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SEDIMENTOLOGICAL STUDIES IN THE LATE PRECAMBRIAN AND LOWER CAMBRIAN ROCKS OF EAST FINNMARK

Volume 1

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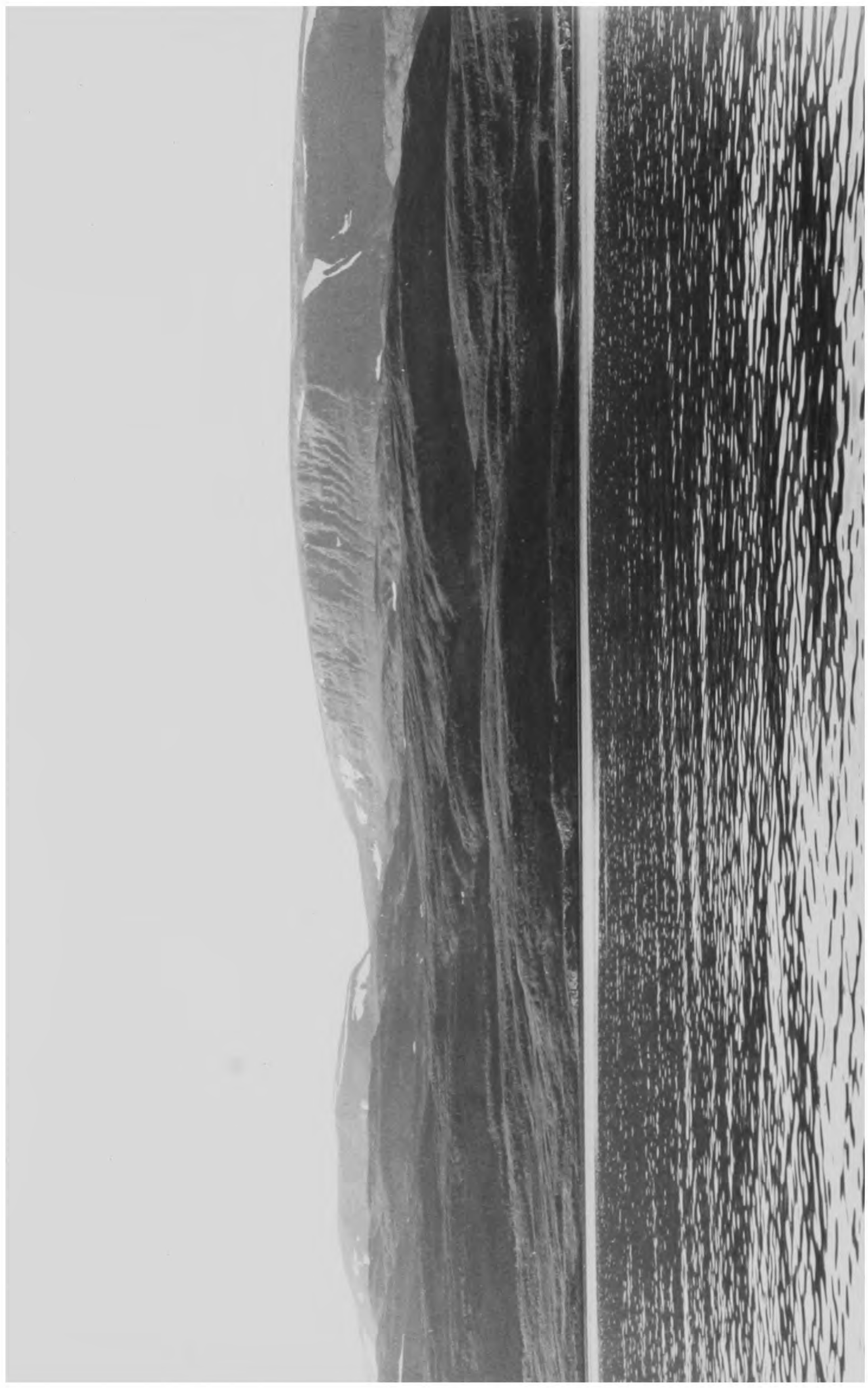
View of the Digermul Peninsula from the head of Tanafjord. The late Precambrian Upper Tillite outcrops along the coast and the overlying succession can be followed up the hill, the highest Lower Cambrian beds forming the escarpment and plateau of the peninsula. The plateau is at about 600m above sea level.



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ABSTRACT

In eastern Finnmark a northward thickening wedge of late Precambrian and Cambro-Ordovician sedimentary rocks extends 200km from east to west, with a maximum north-south width of approximately 50km. The succession lies unconformably upon Precambrian crystalline basement, which crops out to the south. The maximum thickness is between 4,000m and 5,000m with shallow water sandstones and shales predominating. Two distinct tillite formations occur, beneath each of which there is an unconformity shown only by local downcutting and slight regional discordance. Otherwise the succession is conformable, deposition having occurred in a gently subsiding basin.

Structural deformation is very slight in the east but increases towards Laksefjord in the west where the rocks are strongly folded. In the far west of the area two distinct units are present, the Gaissa Nappe consisting of folded Precambrian sandstones and, underlying this, the autochthonous undeformed Dividal Group (previously the "Hyolithus Zone") of late Precambrian and Lower Cambrian age.

This study is concerned with the six members which form almost the entire succession from the top of the Upper Tillite to the top of the Lower Cambrian. These members are the Innerelv amd Manndraperelv Members of the Stappogiedde Formation, the Lower and Upper Members of the Breivik Formation and the Lower and Upper Members of the Duolbas-gaissa Formation

The lower three members are found over a wide area and can be correlated with the Members of the Dividal Group in the extreme southwest of the region. Their maximum thick-ness of about 775m is found on the Digermul Peninsula and gradual thinning occurs towards the southwest, the thickness of the Dividal Group being about 230m. It is possible that

some slight thinning also occurs on the Varanger Peninsula to the southeast of the Digermul Peninsula. The upper three members are confined to the Digermul Paninsula and have a maximum thickness of about 900m.

The majority of the sediments are thought to have been deposited in a variety of offshore shallow marine environments. Within such environments it is considered that there are five main processes by which sand can be transported. These are i) semipermanent currents, ii) tidal currents, iii) wave drift currents, iv) coastal storm surge currents, v) river-generated currents. From the various known and predicted features of the deposits of each type of current it is possible to deduce which was the most important in transporting and depositing sediment in any given sequences of beds.

The base of the Innerelv Member marks a widespread transgression as a result of which quiet water mudstones replaced the coarse fluvial sandstones of the Lillevatn Member. The Innerelv Member consists of mudstones, siltstones and very fine sandstones and has been divided into six intergradational facies, five of which form a series of gradually increasing environmental energy and possibly increasing proximity to a shoreline. In the highest energy facies (I.5) strong currents cut channels but deposited little sand within them, the main sediments being irregularly bedded siltstones and mudstones. In the lower energy facies the beds have a more sheet like form. Palaeocurrent directions where measurable, are unimodal except in the highest energy facies where they are variable.

Coastal storm surge currents are believed to have deposited the majority of the sandstone and siltstone beds in this member but the evidence is insufficient to rule out a river generated or tidal origin.

The Manndraperelv Member consists of three bands of red and white sandstone separated by two bands of green mudstone and siltstone with interbedded graded sandstones. The lowest sandstone band follows gradationally from the Innerelv Member in the east but in the southwest, where it is thinner, it sits sharply and probably disconformably upon the underlying member. Sedimentary structures are poorly seen in it and no conclusions were reached as to the mode of sediment transport except that wave activity was probably important in producing the many irregular bedding surfaces which are present.

The two bands of green mudstones and siltstones with graded sandstones pass up into red and white sandstones similar to those of the lower red sandstone band in two coarsening upward sequences. As suggested by Reading (1965) the graded sandstones are probably turbidites. The overlying red and white sandstones are thought to be mainly shallow marine "shelf" deposits and thus the transitions represent shallowing (regressive) sequences. In the first coarsening upward sequence in the most northeasterly exposure sandstones with low angle cross-bedding are interpreted as beach deposits and the siltstones which overlie them are probably lagoonal. These are in turn overlain by a sequence of sandstones in which bed thickness decreases upwards. These sandstones are interpreted as transgressive offshore deposits. The presence of these beds in the most northeasterly exposure fits with the palaeocurrent evidence in the turbidites to suggest that the shelf and shoreline prograded to the southwest or west although in detail the history is more complex.

The two coarsening upward sequences thin to the southwest of the Digermul Peninsula and in theDividal Group are not developed, these sediments probably having been deposited under continuously deep water conditions.

The Lower Breivik Member consists of a complex alternation of siltstones, mudstones and various types of sandstones. On the Digermul Peninsula and in the Leirpollen area the sandstones occur as beds up to 100cm thick and vary in grain size from very fine to coarse sand. Very fine and fine sandstones predominantly show parallel lamination and cross-lamination and have varying amounts of matrix. Medium to coarse sandstones have little or no matrix and are mostly cross-bedded. The majority of the beds have erosive bases and appear to have been deposited by waning currents. The palaeocurrent distribution in the sandstones is bipolar which suggests a tidal origin for these beds. Fluctuations in current strength produced beds of varying grain size and thickness and were the result of natural variation within the tidal cycle, the affect of storms, and the closeness of the basin to resonance. The Lower Breivik Member in the Laksefjord area and the laterally equivalent Member IV of the Dividal Group show some differences from the eastern sections, red sandstones being a distinctive feature of the parts of the Member around Laksefjord.

The Upper Breivik, Lower Duolbasgaissa and Upper Duolbasgaissa Members form an irregular coarsening upward sequence. The Upper Breivik Member consists mainly of green siltstone with beds of very fine sandstone mostly less than 20cm thick. The sandstone beds commonly have erosive bases and are sometimes graded.

The Lower Duolbasgaissa Member consists mainly of thin to medium-bedded fine sandstones with interbedded siltstones and mudstones. However, it also contains a 20m thick unit of cross-bedded fine sandstone to granule conglomerate.

In the Upper Duolbasgaissa Member 5-50m units of cross-bedded sandstone predominate and thinner bedded sandstones are less common. Sets of cross-bedding up to 4m high occur.

Bipolar palaeocurrent directions are found throughout these three members and all gradations can be seen between the various sandstone types. It is suggested that tidal currents were again the most important mechanism of sediment transport. Two distinct axes of transport are present; NW-SE and SW-NE. The change from one tidal pattern to the other was probably caused by changes in basin physiography.

By comparison with recent sediments in the seas around Great Britain the different types of sandstones are believed to have been deposited at various positions along tidal current transport paths. Down present day paths tidal velocities decrease and there is a concomitant decrease in grain size and change in the sedimentary structures. The thickest cross-beds are interpreted as sand wave deposits and smaller ones formed from megaripples.

Palaeocurrent data and lateral facies variation suggests the possibility of a shoreline to the northeast at least in Manndraperelv Member times and in the upper part of the succession. The sandstone petrography suggests that the source area consisted of acid igneous rocks with subsidiary metamorphic and sedimentary rocks. Consideration of the regional geology suggests that although the nearest shoreline may have been to the northeast the source area must have been the Precambrian shield area to the southeast. Where tidal sediments are present a connection to an ocean is inferred.

In the entire late Precambrian and Cambro-Ordovician succession trace fossils first appear in the Innerelv Member. Their abundance and diversity increases rapidly in the latest Precambrian and early Cambrian sediments. The absence of biogenic activity in older sediments is not due to the lack of suitable sedimentary facies. The incoming and diversification of trace fossils reflects principally the development of the Phyla Annelida, Arthropoda and Mollusca. The build up of atmospheric oxygen is considered the most likely fundamental factor controlling the development of life.

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CHAPTER 1

INTRODUCTION

Geological Outline of the Area

The geology of East Finnmark, Norway's most northerly province, is known largely through the work of Holtedahl (1918, 1931), Føyn (1937, 1960, 1967), Reading (1965), Seidlecka and Seidlecki (1967) and Banks et al. (1971). A northward-thickening wedge of late Precambrian and Cambro-Ordovician sediments lies unconformably upon Precambrian crystalline basement rocks which outcrop to the south (Figs. 1, 2). The crystalline rocks are varied in lithology, consisting of gneisses, schists, acid and basic intrusives, metasediments and metavolcanics (Bugge 1960). Several orogenic episodes have been recognised and age dates ranging from >2600m.y. to 1650m.y. have been recorded (Kratz et al. 1968)

To the northwest the sedimentary wedge is bounded by an overthrust "Caledonian" metamorphic belt which is composed of a number of thrust sheets. The lowest and least metamorphosed of these is the Laksefjord Nappe (Føyn 1967) which extends from Laksefjord to the northwest side of Tanafjord. To the northeast the sediments are separated by a tectonic discontinuity (? thrust) from the Barents Sea Group and the Raggo Group (Seidlecka and Seidlecki 1967). The Barents Sea Group is a thick sedimentary series of probable Precambrian age and the Raggo Group, which overthrusts it, is probably laterally equivalent to the rocks of the Laksefjord Nappe (Laird in press).

The sedimentary wedge can be divided into three parts which are, from oldest to youngest, the "Older Sandstone Series", Vestertana Group and Digermul Group (Table 1).

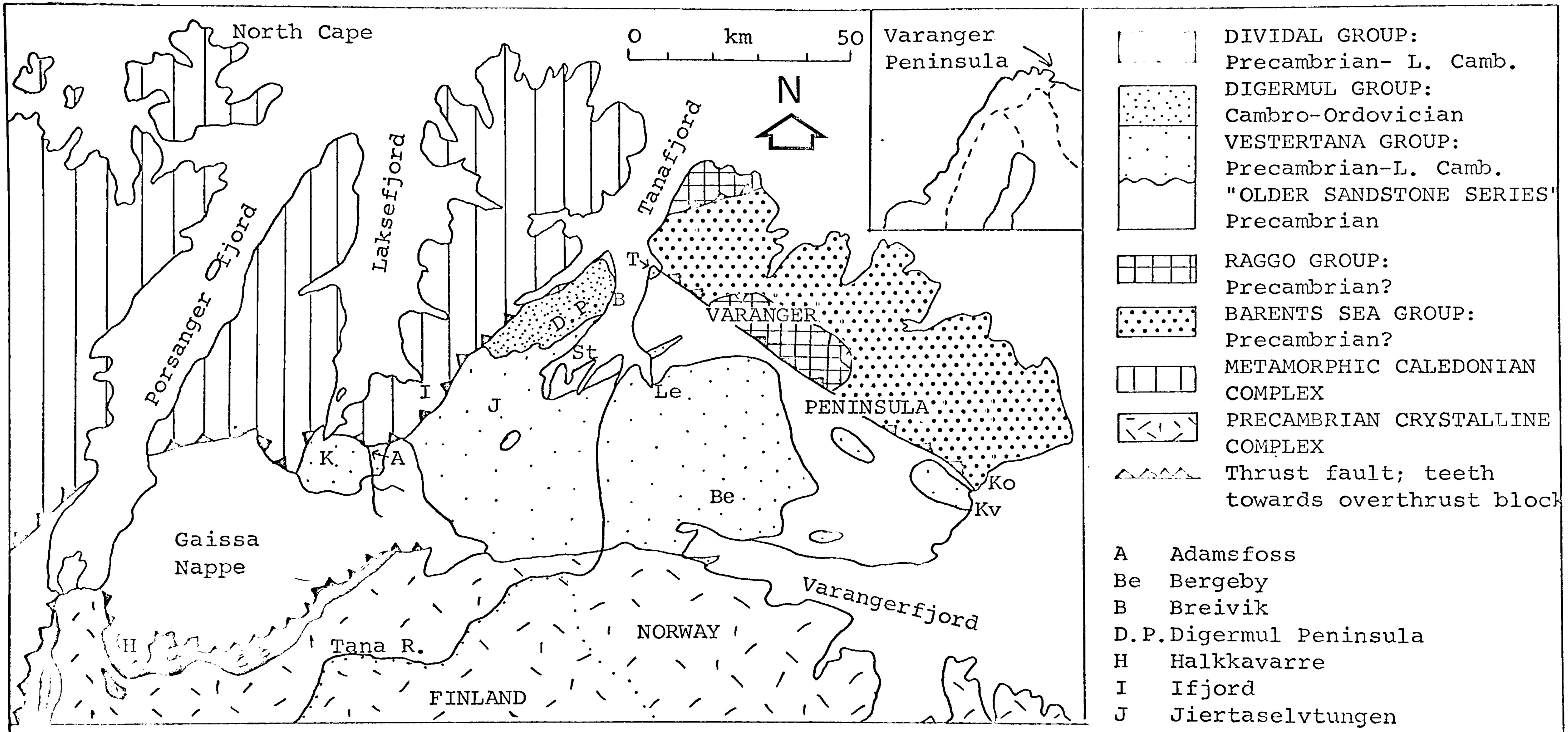


FIGURE 1.

GEOLOGICAL MAP OF EAST FINNMARK

GROUP	FORMATION	MEMBER	THICKNESS average	AGE
DIGERMUL	Berlogaissa		300m	Tremadocian
	Kistedal		720m	M. Cambrian-Tremadocian
	Duolbasgaissa	Upper	350m	L. Cambrian
		Lower	200m	
VESTERTANA	Breivik	Upper	350m	L. Cambrian
		Lower	255m	
	Stappogiedde	Manndraperelv	190m	Precambrian
		Innerelv	300m	
		Lillevatn	40m	
	Upper Tillite		10-50m	
	------------- unconformity ------			
	Nyborg		200-400m	
	Smalfjord Tillite (Lower)		10-50m	
------- regional unconformity of 1-2 degrees ----				
	"Older Sandstone Series"		≥1500m	

TABLE 1. Geological succession of the sedimentary wedge in the Tanafjord area of East Finnmark.

The term "Older Sandstone Series" (Føyn 1937) is used to refer to sandstones and shales with dolomites in the upper part which occur beneath the tillites of the Vestertana Group. The "Older Sandstone Series" crops out in the two main areas, the southern half of the Varanger Peninsula, and between Laksefjord and Porsangerfjord. In the latter area the rocks are allochthonous, forming the Gaissa Nappe of Roberts (1970). In the Varanger region the "Older Sandstone Series", which is >1500m thick, is authochthonous and rests unconformably on the basement at the head of Varangerfjord.

A regional unconformity of 1-2° occurs at the base of the overlying Vestertana Group (Føyn 1937), the erosion being greatest in the south. Reading and Walker (1966) suggested that at Trollfjord, which is the northernmost exposure of the junction, there is a conformable passage from the "Older Sandstone Series" up into the Smalfjord Tillite Fm. of the Vestertana Group. However, M.B. Edwards (personal communication) considers that this junction is also unconformable. At the southern margin of the sediments the Smalfjord Tillite Fm. and the overlying Nyborg Formation overstep the "Older Sandstone Series" and rest directly on the basement. A slight unconformity also occurs at the base of the Upper Tillite Formation (Reading and Walker 1966). The genesis of these glacial and associated sediments have been discussed recently by Reading and Walker (1966), Bjørlykke (1967) and Banks <u>et al</u> (1971).

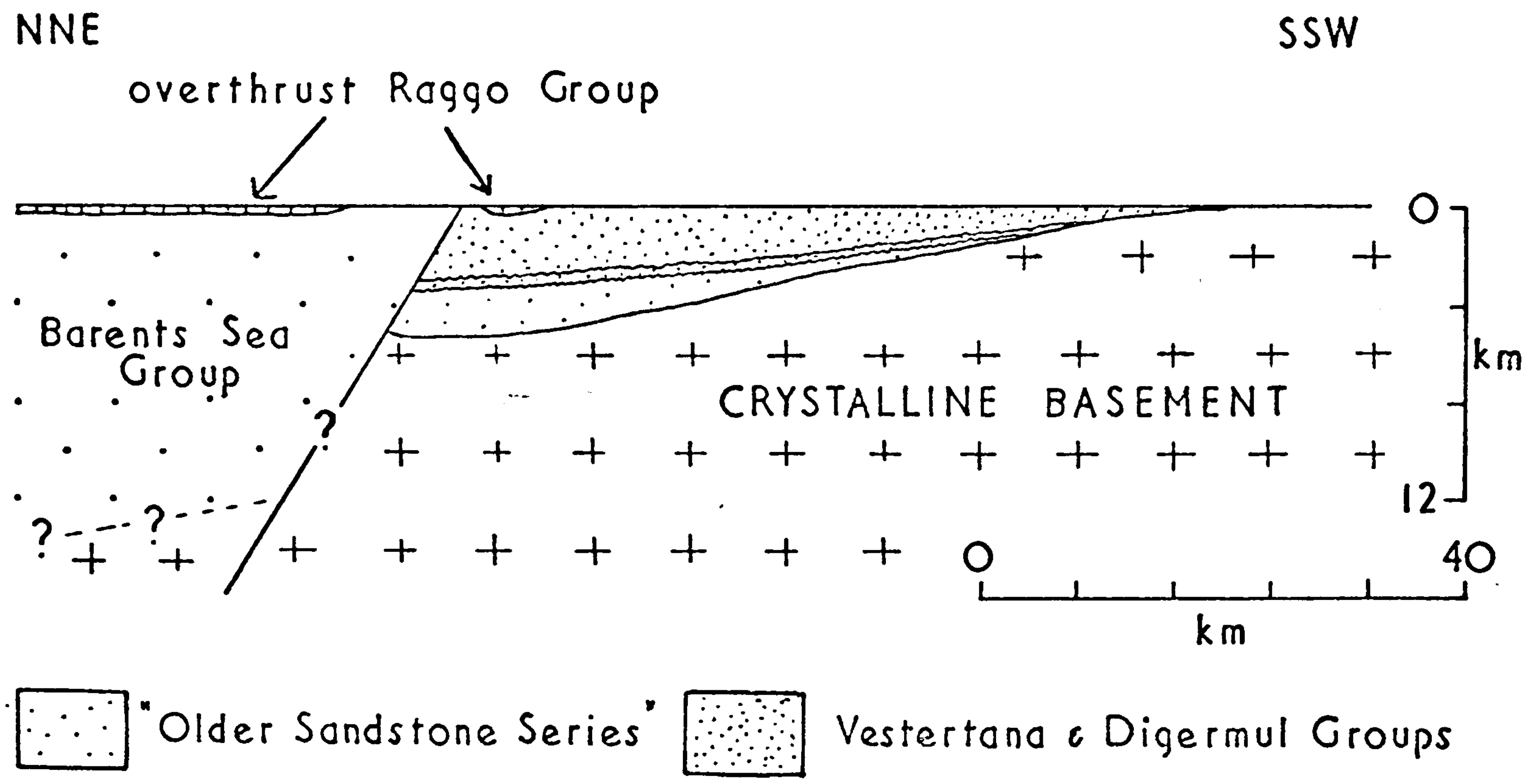


Fig. 2 Hypothetical cross-section of the rocks in the Tanafjord area. No attempt has been made to show the folding of the rocks. Partly based on Siedlecka and Siedlecki (1967).

Above the tillite formations the upper part of the Vestertana Group and the Digermul Group form a continuous sedimentary sequence 2650m thick of late Precambrian, Cambrian and Tremadocian age. The sediments consist of sandstones and shales with no carbonates or volcanics and are predominantly marine in origin. The complete succession was first described in detail by Reading (1965) following the earlier observations of Føyn (1937).

In the southwestern part of the area there is a sequence of autochthonous sediments resting unconformably upon the crystalline basement and overthrust by the Gaissa Nappe. This sequence is known as the Dividal Group (Hyolithus Zone) and has been shown by Føyn (1967) to be the condensed lateral equivalent of the Stappogiedde Formation and the Lower Breivik Member.

Structural deformation of the sediments increases markedly from east to west. On the east side of the Varanger Peninsula the rocks have suffered very little disturbance; the beds dip at angles of less than 5° and no cleavage is present (Pl. 1). In the western part of the peninsula the beds have been thrown into fairly open folds on north-south to northeast-southwest axes. When traced further west towards the metamorphic belt at Ifjord the folds become increasingly tight and small scale thrust and reverse faulting is associated with the folding. The folds are distinctly asymmetrical, facing eastwards (Pl. 2). Slaty cleavage is developed in all fine-grained beds west of the Tana River. The folding is considered by Føyn (1937) to predate the thrusting of the metamorphic and Gaissa Nappes.

Whereas the sediments to the east of Laksefjord are autochthonous or parautochthonous, those between Laksefjord and Porsangerfjord are allochthonous, apart from the Dividal Group. They belong to the Gaissa Nappe and have been subjected to several phases of folding (Gayer and others 1971). This thrust sheet, which is seen to have a minimum displacement of 25km south of Porsangerfjord, apparently dies out south of Laksefjord.

The age of the folding and thrusting is not well known. Both events must post-date the Tremadocian, the age of the youngest sediments. A major phase of folding, metamorphism and igneous intrusion in West Finnmark occurred in late Cambrian or early Ordovician times (Pringle and Sturt 1969), but further movements took place in late Silurian times which involved the folding and large scale thrusting of Ordovician, Silurian and older rocks (Sturt and others 1967). The thrusting of the Laksefjord and Gaissa nappes is probably of the later age.

Rare dolerite dykes intrude the late Precambrian and Cambro-Ordovician sediments and post-date the folding (Reading 1965). They have yielded ages of 320-350m.y. using the K:Ar method (R. Beckinsale, personal communication).

Geography of the Area

In East Finnmark most of the land lying north of the crystalline shield area forms an undulating plateau 300-700m high which is deeply dissected along the fjord coasts. Lowland areas are often densely wooded but the plateau land is either covered with heather or is bare of vegetation, exposing rock debris more or less in situ.

Excellent exposures occur along the sides of the fjords and in the nearby valleys. However, in inland areas large exposures are mainly confined to steep-sided river valleys.

The area is sparsely populated and there are few roads. Some coastal places such as the Digermul Peninsula are best reached by boat, whilst for inland localities one usually has to walk, although the Lapps sometimes drive across country on tractors and motor bikes. Halkkavarre, in the southwestern part of the region, is part of a military training area and permission to visit must be obtained in advance.

Aims of Study and Style of Description

This study is confined to those rocks which occur above the tillites but below the base of the Middle Cambrian i.e. from the base of the Innerelv Member to the top of the Duolbasgaissa Formation (Table I).

The objective was to understand the conditions of sedimentation of the various rock units as they are developed over the whole area and to formulate models of the regional palaeogeography and palaeogeology during the time of deposition of the succession.

Each member is described and interpreted individually since there is little similarity between successive members except in the upper part of the succession. Within each member a rigid facies approach to description has only been feasible in the Innerelv Member. In the other members such an approach would have led to a blurring of much essential detail. Thus, in these members a main section is described and interpreted and then the lateral variation is discussed with reference to this section.

The thesis falls naturally into three parts. In the first part there is a regional description and interpretation of the Innerelv, Manndraperelv and Lower Breivik Members and their lateral equivalent, the Dividal Group. The second part deals with the Upper Breivik Member and the Duolbasgaissa Formation, the outcrops of which are confined to the Digermul Peninsula. Finally, all available information is brought together in an attempt to make palaeogeographical and palaeogeological reconstructions.

Methods and Nomenclature

The main method of collecting data was to measure stratigraphical sections through the various members in selected areas. These sections were initially plotted on scales varying from 1:50 to 1:500 depending on the amount of lithological variation and the amount of detail considered necessary. Since adequate geological maps were available for almost all areas no further mapping was done except where it was necessary to unravel structural complications in order to measure sections. However, lateral tracing of some horizons was undertaken in well exposed areas to test lithostratigraphical correlations between measured sections or to investigate the lateral extent of certain units.

The descriptive nomenclature is based on that of Folk (1968) but is somewhat simplified and McBride's (1963) mineralogical triangle and compositional rock names have been adopted. Ingram's (1954) bed thickness classification is used and the description of bed forms and cross-stratification broadly follows Allen (1968) except that climbing ripples are described as in Allen (1970b). Details are given in Appendix A.

CHAPTER 2

A SHORT REVIEW OF OFFSHORE SHALLOW MARINE PROCESSES

Since almost all the sediments studied are thought to be of shallow marine origin it will be useful to review the main processes of sediment transport in this environment. Two major types of deposit can be recognised; marginal deposits such as beaches, barrier islands and tidal flats, and offshore deposits which today are of far greater extent and therefore should be more common in the geological record. It is believed that the majority of the sediments under discussion fall into the latter category and so this review will deal exclusively with the offshore environment.

Unfortunately less is known about modern offshore sediments than about almost any others, even though some progress has been made in the last decade. The processes which act in shallow seas are many and varied and the final sediment distribution pattern is the result of the complex interaction of all these processes. However, in any environment some processes are much more important than others in transporting sediment and an attempt is made below to define certain processes and suggest what types of sediment they might produce so that their existence might be recognised in ancient seas.

1. Semi-permanent currents

These are currents causing large scale circulation of water. They are driven either by variations in water density between different places as a result of temperature or salinity differences or they can be driven by winds. Curray (1960) has described currents of this type from the Gulf of Mexico and another example is the

Panama current in which current velocities are as high as 0.5-2m/sec (Swift 1969a).

The products of such currents are virtually unknown. If only fine-grained material was available such currents would probably inhibit deposition. If sand were available one might expect current structures to have a unidirectional trend unless the currents are affected by varying wind direction. Whilst some currents flow parallel to coastlines others do not and thus no firm inference of shoreline trend should be made from palaeocurrent directions in ancient sandstones deposited by semi-permanent currents. The sediment would probably be well sorted.

2. Tidal currents

Tidal currents are the dominant sediment moving force on a significant portion of the world's continental shelves and a secondary energy source over the remaining part. In coastal waters tidal currents may be rectilinear or strongly elliptical but on the open shelf they are usually rotary. Two main types of large scale tidal bed form have been recognised, tidal current ridges and sand waves (Off 1963, Stride 1963, Houbolt 1968, McCave 1971). Tidal current ridges as seen in the North Sea are narrow, practically rectilinear structures between 20 and 60km long and 5-10km transversely apart. They are up to 35m high and their crests lie at a few metres depth or shoal at low water. Sand waves are transverse current structures up to 15m high. They generally have an asymmetrical dune-like profile but some are symmetrical. According to McCave (1971) and Terwindt (1971) sand waves in the southern part of the North Sea occur only in water deeper than 17-18m because in shallow water they are destroyed by wave activity. The surfaces of both tidal current ridges and sandwaves are often covered with asymmetrical megaripples (dunes) which can be arbitrarily separated from sand waves on the basis of having heights of less than 1.5m (McCave 1971). As tidal currents become weaker sand waves become gradually finer-grained and smaller and pass eventually into areas of thinner-bedded fine sand and mud accumulation.

Whilst there may be an association of features which typifies inter-tidal areas (Klein 1970) there are fewer criteria for recognising offshore tidal sediments. The best criterion is a bipolar distribution of cross-bedding directions and particularly the presence of herringbone cross-bedding. This can be related to transport by ebb and flood tides in areas of rectilinear or strongly elliptical currents. One need not necessarily expect equal transport in each direction since in any area one direction usually predominates. Some authors (e.g. Reading and Walker 1966) have considered that drapes of mud on the foresets of cross-bedded sandstones are indicative of tidal activity and can be related to deposition during slack water. Whilst this is undoubtedly true in intertidal areas the concentration of suspended mud in offshore waters is insufficient to produce any significant deposit during one period of slack water (McCave 1969). Therefore mud deposits in ancient and modern offshore sediments must reflect rather long periods when sand was not being moved and thus mud drapes are not necessarily diagnostic of tidal currents. In general, tidal sediments are likely to show moderate to good textural and mineralogical maturity. In well exposed areas it may be possible to relate the form of lenticular sandstone bodies to those of tidal current ridges or sand waves.

3. Waves

Whilst normal winds may drive or change the direction of semi-permanent currents the most important waves are probably those generated during storms. Such waves are capable of stirring the sea bottom at considerable depths. Net movement can occur because residual currents are developed but these are usually relatively weak. These currents are often called wave drift currents (Swift 1969a). However, the oscillatory velocities generated may considerably enhance transport by tidal, semi-permanent or other currents (Draper 1967, Johnson and Stride 1969). Wave action may be an important erosive agent in some cases such as in the North Sea where Terwindt (1971) has suggested that winter storm waves lower the height of tidal sand waves.

The most common products of oscillatory wave action are symmetrical ripple marks and the majority of such ripples are probably formed in this way. The deduction of water depth from the dimensions of symmetrical ripple marks is a matter of controversy (Harms 1969, Baker 1970). Some authors have considered that symmetrical ripple axes tend to parallel the shoreline but this is probably true only in the very nearshore zone(see Davis (1967) and reply by Picard). Other bed forms attributed to wave action are the irregularly bedded "silty and sandy streak" facies of de Raaf and others (1965) (Reading 1970) and the irregularly truncated cross-lamination of Campbell (1966).

4. Coastal Storm Surge

These are currents produced by large scale movements of water either onshore or offshore related to the passage of a storm over the area (Seibold 1963, Swift 1969a). Backflows from onshore surges often cause nearshore sand to be transported considerable distances into deeper water. The exact mechanisms are variable. Hayes (1967) in an example from the Gulf of Mexico suggested that sediment was transported as a turbidity current whilst Reineck and others (1968) suggested that the backflow current was aided by tidal currents. In the latter example sand has been carried at least 40km offshore and to depths of 40m.

The beds are characteristically sharp-based and graded. Textural and mineralogical maturity are probably largely controlled by the source of the sediment. If the sediment came from a beach it could be both texturally and mineralogically mature. Current directions in beds produced by this mechmism would tend to be unimodal showing transport away from the shoreline although some onshore transport may be possible in very shallow water.

5. River generated currents.

It is conceivable that rivers in flood can distribute layers of sand far out from their mouths, particularly if the slope is sufficient for turbidity currents to be generated (Walker 1969, Collinson 1970).

Beds produced by this mechanism would be sharp-based, probably graded, and show a unimodal current direction indicating offshore transport. They would be less mineralogically mature than other sandstones within the basin and textural maturity would probably be only moderate.

Exceptional events

The last three processes outlined above will occur infrequently in most seas, perhaps once a year, perhaps once in a hundred years. However, more sediment may be transported during these exceptional events than during the whole of the time when they are not operating.. This same idea of exceptional events can be applied also to tidal currents because it is likely that marked movement of sand only occurs during the strongest spring tides and especially when these are assisted by storm action (Johnson and Belderson 1969).

Thus in the following chapters an attempt is made to interpret offshore shallow marine sediments in terms of the five processes outlined above, each acting over a spectrum of strengths and interacting with each other in various ways.

CHAPTER 3

INNERELV, MANNDRAPERELV and LOWER BREIVIK MEMBERS

3, 1. INTRODUCTION

The three members and the laterally equivalent Dividal Group crop out over a wide area of East Finnmark from Porsangerfjord in the west to the Varanger Peninsula in the east. The succession is thickest in the Tanafjord area being 780m on the Digermul Peninsula where it has been described by Føyn (1937), Reading (1965) and Banks and others (1971) (Fig. 3).

The Innerelv Member consists mainly of red and green mudstones with subordinate rippled siltstones and very fine sandstones. The Manndraperelv Member consists of three feature-forming red and white sandstone bands separated by two bands of green mudstone with interbedded graded sandstones. Reading (1965) considered that the graded sandstones were turbidites and that they passed up into shallow water red and white sandstones in two gradual coarsening upward sequences. The Lower Breivik Member is known to be at least in part of Lower Cambrian age because of the occurrence of <u>Platysolenites antiquissimus</u> Eichwald 140-150m above its base (Banks 1970). It consists of a complex alternation of sandstones and shales, the sandstones occurring as beds from 0.1 to 100cm thick.

To the east of the Digermul Peninsula a similar succession is developed in the Leirpollen area (Beynon and others 1967) but to the west the sequence thins to about 210m at the head of Laksefjord (Fig. 4). The Manndraperelv Member is reduced to one red sandstone unit 20-25m thick but it is thought that the lowest part of the overlying Lower Breivik Member is probably laterally

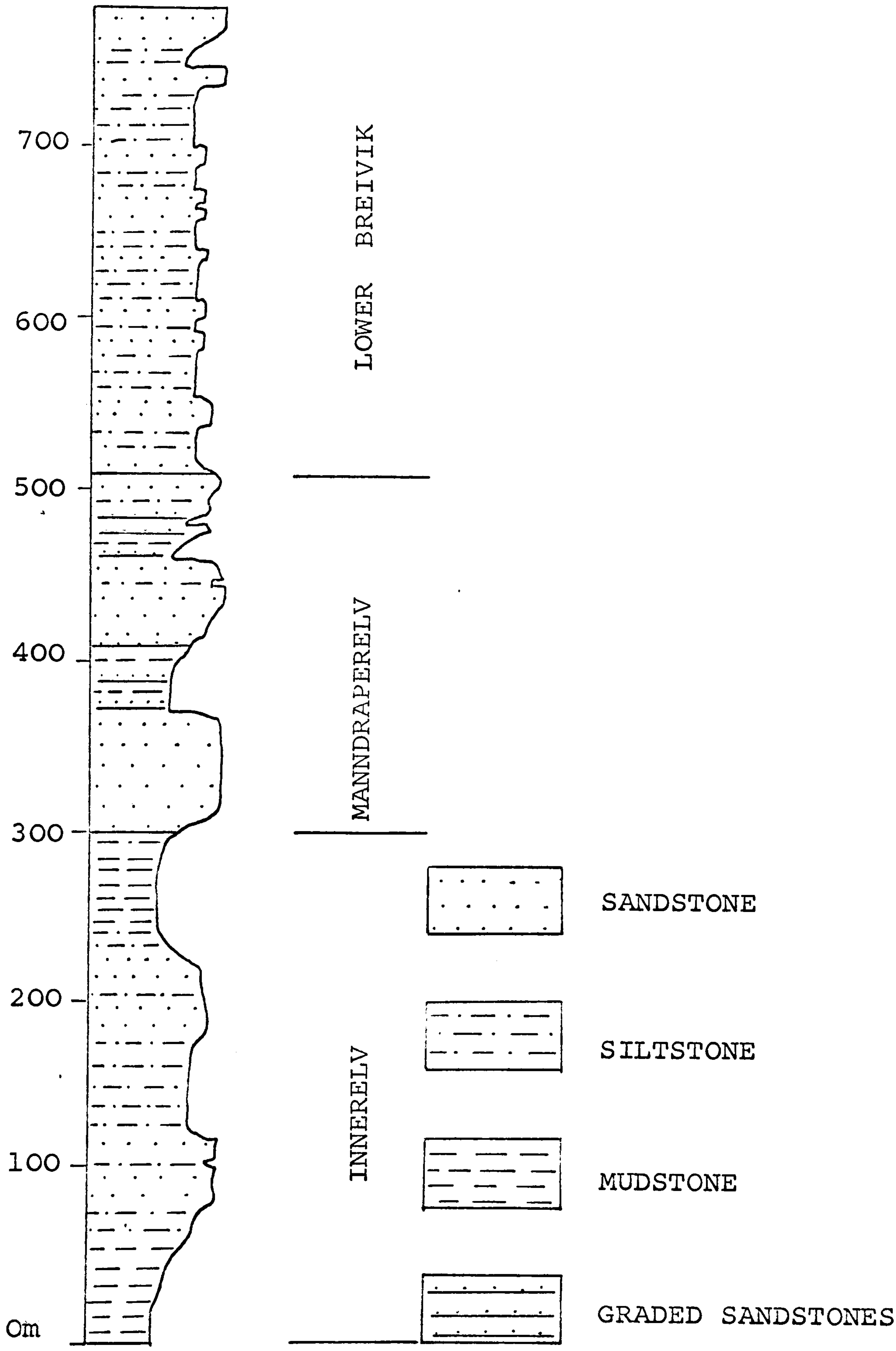


Figure 3. Simplified section through the Innerelv, Manndraperelv and Lower Breivik Members on the Digermul Peninsula, Tanafjord.

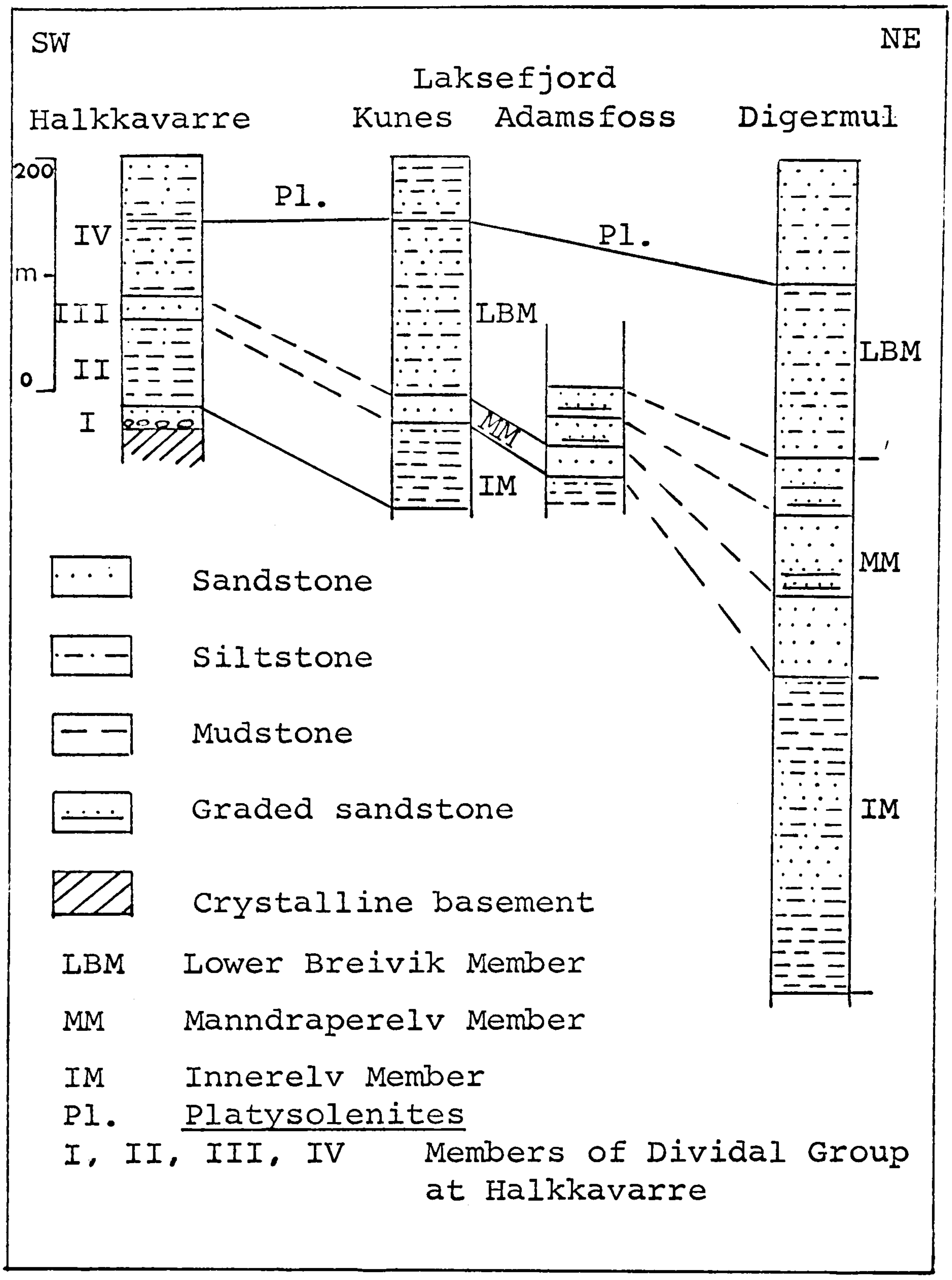


Figure 4. Lateral variation in the Innerelv, Manndraperelv and Lower Breivik Members based on the correlations of Føyn (1967) and Banks and others (1971).

equivalent to the upper part of the Manndraperelv Member on the Digermul Peninsula. The occurrence of _Platysolenites antiquissimus_ in the upper parts of the Lower Breivik Member in the Leirpollen and Laksefjord areas (Føyn 1967, Hamar 1967) confirms the lithostratigraphical correlation with the succession on the Digermul Peninsula. Føyn (1967) has been able to correlate the Laksefjord succession with that in the Dividal Group at Halkkavarre 60km to the south-west on the basis of the lithologies and the finding of further specimens of _Platysolenites_.

3, 2. LILLEVATN MEMBER - INNERELV MEMBER TRANSITION

Digermul Peninsula: Description

Before describing the bulk of the Innerelv Member it is necessary to describe its relationship to the underlying Lillevatn Member, firstly on the Digermul Peninsula and then elsewhere.

On the Digermul Peninsula the junction between the two members is well exposed on the coast opposite the island of Areholmen and the section there was outlined by Reading and Walker (1966, p. 181, "Areholmen" section). The junction, which is taken at the base of the first red bed more than 50cm thick, can be followed laterally for about 100m and there is rapid lateral variation in the beds just above and below it. However, a generalised section is shown in Fig. 5.

The majority of the 40m thickness of the Lillevatn Member on the Digermul Peninsula consists of trough cross-bedded subarkosic sandstones and granule conglomerates. The uppermost 2m of this lithology in this section (0-2m) has increasingly thick intercalations of grey siltstone and these beds are succeeded by 4m of micaceous siltstones with sandy laminae and occasional channels. These channels are partly or wholly filled with sandstones and conglomerates that are mineralogically identical with those below. Closely above these coarse channel deposits coasse sand or gravel laminae are often intercalated with siltstones.

From 6-9m there are cross-sets of siltstone 5-70cm thick with foresets inclined at up to 20° (Pl. 3). Interbedded with these siltstones are lenticular beds up to 30cm thick which are often graded from coarse sand to silt and in rare cases show type B_1 and B_2 climbing ripples. This is the topmost facies of the Lillevatn Member and the lowest 5m of the Innerelv Member consists mainly of red

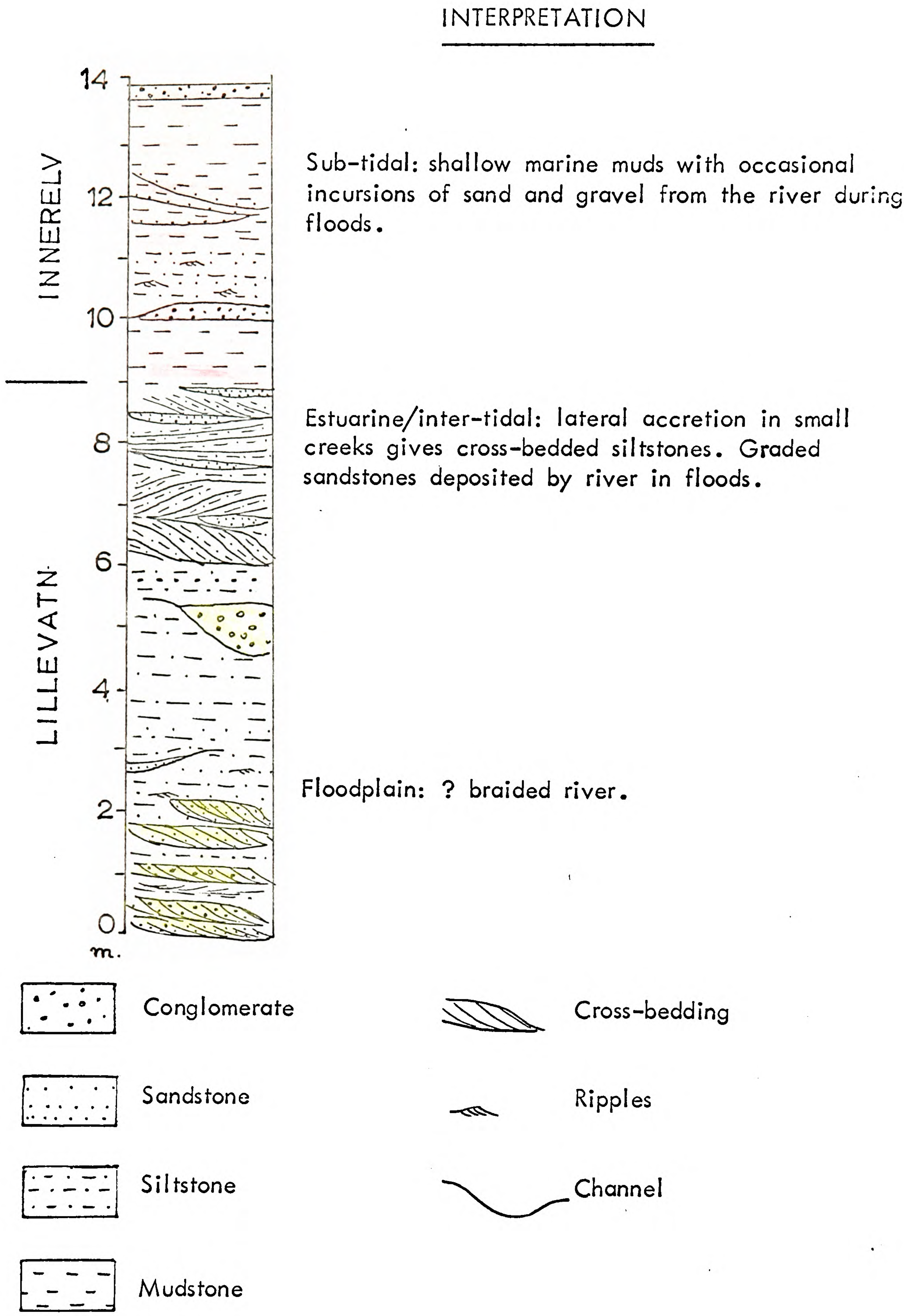


Figure 5. Generalised section of the Lillevatn Member Innerelv Member junction, "Areholmen" section, Digermul Peninsula.

mudstones with thin green bands. However, there are also occasional sharp-based and sharp-topped beds of coarse sandstone and siltstone laminae (Pl. 4) and lenticular, graded sandstones (< 10cm) sitting above low angle erosion surfaces (Fig. 6). Above the basal 5m the sequence consists entirely of red mudstones for several metres.

