (south facing)
9. Sani	29 35' 00" S	2850	2968	no	no	no
10. Mashai	29 42' 25" S	2900	3051	no	no	no
11. Mzimude	29 45' 40" S	3050	3306	no	no	no

Table 6.3 Characteristics of cutbacks investigated for possible debris deposits (also refer to Figure 6.27 for the locality of cutbacks 1 to 11).

6.5.1 Characteristics

The debris extends throughout the higher Nhlangeni cutback, where it has been incised to form distinct debris ridges (Figures 6.28 and 6.29). The deposit is most extensive and reaches its greatest depth on the northern side of the cutback where maximum incision has taken place, whereas the southern side is characterized by occasional bedrock outcrops. Ridge-crest profiles were taken along ridges B and C (Figures 6.28 and 6.30) and 12 cross-ridge profiles through ridges A-B-C-D (Figures 6.28 and 6.31). The mean ridge crest gradient for ridge B is 24° and that for ridge C is 29°. These ridge crests are steepest at their frontal zones where ridge B has a gradient of 34° and ridge C 39°. The ridges are several hundred metres in length, with ridge B attaining a length of 747,36 m and reaching a height of 18.40 m. Numerous cross-ridge profiles show progressive lowering of ridge-crest heights in the down-valley direction, this being particularly evident on profiles 6, 8, 9 and 10 (Figure 6.31).

Most of the ridge sides are presently stable and colonized by *Merx Meullera Distichia* and *Pentaschistis* grasses. The inner sides of ridges B and C are sparsely vegetated and show numerous mass wasting scars as well as the accumulation of larger boulders at the ridge bases, against which smaller material is banking. Fluvial features are absent and the drainage channels are characterized by lichen covered boulders, grasses and *Helichrisum* heath. Lichen size was assessed at the ridge crest and within the inter-ridge basal zone (drainage channels) (Table 6.4). Although *Xanthoparmelia* has a larger average diameter at the ridge crest (mean = 79.93 cm; max. = 190 cm) than within the drainage channels (mean = 46.54 cm; max. = 77 cm), white crustose lichen are larger within the drainage channels (mean = 111.14 cm; max. = 264 cm) than on the ridge crest (mean = 98.77 cm; max. = 193 cm) (Table 6.4). From this, it would appear that the rocks within the inter-ridge basal zone have remained *in situ* for a considerable time. It is apparent that the present fluvial dynamics have insufficient ability to cause significant debris incision. The slope to the west of the escarpment dips westward (Figure 6.32), resulting in a small catchment area with limited drainage coalescence in the cutback itself. This has restricted the ability to fluvially erode and transport the basaltic rock.

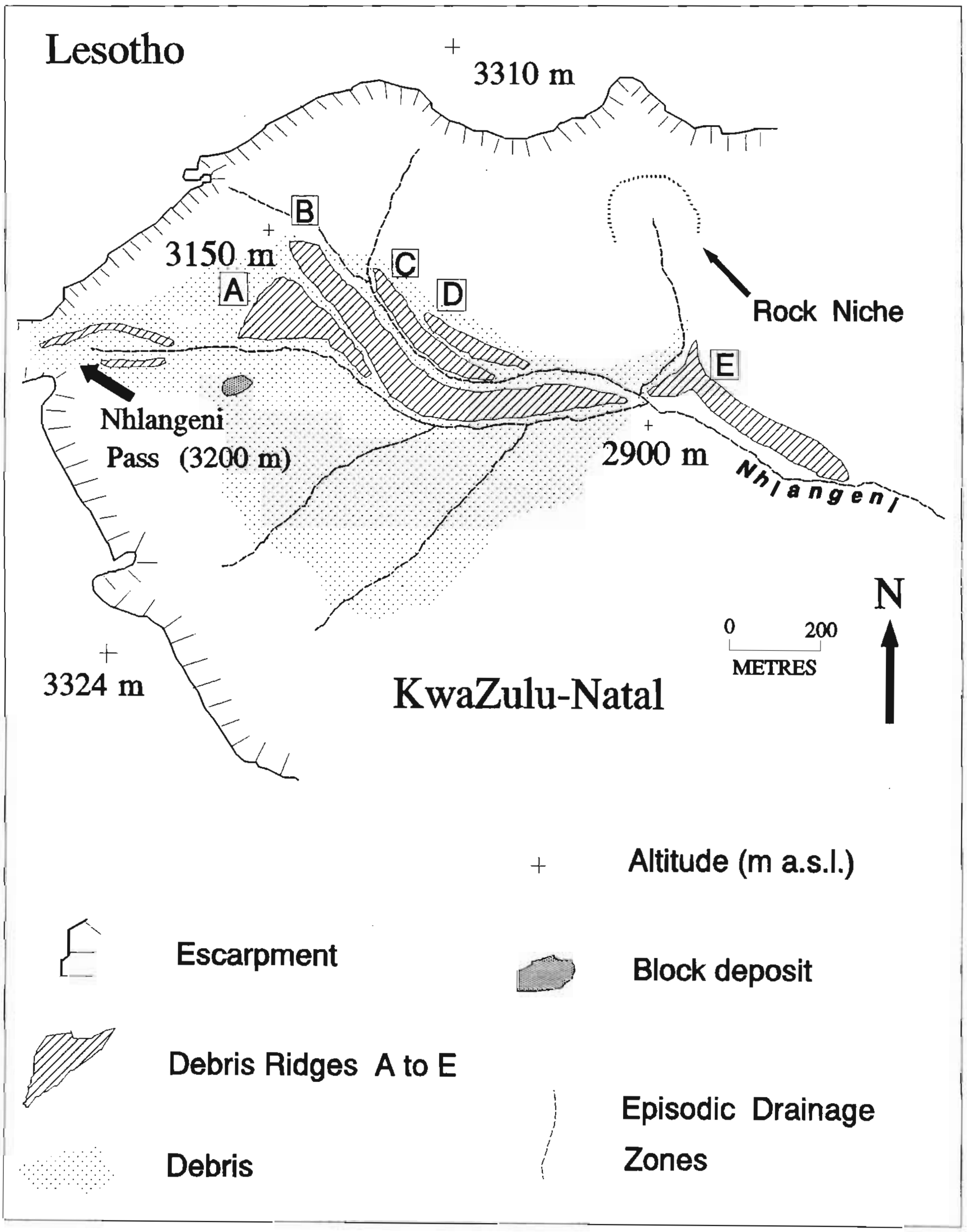


Figure 6.28 Geomorphological map of the Nhlangeni cutback.

Figure 6.29 The debris ridges in the upper Nhlangeni cutback.

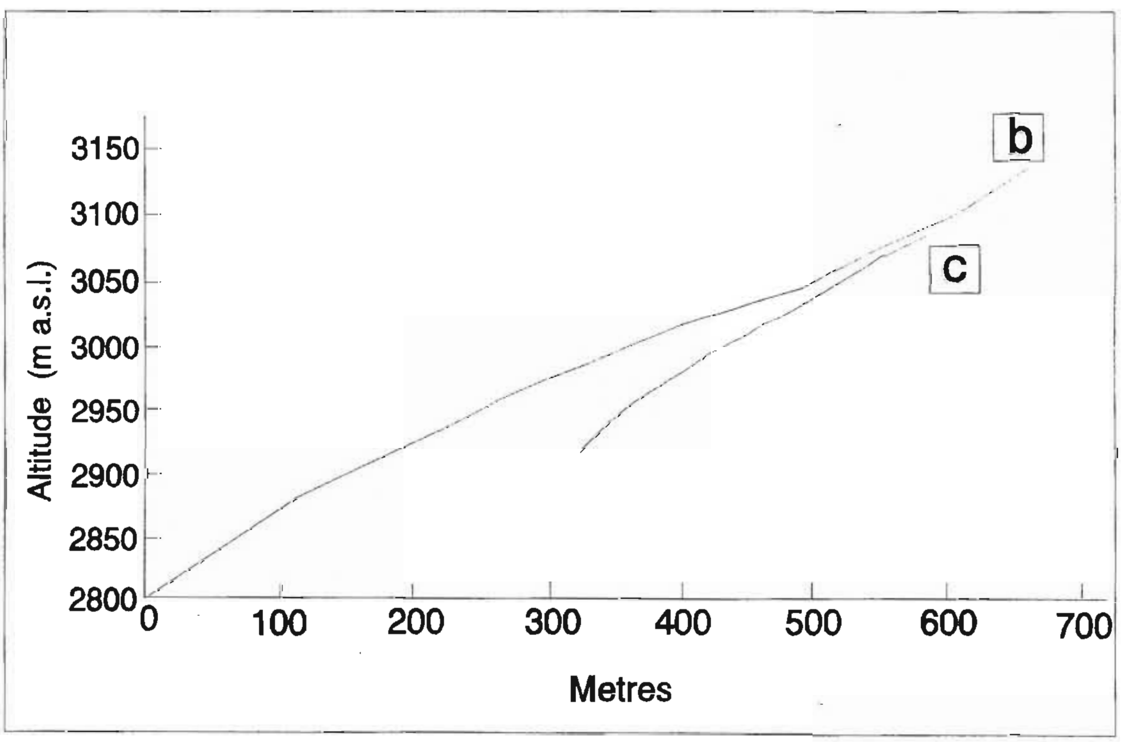


Figure 6.30 Downslope profiles of debris ridges A and B.

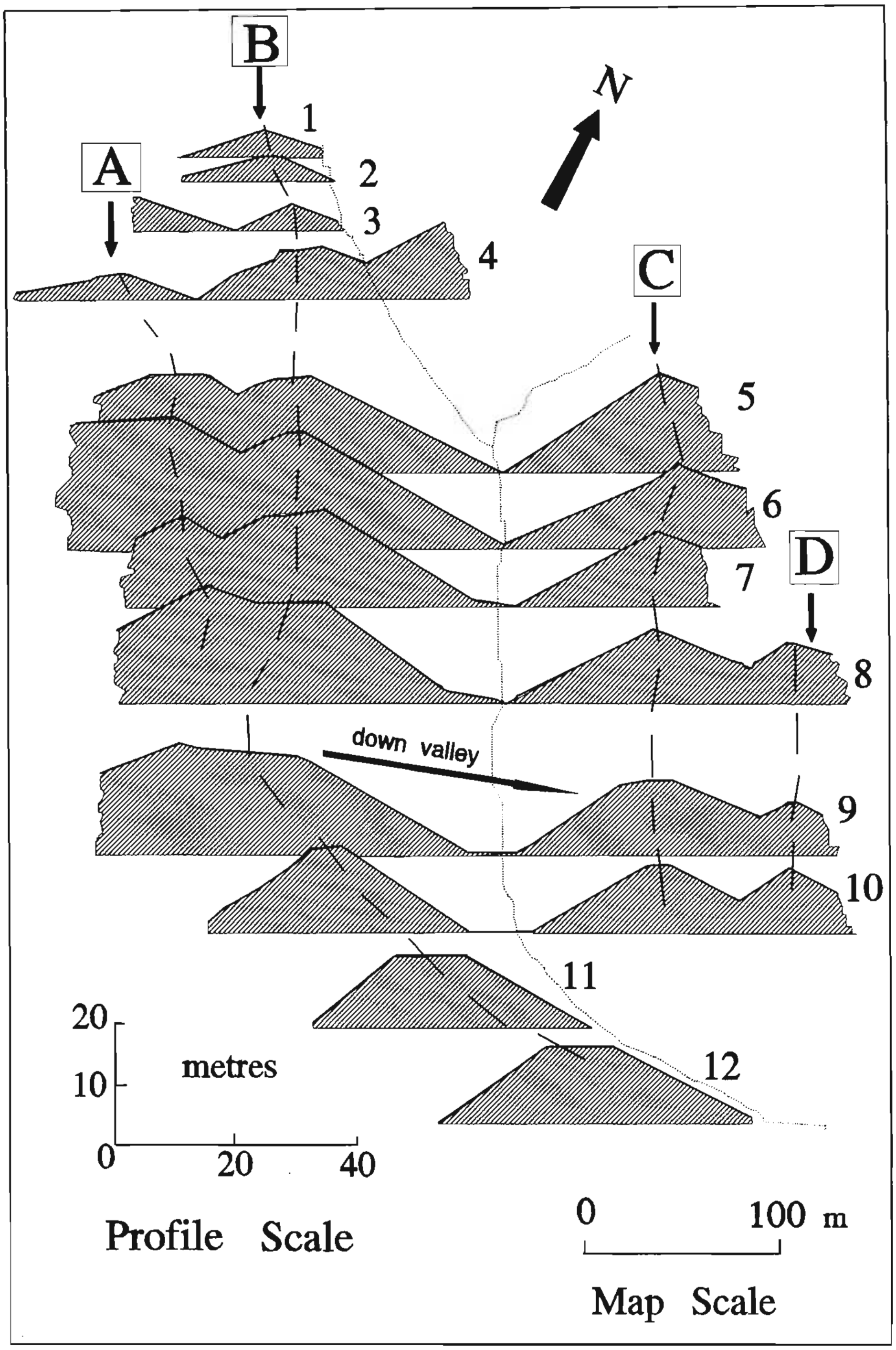


Figure 6.31 Cross-ridge profiles through ridges A-B-C-D.

	Ridge Crest near Backwall				Inter-ridge Basal Zone			
	Sample size	Max. Diam.	Mean diam.	Std. Dev.	Sample size	Max. Diam.	Mean Diam.	Std. Dev.
Xanthoparmelia	42	190 cm	79.93 cm	99.9	22	77 cm	46.54 cm	18.19
White Crustose Lichen	62	193 cm	98.77 cm	106.1	28	264 cm	111.14 cm	63.35

Table 6.4 Comparing lichen size between the ridge crests and the inner-ridge basal zones.

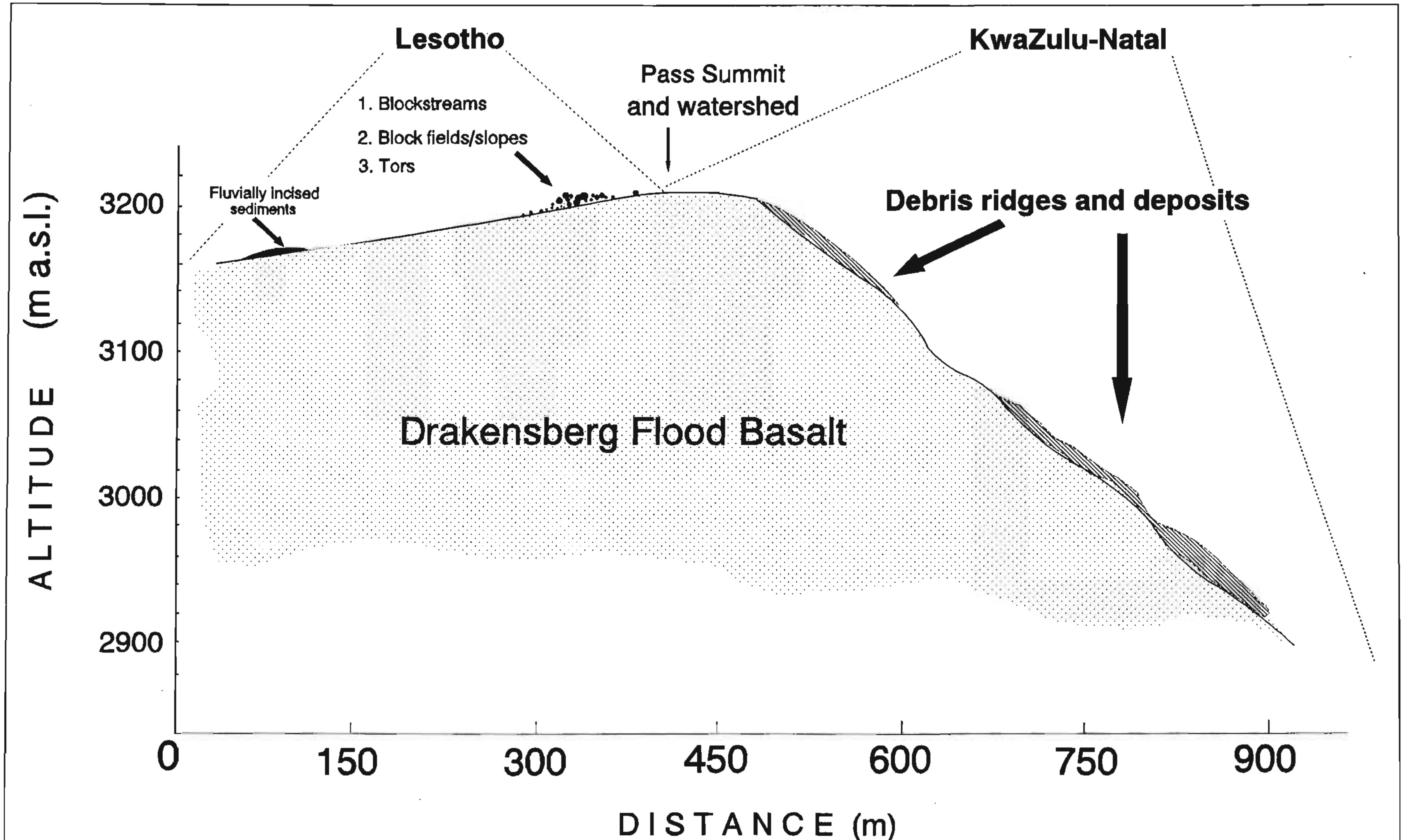


Figure 6.32 Profile through the Nhlangeni cutback and adjacent Lesotho plateau, showing zones in which particular landforms predominate.

The unconsolidated material consists mostly of rocks and boulders ranging from 0.1 to over 2 m in diameter. Although a humic soil cover of up to 0.45 m has formed in places, there is no stratification within the deposit. Very little interstitial material occurs and fines are absent even at depth. Clasts are found to be relatively flat (mean flatness index, site A = 231.0, site B = 244.3) and very angular (mean roundness index, site A = 0.11, site B = 0.14) (Figure 6.33). The two samples, taken from opposite sides of the cutback, were found to be part of the same population at the 0.05 significance level (using the Mann-Whitney U test) and show similar trends of standard deviations and mean values for both flatness and roundness indices (Figure 6.33). The a-axis orientation of stones shows considerable scatter (Figure 6.34). This is also reflected by Chi-square test results (site A: $\chi^2 = 13.47$; site B: $\chi^2 = 13.66$) which do not indicate orientation of clasts at the 0.05 significance level (Figure 6.34). The larger cobbles and boulders have a visual roundness value of 2 to 3 (after Krumbein, 1941).

6.5.2 Fossil Boulder Beds

Extensive boulder beds 1 to 5 m in height and several hundred metres in length are found up to eight kilometres downstream from these cutbacks. These highly rounded basalt boulders have a mean diameter of well over one metre but in many cases have no supporting matrix. The heavy lichen growth and thick weathering rinds are indicative that these boulders have remained stationary for a long time and that some of these channels are now fossil forms. Since some of the stacked basalt boulders are found eight kilometres from the nearest basalt outcrop, it would imply that high levels of discharge transported these boulders.

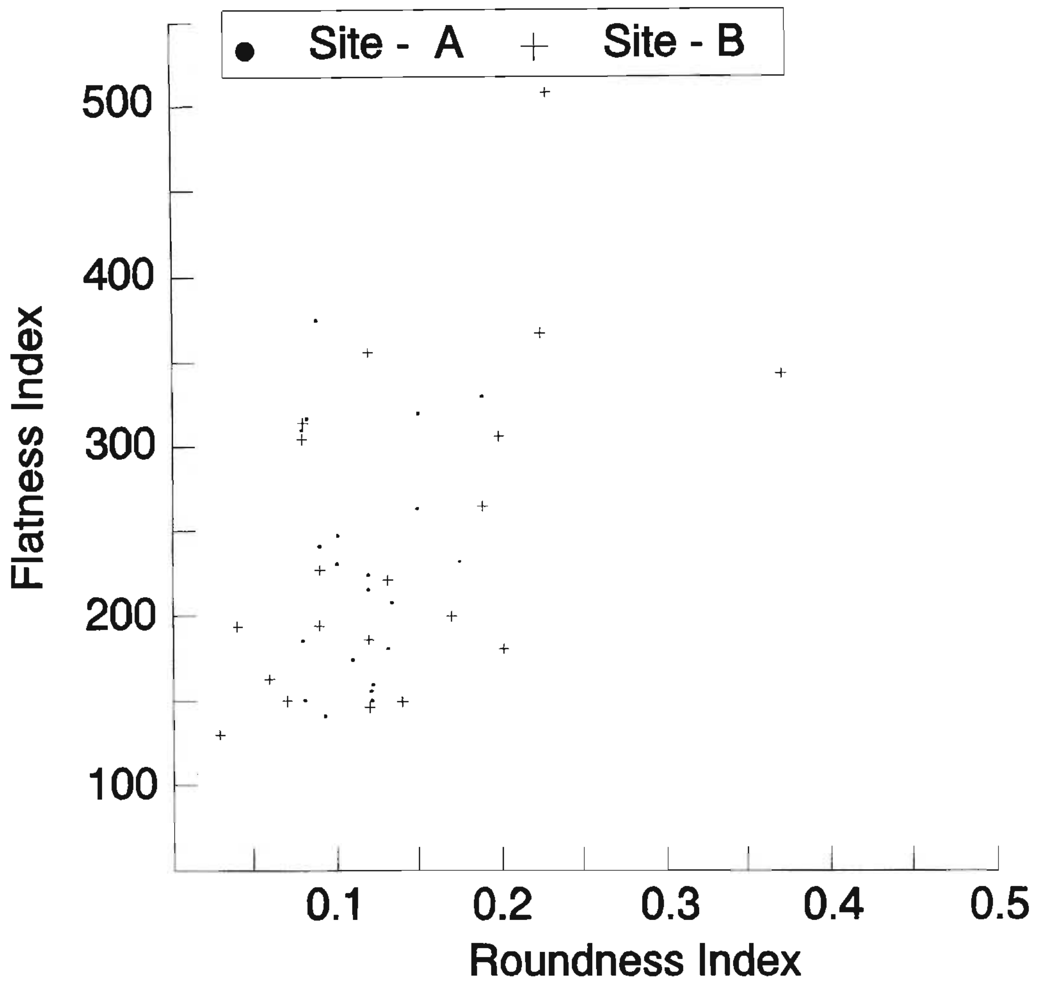
6.5.3 Possible Origin of debris accumulation

Several hypotheses are able to account for the debris accumulation:

1. *High magnitude slope failure*

a. *Debris flow/slide*

Numerous catastrophic slope failures have been known to occur in both bedrock and unconsolidated slope deposits (Wilson, 1990), contributing to large debris accumulations. Morphological characteristics typical of debris flows, including multi-ridges and well defined



	Site A		Site B	
	Mean	Std. Dev.	Mean	Std. Dev.
Flatness Index	231.0	64.0	244.3	95.9
Roundness Index	0.11	0.02	0.14	0.08

Mann - Whitney U test

Flatness Index	Roundness Index
$U_1 = 168.5$	$U_1 = 187$
$U_2 = 231.5$	$U_2 = 213$
Null hypothesis is accepted at 0.05 significance level	Null hypothesis is accepted at 0.05 significance level

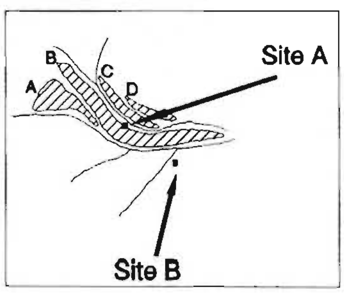


Figure 6.33 Clast size analysis for two sites within the Nhlangeni debris deposit.

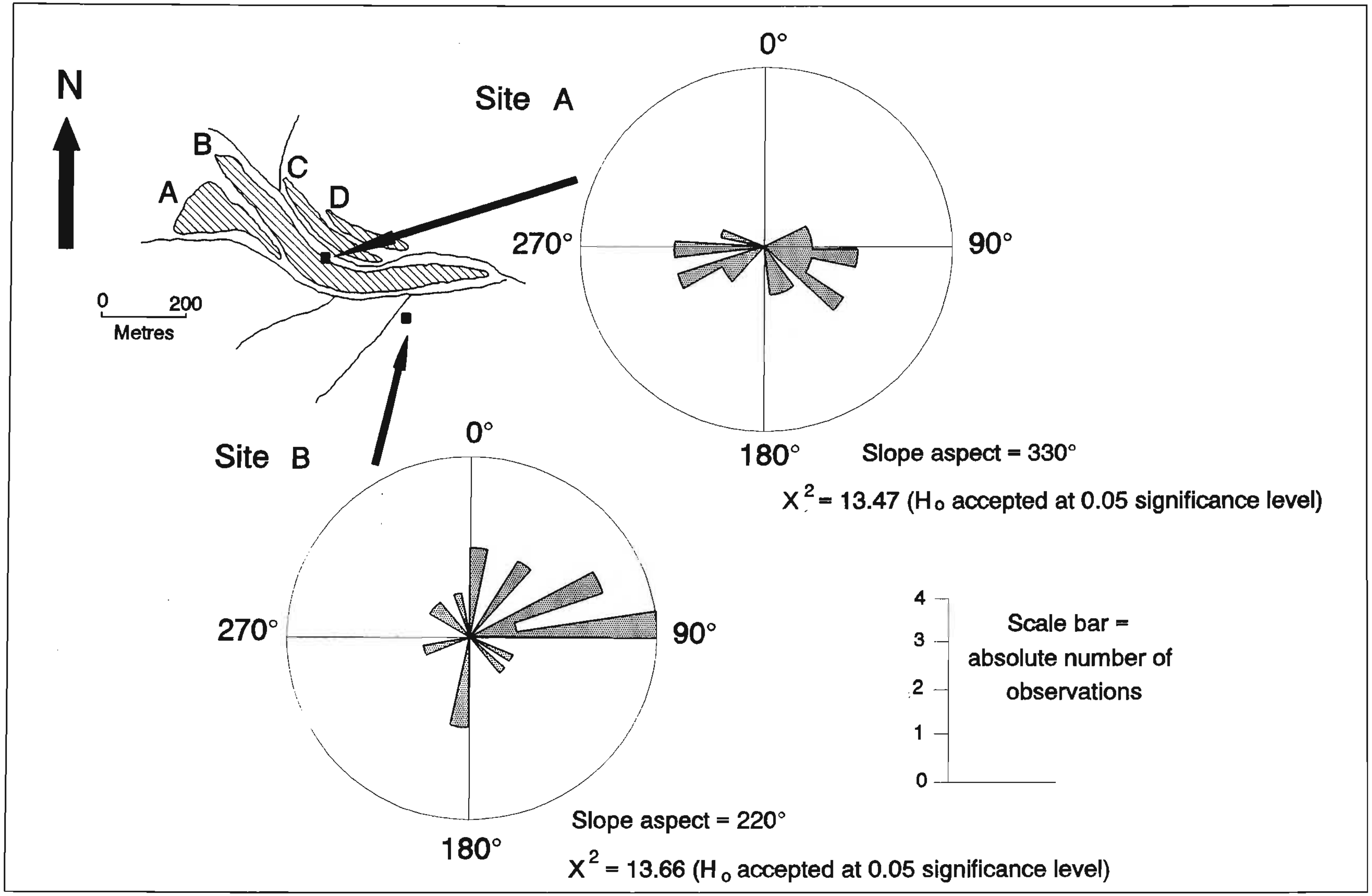


Figure 6.34 Clast a-axis orientation for two sites within the Nhlangeni debris deposit.

front lobes (Rapp and Nyberg, 1981; Whalley *et al.*, 1983; Dawson *et al.*, 1986; Harris and Gustafson, 1993) are displayed in the Nhlangeni deposit. The multi-ridge formation in the Nhlangeni cutback is, however, a product of incision rather than deposition. This is evident from the cross-ridge profiles which show progressive lowering of alternate ridge crest heights in the down-valley direction (Figure 6.31) and from the drainage characteristics which show that the inter-ridge basal zones are a continuum of the bedrock drainage zones above the deposit.

Another possibility is that several debris flows from opposite sides of the cutback contributed to the deposit. Although this should result in some stratification within the deposit, the typical characteristics of such stratification (Van Steijn, 1988) are absent here. The classic "sturzstrom" event described by Heim (1932) and Hsü (1975, 1978), where debris moves towards opposing valley sides is also unlikely because of the large lateral and vertical extent of the deposit and the absence of an extended shear surface. Because debris flows are usually associated with a saturated mass (Varnes, 1978), it is necessary to consider additional sources of water because the catchment above the deposit is too small to permit an adequate volume of water to collect here from precipitation alone. Harris and Gustafson (1993) list several factors including melting snowpacks, icepacks and ice-rich permafrost as additional sources of water for such debris flows. Such suppliers of additional water are, however, absent under present climatic conditions. Further, the fine matrix material which is a common characteristic of debris flows (Hsü, 1975, 1978; Rapp and Nyberg, 1981; Van Steijn, 1988) is also absent here.

b. *Rockslide/Rockfall*

In view of the angle of the rock shear surfaces along the escarpment being very near 90° within similar rock units (Moon and Selby, 1983), it is more likely that macroscale rock falls rather than rock slides would have developed in this and other cutbacks. Microscale rockslides or rockfalls are, however, sometimes active along the escarpment walls. Massive rock failure of a more catastrophic scale would require widespread weakening and fracturing of the bedrock, which is, however, not the case within similar rock units (Moon and Selby, 1983). Although it is possible that processes such as freeze-thaw could have contributed to rock weathering in the past, there are no visible scars within the bedrock and large masses of displaced bedrock are absent. Several authors have indicated that scars associated with Late Glacial and early Holocene slope failures are still clearly visible in some regions (c.f. Wilson, 1993b). The absence of large-scale displacement of bedrock masses in the extensive Nhlangeni deposit, makes it difficult to envisage macro-scale slope failure during late Pleistocene or more recent times. In short, the debris flow or rock fall hypotheses appear unable to explain the lateral extent of the debris across the entire cutback and towards the pass summit, above which bedrock outcrops are absent.

2. Cryogenic processes

a. *Nival processes*

Weathered rocks detached from the free face may fall onto the surface of perennial snow patches and slide, roll or bounce across them, often forming protalus ramparts (French, 1976; Hall, 1985). The meltwater from such snowpacks can also be an important contributor towards chemical weathering and erosion below the snowpacks (Thorn, 1979; Hall, 1985; Strömquist, 1985). However, the nival-related sedimentary features mentioned in the literature (e.g. Ono and Watanabe, 1986; Ballantyne, 1987a; Shakesby *et al.*, 1987) are miniature in comparison to the massive deposit in the Nhlangeni cutback. It is, therefore, doubtful whether snow patch related processes on their own are capable of producing such large scale deposits.

b. *Rockglacier development*

Several characteristics associated with rockglaciers including steep fronts, blocky surfaces and steep side slopes (Barsch, 1977, 1988; Benedict *et al.*, 1986) can be identified in the Nhlangeni deposit and therefore give it the appearance of a fossil rockglacier. Collapsed rockglacier structures such as ridges and gullies may result from subsurface drainage (Johnson, 1981), however, this is not the case with the Nhlangeni deposit where the nature of the drainage patterns indicate that the gullies and ridges are a product of past fluvial incision. The downward fining of rock material within rockglaciers (Barsch, 1977) is not found here. Rockglaciers require an adequate supply of debris for their formation (Whalley and Martin, 1992), and therefore, as discussed above, still have the problem of debris source.

c. *Plateau/Niche/Cirque Glaciation*

Although Groom (1959) has written a detailed paper examining niche glaciers and their associated landforms, this model has not been widely used for the interpretation of fossil landforms. The following attributes discussed by Groom (1959) have, however, been identified in the Nhlangeni and other cutbacks :

1. Steep sided, funnel-shaped hollows with occasionally more deeply incised hollows (Figure 6.35).
2. Short, shallow gully formation in the bedrock with debris deposits at their base.
3. Debris frequently extending up the cliff face.
4. Steep buttresses maintained between hollows.

The cutbacks containing such deposits are all bounded by broad mountain summits well over 3200 m a.s.l. The higher summits reveal some landforms which have not yet been identified. These include pairs of linear ridges on some south-facing slopes, "roche moutonnée" -shaped tors and gently sloping summits with smoothly rounded bedrock surfaces (Figure 6.32). The tors are gently sloping on the upslope side but have displaced angular and rounded rock blocks on the downslope side. Although Sugden and Watts (1977) believe that some roches moutonnée-shaped tors may be of glacial origin, Lindström (1988) cautions that at least some roches moutonée-shaped bedrock features are of preglacial origin. So far, the tors have only been located above the Nhlangeni, KwaNtuba and Bannerman's Pass cutbacks and much further analysis is required of these and other features before unequivocal evidence for their geomorphic origin can be presented.

Cirques, which are "hollows, open downstream but bounded upstream by the crest of a steep slope which is arcuate in plan around a more gently-sloping floor" (Evans and Cox, 1974, p151; Evans and Cox, 1995, p179), appear to occur, but are only found on the south-facing aspects of the highest cutbacks (Table 6.3). Groom (1959) has associated such rounded half funnel-shaped hollows as landforms characteristic of niche glacier development. Two such possible cirques found in the Nhlangeni and KwaNtuba cutbacks, have very distinct debris accumulations immediately bellow the hollow (Figure 6.35).

