SEDIMENTARY MODELS FOR COAL FORMATION IN THE KLIP RIVER COALFIELD

by

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PREFACE

This thesis represents original work by the author and has not been submitted in part, or in whole to any other university. Where use is made of the work of others it is acknowledged in the text.

The research was conducted in the Department of Geology and Applied Geology, University of Natal, Durban, under the supervision of Professor R. Tavener-Smith.

IN MEMORY OF MY FATHER,
DAVID ALEXANDER CHRISTIE

ACKNOWLEDGEMENTS

ABSTRACT

The primary objective of this study was to establish sedimentary models for peat formation in the southern part of the Klip River coalfield during Ecca (Permian) times and to assess palaeoenvironmental controls on coal seam behaviour and distribution. In order to achieve this approximately 2 400 borehole logs and 25 field sections were collected.

The coal-bearing Vryheid Formation records early to late Permian fluvio-deltaic sedimentation within the northeastern main Karoo basin. Three informal lithostratigraphic subdivisions, based on the investigations of Blignaut and Furter (1940, 1952), are proposed : the Lower zone, Coal zone and Upper zone.

An examination of the structural framework and history of the northeastern Karoo basin reveals that the southern and western boundaries of the Klip River coalfield are defined by zones of rapid basement subsidence : the Tugela and Dannhauser Troughs respectively. There is some doubt as to the locality of the source area to the rivers emptying into the Ecca sea. Ryan (1967) postulated the "Eastern Highlands" situated off the present southeast African coast, but it is contended that the Swaziland area, situated no more than 200 to 300 km to the northeast of the Klip River coalfield, constituted a more plausible source area.

The Lower zone represents sedimentation along a westerly to southeasterly prograding coastline dominated by high-constructive lobate or braid deltas, but also showing significant influence by wave processes. The Coal zone, which varies in thickness from 35 to 60 m, represents a major phase of coastal progradation and braided-river deposition on extensive alluvial plains. Significant coal seams formed only during periods of fluvial inactivity, the duration of which was dependent on source-area processes.

Coal seam geometry and behaviour in the Klip River coalfield were not influenced by the depositional environments of associated clastic sediments. The following factors were found to have of profound influence in determining the extent, distribution and rate of peat accumulation:

1.   Platform stability and temporal and spatial variations therein.

2.   The absence or presence of penecontemporaneous clastic sedimentation.

3.   Duration of periods of peat formation.

4.   Lithology and topographic expression of clastic sediments underlying peat-forming swamps.

The peat-forming phase of the Vryheid Formation was terminated by an extensive transgression brought about by an eustatic rise in basin water-level and/or an increased rate of platform subsidence.

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CHAPTER 1

INTRODUCTION

1.1 GEOLOGICAL SETTING AND GEOGRAPHY

The Karoo Sequence of South Africa records late Palaeozoic to early Mesozoic deposition within the intracratonic main Karoo basin (Fig 1.1). In the northeastern part of the basin, namely southeastern Transvaal and northern Natal, some of the main coal resources of South Africa are contained within the Permian Vryheid Formation of the Ecca Group. Arbitrary boundaries have been used to subdivide the area into, amongst others, the Highveld, Witbank, Eastern Transvaal, Klip River, Utrecht and Vryheid coalfields (Fig 1.2).

The Klip River coalfield, the southern part of which this thesis is concerned, traditionally includes the whole of the coal-bearing area between Ladysmith and Newcastle in northern Natal. The part here investigated is situated in the southern part of this coalfield between the villages of Alcockspruit, Elandslaagte and Helpmekaar (Fig 1.3). It is bounded in the east by the Buffalo River and in the west by the foothills of the Drakensberg and includes portions of the Dundee, Glencoe, Klip River, Dannhauser and Newcastle magisterial districts. Dundee, the largest town in the area, as well as Glencoe and Dannhauser are mainly coal mining and agricultural communities.

The national road between Johannesburg and Durban passes through the western portion of the study area and links Newcastle with Ladysmith. Road communication within the area is generally very good, being provided by many tarred and gravel roads and farm tracks. The main Johannesburg - Durban railway passes through Wasbank, Glencoe and Dannhauser. A railway also links Glencoe and Dundee.

The mountainous terrain of the Drakensberg foothills in the west gives way to rolling countryside in the northern and central parts of the study area
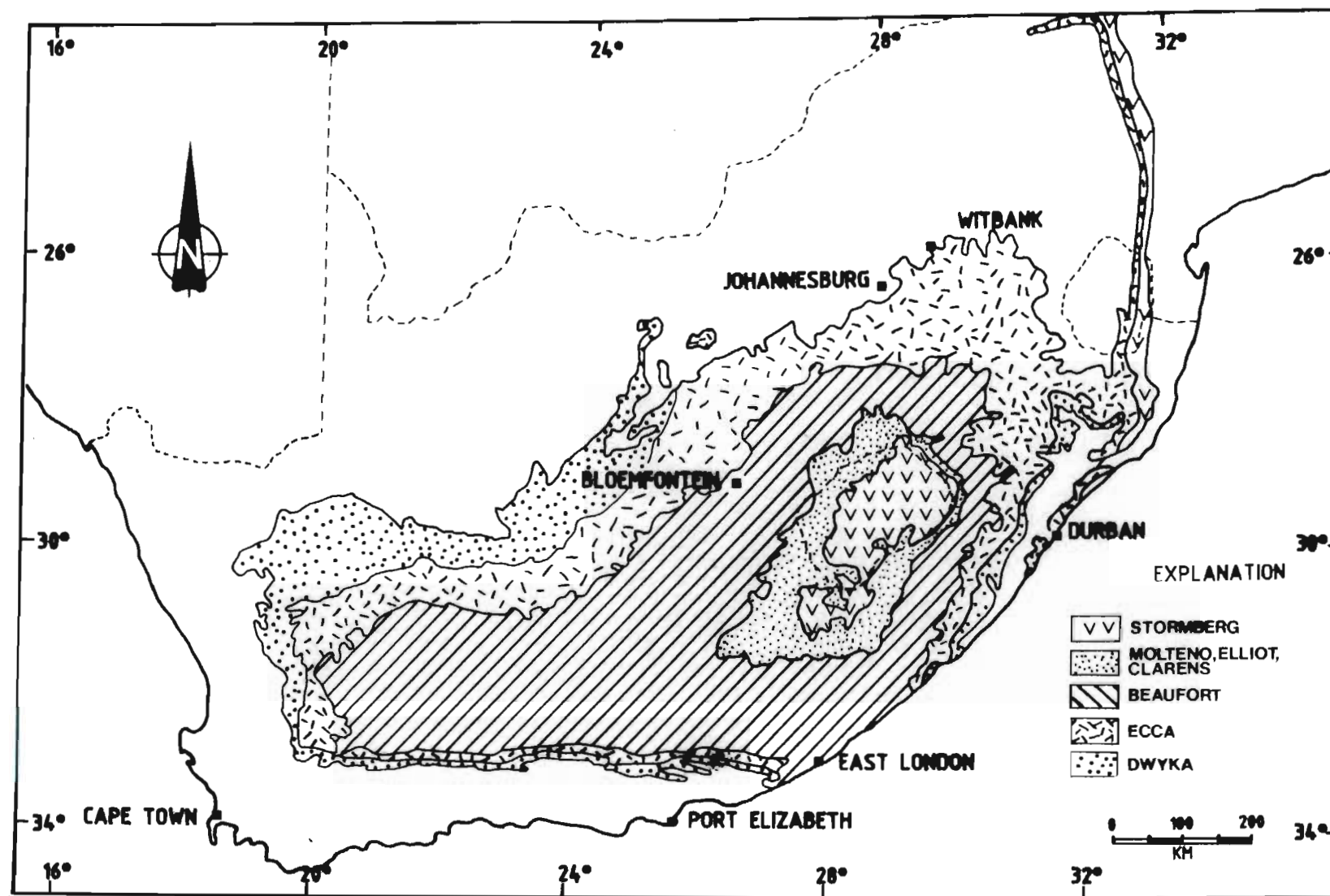
Figure 1.1. Distribution of sediments within the main Karoo basin, South Africa.
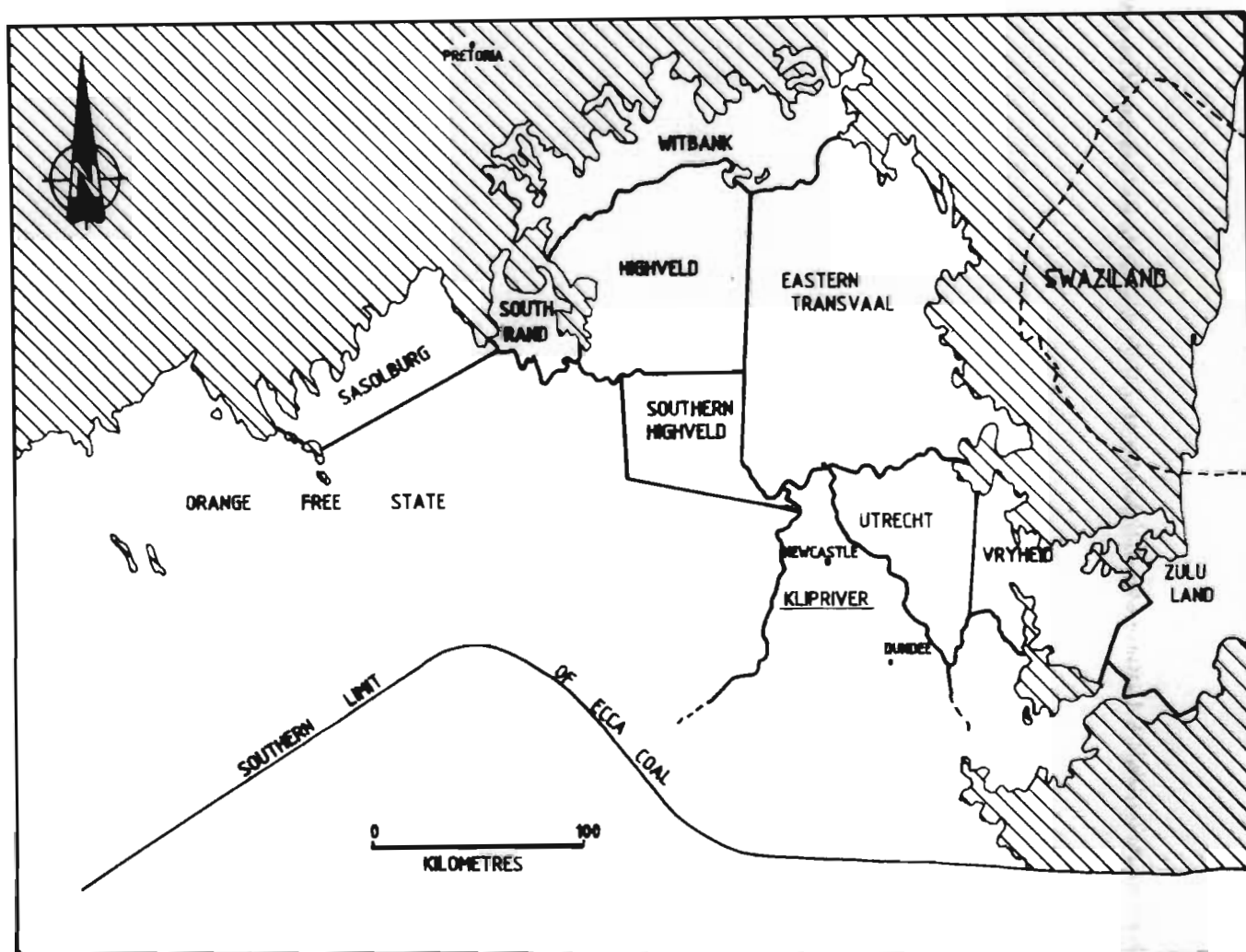
Figure 1.2. Distribution of coalfields within the northern and northeastern parts of the Karoo basin.
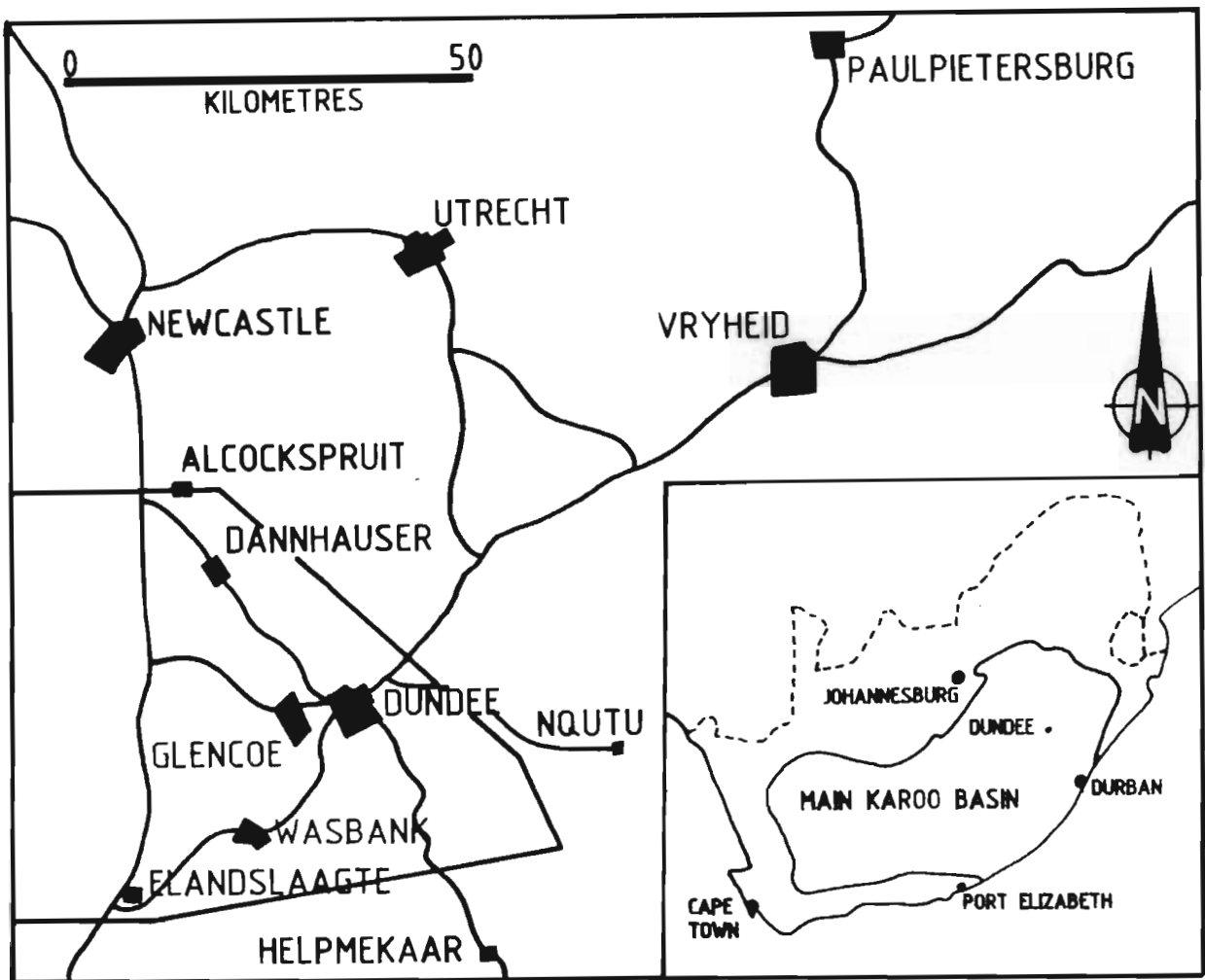
Figure 1.3 General locality map of the thesis area showing its boundaries, main towns and communications.

which are at an average altitude of 1 300 m above sea level. To the west and south of Dundee the Biggarsberg range of mountains, varying in altitude from 1 300 to 1 700 m, form a northwesterly - southeasterly trending spur of the Drakensberg. These mountains are deeply dissected by streams draining into the broad, flat valleys of the Buffalo and Wasbank Rivers. The southerly flowing Buffalo River drains the entire northern, central and eastern portions of the area. Numerous streams drain the remainder of the western and southern portions, the main ones being the Wasbank, Toleni and Sundays rivers. All of these are tributaries of the Tugela River which flows in an easterly direction to the Indian Ocean.

A continental type climate prevails with mean daily temperatures of 23°C during summer and 15°C during winter. Extremes in excess of 30°C and -5°C occur, and snow may fall during particularly cold winters. The area receives an average annual rainfall of 600 mm which falls mainly during the summer months. The vegetation is of a temperate grassland type with dense indigenous bush in river valleys. Acacias are the dominant indigenous trees while aloes are common on mountain slopes, particularly those underlain by dolerite. Exotic poplar, pine, gum and wattle trees are commonly found in small plantations.

1.2 AIMS AND APPROACH

The primary objective of this study was to establish sedimentary models for peat formation in the southern part of the Klip River coalfield during Ecca times and to assess palaeoenvironmental controls on coal seam behaviour and distribution. The economic relevance of this project rests on the premise that an understanding of coal measure stratigraphy and sedimentology in the Klip River and other coalfields will provide a positive contribution to the coal industry. Knowledge gained will be useful as a predictive tool in regional coal exploration and seam correlation. Results relating to palaeodrainage directions, seam splitting and shaling-out, and roof-rock lithology may be

applicable to mine planning and mining procedures in this coalfield and elsewhere.

Between July 1981 and August 1982 data were collected in the form of borehole records from a number of mining and exploration companies, by logging borehole cores and measuring detailed vertical sections from stream valleys, cliffs, and underground and opencast coal mines. Approximately 2 400 borehole logs and 25 field sections were collected and transcribed into a standard format. Each borehole log consists of a graphic log displaying grain size variations and sedimentary structures alongside which are detailed (where possible) descriptions of sedimentary features. Most of the boreholes were originally logged in a superficial manner by company geologists and surveyors; they provided only a rudimentary indication of lithology and only very rarely was any reference made to sedimentary structures. Furthermore, less than 1,5% of the boreholes penetrated more than 50 m below the coal seams and therefore very little information about the underlying strata was available. These shortcomings were overcome to some extent by making reference to field sections measured in great detail and a small number (about 30) of personally logged borehole cores obtained from current drilling projects. This data was used as a standard by which the rest of the borehole logs were calibrated.

Boreholes are distributed throughout most of the area but are commonest in the northern, central and southwestern parts (which coincides with the area underlain by mineable coal). Most of the outcrop sections were measured to the south and southeast of Dundee, some in the vicinity of Glencoe and Wasbank, and two near Alcockspruit (Fig 1.4). Measurement was by hand level and tape measure.

The large and varied data base presented numerous problems with respect to data handling and analytical procedures. It was nevertheless decided at an
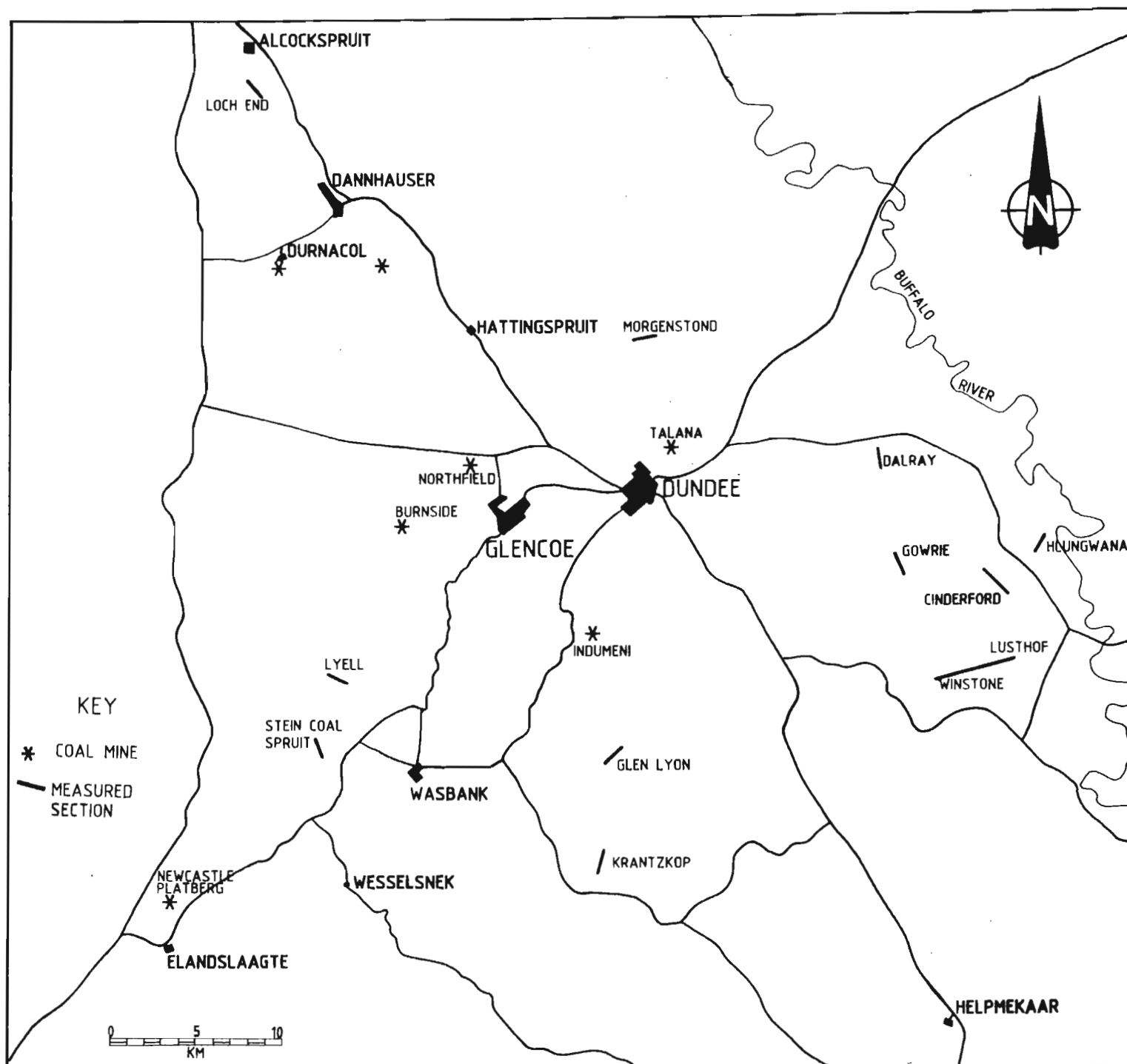
Figure 1.4. Detailed locality map of thesis area showing active and abandoned mines and locality of outcrop sections.

early stage not to resort to computer techniques for data storage and retrieval, lateral correlations or the compilation of isopach maps. This decision was taken despite the fact that computer techniques are widely regarded as essential tools in coal exploration and were used with apparent success by Le Blanc Smith (1980b) in coal seam correlation in the Transvaal coalfields and in compiling isopach maps to illustrate coal seam distribution. The main reasons for this decision were that a large part of the data was already in an easily readable and accessible format and that much time would have been required for data encoding, capturing and verification. Other considerations included the crude appearance of computer drawn graphic logs, inability of the computer to recognize lateral facies changes, the necessity for manual correlation even if a computer was utilised and the ultimate requirement for only a small number of isopach maps which could be more quickly prepared by manual methods.

The following procedure was adopted to determine palaeoenvironmental controls on coal seam distribution and thickness:

1.   Recognition and description of lithofacies from borehole logs and measured sections.

2.   Analysis of vertical and lateral relationships of these lithofacies and the establishment of sedimentary facies models.

3.   Use of isopach maps and cross sections to illustrate the distribution and geometry of certain critical lithological units (including coal seams).

4.   Palaeogeographic and palaeoenvironmental reconstruction showing the relationship between the various sedimentary models with reference to palaeocurrent dispersal patterns and pre-Karoo topography.

5. Correlation of coal seams and the establishment of a uniform nomenclature throughout the coalfield.

6. Establishment of sedimentary models for peat formation.

7. Determination of the principal factors governing peat accumulation and derivation of an understanding of the relationship between depositional milieu and coal seam distribution.

CHAPTER 2

PREVIOUS INVESTIGATIONS

2.1 STRATIGRAPHIC AND SEDIMENTOLOGICAL INVESTIGATIONS

The earliest published account of the Klip River coalfield was that of F.W. North in 1881. He measured 72 vertical sections through the coal measures and inspected numerous coal outcrops in the counties of Klip River, Weenan, Umvoti and Victoria. The principal object of this report was to investigate the feasibility of using local coals rather than importing British coal to power steam engines. North equated the Klip River coals with those of the Stormberg coalfield (Molteno Formation) in the northeastern Cape Province.

E.J. Dunn (1886) was perhaps the first to suggest that the northern Natal coals belonged to the Ecca "Series", a view supported by Anderson in 1901. This was confirmed by du Toit (1918) in his description of the zones of the "Karroo System". He also recognized a three-fold subdivision of the Ecca, the Lower, Middle and Upper "stages", and reported on their distribution and lithology. Use of the terms Lower, Middle and Upper Ecca to subdivide the Ecca Group persisted for many years. With the introduction of a standardised stratigraphic nomenclature system by the South African Council of Stratigraphy (SACS) in 1977 these are now called the Pietermaritzburg, Vryheid and Volksrust Formations respectively (SACS, 1980).

Two comprehensive accounts of the Klip River coalfield were published by F.A. Steart and W.J. Wybergh in 1920 and 1925 respectively. These considered such aspects as the stratigraphy of the "Karroo System", the lithology and thickness of the "Coal Measures" or "Middle Ecca" and the thickness, number, distribution, behaviour and quality of the associated coal seams. A notable advance in understanding the "Middle Ecca" was made by Blignaut and Furter (1940, 1952); they formulated an informal five-fold lithostratigraphic subdivision in the Vryheid, Utrecht and Newcastle areas (Table 2.1).

TABLE 2.1 - STRATIGRAPHIC SUBDIVISION OF THE ECCA GROUP IN THE NORTHEASTERN

KAROO BASIN, WITH PARTICULAR EMPHASIS ON THE VRYHEID FORMATION

KAROO SEQUENCE (SACS, 1980)	ECCA GROUP			
DRAKENSBERG GROUP CLARENS FORMATION ELLIOT FORMATION MOLTENO FORMATION BEAUFORT GROUP ECCA GROUP DWYKA FORMATION PRE-KAROO BASEMENT	Du Toit (1918)	Blignaut and Furter (1952)	SACS (1980)	This Study (1988)
	UPPER ECCA STAGE		VOLKSRUST FORMATION	
	MIDDLE ECCA STAGE	Upper Transition Zone Upper Sandstone Coal Zone Lower Sandstone Lower Transition Zone	VRYHEID FORMATION	Upper Transition Zone Upper Zone Coal Zone Lower Zone Lower Transition Zone
	LOWER ECCA		PIETERMARITZBURG FORMATION	

A regional analysis of the Karoo basin by Ryan (1967) made particular reference to sediment transport directions, major provenance areas and factors controlling sedimentation. On the basis of lithology and palaeocurrent directions Ryan subdivided the Ecca into southern, western, northern (containing the Natal and Transvaal coalfields) and central facies. He also suggested that deposition of the "Middle Ecca" was by fluvial and deltaic processes influenced by extensive marine transgressions and regressions, and recognised the strong control of palaeogeography and palaeotopography on peat formation.

Modern sedimentological techniques were applied by Hobday (1973) in the analysis of the Vryheid Formation in the Muden-Tugela Ferry area. He recognised 7 distinct lithofacies and compiled a detailed composite section showing vertically recurrent upward-coarsening fluviodeltaic cycles. Subsequent sedimentological investigations of the Vryheid Formation have taken a similar approach and focused on the cyclicity and physical controls of sedimentation. These were able to provide some detailed descriptions of depositional environments. Hobday and Mathew (1975) and Hobday et al. (1975) described and interpreted cyclic sequences in the Vryheid-Hlobane area in terms of fluviodeltaic environments; Mason and Tavener-Smith (1978) identified fluviodeltaic deposits in the Empangeni district; Tavener-Smith (1979) published a short account of regional controls on coal seam distribution in the northeastern Karoo basin; Whateley (1980) emphasised structural (tectonic) controls on fluvial and deltaic sedimentation in northern Zululand; Christie et al. (1982) described a 310 m fluviodeltaic sequence from Ceza Mountain in Zululand and compared it to similar deposits at Vryheid and Nongoma. The recognition by Tavener-Smith (1982; in prep.) of coastal beach-barrier and lagoon deposits near Durban and giant cross-beds near Nqutu, Zululand has added significantly to the understanding of wave and tidal influences on deposition, coastline geometry and distributary mouth processes during Vryheid Formation time.

Sedimentological investigations in the Transvaal coalfields (Le Blanc Smith and Erikson, 1979; Cairncross, 1980; Le Blanc Smith, 1980a,b; Cadle, 1982) have resulted in a detailed description of Vryheid Formation stratigraphy there and a realisation that peat accumulation and coal seam distribution are strongly influenced by pre-Karoo topography, depositional environment and differential compaction of underlying strata. However, despite the progress that has been achieved in that part of the Karoo basin, the Klip River coalfield has remained largely neglected. The lack of published data is attributed to limited outcrop within the main coal-bearing area, the confidential nature of borehole data held by the various mining and exploration companies and the gradually waning importance of the area as a coal producer. To date, the most comprehensive account of the coalfield remains the explanation of the Newcastle, Dundee, Elandslaagte and Utrecht geological map compiled by Visser and Bishopp (1976) dealing mainly with coal-seam thickness and quality. Van Vuuren and Cole (1979) and van Vuuren (1981) included the northern and western portions of the Klip River coalfield in a regional subsurface sedimentological investigation of the Ecca Group in the northeastern Karoo basin. They described regionally correlatable regressive cycles of deltaic and fluvial origin and provided broad models for peat formation. They also iterated the strong influence pre-Karoo topography had on sedimentation and the relationship between coal seam behaviour and depositional environment. This investigation confirmed that a subdivision of the Vryheid Formation described from other parts of the basin, comprising upper and lower deltaic sequences separated by a fluvial sequence, is also recognisable within the Klip River coalfield.

2.2 STRATIGRAPHIC SUBDIVISION OF THE VRYHEID FORMATION

The lithology and sedimentology of Vryheid Formation strata within most parts of Transvaal, Natal and Zululand have been rigourously examined in recent years but the recognition and correlation of stratigraphic subdivisions remains

a major problem. All schemes proposed so far have relied to a large extent on the most readily identifiable elements of the Vryheid Formation: coal seams and fluviodeltaic cycles. Attempts to correlate on the basis of cycles are complicated by their diachronous nature and known tendency to lateral discontinuity. The potential of using regional coal seam behaviour as a basis for stratigraphic procedures has, however, never been examined in detail in northern Natal.

Van Vuuren and Cole (1979) recognised up to 8 regressive cycles of sedimentation in the north Karoo basin, the base of each defined by a transgressive surface. They regarded the transgressions to be of regional extent and therefore suitable marker surfaces for correlation. Current research in northern Natal and Zululand has shown, however, that great difficulty is experienced in attempting to correlate using sedimentary cycles and transgressive surfaces at localities only a few kilometres apart. This suggests that transgressions were of only local extent, a view supported by Tavener-Smith (1979, 1983) who argued that local variations in foundational instability within the basin preclude the use of basin-wide stratigraphic markers.

Le Blanc Smith (1980b) used the concept of genetic increments of strata (Busch, 1971) to subdivide the Vryheid Formation in the Witbank coalfield. He regarded the lithostratigraphic subdivision proposed by SACS (1980) to be inappropriate as the Pietermaritzburg Formation is not deposited in the coalfield and the Volksrust Formation is absent because of present day erosion. Ten marker surfaces, presumed to have time-lithologic properties, were selected to divide the sequence into genetically-related stratal increments. The lowest marker surface was defined as either the upper limit of the Dwyka diamictite or the pre-Karoo basement and the remainder comprised coal seams and/or glauconite-bearing sandstones (indicative of regional marine transgressions). Busch (1971), in his original concept of genetic increments of strata, utilised

bentonite-bearing sediments (indicative of volcanic plinean activity) to define time-lithologic marker surfaces. Coal seams, on the other hand, have been shown to be stratigraphically diachronous (Trueman, 1946), a point stressed by Tavener-Smith (1983). Although Le Blanc Smith emphasised the need to take congnizance of the strata between the coals as well as the coals themselves in stratigraphic correlation, this approach appears to be heavily dependent on coal seams as chronostratigraphic markers and the "genetic increments of strata" are essentially sedimentary cycles in a different guise. This approach was, however, successful in recognising lateral variations in sedimentary environments and breaks caused by pre-Karoo basement highs and penecontemporaneous fluvial erosion.

It became evident early in the present investigation that any practical subdivision of the Vryheid Formation in the Klip River coalfield would, like other coalfields, depend to a large extent on the accurate correlation of both the coal seams and the main coal-bearing zone. A complication in this respect was that very little information about the lower and upper boundaries of the Vryheid Formation (with the Pietermaritzburg and Volksrust Formations respectively) was available. There was, however, sufficient data to recognise 3 lithostratigraphic subdivisions within the southern part of the coalfield, each of which was initially named according to the recommendations put forward by SACS (1980). These were, in ascending stratigraphic order, the Dundee, Glencoe and Dannhauser Members (Christie, 1984). The Glencoe Member contained the coal deposits and was defined by the lowermost and uppermost areally persistent coal seams, the No 1 and No 4 Seams respectively (seam nomenclature discussed in Chapter 9). It soon became apparent that the introduction of a new nomenclature, applicable only to a small part of the north Natal coalfields, would introduce further confusion with respect to the regional stratigraphy. Furthermore, these 3 members are of a similar lithology and stratigraphic position to the Lower sandstones, Coal zone and Upper sandstones recognised by Blignaut and Furter (1952) in the Newcastle-Utrecht area. Consequently, it was

decided to use Blignaut and Furter's terminology in a slightly modified form and the informal terms Lower zone, Coal zone and Upper zone were adopted for the purposes of this thesis (Table 2.1). The term "zone" is preferable to sandstones as it appears that the original term applied only to specific sandstones in the lowermost and uppermost cycles. Although the terminology is not in strict accordance with SACS (1980) guidelines, provision is made there for long established stratigraphic names. In the absence of detailed regional stratigraphic descriptions, especially with respect to the lateral continuity of individual coal seams, the informal lithostratigraphic subdivisions proposed for the area under investigation are probably the most suitable as they:

1. Are recognisable in most parts of northern Natal, Zululand and eastern Transvaal,

2. do not rely on poorly understood and documented marker horizons, but rather on easily recognisable vertical changes in lithology,

3. are clearly expressed geomorphologically in many areas,

4. are readily related to the genetic subdivisions proposed by earlier workers,

5. are familiar to most workers concerned with the Ecca Group, and

6. can be easily changed at a later date to accord with a regional scheme of nomenclature.

Detailed descriptions, definitions and correlation of the subdivisions on a regional scale would be required prior to the establishment and acceptance of a formal terminology.

CHAPTER 3

TRACE FOSSILS

3.1 INTRODUCTION

Because of the rarity of invertebrate body fossils in the Ecca Group, locally abundant trace fossils or ichnofossils assume an important role in sedimentological investigations (Hobday and Tavener-Smith, 1975; Stanistreet et al., 1980; Turner et al., 1983; Mason et al., 1984). Trace fossils have been studied by European and North American workers since the turn of the century, but it is only in recent years that they have attracted the attention of South African geologists. The increased interest in biogenic sedimentary structures is due to the environmental or facies approach to the study of sedimentary rocks.

Ichnology may be defined as the study of trace fossils made by organisms including their description, classification and interpretation (Simpson, 1975). A trace fossil is an individually distinctive biogenic structure related more or less directly to the morphological parts of the organism that made it and includes such features as tracks, trails, burrows, borings, gnawings and pellets (Frey, 1978).

Very similar traces may be produced by different organisms but the opposite is also true. An individual species will make different traces when engaged in different activities (crawling, feeding, resting etc.) or moving over or in substrates of differing textures (Howard, 1978). This does not detract from the fact that trace fossils are valuable aids in facies reconstruction and the definition of physical environmental parameters. Detailed studies of modern sediments clearly illustrate the direct behavioural responses or activities of organisms to climatic and hydrodynamic environmental conditions (Schafer, 1972). For this reason Howard (1978) regarded trace fossils as primary sedimentary structures. Additional advantages of using trace fossils as aids to

environmental interpretation are that they have long time ranges and are entirely autochthonous (Frey, 1975, 1978; Seilacher, 1967, 1978).

Trace fossils may be classified by studying aspects of their preservation (stratinomic method), the behaviour of organism which made the trace (ethological method) or by considering the identity of the organism that produced the trace (taxonomic method) (Simpson, 1975). The ethological classification, devised by Seilacher (1953), is the most appropriate, easily applied and generally accepted. It also assists with the interpretation of the depositional environment which prevailed at the time the trace was formed.

3.2 SYSTEMATIC ICHNOLOGY

Ichnogenus SIPHONICHNUS Stanistreet et al., 1980

Remarks: Comparable structures are produced by the bivalve Mya arenaria in response to growth, sedimentation and erosion (Reineck, 1958). Stanistreet et al. (1980) recommended that all these structures be included within the same ichnogenus.

Ichnospecies SIPHONICHNUS ECCAENSIS Stanistreet et al., 1980

Description: Vertical to subvertical unbranched tubes up to 250 mm in length containing concave-downward meniscae pierced by a central tube. The outer tube (cortex) is between 5 and 20 mm in diameter and the central tube is up to 8 mm in diameter (Fig 3.1).

Tubes are commonly closely spaced and may be interpenetrant, but all gradations from dense to sparse distributions occur. Siphonichnus occurs in sediments ranging from dark, sandy mudrock to coarse-grained, pebbly sandstone. It is, however, most common in silty, fine- to medium-grained sandstone which may have been previously heavily bioturbated.
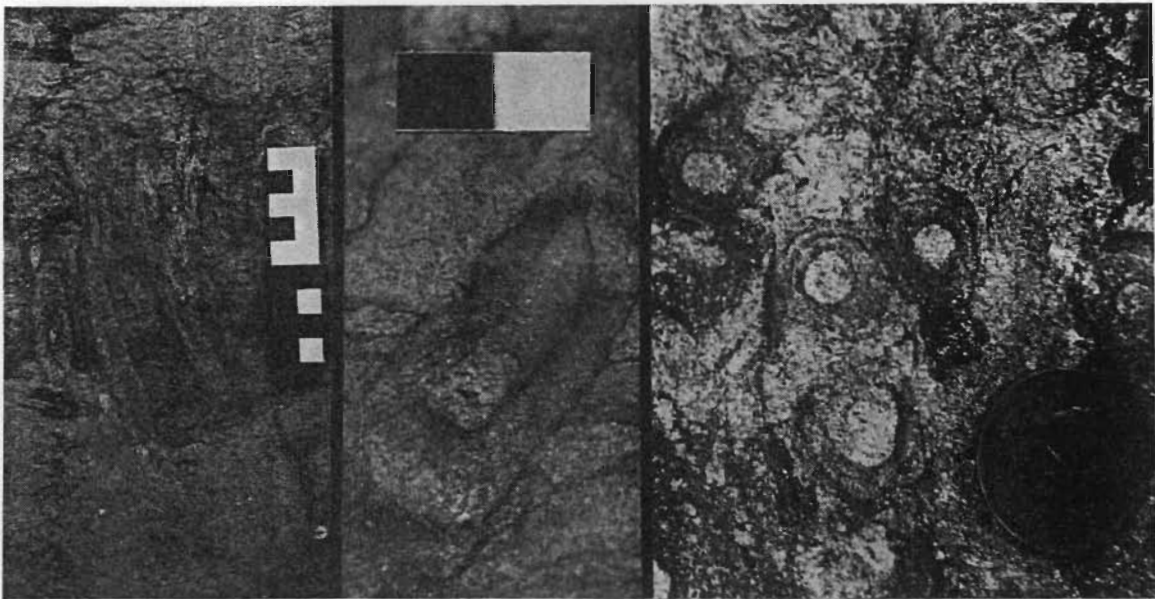
Figure 3.1.   _Siphonichnus_ _eccaensis_. Illustrating, from left to right, elevation, three-dimensional and plan sections of the ichnofossil. Scales divisions: 10 mm.
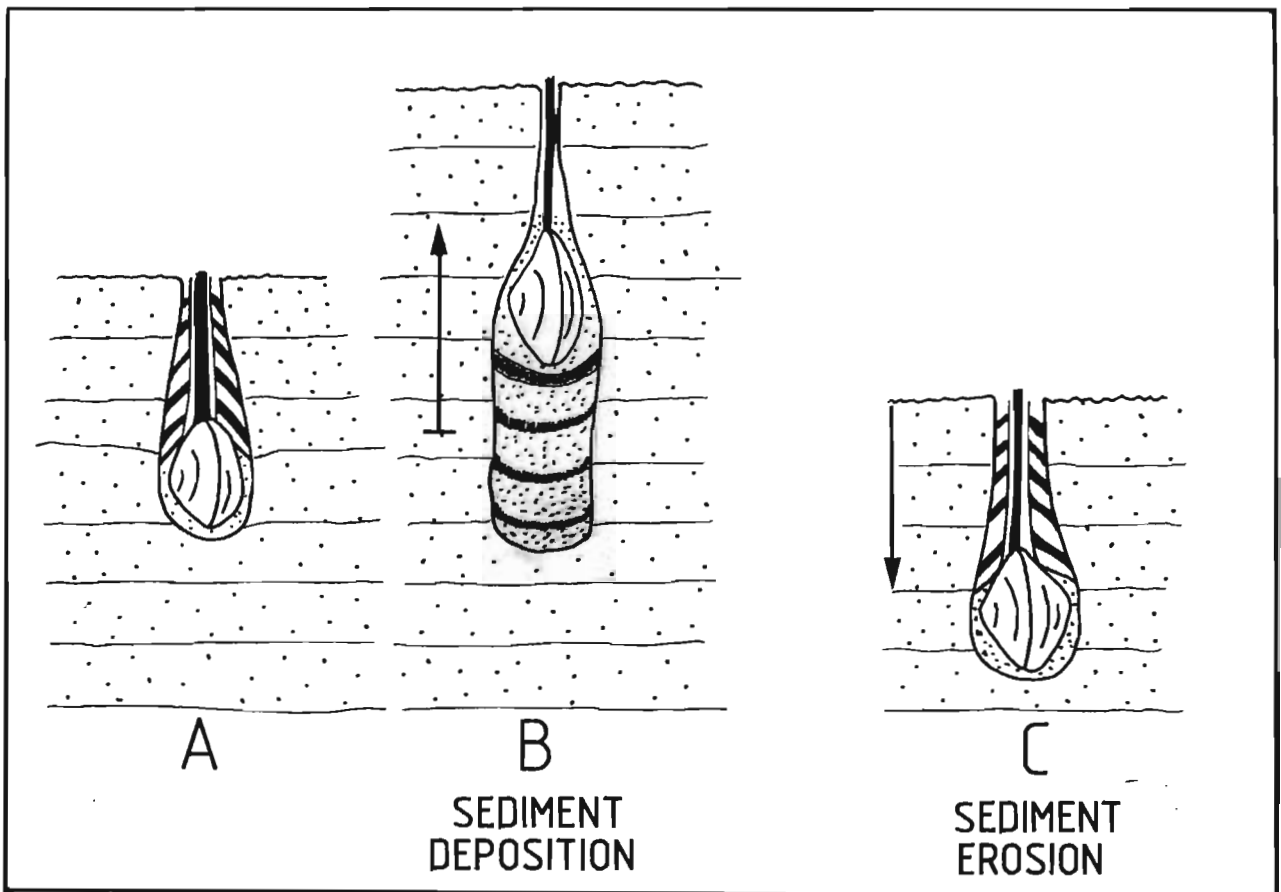
Figure 3.2.   Burrowing patterns of _Mya_ _arenaria_: A. Growning organism gradually burrows deeper. B. With sedimentation, organism migrates upwards. C. In response to erosion organism migrates deeper into sediment (after Reineck, 1958).

Remarks: Traces comparable to Siphonichnus are produced by the downward migration of the siphon-bearing bivalve Mya arenaria in response to erosion (Reineck, 1958) (Fig 3.2). Siphonichnus eccaensis is therefore regarded by Stanistreet et al. (1980) as the domichnium of a suspension-feeding bivalve. The length of the siphon was equal to that of backfill laminae within the outer tube. The occurrence of densely populated colonies of Siphonichnus in previously highly bioturbated sediment commonly produces interpenetrant burrows which confirms that the organism was a suspension feeder. It would be a poor feeding strategy for a deposit feeder to eat sediment ingested previously by other deposit feeders.

Previously this trace was designated Skolithos (Hobday and Tavener-Smith, 1975; Mason and Tavener-Smith, 1978; Tavener-Smith, 1982), but its internal structure does not correspond to the description given by Häntzschel (1975) for that ichnogenus.

Ichnogenus SKOLITHOS Haldeman, 1840

SKOLITHOS ichnosp.

Description: Vertical, unbranched tubes 2 to 8 mm in diameter and 10 to 100 mm in length (Fig 3.3). Tubes show a structureless filling of sandstone and walls may be lined with a thin film of dark organic residue. The burrow normally occurs in sandstone with negligible amounts of argillaceous material and rarely appears to be associated with a distinctive bedding surface.

Discussion: Skolithos burrows were described as domichnia by Seilacher (1964). Hallam and Swett (1966) also regarded Skolithos as a dwelling burrow, probably belonging to a gregarious, suspension-feeding, worm-like organism. Tube-dwelling polychaetes construct analogous dwelling tubes in modern environments along the southeastern Atlantic coast (Curran and Frey, 1977).

Ichnogenus TIGILLITES Rouault, 1850

TIGILLITES ichnosp.

Description: Simple, vertical tubes 1 to 4 mm in diameter and up to 50 mm in length with smooth or annulated external walls. Walls are characteristically defined by a thin film of dark organic material. Tubes are generally closely spaced (only rarely are they crowded) in fine- to medium-grained, silty sandstone which may display signs of earlier bioturbation.

Discussion: Tigillites has received only cursory attention in trace fossil literature and is usually regarded as a synonym of Skolithos (Häntzschel, 1975; Crimes, 1977). In the Vryheid Formation Tigillites is often found in the same situation as Skolithos and the annulated outer surface may be regarded as diagnostic of the ichnogenus. It is generally regarded as a domichnium of a worm-like animal (Häntzschel, 1975).

Ichnogenus DIPLOCRATERION Torell, 1870

Ichnospecies DIPLOCRATERION POLYUPSILON Smith, 1893

Description: Vertical U-tubes showing partly or totally bidirectional spreite (Fürsich, 1974) with dark carbonaceous material (Fig 3.4). Spreite vary considerably, but most define U-in-U tubes. When perfectly preserved they occur on sandstone bedding surfaces as attenuated dumb-bell shaped marks comprising paired circular apertures joined by a narrow strip of spreite-bearing sediment (Fig 3.5). Using the morphology and terminology of Fürsich (1974), the structures have widths of 20 to 150 mm and average aperture diameters of 2 to 4 mm. Diplocraterion occurs as dense colonies in the upper parts of sandstone beds. This ichnofossil is described in greater detail by Mason and Christie (1986).

Discussion: Diplocraterion is the domichnium of a gregarious, suspension-feeding organism (Goldring, 1971; Fürsich, 1974; Knox,
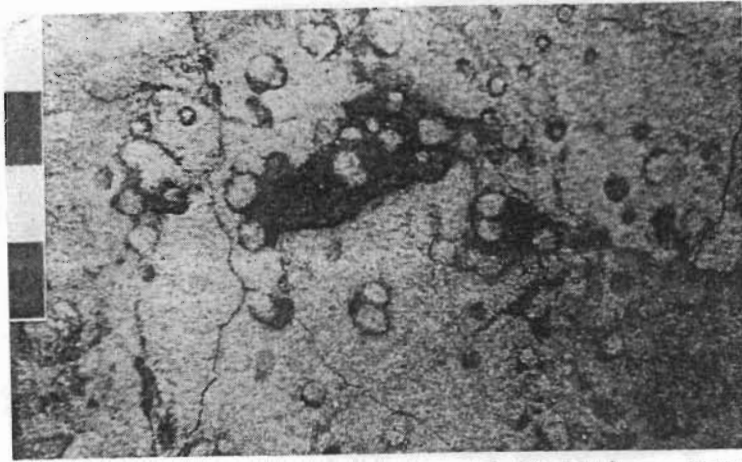
Figure 3.3   Plan section of <u>Skolithos</u>. Note dark organic material lining burrow walls. Scale divisions: 10 mm.
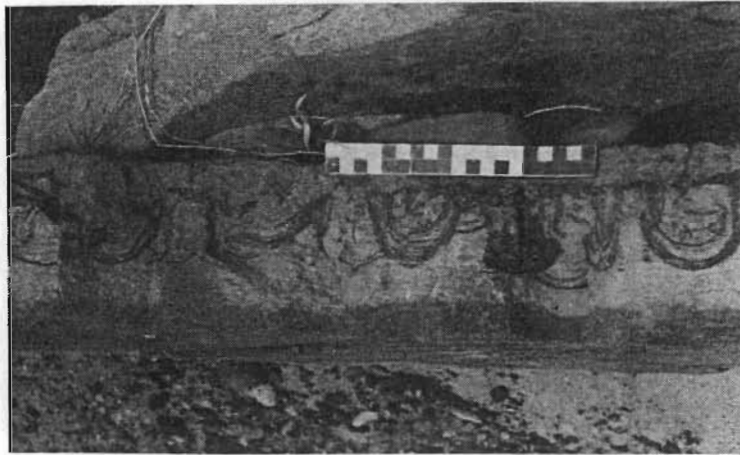
Figure 3.4   Elevation section of <u>Diplocraterion</u>. Scale divisions: 10 mm.
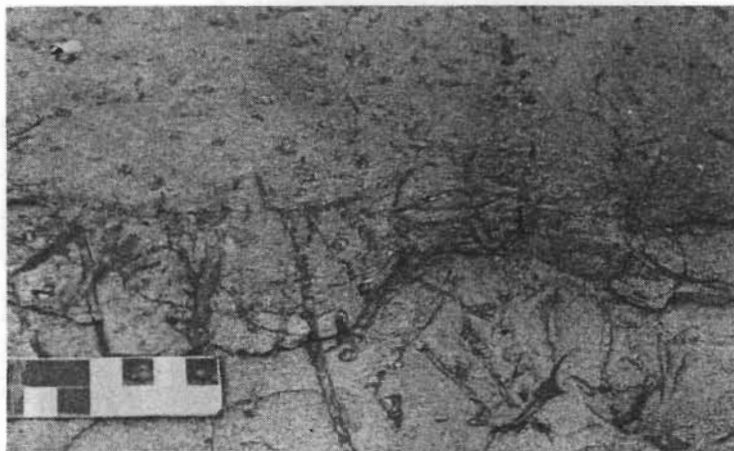
Figure 3.5   Plan section of <u>Diplocraterion</u>. Note spreite joining apertures. Scale divisions: 10 mm.

1973; Mason and Christie, 1986). The suspension-feeding hypothesis is supported by evidence of older, larger burrows penetrated by smaller, juvenile burrows and a weak bimodal orientation of the burrows in relation to associated trough cross-bed foresets (Mason and Christie, 1986). As most of the U-tubes are of the U-in-U type, the spreite may have formed in response to growth of the producer organism (Fürsich, 1974). The wide variation in U-burrow length suggests removal of the upper portions by erosion, while variations in width reflect the presence of both juvenile and mature animals (Mason and Christie, 1986). Fürsich (1974) considered the ichnogenera Diplocraterion and Corophioides to be synonomous as they both display the same diagnostic features (vertical U-tubes containing spreite).

Ichnogenus HELMINTHOPSIS Heer, 1877

HELMINTHOPSIS ichnospp.

Description: Simple, freely winding to meandering, smooth, sandy trails, either flat or showing positive convex epirelief (Fig 3.6). The trails are generally 2 to 3 mm wide and up to 0,50 m long with relief of up to 2 mm. They occur predominantly in fine- to medium-grained sandstone on bedding surfaces.

Discussion: Helminthopsis resembles casts of the modern intertidal polychaete worm Arenicola marina (Schafer, 1972). It is difficult to establish whether this ichnofossil represents trails or faecal casts.

Ichnogenus NEREITES Macleay, 1839

Description: Closely laminated, trilobed, bilaterally symmetrical trails 10 to 20 mm wide which are predominantly straight, but may also curve gently (Fig 3.7A). The median part of the trace is 5 to 10 mm wide comprising either

darker or lighter coloured sediment than the adjacent lateral parts. Sediment in the lateral parts is either identical to the host rock or contains a higher proportion of darker argillaceous material. The trail is commonly punctuated by vertical burrows (resembling Siphonichnus eccaensis) of equal dimensions (Fig 3.7B). These trails are commonly prolific and cross-cutting on bedding surfaces (Fig 3.7B), but gradations to sparse distribution also exist. Although Nereites is predominantly developed at sandstone-siltstone interfaces and on fine- to medium-grained sandstone bedding surfaces, it also occurs in heavily bioturbated silty sandstone in which no sedimentary structures are discernable.

Discussion: Nereites has been interpreted as the foraging trail of a deposit feeder (Seilacher, 1978). Producer organisms that have been suggested include worms (Richter, 1926), gastropods and crustaceans (Seilacher, 1978). The internal structure of the trail probably represents an active backfill of sediment by the animal as it moved through the deposit leaving behind a ribbon of rejected material and a central faecal discharge of excreted sediment. Smith and Crimes (1983) believed that if backfill laminae is the same as the underlying sediment, the trace is probably a trail or furrow. However, a burrow at a sand-mud interface would result in some of the mud being incorporated in the backfill and the lateral zones would then be darker than the host (underlying) sandstone. As Nereites is found mostly at sandstone-siltstone interfaces it is considered to be burrow or "tunnel trail" formed just below the sediment surface (Häntzschel, 1975). Vertical burrows along the trail may be in response to the animal seeking temporary refuge deeper in the sediment, or vertical movement in response to sedimentation. They are considered to be too regular and persistent to be ascribed to accidental piercing by other vertically-burrowing organisms. Nereites may be synonomous with the ichnogenus Scolicia (de Quatrefages, 1849) which also contains trilobed, bilaterally symmetrical trails. More detailed work is required before this trace can be confidently assigned to a specific taxon.
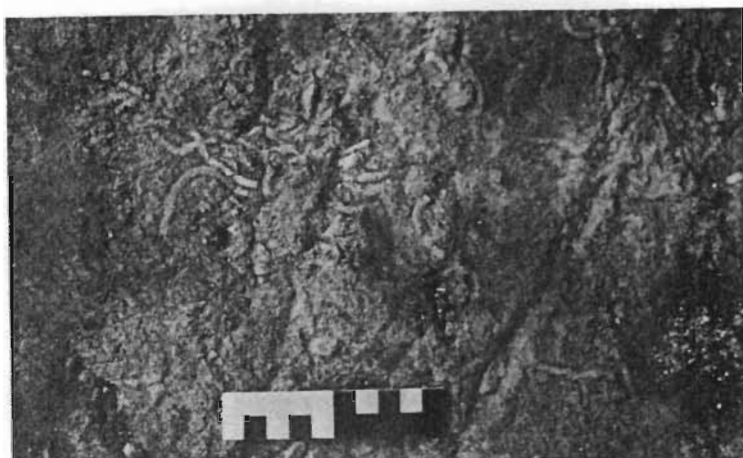
Figure 3.6   Helminthopsis trails. Scale divisions:
10 mm.
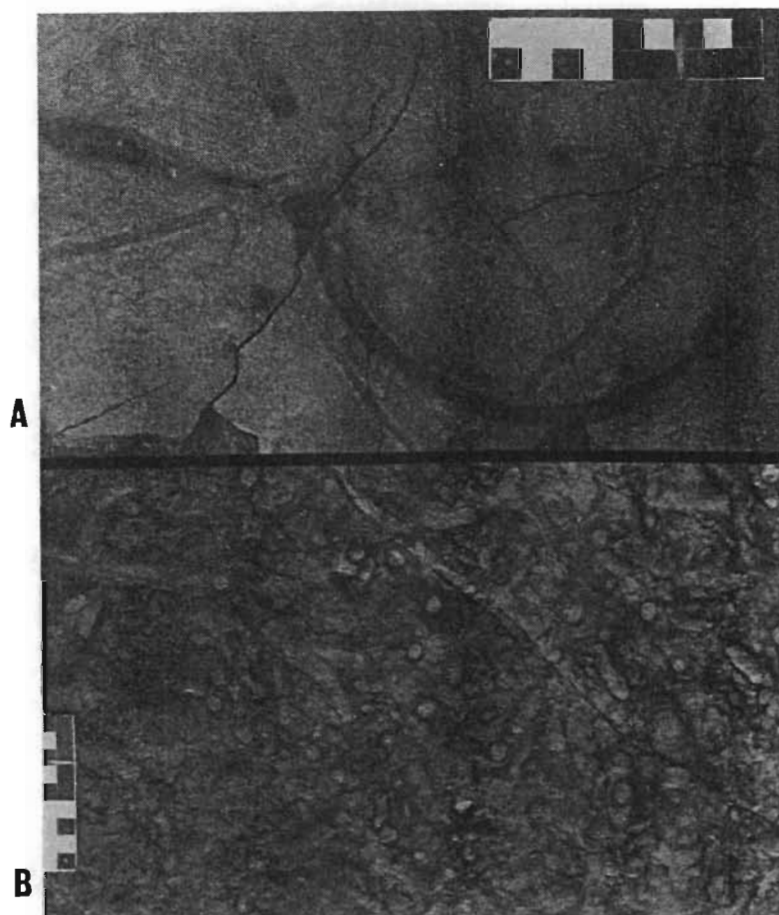
Figure 3.7   Nereites. A. Trilobed straight
and curved character of the trail. B. Dense
and cross-cutting behaviour of Nereites and
presence of vertical burrows. Scale divisions:
10 mm.

Ichnogenus SCOLICIA de Quatrefages, 1849

SCOLICIA ichnosp.

Description: Straight, gently curving or irregularly meandering V- and U-shaped grooves (concave epirelief) commonly with lateral ridges (Fig 3.8). The trails are up to 10 mm wide and 8 mm deep with lengths of up to 1,5 m. They occur on bedding planes of silty, fine- to medium-grained sandstone as isolated occurrences or dense concentrations. In the latter case trails commonly cross-cut one another.

Remarks: These traces are generally interpreted as being produced by a sediment- or suspension-feeding gastropod (Crimes, 1975).

SCOLICIA ichnosp.

Description: A trail of imbricated, stacked laminae up to 10 mm wide (Fig 3.9) which may be straight to slightly sinuous. Traces up to 0,6 m long were observed. This species is generally sparsely distributed in silty, fine-grained sandstone and cuts across earlier formed traces.

Remarks: This trace was probably formed as an active backfill by the peristaltic movement of an animal's foot. There is no evidence to suggest whether the trace is a burrow or a trail.

Ichnogenus PLANOLITES Nicholson, 1873

PLANOLITES ichnosp.

Description: Simple, straight to gently curving, commonly branched burrows, circular to elliptical in cross-section and 5 to 20 mm in width. Burrows are structureless and parallel or subparallel to bedding. The traces are paler in colour than the host rock which is most commonly mudrock to silty,
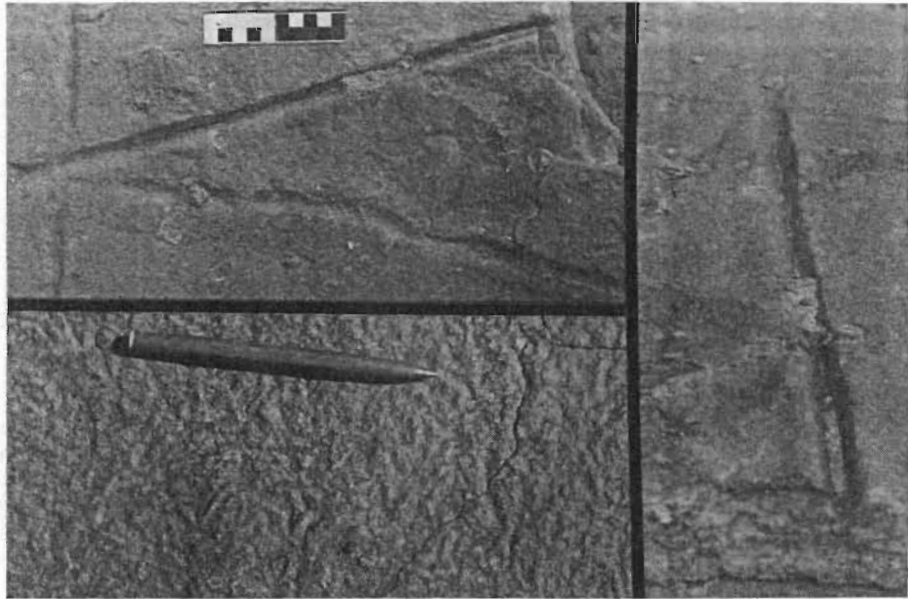
Figure 3.8   Two manifestations of Scolicia ichnosp.
Note V- and U-shaped nature of grooves and lateral
ridges. Scale divisions: 10 mm.
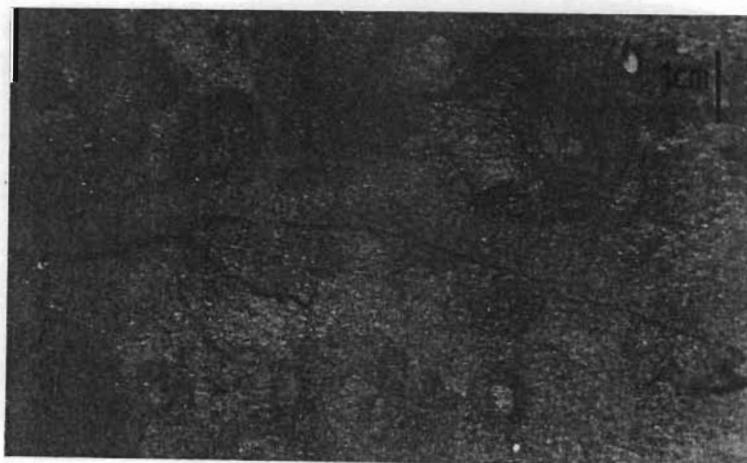
Figure 3.9   Scolicia ichnosp. Trail displays
faint stacked-spreite.

fine- or medium-grained sandstone. <u>Planolites</u> is generally sparsely distributed, but dense concentrations of cross-cutting traces do occur.

<u>Discussion</u>: <u>Planolites</u> represents an active backfilling of sediment in a burrow constructed by a mobile deposit feeder which was either a polychaete or a worm (Pemberton and Frey, 1982). The contrast in colour between the burrow and the host sediment is probably a result of the animal's digestive processes (Nicholson, 1873). Pemberton and Frey (1982) suggested that <u>Planolites</u> could be distinguished from <u>Palaeophycus</u> on the basis of the burrow fill. <u>Planolites</u> has a fill totally different to the lithology of the host rock.

<p style="text-align:center">Ichnogenus <u>PALAEOPHYCUS</u> Hall, 1847</p>

<u>Description</u>: Cylindrical to subcylindrical, straight or sinuous, branched or unbranched, gently curved, commonly undulating, simple burrows 2 to 25 mm in diameter oriented parallel or subparallel to bedding (Fig 3.10). The burrows are preserved as hypichnial ridges or endichnial grooves and may be deformed. They are thinly lined or unlined and contain a structureless fill of the same lithology as the host sediment which is most commonly silty, fine- to medium-grained sandstone. <u>Palaeophycus</u> distribution is moderately dense to isolate.

<u>Discussion</u>: <u>Palaeophycus</u> is considered to be the domichnia of mobile organisms. The burrow fill represents passive, gravity-induced sedimentation (Pemberton and Frey, 1982). Although the same authors regarded the presence of a burrow lining as a diagnostic feature of <u>Palaeophycus</u>, this is not always seen. Osgood (1970) considered predatory polychaetes as modern analogues of the organisms responsible for <u>Palaeophycus</u> burrows. Hypichnial and endichnial preservations suggest that the organism foraged along sediment interfaces (Pemberton and Frey, 1982). This mode of occurrence commonly results in the burrow fill differing from the host rock, but being the same as the superjacent lithology.

Ichnogenus <u>SPIRODESMOS</u> Andreé, 1920

<u>Description</u>: One-way spiral-shaped forms attaining a maximum spiral diameter of 300 mm (Fig 3.11). Coils are an average distance of 50 mm apart with a maximum of two coils in each spiral. The traces vary in width from 3 to 10 mm and may be preserved as U-shaped epichnial grooves and hypichnial ridges. The traces may be preserved in a variety of ways:

1. Unornamented, continuous groove.

2. Unornamented, discontinuous groove consisting of a series of short segments. This may be transitional from a trace comprising a continuous groove.

3. Unornamented, continuous or discontinuous groove punctuated by a number of <u>Siphonichnus</u>-like vertical burrows or small pits. In some traces the spiral may be partially or totally outlined only by these.

4. Trilobed, <u>Nereites</u>-like symmetrical trails or shallow grooves. The traces are characteristically discontinuous and regularly pierced by <u>Siphonichnus</u>-like vertical burrows.

<u>Spirodesmos</u> occurs singly or in dense concentrations. In the latter case the traces commonly intersect. They are found on bedding surfaces of fine- to medium-grained, well-sorted sandstone and are often associated with earlier formed primary current lineation. At many exposures <u>Spirodesmos</u> occurs at a sandstone-siltstone interface.

<u>Discussion</u>: <u>Spirodesmos</u> is discussed in more detail by Mason <u>et</u> <u>al</u>. (1983) who suggested that it may be either a surface trail, a "tunnel trail" (formed just beneath the surface of the sediment) or a burrow. Sedimentary associations strongly suggest that the producer organism lived at or near the sediment surface.

Figure 3.10. Plan section of Palaeophycus. Scale division: 10 mm.
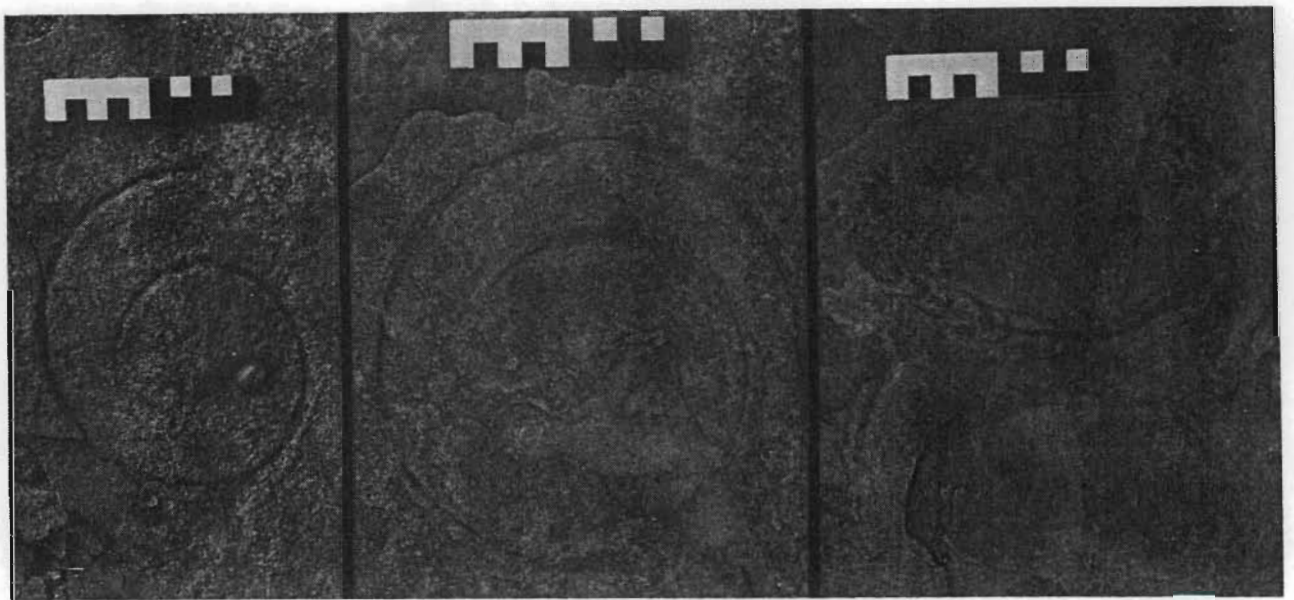
Figure 3.11   Plan section of Spirodesmos illustrating, from left to right, increasing degree of preservation and resulting increasing complexity of internal ornamentation.

The discontinuities present in some of the traces may be due to peristaltic contractions forming an internal pelleted fill (Mason et al., 1983) or by the organism moving through sediment in an undulatory manner. The apparent presence of vertical burrows or small pits along the length of some of the traces may be attributed to the organism changing to an alternative feeding habit. The similarity between the trace described in 4. above and Nereites may indicate similar locomotion and feeding habits of producer organisms. Although Mason et al. (1983) assigned the spiral traces described in 1. and 2. above to the ichnospecies S. archimedes further detailed analysis of these trace fossils is required to differentiate between and classify the various ichnospecies. T.R. Mason (pers. comm.) suggested that the variation in trace morphology may be a function of depth of erosion: those traces which have undergone most erosion display either a series of pits or vertical burrows or a single groove while those that are fully preserved display a trilobed trail.

3.3 ENVIRONMENTAL SIGNIFICANCE OF TRACE FOSSILS

The trace fossil assemblage of the Vryheid Formation within the Klip River coalfield is dominated by representatives of Seilacher's (1967) littoral and shallow water Skolithos, Glossifungites and Cruziana ichnofacies. Deeper water forms (Helminthopsis, Spirodesmos, and Nereites) are also present. This indicates an essentially marine environment and is in agreement with the generally held view that the Ecca sea was marine or at least brackish (Hobday, 1973; Visser and Loock, 1978; Stanistreet et al., 1980; Tavener-Smith, 1982; Mason et al., 1983).