Although the textural maturity is very variable in the sandstones of this sequence they are all mineralogically similar except for slight variations which are functions of grain size and some variations in cement: they are subarkoses with feldspar contents of 5-10% and <5% polycrystalline quartz. There is no bioturbation in the sequence.

Digermul Peninsula: Interpretation

Reading and Walker (1966) interpreted the sequence as a transgressive one. They thought that the cross-bedded sandstones of the Lillevatn Member were possibly fluvial, that the ripple laminated sandstones and siltstontes of the basal Innerelv Member (their C4 facies) were subtidal, and that the succeeding mudstones represented quiet offshore conditions. The fluvial origin of the cross-bedded sandstones of the Lillevatn Member was confirmed by Edwards (in Banks et al. 1970) and it is accepted here that the sequence is indeed transgressive. Burrows are absent because at that time there were few if any animals capable of making them (Banks 1970).

If it is accepted that the lowest beds of the section are of fluvial origin and that the highest are most probably of offshore marine origin it remains to fit in the middle part of the succession. Cross-stratified siltstones are uncommon sedimentary deposits: the fineness of the sediment combined with the size of the foresets suggest that these beds did not form from the migration of dunes and the only modern environment in which similar deposits

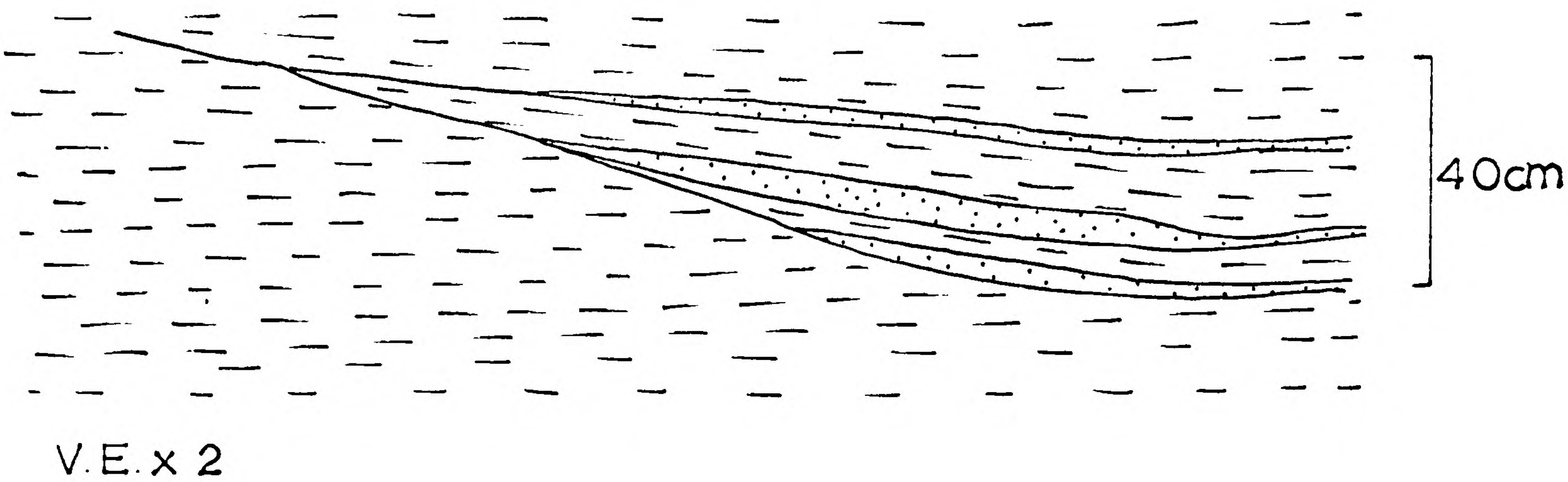


Figure 6. Lenticular, graded sandstones resting above low angle erosion surfaces in red mudstone. Basal part of the Innerelv Member, "Areholmen" section, Digermul Peninsula.

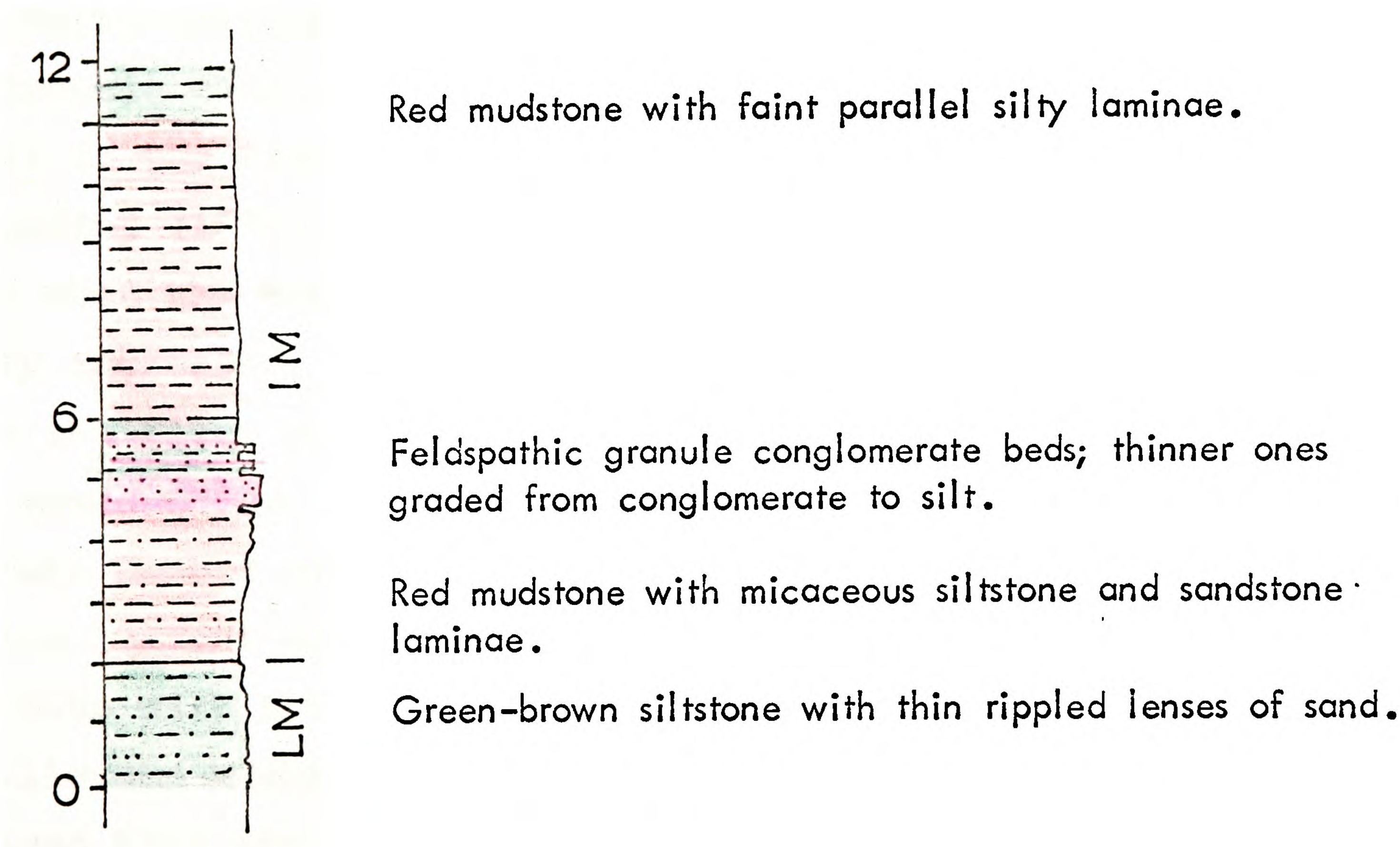


Figure 7. Section through the Lillevatn Member - Innerelv Member junction in the Vestertana area.

are known to form is on tidal flats. There, silts and muds form high angle lateral accretion sets in small meandering gullies (Reineck 1967, Fig 23). This environment is postulated for the cross-bedded siltstones of the Lillevatn Member and possibly these tidal flats were at the margins of an estuary since estuaries are probably common in areas of marine transgression as river valleys become drowned.

Since the mineralogy of all the sandstone and conglomerate beds is the same as that of the fluvial beds it seems that little or no reworking of sand occurred within the basin: thus the deposition of coarse material was probably closely related to fluvial processes. From their sharp and erosives bases and in some cases grading each coarse-grained bed apparently resulted from a single catastrophic event and it is suggested that they are similar to the river-generated sandstones of Walker (1969) and Reading (1970); each bed was deposited during a flood period when the transporting capacity of the stream was greatly increased. In the case of the beds in the Innerelv Member the river swept sand beyond its mouth into the offshore areas whilst the sandstone beds in the uppermost Lillevatn Member were deposited in the estuary. The only definite current direction in these beds is to the NE which fits with the broadly northerly flow inferred for the Lillevatn Member rivers (Banks and others 1971). It is envisaged that the river alternated between these periods of high discharge and periods when only silt and mud reached the estuary.

The only evidence of reworking of the coarse-grained deposits is in the channels within the probably estuarine facies where, above sand lenses, sand of identical grain size is seen finely interlaminated with siltstone. It appears that marine processes are almost absent except that a tidal range is implied by the inter-tidal cross-bedded siltstones. Evidence of slight marine sorting and redistribution of sediment is seen in the ripple laminated sandstones and siltstones (10.5m) which resemble the lenticular and flaser bedding of Reineck and Wunderlich (1969) except that there is no clear separation of coarse and fine material. Either waves or currents may have produced this facies. Thus it seems sea was one in which both waves and, to some extent, tides were dampened to give a low energy coastline.

<u>Other areas</u>

Varanger Peninsula

The transition is sharp both near Leirpollen and further south in the Luoftejokka valley 4.5km north of the head of Varangerfjord. As on the Digermul Peninsula the lowest part of the Innerelv Member consists of red mudstones with some coarse sandstones and rippled laminae of micaceous sandstone and siltstone.

Vestertana

The section seen in the Vestertana area (Appendix B) is shown in Fig. 7. It is similar to the Digermul Peninsula except that there are no cross-bedded siltstones.

Kunes

The upper part of the Lillevatn Member here consists of two units of cross-bedded fluvial sandstones, the upper one commonly being red. Despite this red colour this upper sandstone unit is placed in the Lillevatn Member and the junction with the Innerelv Member taken at its top where there is a very sharp transition to red mudstones which have some micaceous silty bands near the base. The transition is well seen in the south of the area immediately north of the klippe of metamorphic rocks (Appendix B).

Halkkavarre

As shown by Føyn (1967) the Lillevatn and Innerelv Members at Kunes can be correlated with fair certainty with Members I and II of the Didival Group at Halkkavarre, 60km to the SW (Figs. 4, 8). Member I consists of a basal conglomerate (20-30cm) resting on basement rocks followed by about 2m of fissile sandstone and 15m of 10-50cm bedded sandstone with occasional conglomerate layers. These beds are the basal deposits of major transgressive episode but it is not certain whether they are entirely marine or largely fluvial like the Lillevatn Member. Since elsewhere there is evidence of low energy coastlines a fluvial origin may be more likely but as the beds are poorly exposed no definite conclusions can be made. Thus the rapid passage into the mudstones and siltstones of Member II may represent either a fluvial-marine transition or a nearshore-offshore transition.

Summary

A transition at the junction of the Lillevatn and Innerelv Members from coarse fluvial sandstones to quiet water, offshore mudstones marks a period of major transgression over the entire area of East Finnmark. The shoreline, which in Lillevatn Member times must have lain to the north of the area, was pushed far to the south as the sea spread over the crystalline shield.

The transgression was possibly the result of an eustatic rise in sea level as the late Precambrian ice age finally melted away.

KEY to FIGURE 8

Crystalline basement

Conglomerate

Medium to coarse sandstone

Fine and very fine sandstone

Sharp based sandstone with interbedded mudstone

Siltstone

Mudstone

Parallel lamination

Trough cross-bedding

Cross lamination

Flaser bedding

Symmetrical ripples

Mudflake pebbles

Pl. *Platysolenites* horizon

Abrupt litholgical change

Ball and pillow

Descication cracks

Very lenticular beds

B Marked bioturbation

SM Good sole marks (flutes and grooves)

G ?Glauconitic sandstones

Horizons at which bioturbation is present

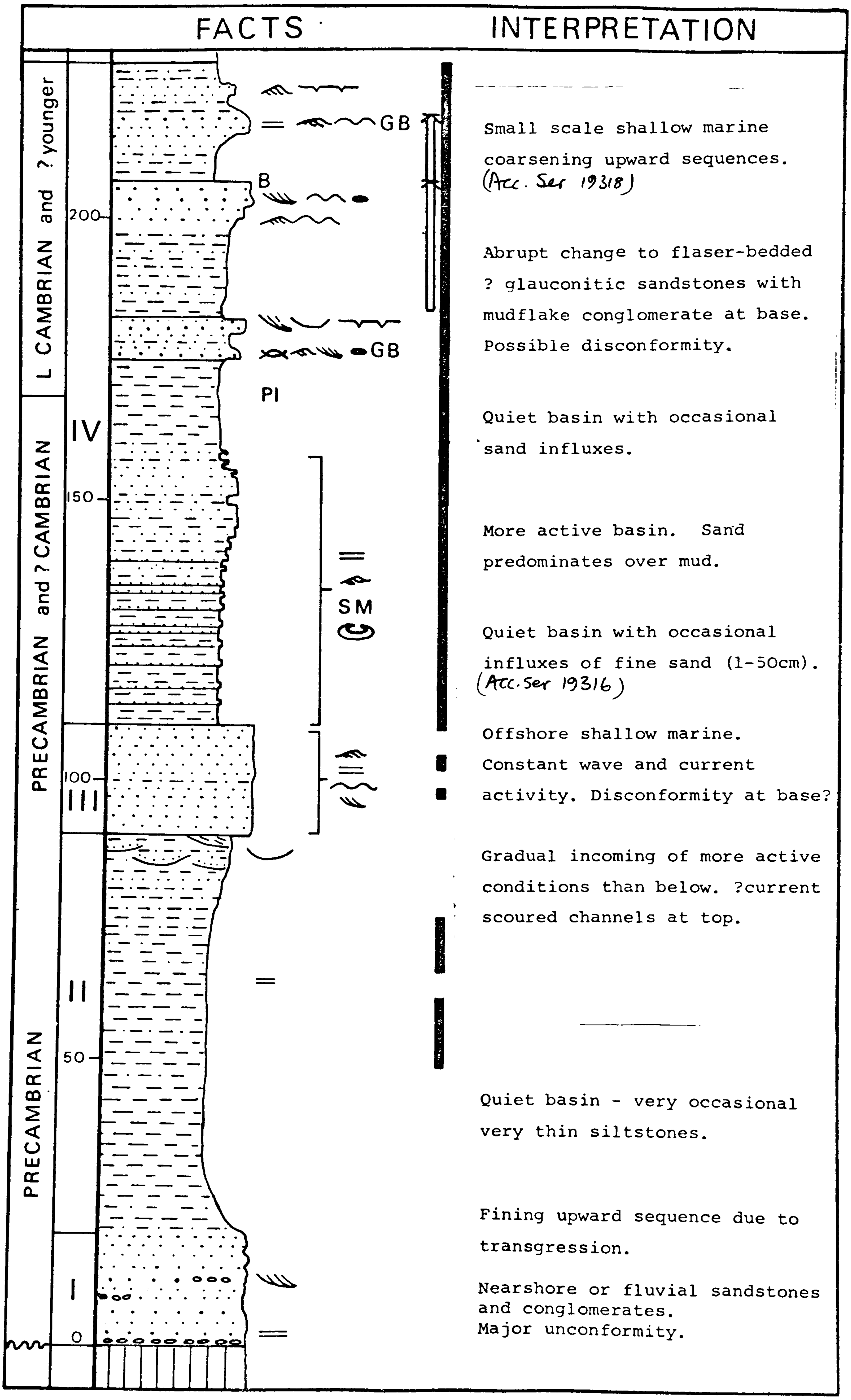


Figure 8. The Dividal Group at Halkkavarre.

3, 3. INNERELV MEMBER

Introduction and Facies

The main section of the Innerelv Member is the coastal section on the SE side of the Digermul Peninsual immediately SW and NE of the mouth of the Manndraperelv. The base of the section is the "Areholmen" section described previously. The main section is described in terms of six facies, the first five of which form a gradational series; these facies are also used to describe all the other sections. Firstly the facies are defined and interpreted, then the main section is described and this is followed by a description of the lateral variation within the Member. Finally the overall interpretation of the Member is discussed.

Facies I 1 Green and red mudstone

Description

This facies consists of green and red strongly cleaved mudstone with very thin parallel laminae of siltstone which are laterally persistent.* Very rarely cross-laminated siltstones and sandstones up to 25cm thick are found. The mudstone is calcareous and occasional very thin, brown-weathering beds within it are composed almost entirely of fine-grained ferroan calcite. In general there is no difference in grain size between red and green coloured beds but where thin green bands occur in a predominantly red sequence they are often slightly coarser or contain some coarser beds.

Interpretation

This facies was deposited under very quiet conditions with deposition of fine-grained material entirely from suspension. Very occasionally terrigenous sedimentation

* See Appendix A for definition of lateral persistence terms

entirely ceased and very thin fine-grained carbonate beds accumulated. The mechanism of carbonate deposition is an enigma but perhaps it was by the settling out of planktonic micro-organisms (Algal or protozoan?). The thicker beds of siltstone and sandstone testify to the occasional presence of strong currents.

Facies I.2 Parallel laminated siltstone

Description

This facies consists of parallel laminated green siltstone interbedded with mudstone (Pl. 5). The laminae are up to 1cm thick and are laterally persistent. Internally the laminae are either massive or show fine parallel lammations; small scale cross-lamination is occasionally seen and its orientation is usually unidirectional. Small burrows are sometimes found in this facies but bioturbation has had virtually no effect on the sedimentary structures. This facies is gradational between I.1 and I.3.

Interpretation

Deposition of the parallel laminated siltstone probably occurred primarily from suspension but occasionally currents were sufficiently strong to cause bed load transport as small scale migrating ripples thus producing cross-lamination.

Facies I.3 Very thin-bedded siltstones and very fine sandstones

Description

This facies consists of laminae and very thin beds of siltstone, silty sandstone and very fine sandstone with subordinate interbedded mudstone (Pl. 6). The beds are less laterally continuous than those of I.2 and pinch and swell in thickness. Internally they mostly have an irregular wavy cross-lamination but thinner beds are parallel laminated.

Bedding surfaces show asymmetrical and very occasional symmetrical ripples, the latter always seeming to have formed as a modification of an earlier bedform. Cross-lamination is apparently unidirectional but it is often difficult to measure accurately. Bioturbation is rarely present in this facies which is intermediate between I.2 and I.4.

Interpretation

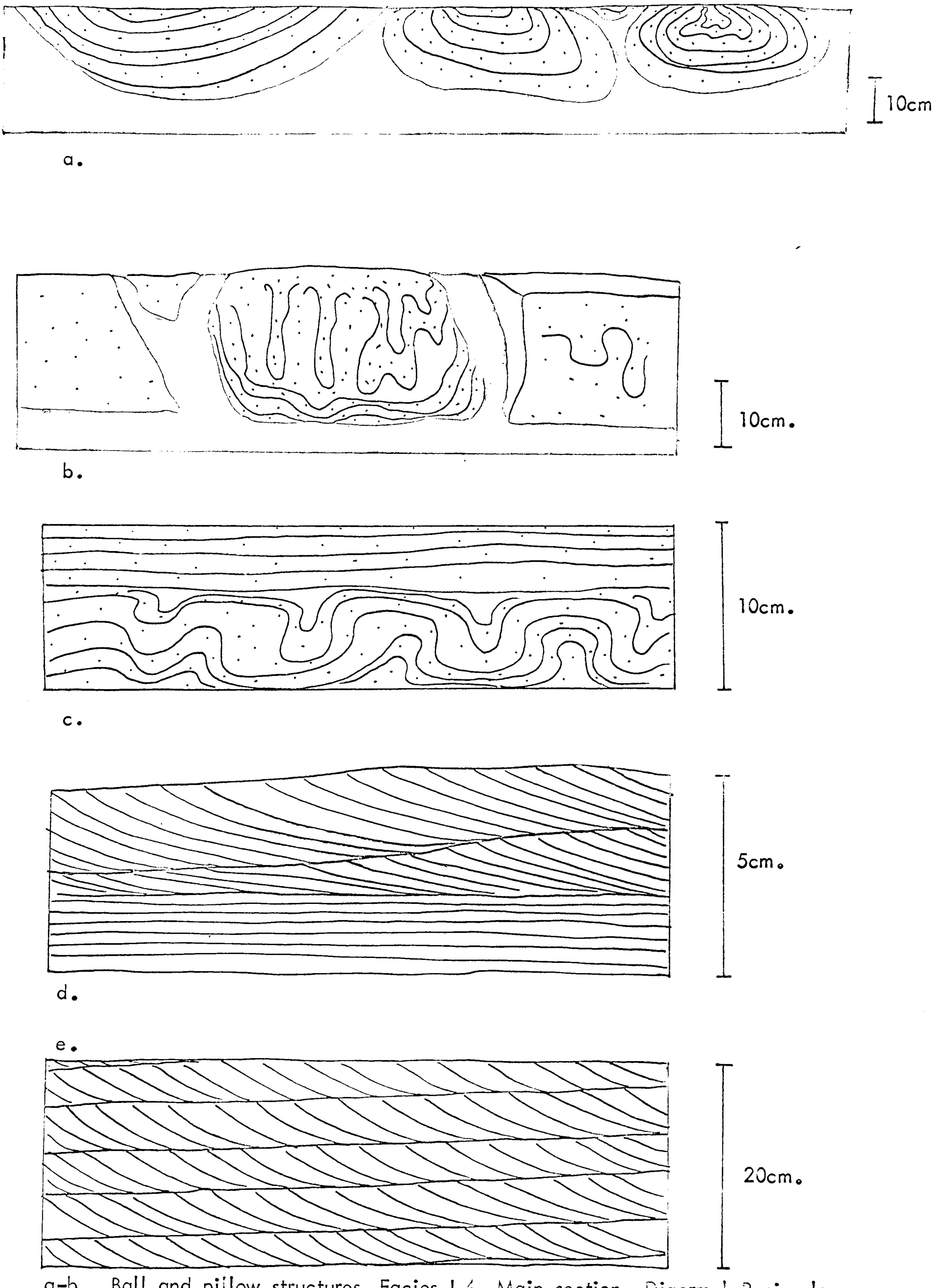
The increased grain size and abundance of cross lamination in these beds suggests that they were deposited from stronger currents than those which deposited the beds of I.2. The presence of symmetrical ripples which are absent in I.1 and I.2 shows that wave activity was capable of modifying the sea floor.

Facies I.4 Thin to thick-bedded sandstones and siltstones

Description

Sharp-based, green-grey, very fine sandstones, silty sandstones, and siltstones (Pl. 7-9, Fig. 9) occur as beds from 3 to 100cm thick interbedded with thinner-bedded siltstones. Whilst the thin beds are mostly parallel sided and moderately laterally continuous, the thicker ones are sometimes markedly lenticular and fill steep sided channels a few metres wide. In one case the beds were seen to be grouped together into discrete packets, all the beds dying out laterally at approximately the same place (Fig. 10).

The bases of beds are poorly exposed but where they were seen no sole marks were found. Internally many beds are slightly graded particularly in their upper parts. Some beds, particularly siltstones, appear massive but others show cross-lamination and parallel lamination.



a–b. Ball and pillow structures, Facies I 4, Main section, Digermul Peninsula.
c. Contorted stratification, Facies I 4, Main section, Digermul Peninsula.
d–e. Stratification within sandstones of Facies I 4, Main section, Digermul Peninsula.
Fig. 9 . Bedding structures in Facies I 4, Innerelv Member.

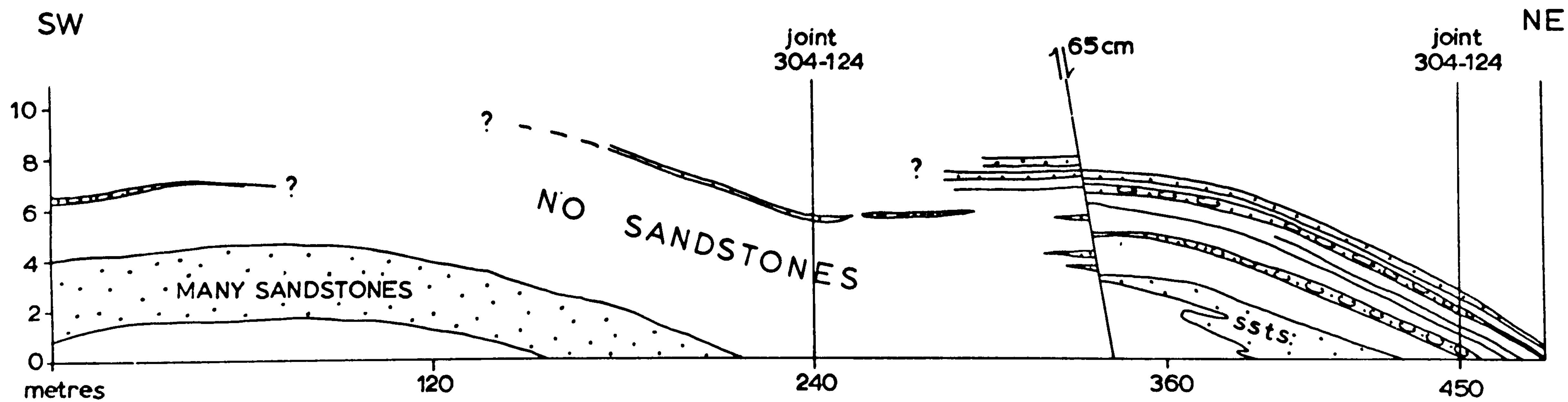


Figure 10. Lateral variation in Facies I 4 sandstones as seen in the coastal exposure immediately north of the mouth of the Manndraperelv, Digermul Peninsula.

Where parallel lamination is present it is usually confined to the lower part of the bed and is overlain by cross-lamination. No primary current lineation was seen associated with the parallel lamination. Alternatively cross-lamination is often present throughout the bed; cosets of type A climbing ripples are the most common structures although type B.1 climbing ripples also occur. Palaeocurrent directions are predominantly unimodal but occasional beds show directions 180^{o} apart. Several beds greater than 20cm thick show various types of contorted stratification as illustrated in Fig. 9 a-c. The nomenclature for these types of structures is as confused as the structures themselves but the term "ball and pillow" structure (Potter and Pettijohn 1963) is convenient for forms such as Fig. 9. a,b. In all the occurrences of ball and pillow structures one ball or pillow encompasses the whole vertical thickness of the bed. The laminae within a ball or pillow may be simple curved up or strongly contorted and in some cases bundles of laminae may be truncated by overlying ones as figured by Sorauf (1965 Fig. 10) in Devonian examples from New York. Although the structures cannot be fully seen in three dimensions the pillows seem to show a strong alignment which is usually perpendicular to the current direction as indicated by cross-lamination. There is no consistent asymmetry to the structures; undeformed beds can often be traced laterally into and out of a zone of deformation. Forms such as Fig. 9c are best described as "convolute lamination with vertical fold axes". In both forms the deformed lamination is often truncated at a planar erosion surface at the top of the bed.

Symmetrical ripples are occasionally seen in this facies but bioturbation is absent. The facies is intergradational between facies I.3 and I.5.

Interpretation

The sandstones and siltstones were deposited from stronger currents than those which deposited the beds of Facies I.3. This is shown by the predominance of sand over silt and the common parallel lamination which is interpreted in most cases as the result of plane bed with movement conditions in the upper flow regime (Simons and others 1965) even though no primary current lineation was seen. Goldring (1966) noted that this lineation is rarely well seen in beds of this fine grain size. In the majority of instances upper flow regime bedforms are overlain by cross-lamination (lower flow regime) and thus the beds were deposited by waning currents. Given the grain size of very fine sand the absence of dune structures is to be expected (Allen 1970a). However, since the angle of the climbing ripples is usually approximately constant throughout the cross-laminated division most of each bed was probably deposited within a fairly narrow range of flow power. The presence of symmetrical ripples shows that this facies was deposited above wave base.

The syndepositional origin of some of the ball and pillow structures is shown by the internal truncations of laminae at the edges of balls. In other cases the truncation at the tops of beds shows that deformation certainly occurred before burial. Thus in contrast to the opinions of Potter and Pettijohn (1963) there is no rigid distinction between ball and pillow structures and load balls. However the amount of loading appears to be relatively slight since where a deformed bed can be traced laterally into an undeformed bed there is little variation in the level of the base of the sand. As suggested by Middleton (1970) the sand was probably "quick" during deformation. The structures are considered to have formed due to slight differential vertical movements in a quicksand which resulted in distinct balls and pillows loading slightly into a mud and the mud being drawn up passively into steep anticlinal folds which finally rupture to give discrete balls and pillows of sand as shown in Fig. 11 and by Selley et al. (1963, Fig. 7A).

The orientation of the pillow axes perpendicular to the palaeocurrent direction is problematical. Either the palaeoslope or current shear could have had an effect on controlling this orientation but there is little evidence for either process.

Contorted stratification in sandstones where there is no loading into an underlying mudstone (e.g. Fig. 9c) can also be explained as the result of quicksand movement as shown by Selley and Shearman (1962).

Facies I.5 Large lenses of irregularly bedded siltstone, sandstone and mudstones.

Description

This facies consists of lenses of irregularly bedded siltstone, sandstone and mudstone up to at least 10m wide and 1.5m deep, each lens being separated from its neighbours by distinct surfaces of discontinuity (Pl. 10 and 11). The lower bounding surface of each lens is usually concave upward; it is smooth or slightly irregular and rarely dips at more than 10° except where sandstones form the margin. The upper surface may also be concave upward, forming the base of the next lens or may be flat when overlain by horizontally bedded sediment. The three-dimensional form of the lenses is not clearly seen but they seem to be channel-shaped rather than dish-shaped.

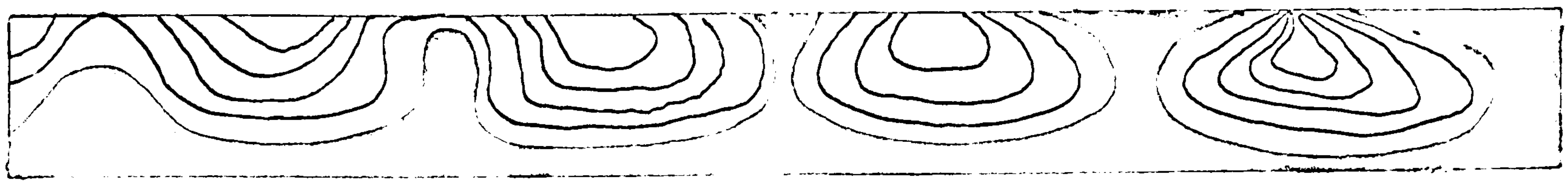


Fig. 11 An illustration of the various stages between convolute lamination where mud is drawn up into anticlinal cores and ball and pillow structures with detached nodules.

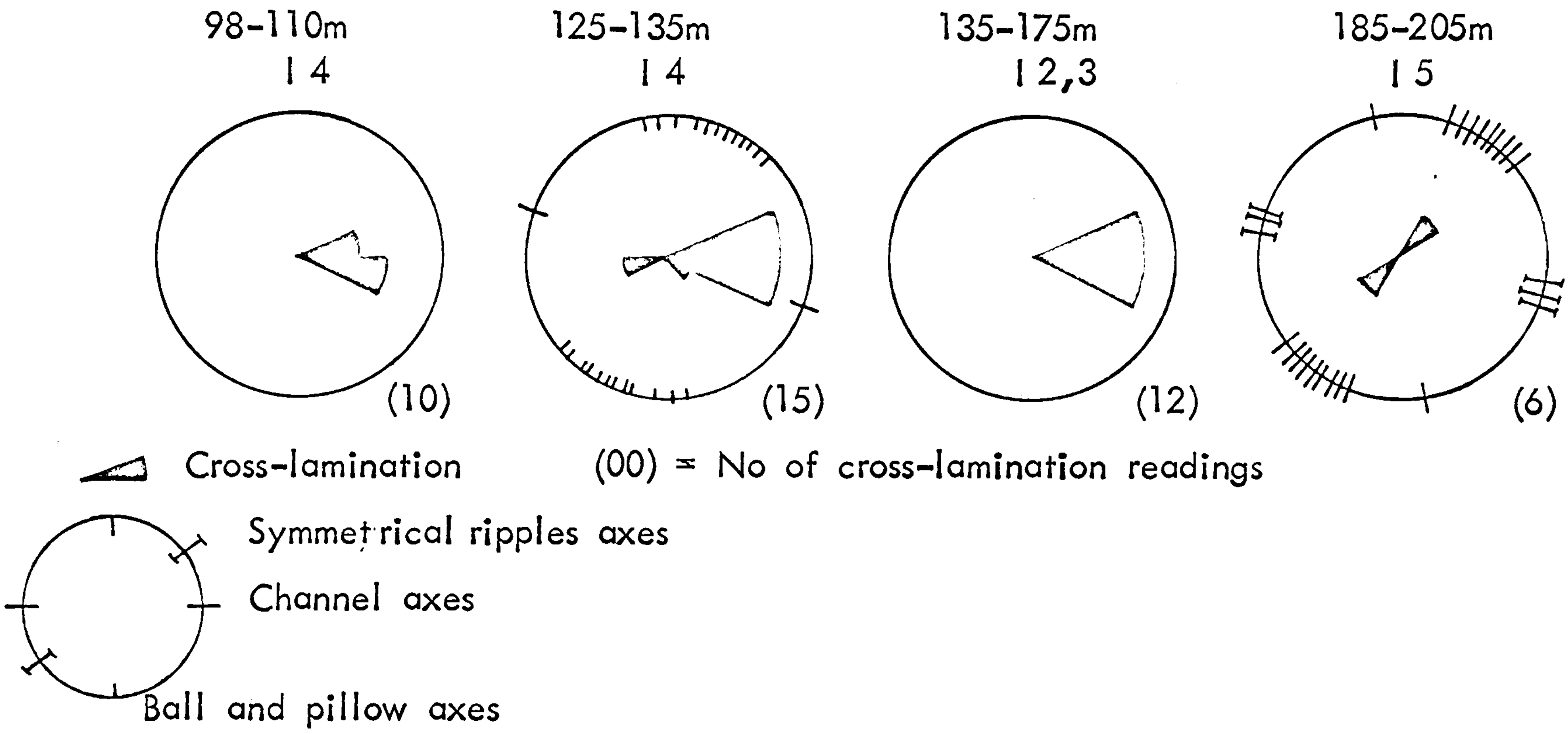


Fig. 12 Palaeocurrent data, main section, Innerelv Member; Digermul Peninsula.

Many different types of sediment occur in rapid alternation within the lenses. These include lenticular rippled laminae and very thin beds of siltstone and very fine sandstone (Facies I.3) and thin to medium-bedded lenticular very fine sandstones (Accession Series 19307). The latter are similar to Facies I.4 but all are lenticular, wedging out towards the margins of the lenses. Parallel lamination is the predominant internal structure; cross-lamination, if present, is confined to the topmost parts of beds except in thin beds; many varieties of cross lamination are found. There are also siltstones with wispy sub-parallel lamination (Pl. 12) and many other irregularly bedded siltstones, sandstones and mudstones, including some beds resembling the silty streak and intermediate silty-sandy streak facies of de Raaf and others (1965, Fig. 9). Sparse palaeocurrent measurements suggest a very varied pattern of sediment transport although an overall bipolar distribution may be present. Symmetrical ripples are apparently orientated roughly perpendicular to the axes of the channel lenses but again the data are meagre. As in other facies the symmetrical ripples formed as modifications of earlier bedforms (Pl. 13).

Within a lens the layers are thickest in the middle and wedge out towards the sides; this is true both for the sandstones, which cut channels within, and parallel to, the larger channel lenses, and also for the finer grained sediments which seem to be draped into the channel lenses. Small rotated slump packets $\leq$ 20cm occur infrequently within the lenses.

Interpretation

The smooth nature of the lower margins of the lenses, their fill which is broadly similar to the surrounding sediment, and the draping effect of material within them might suggest that these lenses are slump scars (Laird 1969). However, other evidence suggests that slumping is not the most important process in the production of the lenses:-

1) There is a lack of associated rotated slump packets except for very small ones preserved within the lenses.

2) There is a lack of shears or contortions in the sediments immediately beneath the "scars".

3) The fill of the lenses is very variable which suggests great fluctuations in environmental energy, in particular the presence of erosively based sandstones suggests that there were currents available to produce scours of the observed dimensions.

Thus it is suggested that the lenses certainly have channel-shaped forms and that they formed by infilling of current eroded scours. In most cases the current which cut the channel deposited little or no sediment within it and the channel was later infilled by sediments mostly deposited under less turbulent conditions. After the initial cutting the margins of the channels were sometimes modified by small scale rotational slumping. The presence of lenticular sandstone beds orientated parallel with the channel axes suggests that the open or partly filled channels acted as funnels for further sediment transport. The predominant parallel lamination in these lenticular sandstones is interpreted as the result of plane bed movement in the upper flow regime and thus these beds were deposited from stronger currents than those which deposited the sandstones of Facies I.4.

The presence of symmetrical ripples indicating wave activity at the sea bed, combined with the generally irregular, rippled nature of the bedding implies a fairly shallow water environment for this facies, and one in which there were great variations in the strengths of currents and possibly waves also.

Facies I.6 Black laminated siltstone

Description

This facies is distinguished from all others by its colour which is due to pyrite which can be seen as small cubes within the coarser laminae. The facies consists mainly of flat, very thin, grey laminae of coarse siltstone intercalated with darker, finer siltstone. However, in some parts this lamination occurs in discontinuous wavy sets (Pl. 14) in which there is frequent evidence of truncation of laminae. There is no evidence of biogenic activity in this facies.

Interpretation

The general environment was obviously one of fairly quiet water but in the beds with truncated sets of laminae scouring of the sea bed is evident. The irregular "scoopy" appearance of this scouring suggests that it might have been produced by wave activity. The bedding may have been produced by the stirring up of the sea floor and suspension of sediment during storms followed by redeposition as the storm died down. The flatly laminated sediment could also have been deposited by settling of storm-stirred material transported from other areas by weak wave-induced currents; however, alternatively it could have been deposited by some other weak current as was the sediment of Facies I.1 and I.2. The factors which caused a reducing environment within these sediments and hence the development of pyrite are not understood. The absence of biogenic activity is probably due to the scarcity of animals capable of burrowing into the sediment at that time (Banks 1970) rather than to the presence of an anaerobic environment.

The Main Section

The Section is shown in Fig. 13 and exact details of the localities are given in Appendix B. In broad outline two coarsening upward sequences (5-130m, 130-205m) are present followed by a reversion to finer-grained sedimentation towards the top of the member.

Palaeocurrent data (Fig. 12) shows a predominantly easterly transport but with some variation.

Lateral Variation

The lateral variation within this member is summarised in Fig. 14. The most obvious feature is the marked thinning southwestwards from the main (Manndraperelv) section such that the member is about 80m at Adamsfoss and 70m at Kunes and Halkkavarre. Eastwards from the main section the thicknesses are not so accurately known but the member does not seem to thin more than a little. Characteristically the lowest 10-40m of the member is red coloured in all the sections and above this it is predominantly green although turning to red again at the top in some places.

First the variation towards the southwest will be described and then that towards the east. Exact localities of measured sections are given in Appendix B.

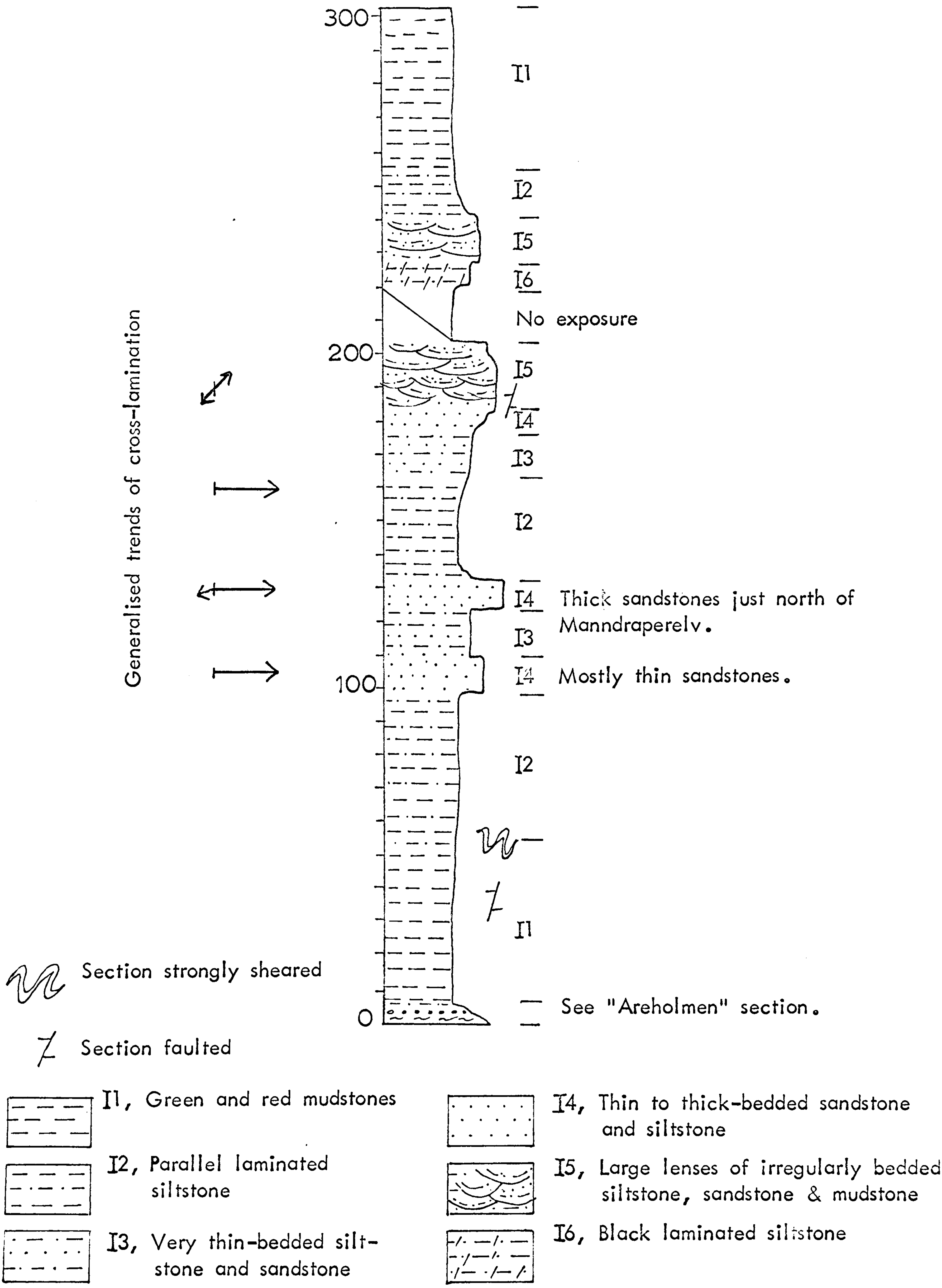


Fig.13 Main section of Innerelv Member; Digermul Peninsula, near Manndraperelv.

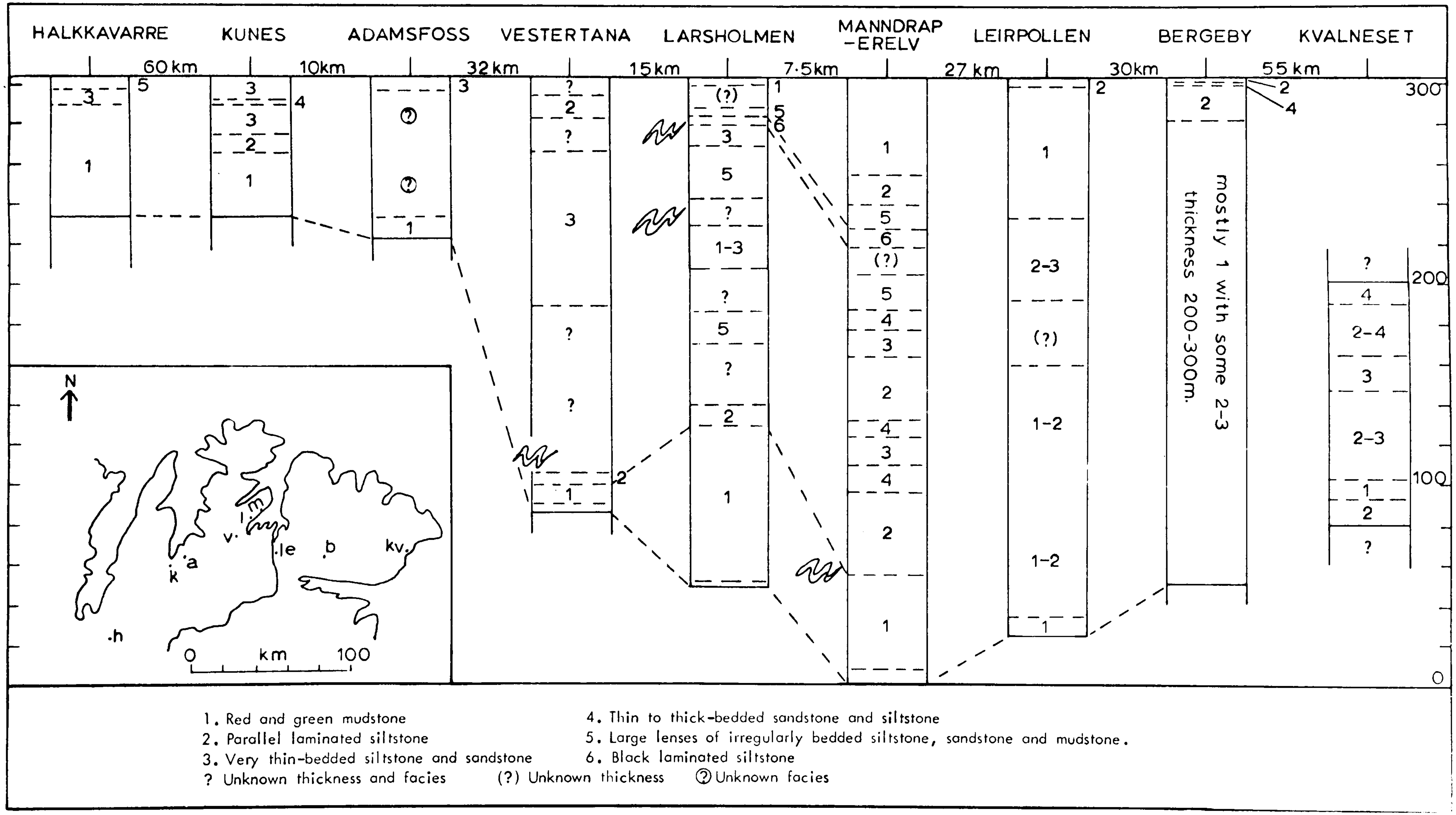


Figure 14. Lateral variation within the Innerelv Member.