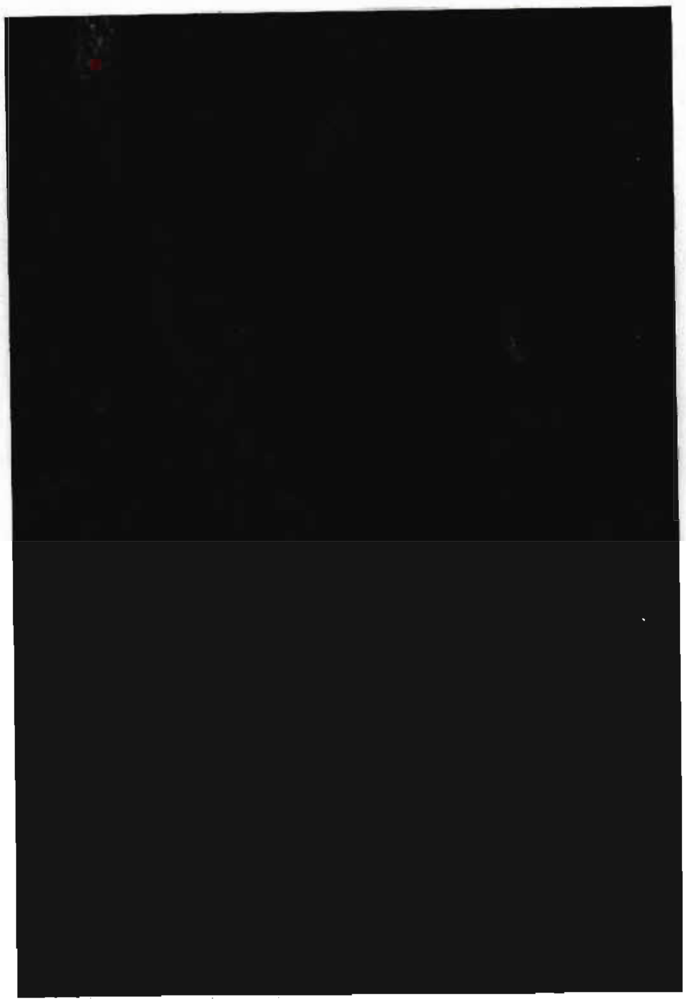


Figure 6.35 A steep sided, deeply incised hollow on the upper Nhlangeni cutback south-facing slope.

The debris within the Nhlangeni cutback consists of randomly orientated angular boulders and clasts, which are concentrated on the south-facing aspects. These sedimentological characteristics would support an hypothesis of glacial dumping over the escarpment walls. The debris, which extends to the pass summit, could be attributed to retreating ice. The other hypotheses are unable to account for the occurrence of debris at the pass summit.

6.5.4 Discussion : a palaeogeomorphic postulation

Although the remnant features identified in this study do not provide unequivocal palaeoenvironmental information, particularly as it is difficult to date them, they do, nevertheless, serve as possible indicators for the localized development of plateau, niche and/or cirque glaciation in the high Drakensberg.

It is postulated that during the Last Glacial Maximum, ± 18 000 yrs B.P. (Van Zinderen Bakker, 1986), localized glacial development occurred at favourable sites in the high Drakensberg and that glaciers overflowed the escarpment walls at insolation-protected sites. This would correspond closely with the last glaciation on Mount Kosciusko, Australia (±20 000 to 15 000 yrs B.P.) (Costin *et al.*, 1967) and Mount Elgon, Uganda/Kenya (±20 000 to 17 000 yrs B.P.) (Hamilton, 1979). Harper (1969) has extrapolated a 5.5 to 9°C Late Pleistocene temperature drop for the highlands of Lesotho, and Talma *et al.* (1974) an 8 to 9.5°C drop for the Wolkberg caves at 19 800 yrs B.P. More conservative estimates for temperatures during the Last Glacial Maximum are 5°C (Deacon *et al.*, 1984) to 6°C (Vogel, 1985) lower than those of today. With an extrapolated modern M.A.A.T. of 4.2°C for Khalong-la-Lithunya (3240 m a.s.l.) (Backéus, 1989), it is conservatively estimated that the higher Lesotho summit regions and Drakensberg cutbacks experienced a M.A.A.T. of between -1 and -3°C during the Last Glacial Maximum. Because the Antarctic polar fronts were displaced northwards from those of today (Howard, 1985), there is a strong possibility that winter precipitation increased during this time (Van Zinderen Bakker, 1975) and consequently brought heavier snowfalls. Conditions during the Late Pleistocene would therefore have been favourable for localized firn development, and particularly so, in pre-existing high altitude cutbacks and south-facing rock niches which remained protected from insolation. The ice would have eroded the escarpment edge and rock niches along the escarpment, thereby enlarging pre-existing gullies and forming funnel shaped hollows and cirques. These ice masses would then have dumped weathered rock material over the escarpment, where debris accumulated in topographically ideal locations. It is most probable that avalanches contributed significantly to the debris accumulation.

Rapid warming, about 16 000 yrs B.P. (Partridge, 1993), may have resulted in much of the snow and ice melting over a short period of time. The increased surface water and runoff during this time (Vogel, 1985) may well have reached catastrophic levels which, in part, can be attributed to rapid melting of snow and ice (Hall, 1994a). This wetter phase, coupled with sufficient meltwater discharge, may have initiated shallow gully formation in the bedrock and produced incision into the deposit below, forming the distinctive ridges found in the Nhlangeni cutback. The fluvial capacity may well have reached levels which enabled boulders to be transported several kilometres downstream.

6.6 SUMMARY

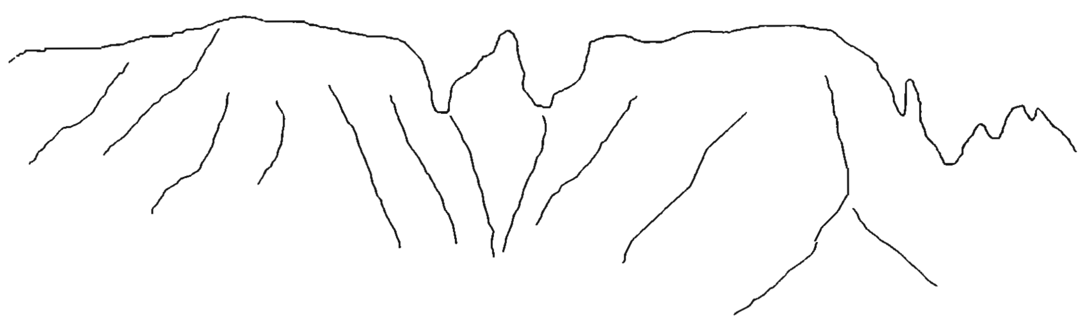
A variety of detrital accumulations including blockfields, blockstreams, stone-banked lobes and debris deposits have been examined. Findings show that these landforms are most frequent at higher altitudes (often above 3200 m) and are predominantly found on south- and southwest-facing slopes. Further implications are that cryogenically induced mass-movement events have been most pronounced on such high altitude slopes.

It is postulated that during the Late Pleistocene, small accumulations of ice developed on localized high altitude south-facing slopes. Full periglacial conditions also appear to have occurred during Holocene Neoglacial events when solifluction processes may have contributed to many of the landforms mentioned in this chapter. Although the larger mass-wasting forms appear to be inactive, several of the smaller landforms show some indication of current activity. Therefore, it would appear that some limited periglacial mass-wasting still prevails, particularly on the higher summits. Such periglacial mass-wasting has also contributed significantly to the slope evolution in the high Drakensberg, as is evidenced on the Mafadi-Njesuthi south slopes.

It has been argued by such as agriculturalists and conservationists that the occurrence of large open stonefields with sparse vegetation cover on the higher Drakensberg slopes is indicative of overgrazing. However, from the findings and discussions outlined in this chapter, it is suggested that a predominantly cold climate with enhanced weathering and cryogenic activity is/has been the primary cause for the stone/rock dominated and sparsely vegetated higher summits. There is, nevertheless, little doubt that erosion has been a contributing factor to many of the more sparsely vegetated sites. However, the extent to which cryogenically-induced erosion processes operate in the high Drakensberg, still requires detailed assessment. The following chapter (Chapter 7) therefore describes several erosion landforms occurring in the high Drakensberg and ascertains the significance of cryogenically-induced erosive processes in their initiation.

CHAPTER SEVEN

TURF EXFOLIATION, NEEDLE ICE PROCESSES AND BANK EROSION



CHAPTER 7

TURF EXFOLIATION, NEEDLE ICE PROCESSES AND BANK EROSION

7.1 INTRODUCTION

Information on soil erosion for the high Drakensberg is extremely limited, yet active erosion is considerable in a number of places (Corté, 1991). Since grazing lands along valley bottoms and within wetlands are an important resource for the Basotho livestock, the rate and extent of turf loss is of obvious concern to many highland farmers. Many of the wetlands are now subject to accelerated erosion, showing deep gullies in the peatlands and sediment deposition on the sedge-meadows (Backéus and Grab, 1995; Morris and Grab, 1997). Overgrazing is considered an important initiating agent to other erosive processes (Morris and Grab, 1997) but the geomorphic processes contributing to sediment mobilization still require further examination.

The extent to which cryogenic processes contribute to erosion, or are overwhelmed by other processes in the high Drakensberg, is unknown. Therefore, this chapter attempts to evaluate the significance of various erosive processes and describe some of their resulting landforms.

7.2 TURF EXFOLIATION

Turf exfoliation is defined as "a denudation process active in periglacial areas which destroys a continuous ground vegetation cover by removing the soil exposed along small terrace fronts" (Pérez, 1992b, p82). Needle ice action, desiccation and aeolian deflation are said to be the primary causes for turf exfoliation (Troll, 1948; King, 1971). Turf exfoliation frequently occurs in cold environments which have an uneven temporal precipitation pattern (Brandt, 1917) and relatively restricted snow cover.

Turf exfoliation has been reported from numerous localities throughout the world, including: Scotland, Iceland, Germany and Poland (Sapper, 1915; Brandt, 1917; King, 1971). However, it is said to be most pronounced in high alpine regions, and quite frequently so on high tropical mountains (Troll, 1973; Pérez, 1992b) such as Mt. Wilhelm, Papua New Guinea (Löffler, 1975), the Venezuelan Andes (Pérez, 1992b), the Mexican volcanoes (Heine, 1977), the Semyen Mountains, Ethiopia (Troll, 1973; Hastenrath, 1974), and East Africa's Mt. Kenya and Kilimanjaro (Hastenrath, 1973, 1978). Despite this widespread occurrence, little detailed work has been undertaken on such an erosion process. The only known measurements on turf exfoliation rates and analyses of soils and plants affected, are those by Pérez (1992b).

In southern Africa, turf exfoliation has only been reported from the high Drakensberg (Troll, 1958; Hastenrath, 1972; Boelhouwers and Hall, 1990; Boelhouwers, 1991a), but data are absent despite its extensive occurrence in some areas. This section aims to provide empirical data on the high Drakensberg turf exfoliation forms and preliminary findings on possible causative mechanisms.

7.2.1 General Characteristics

Turf exfoliation in the high Drakensberg is most common along valley floors and lower valley slopes. Steep slopes (more than 10°) rarely exhibit turf exfoliation features, but are frequently occupied by terracettes. The lower limit for turf exfoliation in the high Drakensberg is said to be between 2000 and 2500 m a.s.l. (Troll, 1944, 1973) and may be controlled by the occurrence of needle ice activity and/or deflation processes. Such denudation processes are, however, actively disrupting vegetation cover to altitudes of over 3400 m, such as at the northern aspects of Mafadi Summit.

The locality of individual turf exfoliation study sites in the Mashai Valley are shown in Figure 7.1. The zones in which turf exfoliation occurs most frequently are the valley bottom (zone of channel flow and adjacent poor drainage), valley floor and lower valley slopes (Figure 7.1). Each of these locality zones hosts particular erosion landforms, resulting from turf exfoliation (Figure 7.1). Brandt (1917) made very similar observations and also reported turf exfoliation features from three zones within river valleys in Europe. At the valley bottom, adjacent to the Mashai Stream, are found wet, incised depressions on a 2° slope (Figure 7.2 and Table 7.1).

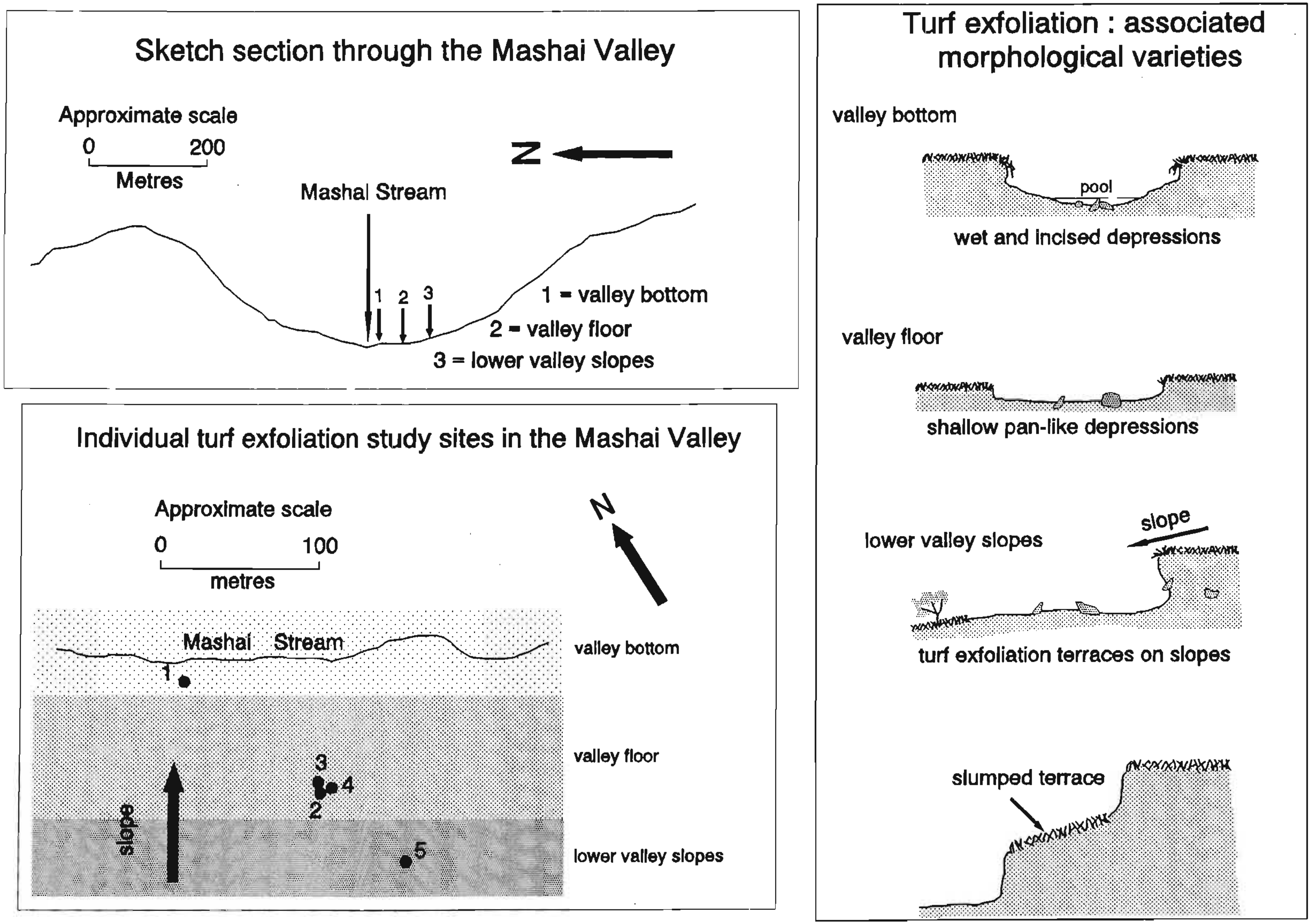


Figure 7.1 Turf exfoliation study sites in the Mashai Valley, with distinct locality zones which host particular turf exfoliated morphological varieties.

Figure 7.2 A turf exfoliated depression at a Mashai Valley bottom site.

Characteristics of individual turf-exfoliated sites							
Locality	Altitude (m a.s.l.)	Slope aspect	Slope gradient	Exfoliated depression			Mean riser height (cm)
				Max. as diameter (m)	Max. ds diameter (m)	Mean diameter (m)	
Valley bottom							
1	~2950	NW	2°	7.50	3.47	5.48	33.0
Valley floor							
2	~2950	NW	2°	2.62	3.40	3.01	15.3
3	~2950	NW	2°	6.75	4.50	5.62	10.7
4	~2950	NW	2°	9.92	1.83	5.87	14.5
Lower valley slope							
5	~2955	NW	6°	-	-		36.0
as = across-slope / ds = downslope							

Table 7.1 Characteristics of individual turf exfoliated sites in the Mashai Valley.

The depressions are usually full of water in the summer months but are reduced to small pools in winter. Because the soils remain wet throughout the dry winter months, needle ice development occurs during most winter nights and turf is exfoliated from the risers around such depressions. The depression measured had an across-slope main axis of 7.5 m and a downslope main axis of 3.47 m (Table 7.1). These depressions are frequently between 60 and 80 cm deep.

Common along the valley floor, are shallow pan-like depressions, usually no more than 20 cm deep (Figure 7.3). Depressions appear similar to those described from the Peruvian and Venezuelan Andes, which have also been described as needle ice pans (Hastenrath, 1977; Pérez, 1992b). These pans only hold shallow pools after heavy rainfall events and become rapidly desiccated during the winter months. Although the pans have no common shape, they are frequently circular or elongate, often with pronounced arcuate shaped risers. The pans reported from Venezuela are between 1 and 4 m in diameter (Pérez, 1992b), while those measured in the Mashai Valley have a long-axis range of between 3.01 and 5.87 m (Table 7.1).

Figure 7.3 Shallow pan-like depressions occur along the Mashai Valley floor.

On the steeper (6° or more) lower valley slopes are found turf exfoliation terraces (Figure 7.4 and Table 7.1), which frequently parallel the contours. Downslope from the retreating terrace is usually found bare ground with rocks and stones (Figure 7.5), as has also been reported by Pérez (1992b) from Venezuela. Some lower valley turf exfoliation sites in the Mashai Valley display slumped terraces with two or three tiered risers (Figure 7.1 and 7.6).

Terracettes are common on slopes throughout the Drakensberg and have been described as a "tiered microrelief on moderately steep grassy slopes" (Higgins, 1982, p460). Although Gallart *et al.* (1993) argue that terracettes "belong to a family of periglacial forms" (p529), others (e.g. Vincent and Clarke, 1976) are not convinced of an "exclusive" periglacial origin. Harper (1969) and Hanvey and Marker (1992) believe that terracettes in the high Drakensberg are primarily the result of needle ice growth and frost heave mechanisms. From this, it could be argued that terracettes are a turf exfoliation landform. However, many terracettes in the high Drakensberg are located on dry north-facing slopes where needle ice very rarely develops, and further, the role of grazing animals in high Drakensberg terracette formation cannot be disputed. The study by Day (1993) suggests that the role of animal disturbance in the initiation of high Drakensberg terracettes may be more significant than that which the literature suggests. From three years of observations, there is little doubt that the terracettes are primarily the result of animal trampling, only slightly modified by a variety of geomorphic processes. Owing to the work already undertaken by Day (1993), there is little need to further examine these landforms in this dissertation, particularly as their origin is largely zoogeomorphic.

7.2.2 Vegetation Characteristics

Pérez (1992b) gave an account of plants growing on retreating micro-terraces along the Mifafi Valley in the Venezuelan Andes. Although plant types growing on such terraces in the Mashai Valley were not identified, general ground cover characteristics appear similar to those described by Pérez (1992b). The vegetation along the terrace edges consists mostly of grasses, sedges and *Helichrysum*. Rooting, which usually extends about 5 to 10 cm down the terrace risers, is able to bind the soil. Therefore, soil is somewhat less disrupted in the upper 5 to 10 cm of the terrace risers, than below. This is a likely reason for the frequently overhanging

Figure 7.4 Lower valley slope turf exfoliated terraces, Mashai Valley.

Figure 7.5 Downslope of the lower valley turf exfoliated terraces is usually found bare ground with rocks and stones.

Figure 7.6 A slumped terrace at a Mashai Valley turf exfoliation site.

swards of vegetation seen at terrace risers (Figure 7.7). The exposed and overhanging roots are, however, subjected to extended periods of desiccation and freezing conditions. Consequently, the lower parts of the terrace risers frequently become littered with dead or withered vegetation towards the end of winter.

The pan-like turf exfoliation depressions ("deflation hollows") along the valley floor are occupied by a seasonal herbaceous mat of *Crassula dependens* and *Sebaea marlothii*. Plant unit density counts were undertaken in three depressions. Results in Table 7.2 show that plant densities reached a peak in March, when plant units averaged between 18.8 and 27.4 per 4 cm^2. A month later the average densities had dropped to between 0.7 and 2.7 per 4 cm^2 (Table 7.2). No live plants were found within the depressions from June to September. All sampling grids recorded plant cover during December and March. However, by May only 20 to 60% of the grids showed plant cover (Table 7.2). From such measurements and two years of observations, it is evident that there is a cyclic pattern of vegetation growth within these depressions. The depression soils slowly become saturated during the spring months, allowing seeds to germinate. Although such sites are frequently trampled by livestock, the continuously

Figure 7.7 Overhanging swards of vegetation are frequently visible at terrace risers.

Depression	Plant unit density counts				Percentage grids with plant cover			
	Dec. 1993	March 1994	May 1994	June 1994	Dec. 1993	March 1994	May 1994	June 1994
1. maximum	27	49	4	0	100	100	20	0
mean	15.7	26.2	0.7	0				
std. dev.	6.6	9.4	1.4	0				
2. maximum	23	39	5	0	100	100	60	0
mean	16.1	27.4	1.6	0				
std. dev.	4.3	11.5	1.7	0				
3. maximum	-	24	8	0	-	100	50	0
mean	-	18.8	2.7	0				
std. dev.	-	4.1	2.9	0				

Table 7.2 Mean seasonal values of individual plant unit density counts in 4 cm^2 grids within three turf exfoliated depressions (n = 10 grids per depression). The percentage grids showing plant cover during the summer, autumn and winter months are also indicated in the table. Plants counted were *Crassula dependens* and *Sebaea marlothii*.

saturated conditions eventually permit a thin mat-like vegetation cover to develop during summer. Such depression soils are then extensively disrupted by needle ice during late autumn, damaging and heaving most of the small herbs and sedges. Eventually, the soils become completely desiccated and the remaining plants wither and die towards early winter. The dead plant matter is, however, likely to play an important role in helping to keep the depression soils bound. This may help to somewhat reduce the rate of aeolian deflation within the pans.

7.2.3 Terrace Morphology and Sedimentology

The mean terrace riser height varies between sites. The highest risers are associated with turf exfoliation terraces on lower valley slopes (mean = 36.0 cm), while the lowest are those on the periphery of pan-like depressions (mean values range from 10.7 to 15.3 cm) (Table 7.1). Owing to the relatively low riser heights at such pan-like depressions, the risers are usually vertical. The higher risers along valley bottoms and lower valley slopes do, however, show a variety of vertical riser profiles (Figure 7.8). Pérez (1992b) has already generated a model for turf exfoliation, which begins with a near vertical terrace riser. The riser is then eroded by needle ice at the base, forming a concavity and an overhanging sward. Eventually, the overhanging sward collapses and the sequence commences again. In the Mashai Valley, a variety of associated depositional micro-forms have been found which conform with this sequence of events (Figure 7.8). Where the terrace riser is still near vertical, a depositional mound of fine earth material sometimes develops against the lower and middle part of the terrace riser, such that only the top half of the terrace riser is left exposed (Figure 7.8). Such depositional mounds develop during winter and early spring, but thereafter are destroyed by rain-wash and overland-flow. Once the concavity is formed, such depositional mounds are replaced by small depositional ridges running parallel to the terrace riser (Figure 7.8 and 7.9). Such ridges are usually no more than 4 cm high and 8 cm in diameter. The relationship between the depth of the concavity (measured as the distance of the overhanging sward from the turf exfoliated riser backwall) and the distance of the depositional ridge from the riser backwall was determined using the Spearman's rank correlation coefficient (Table 7.3). The null hypothesis is rejected at the 0.01 significance level (Table 7.3), thus the deeper the turf

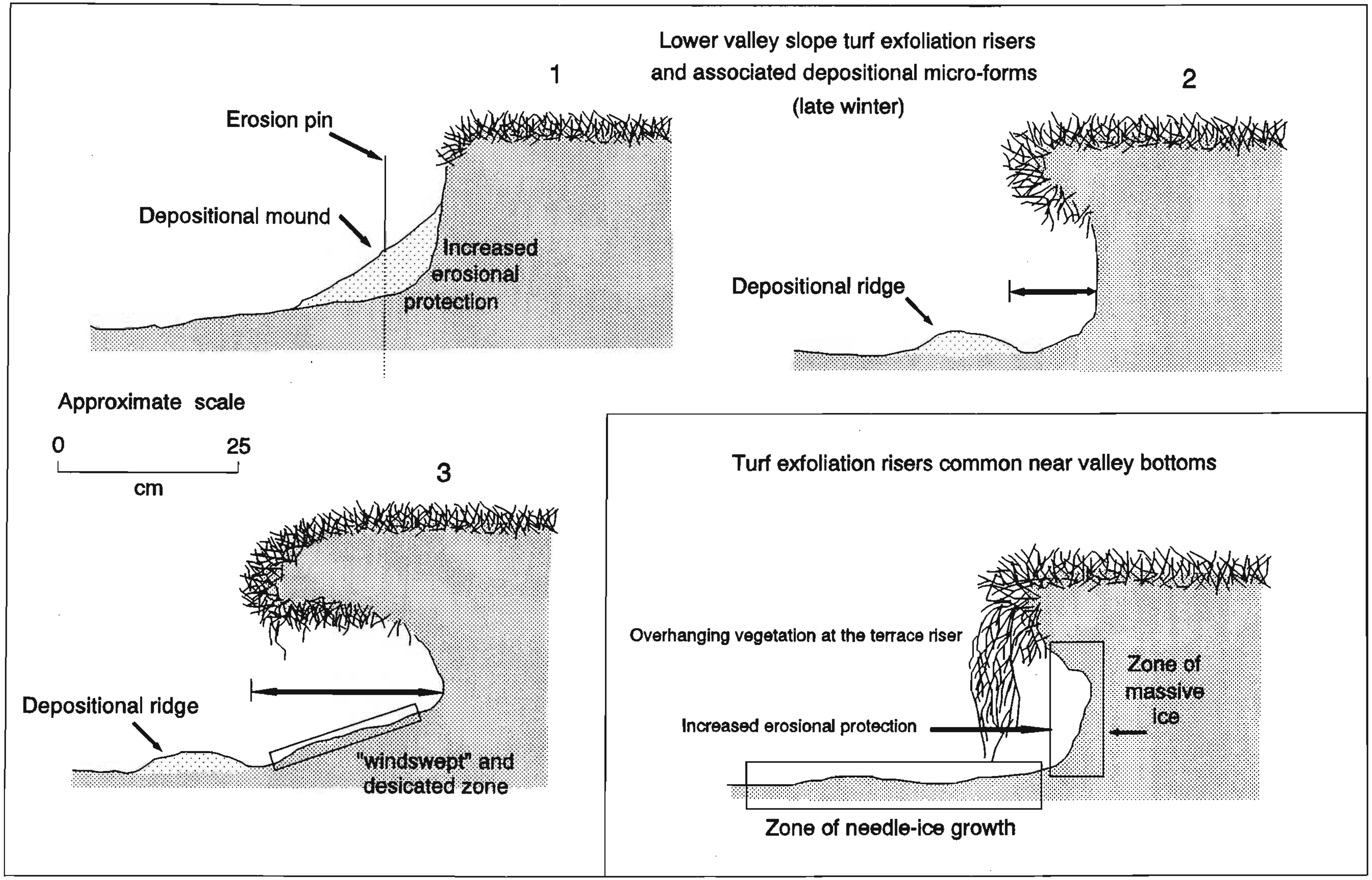


Figure 7.8 A variety of vertical turf exfoliated riser profiles and associated depositional micro-forms.

Figure 7.9 Depositional ridges sometimes run parallel to the terrace riser at lower valley slope sites.

	Distance of overhanging vegetation from scarp backwall (cm)	Distance of depositional ridge from scarp backwall (cm)
	17.5	17.0
	15.5	14.0
	15.0	14.0
	13.5	15.0
	8.5	15.5
	8.0	13.0
	7.5	12.0
	7.0	12.5
	7.0	11.5
	3.5	5.0
	3.0	5.0
	2.0	4.5
mean	10.7	11.6
std. dev.	5.7	4.2
RSQ = 0.72 (null hypothesis is rejected at the 0.01 significance level)		

Table 7.3 Relationship between the distance of overhanging vegetation from the riser backwall and the distance of depositional ridges from the riser backwalls. Results show a correlation, with the deeper turf exfoliated cavities having depositional ridges further from the exfoliated scarp backwall.

exfoliated concavity, the further is the depositional ridge from the exfoliated riser backwall. Figures 7.8, 7.9 and 7.10 show two zones along a terrace riser and an adjacent depositional ridge. Zone one (mid terrace riser) consists mostly of sand and silt, which is loosely held in position (Figure 7.10). Zone two (lower terrace riser) is a "windswept" section which, particularly during the drier months, consists of a compact soil crust with desiccation cracks (Figure 7.10). Although also mostly of sand and fines, this zone has an increased percentage of gravels compared to that of zone one (Figure 7.10). At the base of zone two, gravels and cobbles frequently accumulate as they are eroded from the terrace riser above (Figures 7.9 and 7.10). The particle size distribution of the depositional ridge displays significantly coarser material than that for the retreating terrace (Figure 7.10). This is possibly because finer material is entrained during prolonged periods of high velocity wind action. It is also noticeable that the ridge-side facing the terrace riser has considerably coarser material than the crest- and lee- ridge positions (Figure 7.10). It appears that coarser material is banking against the side facing the terrace riser, while the crest- and lee- ridge positions also support deflated material.

Soil properties for the upper, middle and lower terrace riser positions, as indicated in Figure 7.11, were examined. The bulk density is lowest (0.74 g/cm^3) at the upper position where much rooting occurs, and highest in the middle position (0.86 g/cm^3) (Table 7.4). The organic matter content decreases progressively from the upper position (21.25% of dry weight) to the lower position (18.07%) (Table 7.4). This high organic matter content is likely to have influenced the soil colour, which is mostly olive brown when dry (Table 7.4). The moisture content of surface soils generally increases from the upper to lower riser positions (Table 7.4). However, soils appear to desiccate most rapidly on the middle riser position (Table 7.4). This is possibly because the middle riser position is most exposed to insolation and the desiccating effect of wind. Soils at the upper riser position are protected by roots which draw moisture to them, while those at the lower riser position are better protected from insolation, owing to shading by the terrace riser above it.

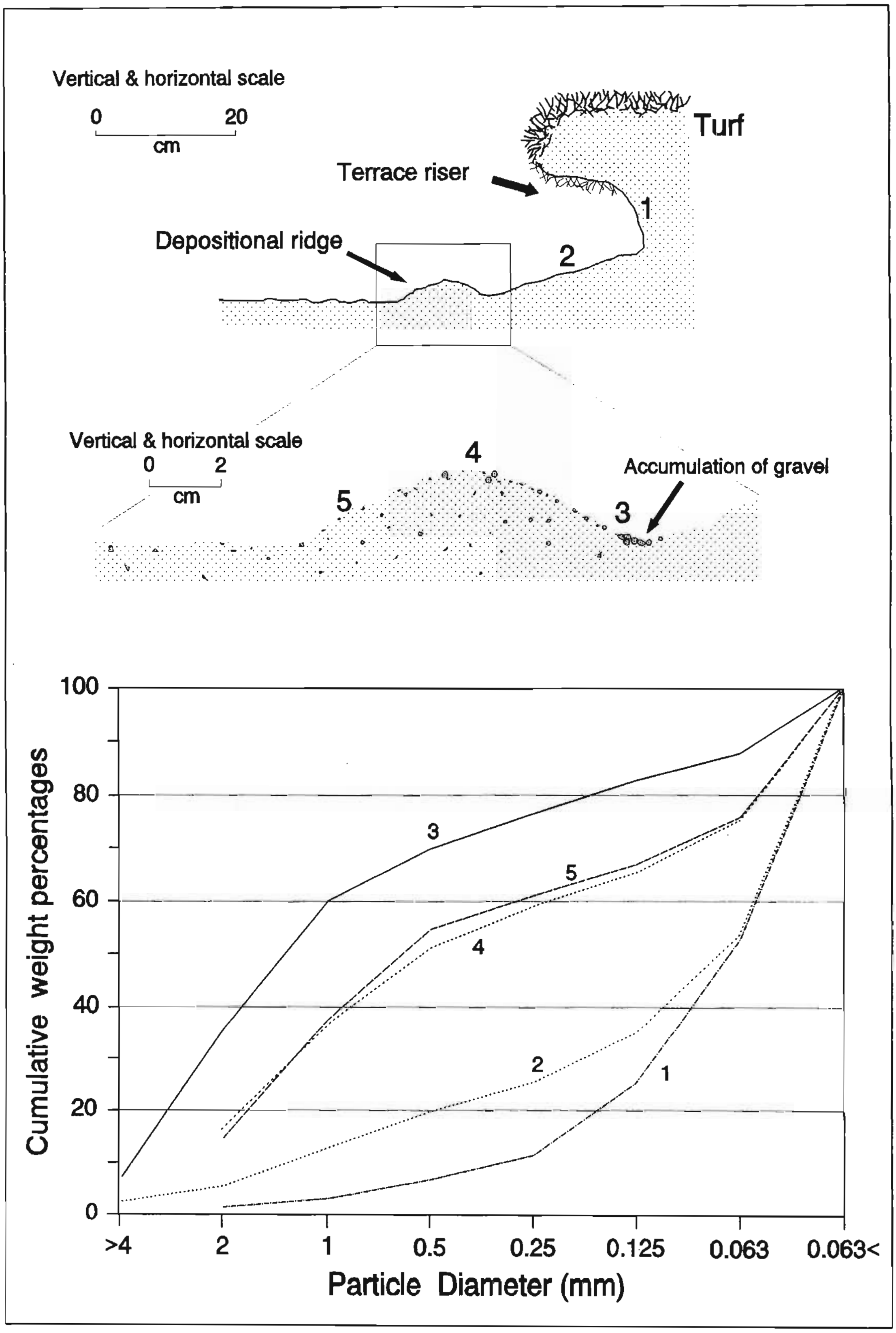


Figure 7.10 Section and particle size distribution through a terrace riser and the adjoining depositional ridge.