The bathymetric zonation of characteristic assemblages of trace fossils was attributed by Seilacher (1967) to differences in feeding habits of organisms; from nearshore suspension feeders to offshore deposit feeders (Fig 3.12). Rhoads (1975) suggested that "food partitioning" is not the only controlling factor but that the physical instability of the substrate, low concentration of dissolved oxygen in deeper waters, light, salinity and temperature variations, sedimentation rates and degree of current or wave turbidity at the sediment
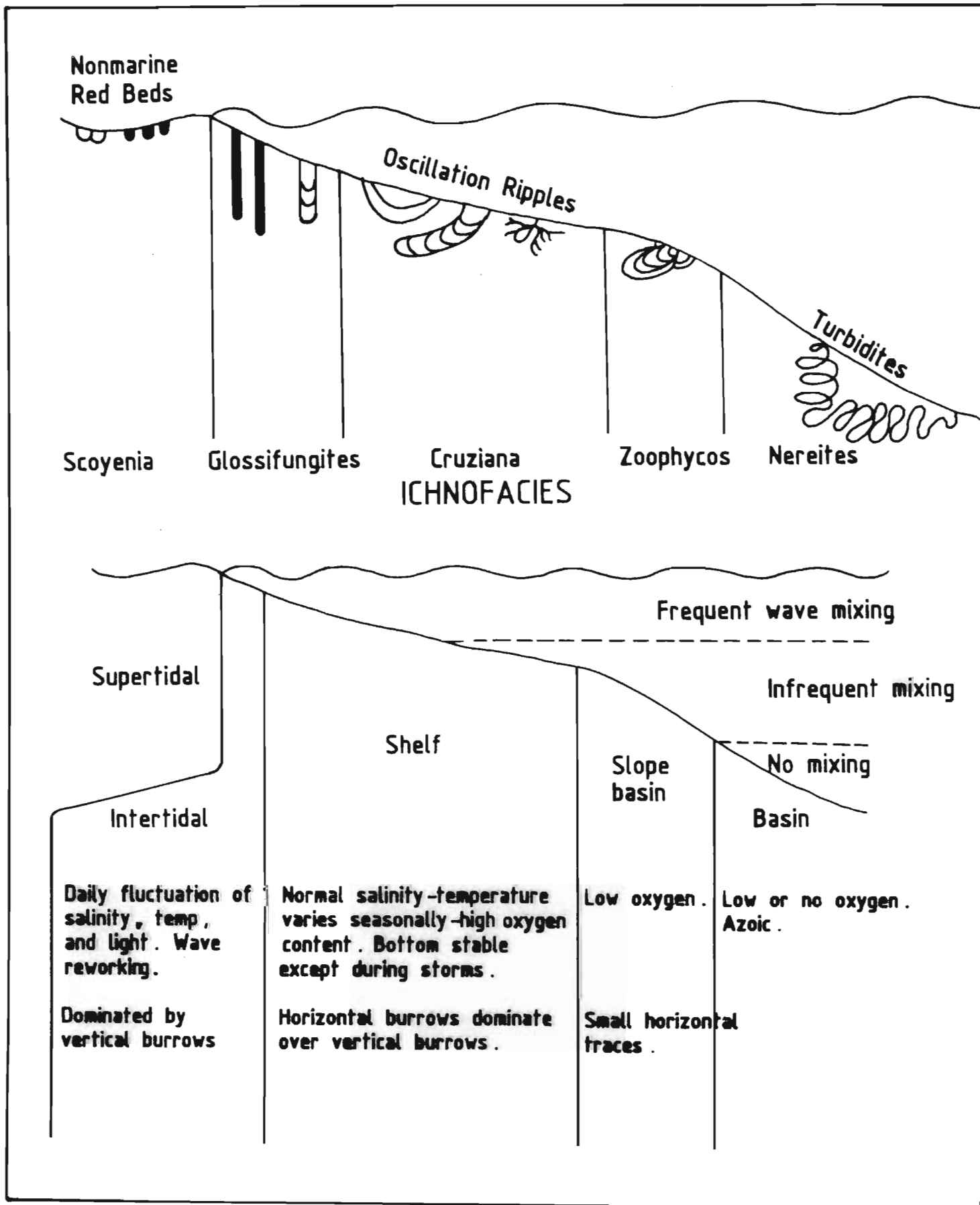
Figure 3.12   Relationship between trace fossil assemblages and ecological parameters across an offshore slope (after Seilacher (1967) and Rhoads (1975)).

surface also played significant roles in trace fossil distribution. Crimes (1975) was of the opinion that trace fossil distribution is controlled ultimately by facies. Thus, any given trace fossil can occur outside its normal bathymetric range. For instance, a deep-water organism may exist in a sheltered, shallow-water, reducing environment where depositional conditions resemble those of a deep-water setting. In certain upward-coarsening sequences within the field area there is an upward increase in the ratio of suspension to deposit feeders. However, trace fossil distribution appears to be largely independent of bathymetry, and "facies crossing" by most ichnogenera is common. A possible reason for this may be the small range in water depths along the coastline (upward-coarsening sequence thicknesses suggest that water depth rarely exceeded 30 to 40 m).

The nearshore environment, especially the littoral zone, is subjected to high-energy current and wave action. Inhabitants of this environment seek refuge in predominantly deep, vertical burrows commonly reinforced by an organic lining of some sort. The overlying sediment not only protects organisms from the unstable substrate but also reduces the extreme variations in temperature and salinity. Environmental fluctations are less common in the shelf or neritic zone and hence horizontal traces (made by deposit feeders) are more common than vertical burrows.

As the concentration of dissolved oxygen decreases with depth, shell-bearing invertebrates, which are confined to levels where oxygen content is greater than 1 ml/1, decrease in abundance. Furthermore, deposit feeders are better suited to an oxygen-deficient aqueous sedimentary environment than are suspension feeders (Theede et al., 1969). The distribution of suspension feeders in deeper water is also limited by their sensitivity to clogging. This can result from deposit feeders producing a fluid, unstable sediment surface that is easily resuspended, even by weak currents (Rhoads and Young, 1970).

Goldring (1964), Rhoads (1975), Howard (1975, 1978), Stanistreet et al. (1980) and many others have shown that trace fossils are valuable aids in the

recognition of erosive and depositional episodes, sedimentation rates, and the energy level of the depositional environment. Similarly, trace fossils are used extensively in this thesis as an aid to environmental analysis and interpretation.

In general, it has been found that the degree of biological activity increases with decreasing current activity (Howard, 1975, 1978). This relationship is shown in Figure 3.13. Slow, continuous deposition may be recognised by complete biogenic reworking and result in a bioturbate texture (Fig 3.14A). The high degree of bioturbation reflects the lengthy time organisms had to rework the sediment rather than animal density or feeding intensity. Continuous, rapid deposition (Fig 3.14B) such as that due to major crevasse splays, turbidity currents or other storm-related depositional events is marked by relatively little bioturbation. Biogenic traces within such rapidly deposited units are usually confined to escape structures, but the maximum thickness of sediment an organism is likely to penetrate is about 300 mm (Howard, 1975). Such catastrophic or episodic events were usually followed by a return to the pre-existing, "normal" conditions and therefore the upper parts of units may be highly burrowed.

Rapidly deposited coarse-grained units commonly alternate with finer-grained beds which represent slow, continuous deposition and contain a different trace fossil assemblage (Fig 3.14C). The difference in ichnofaunal content is believed to reflect textural control. Depending on the thickness of rapidly deposited units, they may contain only escape structures and be densely bioturbated at their tops or, if relatively thin, be entirely reworked. Most sediment accumulation is, however, preceded and followed by erosion and the tops of bioturbated beds commonly show truncated burrows (Fig. 3.14D). Distributary mouth bars and beach foreshores commonly record this type of sedimentary sequence (Howard, 1972; Tavener-Smith, 1982; Mason and Christie, 1986).

The presence of bioturbation also aids in the recognition of "omission"

increasing energy ──────▶

A   B   C
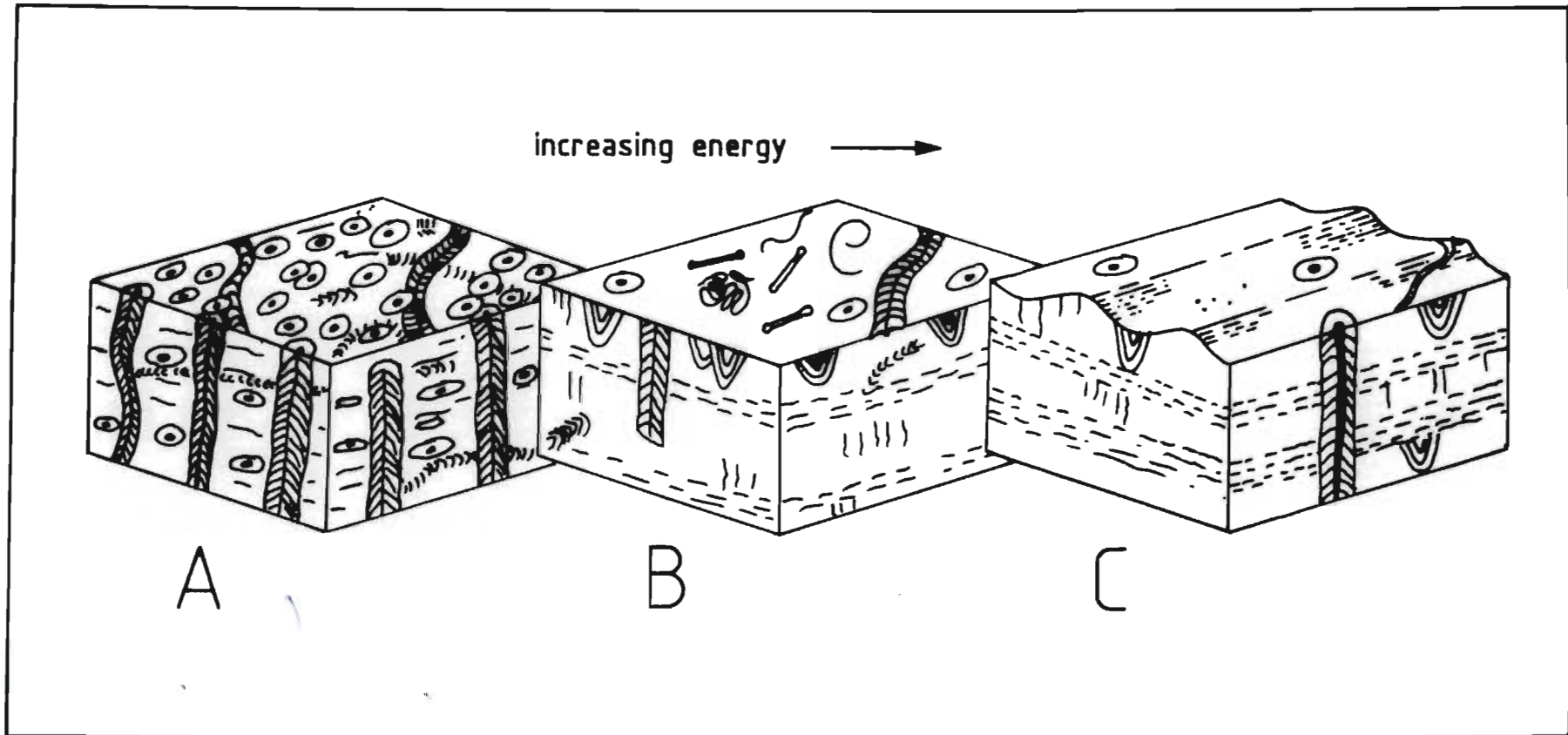
Figure 3.13 Variation in density and type of bioturbation in response to increasing current-energy of the environment (after Howard, 1975).
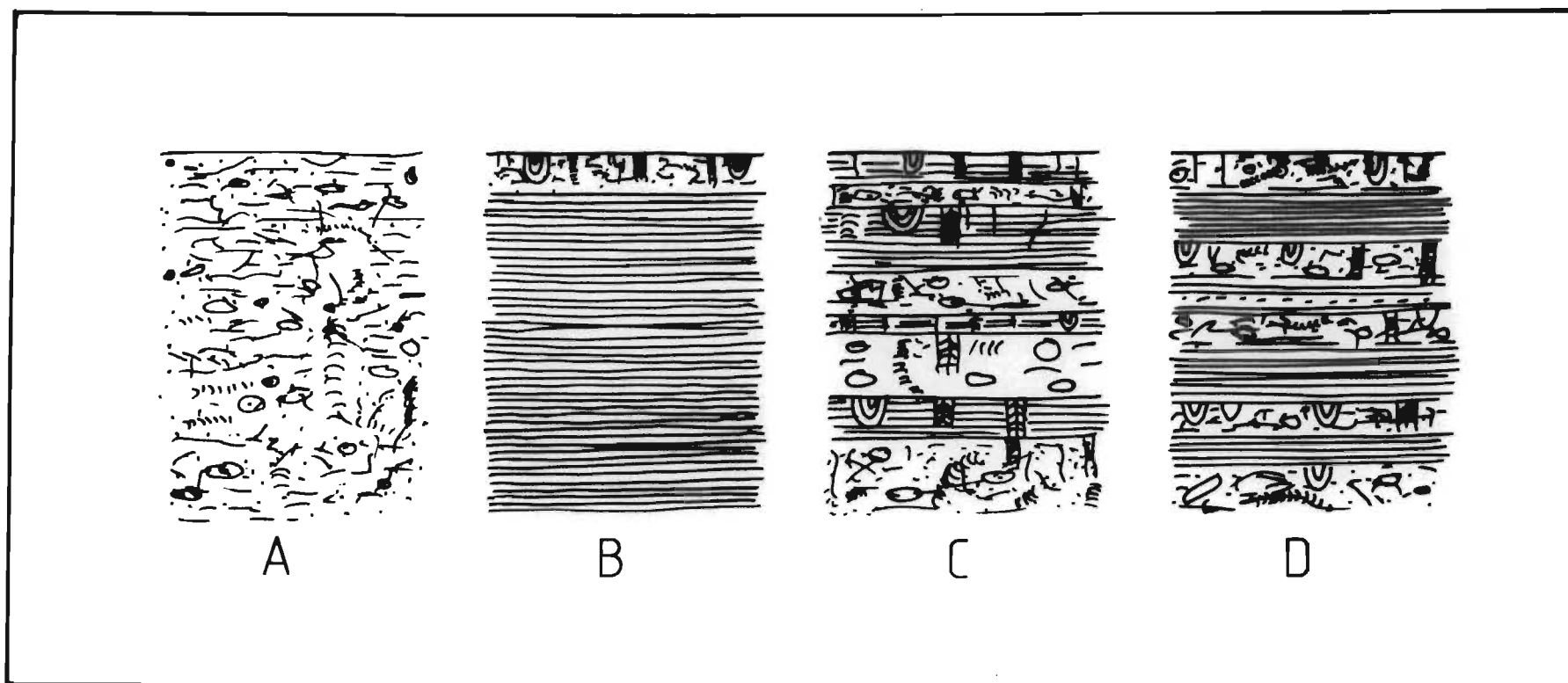
Figure 3.14   Bioturbation patterns developed under differing modes and rates of

deposition (after Howard, 1978).

surfaces; trace fossil density at such surfaces may be indicative of the length of the pause in sedimentation (Miller and Rehmer, 1982). The manner in which some organisms (especially suspension feeders) respond to erosion and deposition is commonly reflected by the internal structure of the trace fossil (Goldring, 1964; Stanistreet et al., 1980). This is especially well illustrated by vertical burrows where retrusive and protusive spreite are indicative of aggradation and degradation of the sediment surface respectively (Fig 3.2).

Sandstones with lithologies and sedimentary structures diagnostic of deposition in a fluvial environment commonly contain trace fossils representative of Seilacher's Skolithos and Glossifungites ichnofacies (Mason and Tavener-Smith, 1978; Stanistreet et al., 1980; Christie et al., 1982; Mason and Christie, 1986). This apparently anomalous association reflects post-depositional water salinities and is attributed to local transgressions or re-establishment of a saline wedge in a distributary channel following flooding. Thus, although trace fossils are used extensively in this thesis as a means of identifying marine conditions, they do not necessarily reflect the water chemistry (ie. salinity concentration) prevailing during the deposition of the host sediment. They commonly relate to post-depositional conditions.

3.4 TRACE FOSSIL ASSOCIATIONS

Vryheid Formation trace fossils form associations which are used to recognise various sedimentary processes.

3.4.1 Siphonichnus (S) Association

Characteristic members include Siphonichnus, Diplocraterion, Scolicia (U- and V-shaped grooves), Nereites, Palaeophycus and Helminthopsis. Siphonichnus, Diplocraterion and, to a lesser extent Scolicia, may occur in dense, single-species colonies. Siphonichnus is the most abundant element of this association and is generally found associated with any or all of the other

members. Diplocraterion only occurs in sparse concentrations when associated with any other ichnofossil while Nereites, although not commonly observed, is always associated with at least Siphonichnus. Characteristically, Siphonichnus and Diplocraterion show protrusive spreite, while Diplocraterion commonly shows U-in-U growth spreite as well.

The Siphonichnus (S) association occurs exclusively within a sandstone substrate (hence the "S" suffix), either at a sandstone-mudrock or sandstone-sandstone interface. Bioturbation is usually confined to the upper portion of the sandstone, but where the bed is less than 200 mm thick, it may be entirely reworked.

Siphonichnus developed within a substrate of sandstone has been reported from numerous sedimentary environments. Hobday and Tavener-Smith(1975) recorded an upward increase in the abundance of Siphonichnus (erroneously called Skolithos) in deltaic sequences and found that they were particularly abundant on sandstone surfaces marking the upper limit of such sequences. Stanistreet et al. (1980) found Siphonichnus associated with sandstones of delta-abandonment phases. A Corophioides-Skolithos association (equivalent to Diplocraterion-Siphonichnus of this thesis) described by Tavener-Smith (1982) occurred within sandstones 120 to 200 mm thick (deposited by high-energy currents) separated by a few millimetres of dark grey siltstone (representing slack-water conditions). Densely populated Diplocraterion colonies were described from a similar sedimentary association by Mason and Christie (1986). Those authors suggested that the producer organisms colonised fluvially-deposited sand during periods of low-energy suspension settling following a return to marine or brackish water conditions. They fed on the organic-rich debris in a well-oxygenated, shallow-water environment. The presence of U-in-U spreite, formed in response to growth of the producer organism, suggests that the periods between high energy events were in the order of weeks or months rather than days. Protrusive spreite, both in Siphonichnus and Diplocraterion, suggests that the environment was also

subjected to periodic erosion. The dominance of suspension feeders in this association implies that a certain amount of turbulence existed. On the other hand, preservation of ridges of sediment commonly observed on either side of V-shaped Skolicia grooves could only have occurred in an environment devoid of strong currents.

The Siphonichnus (S) association is interpreted as representing a low- to moderate-energy, well-oxygenated environment with a low rate of deposition. Extensive bioturbation and evidence of organism-maturity suggests that such conditions may have persisted for lengthy periods.

3.4.2 Siphonichnus (F) Association

Characteristic ichnofossils include Siphonichnus, Nereites, Scolicia (spreite-bearing) and Planolites. Although this association mainly consists of traces produced by deposit feeders, Siphonichnus is the dominant ichnofossil. The association occurs mostly in dark-coloured siltstones, sandy siltstones and silty, fine-grained sandstones (hence the "F" suffix) in which the sedimentary structures are either partially or totally modified by bioturbation.

This association is indicative of a low-energy regime characterised by slow, continuous deposition (Howard, 1975; Fürsich, 1975). The predominance of deposit feeders suggests that food accumulated in the sediment rather than remaining in suspension.

3.4.3 Spirodesmos Association

Characteristic members of this association include Spirodesmos and Skolithos, but Tigillites, Diplocraterion, Siphonichnus, Nereites, Scolicia (V-shaped groove) and Helminthopsis may also be present. Skolithos and Tigillites, although not common, generally occur in sparsely to moderately

populated single-species colonies. Trace fossils of the Spirodesmos association are generally found on bedding surfaces of fine- to medium-grained sandstone and may be overlain by up to a few centimetres of dark-coloured mudrock.

Spirodesmos archimedes has been found in sandstones interpreted as shoreface and embayment deposits (Mason and Tavener-Smith, 1978; Tavener-Smith, 1982). Mason et al. (1983) regarded Spirodesmos as indicative of a shallow, fully marine environment. Stanistreet et al. (1980) suggested that Skolithos represented the trace of an organism which in abited the sandy fringes of an interdistributary embayment. Turner et al. (1981) regarded Skolithos as indicative of a high-energy, shallow-water marine environment; its position at the top of upward-coarsening deltaic sequences suggested that the organism also inhabited distributary mouth bars and distributary channels during temporary or permanent abandonment.

The Spirodesmos association is regarded as being indicative of a truly marine environment, perhaps less influenced by fresh-water fluvial influxes and, at times, of a slightly higher energy regime than that indicated by the Siphonichnus (S) association.

3.5 CONCLUSION

From the above it is clear that trace fossils of the Vryheid Formation are poor bathymetric indicators and single types are not restricted to any particular ichnofacies or depositional environment. However, they are valuable facies indicators in that they provide a means of gauging relative sedimentation rates, energy regimes and oxygen and salinity levels.

CHAPTER 4

SEDIMENTARY FACIES

4.1 FACIES CONCEPT

The facies concept was introduced by Gressly in 1838 to describe sedimentary rocks which displayed similar lithological and palaeontological characteristics (Blatt et al., 1972). Since then the concept has been adopted by biologists and petrologists and used in many different senses. In the sedimentary context the term is generally defined as "... any areally restricted part of a designated stratigraphic unit which exhibits characteristics significantly different from those of other parts of the unit." (Moore, 1949). This definition implies that sedimentary facies are restricted in extent both stratigraphically and geographically, though the same facies may be repeated within the same stratigraphic unit (Blatt et al., 1972).

A sedimentary facies is characterised by sediment texture and composition, the geometry and scale of sedimentary structures and the absence or presence of biological activity. These features enable the hydrodynamic, biological and chemical conditions of the sedimentary environment to be inferred. A facies in isolation, however, gives comparatively little information regarding the depositional environment of a sedimentary sequence; it is the association and mutual relationships of facies that are of greatest significance. Harms et al. (1975) regarded sedimentary facies as the "building blocks" of which stratigraphic sequences are composed. Groups or sequences of facies which are environmentally related are called facies associations, and these associations can be used to construct a facies model. In many stratigraphic sequences facies occur repeatedly, commonly following a pattern which suggests the existence of sedimentary cycles.

Although the palaeoenvironmental interpretation of a sedimentary sequence depends on an examination of both the facies and their lateral and vertical

relationships, it is the last aspect that is most important and easily observed in studying ancient sediments. The study of sediments in vertical profile is based on Walther's Law of Succession of Facies which states: "The various deposits of the same facies area and, similarly, that of the rocks of different facies areas were formed beside each other ... in space, but in a crustal profile we see them lying on top of each other it is a basic statement of far-reaching significance that only those facies and facies areas that can be superimposed, primarily, can be observed beside each other at the present time." (Walther, 1894, translated in Blatt et al., (1972). Blatt et al. emphasised the importance of recognising the nature of the relationship between successive facies for, if Walther's Law is to be applicable, there must be no major breaks in the stratigraphic sequence. Sharp or erosive junctions provide no indication as to whether vertically adjacent facies represent depositional environments that were once adjacent geographically. Furthermore, the vertical sequence observed at any one locality may not display all the facies that were developed laterally (Blatt et al., 1972) owing to the low preservation potential of some deposits (eg. beaches, levees).

The interpretation of depositional environments in ancient sediments relies heavily on the study and definition of present-day environments and sedimentary processes. This aspect of the Principle of Uniformitarianism assumes that geological processes which operate at present also operated in the past to produce similar results. However, this approach has limitations as many of the conditions of sedimentation common in the geological past are not present today, or, if present, are not functioning with the same intensity. Examples of this include variations in the type and rate of erosion consequent upon the widespread establishment of vegetation since Devonian times and differences in the distribution of land, sea and climatic zones. Others could have been consequent upon changes in relief and the distribution of geomorphological features (Reineck and Singh, 1975). Nevertheless, the Principle of Uniformitarianism is valid in broad terms and comparison of ancient and modern

sediments has long established that physical and chemical laws governing sedimentation have remained more or less constant through geological time.

4.2   FACIES DESCRIPTIONS AND INTERPRETATIONS

Although the concept of sedimentary facies has been used as a basis for the construction of palaeoenvironmental models for many years, facies nomenclature has been highly inconsistent and varied. Le Blanc Smith (1980a) illustrated this by listing 7 different kinds of facies names used in describing trough cross-bedded sandstones. Miall (1977a), in a paper reviewing braided stream deposits, was the first to adopt a scheme of nomenclature in which facies were named by using key letters derived from terms describing physical characteristics of the sediment. Le Blanc Smith (1980a) modified this method for application to Karoo sediments in the Witbank coalfield. The parameters he utilised comprise, in order of importance, sediment composition or grain size, sedimentary structures (including ichnofossils) and distinctive mineral components (optional). This scheme was designed primarily for application to borehole cores. Although the coding system is equally suited to facies observed in natural exposures it has been slightly modified and extended in this thesis to accommodate facies showing macro-sedimentary structures.

Trace fossils are generally considered an important aspect of a facies but they commonly relate to conditions prevailing <u>after</u> deposition of the host rock (see Chapter 3). For this reason reference to bioturbation in individual facies is made only where it destroys the original sedimentary structure or is considered an integral component of the facies.

4.2.1 The Conglomerate Facies (C)

Facies Cm: Massive or crudely-bedded conglomerate

Description: Conglomerate is a comparatively rare lithology within the Klip River coalfield. Matrix-supported conglomerates are more common than clast-supported ones. They range in thickness from a single pebble layer to about 1,10m. Clasts are of quartzite, potassium or plagioclase feldspar, and a variety of pre-Karoo basement igneous and metamorphic rocks. They are subangular to rounded pebbles and cobbles up to 150 mm in diameter, but are generally in the pebble size range of 2 to 64 mm (Fig 4.1). Infraformational mudrock and (more rarely) sandstone clasts occurring singly are also present though not common. The matrix is poorly sorted silty, fine-grained sandstone to very coarse-grained sandstone. Matrix-supported conglomerates are characteristically polymodal but in clast supported conglomerates the size distribution is generally more uniform. In the former, variations in clast density are common and in places clasts may be so sparsely distributed that the sediment is more aptly described as a pebbly sandstone. No clast imbrication was detected within either type of conglomerate. Although the conglomerates are predominantly massive, crude flat or cross stratification may be present. Normally this facies has an erosional base.

Interpretation: Pebbles and cobbles are moved only during high-energy discharge in the upper flow regim (Simons et al., 1965). Clast-supported conglomerates displaying a degree of sorting are indicative of currents capable of sorting pebbles and cobbles and winnowing away or holding in suspension sand, silt and clay (Harms et al., 1975). The matrix filtered down into the pore spaces at a later stage following a decrease in current strength. The same authors believed that a matrix-supported conglomerate implies simultaneous transport of fine and coarse particles, possibly as a debris flow. The presence of conglomerate above an erosive surface denotes an abrupt change in energy regime (Collinson and Thompson, 1982) and suggests generation as a coarse, bedload deposit.

The main environments in which conglomerates are found include alluvial fans, braided rivers, shorelines, deep sea submarine fans and glacially influenced environments where till is deposited (Harms et al., 1975). The close association of this facies with coal seams and the Skolithos and Cruziana ichnofacies argues against a deep sea environment, while the absence of tillite excludes a glacial origin.

Facies Ci: Massive or crudely-bedded intraformational conglomerate

Description: Matrix-supported intraformational conglomerates are rare and range in thickness from 0,2 to 0,5 m. Clasts are angular to subangular, disc-shaped fragments of mudrock up to 150 mm across and 10 mm thick. Matrix grain size is medium- to coarse-grained pebbly sandstone. These conglomerates usually lack bedding but occasionally flat and cross stratification may be discerned. They overlie erosive surfaces and occur at the base of upward-fining sequences.

Interpretation: Intraformational conglomerates overlying erosive surfaces are indicative of high-energy currents ripping up and reworking an argillaceous substrate. The composition and shape of the clasts suggests that they were of local derivation and represent cohesive, semi-consolidated clay and silt.

4.2.2 The Sandstone Facies (S)

Facies St: Trough cross-stratified sandstone

Description: Trough cross-stratification is one of the most abundant sedimentary structures of the Vry. nid Formation. This facies is composed of erosive-based, wedge-, tabular- and trough-shaped (festoon cross-bedded) sets which are typically between 50 mm and 0,8 m thick, but range up to 2 m. Set thickness generally increases with grain size. Troughs are up to 40 m wide (average 0,2 to 1,2 m) and traceable in a down-current direction for

up to 25 m (Fig 4.2). Foresets are invariably concave upward with tangential lower contacts. Cosets may be up to 8 m thick. Trough cross-stratification is present in fine- to very coarse-grained, moderately- to poorly-sorted pebbly sandstones. Discontinuous mudrock drapes and concentrations of pebbles may be present along foresets and between successive sets. Trough cross-stratification is commonly associated with facies Sp (planar cross-stratified sandstone) and less commonly with facies Sf (flat stratified sandstone).

Interpretation: Allen (1963) believed trough cross-stratification to be associated with migrating lingoid or lunate dune forms. This supposition was confirmed by work carried out by Briggs and Middleton (1965), Harms and Fanhestock (1965), Harms et al., (1975) and many others. Dunes were observed to migrate by the movement of grains up the stoss side of the structure followed by avalanching down the lee side. Avalanching grains create foresets and are preserved as cross-stratification. Flow becomes separated from the sandy bed at the crest of the dune and a transverse vortex or separation eddy forms in the lee resulting in a hollow scour. Collinson and Thompson (1982) believed that the eddy pattern is closely related to the shape of the bedform and that the eddy pattern in the lee of one dune influences the flow pattern over the next one downstream. Winnowing associated with erosion in the lee of a dune may result in a pebble lag at the base of the scour.

Trough cross-stratification is formed by currents in the upper part of the lower flow regime (Simons et al, 1965; Harms and Fahnestock, 1965). Dune size, and hence set thickness, is related to flow depth. Harms et al. (1975) stated that set thickness is approximately half the flow depth but cautioned that dunes can exist at greater depths. Collinson and Thompson (1982) contended that deep flows generate higher and longer dunes and suggested an approximate ratio of 1:6 for dune height to flow depth. The same authors believed dune height to be independant of grain size; the finer grain size of thinner sets may therefore reflect gentler currents in water more shallow than that in which larger dunes were formed. Le Blanc Smith (1980a) suggested that a complete
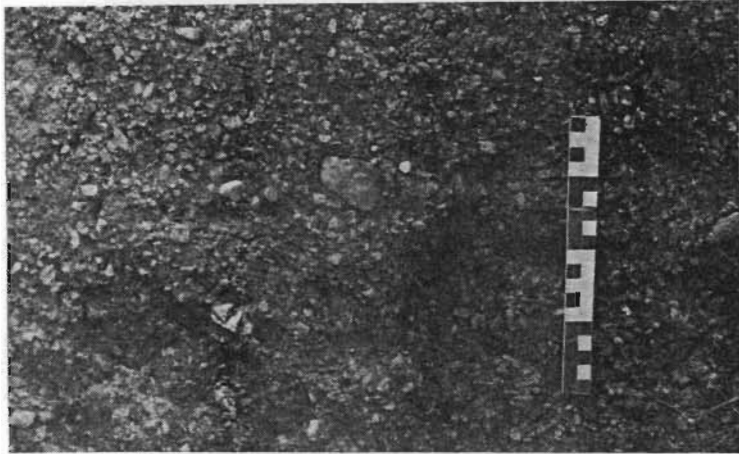
Figure 4.1 Facies Cm: matrix-supported pebble conglomerate. Scale divisions: 10 mm.

Figure 4.2 Facies St: trough cross-bedding, plan section. Scale divisions: 10 mm.

Figure 4.3 Facies Sp: planar cross-bedding. Note numerous reactivation surfaces and overall upward decrease in set size.

spectrum of forms exists between trough and planar cross-stratification; large scale sets up to 40 m wide may thus be a transitional form between dunes and sinuous- or straight-crested bar forms.

Facies Sp: Planar cross-stratified sandstone.

Description: Sets range in thickness from 50 mm to 3 m but average between 0,3 and 0,8 m (Fig. 4.3). Although solitary sets do occur, cosets up to 6 m thick are most common. Sets are usually tabular or wedge shaped, their bases being either slightly erosive with low relief or non-erosive, planar. Some sets persist for up to 150 m in a downdip direction and 60 m laterally. Foresets commonly display angular or strongly discordant basal contacts, but tangential or asymptotic ones may also occur. Reactivation surfaces are common and foresets may be slightly to moderately lingoid in plan.

Grain size and sorting characteristics are generally similar to those of Facies St. Individual foresets show both normal and reverse grading. Thin intercalations (up to 100mm) of flat stratified sandstone, ripple cross-laminated sandstone, or massive mudrock may be developed between successive sets. The sets are commonly incised by trough cross-beds orientated in the same direction as, or at oblique angles to, the underlying planar foresets.

Although Le Blanc Smith (1980a) included cross-stratified sets greater that 3 m thick in Facies Sp, their lithological and structural complexity compels more detailed discussion and they are dealt with in Chapter 5 in connection with facies models.

Interpretation: Planar cross-stratification has been reported from lingoid or transverse bars (Ore, 1964; Smith, 1970, 1971, 1972, Collinson, 1970a; Cant and Walker, 1978). Other terms used to describe similar types of bars are cross-channel (Cant and Walker, 1978), migratory bar avalanche faces (Williams

and Rust, 1969) and sand waves (Harms and Fahnestock, 1975; Collinson and Thompson, 1982). The term transverse bar is used in this thesis.

The origin and growth of transverse bars was described by Jopling (1965) and Smith (1970, 1971). According to these authors a bar is initiated when bedload sediment encounters a depression of sufficient depth to lower the stream velocity below the critical value necessary for continued traction transport. It forms by downcurrent and lateral aggradation of foresets into the depression until equilibrium is established and critical depth and flow velocity are attained. Transverse bars commonly occur in trains in an out-of-phase relationship with one another, such that the convex front of a bar tends to advance into the space between two preceeding bars (Miall, 1977a). Each bar has a separation eddy in its lee but this is dissipitated by mixing into the boundary layer over the stoss side of the downstream sandwave (Collinson and Thompson, 1982).

Planar cross-stratification is indicative of currents in the middle part of the lower flow regime, generally of lower flow intensity than that required for the formation of trough sets in dunes (Harms et al., 1975). The angle of foreset inclination depends of current strength (Harms and Fahnestock, 1965; Jopling, 1965). A steep angle of foreset inclination (close to angle of repose) suggests that a large proportion of the sandstone was transported as bedload by weak currents causing avalanching down the bar face. Flow separation and grain-fall processes become important with stronger flows and the foresets therefore have a lower angle of inclination and are tangential. Reactivation surfaces indicate variations in flow direction, strength and depth (Collinson, 1970a).

Controls on set thickness are poorly understood but it is probable that this increases with increasing flow depth (Collinson and Thompson, 1982). Harms et al. (1975) stated that flow depth must exceed maximum set thickness, while Collinson and Thompson (1982) suggested that a height:flow-depth ratio of 1:2 is generally the rule.

The presence of other sedimentary structures superimposed on planar cross-stratified sets is thought to reflect low-water modification processes during which the bar was out of equilibrium with the flow (Miall, 1977a). Alternatively, shallower flow depths on the upstream surface of a bar may create a higher flow regime where trough cross-stratification and flat stratification are the stable bedforms (Jackson, 1975; Cant and Walker, 1978).

Facies Sx: Cross-stratified sandstone.

Description: This facies was established because of the difficulty in differentiating cross-stratified sandstones in borehole cores. Lithologically, facies Sx is similar in all respects to facies St and Sp.

Interpretation: Cross-stratification is attributed to migrating dune and bar bedforms (Harms et al., 1975).

Facies Sf: Flat stratified sandstone.

Description: Facies Sf comprises flat bedded (essentially horizontal) sandstones that are fine- to medium-grained, pebbly. Stratification varies from parallel, continuous laminae (up to 20 m laterally) to slightly undulating, discontinuous laminae. Low-angle discontinuities are common. Primary current or parting lineation is commonly developed on bedding surfaces. Facies thickness ranges from a few millimetres up to 4 m, but sequences up to 15 m are present. Facies Sf may be intimately associated with facies Sr (ripple cross-laminated sandstone) especially in very fine- and fine-grained sandstones.

Interpretation: Flat stratification is a specific bedform and develops in response to a certain combination of grain size and flow characteristics under two depositional regimes (Harms and Fahnestock, 1965). Fine- and medium-grained flat stratified sandstone (grain size between 0,1 and 0,6 mm) is deposited by currents in the upper flow regime (Simons et al., 1965). Flow velocities are

Simons et al. (1965) suggested that a significant difference between upper and lower flow regime flat stratification is that in the former case little grain size sorting occurs whereas in the latter, size segregation results in alternating coarse and fine laminae and internal grading.

Reineck and Singh (1975) listed a number of depositional environments in which flat stratification is found. These include beaches (formed by swash and backwash action of waves), the coastal shoreface, in certain fluvial environments (on levees, point bars and bar tops) and in deep-water flysch sediments. McKee et al. (1967) observed a preponderance of flat stratified sands in fluvial channel and overbank deposits produced by high-energy flood waters.

Facies Sr: Ripple cross-laminated sandstone.

Description: Asymmetric ripples with or without climbing-ripple (ripple-drift) lamination are included in this facies. Ripple amplitude is less than 50 mm. Ripple crests may be straight, undulatory or lingoid; only rarely are flat-topped ripples observed. Internal stratification is trough or planar cross-lamination commonly enhanced by dark mudrock laminae about 1 mm thick. Occurrences of this facies range from a few tens of millimetres to several metres in thickness. Sand grain size varies from very fine to coarse, but the facies occurs mostly in fine- to medium-grained sandstone.

Interpretation: Ripples are indicative of currents in the lower part of the lower flow regime (Simons et al., 1965; Harms et al., 1975). Straight-crested ripples are formed at slightly lower flow velocities than lingoid ripples while sinuous-crested ripples are an intermediate form (Reineck and Singh, 1975). Straight-crested symmetrical ripples are said to be produced by wave action or, less specifically, by oscillating currents. Asymmetric ripples are the result of unidirectional currents or a combination of waves and currents (Harms et al., 1975). Ripple cross-lamination is formed in numerous sedimentary environments, especially in those involving shallow-water and low-energy conditions (Reineck and Singh, 1975).

Climbing-ripple lamination is formed by the migration and simultaneous upward growth of ripples where there is an abundant suspended sand supply. Such conditions may prevail where streams over-top their banks during flood and form relatively unconfined and slower flows (across levees and flood plains) or where flow depth changes, for example, at deltas or over the crests of large sand waves (Reineck and Singh, 1975; Harms et al., 1975). Sediment laden ephemeral streams (McKee et al., 1967) and turbidity currents (Walker, 1965) may also be characterised by climbing ripples.

Facies Smr: Megarippled sandstone

Description: Megaripples range in length from 0,6 to 30 m and in height from 60 mm to 1,5 m (Reineck and Singh, 1973). Such sedimentary structures are uncommon in the Vryheid Formation and occur in medium to very coarse-grained, pebbly sandstones. They are mostly preserved as straight- and undulatory-crested bedforms with amplitudes of 60 mm to 0,5 m and wavelengths up to 2 m (Fig 4.4). Isolated megaripple trains completely enveloped in mudrock are also present. Trough cross-stratification is the dominant internal structure of megaripples, though some are massive.

Interpretation: Megaripples form in the upper part of the lower flow regime (Simons et al., 1965). In sand with grain size greater than 0,6 mm (medium grained) small ripples are not produced under such conditions and only megaripples form. In finer sandstone small ripples form in preference to megaripples at higher flow velocities (Reineck and Singh, 1975). Straight-crested megaripples form at a lower current velocity than those that have undulatory crests.

Facies Sfl: Flaser-bedded sandstone

Description: This facies comprises ripple cross-laminated (symmetric and asymmetric) very fine- to medium-grained sandstone in which mudrock streaks and drapes are preserved in ripple troughs and crests. The facies may be up to

Figure 4.4. Facies Smr: megarippled sandstone.
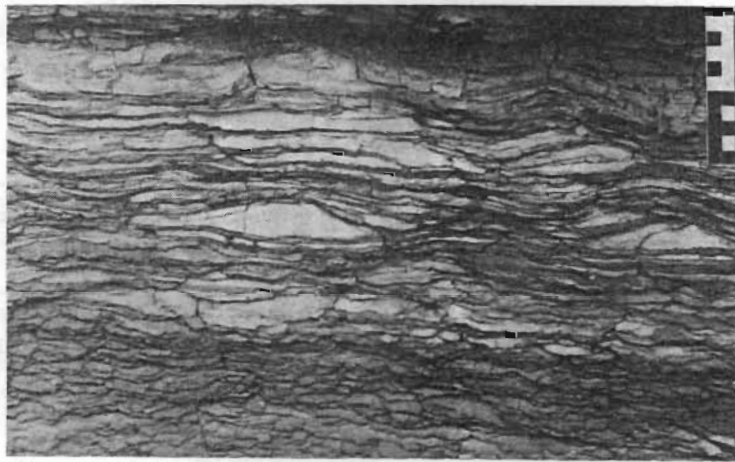
Figure 4.5. Facies Fl overlain by facies W.

Figure 4.6. Facies Al: Alternating sandstone
and mudrock. Note sharp boundary contacts.

several metres thick and is commonly gradational between an underlying
lenticular-bedded unit (facies Fl) and an overlying ripple cross-laminated unit
(facies Sr).

Interpretation: Flaser bedding implies that sand, silt and clay were available
and that periods of current activity alternated with periods of quiesence
(Reineck and Wunderlick, 1968). During periods of current activity sand is
deposited as ripples while mud was held in suspension. During slack water, clay
and silt were deposited mainly in troughs but may also have draped the entire
ripple. Subsequent currents eroded the ripple crests and new ripples buried and
preserved clay and silt in ripple troughs. Flaser bedding is therefore produced
in environments where conditions for the deposition and preservation of sand
were more favourable than for clay and silt (Reineck and Singh, 1975).

Flaser bedding is common in tidal flat deposits but also occurs in any
sedimentary environment which experiences gentle, intermittent currents, for
example, in deltaic, flood plain and lake settings.

Facies Sm: Massive sandstone

Description: This facies comprises massive or structureless sandstone which
may or may not be graded. Sand grain size varies from fine- to very
coarse-grained, pebbly. Graded units vary in thickness up to 400 mm and may be
erosively based and pass upward into stratified sandstone. Massive sandstone
units lacking grading may be up to 2,5 m thick.

Interpretation: The absence of stratification in this facies can be attributed
to grain flow processes caused by rapid deposition from suspension and burial
before bedload movement could take place (Lowe, 1976a); the physical disruption
by liquefaction and movement of waterlogged sediment (Lowe, 1976b); organic
reworking or the deposition of homogeneous sand in which stratification is
concealed. Normal grading reflects the deceleration of a sediment-laden current

and consequent decline in its competence, with coarse grains settling first (Collinson and Thompson, 1982). Graded bedding is attributed mainly to turbidity currents (Kuenen and Migliorini, 1950) but is also produced in shallow-water environments (generally less than 20 mm and rarely more than 200 mm deep) by sedimentation from suspension clouds (Reineck and Singh, 1975). It may also result from deposition in the last phases of a flood or from wave swash (Reineck and Singh, 1975).

Facies Sb: Bioturbated sandstone

Description: Facies Sb is typically a silty, fine- to coarse-grained sandstone in which the original structure has been destroyed by biogenic activity. Both deformative and figurative bioturbation structures are present. In the former case the texture of the sediment is commonly described as "mottled". A wide variety of ichnofossils, including rootlets, has been described from this facies which attains thicknesses of up to 2 m.

Interpretation: Intensely bioturbated units mostly reflect a low but continuous rate of deposition, though according to Howard (1975) thin, rapidly deposited sandstones may be completely reworked during an ensuing period of more gentle deposition. This facies may not indicate the hydrodynamic conditions under which the sediment was deposited, but rather reflects a post-depositional environment favourable for marine organisms.

4.2.3   The Fine-Grained Facies (F)

The terminology of fine-grained rocks in this thesis is that used by Blatt et al. (1972). The lithological terms siltstone, mudstone and claystone refer to the texture of the sediment and not the mineral composition (see table 11.1 in Blatt et al., 1972). "Mudrock" is used as a general term.

Facies Fm: Massive mudrock

Description: Sediments of this facies are light to dark grey in colour, a function of the high proportion of finely disseminated organic material that is present. They are commonly fissile and at places the term shale may be appropriate if the clay content is sufficiently high. Thin, sandy layers a few millimetres to a few centimetres thick are commonly present. This facies may be up to 30 m thick but thicknesses between 50 mm and 2 m are more general. Plant remains and impressions may be preserved.

Interpretation: The complete absence of sedimentary structures in this facies is probably indicative of uninterrupted settling from suspension in an environment free from currents (Pettijohn, 1975). Hobday (1973) suggested that the lack of sedimentary structures may be due to complete reworking and homogenisation by burrowing organisms. On the other hand, the complete absence of trace fossils in such dark coloured sediments was considered a strong indicator of euxinic conditions by Blatt et al. (1972).

Facies Ff and Fr: Laminated mudrock

Description: Sediments of this facies may be flat laminated (facies Ff) or ripple cross-laminated (facies Fr). Although these are distinct facies, they are commonly closely associated, both vertically and laterally. Laminae are generally accentuated by textural and/or colour differences. Sandstone laminae, mostly very fine- and fine-grained but also medium- to coarse-grained, range in thickness from a few grains up to 10 mm and are moderately common.

Laminae within facies Ff are essentially horizontal and parallel. They are generally laterally continuous but also show low-angle truncations. Facies Fr includes asymmetric-, straight-, sinuous-, and lingoid-crested ripple structures with or without climbing-ripple lamination. Ripples are dominantly of siltstone though dark grey claystone laminae may enhance the internal structure.

Interpretation: For sediment sizes finer than about 0,1 mm (very fine-grained sand) the sequence of bed phases with increasing mean flow velocity is: no movement, small ripples and upper flat bed (Harms et al., 1975). Where the current is too weak to permit sediment transport, settling from suspension results in flat lamination. Coleman and Gagliano (1965) found that such laminae resulted from either differential settling of particles initiated by changes in current velocity or from changes in water chemistry. The former generally produces textural variations and the latter colour variations. Colour variations within this facies can, however, also be attributed to textural changes. The presence of low-angle truncations and shallow scour channels suggests periods of erosion followed by deposition from suspension.

McKee (1965) concluded that a vertical change from ripple lamination to flat lamination results from a shallowing of flow depth and a concommitant change from lower to upper flow regime. An alternative and equally affective mechanism for such a change may be a decrease in current strength and subsequent settling from suspension.

Facies Fb: Bioturbated mudrock

Description: This facies comprises mudrock in which the original stratification has been destroyed by biogenic action. A bioturbated admixture of clay, silt and sand results in a mottled texture in which vertical and horizontal burrows may or may not be discernable.

Interpretation: Facies Fb reflects the bioturbation of fine-grained sediment accumulating by slow, continuous suspension settling.

4.2.4 The Heterolithic Facies

Facies Fl: Lenticular-bedded sandstone and mudrock

Description: This facies comprises dark grey mudrock with included lenses of

sandstone (Fig 4.5). Generally more than 75% of the sandstone lenses are discontinuous both vertically and laterally. This facies is commonly transitional into flaser-bedded sandstone (Facies Sf1). The mudrock is generally massive but may be finely laminated. The sandstone is light grey to brown and varies in grain size from very fine to medium. A few lenses may be coarse-grained sandstone. Ripple cross-lamination is the dominant sedimentary structure and sand lenses show either sharp or erosive bases. They range in thickness from a few millimetres up to 50 mm but are most commonly in the 10 to 20 mm range. This facies attains thicknesses of up to 10 m but averages 0,5 to 2,5 m. Bioturbation ranges from rare (individual burrows are visible) to intense (burrows interpenetrant and original sedimentary structure highly modified).

Interpretation: Lenticular bedding is produced when isolated sand ripples are formed on a muddy substrate and subsequently covered by deposition of a clay or silt layer (Reineck and Singh, 1975). It is indicative of an alternation between current or wave action (in the lower part of the lower flow regime) which moved the sand, and slack water conditions during which sediment settled from suspension. Lenticular bedding occurs in intertidal zones, delta fronts, lake bottoms and abandoned river channels (Reineck and Singh, 1975).

Facies A1: Alternating sandstone and mudrock

Description: This facies comprises alternating light brown sandstone and dark grey mudrock on a 50 mm to 0,5 m scale (Fig 4.6). Thicker beds are arbitrarily assigned to their respective facies. The most characteristic feature of this facies is the parallel-sided nature and lateral continuity of the beds. Sandstones are generally fine- to medium-grained and ripple cross-lamination or flat lamination (with and without primary current lineation) are the dominant sedimentary structures. Massive beds, which may show grading, or trough cross-bedding are present to a lesser extent. The beds have planar or slightly irregular bases. The upper junctions are sharp or abruptly gradational into the

overlying mudrock over a few millimetres. They may show ripple marking. Mudrock beds are massive or flat laminated and lenticular bedding is fairly common.

Depending on the relative thickness of sandstone and mudrock three subdivisions can be recognised:

1. Sandstone and mudrock beds of equal thickness - facies Al,
2. Sandstone beds thicker than mudrock beds - facies AlS, and
3. Sandstone beds thinner than mudrock beds - facies AlF.

All three variations may be encountered in the same sequence with facies AlF grading upward into facies AlS through Al.

Trace fossils are plentiful in this facies. Vertical burrows, predominantly Siphonichnus and, to a lesser extent, Diplocraterion are common in sandstone beds whereas horizontal trails and burrows such as Skolicia, Palaeophycus and Planolites are found within the mudrock, particularly at sandstone-mudrock interfaces.

Interpretation: This facies is the same as the Facies 2 described by Hobday (1973) from the Tugela Ferry - Muden area and is similar to Reineck and Singh's (1975) "coarsely interlayered bedding". The sandstone-mudrock interbeds reflect fluctuating energy conditions. Sedimentary structures in the sandstone beds suggest currents in the lower part of the lower flow regime and the upper flow regime. However, as pointed out in the interpretation of Facies Sf, flat stratification may form under lower flow regime conditions by either rapid deposition or a high concentration of fine sediment in suspension. Massive beds showing normal grading were probably deposited by bedload flow related to resedimentation mechanisms (Sanders, 1965; Hobday, 1973). Mudrock beds reflect settling from suspension in an environment in which currents were weak or entirely absent. Bioturbation is probably related to this period of deposition.

The rapid upward transition from sandstone into siltstone suggests that currents responsible for the deposition of the sandstones waned rapidly.

Reineck and Singh (1975) suggested that facies of this kind reflect an environment in which the day-to-day or background deposition of silt and clay from suspension was repeatedly interupted by flood-/storm-related sand incursions. Such deposition is known to take place in a variety of environments including some parts of the marine shelf (Reineck and Singh, 1975), pro-delta or distal delta-front environments (Hobday, 1973), back-barrier environments (Bhattacharyya et al., 1980), and interdistributary embayments (Tavener-Smith, 1977). It is also found on intertidal flats (Reineck and Singh, 1975). A more detailed appraisal of this facies and its environmental significance is given in Chapter 5 in connection with facies models.

Facies W: Wavy-bedded sandstone and mudrock

Description: This comprises alternating ripple cross-laminated, light brown sandstone and dark grey, massive or flat laminated mudrock on a 5 to 30 mm scale (Fig 4.5). The sandstone varies from very fine to medium grained but is generally fine grained. The sandstone component is dominated almost completely by symmetric and asymmetric ripple cross-lamination. Flat lamination, mostly normally graded, occurs to a lesser extent but is intimately associated with the ripples. Facies W differs from Facies A1 in that the upper and lower surfaces of the sandstones are in the form of the ripple troughs and crests. Although this facies could be considered a variation of facies A1, there appears to be no gradational relationship between them and the sandstone morphology is distinctly different. Facies W and F1 are commonly closely associated.

Although Siphonichnus is the dominant trace fossil, this facies contains a greater proportion of horizontal burrows than facies A1. The latter include Scolicia, Planolites and Palaeophycus. The facies may be intensively

bioturbated, commonly to such an extent that it is difficult to discern individual beds.

Interpretation: Wavy bedding is formed under essentially the same conditions as lenticular bedding (Facies F1) but sand supply is more abundant (Reineck and Singh, 1975). Sedimentary association suggests that the sandstones were deposited by tractional currents, generally weaker and carrying less material than those responsible for the deposition of Facies A1 sandstones.

4.2.5   The Organochemical Facies

Facies C: Coal

Description: Up to 5 significant, areally extensive coal seams and a number of subordinate, impersistent and shaly coals are developed within the Klip River coalfield. The seams are flat laminated and comprise alternating bright and dull coal. Intercalations of cannel coal, shaly coal, carbonaceous shale, sandstone and mudrock are common. Some coals rest on root-bearing beds. The major seams range in thickness up to 3,5 m and can be traced laterally for distances exceeding 50 km. Subordinate seams are less than 300 mm thick and up to a few hundred or thousand metres in lateral extent.

Interpretation: Peat, the progenitor of coal, accumulates in luxuriantly vegetated swamp environments (Stach et al., 1982). The "drift theory" of coal formation has been largely discarded and most South African coal seams are considered to have formed in situ (Tavener-Smith, 1983). The presence of plant roots below certain seams strongly supports this belief. The stagnant, anaerobic reducing conditions necessary for peat formation in a swamp environment are provided by a water table high enough to prevent oxidation and the total decomposition of vegetal matter, yet not so high as to prohibit the growth of vegetation (Stach et al., 1982; Galloway and Hobday, 1983). For a suitable relationship between the height of the water table and the surface of

the peat swamp to exist, either subsidence (compactional, isostatic or tectonic) or a rise in the water table equal to the vegetal accumulation rate must occur (McLean and Jerziekywicz, 1978). Cannel coal forms under anaerobic conditions as subaquatic organic muds in swamp ponds and lakes (Stach et al., 1982). The presence of carbonaceous shale, sandstone and mudrock intercalations suggests that the swamps were periodically inundated by sediment-laden waters.

4.2.6 Soft-Sediment Deformation

Facies Sd and Fd: Penecontemporaneously deformed sandstone and mudrock

Description: Deformation structures, while not a common phenomenon in the Klip River coalfield, do occur in a variety of sediments. They are not confined solely to sandstones and are not considered a true sedimentary facies as they:

1.   Affect sediments deposited in a variety of sedimentary environments,

2.   reflect post-depositional processes, and

3.   are considered mainly as accessory structures associated with primary sedimentary features.

Penecontemporaneous deformation structures do, however, give some indication of sediment cohesion, loading, stability and porewater pressure. In a few instances, especially in borehole cores, the primary sedimentary structure is completely modified by penecontemporaneous deformation and so facies Sd and Fd are used. Deformation structures observed include load and flame structures, convolute stratification, water-escape structures and faults.

Interpretation: Soft sediment deformation is discussed in detail by Kuenen (1958), Allen and Banks (1972), Blatt et al. (1972), Reineck and Singh (1975), Lowe (1976a,b) and Neumann-Mahlkau (1976). It is generally attributed to de-watering, overloading or differential loading, seismic disturbances, gravity slumping or current shear.

CHAPTER 5

SEDIMENTARY MODELS

5.1 INTRODUCTION

The identification of depositional environments in the ancient sedimentary record is based on Hutton's "Principle of Uniformitarianism" (processes now operative in the formation of sediments provide the best explanation of past events) and Walter's "Law of Succession of Facies" (facies observed beside each other in the past are associated vertically in the stratigraphic record). Implicit in these two fundamental concepts is recognition of the fact that reconstruction of ancient depositional environments can be achieved only by the detailed study and definition of present-day environments and sedimentary processes. The identification of palaeoenvironments by the recognition of modern counterparts is, however, beset with complications. For instance, it is probable that some environments that existed in the past are not represented today, and geological processes have operated with differing intensities through time. Furthermore, it has been recognised that deposits of the same environment differ in various climatic zones and also with respect to the availability and grain size of sediment (Reineck and Singh, 1975).

In spite of these complications there has, over the last two decades, been an upsurge in the genetic classification of ancient sedimentary deposits and the development of facies models. A major influence on the development and classification of facies models and the use of vertical profiles in palaeoenvironmental reconstruction was the work by Visher (1965). He proposed that a specific sedimentary environment gives rise to a unique vertical profile and on these grounds erected six fundamental models. These idealised models were regarded as standards for interpreting stratigraphic sequences. Subsequent work on palaeoenvironmental reconstruction emphasised the great variability in lithofacies and the impracticability of any environment being characterised by a single diagnostic facies model or vertical profile. A good

example of this was the establishment of vertical facies models by Miall (1977a, 1978) and Rust (1978) to distinguish the various facies associations in braided river deposits. Referring specifically to fluvial deposits, but also to other environments in general, Miall (1983) stated that while the diagnostic approach to lithofacies assemblages and vertical profiles has successfully served to categorise various deposit types and certain geomorphic patterns (eg. channel aggradation), it has been less sucessful in predicting lateral facies relationships and the manner in which various elements of a sedimentary sequence are stacked. In order to appreciate lateral facies relationships or the "architecture" of a sedimentary environment, three-dimensional data are required. A three-dimensional appreciation of facies relationships has been demonstrated by Ferm et al, (1979); Marley et al. (1979); Horne et al. (1979); Flores (1979, 1981); Le Blanc Smith (1980b) and Ryer (1981) to be of the utmost importance in understanding and predicting the control depositional environment had on coal-seam distribution and behaviour.

Vertical profiles are the primary means by which sedimentary successions are illustrated. They provide an easy means of depicting vertical facies relationships and, by the use of cross sections, lateral facies relationships can also be represented. Although the concept of cyclic sedimentation is frequently criticised and commonly open to subjective analysis, it is one of the most fundamental to environmental interpretation. It offers the possibility of recognising some order in a thick and complex succession. The causes of cyclicity, whether by autocyclic or allocyclic mechanisms, are important when considering the basinal evolution of a sedimentary deposit.

Because most depositional scenarios based on the idea of normal, steady sedimentation, Reading (1978) emphasised the importance of not neglecting apparently random or unique events. Normal or day-to-day sedimentary processes persist for lengthy periods of time, but net sedimentation is usually slow. Episodic or "catastrophic" events, on the other hand, occur almost

instantaneously and deposition is rapid. The problem of distinguishing between normal and episodic sedimentation and predicting the impact the respective processes had on the sedimentary record has long been debated in geology (eg. Barrel, 1917; Simpson, 1952; Gretener, 1967). Recently, Dott (1983) posed the question as to whether sedimentary rocks record mainly average, continuous processes or relatively rare, large-magnitude ones separated by long, nondepositional intervals. He considered that an "instinctive abhorrence of unique events" and the "constraints of uniformitarianism" have created a strong bias towards average conditions where continuity rather than discontinuity is assumed. He concluded that sedimentary breaks represent more time than the preserved strata.

The importance of the rare or episodic event can, however, be overstressed. It is possible that normal, everyday sedimentary processes could result in rapidly deposited increments being entirely removed, reworked or modified to varying degrees. The preservation potential of any depositional unit is a largely unexplored facet of sedimentology and introduces a further element of uncertainty in structuring facies models. The preservation of any facies varies considerably and an assessment of how much sediment escapes erosion (both vertically and laterally) by subsequent processes is necessary for the comparison of modern and ancient deposits (Reading, 1978). Despite these limitations and uncertainties, the use of vertical profiles in developing both local and general facies models has remained a powerful tool in environmental reconstruction and interpreting both temporal and spatial variations in depositional environments.

Harms et al. (1975) and Walker (1975) suggested that a facies model should fulfill a fourfold function:

1.   It must act as a norm or standard for comparison with other models.

2.   It should act as a guide for future observations.

3.   It must serve as a predictor in new geological situations.

4.   It must act as a basis for hydrodynamic interpretation of the environment that it represents.

An understanding of depositional environments that prevailed during Vryheid Formation time in the study area relies heavily on the facies model concept despite several limitations. These include the absence in many areas of adequate three-dimensional borehole and outcrop data and reliance on borehole sections which lack detailed sedimentological information both within and below the Coal zone. Although each depositional environment is dealt with in isolation, cognisance is taken of the influence adjacent environments had on sedimentary processes. These models are placed in a regional and stratigraphic context in ensuing chapters.

5.2   FLUVIAL FACIES ASSOCIATION

Deposits interpreted as fluvial in origin, as distinct from distributary channel deposits, are confined to the Coal zone. For purposes of discussion the fluvial facies is divided into a channel subassociation and an interchannel subassociation.

5.2.1   Channel Subassociation

Description

The channel subassociation is developed as sheet sandstones which form distinctive horizons readily recognisable in outcrop and borehole cores. The term "sheet sandstone" is used in the same sense as Potter (1967) to define a multilateral and multistoried sandstone body with a width to thickness ratio exceeding 1 000:1. In terms of lithology and internal characteristics, however, the usage here is synonomous with that of Campbell (1976) which differs from Potter's definition.

Individual sheet sandstones can be traced in the inferred palaeodip direction for distances of at least 60 km and over areas in excess of 1 500 sq km. Thicknesses range from 3 to 30 m, but the average is between 12 and 20 m. Variations in thickness over short distances are common and this is well illustrated in Figure 9.15 (see Chapter 9). The sheet sandstones thin and pinch out abruptly at their margins which are well defined and straight to slightly sinuous. In places lobe-like features up to 5 km wide and 8 km long extend perpendicularly from the edges of sheet sandstones (Fig 9.15). Deposits of the overbank subassociation overlie sheet sandstones with a conformable, erosive base. Borehole data suggests that the amount of lateral interfingering between the two subfacies is limited.

Internally, each sheet sandstone comprises a number of vertically repetitive upward-fining channel sequences (Fig 5.1 a, b). These vary in thickness from 1 to 8 m but only rarely are complete units developed, the upper parts of each sequence are mostly missing. Multilateral and multistoried chanelling results in laterally and vertically truncated channel geometries. Outcrop and borehole data indicate that the channel-fills are between 0,3 and 1 km in width.

Detailed analysis of outcrop sections indicates that the channel subassociation comprises two discrete assemblages of sedimentary facies. Type 1 sequences (Fig 5.1 a) commence with a basal facies Cm or pebbly sandstone (1 to 150 mm thick) overlying a flat to low-angle erosive surface with relief rarely exceeding 100 mm. This is succeeded by a medium- to coarse-grained, pebbly sandstone between 1 and 4 m thick which generally only shows a very slight upward decrease in grain size. Pebble stringers 20 to 50 mm thick occur throughout the sequence. The sandstone is dominated by facies Sp although up to 2 m of facies St (sets 0,4 to 0,6 m thick) may be developed at its base. Facies Sp sets vary in thickness from 100 mm to 1,2 m, are tabular- to wedge-shaped and commonly incised by solitary trough cross-beds up to 250 mm thick. Planar cross-beds generally show an upward decrease in set thickness within individual

sequences (Fig 4.3) but may also be of uniform thickness throughout. In the former case the basal tabular cross-set, generally in the order of 0,8 to 1,2 m thick, is overlain by smaller sets (200 to 400 mm thick). Basal sets between 1,5 and 2,5 m thick, the foresets of which may describe a large trough on bedding plane surfaces, occur to a limited extent. In some instances channels up to 2 m wide and 1 m deep filled with small- to medium-scale facies St are incised into the facies Sp component. Mudrock lenses, generally comprising facies Fm, Ffl or Ff up to 400 mm thick may terminate these sequences.

Type 2 sequences (Fig 5.1 b) are of a similar thickness to those of type 1. They consist of a basal facies Cm or pebbly sandstone overlying an erosive base, fining up to very coarse-grained to coarse-grained sandstone which, in turn, may fine upward into medium-grained sandstone. Scattered pebbles and pebble lenses occur throughout the sequence. In certain outcrops relief at the base of some sequences is up to 1,6 m over less than 50 m. The sequence is dominated by trough cross-bedding (facies St) which shows a progressive upward decrease in set thickness. Medium-to large-scale sets (0,3 to 0,8 m) are present near the base and small-scale sets (100 to 300 mm) occur at the top. Facies Sr and Sf may be closely associated with the latter. Solitary, large-scale planar cross-beds commonly interrupt the sequence. Some sequences culminate in facies Fm, or Ff up to 400 mm thick.

The upper, finer component of both types of sequences is absent in all but a few instances due either to non-deposition or penecontemporaneous erosion. In contrast, the uppermost sequences of each sheet sandstone are invariably fully developed and culminate in a coal seam. Coal seams are rare within the sheet sandstones; they are thin (up to 300 mm) and lenticular and overlie either sandstone or mudrock. In the former case rootlet bioturbation is commonly visible.

KEY
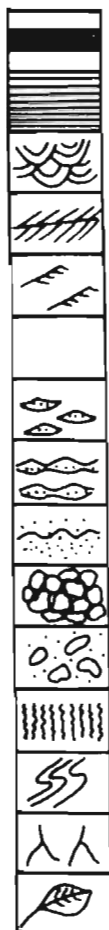
C — COAL

Sf — FLAT STRATIFICATION

St — TROUGH CROSS-BEDDING

Sp — PLANAR CROSS-BEDDING

Sr — RIPPLE CROSS-LAMINATION

Sm/Fm — MASSIVE BEDDING

Fl — LENTICULAR BEDDING

W — WAVY BEDDING

EROSIONAL BASE

Cc — CLAST SUPPORTED CONGLOMERATE

Cm — MATRIX-SUPPORTED CONGLOMERATE

Sb/Fb — BIOTURBATED SEDIMENT

Sd/Fd — DEFORMED BEDDING

ROOTLETS

PLANT FOSSILS

GRAIN SIZE

CLAYSTONE
SILTSTONE
FINE-GRAINED
MEDIUM-GRAINED
COARSE-GRAINED
VERY COARSE-GRAINED

SANDSTONE
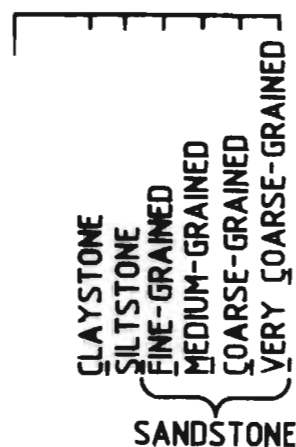
BIOTURBATION
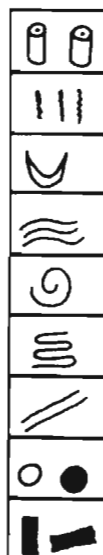
SIPHONICHNUS

TIGILLITES/SKOLITHOS

DIPLOCRATARION

NEREITES

SPIRODESMOS

HELMINTHOPSIS

SCOLICIA

PLANOLITES/PALAEOPHYCUS

UNIDENTIFIED TRACES

KEY TO SECTIONS
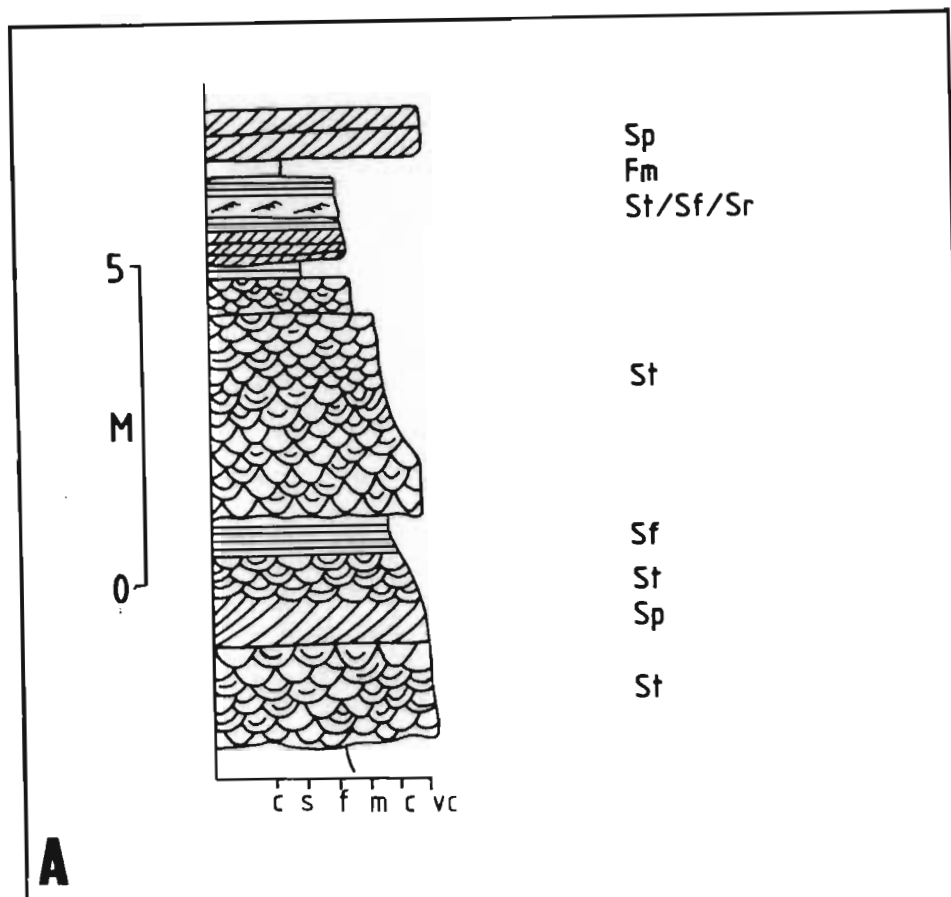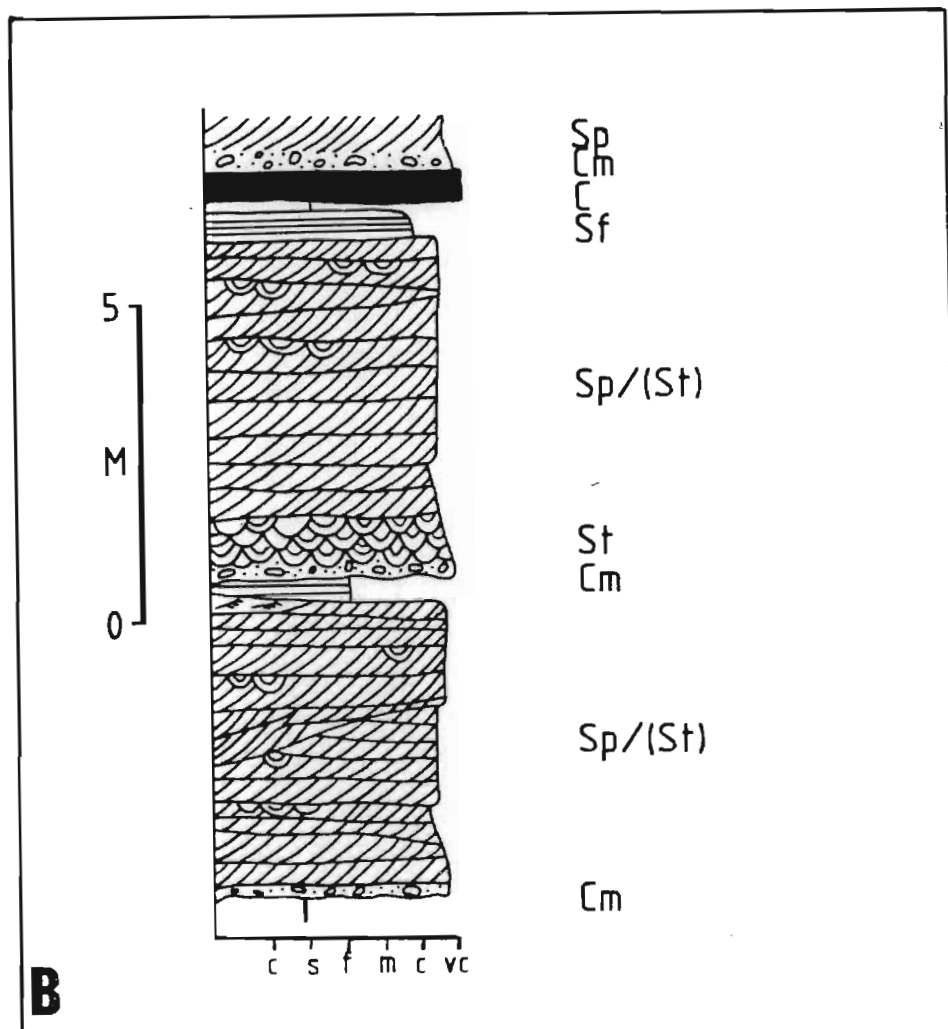
Figure 5.1 A. Vertical section through part of sheet sandstone dominated by Type 1 fluvial sequences. Cinderford section. B. vertical section through part of sheet sandstone dominated by Type 2 fluvial sequences. Krantzkop section.
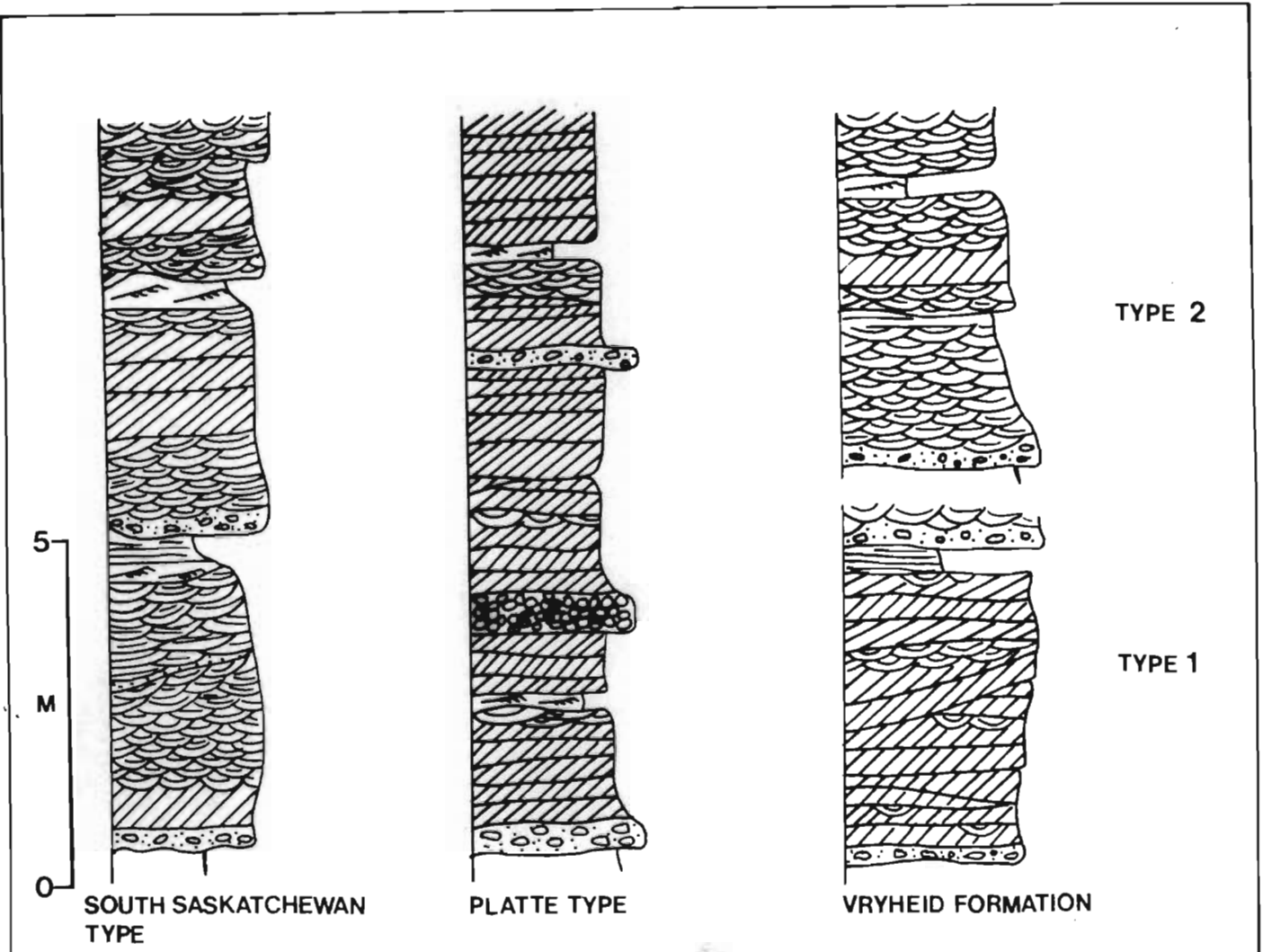
Figure 5.2. Comparison between South Saskatchewan and Platte type braided-river vertical profiles and those of the Vryheid Formation described in this thesis.