Larsholmen:

The member is patchily exposed along the shore south-west of Larsholmen and is strongly folded in its upper half so that some of the rocks are inverted. Facies I.5 is well developed at two levels but I.4 is absent. The uppermost part of the section seems to be very condensed compared with the main section; although this may be partly due to tectonism a similar, relatively condensed section is also present in the Innerelv valley just to the north and there the rocks are less deformed.

Vestertana:

Near the base of the member FaciesI.1 passes up rapidly into Facies I.2 before the section becomes strongly folded. The remainder of the member is poorly exposed except for some localities near the E4 highway and thus no estimate of the thickness can be made.

Adamsfoss:

The member is poorly exposed in the first eastern tributary of the Adamselv about 3.5km SE of Adamsfoss where it is about 80m thick. The lower half is mostly red whilst the upper half is green.

Kunes:

10m of red mudstones are overlain by green beds which become gradually coarser upward. A few ripples in Facies I.3 show northwesterly dipping cross-lamination (Pl. 15).

Halkkavarre:

The member coarsens upward in the top 15m and lenticular sandstones occur within largely silt-filled lenses (channels?) in the top 5m (Pl. 16). Small burrows occur on the bases of a few siltstone layers in the upper 50m.

Leirpollen:

In this area much of the member is distinctly calcareous (Acc.ser. 19308) and weathers massively. It is less sandy than on the Digermul Peninsula and no distinct coarsening upward sequences were seen although the member is coarsest in its middle part. Contrary to the assertion of Beynon and others (1967) no silt-filled channels (I.5) were seen.

Bergeby:

The member is patchily exposed in the valleys of the Bergeby and its westerly tributaries, the uppermost part being well seen in Perledalen. The rocks dip very gently northwards and the thickness can only be roughly estimated from the width of outcrop.

Kvalneset:

Good exposures on the shore and in the raised cliff are found in the most easterly outcrop of the member (Pl. 1). The beds are characterised by the relative abundance of simple burrows above the lowest 30m of the exposure (19309). However, bioturbation is still very slight compared with most Phanerozoic marine deposits. Also, symmetrical ripples are absent and the member consists of laterally persistent siltstones and very fine sandstones from a few millimetres to 20cm thick intercalated with mudstones and fine sandstones (Pl. 17, 18). This locality provides the only good views of bedding surfaces and ripples are seen on the tops of sandstone beds (Pl. 19). All the ripples and cross-lamination show a consistent transport direction to the northwest, a striking contrast to the main section. However there are no obvious petrographic differences between these two areas.

Neither the base nor the top of the member is present in the coastal section and so it cannot be correlated with other sections except that it is probably some tens of metres above the base as no red beds are present except for one thin band.

Summary

In contrast to the two coarsening upward sequences in the Manndraperelv section only one such sequence is present in the thinner western sections. This may be because only one was developed there but a possible unconformity at the base of the overlying member may have removed some material. On the Varanger Peninsula the succession is finer grained than on the Digermul Peninsula except at Kvalneset in the extreme east and no clear pattern of coarsening upward sequences was seen.

Discussion

Facies I.1 - I.5 make up a continuous sequence of increasing current and wave activity (Fig.15). Assuming that wave activity is related to water depth this sequence I.1 - I.5 can be interpreted as a result of shallowing of the sea and it can also be inferred that current activity fell off gradually with depth. In several sections the facies are arranged in more or less regular shallowing, and thus coarsening upward, sequences (e.g. two coarsening upward sequences in the Manndraperelv Section, one at Halkkavarre). By assuming that these coarsening upward sequences represent the superposition of laterally equivalent facies as a result of progradation a model can be developed for the ways in which sediment was transported within the system.

In Facies I.5 sediment was funnelled through narrow channels during periods of strong current activity when transport was largely in an upper flow regime plane bed phase. In Facies I.4 transport was still largely confined to channels (e.g. Fig. 10) but they were much wider and the deposits less lenticular. In Facies I.3 - I.1 deposition occurred effectively as sheets of sand and silt under much lower energy conditions as the currents spread out into deeper water. The sequence of different types of sandstone and siltstone with decreasing current strength are shown in Fig. 15. From its position above and below units of Facies I.5 it seems that Facies I.6 was probably deposited in shallow water, possibly during periods of cut off of supply or in areas sheltered from strong current activity. Given these relationships between the facies it remains to try to estimate values of depth and possibly distance from shore for one facies in order to define the whole system more closely.

Biogenic features of these late Precambrian rocks give no clues since body fossils are absent and trace fossils are very scarce due to the scarcity of burrowing animals at that time (Banks 1970). Another approach to the problem is by comparing the Innerelv Member with other described sediments whose origins have been discussed. Wunderlich (1970) has described a series of Devonian sediments very similar in some respects to Facies I.4 - I.5 and he attributed a shallow marine and tidal flat origin to these beds by comparing them with the modern sediments of the North Sea. The intercalation of lenticular sandstones with irregularly bedded lenses of largely finer material (e.g. Wunderlich's Fig. 27, 28, 37) is very reminiscent of Facies I.5 and his more parallel sided sandstones (Fig. 32)

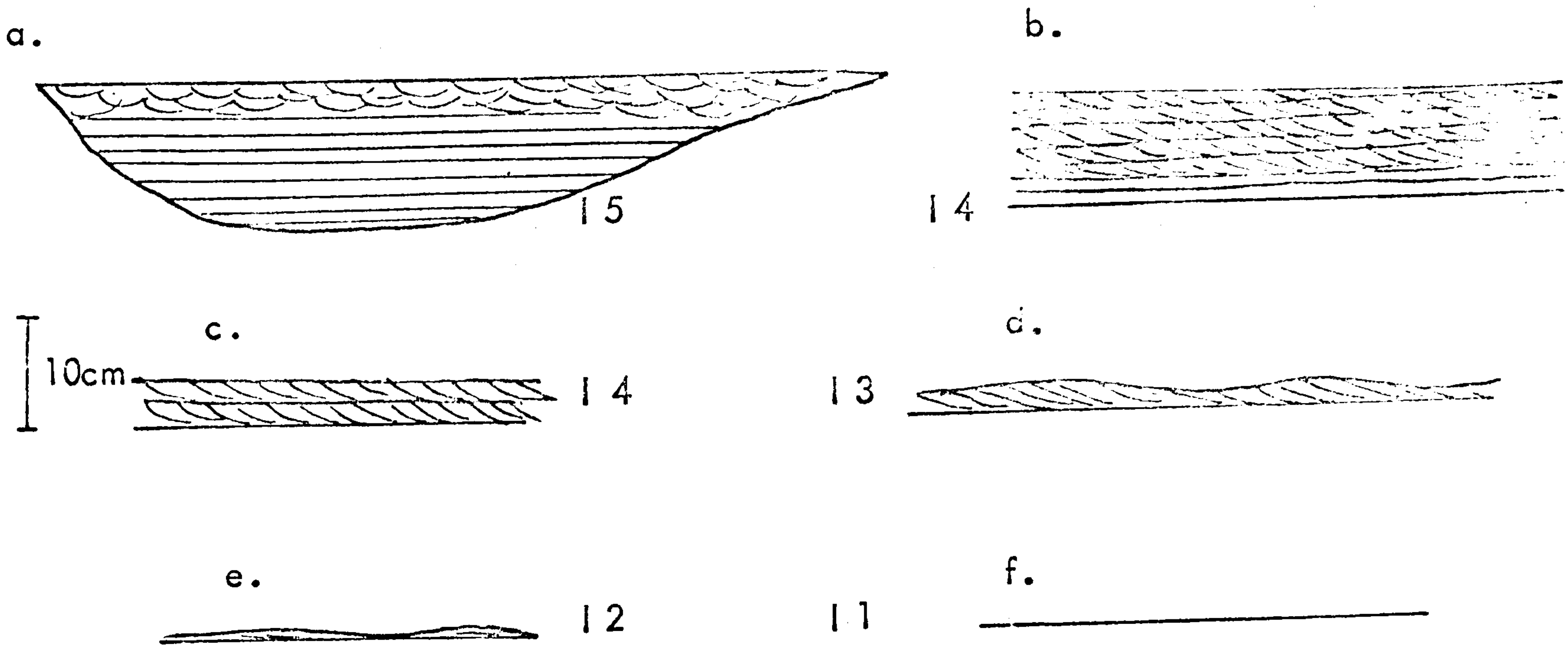


Fig.15 Idealised sketch to show the gradual changes in bed thickness and sedimentary structures in the beds of each facies. a-d = very fine sandstones, e-f = siltstone.

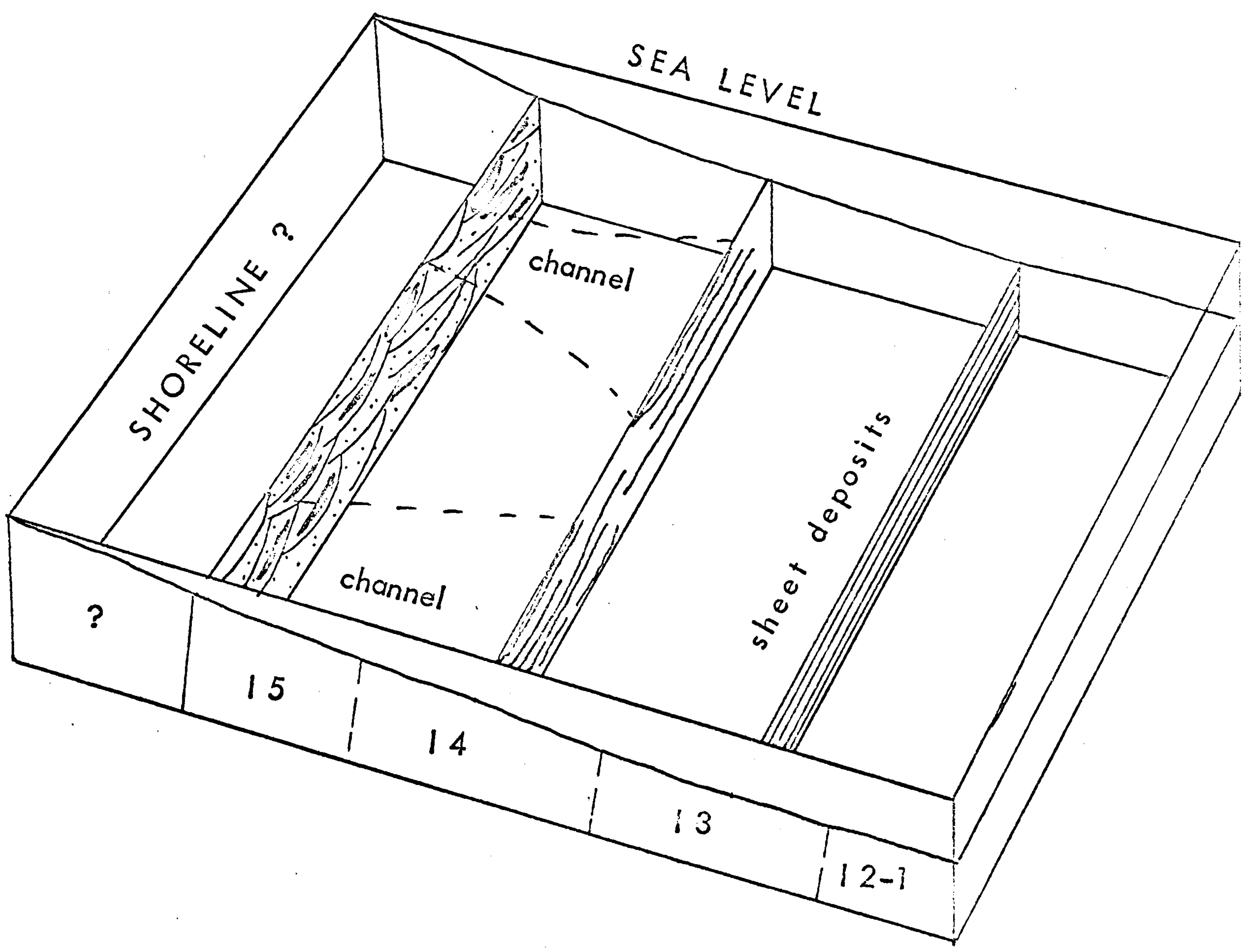


Figure 16. A model of sediment transport in the Innereiv Member and the relationships of the various facies.

are similar to Facies I.4. However, a tidal flat origin for any part of facies I.4 - I.5 can probably be excluded because of the absence of lateral accretion deposits which are so characteristic of gulley sedimentation in inter-tidal areas. Richter (1967) described beds from the Devonian of SW England which are similar to Facies I.4. Lenticular and parallel sided sandstones occur within a shale sequence and the channels are orientated parallel to the inferred shoreline. Shell beds with a diverse offshore fauna and the absence of wave activity led Richter to infer a moderately deep shelf environment and he attributed the presence of the sandstones to tidal currents. Other beds in my experience comparable to Facies I.4 are the Upper Llandovery sandstones of Gullet Quarry in the Malverns and the Upper Wenlock of the Sawdde Gorge, near Llandeilo except that in neither case have ball and pillow structures been seen. The Gullet Quarry beds have been interpreted from fossil evidence to have been deposited in at least a few tens of feet of water (Ziegler 1965, Ziegler <u>et al</u>. 1968) and a similar "shelf" environment is likely for the Wenlock beds.

Thus these comparisons suggest that Facies I.4 - I.5 might have been deposited in a sub-tidal to moderately deep shelf environment.

A more theoretical approach is to attempt to determine what type of current is most likely to have produced the features of the succession. The following features of the currents seem to be important:

1) Transport direction is unimodal except in the shallower water beds.
2) Currents decrease in strength with increasing depth and possibly, though not certainly, have an offshore component of flow.

P= Possible PR= Probable I= Improbable	Currents unidirectional except in shallowest facies	Currents less strong at depth	Channelised flow in shallow water etc.	Great variation in current strength	Currents strong enough to give observed bed-forms
Semi-permanent	P	P	I	I	P
Tidal	P	PR	PR	P	PR
Wave drift	I	PR	I	PR	I
Coastal storm surge	PR	PR	PR	PR	PR
River generated	P	PR	PR	PR	PR

Table 2. Table to show the ability of various types of current to produce the observed structures of the sandstones and siltstones of the Innerelv Member.

3) Sediment transport is channelised in the shallowest water facies but as sheets in deeper water.

4) Currents fluctuate greatly in strength; this is particularly noticeable in the highest energy facies.

From the discussion of Chapter 2 five types of currents might have produced these deposits and Table 2 shows the likelihood of each type producing the observed features. The table shows the difficulty in distinguishing between three possible models:

(A) TIDAL MODEL - in which there is a constant daily movement of material by tidal currents; fluctuations in current strngth would be due to the natural periodicity of tides amplified by storm action (Stride 1965). It is difficult to compare the Innerelv Member directly with modern tidal sediments because in the former case the grain size is restricted to $\leq$ very fine sand whereas in most of the latter (e.g. North Sea) there is a wide range of grain size. This difference strongly affects the relative abundance of various sedimentary structures on a small scale (Allen 1969) and maybe also on a much large scale. No precise model can be proposed except to suggest that Facies I.5 was deposited not far below the inter-tidal zone with Facies I.4 rather further offshore. Although tides are capable of producing localised sequences of sediments with a unimodal palaeocurrent trend where ebb and flood currents are well separated, the general predominance of unidirectional currents suggests that tidal currents are the least likely of the three.

(B) RIVER GENERATED CURRENTS - in which the increase in current strength reflects the increasing proximity to a river mouth, the sand having been deposited directly from outflowing river currents. This model has certain attractions; it would happily account for the only moderate mineralogical maturity of the sediment and perhaps for the very restricted range of grain sizes. In this model the cutting of the Facies I.5 channels would have been done by river currents during flood and much of the finer grained infill would have been due to reworking by relatively weak wave and current action. This reworking might partly explain the divergent palaeocurrent directions in Facies I.5 but does not however explain the divergent directions in the thick sandstones of I.4 although the latter could be due to the wide variation in flow direction of river water from delta distributary mouths.

A questionable part of this model is whether rivers entering a sea are capable of transporting very fine sand and silt very considerable distances beyond their mouths as bed load except in instances where turbidity current flow is generated. If turbidity current flow is to be generated a slope is required and there will usually be a discontinuity between the delta top sands and the base of delta slope turbidites (e.g. Collinson 1970); no such discontinuity is seen in the Innerelv Member. In most rivers flowing into the denser medium of sea water sand is deposited closely around the river mouth. However, in this case the large lateral extent of the member, the small number of facies and the considerable thickness of each suggests that the facies belts must have been quite wide thus implying large distances of sand and silt transport within the basin. As the presence of a few trace fossils most probably indicates marine conditions the wide dispersion of sand and silt is in serious conflict with this river generated model.

(C) STORM SURGE MODEL - in which sediment transport is largely confined to storm surges at a coastline which otherwise experiences only weak tidal and wave activity. It is envisaged that Facies I.5 was developed in an immediately sub-tidal environment, the cutting of the channels being due to a localised back flow of water as the storm surge retreats. The sand was largely derived from a shoreline where very fine sand was accumulating. As sand moved further offshore it was no longer confined to channels and spread out as a thin sheet of material. The finest grained material was carried offshore in suspension and finally settled out to give the siltstone laminae of Facies I.1. The palaeocurrent pattern is explicable if the main mode is considered to be the offshore direction whilst some onshore and locally variable transport is to be expected in the shallowest water facies. During fair weather the channels were infilled by finer material.

One objection to this is the absence of any form of intraformational clasts of mud or sand such as might be expected to form during violent storm erosion; this absence cannot be explained by a lack of cohesiveness of the sediment because at least some of the material must have been relatively cohesive to give the small slump packets seen in Facies I.5.

Taking into consideration the regional geological setting it must be remembered that the base of the Member was a time of major transgression and that the Innerelv Member represents fairly uniform shelf sedimentation over a wide area which probably extended considerably beyond the present limits of outcrop. The presumed broad, open shelf conditions would have tended to keep tidal activity at a fairly low level and so we can probably exclude the tidal model thus leaving river generated and storm surge processes. Of the two the storm surge currents are slightly preferred as the ability of river generated currents to carry sediment far into a shallow marine basin must be considered doubtful.

Accepting the storm surge model as the most likely of the alternatives the relationships of the various facies to each other and to other postulated facies is synthesised in Fig. 16.

This model is rigid in that it demands that sediment supply and the distribution of facies with depth be constant with time. This was probably not the case as might be shown by the development of Facies I.6.

Since no shoreline facies can be seen it is difficult to develop the model further. What sort of shoreline was it? Was it a simple beach or barrier-lagoon system backed by an alluvial coastal plain or did the rivers dominate over the shoreline processes such that the beaches might be fringing arcuate delta lobes. A close association with a river in this case might be inferred from the restricted range of grain size of the material and its submature mineralogy.

If one imagines that a river is dominating the sediment supply the model in Fig. 16 certainly must be modified to allow variation in sediment supply which would occur when the river switches its route to the sea.

3, 3. MANNDRAPERELV MEMBER

Introduction

The base of the Member is taken at the horizon at which sandstone replaces the siltstones and mudstones of the Innerelv Member. It has been divided into three informal units which are described in turn; these are the Lower Sandstone, First Coarsening Upward Sequence and Second Coarsening Upward Sequence.

The Lower Sandstone: Manndraperelv Section

Description

This section (Fig. 17) which is 70m thick, was measured along the side of an un-named stream which flows into Tanafjord 0.7km northeast of the mouth of the Manndraperelv on the Digermul Peninsula. This is the best exposure of the lower part of the unit. The upper part is best observed in the coastal outcrop northeast of the measured section even though there are places there where the beds are tightly folded.

The sequence from the Innerelv Member shows a rapid but gradational transition from mudstone through very thin-bedded siltstones into 1-10cm bedded very fine sandstones with interbedded micaceous siltstone and mudstone. These sandstones make up the lowest 15m of the member (Pl. 20). There is a gradual upward increase in bed thickness within the unit (Pl. 21) with beds up to 1m thick in the upper half. However, in the top 10m there is a reversion to beds 10-30cm thick.

The sandstones are all very fine or fine grained; they are subarkoses or quartzarenites, sometimes slightly micaceous and with a hematitic cement which gives the rocks a vivid red colour. They seem to be well sorted but the grain outlines

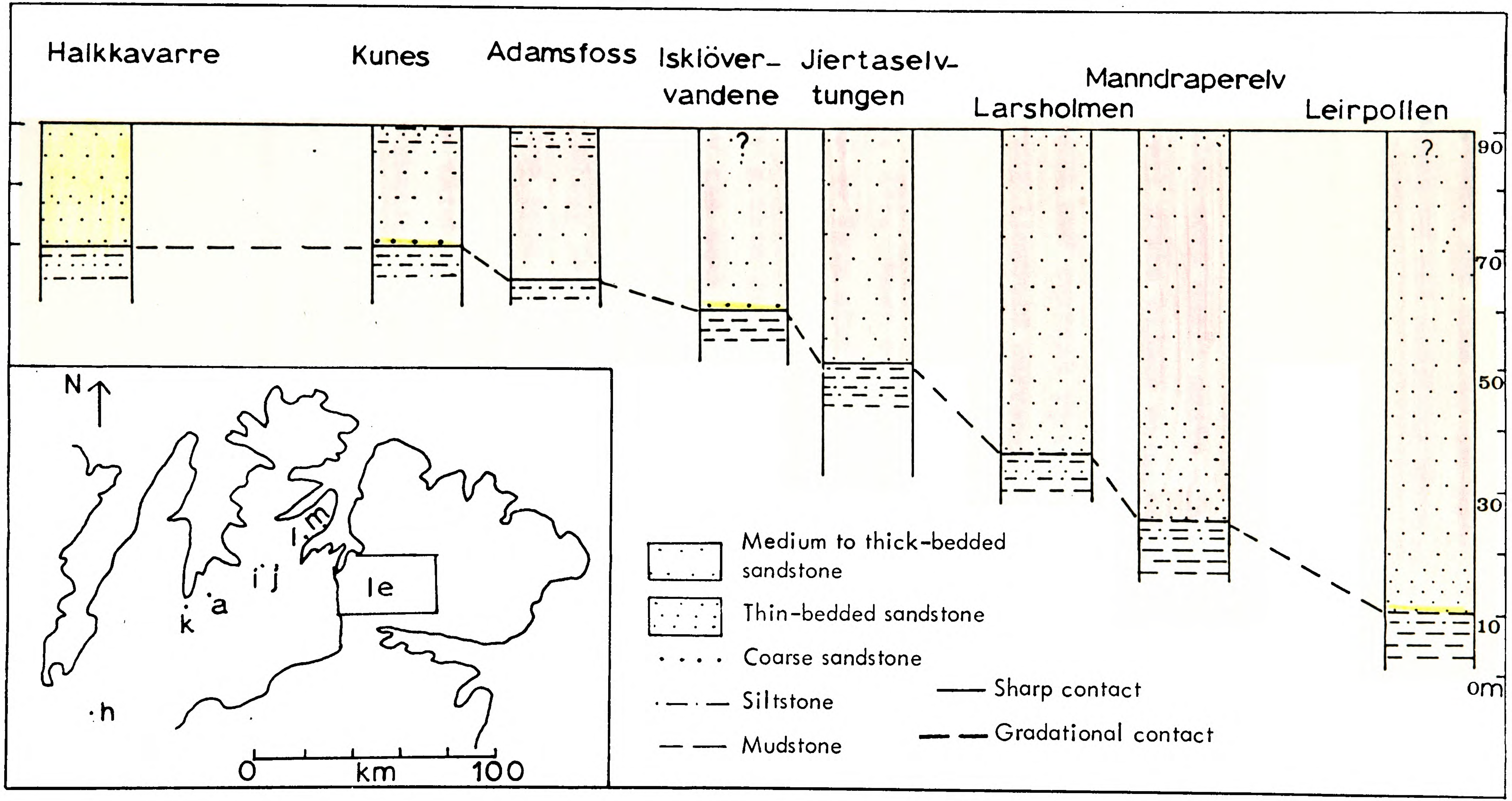


Figure 17. Lateral variation within the Lower Sandstone, Manndraperelv Member.

have often been modified by pressure solution and reaction with the cement and thus no estimate of rounding can be made. Layers of red mudflake pebbles are particularly common at the bases of the sandstone beds.

Internal structures are rarely visible in the beds except for a secondary Leisegang banding but occasionally fine parallel lamination with primary current lineation is seen and also some small scale cross-lamination. Sinuous crested small scale ripple markes are common on upper surfaces and they usually have rather rounded profiles resembling the combined wave and current flow ripples of Harms (1969) (Pl. 22). Less commonly linguoid current ripples and symmetrical ripples are seen. Another upper surface structure consists of roughly equidimensional basins separated by steep sided ridges (Pl. 23). Most of the basins are about 10-25cm across but some up to 75cm also occur. These can be seen to be true bedforms since in cross-section (Pl. 24) they are overlain by a thin drape of finer material which in turn is overlain by the next sandstone bed. In some cases these structures are clearly of erosional origin since parallel lamination within the bed is trucated. This basin and ridge morphology is only one example of the great irregularity in the form of the junctions between successive beds and the rather lenticular nature of the beds, particularly the thicker ones (Pl. 25).

Bioturbation is absent apart from one possible example of a simple surface trail (Pl. 22) found 20m below the top of the unit. The sparse palaeocurrent data (Fig. 18) show that transport was predominantly to the NW.

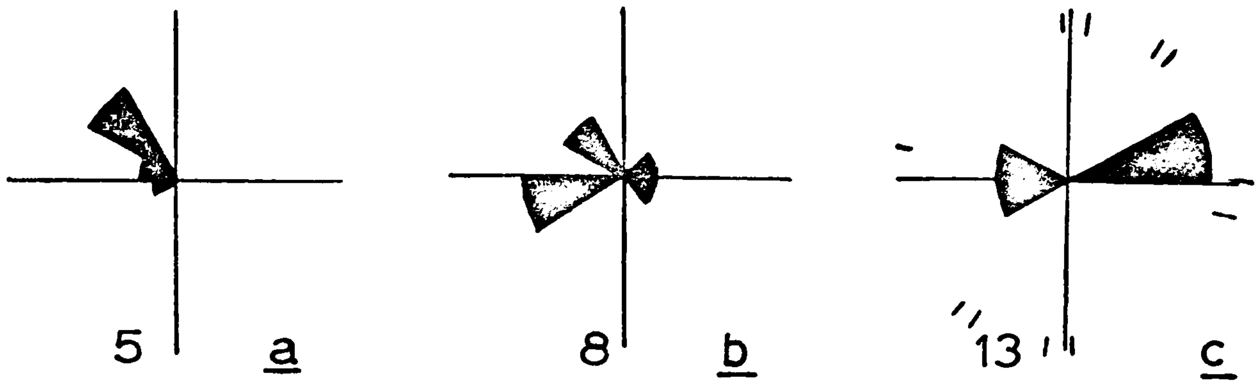


Fig.18. Palaeocurrent data for the Lower Sandstone Unit, Manndraperelv Member. a = Manndraperelv, b = Larsholmen, c = Halkkavarre (Member III). Data from asymmetrical ripples. Axes are symmetrical ripple axes

Interpretation

The increase in grain size upward from the Innerelv Member mudstones into the Lower Sandstone of the Manndraperelv Member shows that there was a gradual increase in environmental energy from that which had prevailed in the upper part of the Innerelv Member. Although there is little variation in grain size in the Lower Sanstone it is inferred from the bed thicknesses that the highest energy and hence probably the shallowest conditions were reached about 50-60m above its base after which there was a reversion to rather lower energy conditions.

The presence of cross-lamination and parallel lamination with primary current lineation shows that moderately strong currents were present whilst the symmetrical and combined flow ripples testify to the influence of wave activity. Indeed the abundance of combined flow ripples suggests that wave activity might have been of considerable importance in facilitating sediment transport in the manner suggested by Draper (1967).

The basin and ridge topography is seen in certain cases to be an erosional feature and in the absence of contradictary evidence this origin is assumed for them all. The presence of undeformed internal lamination precludes a loading or any other type of deformational origin. From their equidimensional form it is unlikely that the basins are current produced erosion structures so it seem s that wave erosion is the only other possibility and perhaps they are a form of erosional interference ripple mark.

Such wave activity is most probably responsible also for much of the irregularity of the bedding planes and where laminated beds drape over the irregular surfaces of underlying beds the lamination may not be a current structure but may be the result of settling of material thrown into suspension by storm waves. Further discussion of the formation of these beds is given after the description of the lateral variation which follows (Fig. 17).

The Lower Sandstone: Lateral Variation

Leirpollen

This section is composite and is based on the data given in Appendix C. At the base there is a sharp gradation up from the fine-grained beds of the Innerelv Member. There is no evidence to suggest that the white sandstone which occurs at the base in the west of the area is of fluvial origin as postulated by Beynon and others (1967); it is in no way different from other sandstones of the unit apart from its colour. Lithologically the section is very similar to the Manndraperelv section and mudflakes are particularly abundant. The top of the unit is not well seen and so it is uncertain whether it becomes thinner-bedded as in the Manndraperelv section.

Larsholmen

Again the Lower Sandstone is very similar to the Manndraperelv section (Acc. ser. 10310) except that it is only 54m thick. Palaeocurrent directions show a more varied distribution (Fig. 18) and in the lowest 5m some possible small burrows are present.

Jiertaselvtungen

At an exposure in the Storelv on the south side of Jiertaselvtungen (Appendix B) the unit is only 39m thick and rests with a sharp planar contact upon red and green mudstones with silty laminae of the Innerelv Member. It is of typical lithology consisting of very fine grained, glassy, red sandstones with beds 10-70cm thick and thin siltstone partings. Parallel lamination is the most common internal structure.

Isklövervandene

The unit again rests sharply on fine-grained beds of the Innerelv Member and the basal bed is a white coarse quartzarenitic sandstone which passes up within 1m into the typical red sandstone facies (Pl. 26, 27).

Adamsfoss

Here the unit, which now constitutes the whole of the member as defined by Føyn (1960, 1967) and Banks and others (1971), is reduced to 25m (Pl. 28). The base is again sharp, the beds are mostly 10-50cm thick and Leisegang banding obscures almost all the original sedimentary structures. In the top 5m the unit becomes thin-bedded and has a large proportion of silt. Some of these beds are graded from medium sand to silt and one shows flutes indicating a current direction towards 030°.

Kunes

In the outcrops south of Kunes the unit is of variable thickness due largely to tectonic complications but also due to some original differences. The maximum measured thickness is 20m, on the hillside overlooking the Stovelv. There, as at Isklövervandene, the lowest sandstone sitting sharply on the Innerelv Member is white but, in contrast, it is fine-grained although a coarse grain size is developed at other localities in this area. Above the basal white zone, which is less than 1m thick, the unit consists of 5-50cm bedded fine and very fine-grained red sandstone before becoming gradually thinner bedded with thicker siltstone partings in the uppermost 4m as at Adamsfoss. In this section the sandstones are similar to those of the Manndraperelv section.

Halkkavarre

The correlation of the Manndraperelv Member at Kunes with Member III of the Dividal Group at Halkkavarre was suggested by Føyn (1967) (Fig. 4). The striking resemblance of the successions in the two areas makes this correlation seem reasonable although it cannot be verified because of lack of outcrop in the intervening 60km.

Member III consists of 20m of grey, 10-50cm bedded, fine sandstones which are mature and supermature quartzarenites and subarkoses. There are subordinate thin-bedded sandstones and siltstones in the middle part of the Member. The contact with Member II is sharp and planar, cutting across the lenticular beds below (Pl. 16). No distinctive basal bed is present; the beds mostly form flat, elongate lenses with length to width ratios of 20 to 1 (Pl. 29). Internally many beds have a complex character with alternations of parallel lamination and small scale cross-lamination but cross-bedding is rarely seen. Bedding surfaces show a variety of ripple types including sinuous crested and straight crested asymmetrical and symmetrical ripples and interference ripples. The few ripples seen have their crests trending approximately north-south (Fig. 18).

The Lower Sandstone: Discussion

It has been shown that the unit thins progressively from 80m in the Leirpollen area to 20m at Kunes and Halkkavarre. In the three thickest sections, as exemplified by the Manndraperelv section, there is a gradual coarsening upward from the Innerelv Member and a gradual increase in bed thickness through the unit suggesting a continual shallowing through this time. Only in the highest beds is there a reversion to slightly deeper conditions.

However, in the western outcrops the sharp change in lithology at the base of the unit and the occasional concentration of coarse sand just above it suggest the presence of a disconformity. There is no evidence of any angular unconformity at the base of the unit although a slight one would not be detected easily. The sharp junction can be explained in terms of submarine reworking and winnowing of the top of the Innerelv Member followed eventually by subsidence and/or lessening of current activity allowing accumulation of sediment. Alternatively, there could have been emergence and erosion followed by transgression.

It is impossible to distinguish between these possibilities since there are so many unknown factors. Above this probable disconformity in the western sections there are no clear trends in bed thicknesses except for the thinning at the top of the Adamsfoss and Kunes sections.

In summary of the regional pattern the Lower Sandstone represents rather uniform shallowing over the whole area.

It is difficult to say how synchronous this was; the base of the unit is almost certainly diachronous particularly if there is a disconformity in the west but the top, at which there is a rapid reversion to finer grained sediment, may well approximate to a time line. Thus similar shallow marine conditions extended over a belt at least 160km long.

There is little evidence from which to make conclusions about the types of current which predominated in this basin. River generated currents can probably be excluded but tidal, semipermanent, coastal storm surge and wave residual currents are all possible.

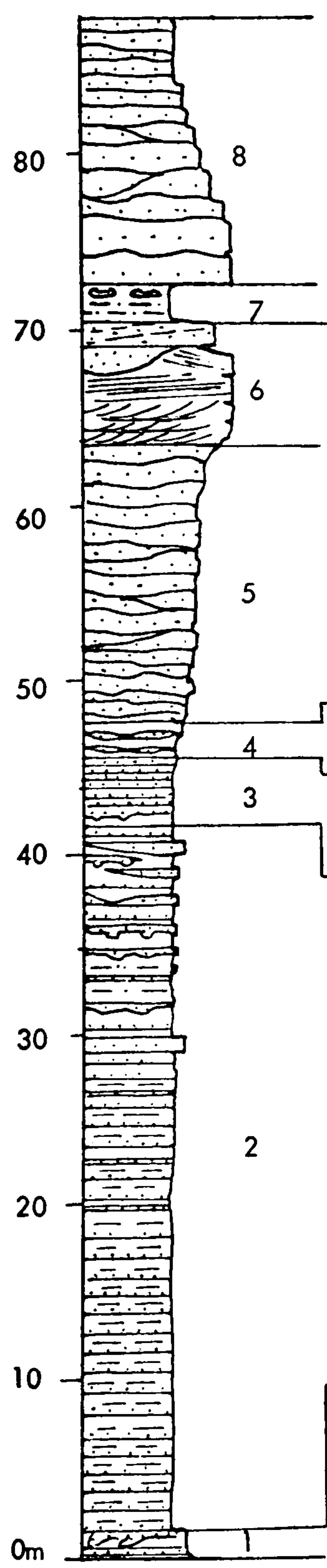
The First Coarsening Upward Sequence: Manndraperelv Section

The section (Fig. 19) was measured in the coastal exposure 2-3km NE of Manndraperelv. The sequence consists of two main parts; a lower part consisting mainly of graded green-grey sandstones and interbedded mudstones, and an upper part composed mostly of red and white sandstones. The sequence has been divided into eight units for ease of description. The junction with the underlying Lower Sandstone is taken at the colour change from red to green which approximately coincides with a sudden change to finer grained sediment.

The main details of the eight units are given in Fig. 19 whilst subsidiary information is given below.

Unit 1

The thickest beds, which occur at the top of the unit, show a clear asymmetry in their ball and pillow structures suggesting slight lateral movement to the WNW. These deformed beds can be traced along the coast (approx. NE) into undeformed beds (Pl. 30). Ripples show that transport was towards the west.



Irregularly bedded red and white sandstones similar to 47.5–63.5m. Bed thickness decreases upward.

Dark red siltstone with sandstone beds towards the top.

Mature and supermature quartzarenite; fine to medium sandstone with trough cross-bedding, low angle bedding and channel sands; some thin lenses of finer material. More details in Fig.22.

Very irregularly bedded red sandstones; beds mostly 20–80 cm, sandstones slightly micaceous, fine and very fine grained, immature to supermature subarkoses and quartzarenites. "Basin and ridge" and large scale hummocky bedding surfaces often draped by thin layers of siltstone. Parallel lamination is the only common internal structure.

1–30cm bedded, lenticular, sandstones and siltstones with grading, climbing ripples, abundant mudflakes, loading and small sandstone dykes.

Numerous parallel laminated and rippled, lenticular, <5cm, green and red sandstones which are micaceous and very fine to fine grained and sometimes graded. Flat symmetrical ripples present.

Graded, micaceous, grey-green coarse siltstones, muddy and silty sandstones andfine and very fine sandstones interbedded with massive green mudstone. Sandstones are immature subarkoses. Beds 1mm to 20 cm in lowest 20m but up to 50 cm between 20 and 40 m. although thin beds still persist. Lowest beds have washed out burrows and small flute and groove marks and are laterally persistent; higher beds show deeper erosion with channels, are increasingly lenticular and slightly coarser grained.
Background sediment changes from mudstone to siltstone upward.
Incomplete Bouma sequences present in all beds: beds less than 1 cm thick usually show cross-laminated division only: lower parallel laminated division predominant in beds thicker than 10cm.

(Acc. Ser. 19311)

1–100cm bedded, greenish-grey, "clean" sandstones and interbedded green mudstone. Sandstones mostly parallel laminated but thinner beds show cross-lamination. Asymmetrical "ball and pillow" structure in places which can be traced laterally into undeformed beds.
Sharp junction with top of Lower Sandstone Unit.

Fig.19. First Coarsening Upward Sequence, Manndraperelv Member, Manndraperelv section.

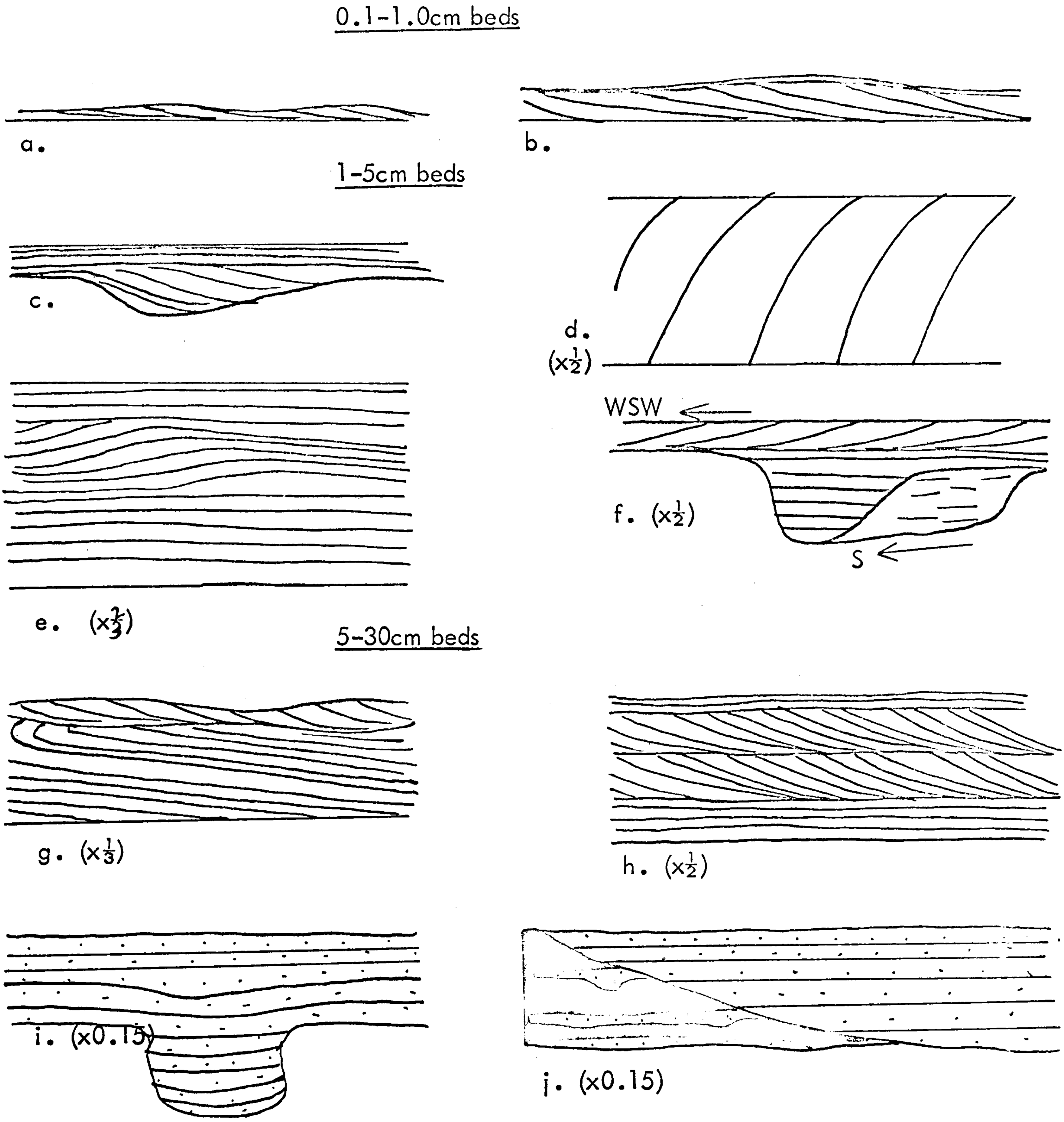


Fig. 20 Typical sedimentary structures in the beds of Unit 2, First Coarsening Upward Sequence, Manndraperelv Member; main section Digermul Peninsula.

a-b. Cross-lamination

c. Cross-lamination filling flute mark and overlain by parallel lamination.

d. Massive bed with grading shown by cleavage refraction.

e. Parallel lamination- cross-lamination- parallel lamination.

f. Very small channel filled with parallel laminated sand and overlain by cross-lamination showing a different current direction.

g. Low angle cross-bedding (dune division) slightly overturned by current shear and overlain by cross-lamination.

h. Parallel lamination overlain by type A climbing ripples and more parallel lamination.

i. Channel, slightly loaded, at base of parallel laminated sandstone.

j. Lenticular parallel laminated sandstone cutting thinner beds.

Unit 2

The matrix content of the sandstones varies from about 20% in the muddy sandstones to about 5% in some of the fine sandstones. Because of diagenetic alteration it is difficult to relate matrix content to the original clay content but this was probably >5% in most cases and the sandstones are thus texturally immature.

Some of the typical sedimentary structures of these beds are illustrated in Fig. 20 and Pl. 31-35. An uncommon feature is the presence of low angle (8-10^{o}) cross-stratification in a few beds between 20-30m (e.g. Fig. 20g). It is interpreted as the result of dune migration in the lower flow regime. It is unlikely to be an antidune structure since the direction of dip is the same as that of the overlying small scale cross-lamination the presence of which also supports a lower flow regime origin. Beds with flute marks exhibit either lower parallel laminated or cross-laminated division of the Bouma Sequence (Bouma, 1962) as their lowest internal structure. As they are of very fine and fine sand grade these observations are contrary to the predictions of Allen (1968) concerning the relationship of flute marks, grain size and initial bed form in graded deposits. However, as pointed out by Middleton (1970 p. 256) Allen's calculations are based on an unrealistic assumption concerning the concentration of the suspension.

Only one bed in this unit does not show evidence of bed load or suspension transport; it is a 2m thick bed occurring at about 30m and has a lower part with intraformational clasts of laminated sandstone in a massive matrix and a completely massive upper part. This is possibly a mass flow deposit of some kind. The lowest 1m of the unit is unique in that is contains a pebbly mudstone bed up to 20cm thick containing granules and pebbles of black mudstone, chert and quartz. Also in this zone even the thinnest beds have very irregular bases and commonly show amalgamation.

The palaeocurrent data (Fig. 21) shows a predominance of southwesterly transport but four beds show indication of transport to the northeast quadrant; two flute marks in the lowest 15m and two sets of cross-lamination between 10m and 20m. These four beds show no differences in petrography or sedimentary structures to others in this zone. These palaeocurrent data contradict Reading's (1965 p. 186) statement that the currents flowed to the north in this part of the succession.

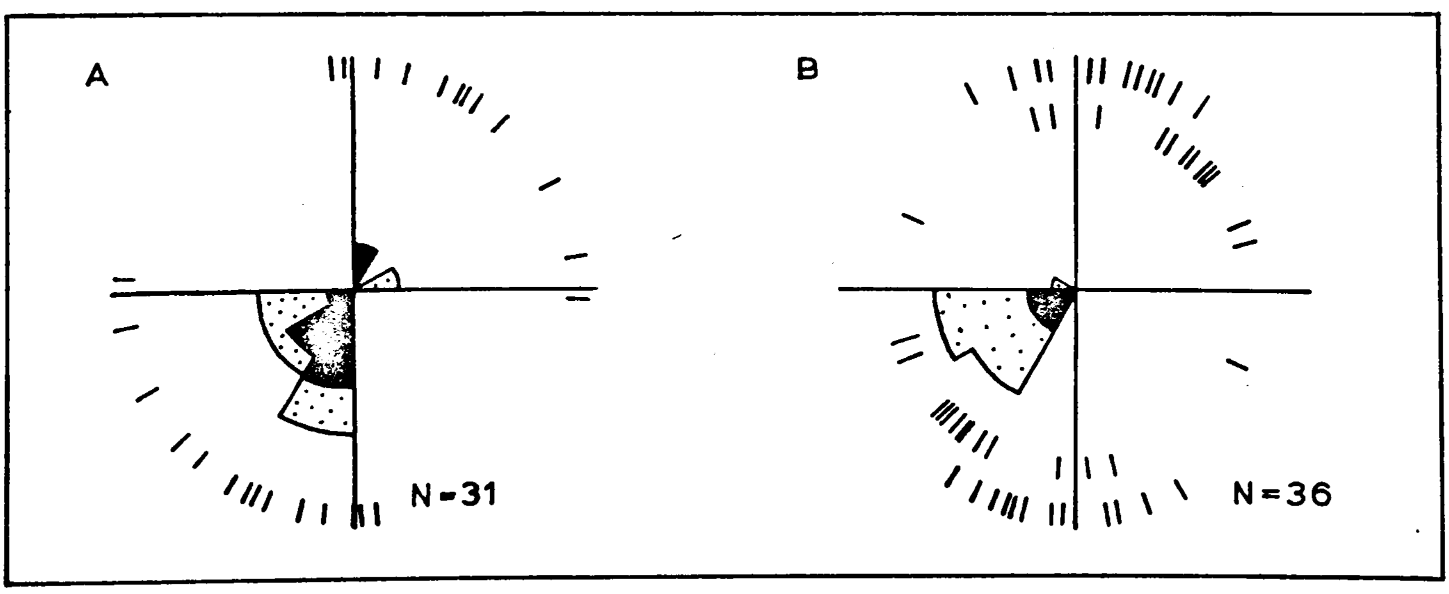


Figure 21. Palaeocurrent data for Unit 2, First Coarsening Upward Sequence, Manndraperelv Member. Black = Flutes, Dots = Cross-lamination, axial lines = groove marks and channels. A = Lower 20 m.
B = Upper 20 m.