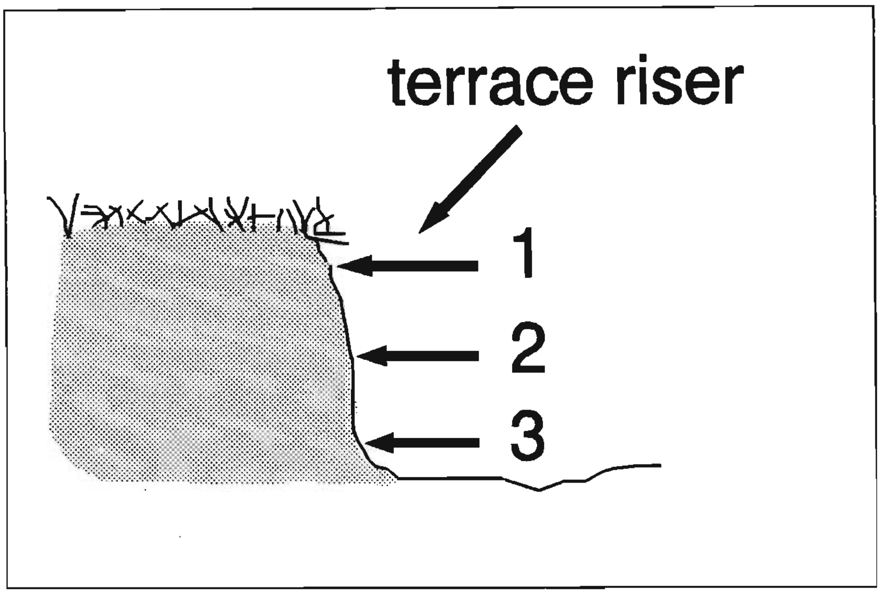


Figure 7.11 The upper, middle and lower terrace sampling positions.

Soil properties at the lower valley slope turf exfoliation site (mean values)					
		Sample number per riser position	Position on the terrace riser upper	middle	lower
Bulk density *(g. per cubic cm)*		20	0.74	0.86	0.80
Field moisture	March 1994	9	35.1%	45.7%	44.9%
(percentage by weight)	June 1994	9	29.6%	31.3%	38.8%
Percentage moisture loss	(March to June)		5.5%	14.4%	6.1%
Organic matter content *(percentage of dry weight)*		18	21.25%	18.83%	18.07%
Soil colour (dry)			2.5Y 4/3 olive brown	2.5Y 4/3 olive brown	2.5Y 5/3 light olive brown

Table 7.4 Mean soil characteristics for the various sampling positions on turf exfoliated risers.

Figure 7.12 a, b & c shows the particle size distribution on different riser positions, in which there is relatively little variation. Generally, the percentage gravel is low, but it does increase from upper to lower riser positions (0 to 7.5%). There is little variation in the distribution of sand and fine material on the different riser positions. Fine fraction particle size results display no significant variations between different riser positions (Figure 7.12 a, b & c). Most of the fine fraction consists of coarse silt (65.6 to 81.8%) (Figure 7.12 a, b & c). The percentage clay is very low (1.2 to 1.8%) throughout the terrace risers (Figure 7.12 a, b & c).

7.2.4 Turf (Terrace) Retreat

Turf retreat at five exfoliation sites, as explained in Section **3.4.2**, was measured from May 1993 to February 1995 (21 months). Of the 85 pins inserted, five were lost by February 1995. Results of the turf retreat are given in Table 7.5 and Figure 7.13 a, b, c, d & e.

The mean rate of turf retreat is greatest at the lower valley slope sites (mean = 16.0 mm/year) and lowest at the valley floor sites (mean values range from 3 to 6.8 mm/year) (Table 7.5 and Figure 7.13 b, c, d & e). However, some differences in the rates of turf retreat occur between turf exfoliation sites. Even at individual sites, there is disparity in the rate of turf retreat (c.f. Figure 7.13 a, b, c, d & e). Pérez (1992b) suggests that such irregular retreat is as a result of large volumes of plant and soil detachment over a short period, while at less disrupted positions retreat is considerably slower.

Preliminary findings indicate that the maximum rate of turf retreat at the valley floor and lower valley slope sites occurs during winter and spring months (Table 7.5 and Figure 7.13 b, c, d & e). However, at the valley bottom site, the rate of retreat during summer and autumn almost doubles that during winter and spring (Table 7.5 and Figure 7.13 a). At the valley bottom and valley floor sites, between 40 and 60% of measurements indicate no turf retreat, whereas at the valley slope site only 5.9% of measurements show no turf retreat (Table 7.5).

Turf retreat at sites along the valley bottom and valley floor is most pronounced on the downslope and downvalley sides of turf exfoliated depressions (Figure 7.13 a, b, c & d). In

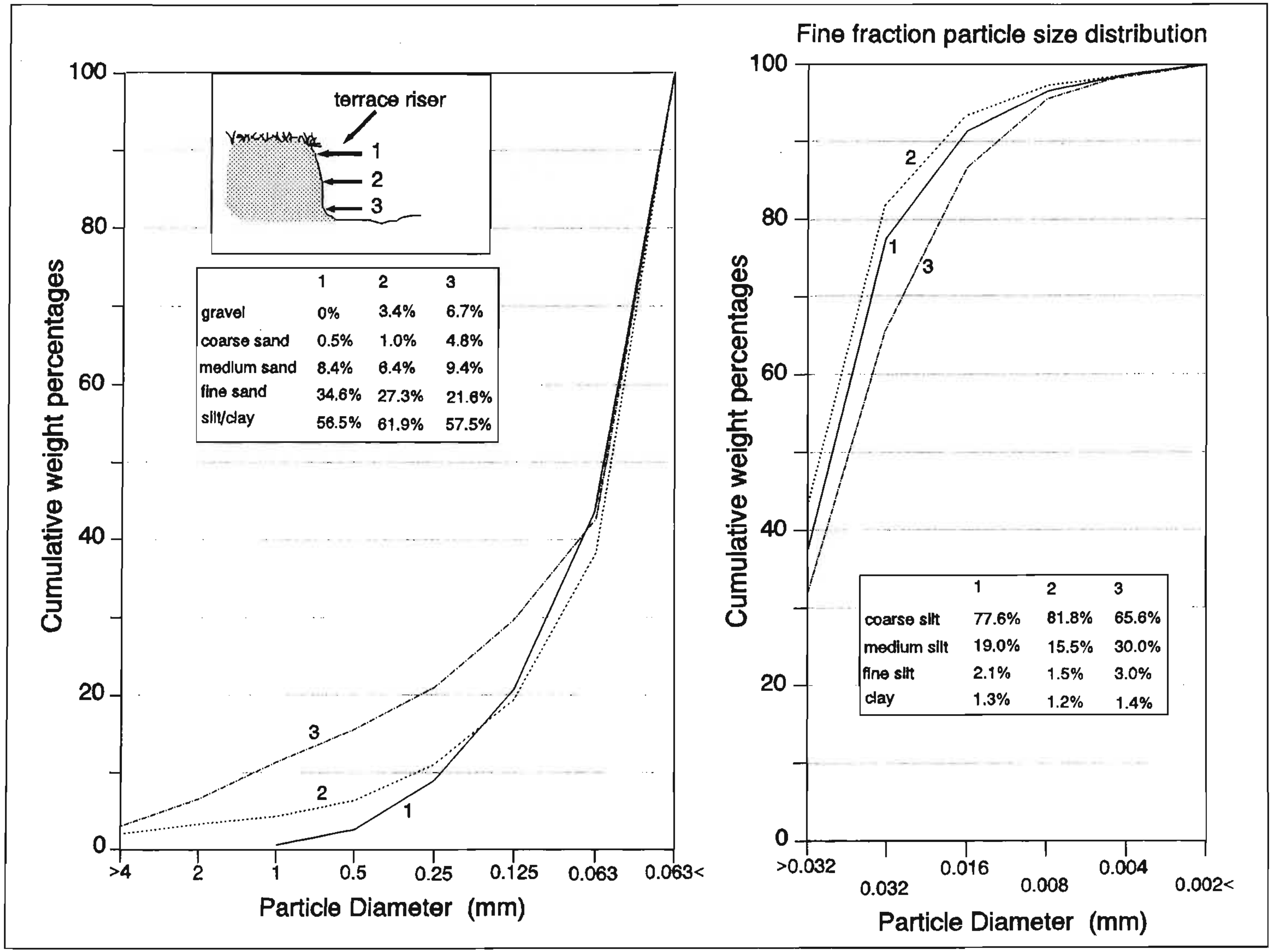


Figure 7.12 a Particle size distribution for the various turf exfoliated riser positions, riser 1 (n = 3 per riser position).

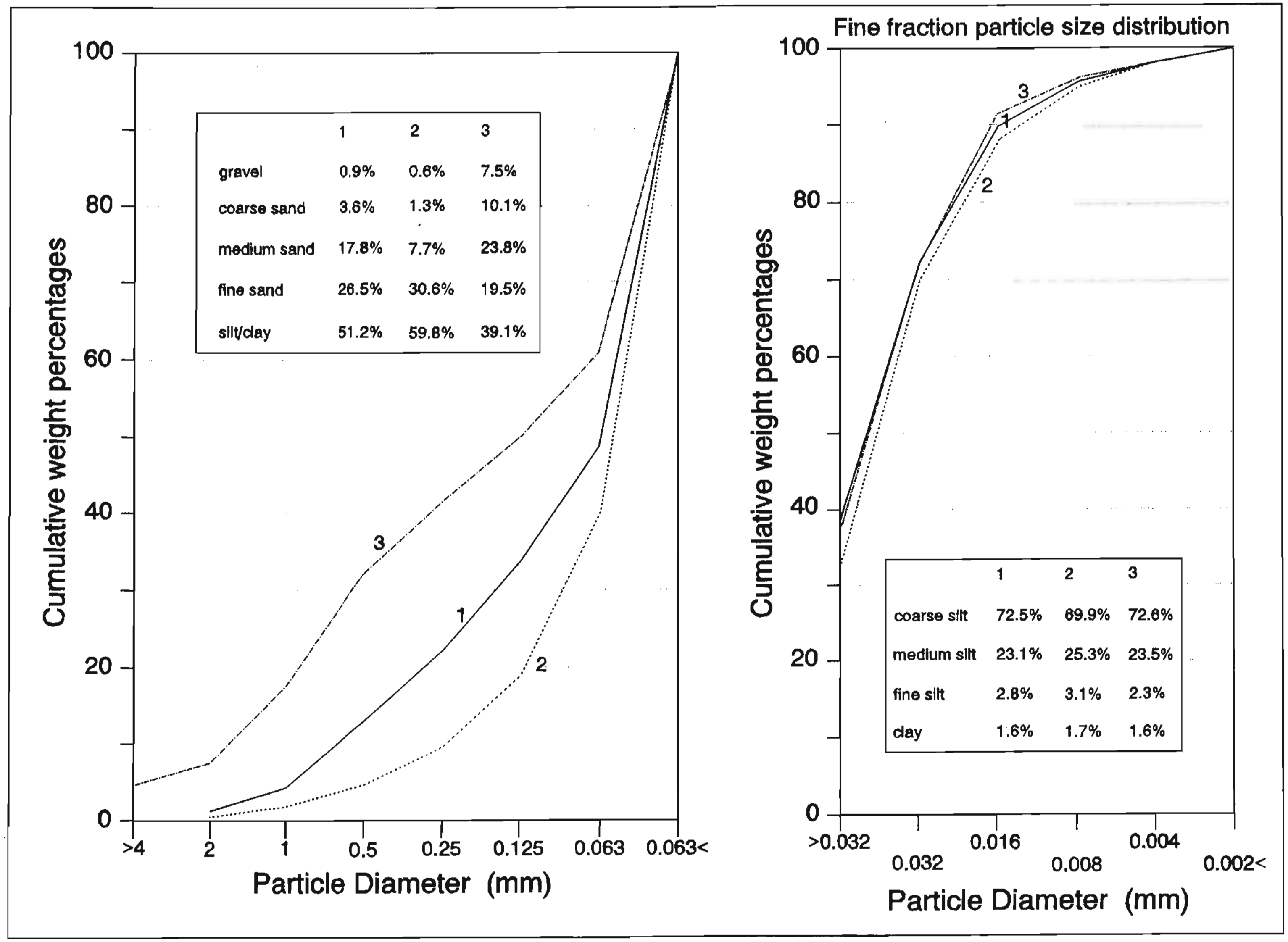


Figure 7.12 b Particle size distribution for the various turf exfoliated riser positions, riser 2 (n = 3 per riser position).

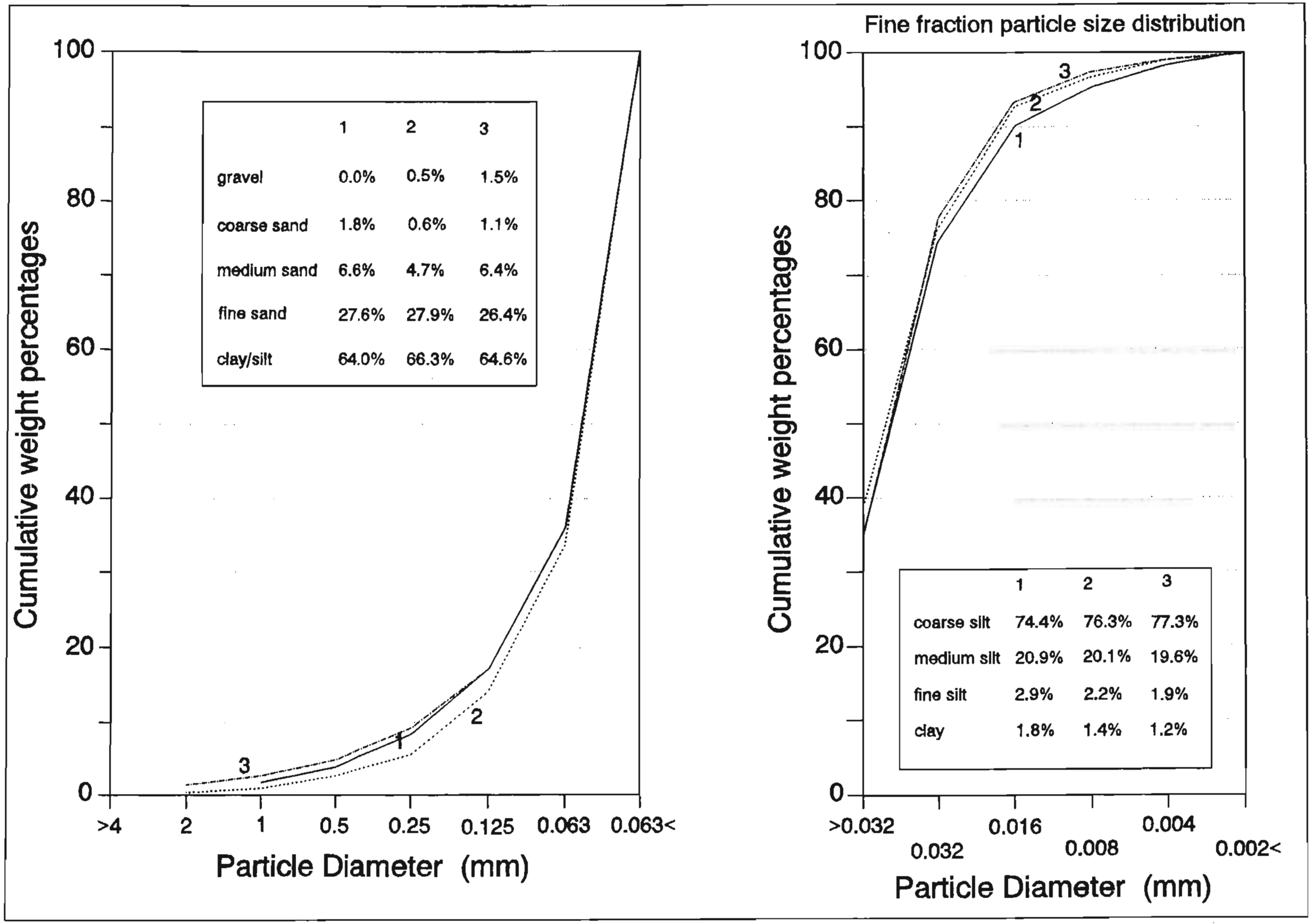


Figure 7.12 c Particle size distribution for the various turf exfoliated riser positions, riser 3 (n = 3 per riser position).

Locality	Number of pins	Period of record	Percentage pins lost	Percentage pins where no turf retreat	Mean turf retreat (mm)			Max. rate of turf retreat (mm/year)	Mean rate of turf retreat (mm/year)
					May'93-Dec'93	Dec'93-May'94	May'94-Feb'95		
Valley bottom									
1	13	21 months	0.0%	46.1%	5.5	10.4	5.1	38.8	11.8
Valley floor									
2	13	21 months	15.4%	54.0%	0.2	2.4	2.7	12.6	3.0
3	21	21 months	9.5%	42.8%	4.5	1.9	3.5	20.0	5.6
4	21	21 months	4.8%	61.9%	5.0	2.0	5.2	37.7	6.8
Lower valley slope									
5	17	21 months	0.0%	5.9%	11.6	5.1	11.3	30.8	16.0

Table 7.5 Measurements of turf retreat from five turf exfoliated sites between May 1993 and February 1995.

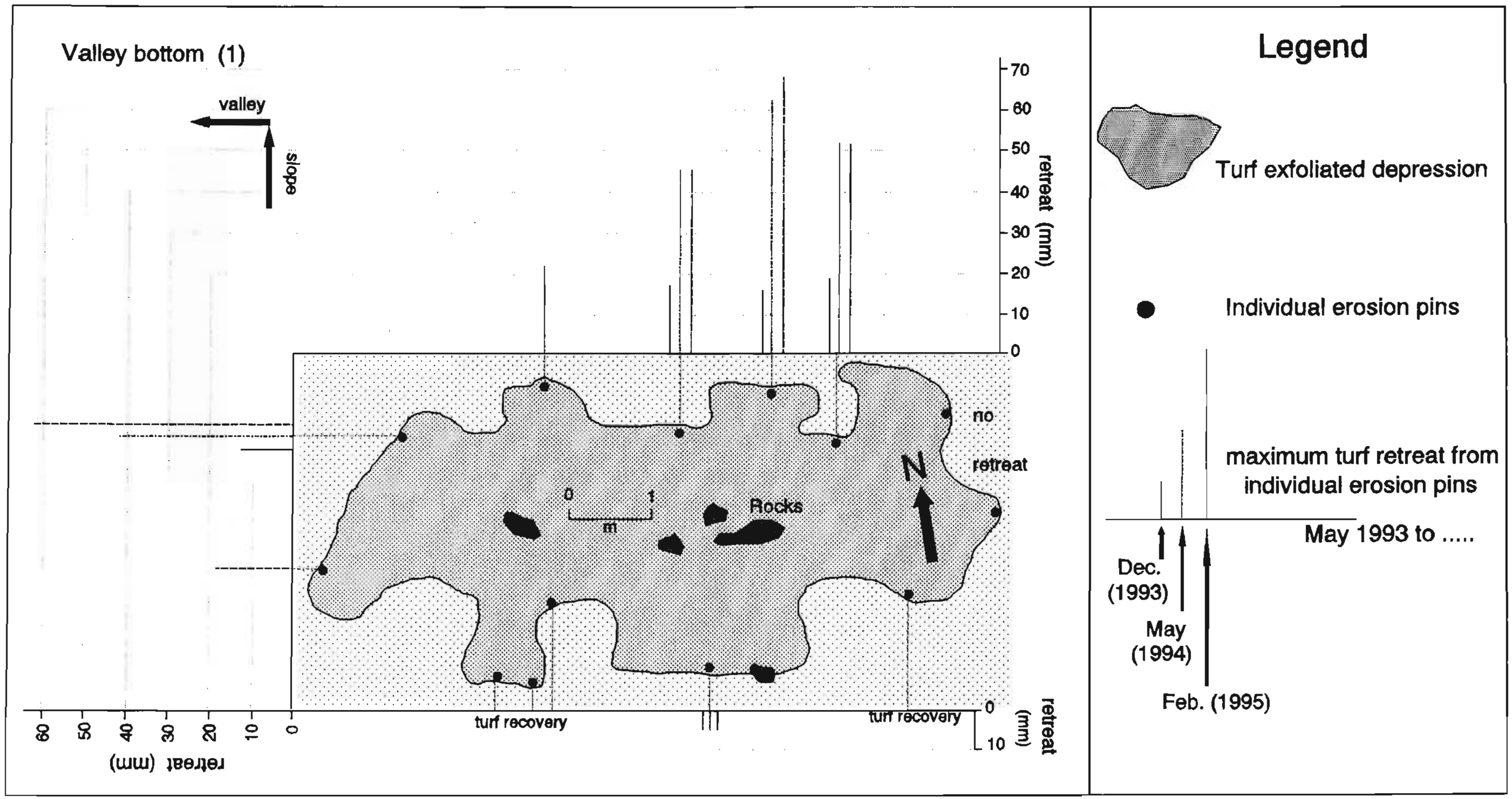


Figure 7.13 a Measurement of turf retreat at the valley bottom, Site 1: May 1993 to February 1995.

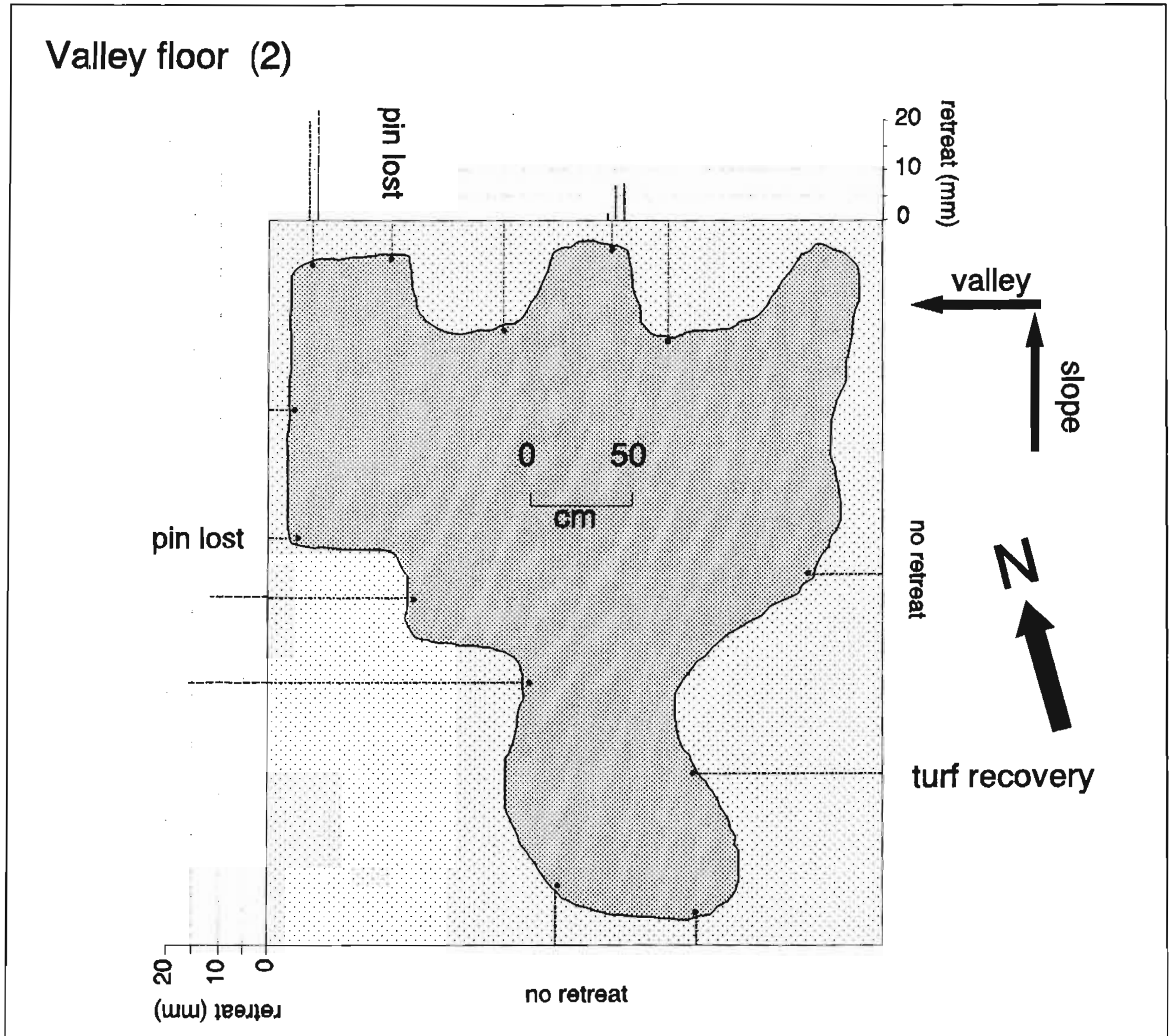


Figure 7.13 b Measurement of turf retreat at the valley bottom, Site 2: May 1993 to February 1995.

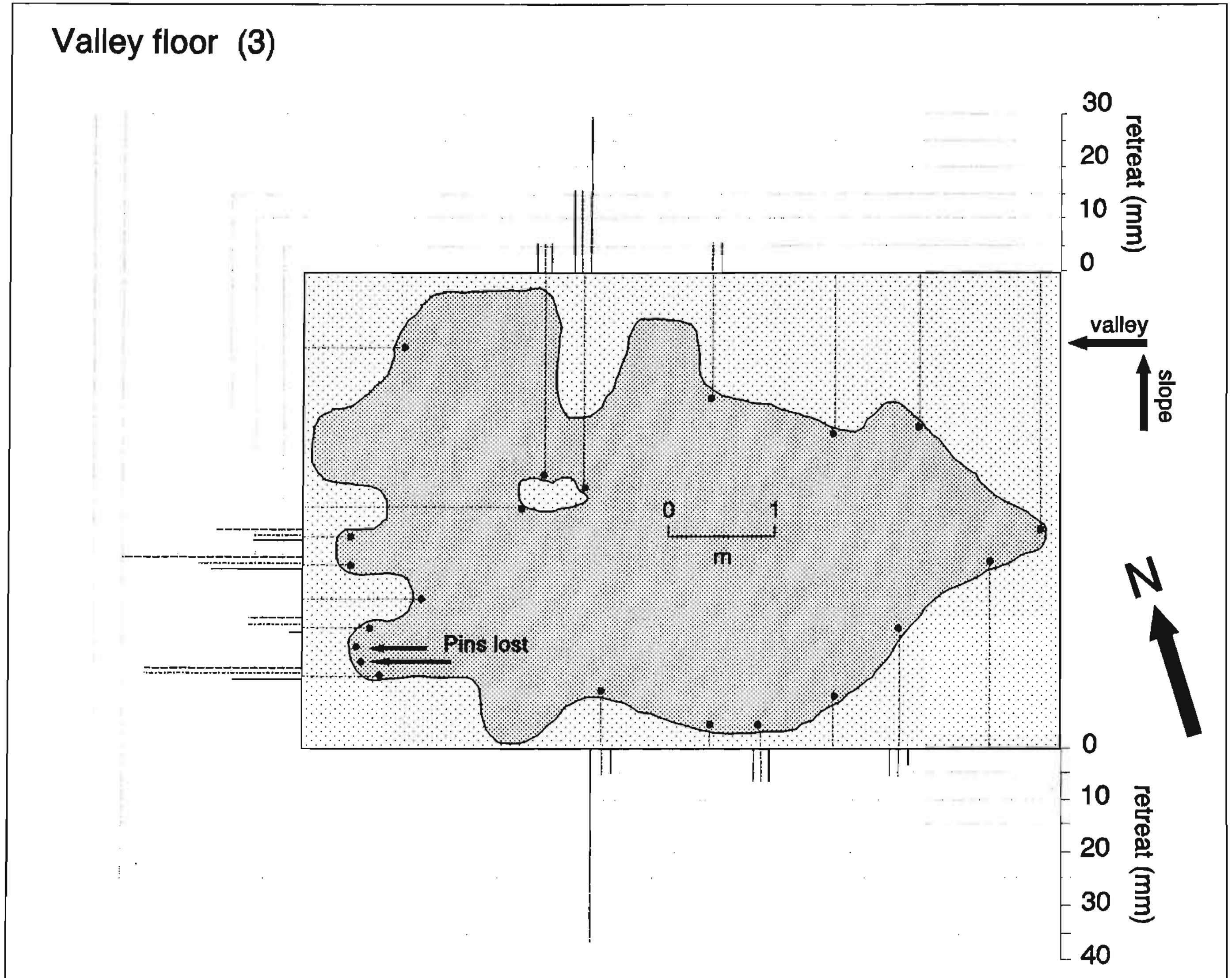


Figure 7.13 c Measurement of turf retreat at the valley bottom, Site 3: May 1993 to February 1995.

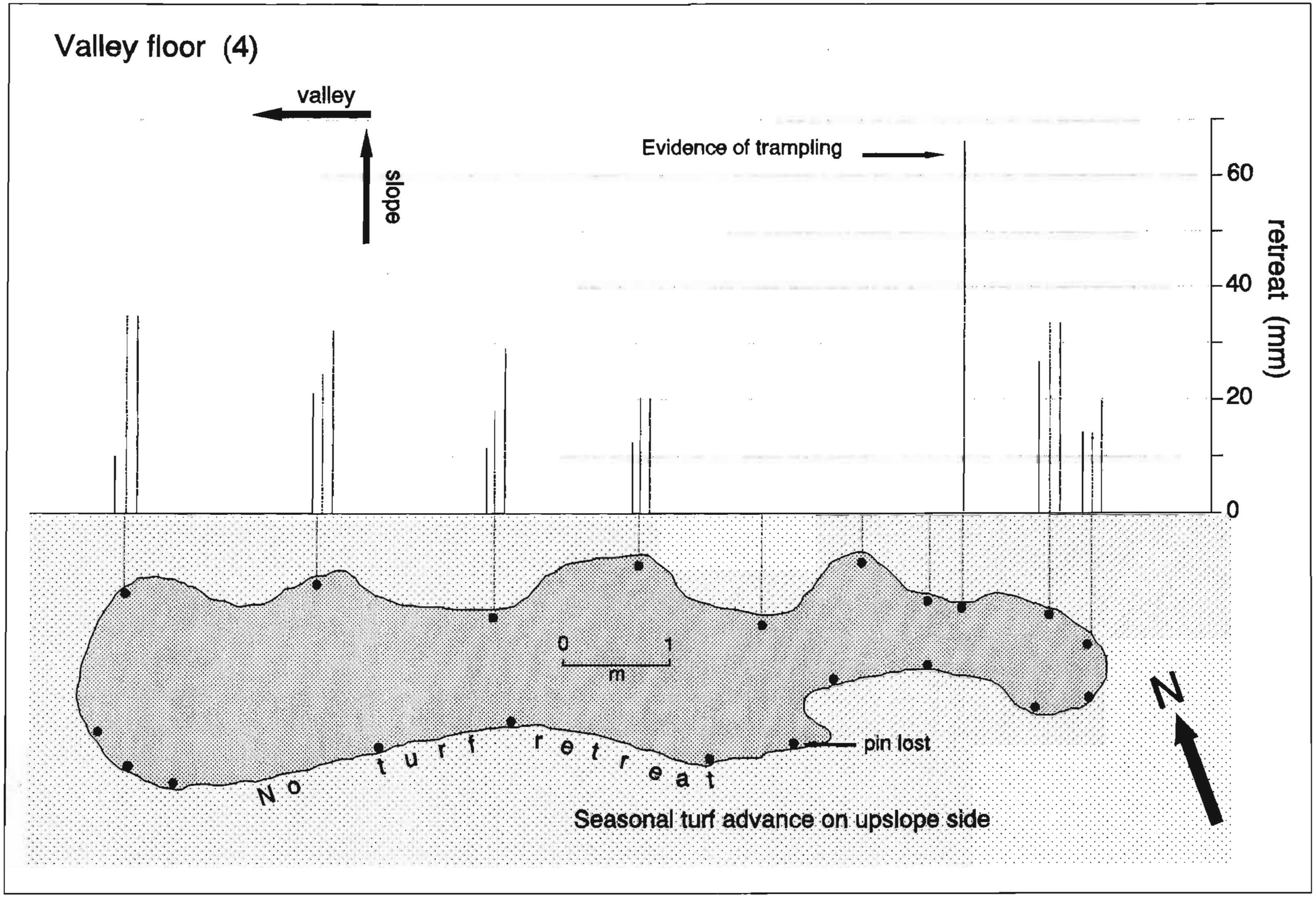


Figure 7.13 d Measurement of turf retreat at the valley bottom, Site 4: May 1993 to February 1995.