Lithologically, the sandstone is a feldspathic to lithic arenite and is characterised by poor sorting, a high silt and clay content (up to 15% by volume) and rapid textural changes. Quartz comprises 50 to 70% of the framework grains while feldspars, including orthoclase, microcline and sodic plagioclase, comprise 5 to 20%. Plagioclase is the most common feldspar and is usually partially to totally altered to sericite. Quartz grains nearly always show slight to moderate undulatory extinction. Rock fragments make up 5 to 15% of the sandstone. Metamorphic rock fragments typically occur in greater abundance than those of sedimentary origin and are predominantly polycrystalline quartz, though small amounts of feldspar and muscovite may also be present. Mudrock fragments are commonly squeezed between and around the more rigid grains.

Accessory grains constitute 1 to 5% of the bulk volume of the sandstone. Heavy minerals, in order of abundance, include detrital muscovite, sundry opaques, red and green garnet, chlorite and zircon. Carbonaceous debris may account for some of the opaque minerals.

Palaeocurrent measurements from individual channel sequences show a well-developed unimodal distribution with r values close to 1 and mean angular deviations of less than 20°. Within the upper, finer sand component of some sequences, however, small-scale cross beds are commonly orientated at high angles (up to 80°) to the mean orientation of the underlying large-scale cross beds. Planar cross-beds within type 2 sequences may also be orientated at high angles to the associated trough cross-beds. With few exceptions there is no significant difference in mean palaeocurrent orientation between sequences within the same sheet sandstone.

Bioturbation within the channel subfacies is rare. It is, however, present to varying degrees within the argillaceous component of the uppermost channel sequence of each sheet sandstone. Siphonichnus, Tigillites, Diplocraterion, Scolicia and Nereites are the dominant ichnogenera present.

Discussion

The vertical association of sedimentary facies, lithological characteristics and unidirectional palaeocurrent distributions suggest the deposits of braided fluvial systems (Williams and Rust, 1969; Smith, 1971, 1972; Cant and Walker, 1976; Miall, 1977a, 1978). The multistoried and multilateral nature of the sheet sandstones indicates deposition by coalescing and aggrading braided-fluvial systems on a broad alluvial plain, an interpretation consistent with those of Ore (1964), Potter (1967) and Campbell (1976).

The conglomerates and pebbly sandstones at the base of both type 1 and 2 sequences were transported and deposited by an upper flow regime current; the

lack of stratification and poor sorting indicates that the sediment was deposited rapidly (Harms and Fahnestock, 1965). The erosive base points to a well-developed and locally extreme current turbulence. Such conditions compare closely with the waning phase of floods recorded from modern braided river channels (Doeglas, 1962; Rust, 1972, 1978; Miall, 1977a).

Facies Sp-dominated type 1 sequences represent sedimentation by migrating transverse or lingoid bars (Smith, 1971; Cant and Walker, 1976; Miall, 1977a). Each sequence is believed to represent a single migrating bar complex. The development of facies St at the base of some sequences reflects dune migration within channels containing deeper and faster flowing water than where transverse bars were able to form. This implies that channel-floor aggradation took place prior to bar formation (Williams and Rust, 1969; Cant and Walker, 1976, 1978; Cant, 1978). The upward decrease in set thickness displayed by some sequences is attributed to waning current processes and the shallowing of water over bar tops (Williams and Rust, 1969; Cant and Walker, 1976). In those sequences displaying no such upward decrease, bars probably originated under the influence of high runoff. Solitary trough cross-bed sets overlying and incised into planar cross-beds reflect dune migration across bar surfaces. Trough cross-bedded channel fills formed by the development of chutes and channels as a consequence of falling water levels. Mudrock facies preserved at the top of some sequences resulted from channel abandonment at the end of a flood or lateral migration phase (Miall, 1977a,b). Alternatively, bar emergence may have caused the development of a sand flat with the accumulation of clay and silt from suspension in shallow water-filled depressions (Cant and Walker, 1978).

Type 1 sequences resemble the deposits of the Platte River (Fig 5.2) which is dominated by wide, shallow reaches with only a few deep channels (Smith, 1971, 1972). Braiding is achieved by subdivision of flow around bars and their dissection at low stage. It is believed that the lack of marked topographic relief within the river course would result in apparently non-cyclic sequences dominated by successive sets of tabular-planar cross-bedding.

Type 2 sequences reflect deposition dominated by migrating dunes within channels where the water was deeper than in type 1 sequences (cf. Smith, 1971, 1972; Cant and Walker, 1978). The progressive upward decrease in both grain size and set thickness is a function of channel aggradation (Cant and Walker, 1976, 1978). Solitary planar cross-beds are believed to be the product of transverse bars that did not develop into a sand flat. Eventual channel abandonment resulted in the accumulation of silt and clay in pools of standing water.

Deposits of this subassociation are similar in character to those of the South Saskatchewan River (Cant and Walker, 1978) but also show resemblances to those of the Platte (Smith, 1971, 1972) and Tana Rivers (Collinson, 1970a) (Fig 5.2). The South Saskatchewan is a braided river characterised by major channels 3 to 5 m deep and up to 200 m wide flowing around and between sand flats and sand-flat complexes up to 2 000 m long. Cross-channel bars are formed at channel junctions or where the channel widens. Bars persist through long periods of varying flow during which ripples, dunes, small bars and upper flow regime flat lamination may form on them. The braided pattern is achieved when cross-channel bar tops become exposed during falling stage. The resulting sand flats, which divide the cross-channel bar into a series of smaller bars, are, in turn, dissected by minor channels. Smaller scale sets form when bars migrate from channels into the shallow water of the flats.

The sparse development of mudrock deposits is believed to be due to both non-deposition and their poor preservation potential. Braided streams are not characterised by large areas of flood plain (Miall, 1977a); clay and silt accumulate in abandoned stretches of a channel and ponds or shallow depressions on sand flats. Frequent channel shifting as a result of avulsion during major floods commonly leads to complete destruction of sand flats while fluctuations in water depth within channels can lead to pronounced bar and sand flat modification (Miall, 1977a; Cant and Walker, 1978; Blodgett and Stanley, 1980). The absence of significant coal seams associated with this subfacies testifies

to the ephemeral nature of features within a braided fluvial system, though Miall (1977a) noted that many modern streams have abandoned areas with a sparse to thick vegetation cover which are inundated at only the highest flood stages. The occurrence of coal in this subfacies is discussed in more detail in Chapter 8.

Palaeocurrent distributions described from the sheet sandstone deposits differ from those recorded in modern rivers (see Williams and Rust, 1969; Collinson, 1970a; Miall, 1977a,b; Cant and Walker, 1978 and many others) which show a much greater directional variability of medium- to large-scale structures. This could be a function of the greater preservation potential of structures formed at high-water stage and orientated at a low angle to channel trend.

5.2.2   Interchannel Subassociation

Description

In terms of volume the interchannel or overbank subassociation is an insignificant component of the fluvial association. However, its palaeoenvironmental and economic importance can be gauged from the fact that most exploitable coal seams are associated with it. The subfacies occurs as discrete sedimentary units rarely exceeding 6 m in thickness characterised by an association of sandstone and mudrock facies. Each unit corresponds laterally with a sheet sandstone. Mudrock facies dominate; the most common being facies Fr (with and without climbing ripple lamination) and Fm in beds 0,2 to 2,5 m thick. Facies W is also common. Sandstones are predominantly fine grained but range to very coarse or even pebbly. Beds are up to 1,5 m thick, but most are less than 0,5 m, repeatedly alternating with mudrocks. Most sequences show facies St (small to medium scale) Sr, Sf and Sfl. The coarser beds are commonly massive (facies Sm).

Upward-coarsening and upward-fining sequences between 0,5 and 2 m are an integral part of this subassociation. The former pass upward from facies Fm or Fr into a sandy facies Fr, facies Sr or facies St. The latter grade from erosively based medium- to coarse-grained facies St through fine-grained facies Sr or Sf into facies Fm or Fr. Many sandstone beds of this subassociation are massive and have a sheet-like morphology. Some upward-fining sequences on the other hand, are straight to slightly sinuous channel-fills tens or hundreds of metres in width. In general it is difficult to ascertain the morphology of these units owing to their limited extent and the low density of boreholes.

Bioturbation is generally rare but dense to moderate concentrations of Siphonichnus, Scolicia and Nereites occur sporadically. Wood and leaf imprints and debris are common, as are root traces.

Discussion

The relative thinness of this subfacies compared with the sheet sandstones, even allowing for the effects of differential compaction, indicates that deposition away from the braid plain was infrequent. Sedimentary processes in the overbank environment seem to have been essentially similar to those of interdistributary embayments which are discussed in more detail elsewhere in this chapter.

The overbank area is envisaged as being covered by swamps, lakes and ponds. Small rivers traversed the plain but most sediment was introduced by periodic flooding of the adjacent major fluvial channels. There is no evidence that the sheet sandstones were lined by levees, so it is probable that most floodwater spilled onto the overbank area as a sheet flood rather than through a single breach point. Upward-coarsening sequences and massive sheet sandstones are thought to have been formed by this process. Upward-fining sequences, on the other hand, reflect deposition by fluvial channel processes. While it is apparent that some channels were straight to slightly sinuous, and bed-load

dominated it is possible that many were meandering. Insufficient three-dimensional data exist to confirm this view. The absence of caliche nodules, pedogenic deposits or any desiccation features argues against the likelihood of extensive sub-aerial exposure. The scarcity of trace fossils suggests an alluvial or upper delta plain setting. Influxes of saline water to provide localised conditions suitable for colonisation by marine or brack-water organisms may have been brought about by water-table fluctuations or storms.

5.2.3   Depositional Model

Any depositional model must take cognisance of the following 3 features:

1. The lateral extent of the fluvial association in general and sheet sandstones in particular,

2. the relationship between the channel and overbank subassociations, and

3. the close association of thick and extensive coal seams with the fluvial deposits.

The importance of understanding depositional processes that operated within the channel subassociation cannot be underestimated, for they provide a basis for understanding the nature of deltaic and other coastal sedimentation, especially with respect to lithological distribution and flow patterns. Controls on the distribution of channel and overbank subfacies and coal seams are discussed in more detail in subsequent chapters.

The sheet sandstones represent deposition by braided rivers on an extensive, essentially unconfined medial to distal alluvial plain characterised by moderate gradients (Miall, 1977a). The rivers displayed features reminiscent of the modern South Saskatchewan and Platte in that they carried a heavy sediment

load, were broad and shallow, and contained transverse to lingoid bars consisting of coarse-grained, granular and pebbly sands. Braiding was achieved by the subdivision of flow around emergent bars (sand flats) and by the dissection of bars at low flow stage. Upward-fining sequences indicate that some channels were up to 5 m deep, but most were shallower. Like most modern braided rivers, flow was probably very variable and dependent on seasonal rainfall fluctuations (see Williams and Rust, 1969; Collinson, 1970a; Smith, 1971, 1972; Cant and Walker, 1978).

The mutually erosional nature of channel sequences (both laterally and vertically) suggests that channels shifted with apparent ease and had easily eroded banks. This is probably a function of the non-cohesive nature of the sand-dominated plain and almost entire absence of vegetation on the braidplains The apparent lack of any significant erosion at the base of individual sheet sandstones is attributed to the ease with which channels could migrate across the unconfined braidplain and their ability to increase their cross-sectional volume by bank erosion rather than downward incision. Scattered, localised swamps existed for short periods, possibly in abandoned channels or ponds on stable islands or sand flats.

The well-defined edges of individual sheet sandstones are attributed to pre-depositional topographic controls (induced principally by differential compaction) and the vegetated and argillaceous (and hence cohesive) nature of the abutting overbank area. It is considered likely (see Chapter 9) that the sheet sandstones occupied the topographically lower areas of the plain. Only when the braid plain experienced severe flooding did water spill over onto the overbank area.

The overbank area is believed to have been covered by swamps much of the time; peat accumulation was interrupted for brief periods by influxes of sediment-laden currents from the adjacent braid plains. These areas were traversed by minor rivers and also contained numerous ponds and small lakes.

5.3   DELTAIC FACIES ASSOCIATION

The deltaic environment can be broadly divided into prodelta, delta front (mouth bar) and delta plain (Fig 5.3). The prodelta and delta front are completely subaqueous while the delta plain is partially subaqueous and partially subaerial.

5.3.1   Prodelta Subassociation

Description

Prodelta deposits are dominated by dark brown to dark grey or black mudrock of facies Fm, Ff, Fl and Fb. They occur at the base of upward-coarsening sequences and overlie sandstones of the immediately preceding sequence with a sharp base. Three distinct types of deposits are recognised on the basis of lithology and degree of bioturbation:

1.   Predominantly black facies Fm with minor Fl and Ff. These vary in thickness from 7 to 36 m and are generally confined to upward-coarsening sequences with thicknesses in excess of 20 m. Reddish-brown iron oxide laminae and silty very fine- to fine-grained sandstone beds (facies Sr and Sf) up to 50 mm thick, but mostly between 5 and 10 mm, are rare but characteristic elements of this subassociation. Bioturbation is sparse to absent; most biogenic features are associated with the sandy intercalations. However, dense concentrations of Siphonichnus, Skolicia and Planolites are commonly present in the lowermost 1 to 3 m of these sequences.

2.   Brown to light grey siltstone of facies Ff. Laminae are up to 10 mm thick, normally graded and display low-angle planar discordances. Bioturbation is rare to absent; only one occurrence of Skolicia was noted.
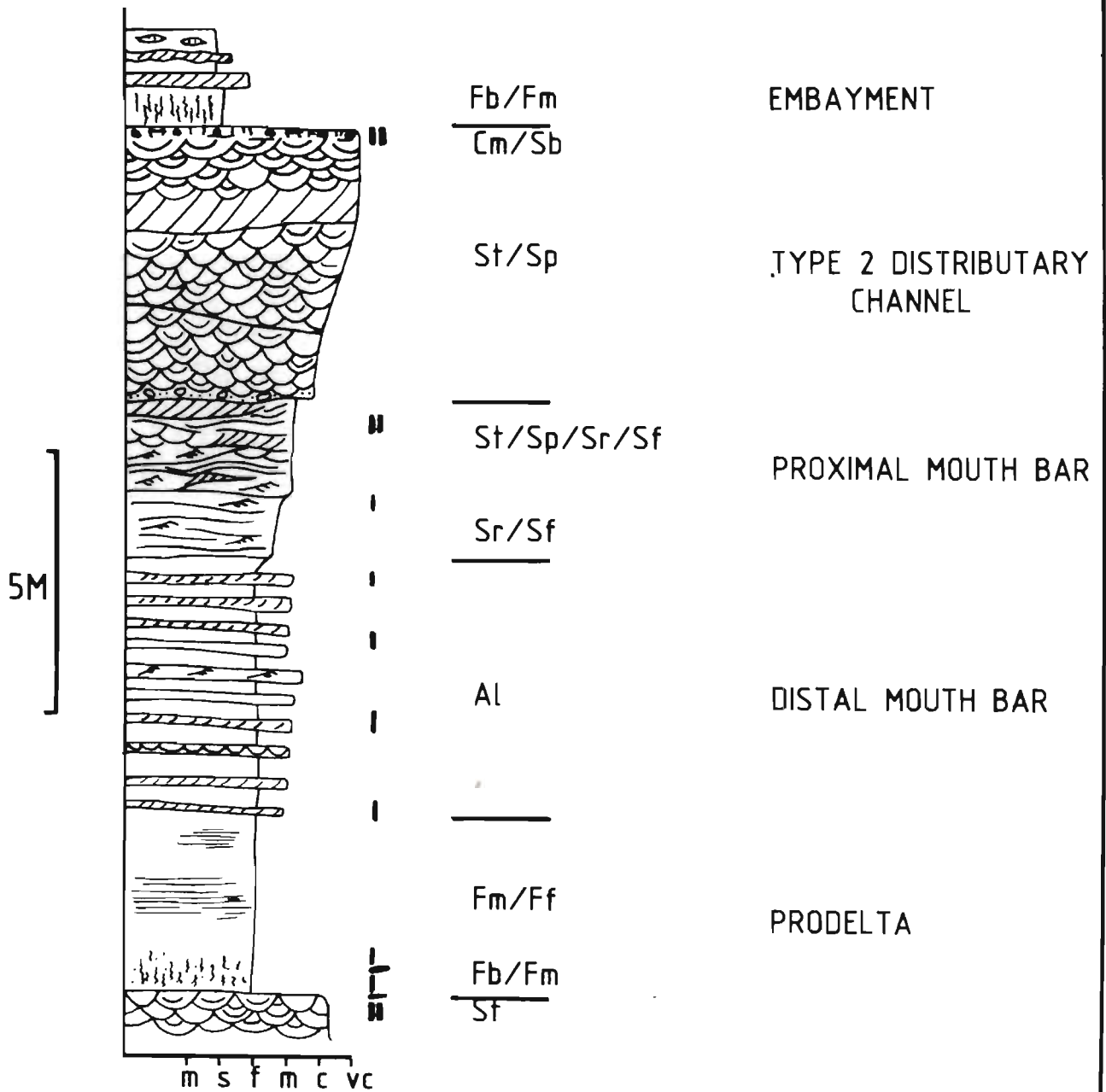
Figure 5.3. Schematic section illustrating the main elements of a prograding deltaic sequence.

3. Extensively bioturbated mudrock of facies Fb and Fl. Dark grey mudrock (commonly sandy) is the dominant sediment and only rarely are sedimentary structures preserved. In some instances subordinate lighter-coloured silty sandstone beds up to 100 mm thick may be discernable. Siphonichnus and Nereites are the dominant ichnofossils, but Planolites and Skolicia also occur.

Facies of the prodelta environment mostly pass upward gradationally into either facies Fl, Al, W or Sr or are overlain by erosively based, coarse-grained sandstone of facies St or Sp. In some sequences facies Fm grades upwards into Facies Fb.

Discussion

The prodelta environment is the most seaward component of the delta system (Blatt et al., 1972; Reineck and Singh, 1975; Galloway and Hobday, 1983). Deposition is dominated by the settling of clay and silt from suspension at depths generally below wave base (Fisher et al., 1969).

Each of the 3 facies associations described above is believed to reflect variations in sediment supply and the position of the prodelta environment with respect to active delta lobes. Dark grey mudrock of facies Fm accumulated in the most distal parts of the prodelta where the influence of distributaries was small. Coarser intercalations represent sporadic influxes of sand during storms or floods. Bioturbation in the basal few metres is probably related to the early stages of the transgression of the underlying abandoned delta lobe. The rarity of trace fossils above this is thought to be a function of either an anaerobic environment, poor substrate conditions or a high sedimentation rate.

The well developed, graded laminae of facies Ff together with the general absence of trace fossils strongly suggests a rapid rate of sedimentation, probably seaward of an active distributary channel. Extensively bioturbated

facies Fb deposits reflect a low rate of sedimentation, plentiful supply of nutrients and marine or brackish waters seldom diluted by influxes of fresh water (see Chapter 3). The position of this facies in upward-coarsening sequences with respect to both underlying (facies Fm in some cases) and overlying facies suggests an environment more proximal than that of facies Fm. However, the relative scarcity of sand-sized material and high degree of bioturbation is indicative of a setting seldom influenced by distributary channel influxes. Thus, it is envisaged that facies Fb represents accumulation in a distal embayment setting between active delta lobes away from the main locus of distributary channel sedimentation. Deposition above fair-weather wave base is unlikely in view of the substantial quantities of clay and silt which must have settled from suspension. Later channel avulsion resulted in this facies being overlain by others representing more proximal environments. Of course, deposition may also have been seaward of a delta lobe fed only by minor, distributary channels carrying a finer sediment fraction than the main distributary trunks.

5.3.2   Distal Distributary Mouth Bar Subassociation

Description

Sediments assigned to this subassociation are dominated by facies Al and W. Typically, however, the two are only rarely associated in a vertical sense. Both facies are described in Chapter 4 but to elucidate specific sandstone depositional mechanisms further detail is provided below.

Facies Al (Figs 5.4 and 5.5) characteristically coarsens upward in most sequences by an increase in the proportion of sandstone to mudrock; facies AlF passes upward into facies AlS through facies AlE. Less conspicuous, though generally present, is an upward increase in sandstone grain size from fine at the base to medium or coarse at the top. The sandstone is generally moderately to poorly sorted and feldspathic. Ripple cross-lamination is the dominant

sedimentary structure. Flat stratification, commonly associated with primary current lineation, is also present and may be the dominant structure in the upper part of facies Al sequences. Ripples are manifested on bedding planes as unidirectional trough laminae up to 100 mm wide and 2 m long. Most beds contain only a single type of sedimentary structure but where two are developed, flat lamination is generally overlain by ripple cross-lamination. Trough cross-bed sets up to 200 mm thick become prominent in the upper part of most facies sequences. Massive beds (facies Sm) are present, but not common. Some sandstone beds amalgamate laterally into a single bed but few were observed to pinch out laterally within a single outcrop.

Facies W intercalations occur on a 5 to 50 mm scale and, like facies Al, display an upward increase in the proportion of sandstone to mudrock in most sequences. These intercalations are commonly poorly defined and have a disorganised appearance owing to their lenticular and/or undulating nature. Although the sandstone component is predominantly ripple cross-laminated (displaying both symmetric and asymmetric ripples), flaser bedding and flat lamination are also developed. The sandstone is rarely coarser than fine to medium, is silty in places and moderately well sorted. The mudrock component is mostly massive but flat lamination and lenticular bedding are developed.

A significant difference between facies Al and W in the distal distributary mouth bar subfacies is that the former is associated with upward-coarsening sequences between 30 and 60 m thick. The latter is present only in thinner sequences.

Discussion

The distal mouth bar environment records background or normal "day-to-day" sedimentation of mud from suspension repeatedly interrupted by rapid, short-lived incursions of coarse detritus from delta distributaries (Elliot, 1978; Galloway and Hobday, 1983). General facies associations and the trace

fossil assemblage leaves little doubt that facies Al and W were deposited in a shallow, sublittoral marine environment. It is probable, however, that each was deposited under different environmental conditions.

Internal sedimentary structures of <u>facies Al</u> indicate that deposition was dominated by offshore-directed unidirectional currents ranging in strength from lower to transitional flow regime. The predominance of ripple cross-lamination suggests that most of the sandstones were deposited by ripple-train or small dune migration. The upward increase in the development of trough cross-lamination and upper flow regime flat stratification and a concomitant overall increase in grain size in individual sequences suggests a shoreward shallowing of the environment and greater proximity to both current and sediment source.

The presence of a few massive sandstone beds suggests that some deposition probably involved debris and grain flow processes, a view supported by van Vuuren (1981). The conditions necessary to sustain such deposition (see Rupke, 1978) were probably uncommon and of a local and short-lived nature.

The main argument centres around which process carried the sand offshore; was the sand driving the current (turbidity current) or was it entrained by the current? A comparison between the sandstones of classical Bouma turbidites and those of Facies Al indicates that while both display a similar morphology (parallel-sided, laterally continuous and interbedded with mudrock) a number of features make it unlikely that deposition was solely by turbidity currents.

These include:

1. A marked difference with respect to internal sedimentary structures (cf. Bouma, 1962). Facies Al sandstones show no sole structures, no significant grading and only rarely display more than two types of sedimentary structure. Significantly, ripple cross-lamination, the dominant sedimentary structure of facies Al, generally forms only a small proportion of the Bouma sequence.
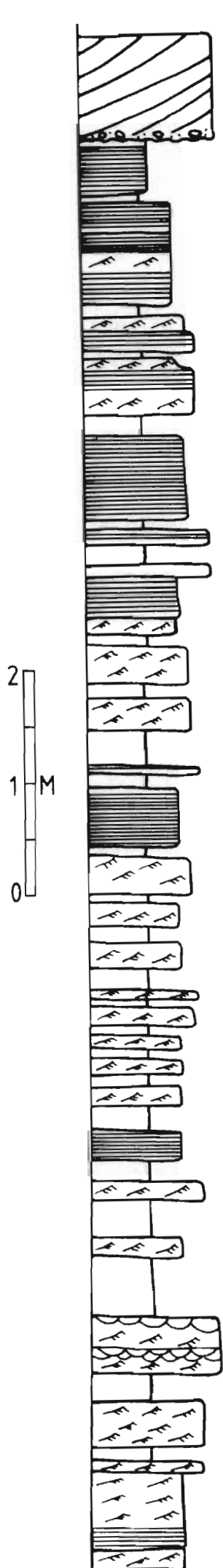
Sf

Sr

Fm
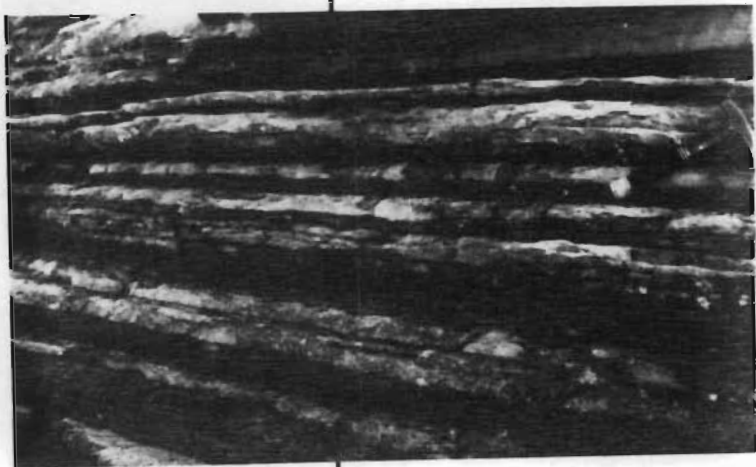
Figure 5.4. Section through and photograph of basal portion
of Gowrie section illustrating distal distributary mouth
bar facies Al deposits.

Figure 5.5. Section through prograding deltaic sequence.
Cinderford section.

Sp/St

Sp/Sr/St

St

AlS

Al

AlF

Sf/AlS

AlS

Al

AlF

Fb/Ff/Fm

Fb

10

M

0

c s f m c vc
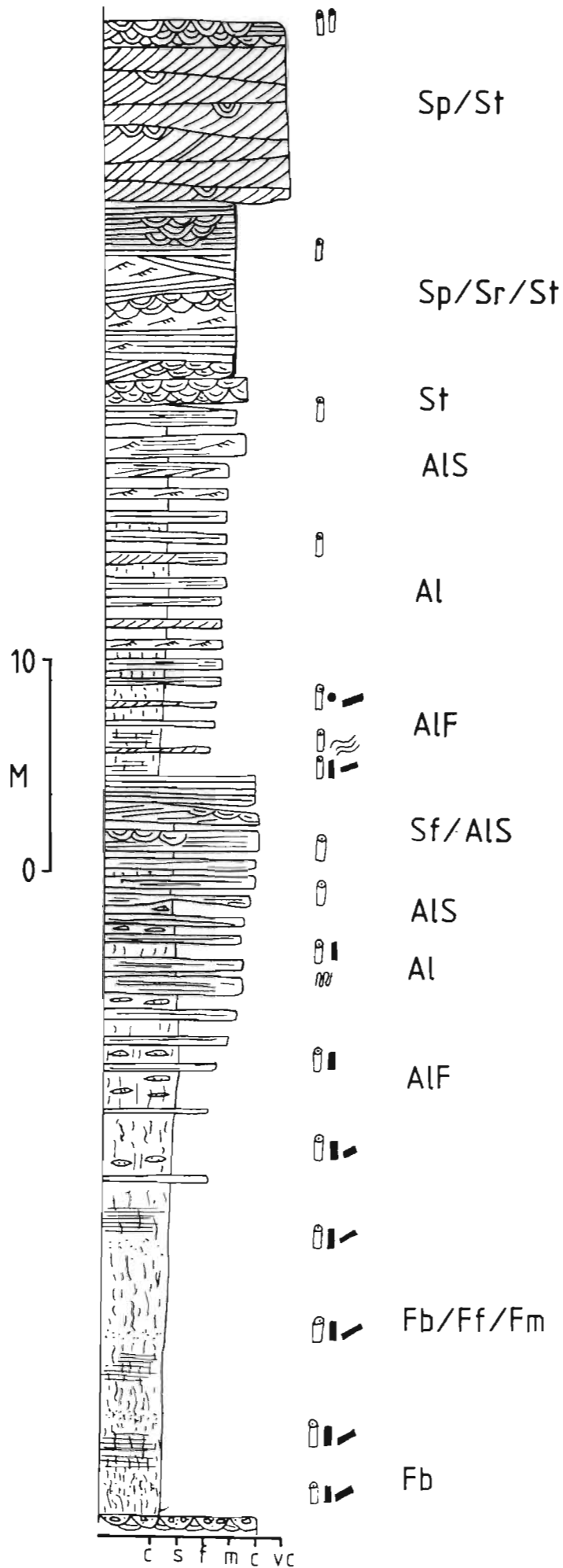
2.   Facies Al sequences show an overall upward increase in grain size. Burke (1972) suggested that the vertical distribution of sand deposited in an environment dominated by turbidity currents should be bimodal, with greatest sand concentrations being at the top and base of the progradational sequence.

3.   The thickness of upward-coarsening cycles suggests that delta progradation was into water seldom deeper than 40 to 60 m and that the distal mouth bar occurred in water no deeper than 20 to 40 m. Although turbidites have been described from delta-front settings (Heezen, 1955; Walker, 1965; Collinson, 1968b; Flores, 1972; Link; 1975; Massari, 1978) they are mostly associated with a continental slope environment (see Reineck and Singh, 1975; Heezen, 1974; Rupke, 1978; Galloway and Hobday, 1983), ie. much deeper water. Tavener-Smith (1985) reported turbidites from the basal prograding sequence of the Vryheid Formation in Zululand where water depths were probably about 100 to 150 m. Tavener-Smith and Mason (1985) reported turbidites deposited under similar conditions from the Tugela Ferry area in Zululand.

4.   Mass-gravity transport (including turbidity currents) is regarded as one of the major mechanisms for removing sediment from actively prograding delta fronts onto the basin floor (Galloway and Hobday, 1983). The initiation of this processes is reflected by the presence of various deformational features, including growth and graben faults, slumps and diapiric mud intrusions (Coleman, 1976; Edwards, 1976; Brown et al., 1973; Flores and Erpenbeck, 1981). While the limited extent of natural outcrops in the field area may mask many of the above features, facies Al and the overlying mouth bar sequences appear to be largely devoid of such deformational structures.

The possibility that bottom-hugging traction currents emanating from distributary channels were the main transporting agent is examined below. Beds

almost identical to facies Al sandstones with respect to internal sedimentary structures and vertical facies associations were suggested by Hobday (1973) to have been introduced by unusually high river discharge. He also recognised that some beds accumulated from turbidity flows or related resedimentation mechanisms triggered by slumping due to depositional oversteepening. Flores (1974) and Flores and Tur (1982) also attributed the deposition of sandstones on sloping delta fronts to traction currents issuing from distributary channels. Elliot (1978) regarded the coarser beds of the distal bar environment as representing flood-generated sediment incursions from the distributary mouth. He thought it probable that sediment-laden traction currents discharged from delta distributary channels during floods were periodically denser than the basin waters and entered the basin as bottom-hugging currents. Some of these may have developed into turbidity currents. Collinson (1970b) and McCabe (1978) suggested that high-discharge flows emanating from distributaries as heavily-laden suspension currents deposited sand on the slopes of the Kinderscoutian Delta.

However, by ascribing distal distributary mouth bar sand sedimentation to traction currents raises a number of problems with regard to discharge characteristics. Bates (1953), using the jet theory to explain phenomena observed off mouths of sediment-laden rivers, noted inter alia that river water is rarely denser than saline coastal waters and discharges are of the hyperpycnal rather than hypopycnal type. He maintained that even rivers carrying appreciable amounts of suspended material rarely have a fluid density approaching or exceeding that of the relatively undiluted saline bottom-waters (about 1,02) immediately offshore. Church and Gilbert (1975), on the other hand, contended that many of the assumptions Bates used in applying the jet theory to river-mouth processes are frequently not met with under actual conditions. In a synthesis on sediment transport and deposition at river mouths, Wright (1977) regarded inertia and associated turbulent diffusion, turbulent bed friction and buoyancy as the three main forces influencing river-mouth processes. The role played by each force depends on factors such as

river discharge volume and outflow velocity, relative water depths in and seaward of the river mouths, grain size of sediment load, and density contrast between the river and basin waters. Where bed-load transport of distributary channels is large, basin depths just seaward of the outlet are seldom greater (more commonly shallower) than outlet depth. Under these conditions turbulent friction dominates and resultant shoaling causes the effluent to decelerate and spread.

From the above it is apparent that the generation of hypopycnal or density undercurrent outflows in a marine environment requires exceptional conditions. However, these conditions are apparently commonly developed where rivers enter lakes (Lambert et al., 1976). In the Walensee of Switzerland sediment-laden currents emanating from rivers are between 1,5 and 2 m thick on average and attain speeds of at least 30 cm/s on delta slopes greater than 2° and 6 cm/s on slopes of less than 2° (Lambert el al., 1976). Underflows were able to develop where density contrasts of as little as 0,0025 g/cm³ were present. Underflows generated by the Colorado and Virgin Rivers in Lake Mead, USA, travelled distances in excess of 100 km over a period of 8 days. While these currents were barely capable of transporting sand, this example does serve to illustrate their longevity.

The following circumstances may have contributed to the generation of offshore-directed density undercurrents:

1. Rivers draining the source area, which was probably snow covered and blanketed by a considerable amount of unconsolidated glacial till (see Chapter 6), were cold and had a high mixed-sediment load.

2. Influxes of large volumes of fresh water from the numerous rivers along the coast, combined with moderate wave turbulance, may to some extent have diluted the saline water (thus decreasing its density) in the inshore area. In addition, the presence of giant (Gilbert-type) cross-beds (see sections 5.3.4.2 and 5.3.5) confirms that bottom currents were generated close to some river mouths under certain conditions.

3. It is possible that the Ecca sea was not fully marine (see Chapter 3).

It would, however, be unrealistic to totally discard the possibility that turbidity currents played a role in sand transport and deposition in view of the frequency with which such processes have been reported from ancient and modern delta fronts and the large amount of coarse detritus deposited in the mouth bar environment. It is also possible that the development of typical Bouma sequences was suppressed by the limited nature of the offshore slope.

Facies W was probably deposited in a similar manner to facies Al, but in environments fringing the main sources of sand supply, perhaps embayments. In view of the finer grain size of sandstone intercalations compared to those of facies Al it is possible that sediment may also have been supplied by longshore drift. Symmetrical ripples and the generally confused appearance of the facies strongly suggests wave reworking. This implies deposition in an environment shallower than that suggested for facies Al.

5.3.3   Proximal Distributary Mouth Bar Subassociation

Description

Deposits of this subassociation are not always developed; they are present in only 50 to 60% of the upward-coarsening deltaic sequences recorded. Where present, they have a gradational relationship with underlying mudrock facies but may overlie facies Al deposits with a sharp, planar base. They are invariably succeeded by erosively-based facies Sp or St and only rarely do they form the uppermost component of an upward-coarsening sequence in the study area.

The subassociation ranges in thickness from 4 to 10 m and in lateral extent up to an observed maximum of 5 km. Lateral thinning of sequences by as much as 5 to 6 m over a few hundred metres is common. The sandstone is moderately

sorted and fine- to medium-grained with subordinate, discontinuous mudrock intercalations and pebbly lenses. It has a characteristically disorganised appearance due to numerous low-angle discordances and minor channel scours (Fig 5.5). The subassociation is dominated by facies St and Sr (symmetrical and asymmetrical ripple crests), though substantial thicknesses of facies Sf, Sfl and Sm are also developed. Most sequences contain hummocky cross-stratification comprising erosively-based beds 50 to 100 mm thick showing flat lamination overlain by slightly silty ripple cross-lamination. A few sequences comprise only facies Sf in which the laminae are parallel and horizontal to subhorizontal. Most sequences show a concomitant upward increase in grain size and scale of sedimentary structures (facies St in particular).

Cross-bed azimuths are generally directed away from or oblique to the inferred coastline trend and have an angular range varying from 80° to 200°. Most distributions are unidirectional bimodal or polymodal. Only few are bidirectional. Bioturbation is rare, confined to discrete horizons within the sandstone and commonly associated with mudrock laminae. Trace fossils are generally of the Siphonichnus (S) assemblage.

Discussion

The distributary mouth bar is an environment where detritus is deposited following an abrupt reduction in velocity and transporting competence of distributary channels as they debouch into a body of standing water (Coleman, 1976). Modern proximal distributary bars are zones extending a few kilometres offshore and up to 10 m thick (Coleman, 1976; Davis, 1985). They experience fairly high rates of continuous sedimentation (Galloway and Hobday, 1983). This is reflected in the ancient deposits by the predominantly sandy nature of the sediments and their relatively well-sorted, mature texture compared with distributary channel sandstones. Periods of reduced flow (also a feature of distributary channels) allowed the deposition of silt and clay from suspension and limited bioturbation. The rarity of bioturbation may also be attributable

to the inhibiting effect the almost constant influx of fresh water would have had on organisms. Densely bioturbated examples of this subfacies suggest deposition marginal to the main locus of sedimentation. The fact that most of the organisms were suspension feeders (as indicated by vertical burrows) attests to the turbid nature of the environment.

Sedimentary structures are consistent with predominantly lower flow regime (offshore and longshore) currents while hummocky cross-stratification comprising facies Sf/Sr motifs and symmetric ripples suggests that wave reworking played an important role in the proximal distributary mouth bar environment (Reineck and Singh, 1975; Kumar and Sanders, 1976; Kreisa, 1981; Dott and Bourgeois, 1982). The generally confused appearance of these sequences is probably in response to a complex interplay between fluvial and marine processes.

Sequences dominated by facies Sf underlying erosively based distributary channel-fills have been described by Christie (1983) and Tavener-Smith (1985) as being indicative of reworking of sand shoals by shallow wave-surge in a nearshore/distributary mouth bar environment. Similar sedimentary structures are also formed by wave surge along the delta front of the modern Burdekin (Coleman, 1976) and Rhone (Oomkens, 1970) deltas. Although laterally extensive mouth bar sands are regarded as indicative of sediment redistribution by wave currents (Galloway and Hobday, 1983; Davis, 1985), a number of closely space distributary channels and associated coalesced mouth bars would probably result in a similar geometry (Flores and Tur, 1982).

The marked rarity of sediment coarser than medium-grained sandstone is significant, especially in view of the fact that sandstone of this subassociation is finer than that of overlying distributary channel sandstones and the upper few metres of underlying facies Al sandstone interbeds. This, and the absence of deformational structures, seem to support the hypothesis that sand in the distributary mouth bar environment was deposited predominantly by

river-generated density currents rather than turbidites (refer to previous section). The proximal distributary mouth bar setting reflects average or day-to-day sedimentation; coarse sand and pebbles discharged during floods by-passed it and were deposited more distally.

5.3.4   The Delta Plain

The delta plain environment comprises active and abandoned distributary channels and/or delta lobes separated by stretches of shallow water and emergent or near-emergent surfaces (Elliot, 1978). These interdistributary interlobe areas, referred to as bays or embayments in this thesis, were low-energy environments frequently interrupted by flood-generated influxes from adjacent active distributary channels and delta lobes.

5.3.4.1   Distributary Channel Subassociation

Description

Distributary-channel deposits are identified chiefly by the position they occupy within an upward-coarsening sequence (ie. the uppermost component) and their textural characteristics. They comprise poorly-sorted medium- to coarse-grained, pebbly sandstones and are dominated by facies St and Sp. The basal surface is invariably erosive and overlain by up to 200 mm of facies Cm. Individual sequences vary in thickness up to 25 m. Some can be traced for distances of up to 20 km but most appear to extend no more than 5 km along palaeostrike. They show a complex internal geometry owing to their multilateral and multistoried nature and tend to vary in thickness over short distances.

Sequences dominated by facies St with minor facies Sp, Sf and Sr are most common; they comprise cosets 1 to 3 m thick bounded by low-angle erosional surfaces showing relief up to 0,6 m (Fig 5.6). Sets vary in thickness from 0,3 to 0,8 m and each sandstone coarsens upward from medium- to coarse- and then to very coarse-grained, pebbly texture and there is a concomitant increase in set

thickness. Bioturbation (<u>Siphonichnus</u> (S) assemblage) is abundant in the upper 100 to 500 mm of the sequence and is generally associated with a pebble layer on its upper surface.

A second type of sequence is dominated by facies Sp (though St is also moderately common) and forms upward-fining units 1,5 to 2,5 m thick (Figs 5.6 and 5.7). Each is erosively based (relief up to 200 mm), grades up from coarse-grained, pebbly sandstone into sandstone of medium-grain, and is bioturbated (<u>Siphonichnus</u> (S) assemblage) in the top 50 to 100 mm. Most sequences are capped by up to 200 mm of facies Fm or Ff. Some show no upward-fining tendency and are erosively-based with cosets 0,4 to 1,2 m thick. Each comprises 1 or 2 sets of planar cross-beds, commonly incised by solitary trough cross-beds and intercalated by up to 100 mm of dark grey mudrock (facies Fm or Ff). Abundant <u>Diplocraterion</u> and/or <u>Siphonichnus</u> are present in the upper parts of these sandstone beds.
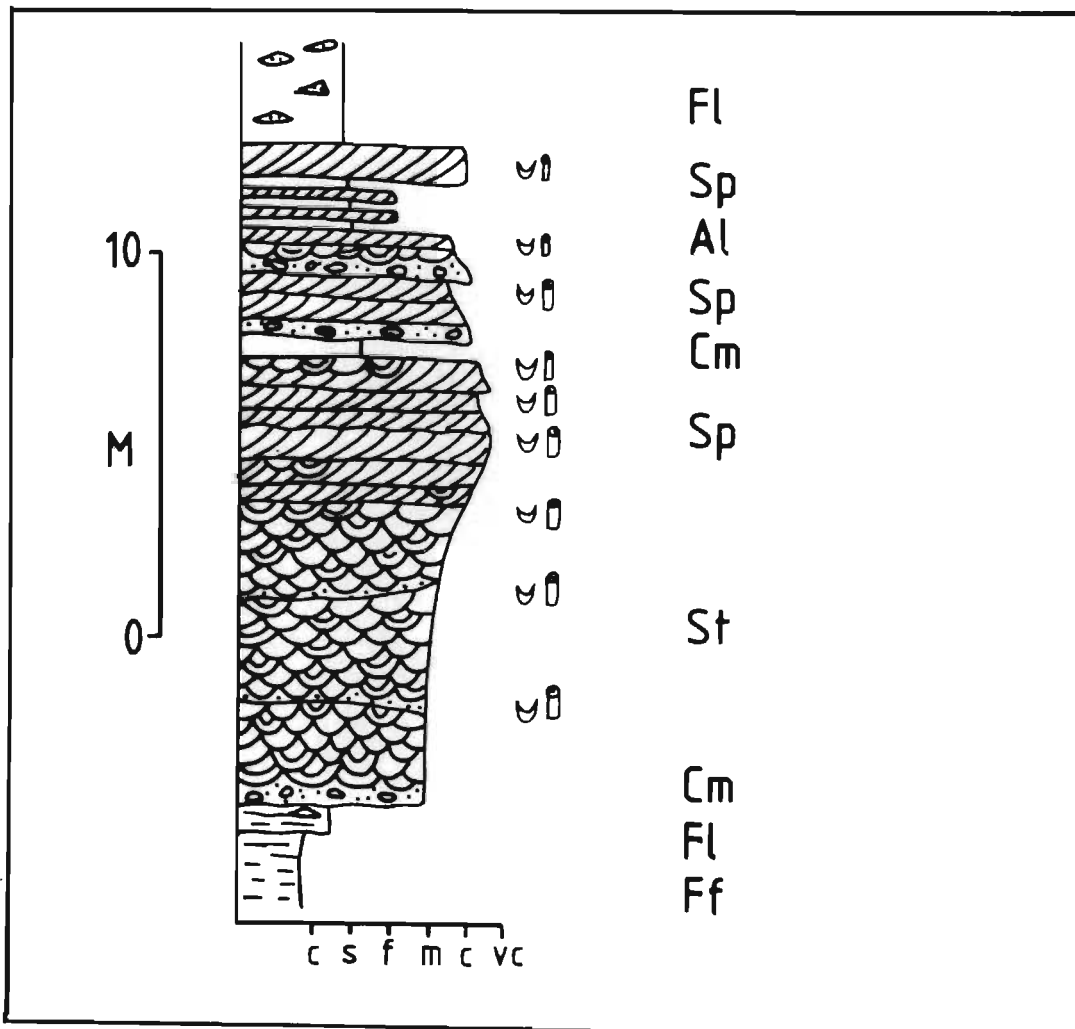
Figure 5.6. Facies St-dominated distributary-channel sequence overlain by facies Sp-dominated sequence. Basal part of Stein Coal Spruit section.
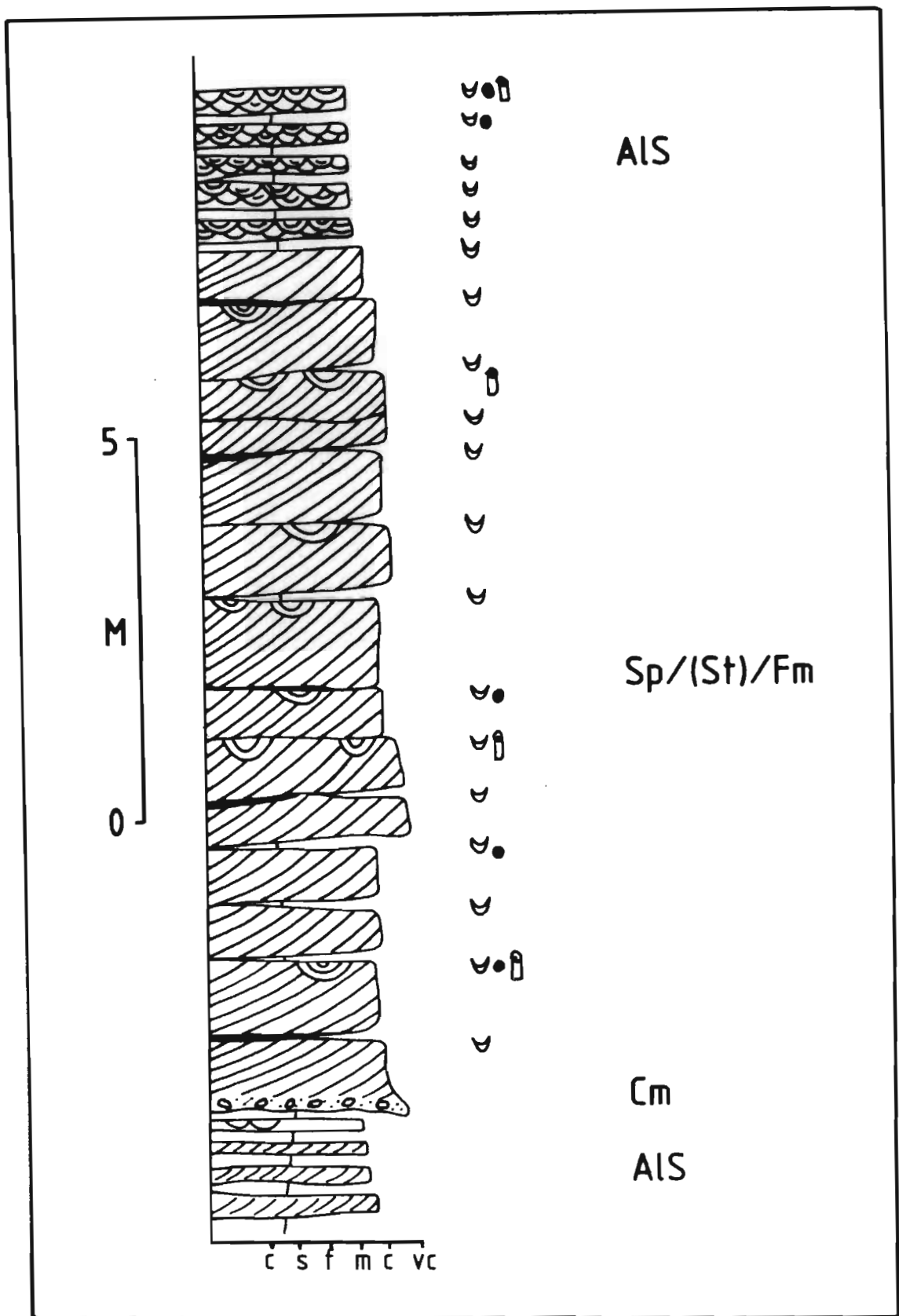
Figure 5.7. Section through a facies Sp-dominated distributary-channel sequence. Lower portion of Winstone section.

Both types of distributary channel fills overlie either prodelta or distributary mouth bar sediments, but the two are commonly developed within the same upward-coarsening sequence. In this case the first type (facies St-dominated) is always overlain by the second type (facies Sp-dominated). Some sequences of this subassociation contain intercalations of facies Fm, Ff, Fl or W up to 2 m thick and a few hundred metres across.

Palaeocurrent distributions of individual sequences are invariably directed offshore and unimodal with an angular spread of up to 110°. Current reversals are not common, but where present, are associated with facies St-dominated sequences.

Discussion

Distributary channels serve as the primary paths for discharge of sediment-laden water into a basin and are sites of deposition for the coarsest sediment that is carried to the delta (Galloway and Hobday, 1983). Distributary channel-fill sandstones have essentially the same lithology and sedimentary structures as the fluvial channel-fills. The predominance of facies St-dominated sequences, however, suggests that channels containing migrating dunes became deeper (2 to 4 m) as they approached the coast. This was possibly a function of the decreased gradient and preferential downward incision of channels in response to the more confining nature of the delta plain environment. The absence of upward-fining sequences within these sandstones may be attributable to in-channel fluvial processes or the removal of the finer, upper sediments by marine processes (Elliot, 1978). Tidal or, more likely, up-channel currents induced by waves were probably responsible for observed current reversals.

Facies Sp-dominated sequences, bearing a strong resemblance to their fluvial counterparts, reflect the development of broad, shallow (1 to 2 m depth) channels dominated by transverse bars and sand flats (Miall, 1977a). Bioturbation on upper surfaces of sandstones indicates that during low discharge the saline basinal waters extended far up distributary channels.

This, together with the cyclic nature of the sequence testifies to the episodic nature of fluvial activity. It is suggested that these deposits represent upper delta plain distributary channels while facies St-dominated channel-fills represents lower delta plain distributaries. This is supported by the fact that when the 2 sequences are associated vertically the former always overlies the latter.

The multichanneled nature of these sandstone bodies suggests that the delta plain comprised braided channels undergoing repeated switching or avulsion. Confirmatory, though circumstantial evidence is the absence of associated levee deposits or sediment likely to have provided cohesive banks. Like the more proximal fluvial channels, distributary channels avulsed or perhaps even bifurcated to take advantage of the smallest variations in gradient. The rarity of significant mudrock deposits may be due either to their poor preservation potential on a delta plain or the predominantly subaerial nature of the interdistributary areas; only minor ponds or embayments being present.

Distributary channel and eventual delta-lobe abandonment is indicated by pebble lags and bioturbation at or near the top of the subfacies. The former are attributed to current winnowing and/or wave reworking while the bioturbation reflects renewed influence of the saline basinal waters.

Evidence of only limited wave action, micro-tidal conditions (see section 5.4; Vos and Hobday, 1977; Tavener-Smith, 1982) and rapid delta progradation suggests that the rate of sediment influx was substantially greater than the rate at which the sediment was redistributed by wave and tidal processes. These deltas are therefore described as fluvially-dominated (Galloway, 1975) or high-constructive (Fisher, 1969; Fisher et al., 1969). The distinction between lobate and elongate high-constructive deltas is based mainly on sand-body (delta plain) geometry (Fisk, 1955) and to a lesser extent on the thickness of prodelta deposits. As most delta plain sandstones have a sheet-like geometry and overlie relatively thin prodelta deposits, such deltas are classified as

being of the high constructive, lobate type. The implied close spacing of distributary channels supports the suggestion that delta lobes had a lobate, smooth outline.

The distributary channel sandstone bodies described here are similar in many respects to the Ordovician "fan delta" complex of the Haouaz Formation, Libya (Vos, 1981) and the short-headed stream delta of the Pennsylvanian Haymond Formation, Texas (Flores, 1975). Referring to use of the term "fan delta" by Vos, the original definition of a fan delta is an alluvial fan that progrades into a standing body of water from an adjacent highland (Holmes, 1965; McGowen, 1970). McPherson et al. (1987) reached a similar conclusion regarding usage of the term "fan delta", describing coarse-grained deltas formed by the progradation of a braided alluvial-plain system into a body of standing water as braid deltas. The delta plain of braid deltas, like the deposits described here, is composed entirely of braided fluvial distributary channels. Accordingly, use of the term to describe any fan-shaped, coarse-grained delta is incorrect and the terms "lobate delta" and "braid delta" are preferred in this thesis.

Flores (1975) regarded the distance between the fluvial headwaters and delta as an important control on delta plain (channel) grain size. Like the Haymond Formation delta deposits, the distributary channels considered here are much coarser than, for example, those of the Mississippi, Fraser or Rhone deltas and were probably also short-headed with a source no more than 200 or 300 km upstream from the delta (this aspect is considered in more detail in Chapter 6).

5.3.4.2   Giant Cross-Bed Subassociation

Description

Giant cross-beds in the study area are identifiable only in outcrop and not in borehole core or logs. They are thus described from the southern and eastern parts of the area only. Set thickness varies from 3 to 20 m (the lower limit was arbitrarily set) and they occupy the same stratigraphic position in upward-coarsening sequences as the distributary channel subassociation. Giant cross-beds are relatively rare in the field area (they occur in less than 10% of upward-coarsening sequences) but visits to the Vryheid, Utrecht and Nqutu areas revealed that they are a common component of the basal sandstone of the Vryheid Formation. In this study they are considered an important element of the deltaic environment as their scale and unique association of facies provide crucial clues to the relationship between fluvial and basinal processes. Two occurrences of giant cross-beds are described to illustrate vertical and lateral facies associations.

The first is situated on the eastern slopes of Hlungwana, a prominent hill approximately 22 km southeast of Dundee alongside the Dundee-Nqutu road. The set is 19 m thick and the upward increase in grain size from medium- to coarse-grained sandstone at the base to coarse-grained, pebbly sandstone at the top is reflected in the downdip decrease in grain size of individual foresets and bottomsets (Fig 5.8). Bottomsets overlie facies Al (which also shows a net upward increase in grain size) with a sharp, planar junction. Nowhere is there any evidence of significant erosion. Individual foresets are planar to gently arcuate (concave upward). Dips of up to 26° near the top of the set decrease progressively to less than 5° near the base. Where foresets intersect the horizontal or near-horizontal erosive surface defining the top of a set they describe straight or gently arcuate patterns and can be traced for tens of metres along strike (where outcrop permits it). Azimuth orientations are unimodal and display an angular range of 120° (from 110° to 230°, mean = 165°). Numerous concave upward internal erosional surfaces or reactivation surfaces separate bundles of foresets having differing dips.

Foresets are 50 to 300 mm thick and predominantly normally graded. Most are internally structureless or show poorly to well-developed flat lamination (parallel to foreset bedding surfaces). Facies St intrasets are common and generally orientated in the same direction or slightly oblique to the dip of the foresets containing them. There is a progressive increase in intraset frequency down the dip of individual foresets. The planar upper surface of the giant cross-bed set is incised by similarly orientated trough cross-beds up to 0,5 m thick. No flat-lying topset beds are present and the upper surface is overlain by a sharply-based facies Ff/Fm mudrock. Siphonichnus, Diplocraterion and Helminthopsis burrows are present in the upper parts of individual giant foresets and are particularly dense on the upper surface of the set. About 2 km to the west, at Cinderford (Fig 5.5), a sandstone occupying the same stratigraphic horizon as this giant cross-bed set comprises only facies St.

A second giant cross-bed set is situated on the farm Glen Lyon, approximately 11 km east of Wasbank. It crops out in a tributary of the Wasbank River a few hundred metres east of the homestead. The river valley follows the dip of the foresets, so they can be traced for about 500 m downstream. The set is about 18 m thick and asymptotically based (Fig 5.9). Foresets are planar to strongly concave-upwards with dips of up to 20° toward the southwest (mean = 205°). This angle decreases rapidly downdip. The bottomsets dip at less than 4° and can be traced down dip for approximately 500 m. Locally the dip may increase up to 8° and the bottomsets drop down in a series of extended steps.

The main differences between this set and the one exposed at Hlungwana are:

1. The foresets of this set pass distally from fine- to medium-grained sandstone (with sporadic mudrock intercalations up to 10 mm thick) into facies AlS, Al, AlF and eventually facies Ff and Fm (Fig 5.10). Sandstone bottomsets pinch-out in a downdip direction and may also grade into facies W.

Figure 5.9. Upper and proximal portion of giant cross-beds. Note planar nature of top of sets and internal discordances (arrowed). Hammer (circled) for scale. River valley, Glen Lyon farm.
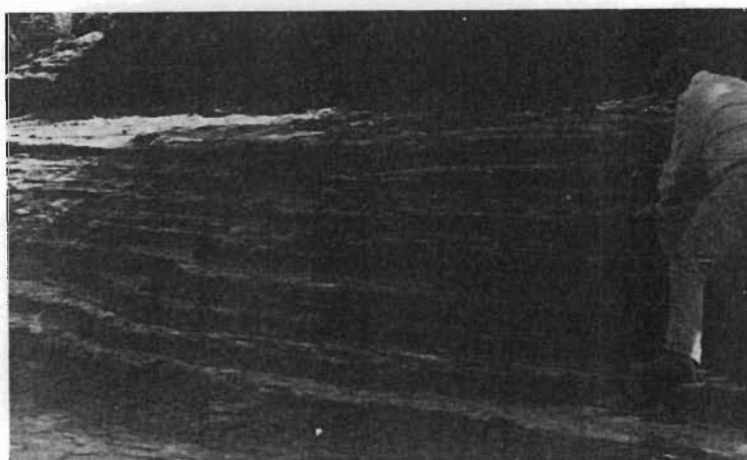
Figure 5.10. Bottomsets of giant cross-beds approximately 400 m down-dip of locality shown in Figure 5.9. Note predominance of mudrock. River valley, Glen Lyon farm.

2.   The bottomsets, in addition to being flat laminated, are rippled and contain small-scale trough cross-beds (Facies St). Large numbers of fossilised tree trunks, some coalified, are scattered on these bedding surfaces.

3.   The upper surface of the giant cross-bed is overlain by 1,6 m of erosively-based medium- to coarse-grained facies Sp and St sandstone. Sets are 300 to 400 mm thick with a similar orientation and angular distribution as the foresets. The sandstone is overlain by a 0,9 m thick sequence of facies Ff and Sfl containing abundant Skolithos, Siphonichnus and Diplocraterion burrows.

Discussion

Giant cross-beds are a relatively rare phenomena in the geological record and have accordingly received relatively little attention in the literature. Gilbert (1884) was the first to describe giant cross-beds formed by deltas prograding into a standing body of fresh water. Collinson (1968a) interpreted giant cross-bed sets between 4 and 40 m thick from the Kinderscout Grit (Upper Carboniferous) in northern England as Gilbert-type deltas and also discounted the possibility of them being large aeolian dunes or marine sand waves. This interpretation was rejected by McCabe (1978) who believed they represent large alternate bars within deep fluvial channels. Planar cross-beds up to 5 m thick at the top of the Dwyka Formation within the Witbank Coalfield were considered by Le Blanc Smith and Erikson (1979) to represent outwash deltas in a fluvio-glacial lacustrine setting.

The idea that giant cross-beds described in this thesis formed as fluvial sand waves or large side bars is rejected owing to the absence of any associated channel sides, and in the light of vertical and lateral facies associations.

The upward coarsening nature of the sets and their sedimentary context makes it more likely that they were deposited by a prograding delta. This hypothesis is supported by the lateral transition from proximal sandstones to distal mudrocks at Glen Lyon. That suspension setting was the dominant sedimentary process in the distal parts of the giant cross-bed set provides convincing evidence for deposition in a body of standing water.

The hydrodynamic control of sediment distribution and deposition on deltas and at distributary channel mouths has been speculated on to some extent in the literature (Bates, 1953; Church and Gilbert, 1975; Wright, 1977) and in section 5.3.2 of this thesis. Homopycnal flow, where river and basin waters are of equal density, leads to thorough three-dimensional mixing of the respective water masses and results in rapid dumping of bedload material at the river mouth and settling of suspended load in a more distal setting (Oomkens, 1970; Church and Gilbert, 1975). This regime (homopycnal flow) is believed to promote the development of Gilbert-type deltas with steeply dipping foresets and is equivalent to the "inertia-dominated effluent" of Wright (1977) where the effects of buoyancy and bed friction are negligible. Wright and Coleman (1974) regarded Gilbert-type deltas to be the result of fully turbulent outflow with negligible bottom interference, ie. high outflow velocity, small density contrasts and deep water immediately seaward of the river mouth.

Church and Gilbert (1975) noted that "...appreciable quantities of bed material reach the bottomsets by slipping down the foresets slope..." on small lacustrine deltas. Gilbert (1975) also regarded slumping as an important process in lacustrine (homopycnal-dominated) delta formation. No evidence of slumping was observed in any of the sets described here.

Evidence not favouring homopycnal outflow is provided by the presence of intrasets and asymptotically-based bottomsets. The intrasets are, on the basis of their mean orientation with respect to that of the giant cross-bed foresets, believed to have been formed by inflowing fluvial currents. These currents were

sufficiently strong to move bedload material down the foreset slope thus reducing the angle of foreset dip in the lower parts of the foreset (Collinson, 1968a, 1970b; Axelsson 1967). The increase in the number of intrasets near the base of the foresets or within the bottomsets, and basinward change in the junction of the foresets with the underlying deposits (angular to asymptotic) may be indicative of flow separation with the current only impinging on the foreset slope some distance beyond the crestline of the giant cross-beds. The foregoing features are suggestive also of an element of hypopycnal flow whereby denser inflowing water sinks below the basinal waters and carries sediment away from the point of discharge preventing the development of steep foresets (Church and Gilbert, 1975).

The arguments cited above suggest that depositional mechanisms responsible for the formation of giant cross-beds are more complex than is commonly appreciated. Evidence supporting the existence of hypopycnal flow-conditions at distributary channel mouths has been cited in the section dealing with the distal distributary mouth bar environment. Homopycnal conditions were just as likely to have existed and there is reason to suppose that temporal and spatial variations in basinal water density did exist (as implied by the lateral transition from giant cross-beds at Hlungwana to facies St/Sp - dominated distributary mouth bar and channel sandstones at Cinderford).

In the opinion of the author (based on the above observations and considerations) distributary-channel avulsion is the main mechanism controlling the formation of giant cross-beds. Avulsion allows the distributary channel to discharge into relatively deep water (5 to 30 m, as indicated by the thickness of individual sets). Intrasets indicate that flow separation probably did occur. Internal discontinuity surfaces and bioturbated foreset surfaces reveal evidence of stage fluctuation and episodic bedform progradation. Trace fossils along foreset bedding surfaces are indicative of periodic cessations in current activity. This is confirmed by the mudrock intercalations which
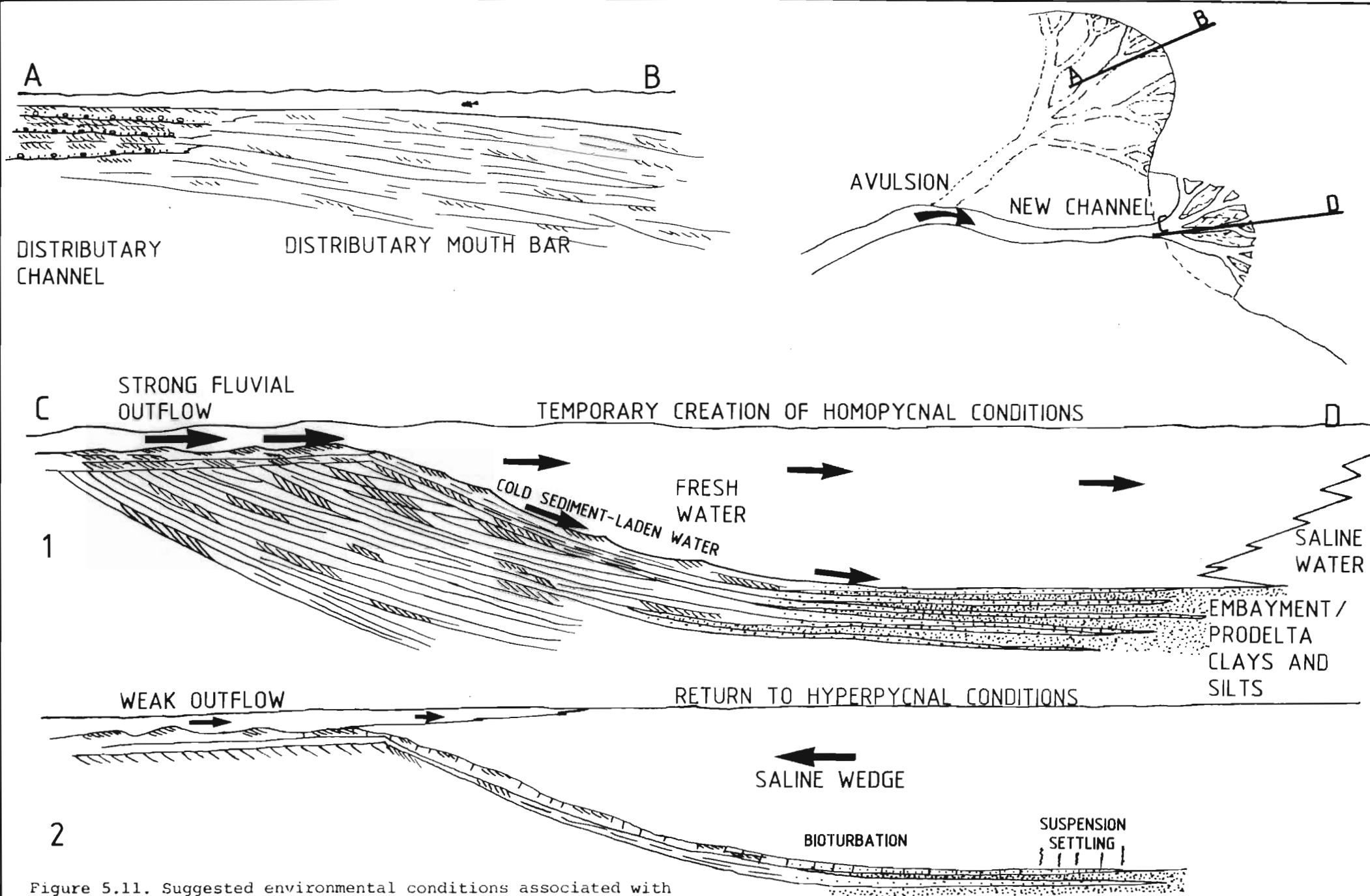
Figure 5.11. Suggested environmental conditions associated with the formation of giant cross-beds.

represent periods of depleted bedload supply when deposition was dominated by suspension setting. Subsequent renewal of bedload supply and the influx of fresh water curtailed further faunal activity. Concave-upward erosion surfaces, indicating the removal of substantial amounts of sediment, are believed to have formed either under steady state conditions by changes in the geometry of the leeside flow (from 3- to 2-dimensional flow separation) and by fluctuations in flow depths and velocities (McCabe and Jones, 1977). Changes in leeside flow geometry following a pause in current activity are also indicated by surfaces of discordance separating asymptotically-based foresets from overlying tangentially-based foresets.