Unit 3

The passage from the thick lenticular sandstones below is marked by an increase in the number of lammae and very thin beds of very fine and fine sandstones (Pl. 36) which make up most of the rock. Near the base of the unit some of these thin beds can be seen to pass laterally into thick sandstones identical with those below. Trace fossils are absent in this unit and in all subsequent units of this sequence.

Unit 4

This unit of erosively based sandstones occurs in a strongly tectonised zone but seems to be thickest at the NE end of its coastal outcrop and to thin southwestwards over about 50m.

Unit 5

The lowest few metres of this unit are mostly thin-bedded but this facies soon gives way to medium to thick-bedded red and white sandstones (Pl. 37) which are very similar to those of the Lower Sandstone Unit in their petrography and irregular, undulatory bedding. The three dimensional form of the large undulations such as those of Pl. 37 can only be seen in inland exposures; there, bedding surfaces consist of broadly equidimensional rounded hummocks and depressions forming an egg-carton pattern. Internal structures of the beds are poorly seen but consist mainly of parallel lamination which is either horizontal or curves around the hummocks and depressions of the underlying bed. Very occasionally cosets of small scale cross-lamination are seen within the beds. Leisegang banding is a common feature of the red beds.

Units 6 and 7

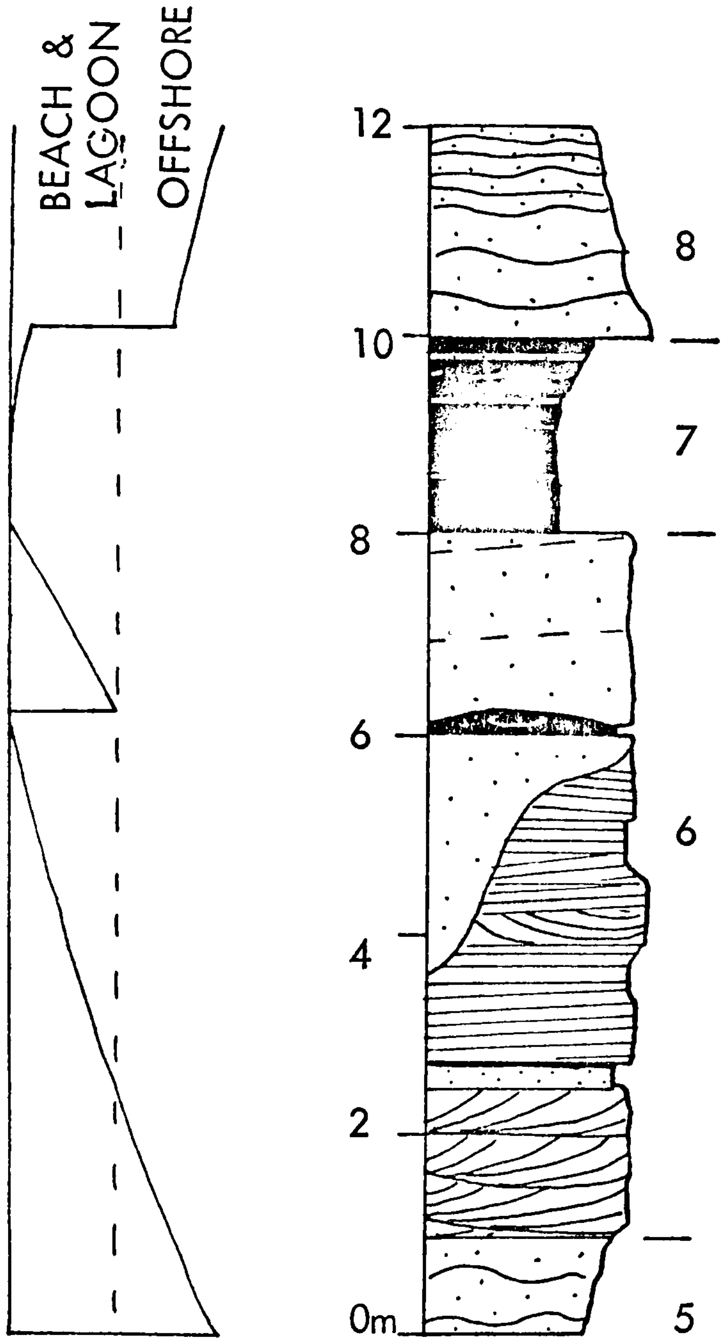
Details are given in Fig. 22. The units are illustrated in Pl. 38-40.

Unit 8

This unit sharply overlies Unit 7 and is lithologically identical to Unit 5 (Pl. 40, 41). Some parallel laminated beds also show primary current lineation but generally sedimentary structures are obscure. There is a gradual upward decrease in bed thickness and a colour change from red to white near the top; also the bedding becomes more regular as the beds become thinner (Pl. 42).

Interpretation

Reading (1965) interpreted this coarsening upward sequence as a regressive cycle passing up from relatively deep water turbidites into shallow water sandstones which were probably marine but possibly continental. Walker (1969) quoted this sequence as a possible example of a small scale turbidity current to "agitated basin" regressive cycle in which the turbidites were generated by river flood processes. In this paper Walker notes the importance of a fully developed regressive sequence in that all the lithofacies can be interpreted in the context of progressively changing conditions of shallowing and progradation of the shoreline; thus if one or two of the lithofacies can be interpreted there are then constraints on the interpretations of the others. This idea is applied to the present sequence and so rather than trying to interpret the units in the order in which they were laid down an attempt is made to deal with the uppermost part of the cycle first and then to try to fit in the lower part. Units 6,7 and 8 will be discussed first.



Unit	FACTS	INTERPRETATION
8	Irregularly bedded sandstones; bed thickness decreases upward.	Offshore marine
7	Dark red poorly sorted siltstone with sandstones increasingly common upward; some sandstones show ball and pillow structures.	Transgression; no beach or immediately sub-tidal facies preserved Lagoon with washover sands
6	Massive sandstone except for vague low angle bedding dipping towards the SW.	Temporary transgression at base; then deposition of another beach sand.
	Lens of red siltstone with ball and pillow sandstone.	Lagoon and washover sand
	Low angle cross-stratified sandstone (Acc.ser. 19312) with fine parallel lamination within sets which dip to N and W. Some trough cross bedding also. Cut out progressively to SW by massive channel sandstone which has a ? N-S trend.	Beach sand (? foreshore and backshore) truncated by a tidal channel sand
	Flat-bedded sandstones with ripples to WNW Overlain by mudstone 3cm thick.	
	Sandstone with trough cross-beds to WSW.	Shoreface sands
5	Irregularly bedded, slightly micaceous sandstones with siltstone partings.	Offshore marine

Fig.22 . Description and interpretation of Units 6 and 7 and parts of Units 5 and 8. First Coarsening Upward Sequence, Manndraperelv Member, Manndraperelv Section.

Low angle cross-bedding overlying trough cross-bedding has often been found in modern beach sequences (Thompson 1937, Ball 1967) and in ancient sequences of probably beach origin (Masters 1967, Lane 1963). This is the favoured interpretation for the lower part of Unit 6 (Fig. 22) and supporting evidence comes from the textural amd mineralogical maturity of the sediment. The variable dip directions of the low angle cross-bedding suggests that both foreshore and backshore deposits may be present. The lenticular sandstone which truncates these beds represents the fill of a channel cut through the beach sediments. Possibly this could have been a well established tidal inlet but more probably, since the sandstone is apparently massive, it was a rapidly cut and filled channel formed where the beach was temporarily breached by storm waters.

The red siltstone facies is interpreted as lagoonal because of its obviously low energy origin compared with the immediately adjacent beds and because of its position above the presumed beach sediments. The clean sandstones occurring within the siltstone were probably deposited as lobes of material washed over the beach ridge during storms. There is no sign of any aeolian deposits so the beach ridge must have been of low relief. However it is inevitable that beach sand must have been widely blown by the wind especially as there was no vegetation at that time to trap it; thus much of the coarser fraction of the lagoonal siltstones may be wind-carried. This would explain poor sorting of these siltstones although this feature may be also a diagenetic effect caused by corrosion of quartz grains by hematite cement.

Above the lower lagoonal unit the sandstone with low angle bedding may be another beach deposit or, alternatively it could have formed by a lateral accreting channel of a tidal delta building into the lagoon. The beach hypothesis is preferred because of the uniformity of grain size within the sandstone and the southwesterly dip of the cross-bedding which fits with the regional pattern of sedimentation (as discussed later). If this sandstone is a beach deposit a temporary transgression must have occurred causing partial erosion of the lagoonal sediments. This was followed shortly by further outbuilding of beach. This transgression could have been a local phenomena caused by a temporary lack of sediment supply related to changes in the pattern of longshore drift or tidal inlets as is seen in the barrier sands of the Niger delta (Allen 1965) and elsewhere. Alternatively the transgression might have been a widespread event.

The upper lagoonal unit, is overlain by Unit 8 which is similar to the facies below the beach sequence (Unit 5). By its position Unit 5 must have been deposited in a sub-tidal offshore environment and thus the same origin is inferred for Unit 8. Therefore the passage from Unit 7 to Unit 8 marks a transgression during which the beach and immediately sub-tidal facies were not preserved presumably due to erosion just below the beach zone. Modern examples of such a transgressive sequence have been given by Oomkens (1967) and Fischer (1961) and modern and ancient examples by Swift (1968). The inferred situation is speculatively sketched in Fig. 23.

Two main problems remain concerning this part of the succession. Firstly, when did the lagoons form? Did they

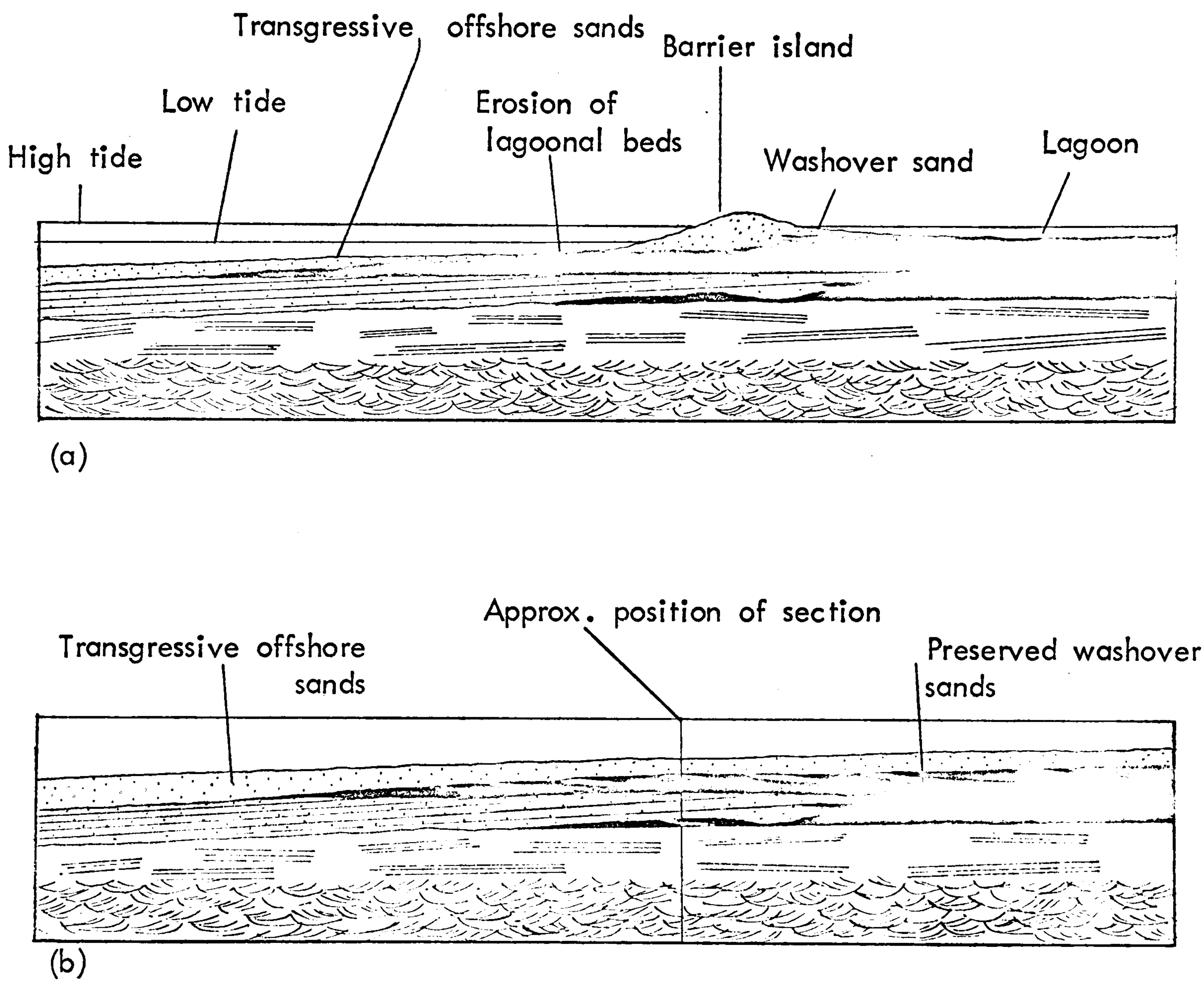


Fig.**23**. Diagramatic sketch of a transgressive barrier island system
a) Early stage, b) Late stage.
Vertical thickness of beds is about 10m and vertical exaggeration about x50.

form behind the beaches as the shoreline prograded and the former beach ridge surface subsided or did they only form after the initiation of transgression when, with drowning of the coastline, the beach ridges became engulfed to form barrier islands? The latter origin of barrier-lagoon systems has been much favoured by Hoyt (1969) for modern examples but his views have beencchallenged by Otvos (1970).

The only evidence upon which to base a conclusion in this section is that the sandstones within the upper lagoonal sequence become more common upward. If these are washover sands this implies that the beach zone was coming progressively nearer to that part of the lagoon preserved in the section; that is it must have been transgressing back over the lagoon at the time of deposition of the bulk of the lagoonal sediments. This suggests that most if not all of the preserved lagoonal sequence is a transgressive deposit and it therefore favours the development of the lagoons by drowning of the coastline. There is no way of deciding from this section whether the transgression was due to a reduction in sediment supply, eustatic rise in sea level or increased rate of basin subsidence but this problem will be returned to later.

The second problem is to decide the direction in which the shoreline was prograding. As the main section is the most northeasterly exposure of the Member and, as will be shown later, it is the only one containing a beach/lagoon system it is inferred that the shoreline prograded from NE to SW.

In summary, it is believed that Units 6 and 7 represent the progradation of a low relief beach ridge system behind which lagoons developed possibly only during times of transgression. Two transgressive events can be seen; the first was only shortlived but the second resulted in a transgressive sequence which continues to the top of the cycle and in which offshore conditions soon became re-established to give deposition of Unit 8.

Passing now to Unit 5 its position in the sequence assigns it to a sub-tidal shallow-marine origin but it is difficult to say what processes were important in the deposition of the sediment. The presence of moderately strong currents is implied by the parallel laminated beds if it is assumed that at least some of these formed under upper flow regime plane bed conditions. River generated current activity can be excluded for these beds because of the absence of fluvial deposits higher in the sequence and because of the mineralogical maturity of the sediments. The strength of tidal activity within the basin can be estimated indirectly. Allen (1965) noted that where tidal ranges are high (e.g. Niger delta, North sea) tidal flats are usually developed behind barrier islands; where tides are low (e.g. Gulf of Mexico) the lagoons have a low-energy character. Since the lagoons in this case are of the low-energy type it can be inferred that the tidal range in the open basin was relatively small and thus that tidal currents were weak although not necessarily insignificant. Wave activity seems to have been strong (see below) and so both wave residual and coastal storm surge currents may have been important in transporting and depositing sediment. However, there is insufficient evidence to draw any firm conclusions. Since the lower part of the unit is thinner bedded and more rippled current energy probably fell off with depth.

The hummocky bedding is thought to be a wave produced feature since its form is unlike any current produced structure. It is probably that, like the "basin and ridge" structure into which the hummocky bedding seems to grade it is a sort of interference erosional ripple pattern. As in the Lower Sandstone some laminated beds which drape over hummocky surfaces may have formed by the settling of storm-suspended material, rather than by current action. However, where a thin siltstone layer separates the hummocky and draping sandstone beds the production of the hummocks and the subsequent drape must be the results of different events.

Having interpreted the upper half of the sequence we can now turn to the lower half and try to fit it into a comprehensive model. The incoming of mudstone at the base of the sequence shows that there was a rapid change to a quieter water environment than had existed during the deposition of the Lower Sandstone Unit. This was probably the result of deepening of the basin, a continuation of the trend seen in the uppermost beds of the Lower Sandstone Unit. However, occasional westerly flowing currents of unknown origin deposited beds of "clean" sand in the lowest 1.5m to give Unit I and from the asymmetry of the ball and pillow structures in these beds the palaeoslope seems to have been towards the WNW.

The sandstones of Unit 2 were deposited from episodic waning currents as is shown by their erosional bases, grading and Bouma sequences of sedimentary structures. The upward increase in bed thickness, lenticularity, grain size and the increasing predominance of parallel lamination over cross-lamination indicate that current activity was stronger in the upper part of the Unit. As the background sediment changes from mud to silt upwards this increase in current activity can probably be correlated with a decrease in the depth of deposition. The association of features described above, plus their high matrix (mud) content, suggests that these sandstones were deposited by turbidity currents and that the changing bed thickness etc. within the unit reflects a transition from a distal to a more proximal environment of deposition (Walker, 1967).

Support for the turbidity current theory comes from the rather abrupt tsansition at the top of Unit 2 from thick, lenticular sandstones into the thinner bedded, lower energy deposits above. Thus there seems to be a discontinuity between the processes operating in the shallow marine sandstone facies of Unit 5 and graded sandstone facies of Unit 2. This is explicable in terms of a turbidity current model if the intervening units (Unit 3 and 4) can be considered as a slope facies; the turbidity currents accelerating down this slope before being deposited at its base. Unit 3 could be interpreted as a slope facies where currents were strong enough to produce marked erosion of the sea floor but the slope was sufficient to ensure that little material was deposited; thus lenticular beds were formed.

What processes could have initiated the development of the turbidity currents? There is no evidence in this case to support the river generated hypothesis of Walker (1969), nor is there evidence of a zone from which they might have been generated by slumping. The most likely explanation is that they could have developed from the suspension and offshore transport of large amounts of sediment during storms on the shelf, a mechanism suggested by Passega (1962). Possibly some sort of valley system existed on the slope to concentrate the flow of material into channels.

Whilst this model has many attractions there is however, some contradictory evidence. If one accepts the evidence from the beach/lagoon sequence that the shoreline prograded to the SW this fits with the predominantly SW flow of the currents but there is then the problem of explaining the few NW flowing currents. Either one must accept the possibility of currents flowing from the other side of the basin up the opposite slope to some extent, or else conclude that the currents were not primarily slope-controlled and therefore were not turbidity currents. It is, however, difficult to think what other type of current could have produced these beds.

With all these ideas and problems in mind the lateral variation of this coarsening upward sequence can now be dealt with.

First Coarsening Upward Sequence: Lateral Variation

The lateral variation (Fig. 24) shows a similar pattern to that of the Lower Sandstone Unit with a gradual thinning towards the southwest. However, in this case the Manndraperelv section is thicker than that in the Leirpollen area. In all areas the Lower Sandstone is sharply overlain by finer grained beds.

Leirpollen.

The composite section of the first coarsening upward sequence (Appendix C) is 80m thick and although broadly similar to the main section it differs from it in several details. The lower part of the sequence consists of sharp-based, slightly graded, grey, micaceous sandstones mostly 5-40cm thick with interbedded green mudstones and siltstones. The thicker sandstones are often lenticular with channelled bases. Internally the beds mostly show parallel lamination which is sometimes overlain by irregular cross-stratification in the thickest beds. The main differences from the main section are as follows:

(i) There is not such a distinct transition from thin-bedded sandstones at the base to thicker sandstones at

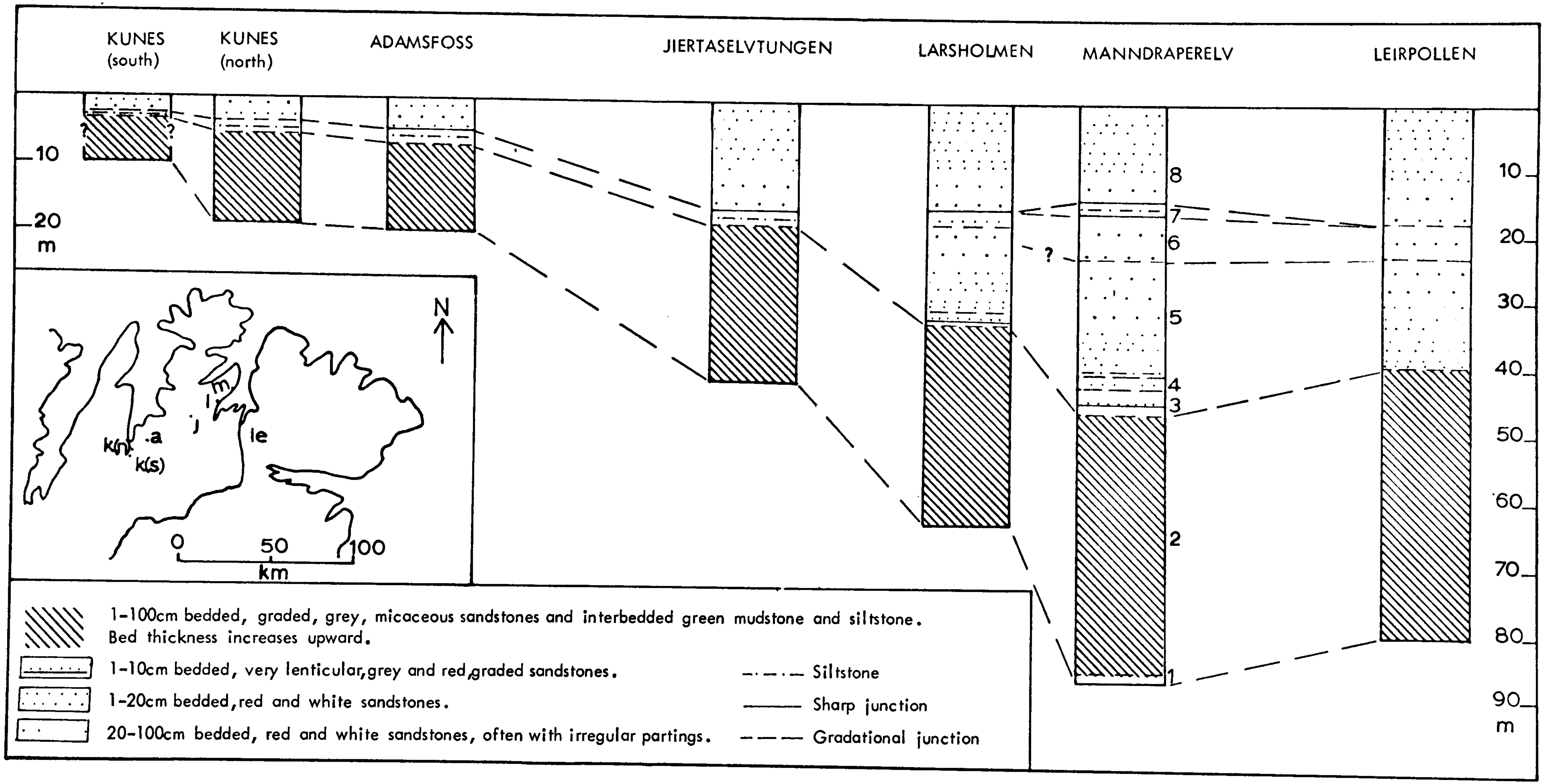


Figure 24. Lateral variation in the First Coarsening Upward Sequence, Manndraperelv Member.

the top; thick beds commonly occur near the base of the sequence.

(ii) There are many fewer <5cm thick beds.

(iii) There is not such an abrupt end to the sedimentation of graded, micaceous sandstones and upward passage into finer-grained beds clearly deposited by other processes; rather these sandstones seem to merge more gradationally into red and white quartzose sandstones of the upper part of the sequence by gradual reduction of the mica and matrix content. This feature is especially well seen in the Vaderelv section although the section north of Lievlamfjeldet shows a sequence which is more similar to the Manndraperelv section.

(iv) Palaeocurrent data (Fig. 25), although very meagre due to the scarcity of flute marks and cross-lamination, indicates a N or NNW transport of sediment as noted by Beynon et al. (1967).

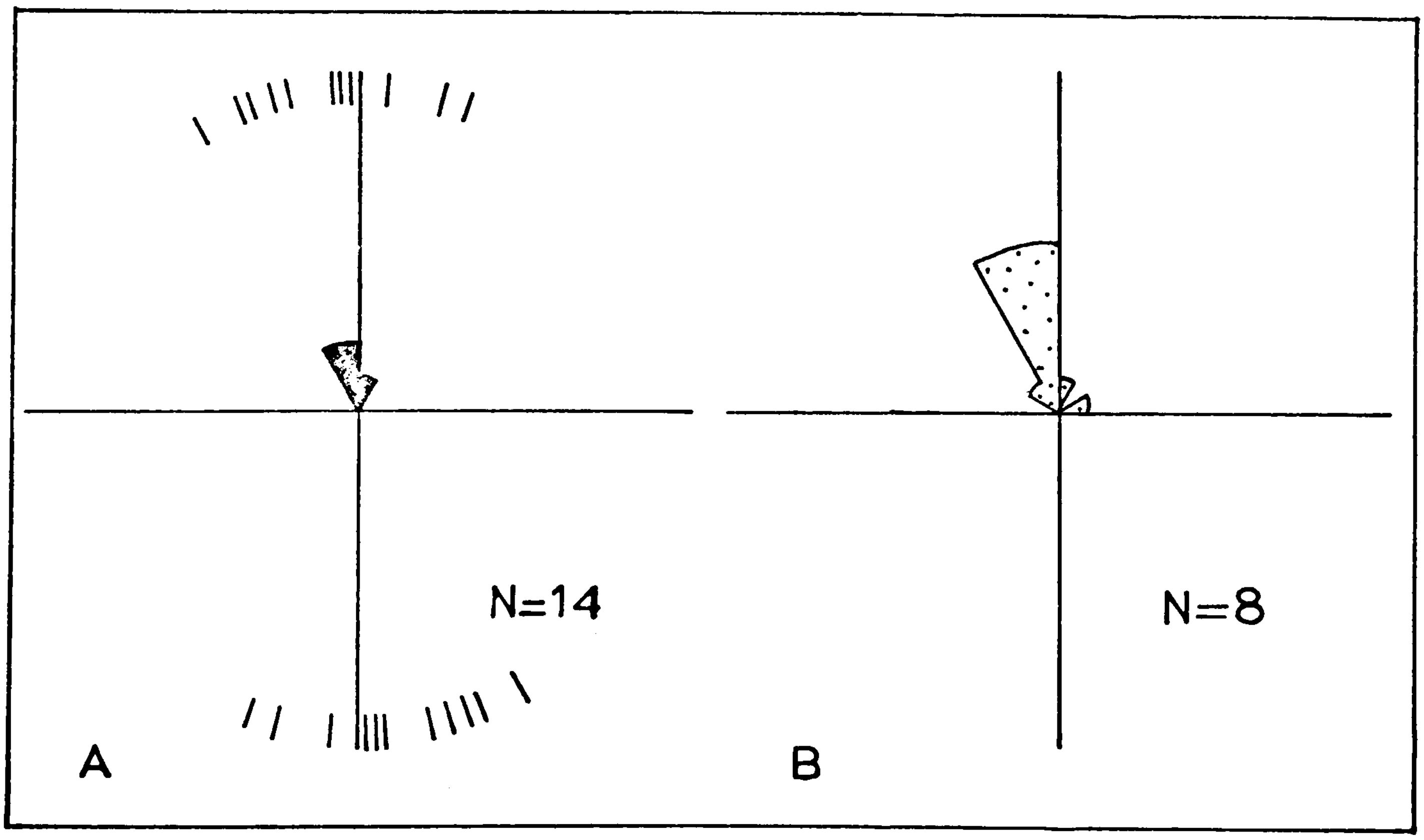


Fig. 25 Palaeocurrent data for the Leirpollen area, First Coarsening Upward Sequence, Manndraperelv Member. Black = flutes, Dots = cross- lamination, axial lines = grooves and channels.
A. Graded, grey sandstones
B. "Shallow marine" sandstones

The red and white sandstones in the upper part of the sequence show more sedimentary structures than those of the main section although they are still not common. Cross-lamination, parallel lamination and occasional cross-bedding are seen as well as "basin and ridge" structure, symmetrical and asymmetrical small-scale ripples and primary current lineation. A few beds near the base of the unit show parallel lamination overlain by cosets of small-scale cross-lamination suggesting that they are waning current flow deposits. Significantly the palaeocurrent directions in these beds are similar to those of the graded sandstones below (Fig. 25). Bed thickness is greatest near the middle of the unit where white and red fine quartzarenites occur in beds up to 1m thick and show parallel lamination, which is possibly within low angle cross-sets, and large ball and pillow structures. Above and below this horizon bed thickness decreases and the sandstones become slightly micaceous and have silty interbeds. By analogy with the main section the thick-bedded sandstones probably represent the point of maximum regression and formed close to or at the shoreline and the sandstones above were deposited during the transgressive phase.

Larsholmen

The section exposed in the fault gulley above Larsholmen (see Appendix B) is only one of many sections which can be measured along the outcrop between the Manndraperelv section and the E.4 highway (Jiertaselvtungen). Good exposures are present in a stream just northeast of Stappogiedde, in the Innerelv valley, and in the more strongly folded area between Rasmuselven and Moskeviken. However, none of these show any features substantially different from that at Larsholmen.

The lowest 31m corresponds closely to Unit 2 of the main section and 31-33m corresponds with Unit 3 in which many thin lenticular beds replace the thicker ones below. Above 33m thin ribs of cleaner, red sandstone are interbedded with red siltstone and there is a gradual increase in bed thickness so that by 36m the beds are mostly 10-30cm thick. This facies corresponds to Unit 5 of the Manndraperelv section and continues to 46m where it is mostly thick bedded. Above this comes a 2.3m unit which is composed mainly of thin-bedded, rippled, grey sandstones but which, in the most southwesterly part of the exposure, consists of very thin-bedded green-grey silty sandstones. This unit is sharply overlain by thick-bedded, white sandstones which, however, become thinner bedded upward and therefore correspond to Unit 8 of the Manndraperelv section.

Thus in the Larsholmen Section the sequence can be considered as regressive up to the unit of thin-bedded, grey sandstones and the beds above, corresponding to Unit 8, form the transgressive sequence. Although the unit of thin-bedded grey sandstones which pass laterally into silty sandstones has no lithological similarity to the lagoonal sediments of the Manndraperelv section, it occupies a similar position in the sequence. Possibly it formed in a sheltered offshore environment, perhaps behind a sand bar. No similar fine-grained horizons occur in the exposures between this section and the Manndraperelv Section.

The palaeocurrent data shown in Fig. 27 were collected in the Larsholmen section and in the Rasmuselven-Moskeviken area. The information is sparse due to poor exposure of bedding planes but there seems to be a divergence between the westerly directed cross-lamination and the roughly N-S

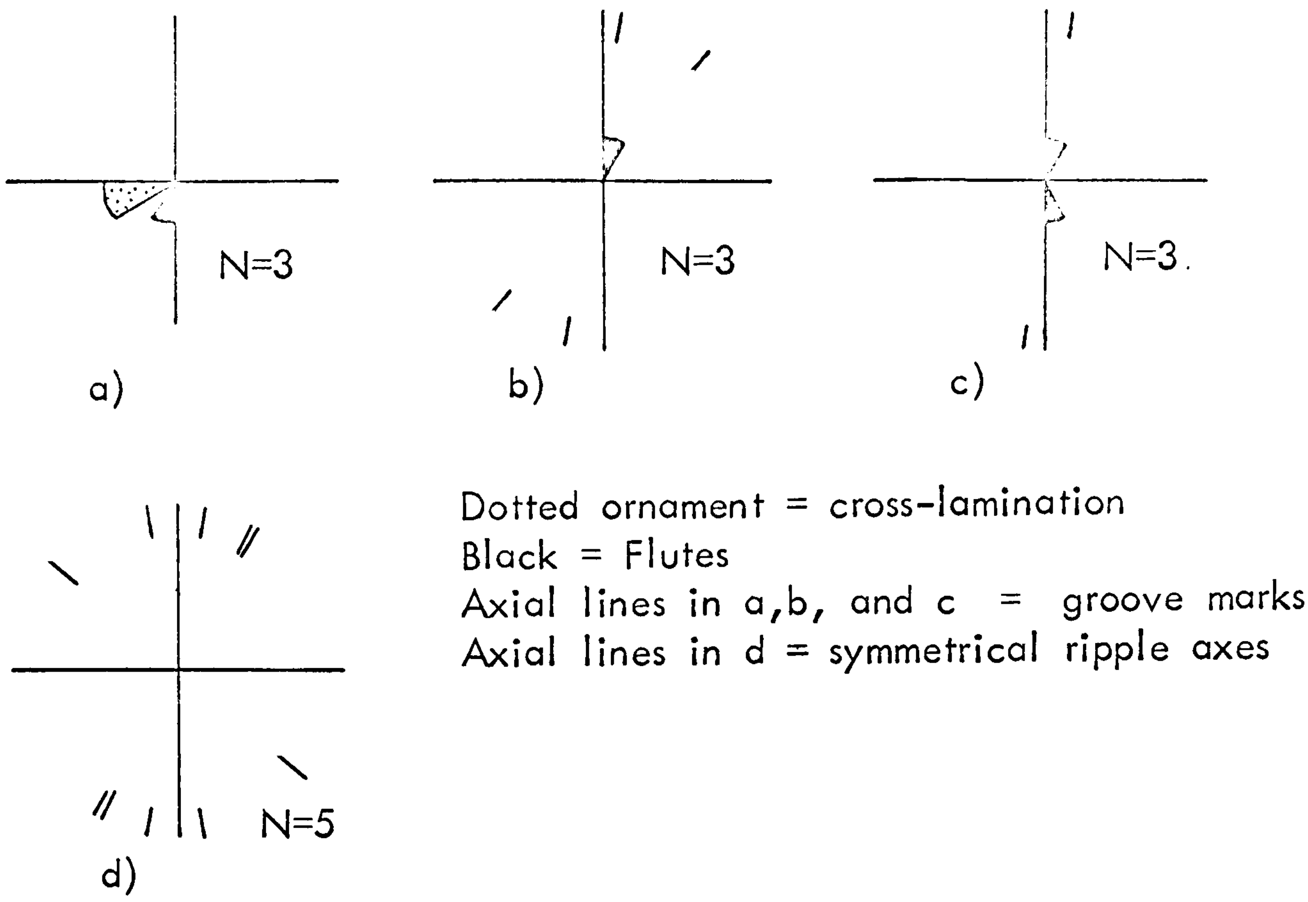


Fig. 26. Palaeocurrent data for western sections of the First Coarsening Upward Sequence, Manndraperelv Member. a,b and c are from the grey, graded sandstones at Jiertaselvtungen, Adamsfoss and Kunes (north) respectively and d is from the red and white sandstones at Jierta selvtungen.

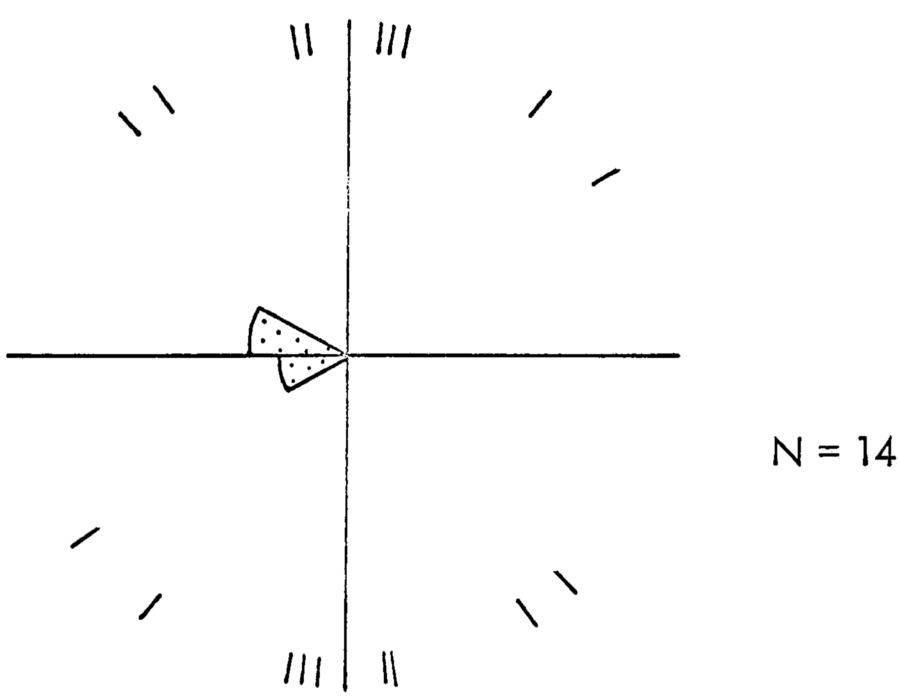


Fig. 27. Palaeocurrent data for sections of the First Coarsening Upward Sequence (graded, grey sandstones) in the Larsholmen area. Dotted ornament shows cross-lamination, axial lines show groove marks and channels.

orientated channels. The divergence is thus between the depositional and erosional phases of the current flow.

Tiertaselvtungen

The section shown in Fig. 24 is a composite of two closely spaced sections (Appendix C) occurring in this area. Tectonic complications make the thickness of the lower part of the sequence uncertain but that of the upper part is well established. The lower part consists of the usual graded, grey sandstones which increase in bed thickness upward (Unit 2) before thinning rapidly and becoming red (Unit 3). Palaeocurrent data are shown in Fig. 26.

The main difference from the more easterly sections is that the red and white sandstones of the upper part of the sequence sit sharply on the underlying beds and show a gradual decrease in bed thickness upward rather than first increasing then decreasing in thickness. The lowest beds are 20-50cm bedded, clean, pink sandstones which are internally massive apart from occasional cross-lamination but have undulatory partings. They pass up into redder beds which are slightly micaceous in places and show abundant symmetrical ripples of variable trend. The top two metres of the sequence consists of white, 10-30cm bedded sandstone with shaly interbeds.

Adamsfoss

Although the sequence is now less than 25% of its maximum thickness, the same major divisions can be seen. The lowest 13m, corresponding to Unit 2, contains some medium sandstones near the base but most beds are very fine sandstones or muddy and silty sandstones. Beds up to 15cm thick occur near the top of the unit which is abruptly overlain by a coarsening upward unit of green and red laminated siltstone and very fine sandstone. This unit is overlain by 10cm bedded white sandstones with abundant wavy ripple cross-lamination but the junction between the two units is not exposed. Bed thickness increases to 50cm in the middle of this sandstone unit but decreases again towards the top of the sequence. The thickest beds show parallel lamination.

Kunes

In the northern outcrops the succession is very similar to that at Adamsfoss, except that the white sandstones rest gradationally on the underlying siltstones and occur as beds 5-50cm thick.

However, in the southern outcrops the coarsening upward sequence is only 10m thick. The section is rather uncertain owing to complex small-scale shearing and folding but since the white sandstone unit, which might be expected to retain its thickness, is substantially thinner than in the north, it may be that there was a significant initial difference between deposits in the north and south of the Kunes area.

Halkkavarre

No coarsening upward sequence is developed here, the probably equivalents consisting of 1-50cm bedded sandstones intercalated with grey mudstones resting abruptly upon Member III (≃ Lower Sandstone Unit). These beds are fully described in the chapter on the Lower Breivik but it can be noted in passing that they were probably deposited in a quiet basin of moderate depth and that the current directions are towards the SW. Whilst it is probably that these beds are laterally equivalent to the first coarsening upward sequence, it is also possible that another non-sequence exists at the top of Member III and that no lateral equivalent exists. Such an interpretation would significantly alter the palaeogeographic reconstructions.

In summary, six main points can be noted.

(i) The sequence thins from 87m at Manndraperelv to 10-20m in the Kunes area but is not developed at Halkkavarre where (?) deep water conditions were probably continuous.

(ii) The sequence of facies graded sandstones - siltstone - "shallow marine" sandstones is present everywhere except in parts of the Leirpollen area where the graded sandstones pass up gradationally into the"shallow marine" beds.

(iii) In the three most easterly outcrops a sequence of increasing and then decreasing bed thickness within the "shallow marine" sandstones probably reflects regression followed by transgression.

(iv) In western outcrops the "shallow marine" sandstones show no consistent variation in bed thickness.

(v) Shoreline sediments only occur definitely in the Manndraperelv section but are possibly also present at Leirpollen.

(vi) Palaeocurrent data are meagre but the direction of current flow was variable.

First Coarsening Upward Sequence: Discussion

The available information in the poorly exposed and structurally complex western outcrops is insufficient to provide firm evidence about the development of the First Coarsening Upward Sequence and so this discussion must be rather speculative. The main problems are:

(i) The origin of the currents which deposited the graded sandstones in the lower part of the sequence and associated with this are problems as to the shape of the basin, its submarine geomorphology and the variation in subsidence within it.

(ii) In the western outcrops it is difficult to say what proportion of the "shallow marine" sandstones were deposited in the regressive phase and what in the transgressive.

Dealing first with the graded sandstone problem it is clear that in most sections these beds are quite distinct from the overlying red and white sandstones of presumed shallow marine origin and are separated from them by finer grained (siltstone) beds. But in the Leirpollen (Vaderelv) section the more gradual passage from one type of bed into the other suggests that there is a close link between the mechanisms producing the graded sandstones and the shallow water sandstones. If one assumes that the deposition of all graded beds was related to currents originating in shallow water, one can explain the break between them and the "shallow marine" sandstones by a slope facies between them as suggested for the Digermul section.

Thus the most plausible hypothesis is to invoke small-scale turbidity current action to augment the transport of sand and silt into the deeper parts of the basin. Material was brought to the shelf edge by some other type of current and then the slope was sufficient for autosuspension (Bagnold 1962) to develop to some extent thus considerably boosting the power of the current. The lack of an obvious slope facies in parts of the Leirpollen area suggests that the slope was locally slight there, autosuspension probably of minor importance, and the graded beds largely the product of the initial type of current. Elsewhere the turbidity current mechanism probably developed sufficiently to continue sediment transport long after the other, initial current had ceased.

If one is to accept this hypothesis one must also accept some of the puzzling features of the palaeocurrent pattern. One must accept that the northeastward flowing currents in the turbidites of the Manndraperelv section were derived from the opposite side of the basin and flowed uphill for the last part of their journey. One must also accept that the northerly flowing currents in the Leirpollen area probably turned sharply around to the west after passing through the area, otherwise it is very difficult to reconstruct a realistically shaped basin.

On the basis of the palaeocurrent data and the distribution of shoreline sediments the general trend of infilling of the basin has been speculatively sketched in Fig. 28.

One feature which requires explanation is the very sudden transition from thick turbidite sandstones into shallow marine sandstones in the western sections with no slope facies in between. Imagine that the basin had an initial topography similar to lower boundary of the cross-section shown in Fig. 30 (i.e. sediment simply infilled a depression with little additional subsidence during sedimentation): then, by plotting the thicknesses of the various facies and giving approximate slopes to each, it is possible to get some idea of the original topography of the basin. This technique shows that, taking a roughly E to W section through the basin, the slope must have become less and less during infilling and must have been virtually absent at the time of formation of the upper parts

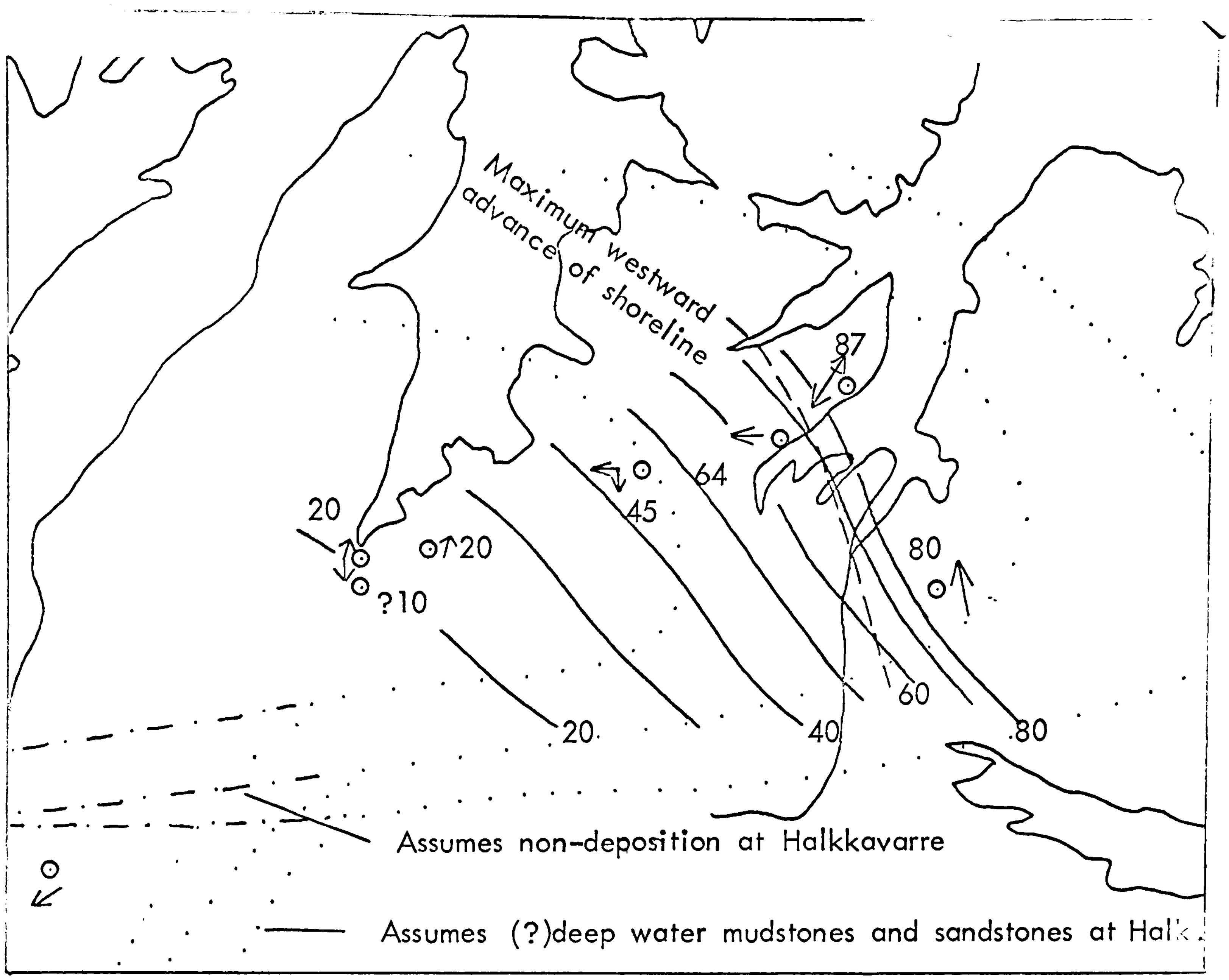


Fig.28 . Speculative sketch showing the development of the First Coarsening Upward Sequence, Manndraperelv Member. Arrows show palaeocurrents in graded sandstones, numbers refer to thicknesses and full lines are tentative isopach lines. The dotted lines suggest the position of the shelf edge during successive stages of infilling of the basin. The dashed line shows the western limit of shoreline progradation.