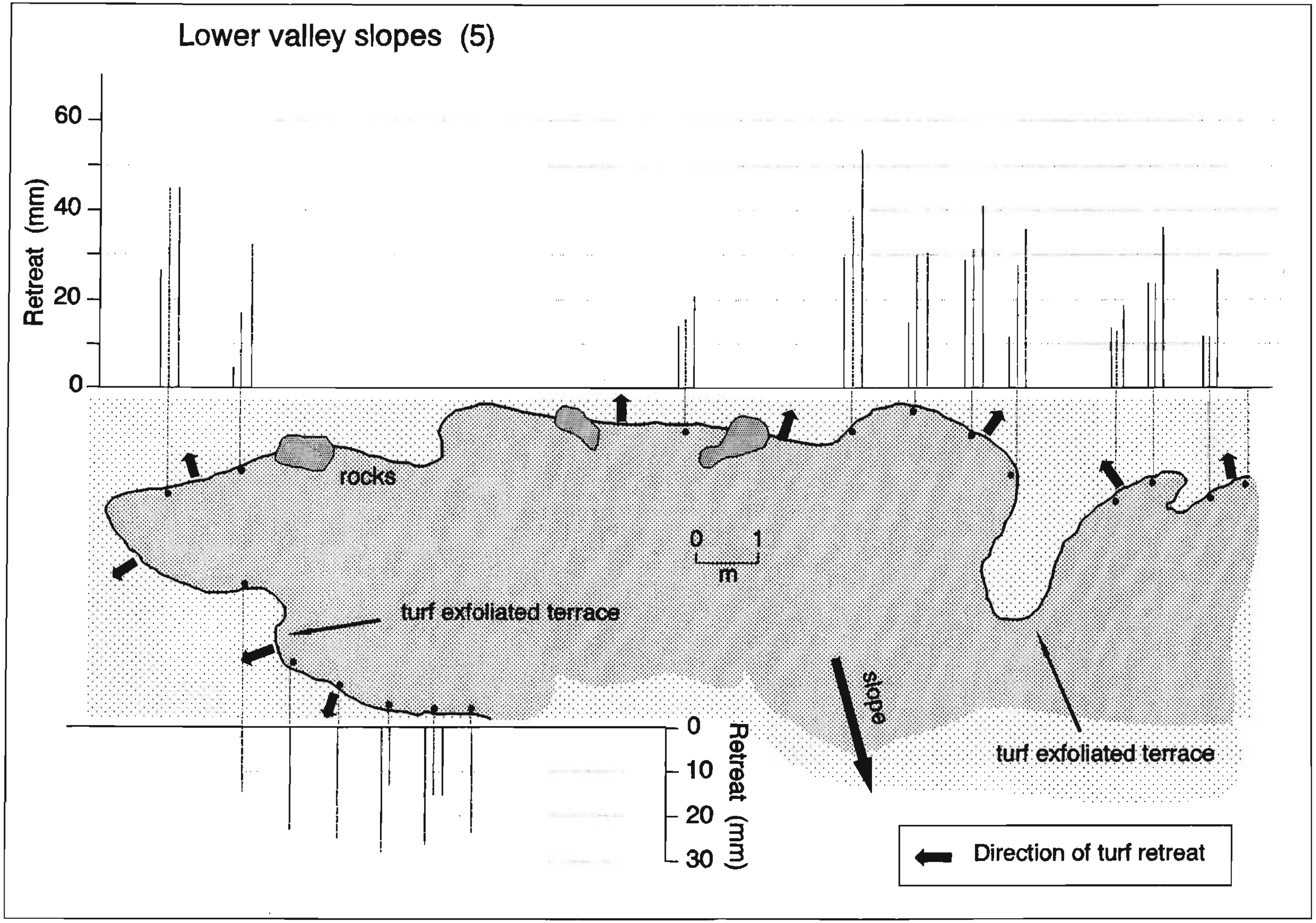


Figure 7.13 e Measurement of turf retreat at the valley bottom, Site 5: May 1993 to February 1995.

fact, very little recession was measured on upslope and upvalley sides. Instead, there are some instances where turf advanced towards depression centres from upslope sides (Figure 7.13 a & b). For instance, three pins on the upslope side of an incised valley bottom depression became overgrown by advancing turf. The findings in this section indicate that turf exfoliation is active at all of the study sites in the Mashai Valley.

7.2.5 Possible Processes Affecting Turf Exfoliation in the High Drakensberg: Discussion

Not only may there be several processes operating synergistically to initiate turf exfoliation, but the importance of different factors may be somewhat varied between sites. Needle ice activity is frequently regarded as a primary (and sometimes the primary) factor causing turf exfoliation (e.g. Troll, 1944, 1958, 1973; King, 1971; Hastenrath, 1973, 1978; Hastenrath and Wilkinson, 1973; Heine, 1977; Pérez, 1992b). To this end, needle ice development was observed at all three localities (Figure 7.1) in the Mashai Valley. However, temporally and spatially, the extent of needle ice activity between such sites is very often markedly different. Turf exfoliation sites along the valley bottom remain saturated, even towards the end of winter. Consequently, terrace risers at such sites usually become extensively frozen. Below, on the depression soils, needle ice activity occurs from late autumn to mid-spring (Figure 7.14). The needle ice lifts material from near the lower part of the riser, thereby contributing towards retreat at this point. The extensive freezing and wet conditions also reduce aeolian deflation at these sites. The better protected upper parts of terrace risers facilitate overhanging swards and strands of vegetation, which frequently reach the terrace treads (Figures 7.7 and 7.8). However, at the valley floor and lower valley slope sites, conditions are somewhat different. Here, owing to drier conditions than at the valley bottom sites, needle ice activity is usually restricted to late autumn and early winter. At such sites, needle ice activity primarily occurs on the tread and basal areas of risers, decreasing with height up the riser face (c.f. Gradwell, 1960; Troll, 1973; Pérez, 1992b). Needle ice activity loosens the soil particles and thus weakens the soil layer (Troll, 1944, 1958, 1973; King, 1971), making it easier for other soil erosion agents to remove soil particles from the exposed terrace risers. The soils within shallow pan-like depressions are regularly disrupted by needle ice during late autumn and early

Figure 7.14 Needle ice frequently occurs below terrace risers at valley bottom turf exfoliation sites.

winter. These soils eventually become increasingly desiccated, thin, crumbly and "puffy" (Figure 7.15), as described by Soons (1967) and Pérez (1987c, 1988).

Needle ice activity on terrace treads may also actively move collapsed clumps of turf and sediment away from the terrace riser, leaving behind coarser material such as cobbles and boulders. The coarse material which is strewn along retreating terrace risers (Figure 7.16) has been referred to as "rubble pavements" by such as Pérez (1992b).

Although aeolian deflation is sometimes considered an important component in the turf exfoliation process (e.g. Troll, 1958, 1973; Heine, 1977), this has not previously been suggested for the high Drakensberg. However, from this study, aeolian deflation does seem to be a primary cause for turf exfoliation in the high Drakensberg. It is suggested that deflation is most effective in dry, uncohesive soils (Spönemann, 1977). The high Drakensberg climate of wet summers and dry winters is favourable for such desiccation and subsequent deflation. Figure 7.17 indicates the mean monthly wind speed and precipitation data for Garden Castle (1850 m a.s.l.) in 1993. Although the recording station is about 1100 m below

Figure 7.15 During the winter months the pan-like depressions dry out and the soils become increasingly desiccated, thin, crumbly and "puffy".

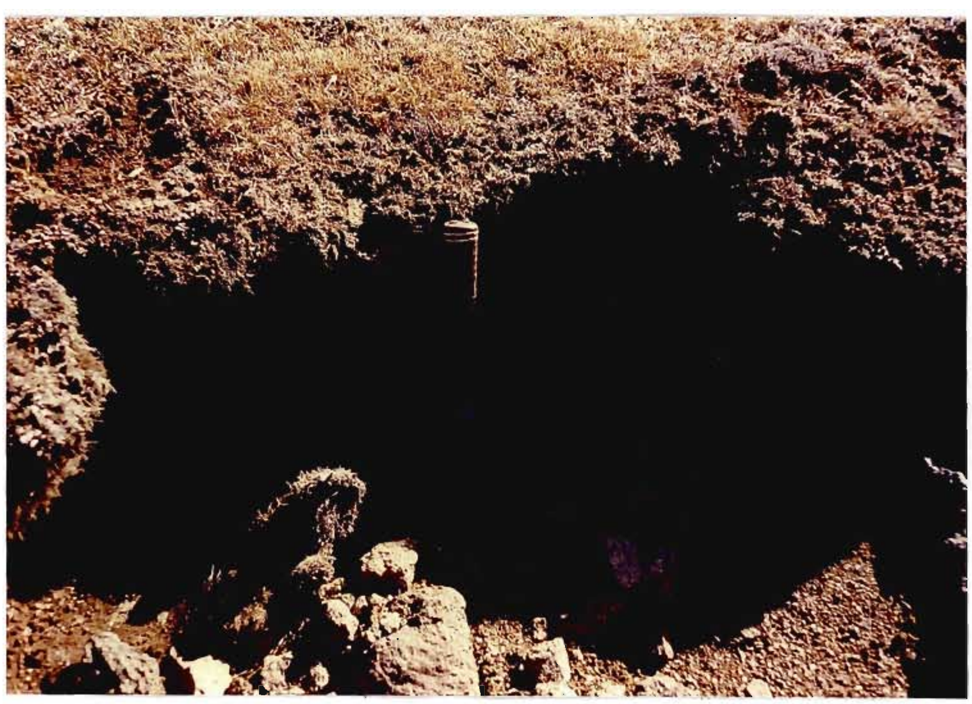


Figure 7.16 Turf, coarse material and rocks frequently accumulate along retreating terrace risers.

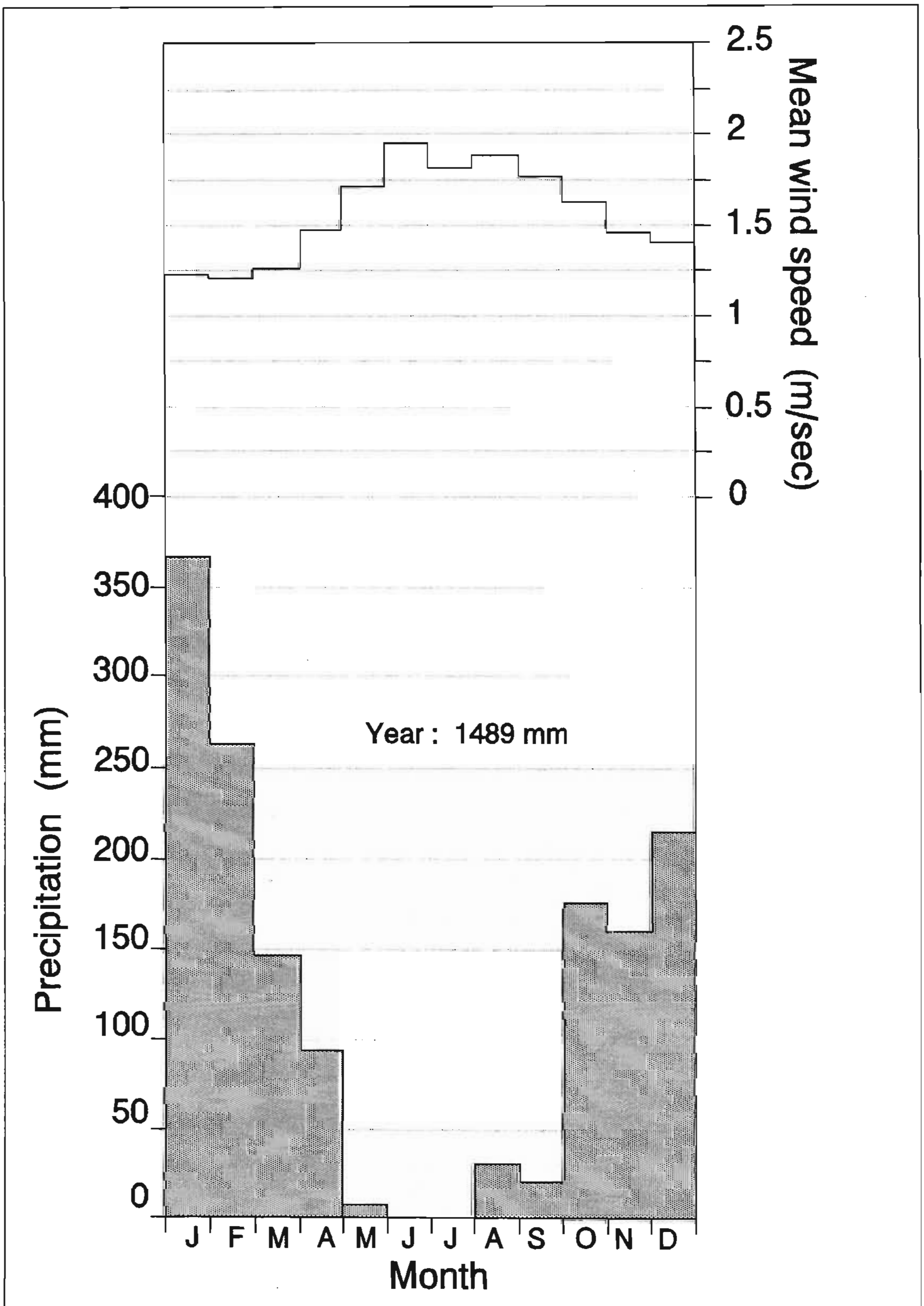


Figure 7.17 Mean monthly wind speed and precipitation values for Garden Castle (1850 m a.s.l.) during 1993.

and 8 km from the Mashai turf exfoliation sites, general trends are discernable and are applicable to the study sites. It can be seen in Figure 7.17 that the mean monthly wind speed is lowest during the wet summer months and highest during the cold, dry winter months. As a consequence, when the soils are at their driest, monthly wind speeds are at their highest. Although the average monthly wind speeds range between 1.81 and 1.95 $m.sec^{-1}$ during the winter months at Garden Castle recording station, the wind speeds on the summit are commonly much higher (pers. obs.). For instance, the author was regularly unable to walk upright, owing to the extremely powerful wind speeds. Exposed soils, initially disrupted by needle ice, are easily entrained by such strong and prolonged winds. Fine soils may be deflated from the shallow pan-like depressions or removed from exposed terrace risers. Evidence for such deflation processes are the depositional mounds and ridges along the base of terrace risers, as described in Section **7.2.3**. During times of intense wind action, gravel particles could be observed rolling down lower terrace risers to form such depositional micro-forms.

Rain-splash and rain-wash are also important contributing processes to turf exfoliation throughout the wet summer months. These processes disrupt and remove debris that accumulates at the base of terrace risers during the dry winter and early spring months. Rain-wash also actively erodes terrace treads, as is evidenced by small rivulets in Figure 7.18. The importance of water as an erosional agent at valley bottom turf exfoliation sites is possibly greater than needle ice and deflation processes. This may explain why the rate of terrace retreat at such sites is greater during the wet period than during the dry period. Water induced erosion and solution processes are likely to be the most active causes for turf exfoliation at valley bottom sites. However, the role of water as an erosional agent at valley floor and lower valley slope sites is considered to be less significant than at valley bottom sites. In contrast to the valley bottom sites, it would appear that the drier sites experience greatest retreat during the cold, dry period, and are less influenced by water erosion.

Basotho livestock (cattle, sheep, goats, horses) are regularly seen trampling the terrace edges from October to May. Such grazing pressure frequently results in edge failure (Fredriksson, 1972; Pérez, 1992b), with clumps of turf left to desiccate at terrace treads or within turf

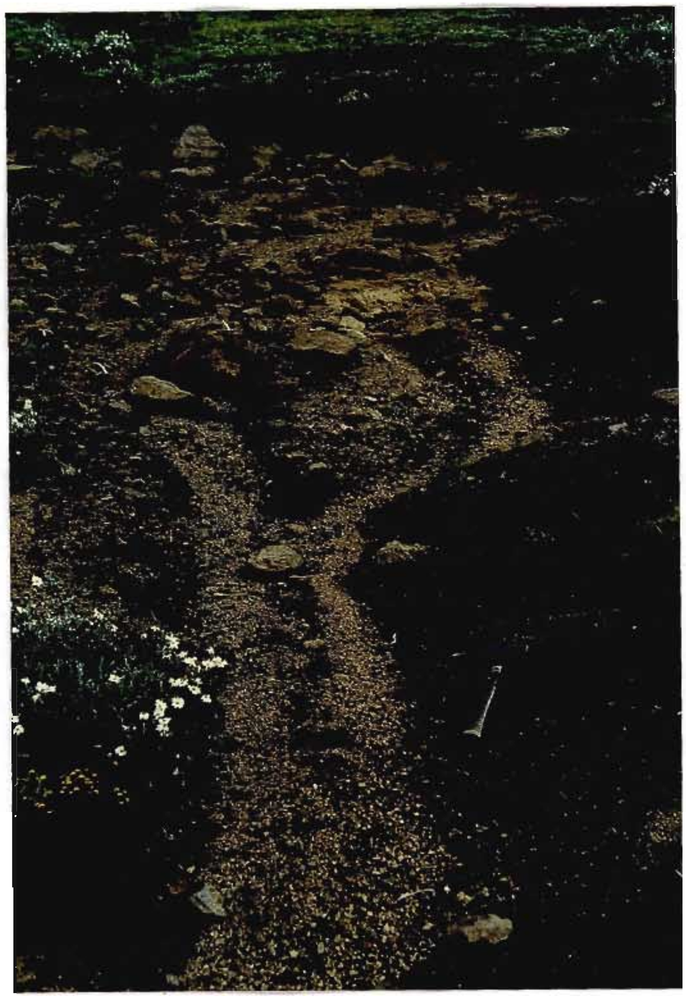


Figure 7.18 Rain-wash actively erodes terrace treads, forming small rivulets.

exfoliated depressions. The freshly exposed soil is usually less compact and is rapidly eroded by needle ice, wind and runoff processes. Trampling has also been shown to reduce infiltration, thereby decreasing the root zone soil moisture (Jacobsen, 1987). This may enhance surface runoff processes during summer. Further, trampling along terrace risers may not only cause vegetational changes (Jacobsen, 1987), but considerably deplete the protective turf cover. Animals trampling pan-like turf exfoliation depressions kick and carry out wet sediment on their hooves. As the depressions have no outlet, such zoogeomorphic activities during summer and aeolian deflation during winter and early spring, are the only mechanisms envisaged, which are able to remove sediment from within the depressions. Clearly, such grazing animals are important agents which help accelerate other erosive processes such as needle ice action, deflation and rain wash.

Several important processes, operating synergistically, are responsible for turf exfoliation in the high Drakensberg. However, it is evident that the strong seasonality from mild, wet summers to cold, dry winters has helped induce a cyclic pattern of dominating processes (Figure 7.19). Needle ice (cryogenic) activity becomes an important soil/turf disruptive agent from mid-autumn (April), but may continue to operate until mid-spring (October). Meanwhile, deflation processes become increasingly prevalent towards late winter (August) and early spring (September). Both deflation and cryogenic processes are usually abruptly halted during the onset of late spring (October/November) rains and rising air temperatures, when rainsplash and rain wash become the primary turf exfoliation processes. These rain-induced processes are enhanced by grazing pressure throughout summer and early autumn (March). When precipitation diminishes and air temperatures drop towards mid-autumn, the livestock are usually taken down from the alpine belt. Cryogenic processes once again become prevalent and the seasonal cycle of turf exfoliation repeats itself (Figure 7.19).

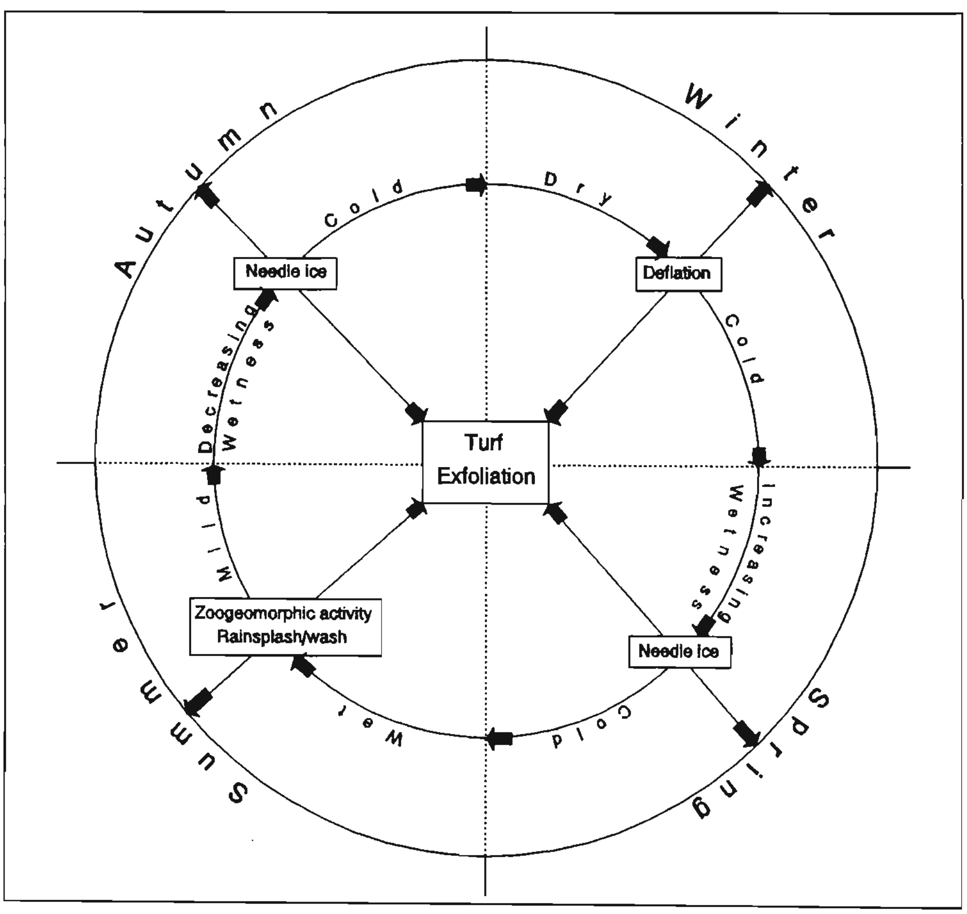


Figure 7.19 A model depicting the cyclic pattern of dominating processes contributing to turf exfoliation in the high Drakensberg.

7.3 NEEDLE ICE PROCESSES AND BANK EROSION

Stream or river bank erosion exerts an important control on the channel morphology and hydrology within an area (Leopold, 1973; Hooke, 1979; Church and Slaymaker, 1989; Lawler, 1993). The processes and rates of bank erosion along high Drakensberg streams are unknown, yet escalating problems of erosion and silt accumulation are imminent (Morris and Grab, 1997).

Although several workers have examined the role of frost action on sediment yield and bank erosion (e.g. Hay, 1943; Wolman, 1959; Soons, 1967; Hill, 1973; Imeson and Jungerius, 1974; Gardiner, 1983; Lawler, 1985, 1986, 1987, 1993), Lawler (1993) expressed concern that too many workers have ignored the impact of freeze-thaw activity when assessing bank erosion. Further, it has been shown that there may be a high statistical correlation between frost frequency and bank erosion rates (e.g. Hill, 1973; Gardiner, 1983; Lawler, 1986, 1993).

This section attempts to determine the significance of cryogenic processes as contributors to stream bank erosion along high Drakensberg streams. This assessment of stream bank sediment mobilization must necessarily be regarded, in many respects, as exploratory and rather to serve as a pilot study for future projects. As needle ice is the most commonly observed form of frozen ground in the high Drakensberg, and a contributing factor to stream bank sediment mobilization, its occurrence is examined in some detail.

7.3.1 Needle Ice

Washburn (1979, p92) defines needle ice as the "accumulation of slender, bristle-like crystals (needles) practically at, or immediately beneath, the surface of the ground". Despite ongoing interest in needle ice processes, Lawler (1988a) has identified several research gaps and has made a call for further research on the environmental controls, distribution and effects of needle ice. Lawler (1988b) has listed no fewer than 43 terms describing needle ice, of which 21 have been used in the English language. Needle ice has a global distribution but is particularly well represented in high altitude mountain ranges such as the Rockies, Andes and

East African mountains (Lawler, 1988a). Needle ice lengths for a single nights growth commonly vary from 10 to 50 mm (Lawler, 1988a) but exceptional lengths of up to 100 mm have been reported by Schramm (1958) and Mackay and Mathews (1974b). Multi-layered needle ice may develop during the night when there is moisture stress (Soons and Greenland, 1970), or alternatively, it may be the result of several days growth. Such multi-layered needle ice has been referred to as "polycyclic" needle ice (Outcalt, 1971b) and an apparent maximum recorded length of 400 mm has been reported by Krumme (1935).

Past needle ice studies have focussed attention on:

1. The environmental controls on needle ice, such as soil moisture and textur (e.g. Meentemeyer and Zippin, 1981).

2. The growth of needle ice in the natural environment (e.g. Brink *et al.*, 1967; Soons and Greenland, 1970).

3. Needle ice as a soil erosion agent or vegetation disruptive agent (e.g. Haasis, 1923; Hay, 1936; Soons, 1967; Brink *et al.*, 1967; Brockie, 1968; Lawler, 1986, 1987).

4. Laboratory studies and simulation (e.g. Outcalt, 1969; Higashi and Corte, 1971).

5. The role of needle ice in patterned ground and other miniature landform development (e.g. Fahey, 1973; Beaty, 1974; Mackay and Mathews, 1974a, 1974b).

6. Bibliographic reviews (Lawler, 1988a, 1988b).

As noted above, needle ice is a common soil transportation and vegetation disruptive agent in many cold environments, yet, the environmental impact of needle ice still appears to be little understood. Despite the frequent occurrence of needle ice in the high Drakensberg, no needle ice studies have been forthcoming from this region. The Drakensberg and Lesotho mountains offer a natural setting in which some aspects of the environmental controls on needle ice development can be tested. To this end, the needle ice growth and ablation cycles were examined in the Mashai Valley on three separate occasions.

7.3.2 Needle Ice Growth and Ablation

The occurrence of needle ice depends on the correct combination of soil and microlimatic variables such as soil physical properties, soil moisture, freezing duration and cooling rate (e.g. Lawler, 1988a). Various environmental controls affecting needle ice development and growth have been studied for at least three decades (e.g. Outcalt, 1970, 1971a, 1971b; Soons and Greenland, 1970; Meetemeyer and Zippin, 1981; Lawler, 1988a, 1993). However, it appears that the defining of rigid environmental controls has not been successful. For instance, there is still much uncertainty regarding the climatic controls on the growth and ablation cycles of needle ice. It has been suggested that for needle ice to develop, the air temperature should commonly be -0.7°C or less (Lawler, 1993), while the equilibrium surface temperature should be at least as low as -2°C (Outcalt, 1971b). This being the case, then needle ice could have, given adequate moisture, developed in the upper Mashai Valley (2950 m a.s.l.) on 127 nights from April to September of 1993. Needle ice development along stream banks was observed during most nights whilst visiting the Mashai site from May to September (during the years of 1993 to 1996).

Although ground frost heave activity has been measured in the field (e.g. Fahey, 1973, 1974; Smith, 1987b) and simulated in laboratories (e.g. Coutard *et al.*, 1988), these studies have not examined the role of needle ice growth/ablation. Ellenberg (1974) did examine needle ice growth/ablation on two consecutive days and, based on this, produced a 4-phase needle ice sequence of growth-stagnation-meltout-ice free period. Eight such needle ice events were investigated in the Mashai Valley during 1995/6, with the specific aim to examine temperature as an environmental control.

7.3.2.1 Observations from the high Drakensberg

Needle ice development in the high Drakensberg is primarily controlled by the pronounced diurnal temperature fluctuations from April to September. During clear and calm conditions, needle ice would normally develop within one or two hours after sunset when air and ground temperatures rapidly drop to below 0°C. On 50% of occurrences, the air temperature (@ +

1.5 m) had dropped to at least -3°C before needle ice developed. However, on three occasions needle ice developed when the air temperature was above -0.7°C (the minimum air temperature suggested by Lawler, 1993). On the 4th June 1996, needle ice developed in the Mashai Valley when the air temperature was +1.5°C and the surface temperature -0.3°C (@ + 2 cm). Such findings confirm the great difficulty of placing strict environmental constraints on needle ice development and imply that needle ice may develop even when air temperatures are above freezing. The existing near-surface ground thermal regime is an important determining factor for needle ice development. For instance, when the near-surface ground temperature was only 1.8°C (@ - 1 cm) at 16h00 on the 4th June 1996 (Figure 7.20 e), the air temperature was 6.4°C. Between 16h00 and 17h00, the air temperature dropped by 2°C while the surface temperature dropped by 3.8°C, consequently permitting the ground surface temperature to drop 2.4°C. Needle ice was thus able to develop owing to the ground thermal regime being well below 0°C, despite a positive air temperature. When the initial near-surface ground temperature is reasonably high (3 to 4°C) (Figure 7.20 a), it delays needle ice development until the lower air temperatures or increased heat radiation sufficiently reduce ground temperatures. Needle ice develops most frequently when the near-surface ground temperature approaches -1.0°C but was observed to develop when the ground temperature was as high as -0.4°C (Figure 7.20 b).