The composition, texture and unipolar and unimodal palaeocurrent distribution of the cross-bedded sandstone overlying the giant cross-beds at Glen Lyon strongly suggests a fluvial origin. This is confirmed by their striking similarity to distributary channel sandstones. Coset thickness indicates shallow-water origin and the erosive base to the coset suggests deposition either within channels or perhaps just seaward of a river mouth. Although these sets appear to truncate foresets, it is probable that this sequence is equivalent to the topsets of Gilbert's (1884) nomenclature (Collinson, 1968a). The absence of beds of this type at Hlungwana is attributed to delta-lobe abandonment prior to the fluvial-dominated topsets prograding over that part of the sequence.

Figure 5.11 illustrates and summarises the envisaged sequence of events leading to the deposition of the giant cross-bed subassociation.

5.3.4.3   Embayment Subassociation

Description

Embayment deposits range from 1,5 to 22 m thick and may be superimposed on

distributary channel mouth bar or coastal foreshore deposits. They are mostly developed near the top of the Lower zone, within the Coal zone, and in the lower part of the Upper zone. Coal seams are an important component of the embayment subfacies, and are a useful aid in identifying emergent surfaces.

Up to 7 distinct sedimentary units were recognised within the embayment subassociation during the course of this investigation. Although each is associated with a characteristic facies assemblage, it is the geometry and lithology of the sandstone component that is the most distinctive feature. Detailed lithological, compositional and morphological descriptions of each unit are given in Figure 5.12 and various aspects are commented on below.

The subassociation is dominated by sediments from the fine-grained and heterolithic facies. The entire spectrum of fine-grained facies is represented, with facies Fm, Ff and Fl being most common. Facies W is often present, commonly in association with facies Fl. Facies Al may occur, but not to the same extent as facies W.

Type 1 and type 2 sequences are characterised by an upward increase in grain size. The top sandstone component may have a sharp and planar or erosive base but more commonly displays a gradational relationship with the underlying facies. Type 1 sequences are typically lenticular and difficult to trace more than two or three hundred metres .

Type 3 sequences comprise erosively-based, poorly-sorted, coarse-grained, pebbly sandstones which may or may not show a slight upward increase in grain size. Unidirectional trough-cross bedding (facies St) is the dominant structure though rare cross-bed reversals were observed. Many of the thinner units (ie. less than 0,5 m) are structureless or display weakly-defined, slightly undulating flat stratification. The geometry of these sandstones could not be ascertained in the field or by borehole logs, but the impossibility of tracing them over more than very short distances suggests that they are lenticular.

1		C St Sr Fr/Ff St	1-4 m thick Lenticular.	Prograding crevasse splay/sub-delta.
2		C Sr Al/W Cm	1-4 m thick. Lenticular.	Intermittently prograding crevasse splay/sub-delta.
3		Cm St Ff/Fm	,2-2 m thick Lenticular.	Crevasse splay (medial).
		C/St Sr/Fr Ef Cm Fm Sp/St	Variable thickness and lateral extent	Bay fill, prograding bay margin. Deposition from suspension settling.
4		Fr Sm/(Sp) Fm/Ff	,2-2 m thick Lenticular.	Crevasse splay (proximal). Abrupt decrease in current energy.
5		Fm C Sr/Sf Fm	,2-2 m thick Sheet like geometry	Overbank sheet-flood sandstone. Later colonisation by swamp/marsh.

Figure 5.12. Various types of facies associations characterising delta-plain deposition (continued over page).
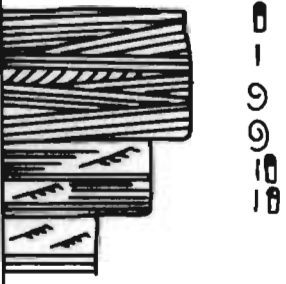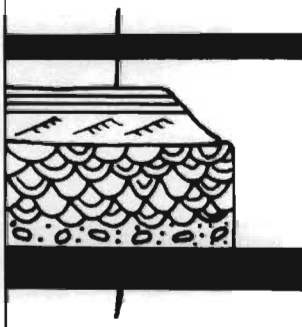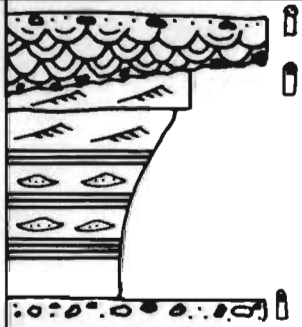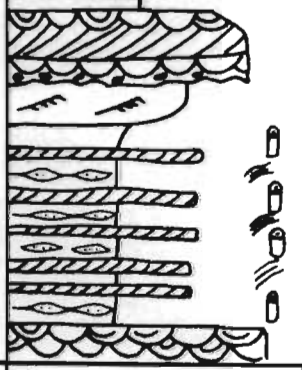
6		Sf Sr/Sf Ff/Fr	1-2m thick. ? extent.	Wave reworking of sands into (?) spits or bars.	
7		C Fm Sr/Sf St Cm C	,5-3 m thick. Lenticular (10-500 m)	Minor distributary/ crevasse channel.	

TYPE ASSOCIATIONS

3 1		St Sr Ff/Fl Fm Cm	Crevasse-splay lobe incised into a more distal deposit.	
7 2		St/Sp Fr Al/W St	Crevasse channel incised into prograding crevasse-splay.	
6 5		Sf Sr/Sf Fm	Wave reworking of earlier deposited sheet-flood sandstone.	

Figure 5.12. Continued from previous page.

Type 4 sequences are characteristically erosively-based, very coarse-grained, pebbly sandstones with a sheet-like geometry. They are predominantly massive to weakly cross-stratified. Lenses of mudrock up to a few centimetres thick are common.

Type 5 sequences comprise a distinctive, fine- to medium-grained sandstone dominated by facies Sf and Sr (asymmetric, including climbing ripples). The sandstone can be traced normal and parallel to paleoslope for distances up to 3 km and rarely shows an upward decrease in grain size. This is the most common sequence developed in this subassociation.

Type 6 sequences are distinctive in that the upper sandstone component is fine- to medium-grained, and moderately to well sorted. Facies Sr (asymmetric and symmetric ripples) and facies St are present, but facies Sf, showing low-angle discordances, is strongly predominant.

Type 7 sequences commence with a medium- to very coarse-grained, pebbly sandstone (with erosive base and commonly a pebble lag) passing upward into fine- to medium-grained sandstone. Facies St and rare facies Sp show an upward decrease in set size and are overlain by facies Sr and Sf. The sandstone is commonly overlain by sharply based facies Fm or Ff.

Composite sandstone components comprising sandstone from two types of sequences are common. Thus, type 6 sandstones commonly overlie type 5 sandstone, and a type 1 sequence may overlie a type 3 sandstone.

Bioturbation is common in the embayment subassociation; Siphonichnus (S) and Siphonichnus (F) being developed within sandstones and mudrocks respectively. Burrows are found throughout the fine and heterolithic facies but are usually confined to the upper 50 to 100 mm of sandstones. In some instances, however, bioturbation may be present throughout a sequence (particularly types 1 and 2) or associated only with erosive surfaces (types 3 and 4). Thin, persistent

bands of diagenetic siderite are commonly associated with densely bioturbated horizons on sandstone bedding surfaces.

In general, the sandstone components of these sequences show cross-beds parallel, oblique or at right angles to the trend of associated distributary channel fills.

Discussion

Flood-generated processes dominate interdistributary bay sedimentation (Elliot, 1974, 1978; Coleman, 1976; Horn et al., 1978) and the varying assemblages reflect different processes by which embayments were filled. The thickness of upward-coarsening sequences capped by coal indicates that embayment water depth rarely exceeded 5 m, though the embayments commonly underwent repeated subsidence and deepening. The ichnofaunal assemblage is diagnostic of predominantly brackish waters (see Chapter 3).

Evidence that suspension settling and gentle currents were predominant is provided by the high proportion of argillaceous sediments and by the trace fossil distribution (predominantly near the top of sandstone units - see Chapter 3). Elliot (1974) contended that most sediment is introduced into embayments by water spilling over fluvial (distributary) channel banks as sheet flows or through crevasses. It is also probable that some argillaceous sediment was brought into embayments by longshore drift.

Facies F records the deposition of suspended sediment in quieter or deeper parts of bays beyond the direct influence of flood processes. Facies F1 is believed to have originated from the reworking and concentration of sandy material by wind-generated waves (Reineck and Singh, 1975, Coleman, 1976). Facies A1 and W were deposited from sand-laden currents (possibly related to seasonal flooding) followed by a return to "normal" sedimentation from suspension. Although these sandstones may represent distal crevasses or

overbank sheet-flood events, Coleman (1976) described comparable sequences deposited at river mouths advancing into embayments of the Mississippi delta. Similar sequences have been attributed to levee sedimentation (Elliot, 1974, 1975) but facies associations, absence of evidence emergence (desiccation cracks and root bioturbation) and only minor development of climbing ripple cross-lamination (Reineck and Singh, 1975) argue against this. Facies W sandstones were also subjected to wave reworking though the mudrock intercalations suggest that this was only of an intermittent nature. The progressive upward increase in grain size displayed by many facies F, Al and W sequences reflects a progressive increase in sediment supply and current activity as bay margins prograded. This is a well-documented characteristic of modern delta distributary systems (Oomkens, 1970; Elliot, 1974; Coleman, 1976).

Type 1 sequences imply a depositional event with a steadily prograding sediment source. Similar sequences are produced by the progradation of small mouth bars associated with minor distributary or crevasse channels emptying into interdistributary embayments of the Mississippi (Coleman et al., 1964) and Rhone (Oomkens, 1970) deltas. Coleman and Gagliano (1965) demonstrated that closely spaced mouth bars may coalesce and produce a laterally continuous advancing front which fills a bay. Type 2 sequences may represent a prolonged, but intermittent, depositional event.

Type 3 sequences appear to record the sudden influx of powerful sand-laden currents into an embayment and their lithological similarity to distributary-channel deposits suggests deposition as crevasse-splay lobes (Elliot, 1974; Coleman, 1976; Horne, 1979). Type 4 sequences indicate rapid deposition, probably following an abrupt decrease in gradient and current velocity. The presence of thin siltstone lenses, internal discordances and associated burrows points to periodic reactivation of the lobe. These features are typical of a proximal crevasse splay where bedload is deposited rapidly in close proximity to the channel breach (Elliot, 1974). Although there is no concrete proof that types 1 (and 2), 3 and 4 sequences are distal to proximal equivalents

respectively, investigations by Reineck and Singh (1975), Elliot (1974, 1975), and Coleman (1976) suggested that this may be the case.

Type 5 sandstones were deposited by overbank sheet floods. The absence of coarse-grained, pebbly material typical of fluvial channels and the great lateral extent of these sandstones suggests that deposition was due to spilling over a long stretch of the channel banks rather than breaching at a single point (Elliot, 1974).

Lithological and textural characteristics of type 6 sequences demonstrate that to some extent wave processes played a role in the embayment environment. Wave reworking of crevasse-splay sediment in inter-distributary bays of the Mississippi delta (Fisk et al., 1954) produced sand spits of a similar scale and internal morphology to those described here. Elliot (1974) described similar sequences from the mouths of inter-distributary embayments where they are believed to result from the redistribution of sediment from adjacent distributary mouth bars by wave-induced longshore currents.

Deposition by high-bedload rivers which incised into bay-fill and crevasse-splay sandstones is indicated by type 7 sequences (Elliot, 1974). The composition, texture and erosive base of these channel deposits suggests avulsion, probably related to crevassing from a major distributary channel. The upward decrease in grain size resulted from gradual channel filling and subsequent abandonment. Channel elevation is brought about by the tendency of river channels to construct alluvial ridges by channel aggradation and levee development (Fisk, 1955). The apparent absence of levee deposits, however, is confirmed by the relative abundance of overbank sheet sandstones; distributary flood-waters were more likely to spill over the river bank than breach it.
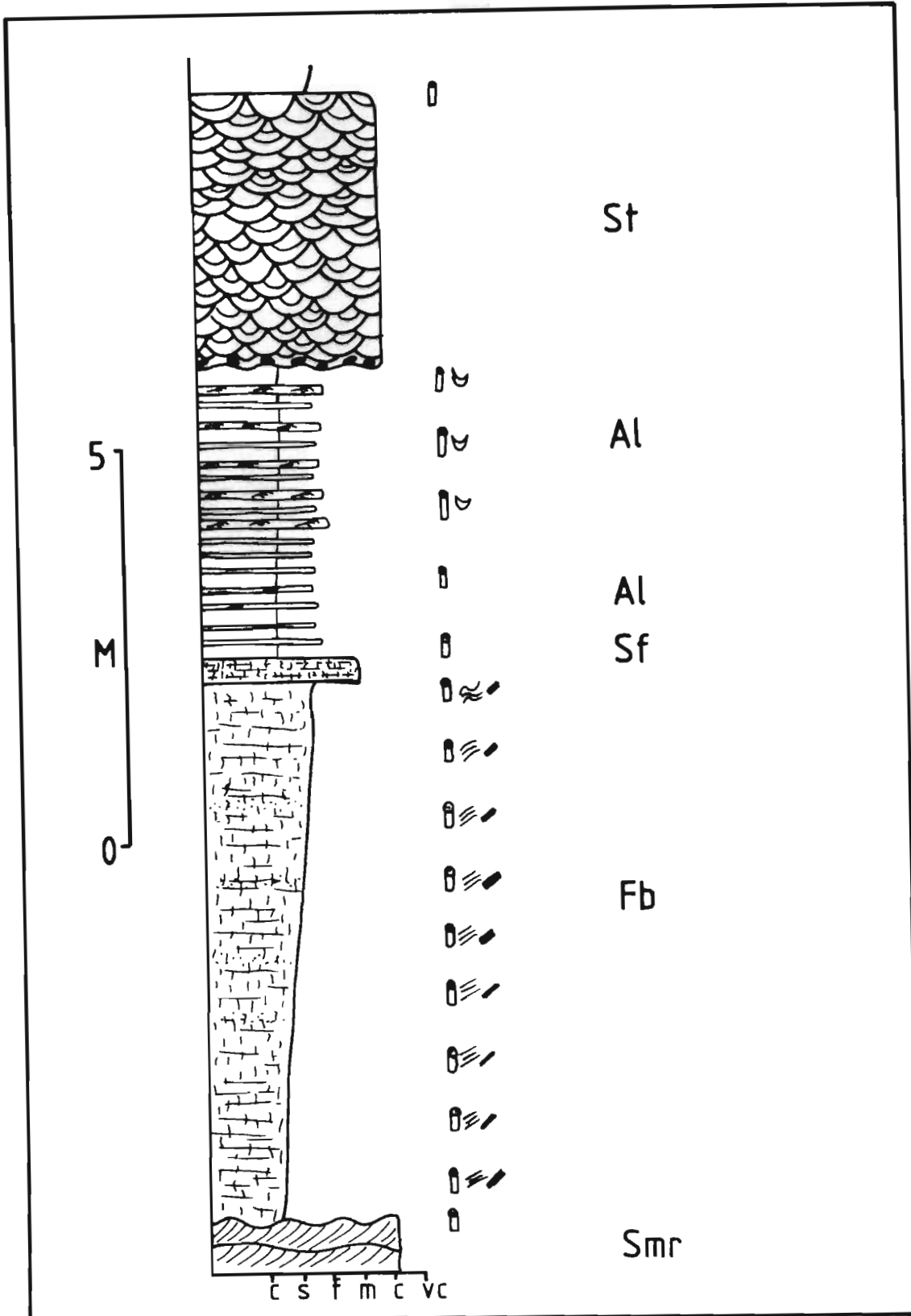
Figure 5.13. Section through extensively bioturbated prodelta mudrocks and distal distributary mouth bar deposits overlain by an erosively-based distributary-channel sandstone. Winstone section.

Coals record the development of marshes and swamps where the water was sufficiently shallow to permit colonisation by plants. Where the water was not sufficiently shallow to permit this the subaqueous surface (usually sand deposited by a flood event) was colonised by organisms and a burrowed, sideritic horizon resulted (Horne et al., 1978). Thus, coals and burrowed, sideritic horizons represent periods of diminished detrital influx (Baganz et al., 1979). Increases in water depth, probably brought about by delta lobe subsidence, are implied where coals are overlain by sediments (facies Fm, Ff or Fl) indicative of a deeper water environment.

5.3.5   Model for Deltaic Deposition

Vertical facies associations

Although deltaic sequences display a varied and sometimes complex vertical facies association, all have the following general characteristics:

1.   There is a net upward increase in sediment grain size.

2.   Lateral correlation of individual sequences over distances in excess of 5 km is generally difficult owing to variations in facies associations and thicknesses.

Ideally, all elements of the classical vertical arrangement of depositional environments, in ascending order, are represented: prodelta, distal and proximal distributary mouth bar, distributary channel and delta plain (Figs 5.3 and 5.5). This association is subject to variation and many sequences are either incomplete or complex. An incomplete sequence may, for example, comprise a distributary-channel sandstone resting directly on distal distributary mouth bar or even prodelta sediments (Figs 5.13 and 5.14). More complex patterns are developed where an upward-coarsening sequence terminates prematurely, recommences ab initio and then proceeds to completion (Figs 5.5 and 5.13). Incomplete sequences are the most common type of deltaic assemblage. Most sequences show at least some degree of internal complexity.

Figure 5.14. Photo illustrating facies Sp-dominated
distributary channel sandstone overlying either
prodelta or distal embayment deposits. Height of
cliff 8 m. Adjacent to river valley, Gowrie farm.

Thickness of deltaic sequences

The range in thickness of upward-coarsening sequences (average from 5 to
35 m) reflects the depth of water into which lobes were prograding and
subsidence rates within the basin (Galloway and Hobday, 1983). While
compactional subsidence allows the accumulation of complete deltaic sequences,
it introduces an element of instability into delta lobe-development by creating
gradient advantages on the delta plain. Regional or basin subsidence due to
tectonic or isostatic movements can lead to the accumulation of thick deltaic
sequences, especially where sedimentation keeps pace with subsidence.

The wide range in delta front thickness, even within a single lobe is believed to be a function of:

1. The position of the measured section within the delta lobe,
2. the rapid and frequent avulsion of channels feeding lobes, and
3. local variations in the rate of compactional subsidence.

The common development of complex upward-coarsening sequences suggests that their thickness is not a reliable indication of water depth. Giant cross-beds indicate that water depths in front of some distributaries varied from 5 to 25 m. This was probably the typical depth range for water along the coast.

Distributary mouth processes

This aspect of deltaic sedimentation has already been discussed in relation to delta-front and giant cross-bed deposits. Evidence suggests that temporal and spatial variations in channel-mouth dynamics existed. During "normal" or fair-weather conditions it is suggested that hyperpycnal conditions existed with buoyant and inertial factors dominating discharge at distributary mouths. During floods sediment-laden, high-velocity fluvial currents entered the basin as bottom-hugging density currents which flushed sand beyond the mouth bar (hypopycnal conditions). Giant cross-beds reflect a unique combination of conditions. Distributary channels debouched into relatively deep water where density contrasts between the effluent and receiving basin waters were negligible (homopycnal conditions) and inertial factors dominated foreset geometry. Intrasets indicate that hypopycnal conditions also existed.

The influence of waves

Sedimentary characteristics of the distributary mouth bar subfacies suggest that waves played an important role in deltaic processes. Their influence is confirmed by the presence of wave-dominated stretches of coastline. On the

other hand, distributary-channel sandstone morphology, composition and texture indicate <u>wave-influenced</u> rather than <u>wave-dominated</u> deltas. The effect of waves on delta development is believed to have been negated, especially in the nearshore area, by the rapid introduction of coarse sediment and the associated low gradient offshore slope (Coleman, 1976; Galloway and Hobday, 1983; Davis, 1985).

Waves are believed to have influenced deltaic sedimentation in the following ways:

1.   Waves breaking in the vicinity of channel mouths promoted vertical mixing, broke down density stratification and reduced the role of buoyancy (Davis, 1985).

2.   Wave action and associated longshore drift redistributed sand along the coast.

3.   Wave-induced shoreward migration of mouth bars probably constricted outlets during reduced flow (Davis, 1985).

Summary

Most deltaic sequences within the Vryheid Formation have been likened to the Mississippi delta (Hobday, 1973; Mason and Tavener-Smith, 1978). However, from the foregoing discussion it is evident that in contrast to the Mississippi delta, the deltaic sequences described relate to lobate, high-constructive, <u>braid</u> deltas fed by mixed-load braided rivers. Wave reworking also had a significant influence on deposition. It is apparent that these deltaic sequences are, with respect to texture and sedimentary processes, more akin to the short-headed stream delta model proposed by Flores (1975), the "fan" delta complex described by Vos (1981) and the braid deltas of McPherson <u>et al.</u> (1987).

5.4   WAVE-DOMINATED COASTLINES

5.4.1   Delta-Associated Coastlines

Description

A sedimentary succession located at Alcockspruit Quarry, approximately 12 km northwest of Dannhauser, provides the type section for this depositional setting. The quarry face provides excellent outcrop over a distance of about 600 m and a thickness of 34 m, but neither the upper nor lower limit of the sequence are exposed. The succession comprises 3 components (Fig 5.15).

The lower component comprises 5,8 m dark grey facies Fm. Rare, lighter coloured fine- to medium-grained, normally-graded sandstone laminae up to 3 mm thick are present. The upper 0,6 m of the mudrock becomes progressively sandier and contains dense concentrations of Siphonichnus.

Sediments of the middle 10,7 m thick component follow abruptly. They comprise upward-fining sandstone beds capped by mudrock varying in thickness from 50 to 300 mm (Fig 5.16). Typically, the sandstone is moderately well sorted and comprises 60% quartz, 15% K-feldspar, 15% decomposed plagioclase and 10% argillaceous and calcite matrix. Grains are sub-angular to rounded. Bases of beds are nearly always sharp and commonly erosive. Most are planar to gently undulating with wavelengths up to a few metres and erosive relief up to 200 mm, though mostly less than 10 mm. The lower 70 to 90% of each bed comprises fine- to medium-grained facies Sf and Sm sandstone. Individual laminae are 5 to 10 mm thick and commonly emphasised by thin (1 to 2 mm) argillaceous partings. They are predominantly horizontal and parallel, though many laminae display gentle undulations. Primary current lineation is common on bedding planes. The sandstone grades upward over a few millimetres into a zone of darker coloured, silty, fine-grained facies Sr sandstone. The ripples are generally poorly defined, predominantly asymmetric and manifested as
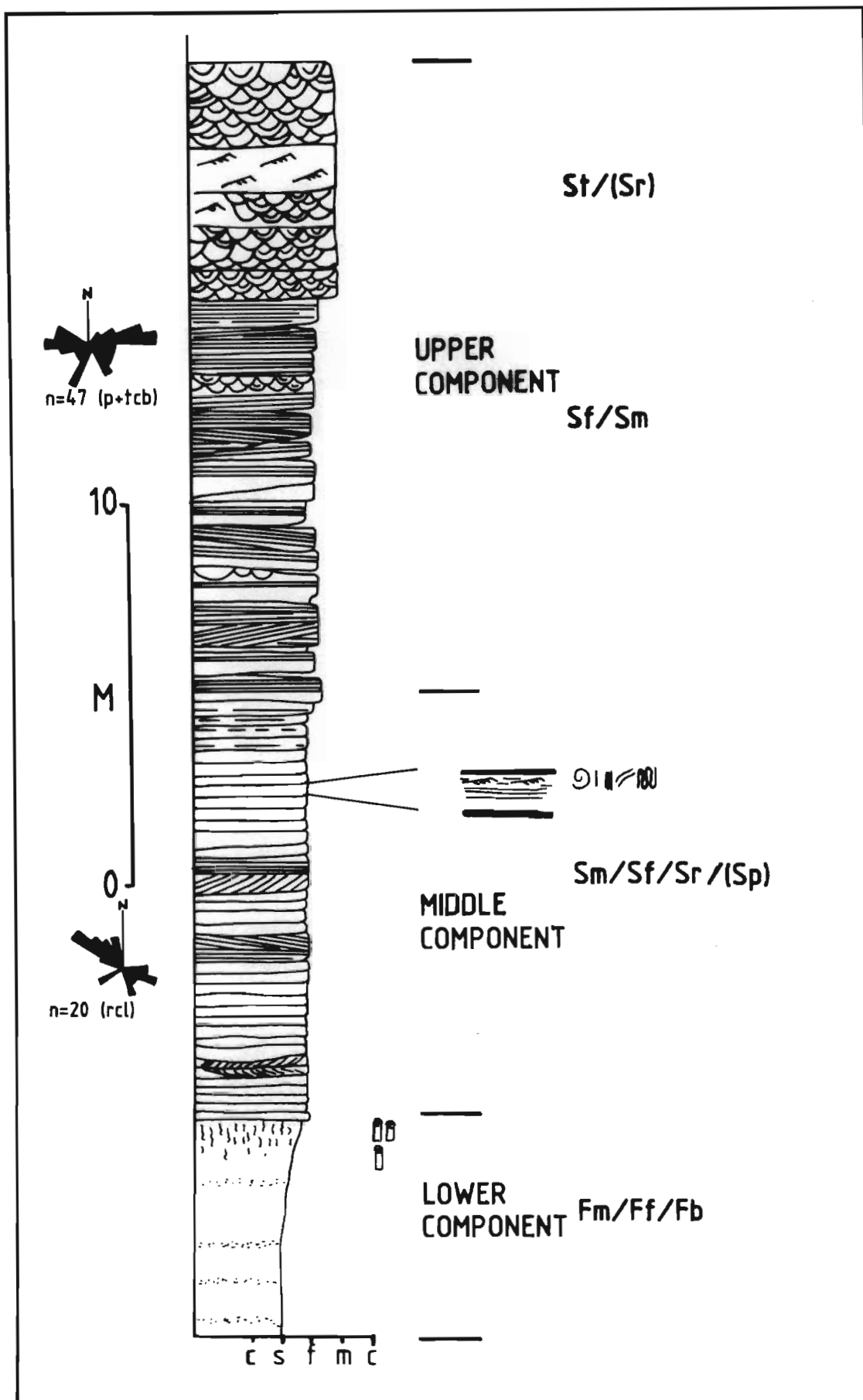
Figure 5.15. Vertical section through the Alcockspruit shoreface deposits. Alcockspruit Quarry.

unidirectional small-scale troughs on bedding surfaces with mean orientations of 306° and 109° respectively. The orientation of primary current lineation associated with facies Sf matches these directions. A few symmetric wave-ripples are also present. In rare instances facies Sr is overlain by a thin (5 to 10 mm) layer of silty, fine-grained facies Sf. Most beds are mantled by between 2 and 20 mm of sharply-based facies Fm. Some of these argillaceous partings may be as much as 150 mm thick and contain a number of fine- or medium-grained facies Sf sandstone partings 10 to 20 mm thick. Within the upper 3 to 4 m of the sequence laminae of facies Sf are less well defined than lower down and consequently the sandstone displays only faint flat laminae or is massive (facies Sm).

Approximately 1 m above the base of this middle component a 1 m deep channel structure contains medium-grained facies Sp sandstone with cross-beds orientated towards the west and southwest (Fig 5.17). Sets are 100 to 300 mm thick and wedge-shaped. Some foresets describe shallow sigmoidal curves but most have truncated tops and asymptotic bottoms. Reactivation surfaces are common, as are siltstone laminae draping foresets. Coarse-grained, pebbly sandstone lenses up to 15 mm thick are present at the base of individual sets. Two sandstone beds, 0,9 and 1 m thick, occur 4,1 and 5,9 m above the base of this division. When traced laterally it is apparent that their abnormal thickness is due to the amalgation of a number of sandstone beds.

Trace fossils are concentrated at sandstone-mudrock interfaces as concave or convex hyporelief burrows and tracks or as vertical burrows penetrating both mudrock and sandstone. Ichnospecies present include Spirodesmos, Nereites, Skolicia, Helminthopsis, Diplocraterion, Skolithos and Siphonichnus.

Beds within the upper component, which is 16,5 m thick, are essentially similar to those of the underlying one. There is, however, a marked increase in bed thickness, the average being 0,3 to 0,9 m (Figs 5.15 and 5.17). Another prominent feature of this component is the 2° to 5° northwesterly dip and
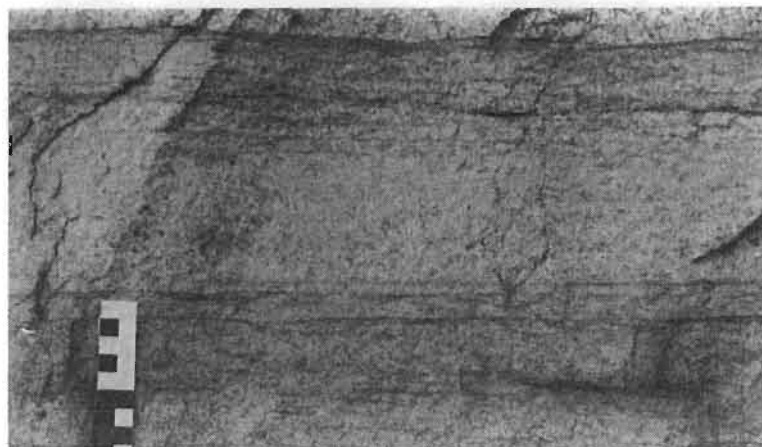
Figure 5.16. Lower shoreface bed showing facies Sm,
Sf and Sr (silty) overlain by veneer of mudrock.
Alcockspruit Quarry. Scale divisions: 10 mm.
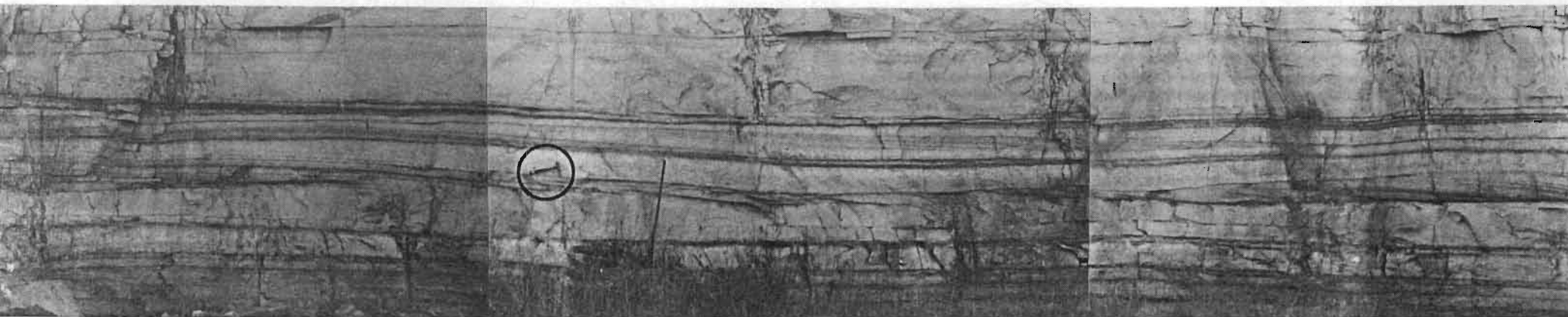
Figure 5.17. Photo-mosaic of quarry wall illustrating upward increase in bed
thickness between middle and upper components and undulating nature of beds.
Hammer for scale. Alcockspruit Quarry.

presence of numerous large-scale bedding undulations and low-angle truncations. Individual beds comprise medium-grained sandstone and are either massive (facies Sm) or display poorly-defined parallel to sub-parallel and gently undulating laminae of facies Sf. This is overlain gradationally by a silty sandstone of facies Sr which is rarely thicker than 30 mm and mostly between 5 and 20 mm. A veneer of dark grey mudrock caps some beds. Planar cross-beds (facies Sp), many showing foresets dipping at less than 15° in a southeasterly to southerly direction, occur as isolated sets between 200 mm and 1,5 m thick (average 20 to 500 mm). Some sets can be traced for up to 40 m in a downdip direction, eventually grading into facies Sf. Reactivation surfaces and siltstone laminae mantling foresets are common. A 0,85 m thick, sharply-based, medium-grained facies Sr sandstone occurs 4,5 m below the top of the sequence. Ripples are asymmetric and vary in amplitude from 20 to 50 mm. On bedding surfaces they are manifested as small troughs with azimuths between 110° and 150°. This division is capped by a 2 m thick, erosively-based, facies St medium-grained sandstone. Sets are between 200 and 400 mm thick orientated between 90° and 120°. Bioturbation is sparse and confined to Siphonichnus, Spirodesmos and Skolithos.

A prominent feature of the upper component is the manner in which individual beds undulate (Fig 5.17). Dip inclinations of these undulations appear to be random and are generally less than 10°. Wave lengths are typically 2 to 25 m and amplitudes vary from 100 mm to 1,2 m. Some undulations are clearly a result of erosion by the overlying strata, but most display form-concordant internal stratification. In the latter case overlying beds may wedge out against the positive relief features.

A sequence similar in most respects to that at Alcockspruit is exposed at Krantzkop approximately 18 km south Dundee. This is 28 m thick and forms a major cliff feature which can be traced at least 4 km. The basal 7,2 m comprises highly bioturbated, silty, fine-grained sandstone grading upward over about a metre into a dominantly sandstone unit 7,5 m thick. This comprises

flat-lying, parallel-sided beds which thicken upward from about 80 to 100 mm at the bottom to 200 mm near the top. Beds within the lower 4 m pass upward from fine- to medium-grained facies Sf sandstone into silty facies Sr. A thin veneer of mudrock caps each bed. As at Alcockspruit, facies Sf makes up the greater part of each bed. Within the upper 3,5 m of this unit beds comprise a basal facies Sf (40 to 50 mm thick) overlain by an equal thickness of facies Sr, facies St (small-scale) or a combination of both. Up to 20 mm of silty facies Sr is present at the top of each bed. Skolithos and Siphonichnus burrows are common. These beds are transitional upward into 10,7 m of facies St through a 2 m thick zone comprising facies St and Sf. The upper 8 m of this component consists of medium- to coarse-grained sandstone with scattered granules. Cross-beds are 150 to 300 mm thick, though some sets may be as much as 0,6 m. Current reversals are common and measurements indicate a strong bipolarity with northeast and southwest components. A maximum towards the south (234°) exists. Siphonichnus burrows are rare but Diplocraterion is moderately common on the top bedding surface of the sequence. The cycle is overlain by densely bioturbated (Siphonichnus) facies Fm. This is succeeded by a 1,5 m sharply-based facies St which coarsens upward from fine-grained sandstone into coarse-grained, pebbly sandstone. Siphonichnus is present throughout and associated with Diplocraterion in the upper 0,5 m.

Interpretation

The upward-coarsening sequence exposed at Alcockspruit, with associated sedimentary structures is similar in many respects to modern and ancient prograding wave-dominated shoreline sequences documented by Bernard et al. (1962), Howard (1972), Davidson-Arnot and Greenwood (1976), Kumar and Sanders (1976), Vos and Hobday (1977), Brenchley and Newall (1982) and Tavener-Smith (1982). However, many features seem to indicate that although wave action played an important role in sedimentation, it was not the only process. The trace fossil suite attests to a shallow marine environment.

Sedimentary characteristics of the lowermost argillaceous component suggest an environment dominated by the deposition of clay and silt from suspension with infrequent and short lived periods of sand influx. Similar depositional regimes are described from shelf environments (Hayes, 1967; Banks, 1973, Reineck and Singh, 1975; Vos, 1977).

Graded sandstone beds overlying shelf mudrocks, regarded as "storm-sand layers" by Reineck and Singh (1972) were probably generated by single depositional events and are believed to have formed when storm-wave energy declined sufficiently to permit suspension fallout of fine sand transported away from the turbulent nearshore zone. Deposition was rapid and, with decreasing wave energy, graded from sand into silt and clay. This interpretation implies an environment below normal wave base where day-to-day deposition from suspension was interrupted by periodic storm-derived influxes. The setting is comparable to the transition zone of Reineck and Singh (1971) and Reineck and Singh (1975) which occurs seaward of the lower shoreface . An alternative mechanism was proposed by Nelson (1982) (discussed below) whereby sand is transported offshore by bottom currents for distances in excess of 100 km. The intensely bioturbated sandy siltstone contrasts sharply with the barren underlying beds. The abrupt appearance of Siphonichnus could be a function of either an increase in water turbidity, oxygen levels, the development of a favourable substrate or a combination of all of these.

The well developed, repetitive vertical sequence of sedimentary structures and vertical grading of grain size in beds of the middle component suggests deposition by waning currents. Beds displaying a similar motif have been recorded in a modern lower shoreface environment by Kumar and Sanders (1976) and in a shallow, epicontinental shelf setting by Nelson (1982). Although the respective authors agreed that each bed relates to a single storm event, they each proposed a different mechanism for the formation of the observed sedimentary structures.

Kumar and Sanders described beds which recorded falling energy levels related to waning storm-wave activity off Long Island, New York in water depths ranging from 5 to 21 m. Flat lamination and associated primary current lineation overlying a sharp, erosive base is believed to reflect sea floor scouring followed by deposition under conditions of intense bottom shear comparable to currents within the transitional lower/upper flow regime. Waning stages of the storm resulted in the formation of ripples. Extended periods of "fair-weather" sedimentation from suspension below wave base produced the argillaceous capping and allowed extensive bioturbation. Davidson-Arnot and Greenwood (1976) regarded composite sets of "plane-to-ripple" bedding within their seaward slope facies (zone of shoaling waves) as being an indication of storm-induced sedimentation. Clifton (1976) described parallel flat lamination (from a similar setting) generated by sheet flow as the oscillatory current equivalent of the upper flow regime. Ancient examples considered to have formed by similar processes in a lower shoreface environment have been described by Howard (1972), Clifton (1976), Vos and Hobday (1977) and Tavener-Smith (1982).

Several recent studies have emphasised the importance of storms in nearshore environments. However, details of the results of such events in the rock record remain obscure because of the lack of sedimentary indicators for identifying them. Hummocky cross-stratification, described originally by Campbell (1966) as "large-scale truncated wave ripple lamination" and later by Harms et al. (1975) and Bourgeois (1980), is commonly regarded as the most distinctive sedimentary structure attributable to storm waves. It has been described mainly from ancient lower shoreface sequences dominated by very fine- to fine-grained sandstone by Bourgeois (1982), Dott and Bourgeois (1982) and Kreisa (1981) but still remains a contentious issue (see Walker et al., 1983; Dott and Bourgeois, 1982; Allen, 1985). The apparent absence of hummocky cross-stratification at Alcockspruit and related sequences is attributed to the predominance of medium-grained sand. Dott and Bourgeois (1982), however, contended that beds displaying a F-X-Mb sequence (facies Sf-Sf-Fm respectively of this thesis) form a minor component of a continuum associated with hummocky cross-stratification.

Dott and Bourgeois believed that flat lamination (passing upward into hummocky cross-stratification) may represent storm-wave stirring not severe enough to scour the bottom but still capable of suspending much fine sediment: deposition simply draped a flat rather than a hummocky surface. An alternative proposal by these authors was that flat lamination was formed by storm-rip or wind-driven currents under unidirectional upper flat-bed flow. In reply to Walker et al. (1983), Dott and Bourgeois (1983) suggested that flat lamination might represent either lower flat-bed deposition or upper flat-bed oscillatory flow. However, they concluded that the maximum orbital velocity necessary for the latter (at least 60 cm/sec for fine sand and 1 m/sec for medium sand at a wave period of 10 sec) is unlikely to occur at depths greater than 10 to 20 m. Walker et al. (1983) attributed flat lamination in such beds to the action of turbidity currents; later wave reworking resulted in hummocky cross-stratification and ripple cross-lamination. Nelson (1982) also stressed the similarity between graded sand layers of the Yukon Delta and Bouma's (1962) turbidite sequence but argued that the observed proximal-distal variations in vertical structures militates against such a mode of emplacement.

Nelson described a sequence of graded sands interbedded with mud extending offshore for over 100 km from the Yukon Delta (North Bering Sea, Alaska) across a shallow (<20 m) epicontinental shelf. Evidence suggests that in severe storms where there is a sea-level setup against the coast, rapidly waning offshore-directed currents develop subsequent to wind-driven surface currents (ie. when storm-wave energy is on the decline) and are the important sediment transporting agents responsible for the observed grading and sedimentary structures. In addition, Nelson noted that the greater energy of wave oscillation currents in shallower inshore water results in predominantly trough cross-lamination in contrast to current-ripple lamination formed in deeper offshore areas. Although the offshore is dominated mainly by bottom-current flow and shows little influence by wave currents, significant amounts of sediment have been removed from the sea floor by resuspension and current advection during storm events.

These processes raise two questions: by what mechanism is sediment transported away from the nearshore zone and do the observed sedimentary structures represent deposition by offshore-directed bottom currents or by wave-induced, waning oscillatory currents? During fair weather sand is maintained in the nearshore zone by dominantly shore-parallel currents and shoaling waves, but by what mechanism it is transported seaward is not yet fully understood. Although Nelson (1982) measured seaward-flowing bottom currents following a moderate storm, no sand was being moved. He concluded that this only occurs after severe storms, a view supported by Murray (1970). Other mechanisms proposed include sediment gravity flows (turbidity currents) (Hayes, 1967), fluid gravity flows (Reineck and Singh, 1972), a downwelling coastal jet induced by onshore winds in conjunction with rip currents (Swift, 1976) and laterally restricted offshore currents analogous to, but on a larger scale and more powerful than, rip currents (Brenchley and Newall, 1982).

The offshore-directed bottom currents described by Nelson (1982) appear to offer the most satisfactory mechanism to explain the observed sequence of sedimentary structures at Alcockspruit. Such currents were responsible for transporting the sediment offshore. This does not, however, preclude later modification by wave processes. It is envisaged that the sandstone beds were emplaced by upper flow regime unidirectional currents during the waning phase of a severe storm. The presence of numerous beds comprising a massive or poorly developed facies Sf basal component argues against deposition by oscillation currents; rapid deposition by unidirectional currents in which individual grains were unable to adjust to the hydrodynamic regime is inferred. Symmetric and asymmetric ripples and small-scale trough cross-laminae displaying opposing current directions in association with an increased argillaceous matrix at the top of individual beds is thought to indicate a decrease in offshore flow intensity and an increasing influence of wave processes. Wave reworking during periods of fair weather caused minor rippling and reworking of the substrate, but bioturbation and the settling of clay and silt from suspension predominated. Thin sandstone beds contained in thicker than average

fair-weather (mudrock) components are considered representative of infrequent, minor storms where only small quantities of sand were deposited. However, the relatively thin development of facies Fm and presence of amalgamated facies Sf/Sf-Sr beds suggests that fine sediment deposited during fair weather was occasionally resuspended during storms by wave currents and/or removed by unidirectional bottom currents.

Sedimentary characteristics of the upper component of the sequence indicate that deposition was dominated by essentially the same processes as the lower part. The overall increase in bed thickness and grain size are taken to indicate shallower water closer to the coast. Numerous low-angle truncations between individual beds and large-scale undulations suggest an environment where the sea floor was regularly scoured by vigorous currents, probably at the peak of storm-wave activity when offshore-directed currents were strongest. The presence of amalgamated beds and beds capped by thin veneers of facies Fm supports such an interpretation. Trough cross-beds at the top of the quarry section are thought to represent the base of the upper shoreface, but insufficient exposure precludes positive identification.

The relationship between directional structures and coastline palaeogeography is commonly indirect because of longshore currents (Horne, 1979). For example, current azimuths oblique to the coastline (the trend of which was deduced from swash lamination and a berm) were reported by Tavener-Smith (1982) from near the base of the Vryheid Formation at Durban. The best indication of the coastline trend existing during the deposition of the sequence under discussion is given by the attitude of the gently dipping major bedding planes. These suggest a coastline orientated northeast-southwest and situated some distance to the southeast (see Davidson-Arnot and Greenwood, 1976). This inferred trend is supported by the westerly and south-westerly orientation of cross-bed azimuths within distributary-channel deposits at stratigraphically higher levels. Offshore-directed trough cross-beds and ripple cross-laminae in the middle component indicate shore-normal currents. The

predominant onshore orientation of these structures, formed during the waning phase of storm activity, reflects the stronger landward component of wave-induced oscillatory currents (Clifton, 1976). The orientation of cross-beds within a channel-like structure near the base of the sequence is probably related to offshore-directed currents. Solitary tabular planar cross-beds in the upper component with foresets extending considerable distances down dip, reflect deposition by shoreward and longshore migrating bars (Hunter et al., 1979) active for relatively lengthy periods of time. Indications that these bars prograded intermittently are provided by numerous mudrock laminae separating foresets.

Sedimentary characteristics of the lower two components of the Krantzkop sequence are in accordance with the depositional processes proposed above. Facies St deposits at the top display characteristics consistent with a prograding upper shoreface regime (Clifton et al., 1971; Vos and Hobday, 1977). Clifton et al., (1971) observed that during fair weather, shoreward migrating dunes and megaripples in the nearshore are deposited in the zone of wave buildup seaward of the breaker zone. The shoreward transition from ripples (of the lower shoreface) to larger bedforms is attributed to a shoreward increase in wave orbital velocity on the sea-floor. The dominantly seaward orientation of cross-bedding is interpreted as a product of rip currents. This is well documented in modern nearshore settings during periods of high wave energy (Vos and Hobday, 1977; Komar, 1976). The predominance of rip current deposits is thought to reflect preferential preservation of high energy events. A smaller number of easterly-directed trough cross-beds indicates that longshore currents were active to a lesser degree.

The absence of foreshore or beach deposits capping the cycle at Krantzkop or any other sequence examined is considered significant. Their complete absence over the entire study area suggests that such beds were not developed, rather than that they were subsequently eroded. Also significant is the fact that each of these sequences is overlain by deposits suggesting formation in an

embayment. Although a lack of outcrop and borehole data precludes close lateral correlation, wave-dominated sequences in the study area are associated laterally with fluvial-dominated deltaic deposits. This suggests that delta lobes were separated by stretches of wave-dominated coastline. The absence of foreshore deposits in such a setting seems to indicate that waves did not normally reach the coast and it is possible that they were prevented from doing so by either a very low coastal gradient (see Coleman, 1976) fronting the interlobe area or a shallow subaqueous offshore bar. The former could have been brought about by the large volume of sediment discharged from the deltas. There is no evidence to support the latter suggestion. The proposal that sequences described in this section occurred seaward of an embayment and reflect deposition by storm-surge offshore currents is supported by the presence of overlying embayment deposits and Nelson's (1982) observation that a coastal re-entrant is necessary to focus storm surge. Field evidence from the study area suggests the probability that such re-entrants would not have exceeded 10 to 15 km across.

To summarise, the model proposed relates to sedimentation along coastlines between major delta lobes. Deposition was primarily by waning storm-surge return currents. Reworking by wave action was also an important process. Water depth in the lower shoreface zone probably did not exceed 20 m. The essential elements of this model are the proximity to active distributary channels, availability of large volumes of fine- to medium-grained sand and short periods of rapid sedimentation. As there is no evidence of deposition in a foreshore environment it is postulated that wave energy was effectively diminished by a gently dipping offshore slope before waves reached the coastline.

5.4.2   Lateral-Accretion Coastlines

Description

Upward-coarsening sequences dominated by facies Sf are confined to the upper

80 to 100 m of the Lower zone. Detailed cross sections east and southeast of Dundee suggest that, unlike fluvial-dominated deltaic deposits lower in the sequence, these cycles are laterally extensive and can be traced with confidence over distances up to 15 km (see Chapter 7). Sedimentary characteristics are generally consistent over the entire field area but the degree of exposure varies considerably. For this reason a detailed description is given of a well-exposed composite sequence at Gowrie and reference made to sections from the Dalray, Winstone, Loch End and Cinderford localities.

The most striking feature of the 19 m thick sequence at Gowrie is its perfectly upward-coarsening nature (Fig 5.18). The basal 4 m comprises a sharply-based, massive to faintly flat laminated siltstone which passes up into a 2 m zone of alternating sandstone and siltstone and then into 3 m of very fine-grained facies Sf sandstone (Fig 5.19). The laminae of this zone are from 5 to 10 mm thick and grade up from very fine-grained to fine-grained sandstone at the base into dark grey or brown siltstone. Above this zone the sandstone becomes coarser and less silty. It grades from fine grained in the lower part to medium grained in the upper 10 m of the sequence. Petrographically, the sandstone is moderately well sorted, most grains being subangular to subrounded. Sutured contacts between quartz grains are common. Most of the plagioclase is moderately to highly decomposed. The matrix is predominantly clay though a small amount of calcite is present. Some of the clay matrix was probably derived from decomposed feldspar grains.

Facies Sf laminae are well-defined, parallel and laterally continuous for at least 10 to 15 m (Fig 5.20). They are normally-graded and primary current lineation is common. Isolated sets of northwest-dipping (325°), low-angle (2 to 10°) planar cross-beds pass laterally into facies Sf over about 15 m. Within the upper 2 to 4 m of the sequence similarly orientated wedge-shaped sets of facies Sp up to 400 mm thick are present. The sequence is capped by a 350 mm thick coal. Sandstone immediately underlying this is root bioturbated and light grey. The coal is overlain by 4,5 m of bioturbated facies Al and Fl followed by erosively based, coarse-grained facies St.
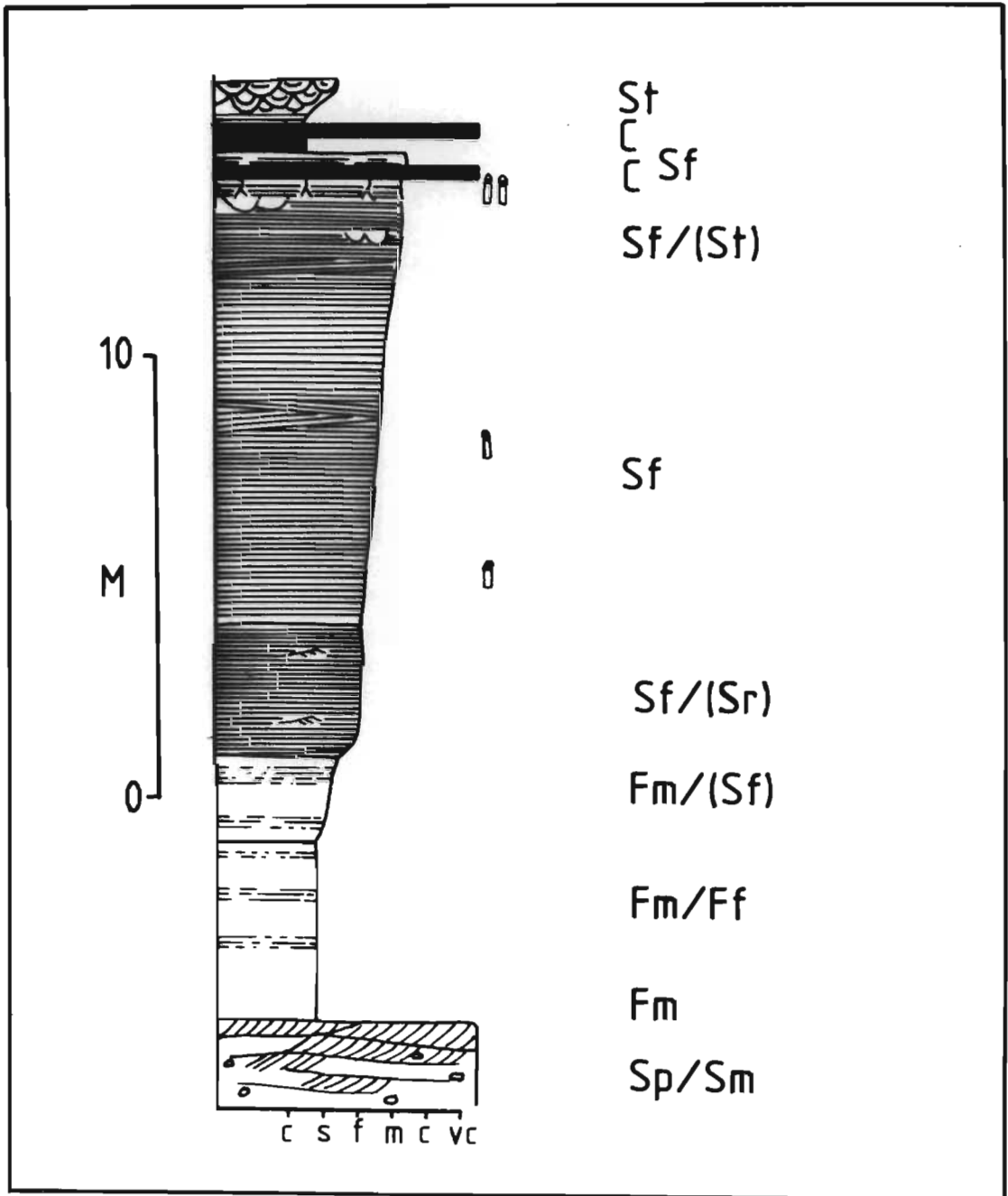
Figure 5.18. Vertical section through lateral accretion-coastline sequence and associated back-barrier coal seams. Gowrie section.

Figure 5.19. Basal portion of facies Sp-dominated lateral accretion coastline deposit. Note upward transition from mudrock- into sandstone dominated sediments. Gowrie section. Scale divisions : 10 mm.
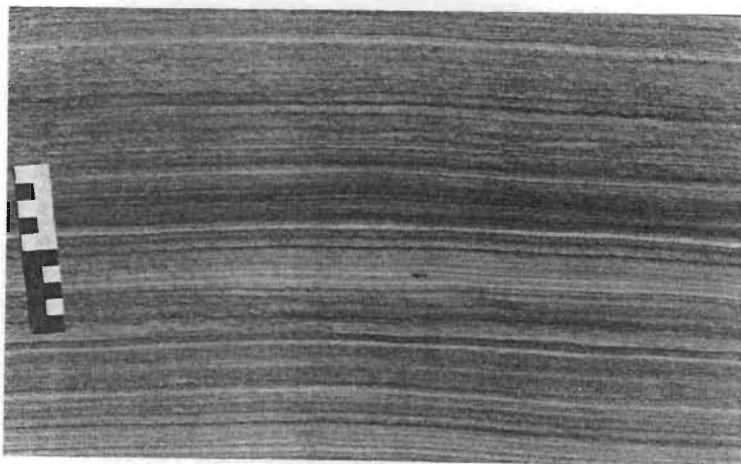
Figure 5.20. Facies Sf-beds. Gowrie section. Scale divisions : 10 mm.

At Dalray a cycle similar to that at Gowrie and at the same stratigraphic level displays sedimentary characteristics which significantly aid interpretation (Fig 5.21). The transition from the basal argillaceous facies is through a 1,1 m zone comprising 50 to 150 mm thick, fine-grained sandstone units interbedded with siltstone. The vertical sequence within units is: sharply-based facies Sf, facies Sr, facies Fl and facies Fm/Ff. Facies Sr shows symmetric and asymmetric ripples commonly mantled by a thin (up to 50 mm) layer of mudrock.

Although facies Sf dominates the overlying sandstone, there are far more sedimentary structures than at Gowrie. In addition, lenticular intercalations of facies Fm/Ff (up to 120 mm thick and 20 m across) are present. Facies Sr (asymmetric and symmetric ripples) is intimately associated with facies Sf. Low angle truncations, hummocky and swaley cross-stratification (Duke et al.,1980) and low-angle (2 to 8°) planar cross-beds are present throughout (Fig 5.22). Isolated sets of facies St (sets up to 200 m) are developed in a 4 m thick zone 2 m below the top of the cycle. Mean cross-bed orientation is towards the northeast, with a range from 350° to 80°. Low-angle truncations are common within the uppermost 1 m of the cycle and major bedding planes dip at 2 to 5° toward the west (280°).

Two significant variations on the above theme occur at Gowrie and Loch End. At Gowrie a cycle immediately underlying the one described above coarsens upward from facies Fm/Ff into a 6 m facies Sf sequence overlain by an erosively-based, coarse-grained, pebbly sandstone of facies St. It is probable that the last sandstone grades laterally into facies Sf over a distance of between 1 and 4 km. This has implications with respect to the interpretation of the depositional setting of this suite of cycles.

At Loch End facies Sf, with abundant primary current lineation and pebble lenses (facies Cc) up to 200 mm wide and 20 mm thick, is incised by a 7,2 m thick, erosively-based sandstone which fines up from medium- to coarse-grained,
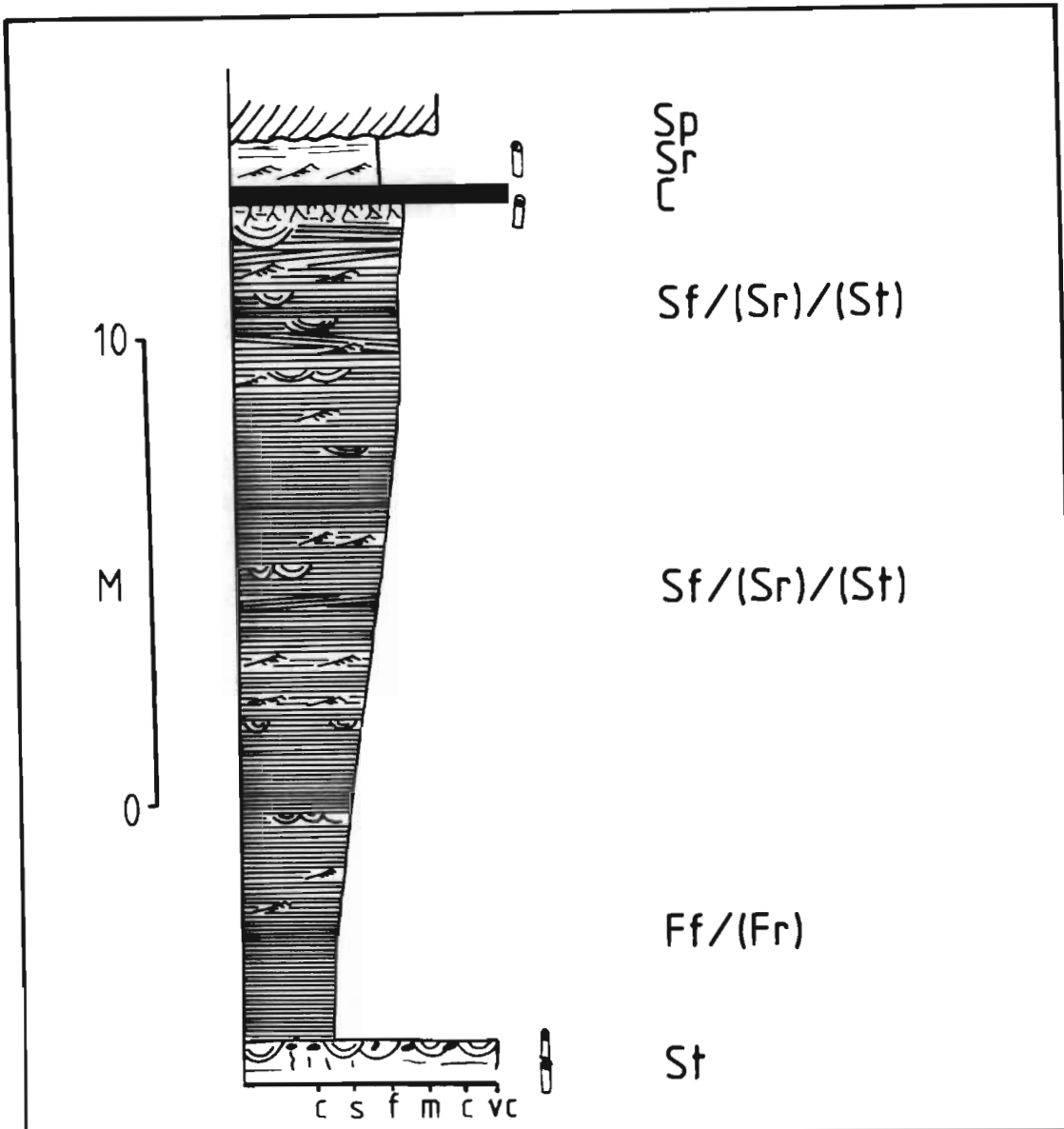
Figure 5.21. Section illustrating lateral-accretion coastline sequence and associated back-barrier coal seam. Basal part of Dalray section.

Figure 5.22. Hummocky cross-stratification, basal portion of Dalray section. River bed, Dalray farm.

pebbly sandstone into fine- to medium-grained sandstone (Fig 5.23). The basal 3 m of the last comprises facies St which passes up into trough-cross beds with a low-angle, semi-planar foreset configuration. In the basal part cross-beds dip to the west. Foresets near the top are similarly orientated but show a weak bipolarity toward the northeast and east. Reactivation surfaces, as well as thin mudrock partings between sets and foresets, are common. Overlying the cross-bedded sequence is a 1 m thick, fine-grained facies Sf sandstone with many low-angle truncations. This is succeeded by a fine-grained facies Sr orthoquartzite with rootlet remains and bright coal laminae up to 5 mm thick. Convolute laminae and Siphonichnus burrows are common. Resting on this are 250 mm of alternating medium- to coarse-grained, massive or crudely-laminated sandstone (up to 40 mm thick) and dark grey, facies Fm mudrock (up to 5 mm thick). The overlying sequence comprises a 3,7 m thick siltstone (displaying facies Fl, Fm and Fr) followed by a coal seam.

Bioturbation is sparse within this suite of sequences but a distinctive trace fossil assemblage is present. At Loch End, facies Sf bedding planes show Spirodesmos and Siphonichnus burrows. Rare Diplocraterion and Skolithos burrows

are present at Gowrie and sparse <u>Skolithos</u> burrows are present at Cinderford and Winstone. <u>Siphonichnus</u> occurs in sandstones immediately underlying the coal seams at both Gowrie and Dalray. The basal argillaceous component of these cycles is mostly devoid of bioturbation, though the lowermost 0,5 to 1 m may contain dense concentrations of <u>Siphonichnus</u>, <u>Scolicia</u> and <u>Nereites</u>. An exception occurs at Winstone where the 16 m-thick lower argillaceous component has been intensely bioturbated by <u>Siphonichnus</u>, <u>Nereites</u>, <u>Scolicia</u>, <u>Planolites</u> (<u>Siphonichnus</u> (F) assemblage) and many other unidentifiable horizontal burrows.

Interpretation

Upward-coarsening sequences dominated by facies Sf are an uncommon feature of the sedimentary record. However, studies of modern coastal depositional processes provide a framework for interpreting the lithologic components of this suite of sedimentary sequences. Their upward-coarsening nature and ichnofossil assemblage suggests progradation into a shallow marine environment.

The basal component of the Gowrie cycle, facies Fm, reflects deposition in an environment dominated by suspension settling. The scarcity of biogenic structures, normally a characteristic feature of this environment (Coleman, 1976; Reading, 1978) shows that conditions were unfavourable for burrowing organisms. Bioturbation in the lower part of this facies is believed to reflect faunal activity immediately following the abandonment and transgression of underlying deposits. Such vertical variations in faunal activity and the presence of sporadic, extensively bioturbated sequences (such as at Winstone) indicate that both spatial and temporal variations in conditions controlling faunal distribution existed.
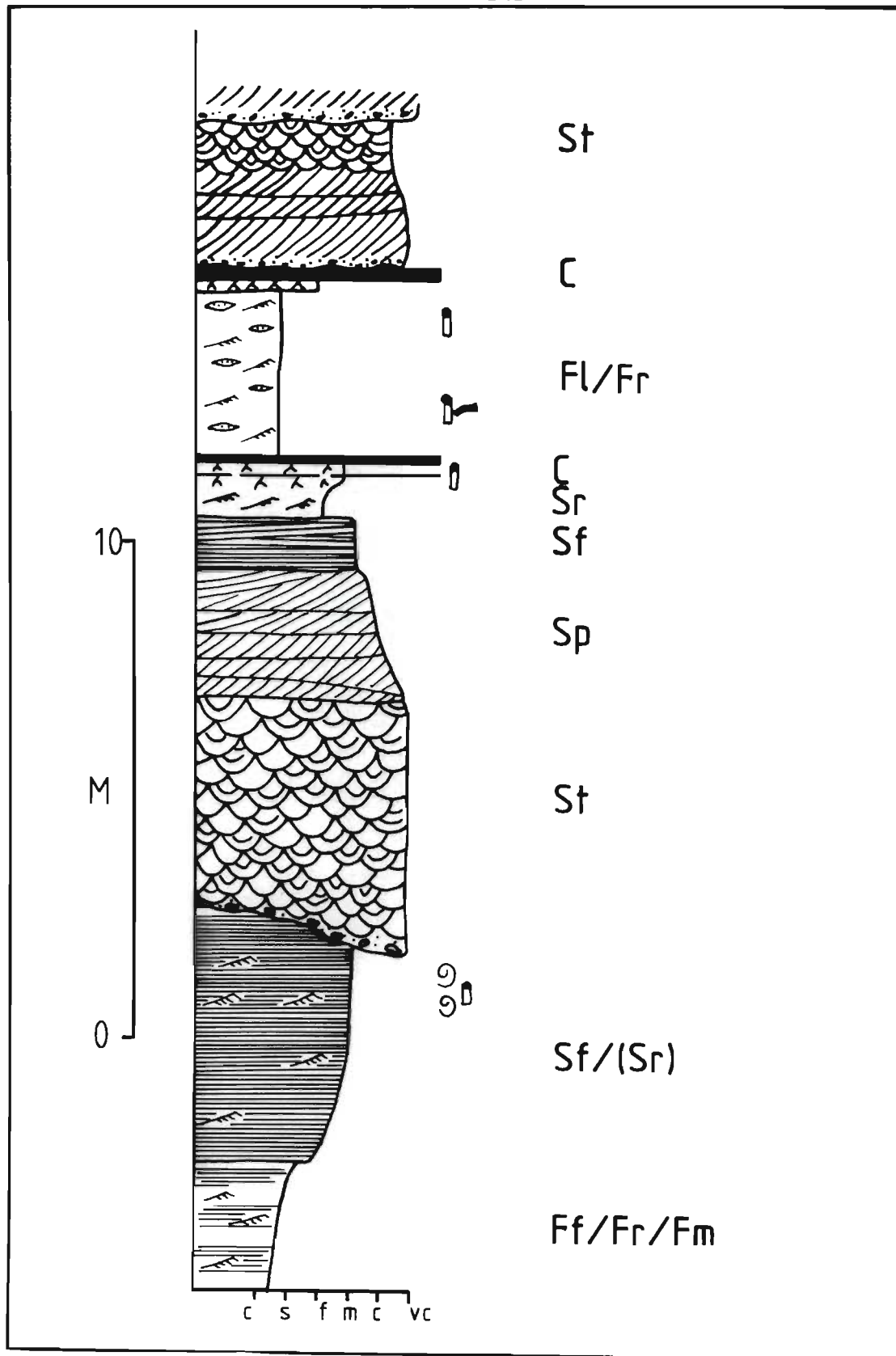
Figure 5.23. Section illustrating shoreface deposits overlain by a tidal-inlet sequence and associated back-barrier sediments. Basal part of Loch End section.

Normally-graded sandstone laminae intercalated with mudrock occupy the same
environmental position as the storm-sand layers at Alcockspruit. A similar
depositional mechanism is implied. However, deposition by offshore-directed
bottom currents is unlikely. The absence of ripples indicates deposition below
wave base and probably seaward of the lower shoreface. The progressive upward
increase in sand content, grain size and traction-generated sedimentary
structures implies decreasing water depth and closer proximity to sediment
source.

Beds at Dalray overlying facies Fm offshore sediments are similar to
storm-dominated lower shoreface deposits described by Kumar and Sanders (1976)
and Tavener-Smith (1982). Distinct, normally-graded laminae of facies Sf at the
base of each unit suggest rapidly alternating flow regimes more typical of wave
surge than unidirectional currents. Symmetric ripple cross-lamination (facies
Sr) also suggests reworking by wave currents (Clifton et al., 1971; Reineck and
Singh, 1975). Facies Fm/Ff reflects a return to "normal" conditions where
day-to-day deposition was dominated by suspension settling below fair-weather
wave base.

Although the upper sandstone component of these sequences is dominated by
facies Sf, it is the subordinate facies that reveal most about sedimentary
processes. The relative lithological maturity of the sand and presence of
wave-formed symmetric ripples (Reineck and Singh, 1975), onshore- and
longshore-orientated cross-beds (inferred from local and regional palaeocurrent
distributions) and hummocky and swaley cross-stratification are strong evidence
for deposition in a wave-dominated environment (Clifton, 1976; Dott and
Bourgeois, 1982; Heward, 1981). Well-defined, graded laminae of facies Sf
confirm this interpretation. Flat laminated sandstone displaying primary
current lineation indicates deposition under conditions of intense bottom shear
(Kumar and Sanders, 1976) comparable to that caused by currents of the upper
flow regime. Low-angle planar cross-beds also indicate a high energy flow
regime (Harms and Fahnstock, 1965; Rust, 1978). Oscillatory currents generated

by waves readily initiate sand movement by sheet flow over a planar surface (Clifton, 1981). The parallel, extremely regular nature of laminae and occasional development of swaley cross-stratification suggests that currents exceeded the velocities required for hummocky cross-stratification (Allen, 1985). Swaley cross-stratification has been interpreted as representing deposition in a storm-dominated upper shoreface environment, probably above fair-weather wave-base (Walker et al., 1983; Leckie and Walker, 1982). The close association between facies Sf and Sr, particularly at Dalray, reflects a fluctuating flow regime probably induced by oscillatory wave surge (Clifton et al., 1971; Komar and Millar, 1975). The rarity of other forms of cross-stratification is probably due to prolonged periods of wave reworking which obliterated evidence of short-lived, lower-energy currents. However, the presence of mudrock intercalations suggests a few pauses in sedimentation. The sporadic occurrence of trace fossils and prevalence of domichnia points to a high, fairly continuous sedimentation rate and an unstable substrate, a feature common to many high energy marine regimes (Clifton, 1976, Bourgeois, 1980; Graham, 1982). The trace fossil assemblage indicates a shallow marine environment: Spirodesmos and Skolithos would probably not have tolerated an environment in which salinity concentrations were diluted by fresh water influxes (see Chapter 3).