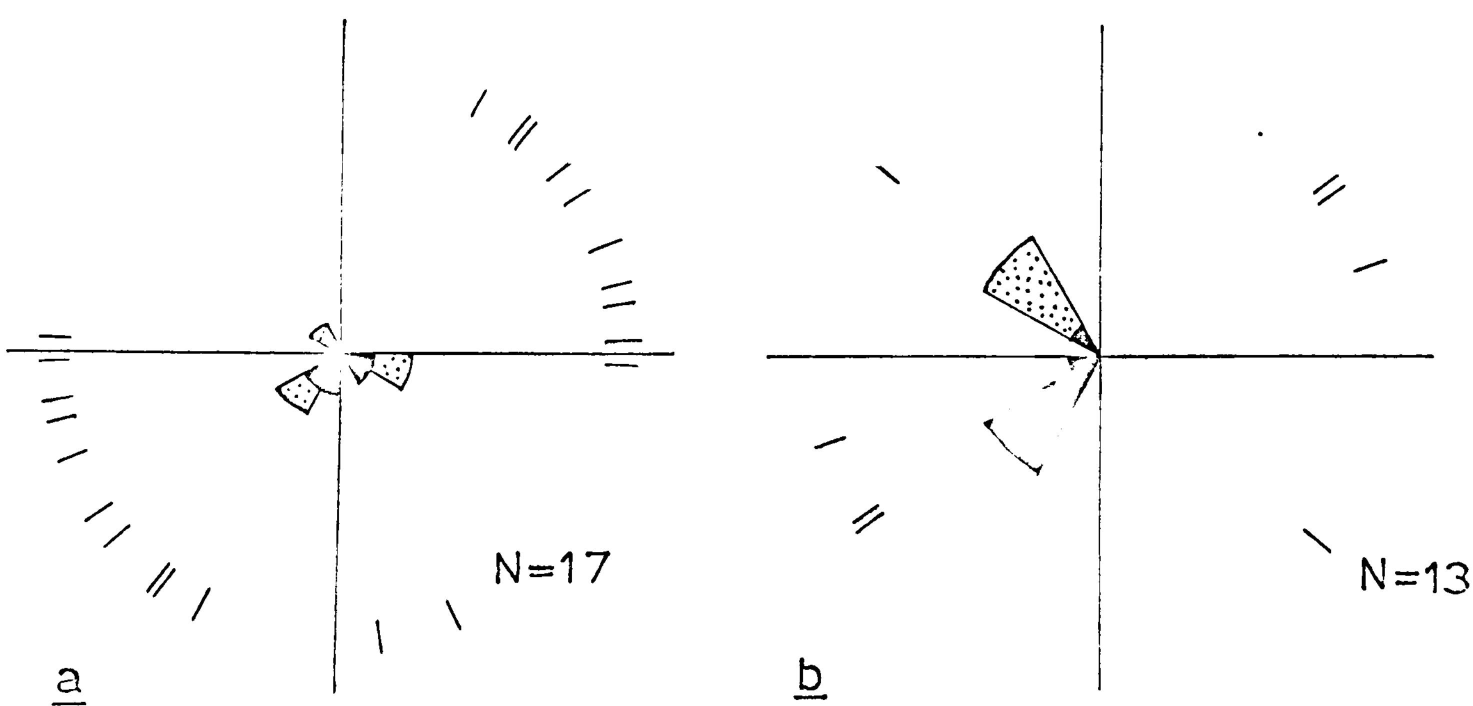


Fig.29. Palaeocurrent data for the Second Coarsening Upward Sequence, Manndraperelv Member, Manndraperelv section. a = 0-18 m, b = 18-30 m. Black = flute marks, dotted ornament = cross-lamination, axial lines = groove marks, channels and primary current lineation.

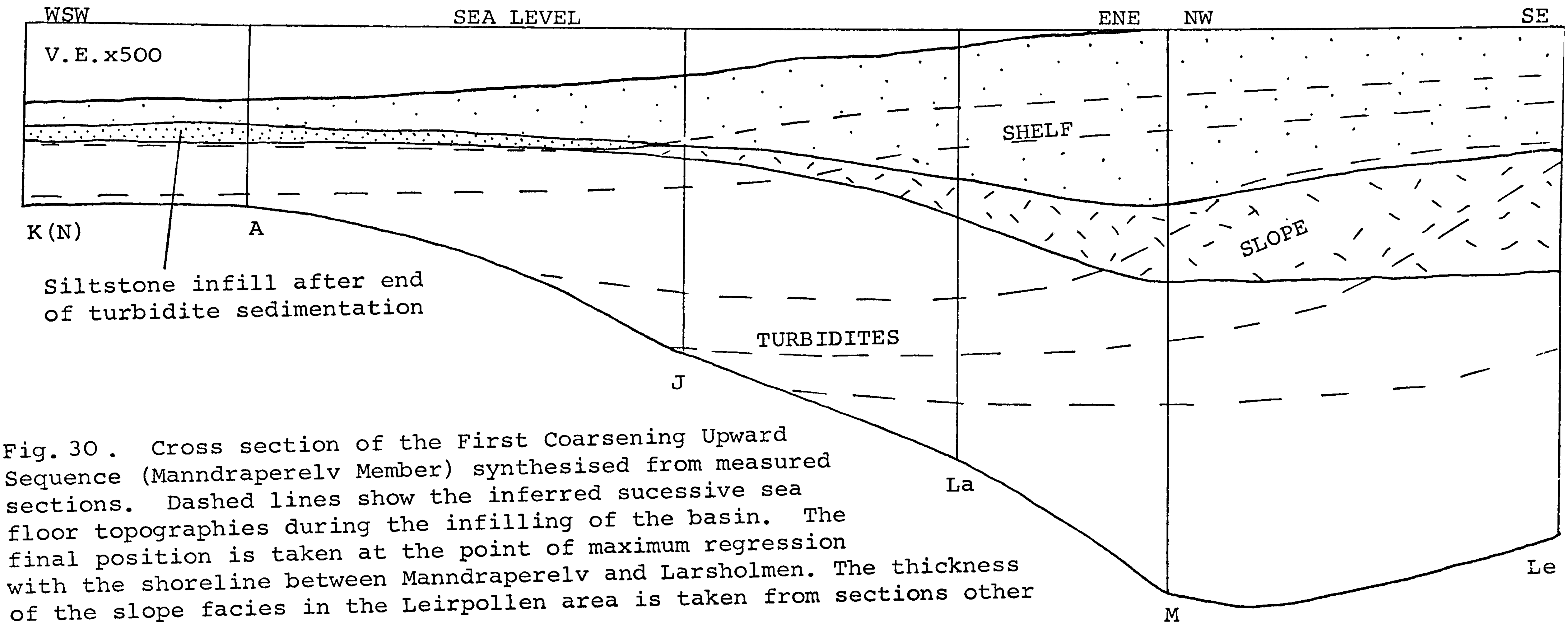


Fig. 30. Cross section of the First Coarsening Upward Sequence (Manndraperelv Member) synthesised from measured sections. Dashed lines show the inferred sucessive sea floor topographies during the infilling of the basin. The final position is taken at the point of maximum regression with the shoreline between Manndraperelv and Larsholmen. The thickness of the slope facies in the Leirpollen area is taken from sections other than that in the Vaderelv.

of the western sections. This cross-section is not strictly applicable since the scant evidence suggests N and S flowing currents in the western sections, but the same situation probably exists for a N-S section also. The stage when the slope became insufficient for turbidity current generation corresponds to the sharp top to the turbidite sediments. The sediments above mark the infilling by other processes of a basin in which the sea floor had no marked variation in slope. Turning now to the second problem it was noted earlier that a considerable proportion of the shallow marine sandstone facies in the eastern exposures was probably deposited during transgression rather than regression but that in the western exposures there was no clear pattern in the deposits. The normal interpretation would be that the majority of beds were deposited in the regression. However, is it possible that the last stages of regression were a time of winnowing and non-deposition or even erosion (analogous to the disconformity at the base of the Lower Sandstone Unit in this area) and that deposition of the majority of beds occurred only when the basin began to deepen. For example, the Jiertaselvtungen section, in which the thickest "shallow marine" beds rest sharply on siltstones and the sandstones become progressively thinner bedded upward, could be explained by this model. However, the evidence from the other sections is contradictory and so no conclusions can be made.

In summary, the main conclusions confirm the interpretation of Reading (1965) that the sequence passes up from turbidites into shallow marine sandstones but it has been shown that, at least in the east, the sequence is not wholly regressive, the upper part having been deposited during transgression.

The Second Coarsening Upward Sequence: Manndraperelv Section

Description

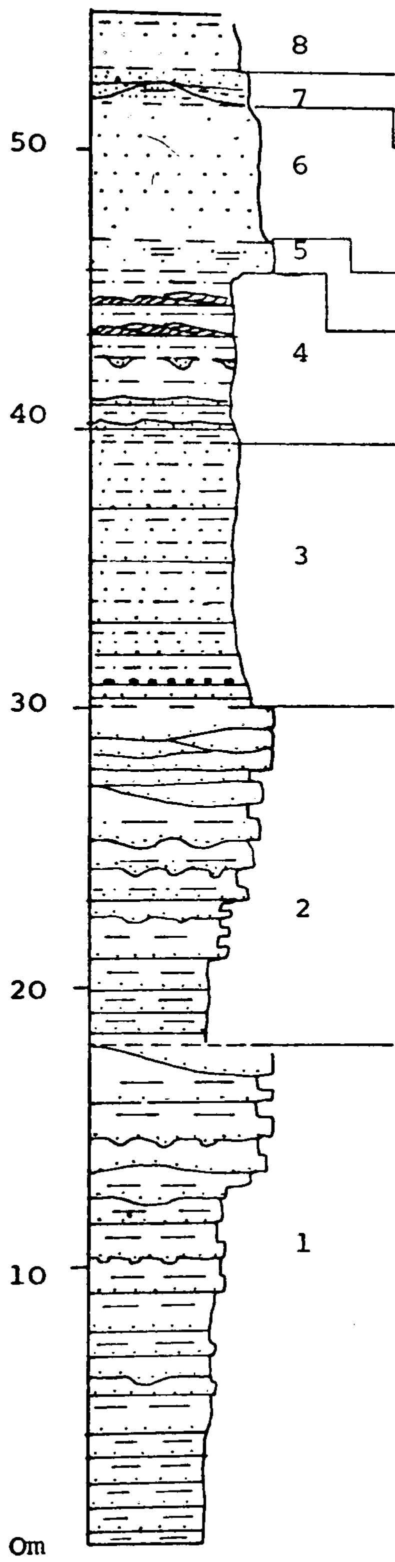
The Second Coarsening Upward Sequence is very similar to the First although its maximum thickness is only 55m as opposed to 87m.

This section was also measured along the coast NE of the mouth of the Manndraperelv and is a continuation of the section through the first sequence. It is cut by a fault near its base which downthrows the beds on the NE side by 6-7m. The beds are again subdivided into units and a detailed lithological description is shown in Fig. 31. As in the first sequence the sandstones are mainly micaceous subarkoses in the lower part and subarkoses and quartzarenites in the upper part.

Unit 1

This unit sharply overlies Unit 8 of the preceeding sequence (Pl. 42). It has many similarities to Unit 2 of the First Coarsening Upward Sequence in its bedding characteristics (Pl. 43, 44). Sole marks consist mainly of groove, prod and brush marks with occasional flute marks. Several of the 20-40cm beds have narrow channels at their bases which are often sculptured with various sole marks (Fig. 32a). J.McD. Whittaker (pers. comm.) has coined the term gutter casts for these narrow channels and this term could be used also for the channels shown in Fig.20f,i. Trace fossils are abundant but only on the soles of the thinnest beds. A few of the thinner beds at about 10-12m in the section have numerous small sandstone dykes extending down into the underlying mudstone. One example of convolute lamination was seen and several examples of low angle cross-bedding similar to that shown in Fig. 20.

Palaeocurrent directions within this Unit are rather variable (Fig. 29).



5-20 cm bedded, rippled sandstones with thin siltstone interbeds. Few beds traceable for more than 25m.

Eroded hump of sandstone draped on one side by siltstone and on the other by sandstone.

Very thin-bedded, parallel laminated and rippled sandstone which is mostly very fine or fine grained. Some 10-20 cm sandstone beds.

Parallel sided, 5-20 cm bedded, fine to medium, red sandstone. Most beds parallel laminated. Burrows at base.

Red and green siltstone with intercalated 1-15 cm bedded, sharp based sandstones, some with types A and B1 climbing ripples. Symmetrical ripples and some burrows present.

Irregularly laminated and very thin-bedded, grey-green, micaceous siltstone and very fine sandstone with some graded beds. Also 5-40cm bedded, lenticular, graded, very fine to medium sandstones with matrix varying from 0-10%. Many beds near base have intraformational mudstone (ironstone) and sandstone pebbles. Symmetrical ripples and trace fossils present.

Identical sequence to Unit 1 except that some of the sandstones are medium grained and the background sediment at the top is siltstone with lamination and low amplitude symmetrical ripples. The unit culminates in a group of amalgamated, lenticular sandstones with primary current lineation, cross-bedding and abundant mudflakes.

Sharp- based, graded, grey-green, micaceous, coarse siltstones, muddy and silty sandstones and very fine and fine sandstones with interbedded massive mudstone and siltstone. Average bed thickness increases from c. 1cm near the base to 10-50 cm near the top. Beds become coarser and less laterally persistent upward. Bouma sequences present with the ripple laminated division common in thin beds and the lower parallel laminated division common in thicker beds. Horizontal burrows common.

(Acc. ser 19313)

Fig.31. The Second Coarsening Upward Sequence, Manndraperelv Member, Manndraperelv section.

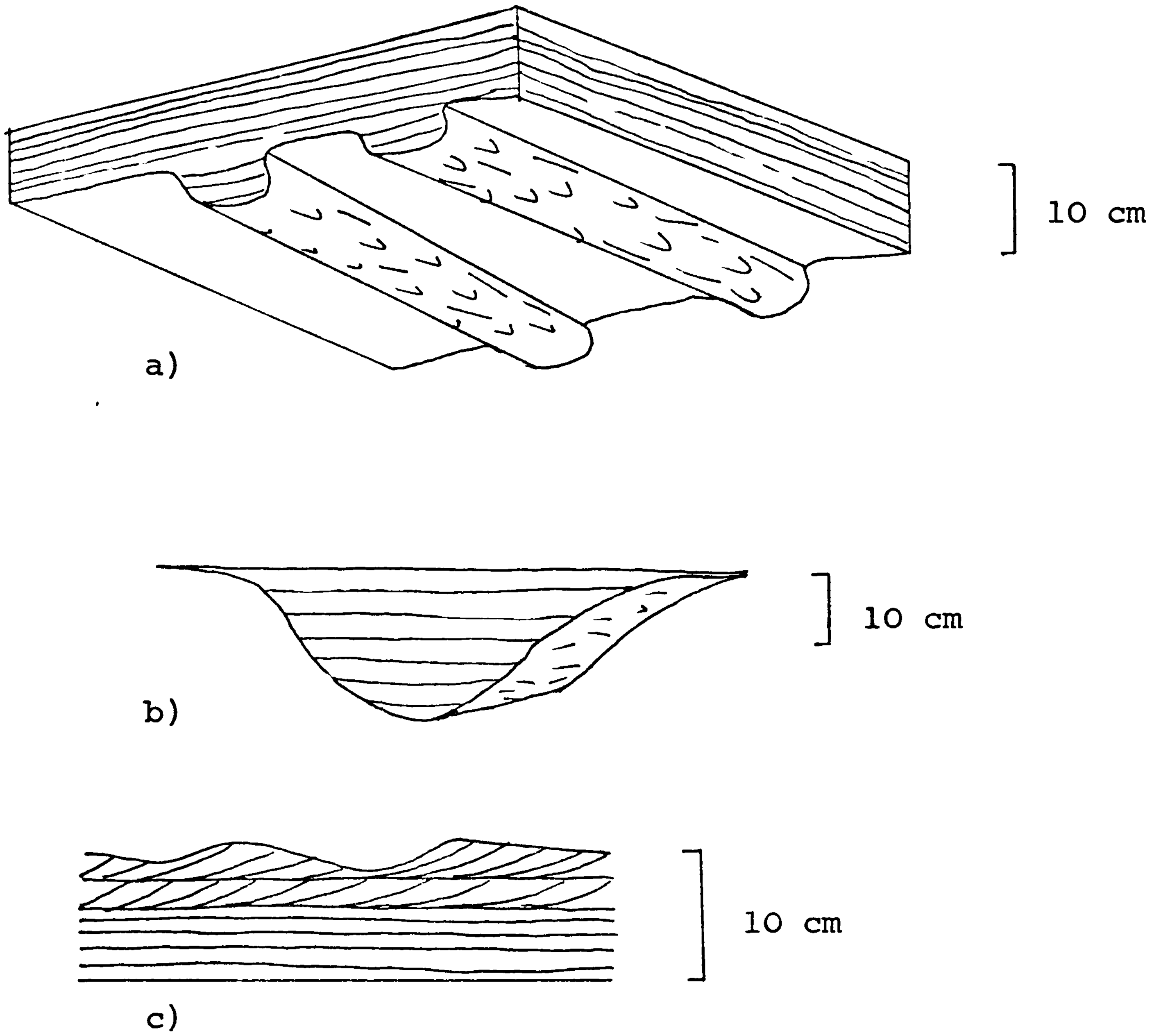


Fig.32. Sedimentary structures in the Second Coarsening Upward Sequence, Manndraperelv Member, Manndraperelv section.
a). Narrow channel sculptured with flute and groove marks on the bases of sandstones in Units 1 and 2.
b). Discrete small channels of sandstone in Unit 4.
c). Matrix-free, sharp-based sandstone showing parallel lamination overlain by ripple cross-lamination and with asymmetrical ripples on the top surface.

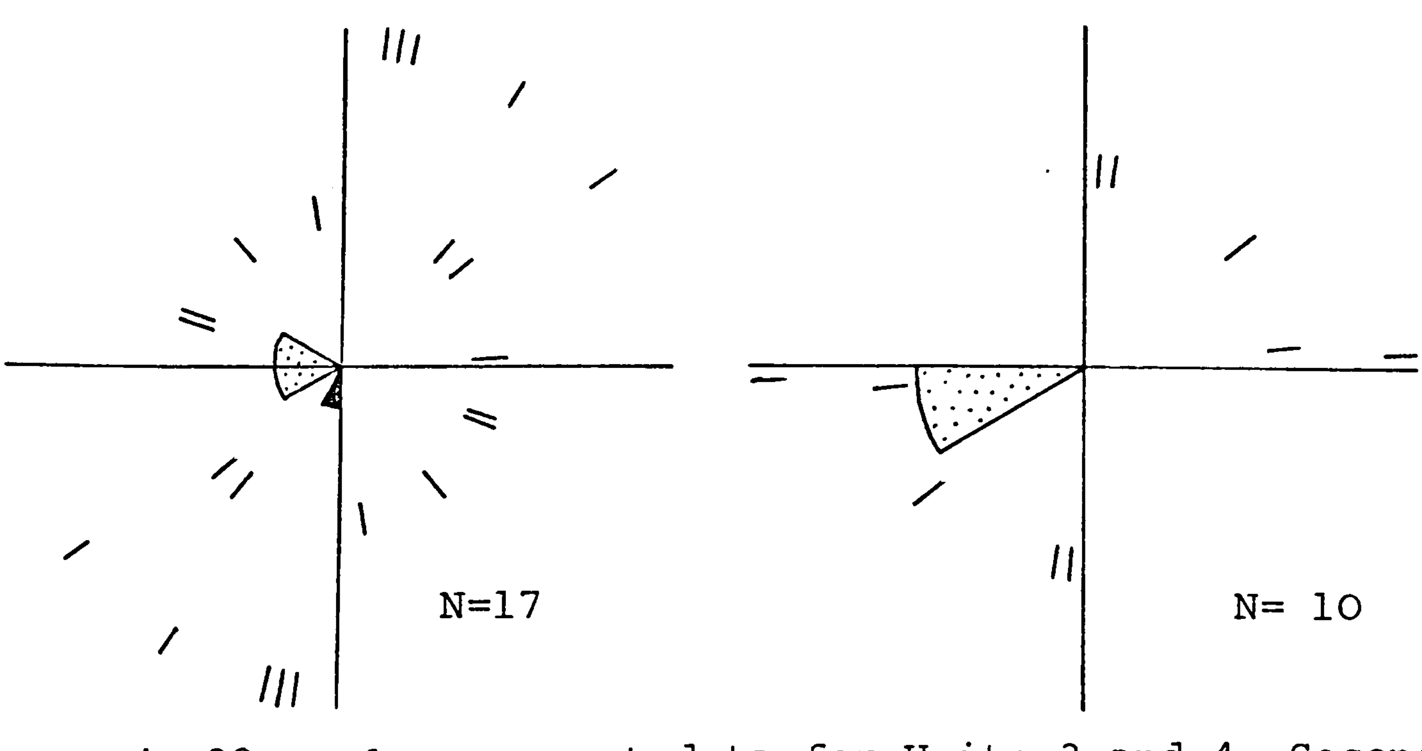


Fig.33. Palaeocurrent data for Units 3 and 4, Second Coarsening Upward Sequence, Manndraperelv Mbr, Manndraperelv section. Black=flutes, dots=cross-lamination, outer axial line = groove marks, inner axial lines= symmetrical ripple axes.

Unit 2

The base of this unit is marked by an abrupt reversal to mudstone which contains only very thin-bedded, graded sandstones and siltstones. However these beds pass gradually up into thicker and more lenticular beds. Some thick beds in the upper part of the unit can be traced for the length of the exposure (c. 50m) but most die out in one or other direction within this distance. Whilst most beds contain >5% clay matrix the lower parts of several of the thicker beds are matrix-free and also moderately to well sorted. At the top of the Unit the group of amalgamated sandstones (Walker 1966) form a distinctive feature (Pl. 45). Several beds wedge out towards the SW, which is approximately the palaeocurrent direction, the lenticularity being in part due to erosion by later currents.

The background sediment at the top of the unit is mostly micaceous siltstone but some laminae and very thin beds of rippled sandstone are also present not all of which show grading or sharp bases.

The palaeocurrent pattern again shows a wide spread in transport directions (Fig. 29).

Unit 3

This unit is of a much thinner bedded aspect than those below (Pl. 46, 47). The thicker sandstones which are present are mostly similar to those below but there are also fine to medium sandstones in which mica and clay matrix are almost entirely absent. These beds are often erosively based, have asymmetrical ripples on their upper surfaces and internally show parallel lamination overlain by cross-lamination (Fig. 32C). Mudstone pebbles, often partly or wholly diagenetically altered to ironstone are common in these "clean" sandstones. Some of the intraformational samdstone clasts near the base of the unit are up to 12cm long (Pl. 48) and lenses and layers of this coarse material are often draped by laminated very fine sandstone.

Palaeocurrent directions (Fig. 33) are similar to those in Units 1 and 2 and symmetrical ripples, which are common (Pl. 49), show a rough E-W alignment.

Unit 4

This unit has some similarity to Unit 4 and the lowest part of Unit 5 in the First Coarsening Upward Sequence (p. 50). The siltstone is mainly red and has a faint brown laminae within it due to variation in the cement rather than to any differences in grain size. Most of the sandstones (Pl. 50) are laterally continuous (at least 30m) but a few are remarkably lenticular forming small narrow channels almost as deep as they are wide (Fig. 32b). The internal lamination shows these are true channels and not load structures. A few beds have well developed horizontal burrows on their soles (Banks 1970, Pl. 1c) and several have concentrations of small mudflake pebbles near their bases. The cross-lamination shows a consistent sediment transport direction to the WSW (Fig. 33).

Unit 5

Unit 4 passes gradationally up into Unit 5 (Pl. 51) in which siltstone is confined to very thin partings between the sandstone beds, the latter showing mainly parallel lamination but also some small scale cross-bedding near the top. Horizontal burrows found at the base of the unit have been figured by Banks (1970, Pl. 2d).

Unit 6

Unit 5 grades up into Unit 6 which is similar to the upper part of Unit 8 of the First Sequence.

Unit 7

This remarkable unit is shown in Pl. 52 and Fig. 33a.

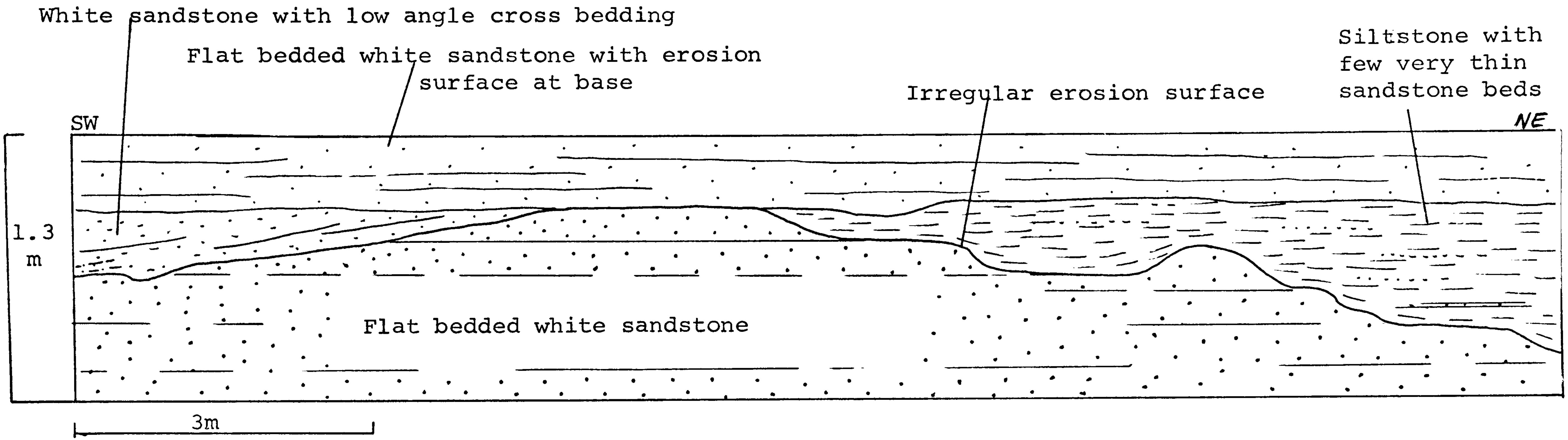


Fig.33a. Sketch of the beds of Unit 7 in the Manndraperelv section, Second Coarsening Upward Sequence, Manndraperelv Member.

The sequence of events seems to have been as follows:

(i) Deposition of flat-bedded white sandstone.

(ii) Erosion to produce an irregular surface with a relief of at least 75cm.

(iii) Deposition of drapes of siltstone on the NE side and sandstone on the SW side; the age relationships of the two deposits is unknown. Some possible modification of the erosion surface on the SW side possibly occurred since it is more planar than the NE side.

(iv) Truncation of siltstone and (?) sandstone by flat-bedded sandstones.

Unit 8

The uppermost sandstones of Unit 7 pass gradationally up into those of Unit 8 which in turn pass up into the thinner bedded sandstones and siltstones which are taken as the basal beds of the Lower Breivik Member. A few poorly preserved burrows occur in this unit.

Interpretation

The similarity of this sequence to the underlying one allows many of the same arguments to be used again. The sandstones of Units 1 and 2 can best be explained as turbidites. The abrupt base of the sequence is interpreted as the result of a deepening of the basin at a time of minimal sediment supply. Thus no slope facies was deposited between the presumed outer shelf sandstones at the top of the underlying sequence and the deep water mudstones and thin turbidites at the base of Unit 1. Since medium to thick-bedded sandstones appear much lower in this sequence than in the first it is probable that the water was never as deep nor the shelf so far away. This fits with the reduced thickness of the turbidites (30m as opposed to 40m).

The rapid reversion to thinner beds at the base of Unit 2 could be due to a sudden widespread deepening of the basin but more probably it reflects a local switch in the direction of sediment transport. If the turbidites were building out as small submarine fans this switch could be due to the abandonment of one established fan or part of a fan in favour of another area where there was a greater slope in a manner analogous to the imbricating delta lobes of Scruton (1960). Unit 2 then reflects the later return of sediment to the site of the former fan once conditions favoured deposition there again. Alternatively the whole system of funneling sediment down from the shelf might have been abandoned and then later re-opened or a new one formed nearby.

The symmetrical ripples at the top of Unit 2 presumably formed as the result of wave activity and suggest shallower water conditions than had existed below.

By analogy with the First Coarsening Upward Sequence Unit 3 ought to be a slope deposit and indeed there are several features which support this hypothesis. The lenses and layers of intraformational pebbles testify to strong, erosive, current activity but the thinness of the beds shows that little sediment was deposited. It is to be expected that on a slope turbidity currents will deposit only the coarsest fraction of the load although some finer material deposited by the tail of the current is also likely. This "tail" material is probably seen in the laminae and very thin beds of siltstone and sandstone which make up most of this unit, although these beds could also have been deposited from other very weak (low density?) turbidity currents or from nepheloid layers as envisaged by Stanley (1969).

It must be noted, however, that the beds in this postulated slope facies are moderately laterally continuous and no large scale lenticularity is seen. This is evidence that the slope environment was not deeply dissected by valleys although it must be admitted that this section could show an area of smooth ground between valley systems.

The remainder of the sequence is rather difficult to interpret in any definite way. One can see abundant evidence in Unit 4 and to a certain extent in the higher units of episodic transport of sand producing sharp based beds intercalated with siltstone or very fine sandstone. The presence of abundant symmetrical ripples probably indicates moderately shallow water although no limits can be placed on this depth. The red siltstone of Unit 4 is similar to the lagoonal siltstone of the First Coarsening Upward Sequence but none of the associated beds suggest the presence of shoreline deposition. The palaeocurrent directions in the sandstones of Unit 4 are consistently to the west and thus approximately the same as those in the underlying turbidites. This presumably is an offshore direction and the currents which deposited these beds may have become turbidity currents on reaching the slope.

In Units 5 and 6 the scarcity of fine-grained sediment suggests more constant agitation of the sea floor but the grain size and bed thickness suggest that currents were relatively weak, particularly in Unit 6.

Unit 7 is a puzzling feature. One major problem is how to explain the formation of the initial very irregular erosion surface. It seems that this must have beem produced by some exceptional event such as a major storm. In modern environments storms can profoundly modify the sediment surface in the nearshore, inter-tidal and supra-tidal environments both by erosion and deposition (Ball and others 1963, Hugh 1967). If the sandstone and siltstone on either side of the "hump" were deposited simultaneously the hump is likely to have been an emergent feature, the siltstone forming in a protected slough behind it and the sandstone forming as a very low energy beach in front of it. Later, as a result of a rise in sea level, sands were deposited over both.

Another possibility is that siltstone was deposited on both sides of the mound but as current strengths later increased the siltstone on the SW side was eroded and sand deposited but that on the NE side was protected by the hump. Whether this could have happened in some depth of water or only in very shallow water is debatable.

Tentatively concluding that Unit 7 was probably deposited in rather shallow water it is likely that Unit 8 was probably deposited in deeper water, a trend which continued with the passage from Unit 8 into the siltstones and sandstones of the Lower Breivik Member.

In general it seems that the shelf environment in the Second Coarsening Upward Sequence was of a lower energy character than that in the First Sequence. The sandstone beds are rarely >20cm thick, there is an absence of the hummocky bedding and basin and ridge structures seen previously and if Unit 7 does represent near shore conditions then wave and current activity were certainly weak except for occasional exceptional events.

Second Coarsening Upward Sequence: Lateral Variation

The lateral variation follows much the same pattern as in the First Coarsening Upward Sequence, the main exception being the substantial thinning in the Leirpollen area (Fig. 34).

Leirpollen

The complete sequence was only seen in the Vaderelv section, although the red and white sandstones of the upper part are well exposed in many places. The sequence is only 28m thick and no true turbidite facies is present except perhaps in the lowest few metres. The succession being broadly similar to the upper 20m of the Manndraperelv section. The lowest 12m consists of thin to medium-bedded green and red micaceous sandstones with interbedded red and green siltstones. The sandstones are sharp-based, some are graded but most appear massive. Sand becomes predominant upward and there is a passage through very fine, red, micaceous sandstones into red, then white, rippled clean sandstones which are thicker bedded than in the Manndraperelv section, being mainly 20-50cm thick. There are no obvious transgressive deposits except perhaps in the uppermost few metres.

Larsholmen

In the lower part of the section the turbidites are not well exposed but instead of the two units showing upward increases in bed thickness there seems to be only one. Above these beds the sequence is very similar to the Manndraperelv section except that the red sandstone unit is substantially thinner. This may be due to tectonic factors since this unit is considerably

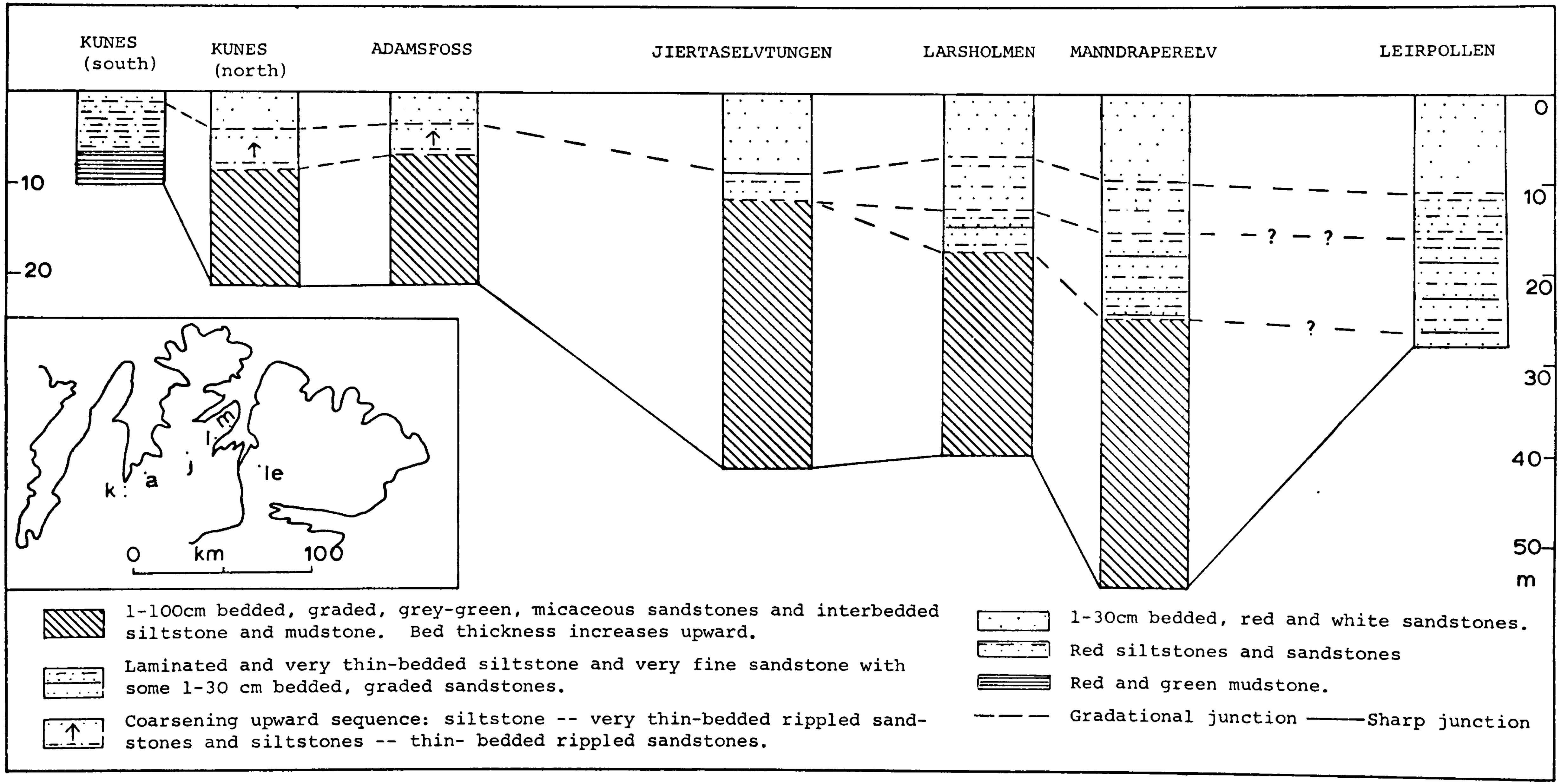


Figure 34. Lateral variation in the second coarsening upward sequence, Manndraperelv Member.

thicker again a few kilometres to the SW in the Rasmuselven-Moskeviken area.

Jiertaselvtungen

This section is a composite of three sections measured in this area which, although closely spaced, are of substantially different thickness due to tectonic thickening and thinning (Appendix C). The First Coarsening Upward Sequence is sharply overlain by green mudstone with a few relatively clean sandstones but these soon give way to micaceous dirty sandstones of typical turbidite apsect. The turbidite sandstones and siltstones show a gradual upward increase in bed thickness and one flute mark and one set of cross-lamination showed current directions towards the south. Bioturbation is clearly seen near the top of the turbidite facies. The grey-green turbidite sandstones pass up into similar looking dark red sandstones interbedded with dark red and brown siltstone. This lithology, which is 3m thick, is sharply overlain by clean red sandstones, 5-15cm thick, which have symmetrical ripples. These beds are overlain by white sandstones which continue to the top of the sequence.

Thus the main feature of this section is the absence of a unit of thinner bedded grey-green sandstones above the thickest turbidites and below the red and white sandstones.

Adamsfoss

A few clean sandstones again occur interbedded with mudstone at the base of the sequence but these give way to dirty sandstones upward. The thickest turbidites are up to 1m thick and the top of this facies is overlain by red and green siltstones. There is then a gradual coarsening upward sequence over 3m of siltstone with isolated sandstone rippled - very thin-bedded rippled sandstones and siltstone - white rippled sandstone with beds 1-15cm (Pl. 53). The uppermost facies continues to the top of the sequence.

Kunes (north)

If the structural interpretation shown in Appendix C is correct the sequence is similar to the Adamsfoss section except that the turbidite sandstone beds are rarely thicker than 10cm.

Kunes (south)

Again the section here seems to be substantially different to that in the north. It begins with green and red banded mudstones and siltstones which are followed by a unit of 1-10cm bedded clean sandstones with interbedded rippled siltstones occurring about half way up the sequence (Pl. 54). This unit is overlain by siltstones which pass gradually up into thin bedded clean sandstones and isolated ripples of sandstone in siltstone and finally into a brown (limonite) spotted medium to coarse white sandstone which is massive except for occasional vague parallel lamination.

Second Coarsening Upward Sequence: Discussion

The substantial thinning of the Leirpollen section and the virtual absence of turbidites there fits with the deductions from the Manndraperelv section that the shelf did not retreat so far in this sequence as it did in the first (Fig. 35). The Leirpollen section can be considered as showing a transition from a slope to an inner shelf invironment. This evidence of a shelf in the east combined with the generally westward palaeocurrents in the turbidites of the Manndraperelv section suggest that the shelf was prograding from east to west in that area. The lack of palaeocurrent data for the rest of the basin makes it impossible to develop a more general picture but the best guess is that the infilling of the basin was similar to that for the First Coarsening Upward Sequence.

As in the previous sequence the thickest turbidites in the western sections are abruptly overlain by siltstones which pass up within a few metres into shallow water sandstones. Once again the absence of a slope facies in this western area can be explained as the result of a gradual lessening of the slope during infilling of the basin so that after a certain stage turbidity currents were no longer generated (Fig. 36).

Unlike the first sequence there is no evidence of any shoreline deposits in this sequence, except possibly for the "hump" in the Manndraperelv section. The thickest beds in the "shelf" sandstones occur in the Leirpollen section which would be expected to contain the shallowest water deposits. Also there is little evidence that transgressive deposits form the top of the sequence although they may occur in the Manndraperelv and Leirpollen sections. Thus the whole of this Second Coarsening Upward Sequence except maybe for a few centimetres at the base and a metre or two at the top is a regressive deposit.

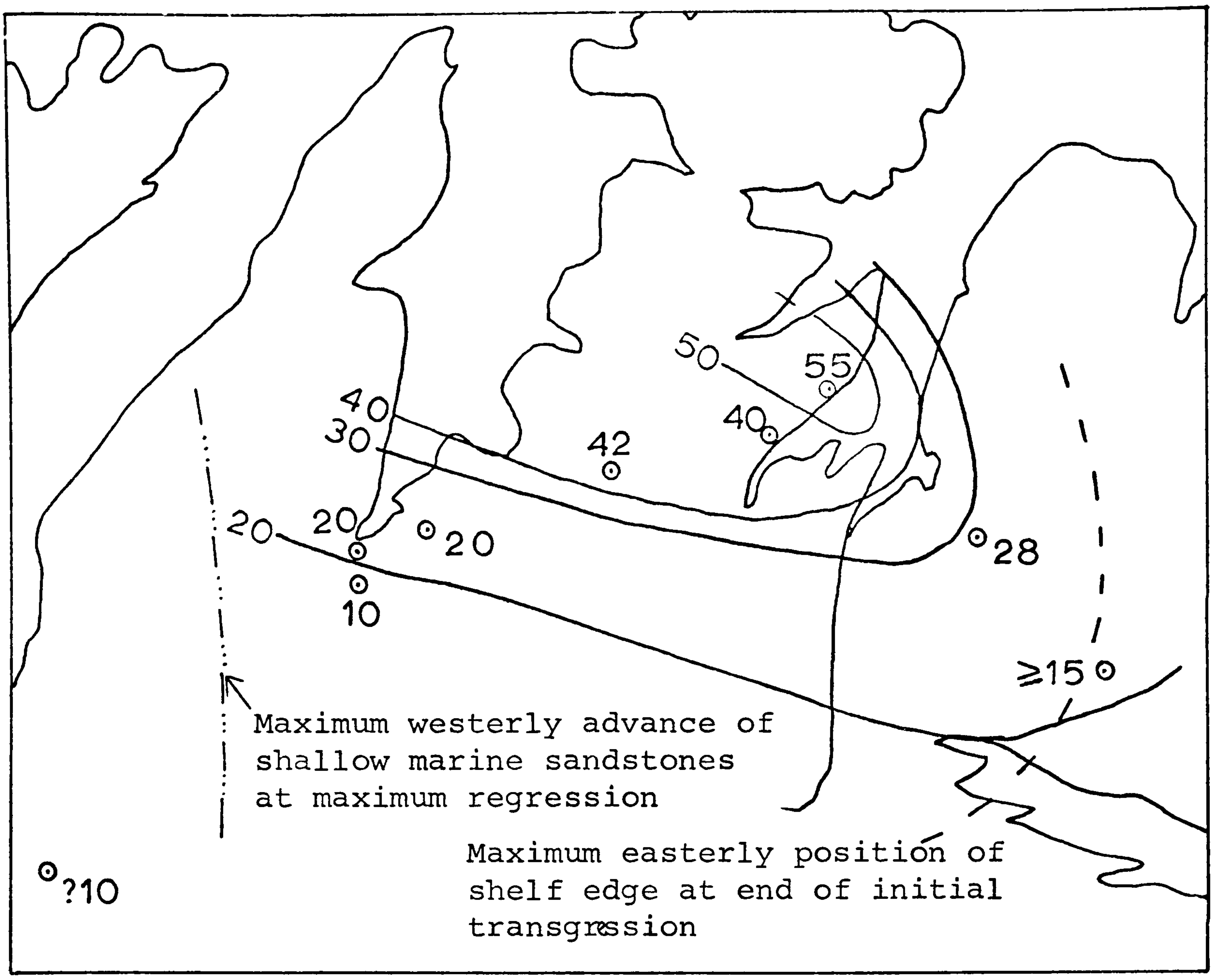


Fig.35. Speculative sketch to show the development of the Second Coarsening Upward Sequence, Manndraperelv Member. Numbers refer to thicknesses and the full lines are tentative isopachs. The most south-easterly locality is the Perledalen section (see Appendix C).

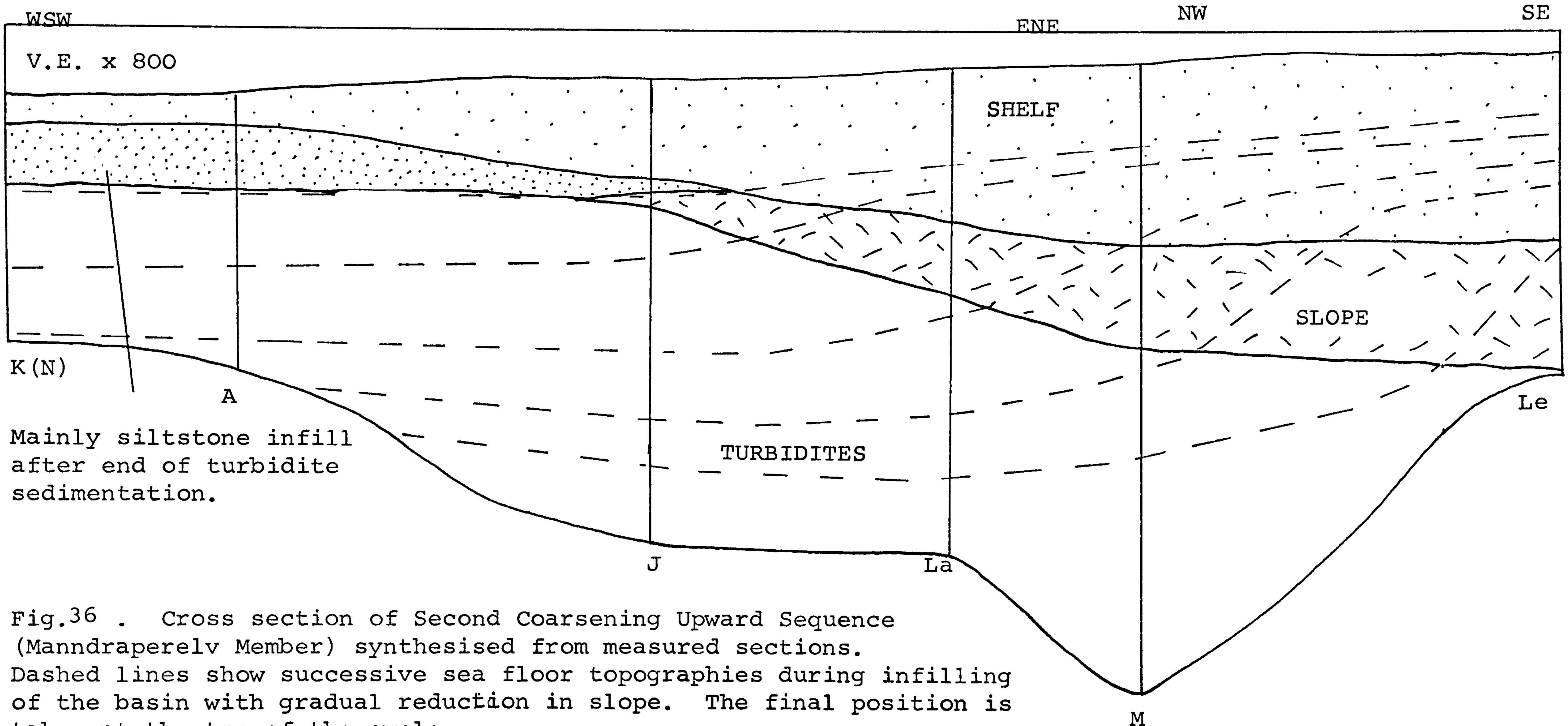


Fig.36 . Cross section of Second Coarsening Upward Sequence (Manndraperelv Member) synthesised from measured sections. Dashed lines show successive sea floor topographies during infilling of the basin with gradual reduction in slope. The final position is taken at the top of the cycle.

Summary of the Manndraperelv Member; discussion on the causes of cyclicity and the development of turbidites

The main conclusions are summarised in Figs. 37, 38.