Lawler (1993) devised a slope coefficient of an equation which suggests an average needle ice development rate of almost 1.0 mm h^{-1}, with a total range from 0.17 to 1.71 mm h^{-1}. An extreme rate of 8 mm h^{-1} has been reported by Soons and Greenland (1970). Field observations at the Mashai Valley site recorded needle ice growth rates considerably higher than those reported from other regions. Average growth rates during the growth phase varied from 3.6 mm h^{-1} to 5.0 mm h^{-1}, with a maximum of 16 mm h^{-1} recorded between 22h00 and 23h00 on the 27th May 1995. The rate of needle ice development during the growth phase was tested for correlation against changes in air temperature (@ + 1.5 m), surface temperature (@ + 2 cm) and near-surface ground temperature (@ - 1 cm). The results (Figure 21 a-c) show no correlation between rates of temperature change and rates of needle ice growth. Therefore, it would appear that the rate of growth is more dependent on factors such as moisture availability, the development of massive ice at the surface, and vertical fluxes owing

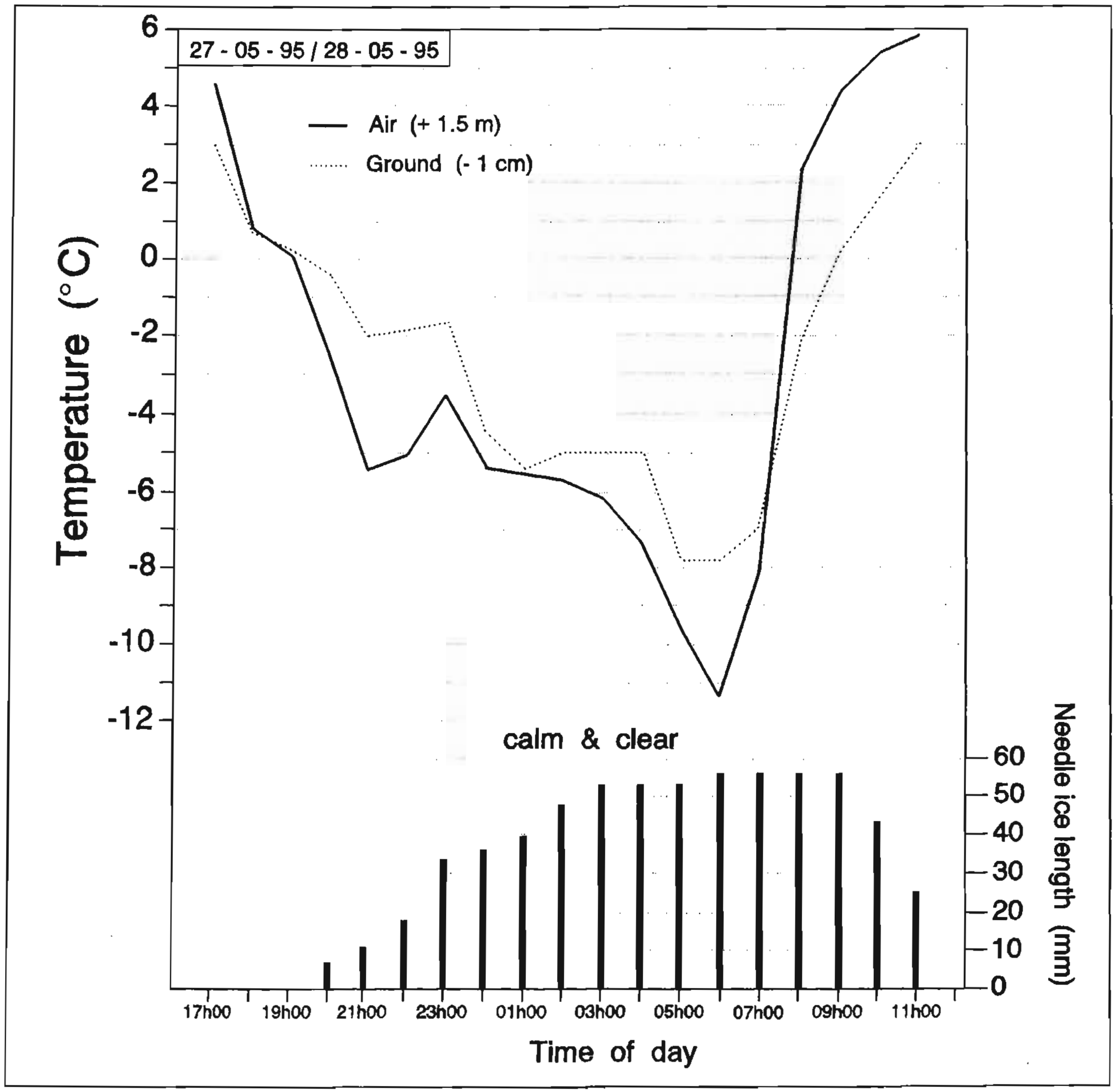


Figure 7.20 a Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (27-05-1995 to 28-05-1995).

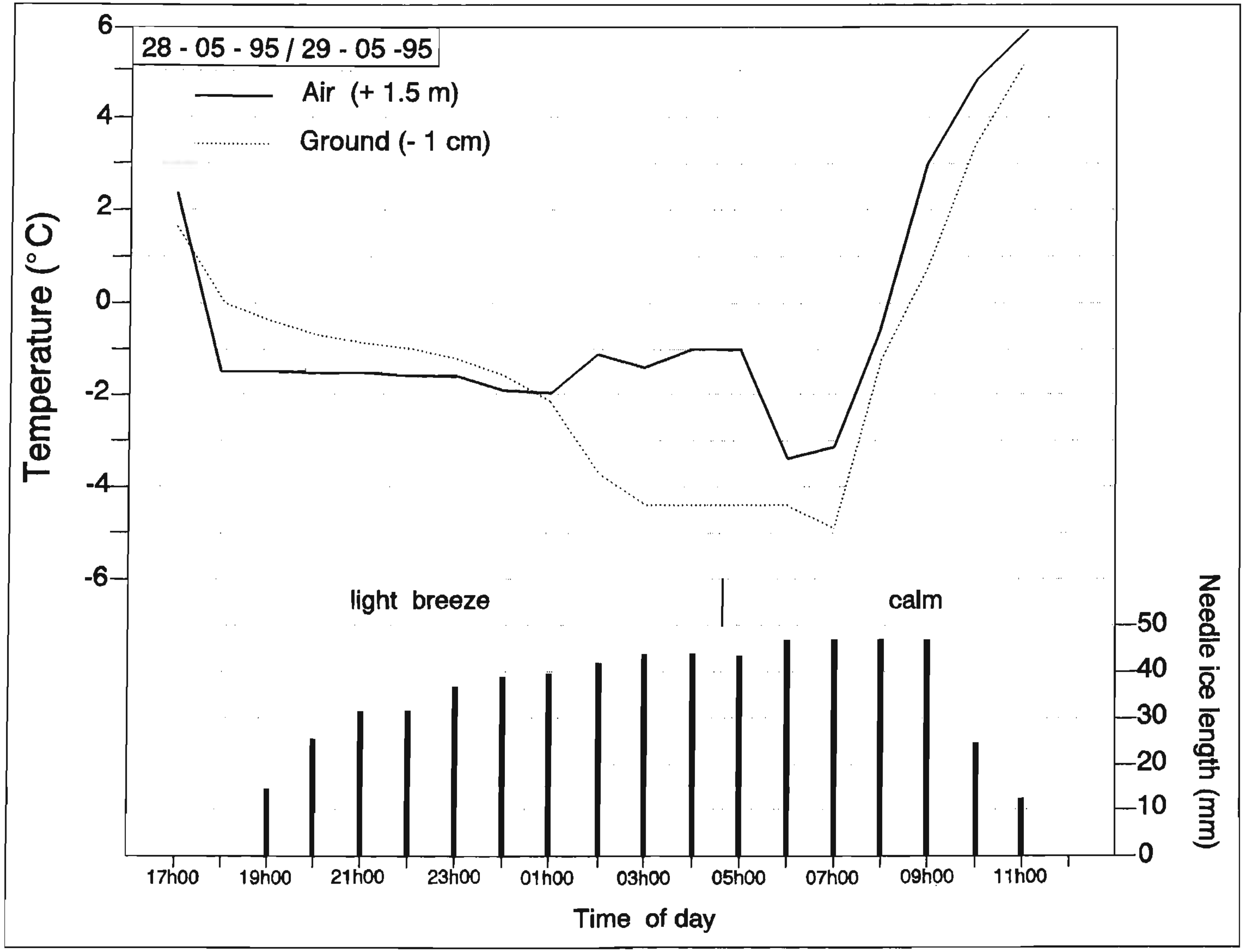


Figure 7.20 b Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (28-05-1995 to 29-05-1995).

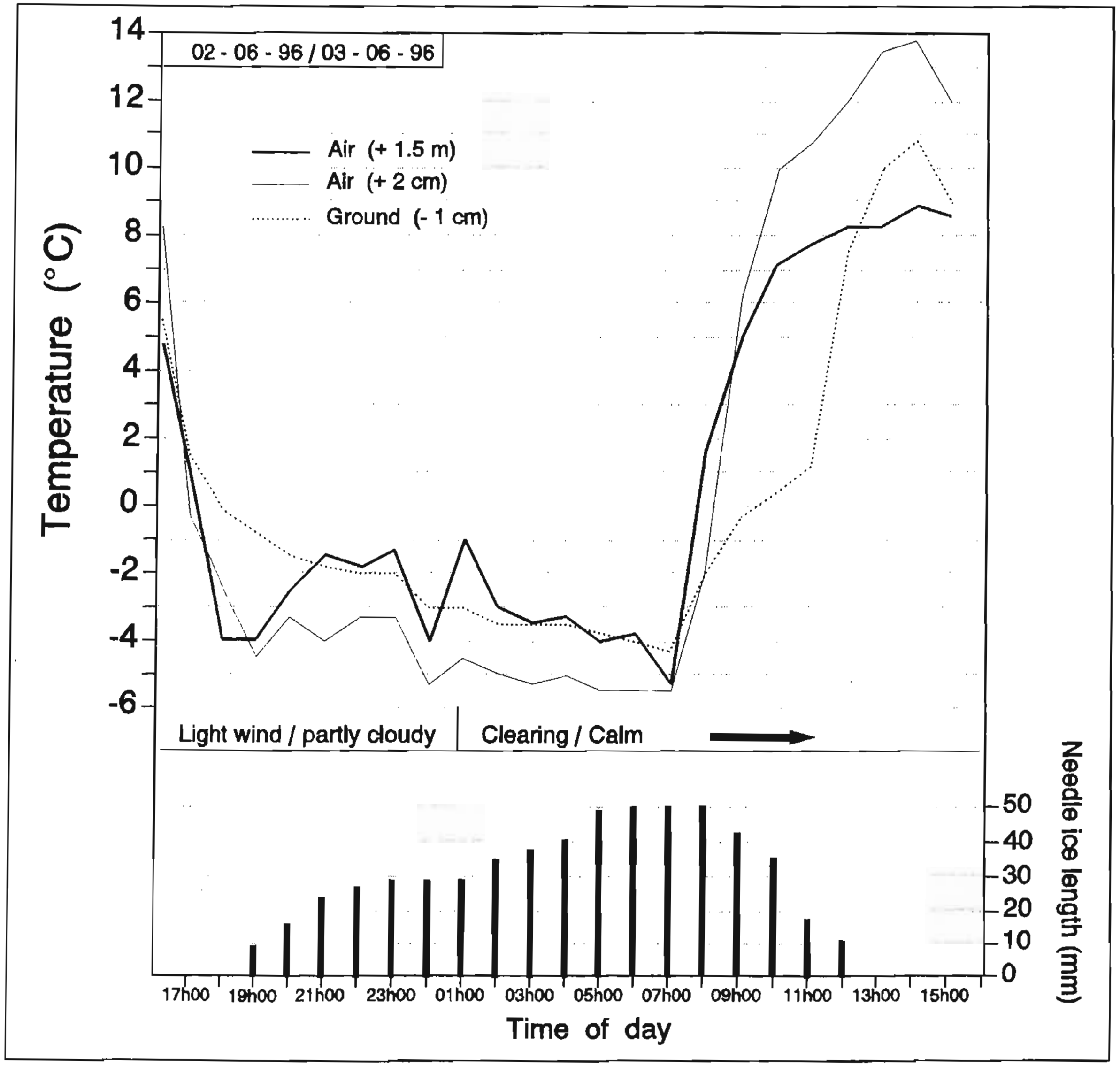


Figure 7.20 c Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (02-06-1996 to 03-06-1996).

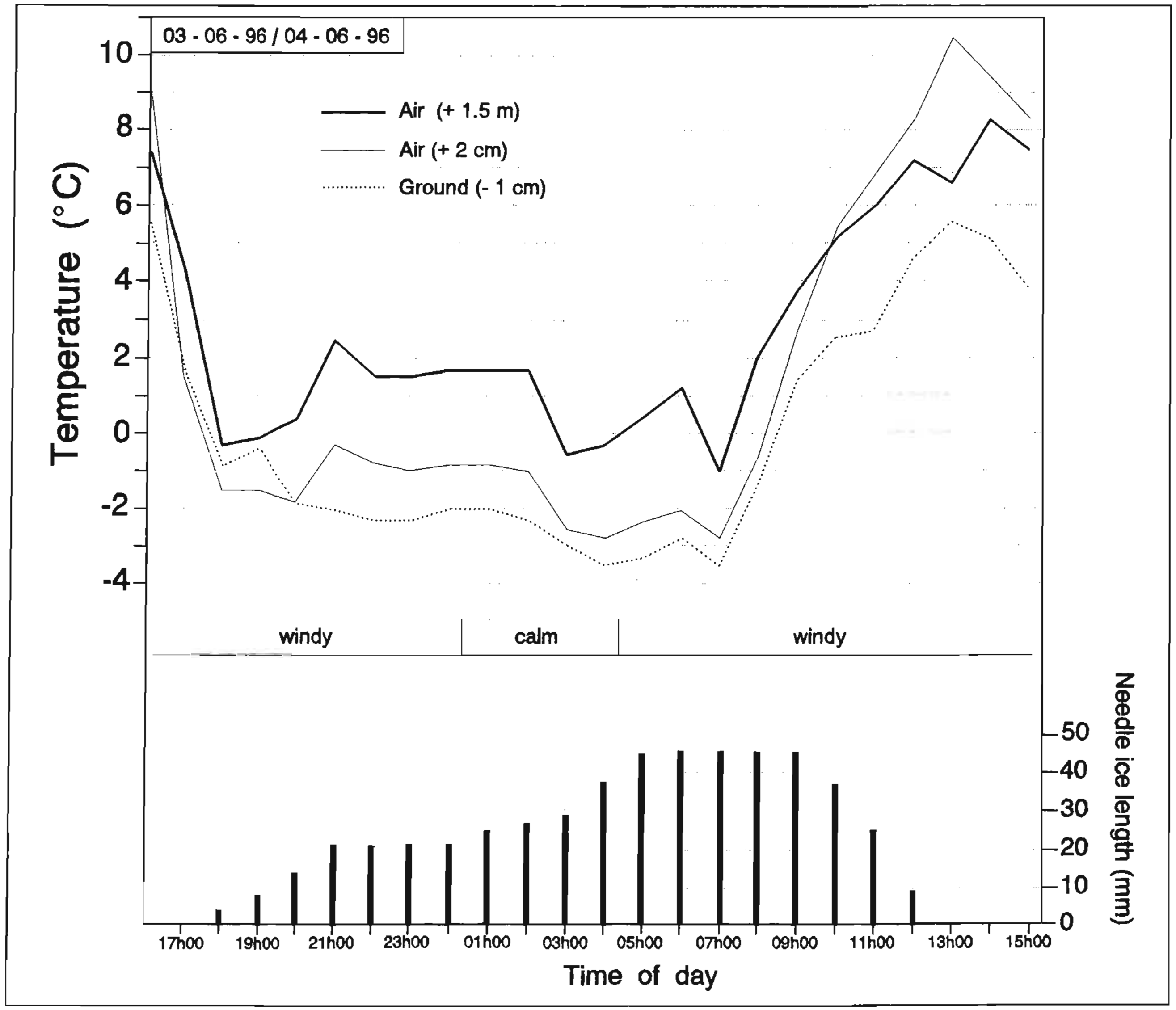


Figure 7.20 d Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (03-06-1996 to 04-06-1996).

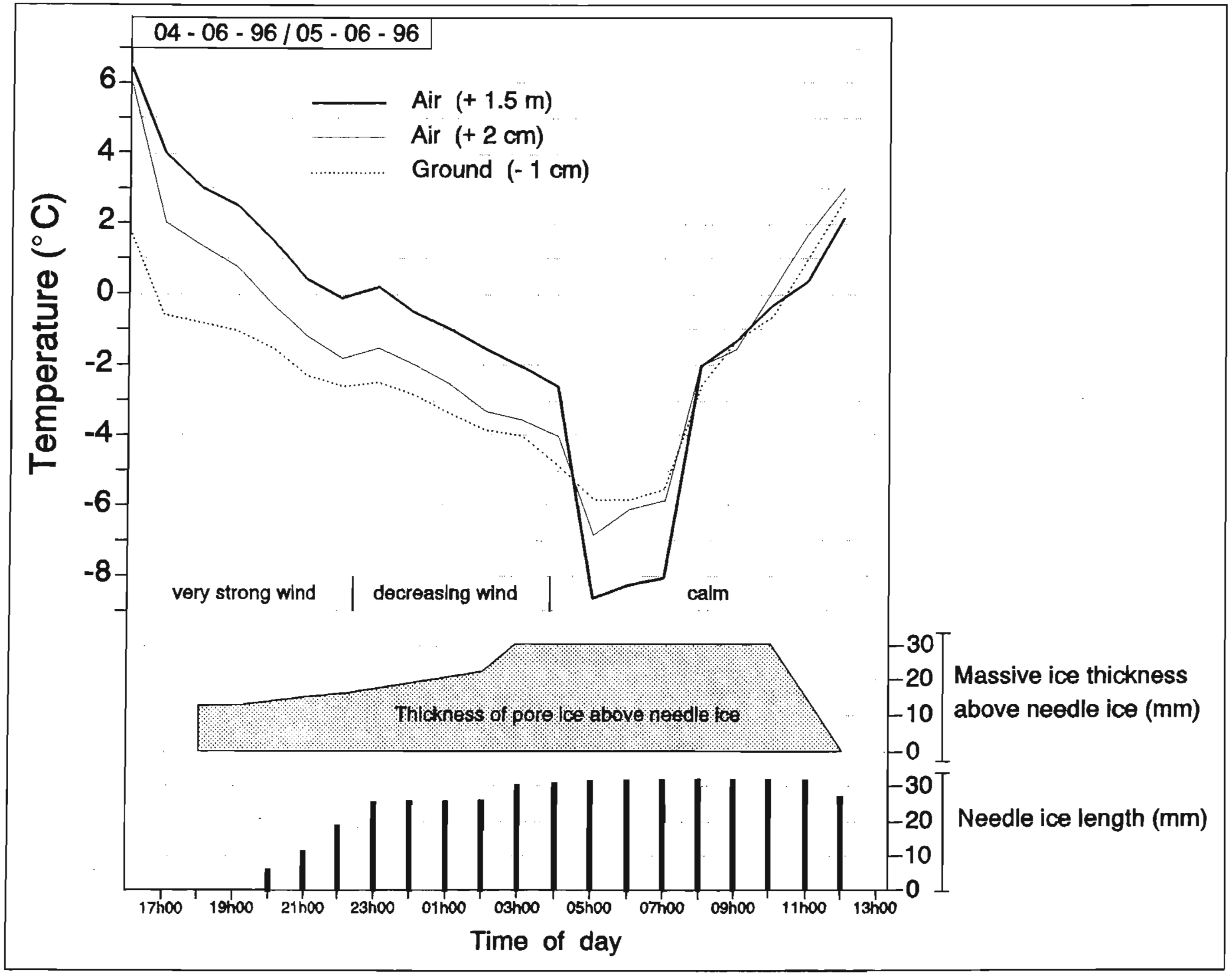


Figure 7.20 e Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (04-06-1996 to 05-06-1996).

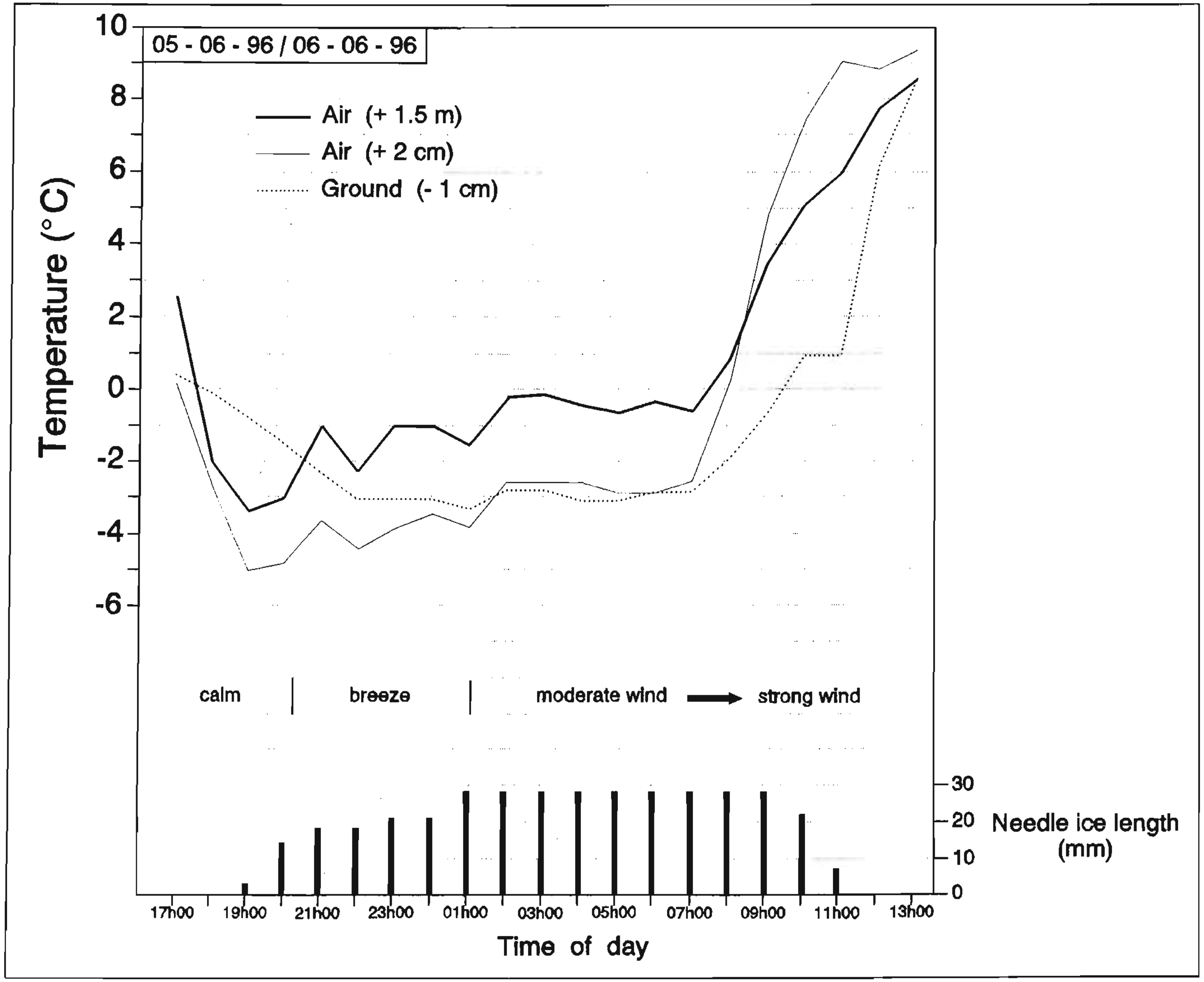


Figure 7.20 f Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (05-06-1996 to 06-06-1996).

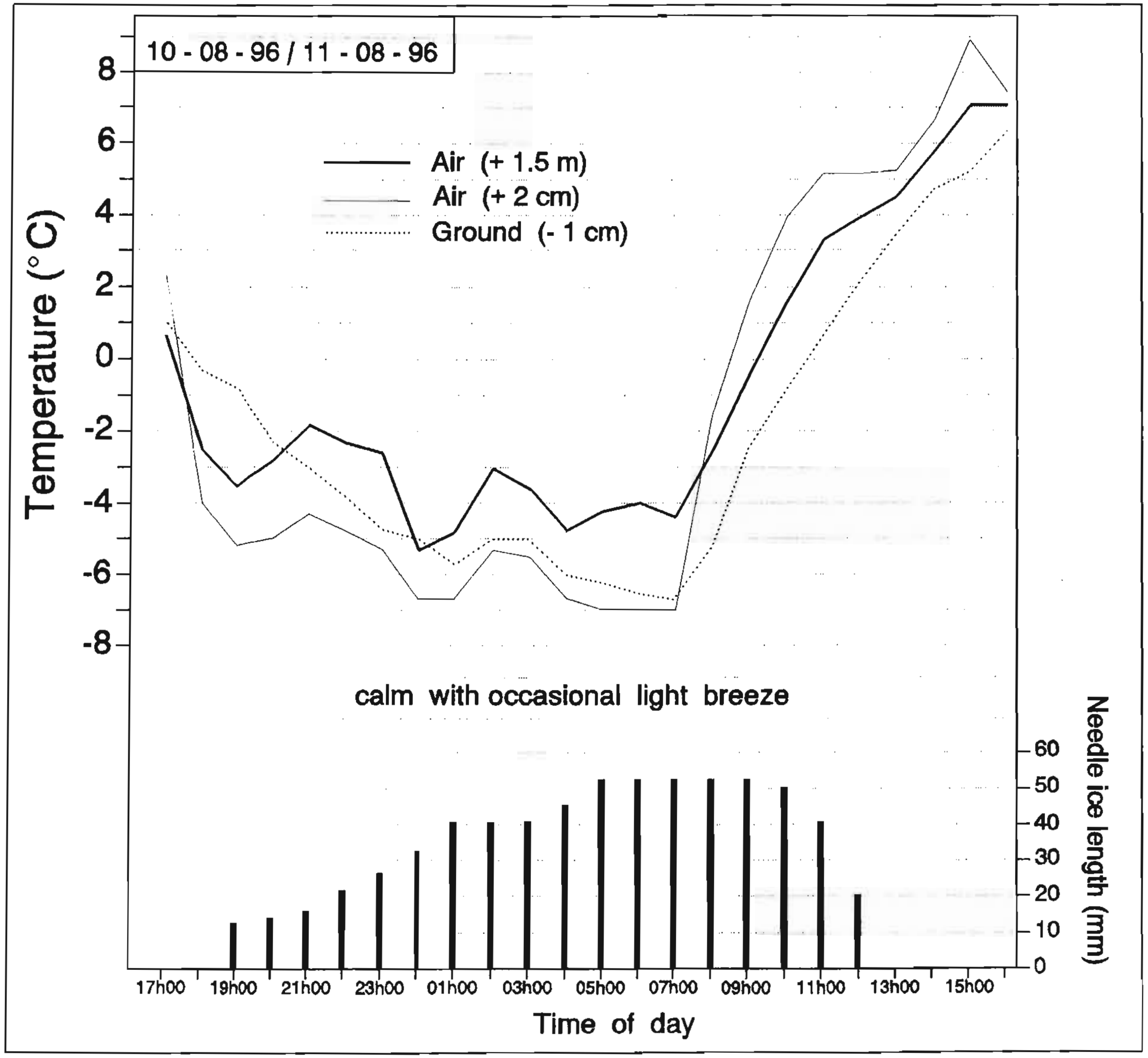


Figure 7.20 g Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (10-08-1996 to 11-08-1996).

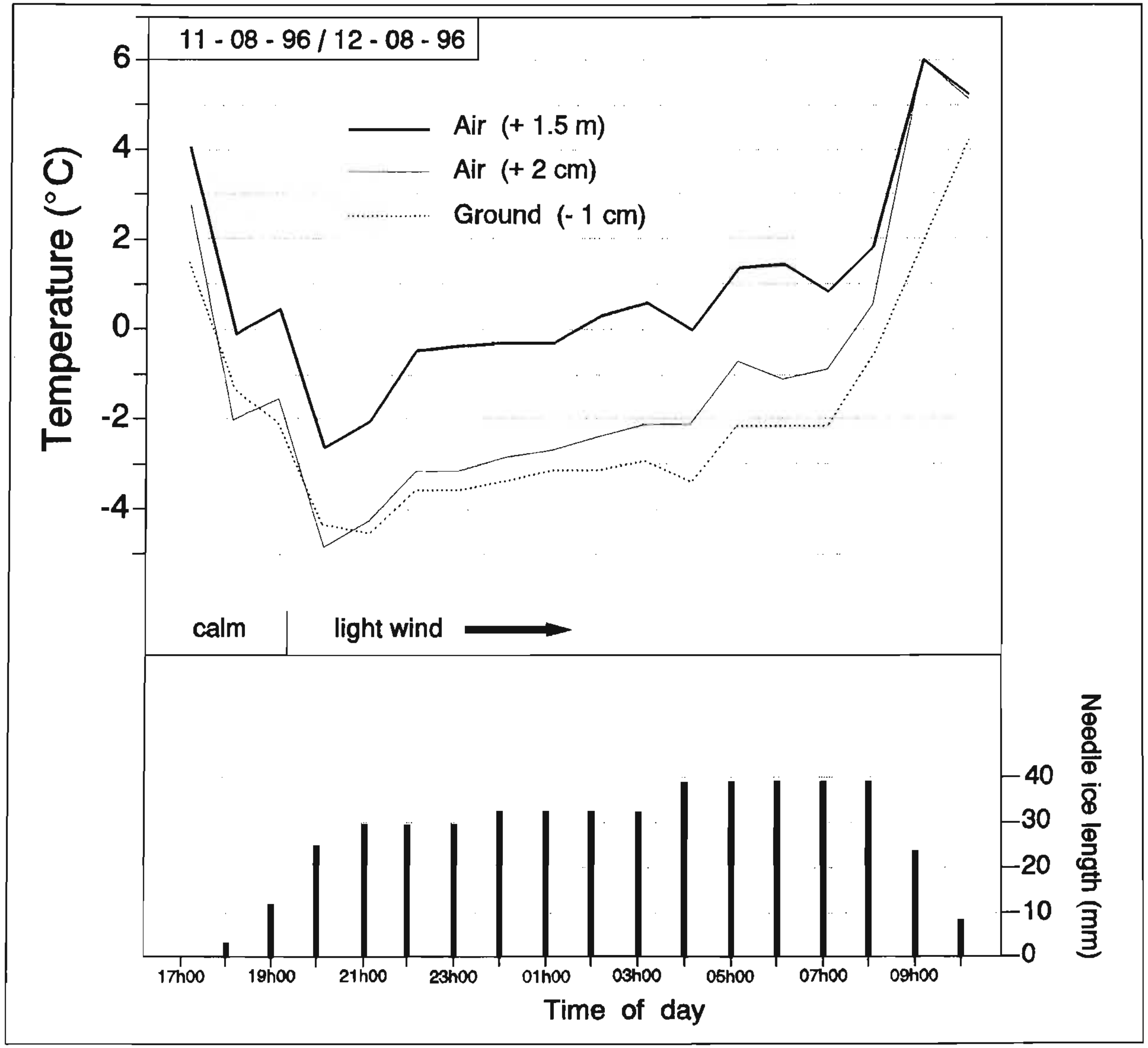


Figure 7.20 h Needle ice growth and ablation cycles against air and ground temperatures, Mashai Valley (11-08-1996 to 12-08-1996).

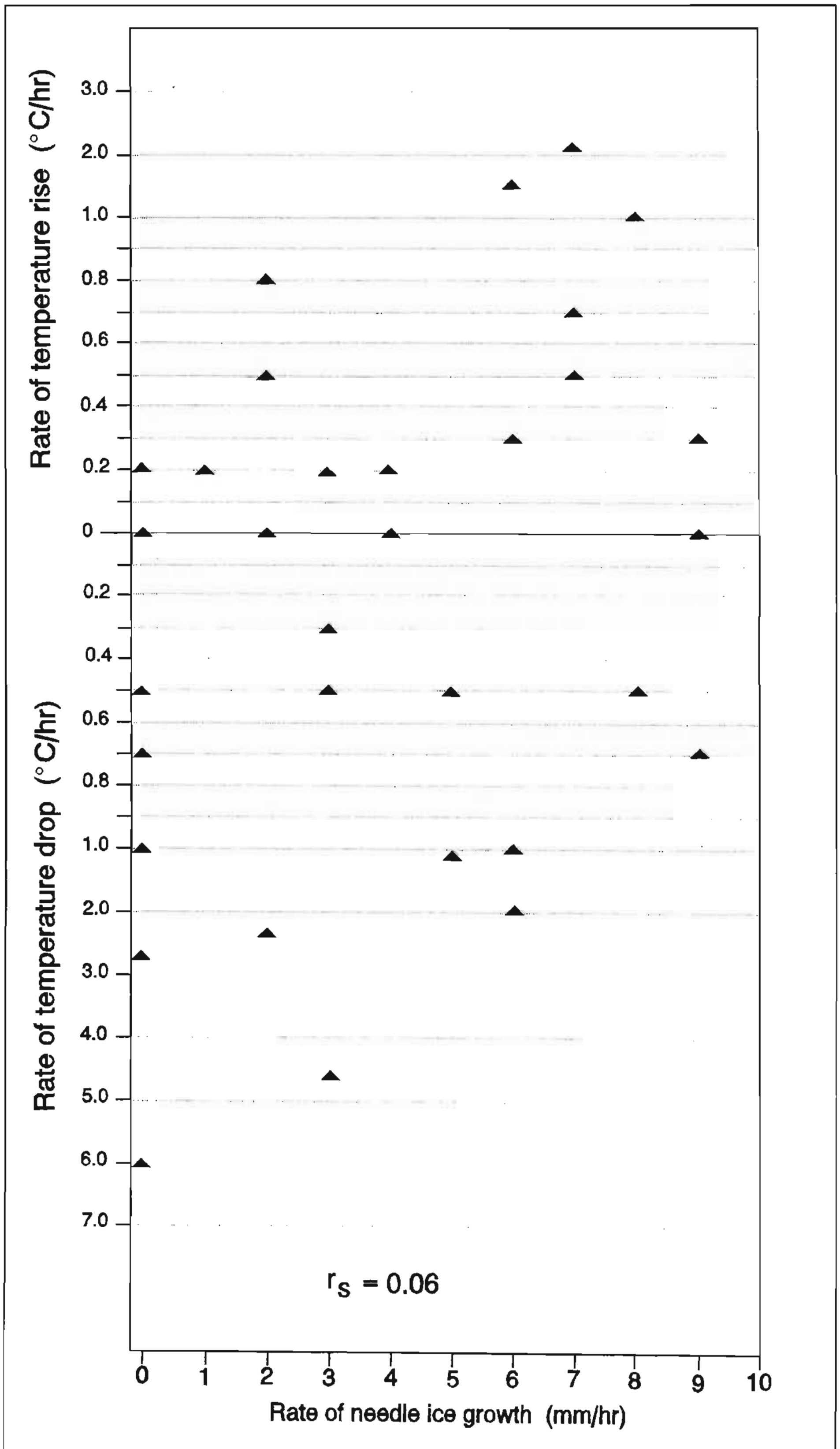


Figure 7.21 a Correlation of rates of air temperature change (+ 1.5 m) with rates of needle ice growth.