Cross-bedding orientations suggest that onshore and obliquely onshore or longshore currents dominated deposition. The absence of offshore-directed structures may be a result of non-deposition but is more likely to indicate a lack of rip-current activity.

The above describes in general terms the environmental setting and sedimentary mechanism responsible for the formation of facies Sf. Somewhat more speculative is the situation within the nearshore where it formed. Flat lamination in modern environments appears to be commonest in the shoreface

where water depths are generally not more than 4 m (Emery, 1960; Bernard et al., 1962; Reineck and Singh, 1971, 1975; Howard and Reineck, 1972). Clifton et al. (1971), in their investigation of a high-energy shoreface, described an "outer planar zone" which approximates to the upper/middle shoreface. It occupies the outer part of the surf zone (shoaling waves) where water depths are 1 to 3 m. Planar parallel lamination (facies Sf of this thesis) was observed to form only as a result of intense wave activity close to the shoreline under waves of short or intermediate periods. To reconcile depths described above with the observed thicknesses of the facies Sf component of each cycle (between 10 and 12 m) one has to assume that deposition took place under conditions of gradual but continuous subsidence and that sedimentation kept pace with subsidence. However, the perfectly gradational upward increase in grain size and absence of noticeable breaks in the sequence suggests that the shoreline underwent steady progradation and aggradation exceeded subsidence. Thus, assuming no significant changes in sea level, maximum wave-base depth (as indicated by the thickness of the facies Sf components (Heward, 1981)) was about 12 m. Following the approach outlined by Clifton (1976), the generation of sheet-flow deposits in fine and medium sand requires orbital velocities in the order of 1 m/sec. Such velocities are not likely to be produced at a depth of 10 m by short-period waves (periods of 5 seconds or less) as the necessary wave height of 3 m would be unstable. However, orbital velocities of 1 m/sec are readily induced at a depth of 10 m by waves approximately 2 m high with periods of 8 to 12 seconds (intermediate- to long-period).

Gently inclined flat bedding, numerous low-angle truncations, and generally less well defined stratification within the upper 1 to 2 m of the cycles at Dalray and Gowrie are some of the features normally associated with swash or foreshore deposits of a beach (Clifton et al., 1971; Kumar and Sanders, 1976; Clifton, 1981; Heward, 1981). Petrographically, these sandstones differ little from those of the underlying deposits and there is no significant change in gross sedimentary characteristics. Although the absence of aeolian (beach)

deposits argues against subaerial exposure, coal seams overlying root-bioturbated sandstone provide indisputable evidence for in situ accumulation of plant material in virtually terrigeneous conditions. It is probable that the swamps and/or marshes existed in a back-barrier setting. (Bernard et al., 1962; Elliot, 1975, 1978; Barwis and Hayes, 1979). The barrier may have only been slightly emergent, but of sufficient stability to maintain a protected environment necessary for peat accumulation. Deposits landward of the back-barrier coals (that is, stratigraphically above them) are similar to those found in the embayment environment (see section 5.3.4.3) and include minor distributary channels, crevasse splays, sheet-sandstone deposits, and prograding sub-deltas. In this sedimentary context, however, deposition was into a lagoon separating the beach barrier from the mainland.

The sedimentary succession at Loch End displays features not encountered to the east and southeast of Dundee. Facies Sf is interpreted as having been deposited in an upper shoreface environment, while the pebble lenses represent post-storm lags as described by Clifton (1981). The overlying erosively-based, cross-stratified sandstone could have been deposited in a number of environments. The dominantly offshore-directed cross-bedding may have resulted from rip currents (Davidson-Arnot and Greenwood, 1976), fluvial activity (Clifton et al., 1971), tidal currents (Hobday and Horne, 1977) or the interaction of surf and swash processes (Clifton et al., 1971). The upward-decrease in grain size, variation in cross-bed morphology and facies associations are, however, compatible with descriptions of modern and ancient tidal inlet deposits (Kumar and Sanders, 1974; Hobday and Horne, 1977; Elliot, 1978; Reinson, 1984). A modern inlet-fill sequence described by Kumar and Sanders has a channel lag-gravel overlain by a deep-channel facies containing ebb-orientated cross-stratification with flood-generated reactivation surfaces. This passes up into a shallow-channel facies comprising flat lamination and ripple-cross lamination. Van Beek and Koster (1972) described cross-beds from a tidal inlet that were consistently ebb-orientated in the lower part and weakly bimodal in the upper.

The mature textural and compositional characteristics of the overlying sandstone strongly suggests wind-generated depositional processes (Reineck and Singh, 1975). Extensive root bioturbation, thin and discontinuous coal seams, abundant plant debris and highly disturbed and almost obliterated sedimentary structures are all features characteristic of backshore dune deposits (Davies et al. 1971; Barwis, 1978; Reinson, 1984). Thin coal laminae intercalated with orthoquartzitic sandstones indicate that plant growth was not continuous but probably interrupted by periodic influxes of sand from storm washovers. Certainly, the combination of features displayed by the overlying medium- to coarse-grained sandstone suggests deposition by storm-washover processes (Schwarz, 1975; Tavener-Smith, 1982). This interpretation implies inlet abandonment and the establishment of a beach-barrier complex. Evidence that the environment landward of the barrier complex was a lagoon dominated by gentle currents and suspension settling is provided by the lithology, sedimentary structures and-trace fossil assemblage of the overlying facies.

Lateral and vertical facies associations strongly suggest that facies Sf-dominated cycles represent laterally extensive accretion-ridge barrier sands (Marley et al., 1979; Flores, 1979) which resulted from wave reworking of distributary channel/mouth bar sands. Sediment (predominantly sand) was distributed along the coast by longshore currents and formed a physical barrier between the open sea and the embayment environment. The back-barrier environment was thus essentially a confined embayment. Long stretches of the coastline (distances of at least 10 to 15 km) were dominated by wave processes and only rarely were beach barriers dissected by distributary channels (as demonstrated by the lower cycle at Gowrie). Significant compositional and textural differences between facies Sf sandstones and those of fluvial-dominated sequences, presence of marine ichnofossils (Spirodesmos and Skolithos), and absence of offshore-orientated cross-beds and facies Al delta-front deposits confirms this opinion. Microtidal conditions are indicated by the presence of tidal inlets incised into barrier ridges (Heward, 1981).

Sequences in which sediment is supplied by longshore drift away from distributary channel-mouth influences were termed "holomarine" by Oomkens (1970). The closest modern analogues to this style of sedimentation are interdistributary, wave-dominated coastlines associated with the Rhone delta complex of southern France (Oomkens, 1970), the Niger delta of West Africa (Allen, 1965; 1970) and the Burdekin of northeastern Australia (Coleman, 1976). This type of delta is characterised by relatively smooth, strike-continuous delta front sands cut locally by distributary channels as well as by microtidal conditions and moderate wave energy.

CHAPTER 6

STRUCTURAL FRAMEWORK AND PALAEOGEOGRAPHY OF THE NORTHEASTERN KAROO BASIN

6.1   STRUCTURAL ELEMENTS AND HISTORY OF THE KAROO BASIN

The main Karoo basin of South Africa is situated on the southern and eastern flanks of the Kaapvaal Craton, an ancient craton nucleus which has been in existence for at least 2500 million years and remained essentially stable for the last 1100 million years (Clifford, 1970). The basin was described by Falcon (1986a) as being a relatively stable, cratonic feature with coal seams developed on the wide southern flank of the Kaapvaal Craton (Figs 6.1 and 6.2A). The initiation and evolution of the basin is, however, an enigma. Tankard et al. (1982) listed a number of alternatives, but regarded none of them as satisfactorily providing an explanation of the mechanism by which large-scale subsidence necessary to create the basin was initiated. These included subcrustal erosion or asthenospheric deflation, mantle phase changes, and lag in isostatic rebound after melting of the Dwyka ice sheets.

Winter (1984) believed the Karoo basin to have been initiated by the collision of two continents during the early Carboniferous (Fig 6.2B). The asymmetric shape of the basin was regarded by him as being typical of a peripheral foreland cratonic basin. The lack of high-temperature metamorphism and volcanism normally associated with continental collision was attributed to the southward subduction of oceanic crust along the collision zone (Dickinson, 1974). Although orogenic effects of collision persisted until the end of the Permian or early Triassic, the northern and eastern parts of the basin remained relatively stable and passive, and sedimentation was largely unaffected by the collision.

An important aspect of the evolution of the Karoo basin was the isostatic response of the crust to firstly, loading and unloading as a result of the Dwyka glaciation and deglaciation and then, further loading within the basin
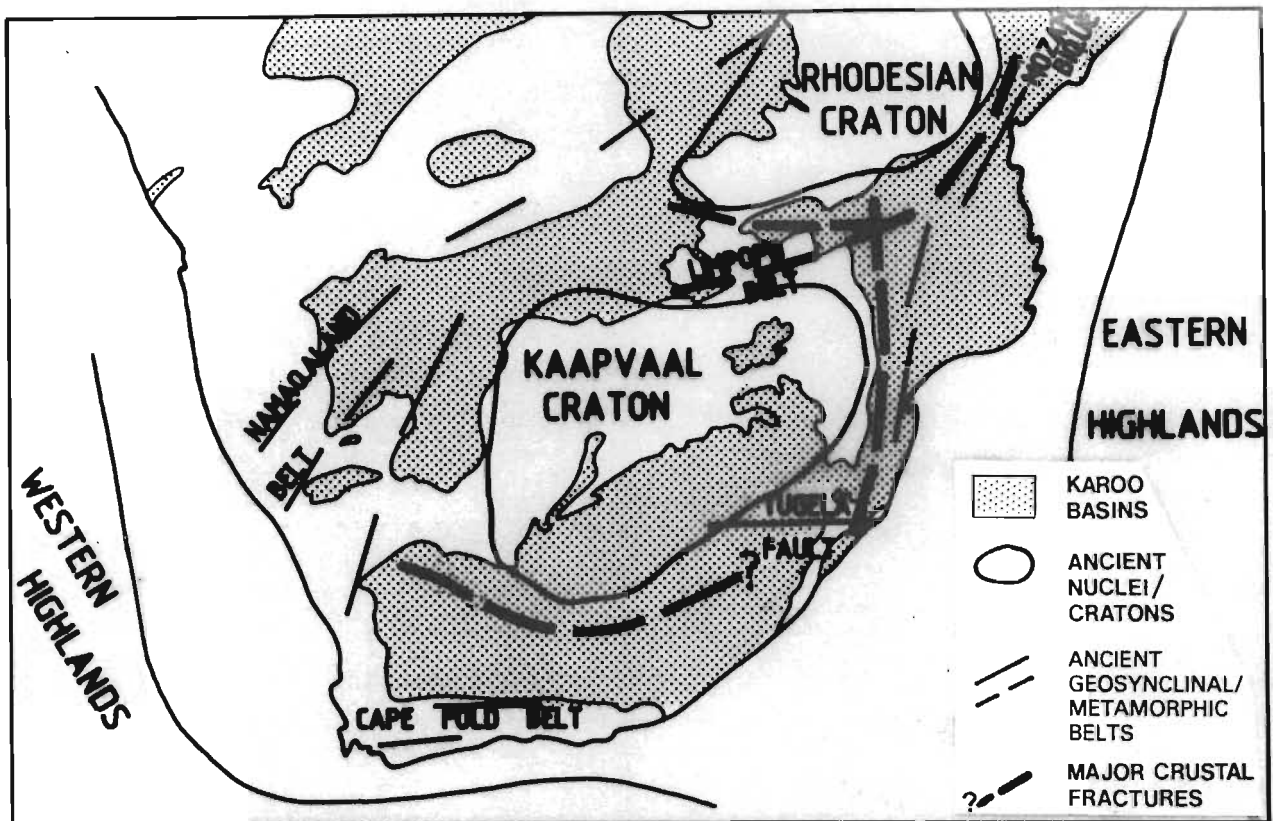
Figure 6.1. Major structural features affecting Karoo sedimentation in southern Africa during Palaeozoic times (after Falcon, 1986a).
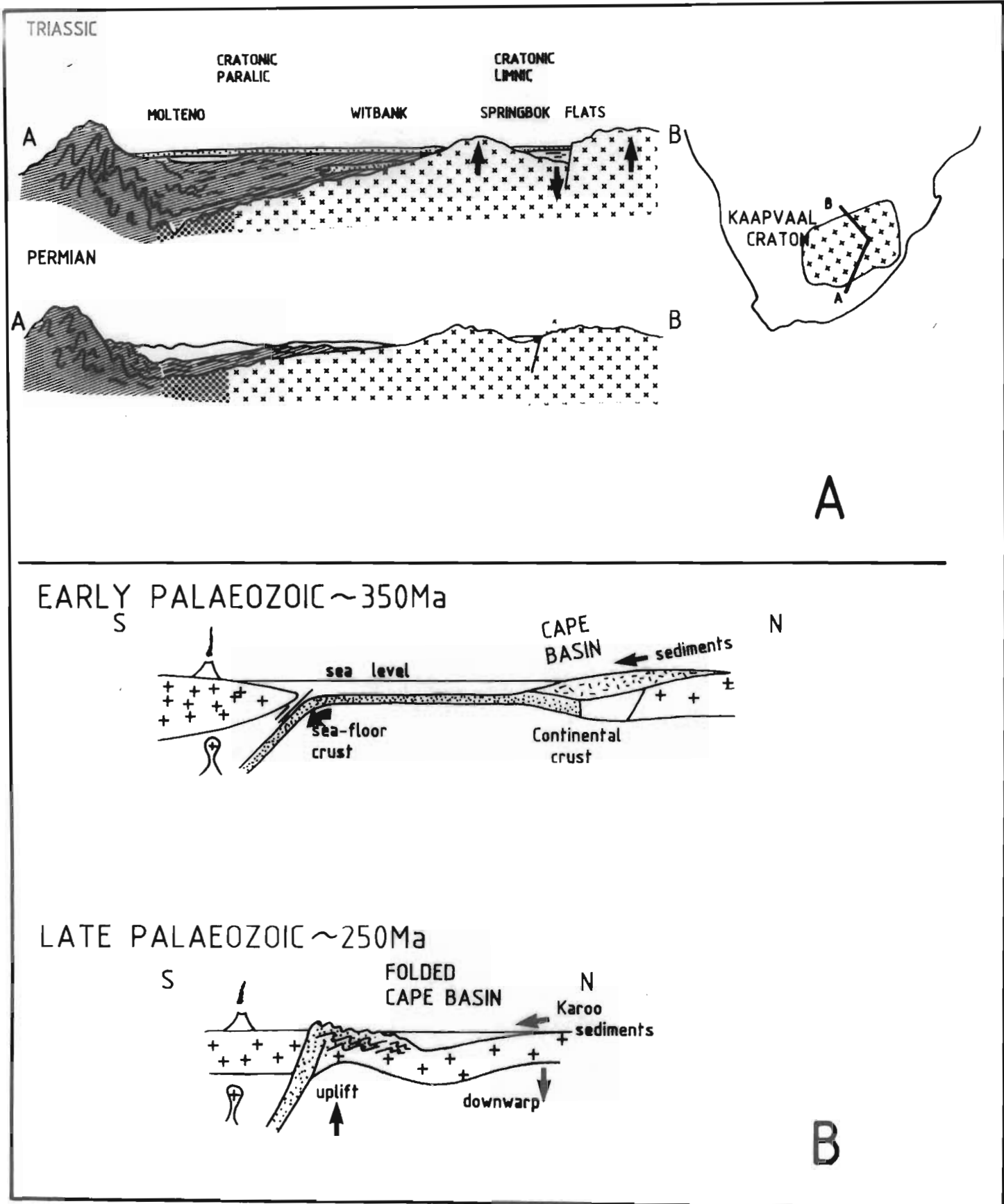
Figure 6.2. A: cross-section through the coal-bearing basins in South Africa during the Triassic and Permian (after Falcon, 1986a). B: suggested evolution of the Karoo basin based on southward subduction of oceanic floor (after Winter, 1984).

during Ecca deposition. Walcott (1970) showed that isostatic response of the crust to unloading due to glaciation takes place exponentially, with uplift most rapid initially, and slowing with time. Likewise, subsidence due to loading (sediment or otherwise) is initially rapid and thereafter slows. The continental lithosphere has a high flexural rigidity and reacts to loading in an elastic manner; loading produces subsidence extending several hundred kilometres from the point of the load (Walcott, 1970, 1973; Murrel, 1976).

Relatively little is know about the structural history of the areas to the immediate north and east of the basin; vague reference is made to structurally positive (highland) provenance areas situated north of Witbank (Witwatersrand Arch) and off the Natal east and northeast coasts (Eastern Highlands) by Ryan (1967). Mathew (1974) and Whately (1980), whose investigations focussed on the Vryheid and Nongoma areas of northern Natal respectively, referred to adjacent source areas to the north which were elevated as a result of isostatic rebound following deglaciation.

A number of intrabasinal structural features have apparently influenced sedimentation in the northeastern part of the basin significantly (Fig 6.3). Ryan (1967) identified the Natal Trough, a northeast-southwest orientated feature which played an important role in Ecca sedimentation. Tavener-Smith (1979) focussed on this feature, confirming its influence on the thickness and palaeodrainage of the Ecca sequence. Confirmation that this was a structural element of long standing was provided by Matthews (1970), who reported that it existed during the deposition of the Natal Group sediments and was still a topographic lowland during Dwyka times, and Stratten (1970) who measured anomalously thick sequences of Dwyka sediments along the trough axis.

A possibly contiguous feature to the Natal Trough, a north-south orientated graben, was recognised by Whately (1980) in the vicinity of Nongoma, Zululand. It has not been established if the graben was active prior to the Permian, but its development is attributed to incipient rifting in response to crustal
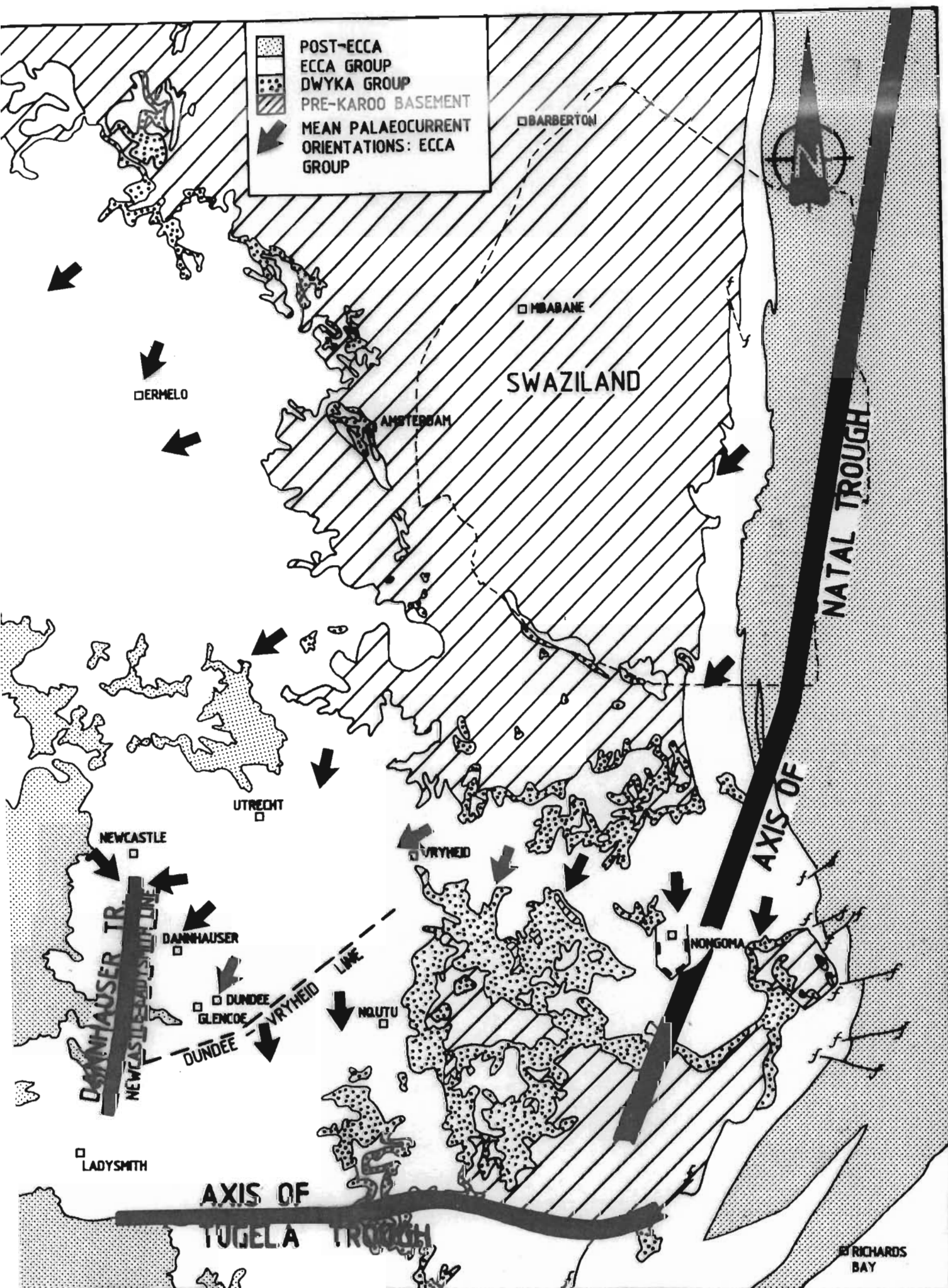
Figure 6.3. Structural framework of the northeastern Karoo basin during

thinning induced by tensional stresses prior to the break up of Gondwanaland. Subsidence within the graben during deposition allowed at least 1 029 m of Ecca sediments to accumulate. Christie (1984) illustrated the increase in thickness and complexity of the Vryheid Formation sequence eastwards from Vryheid, through Ceza to southeast of Nongoma, probably in response to this feature,by means of a cross section.

Tavener-Smith (1986) has shown, by means of a north-south cross section (Fig 6.4), that the complexity and thickness of the Vryheid Formation increases southwards from near the northern margin of the basin (Paulpietersburg) to Tugela Ferry. This cross-section, showing vertically-stacked progradational sequences, demonstrates that deposition was accompanied by progressive basin subsidence. The abrupt increase in floor gradient southeast of Vryheid is clearly illustrated and coincides with coal-seam thinning and splitting. Tavener-Smith regarded the region to the south of this hinge line (the "Dundee-Vryheid" line) to be within Ryan's Natal Trough. In view of this trough's orientation and position as indicated by Ryan (1967), however, it is more likely that this is the northern flank of Ryan's Tugela Trough, the axis of which is defined by the east-west striking Tugela Fault. Tavener-Smith's conclusions nevertheless remain valid and are in agreement with Ryan's (1967), Hobday's (1973) and van Vuuren and Cole's (1979) observations that the Tugela Trough was a zone of rapid and prolonged subsidence during Ecca sedimentation. This resulted in less extensive shoreline progradation and the deposition of thick sequences of deltaic sediments.

A north-south orientated palaeovalley situated just west of the Ladysmith-Newcastle road and south of Newcastle, referred to here as the Dannhauser Trough, was delineated by van Vuuren and Cole (1979). It is believed to be a continuation of a linear zone of rapid subsidence identified by Ryan (1967) to the northeast of Newcastle. Figure 6.5 illustrates the geometry of this trough and shows that although glacial deposits had a subdueing effect on relief, it was still a prominent feature during early Ecca sedimentation. Christie (1985) confirmed the presence of this trough by recognising that while
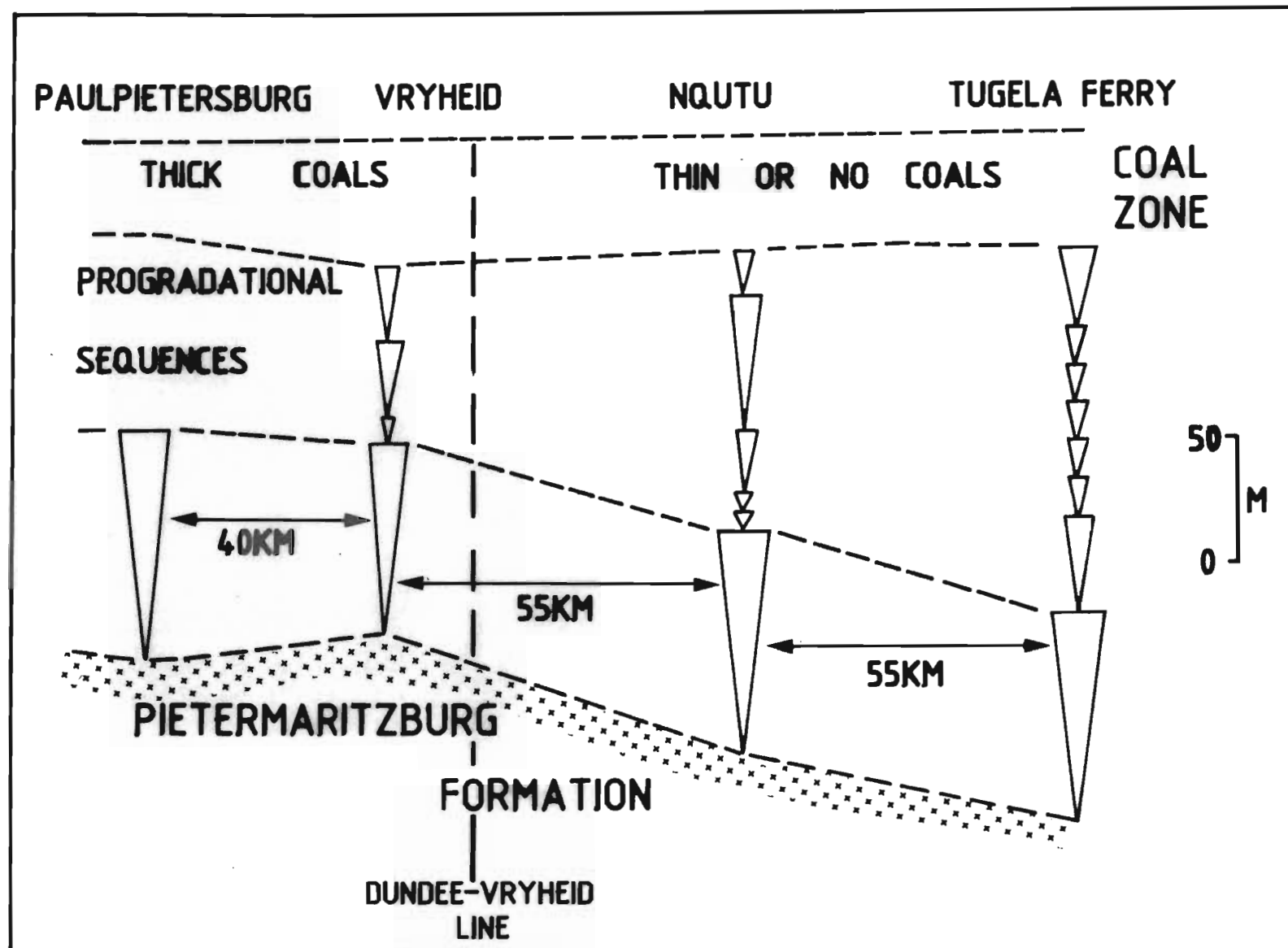
Figure 6.4. Diagrammatic representation of prograding sequences along a north-south section between Paulpietersburg and Tugela Ferry (after Tavener-Smith, 1986).
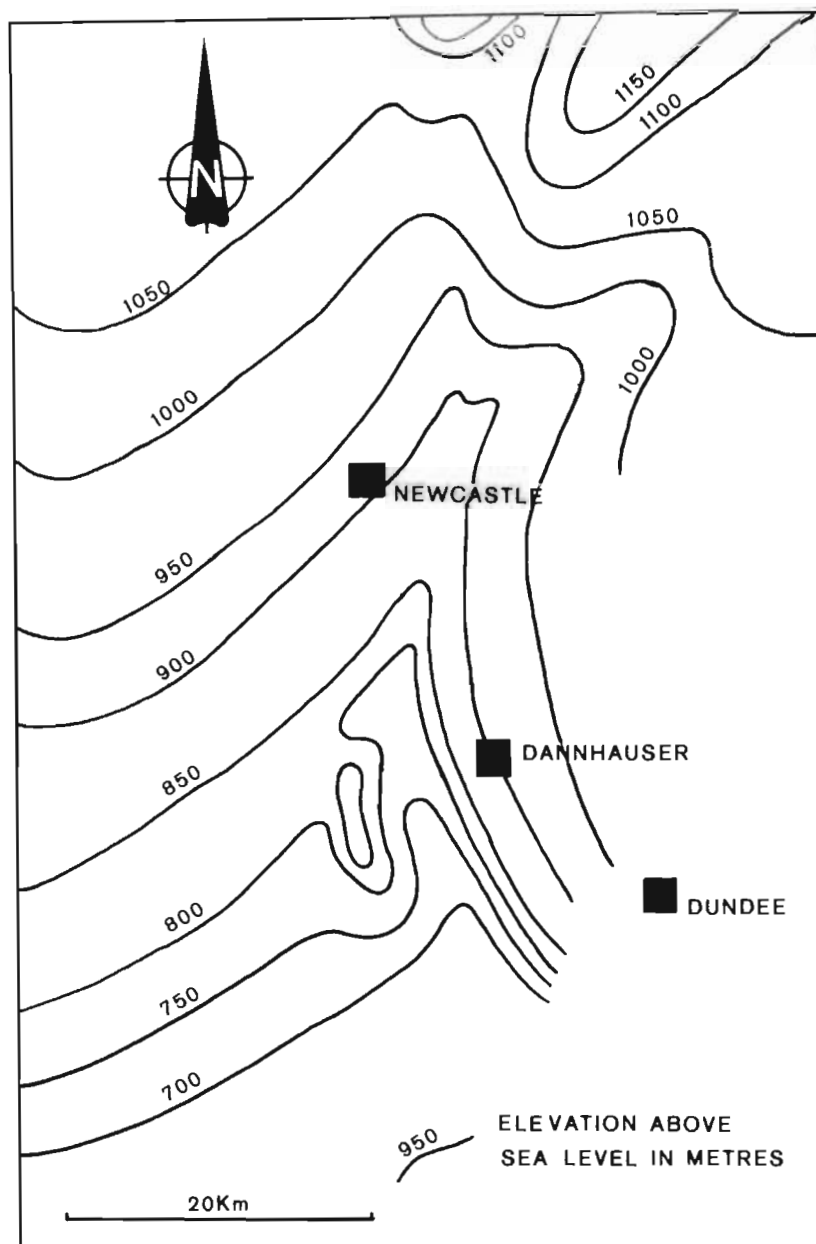
Figure 6.5. Structural contour map, top of Dwyka Formation, of the Klip River coalfield (after van Vuuren and Cole, 1979).

the Coal zone thickened rapidly west of Dannhauser, coal seams split and thinned in the same direction (Fig 9.18). Similarly, Roberts (1985) identified this trough axis to the northeast of Newcastle as a zone where coals are thin or absent. This feature is discussed in more detail in the chapter dealing with controls on coal distribution.

The foregoing observations indicate that basinal tectonics and palaeorelief had a profound effect on Karoo sedimentation, but there is no doubt that erosion before and during Dwyka glaciation (partially influenced by the structure and lithology of the basement rocks) modified pre-Karoo topography extensively (Ryan, 1967; Stratten, 1970; Matthews, 1970; Crowell and Frakes, 1972; Hobday and Von Brunn, 1979; Von Brunn and Stratten, 1981). Le Blanc Smith (1980b) convincingly illustrated the very localised control pre-Karoo erosional topography had not only on fluvial drainage patterns, but also on coal-seam distribution in the Witbank coalfield. Although not very much detailed data relating to the topography of the basement underlying the northern Natal coalfields exists, it is probable that relief features were not propagated through the relatively thick sequences of the Dwyka and Pietermaritzburg Formations.

6.2 CLIMATE

Climatic factors, including temperature, rainfall and humidity have a profound effect on sedimentation. They determine rate and type of weathering and rates of plant growth and decay, and hence the volume of sediment passing through a drainage system into its receiving basin. The general consensus is that the climate prevailing during Vryheid Formation sedimentation was cold to cool temperate following the widespread arctic to sub-arctic conditions experienced during the Permo-Carboniferous Dwyka glaciation. Evidence supporting this includes:

1.   The stunted, broad-leafed Glossopteris-Gangamopteris flora which is
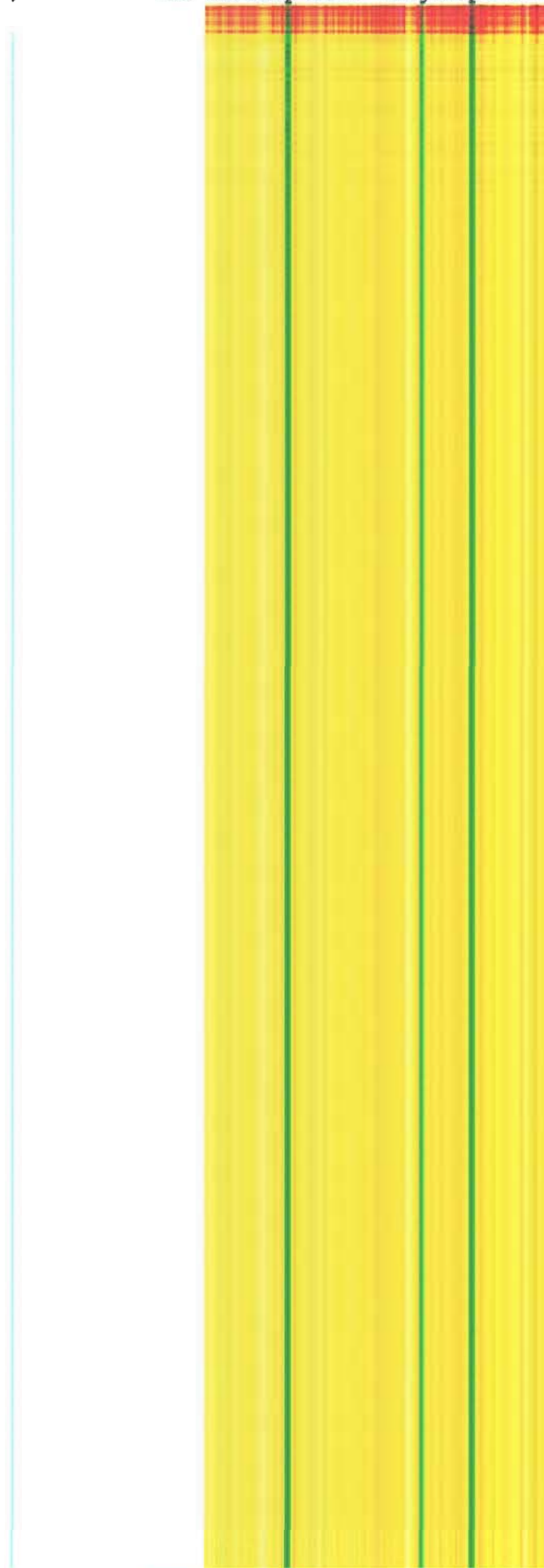
1. The stunted, broad-leafed Glossopteris-Gangamopteris flora which is similar to that found at present at high latitudes (Plumstead, 1957; Falcon, 1975, 1986a),

2. the presence of fresh feldspar in arenites (Chandra and Taylor, 1982),

3. indirectly, the absence of leached seat-earths below coal seams (Plumstead, 1962), and

4. the inferred high latitude occupied by Gondwanaland during the deposition of the Karoo Sequence (Martin, 1961, 1981; Anderson and Anderson, 1985; Tankard et al, 1982).

Seasonal variations in temperature and rainfall are indicated by plant fossils displaying marked seasonal growth rings (Plumstead, 1962) and fluctuations in sedimentation rates implied by facies Al (see also Hobday, 1973) respectively. Falcon (1986a), on the basis of present-day vegetation-zone requirements estimated a mean annual precipitation of 1 000 mm during Vryheid Formation times. The relatively high proportion of dispersed clay and silt in Gondwana coals (compared to Laurasian coals) was suggested by Plumstead (1962) to reflect windy conditions during peat accumulation. Windy conditions, as implied by the effect waves had on coastal sedimentation in the Lower zone, are a feature of present-day temperate zones. The intimate association that frequently exists between fresh and weathered (to kaolin) feldspar grains observed in most sandstones strongly suggests varying degrees of pre- and syndepositional weathering and a humid atmosphere.

6.3 PROVENANCE AND PALAEODRAINAGE

Regional palaeocurrent distributions indicate that most drainage within the

northeastern Karoo basin was to the south and southeast. Within the Klip River coalfield the pattern is similar but a distinct spatial variation in palaeodrainage patterns exists (Fig 6.6). There are no significant variations in mean palaeocurrent orientations between the Lower and Coal zones at any specific locality. Palaeocurrent distributions throughout the area relate closely to pre-Karoo basement gradients and indicate a westerly to southwesterly prograding coastline in the vicinity of Alcockspruit, west of Dannhauser and west and southwest of Dundee. South and southeast of Dundee progradation was towards the south and southeast respectively. The orientation of edges of fluvial sheet sandstones within the Coal zone lends further support to the above palaeodrainage distributions (Fig 9.15). North and west of Alcockspruit drainage during deposition of the Coal zone was toward the southeast. This direction is also confirmed by the southeasterly orientated edge of a fluvial-channel sandstone.

The coarse, immature nature (compositional and textural) of fluvial-channel sandstones points to rapid deposition from a provenance area characterised by significant relief and weathering rates. Two schools of thought as to the location of the source area for the Vryheid Formation in this part of the basin exist. Ryan and Whitfield (1979) and Stratten (1970) postulated an "Eastern Highland" source area off the present southeast African (Natal/Mocambique) coastline. Roberts (1985) also regarded this as the probable provenance (of granitoid composition) to Vryheid-area sediments. R. Tavener-Smith (pers. comm.), on the other hand, is of the opinion that, in view of the texture of fluvial sandstones and implied short-headed nature of the deltas, the source area was situated further to the southeast in the Barberton-Swaziland area. Whateley (1980) proposed a source area situated in a similar locality, the "Swaziland Highlands", to explain depositional patterns in the Nongoma area.

Figure 6.6. Palaeodrainage patterns existing during deposition of the Lower and Coal zones (continued over page).
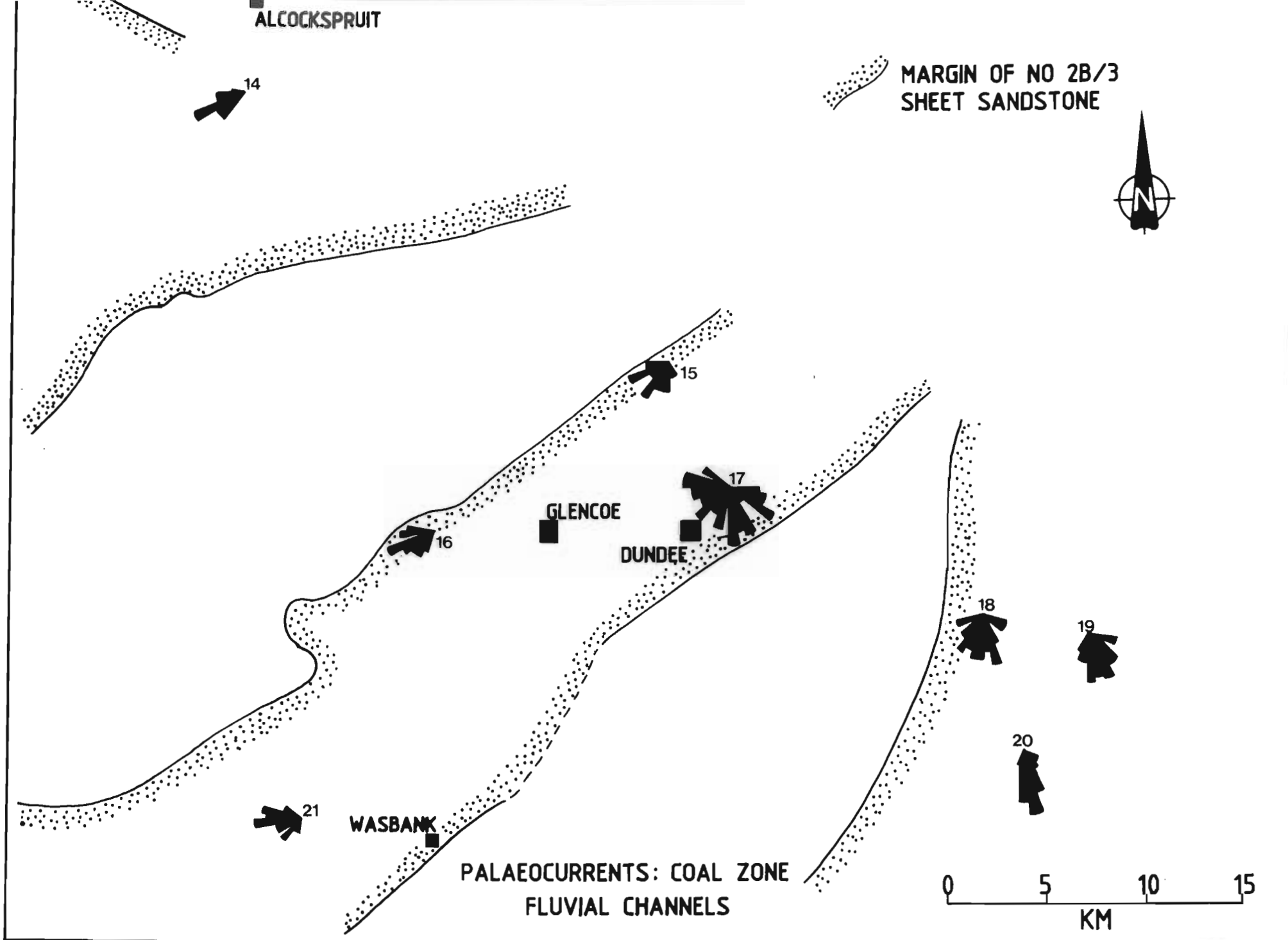
Figure 6.6 (continued). Numerals refer to localities and statistical data presented in Appendix 1 (back folder).

Features having a possible bearing on the location of the source area include:

1. The existence of easterly dipping (5° to 10°) Karoo sediments crop-
    ping out along the eastern border of Swaziland and Kangwane. In
    Kangwane, coals in the Vryheid Formation generally occur less than
    10 m above the pre-Karoo basement and in some areas they rest
    directly on it (similar to the Witbank coalfield). The pre-Karoo
    surface dips towards the east at 3°, a manifestation of the
    north-south trending Natal Trough (Ryan, 1967). The easterly
    thickening of the Vryheid Formation in response to this is
    accompanied by an overall thinning of coal seams in the same
    direction.

2. Cross-bedding orientations from within and below the Coal zone of the
    Vryheid Formation in the above outcrop area are towards the west,
    southwest and south (Ryan, 1967).

3. Dwyka and Ecca sediments occur as outliers within the Swaziland
    highland area along the Swaziland-Natal border and further towards
    the northwest to around Amsterdam. No Karoo rocks are present to the
    northeast and north of this line in the Swaziland-Barberton area (see
    Fig 6.3).

4. There is evidence that the Swaziland-Barberton area was covered by
    southwesterly migrating ice sheets during the Dwyka glaciation
    (Stratten, 1970).

5. Christie and Tavener-Smith (1979) identified a westerly-flowing
    drainage system within coal-bearing Vryheid Formation sediments along
    the Natal north coast (between Ballito Bay and Sheffield Beach). This
    corresponds to similar evidence presented by Ryan (1967) supporting
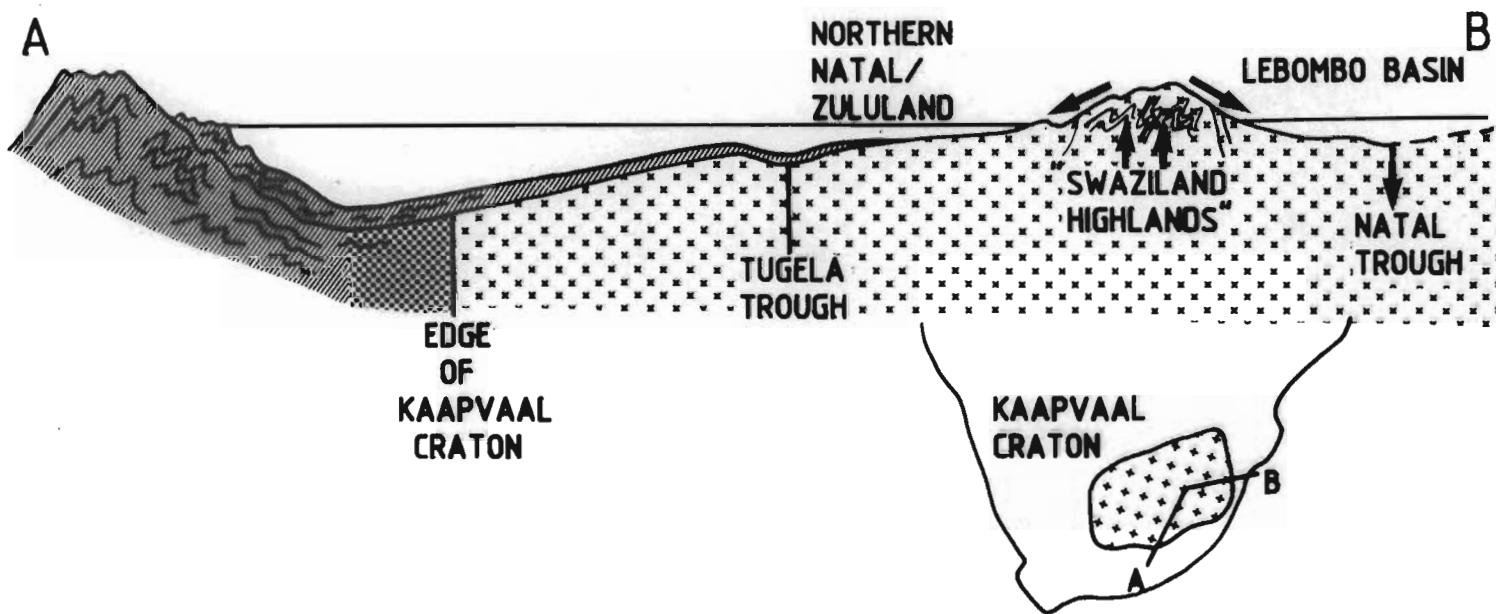
Figure 6.7. Proposed model for provenance of Ecca sediments in the northeastern Karoo basin.

the existence of an easterly-situated provenance.

6. Whateley (1980) measured southerly-directed palaeocurrents in the vicinity of Nongoma.

7. Investigations by van Vuuren (1981) and Le Blanc Smith (1980b) have established that the present position of the northeastern margin of the Karoo basin is little changed from Vryheid times.

From the above it is evident that the Swaziland-Barberton area was a structurally positive or at least stable part of the basin during the deposition of the Vryheid Formation. There are, however, conflicting data as to whether it constituted a source to the sediments of the northeastern part of the Karoo basin. While Whately (1980) found convincing evidence that it controlled the palaeoslope to the south (he suggested that the area was undergoing active uplift due to isostatic rebound and incipient rifting), palaeocurrent distributions along its eastern flank as measured by Ryan (1967) indicate no such control. Indeed, most currents were towards the southwest (ie. towards the basin margin) and south (Fig 6.6). On the other hand, there is clear evidence that pre-Karoo basement topography and subsidence rates influenced peat accumulation and the thickness of the Vryheid Formation in this eastern area. From the above it is also clear that the Natal Trough was active during the Permian. If Ryan's Eastern Highlands did exist, one is faced with the complexities of explaining sediment transport across the actively subsiding Natal Trough and then the relatively stable (topographically positive) or uplifting Swaziland-Barberton area. Figure 6.7 illustrates the proposed source area to Ecca sediments in the northeastern Karoo basin.

Although the presence of Ecca sediments on the eastern and southwestern borders of Swaziland does place some restrictions on the extent of a possible source area, the position of the Swaziland-Barberton area as a provenance is considered more suitable than the "Eastern Highlands" for the following reasons:

1. The texture and depositional pattern of fluvial deposits is consistent with the proximal to medial reaches of present-day high-latitude rivers (such as the Platte and Sasketchewan - see Chapter 5); coarse sediment is transported down a steep gradient fronting the source area by low sinuosity, bed-load dominated rivers.

2. It favours the interpretation that the deltas were short headed with a source-to-headwater distance in the order of 200 to 300 km.

3. The composition of Vryheid Formation sandstones, ie. predominantly quartz, feldspar and rock-fragment grains, is compatible with the lithology of the Archean and Proterozoic granitoid rocks of the Swaziland-Barberton area.

CHAPTER 7

THE LOWER ZONE

7.1   STRATIGRAPHY AND DEPOSITIONAL SETTING

The Lower zone is dominated by upward-coarsening sequences varying in thickness up to 60 m with an average of between 8 and 40 m. Detailed annotated vertical sections are presented in Figures 7.1 to 7.10 (see back folder). From these illustrations it is evident that many of the sequences within the lower part of the Lower zone (lower component) do not extend for more than 4 to 5 km normal to palaeodip but have a lenticular morphology (Figs 7.11 and 7.13). Parallel to palaeodip some of these sequences can be traced with confidence for at least 6 km (Fig 7.12), but beyond this appear to split basinward into two or more sequences. The top of the Lower zone (upper component) to the northwest of the Dundee-Vryheid line is defined by a 10 to 40 m thick fluvial sheet sandstone which extends over an area in excess of 1 500 sq km (Figs 7.11 and 7.12). To the southeast this sheet sandstone is transitional into a succession of upward-coarsening, wave-dominated sequences which can be traced laterally for distances of up to 15 km. The paucity and low density of outcrop and borehole data prevents the compilation of any meaningful isopach maps.

Included in this description are black siltstones of the Pietermaritzburg Formation which were either completely or partially intersected in a total of 11 boreholes in the vicinity of Dannhauser only (Figs 7.11 and 7.12). The thickness varies from 40 to 85 m and appears to be strongly influenced by pre-Karoo basement relief.

7.1.1   The Lower Component.

The lower component of the Lower zone represents deposition along a shoreline dominated by high-constructive (fluvial-dominated) lobate or braid deltas. Although delta sedimentation was influenced to varying degrees by wave processes, only short stretches of the coastline (between delta lobes) were wave dominated. In general, wave energy was insufficient to produce laterally continuous delta-front sheet sands.
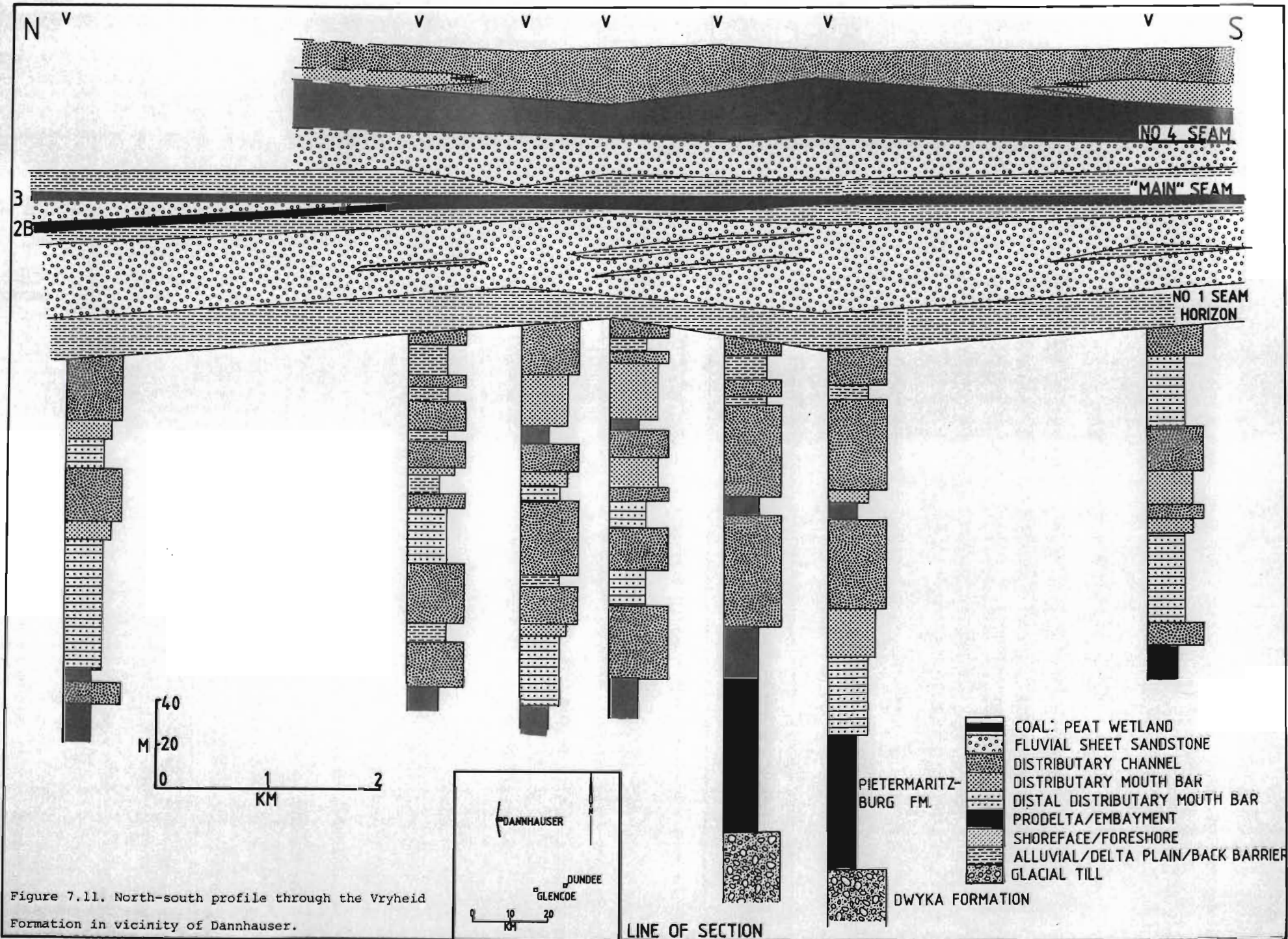
Figure 7.11. North-south profile through the Vryheid
Formation in vicinity of Dannhauser.

N

S

NO 4 SEAM

"MAIN" SEAM

3

2B

NO 1 SEAM
HORIZON

40

M -20

0   1   2
         KM

DANNHAUSER

DUNDEE
GLENCOE

0   10   20
        KM

LINE OF SECTION

PIETERMARITZ-
BURG FM.

DWYKA FORMATION

COAL: PEAT WETLAND
FLUVIAL SHEET SANDSTONE
DISTRIBUTARY CHANNEL
DISTRIBUTARY MOUTH BAR
DISTAL DISTRIBUTARY MOUTH BAR
PRODELTA/EMBAYMENT
SHOREFACE/FORESHORE
ALLUVIAL/DELTA PLAIN/BACK BARRIER
GLACIAL TILL

W v v v ' v v v E

NO 4
SEAM

3 "MAIN"
2B SEAM
1
NO 1 SEAM
HORIZON

?
?
?
?
?
?

40
M 20
0 1 2
KM

DANNHAUSER

DUNDEE
GLENCOE

0 10 20
KM

Figure 7.12. East-west profile through the Vryheid Formation in vicinity of Dannhauser.

LINE OF SECTION.

LYELL   MORGENSTOND   DALRAY   GOWRIE   CINDERFORD   WINSTONE

DUNDEE-VRYHEID LINE

Figure 7.13. Profile extending from the vicinity of Dundee in the north to Winstone in the southeast.

While progradation was one major aspect of delta sedimentation, delta-lobe switching a result of distributary-channel avulsion was another. That avulsion was a frequently occurring event is reflected by the high proportion of sequences showing distributary-channel sandstones incised into embayment and distal delta-front sediments. It is suggested that, in view of the implied rapid rates of progradation and braided nature of distributary channels, aggradation was an important component of delta-lobe development. Gradient advantages thus induced resulted in the abrupt abandonment of certain delta lobes, perhaps as a result of avulsion during flooding, and sediment being diverted into an adjacent area of open water. It is likely that localised subsidence due to differential compaction of the underlying thick pile of sediments also played a major role in creating gradient advantages and encouraging channel switching.

Further evidence supporting the theory that lobe switching was an important process during coastal progradation is derived from the observation by Coleman (1976) that high rates of sediment accumulation and rapid delta progradation result in lower offshore-slope angles than those of other coastal environments. Continental shelves fronting many modern deltas are purely depositional and play an important role in determining the pattern of delta-lobe switching. Lobe switching is most common where offshore slope, wave power and tidal range are low.

Sequences displaying a complex progradational pattern within prodelta and distal distributary mouth bar sediments reflect avulsion events that occurred in more proximal settings. For example, an upward-coarsening sequence that terminates abruptly within distal mouth bar sediments and is then succeeded by prodelta mudrocks (eg. Fig 5.5 and lower part of Cinderford section, Fig 7.9) indicates delta-lobe progradation, lobe abandonment and progradation of the same or a different lobe in the same vicinity. The relative position of the measured section with respect to the delta lobe when it was abandoned would also give rise to the numerous permutations of sedimentary sequences preserved at any given locality.

Transgressive events are preserved as abrupt vertical facies changes from delta plain environments into deposits of deeper water (embayment and prodelta) settings. The various transgressive associations depicted in Figure 7.14 are believed to reflect the relative proximity of the abandoned delta lobe to prograding deltas; an abandoned lobe isolated from sedimentation is depicted in Figure 7.14A whereas one proximal to an active delta would be subjected to periodic, if minor and short-lived, sandy influxes (Fig 7.14C). The absence of destructive barrier sandstones and coal seams commonly associated with transgressive events (Fisher, 1969; Heward, 1981) is attributed firstly, to the rapidity with which lobes were drowned once abandoned and secondly, to low wave-energies induced by a shallow, low-gradient offshore slope of what was previously a high-constructive coastline. Rates and controls on shoreline migration, offshore water depths and cyclicity are discussed in section 7.2.

On the basis of the above observations it is postulated that shoreline advance was non-uniform and episodic. The shoreline had an undulating pattern; prograding deltas formed prominent lobes extending into the basin, the embayed intervening stretches were occupied by abandoned, drowned delta-lobes and shores dominated by wave processes (see Figure 9.4). It is probable that, owing to the large volume of fresh water flowing into the basin, water salinity along the coast was to some extent lower than that of the Ecca sea as a whole. Furthermore, water salinities along the coast varied considerably both spatially and temporarily and were a function of proximity to sites of inflowing fresh water and seasonal fluctuations in river discharge. The presence of distinctive trace-fossil assemblages, as discussed in Chapter 3, confirms this. Studies from other parts of the basin have shown that microtidal conditions prevailed along the coast (Vos and Hobday, 1977; Tavener-Smith, 1982). There are no significant structures within this lower component suggesting that tides played a major role in sedimentation though barrier-beach deposits exposed on the farm Loch End (Fig 7.1) and developed within the upper component of the Lower zone support a low tidal range.

7.1.2   The Upper Component.

The northwestern portion of the upper component (between Dundee and Alcockspruit) represents deposition almost entirely by coallescing braided distributary-channels on an extensive lower delta plain (Figs 7.11 and 7.12). These deposits contrast strongly with those of the lower component in that they are laterally continuous over a wide area.This represents a phase of prolonged progradation along the western and southwestern coastlines and the coallescing of delta-lobes to form an extensive delta plain dominated by fluvial sedimentation. Areas of the plain abandoned briefly, as at Talana for example (Fig 7.4), were frequently inundated by basinal waters and hosted dense colonies of burrowing organisms.

To the southeast (of the Dundee-Vryheid line) the coast was dominated by lateral-accretion barrier beaches that were periodically breached by rivers (Fig 9.4). Back-barrier lagoons or barred embayments underwent prolonged periods of clay and silt deposition from suspension interrupted by sporadic high-energy influxes of sand introduced by flooding rivers or barrier-beach washovers during storms. Marley et al.(1979) suggested that a similar type of coastline represented in the Blackhawk Formation (Lower Cretaceous), Utah is indicative of a stable, slowly prograding shoreline. The lateral continuity of the upward-coarsening shoreface sequences compared to the underlying sequences supports this observation.

Palaeocurrent distributions do not differ significantly from those of the lower component (see Chapter 6, Fig 6.6) but a discernable southeasterly to easterly mode displayed by most sequences is attributed to longshore currents. That most sediment was supplied by longshore currents is supported by the relatively mature texture of the sandstones in comparison to delta front and fluvial sandstones.

An important aspect of the component is that in this area it contains some

of the earliest formed coals within the Vryheid Formation (Figs 7.7, 7.8 and 7.9). They are rarely more than a few tens of metres across. Some of these coals formed in marshes or swamps on the back slopes of barrier beaches while others accumulated on the landward fringes of barred embayments and lagoons and emergent surfaces within these environments.

The lateral relationship between the northwestern and southeastern parts of the area is depicted graphically in Figure 7.13.

7.2   CYCLICITY AND CONTROLS ON SEDIMENTATION

The concept of cyclic sedimentation has been the subject of controversy for more than a century (Duff et al., 1967), especially with respect to terminology. Many interpretations and definitions of cyclic and rhythmic sedimentation exist within the literature (see Wanless and Weller, 1932; Sander, 1936; Fearnside, 1950; Fiege, 1952; Beerbower, 1961; Duff et al., 1967; Schwarzacher, 1969, 1975). In spite of the varied usage of the terms rhythm, cycle and cyclothem, it is clear that they refer to a sedimentary sequence in which the strata are arranged in a recognisable, non-random pattern related to a series of genetic events. In this thesis the term "cycle" is preferred for such arrangements.
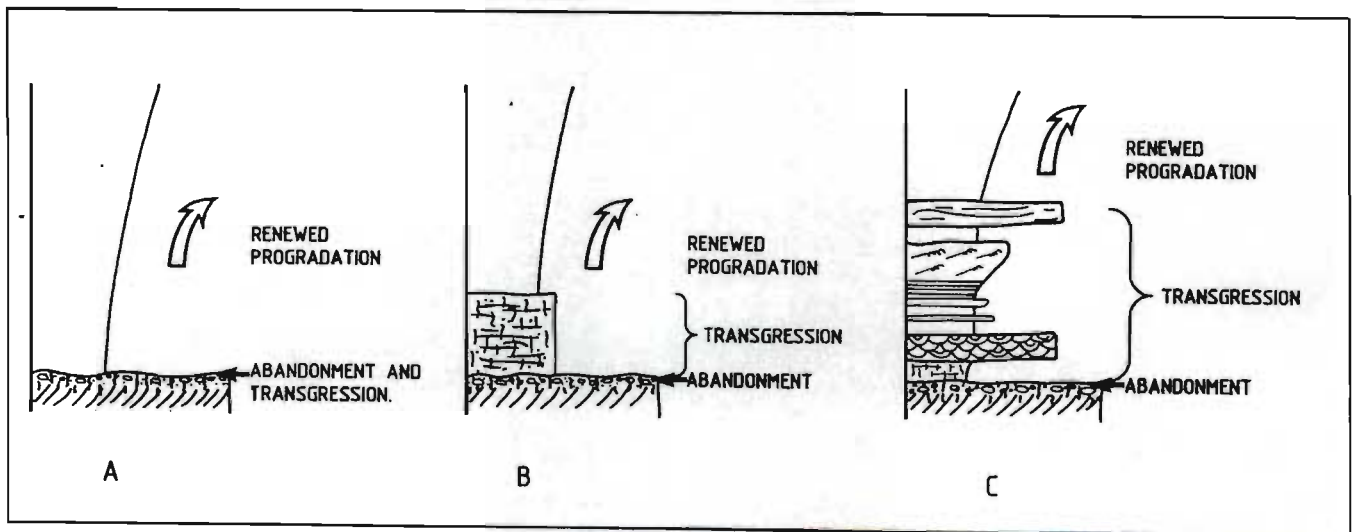
Figure 7.14. Variation in facies associations indicative of delta-lobe abandonment, transgression and renewed progradation.

The most obvious and common cycles related to transgressive-regressive events within the Lower zone coarsen upward and are on average between about 8 and 40 m thick and correspond in scale and genesis to Wanless and Weller's (1932) and Heckel's (1977) cyclothems, Fischer's (1982) megacyclothems and Busch and Rollins' (1984) 5th-order transgressive-regressive units or cyclothemic "punctuated aggradational cycle" sequences. These cycles form the basic framework for understanding mechanisms and controls of sedimentation in the Vryheid Formation over the entire northeastern Karoo basin. Possible controls on cyclic sedimentation include autocyclic (switching of sediment sources due to sedimentary processes) and allocyclic (basin and source-area tectonics) mechanisms and eustatic sea-level variations.

One can model two extreme sequences on the basis of the above mechanisms (cf. Busch and Rollins, 1984). If all the transgressive-regressive cycles are allocyclic, there should be a constant number of synchronously formed sequences at all localities within the basin. On the other hand, if sequences are due to autocyclic mechanisms there should be a variable number of randomly formed units throughout the basin.

Van Vuuren and Cole (1979) regarded the cyclicity displayed by the Vryheid Formation to be the result of repeated rapid subsidences and/or rises in sea-level, followed by delta progradation. Implicit in this mechanism was that cycles could be correlated basin-wide and that transgressive events were synchronous. This proposal was recanted by van Vuuren (1981) who recognised that most upward-coarsening sequences are of limited lateral extent but that the upper boundaries of certain major regressive sequences (commonly comprising two or three deltaic sequences) can be correlated with confidence over large parts of the basin. Eustatic variations in sea level during Vryheid Formation deposition have never been conclusively proven, but it has long been held that seasonal melting of ice after the Dwyka glaciation would have released vast volumes of water into the Karoo basin (Crowell, 1983; Tankard et al., 1982). On a larger time scale, sea-level oscillations in response to freezing and thawing

of the ice cap during a series of waning glacial and interglacial periods during Ecca sedimentation was proposed by Falcon (1980, 1986a) and Crowell (1983).

Although it is clearly evident that autocyclic mechanisms (avulsion related to delta-plain aggradation, differential compaction of sediments, and river flooding) accounted for the cyclicity observed in the Lower zone, it is contended that this pattern was superimposed on allocyclic mechanisms related to eustatic variations in sea level and basin and source-area tectonics. The variation in the pattern of cyclicity between the upper and lower components of the Lower zone reflects the dominant control allocyclic mechanisms had on deposition. In order for the Karoo basin to have accommodated the thick pile of transgressive-regressive cycles there would have had to have been a progressive (and episodic) rise in mean water level or basin-floor subsidence or a combination of both. A clue as to which process was dominant is afforded locally by the lateral transition of single upward-coarsening sequences into multiple sequences towards the west and southeast of the Newcastle-Ladysmith and Dundee-Vryheid lines respectively. Regionally, this pattern is echoed by proximal-distal profiles through the Vryheid Formation in northern Natal (Tavener-Smith, 1979; Tavener-Smith, 1986, see Figure 6.4). Such lateral relationships can only be related to spatial and temporal variations in rates of basin-floor subsidence (Fig 7.15). In addition, the relationship between basinal variations in pre-Karoo, or even post-Dwyka, floor elevations and regional sedimentary patterns do not correspond to a process of basin deepening by eustatic rises in water level alone.

While basinal evolution was a function of continental tectonic architecture, local tectonic features (such as described in Chapter 6) related to faulting and/or gentle warping strongly influenced variations in rates of subsidence during Vryheid Formation deposition. There can be little doubt either that crustal loading induced isostatically by sedimentation contributed to basin subsidence (see Chapter 6). Bloom (1967) indicated that even small loads will
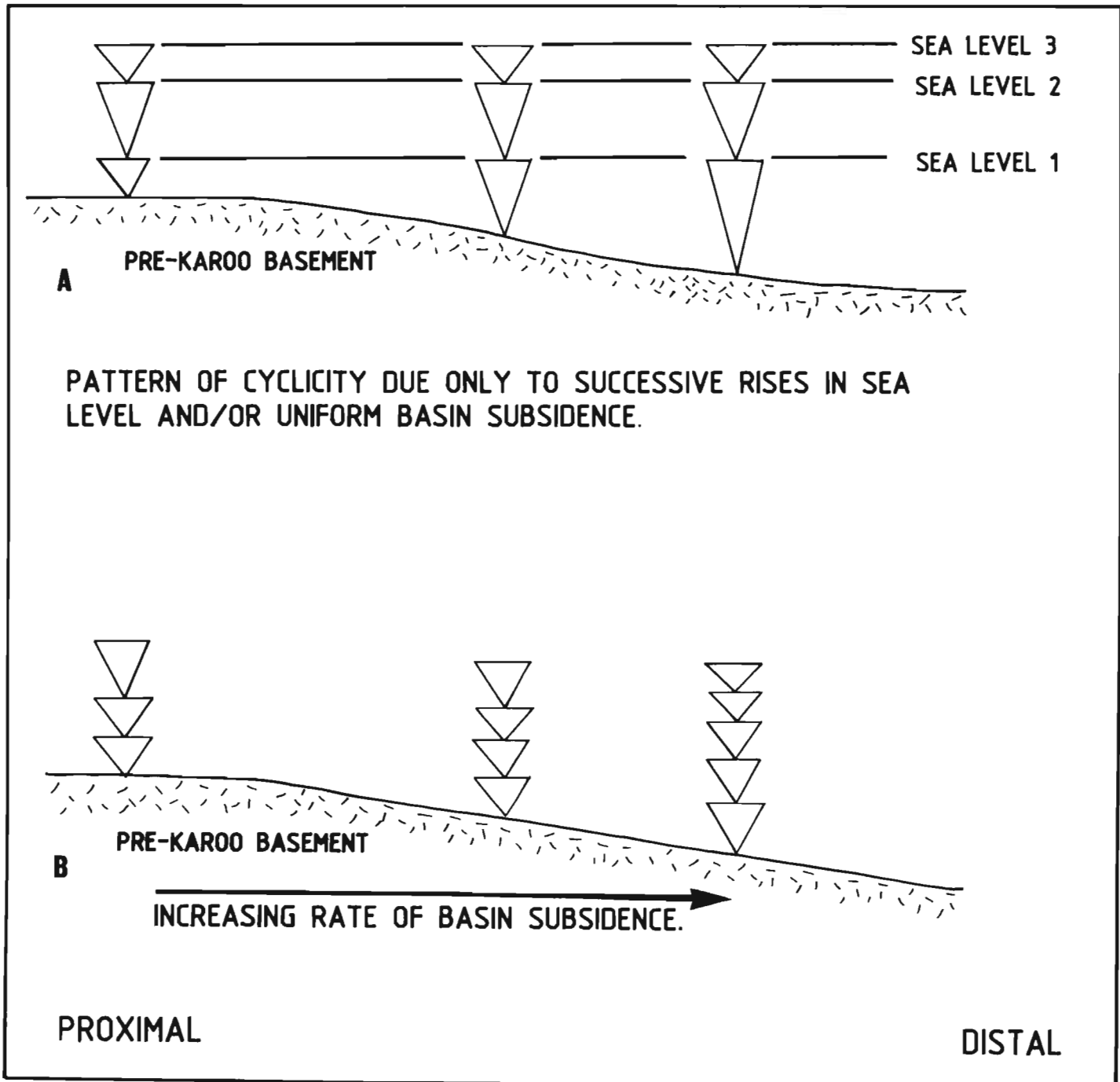
Figure 7.15. Illustration depicting response of cyclicity to A, only eustatic variations in sea level and B, spatial variations in basin or basement subsidence.

be compensated isostatically, so subsidence due to sediment loading was probably fairly continuous though not uniform (variations being due to changes in rate of sediment supply). Differential crustal loading would have amplified spatial variations in rates of basin-floor subsidence that were controlled primarily by tectonic features.