A basic problem in all cyclic sequences is to determine the cause of the cyclicity. As has been stated earlier there are three major variable; absolute sea level, subsidence and sediment supply. These can be subdivided as follows:

Factors causing shallowing (regression)	Factors causing deepening (transgression)
A. Sediment supply	C. Subsidence
B. Eustatic fall in sea level	D. Eustatic rise in sea level

The fact that we have cyclicity shows that the ratio of the rates A + B to the rates of C + D was variable. Unfortunately these factors are not wholly independent in that an eustatic rise or fall would tend to produce a decrease or increase in sediment supply. Also sediment is inevitably attracted to the areas of greatest depression but once an area has been infilled sediment is switched to another. Thus there is a natural cycle of sediment following subsidence.

What evidence is there for these various factors in the Manndraperelv Member? The fact that the thickness of the Member varies so much over the area shows the importance of tectonic subsidence. Since this differential tectonic subsidence was a major feature any eustatic changes must have been relatively minor. Such changes are always hard to substantiate even when good fossil zonation is available and in this case there are no criteria by which they could be recognised or be shown to be absent. Therefore, ignoring for the present the

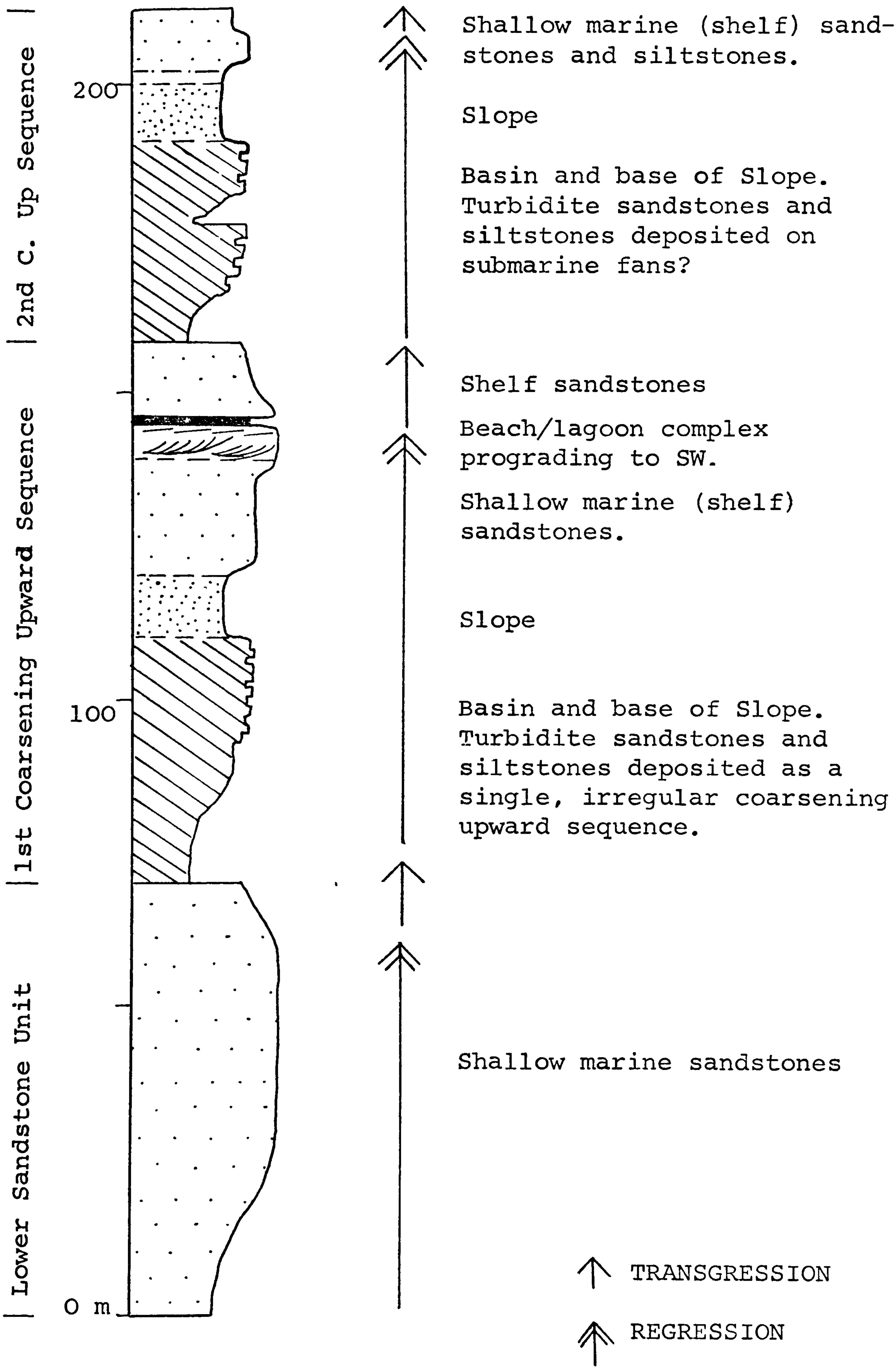


Fig.37. Summary of the environmental interpretation of the Manndraperelv Member as seen in the Manndraperelv section, Digermul Peninsula.

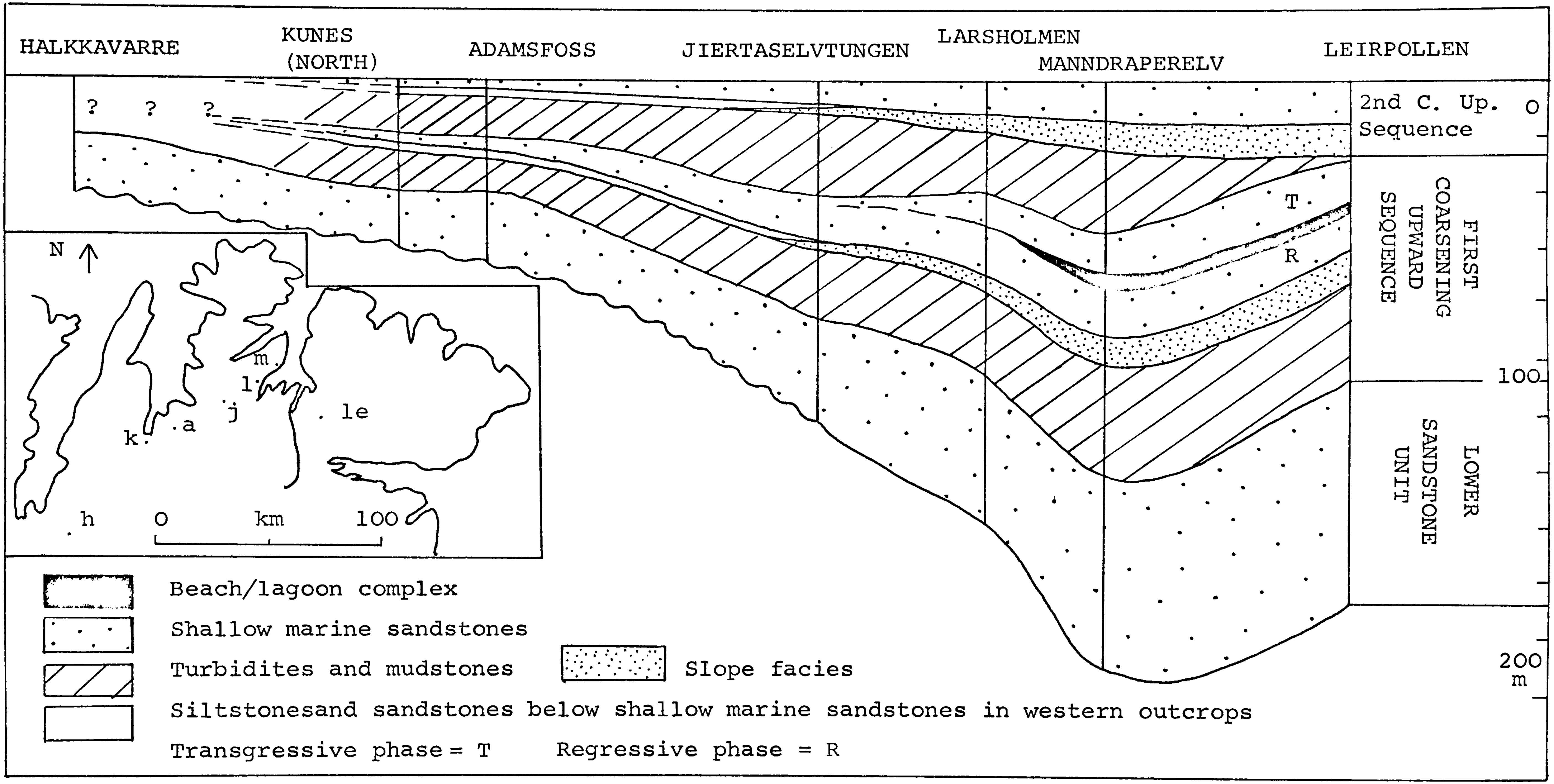


Fig.38 . Summary of the lateral variation within the Manndraperelv Member.and the laterally equivalent lower parts of the Breivik Fm in the western outcrops.

possibility of eustatic changes, three possible relationships between subsidence and sediment supply can be considered (Fig. 39) as causes of cyclicity.

(i) Subsidence was fairly steady but sediment supply was cyclic due to switching of the direction of transport between the area under discussion and other areas not now exposed.

(ii) Subsidence was unsteady whilst sediment supply was steady.

(iii) Both subsidence and sediment supply were unsteady but out of phase with each other, sediment supply being attracted after periods of high subsidence.

There are no criteria upon which to decide between these alternatives but (i) seems intuitively the most likely.

Turning now to the problem of turbidite deposition Fig. 38 emphasises the marked variation in subsidence over the area and incidentally that the centre of deposition moved westwards with time. It is believed that this variable subsidence was a primary cause of turbidite sedimentation in the lower parts of the two coarsening upward sequences as will now be explained.

The evidence suggests that before each coarsening upward sequence was deposited the sea floor subsided unevenly to give a basin with a maximum depth close to the maximum thickness of the sequence. When the rate of sediment supply became greater than the rate of subsidence the basin began to fill but it is envisaged that current and wave activity were only strong enough to move and deposit material in the shallower parts of the basin and little was deposited in deeper water. Thus with abundant sediment supply an equilibrium depositional and morphological profile (Moore and Curray 1964) became established in which the shelf break occurred approximately at the lower limit of wave and current action. Based on the thickness of the shelf sediments in the First Coarsening Upward Sequence this break occurred at a depth of about 20m. Counter-balancing the continued build up of the shelf by influx of sediment the equilibrium profile was maintained by the sweeping of sediment down into the deeper parts of the basin as turbidity currents. In this way the shelf and slope facies prograded across the basin in the manner sketched in Figs. 30, 36. The equilibrium profile need not necessarily have remained the same during each infilling since the changing physiography of the basin would have affected both current and wave strengths and indeed the profiles may have differed substantially from one sequence to another.

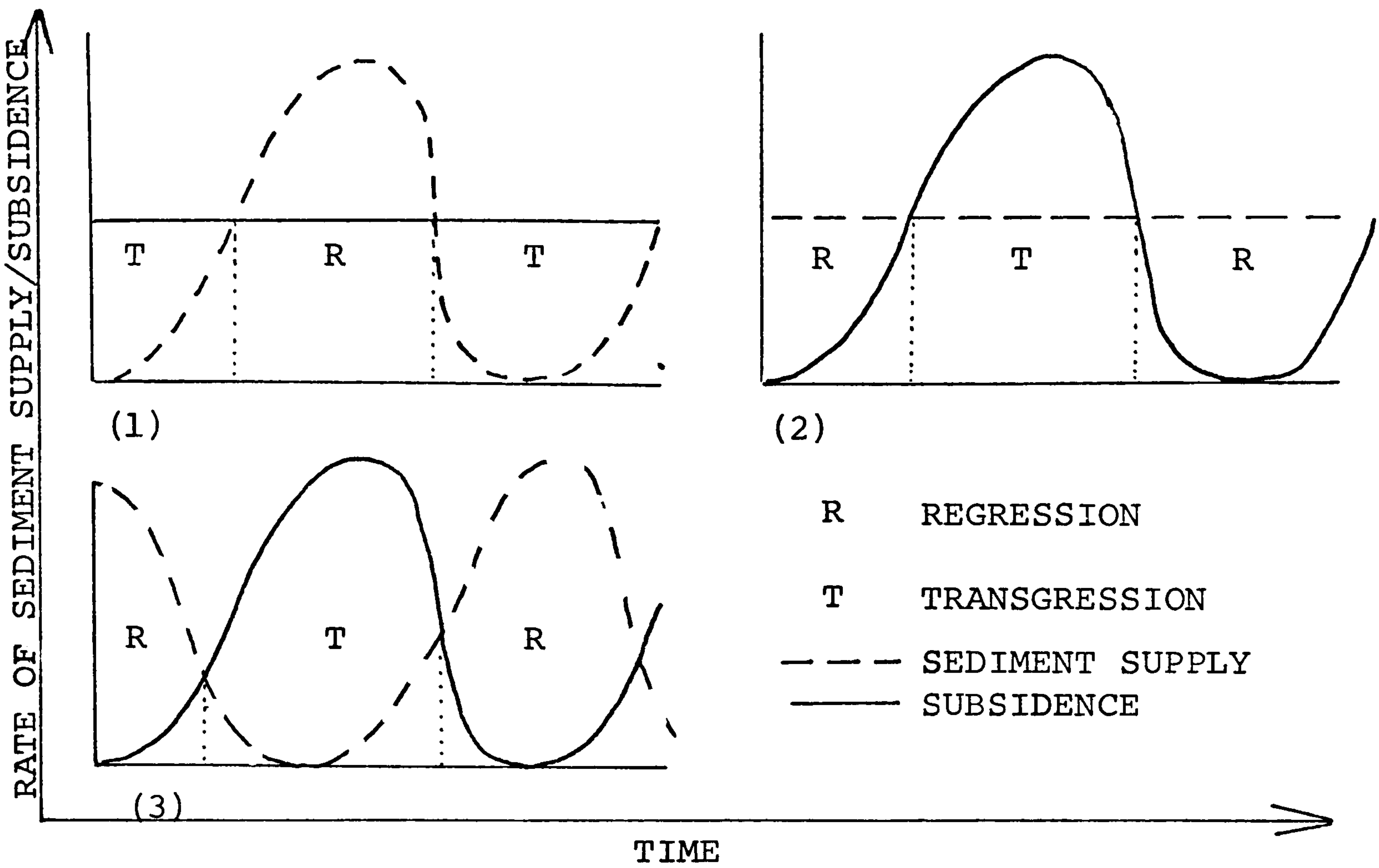


Fig.39. Simplified graphs to show various relationships of the rates of subsidence and sediment supply producing regression and transgression.

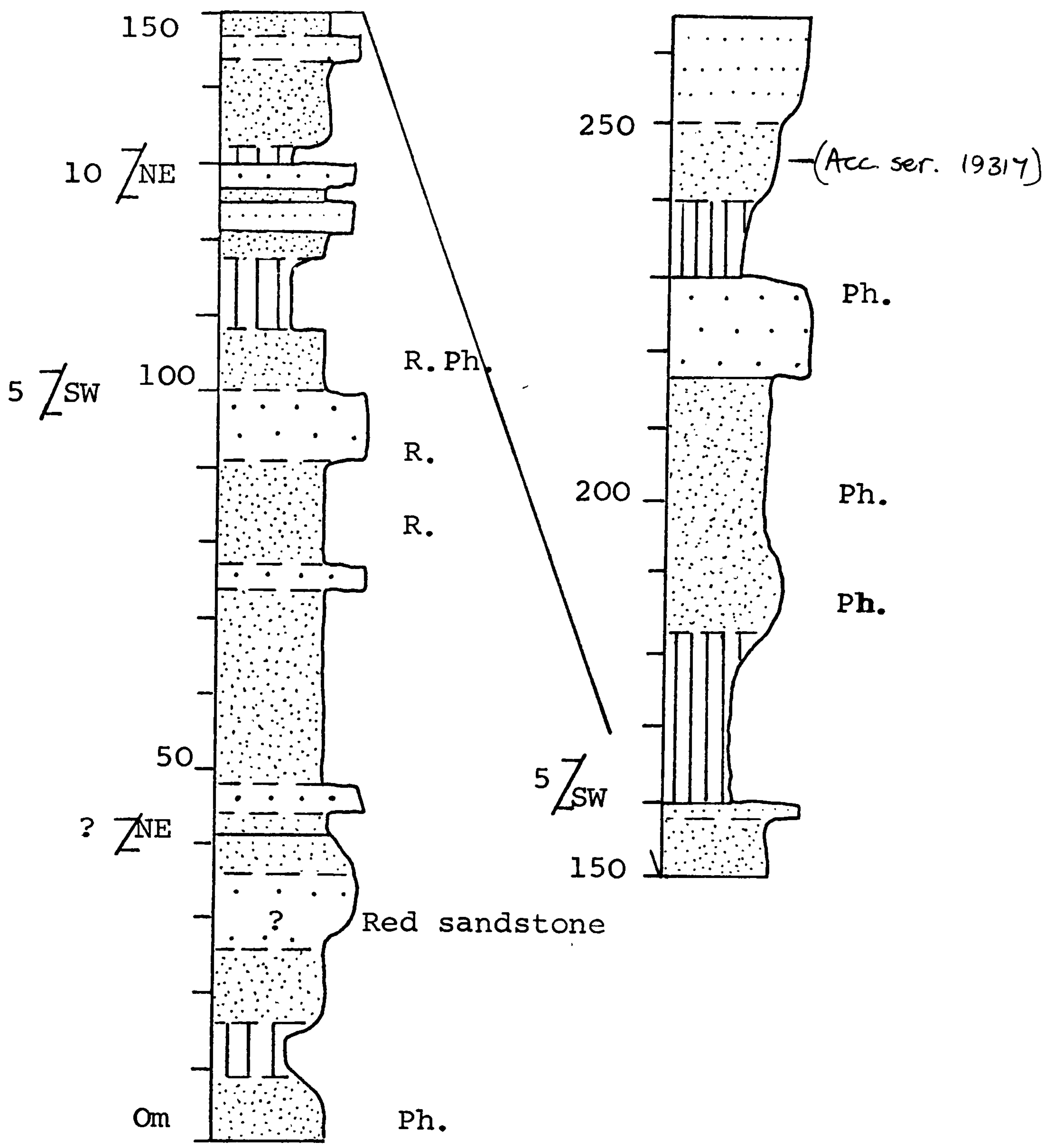
3, 5. LOWER BREIVIK MEMBER

Introduction

The Lower Breivik Member, which is at least partly of Lower Cambrian age, is the highest stratigraphical unit remaining on the Varanger Peninsula and in the Laksefjord area. In addition, the laterally equivalent Member IV of the Dividal Group is the youngest unit in the west of the area. Only on the Digermul Peninsula are younger rocks found. Not only is this the first horizon at which body fossils occur but it is also marked by an abundance and diversity of trace fossils unknown in the older rocks (Banks 1970). As in the previous section the coastal section on the Digermul Peninsula is described and interpreted first and then other areas are discussed.

Main Section: Description

This section (Fig. 40, see also Appendix C) was measured along the coastal outcrop south of Breivik. The thickness is at least 264m but may be more because there is a small fault near the base of the section across which no definite correlation could be made. The throw of other faults in the section are more easily estimated (e.g. Pl. 55). The base of the member was defined by Reading (1965) as the top of the last red sandstone of the Manndraperelv Member. This is not a good choice since another red sandstone horizon occurs a short distance above what Reading clearly mapped as the base of the Lower Breivik Member. The base is taken here at the place where the red and white sandstones of the Second Coarsening Upward Sequence pass up into interbedded sandstones and siltstones. This passage is often



Facies 4. 10-30 cm bedded medium-coarse cross-bedded sandstones

Facies 3. 10-100 cm bedded fine and very fine sandstones

Facies 2. 0.1-10 cm bedded sandstones with subordinate siltstones

Facies 1. Siltstone with 0.1-10 cm bedded sandstones

Fault: Downthrown side and amount of throw in metres shown.

Ph. <u>Phycodes pedum</u>
R. <u>Rusophycus</u>

Horizons where these trace fossils are present or common.

Figure 40 . Simplified section through the Lower Breivik Member as exposed on the coast south of Breivik.

gradational over a few metres but the beds form an easily recognisable horizon in all sections. The top of the member is easily defined (Reading 1965) by the abrupt transition from the white sandstones which form the uppermost 10-20m of the member to the green siltstones of the Upper Breivik Member.

The member consists of a complex alternation of sandstones and shales, the sandstones occurring as beds from 0.1-100cm thick, varying in grain size from silty and muddy very fine sand to coarse sand, and showing abundant sedimentary structures (Acc.Ser 19315). At some levels these sedimentary structures are strongly modified by bioturbation but in most cases they are well preserved. Phycodes pedum is the most common trace fossil (Pl. 56)(Acc.Ser 19314) and in association with Rusophycus suggests that deposition occurred within the Cruziana facies of Seilacher (1967). Symmetrical ripples are a common feature of bedding surfaces throughout the member. Reading (1965) considered that there were two main types of sandstone; (i) Cross-bedded and rippled orthoquartzites and lithic sandstones; (ii) Graded greywackes. His lithic sandstones were apparently identical to the orthoquartzites except for the presence of intraformational clasts of fine-grained sediment. However, no clear divisions of either mineralogy or sedimentary structures really exist within the sandstones. It is thought that the various sandstone beds form a gradational sequence in which the grain size was the major factor influencing the sedimentary structures and general appearance of the beds (Pl. 57).

In parts of the member (e.g. 108-123m, 131-147m, 147-158m, 230-264m) there are distinct sequences in which bed thickness and, to a limited extent, grain size, gradually increase upward. These sequences are followed by abrupt transitions back to finer sediment. However, in other places various lithologies alternate without apparent pattern. Since little significance can be seen in the vertical sequence of lithologies the section is not described in detail but four broad facies are described and hence a generalised model of the processes of sedimentation and environment of deposition is developed. The facies are all intergradational and the divisions on the section only attempt to show the most common type of bed at any given level.

In all the divisions of each facies the palaeocurrent directions are usually bipolar although there is often a widespread of directions and the number of measurements too few to give an accurate picture. Details are given in Appendix C but the totals of readings are shown in Fig. 41.

Facies 1: Siltstone with 0.1-10cm bedded sandstones

Description

This facies consists of massive green siltstone with intercalated laterally persistant laminae and beds of green and grey coarse siltstone, silty and muddy sandstone and very fine to fine sandstone. The beds are mostly less than 10cm thick but a few beds up to 60cm are also present. The majority of beds aresharp-based and some show flute and groove marks on their soles. Internally they are either massive apart from slight grading or show cross-lamination associated with asymmetrical ripples which have a chord length of 10-20cm (Pl. 58). The occasional thick beds are frequently

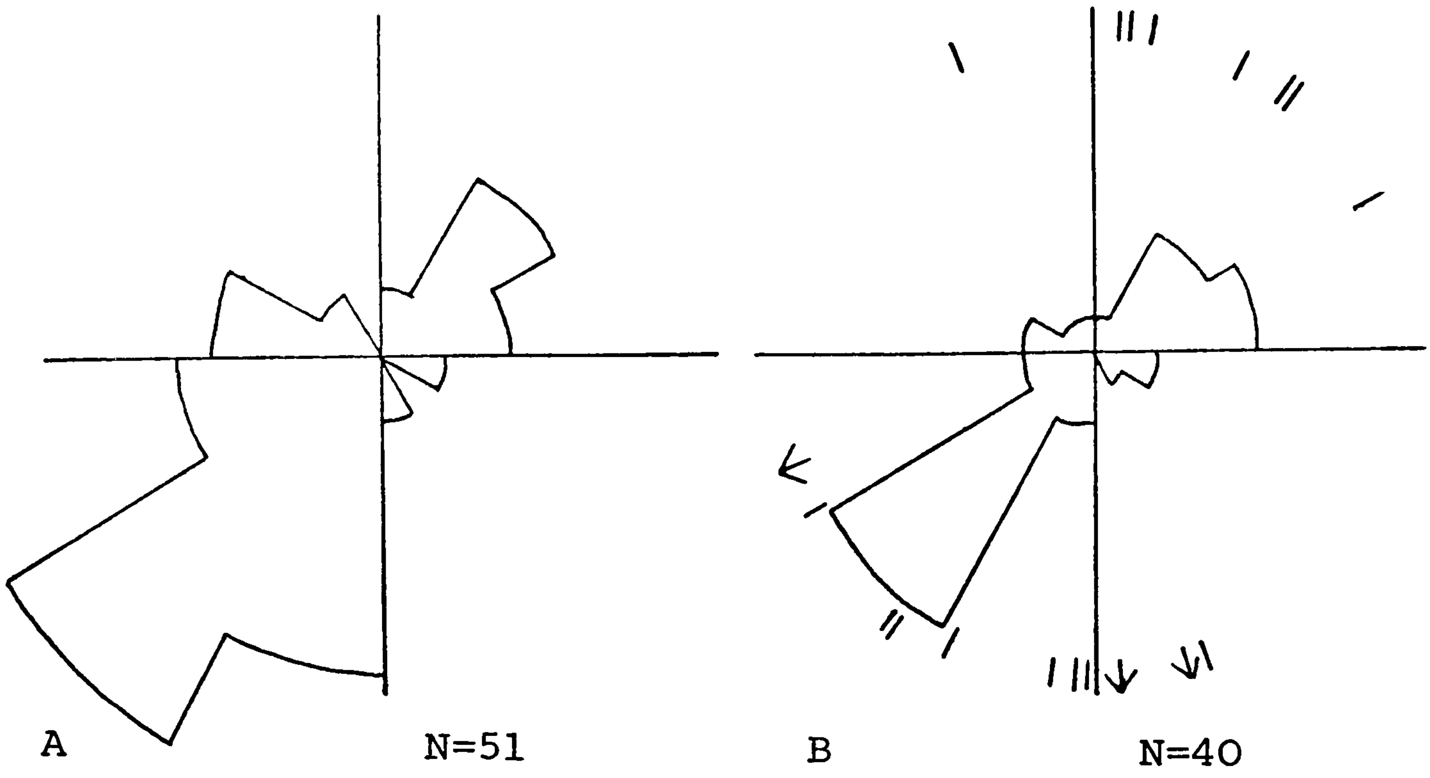


Figure 41 . Palaeocurrent data from the Lower Breivik Member, main section. A = Cross-bedding (mainly Facies 4) B= Cross-lamination (mainly Facies 1,2 and 3) Axial lines in B = Channels (mainly Facies 3) Arrows in B = Flute marks (Facies 1 and 2).

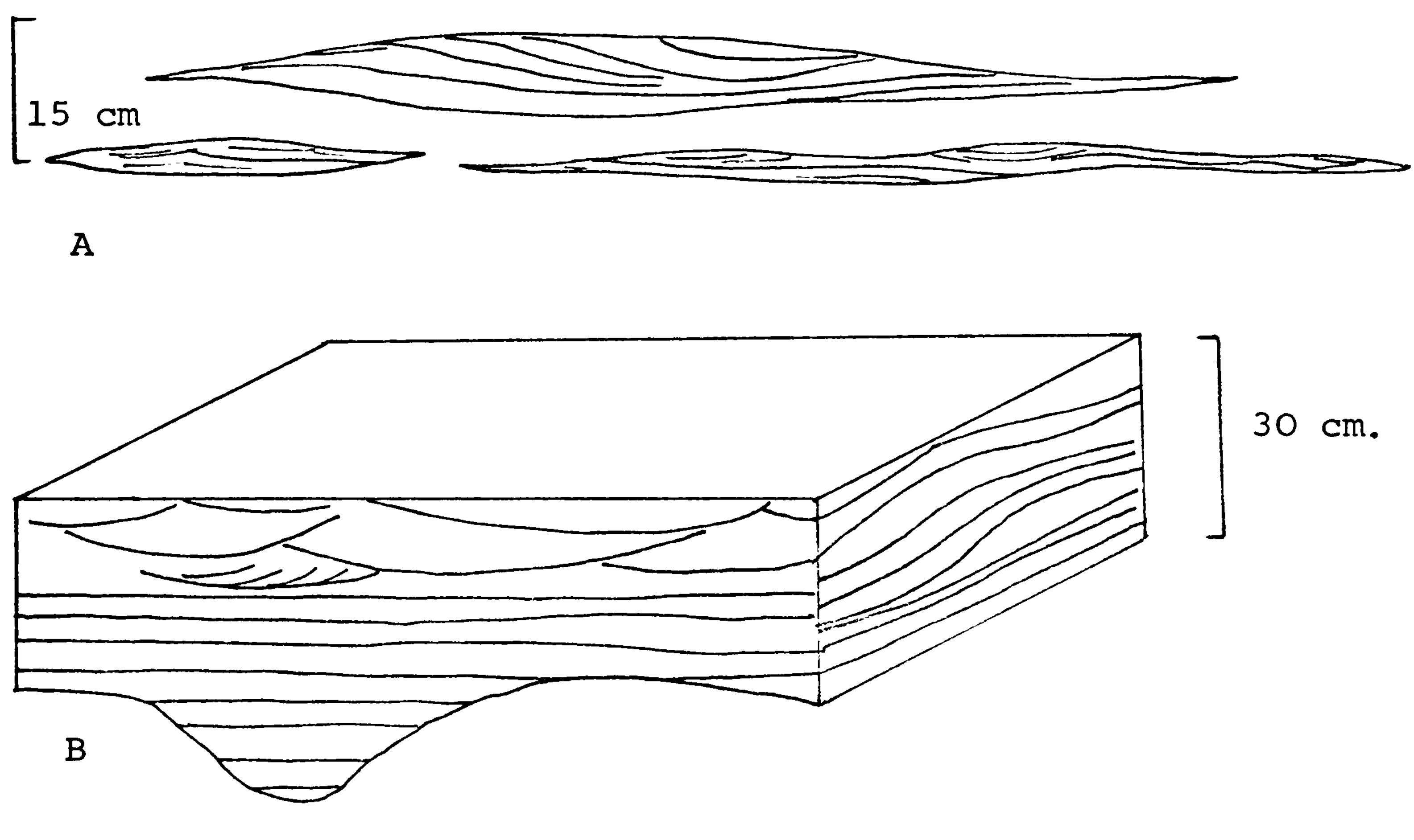


Figure 42 . Sedimentary structures in the Lower Breivik Member. A). Very lenticular sandstones of Facies 2 with irregular internal stratification.

B). Typical sandstone of Facies 3 with channeled base and cross-bedding with preservation of stoss side beds.

deformed into ball and pillow structures.

The "chloritic" matrix content of the sandstones varies from 0 - >10% but otherwise they are moderately to well sorted. This matrix is intimately associated with a cement of chert and fine-grained carbonate. Mineralogically the beds are slightly micaceous subarkoses and quartzarenites. The level of bioturbation is variable in this facies but generally rather low.

Interpretation

This facies was deposited in a quiet water environment in which siltstone was deposited from suspension. Periodically stronger currents brought in coarser sediment which was deposited as the currents waned as is shown by the grading of the sandstone beds. Whilst this facies has much in common with a distal turbidite facies the bipolar palaeocurrents and the general setting argue against such an origin.

Facies 2: 0.1cm bedded sandstone with subordinate siltstone

Description

Facies 1 grades into Facies 2 as sandstones become more common. The sandstones are texturally and mineralogically similar to those of Facies 1 except that they generally have less "chloritic" matrix and are mainly very fine, fine or even medium-grained. At some horizons the sandstones are laterally persistent (Pl. 69) and show sedimentary structures similar to those of Facies 1 with the addition that some of the thicker beds have a lower parallel laminated division with primary current lineation overlain by a cross-laminated division.

However, at other horizons, mainly in the upper half of the member, the sandstones are less regularly bedded. They commonly have strongly erosive bases, pinch and swell in thickness and can be laterally impersistent (Pl. 60). The internal structure of these beds is often a wavy cross-stratification from which it is usually impossible to determine any current direction (Fig. 42A). These beds often grade up into silt at their tops. Even very thin rippled beds are laterally impersistent in this facies at horizons where many sandstones are stacked together with little intervening siltstone (Pl. 61). The level of bioturbation is variable but burrowing is very marked at some levels.

Interpretation

This facies was deposited in a more agitated environment than was Facies 1. Currents capable of transporting sand were frequent although they were again of an episodic waning nature as is shown by the erosive bases and grading of the beds. At some levels these currents seem to have been all of similar strengths, thus giving sequences of beds of rather constant thickness (e.g. Pl. 61); however, at other levels the alternation of beds of very different thicknesses points to more variable current strengths. Lenticular beds probably formed when there was insufficient sediment to cover the whole sea floor and sand was distributed as isolated patches.

Facies 3: 10-100cm bedded fine and very fine sandstones

The beds of this facies (Pl. 62) are gradational into the lenticular very fine and fine sandstones of Facies 2. They occur either as single beds or as a series of amalgamated beds intercalated with beds of Facies 2. They are very erosively based often filling steep sided channels up to 30cm deep. Whilst the thinnest beds may be lenticular the thickest ones are laterally persistent. Internally parallel lamination predominates but they also have a wavy type of trough cross-bedding in which the stoss side bedding is commonly preserved (Fig. 42B). Many of the beds grade up into massive siltstone.

Texturally many beds are supermature in their lower parts but have appreciable amounts of matrix towards their tops. Like the sandstones of Facies 1 and 2 they are subarkoses and quartzarenites. Bioturbation is an uncommon feature of this facies.

Interpretation

These beds were obviously deposited by episodic currents of similar type but much stronger than those which deposited the sandstones of Facies 2. The fact that they are fine grained suggests that this was the maximum grain size available for transportation. The predominance of parallel lamination is probably due to the small range of flow power over which dunes form at this grain size (Guy and others 1966) particularly if the beds are considered as the result of waning flows (Allen 1970a) (Fig. 44).

Facies 4: 10-30cm bedded medium to coarse sandstones

Description

Facies 4 (Pl. 63, Fig. 43) is gradational into Facies 2 and 3 and is distinguished from them by the coarser grain size of its beds. It corresponds to Reading's (1965) "cross-bedded orthoquartzites and

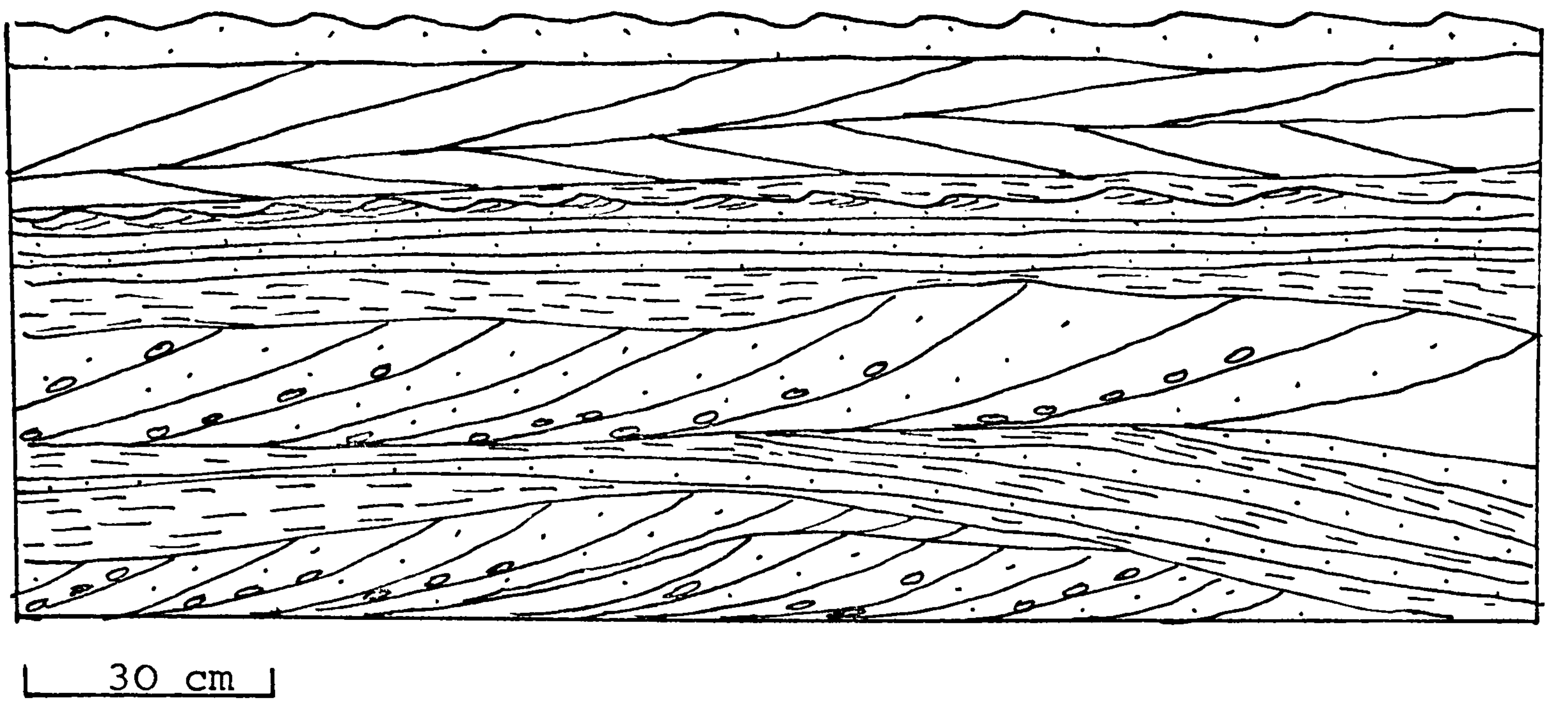


Figure 43. Drawing from a photograph of beds of Facies 4, Lower Breivik Member. Cross-bedded coarse sandstones are interbedded with thinner parallel laminated sandstones and mudstones.

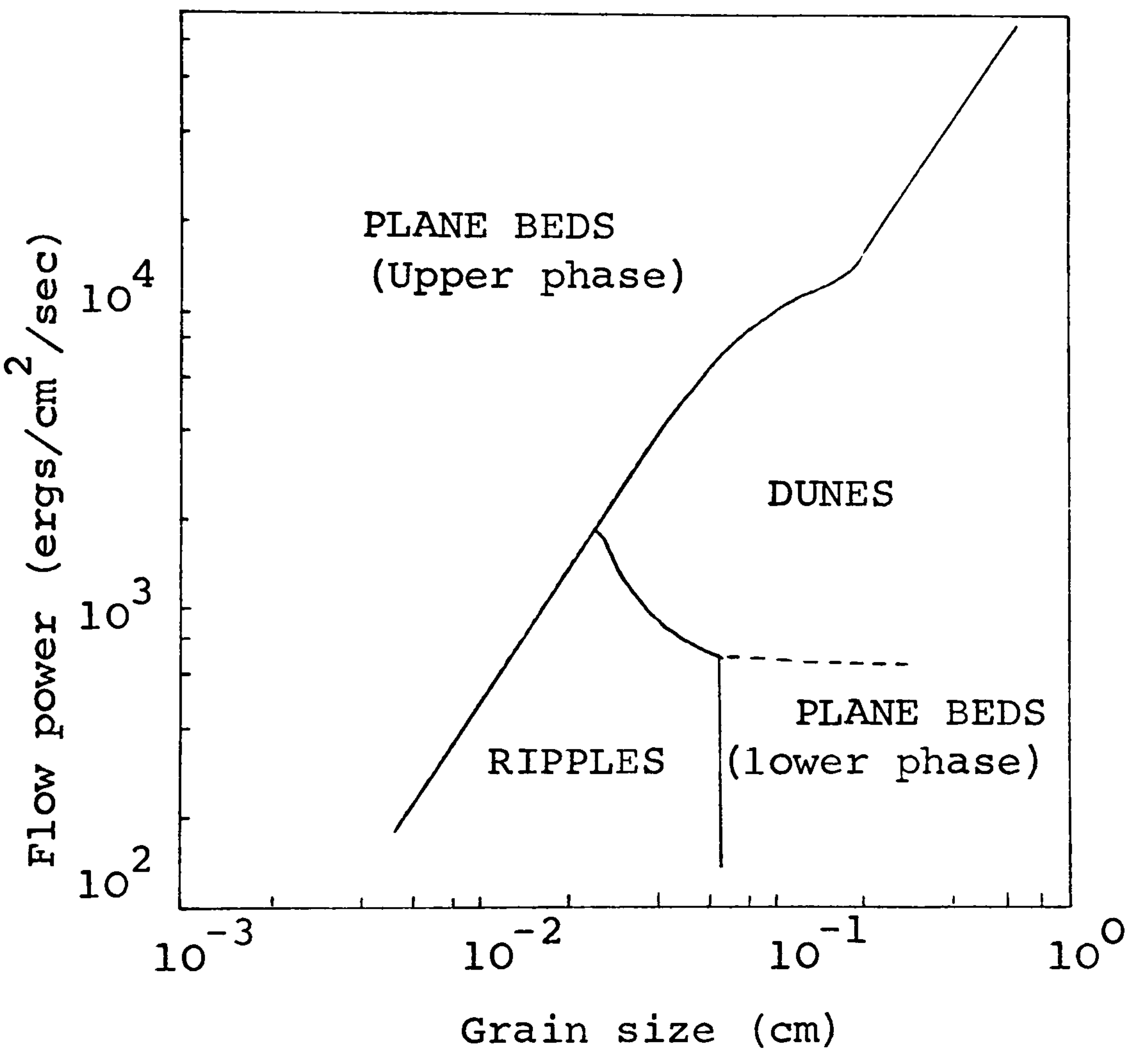


Figure 44. Bed form as a function of flow power and grain size in the case of a current whose power decreases with time at a fixed point. After Allen (1970a).

lithic sandstones". The beds characteristically weather white and are siliceous supermature subarkoses and quartzarenites except where the abundance of intraformational clasts makes them less well sorted. The intraformational clasts are of green siltstones and mudstone and many have been partly or wholly diagenetically altered to a dark brown "ironstone".

Trough cross-bedding is the dominant feature of these beds although parallel lamination with primary current lineation is common in the thinner and finer beds where it is often overlain by cross-lamination. The dips of the cross-sets range from 15-26° and are highest in the coarsest beds. As in Facies 2 and 3 the beds are commonly erosively based and lenticular although the depth of erosion is generally less. A remarkable feature of this facies is that the sediments interbedded with these sandstones are mudstones and siltstones which are the finest sediments within the member. These fine-grained beds, which are up to 30cm thick, are massive and show no signs of bioturbation (Pl. 64). Indeed bioturbation is a rare feature of this facies except occasionally at the tops of units.

Interpretation

Since there are all gradations between beds of this facies and those of Facies 2 and 3 it is inferred that they were deposited by similar currents. The abundance of cross-bedding is due to the far greater range of flow power over which this structure forms in medium to coarse sand compared with finer sand (Guy and others 1966; Allen 1970a). The lenticularity of the beds can again be related to a shortage of sediment supply.

Because the interbedded fine material was not bioturbated it is suggested that it must have been deposited rapidly implying that there were calm periods when the water had a high concentration of suspended material. Such conditions could have arisen after a strong current had passed over a muddy area and taken a large amount of sediment into suspension.

Main Section: Discussion

The following facts and deductions are pertinent to a discussion of the type of current which deposited the sandstone beds.

(i) The currents were episodic, deposition occurring as they waned.

(ii) They had a bipolar flow pattern.

(iii) They were strong enough to produce flat bedding with primary current lineation in medium sand and of causing considerable erosion of the sea floor.

(iv) Textural maturity of the sediments is variable being lowest in the finest sandstones.

(v) The sediments are moderately mineralogically mature.

Table 3 is based largely on the criteria discussed in Chapter 2. It shows that the currents most likely to have produced the sandstones of the Lower Breivik Member are tidal ones. The main criterion for this conclusion is the bimodal direction of transport throughout the member. Each bed in Facies 1, 2 and 3 is thought to have formed from a single exceptional tidal event in the way envisaged by Stride (1965) although in Facies 4 it is more difficult to distinguish individual events. The presence of symmetrical ripples suggests that deposition occurred above wave base but there is no

P = Possible

PR = Probable

I = Improbable

	Epidodic waning flows	Bipolar current directions	Strong enough to give observed structures	Textural maturity variable	Mineralogical maturity moderate
Semi-permanent	I	P	P	P	PR
Tidal	PR	PR	PR	P	PR
Wave Drift	PR	I	I	PR	PR
Coastal storm surge	PR	I	PR	PR	PR
River generated	PR	I	PR	P	I

Table 3. Table showing the abilities of various types of current to produce the observed structures and other features of the sandstones of the Lower Breivik Member.

variation in the form of frequency of the ripples to suggest that there were any marked changes in water depth during deposition of the member. There is no evidence of any sequences of marginal sediments such as form on tidal flats (Reineck 1967) or beaches (Lane 1963).

Therefore it is concluded that the member was deposited in a shelf environment, of perhaps some tens of metres depth, and that there were probably no great variations in depth throughout the time of deposition. The persistent palaeocurrent pattern suggests that the basin was partly enclosed but the strength of the tidal currents necessitates some opening to a large sea or ocean. The wide spread of current directions within the bipolar pattern shows that the tidal currents had elliptical rather than rectilinear paths.

Large scale tidal bed forms such as sand waves or tidal current ridges are absent. However since many beds are lenticular due to a shortage of sediment supply this patchy distribution of sand might be manifested on a larger scale by the presence of sand ribbons (Kenyon 1970) which characterise areas of little sediment but strong currents in present seas. This is particularly possible for beds of Facies 4 but the exposure is insufficient to prove a sand ribbon form.

Although there are considerable periodic variations in the strengths of tidal currents due to the changing positions of the earth, moon and sun, these are insufficient to account for the great differences in grain size and bed thickness which occur. Johnson and Stride (1969) amongst others, have pointed out the greatly increased sediment transport that occurs when tidal currents are aided by storms.

This storm plus tidal origin is very likely for many of the thicker beds in the succession but does not readily explain how the thicker beds tend to be grouped in packets forming units of Facies 3 and 4. This packaging of beds suggests a number of longer term changes in current strengths. Johnson and Belderson (1969) have discussed several ways in which changes in physiography of a basin can cause either abrupt or gradual fluctuations in tidal current strengths. These fluctuations need not be related in any consistent way to changes in water depth or distance from the shoreline. One example is when the changes in basin physiography cause the natural oscillation of the basin to come close to an important tidal period. The resulting resonant amplification will greatly increase tidal velocities over the whole sea. Although such a resonant amplification may be a comparatively rare phenomenon it is nevertheless important since its effects would be great. Most work would be done during a relatively short period before the changes in bed roughness and basin dimensions resulted in a reduction of resonance and hence current velocities. Resonance thus tends towards self-destruction.

The idea of resonance can possibly be applied to the beds of Facies 4 where there were strong currents giving the medium to coarse sandstones and rapid deposition as evidenced by the scarcity of bioturbation. However, although resonance might cause rapid deposition in some places, it could cause erosion in others and so even if units of Facies 4 resulted from resonance they need not be laterally persistent over a wide area.

In conclusion, whereas some depth changes are not ruled out, the Johnson and Belderson model of tidal velocities fluctuating independently of depth and proximity to a shoreline best explains the irregular alternation of lithologies in this member. In addition storm action was very probably important in enhancing tidal currents and greatly increasing their capacity to transport sediment. Although this tidal model best fits the available information the origin of a small minority of beds by processes such as coastal storm surge currents or even semipermanent currents cannot be ruled out. For any individual bed a number of origins are possible but when all the information from all the beds is grouped together a tidal origin is the most probable.

Lateral Variation

The pattern of lateral thickness variation shown in Fig. 45 is similar to that in the underlying members but it must be remembered that the sections are incomplete except on the Digermul Peninsula. The horizon of *Platysolenites antiquissimus* has been taken as a datum line but it is not really known how closely this line corresponds to a time line. According to Hamar (1967) *Platysolenites* is characteristic of zone 1aβ of the Norwegian Lower Cambrian zonal scheme. Although the three easterly sections are very similar in their lithologies and thicknesses no detailed correlation was established (Fig. 46).