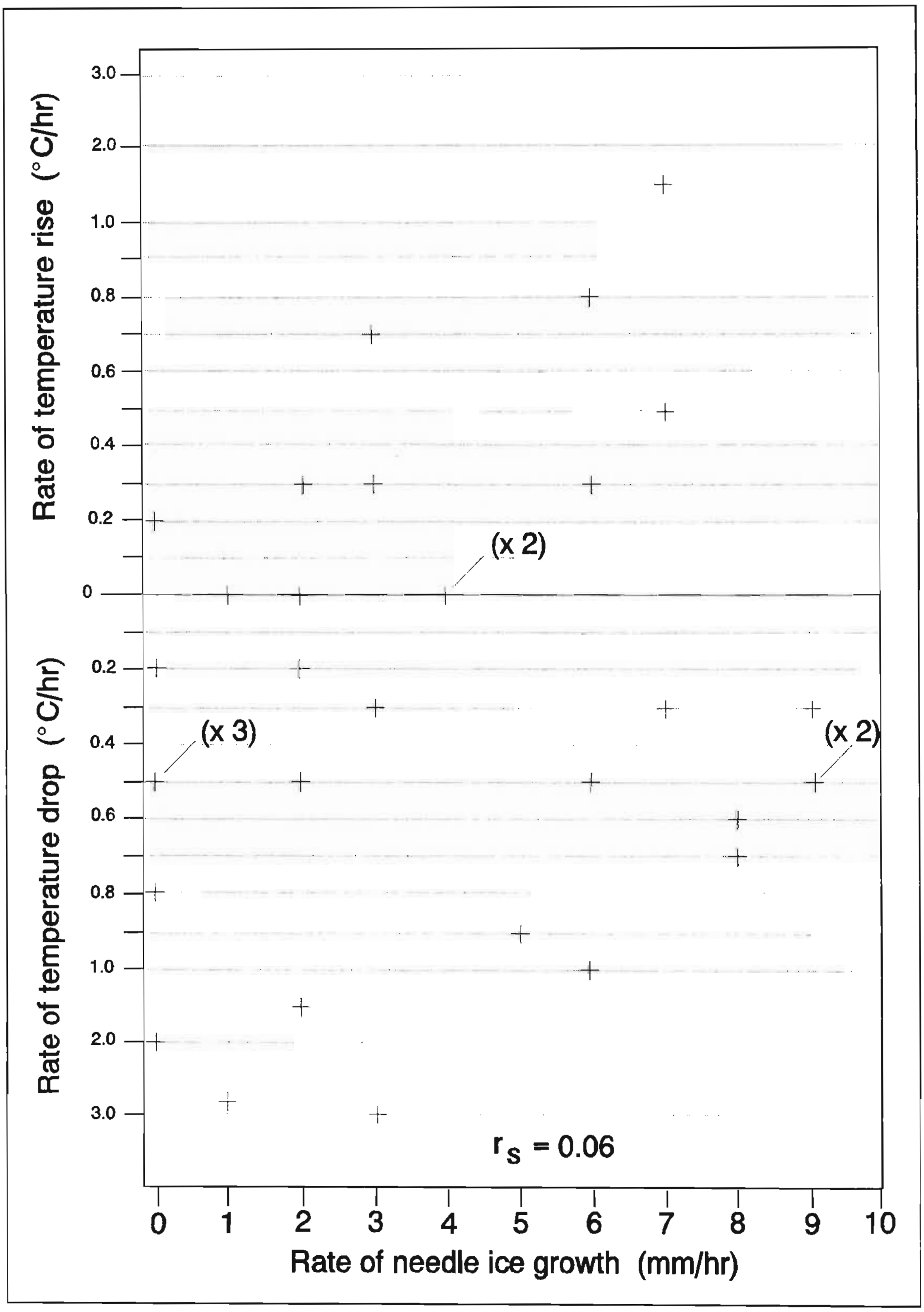


Figure 7.21 b Correlation of rates of surface temperature change (+ 2 cm) with rates of needle ice growth.

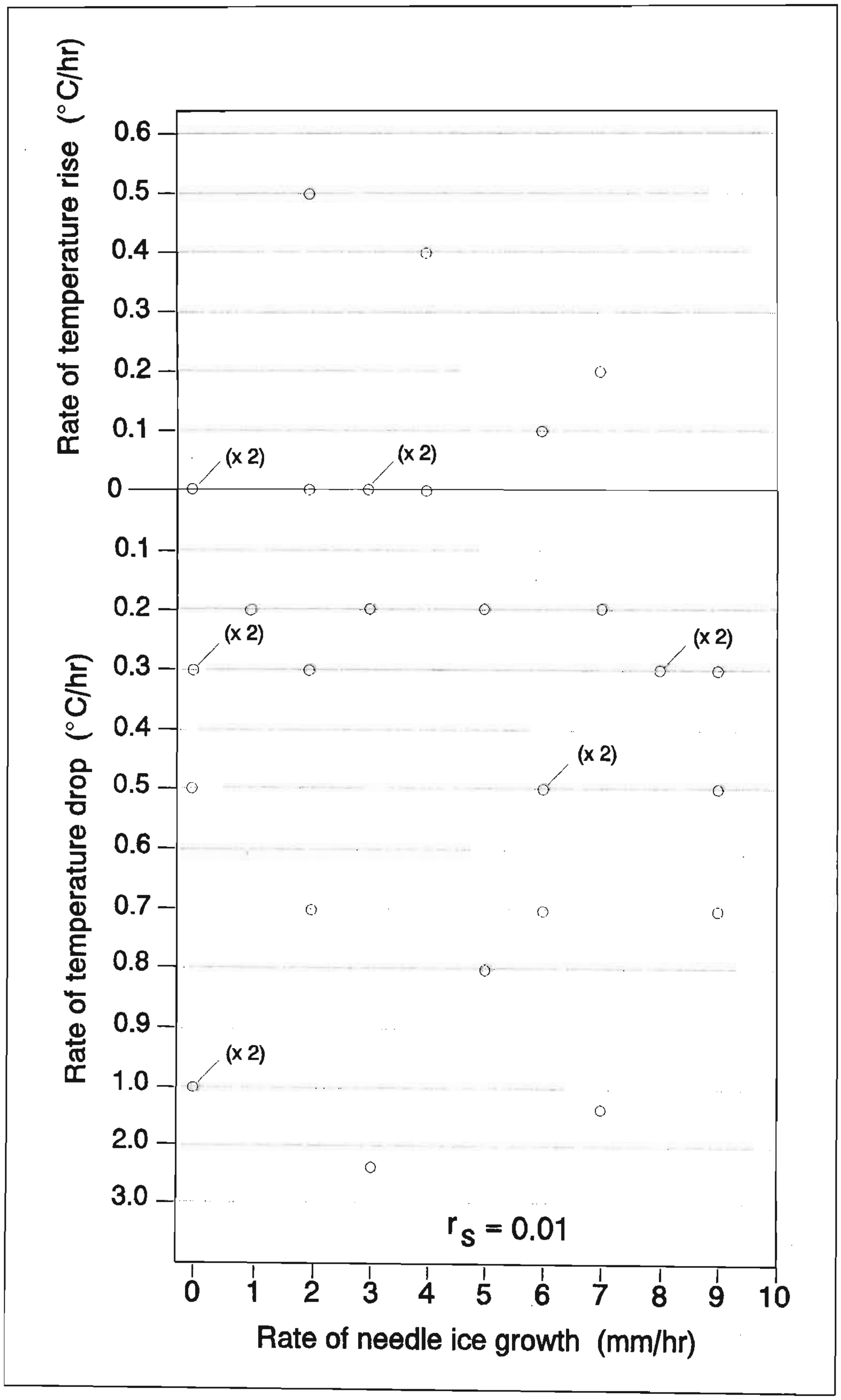


Figure 7.21 c Correlation of near-surface ground temperature (- 1 cm) with rates of needle ice growth.

to wind. Needle ice growth commonly occurs even when temperatures rise (provided these remain below 0°C). Figure 21 a-c represents 33 hours of observations to assess the rate of needle ice growth against changes in temperature. From Figure 21 a, it is evident that during 7 of the observation hours no needle ice growth was recorded. However, needle ice growth was recorded during 12 hours when air temperatures increased, 10 hours when air temperatures decreased and 3 hours when air temperatures remained constant. During the same observation period, needle ice growth occurred during 7 hours of increasing surface temperatures and 15 hours of decreasing surface temperatures (Figure 21 b). Ground temperatures usually decrease during needle ice growth (total recorded = 17 hours) but phases of growth during ground temperature rise are evident from 4 of the observation hours (Figure 21 c).

Rapid needle ice development is frequently noticeable during the initial stages of growth when temperatures drop quickly to below 0°C. Decreasing air and ground temperatures later during the growth phase do not always initiate further growth, possibly owing to other environmental constraints such as available moisture. However, needle ice growth may sometimes occur during a delayed ground temperature response to changing air temperature. This permits growth during increasing air temperatures when the ground temperatures are still decreasing (e.g. Figure 20 d). Similarly, advanced growth may sometimes occur one or two hours after a significant drop in air and/or ground temperature (e.g. Figure 20 a).

It has been argued (Outcalt, 1971a; Lawler, 1993) that when the rate of ground cooling is particularly high, this may promote pore-freeze and thereby prevent needle ice development. Results from the Mashai Valley show that rapid rates of cooling after sunset produce needle ice development even when air temperatures drop by as much as 4.6°C h^{-1} and ground temperatures drop by 2.4°C h^{-1}. Pore ice development at the ground surface appears to be primarily in response to increased vertical fluxes owing to wind, as was observed on the 4/5th June 1996. Very windy conditions did not induce rapidly lowered near-surface ground temperatures but possibly caused rapid temperature decreases through the soil pore areas and soil profile, thereby preventing the development of a suction gradient. Despite this, needle ice developed in the Mashai Valley even during extremely windy conditions. This was possible, owing to the layer of pore ice (30 mm thick) that developed on the 4/5th June, which appears to have protected the underlying soil against excessive vertical fluxes of wind and so permitted needle ice (32 mm) to develop below the frozen sediment cap.

Despite the needle ice growth phase in the Mashai Valley usually lasting 12 hours or less, it nevertheless permits needle ice lengths of up to 55 mm. According to Lawler (1993), needle ice lengths usually range from 2 to 31 mm for individual growths. In the Mashai Valley, 87% of observations during the winter period recorded needle ice lengths exceeding 31 mm for individual growths. The longest measured needle ice length after a single nights growth in the high Drakensberg is 90 mm (26 May 1993). Although the duration of freeze (growth phase) is not the only criteria determining needle ice growth, shorter growth phases of 7 to 8 hours produced shorter needle ice lengths (28-32 mm) than longer growth phases (11-12 hours) where lengths usually exceeded 40 mm (Table 7.6). From Figure 20 a-h, it is evident that the most pronounced growth occurs during calm, clear conditions, while windy and overcast conditions retard needle ice growth. Maximum needle ice length after one nights growth is therefore dependent on several factors including freeze duration, wind frequency and speed, duration and quantity of cloud cover and surficial ground properties (e.g. presence of pore ice, particle size distribution and moisture availability).

The four needle ice phases (growth-stagnation-ablation-ice free period) identified by Ellenberg (1974), are recognizable attributes of the diurnal needle ice cycles observed by the present author in the high Drakensberg. The duration from initial needle ice growth to complete ablation varied between 17 to 20 hours for the winter recording periods. Although no quantitative measurements are available for autumn (April/early May) and spring (September) months, it is envisaged that the diurnal duration of needle ice activity during the milder months is somewhat reduced owing to rapid ablation which also begins at an earlier hour. The needle ice growth phase usually begins between one and three hours after sunset and frequently continues until 05h00. The recorded growth phases varied from 7 to 12 hours but may include periods of stagnation between individual growth cycles. After the growth phase, there is a period of stagnation which may last from 4 to 10 hours. The period of stagnation is extended during windy conditions when needle ice growth is halted (e.g. Figure 20 d). The protective

DATE	GROWTH PHASE		NEEDLE ICE PHASES (Hours & % of Total Hours)				MAX. LENGTH	MINIMUM TEMPERATURE (°C)		
	Av. rate of growth (mm/hr)	Max. rate of growth (mm/hr)	Growth (including phases of stagnation)	Stagnation	Ablation	Total	(mm)	Air (1.5m)	Air (2cm)	Ground (- 1 cm)
27/28-05-95	5.0	16	11 (65%)	5 (29%)	3 (18%)	17	55	-11.4	-	-7.8
28/29-05-95	3.8	14	12 (67%)	6 (33%)	3 (17%)	18	46	-3.4	-	-4.9
2/3-06-96	4.5	9	11 (58%)	4 (21%)	5 (26%)	19	50	-5.3	-5.5	-4.3
3/4-06-96	3.8	9	12 (60%)	6 (30%)	4 (20%)	20	46	-1.0	-2.8	-3.5
4/5-06-96	4.0	8	8 (------)	10 (------)	-	-	32	-8.6	-6.8	-5.8
5/6-06-96	4.0	13	7 (39%)	10 (56%)	3 (17%)	18	28	-3.4	-5.0	-3.3
10/11-08-96	4.8	13	11 (58%)	6 (32%)	4 (21%)	19	53	-5.3	-7.0	-6.7
11/12-08-96	3.6	14	11 (61%)	9 (50%)	3 (17%)	18	40	-2.6	-4.8	-4.5

Table 7.6 Various properties of needle ice growth and ablation from the Mashai Valley.

cover of pore ice (e.g. Figure 20 e) may delay the ablation phase and consequently prolong the stagnation phase. Needle ice ablation usually lasts 3 to 5 hours but may take considerably longer during mid-winter when air temperatures remain below 5°C.

Because of the high frequency of diurnal freeze events and the general absence of snow in the high Drakensberg, needle ice is a common cryogenic occurrence for at least six months of the year. According to Lawler (1988a), the highest documented needle ice frequency is 70 to 82 events per annum in the southern Polish mountains (Gerlach, 1959). From the needle ice observations and available climatic data (see Section **2.6.4**), it is conservatively estimated that between 120 and 150 annual needle ice events occur in the high Drakensberg. Such a high frequency of needle ice has important implications for sediment mobilization, vegetation disruption and soil patterning. However, during years of exceptional snowfalls when snow may cover the ground for over 50% of the winter period (e.g. during the 1996/7 winters), annual needle ice frequency would be significantly reduced. Therefore, the occurrence/absence of snow has an important indirect control on the frost efficacy in the high Drakensberg.

7.3.3 Cryogenic Processes on Stream Banks

Snow, sheets of stream ice and needle ice may frequently be observed along high Drakensberg stream banks from May to September. It is possible that all the above mentioned phenomena contribute in some way towards sediment mobilization and redeposition along stream banks. Needle ice is, however, by far the most common and widespread cryogenic processes operating on stream banks in the high Drakensberg. Because of the irregular occurrence of snow and sheets of stream ice, the effects of such have not been examined.

Observations regarding moisture and texture controls on needle ice growth have shown that the moisture required increases as the percentage fines in the soil decreases (Meentemeyer and Zippin, 1981). Further, Meentemeyer and Zippin (1981) suggested an optimal efficiency composition of 12 to 19% fines for needle ice development. No soil samples collected in the Mashai Valley met such criteria, yet needle ice developed prolifically. The bank material along high Drakensberg streams consists mostly of gravel and sand. Consequently, needle ice in the Mashai Valley usually develops where there are only 5 to 10% fines. Sediment within 40 cm of the channel consisted predominantly of gravel and remained saturated throughout much of winter. Therefore, pore ice rather than needle ice develops in this area. The percentage of gravel generally decreases away from the stream channel (Figures 7.22 and 7.23). It was also found that surface moisture increases away from the channel, owing to an increased percentage of finer material with smaller pore spaces (Figure 7.22).

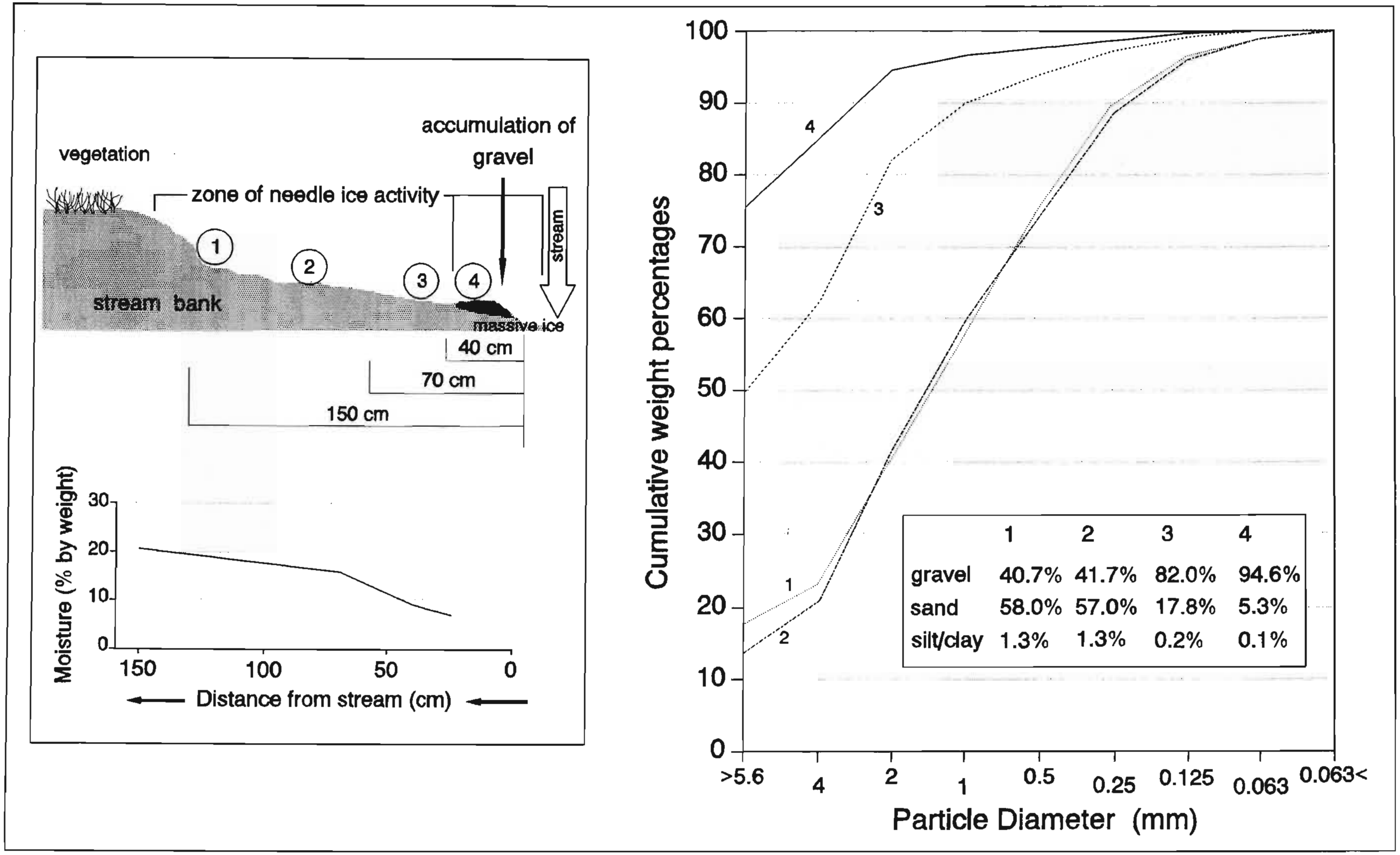
The length of needle ice and its ability to heave material was assessed as a function of distance from the Mashai Stream channel (Figure 7.23). It was found that needle ice length is greatest just above the gravel zone, where it may grow to 4.8 cm in length (Figure 7.23). This is also the zone where the greatest mass of material is uplifted (e.g. 50g/25cm^2). The length of needle ice then progressively diminishes with increasing distance from the channel, but increases again in zones of finer material (Figure 7.23). The mass of uplifted material depends largely on the gravel component, such that where the percentage gravel is higher, the mass uplifted increases (Figure 7.23). Although Meentemeyer and Zippin (1981) found that needle ice does not preferentially uplift any particle size class, comparisons between surface (heaved) and sub-surface material in the Mashai Valley show preferential heave of finer material (Figure 7.24).

Lawler (1987) identified five zones of frost-related stream bank activity with distance down the stream bank:

1. The upper stream bank which is severely frost disrupted.
2. Below (1) there is a zone of short needle ice growth.
3. A zone of debris accumulation occurs below (2).
4. Much needle ice activity is found on a gentle gradient, below (3).
5. The lower zone, adjacent to the flow channel, has limited needle ice activity.

Preliminary findings from the Mashai Valley also reveal a zonation of frost-related stream bank activity with distance down the stream bank (Figure 7.22):

1. The upper stream bank with medium length needle ice growth.



	1	2	3	4
gravel	40.7%	41.7%	82.0%	94.6%
sand	58.0%	57.0%	17.8%	5.3%
silt/clay	1.3%	1.3%	0.2%	0.1%

Figure 7.22 Mean particle size and moisture distribution with distance from the Mashai Stream, winter 1994 (n= 10 per sampling position).

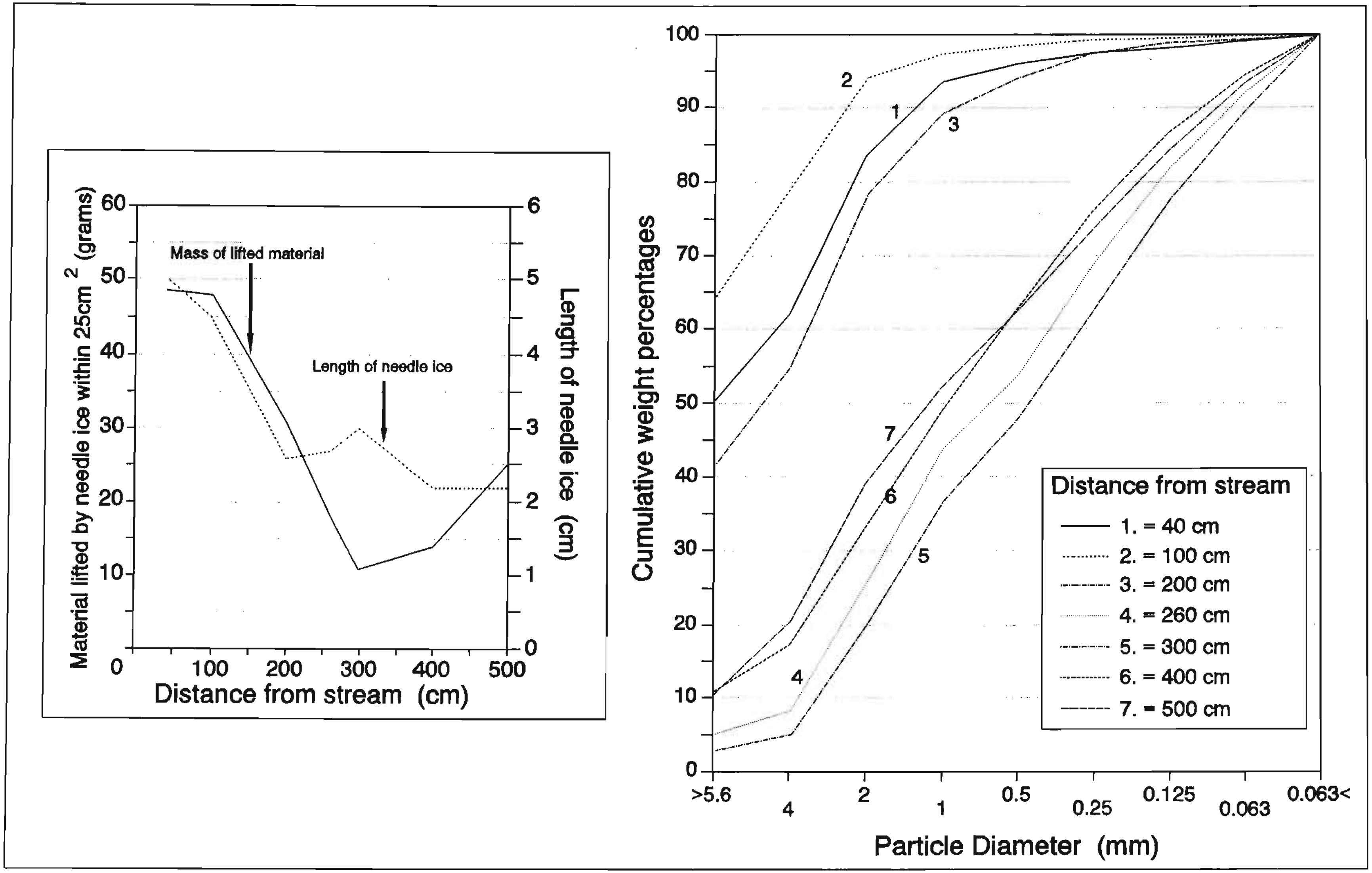
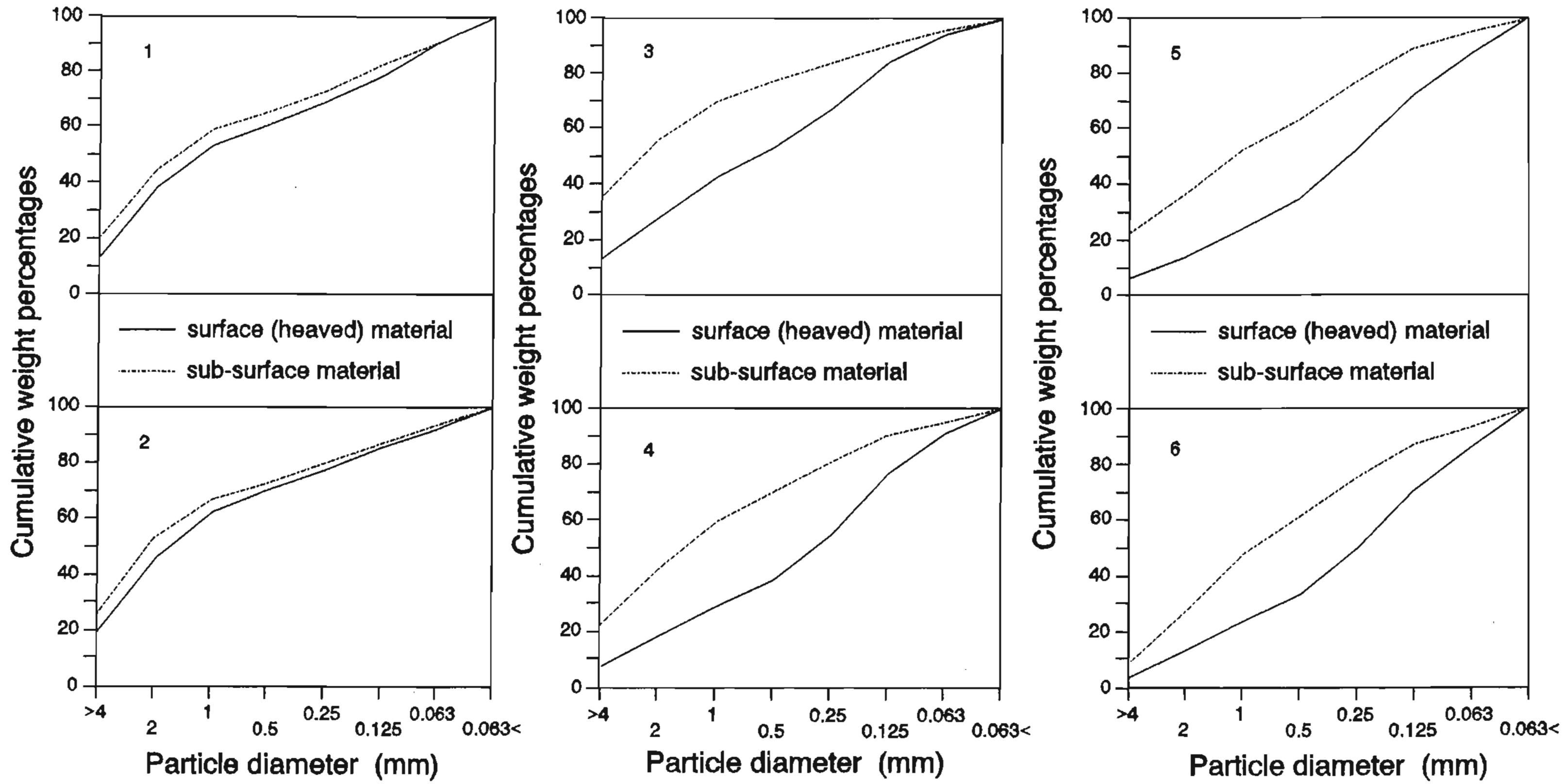


Figure 7.23 Mass of needle ice lifted material against needle ice length, with distance from the Mashai stream. Particle size distribution of needle ice lifted material, with distance from the Mashai Stream, is indicated on the graph to the right.



Mean Values

	Sample N°	Gravel	Sand	Silt/Clay
Surface (heaved) material	12	26.6%	64.0%	9.4%
Sub-surface material	12	43.9%	49.9%	6.2%

Figure 7.24 Comparing the particle size distribution of needle ice lifted material against the sub-surface material (n = 2 per sampling position).

2. Below (1) there is a zone of long or short needle ice,depending on the predominance of gravel within this zone.
3. A zone of well developed needle ice growth occurs below (2).
4. A gravel accumulation zone below (3).
5. A zone of pore ice is usually found adjacent to the flow channel.

7.3.4 Mechanisms of Needle Ice Sediment Transfer and Needle Ice Collapse/Ablation

Lawler (1993) identified four principal processes which transport lifted sediment during needle ice ablation:

1. (a) Direct particle fall; especially where the sediment cap becomes desiccated. (b) It was observed that on windy nights curved needle ice does not necessarily lift particles at right angles to the ground.
2. Downslope transfer by sediment-laden meltwater rivulets.
3. Downslope sliding owing to gap developement at the base of the ice needles.
4. Toppling failure.

All the above processes were observed at the high Drakensberg study sites. In addition, several new observations were made during a detailed investigation in the Mashai Valley.

On some surfaces adjacent to the stream channel, zones of enhanced (longer) and impeded (shorter) needle ice growth were observed (Figure 7.25). Zones of impeded needle ice are attributed to a greater absence of fines and more rapid desiccation. On steep west- and south-facing stream banks, the area of enhanced needle ice offers early morning shading on the shorter needle ice immediately below (Figure 7.25). Consequently, needle ice ablation is somewhat delayed and slower on the lower sections of stream banks. On relatively flat surfaces, the zone of maximum ablation is frequently in the middle of the ice filament. During the onset of ablation, when the sun is at a low azimuth, the upper part of the needle ice cluster is protected against ablation by the sediment cap above the ice filaments (Figure 7.25). Similarly, the lower part of the needle ice cluster is protected against ablation through shading from the needle ice cluster itself (Figure 7.25). The less protected central parts of the

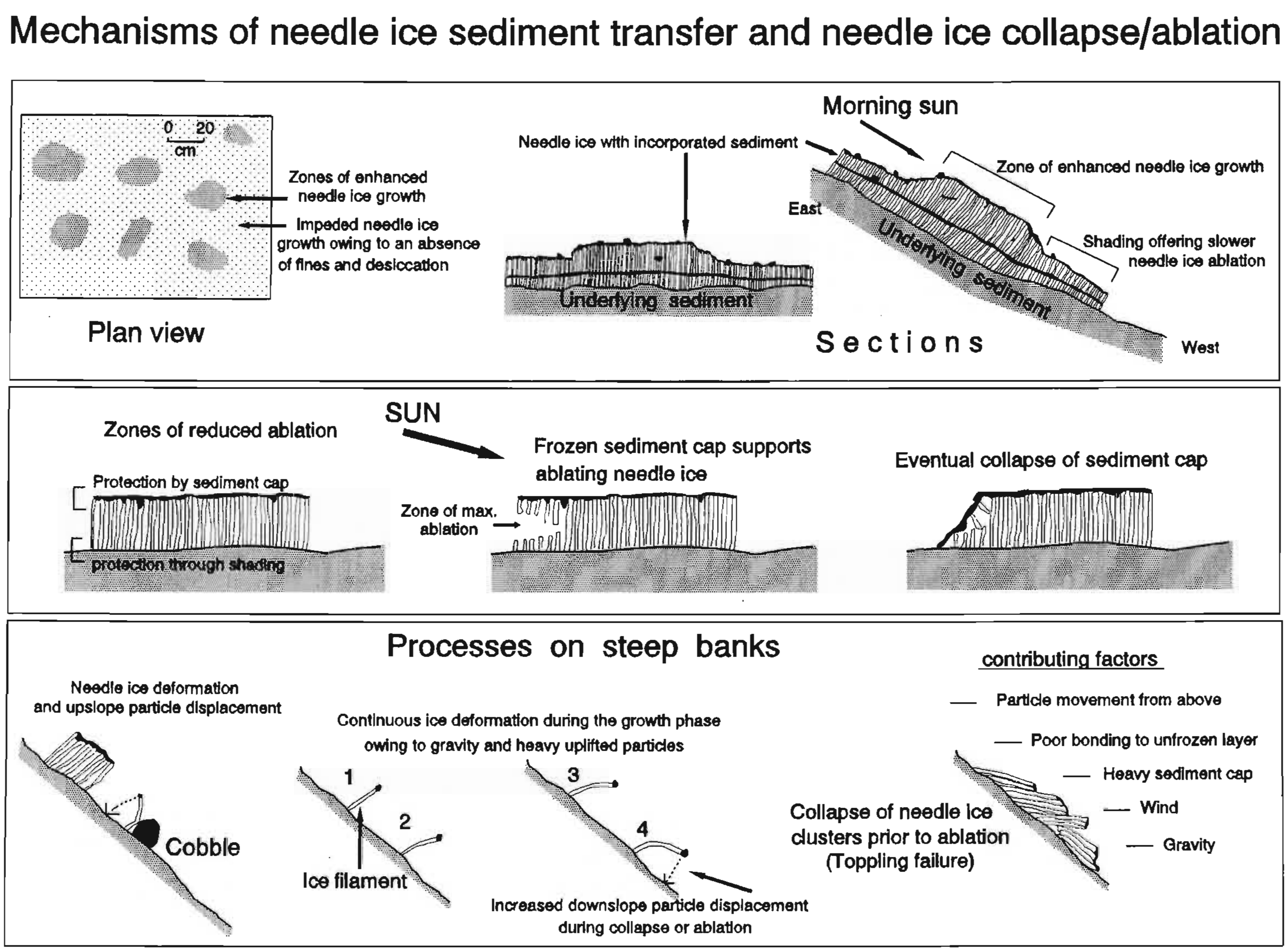


Figure 7.25 Mechanisms of needle ice sediment transfer and needle ice collapse / ablation.

ice filaments are therefore subjected to maximum ablation. Individual ice filaments then dissipate (ablate) in the central parts but remain attached to the frozen sediment cap and underlying sediment. As ablation progresses further into the needle ice clusters, the sediment cap (with some shortened needle ice still attached) eventually collapses under gravity (Figure 7.25).