Although in previous discussions it has been assumed that individual upward-coarsening sequences reflect existing water depths, this may not have always been the case. The upward shallowing of depositional environments represented in most sequences indicates only that sea-floor aggradation exceeded relative rises in water level; water depths may indeed have been less than indicated by upward-coarsening sequences. In spite of this, water depths along the coast probably did not exceed 80 m and were on average not more than about 30 m.

Implicit in the pattern of transgressive-regressive cyclicity is that the shoreline oscillated within a fairly narrow limit. It was only towards the end of Lower zone deposition that significant basinward progradation of the shoreline to the northwest of the Dundee-Vryheid line occurred and an extensive and enduring delta plain was established. During the same period the southern and southeastern shore received sediment mainly via longshore currents and very little directly from fluvial sources. This is believed to reflect source-area control in depositional patterns, ie. rates of uplift, erosion and, ultimately, sediment supply varied geographically.

The above suggestions relating to mechanisms controlling patterns of sedimentation have a significant bearing on the creation of environments suitable for the accumulation of thick peat deposits. This aspect is discussed in greater detail in the following chapters, but the controls on establishing such settings are commonly manifested in vertical and lateral facies associations within the Lower zone.
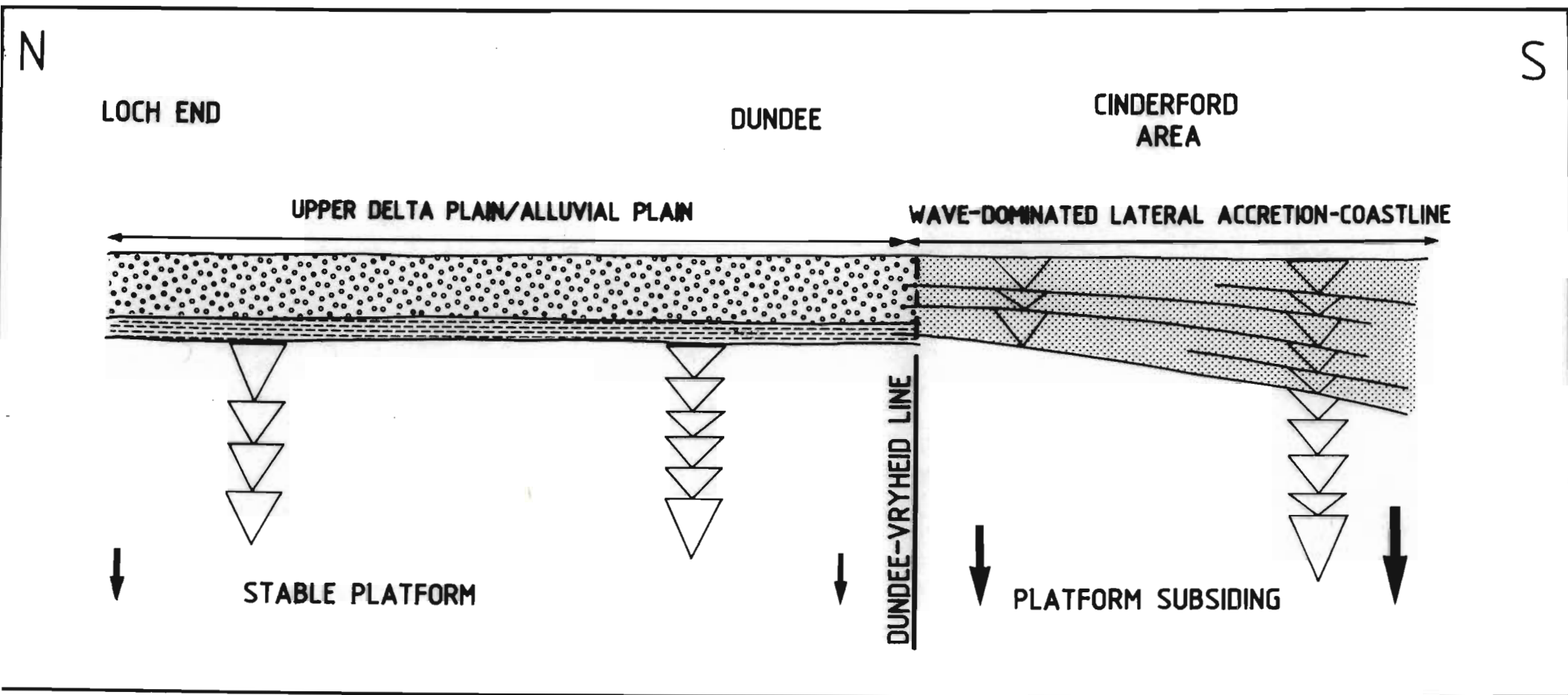
Figure 7.16. Schematic profile through the Lower zone illustrating factors influencing deposition.

7.3   DEPOSITIONAL MODEL

The depositional history of the Lower zone can be summarised as follows (Fig 7.16):

1.   The random pattern of cyclicity displayed by the lower component of the Lower zone reflects repeated shoreline regression and transgression as a result of rapid and frequent autocyclic switching of delta lobes superimposed on a high rate of basin-floor subsidence and possibly also eustatic rises in sea level. The manner in which the position of the shoreline oscillated is indicative of an imbalance between sediment supply (coastal aggradation and progradation) and basin-floor subsidence (transgression).

2.   During the closing stages of Lower zone deposition sediment supply to the western and southwestern coasts was sufficient to offset the decreased rates of basin subsidence and a major phase of shoreline progradation saw the establishment of an extensive delta plain dominated by coalescing braided fluvial-channels.

3.   While the decrease in rate of basin-floor subsidence in the southeast is reflected in the lateral continuity of lateral-accretion shoreline deposits, this pattern of sedimentation suggests a substantially diminished supply of sediment to the area.

4.   The areas to the west and southeast of the Newcastle-Ladysmith (Dannhauser Trough) and Dundee-Vryheid lines were undergoing more rapid rates of subsidence than the central parts (ie. that part of the coalfield in which the major economic coals accumulated). This is reflected by the tendency for upward-coarsening sequences to split and display more complex facies associations in a basinward direction.

CHAPTER 8

PEAT-FORMING ENVIRONMENTS - PHYSICAL AND CHEMICAL CONSIDERATIONS

8.1   INTRODUCTION

The introductory paragraphs to Chapter 5 identified numerous problems and shortcomings associated with palaeoenvironmental interpretations; these are magnified when dealing with coal seams owing to the relatively slow nature of the processes operative on the peat-forming environment. In addition, the geographic and sedimentological relationships which existed between ancient organic and clastic facies is poorly understood. This chapter highlights some of these problems and examines concepts which will form the basis for interpreting the coal and coal-bearing sediments of the Vryheid Formation within the Klip River coalfield.

8.2   TERMINOLOGY

A comparison of the nomenclature of peat-forming environments used by various authors reveals that no standard terminology exists, frequently resulting in confusion when attempting comparisons.

For example, Stach et al. (1982) referred to 4 swamp types which can be distinguished on the basis of plant communities:

1.   Open water areas with water plants,

2.   open reed swamps,

3.   forest swamps, and

4,   moss swamps or bogs.

Falcon (1986a) used a similar terminology but omitted from the coal-forming facies of Gondwanaland the moss swamps or high moor environment (Fig 8.1). Martini and Glooschenko (1985) regarded a "swamp" as a particular type of wetland, and described wetlands as "...areas where the water table is near or

above the mineral soil for most part of the season, supporting hydrophylic vegetation. Pools of water less than 2 m deep may be present". The five classes of wetlands used in Canada are:

1. Bog (highmoor) - peatland raised above the surrounding terrain, nutrients exclusively from precipitation. Vegetation includes moss, lichens and shrubs. May contain trees.

2. Fen - peatland directly affected by, and receiving nutrients from, mineral soil. Vegetation rich and varied: grasses, sedges, herbs, shrubs and trees.

3. Swamp - wooded minerotrophic wetland. Vegetation ranges from coniferous to deciduous trees and also shrubs, herbs and mosses. May or may not produce peat.

4. Marsh - periodically inundated by fresh or saline water. Vegetation: grasses, sedges and sporadic shrubs.

5. Shallow, open waters - shallow (depth less than 2 m), small ponds to wide lakes occurring within bogs, fens and marshes. Some contain barren, acidic waters, others develop algae blooms.

The above classes can be modified by wetland form (surface morphology), type (physiognomy of plant cover) and variety. Fens and swamps are the most important peatlands in Canada, swamps may form peat but marshes form no significant peats.

McCabe (1984) restricted his terminology to floating, low-lying (low moor) and raised (high-moor) swamps, each part of a continuum in the evolution of a peat swamp's morphology (Fig 8.2). Low-lying peat swamps are the type usually

	DRY FOREST	WET FOREST	REED MOOR	OPEN WATER
Water cover	NONE	OSCILLATING	ALMOST COMPLETE	COMPLETE
Acidity	HIGH	MEDIUM	LOW	LOW
Atmospheric O_2	PRESENT	PARTLY PRESENT	LARGELY ABSENT	ABSENT
Chemical reactions	OXIDATION	OXIDATION and REDUCTION	MAINLY REDUCTION	REDUCTION
Organic activity	FUNGI, INSECTS AEROBIC BACTERIA	ACTINOMYCES, AEROBIC and ANAEROBIC BACTERIA	MAINLY ANAEROBIC BACTERIA	ANAEROBIC BACTERIA
Mode of plant decomosition	ROTTING	MAIN ZONE OF PEATIFICATION		PUTREFACTION
Peat types	WOODY WITH RESIN BODIES	WOODYwith inc. HUMIC COLLOIDS	FIBROUS TO EARTHY	ORGANIC MUD
Microlithotype	SEMIFUSITE, FUSITE, DURITE	VITRITE, CLARITE, TRIMACERITE, VITRINERTITE	DURITE N-E, INERTODETRITE, MACROITE (SEMIFUSITE)	TRIMACERITE ε, DURITE ε LIPTITE, CARBOMINERITE
Lithotype	FUSAIN (fibrous coal) some DURAIN (dull coal)	VITRAIN (bright coal), CLARAIN (banded bright coal), DURO-CLARAIN(banded coal)	CLARO-DURAIN (banded dull coal, some DURAIN (dull coal)	DURAIN (dull coal), CANNEL and BOGHEAD COAL, OIL SHALE
Coal type	HUMIC			SAPROPELIC

Figure 8.1. Coal facies in a typical Vryheid Formation peat swamp (after Falcon, 1986a).

referred to in clastic facies models; they are generally slightly acidic (pH-4,8 to 6,5), rich in nutrients, contain a great diversity of plant types and typically have very wet surfaces with a telmatic flora. Raised swamps can only develop in areas where precipitation exceeds evaporation. Their vegetation in modern environments ranges from a herbaceous flora with abundant <u>Sphagnum</u> moss in humid, temperate areas to dense forests comprising a limited number of tree species in the tropics. Their waters are generally highly acidic (pH-3,3 to 4,6) and low in plant nutrients. Raised swamps are only rarely invoked in palaeoenvironmental models (see Smyth, 1980; Flores, 1981).

In their description of coal-forming environments Galloway and Hobday (1983) refer rather informally to open marshes, swamps and raised bogs:

1. Open marshes - contain reeds and other herbaceous plants, no woody material. Saline marshes do not produce peat. Fresh-water marshes have potential to form coal and may, in addition to rooted plants, support floating mats of vegetation. Marshes grade landward into swamps.

2. Fresh-water swamps - contain variable amounts of woody vegetation in forest swamps and drier forests. Have potential for thickest peat accumulations and economically important seams.

3. Raised bogs - regarded as only a locally important peat-forming environment. Vegetation as per McCabe.

These environments reflect the same landward zonation in flora described by Falcon (1986a).
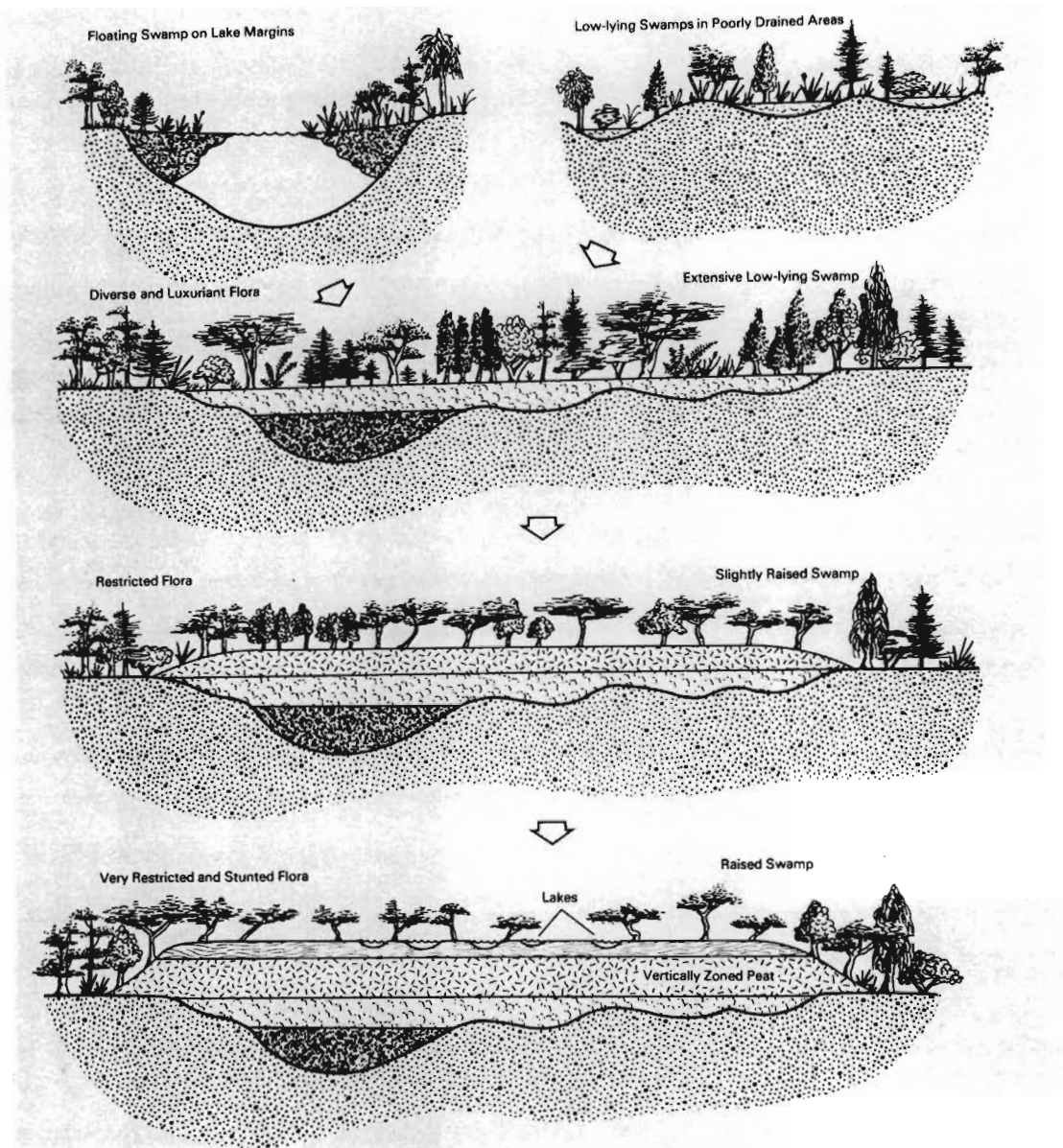
Figure 8.2. Evolution of swamp types showing the development of a raised swamp (after McCabe, 1984).

TABLE 8.1   SUMMARY OF PEAT-FORMING ENVIRONMENTS

STACH ET AL (1982)	OPEN WATER	OPEN REED	FOREST SWAMP		MOSS SWAMP/BOG
FALCON (1986a)	OPEN WATER	REED MOOR	WET FOREST	DRY FOREST	—
MARTINI AND GLOOSCHENKO (1985)	SHALLOW OPEN WATER	MARSH	FEN ? SWAMP		BOG
McCABE (1984)	FLOATING	LOW - LYING SWAMP			RAISED SWAMP
GALLOWAY AND HOBDAY (1983)	? OPEN MARSHES	FRESH-WATER SWAMP			RAISED BOG
WATER COVER	COMPLETE	ALMOST COMPLETE	FLUCTUATING	NONE	NONE (RAISED WATER TABLE)

This brief discourse gives some idea of the varied terminology that has crept into descriptions of peat-forming environments over the last few years (Table 8.1). Terminology used in this thesis will conform to that of McCabe (1984) with modifications. The term "wetland" will be used as a general term to describe all peat-forming environments, these being openwater, marsh (to differentiate from fresh-water wetlands), low-lying swamps and raised swamps.

8.3 THE INFLUENCE OF TECTONIC SETTING, BASEMENT RELIEF AND SUBSIDENCE RATES

One of the most important factors influencing the areal distribution of peat is the nature of the platform upon which it accumulates (Ferm et al., 1979). The degree of tectonic activity and basement palaeotopographic relief influences the number, thickness and distribution (extent and splitting) of coals. The lithology (different sediments exhibit different rates and degrees of compaction) and relief of the underlying sedimentary surfaces constitutes a more local but equally important control. Thus, for the accumulation and preservation of peat deposits to form economic coal seams the above controls must contribute to maintaining a delicate balance between platform subsidence, peat-surface aggradation and a sufficiently high water table. Required is a water table high enough to cover the decomposing vegetation (and prevent oxidative degradation of plant material), but sufficiently low that it does not prevent plant growth (see McLean and Jerzykiewicz, 1978; Stach et al., 1982; Galloway and Hobday, 1983; McCabe, 1984). To illustrate the far-reaching effect of this control, one needs only compare the Permo-Carboniferous Gondwanan and Laurasian coals. Most European and North American coals were deposited in a rapidly-subsiding foredeep basin (Stach et al., 1982) while South African coals accumulated on the stable margin of the Kaapvaal Craton. Although climate (as will be shown) played an important role in determining coal composition, the rapid subsidence to which the Laurasian peats were subjected ensured minimal oxidative degradation and vitrain-rich coals resulted. The low subsidence rates of the Karoo basin peat-forming platforms assisted in producing coals of variable composition with a high proportion of inertinite-group macerals (Falcon, 1986b).

The effects of differential compaction and pronounced pre-Karoo basement relief on coal-seam thickness has been well documented by Le Blanc Smith (1980b) in the Witbank coalfield. Similar pre-Karoo basement-relief effects in the Klip River coalfield have been subdued by the thickness of sediments between the pre-Karoo basement and Coal zone. The influence basin tectonics had on platform subsidence has been discussed in previous chapters. Its effect on coal-seam distribution is discussed in Chapter 9. Briefly, however, coals formed in a stable cratonic environment, such as the northern and northeastern margins of the Karoo basin, may be of variable thickness, but individual seams may cover areas in excess of 10 000 sq km (Galloway and Hobday, 1983). While the coal-bearing horizon within the Vryheid Formation is relatively thin and uniform, differential rates of basin-floor subsidence, especially in areas subjected to pronounced basement downwarping, resulted in local variations in the number and behaviour of seams (Tavener-Smith, 1979; Turner and Whateley, 1983).

8.4   THE INFLUENCE OF CLIMATE AND FLORA ON COAL TYPE

A cold-temperate climate, as experienced during the Permian Vryheid Formation peat-forming period, is characterised by low winter (often below freezing) and moderate to high summer temperatures. This is confirmed by fossil trees displaying marked seasonal growth rings and layers of leaves, presumably shed during autumn (Plumstead, 1957). Although there is no indication as to rainfall patterns (despite Krausel's (1961) assertion that alternating wet and dry seasons existed), peat-forming environments in similar modern climatic zones (along Hudson Bay, Canada and Siberia) receive between 400 and 500 mm pa. Falcon (1986a) estimated a mean annual rainfall of 1 000 mm during Vryheid Formation sedimentation. The Glossopteris-Gangamopteris flora, which was more diverse than those of the Dwyka and early Ecca, comprised abundant arborescent and herbaceous Glossopteridophyta and minor populations of lycopods, ferns, cordaites and early gymnosperms (Falcon, 1986a). In modern-day equivalents, temperate peat bogs and lakes are colonised by dense, broad-leaved, deciduous

woodland floras which are characterised by different layers of plant communities: trees, shrubs, herbaceous plants and mosses and lichens.

The significance of the above data is that the parental vegetal components of peat swamps are influenced to some extent by the type of decaying humic matter provided by differing floral communities. Climate, on the other hand (by way of variations in temperature, humidity, evaporation and precipitation), controls flora species and densities and affects the rate and type of plant decay (Stach et al., 1982; Falcon, 1986a). At present, major peat-accumulating environments exist mostly in high-latitude cool climates (McCabe, 1984; Martini and Glooschenko, 1985). This is in spite of a growing season much shorter and cooler than the hot, wet tropical climates where plant decay is high due to high evapo-transpiration rates (McCabe, 1984). Vitrinite and inertinite, the most important constituents of humic coals, are derived from essentially the same plant substances, namely lignin and cellulose of cell walls. Degradation of vegetal matter in cold, relatively dry environments, where water-table level fluctuates and plant growth is seasonal (where conditions for biogenic and oxidative vegetal degradation exist periodically, such as is thought to have existed during Vryheid Formation peat accumulation), is thought to produce coals of variable composition rich in inertinite-group macerals and generally high in mineral-matter content (Fig 8.3) (Adams, 1960; Stach et al., 1982; Falcon, 1986a). Northern Hemisphere (Laurasian) coals formed in hot and humid equatorial swamps from rapidly growing plants in rapidly subsiding basins. Consequently, rapid plant degradation under predominantly reducing conditions produced vitrinite-rich coals with a low mineral matter content. Although these physical processes have the strongest influence on maceral type, plant type also plays a role (Smyth, 1980; Stach et al., 1982). Reed peats, which are poor in lignin and undergo intense structural decomposition, give rise to durain-rich coals with the dominant vitrinite being desmocollinite. Forest swamps, ie. wood-rich peats with high lignin contents, produce vitrain-rich coals with vitrites and exinite-poor clarites.

Sapropelic coals (cannel coal and torbanite) accumulate as subaquatic muds under reducing, anaerobic conditions in an open-water environment (ponds and lagoons) (Hutton et al., 1980; Stach et al., 1982;). Cannel coals consist of the remnants of degraded vascular-plant material (including pollens, spores, resins and leaf cuticles) and fine mineral-matter which has been washed or blown into open water. Torbanites may also comprise some of the above macerals, but are generally dominated by algae equivalent to the extant Botryococcus braunii.

8.5   CHEMISTRY OF THE PEAT ENVIRONMENT

Coal-forming environments of the Vryheid Formation were essentially paralic in nature and locally subjected to marine- or brackish-water inundations during peat accumulation. This had a profound effect on prevailing pH, Eh and salinity levels and resulted in a variety of geochemical subenvironments (Falcon, 1986a). In addition to the contraints placed on peat-accumulation by water-table levels, water pH is critical. Acidity controls decomposition of plant remains by bacteria (Stach et al., 1982); microbial degradation is reduced where pH of peat waters is less than 4,5. Generally, low-lying swamps have a water pH of between 4,8 and 6,5 and raised-swamp waters vary from 3,3 to 4,6. The influx of marine or any other alkaline waters raises pH and results in severe structural decomposition of plant material. For this reason peat accumulation in environments directly influenced by marine waters (such as mangrove swamps) is uncommon.

Redox potential (Eh) plays an important role in peat formation and is a function of available oxygen within the peat waters (Stach et al., 1982). Aerobic conditions within a peat environment created, for instance, by the inundation of the wetland environment by aerated water, or a lowering of the water table, result in the disintegration of plant matter or the production of inertinite-group macerals. Under low Eh, anaerobic (reducing) conditions bacterial activity is reduced and conditions conducive to the formation of vitrinite-group macerals (vitrains) prevail.

PHASE 1   DURAIN AND FUSAIN FORMATION

AEROBIC DECOMPOSITION

LOW WATER TABLE

PHASE 2   BANDED DURAIN/ CLARAIN / VITRAIN FORMATION

pH FLUCTUATING
AEROBIC / SUBAEROBIC DECOMPOSITION

Direction of water flow

RISING WATER TABLE

PHASE 3   VITRAIN AND CLARAIN FORMATION

pH LOW
ANAEROBIC DECOMPOSITION

FLOODED SWAMP

PHASE 4   COAL SEAM COMPOSITION AFTER PEATIFICATION

P
H   3
A   2
S   1
E
S

Coal
seam

DURAIN
FUSAIN

INTERMEDIATE ASSOCIATION
BANDED DURAIN/CLARAIN

CLARAIN / VITRAIN
(possibly sapropelic clay rich coals above)

Figure 8.3. Variations in coal composition as a function of
fluctuating water level (Falcon, 1986a).

Sulphur is an important component of paralic coals and forms where there is a sufficient supply of iron and sulphur. Sulphur is introduced as a sulphate from saline waters or may also be of organic (plant) origin (Stach et al., 1982). The formation of FeS_2 (pyrite or marcasite) requires reducing conditions and a catalyst in the form of bacteria; Neavel (1966) has shown that there is insufficient energy for a chemical reaction of sulphates to disulphides. Williams and Keith (1963) and Mansfield and Spackman (1965) provided convincing evidence that much of the sulphur in pyrite-rich coals is introduced by the inundation of peat swamps by marine waters. Such coals generally display concentrations of pyrite at their tops and are overlain by marine rocks. Low sulphur (pyrite) coals are generally overlain by continental deposits. Horne et al. (1978) regarded the environments of deposition of the sediments immediately overlying the coal as having a greater bearing on sulphur concentration than the environments of deposition of the peat and the sediment on which it accumulated. D.L. Roberts (pers. comm.) has shown that vitrains are richer in sulphides than fusains and durains in Vryheid Formation coal seams owing to the affinity of both vitrinite and pyrite formation to a reducing, anaerobic environment.

8.6   THE ORIGIN OF MINERAL MATTER IN COALS

Mineral matter (ash) in coal is either of detrital (air- and water-transported), plant-derived or authigenic origin. Plant-derived inorganic material was suggested by Renton and Cecil (1979) to make up the majority of non-epigenetic minerals in coal by the total degradation of peat material (to form, for example, carbonaceous shale partings). Authigenic minerals are introduced into a peat during or after its deposition or during the coalification process (McCabe, 1984). Typical minerals include carbonates, silicates and pyrite.

8.7 PEAT ACCUMULATION AND COMPACTION RATES

Numerous estimates of modern and ancient peat accumulation rates and compaction ratios have been provided in the literature (see Ryer and Langer, 1980; Stach et al., 1982; Galloway and Hobday, 1983; McCabe, 1984; Martini and Glooschenko, 1985). From the foregoing discussions it is obvious that any estimate of a peat-accumulation rate is a function of a number of interrelated parameters, but particularly climate. In addition, peat compaction ratios are not constant and are dependent on peat composition, mineral-matter content and number and thickness of clastic partings (McCabe, 1984).

Rates of peat accumulation of between 0,1 and 2,3 mm yr^{-1} were tabulated by McCabe (1984) while in Canada, Martini and Glooschenko (1985) gave average rates of accumulation of 0,1 to 0,2 mm yr^{-1} in arctic regions, 0,5 to 0,7 mm yr^{-1} in subarctic-boreal areas and greater than 1 mm yr^{-1} in cool-temperate, wet regions. Galloway and Hobday (1983) stated that peat accumulation is four times more rapid in tropical forests than in swamps with lower rainfall. Ryer and Langer (1980) reviewed a number of estimates of peat-to-coal compaction ratios. These varied from 1,4:1 to 30:1, but the average is between about 3:1 and 10:1.

The fact that peat undergoes a significantly greater degree of compaction than clastic sediments and is especially prone to autocompaction (eg. Mallet, 1986) provides a basis for understanding its role in relation to clastic depositional systems. It is obvious that only by decompacting coal seams to their original (peat) thickness can pre-, syn- and post-peat accumulation depositional processes be fully understood.

8.8 PEAT ACCUMULATION IN RELATION TO CLASTIC DEPOSITIONAL MODELS

The concept of coal-seam depositional modelling, ie. the erection of a model that ascribes variations in coal properties and seam distribution to changes in

the depositional environment of associated clastic facies, was evolved and brought to the fore by, amongst others, Dapples and Hopkins (1969), Ferm (1970) and Wanless et al., (1970). Although the concept was germinated in the 1950s and 1960s (Young, 1955; Fisk 1960; Fisher, 1964, 1968) and was a logical follow-on of the cyclothem era (involving mainly the recognition and interpretation of cyclic sequences in coal-bearing strata - Rahmani and Flores, 1984) its expansion and dominance in the field of sedimentology during the early 1970s can be attributed to the world energy crisis and an upsurge of research into modern clastic and peat depositional processes and environments. Perhaps one of the most significant publications dealing with this concept was that of Ferm et al. (1979) who described Carboniferous depositional environments in the Appalachian region of the USA. Like many other investigators, they emphasised depositional environment of associated clastic facies as being the primary control on coal-seam quality, geometry and roof-rock conditions (eg. see Ryer, 1981; Flores 1981; Gersib and McCabe, 1981; Galloway and Hobday, 1983; Flores et al., 1984; Levey, 1985, Fielding, 1985b).

McCabe (1984) has questioned the validity of using depositional models to predict or account for certain characteristics of coal seams. He further contended that the accumulation of economic coals cannot be reconciled with the view that active clastic sedimentation occurs in close vicinity to peat-forming swamps. Fielding (1987) proposed that existing coal depositional models have only limited predictive capability; he regarded subsidence regime (compactional and tectonic) and, to a lesser degree, sediment supply as the dominant controls on coal distribution rather than environment of accumulation.

A comparison of a few of the more recent investigations into depositional modelling of coal seams (Table 8.2) shows that the relationship between implied depositional settings of coals and their geometry is in fact highly variable. Clearly, the models established by Horne et al. (1978) are not universal and other factors are required to explain coal-seam geometry and behaviour. Furthermore, few of the current models give any indication as to the extent depositional environment influences the mineral-matter content of peats.

McCabe's (1984) central argument - organic versus clastic accumulation - arises from an examination of modern peat-forming environments associated with active clastic depositional systems, especially deltaic and other coastal environments. Because peat accumulates slowly and continuously over many thousands of years (if it is to form an economic coal seam) and has a relatively high compaction ratio during coalification compared to clastic sediments, even rare influxes of sediment into a peat swamp are significant with respect to the composition of the coal ultimately produced. Even at peat accumulation rates of 1 mm yr^{-1}, any clastic sediment introduced into a swamp by, for example, a 10 or 100 year flood-event can be regarded as a normal part of sediment supply. Such events may result in the deposition of a few centimeters or tens of centimeters of sand, silt or clay over a period of a few hours or days, thus upsetting the entire equilibrium of the ecosystem. In the opinion of McCabe, most sediment introduced, however, is incorporated into the peat matrix rather than forming a discrete parting. The influx of fresh or saline water with relatively high pH and Eh values also results in varying degrees of biogenic and/or oxidative peat degradation. Peats accumulating in close association to active clastic-sediment environments (such as the Mississippi River Delta, Snuggedy Swamp, and Dismal Swamp) were shown by McCabe to have mineral matter contents so high that it could be assumed carbonaceous shale would be the ultimate product. A strong relationship exists between clastic sediment influx and style or morphology of autochtonous peat environments: floating and raised swamps produce low mineral-matter peats, low-lying swamps have high mineral-matter contents unless isolated from areas of clastic deposition.

While conceding that low-ash coals may originate from high-ash peats depleted in mineral matter during peatification and coalification, McCabe listed 3 factors that contributed to the formation of low mineral-matter peats:

TABLE 8.2   COMPARISONS OF IMPLIED DEPOSITIONAL ENVIRONMENT AND COAL SEAM GEOMETRY

Depositional Modeller/s, Formation, Age	IMPLIED ENVIRONMENT OF PEAT ACCUMULATION AND COAL-SEAM GEOMETRY				
	Back Barrier	Lower Delta Plain	Upper Delta Plain	Alluvial Plain	Abandoned/Transgressive Plain
Horne et al. (1978) Appalachian Basin, Carbon	Thin, LD	Thin, LE	Thick, LE	Locally Thick, LD	—
Levy (1985) Green River Basin, Upper Cret.	—	Thick, LE		Locally Thick, LV	Thin, LD
Ryer (1981) Ferron Sandstone, Cret.	—		Thick, LD	—	—
Cavaroc and Flores (1984) Crevasse Canyon Fm., Upper Cret.	Thin, LD	—	—	Thin, LD	Thin-Thick, LE
Merritt and McGee (1986) Chignik Formation, Cret.	—	—	—	Thick, LD	—
Carlson (1979)	Thick, LD	—	—	—	—
Turkowski and Rahmani (1984) Horseshoe Canyon, Upper Cret.	—	—	—	Thick, LD	—
Clemmensen and Surlyk (1976) ?, Jurassic	Thick, LD	—	—	—	—

KEY : Thin - Coal seams < 1M

Thick - Coal seams > 1M

LD   - Coal seams laterally discontinuous, lenses usually parallel to channels or beach barriers

LE   - Coal seams laterally extensive

1. Chemistry of swamp waters,

2. raised or floating swamps, and

3. peat accumulation temporally distinct from local clastic sedimentation.

Of the 3, the last factor was regarded as the most important, especially as most economic coals (those with favourable thickness and ash parameters) described in the literature are attributed to formation in low-lying swamps.

The first factor cannot be totally ignored, especially in the light of the mechanism proposed by Staub and Cohen (1979). They maintained that the acidic waters of a swamp rapidly flocculate the clay fraction from waters entering the swamp. Peats away from the swamp margin are relatively low in ash. Support to such a process is provided by the observation that coals purported to have accumulated in an upper delta/fluvial plain setting in association with active fluvial channel sedimentation are elongate parallel to channel trend and adjacent to channels are thin, show multiple splits interfingering with clastic sediments, and contain a high proportion of mineral matter (Horne et al., 1978; Howell and Ferm, 1980; Belt et al., 1984; Fielding, 1985a,b; Nelson et al., 1985). Obviously, there would have to be a balance between peat-surface and channel-area aggradation. In contrast, in a description of coal-forming environments of the Gallup Coal Field (Upper Cretaceous), Cavaroc and Flores (1984) described two alluvial plain environments, one characterised by active fluvial channels and the other partially abandoned. In the former, persistent, poorly-drained reducing environments in which substantial thickness of peat could accumulate were continually flooded with detritus from adjacent channels. In addition, a low water table between river flood stages resulted in oxidation of much of the organic material. The alluvial plain partially abandoned by major channel activity contained peat swamps which were subjected to much reduced influxes of detrital sediment. Even limited fluvial activity, however, produced a series of discontinuous peat swamps.

Implicit in point 3 above is that many coals are formed in an environment distinctly different (ecologically as well as temporally) from that indicated by the underlying and overlying sediments. It follows that a coal cannot always be classified on the basis of the environmental setting of the underlying sediments. It is suggested that the control the clastic depositional environment preceding the period of peat accumulation provided was by the morphological, topographical and lithological expression of its depositional processes. It is well documented that topography exerts a strong influence on peat thickness and geometry (Ferm et al., 1979; Staub and Cohen, 1979; Flores, 1981; Le Blanc Smith, 1980b) and it is intuitive that lithology (and vertical variations therein) of the sediments underlying a peat swamp will have a significant control on water-table levels and thus peat accumulation rates.

Fielding (1984, 1985b) also recognised that areally extensive and thick allochthonous peats form in abandoned areas of delta plains unaffected by contemporaneous channel activity. Fielding (1987) placed more emphasis on the role of platform subsidence during peat accumulation on alluvial and delta plains. Coal-seam splits record the interruption of peat accumulation by clastic sedimentation and reflect localites of accelerated subsidence. Splits may be caused by:

1. Different compaction rates of peat, clay and sand,

2. syndepositional faulting within the sediment,

3. fluvial sedimentation contemporaneous with peat accumulation (eg. channel position controlled by enhanced subsidence), and

4. basinward increase in epeirogenic subsidence rate.

TABLE 8.3   CONTROLS ON COAL FORMATION, GEOMETRY AND COMPOSITION

RANK OF CONTROL	GLOBAL	LOCAL	
FUNCTION OF CONTROL	Ability of Environment Accommodate Peat Facies	Geometry (Thickness and Distribution of Coal Seam	Composition
T Y P E O F C O N T R O L	1. Climate. 2. Subsidence (Tectonic and Compactional) Regime.	1. Subsidence - rates and spatial variations 2. Presence or absence of contemporaneous clastic sedimentation 3. Topography and morphology of peat-forming surface 4. Lithology of sediments underlying peat environment. 5. Type of swamp - raised or low-lying 6. Duration of peat accumulation	1. Climate 2. Vegetation 3. Water-table fluctuations 4. Salinity, Eh, pH of swamp water 5. Proximity to active clastic sedimentation 6. Type of swamp - raised or low-lying NB Most of above factors are interdependent.

8.9   SUMMARY

It is evident that coal seam behaviour and composition are not only a function of the depositional environment of contemporaneous clastic sedimentation. A complex interrelationship exists between a number of parameters, both in a global and local sense. An attempt has been made to identify and create a hierachical scheme of controls on coal-seam characteristics (Table 8.3). It is, however, recognised that each of these controls affects the evolution of any coal seam in a number of different ways at more than one level. It is only by appreciating the manner in which each parameter influences coal-seam geometry and thickness, and applying these to the coal seams of the Vryheid Formation Coal zone, that meaningful and widely applicable models for coal-seam formation may be constructed.

CHAPTER 9

THE COAL ZONE

The Coal zone comprises a well-defined stratigraphic unit varying in thickness from 35 to 65 m. It is subdivided into 5 stratal (clastic) increments, each separated by a laterally persistent coal seam. In general, there is an inverse relationship between Coal zone thickness and aggregate coal-seam thickness. This is particularly evident to the west and southeast of the Newcastle-Ladysmith and Dundee-Vryheid lines respectively. In this chapter, reference to the "main" coalfield describes that area of the southern Klip River coalfield enclosed by the Newcastle-Ladysmith and Dundee-Vryheid lines, ie. that part of the coalfield containing the economically exploitable coals.

The main objective of this study was to relate coal-seam geometry and composition to the environment of peat accumulation of the respective seams. The genetic relationship so established should have formed the basis of a predictive model with application in exploration strategy and mine planning. However, as elucidated in the previous chapter, this relationship is not always evident and it will be shown that other factors, such as spatial and temporal variations in platform stability and source-area mechanisms, exerted a strong control on coal-seam geometry. Of secondary importance, but essential to understanding and defining a stratigraphic and genetic framework of the Coal zone, was the correlation of coal seams throughout the thesis area and establishment of a uniform seam nomenclature.

The large number of boreholes which penetrated the Coal zone permitted the compilation of a number of isopach maps detailing the geometry of specific horizons. This technique was utilised to compare and relate coal-seam geometry to the lithology and geometry of adjacent clastic units. However, as most boreholes penetrated to only a few metres below the No 2B Seam, little thickness and lithofacies data was available for the No 1 and 2A Seams and associated strata.

No proximate or ultimate coal analyses were utilised (as they were not made available by the companies approached for data) and the scope of this study did not encompass petrographic analysis of coal. Only the macroscopic composition of individual coal seams is described and brief reference is made to sulphur content and distribution at the end of this chapter. In order to protect company confidentiality, no specific borehole or cross-section localities are stated.

9.1   COAL-SEAM NOMENCLATURE AND CORRELATION

On examining borehole logs during the early stages of this project it soon became apparent that the existing coal-seam nomenclature within the Klip River coalfield was in disarray and lacking uniformity. This can be attributed to early mining pioneers naming coal seams in vertically descending order - Top, Middle, Bottom, etc - and also to coal seams being named without reference to the existing nomenclature or their lateral distribution. It was therefore decided to number each coal numerically in ascending order and identify splits by appending the letters "A", "B" and "C". The nomenclature proposed here and how it corresponds to the traditional usage is shown in Figure 9.1.

Correlation of the respective coal seams depended to a large extent on the close spacing of boreholes on individual coal-mine properties. However, once the stratigraphic and sedimentological framework of the Coal zone was established and the relationship between coal-seam behaviour and pre-Karoo basement structure understood, seam correlation in areas with sparse borehole information could be carried out with a high degree of confidence. Suggested coal-seam correlation within the coal field is presented schematically in Figure 9.2. A striking feature of all coal seams is that they are laterally persistent and most cover areas in excess of 1 500 sq km.

An important aspect of seam correlation and understanding seam behaviour is that firstly, the No 2 and 3 Seams are amalgamated (to form the "Main" Seam) in
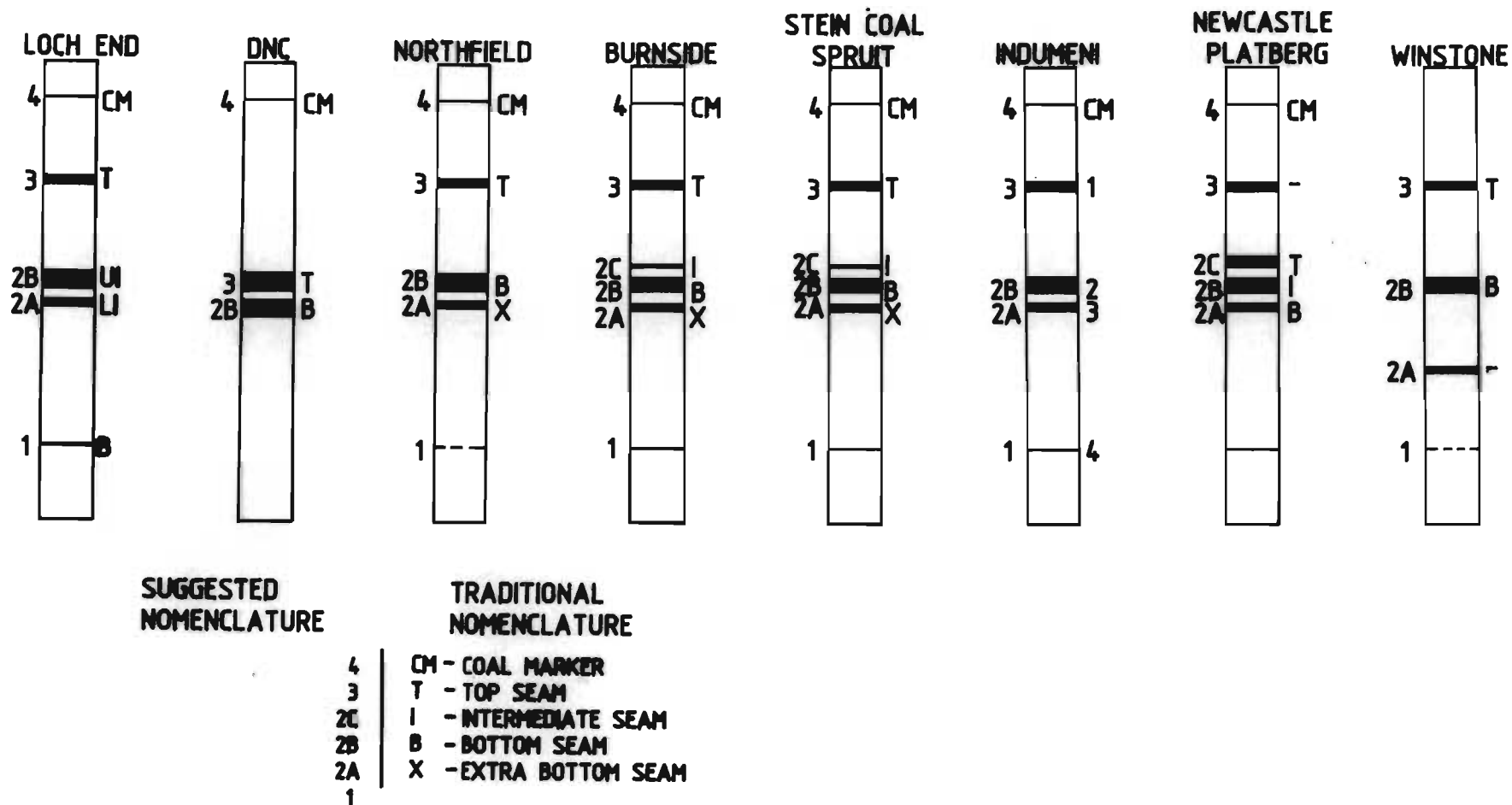
Figure 9.1. Coal seam nomenclature, southern Klip River coalfield, comparing traditional nomenclature with that used in this thesis.
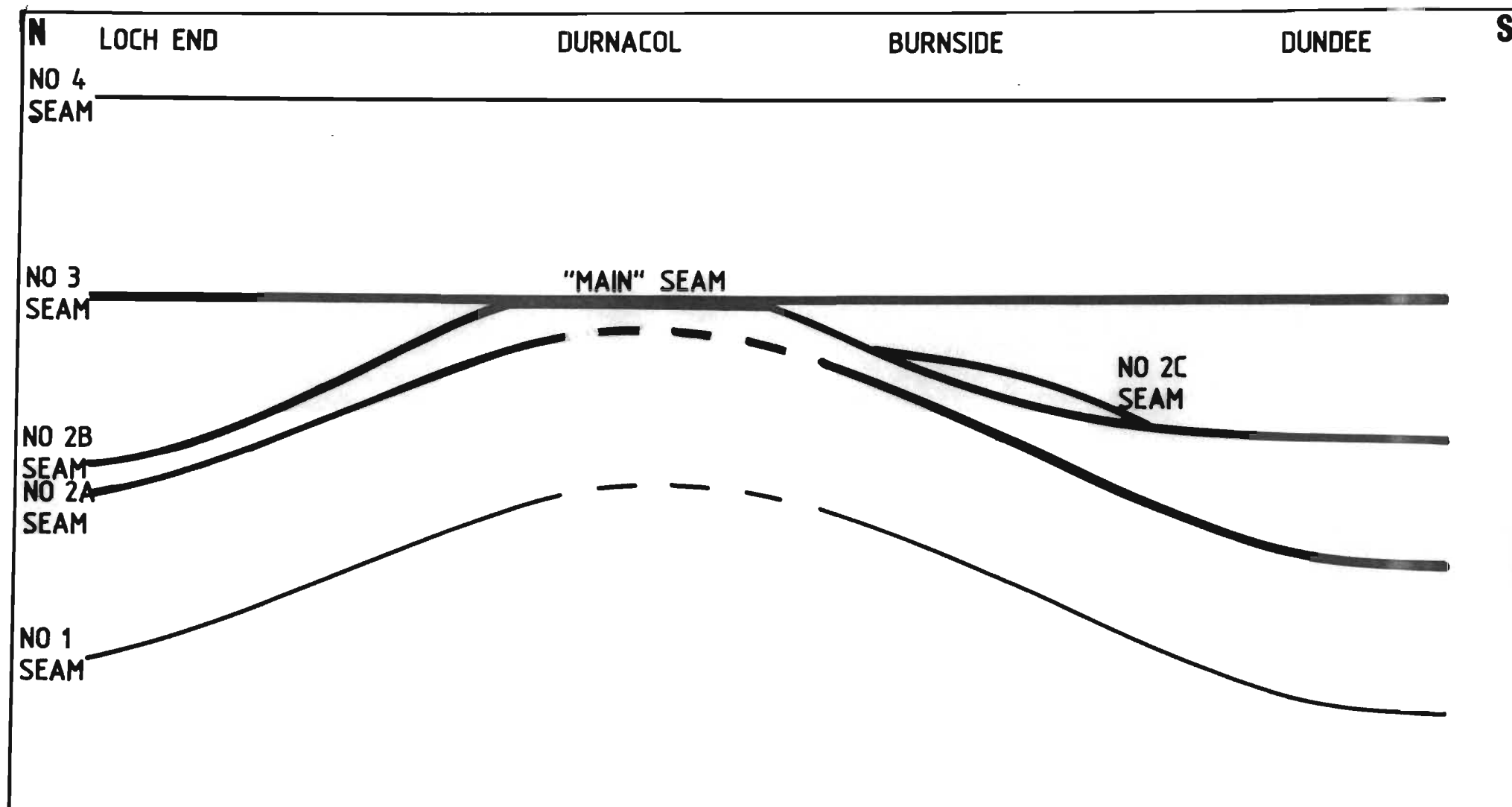
Figure 9.2. Schematic north-south section illustrating distribution and nomenclature of coal seams in southern Klip River coalfield.

a northeast-southwest trending zone located approximately between Glencoe and Dannhauser and secondly, the Dundee-Vryheid and Newcastle-Ladysmith lines delineate to the north and east respectively the stable, main coal-bearing area of the Klip River coalfield.

9.2   THE NO 1 SEAM

9.2.1   Description and Distribution

This seam consititutes the earliest recognisable coal horizon within this part of the coalfield. Although the seam, except to the north and northeast of Dannhauser, is rarely penetrated by boreholes, its distribution is moderately well understood. The seam attains its best development to the north and northeast of Durnacol village where it can be traced with confidence over a distance of about 20 km. Immediately north of Dannhauser it comprises a single coal up to 0,60 m thick, but splits into 2 seams along an east-west orientated line towards the north. Each is up to 400 mm thick separated by between 0,50 and 1,80 m facies Sr or St.

Although borehole data is sparse between Dannhauser and Dundee the seam appears to be developed only sporadically. However, in the vicinity of Dundee and Wasbank and immediately to the southeast of these towns the No 1 Seam is developed as a zone of thin, discontinuous, dull and shaly coals, each up to 350 mm thick, but rarely exceeding 150 mm. The seam is associated with highly bioturbated (Siphonichnus (S) association) sediments (Figs 7.2, 7.3 and 7.13). This situation persists towards the southeast until the approximate position of the Dundee-Vryheid line is intersected, beyond which the seam horizon cannot be traced with any certainty. At approximately the same horizon in the Cinderford vicinity (Fig 7.13) a number of thin coals (each less than 100 mm thick) are developed over an interval of up to 25 m. The seam could not be identified west of the Newcastle-Ladysmith line within the Dannhauser Trough.

9.2.2   Environmental Association and Controls on Accumulation

The No 1 Seam provides an excellent illustration of the relationship between coal seam distribution, environmental association and platform stability. The presence of the No 1 Seam signals the existence of an environmental milieu in which subsidence (probably predominantly tectonic) of the sedimentary platform was sufficiently retarded for the first time over a widespread area to allow the establishment of peat-producing wetlands. To the north and northeast of Durnacol village these wetlands existed on an extensive delta plain devoid of clastic sedimentation for much of the time. The northward split of the seam is attributed to a greater rate of platform subsidence and subsequent influx of fine sand. This split may correspond chronologically to the transgression that terminated peat accumulation in the southern part of the area (described later), renewed peat accumulation reflecting local platform stability. Alternatively, the split may have originated as an overbank sheet-flood from a river system situated to the north of Alcockspruit.

The complex and discontinuous nature of the coal seam in the vicinity of Dundee and Wasbank and southeast of the Dundee-Vryheid line reflects peat accumulation in shallow embayments, back-barrier lagoons (or barred embayments), abandoned channels and shallow depression on a lower delta plain. The swamps and marshes were frequently inundated and drowned, hence the lenticular nature of the coals. The predominantly dull, shaly nature of the coals, attributed to high pH and Eh of the basinal waters and possibly a reed-dominated parent vegetation, and associated intense bioturbation supports a coastal setting. The greater number of coals and associated clastic partings (vertically and laterally) than within the main coalfield is considered to reflect peat accumulation in association with active clastic sedimentation and a lesser degree of platform stability.

Poor and sporadic development of the seam between Dundee and Dannhauser (over what was later to prove the main peat-forming area of this part of the coalfield) can be attributed to non-development of the peat facies due to either:

1.   A rapid rate of platform subsidence,

2.   a low rate of platform subsidence relative to adjacent areas (resulting in a slight elevation of the platform and consequently lower water table), or

3.   continued clastic deposition.

Alternatively, but less likely (considering the seam's sporadic development), the peat may have been removed by subsequent fluvial erosion. In view of the relative stability of the platform over this area with respect to later peat-forming episodes, the second of the above 3 alternatives appears most likely.

9.3   THE NO 1/2A CLASTIC PARTING

Irrespective of the nature of sedimentary facies contained within this interval, its thickness over the central part of the coalfield is between 18 and 30 m and mostly in the order of 20 m. Sedimentary associations indicate that No 1 Seam peat accumulation was terminated by a combination of platform subsidence and the influx of fluvially-derived sand (Figs 7.11 and 7.12). Although there is no physical evidence of a transgression, abundant bioturbation suggests deposition on an extensive coastal plain fronted by wide, shallow embayments. This environment supported a number of minor swamps and marshes but these were generally short-lived.

In the northern part of the coalfield the No 1 Seam is directly overlain by a braided-river sheet sandstone which can be traced as far south as Dundee and Wasbank where it overlies lower delta plain deposits described above. This was

a period of major shoreline progradation, with rivers debouching into the basin to the west and south of the coalfield. The upper 0,50 to 1 m of this sheet sandstone is commonly bioturbated (Siphonichnus (S) association) signifying the re-establishment of a brackish-water environment following the abandonment of the fluvial system. This suggests that these rivers were not a great distance from the coast, and probably in a lower/upper delta plain setting.

To the south and east of the Dundee-Vryheid line enhanced platform subsidence is reflected by the continued presence of offshore (shoreface) conditions and repeated transgressions (Fig 7.13). Minimal sediment supply (particularly sand) from rivers to this sector resulted in extensive stretches of wave-dominated shorelines and, by longshore drift, the deposition of lateral-accretion coastlines fronted by low beach barriers.

9.4   THE NO 2A SEAM

9.4.1   Description and Distribution

The No 2A Seam signals the start of the main peat-forming phase within this part of the coalfield. While it is by no means an economic coal, it is used extensively in coal exploration as a bottom marker to the economic coal seams. The seam rarely exceeds 350 mm in thickness and is more commonly between 50 and 200 mm. Its composition over the field is varied, but comprises predominantly dull coal (fusain) with bright coal bands and laminae to mixed bright and dull coal. In many borehole intersections the basal part of the seam is described as shaly, dull coal. Typically, the seam comprises numerous clastic partings in the vicinity of and to the southeast of the Dundee-Vryheid line. These clastic partings are commonly bioturbated (Siphonichnus (S) association) and vary from silty, fine- to very coarse-grained sandstone (Fig 7.3).

It was originally thought that the No 2A Seam was a split of the No 2B Seam

(Christie 1983, 1985) and that it was represented at Durnacol by what was termed locally the No 1 coal at the base of the "Main" Seam (Fig 9.3). However, as will be shown, this is not so. It appears that the No 2A Seam is a discrete coal seam and distinct from the No 2B Seam. The seam is well developed to the north and northeast of Dannhauser and in places attains a thickness of 0,60 m. The northerly and northeasterly extent of the seam is not known, but in the central sector of the main coalfield the seam is only sporadically developed. Where it is present, it is represented by a carbonaceous shale with rare coal laminae. This area of non-development coincides almost exactly with the area where the No 1 Seam is only sporadically developed. To the south and southeast of Burnside colliery the seam is extensively developed. Southwest of Burnside the seam steadily increases in thickness and quality from a thin (100 to 200 mm) dull, shaly coal associated with numerous clastic partings to a mainly mixed bright and dull coal with thicknesses ranging on average from 0,20 to 0,60 m in the vicinity of Elandslaagte. To the southeast of the Dundee-Vryheid line the seam contains numerous bioturbated clastic intercalations and may be developed over an interval of up to 1 m (Figs 7.7 and 7.8). Individual seams are lenticular, not more than 150 mm thick and aggregate coal thickness rarely exceeds 300 mm.

9.4.2   Environmental Association and Controls on Accumulation

The geometry and development of the No 2A Seam closely resembles that of the No 1 Seam. Broadly, the seam records peat accumulation on a delta plain to the north and west of the Dundee-Vryheid line, to the south and east peat formed in barred embayments and/or lagoons behind prograding beach barriers (Fig 9.4). The non- development of the seam in the central part of the area is attributed to the inherently stable nature of that part of the platform; the level of the water table within the predominantly arenaceous sediments was insufficient to support or maintain a hydrophilic flora and promote conditions conducive to peat formation. In contrast, the adjacent northern and southern areas appeared to have supported extensive peat-forming swamps. The basinward increase in
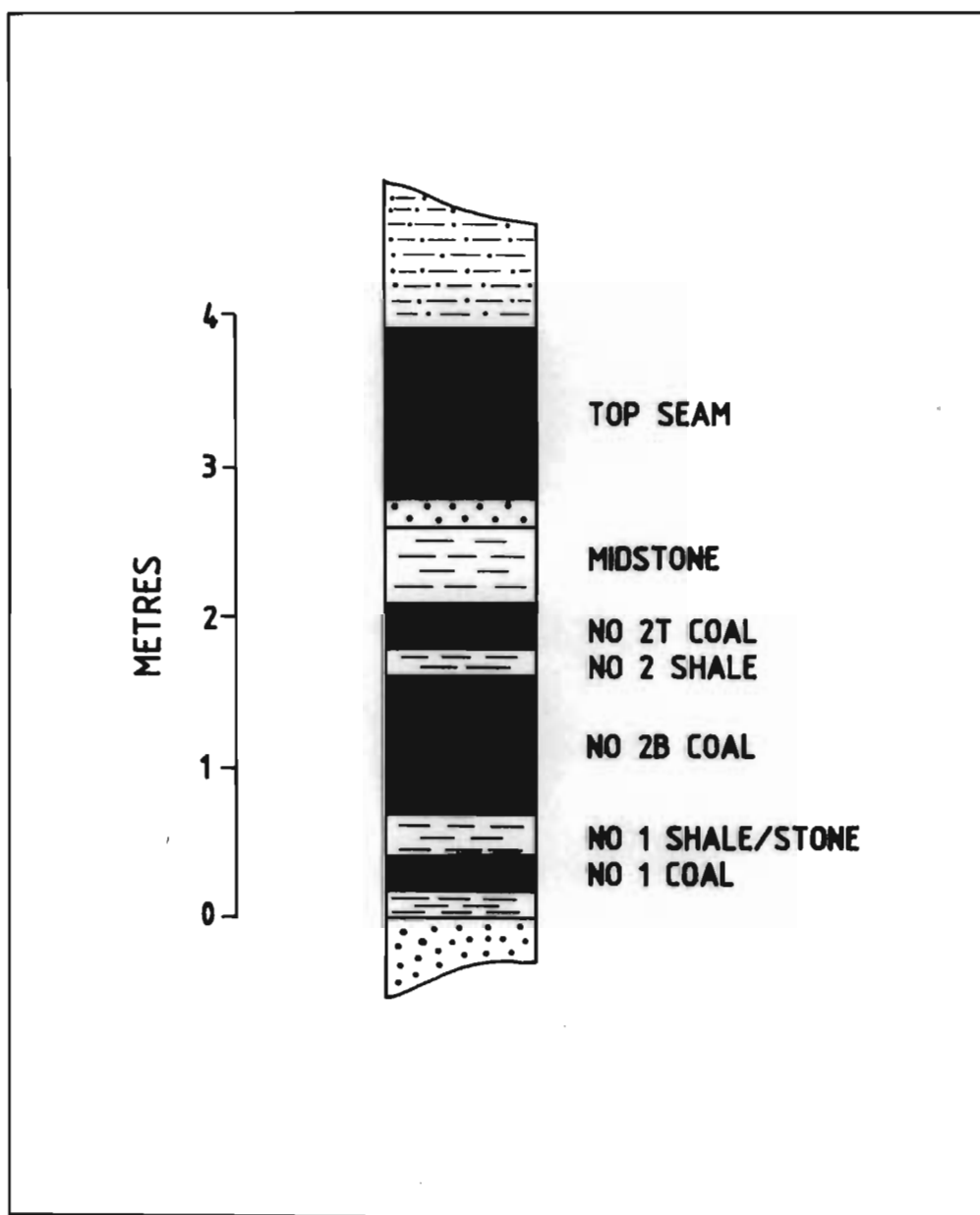
Figure 9.3. Coal-seam nomenclature, Durnacol mine.

coal-seam thickness southwest of Glencoe, ie. towards a less stable part of the coalfield, strongly suggests the existence of an optimal relationship between subsidence (mainly tectonic), peat aggradation and water table level.

The extensive development of the seam over the main coalfield is in stark contrast to its mode of occurrence to the southeast of the Dundee-Vryheid line where peat accumulated in association with a prograding coastline. The lenticular nature of coals is a function of the distribution of wetlands associated with barred embayments and lagoons periodically inundated by sand (Horne et al., 1978; Galloway and Hobday, 1983). The implied coastal setting suggests that marshes as well as swamps existed, and possibly even floating mats of vegetation. Controls on peat swamp/marsh development and distribution in embayments and back-barrier lagoons included:

1. Swamps or marshes would only exist on the fringes of these bodies of water in areas where the water was shallow enough to promote plant growth.

2. Influxes of sediment predominantly from rivers was probably the single most important mechanism by which peat accumulation was terminated at any one locality. Crevasse splays and sheet-flood deposits would also have contributed to the shallowing of the environment, thus creating sites suitable for the establishment of swamps and marshes.

3. Episodic coastal aggradation would also have played an important role in terminating swamp environments and creating sites for new swamps elsewhere.

The predominantly dull nature of the coal can be attributed to a number of factors, including peat accumulation in an environment in which both surface waters and groundwaters were to some extent saline (high pH and Eh), and rarity
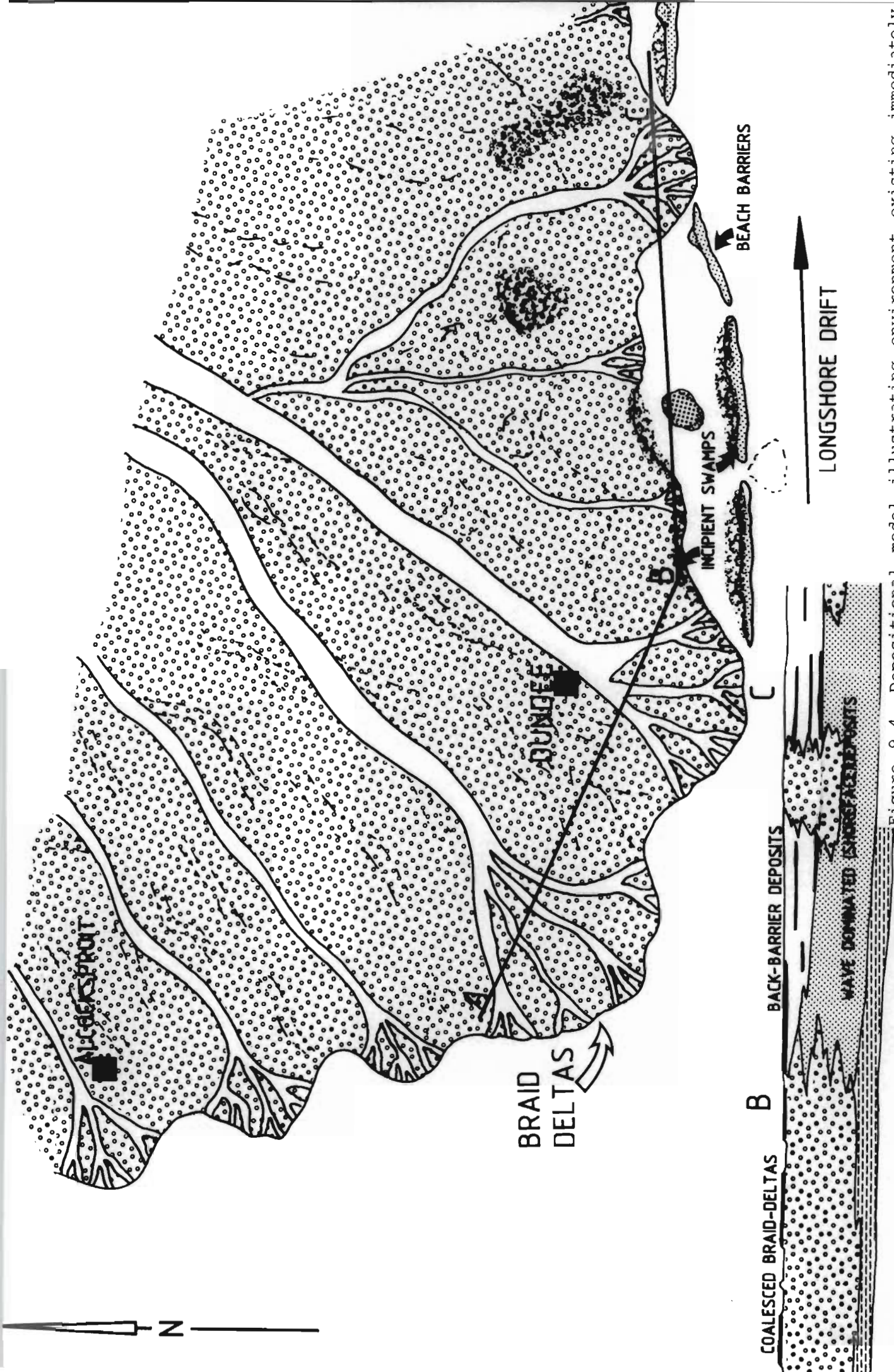
Figure 9.4. Depositional model illustrating environment existing immediately prior to the establishment of the No 2A Seam peats.

BEACH BARRIERS

LONGSHORE DRIFT

INCIPIENT SWAMPS

BRAID DELTAS

COALESCED BRAID-DELTAS

BACK-BARRIER DEPOSITS

WAVE DOMINATED (SHOREFACE) DEPOSITS

ALBERTSPRUIT

DUNDEE

N

LOCALITY OF CROSS SECTIONS.

KEY

BRIGHT COAL
MIXED MAINLY BRIGHT COAL
MIXED BRIGHT AND DULL COAL
MIXED MAINLY DULL COAL
DULL COAL
CARBONACEOUS MUDROCK
MUDROCK(Facies Fm, Fr, Ff, Fl)
ALTERNATING SANDSTONE AND MUDROCK (Facies Al)
FINE-GRAINED SANDSTONE
MEDIUM-GRAINED SANDSTONE
COARSE-GRAINED SANDSTONE
V   DATA POINTS

of woody material (see previous chapter). Evidence that the environment was influenced by saline waters is provided by floor and roof facies associations and bioturbated clastic partings. An imbalance between water-table level and the peat surface (thus resulting in oxidative degration) may also have been a contributory factor, but this is thought unlikely in view of the relatively high rates of subsidence which are thought to have prevailed over this part of the platform. Vertical facies associations suggest that peat accumulation terminated as a result of peat-swamp and -marsh aggradation being unable to keep pace with platform subsidence.

9.5   THE NO 2A/2B CLASTIC PARTING

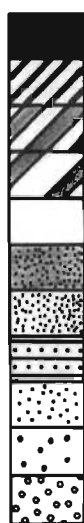A characteristic feature of this parting is that where the No 2A Seam is developed, its thickness over the main coalfield is consistently between 3 and 6 m (Figs 9.5, 9.6 and 9.7). To the southeast of the Dundee-Vryheid line it displays on equally consistent thickness, being in the order of 18 to 20 m but thickening to 30 m in a southeasterly direction (Fig 7.13).

Initially, deposition was dominated by minor braided distributaries flowing across an extensive delta plain and probably emptying into shallow embayments and lakes. The influence of saline basinal waters is reflected by the high degree of bioturbation, especially in the southeast. The second phase of sedimentation was by braided rivers which coalesced to form a sheet sandstone covering almost the entire area. In places these channels eroded through the underlying sediments to rest directly on the No 2A Seam peats. Figure 7.13 indicates that the locus of fluvial sedimentation was located southeast of the Dundee-Vryheid line, due probably to this sector being topographically lower than the main coal-forming platform. Although channel distribution was to a large extent controlled by platform topography, the influx of sand owes its origin to a major phase of erosion in the source-area.
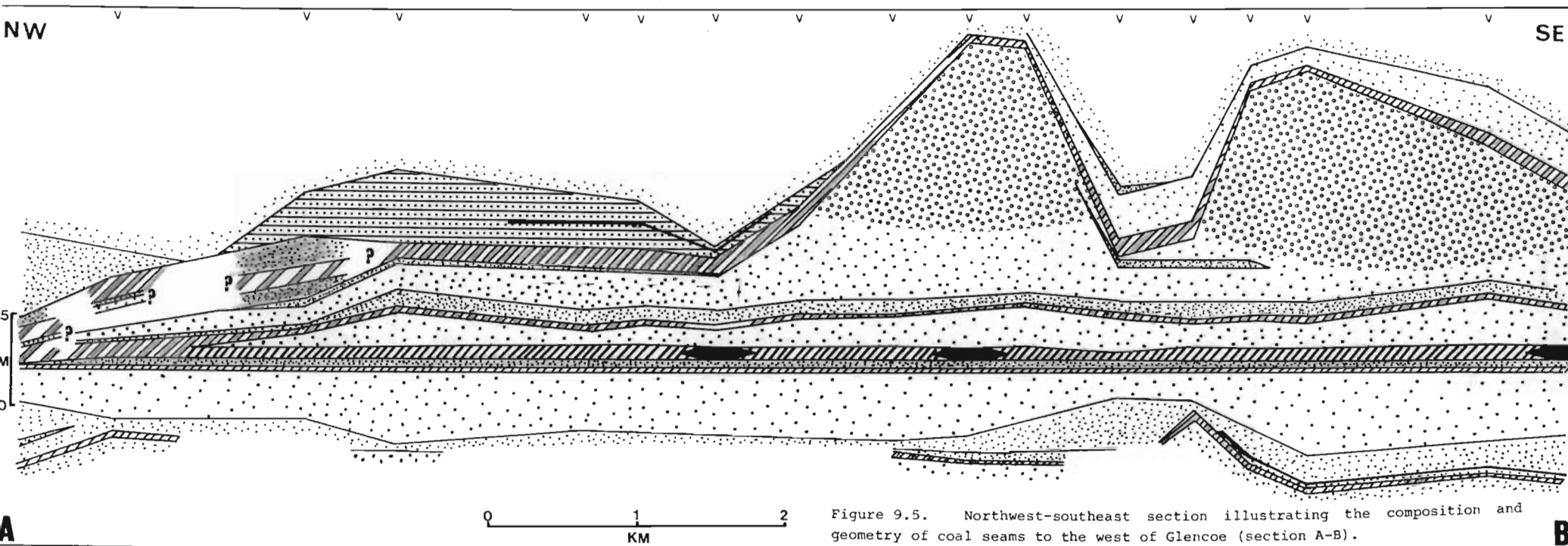
Figure 9.5. Northwest-southeast section illustrating the composition and geometry of coal seams to the west of Glencoe (section A-B).