Leirpollen

The Member is best exposed in the Vaderelv valley but folding and faulting make measurement difficult in its lower part (Pl. 65). Although the lowest 130m consist largely of Facies 2 there are also a number of grey, very fine and fine sandstone beds up to 50cm that

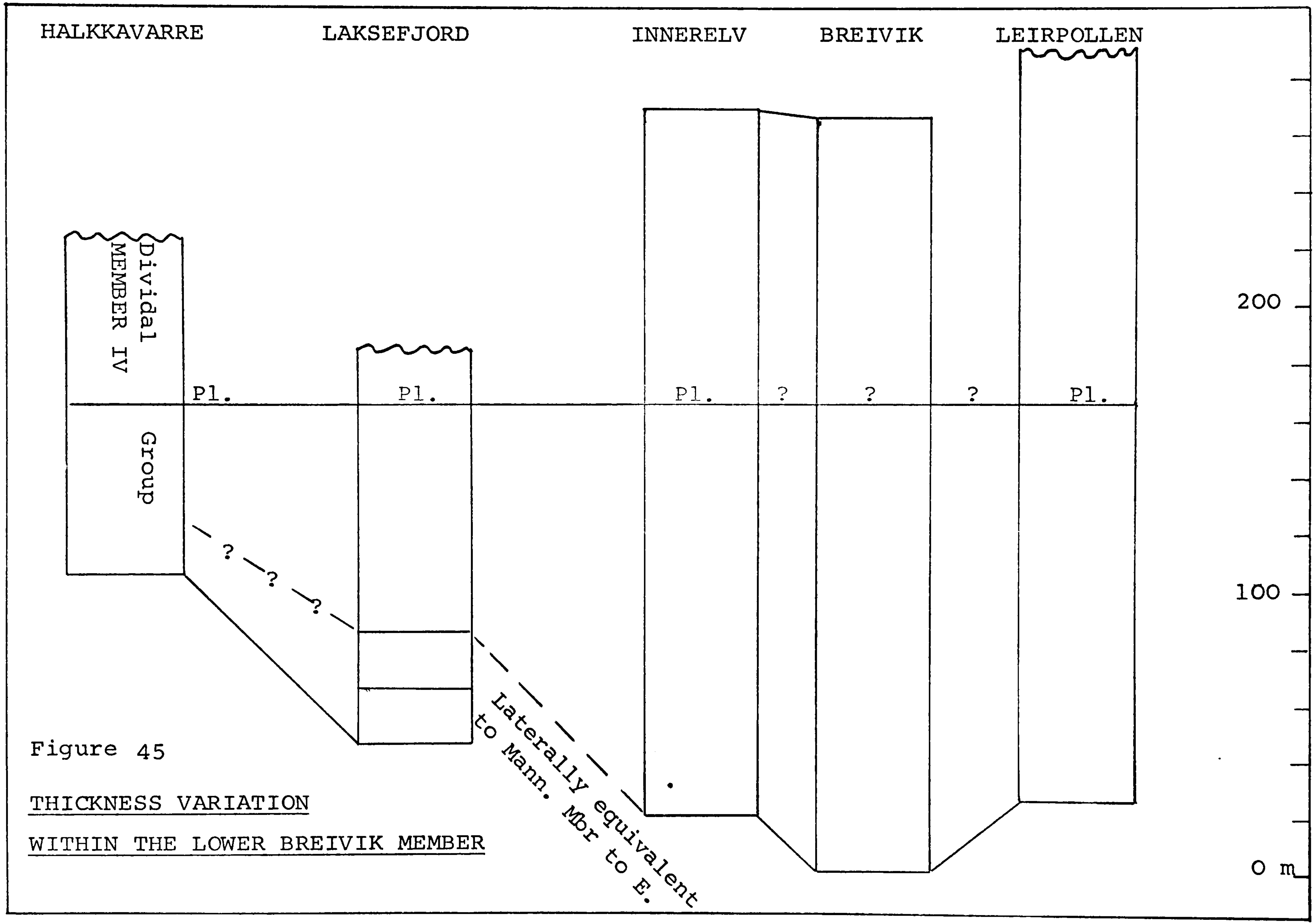


Figure 45

THICKNESS VARIATION

WITHIN THE LOWER BREIVIK MEMBER

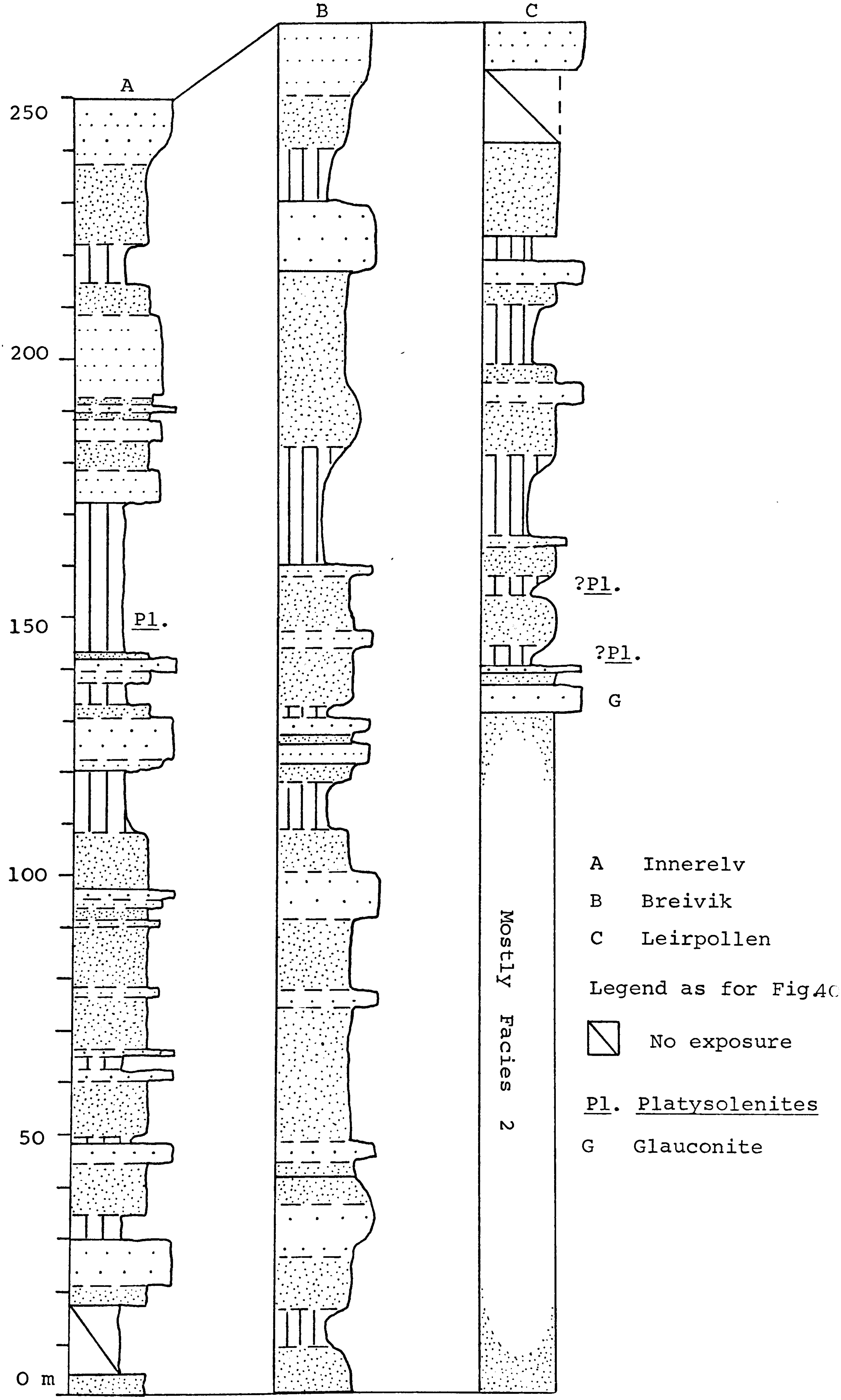


Figure 46. Lateral variation in eastern sections of the Lower Breivik Member

are frequently deformed into ball and pillow structures. A distinctive unit of Facies 4 occurs at 130-135m and the sandstones contain 1-5% of a green mineral which is probably glauconite (Appendix D). A short distance above this unit came two units of Facies 1. From Føyn's (1967) map it is believed that his specimens of Platysolenites come from mudstones within one of these units but no further specimens were found to confirm this. The remainder of the succession is similar to the Breivik section in consisting of a number of poorly defined coarsening upward sequences. It is probable that the white sandstones of Annecaerro, which are the highest beds of the succession, can be correlated with those at the top of the Breivik section as suggested by Reading (1965). Apart from the glauconitic horizon the sandstones are petrographically similar to those of the Breivik section being subarkoses and quartzarenites.

Innerelv

This section is about 12km SW of the Breivik section but again there is no detailed correlation between the two. Although there appears to be some similarity between the positions of units of Facies 1 in each section it would need detailed measurements in the relatively poorly exposed intervening area to prove a correlation. An interesting feature is that the palaeocurrent directions are again bipolar in the Innerelv section but the orientation is NW-SE rather than SW-NE as it is at Breivik (Fig. 47).

Ifjord-Vestertana area

The member can be traced southwestwards from the Innerelv section as far as the Ifjord-Vestertana road. The top and bottom are well exposed in Engdalen where

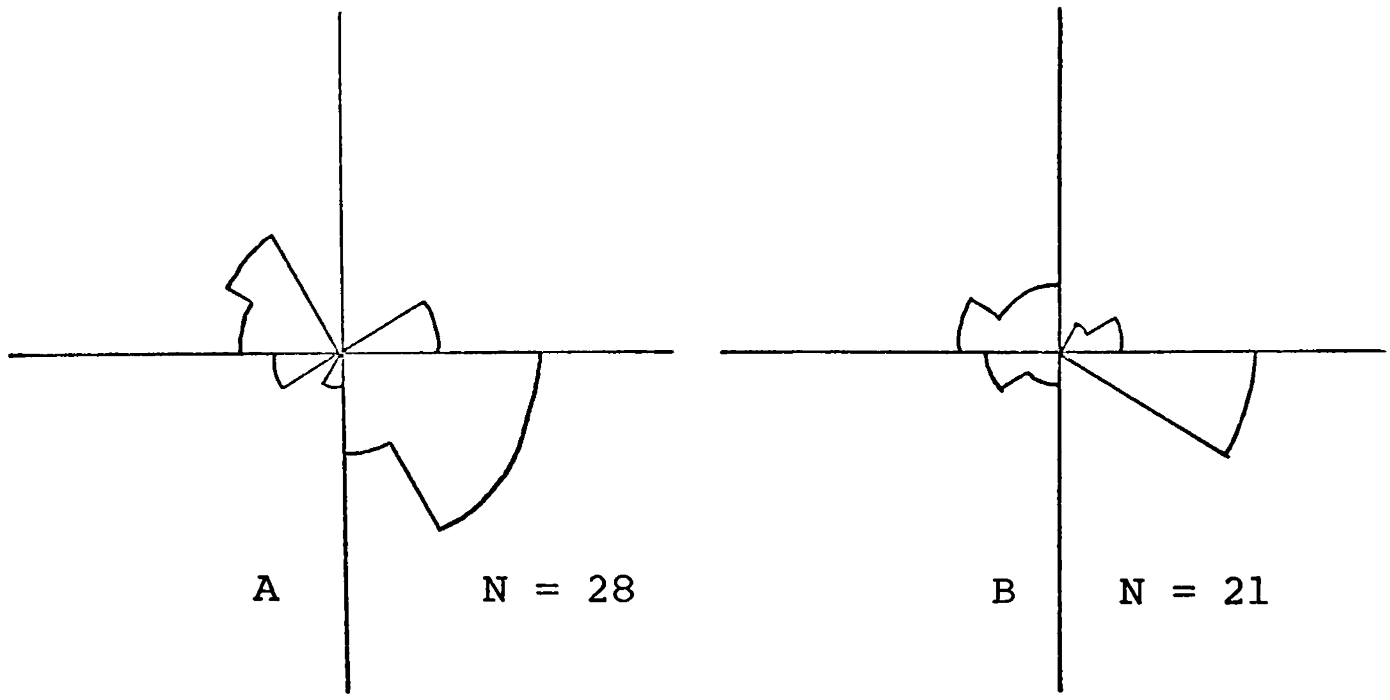


Figure 47 . Palaeocurrent data from the Innerelv section, Lower Breivik Member. A = cross-bedding, B = cross-lamination.

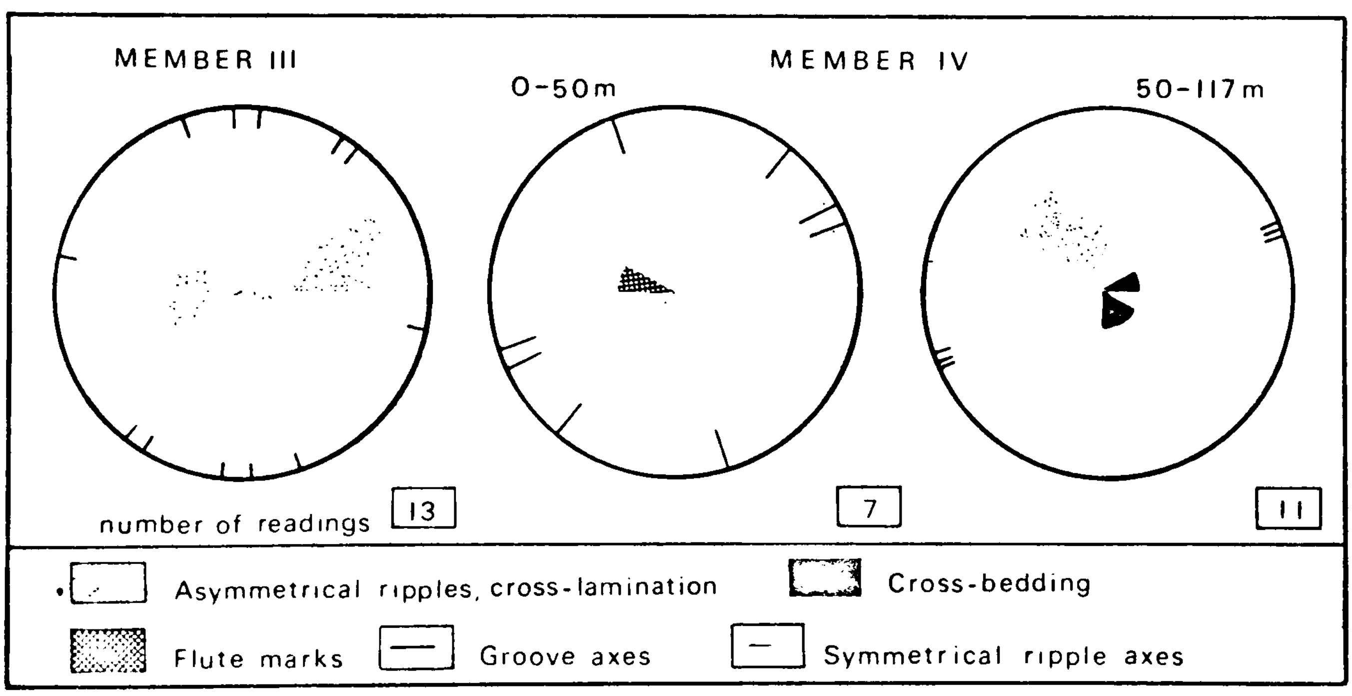


Figure 48. Palaeocurrent data from Members III and IV of the Dividal Group, Halkkavarre.

the member is broadly similar to the Innerelv section although detailed measurements were not made.

Laksefjord (Adamsfoss and Kunes)

The lowest 40m of the member has already been described as the lateral equivalent of the two coarsening upward sequences of the Manndraperelv Member to the east. Above these beds the member is poorly exposed and strongly tectonised and can only be described in general terms. Details are given in Appendix C.

At Adamsfoss a further coarsening upward sequence 30m is present and this is followed by about 30m of siltstones and sandstones some of the latter resembling Facies 4. In the northern part of the Kunes area the thickness is estimated to be about 150m although it could be more. Above the two coarsening upward sequences the member consists of at least 40m of dark red and green thin-bedded sandstones and siltstones. Above these beds there are a number of red sandstone units (up to 15m) intercalated with units of green siltstones and mudstones (Up to 50m?), the sandstones being similar to those of the Manndraperelv Member. The Platysolenites horizon (Føyn 1967) occurs at top of this facies and above it come 20m of thin to medium-bedded sandstones and siltstones before the member is truncated by overthrust metamorphic rocks.

In the southern part of the Kunes area four units of white sandstone, each 1m thick, occur within an otherwise thin-bedded sandstone and siltstone facies which is 20m thick. Above this come predominantly dark red sandstones and siltstones.

Halkkavarre

Member IV of the Dividal Group was thought by Føyn (1967) to be 130m thick and to be overlain by further beds of the Dividal Group (Members V and VI). However the accurate location of the thrust plane between the Dividal Group and the overlying Gaissa Nappe has shown that Føyn's Member V and VI and the top 10m of Member IV actually belong to the Nappe (Pl. 66). The true thickness of Member IV is thus 12om.

The exact junction between Members III and IV is obscured by scree but it appears to be sharp. A detailed section through Member IV is given in Fig. 8. The sandstones are slightly micaceous subarkoses in the lower part and quartzarenites and subarkoses in the upper and the cross-bedded sandstones of the upper part strongly resemble Facies 4 of the Breivik section. Palaeocurrent data are given in Fig. 48. The abrupt facies change at 175m and an abundance of (?) glauconite in the immediately overlying beds suggests the possibility of a disconformity at this level.

Discussion

The comparison of the Innerelv and Leirpollen sections with the Breivik section suggests that sedimentation was rarely uniform over the area since no definite correlation of any facies units except the top sandstone is apparent. This is explicable in terms of the varied sediment distribution patterns that are common in tidal seas. This lack of correlation may be also partly due to a shortage of sand supply in the basin which led to the formation of patchy deposits even where conditions were uniform.

The marked difference in palaeocurrents between the Innerelv and Breivik sections is difficult to explain. The consistency of the directions in both sections suggests some local control on the tidal pattern such as either a nearby shoreline or perhaps more likely, a "swell" area. Certainly such a persistent difference would not be expected in a uniform open sea.

The relationship of these eastern sections to those in the west is difficult to see. There is little evidence as to the origin of the beds in the Laksefjord area except that they were probably deposited in a "shelf" environment. At Halkkavarre the upper part of Member IV shows similarities to the eastern sections but, remembering the possibility of an disconformity at 175m in that section, these beds may be younger than those in the east.

In conclusion, in the eastern area the Lower Breivik Member was deposited in an offshore "shelf" environment with active tidal currents; this was probably in a partly enclosed basin but its dimensions are unknown. Sand was often in short supply which resulted on a small scale in many lenticular beds, and on a larger scale it contributed to the lateral impersistence of the sandstone facies. The sea extended at least as far as Halkkavarre in the SW where shelf sedimentation also prevailed.

CHAPTER 4

THE UPPER BREIVIK MEMBER AND DUOLBASGAISSA FORMATION

4, 1. INTRODUCTION

The Upper Breivik Member and the two members of the Duolbasgaissa Formation form the remainder of the Lower Cambrian succession. The only recorded fossil from these rocks is Holmia sp. from the Upper Duolbasgaissa Member but a rich Paradoxides fauna occurs at the base of the overlying Kistedal Formation (Reading 1965) and thus the junction between the Duolbasgaissa and Kistedal Formations approximates to the Lower Cambrian-Middle Cambrian boundary.

The boundaries between the three members are more gradational than those between the lower members. There is considerably lithological continuity from one member to another, the sequence gradually but irregularly coarsening upwards (Fig. 49). Sedimentologically the members form a distinct entity and thus it is very convenient to discuss the three members together.

The members are confined to the Digermul Peninsula where they form part of a broadly synclinal structure. Only the outcrops on the eastern side of the peninsula were studied because those on the west are very strongly tectonised and not easily accessible.

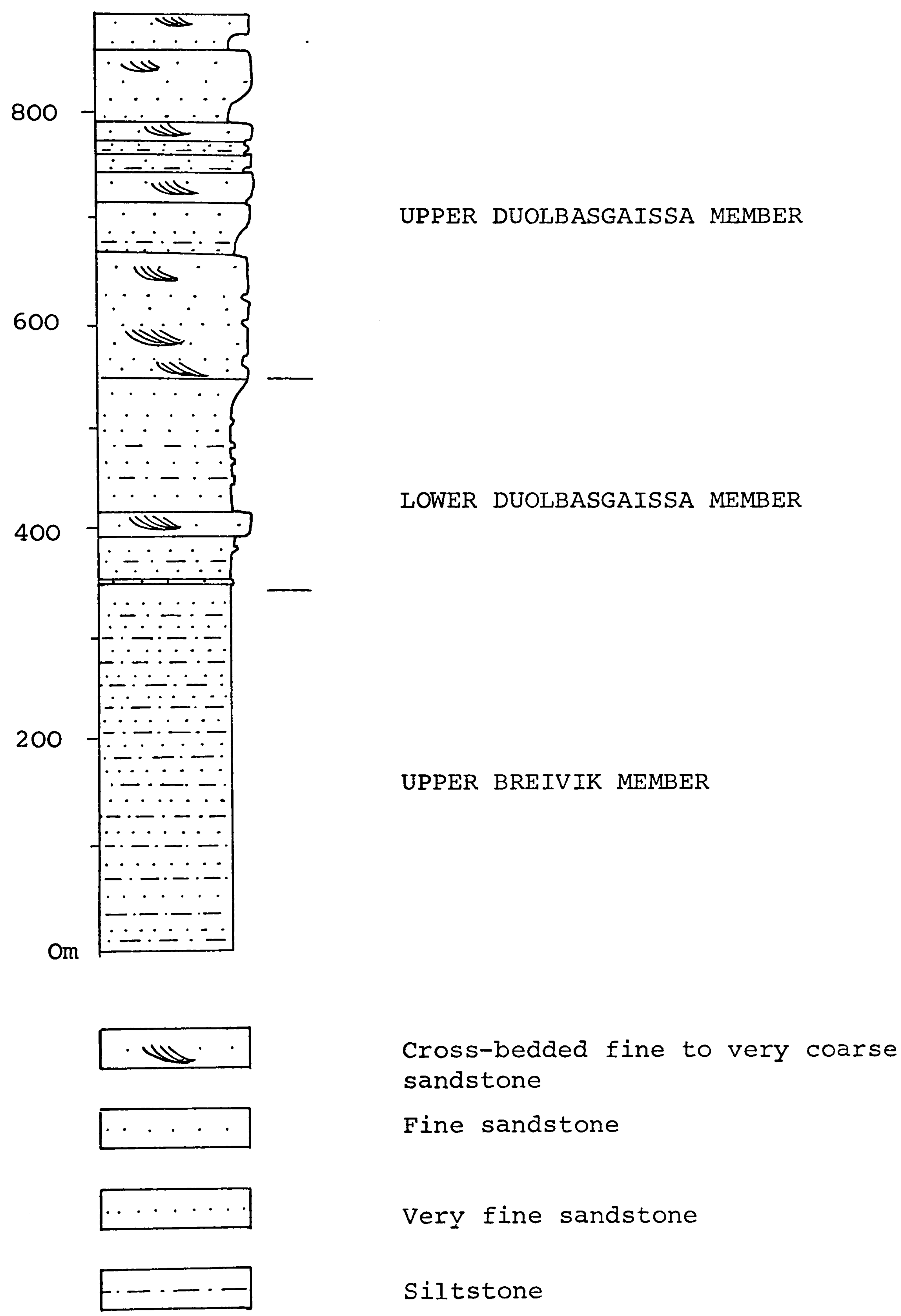


Fig.49. Simplified section through the Upper Breivik Member and Duolbasgaissa Formation as seen near Breivik, Digermul Peninsula.

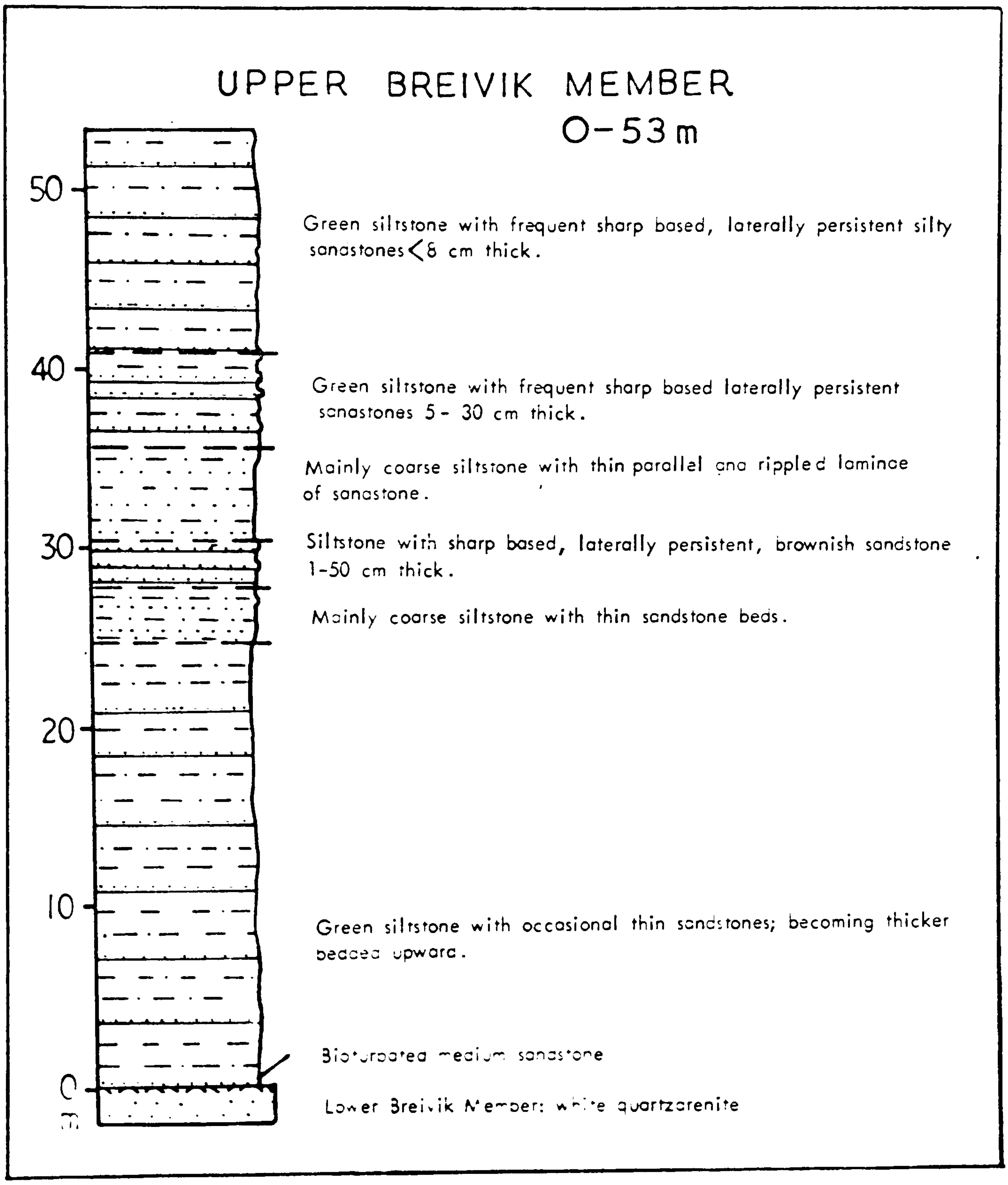


Figure 50. Section through the lowest 53m of the Upper Breivik Member as seen on the coast south of Breivik.

4, 2. UPPER BREIVIK MEMBER

Introduction

The Upper Breivik Member consists largely of green siltstones alternating with laminae and thin beds of sandstone. Its thickness is estimated to be 350m but accurate measurement is hindered by extensive faulting combindd with the homogeneous nature of the member. It is best exposed along the coast on both sides of Breivik and this is the section described below and shown in Figs. 49, 50.

Section at Breivik

The junction with the Lower Breivik Member is extsemely sharp (Pl. 67). Reddish grey, poorly sorted, medium sandstone is piped down in irregular burrows at least 15cm into the white quartzarenite which forms the top of the lower member. The contact is planar and non-erosive along the 30m over which it is exposed. Reading (1965) traced this sharp contact over the entire peninsula and it is also present in the southernmost outcrops in Engdalen (Appendix B).

The basal sandstone horizon, which is markedly bioturbated, is only a few centimetres thick and is succeeded by massive green siltstones with occasional sharp based, laterally continuous sandstone beds up to 15cm thick. The lowest 53m of the section are summarised in Fig. 50. The sharp-based sandstones are mostly of fine or very fine sand grade but may show marked grading in their lowest parts with grains up to very coarse sand size. The thicker sandstones often have erosive bases and some are amalgamated giving "beds" up to 150cm thick. A variety of undiagnostic burrows occur on the bases of the beds, the majority being predepositional in origin. Within the beds parallel lamination is common and Bouma sequences are sometimes present. Fig. 51 shows some structures seen within one such sandstone bed. The sandstones are immature to mature subarkoses to quartzarenites. Up to 10% of disseminated chlorite is present as matrix.

Above the lowest 50m the member gradually becomes very uniform in lithology consisting of green or occasionally reddish-grey siltstones alternating with laterally persistent green-grey sandstone layers which show all gradations from thin laminae to beds 20cm thick (Pl. 68 (Acc. ser. 19319)). This facies continues to the top of the member. The great majority of the sandstones have sharp bases and evidence of erosion is seen in the irregular and sometimes fluted and grooved nature of these bases. Loading of flute marks and ripples is a common feature.

Internally many sandstone layers exhibit slight grading. Sandstone units less than 1cm thick show a fine parallel lamination and occasional low angle cross-lamination associated with low amplitude ripple marks. Very thin beds predominantly show small scale cross-lamination associated with asymmetrical ripple marks and a few horizons consist of rows of isolated ripples. Thin and medium-bedded sandstones are far less common than thinner layers; they tend to be more lenticular and show vague wavy or parallel lamination. Some climbing ripples are present but the angles of climb are extremely small (Fig. 52) corresponding to the type A of Allen (1970b). Petrographically these sandstones are similar to those of the lowest 50m of the member except that they are mostly of very fine sand grade.

The current directions indicated by cross-lamination throughout the member are shown in Fig. 53. Two flute

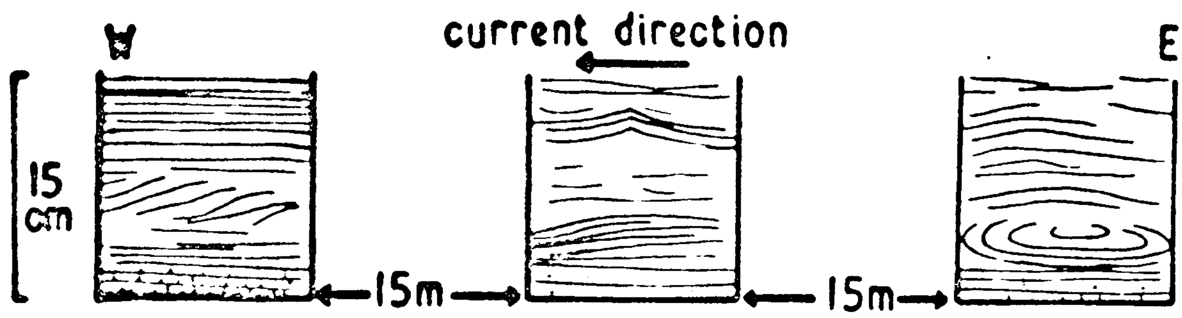


Figure 51. Sedimentary structures seen within one sandstone bed about 15m above the base of the Upper Breivik Member. Coast south of Breivik.

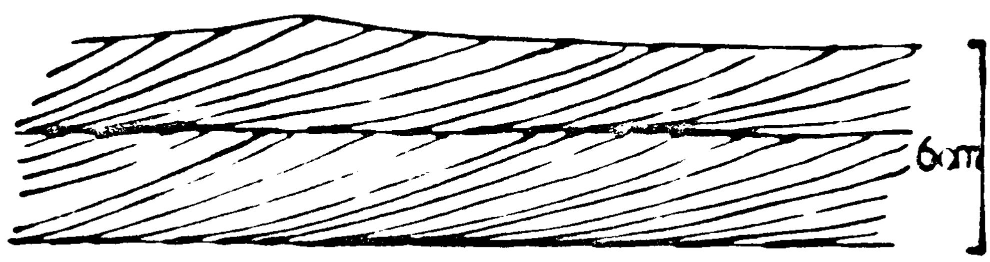


Figure 52. Type A climbing ripples in which the angle of climb is so small as to be unrecognisable over a short lateral distance. Upper Breivik Member on the coast near Breivik.

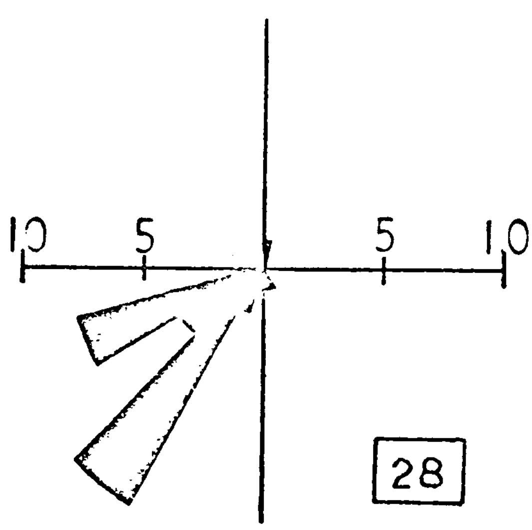


Figure 53. Palaeocurrent data based on cross-lamination from the Upper Breivik Member in the coastal exposure near Breivik.

marks observed suggested current flow towards 190°. Bioturbation is intense at some levels and strongly modifies the original lamination; this is well seen on the coast just north of Breivik (Pl. 69). However, in general, burrowing has not greatly disturbed the stratification. A variety of burrows is present, the most common ones being simple oblique and horizontal compacted tubes up to 5mm across. Phycodes palmatum and Teichichnus are present, the latter showing a lamination which in vertical section can be confused with current-formed cross-lamination as noted by Sellwood (1970).

Lateral Variation

No other sections were measured through the member but at the southern end of the peninsula, south of Hill 551 (Appendix B) the member consists of rather massive green siltstones with sandstone layers only becoming frequent towards the top.

Interpretation of Upper Breivik Member

The base of the member appears to mark a considerable change in the sedimentary regime but there is no evidence of either subaerial or submarine erosion at the junction with the underlying member. This basal bioturbated bed is interpreted as having accumulated slowly during a phase of reduced sediment supply and, since it is overlain by fine-grained beds, it probably marks a period of increasing water depth. This phase of deepening was at least as widespread as the present area of outcrop and probably nearly synchronous over this area.

The sharp-based sandstones of the lowest 50m represent intermittent rapid influxes of substantial amounts of sand into an area of otherwise fine-grained sedimentation. The presence of parallel lamination below cross-lamination suggests deposition from waning currents which flowed southwestwards and which were strong enough for upper flow regime bedforms to be developed. The absence of dunes is due to the fine grain of the sand (Allen 1970a).

Much the same situation seems to have been present throughout the remainder of the member although the deposits of sand were smaller and in rare instances the currents also flowed northeastwards. Deposition was confined to lower flow regime conditions and the type A climbing ripples suggest a low rate of deposition relative to ripple migration (Allen 1970b).

The trace fossils _Teichichnus_ and _Phycodes_ were included by Seilacher (1965) as members of his _Cruziana_ facies and thus suggest a neritic environment of deposition in the main section. The rather restricted trace fossil assemblage is probably indicative of conditions in the quietest, and in this case deepest, part of the _Cruziana_ facies. The finer grained aspect of the beds in the south fits in with the predominant southwestward transport of sand and possibly indicates that these beds were deposited in somewhat deeper water. The origin of the currents which deposited the sandstone beds will be discussed after the Lower Duolbasgaissa Member has been described.

4, 3. LOWER DUOLBASGAISSA MEMBER

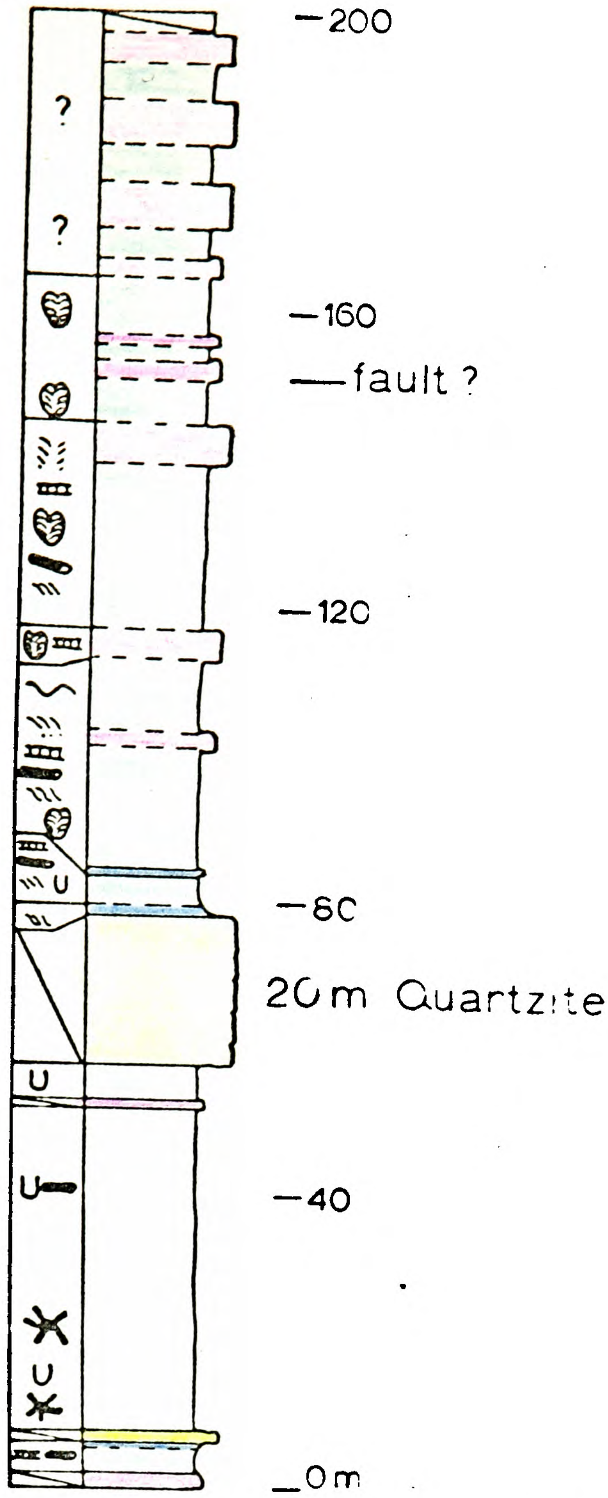
Introduction

The Lower Duolbasgaissa Member consists mainly of very thin to medium-bedded fine sandstones with interbedded siltstones and mudstones. Most of the member is a slightly coarser and thicker bedded continuation of the Upper Breivik Member facies and the trend towards coarser and thicker beds continues throughout it. It also includes a 20m thick unit of cross-bedded coarse sandstones and granule conglomerates and several similar but thinner units. The member is 200m thick and is best exposed in a bay 2km north of Breivik. The section at the southern end of the bay is described below and shown in Fig. 54. The beds above the 20m Quartzite were measured in the raised cliff.

Section exposure north of Breivik

There is no clear boundary between this member and the Upper Breivik Member but the Lower Duolbasgaissa Member may be distinguished by the common reddish-grey colour of its lower beds, by the presence of some thick sandstone beds and occasional granule conglomerate beds and by the occurrence in places of large horizontal burrows. The base of the member is arbitraily taken at the base of a 2m thick reddish-grey, graded sandstone unit which crops out on the coast just south of the bay. The lowest 10m of the section is shown in Fig. 55, partly also in Pl. 70.

Sharply above the coarse sandstone, but without a burrowed contact, come laminae and very thin beds of reddish-grey fine sandstone with interbedded siltstones and mudstones, the sandstones being similar to those of the Upper Breivik Member. They have sharp bases, show parallel and cross lamination and their palaeocurrent pattern is identical showing a dominant mode to



Mainly laminated and 1-10 cm bedded, bioturbated fine sandstones with interbedded siltstones and mudstones. Occasional thicker sandstones and 1-15 cm bedded "grit" beds.

Mainly medium to thick bedded fine sandstones with flat bedding, low angle cross-bedding and ball and pillow structures. Subordinate siltstones and mudstones.

1-10 cm bedded fine to coarse sandstones and granule conglomerates with interbedded siltstones.

Fine to very coarse sandstones and granule conglomerates with cross sets 25-150 cm thick.

Unexposed.

Sharp junction.

Gradational junction.

TRACE FOSSIL SYMBOLS

Rusophycus

Monomorphichnus

Dimorphichnus

Diplichnites

No trace fossils

Plagiogmus

U Arenicolites

Sinuous horizontal burrows

Radial burrow systems

Burrows as above but not radial.

Figure 54. Lower Duolbasgaissa Member section north of Breivik.

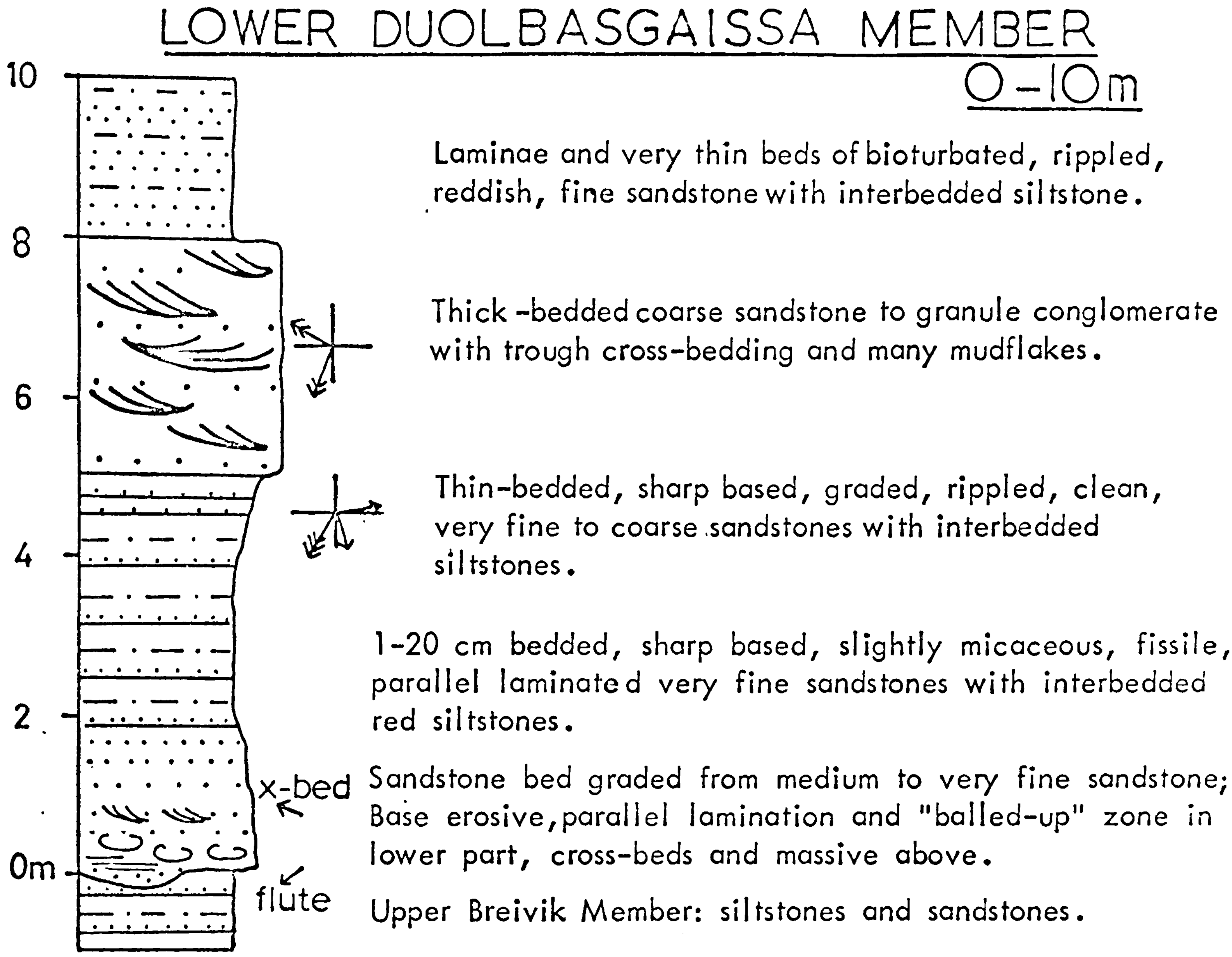


Figure 55. Lower Duolbasgaissa Member section, 0-10 m.

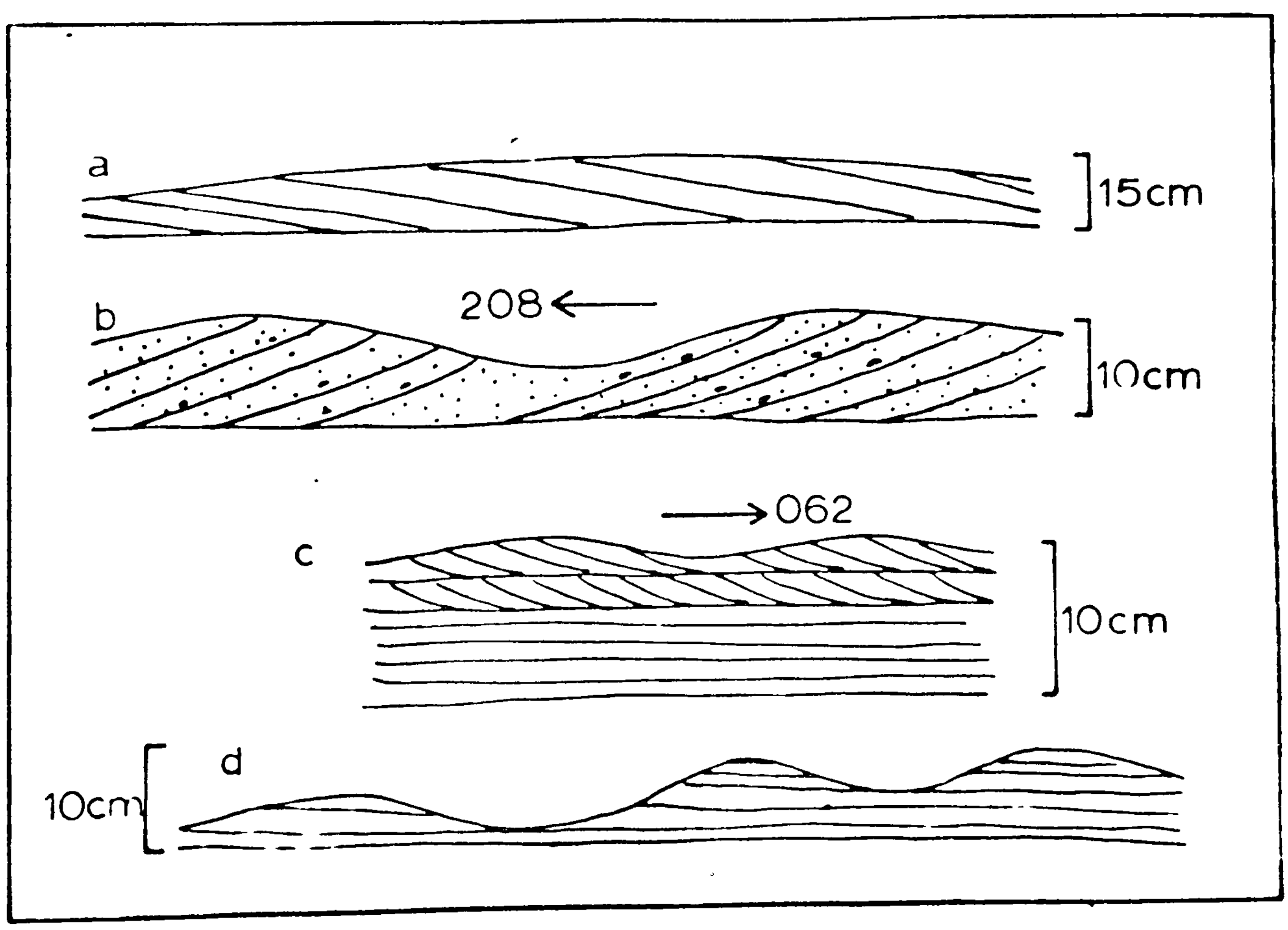


Figure 56. Sedimentary structures in the Lower Duolbasgaissa Member. a). Low angle cross-bedding in fine sandstones.

b). Small dunes in "grit" beds.

c). Parallel lamination overlain by cross-lamination in fine sandstone.

d). Irregular eroded top to sandstone bed.

the southwest and a much weaker one to the northeast (Fig. 57). Most beds show considerable lateral continuity; a few are slightly graded. The distinguishing feature of these beds apart from their colour, is the presence of 2cm wide horizontal burrows which in places form a radial pattern (Pl. 71). The sandstones are moderately to well sorted and of fine sand grade; they are subarkoses and quartzarenites with siliceous and iron carbonate cements and a minor amount of disseminated chlorite and microcystalline quartz.