On steep stream banks comprised of cobbles, needle ice growing immediately above the cobbles are sometimes deformed and grow in an upslope direction (Figure 7.25). In such instances, particle displacement during ablation is frequently in an upslope direction (Figure 7.25). Continuous ice deformation during the growth phase may occur in the downslope direction, owing to gravity and the mass of the uplifted particles, thereby permitting increased downslope particle displacement during needle ice collapse or ablation (Figure 7.25). Collapse of needle ice clusters prior to ablation (including toppling failure) is attributed to: (a) poor bonding to the unfrozen layer, (b) a heavy sediment cap, (c) wind, (d) gravity and (e) particle movement from above.

7.3.5 Sediment Mobilization on the Mashai Stream Banks

Soil Trough Results

Bank debris were collected from three troughs during the winter (May to August) of 1993, and from one trough for the spring (August to October) of 1993 (see Section **3.4.2**). The first winter recording period (29 May to 24 July) experienced approximately 24 frost days and received no precipitation. During this time, sediment mobilized ranged between 2.4 and 3.2 grams/day/0.5 m along the bank. The second winter recording period (24 July to 25 August) received 27 frost days and 36 mm of precipitation. During this period, sediment mobilized ranged between 1.7 and 2.7 grams/day/0.5 m (Table 7.7). The sediment mobilized during this winter period appears, therefore, to be predominantly a result of needle ice action. It is also evident (Table 7.7) that sediment on steeper banks (18° to 23°) is more effectively mobilized than that on shallower banks (11°). During the period from 25 August to 25 October 1993, approximately 213 mm of precipitation fell in the region (Table 7.7). Sediment mobilized on

	Bank Gradient	Bank Aspect	N° of Days	~ Frost Days	~ Precip. (mm)	Sediment mobilized mass (g)	grams/day/0.5m	Particle size distribution gravel	sand	silt/clay
Trough 1	23°	245°								
29 May - 24 July			56	52	0	167.2	3.0	92.3%	7.5%	0.2%
24 July - 25 Aug.			32	27	36	63.9	2.0	87.5%	12.1%	0.4%
25 Aug. - 25 Oct.			61	20	213.4	-	-	-	-	-
Trough 2	11°	200°								
29 May - 24 July			56	52	0	133.5	2.4	78.0%	20.8%	1.2%
24 July - 25 Aug.			32	27	36	56.0	1.7	76.9%	21.7%	1.4%
25 Aug. - 25 Oct.			61	20	213.4	1473.1	24.1	39.0%	55.9%	5.1%
Trough 3	18°	205°								
29 May - 24 July			56	52	0	179.4	3.2	75.2%	23.7%	1.1%
24 July - 25 Aug.			32	27	36	87.8	2.7	78.9%	20.7%	0.4%
25 Aug. - 25 Oct.			61	20	213.4	-	-	-	-	-

Table 7.7 Characteristics of sediment mobilization and particle size distribution from three experimental sites along the Mashai Stream.

an 11° stream bank amounted to 24.1 grams/day/0.5 m during this period (Table 7.7). Clearly, the rate of sediment mobilization as a result of rain-splash and rain-wash is several times greater than that mobilized as a result of needle ice action.

The particle size distribution of sediment collected in the troughs was predominantly gravel (75.2 to 92.3%) for the months of May to August (Table 7.7 and Figure 7.26 a-c). This is also an indication that the gravel component of uplifted material is displaced streamwards more quickly than sand and fines. Further, it is possible that the low percentage of fines (0.2 to 1.4%) is, in part, a result of deflation. The particle size distribution for sediment collected after a wet period (August to October) is very characteristic of the soil as a whole, with a more even distribution of gravel (39%), sand (55.9%) and fines (5.1%) (Table 7.7 and Figure 7.26 a-c).

Particle Displacement Results

Needle ice induced movement of painted clasts has been measured on talus slopes by Pérez (1985, 1987b, 1987c, 1988). However, rates of needle ice induced particle displacement on stream banks appears only to have been examined by Lawler (1993) and data are limited. No particle movement data have been available from the high Drakensberg prior to the present study.

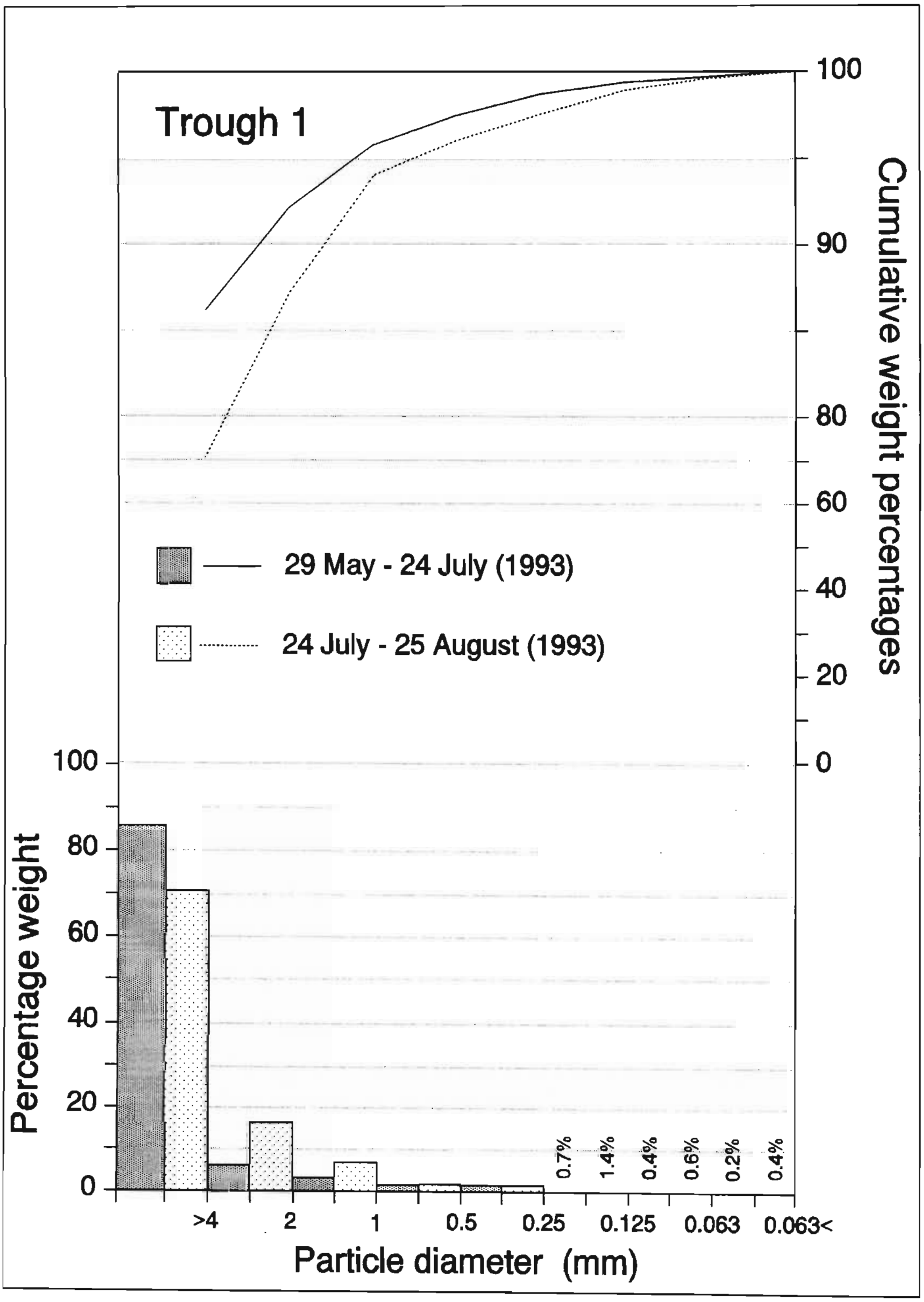


Figure 7.26a Particle size distribution for sediment trapped in trough 1.

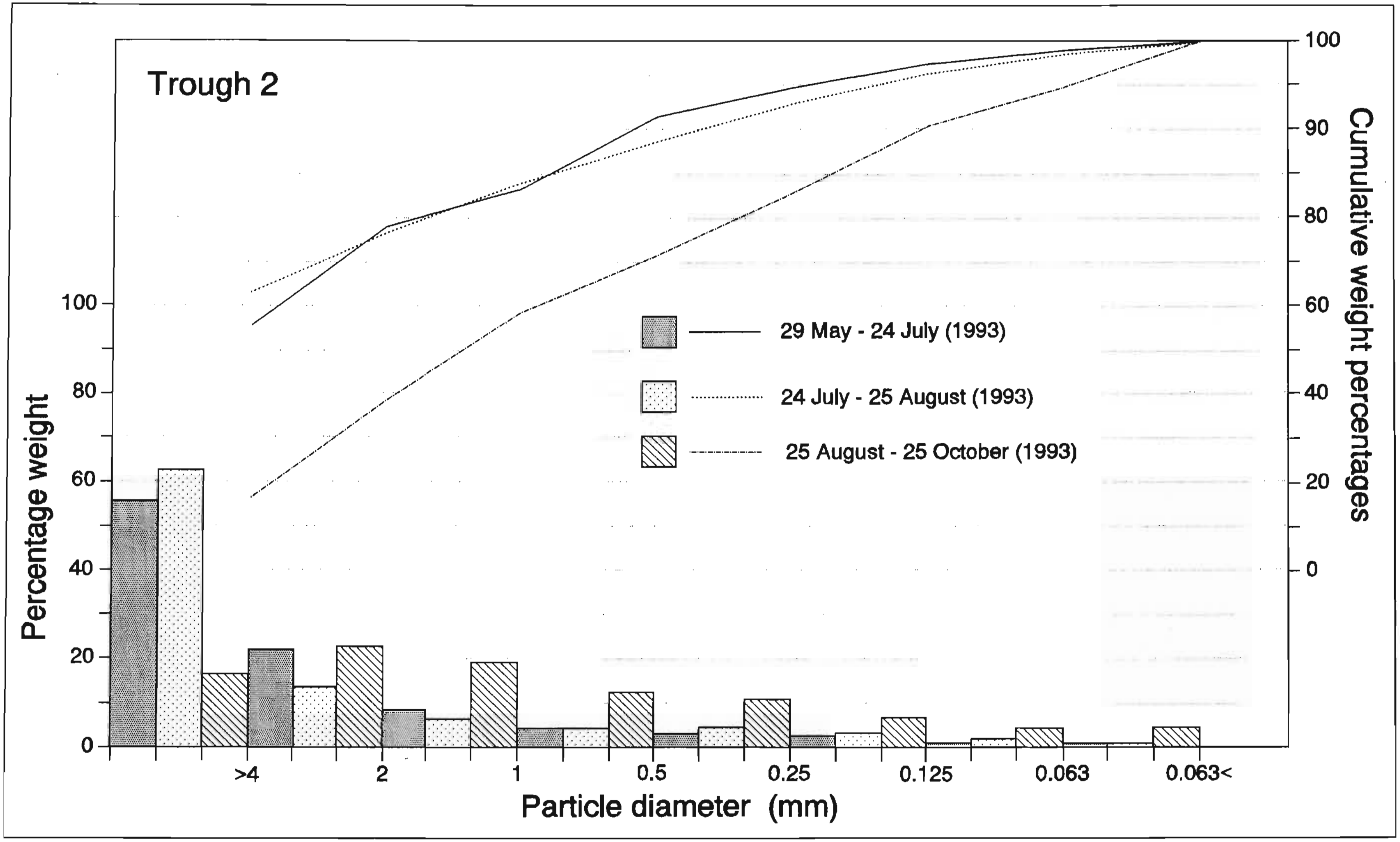


Figure 7.26b Particle size distribution for sediment trapped in trough 2.

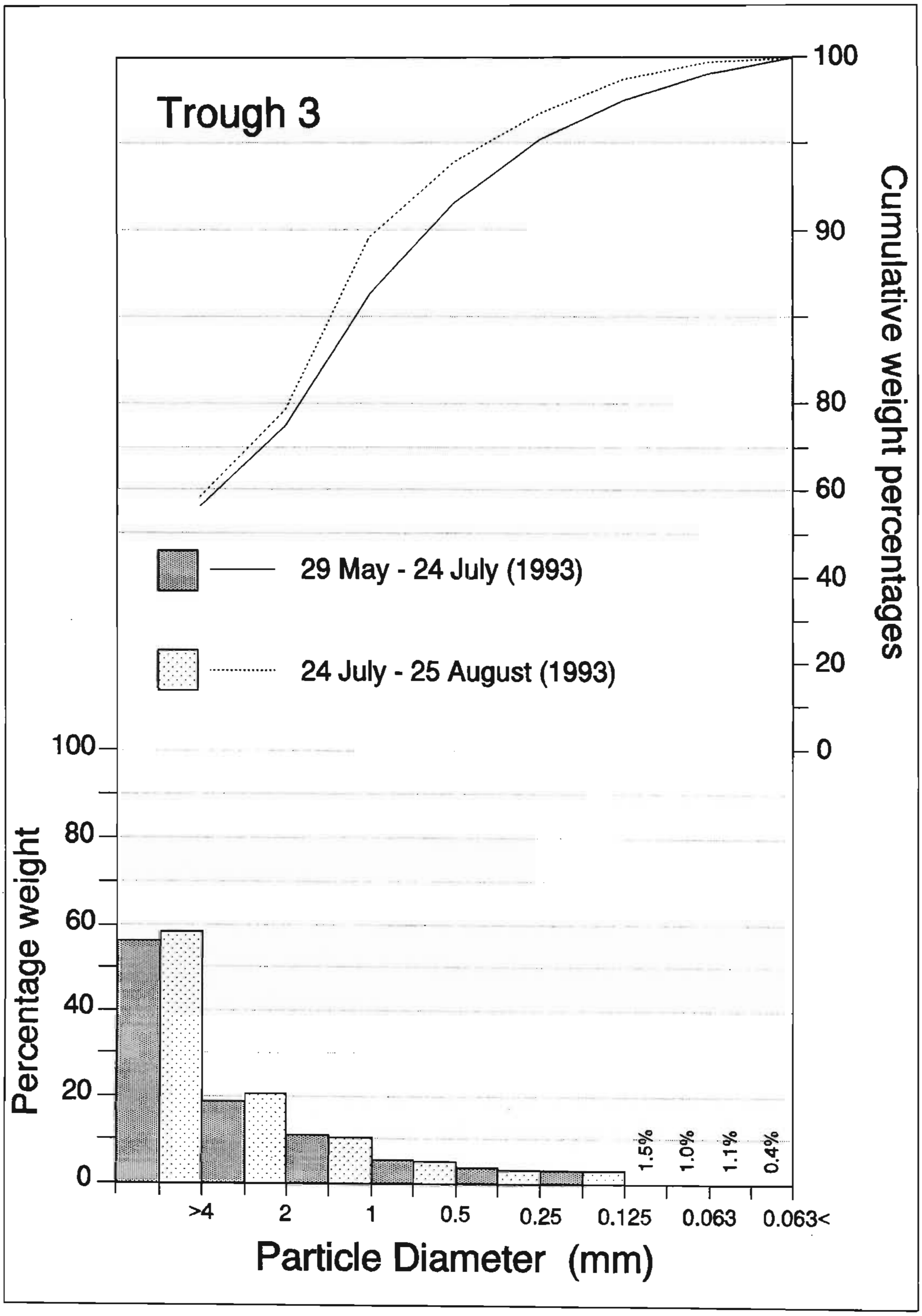


Figure 7.26c Particle size distribution for sediment trapped in trough 3.

1995. Twenty-four particles spaced 5 cm apart were painted along each of the four transects. Two transects were 40 cm from the stream channel and another two transects were 2 m from the channel (slope gradient was uniform). Results (from Figure 27 a) show that the rate of particle displacement is substantially increased towards the stream channel. Downslope displacement for the two lower transects total 41 and 99 mm while the displacement from the transects located 2 m from the stream total 0 and 15 mm. Maximum downslope displacement for an individual particle was 18 mm. Ten particles were displaced upslope for the three transects that encountered particle movement. Maximum upslope displacement for an individual particle was 7 mm. Average downslope displacement per particle was 1.0 and 2.79 mm for the lower transects and 0 and 0.37 mm for the upper transects. The findings indicate accelerated particle displacement towards the lower reaches of stream banks, possibly owing to enhanced needle ice activity. However, rates of particle movement along stream banks change considerably, depending on bank gradient, particle size distribution and moisture supply.

Particle displacement along three transects was also measured over a period of nine weeks during the 1995 winter. Fifty particles spaced 2 cm apart were painted along each transect. Transects occupied stream banks of variable gradient and were positioned one metre from the stream channel. Results indicate substantially enhanced particle movement on the steeper (23° of plot 3 and 18° of plot 1) stream banks than on the shallower (11° of plot 2) stream bank (Figure 27 b-d). For instance, total downslope displacement of 50 particles on a steep (23°) Mashai stream bank amounted to 854.1 cm while on a low gradient (11°) bank it only amounted to 131 cm (Figure 27 c/d).

Average downslope displacement per particle varied from 2.67 cm (11° bank gradient) to 17.1 cm (23° bank gradient) for the nine week winter period. A similar finding has been reported by Pérez (1987b) who explained that slope gradient significantly affects movement rates. This is possibly owing to sudden and/or successive needle ice collapse (or toppling failure) on steep gradients. Such processes appear less pronounced and less frequent on shallower gradients.

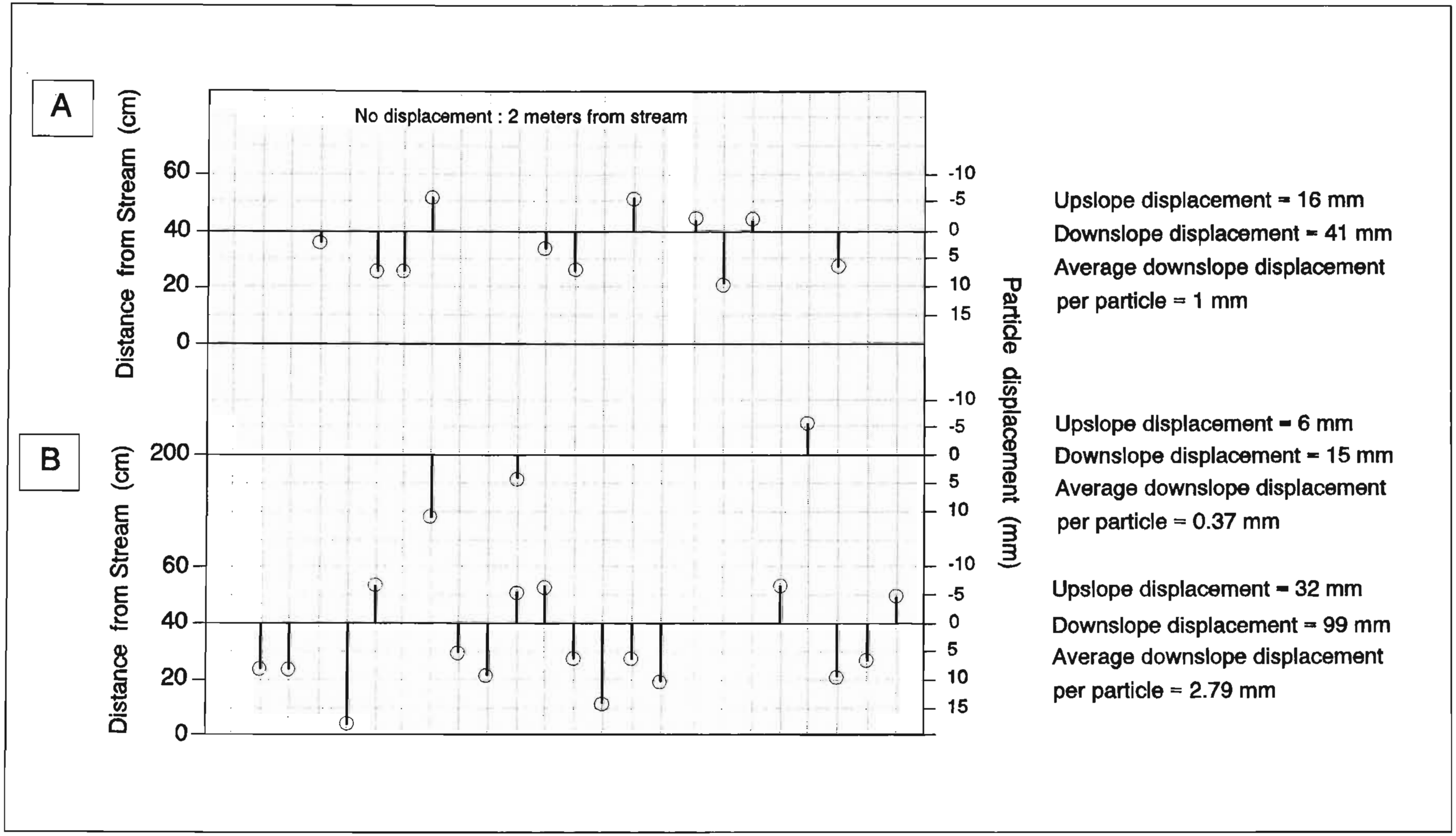


Figure 7.27 a Particle displacement with variable distance from the Mashai Stream.

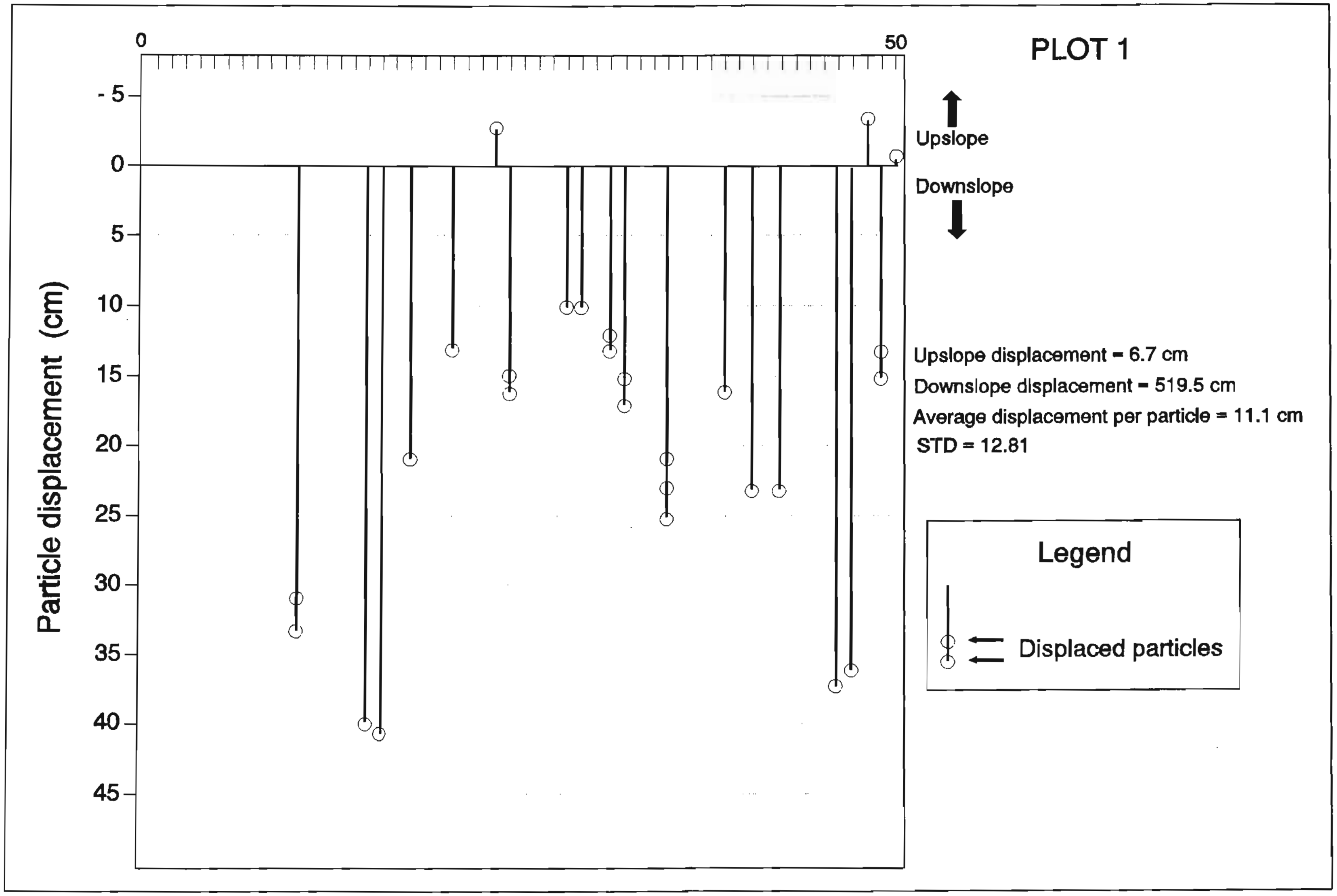


Figure 7.27 b Particle displacement on an 18° stream-bank gradient.

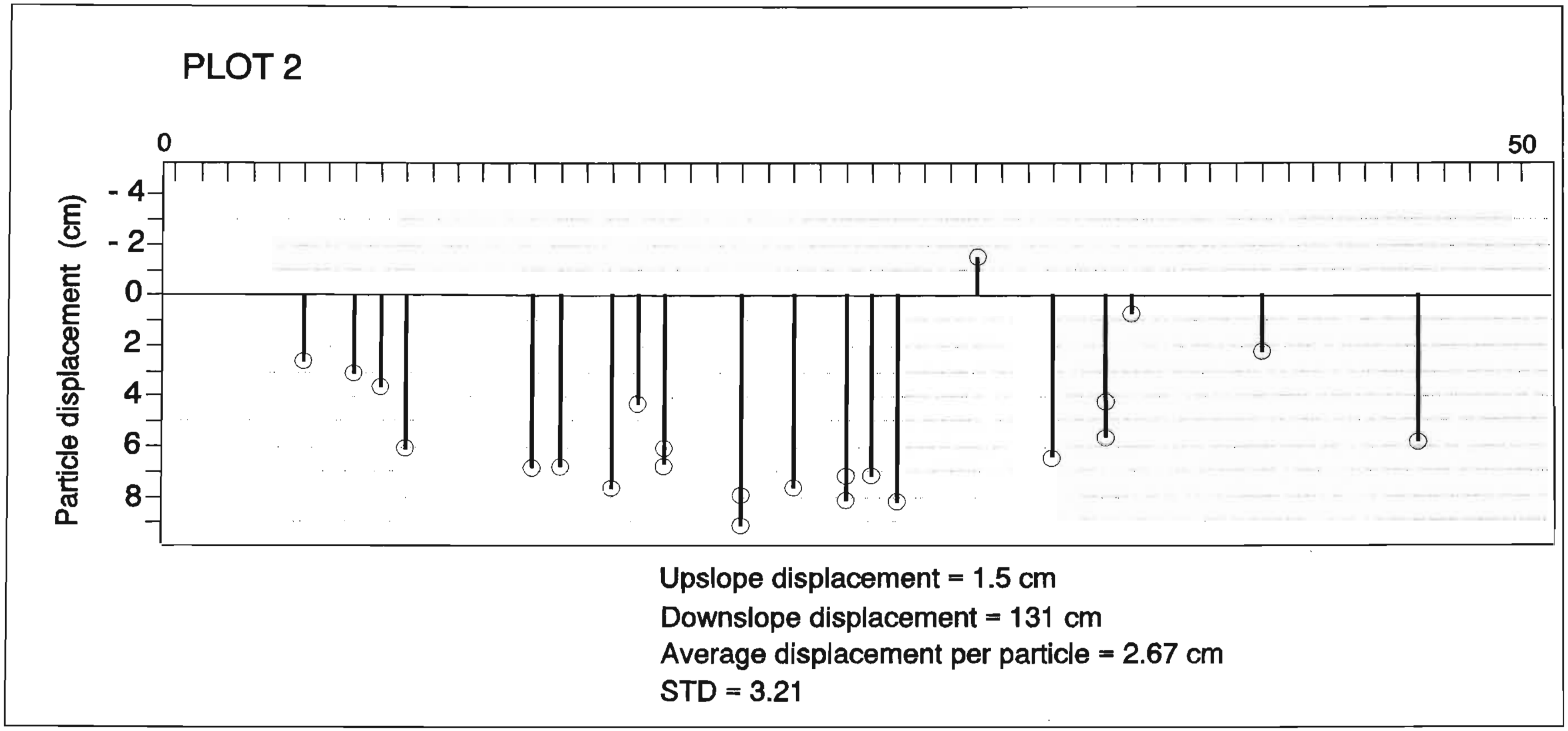


Figure 7.27 c Particle displacement on an 11° stream-bank gradient.