Figure 9.6. Northwest-southeast section to the southwest of Glencoe illustrating distribution of coal seams and geometry and lithology of associated clastic partings (section C-D).
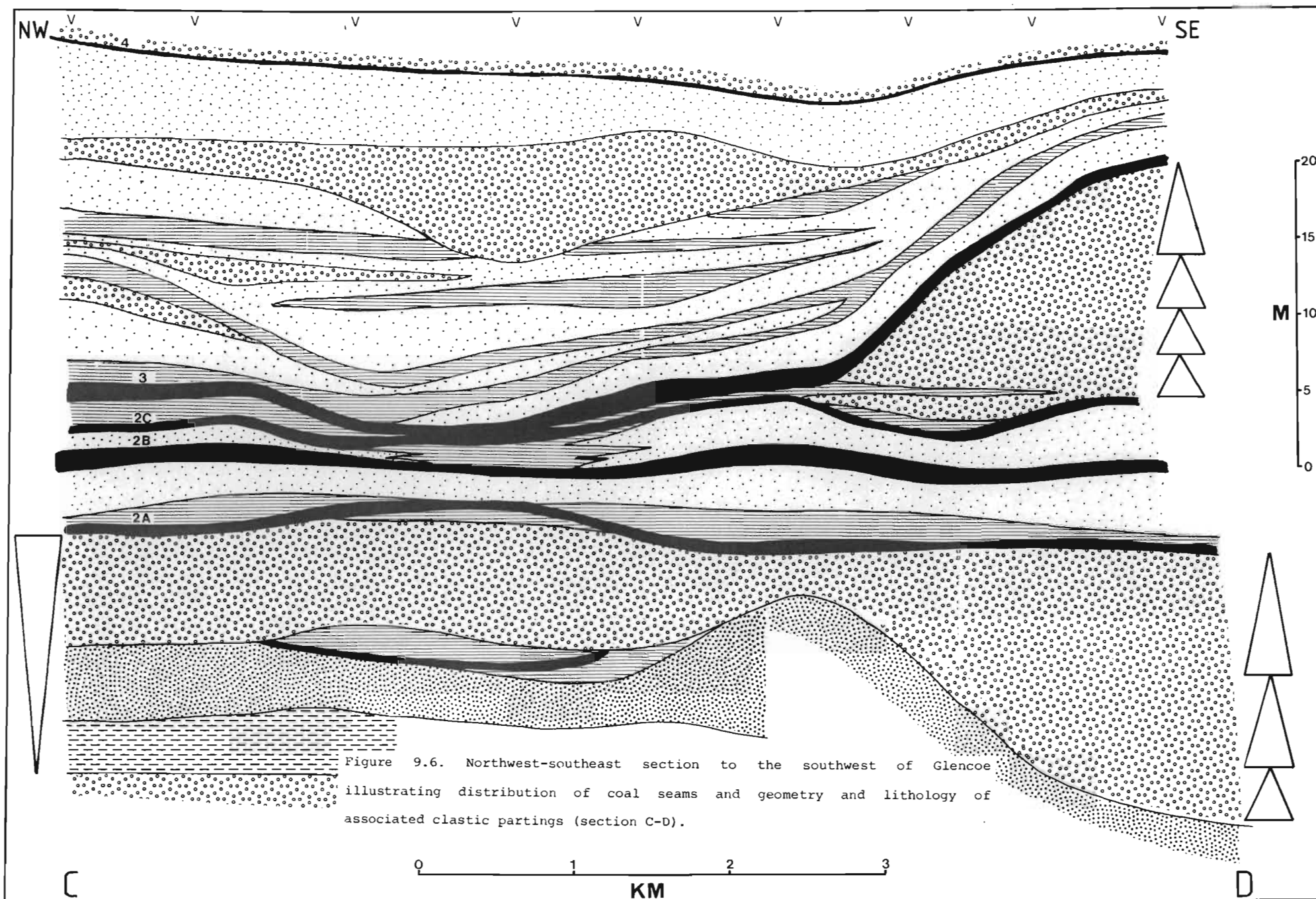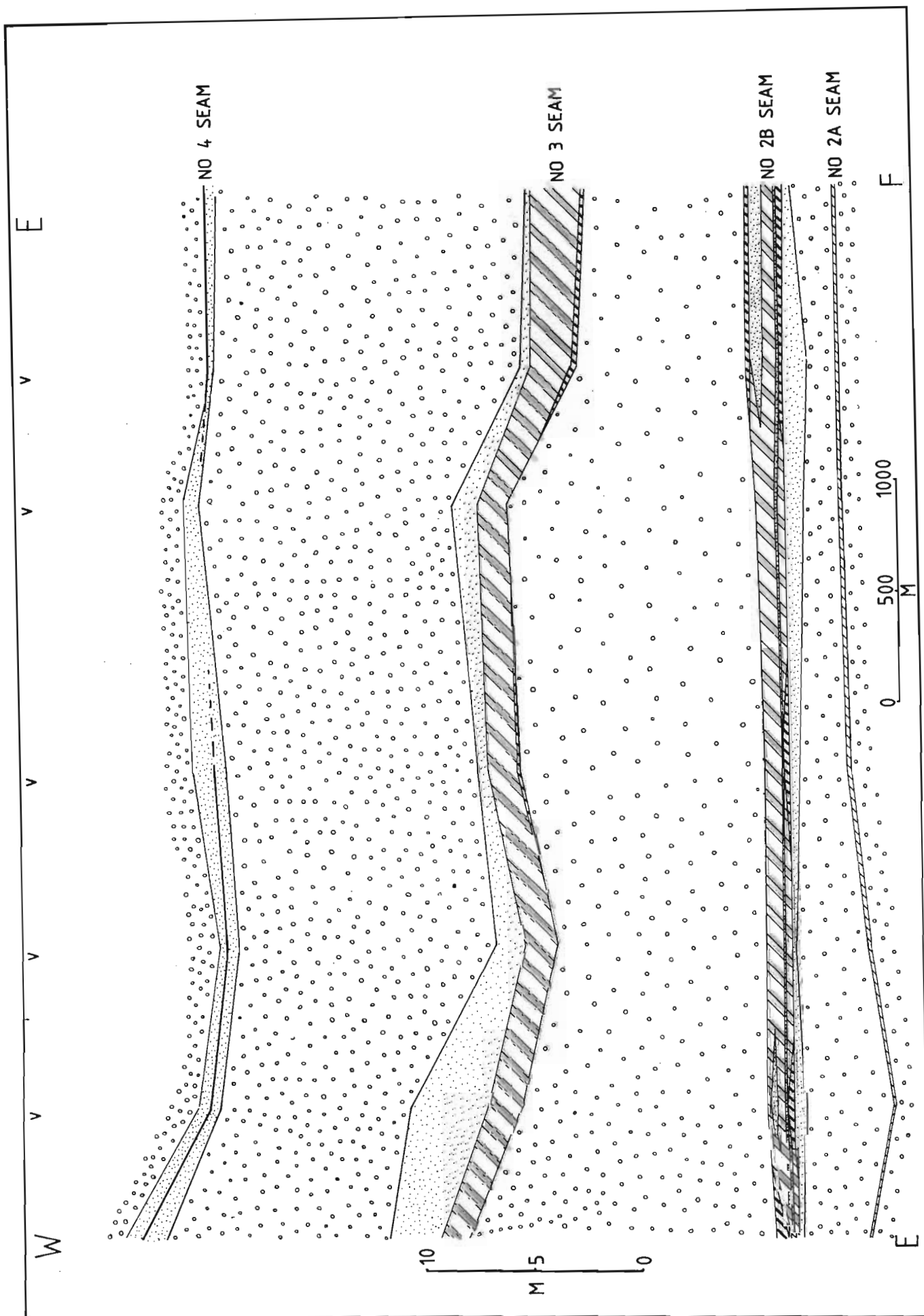
Figure 9.7. West-east section illustrating geometry and composition of coal seams in the northern part of the thesis area (section E-F).

9.6   THE NO 2B AND 2C SEAMS

9.6.1   Description and Distribution

The No 2B Seam is one of the prime economic coals of the Klip River
coalfield and as such its geometry and composition are well documented. Where
this seam amalgamates with the No 3 Seam the two are referred to as the "Main"
Seam, although a thin parting is always distinguishable. The areal variation in
seam thickness is depicted in Figure 9.8. Attention is drawn to the tendency
for the seam to thin towards the west and southwest, ie. towards the unstable
margin of the platform. The thickest development of the seam is in the vicinity
of Dundee and Glencoe.

The seam displays a characteristic internal macroscopic composition and
structure which persists over the entire area north of the Dundee-Vryheid line
(Figs 9.5, 9.7, 9.9 and 9.10). This uniformity provided an invaluable aid in
confirming seam correlations, especially to the southwest of Glencoe. Some
discrepencies do exist, but these are attributed to the inherent variations in
coal descriptors applied by the numerous geologists involved in logging
borehole core.

Between Dundee and Elandslaagte and as far north as Dannhauser the lower
part of the seam comprises between 180 and 250 mm of mixed, mixed mainly bright
or bright coal. This is succeeded by a clastic parting which varies in
thickness from 5 mm to 0,60 m (average 200 mm). At its thinnest it is a fine-to
coarse-grained sandstone, but is mostly composed of facies Fm carbonaceous
mudrock overlain by a thin sandstone. On Durnacol mine this parting is referred
to as the No 1 shale or stone band (Fig 9.3). North of Dannhauser the seam's
lower component is described as mixed mainly dull to dull coal and is
invariably overlain by a fine-to coarse-grained, pebbly sandstone 3 to 20 mm
thick.

The upper part of the seam is more complex than its lower interval. Between
Dundee and Dannhauser it comprises bands of mixed to mixed mainly bright

coal separated by a clastic parting. The geometry of this parting is critical in understanding the behaviour of the No 2B Seam. Over most of this area it occurs as a sandy mudrock not more than 400 mm thick and generally in the order of 50 to 100 mm (No 2 shale at Durnacol mine, Figure 9.3). However, to the southwest of Glencoe the parting increases both in thickness and grain size and the overlying coal is named the No 2C Seam (Figs 9.5, 9.11 and 9.18). Although the northern and northeastern boundary of the split are clearly delineated, the southeastern boundary cannot be determined with any accuracy. It does, however, appear to be situated to the southeast of the Dundee-Wasbank road as the No 2C Seam cannot be discerned further eastward within the Indumeni colliery property. Figure 9.12 depicts the distribution and thickness of the No 2C Seam. In the northern part of the coalfield the upper part of the No 2B Seam comprises a higher proportion of dull coal than in the central and southern parts.

To the southeast of the Dundee-Vryheid line the No 2B Seam is extremely variable in composition (Fig 9.13) and in places it contains highly bioturbated (Siphonichnus (S) association) sandy partings. West of the Newcastle-Ladysmith line the seam becomes increasingly shaly over a few kilometres and is eventually preserved only as a carbonaceous mudrock with a few coal laminae. The seam is similarly developed in an area to the west of Glencoe which forms an "embayment" into the main coalfield (Fig 9.8). Within this area the No 2 and 3 Seams are represented by a sequence of predominantly dull coal and carbonaceous mudrock with minor coal and sandstone intercalations. Figure 9.13 illustrates the transition into this area from the southeast.

Southwest of Dannhauser, on the Durnacol property, the seam commonly comprises only carbonaceous mudrock in places. These occurrences are up to 1 km across and have either an irregular, quasi-circular outline or two sides may be linear and parallel and show bifurcations (Fig 9.14). In section these mudrock bodies have a channel-like geometry and the junction with the coal is sharp and planar. Microscopic examination reveals that the high mineral-matter content is
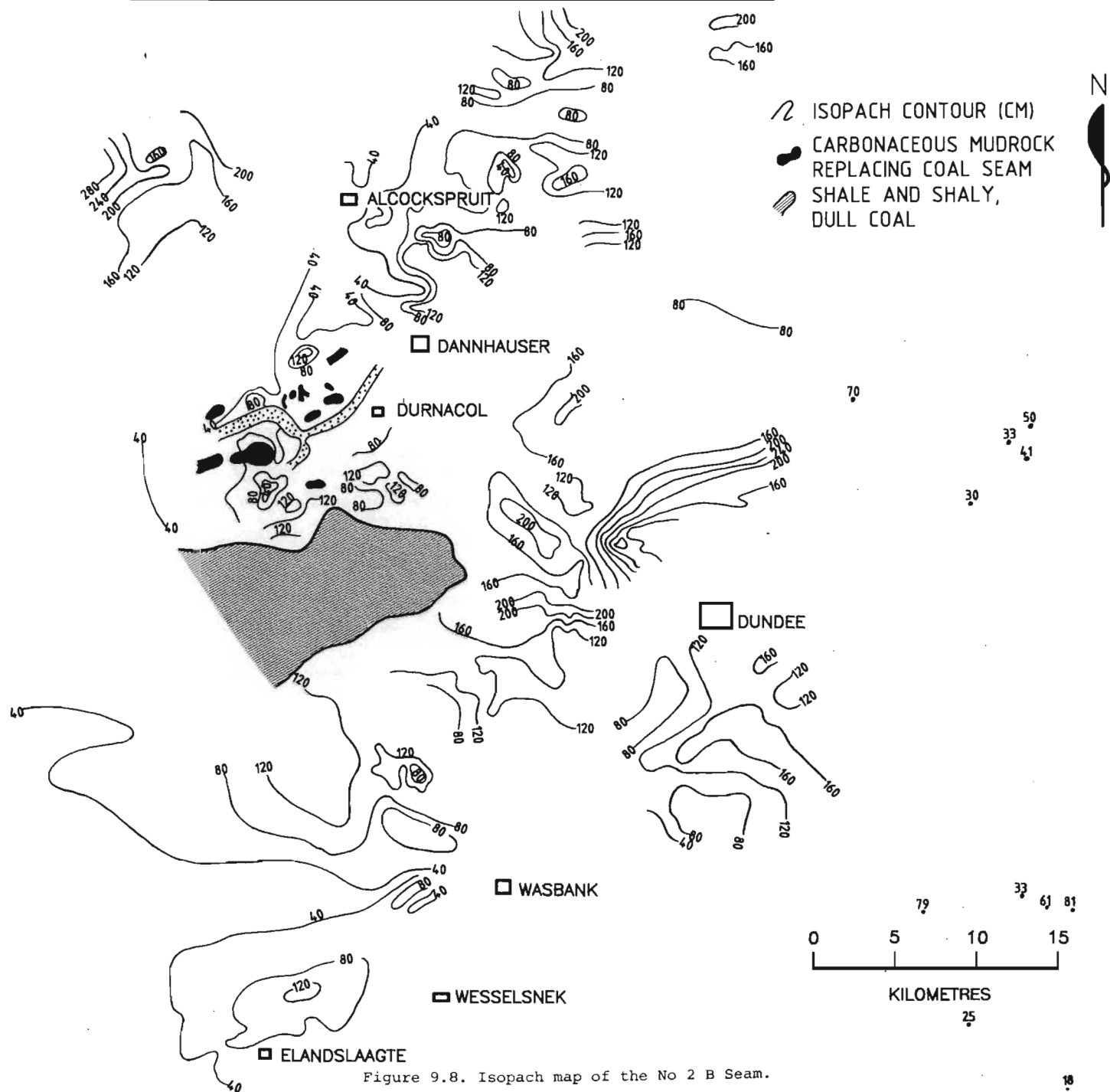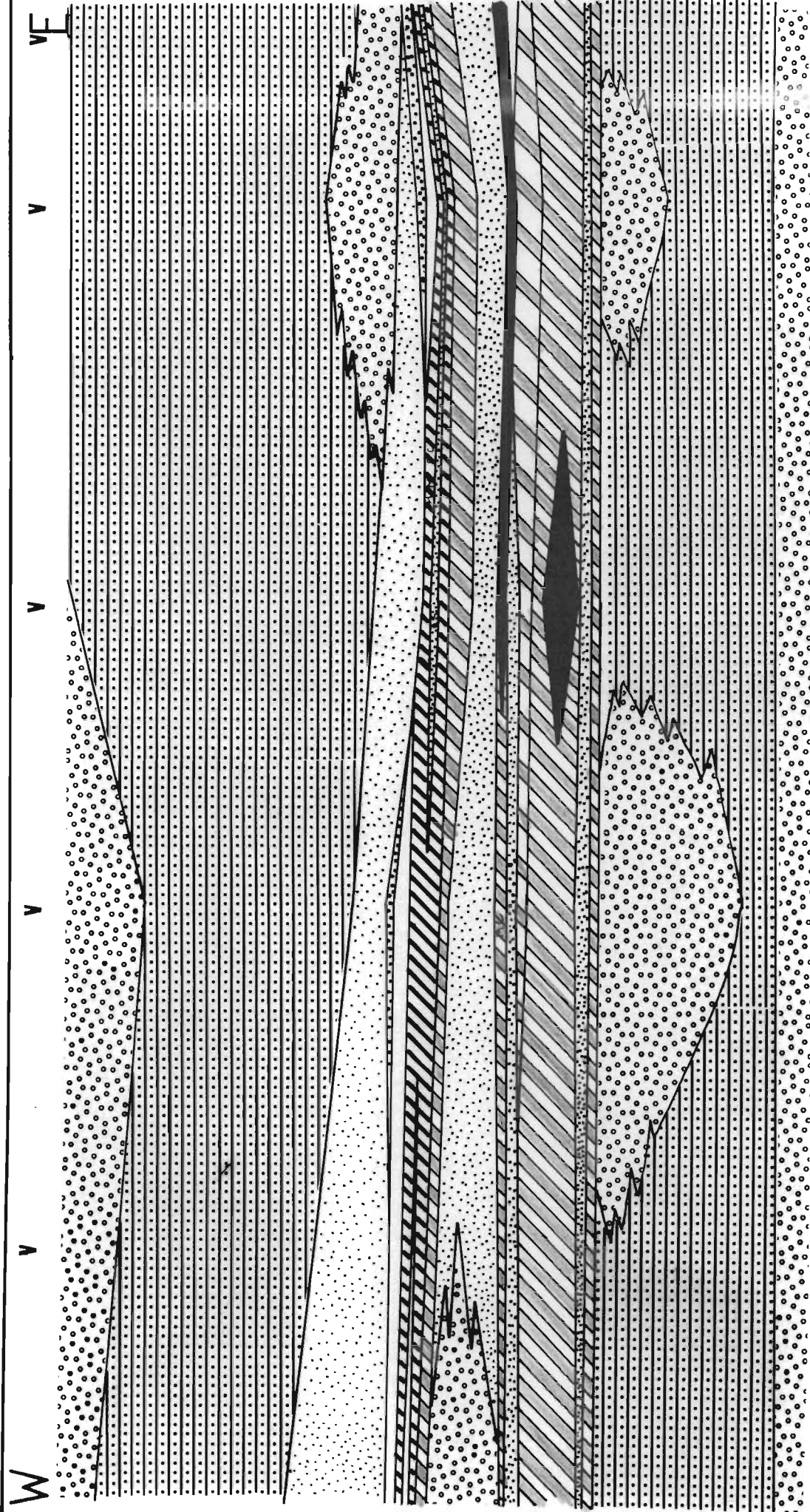
Figure 9.8. Isopach map of the No 2 B Seam.

Figure 9.9. West-east section through the No 2 B and 3 Seams illustrating their composition and lithology of the No 2 B/3 parting in the vicinity of Dannhauser (section G-H).
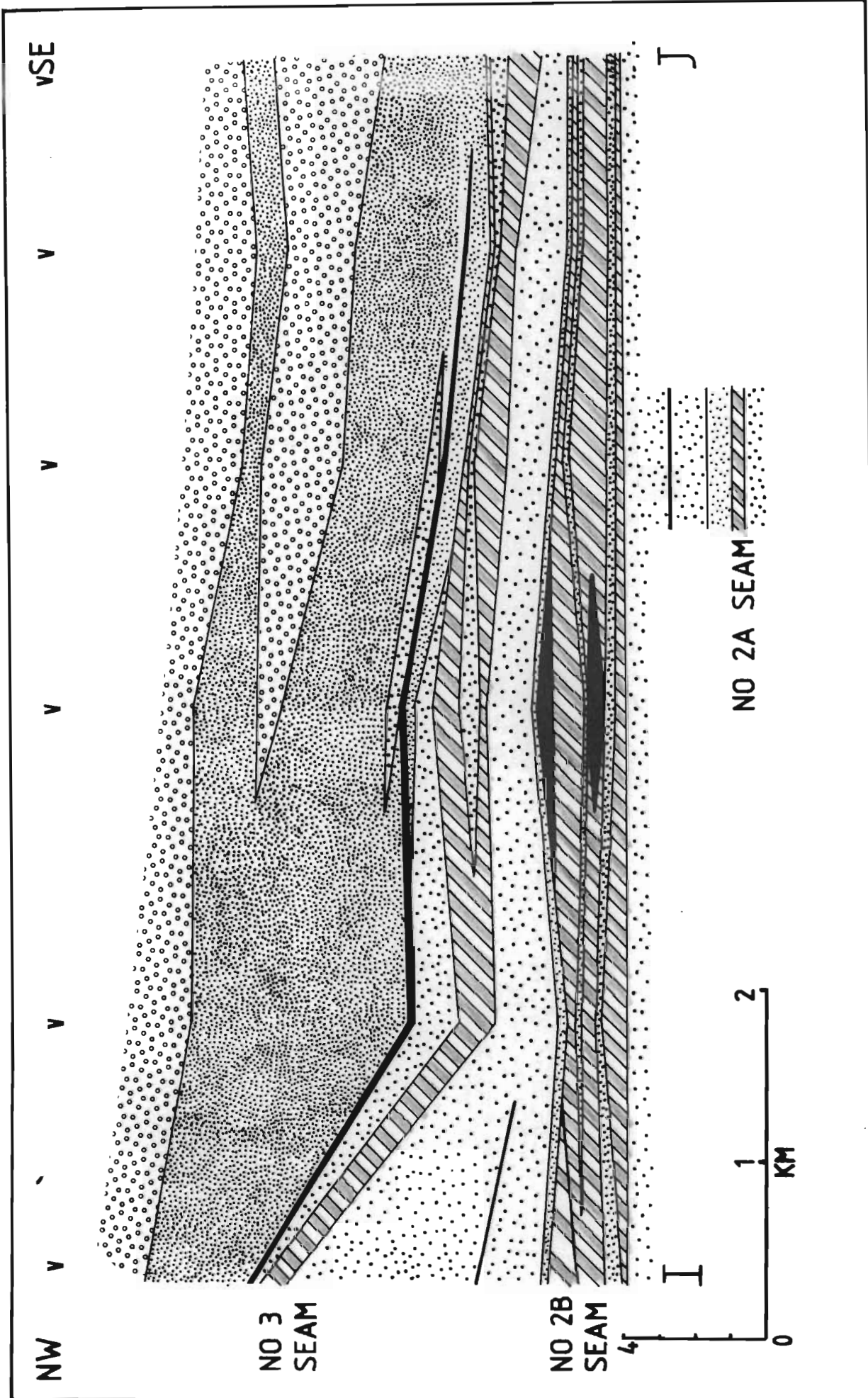
Figure 9.10. Northwest-southeast section illustrating the geometry and composition of coal seams to the north of Glencoe (section I-J).

complemented by up to 25% inertodetrinite, sporinite and cutinite. Figure 9.14 indicates the presence of a sandstone channel <u>within</u> the No 2B Seam. Its orientation is parallel to the edge of the sheet sandstone separating the No 2B and 3 Seams to the north and it dissects a number of the mudrock bodies.

9.6.2   Environmental Association and Controls on Accumulation

The wide lateral extent and compositional homogeneity of the No 2B Seam is indicative of peat accumulation on a coastal plain devoid, except for short periods, of clastic sedimentation. The predominantly mixed, mainly bright nature of the coal in the southern and central parts of the coalfield suggests accumulation in a low-lying, treed swamp in which fluctuations in water-table level led to periodic degradation of the peat. The higher proportion of dull coal in the north is attributed to a lower rate of platform subsidence and consequent imbalance in water-table and peat-surface levels. Basinward thinning of the seam is believed to be due mainly to the deleterious influence of saline basinal waters on plant growth and peat production. No doubt, platform instability in proximity to the coalfield margin also adversely influenced peat formation. The apparent decrease in No 2B Seam thickness southwest of Glencoe is also due in part to the exclusion of the No 2C Seam in the measurement.

The shaly nature of the seam west of the Newcastle-Ladysmith line and within the embayment or lake west of Glencoe suggests major zones of platform instability. Consequently, these areas of the swamp were almost perpetually covered by water too deep for significant plant growth to occur. While subsidence within the Dannhauser Trough corresponds to a pre-Karoo basement structure, the morphology of this feature can only be reconciled with compactional subsidence processes, possibly related to the position of an earlier formed inter-distributary embayment.
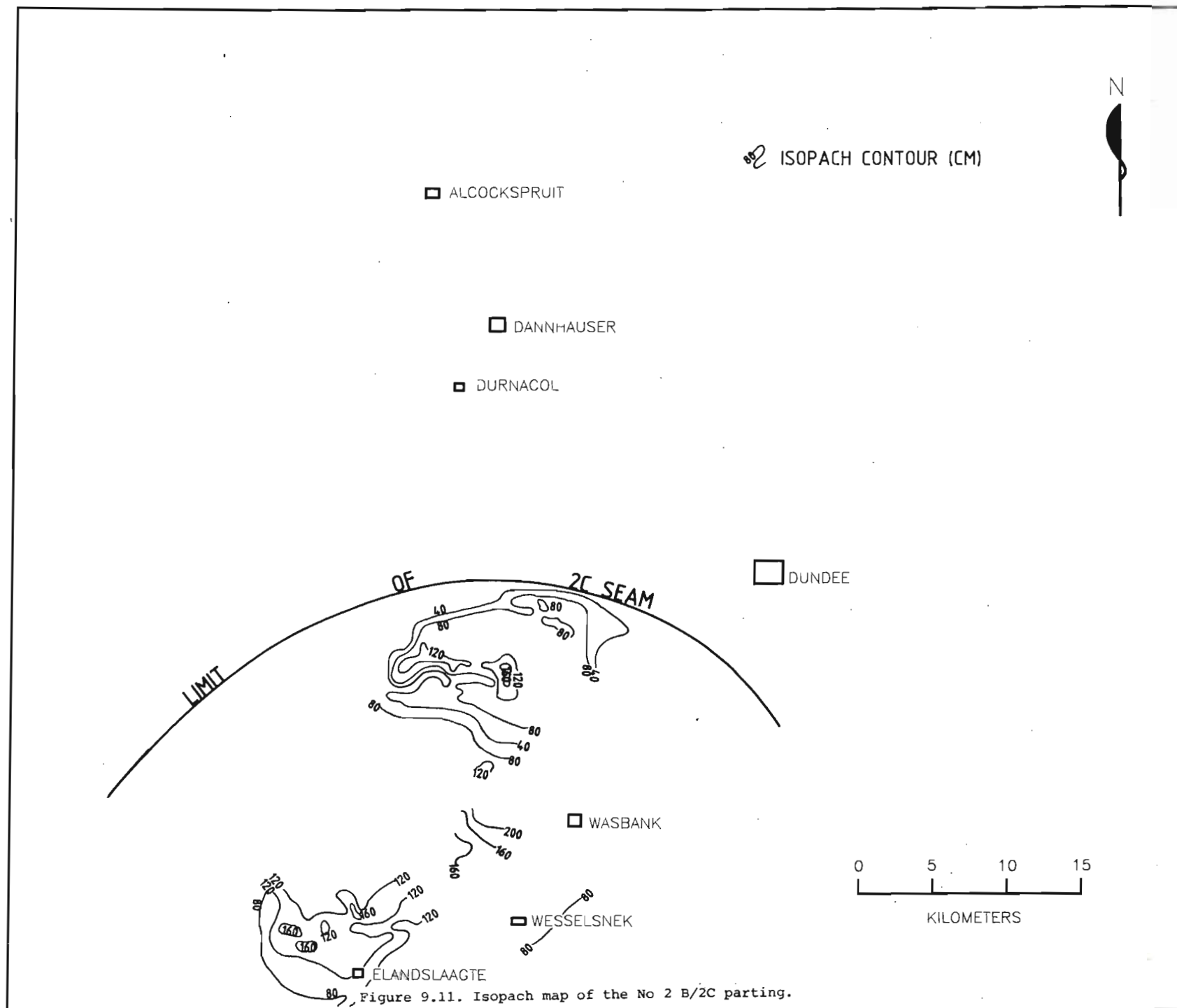
Figure 9.11. Isopach map of the No 2 B/2C parting.

□ ALCOCKSPRUIT

$\mathcal{L}_{80}$ ISOPACH CONTOUR (CM)

N

□ DANNHAUSER

◻ DURNACOL

LIMIT OF 2C SEAM

□ DUNDEE

30

20

30

10

20

20

20

20

40

□ WASBANK

40

60 60 60 50 60 60

50 50 50

50 60

50 30

30
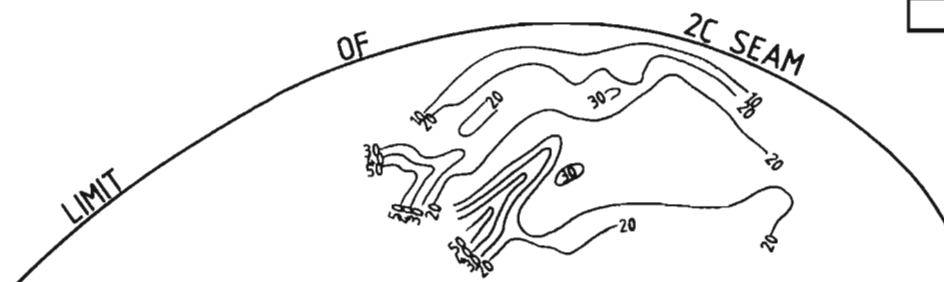
□ WESSELSNEK

□ ELANDSLAAGTE

0   5   10   15

KILOMETERS

Figure 9.12. Isopach map of the No 2 C Seam.

NO 3 SEAM

PARTING NOT
TO SCALE

NO 2B SEAM

Figure 9.13. Composition of the No 2 B and 3 Seams in the area southeast of the
Dundee-Vryheid line (Cinderford area).
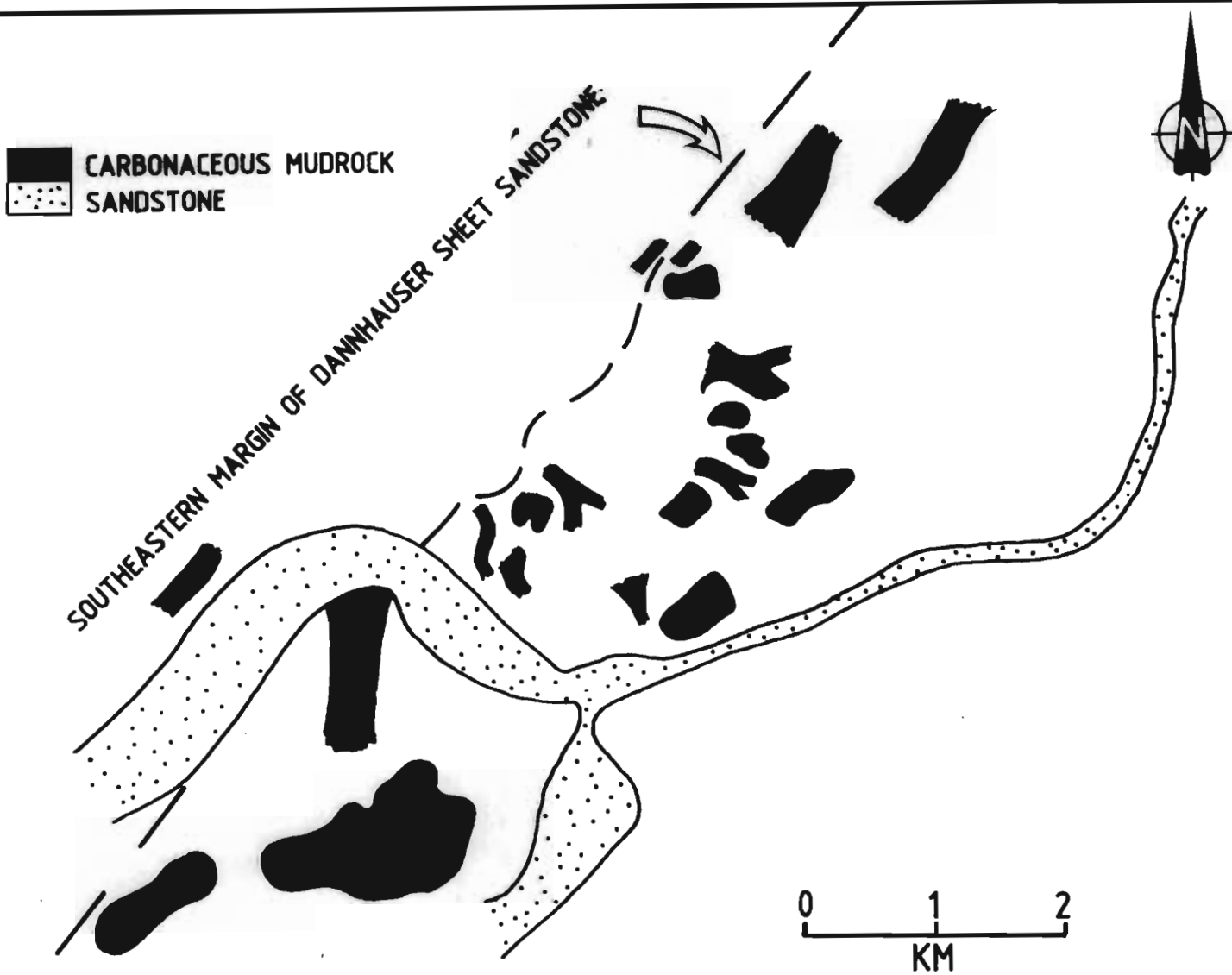
Figure 9.14. Distribution and morphology of clastic partings within the No 2 B Seam in the vicinity of Dannhauser. Data derived from borehole logs and underground mapping.

Peat accumulation to the southeast of the Dundee-Vryheid line was on delta plain sediments. Periods of enhanced platform subsidence are indicated by the presence of bioturbated clastic partings; peat accumulation being periodically interrupted by sand incursions and minor transgressions by basinal waters. The wide variation in seam composition reflects these events and lateral variations in subsidence rates over this part of the coalfield.

The association of sand with the basal clastic parting of the No 2B Seam indicates a clastic origin rather than formation by degradation of plant matter (cf. Renton and Cecil, 1979). The extensive development of the parting suggests that the swamp was founded on a flat, featureless platform. There is no doubt that the swamp was inundated by water carrying mainly suspended sediment, but the mechanism by which the water was introduced is not clear. Except for the intraseam sandstone channel near Dannhauser, no other features to suggest the origin of a sheet flood are obvious. It is evident, however, that ponds, small lakes and minor channels meandering (in a southwesterly direction) through the swamp existed in the vicinity of Dannhauser. Some of these ponds may have been remnants of abandoned channels. The sharp, planar junction between the coal and mudrock cannot be reconciled with channel erosion features; an irregular, jagged junction with phytoclasts would be the more likely result. It is possible that the channels and ponds were in existence since the inception of the swamp and aggraded at the same rate as the swamp surface. The argillaceous content of these features does not 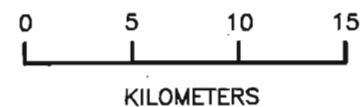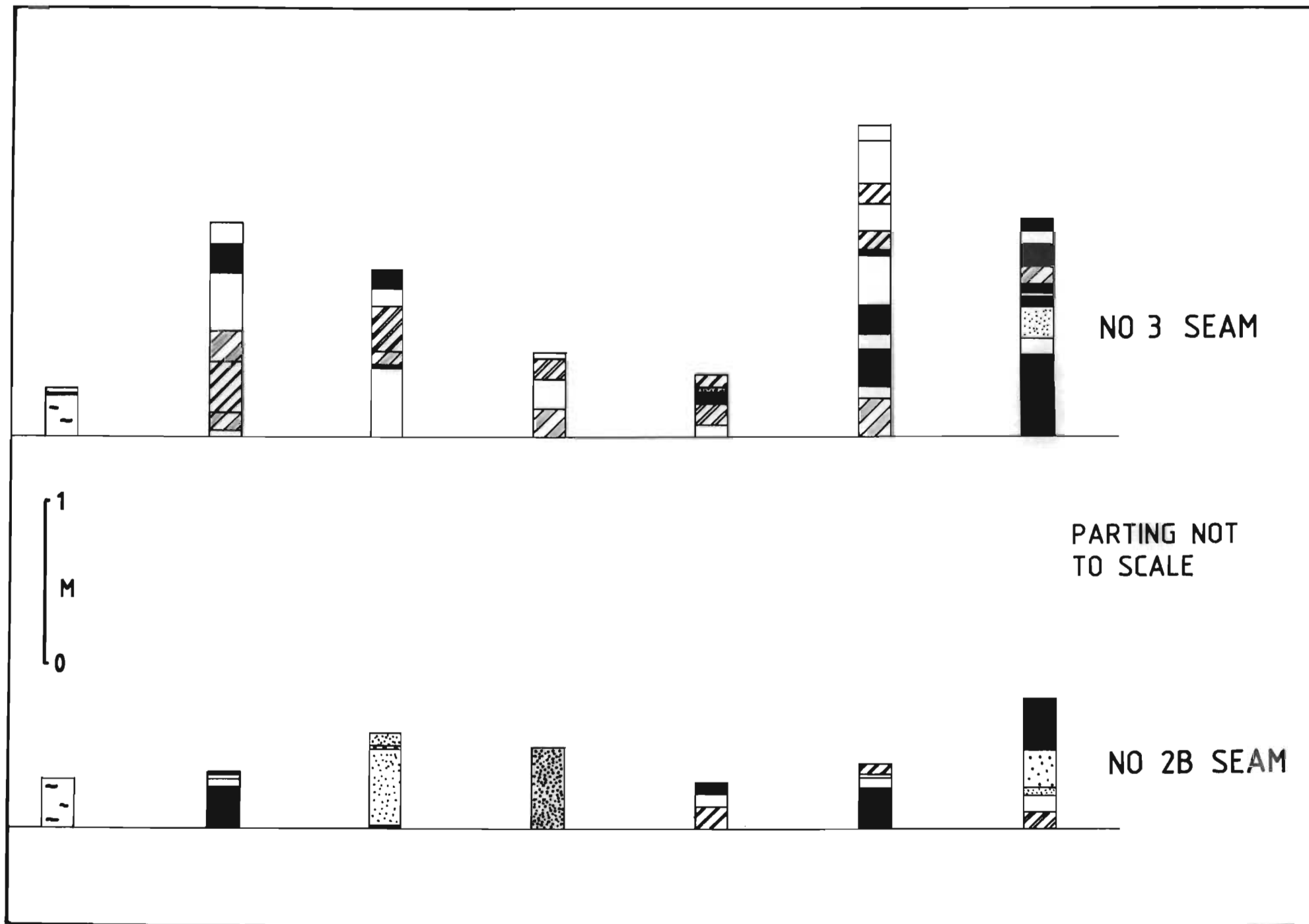signify that currents were insufficient to transport sand-sized particles, rather that the only sediment available was silt, clay and organic matter. These channels resemble the anastomosing river systems described by Smith and Smith (1980): river courses stabilized by peat and overbank sediments produce relatively thick, narrow channels dominated by vertical aggradation. The channels were probably swamp drainage features and did not originate beyond its boundaries. In contrast, the sandstone-filled channel could only have originated beyond the margins of the swamp. Subsurface mapping strongly suggests that this channel foundered within the swamp. Its erosional relationship with the underlying intraseam mudrock bodies indicates

that it was a late event in the depositional history of the swamp and may be the same age as the sheet sandstones overlying the No 2B Seam.

The well-defined and sharp proximal boundaries of the 2B/2C Seam split and its limited lateral extent are indicative of inception by differential-compaction subsidence rather than basement tectonics. As there is not any obvious northeastery source to the sediments comprising the parting it must be presumed that they were derived either by overbank processes from adjacent fluvial channels or as a result of basinal transgression subsequent to the subsidence of this part of the platform which gave rise to the split. The thickening of the split to the southeast reflects decreasing platform stability towards the edge of the main coal-forming platform.

9.7   THE NO 2B/3 CLASTIC PARTING

The parting between the No 2B (and in places the No 2C Seam) and the No 3 Seams is relatively well documented and comprises the deposits of alluvial plain sheet sandstones formed by coalescing braided fluvial channels and their associated overbank flood deposits. Figures 9.6 and 9.15 and also 5.1 and 5.2 illustrate the thickness, distribution and lithologies of the parting.

Currents within the "Glencoe" sheet sandstone were towards the southwest. Approximately where the sheet sandstone intersects the Dannhauser Trough, its width increases markedly, signalling possibly transition into a delta lobe. The "Dannhauser" sheet sandstone displays a similar tendency; palaeocurrent orientations are towards the west in the northern part of the area, but swing towards the southwest where it intersects the Newcastle-Ladysmith line. It appears that the two sheet sandstones intersect near the western boundary of the thesis area, but with no appreciable increase in aggregate thickness of the parting. The margins of the Glencoe sheet sandstone, particularly on its southeastern edge, are interrupted by what are interpreted as sandstone crevasse-splay lobes. In the interchannel area overbank sheet-floods are
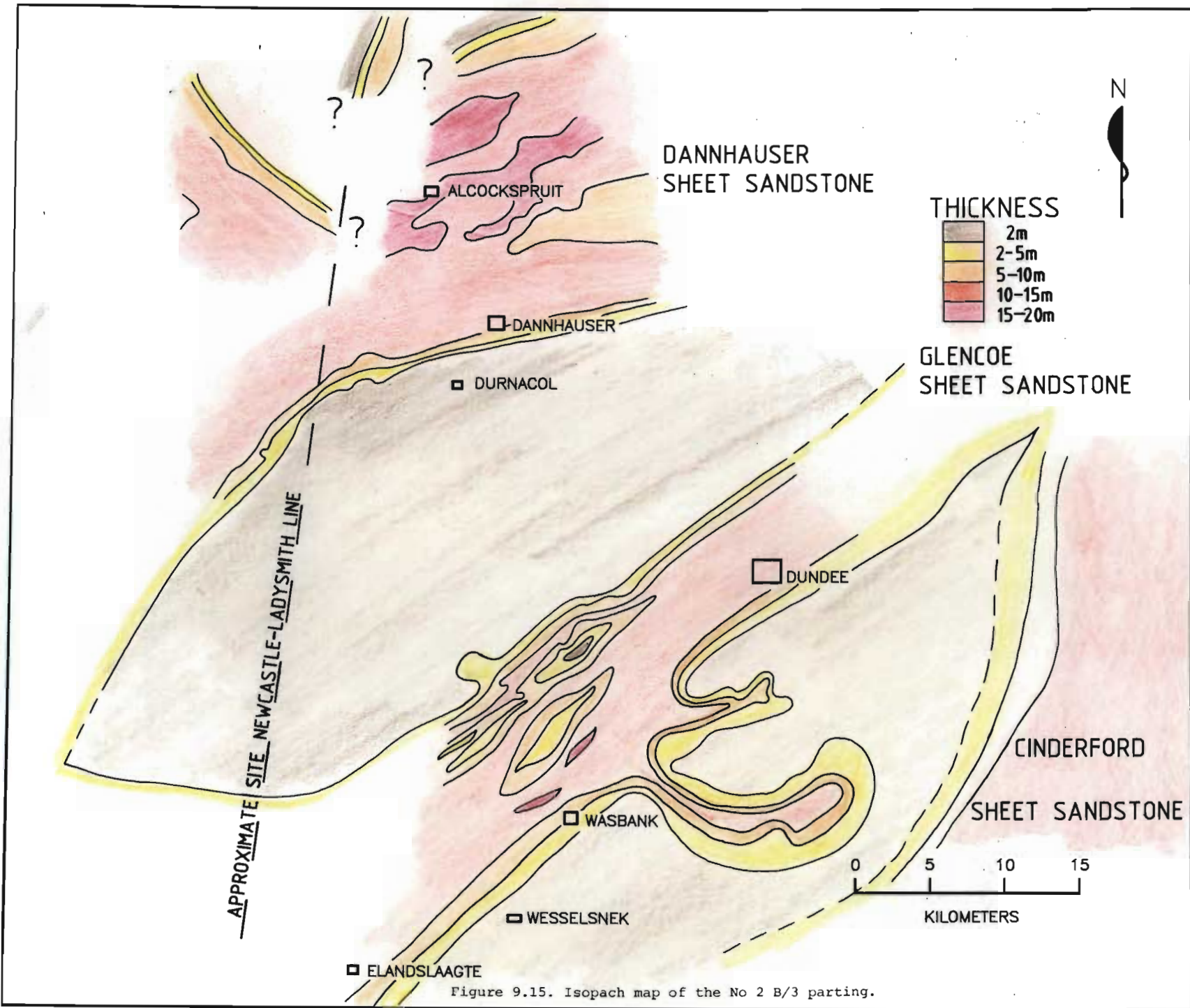
Figure 9.15. Isopach map of the No 2 B/3 parting.

indicated by the predominantly argillaceous nature of the parting and thin (5 to 200 mm), tabular sandstones of wide lateral extent (Fig 9.5 and 9.6).

While the initiation of this phase of fluvial activity is attributed to source-area controls, controls on the distribution of the sheet sandstones can only be speculated on. Two possibilities are discussed below:

1. The earliest fluvial channels chose a random path through the peat swamps. Localised peat compaction thus induced created depressions which served as loci for further fluvial sedimentation.

2. Rivers forming the sheet sandstones took advantage of pre-existing topographic depressions within the peat swamp.

Of the two, the second alternative appears more attractive since the margins of the sheet sandstone coincide with previously identified structures or areas of instability within the sedimentary platform. For example, the position the Glencoe sheet sandstone coincides almost exactly with that of the 2B/2C Seam split. The relative thinness (even allowing for the differing compaction ratios of mudrock and sandstone) and dominance by mudrock facies suggests that the interchannel areas were only rarely inundated by sediment-laden floodwaters. This, the sharply-delineated margins of the sheet sandstones, the fact that channel sedimentation was confined to well-defined conduits, and the only sporadic development of the No 1 and 2A Seams over this area (suggested to be a function of the relative stability of the platform underlying this area) provide convincing evidence that the interchannel area between Dannhauser and Glencoe was slightly elevated with respect to the braided channel courses (Fig 9.16). The implied absence of levees (see Chapter 5) along the edges of these sheet sandstones means that sheet floods rather than channel bank breaching (to form crevasse splays) was the primary mechanism by which sediment was introduced into the interfluvial area.
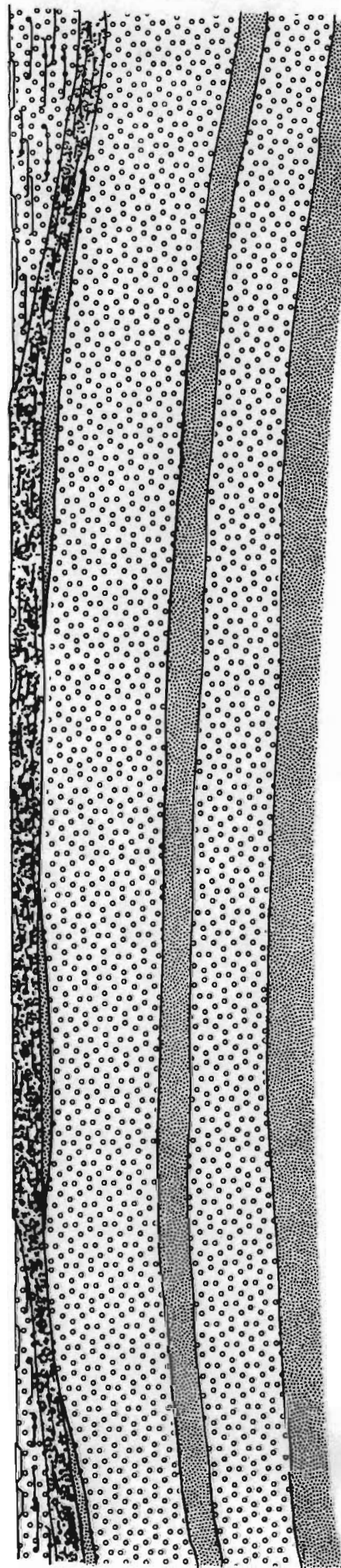
Figure 9.16. Envisaged controls on the development of the No 2 B/3 sheet-sandstone parting and No 3 Seam peats.

The northwestern edge of the Cinderford sheet sandstone appears to coalesce with the Glencoe sheet sandstone (Fig 9.15). The wide extent of the Cinderford sheet sandstone is attributed to the enhanced rate of subsidence this part of the platform experienced. Cognisance should be taken of the orientation of edges of the sheet sandstones, identified only from borehole data, situated to the north of the thesis area (ie. north of Alcockspruit) and their relationship with the Dannhauser sheet sandstone. This area forms the bridge between coal development to the east and to the west of the Newcastle-Ladysmith line. Although it is possible that the sheet sandstones do coalesce at this horizon, it is equally possible that they, and their associated coal seams, may be developed at slightly different stratigraphic levels. At present insufficient data precludes a satisfactory explanation to the observed geometries. It is, however, evident from the orientation of these bodies and implied flow directions that the Dannhauser Trough exerted considerable influence on palaeodrainage in this area and further to the south.

Evidence that fluvial currents did not appreciably erode the underlying No 2B Seam peat is provided by the consistent thickness and internal composition of the seam whether overlain by a sheet sandstone or overbank mudrocks. This is attributed to the peat's interwoven texture, which makes it highly resistant to mechanical erosion (McCabe, 1984), and the tendency for braided rivers to form wide, shallow channels rather than induce stream-bed erosion (Miall, 1977a; Schumm, 1977).

9.8   THE NO 3 SEAM

9.8.1   Description and Distribution

The No 3 Seam is the uppermost economic coal within the coalfield. Figure 9.17 depicts the seam's distribution and thickness over the area. Notable features are:

1. The seam attains its maximum thickness (between 1,6 and 3,6 m) to the northeast of Alcockspruit and Dannhauser.

2. Southeast of the Dundee-Vryheid line average seam thickness is 1,20 to 1,80 m.

3. Southwest of Wasbank the seam is developed only sporadically and in the vicinity of Elandslaagte the seam is either not present or represented (rarely) as a carbonaceous mudrock (Fig 9.18). The absence of the No 3 Seam at Newcastle Platberg colliery resulted in the erroneous identification and naming of the seams (see Figure 9.1).

The No 3 Seam comprises mixed and mixed mainly dull coal (Figs 9.5, 9.7, 9.10 and 9.20). In places the entire seam may consist only of dull coal but, although this is a common component of the seam, it rarely occurs in bands thicker than 400 mm. Within the main coalfield, with exception of the northeastern central area (Fig 9.9), bright coal bands thicker than 200 mm are rare. In contrast, a large proportion of the seam to the southeast of the Dundee-Vryheid line consists of mixed mainly bright and bright coal bands (Fig 9.13). Clastic partings within the seam are rare and only present to any extent immediately north of Dundee (Fig 9.10). A "rider" coal up to 150 mm thick occurs between 0,5 and 1,5 m above the No. 3 Seam in the vicinity of Dundee and Glencoe (Fig 9.10) and to the east of Alcockspruit. This seam is widely developed over the former area, but in the latter area its occurrence is sporadic. Like the No 2B Seam, the No 3 Seam comprises predominantly carbonaceous mudrock between Durnacol and Burnside collieries (Fig 9.17). Lenses of carbonaceous mudrock of similar geometry and composition to those developed within the No 2B Seam are prevalent near Durnacol only where the No 2B/3 parting is less than 2 m thick.

There is a distinct though complex relationship between the thickness (and thus lithology) of the No 2B/3 parting and that of the No 3 Seam (Fig 9.19). Briefly, in the northern and central parts of the coalfield the seam is thickest where it overlies the Dannhauser sheet sandstone and thins noticably
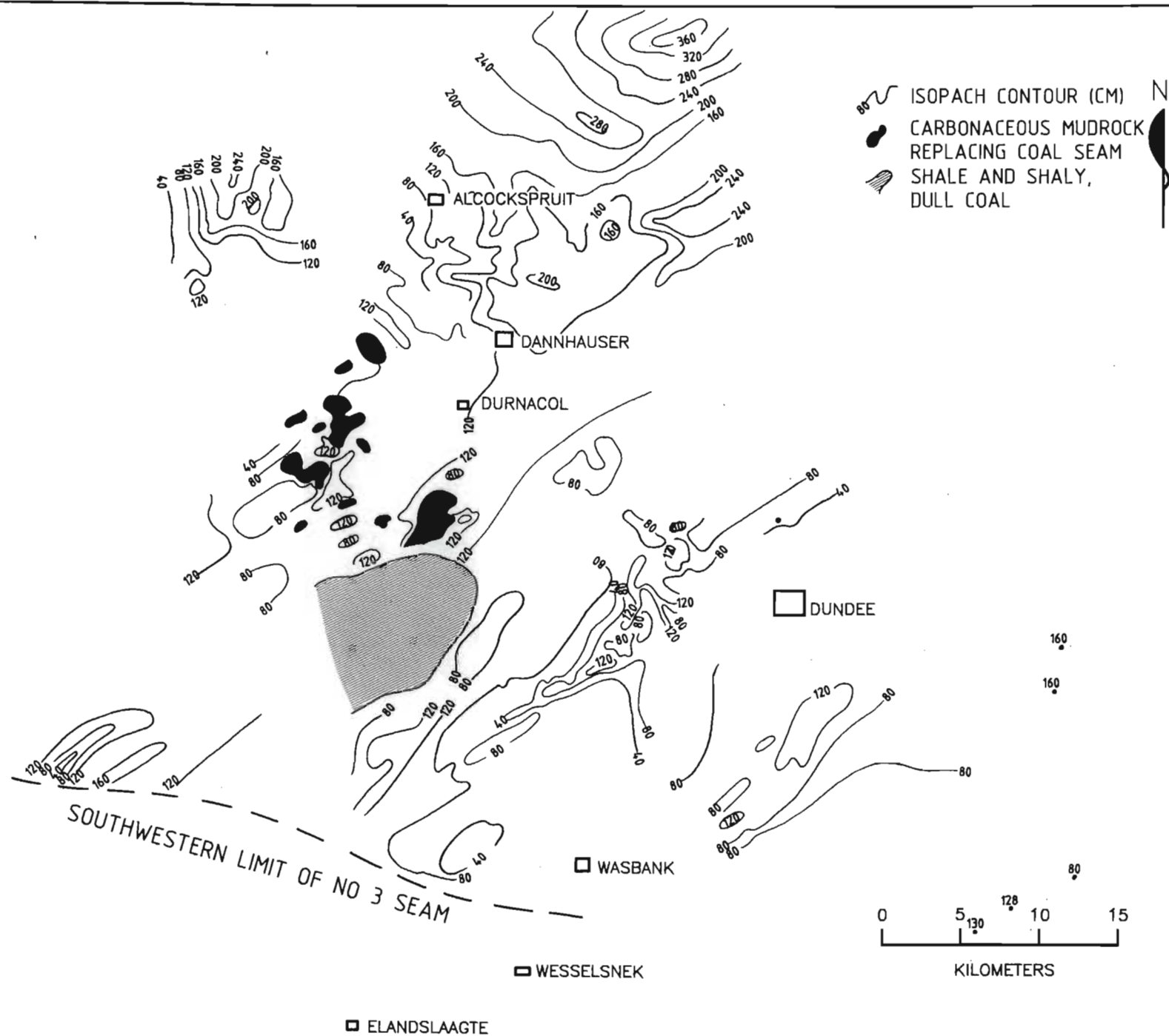
Figure 9.17. Isopach map of the No 3 Seam.
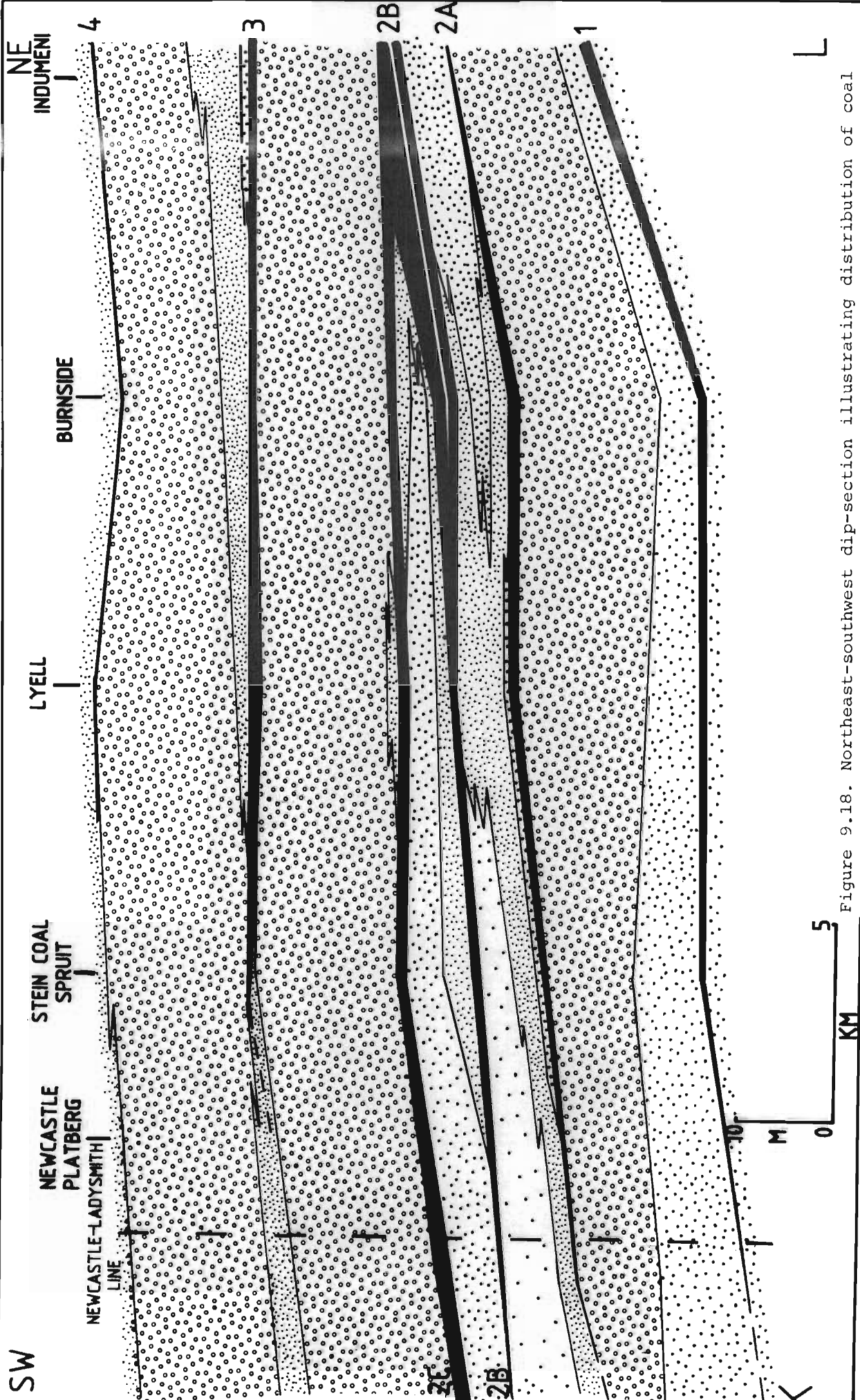
Figure 9.18. Northeast-southwest dip-section illustrating distribution of coal seams (especially the No 2 C Seam) between Indumeni and Newcastle Platberg collieries and beyond the Newcastle-Ladysmith line (section K-L).

Figure 9.19. Isopach map illustrating relationship between thickness of the No 2 B/3 parting and No 3 Seam.
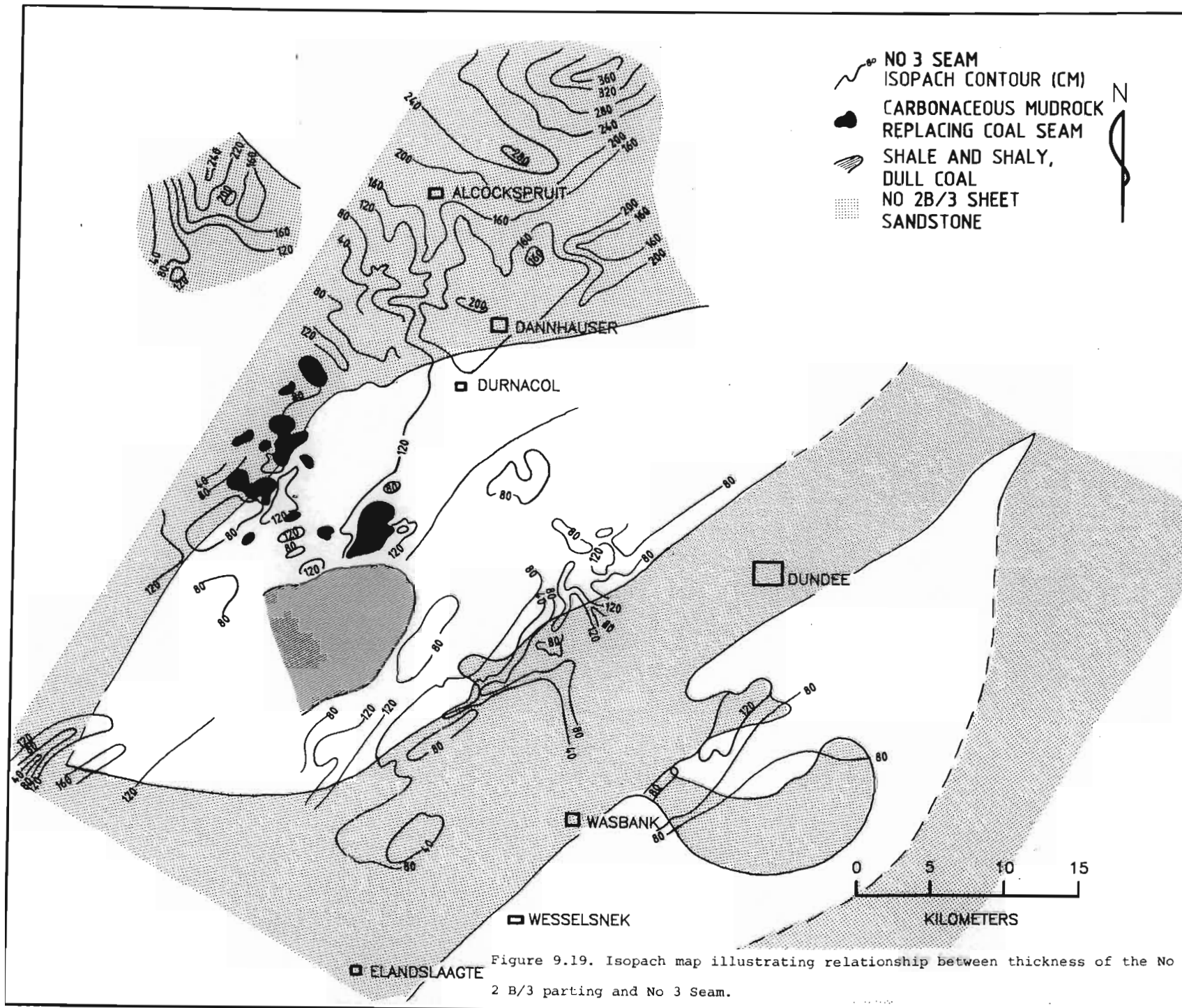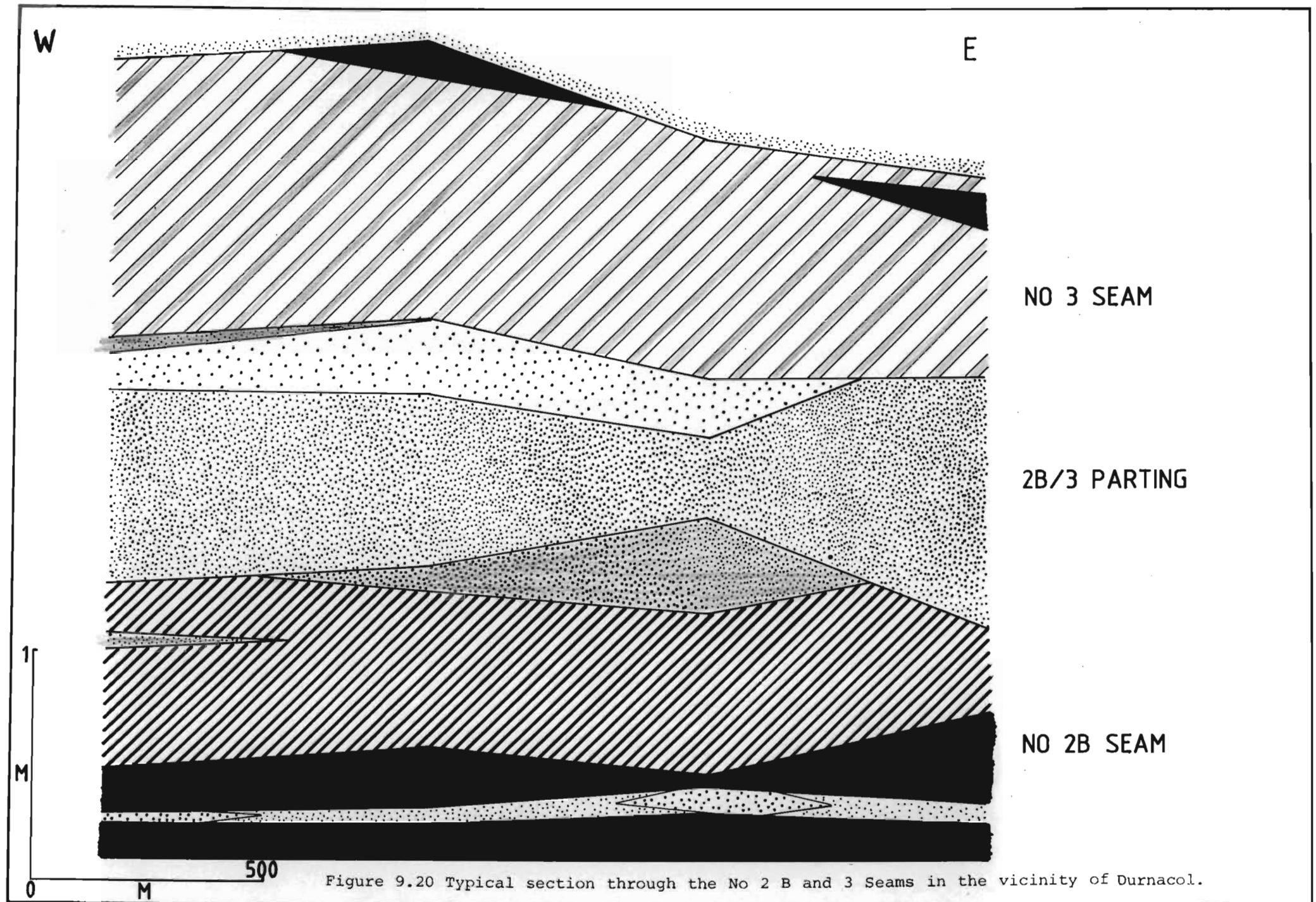
Figure 9.20 Typical section through the No 2 B and 3 Seams in the vicinity of Durnacol.

where the parting is thinner and argillaceous in composition. Although the seam
is generally thinner in the southern part of the coalfield, this also holds
true to the west and north of Glencoe. However, between Burnside colliery and
Wasbank the seam thins where it overrides the Glencoe sheet sandstone (Fig
9.6).

9.8.2   Environmental Association and Controls on Accumulation

Sedimentary associations suggest that the No 3 Seam accumulated on an
abandoned alluvial plain. Evidence supporting this includes:

1.   The composition of the coal seam, particularly in the vicinity of
     sheet-sandstone margins, gives no indication of peat accumulation
     being interrupted to any extent by clastic incursions.

2.   The thickest parts of the seam invariably overlie sheet sandstones.
     This indicates that either peat swamps were established on fluvial
     channels before the overbank area (ie. complete cessation of fluvial
     activity), or peat swamps were established over the entire area
     simultaneously but peat accumulated more rapidly over the channels.

Although the seam's high proportion of dull coal and sparse clastic partings
may be attributed to peat accumulation in an ombrogenous, raised swamp,
these features may be equally explained by formation in a low-lying swamp on a
stable platform undergoing a low rate of subsidence and devoid of clastic
deposition (cf. Smythe, 1980; Flores, 1981). Considering the history of the
platform with respect to spatial variations in stability and subsidence rates,
it is probable that both conditions prevailed: a stable platform combined with
slight doming of the swamp within the main coalfield (Fig 9.16) contributed to
a generally low water-table and preferential formation of inertinite-group
macerals under mildly oxidizing conditions. Ponds and small lakes (near
Dannhauser) testify to a water-table level at or near the surface of the swamp.

In contrast to the No 3 Seam, most "fluvial" coal seams are reported to thin when ascending fluvial-channel deposits (Ferm and Cavaroc, 1968, Cairncross, 1980; Howell and Ferm, 1980; Le Blanc Smith, 1980b). This is generally attributed to a low water-table over topographically positive channel fills undergoing proportionally less compaction than adjacent overbank mudrocks. The apparently anomalous behaviour of the No 3 Seam is believed to be a function of differential platform loading and pre-existing controls on platform stability and subsidence:

1. In the previous section (9.7) it was proposed that sheet sandstones preferentially occupied topographically lower parts of the alluvial plain. Implicit in this statement is that the intervening overbank stretches were more stable and thus slightly elevated (the same mechanisms were put forward to explain the absence of the No 1 and 2A Seams in this area). It is thus possible that the water-table level over the topographically lower sheet sandstones was higher than adjacent areas.

2. Enhanced compactional subsidence of the platform underlying the sheet sandstones induced by sediment loading on a generally stable alluvial plain contributed to higher rates of peat accumulation over topographically depressed areas of the platform.

3. The greater permeability of the sheet sandstones and regular recharge by source-area precipitation would have ensured a water-table level adequate for thick peat accumulations. Water-table recharge of the argillaceous overbank deposits (proximally and laterally) would have been inhibited by their low permeability and porosity.

The inverse relationship between the No 3 Seam and 2B/3 parting thickness, ie. the seam thins as it ascends the sheet sandstone, between Burnside and

Wasbank (Fig 9.19) is probably linked to the seam's shaling-out and non-development towards the southwest. Both are believed to be a function of rapid subsidence rates and increased groundwater salinities in the vicinity of the margin of the coalfield. It is suggested that the widespread establishment of swamps was inhibited by a combination of a greater rate of platform subsidence (see Le Blanc Smith, 1980b) in the distal parts of the plain, and the frequent incursion of saline surface and ground waters (the latter promoted by high permeability of the sheet sandstone).

Relatively thick coal to the southeast of the Dundee-Vryheid line (especially with respect to the earlier-formed seams) is a function of increased platform stability, and apparently no influxes of saline water. The preponderance of bright coal reflects a higher water-table level than further to the north but the great variability in coal seam composition may be attributable to temporal and spatial variations in platform stability, groundwater salinities and vegetation species. Figure 7.13 indicates that the No 3 Seam sedimentary platform in this area comprises 2 sheet sandstones with an aggregate thickness of approximately 35 m. It is proposed that the normally high gross rate of subsidence (compactional and tectonic) experienced in this area was retarded to some extent as a result of the low compactibility of this coarse-sand platform.

9.9   THE NO 3/4 CLASTIC PARTING

The No 3/4 clastic parting maintains a fairly consistent thickness of about 25 m over the study area, but thicknesses of between 15 and 25 m have been recorded (Figs 7.11 and 7.12). The parting comprises alluvial plain deposits and, as such, no bioturbation is present. Two sedimentary components are recognized : braided fluvial-channel sandstones (up to a few hundred metres in width) and sheet sandstones, and overbank deposits consisting of sheet-flood and crevasse-splay sandstones, pond and lake fills, and rare coals. Over most

of the coalfield the parting shows a sheet sandstone between 8 and 15 m thick overlying overbank deposits of 0,5 to 10 m thickness. However, in many boreholes this pattern is not evident and there is an apparently random interfingering of overbank mudrock and fluvial sandstone deposits. There appears to be only a weak en echelon relationship between the position of sheet sandstones of the No 2B/3 parting and fluvial deposits of this interval: distribution of fluvial channels was apparently not controlled to any extent by pre-existing topographic or subsidence effects.

The inundation of the No 3 Seam peat swamps by braided rivers was undoubtedly related to renewal of source-area fluvial activity. The maintainance of alluvial plain conditions indicates that sediment aggradation exceeded platform subsidence. The absence of any major sedimentary trends is believed to reflect an overall decrease in platform subsidence rates and perhaps also reflects the rapidity with which this depositional event took place; sedimentation was not in equilibrium with basement and platform movements.

9.10   THE NO 4 SEAM

The No 4 Seam represents the final phase of regional peat accumulation in the field area. Although the seam rarely exceeds 100 mm thickness, and is mostly between 10 and 20 mm, it displays great lateral persistence over the entire coalfield (Figs 7.11, 7.12, 9.6 and 9.7). Obviously, it has no economic significance whatsoever but its wide lateral extent serves as an invaluable upper marker to the Coal zone in exploration drilling. This feature gave rise to it being named the "Marker" Seam.

The formation of a thin coal over a wide area at a single horizon can only be reconciled with peat accumulation in swamps occupying a stable, almost featureless plain devoid of clastic sedimentation. Its thinness is attributed to the short period of time allowed for the seam's development on a platform

that was undergoing only limited subsidence. Peat accumulation was terminated by the initiation of a further episode of alluvial-plain sedimentation.

9.11   THE UPPER ZONE

The Upper zone sequence reflects the final phases of extensive terrestrial sedimentation and a change in the relationship between platform subsidence and basinal tectonics. A further extensive phase of alluvial-plain sedimentation led to the deposition of between 8 and 40 m of fluvial channel and overbank deposits. There followed a complete cessation in sedimentation during which the entire area was transgressed. The wide extent of this transgression, which is reflected by prodelta mudrocks overlying fluvial sandstones, may have been as a result of an eustatic rise in basinal water-level. However, a mechanism by which platform subsidence exceeded sediment supply, and thus platform aggradation, and resulted in a relative rise in basinal water level is possibly a more plausible mechanism (See Hobday, 1973, Mason and Tavener-Smith, 1978). As the Upper zone was intersected only in boreholes, no comment on wave processes accompanying the transgression can be made.

The transgressive surface is overlain by up to 4 prograding coastal sequences. Thicknesses vary from 15 to 60 m. In borehole core most of these appear to be of deltaic origin but R. Tavener-Smith (pers. comm.) has identified wave-dominated shoreface sequences in the vicinity of Nqutu to the east of the Buffalo River. This period of renewed sediment supply and coastal progradation saw the periodic establishment of delta plain environments over the field area. Sediment supply was, however, unable to keep pace with basin subsidence over an extended period. A thick sequence (in excess of 60 m) of offshore and/or prodelta deposits, generally extensively bioturbated and termed the "Upper Transition zone" by Blignaut and Furter (1940), reflects the extent to which the basin deepened at the end of Vryheid Formation times.

CHAPTER 10

SUMMARY OF CONTROLS ON PEAT ACCUMULATION IN THE SOUTHERN KLIP RIVER COALFIELD

It is clear from the foregoing discussions that coal-seam distribution, thickness and quality are not influenced by the depositional style of associated clastic facies. However, characteristics of the coal seams are certainly related to the sedimentary history of the Coal zone. In contrast to the numerous authors listed in Chapter 8 who categorised coal-seam behaviour and geometry on the basis of supposedly contemporaneous clastic depositional environments, only two broad groups of coal are recognised here:

1. Those coals that accumulated on abandoned delta or alluvial plains essentially devoid of clastic sedimentation, and

2. coals formed in swamps located on a prograding coastal platform (on the fringes of lagoons and embayments) or in temporarily abandoned river channels on an active alluvial plain, ie. associated with clastic sedimentation.

The former group is characterised by coals that have an areal extent in excess of 2 500 sq km and includes the thickest and economically most significant coals, the No 2B and 3 Seams. Such coals were referred to as "transgressive coals" by Fisher et al. (1969), Le Blanc Smith (1980b) and Galloway and Hobday (1983). The latter group comprises thin, laterally discontinuous coals that are commonly shaly and contain numerous clastic partings.

A comparison of Lower zone and Coal zone sedimentary facies associations suggests that the prime factor responsible for creating an environment capable of supporting prolonged periods of peat accumulation over a wide area was increased platform stability and consequent shoreline progradation to form coastal or alluvial plains of considerable lateral extent.

Economic coal-seam formation over the main coalfield occurred only in the intervals between fluvial sedimentation; successive periods of peat-swamp activity were terminated by the influx of fluvial sand onto the alluvial and delta plains. Thus the <u>duration</u> of the respective periods of peat accumulation was a function ultimately of the periodicity of sediment supply from the source area. Probable source-area processes which exerted a control on sediment supply included tectonic and/or variations in climatic patterns. Peat accumulation was terminated by sediment influx and <u>not</u> by the development of an unfavourable balance between platform subsidence and peat-swamp aggradation.

There is no doubt, however, that spatial variations in subsidence rate were <u>the</u> major control on coal-seam distribution and thickness. While variations in basement subsidence provided a primary control on regional peat distribution, differential compactional subsidence is manifested by local variations in seam geometry. The relationship between the thickness of the No 3 Seam and the underlying No 2B/3 parting perfectly illustrates this mechanism, though superficially, the relationship appears to contradict the general rule. This also serves to focus on two other aspects of sediment-compaction mechanisms:

1. Lateral variations in lithology, and thus compactional subsidence of sediments <u>immediately</u> underlying peat swamps, did not appear to significantly affect peat accumulation rates owing to their limited degree of loading. Earlier-deposited sediments exert a greater influence on compaction rates.

2. The mass of any sedimentary unit (a function of lithology and thickness) at surface or immediately below a swamp provokes a more immediate response from compactional subsidence by way of loading. The manner in which underlying sediments react to this loading determines local compactional subsidence rates and lateral variations therein.

Variations in sediment compaction were reflected by platform topography and this, to a large degree, influenced groundwater levels. Thickest coals formed where enhanced platform subsidence was matched by peat accumulation and swamp aggradation. Areas where platform subsidence exceeded peat accumulation throughout most of Coal zone sedimentation, such as existed mostly to the southeast and west of the Dundee-Vryheid and Newcastle-Ladysmith lines respectively, swamps foundered and are represented as carbonaceous lake or embayment mudrocks. The lithology of sediments on which swamps were founded effected peat accumulation rates in a number of ways. Not only is the permeability and porosity of the sediment important, but so is its nutrient content and ability to promote plant growth. Although fluvial-channel sandstones provided an ideal aquifer by which swamp groundwater levels were recharged, the continuous influx of fresh, oxygenated water may have had adverse reactions on peat maturation. This may have been one of the factors responsible for the predominance of coals with a high inertinite content (dull coals).

Probably the most significant control on seam thickness was the length of time available for each successive period of peat accumulation. As stated in an earlier chapter, it is futile to attempt to provide an absolute estimate of the duration of coal formation; the potentially large number of variables which ultimately controlled peat accumulation make such an exercise impractical. Lateral variations in thickness of individual coal seams testify to the substance of this argument. However, it is tentatively proposed that while the coal seams represent fairly lengthy periods of peat accumulation (in the order of tens of thousands of years), the intervening periods of clastic sedimentation were relatively short-lived.

Erosion of peat by fluvial currents does not appear to have been a major process in determining coal-seam thickness or distribution. This, as indicated earlier, is attributed to the unconfined nature of the alluvial and delta plains and tendency for bed-load dominated rivers to widen rather than deepen their channels.

Parameters influencing the composition of coal seams are similar to those outlined in Chapter 8. The mechanism by which these controls determined coal composition are enumerated below:

1. The predominance of vitrain or fusain in a coal seam is a function of water-table level which is in turn determined by the lithology and topography of the peat-forming platform. Equally important is the balance between water-table level and net subsidence rate.

2. Eh and pH of groundwater appear to have exerted a significant influence on coal-seam composition as well. Coals formed in a marginal swamp or marsh are invariably dull and shaly. A possible contributory factor to the predominantly dull nature of the No 3 Seam is the continual recharging of swamp waters by fresh, neutral to alkaline water introduced via the underlying sheet-sandstone aquifer.

3. Coals formed in raised-swamps, possibly such as the No 3 Seam, contain few clastic partings. In contrast, parts of the No 1 and 2A Seams were formed in association with an actively prograding coastline contain a large number of clastic partings and have a high mineral-matter content.

Structural features of the sedimentary platform were invariably propagated upwards through the Coal zone. Examples of this include the similarity in the position of the No 2B/2C Seam split, the position of the northwestern margin of the Glencoe sheet sandstone and the shaling out the No 3 Seam southwest of Glencoe; the position of lake/embayment deposits in both the No 2B and 3 Seams; the similarity in the distribution of the No 1 and No 2A Seams; and the amalgamation of the No 2B and 3 Seam over an area where No 1 and 2A Seams are only sporadically developed.

10.1   COMPARISON WITH OTHER COALFIELDS

The Klip River coalfield is unique in that it records coal formation on the margin of the stable sedimentary platform within the northeastern Karoo basin; to the south and west of the coalfield coals thin and shale-out rapidly (Fig 9.18). One or two thin and impersistent seams are developed in the Muden-Tugela Ferry area (Hobday, 1973) and there is no record of any significant coals west of the Dannhauser Trough (van Vuuren and Cole, 1979). Borehole core from the vicinity of Ladysmith examined by the author shows a similar pattern. On the other hand, coalfields with a more proximal aspect display a greater number of coal seams with a greater aggregate thickness than the Klip River coalfield. In order to illustrate differences in coal-seam behaviour and distribution, some aspects of factors governing peat formation in the Vryheid and Utrecht coalfields (Fig 10.1) and the Witbank coalfield (Fig 10.2) are examined below.

10.1.1   Comparison With The Vryheid and Utrecht Coalfields

The most striking differences between the Klip River coalfield and these more proximal coalfields are the generally greater thickness of the Coal zone, greater number of coal seams and more complex coal-seam behaviour (Roberts, 1985; Smith, 1985). With respect to the latter difference, although coal seams appear to be of equal lateral extent to those of the Klip River coalfield, they are subject to a much greater degree of splitting. The greater thickness of the Coal zone and greater number of coal seams suggests that coal formation started earlier and extended over a longer period than the Klip River coalfield. This can be ascribed to the generally greater stability of the more proximal peat forming platforms which was, in turn, a function of:

1.   basement stability; the coalfields were situated further from the loci or axes of basement subsidence (tectonic and isostatic), and

2.   the lesser thickness of the sediment pile underlying the coal platform (see Fig 6.4) meant that it underwent less compactional subsidence than the Klip River coalfield sedimentary sequence.
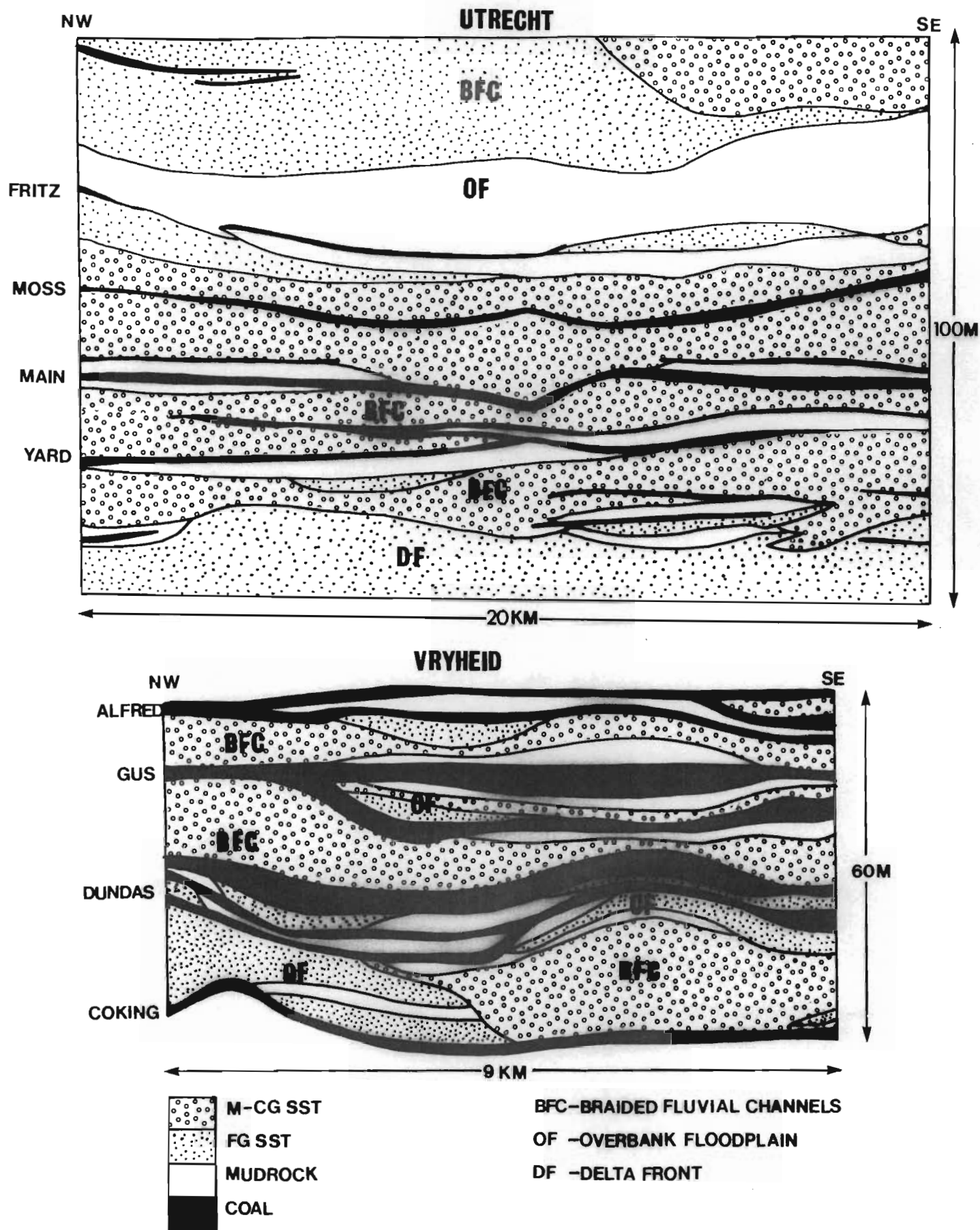
Figure 10.1. Sections illustrating coal seam nomenclature and inferred depositional environments in the Utrecht and Vryheid coalfields (after Roberts, 1985 and Smith, 1985).
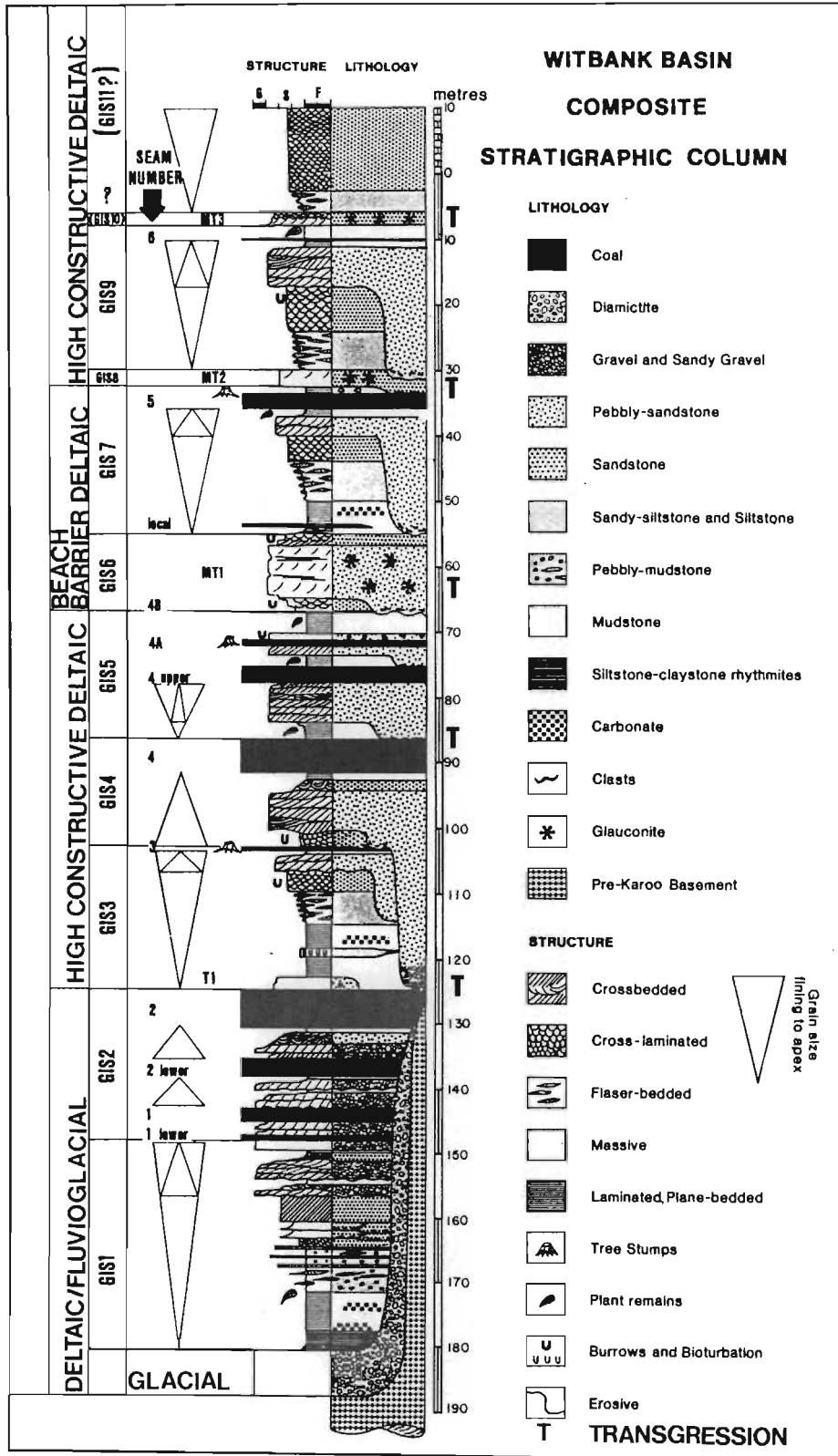
Figure 10.2. Section illustration coal seam nomenclature and depositional environments of coal seams in the Witbank coalfield (after Le Blanc Smith, 1980b).

It is thus apparent that the more complex seam behaviour displayed by these coalfields is <u>not</u> a function of platform stability but rather some other factor. This is believed to be the style and distribution of fluvial sedimentation which was a function of proximity to the source area, higher platform gradients and greater fluvial-current energies. Although fluvial sheet-sandstones are ubiquitous within these proximal areas, it is suggested that fluvial currents were restricted to a relatively small number of major channels at any one time. This allowed peat accumulation to continue unabated over some parts of the plain and it appears that in many instances large tracts of the fluvial system, temporarily abandoned, were colonised by swamp vegetation (Roberts, 1985). Over the area now occupied by the Klip River coalfield, lower alluvial-plain gradients allowed the rivers to spread and form an extensive network of coalescing, weakly-confined river channels (within the major fluvial tracts). The sheet sands thus deposited terminated peat accumulation over the entire plain. Only rarely were swamps established within fluvial channels. In most other respects the controls on peat accumulation were similar to those recognised in the Klip River coalfield.

10.1.2   Comparison With The Witbank Coalfield

The Witbank coalfield is situated on the northern edge of the Karoo basin (Fig 1.2). It comprises up to 180 m of sediments that contain 6 coal seams (Le Blanc Smith, 1980b). Most of these coals were deposited in proximal glacio-fluvial and deltaic settings. A comparison of controls on peat accumulation revealed the following:

1.   Exploitable coal seams within the Klip River coalfield are restricted essentially to a single horizon. Within the Witbank coalfield significant coals occur throughout the Vryheid Formation. This difference is attributed to a greater degree of basement stability and the fact that the successive coal-forming platforms overlaid considerably thinner sequences of sediments than the platform of Klip River coalfield (and were thus subject to a lesser degree of compactional subsidence).

2.   Periods of peat accumulation in the Witbank coalfield were invariably terminated by transgression and swamp drowning. In the Klip River coalfield peat accumulation was brought to an end by the influx of river-borne sand. This illustrates a fundamental difference in mechanisms controlling peat formation, one related to intrabasinal stability and the other, the Klip River coalfield platform, to relatively rapid basin subsidence with the duration of peat accumulation manipulated by source-area controlled fluvial activity.

3.   Seam distribution and thickness within the Witbank coalfield is strongly influenced by pre-Karoo topographic relief: a mountainous terrain characterised by glacially-eroded valleys. The influence basement relief had on differential compaction is two-fold. Firstly, by inducing lateral variations in thickness of sediments underlying the respective coals and secondly, by the tendency for fluvial-channel sandstones to be preferentially deposited along pre-Karoo valleys. Although data pertaining to basement palaeorelief of the Klip River coalfield is sparse, it is likely that it was more subdued and that its influence on peat accumulation was almost entirely negated by the thickness of the underlying sediment pile. Differential basement subsidence and differential compaction rates were the chief mechanisms by which coal distribution and thickness were controlled in this coalfield.

4.   Northern Natal coals generally have a higher vitrinite content than those coals deposited in the Witbank coalfield (Falcon, 1986b). This suggests that northern Karoo basin peats underwent a greater degree of oxidative degradation than their Natal counterparts. The following characteristics of the Witbank coalfield may have contributed to this:

a)   A cooler prevailing climate (Le Blanc Smith and Eriksson, 1979; Falcon, 1986a)

b)   greater stability of the peat-forming platform led to an imbalance in peat-swamp aggradation and water-table levels,

c)   proximity of swamps to influxes of cold, oxygenated surface- and groundwater, and

d)   fluvial-channel activity contemporaneous to peat formation.

5.   Peat erosion by fluvial channels confined by pre-Karoo valleys played an important role in coal-seam distribution and thickness in the Witbank coalfield. The rarity of peat erosion in the Klip River coalfield is attributed to the flat, featureless nature of the alluvial plains during fluvial sedimentation and tendency for the weakly-confined channels to widen laterally rather incise vertically.

6.   In the previous section it was tentatively proposed that the duration of peat-forming periods was longer than intervening periods of fluvial sedimentation. Le Blanc Smith (1980b) believed the opposite to be true in the Witbank coalfield. He also attributed variations in No 4 Seam thickness over the coalfield to variations in the duration of peat accumulation. In this respect he makes no reference to the numerous other variables effecting seam thickness, in spite of recognising the role of differential compaction in determining coal-seam geometry.

10.2   CORRELATION OF COAL SEAMS WITHIN THE NORTHEASTERN KAROO BASIN

Correlation of coal seams between coalfields located in northern Natal and over the entire northern and northeastern Karoo basin has been the subject of relatively few investigations (van Vuuren and Cole, 1979; Cadle, 1982; Tavener-Smith, 1983) but has, on numerous occasions, been the subject of heated debate. Although a basin-wide correlation of coal seams is beyond the ambit of this thesis, a number of factors having a possible bearing on the subject are speculated on below:

1. Most major coal seams developed in northern Natal accumulated in swamps on abandoned alluvial plains. Peat accumulation at any one time was probably terminated almost simultaneously over the entire region by influxes of fluvial sand. It is thus proposed that successive periods of peat accumulation in the Vryheid, Utrecht and Klip River coalfields are now represented by laterally extensive coal seams that can be traced beyond the boundaries of the respective coalfields and represent chronostratigraphic marker horizons within the Coal zone.

2. Sedimentation in the northern Natal and Transvaal coalfields was controlled by source areas that were geographically, climatically and tectonically distinct (Ryan, 1967; Falcon, 1986a). Significant differences in platform stability and controls on coal-seam distribution also prevailed. In view of the fundamental differences in mechanisms controlling the timing and duration of peat accumulation it appears to be highly unlikely that individual coal seams can be traced over the entire northeastern part of the Karoo basin.

10.3   APPLICATION OF FINDINGS TO EXPLORATION AND MINING

Over the last decade the relationship observed between sedimentary facies and potential coal-mining problems and coal quality has been applied with varying degrees of success (Horne et al., 1978; Carrucio and Geidel, 1978; Houseknecht and Iannacchione, 1982). The application of the findings of this investigation to exploration strategy, resource evaluation and mine planning take due cognisance of these principles and are outlined below:

1. At the very least this investigation provides a framework by which coal seams can be identified, named and correlated over the coalfield. The value of understanding coal seam behaviour, distribution and thickness on a regional scale during exploratory drilling or for the purposes of a resource evaluation cannot be underestimated.

2. An understanding of the distribution of the sheet sandstones forming the No 2B/3 parting give an accurate indication of where the No 2 and 3 Seams amalgamate to form the "Main" Seam. Similar patterns can be expected to the north and east of the coalfield. Neither the No 2B nor 3 Seams approach mineable thicknesses to the west of the Newcastle-Ladysmith line. However, it is suggested that to the west of this line, where the two seams are amalgamated in a triangular shaped area bounded by the margins of the Dannhauser and Dundee sheet sandstones to the north and south respectively, a potentially economic "Main" Seam exists.

3. Southeast of the Dundee-Vryheid line only the No 3 Seam has economic potential.

4. Due consideration should be given to the locality of pond and channel deposits within the No 2 and 3 Seams in the vicinity of Dannhauser; inadequate borehole density-spacing could result in an over-estimation of in-situ reserves with serious consequences. The locality of these instraseam clastics should be taken into consideration during mine planning to facilitate mining methods and strategy, shaft siting and tunnel routes.

5. On the basis of the sedimentological history of peat accumulation it is expected that sulphur content of the coal seams should decrease upwards through the Coal zone, and from the eastern and southern

(distal) margins towards the central (proximal) parts of the
coalfield. Likewise, it is suggested that the Klip River coalfield
has a generally higher sulphur content than the more proximally
situated Vryheid coalfield.

6. The internal consistency of individual coal seams over extensive
    areas with respect to composition and clastic partings provides an
    excellent tool for seam correlation over and above normal
    stratigraphic and sedimentological methods. The utilization of this
    characteristic in exploration is obvious; at mine-planning stage it
    provides an indication of the reliability of interpolated geological
    data.

7. Complex and dangerous mining conditions can be expected to develop
    adjacent to areas where the parting between the No 2 and 3 Seams
    changes thickness and lithology (cf. Horne et al., 1978; Houseknecht
    and Iannachione, 1982). The following problems, related to
    differential compactional mechanisms, may arise:

    1) slickensided contacts between and within sandstone and mudrocks,
        propogated into the coal seams,

    2) severe jointing, and

    3) steep gradients within the coal seams as they ascend and descend
        in the vicinity of the margins of sheet sandstones.

    The first two of these problems will give rise to severe roof
    instability, while the third will impose some limitations on
    underground transport and mining methods.

8. The greater permeability of the sheet sandstone forming the parting
    to the No 2 and 3 Seams may have implications with respect to
    underground-water drainage and the escape or infusion of methane gas.

9. For the same reason cited in No 8 above, coals enclosed either partially or totally by sheet sandstone deposits are predicted to contain a higher diagenetic mineral-matter content than those associated with predominantly argillaceous roof and floor rocks.

10. The coalfield can be divided into a number of "mining provinces". On the basis of geometry, behaviour and composition (or lithology) of coal seams and clastic partings, similarities or differences in mining conditions or problems can be foreseen and extrapolated.

CHAPTER 11

SUMMARY

The coal-bearing Vryheid Formation of the Klip River coalfield records early to late Permian fluvio-deltaic sedimentation within the northeastern main Karoo basin. Three informal lithostratigraphic subdivisions, based on the work of Blignaut and Furter (1940, 1952), are proposed: the Lower zone, Coal zone and Upper zone.

An examination of the structural framework and history of the northeastern Karoo basin reveals that the southern and western boundaries of the Klip River coalfield are demarcated by zones of rapid basement subsidence: the Tugela and Dannhauser Troughs respectively. There is some doubt as to the geographic locality of the source area to the rivers emptying into the Ecca sea. Although Ryan (1967) postulated the "Eastern Highlands", situated off the present southeast African coast, as a source area, it is here contended that the Swaziland area constituted a more plausible source area. This implies that the source area was situated some 200 to 300 km to the northeast of the Klip River coalfield.

The Lower zone represents deposition along a westerly to southeasterly prograding coastline dominated by high-constructive lobate or braid deltas. During early stages of deposition only short stretches of the coast, situated between active delta lobes, were influenced by wave processes. However, during the closing stages of Lower zone sedimentation, long stretches of the southern and southeastern coast were fronted by low beach-barriers. This pattern of sedimentation was in response to substantially reduced sediment supply to that part of the coast and increased basement stability. In contrast, increased basin stability along the western and southwestern coastlines was manifested by major phase of progradation and an extensive delta plain, dominated by coalescing braided fluvial-distributaries, was established. The observed cyclicity within the Lower zone is attributed to repeated shoreline regressions and transgressions as a result of rapid and frequent autocyclic switching of delta lobes superimposed on a high rate of basin-floor subsidence.

The Coal zone represents a major phase of coastal progradation and deposition on extensive alluvial plains over the main part of the coalfield. Sedimentation was characterised by periods of high-energy, bed-load dominated rivers (that formed sheet sandstones) followed by a total cessation of fluvial activity and the establishment of peat-forming swamps. Significant coal seams formed only during these periods of inactivity, the duration of which was dependent entirely on source-area processes. Early Coal zone sedimentation to the southeast of the Dundee-Vryheid line was dominated by wave processes and peats accumulated in lagoons and barred embayments.

A review of modern peat-forming environments reveals that the terminology is non-uniform and confused, especially with respect to the term "swamp". This frequently makes it difficult to compare the numerous peat-forming models which have appeared in the literature in recent years. A comparison of some of the more recent investigations into coal-forming models was undertaken in an attempt to identify some of the fundamental mechanisms controlling coal-seam distributions and thicknesses in sequences ranging in age from Permian to Cretaceous. This showed that the relationship between implied depositional settings of coals and their geometry and composition is, in a global sense, highly variable. The models established by Horne et al. (1978) and later investigators can not be universally applied and other mechanisms are required to explain coal-seam characteristics. What emerges is that the presence or absence of clastic deposition contemporaneous to peat accumulation, the lithology and topography of the floor to peat-swamps and the degree and rate of platform subsidence play a far greater role in determining coal-seam geometry and composition than depositional processes of associated clastic environments. Also, many coals are formed in an environment distinctly different from that indicated by underlying and overlying sediments.

It is evident that the depositional history of vertically-associated clastic sediments had little or no influence on the distribution and thickness of coal seams in the Klip River coalfield; factors influencing coal seam geometry were

complex and inter-related. The following were found to have played a major role in determining the extent, distribution and rate of peat accumulation:

1. Platform stability; temporal and spatial variations in platform subsidence (basement and compactional) rates.

2. The absence or presence of penecontemporaneous clastic sedimentation.

3. The duration of periods of peat formation.

4. Lithology and topographic expression of clastic sediments underlying peat-forming swamps.

Generally, the economically significant coal seams, the No 2 and 3 Seams, accumulated on alluvial plains devoid of clastic sedimentation and undergoing very low rates of subsidence. As such, they are laterally persistent and cover areas in excess of 1 500 sq km. In contrast, those seams associated with penecontemporaneous sedimentation in a marginal and rapidly subsiding area of the platform, such as the No 2A and 2B Seams to the southeast of the Dundee-Vryheid line, are thin, lenticular and contain numerous clastic partings.

The peat-forming phase of the Vryheid Formation was brought to an end by increased rates of basement subsidence and/or a eustatic rise in basinal water-level. Sediment supply was unable to keep pace with this relative rise in basin water-level and an extensive transgression ensued. Upward-coarsening deltaic and wave-dominated sequences testify to the re-establishment of a coastal setting at the close of Vryheid Formation deposition.

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Figure 7.1. Loch End outcrop section, river valley, Loch End farm, 5 km south of Alcockspruit.

Figure 7.2. LYELL outcrop section, river valley, Lyell farm, 7,5 km west of Wasbank.

Figure 7.3. STEIN COAL SPRUIT outcrop section, river valley, Stein Coal Spruit farm, 7 km west of Wasbank.

Figure 7.4. TALANA outcrop section, river valley immediately southwest of Talana mine.

Figure 7.5. MORGENSTOND outcrop section, river valley northeast of homestead, Morgenstond farm, 11 km northeast of Dundee.

Figure 7.6. DALRAY outcrop section, river valley north and east of homestead, Dalray farm, 10 km northeast of Dundee.
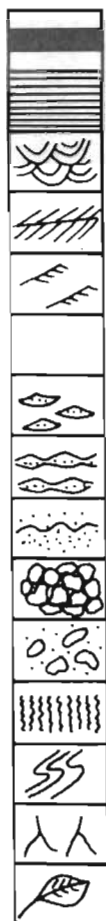
Figure 7.7. Gowrie outcrop section, river valley, 15 km eastsoutheast of Dundee.

Figure 7.8. WINSTONE outcrop section, river valley, Winstone farm, 20 km southeast of Dundee.

Figure 7.9. CINDERFORD outcrop section, river valley, northeast of homestead, Cinderford farm, 19 km southeast of Dundee.

Figure 7.10. KRANTZKOP outcrop section, river valley and cliff faces, west of homestead, Krantzkop farm, 12,5 southeast of Wasbank.
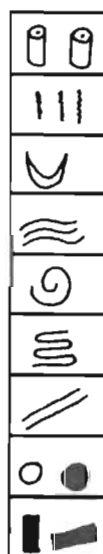
KEY

C – COAL

Sf – FLAT STRATIFICATION

St – TROUGH CROSS-BEDDING

Sp – PLANAR CROSS-BEDDING

Sr – RIPPLE CROSS-LAMINATION

Sm/Fm – MASSIVE BEDDING

Fl – LENTICULAR BEDDING

W – WAVY BEDDING

EROSIONAL BASE

Cc – CLAST SUPPORTED CONGLOMERATE

Cm – MATRIX-SUPPORTED CONGLOMERATE

Sb/Fb – BIOTURBATED SEDIMENT

Sd/Fd – DEFORMED BEDDING

ROOTLETS

PLANT FOSSILS

GRAIN SIZE

CLAYSTONE
SILTSTONE
FINE-GRAINED
MEDIUM-GRAINED
COARSE-GRAINED
VERY COARSE-GRAINED

{ SANDSTONE

BIOTURBATION

SIPHONICHNUS

TIGILLITES/SKOLITHOS

DIPLOCRATARION

NEREITES

SPIRODESMOS

HELMINTHOPSIS

SCOLICIA

PLANOLITES/PALAEOPHYCUS
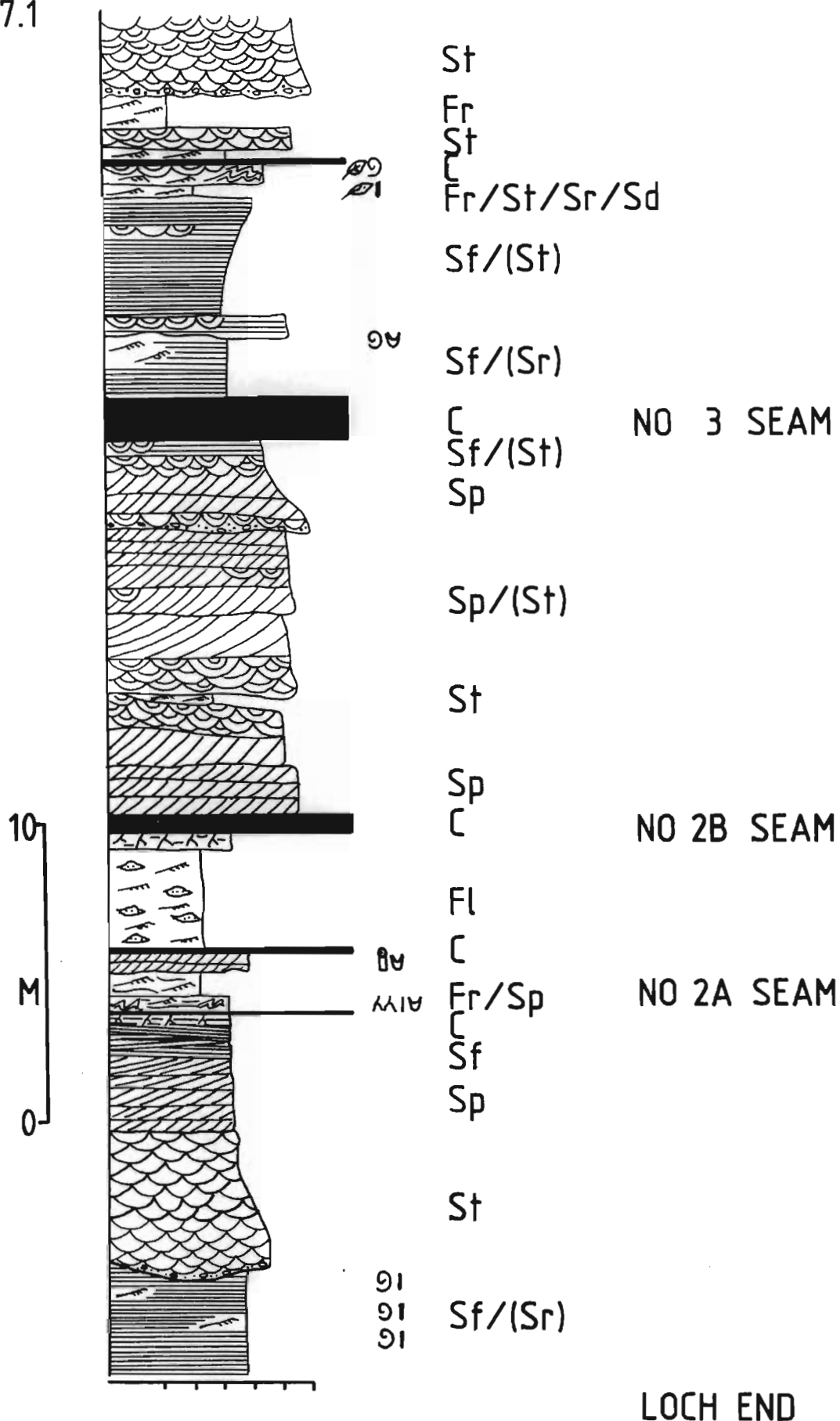
UNIDENTIFIED TRACES

KEY TO SECTIONS

FIGURE 7.1

St
Fr
St
C
Fr/St/Sr/Sd
Sf/(St)
Sf/(Sr)
C   NO 3 SEAM
Sf/(St)
Sp
Sp/(St)
St
Sp
C   NO 2B SEAM
Fl
C
Fr/Sp   NO 2A SEAM
C
Sf
Sp
St
Sf/(Sr)

LOCH END

FIGURE 7.2

Er
Sf
Er
St
Ff/ Fr
St
Sr/Fr
Fm/Fr
C   NO 3 SEAM
Sp/(St)
St
Sp
Cm
C   NO 2C SEAM
Fm
Fr
Ff
Fm
C   NO 2B SEAM
Fr/Sr
Al/W
Sp
Sf/Sm
C/Fm
FrSd
C   NO 2A SEAM
Fr/Ff
Sf
St
Sf/Sm
GCB
Ff

10
M
0

c s f m c vc

LYELL

FIGURE 7.3

Sp/(St)

Fm
St — NO 2C SEAM

Fr

Sf — NO 2B SEAM
C

10

M

0

Ff/(Fr)

St — NO 2A SEAM
C

Sp/(St)/(Fm)

Ff/Al
St
St/Sr

Fm
St C — NO 1 SEAM
Al

St/(Sf)

Sp

St

Ff
Fl   STEIN COAL SPRUIT

FIGURE 7.4

Sp/St

St

Sp

Sf

Sp/St
Cc

Al/W

Sp
St
Sr/Sf

Sp
Cm

Sp/St

Cm
Sf
Cm

TALANA

10

M

0
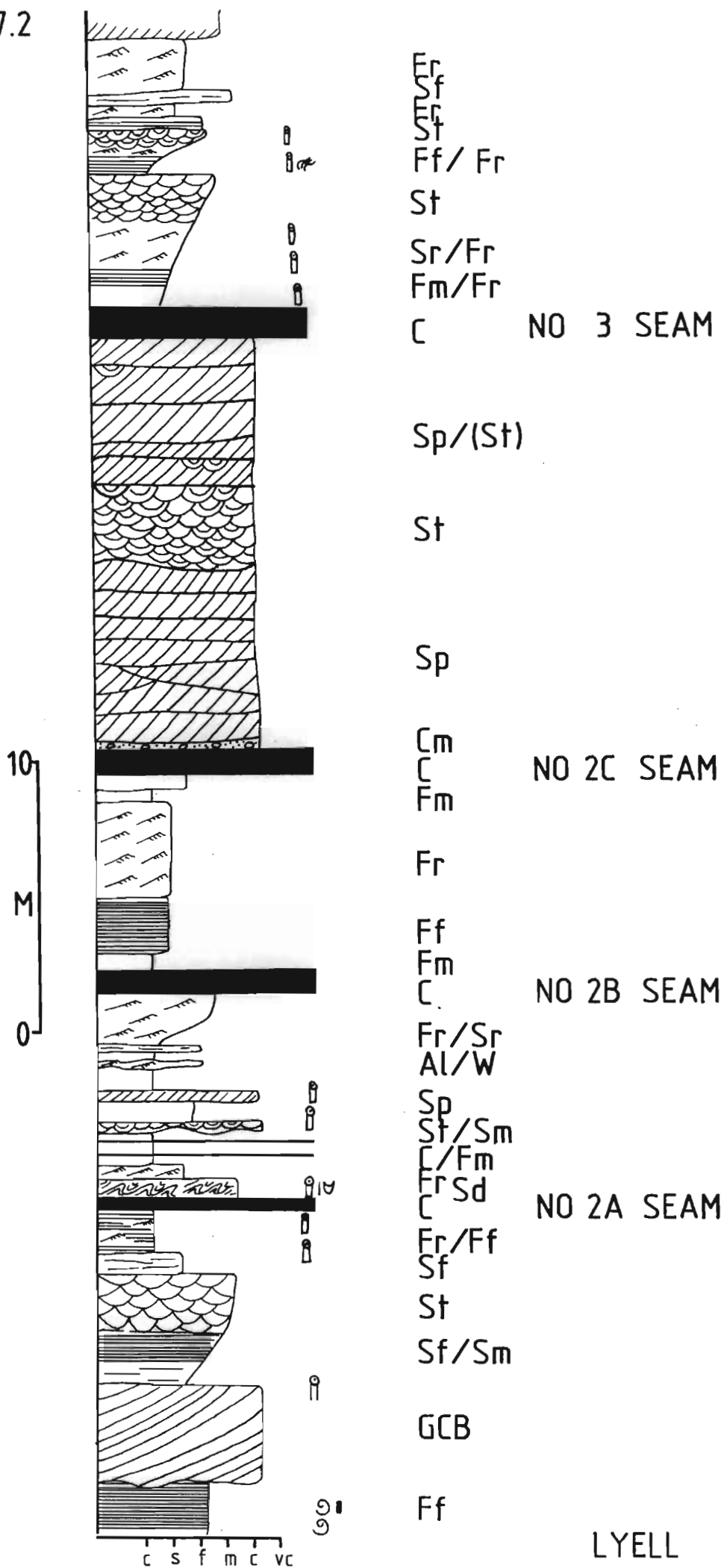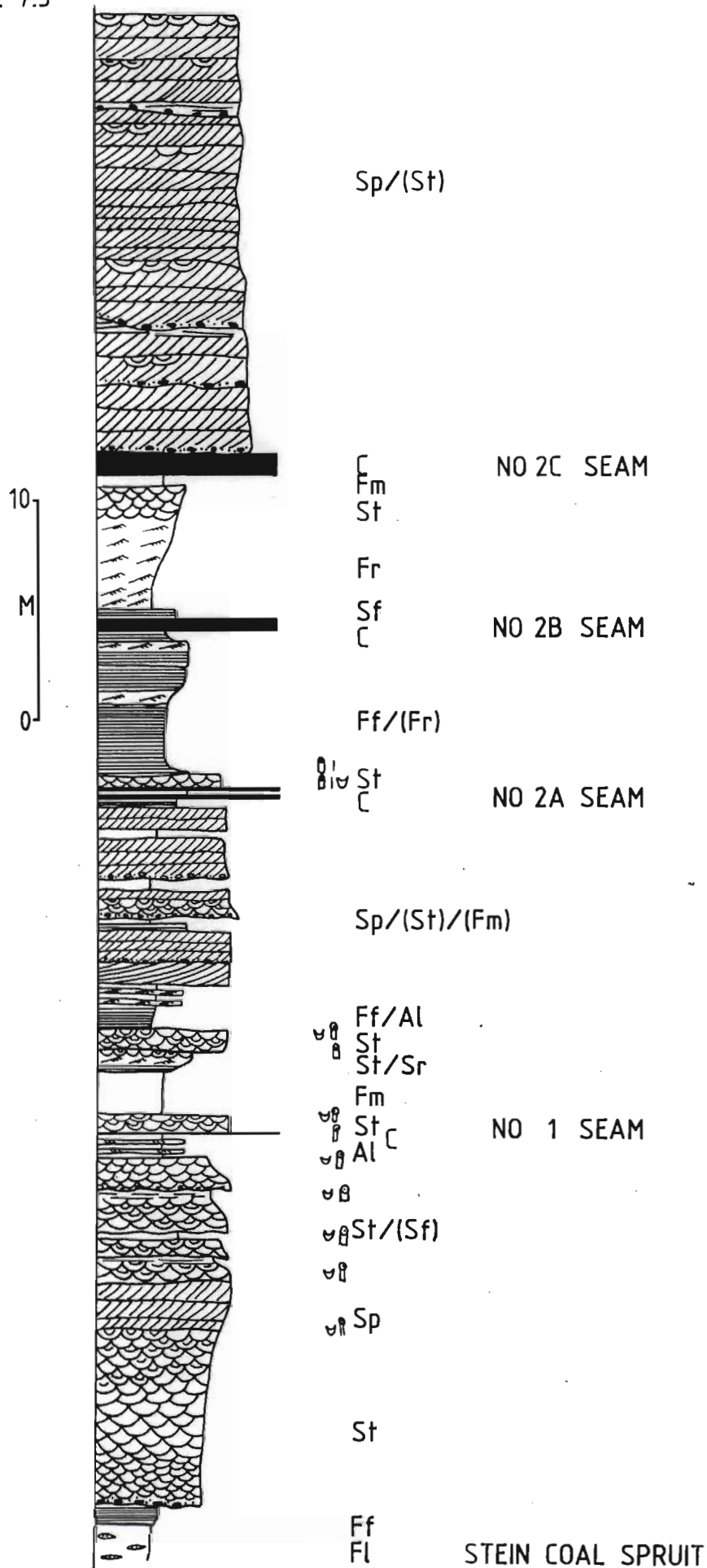
c s f m c vc

FIGURE 7.5
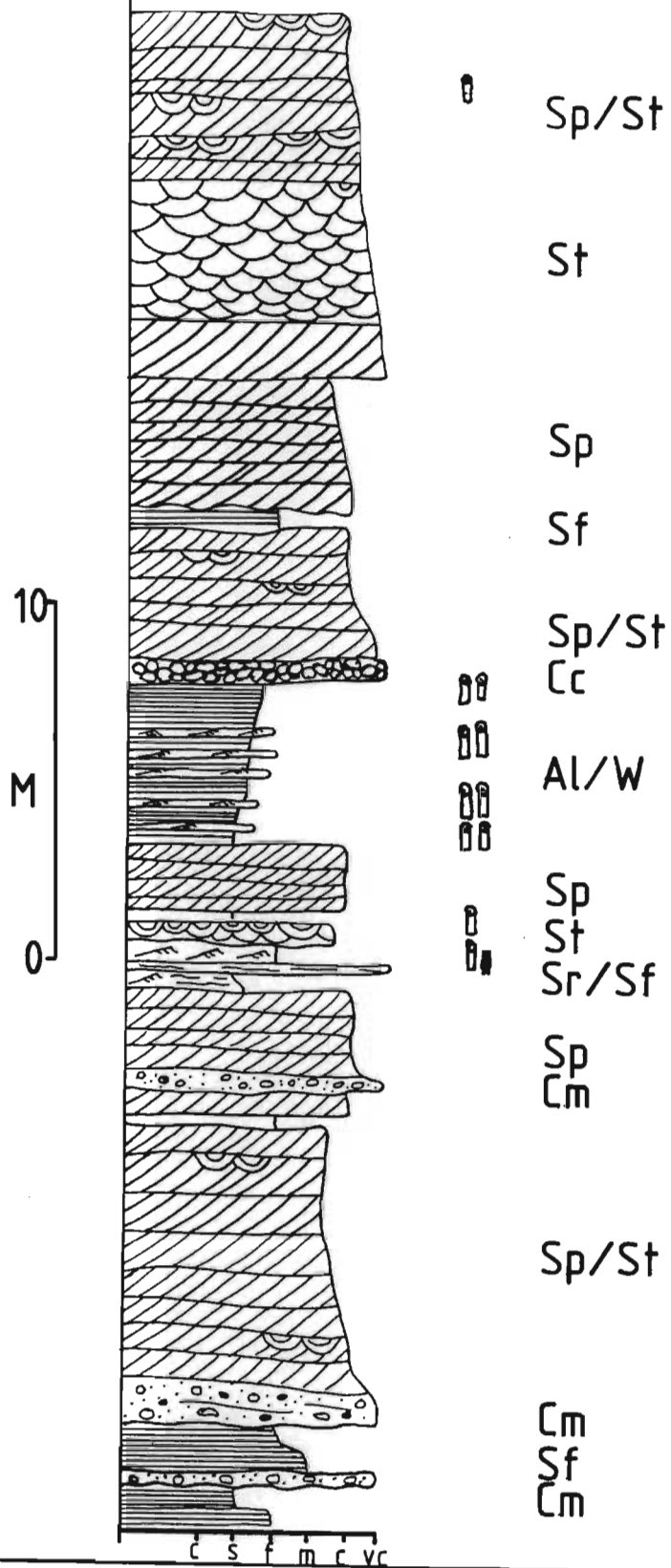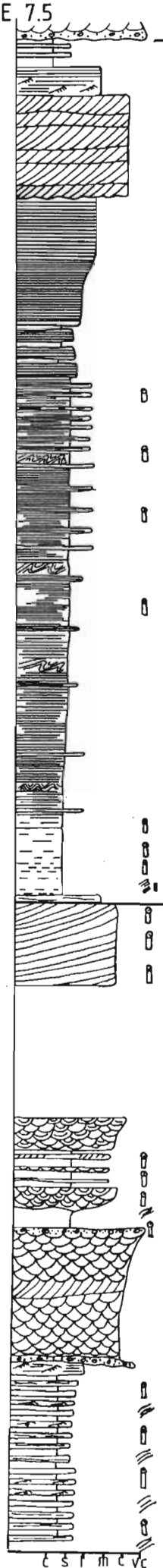
Left column (labels, top to bottom):
Al
Sf/Sr
Sp
Sf
AlS
Al
AlF(Sf/Ff)
Fm/Ff
C
GCB
St
Al
St
Fm
St/(Sp)
Cm
Al

Scale: 20 — m — 0

Right column (labels, top to bottom):
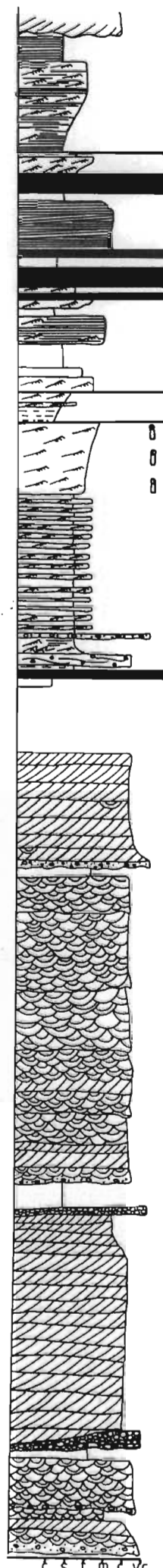Sp
Ff
Fr/Ff
C
Fr   NO 3
C
Sf
C   NO 2B
Sf/Sr
Fm
Sr
C   NO 2A
Fr
Al
Cm
Sm/Sf   NO 1
C
Sp
Cm
St/(Sp)
Cc
Sp
Cc
St
St/Sp

MORGENSTOND

FIGURE 7.6

Sp
C   NO 2B SEAM
Sr/Fm

Sp/(St)

Cm
C   NO 2A SEAM
St
Sr
Sf

Sf/Sr
Fm/Ff
St
Sf/Sr

Sf

Ff/Fm
Fm

St
C
Ff
C
Fm/Ff/Fr
C
Sf/Sp/Fr
C   ?NO 1 SEAM
Sf/Sr/St

Sf/(Sr)/(Sp)

Ff/(Fr)

Fm
St   DALRAY

10

M

0

FIGURE 7.7

St
GCB
Sf

St /(Sp)

Sf
St
Sf

AlS

Al

AlF

GCB

St /(Sr)

St /(Sr)

Sp

Al

20

M

0

Sp /(Sf) /(Sm)
Al
Sb
C
St /Sp
Sp

St /(Sp)

Sf

AlS

Fm /(Ff)

NO 3

NO 2B

NO 2A

C

C

C
C

St /
(Sp)

St
Sp
St

C
Sf /Ff
Sp
St
Ff /Fr
Sr
St
Sm

Sf /
(St)

Sf

Ff /
Fr

GOWRIE

FIGURE 7.8

NO 1

NO 2B

NO 2A

WINSTONE

FIGURE 7.9

CINDERFORD

St

Ff/Fr/Al

Sp/St

Sf/Sr/St

AlS

Al

AlS

Al

AlF

Fm/Fb

20

M

0

St

St

Sp/Sr/Sf

Sp

Fm/Fl

Sp/St

Fm/Fl/Fr

Sp/St

NO 2A

C

Fr

Sf/St

Sf

Ff
Sf

Ff

Al

Sf/Sp

Sf

Sf

Ff

Ff/Fr

Fm/Ff

NO 3

NO 2B

Sp/St

C

Sf

Sp/St

C.

Sf/Ff
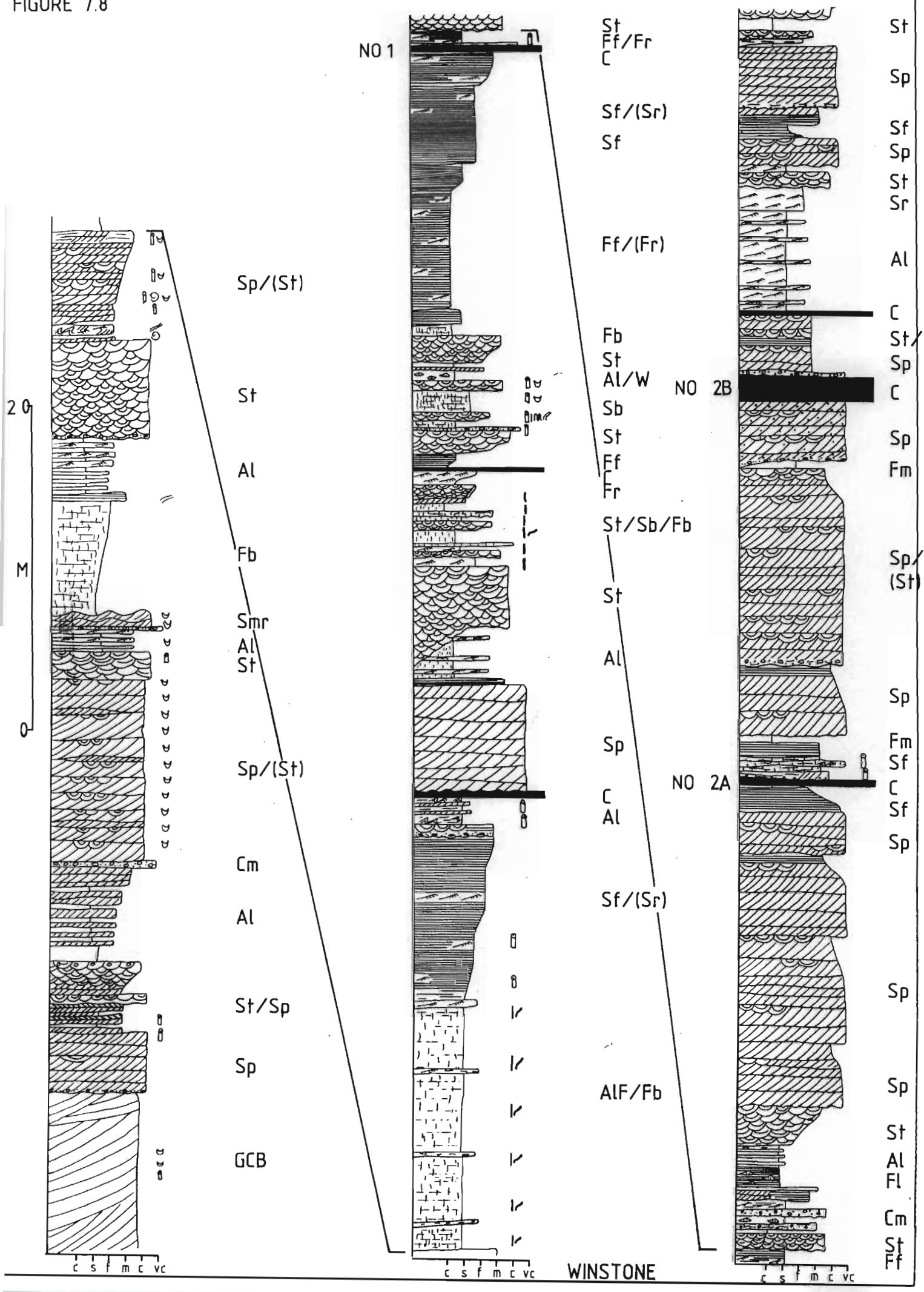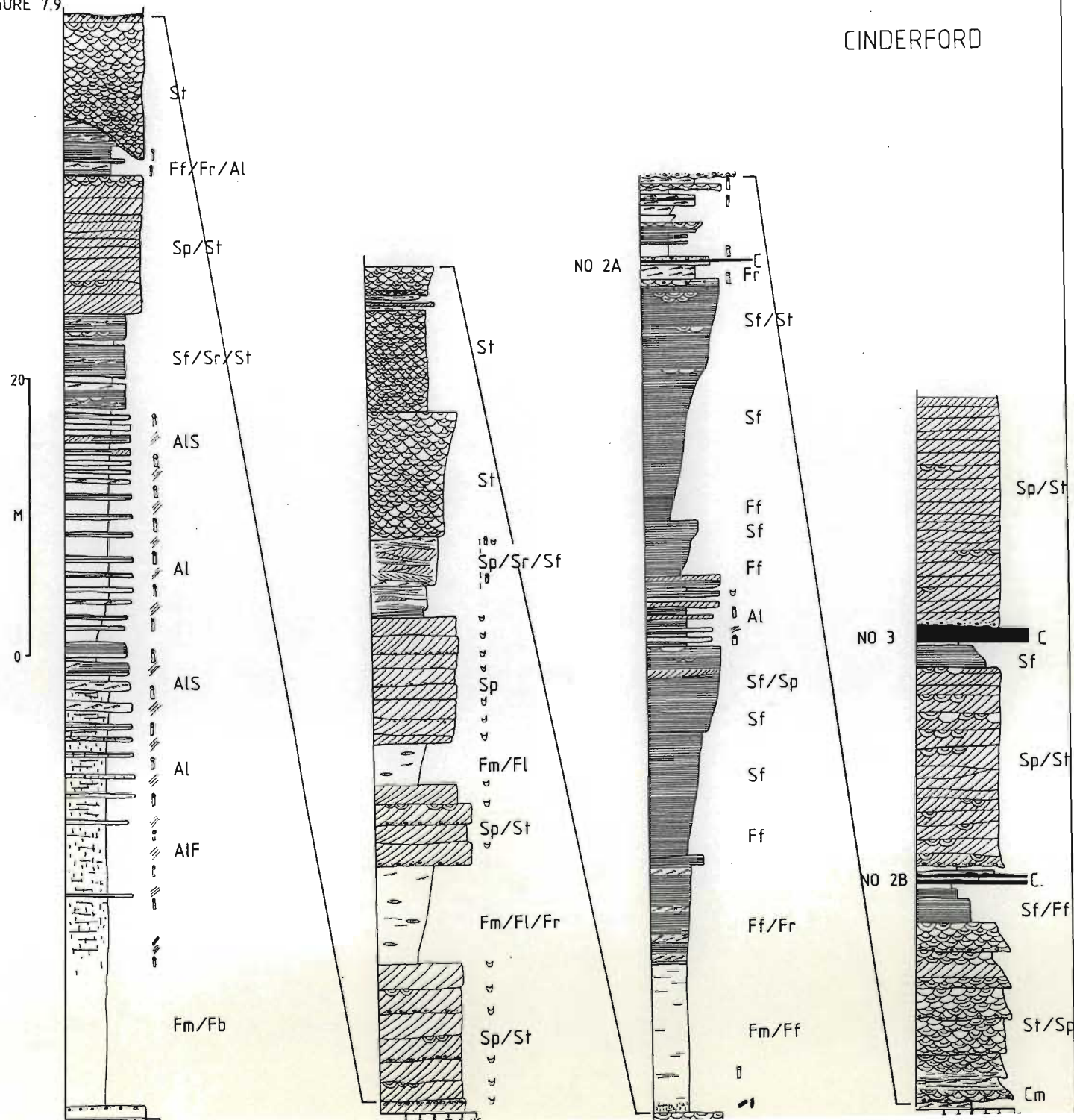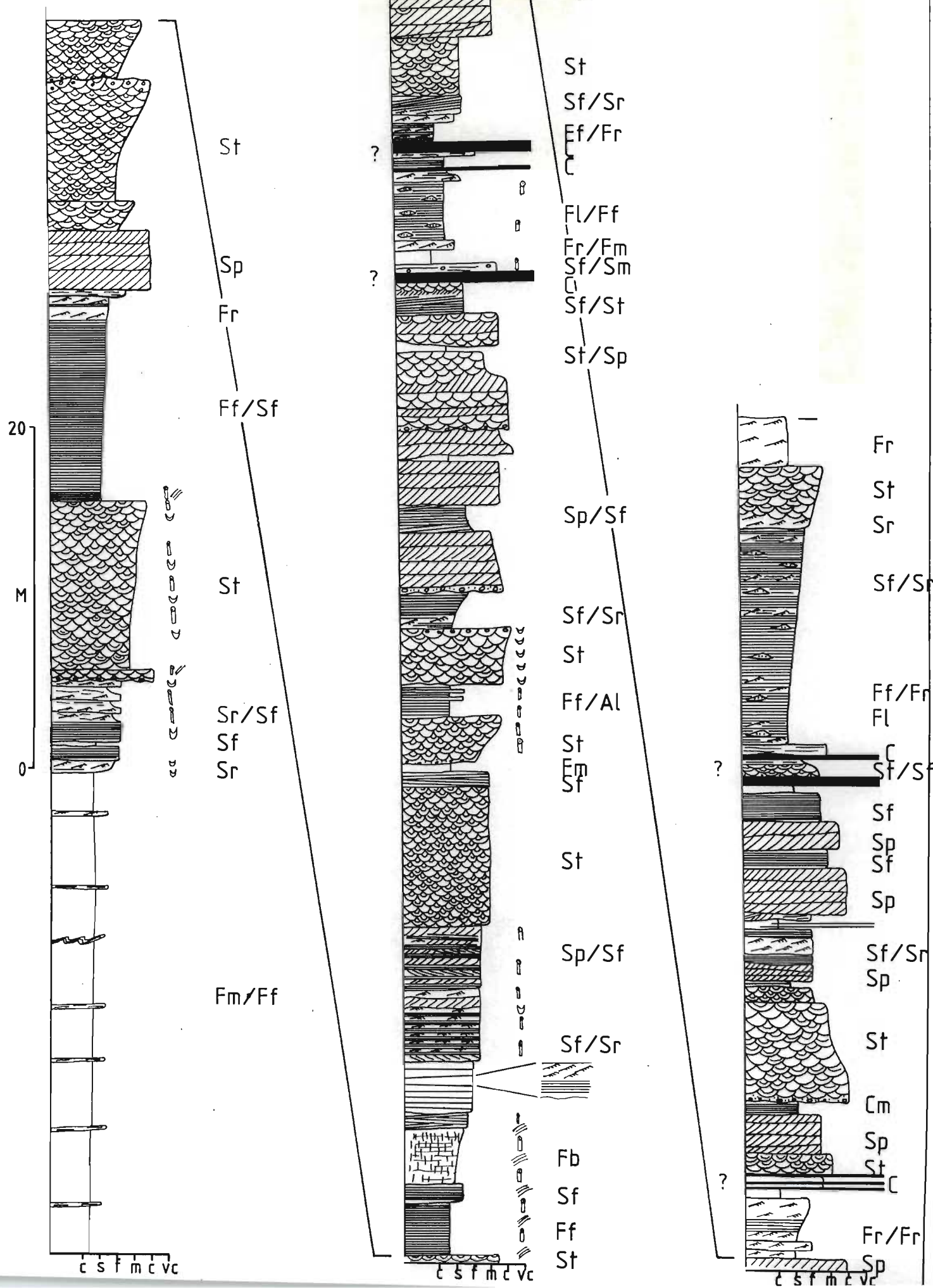
St/Sp

Cm

FIGURE 7.10

KRANTZKOP.

APPENDIX 1

ADDENDUM TO PALAEOCURRENT DATA PRESENTED IN FIGURE 6.6

NUMBER	STRATIGRAPHIC HORIZON	LOCALITY	TYPE OF CROSS BEDDING (SUBORDINATE IN BRACKETS)	DEPOSITIONAL ENVIRONMENT	NUMBER OF READINGS	MEAN ORIENTATION
1	LOWER ZONE	LOCH END	TROUGH AND PLANAR	DELTAIC	28	256°
2	LOWER ZONE	MORGENSTOND	TROUGH AND PLANAR	DELTAIC	56	205°
3	LOWER ZONE	BURNSIDE	TROUGH AND PLANAR	FLUVIAL, DELTAIC	85	250°
4	LOWER ZONE	TALANA	PLANAR AND (TROUGH)	FLUVIAL	38	185°
5	LOWER ZONE	INDUMENI	TROUGH AND PLANAR	DELTAIC	39;22	148°;225°
6	LOWER ZONE	STEIN COAL SPRUIT	TROUGH AND (PLANAR)	FLUVIAL, DELTAIC	12;13	165°;224°
7	LOWER ZONE	GLEN LYON	PLANAR AND TROUGH	DELTAIC	36	202°
8	LOWER ZONE	KRANTZKOP	TROUGH AND (PLANAR)	FLUVIAL, DELTAIC	103	185°
9	LOWER ZONE	DALRAY	TROUGH	DELTAIC	17	187°
10	LOWER ZONE	GOWRIE	TROUGH AND (PLANER)	DELTAIC	69	134°
11	LOWER ZONE	HLUNGWANA	TROUGH AND PLANAR	DELTAIC	56	145°
12	LOWER ZONE	CINDERFORD	TROUGH AND PLANAR	FLUVIAL, DELTAIC	98	140°
13	LOWER ZONE	WINSTONE	TROUGH AND (PLANER)	FLUVIAL, DELTAIC	185;45	132°;194°
14	COAL ZONE	LOCH END	PLANAR AND (TROUGH)	BRAIDED FLUVIAL	27	229°
15	COAL ZONE	MORGENSTOND	PLANAR AND (TROUGH)	BRAIDED FLUVIAL	32	212°
16	COAL ZONE	BURNSIDE	TROUGH AND PLANAR	BRAIDED FLUVIAL	37	248°
17	COAL ZONE	TALANA	PLANAR AND (TROUGH)	BRAIDED FLUVIAL	49;57	257°;161°
18	COAL ZONE	GOWRIE	TROUGH AND (PLANAR)	BRAIDED FLUVIAL	27	202°
19	COAL ZONE	CINDERFORD	TROUGH AND PLANAR	BRAIDED FLUVIAL	29	176°
20	COAL ZONE	WINSTONE	PLANAR AND (TROUGH)	BRAIDED FLUVIAL	66	168°
21	CIAL ZONE	STEIN COAL SPRUIT	PLANAR AND (TROUGH)	BRAIDED FLUVIAL	18	270°

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