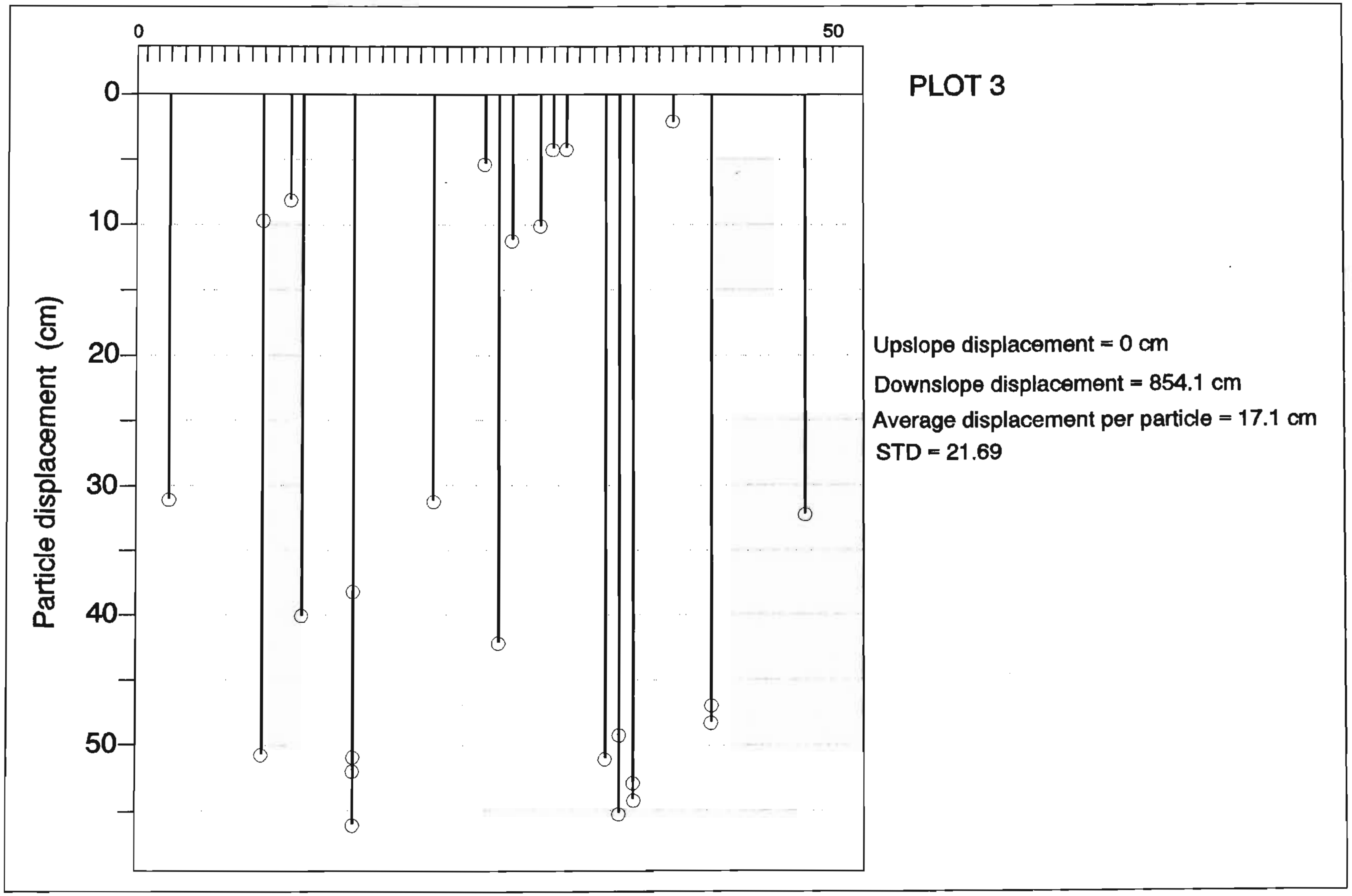


Figure 7.27 d Particle displacement on a 23° stream-bank gradient.

movement rates (23° gradient = STD of 21.69 and 18° gradient = STD of 12.81) than the shallower 11° bank (STD = 3.21). The greater variability of movement on the steeper gradients may again be attributed to the occurrence of toppling failure. Toppling failure and mini slides (micro solifluction) will rapidly transport particles downslope over a given transect zone (Figure 27 b/d). Adjacent to the area affected by toppling failure or micro solifluction are zones subjected to more conventional needle ice collapse (ie. direct particle fall) with consequent reduced rates of down slope movement. The greater absence of toppling failure and micro solifluction on shallower gradients appears to reduce the variability of particle movement along such stream banks.

7.3.6 Discussion

A model has been produced to summarize and explain the sediment mobilization along high Drakensberg stream banks for the 6 month cold/dry period (April to September) and the 6 month mild/wet period (October to March) (Figure 7.28). Needle ice action along high Drakensberg stream banks is an almost daily occurrence from May to mid-September. This prolonged period of activity contributes to bank erosion and sediment transport on stream banks. Sediment lifted by needle ice becomes loose, with a low bulk density and reduced particle bonding capacity (Lawler, 1993). Therefore, any subsequent stream level rise or rain-wash process during the oncoming wet period will more adequately be able to entrain such sediment. Needle ice action throughout winter also causes the build-up of gravel along lower stream banks (Figure 7.29). From this, it becomes apparent that needle ice induced particle displacement is important as an indirect erosion agent as well as in preparing material and rendering it friable on stream banks (Lawler, 1987). Needle ice also produces erosional notches (Lawler, 1993) and causes turf exfoliation along stream banks, thereby contributing towards bank retreat (Figure 7.30). Mini mudflows, as described by Soons and Greenland (1970) and Lawler (1993), may sometimes be observed on steeper banks. Such mini mudflows usually occur after needle ice collapse, and are associated with rapid melt.

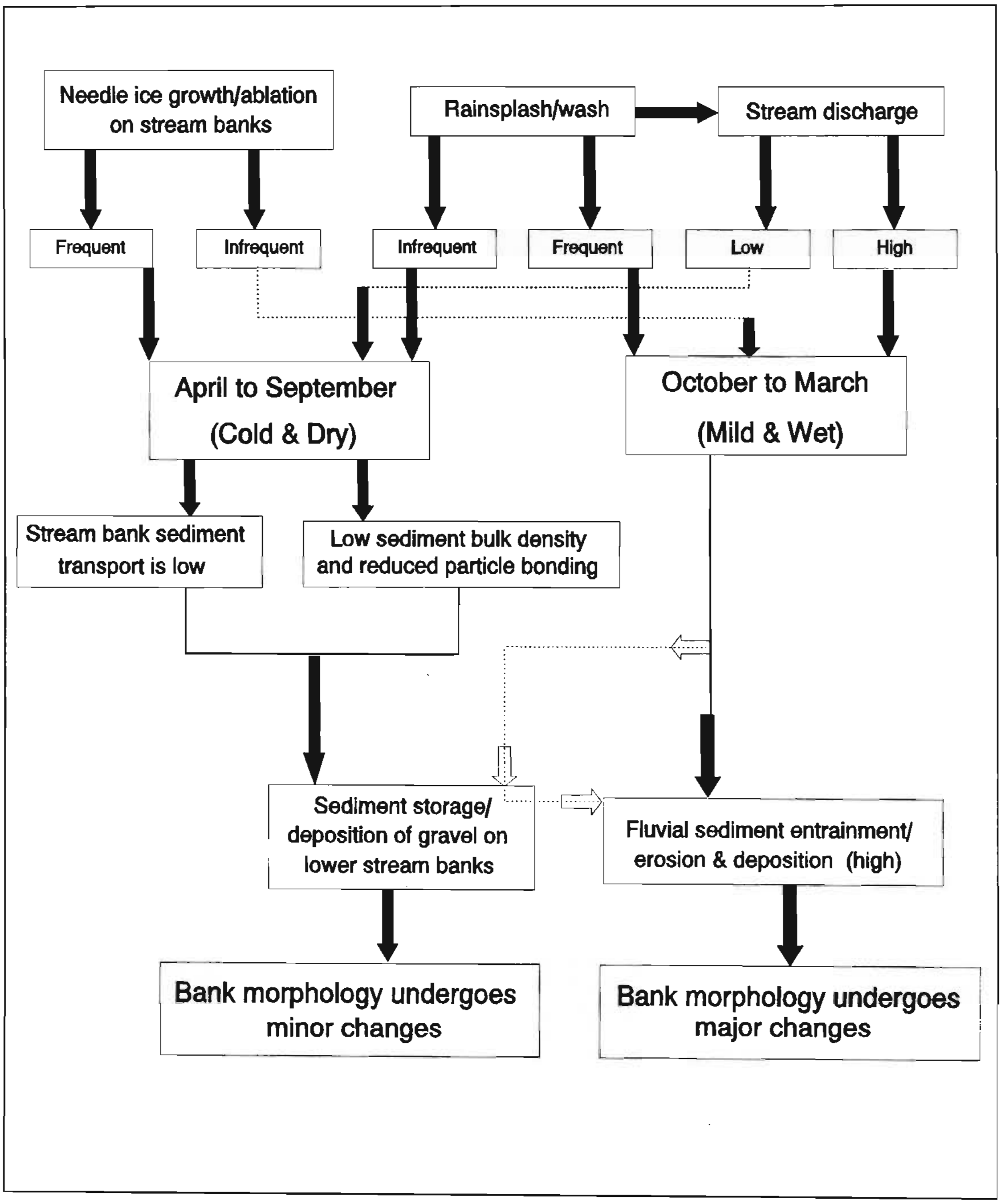


Figure 7.28 A model summarizing the sediment mobilization processes along high Drakensberg stream banks.

Figure 7.29 Needle ice action throughout winter frequently causes the build-up of gravel along lower stream banks as shown here.

Figure 7.30 Needle ice action contributes towards bank retreat and produces erosional notches as in the example shown here.

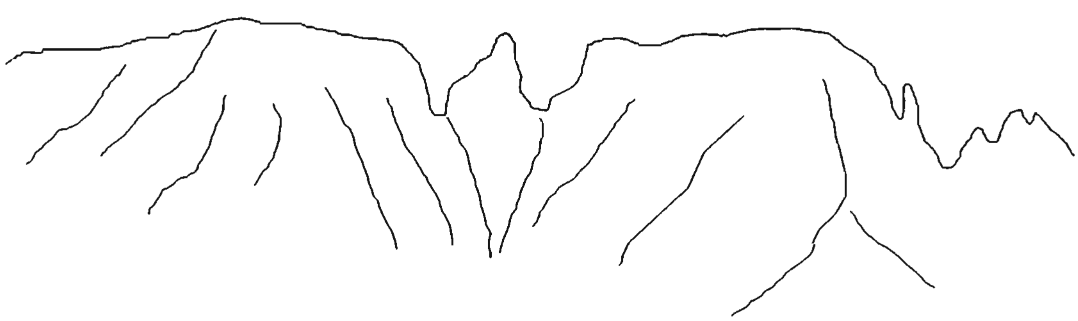
From the presented investigations and the proposed model (Figure 7.28), it would appear that, despite the significance of cryogenic activity during winter, rain-wash and stream discharge are more effective in sediment mobilization on stream banks than are needle ice processes. Water-induced erosion not only transports a greater mass of sediment, but it also mobilizes a more uniform particle size distribution. Stream banks undergo relatively minor morphological changes during the six month frost period (April to September), however, the stream bank morphology may be rapidly altered through accelerated erosion and sediment deposition during the six month wet period (October to March).

7.4 SUMMARY

This investigation has identified several mechanisms contributing to turf exfoliation and stream bank erosion in the high Drakensberg. The pronounced seasonal weather patterns including cold/dry winters and mild/wet summers have promoted a cycle of geomorphic process events which operate synergistically and initiate particular erosion landforms. During the cold/dry period, needle ice activity and aeolian deflation are the primary erosional mechanisms operating in the high Drakensberg. Turf is frequently severely disrupted and sediment becomes increasingly friable during this period. Subsequently, during the mild/wet period, the exposed sediments and loosely held turf are easily removed by rain wash and stream flow. From the above, it becomes evident that cryogenic activity during the colder period initiates erosion and "prepares" material, such that turf and sediment are more easily removed during the fluvial erosional period. There is strong evidence, however, that cryogenic soil erosion processes are overwhelmed by water induced erosion processes in the high Drakensberg.

CHAPTER EIGHT

DISCUSSION AND CONCLUSIONS



CHAPTER 8

DISCUSSION AND CONCLUSIONS

8.1 THE HIGH DRAKENSBERG AS A PERIGLACIAL REGION

A key question that has either been ignored or unsatisfactorily answered in the past is whether or not the high Drakensberg qualifies as, or has qualified as, a periglacial region. Clearly, there has been uncertainty surrounding this question, which has prompted descriptive terms such as "marginal periglacial zone" (Lewis, 1988a, 1988b; Boelhouwers, 1991a, 1994) and "subperiglacial zone" (Hanvey and Marker, 1992). Although Hanvey and Marker (1992) define the subperiglacial zone as one where seasonal freezing and frequent frost cycles occur, this has not been quantitatively shown for the high Drakensberg and does not correspond with Karte's (1983) defining attributes which require a MAAT between -2 and -8°C.

The absence of an internationally agreed definition for the term "periglacial" has surely contributed to some of the uncertainty expressed amongst southern African periglacial geomorphologists. Should permafrost be considered as a necessary attribute of periglacial regions, as has been argued by such as Péwé (1969) and Harris (1988), then the high Drakensberg might never have qualified as a periglacial region. Both this and other studies have found no permafrost-related landforms (see Table 8.1) in the high Drakensberg, and it would thus appear so far that this region has been absent of permafrost during the Quaternary. Thorn (1992) has recently argued that permafrost may indeed be a *sufficient* attribute to entitle a region as periglacial, but warns that it is inappropriate as a *necessary* attribute. Although a "hard and fast" definition of the term periglacial has still not been forthcoming, Thorn's (1992) revised definition outlined in Section **1.1** appears to have received little criticism.

FEATURE	ACTIVE	RELICT	NO RECORDING
Needle Ice Landforms	X		
Cryo-deflation Landforms	X		
Sorted Patterned Ground	X	X	
Non-sorted Patterned Ground :			
Mudboils			X
Stripes			X
Circles	?		
Steps	X	X	
Thufur	X	X	
Earth Hummocks			X
Palsas			X
Pingos			X
Blockfields/Blockslopes	?	X	
Blockstreams	?	X	
Asymmetric Valleys	?	X	
Stone Pavements			X
Cryoplanation Landforms		?	X
Nivation Hollows		?	
Protalus Ramparts			X
Slushflows/Thaw Slumps			X
Turf-banked Lobes	X	X	
Stone-banked Lobes/Sheets	X	X	
Solifluction Lobes	?	X	
Ploughing Blocks			X
Rockglaciers			X
Ice-wedge-casts			X
Sand-wedge-casts			X
Ventifacts		?	X
Moraine			X
Striated Bedrock (Glacial)			X

? = No conclusive evidence

Table 8.1 The recording of various cryogenic (including periglacial and glacial) landforms in the high Drakensberg.

Thorn's (1992) definition is similar to that suggested by French (1988) a few years earlier:

> "Periglacial geomorphology is concerned with the landforms and processes typical of the cold non-glacial regions of the world. These regions are mainly dominated by cryogenic, or frost action, processes, often in association with the formation of permafrost". (French, 1988, p xvii)

Some of the findings outlined in this dissertation are reviewed and discussed so as to ascertain whether the high Drakensberg qualifies as a periglacial region, following the definitions of French (1988) and Thorn (1992).

Despite the absence of permafrost, several cryogenic landforms have now been found and described from the high Drakensberg. Table 8.1 lists the variety of cold region landforms commonly found in periglacial regions and indicates those which have been recorded (active or relict) in the high Drakensberg. From this, it is now evident that almost all cryogenic landforms which are independent on glacial ice, permafrost and nival-associated processes, occur in the high Drakensberg.

8.1.1 The Past

Relict cryogenic landforms found in the high Drakensberg include sorted and non-sorted patterned ground, blockfields and blockslopes, blockstreams, turf-banked lobes, stone-banked lobes, solifluction lobes and asymmetric valleys. Given the large size and considerable depth of sorting of some of these landforms, there can be little doubt that they were initiated during a much colder and possibly wetter climate than is encountered today. Although permafrost may not have been present, deep and prolonged seasonal ground freezing would have been necessary to develop some of the features. Possible processes associated with such landform development may well have included soil convection processes and gelifluction. The relict landforms are, however, restricted to the higher summits and cooler south- and southwest-facing slopes. For instance, the large sorted circles are found above 3400 m a.s.l., while block-accumulations and solifluction lobes predominate above 3100 m a.s.l. on south- and southwest-facing slopes. As the relict cryogenic features form a significant component of the overall geomorphology above 3100 m a.s.l., there is sufficient evidence to support a former periglacial belt which extended to the highest summits.

8.1.2 The Present

Several active cryogenic landforms and processes have been identified in the high Drakensberg. Miniature varieties of sorted patterned ground emerge over a few weeks and are recurrently forming, seasonal, frost related patterns. The occurrence of such patterns demonstrates the effect of diurnal freeze-thaw cycles during contemporary winters. Many thufur may be actively forming today while others are in a state of quasi equilibrium or have decayed to an end point. Various phases of stone-cored thufur found at several sites may indicate that frost push/pull processes have been operative during recent times. Pegs inserted to 5, 10 and 20 cm depth at a thufur site showed evidence of heave during 1994, and thus may support the contention of recent frost push/pull processes. A detailed study of ground temperatures at the Mashai Valley site has demonstrated that seasonal ground freezing may penetrate to below 12 cm depth and induce thermal pressure differentials during contemporary winters. Needle ice is found to be a common and widespread soil disruptive agent from mid-autumn (April) to mid-spring (October). It has also been shown that needle ice is an important contributing agent to turf exfoliation and bank erosion. In addition, needle ice is responsible for the breakup of other cryogenic landforms such as thufur.

Notwithstanding the above mentioned cryogenic evidence, there are several factors which inhibit a more severe frost action environment in the high Drakensberg. The period of continuous ground freezing is restricted to the three winter months, namely June to August. Cool to mild conditions prevail for at least half the year (October to March), which are not conducive to cryogenic activity. In fact, the present study has found that most miniature cryogenic landforms are not maintained below 3200 m a.s.l. during the warmer, wetter months. Needle ice induced landforms and miniature sorted patterns are destroyed by fluvial processes during the wetter months. To this end, the study has clearly demonstrated that the cryogenic component in turf exfoliation and bank erosion is overwhelmed by other processes such as deflation, rain-splash, rain-wash and stream flow. Preliminary results of sediment yields on a Mashai Steam bank indicate a seven to eight fold increase in erosion rates during the wet period over those during the dry frost period. This is reiterated when examining bank morphology, which undergoes few changes during the winter months but may be radically altered during the wet period. The absence of adequate soil-moisture to permit frost action during the colder months is a primary cause for the restricted spatial occurrence of soil-frost phenomena. A further limitation is the general absence of fines and deep soils on the steeper slopes. Because the high Drakensberg receives much precipitation during the mild summer months, this has permitted a relatively well vegetated ground cover to about 3380 m a.s.l., which offers considerable insulation.

From the above discussion, it is apparent that the present-day high Drakensberg environment is not "mainly dominated by cryogenic, or frost action processes", an attribute which could qualify a region as periglacial (French, 1988, p xvii). Further, the processes associated with seasonal frost action and seasonally frozen ground, which occur in the high Drakensberg, are not necessarily restricted to periglacial regions (French and Karte, 1988). In conclusion, the high Drakensberg does not qualify as a periglacial region today. Only the higher summits (possibly above 3380 m a.s.l.) are geomorphically dominated by frost action processes *per se* and could, therefore, qualify as small high altitude periglacial zones within the high Drakensberg. Nevertheless, the severe frost action which is clearly the overwhelming process during the winter months, could make the high Drakensberg eligible as a marginal periglacial region, as has been agreed on, but not explained, by others (Lewis, 1988a, 1988b, Boelhouwers, 1991a, 1994).

8.2 ENVIRONMENTAL IMPLICATIONS

An objective of this study has been to formulate some broad environmental implications from the findings, some of which may be of interest and potential value to such as botanists, ecologists, geographers, geologists, range scientists and Quaternary scientists. Although discussion sections in each chapter have focused on several environmental implications pertinent to that chapter, this section aims to summarize and correlate some of the findings made. As dates have not been forthcoming for landforms which are of potential palaeoenvironmental value, much of such discussion is still somewhat speculative. Further, much concern has been expressed on the relative importance placed on relict landforms as palaeoenvironmental indicators (e.g. Hall, 1991, 1992, 1994b; Lautridou *et al.*, 1992; Barsch, 1993), particularly as some authors are flawed in their interpretations. However, if dealt with cautiously, several relict cryogenic landforms may provide valuable indications of past environments.

8.2.1 Palaeoenvironments

8.2.1.1 The Late Pleistocene

Numerous studies have argued for periglacial (and in some cases marginal glacial) conditions in the high Drakensberg during the Pleistocene (Sparrow, 1967b, 1971; Harper, 1969; Hastenrath, 1972; Butzer, 1973; Dyer and Marker, 1979; Marker, 1991; Meiklejohn, 1992, 1994). In fact, Harper (1969) proposed the existence of two cold stages during the Pleistocene, the older being of unknown age and the younger of Late Pleistocene age (i.e. Last Glacial Maximum). Many of the studies are flawed in their research approach, primarily as they lack qualitative data and are often based on presumptions (see Section **1.2.2.2**; Butzer, 1973; Hall, 1992). No pre-Holocene dates of any geomorphic landforms or sedimentary deposits from the high Drakensberg have been forthcoming. Until such dates are obtained, there can only be speculation on the age of such landforms!

The present study has identified several large relict landforms that, owing to their deep weathering and extensive vegetation cover, could be remnant periglacial features from the Last Glacial Maximum. However, it is possible that many periglacial landforms that developed during the Late Pleistocene have since decayed to an end point, owing to weathering and erosion. It is postulated that most of the blockfields/slopes, blockstreams and debris deposits are of Late Pleistocene age. Observations indicate that most of the relict landforms are structurally controlled where there is apparent high joint density in the basalt. Such areas appear to hold considerable quantities of moisture, even in the drier winter months (Boelhouwers, 1988b). Enhanced mechanical weathering at such zones may have initiated many of the block-accumulations during the Last Glacial Maximum. The so called basalt steps (Harper, 1969) are a prominent feature on the higher south-facing slopes and many of these exhibit block-accumulations on their treads. Attributes typical of cryoplanation terraces (c.f. Priesnitz, 1988; Nelson, 1989; Czudek, 1995) are present but no evidence has yet been found to support the former prevalence of nivation at such sites (Boelhouwers, 1988b).

Meiklejohn (1992, 1994) has argued for enhanced weathering and transport on north-facing slopes during the Late Pleistocene. The better insolation-protected south-facing slopes received less temperature variations and consequently underwent less weathering than north-facing slopes (Meiklejohn, 1994). However, the present study has found that both weathering and mass-wasting have been significant on south-facing slopes during colder times (see Section **6.4.2**). It would therefore appear that structural control has determined much of the morphology, including the steep south-facing slopes and the hollows on north-facing slopes. Geological structure helps control the moisture supply near or at the rock surface, which ultimately influences weathering processes and their intensity.

Based on some of the discussion above, it would appear that the Drakensberg plateau experienced periglacial conditions over much of its range during the Late Pleistocene. The block accumulations, slope and valley forms suggest enhanced mechanical weathering and frost action. The occurrence of these features would not necessarily imply perennial snow or ice at such sites during the Late Pleistocene. However, as the higher summits above 3380 m a.s.l. are today very sparsely vegetated, it is conceivable that higher altitudes during the Late Pleistocene were mostly devoid of vegetation. This may be a reason why the high altitude bogs of Lesotho only formed during post-glacial times (Van Zinderen Bakker and Werger, 1974).

It has also been postulated that localized glacial development occurred along the high Drakensberg escarpment during the Late Pleistocene, owing to favourable firn accumulation sites and considerable snowfalls during this time (see Section **6.5.4**; Hall, 1994a; Grab, 1996a). It would seem that the Drakensberg escarpment zone was considerably wetter than the high plateau on the leeward side, hence the somewhat different palaeoenvironments outlined above. Contemporaneous snowfalls are still most pronounced along the escarpment and rapidly dissipate westwards (e.g. Carter, 1976).

8.2.1.2 The Holocene

The Early Holocene was accompanied by a rapid rise in temperature and increase in the available moisture over much of southern Africa (Scott, 1982; Broecker and Denton, 1989; Partridge *et al.*, 1990). It is believed that mechanical weathering was reduced in the high Drakensberg during this time (Boelhouwers, 1988b). It would appear that vegetation stabilized on the high Drakensberg during the Early Holocene, as Marker (1994, 1995b) has identified the first organic phase between 13 500 and 9 000 yrs BP. Many older block-accumulations show rock discolouration and extensive lichen and vegetation growth, suggesting that chemical and biological weathering have been prevalent throughout much of the Holocene. From the organic deposits found at high altitudes (e.g. Hanvey and Marker, 1994; Marker, 1994, 1995b), it would appear that the high Drakensberg did not have a periglacial zone for extended periods during the Holocene.

An extensive literature review by the present author could not find a single study that has argued for Holocene cold phases in the high Drakensberg. It has been customary to presume that relict landforms in the high Drakensberg are of Pleistocene age. Yet, as reviewed in Jerardino (1995), Late Holocene Neoglacial episodes were important events in southern Africa, with possible temperature departures of up to 3°C. It has now been argued (see Sections **4.6.2** and **6.3.2.3**) that some relict periglacial landforms such as sorted circles and stone-banked lobes are remnants of such Late Holocene Neoglacial events. This would imply a re-emergence of a periglacial zone, possibly down to approximately 3200 m a.s.l. This implies a slight depression in the vegetation belts and a possible absence of vegetation above about 3300 m a.s.l. during some of the cooler Neoglacial episodes. It is conceivable that mechanical weathering was enhanced during these cold periods, which may account for the relatively fresh appearance of stone-fields on Mafadi Summit. Clearly, however, there is an urgent need to substantiate some of the inferences made, possibly by providing more specific dates.

8.2.2 Present-day Environments

Several present-day environmental implications have emerged from the current project:

1. It is found that altitudinal differences do not determine the size and development of active cryogenic landforms such as sorted patterned ground and thufur. However, it is evident that sorted patterns below about 3200 m a.s.l. are destroyed during the summer months. Such findings could have valuable implications on the present high Drakensberg climate, particularly as this is poorly understood. It would appear that temperatures are only marginally depressed from about 2900 to 3482 m a.s.l. during the winter months, owing possibly to inversion. However, during the warmer months the lapse rates are likely to be strengthened considerably, allowing for the preservation of cryogenic phenomena at higher altitudes. This could have implications for vegetation establishment and growth, hence also the abrupt decline in ground cover from about 3380 m a.s.l.

2. The distribution of relict and active cryogenic landforms, together with the studies on thufur elongation/coalescence, thufur breakup and turf exfoliation, have shown that aspect exercises considerable control on the geomorphic processes. In fact, as mentioned in Section **2.6.3**, Granger and Schulze (1977) have measured radiation receipts as much as 470% more on north-facing than on south-facing slopes in mid-winter. Because the high Drakensberg is today a marginal periglacial region, aspect significantly controls freezing and thawing processes. This is well illustrated in Figure 8.1 which shows snow patches on a south-facing slope, while the north-facing slope is almost completely dry.

3. The research findings suggest pronounced temperature differentials between thufur apexes and depressions in this marginal periglacial environment. Temperature records provide valuable evidence that such temperature differences are most pronounced in winter when thufur apexes are frozen and depressions generally unfrozen at depths below 5 cm. It is possible that, owing to the restricted freeze intensity and duration, thufur formation in marginal periglacial environments develop under a different suite of ground thermodynamics than do similar landforms commonly referred to as "earth hummocks" in permafrost environments. It appears that existing micro topography (e.g. thufur) is an important factor controlling the maintenance

Figure 8.1 Aspect significantly controls the freezing and thawing processes in the high Drakensberg. For instance, snow lasts considerably longer on the south-facing slopes than on the north-facing slopes, as is displayed in the photograph. The photograph was taken near the Mashai Valley in June, 1990, three weeks after a snowfall.

or further development of landforms in marginal periglacial regions by creating "frozen pockets" in an otherwise predominantly unfrozen environment. Further long-term thermal studies on thufur and earth hummocks are essential to better understand the dynamics of similar landforms in apparently dissimilar ground thermal environments.

4. As snowfalls are usually both infrequent and light during the high Drakensberg winter period, this enables considerable ground freezing to occur to depths of 30 cm or more. However, during years when a thick snow cover insulates the ground for over 50% of the winter period, this significantly reduces the frequency of needle ice events, and consequently, the potential for disruptive ground frost activity. Thufur thermal properties during the four year (1993 - 1996) logging period have indicated that snow exercises considerable control on the freezing processes within thufur. It is concluded that the presence of snow in marginal periglacial regions such as the high Drakensberg is a primary control on the efficacy of frost action processes.

5. It has emerged from this study that lithology (i.e basalt joint structures) significantly determines the distribution of geomorphic forms and wetlands. Because of the pronounced valley asymmetry (Meiklejohn, 1992, 1994), wetlands most frequently emerge on the gentler north-facing slopes (Grab, 1994). Here, moisture is readily available even during the winter months, enabling the establishment of cryogenic phenomena such as thufur and elongate ridges. The restricted occurrence of thufur to wetlands is largely a function of arid climatic conditions which have impeded widespread cryogenic landform development in the high Drakensberg. The cryogenically-induced wetland microtopography has an important function to disperse and absorb flow when the water table is above the surface (Morris and Grab, 1997), thus preventing excessive soil and water loss from the wetland systems. The breakup of thufur and eventual disappearance of such micro-forms could severely disrupt the wetland systems by promoting channel flow, which consequently enhances soil loss and irregular flow out of the wetlands. Further, the thufur may potentially provide valuable indications of present-day environmental change. However, it appears that anthropogenic interference, rather than climatic change, is currently causing most thufur disruption. Although the value of thufur has been little understood, there is now little doubt that the high Drakensberg thufur are a priceless ecological asset to the region and its surrounds.

6. It has been shown that the high Drakensberg has an exceptionally high frequency of needle ice development, with possibly between 120 and 150 annual events, despite an absence of fines at most places where it occurs. Results from the Mashai Valley confirm that needle ice usually develops when the near-surface ground temperature approaches -1.0°C, yet, needle ice development was recorded when ground temperatures were as high as -0.4°C and the air temperature was above 0°C. It may be concluded that rigid environmental parameters are very difficult to define for needle ice development, as factors such as existing ground thermal properties, soil texture and moisture and wind velocity, may exercise a synergistic control on needle ice development.

7. The study has found that needle ice processes contribute towards particle movement and soil erosion in the high Drakensberg, albeit secondary to water-induced erosion. Turf exfoliation is another natural soil erosion process, which is typically found in alpine regions

with temporal precipitation patterns. Zones of accelerated turf exfoliation have been identified and attributed to a greater moisture supply near valley bottoms. It has frequently been presumed that the sparsely vegetated and gravel/rock dominated higher summits are a product of overgrazing. The presence of active perennial soil frost phenomena on such summits would suggest that a sufficiently cold climate throughout the year has impeded the establishment of vegetation. In fact, the restricted occurrence of Basutos and their livestock on such higher slopes appears to have had little repercussion on erosion rates there.

8.3 THE FUTURE OF SOUTHERN AFRICAN PERIGLACIAL GEOMORPHOLOGY

Corté and Hall (1991) and Hall (1992) have emphasised the potential of southern African periglacial geomorphology of providing a valuable contribution to the understanding of the world's periglacial regions. However, Hall (1992) has shown that knowledge pertaining to the southern African periglacial environment is rudimentary, and that greater rigour is required to avoid unqualified presumptions of cryogenic processes and resultant landforms. More recently, Boelhouwers (1995a) has reiterated Hall's concerns:

> "Unless field studies become more quantitative and process orientated, we can have little hope that our understanding of local frost-action processes, past or present, will substantially improve". (Boelhouwers, 1995a, p164)

To this end, it has been an objective of this study to identify some of the research shortcomings (see Section **1.2.2**) and to improve on these by providing quantitative assessments on landforms and processes operative in the high Drakensberg.

Barsch (1993) and Haeberli (pers. comm.- 1997) have identified several research needs which would ensure a prospective future for international periglacial geomorphology:

1. More process studies.
2. A greater contribution examining geomorphic response to environmental change.
3. More information on relict periglacial landforms so as to improve palaeoclimatic reconstructions.
4. The prediction of future development in periglacial regions in the context of projected global change.

Similarly, Boelhouwers (1995a) has followed international trends to identify the following topics as the main objectives of southern African endeavours: mass wasting, weathering, patterned ground formation, frost wedging and cracking, ground ice, zonation, modelling and palaeoenvironmental reconstructions. The present study has already focused on several of these to a greater or lesser extent. However, there remains much potential in the high Drakensberg and Lesotho mountains for periglacial research that focuses on the needs identified by Barsch (1993), Haeberli (pers. comm.- 1997) and Boelhouwers (1995a).

Future process studies on such as thufur development and the growth and decay of needle ice in the high Drakensberg could make a valuable contribution to global periglacial research. Further, the region is occupied by landforms such as blockstreams that are internationally considered as problem features owing to the uncertainty of their developmental origin (Harris, 1988). The relative abundance of blockstreams in some high Drakensberg areas could make these potential research targets of substantial international value. The variety of relict cryogenic landforms in the high Drakensberg have a potential value in palaeoclimatic reconstructions, and hence also to improving the knowledge of Quaternary environments in highland Lesotho. Similarly, it is still not certain how the high altitude southern African climates and environments are changing in response to global changes. As cryogenic landforms such as thufur are useful in the reconstruction of short-term climatic fluctuations (Priesnitz, 1988), these could contribute towards improving the knowledge of localized changes.

There is an increasing need for environmental impact assessments in Lesotho, owing to several projects including:

1. The Highlands Water Project
2. The Highlands Road Construction Project
3. Plans for hotel construction and the development of a tourist industry.

Important information on the distribution, rates, quantities and processes of soil erosion are necessary for many highland regions where such projects are being planned. Further, the loss of Lesotho's alpine wetlands is a potentially serious ecological and economic problem to Lesotho and South Africa (Backéus and Grab, 1995; Morris and Grab, 1997). The importance of cryogenic processes and landforms within the context of soil erosion and wetland ecology has briefly been examined in this dissertation. Clearly, it emerges that future environmental impact assessments must take cognisance of such cryogenic processes and landforms if they are to fulfil their objectives adequately. In marginal periglacial areas such as the high Drakensberg, the present geoecology is influenced and in some way dependent on cryogenic processes and landforms (e.g. Barsch, 1993), and therefore, cannot be ignored in future projects planned for highland Lesotho.

Given the above mentioned research needs and potential for applied work in the marginal periglacial regions of the high Drakensberg and Lesotho mountains, geomorphologists with a sound periglacial knowledge should promote themselves better to create research and work opportunities in such areas. Should southern African workers follow some of the guidelines outlined in this section, then there is certainly a future ahead for southern African periglacial geomorphology.

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