In the sequence up to the base of the 20m Quartzite unit the average bed thickness gradually increases and parallel lamination becomes the most common sedimentary structure particularly in the thin to medium scale beds. These thicker beds also show a low angle cross-stratification in which the foresets are characteristically straight, variable in direction of dip within any one bed, and dip at less than 15° (Fig. 56a). Primary current lineation is commonly seen in the parallel laminated beds (Pl. 71) and occasionally on the low angle foresets where it is orientated parallel to the dip of the foresets. Cross-lamination frequently occurs above parallel lamination. but very rarely below it (Fig. 56c, Pl. 72).

Just below the 20m Quartzite a 1m thick sandstone unit occurs which shows a ball and pillow type of deformation (Pl. 73). Thin shale partings within this unit suggest that several sandstone beds were deposited in discrete events but then they all became "balled-up" together before the deposition of any overlying bed.

As the average bed thickness increases upward the beds become more lenticular and irregular erosion of their top surfaces becomes marked (Fig. 56d). This

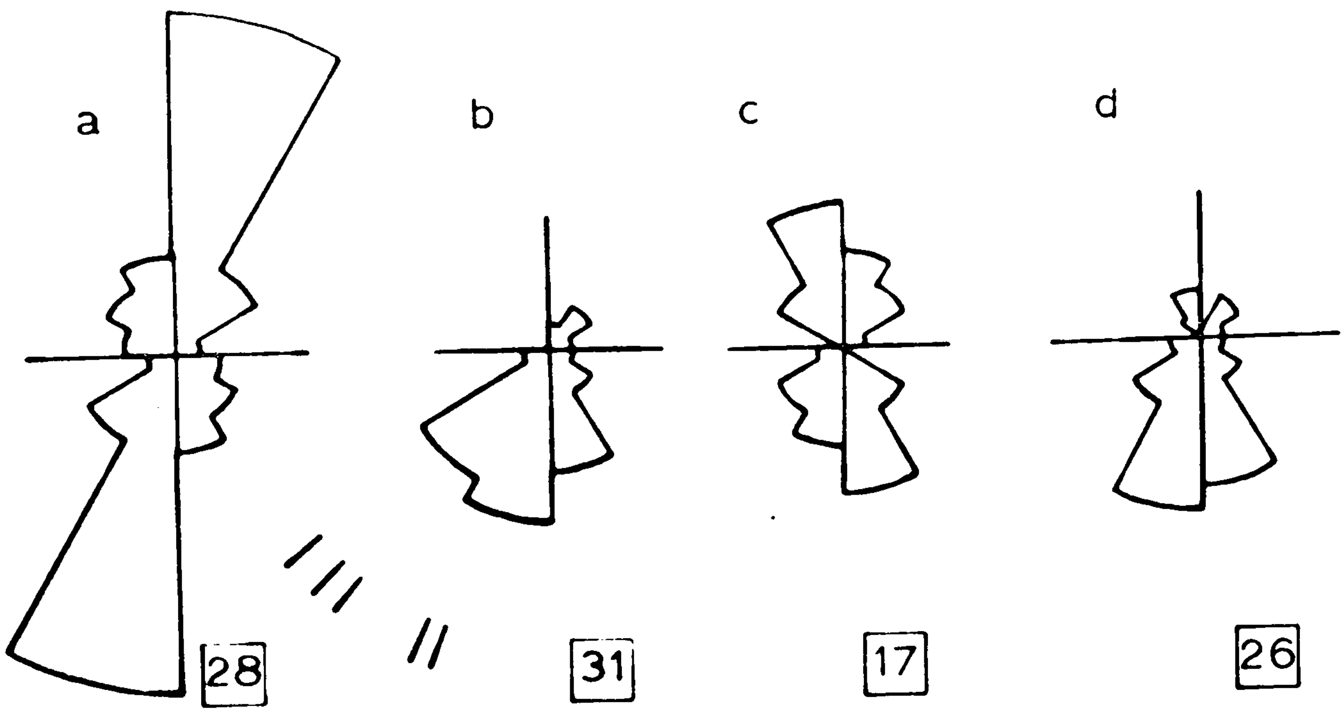


Figure 57. Palaeocurrent data for the L. Duolbasgaissa Mbr.
a). Primary current lineation in fine sandstones.
b). Cross-lamination in fine sandstones.
c). Symmetrical ripple axes.
d). Cross-stratification in "grits".

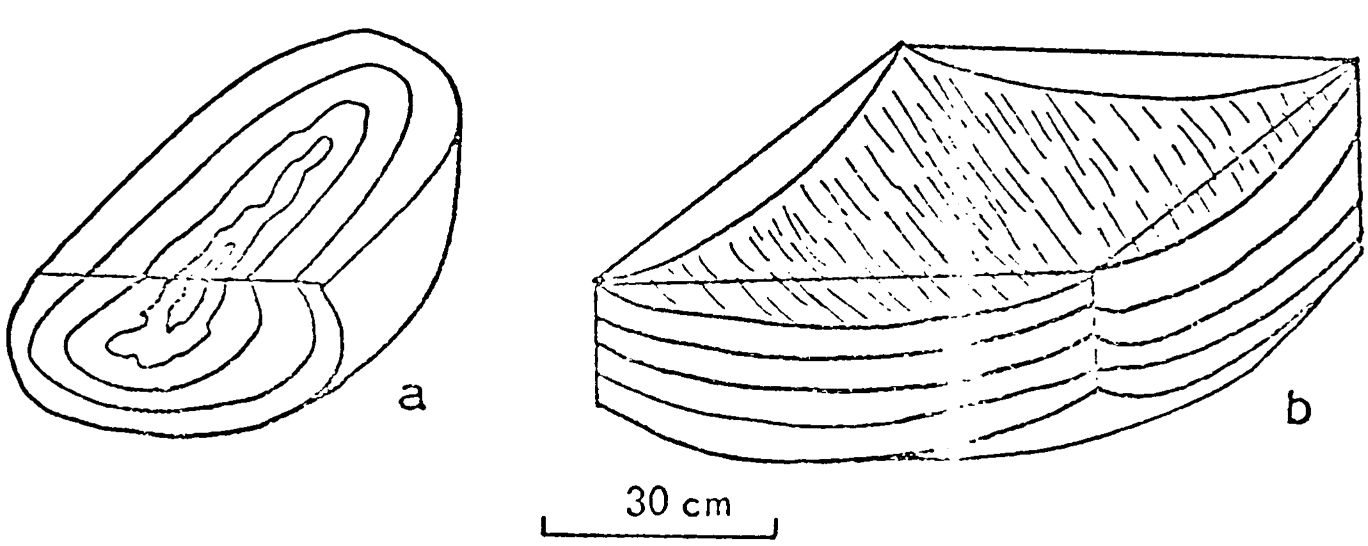


Figure 58. Penecontemporaneous deformation structures in the fine sandstones of the Lower Duolbasgaissa Member.

a). Ball and Pillow structure

b). Slightly deformed bed whose curved bedding could be confused with current formed cross-bedding except that there is primary current lineation on the bedding surfaces.

erosion probably reflects the increased ability of currents to erode the substratum. Also in this sequence symmetrical ripples become more abundant upwards and, concomitantly, beds of coarse to very coarse sand and granule conglomerate become common (Pl. 74). These "grit" beds are up to 15cm thick and their bases are occasionally slightly loaded; internally each bed usually consists of a single row of large scale ripple marks (Fig. 56b) although some of the thinner beds form graded units. Cross-stratification directions in the "grits" are variable but mainly dipt to the south (Fig. 57d). The grits are siliceous, supermature, quartzarenites.

The trace fossil assemblage also changes as the 20m Quartzite is approached. The large horizontal burrows, which are common in the thinner bedded sandstones are replaced upward by dominantly vertical burrows, some of which are U-tubes. These burrow-tubes are less than 5mm in diameter and no septa occur within the "U".

The 20m Quartzite consists of cross-bedded, medium to very coarse sandstones and granule conglomerates which are siliceous, supermature quartzarenties and mineralogically identical to the "grits" below (Acc. ser 19323). At the base of the unit there is a sharp, planar contact with the underlying facies (Pl. 75) which can be traced for at least 50m. No evidence of major erosion is seen although the junction is somewhat irregular on a small scale. The lowermost beds are granule conglomerates and contain numerous chert and hematite clasts and many mudflake pebbles. Within the unit the thickness of the cross-sets varies from 25-150cm and is correlated with grain size, the thickest sets being formed of the coarsest material. Trough cross-beds predominate but the largest sets sometimes approach a tabular form. Cross-bedding dips suggest that current flow was mainly to the northwest but southeasterly and southwesterly flowing currents were also present (Fig. 59 Section 3). Grain size and cross-bed thickness decrease gradually upwards. Trace fossils are absent and there are no drapes of fine-grained material on the cross-beds although mudflake pebbles are common. One 30cm thick horizon of primary current lineated fine sandstone similar to the sandstones below occurs 2.7m above the base.

The top of the 20m Quartzite is sharp and the overlying facies consists of 1-15cm bedded medium to coarse sandstones and interbedded siltstones similar to those at 4.5-5m.

The remainder of the member is broadly similar to the beds just below the 20m Quartzite. Green-grey fine sandstones are interbedded with green siltstones and mudstones and the typical lithology is shown in Pl. 77. Several sandstones have well developed flute marks on their soles and symmetrical ripples are common on the tops of beds. Flaggy sandstone beds up to 1.5m thick are found and several of the thickest beds have been deformed into ball and pillow structures (Pl. 78, Fig. 58). A syn-depositional origin for these structures can be proved in most cases using the following criteria:-

(i) The lower part of the bed is more deformed than the upper part.

(ii) The truncation of balls or upturned laminae by less deformed laminae within the bed.

Where laminae have been upturned less than 30° a cross-stratification is produced which might be confused with current formed structures or a scour fill such as a wave scour. The circular nature of these load structures in plan view distinguishes them from current formed cross-stratification and the common unidirectional primary current lineation is a feature unlikely to be associated with wave scours (Fig. 58b). In each of these units deposition seems to have occurred as a single more or less continuous event rather than as separate events with intervening periods of mud deposition as happened in the bed just below the 20m Quartzite.

The thick-bedded sandstone become increasingly frequent towards the top of the member where they are slightly coarser, several beds being of medium sand grade, and more glassy looking, this fact reflecting the presence of a wholly siliceous cement. Flat beds and low angle cross-beds both with primary current lineation are the main internal structures. Grit beds are a persistent but minor feature above the 20m Quartzite and they are commonly graded. In this part of the succession there is every gradation between the grits and the fine sandstone beds.

All the trace fossils found beneath the 20m Quartzite occur again above it but the assemblage is dominated by forms such as Rusophycus, Cruziana and Dimorphichnus which are probably attributable to trilobites. Bioturbation is a distinctive feature of the thinner bedded parts of the succession but no burrows are seen in the thick-bedded sandstone.

Lateral variation

The most obvious lateral variation within the member occurs in the 20m Quartzite (Fig. 59). Reading (1965) noted that this unit dies out when traced southwest from Breivik. North of Locality 1 it continues to the north end of the peninsula with a more or less constant thickness and lithology and southwest of

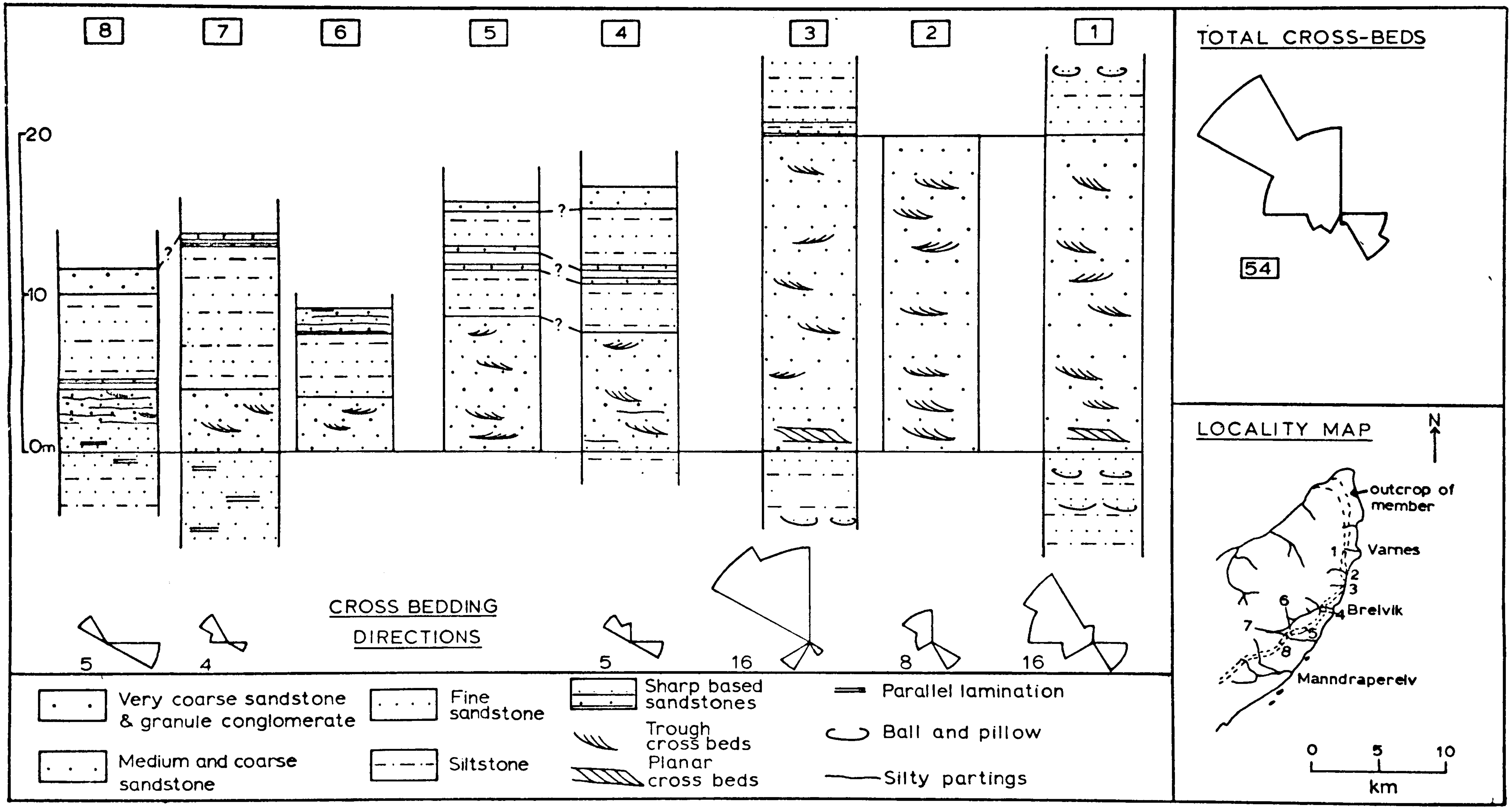


Figure 59. Lateral variation within the 20m Quartzite.

Locality 8 it gradually thins so that at Duolbasgaissa, 7.5km to the south west, the horizon is marked only by a few thin-bedded "grits". Although the average grain size decreases southwestwards some granule conglomerate beds are present as far south as Duolbasgaissa.

Sections 1 and 2 are similar to the main section (3) in grain size, petrography, in the style and orientation of cross-bedding and in having sharp bases and tops (Pl. 79, 80). The cross-bedding directions are distinctly bipolar and herringbone cross-beds are common. The foresets dip at up to 30°, are asymptotic to tangential in shape and in the largest sets they show slight grading from bottom to top. However, the sections differ from the main section in that there is neither a marked correlation of grain size and cross-bed thickness nor any regularity in the grain size distribution within the unit.

To the south of section 3 the 20m Quartzite splits up into a number of units intercalated with the typical Lower Duolbasgaissa facies of thin-bedded fine sandstones with interbedded siltstones and mudstones (Pl. 81). The basal unit always sharply overlies thin to medium-bedded fine sandstones which show flat bedding or low angle cross-beds and which have thin siltstone interbeds (Pl. 82). Medium to thick-bedded units of the Quartzite show cross-bedding and the bipolar orientation is maintained. This beds may show grading or be massive and many are poorly sorted. The coarse beds in the upper parts of sections 4-8 are commonly purple in colour whereas most other beds are white or buff coloured.

Little lateral variation is present within the remainder of the member. The succession at Varnes is very similar to that in the main section. To the south, the section at Duolbasgaissa is also similar apart from the absence of the 20m Quartzite. The thin-bedded sandstones show consistent cross-lamination directions towards the southwest quadrant and the member becomes gradually thicker bedded upward. At the southern end of the peninsula, on the south side of hill 551, the member is rather thinner bedded and is slightly finer-grained than elsewhere.

Interpretation of Lower Duolbasgaissa Member

The episodic nature of the deposition of the fine sandstone beds is less obvious than in the underlying member but a number of beds show clear evidence of having been deposited from waning currents (e.g. Fig. 56c).

The abundance of flat bedding with primary current lineation in the sandstones shows that these beds were deposited by stronger currents than most of the beds in the underlying member. The irregular low angle cross-stratification which is usually found above flat bedding but below small scale cross-lamination is thought to have formed from the migration of the "washed-out dunes" of Simons and others (1965) in the transition zone between the upper and lower flow regimes. The absence of true dunes may be due to the narrow range of flow strength in which dunes form at this grain size (fine sand) (Guy and others 1966). However, since at least some beds were deposited from waning currents Allen's (1970a) deductions may be applicable (Fig. 44).

The trace fossil assemblage, which is more diverse than in the Upper Breivik Member, is representative of the Cruziana facies but it is not clear whether this increased diversity is related directly to a decrease in water depth or merely to the increased energy of the environment. However, the abundance of symmetrical ripples demonstrates an increase in wave activity at the sea bed and this suggests that the sea was probably shallower than in Upper Breivik times. Thus it seems that current strength varied proportionately with water depth. As before the sea was probably deepest in the south at this time and from this it is suspected that the boundary between the Upper Breivik Member and the Lower Duolbasgaissa Member is diachronous, being youngest in the south west.

What type of current deposited the fine sandstones of the Lower Duolbasgaissa Member and, by extension, the very fine sandstones of the Upper Breivik Member? Any postulated current must have been able to:

a) Produce sediments with a bipolar palaeocurrent pattern in which the minor mode is almost absent in the deepest water facies.

b) Produce mineralogically mature sediments.

c) Produce texturally immature to mature or supermature sediments (rounding not estimable).

d) Produce sediments whose deposition was episodic with some beds clearly deposited from waning currents.

e) Produce primary current lineation and flat bedding in fine sand.

f) Become weaker with increasing depth.

From the discussion in Chapter 2 it is considered that tidal currents and coastal storm surge currents are the most likely contenders (Table 4). If storm surges were responsible the bipolar current pattern would be the result of onshore and offshore transport with the offshore mode predominant. The main objection to this process is that the amount of onshore transport (reflected in the minor mode) is rather greater than might be expected. Thus it is believed that these

P = Possible PR = Probable I = Improbable	Bipolar palaeocurrents	Mineralogically mature	Textural maturity variable	Episodic deposition	Strong enough to give p.c.l. in fine sand	Strength decreases with depth
Semi-permanent	P	P	P	I	P	P
Tidal	PR	PR	P	PR	PR	PR
Wave drift	I	P	P	PR	I	PR
Coastal storm surge	I	P	PR	PR	PR	PR
River generated	I	I	P	PR	PR	PR

Table 4. Table to show the likelihood of various current types producing the observed features of the fine and very fine sandstones of the Upper Breivik and Lower Duolbasgaissa Members.

sandstones were probably deposited by tidal currents.

As was suggested in the Lower Breivik Member storm conditions were probably important in strongly augmenting tidal currents as they do in modern seas (Stride 1963, Hadley 1964, Houbott 1968, Johnson and Stride 1969).

Rectilinear or strongly elliptical tidal currents would give the bipolar current pattern and the imbalance between the modes resulted from the fact that south-westerly flowing currents were generally stronger than the northeasterly flowing ones; from this imbalance it is inferred that there was a net transport of material to the southwest. The variable thickness of the sand-stones reflects the combined effect on current strengths of natural variation of tides and the effects of the weather. Johnson and Stride (1969) suggest that in the North Sea transport rates during spring tides enhanced by storms would be about one hundred times that of neap tides under light winds.

In the deepest water facies, represented by the Upper Breivik Member, it is probable that tidal currents were frequently too weak to carry sand and only trans-ported suspended silt and clay. Under these conditions fine-grained sediment would have accumulated slowly.

It is thought that the environment in which these very fine and fine sandstones were deposited was similar to the zone of fine sand and muddy sand described by Belderson and Stride (1966) from the Celtic Sea. This zone occurs towards the end of major tidal sediment transport paths down which sediment grain size and maximum current velocities gradually decrease. The fine and muddy sand zone occurs beyond a zone in which sand waves of coarser sand and gravel are accumulating.

If this comparison is valid it is likely that a sand wave facies was deposited contemporaneously with the fine sandstone facies and probably to the northeast of it since that is the direction from which sediment was transported.

It might be thought that the presence of upper flow regime flat bedding in the fine sandstones implies stronger currents than the cross-bedding seen in coarser beds. However, Fig. 44, shows that this need not be true.

The 20m Quartzite

Although the 20m Quartzite is seen to die out to the soutwest its three dimensional form is unknown. However, it is likely to have a fair lateral persistence along the axis of the palaeocurrent directions (i.e. NW-SE). If that is so it is a tabular body whose margins are unknown except for the ragged south-westerly one where it rapidly passes laterally into finer grained beds.

The ubiquitous cross-bedding suggests that the Quartzite formed as a result of the migration of dunes of medium sand to gravel in response to currents which flowed northwestwards and to a lesser extent southeastwards. Since the maximum cross-set thickness is 1.5m the dunes were probably not much more than 2m high. The scarcity of fine-grained interbeds suggests that there were few quiet water periods of sufficient length to allow the settling of fine material. The absence of bioturbation is probably due to a combination of an inimical environment for infaunal animals and a low preservation potential for any burrows which were produced.

A shallow marine origin is postulated for the 20m Quartzite because of the textural and mineralogical maturity of the sediments, because of its presence within a marine sequence and because of its bipolar palaeocurrent pattern. An estuarine origin is unlikely since the Quartzite interfingers with deeper water sediments transverse to the directions of sediment transport. In an estuary deposits of tidal flat and beach ridge origin usually lay at the margins of the estuary transverse to the directions of sediment transport.

If the Quartzite is of marginal origin it must have formed as the shoreline prograded southwestwards during a regressive phase. The bimodal current pattern could be attributed to longshore currents. However, the sharp lithological change at the base of the Quartzite and the lack of any consistent changes in either grain size or sedimentary structures within it make this interpretation improbable and an offshore tidal origin is favoured.

If sand waves were present they would have been on a small scale, hardly more than dunes (mega-ripples) since the maximum cross-set thickness in the Quartzite is 1.5m. Tidal current ridges attain heights of 35m (Houbolt 1968) and they are often covered with dunes which produce cross-stratification of the scale seen in the Quartzite. On the other hand Houbolt's work suggests that the cross-bedding dips formed on tidal current ridges are more variable than those seen in the Quartzite. Furthermore, since the Quartzite interfingers with laterally equivalent finer grained beds it must have accumulated over a considerable period of time; a rough estimate is 1m.y. Thus the 20m Quartzite could not represent a single fossilised set of tidal current ridges since such ridges stand proud above the surrounding clean swept sea floor and would not show interfingering relationships with finer grained sediment. Using the same argument the Quartzite is also unlikely to represent an accumulation of the eroded remnants of several ridges.

The best interpretation is that the 20m Quartzite represents a gradual accumulation of sediment in an area where dunes and small sand waves up to 2m high migrated under a NW-SE orientated rectilinear tidal current system. The direction of overall sediment transport was to the NW. From the previous discussion of the fine sandstones the 20m Quartzite facies might be expected to pass gradationally into a fine sandstone facies down the sediment transport path, i.e. to the NW. However, this cannot be proved.

This current system became rapidly weaker transverse to the transport path towards the southwest where the water was presumably deeper. This is shown by the southwesterly decrease in average grain size within the Quartzite and by its interfingering with finer, thinner-bedded sediment. In this southwestern area coarse material was only occasionally introduced to form grit beds, probably at times when the tidal currents were enhanced by storms. Each grit bed thus represents a single tidal event in the manner envisaged by Stride (1965). Although the 20m Quartzite comes at the top of a sequence in which both bed thickness and the number of grits increases upward the sharp change in lithology and in palaeocurrent directions at its base points to an abrupt change in the current system within the basin

The overall interpretation of the 20m Quartzite is shown in Fig. 60.

Other coarse-grained horizons

The "Quartzite" unit near the base of the Member (Fig. 55) has a series of thin-bedded fine to coarse sandstones below it and thus the sequence shows a gradual encroachment of coarser material into the area. Tidal currents are again postulated for these sandstone beds.

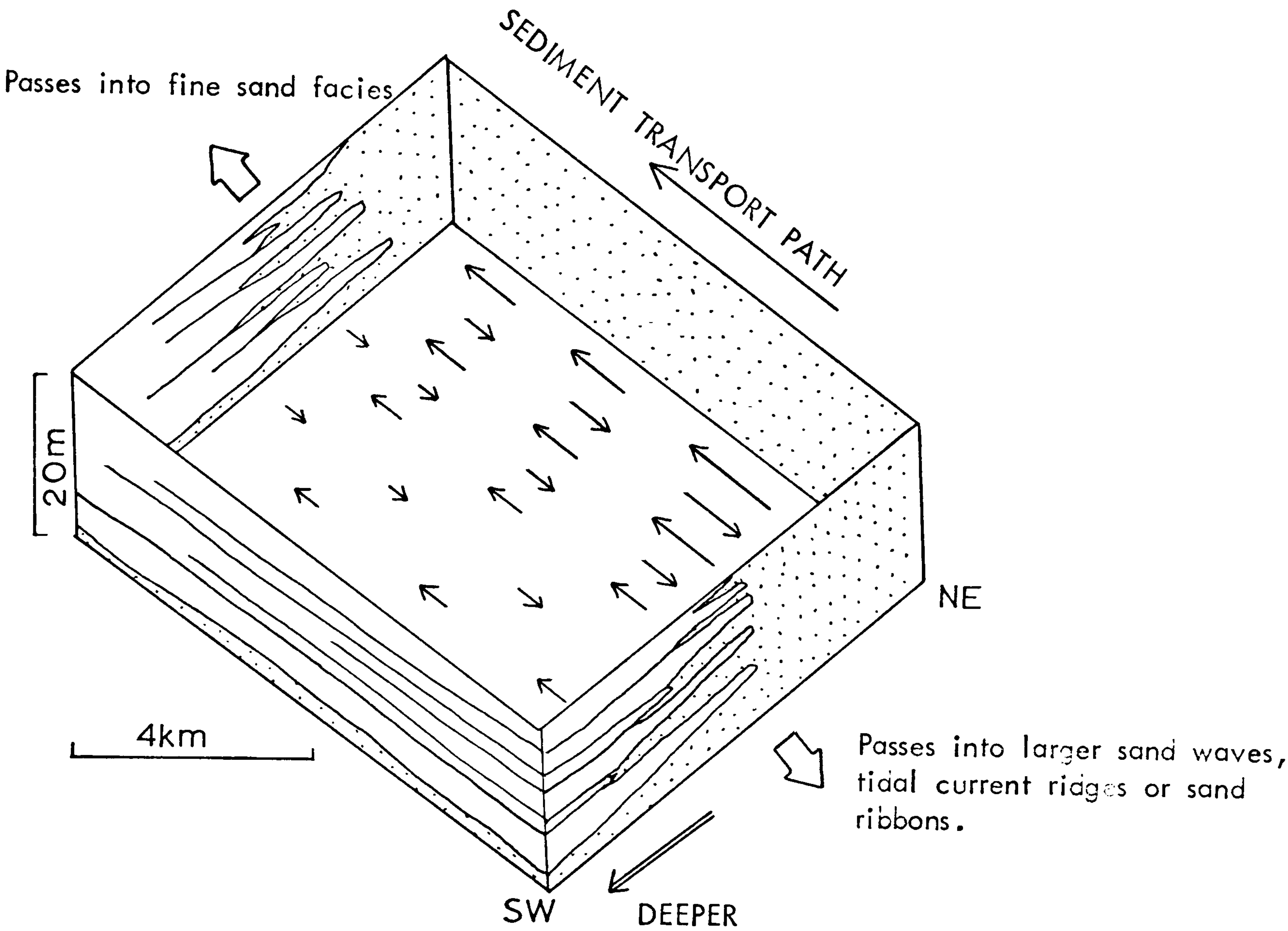


Figure 60. Speculative interpretation of the 20 m Quartzite based on the data of Fig. 59 which is represented in the nearer SW-NE face of this diagram.

The fact that the 20m Quartzite passes laterally into a series of thin grit beds provides an explanation for the grits seen in the remainder of the Member. These were probably carried by exceptionally strong tidal currents from areas of coarse sand and gravel accumulation and finally deposited in an area of otherwise finer-grained sedimentation.

Below the 20m Quartzite the grit beds which show grading are usually thinner than those showing cross-stratification and were probably deposited from weaker currents. In such coarse material grading is unlikely to form by fallout from suspension. However it might be formed in the "plane bed with movement" phase which occurs at <u>lower</u> current strengths than the dune phase in coarse sand (Williams 1967). Guy and others (1966) noted that under these plane bed conditions the coarsest grains gradually became buried and so a graded bed might eventually be formed, especially under waning flow conditions.

In summary, it is concluded that the Upper Breivik Member and Lower Duolbasgaissa Member were deposited under the influence of a rectilinear or strongly elliptical tidal current system which was usually orientated NE-SW but for a time during the Lower Duolbasgaissa Member changed to a NW-SE orientation resulting in the deposition of the 20m Quartzite.

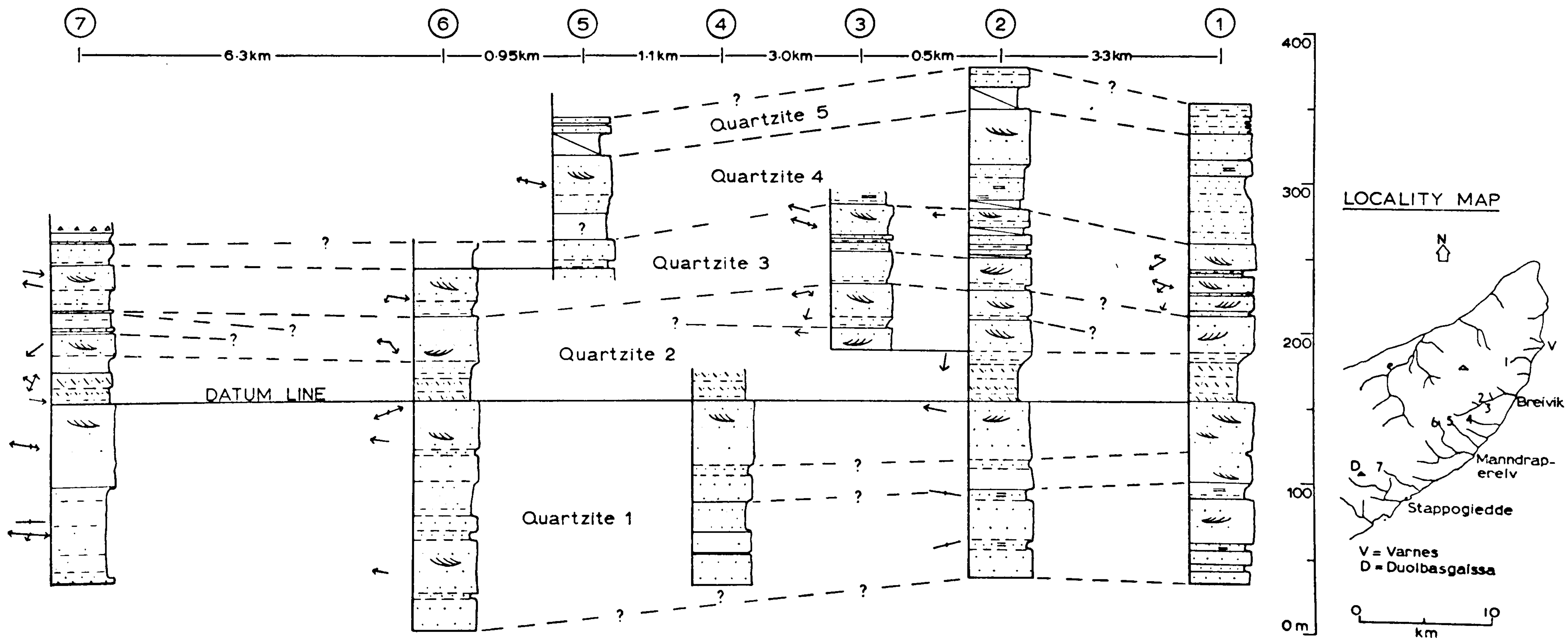
4. 4. UPPER DUOLBASGAISSA MEMBER

Introduction

The Upper Duolbasgaissa Member crops out over much of the plateau of the Digermul Peninsula and forms most of the marked escarpment around the plateau. (see Frontispiece). It is well exposed in this escarpment and many lithological units can be traced out for several kilometres. Reading (1965) gave the thickness as 300m but it is now known to be up to 350m thick. The member consists of 5-50m thick units of medium and thick-bedded quartzarenitic sandstones with subordinate intercalated units of thinner bedded sandstones, siltstones and mudstones.

The junction with the Lower Duolbasgaissa Member is taken at the base of the first major sandstone unit and the upper junction is taken at the top of the last sandstone unit which is >5m thick, the overlying Kistedal Formation being composed of thin-bedded sandstones and shales in its lower part. Lateral variation is more marked within this member than within any other (Fig. 62) and its study provides some of the major lines of evidence which lead to the understanding of the environmental history of the member. Logistic reasons confined the study to the outcrop between Duolbasgaissa and a valley 1.5km south of Varnes although the lowest part of the member was also seen at the southern end of the peninsula, on Hill 551. There is a slight tendency for thin-bedded sandstones with fine-grained interbeds to become more common southwestwards at the expense of the cross-bedded sandstones. Some general views of the succession are shown in Pl. 83, 84.

The section in Breivik Valley has been chosen as the main section because it is the most complete. The



KEY As for Figure 61 plus ▵ ▵ ▵ = conglomeratic sandstone.

Figure 62. Lateral variation in the Upper Duolbasgaissa Member.

lowest 150m of the section were measured on the floor of the valley, 150-258m on the southern slope of the valley, and the remainder on the northern slope (Fig. 61, Pl. 85, 86). The section on the southern slope equates with Section 3 of Fig. 62, the remainder equates with Section 2 of Fig. 62. The Member has been divided into informal numbered units (Quartzites 1-5) which are described and discussed individually. The interpretation largely follows from the arguments developed for the Lower Duolbasgaissa Member. It will be shown that the vertical and lateral facies distribution again favours an offshore tidal origin.

<u>Quartzite 1</u>

Main section

Quartzite 1 is 122m thick and consists largely of medium to very thick-bedded, white or pink fine sandstones to granule conglomerates which are siliceous, supermature quartzarenites. Siltstones and mudstones are confined to thin partings. Sedimentary structures are often obscured by diagenetic and tectonic modifications of the rocks but where they are present trough cross-bedding is characteristic and the lithology is very similar to that of the 20m Quartzite of the Lower Member except that herringbone cross-bedding is less common and a few trace fossils, notably <u>Skolithos</u> and <u>Syringomorpha</u> are present. Symmetrical ripples are occasionally seen and mudflake conglomerate horizons are frequent.

At 20-40m this lithology is interbedded with, and appears to grade into, thin to medium-bedded, flaggy grey, fine sandstones which are of typical Lower Duolbasgaissa facies in showing primary current lineation associated with flat bedding and low angle cross-bedding. This Lower Duolbasgaissa lithology occurs again at 54-60m (Pl. 87) where a sequence in which bed thickness gradually increases upward is sharply overlain by the typical cross-bedded sandstone lithology. The orientation of the primary current lineation in these fine sandstones is roughly E-W.

At the top of Quartzite 1 large (3-4m) cosets of cross-bedding can be seen on the southern face of the valley, the cosets consisting of a lower set about 2m thick with smaller sets above (Pl. 88, Fig. 63). Cross-bedding in these sets is towards the NW. Bedding planes at this horizon show 30cm high lunate bars, shallow troughs (c. 1m wide) with small symmetrical ripples within them and surfaces which are very irregularly sculptured and are similar to the basin and ridge structures of the Manndraperelv Member.

Lateral Variations

The thin horizons of Lower Duolbasgaissa facies found in Breivik Valley (Section 2) thicken markedly southwestwards at the expense of the cross-bedded "quartzite" facies. Concomitantly they also become thicker bedded and coarser so that in Sections 4, 6 and 7 they are no longer strictly similar to beds of the Lower Duolbasgaissa Member although there are many points of resemblance.

A thick unit of these "Lower Duolbasgaissa Facies" beds is found in Section7. It consists of bands of 1-10cm-bedded, rippled, grey, fine sandstones with green mudstone partings alternating with bands of 10-100cm -bedded flat-bedded and cross-bedded reddish-grey fine to coarse sandstones with infrequent mudstone partings (Pl. 89). Cross-bedding dips are less than 20° and the foresets are usually straight; flat beds often show primary current lineation. Ball and pillow beds up to 1.8m thick are present and a number of these

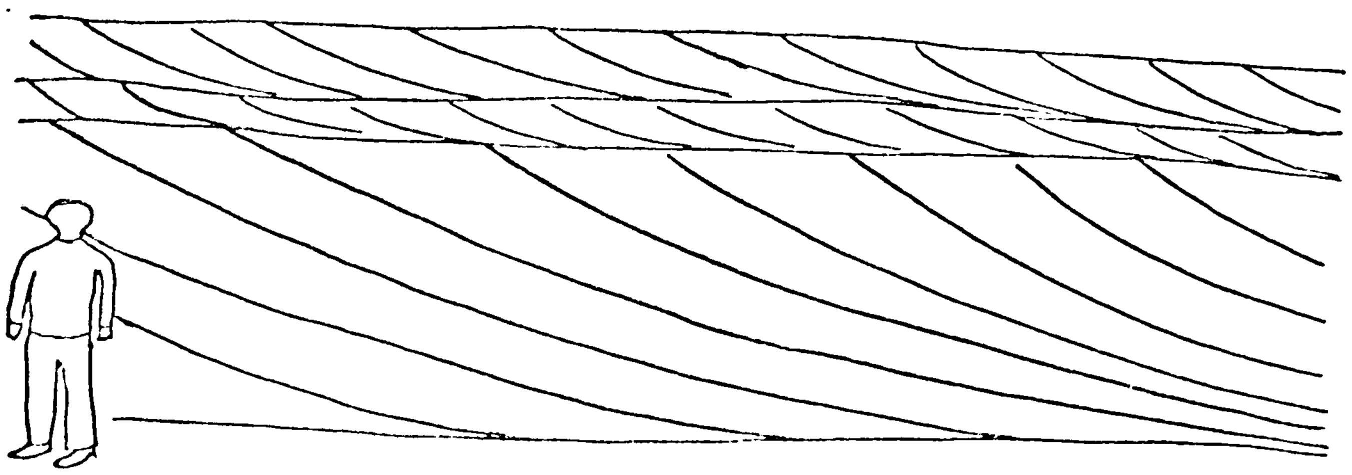


Figure 63. Large scale cross-bedding at the top of Quartzite 1, Upper Duolbasgaissa Member section, Breivik valley.

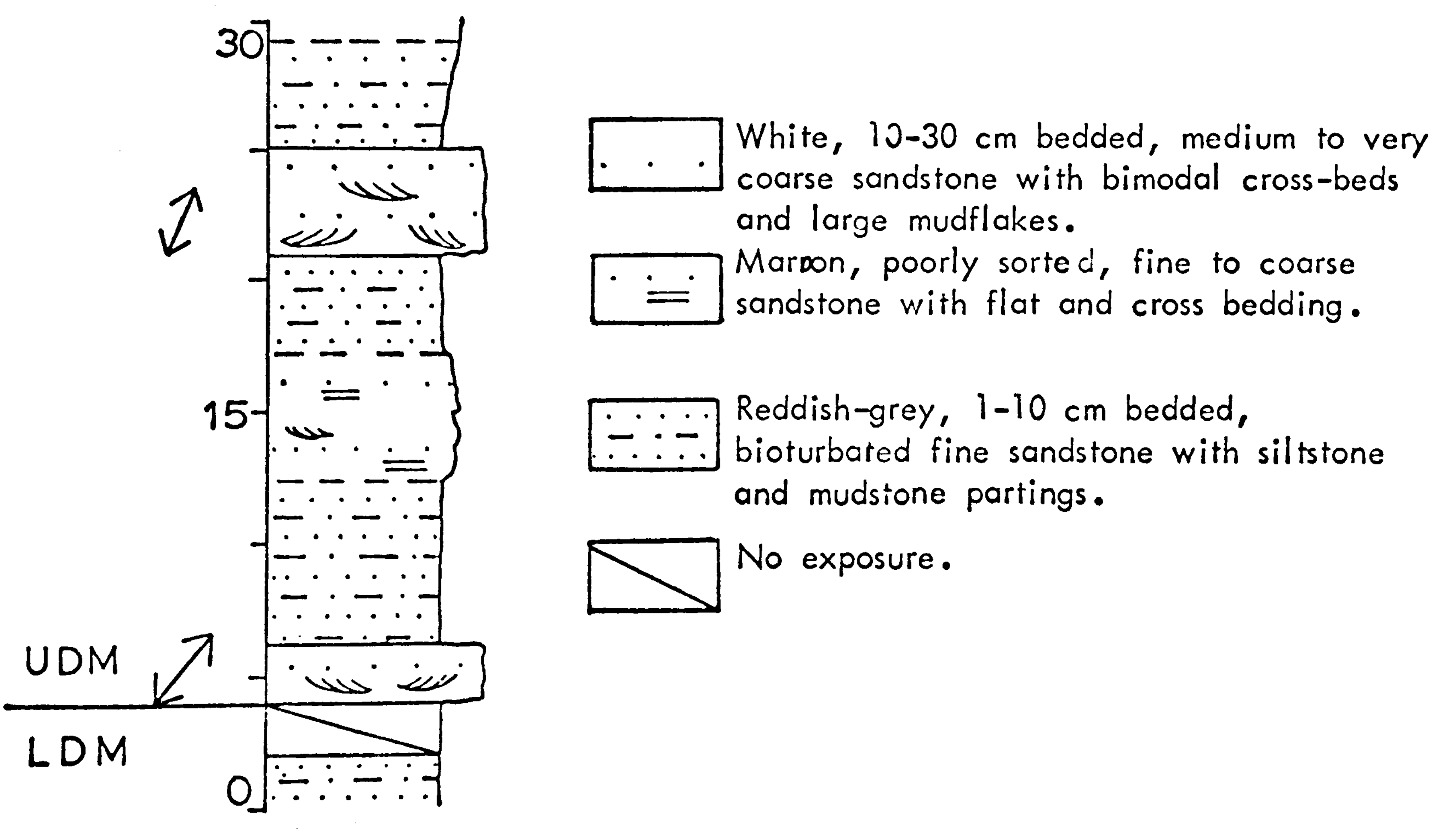


Figure 64. Beds at the junction between the Lower and Upper Duolbasgaissa Members on the south side of Hill 551, at the south end of the Digermul Peninsula.

can be traced laterally (within 100m) into undeformed flat-bedded sandstones. Whilst some of the thick beds seem to represent one phase of deposition others are undoubtedly composite beds. Occasional "grit" beds occur and symmetrical ripples are ubiquitous. Trilobite markings of all sorts are very common and Diplocraterion and Gyrochorte are also found. Primary current lineation orientations correspond well with those in Section 2 and the associated cross-bedding shows a bimodal pattern.

The medium to thick-bedded cross-bedded sandstones and granule conglomerates which make up the remainder of Quartzite 1 in sections 1, 4, 6 and 7 show similar features to those of the Breivik Valley Section. The large scale trough cross-beds seen at the top of Quartzite 1 in the main section were not observed elsewhere but in many cases the rocks are structureless due to diagenetic and tectonic modification. In sections 4, 6 and 7 the contacts between the two facies are mostly gradational.

On Hill 551 at the southern end of the peninsula the succession at the base of the member is shown in Fig. 64.

The bipolar cross-bedding in the white sandstones has an unusual orientation compared with all the other sections. The thin-bedded sandstones are of Lower Duolbasgaissa facies and the beds at 13-17m are similar to the reddish-grey sandstones described from Quartzite 1 in section 7. The thin-bedded sandstone facies continues above 30m for some considerably thickness.

Interpretation

By comparison with the 20m Quartzite the cross-bedded sandstones of this unit are thought to represent dune and small scale sand wave accumulations under tidal conditions. However, the presence of larger scale bed forms such as tidal current ridges cannot be excluded. The large cross-sets at the top of Quarzite 1 in Breivik Valley formed from sand waves 3-4m high which probably had dunes on their backs producing the smaller cross-sets seen above the large set. The thinner bedded, finer grained horizons are interpreted as lower energy tidal deposits similar to the fine sandstones of the Lower Duolbasgaissa Member. The commonly seen gradations between the two facies in Quartzite 1 confirms that they were both deposited by the same type of current; although a similar relationship between these two facies was postulated in the Lower Duolbasgaissa Member it is not so clearly seen in the field as it is here.

Since the cross-bedded facies becomes less common towards the southwest it is concluded that the tidal currents were generally weaker in the southwest. This trend was observed in the NE-SW trending outcrop which is at 45^{o} to theE-W orientated current system. Also it is probable that the base of the member, which is defined on the first appearance of a cross-bedded sandstone unit, is diachronous and becomes progressively younger towards the southwest. This diachronism is particularly suspected between Sections 6 and 7 where Quartzite 1 thins markedly.

If it is assumed that these sand waves had more or less straight crests it is possible to estimate the minimum depth of deposition for the uppermost part of Quartzite 1 (Allen 1970c, p. 78, 79); a value ofaproximately 20m is obtained which corresponds well with the depths of sand waves in the North Sea (McCave 1971). However if the sand waves were strongly three-dimensional the water depth could have been considerably less although they might then have been destroyed by wave activity.

Quartzite 2

Main Section

The lowest part of Quartzite 2 (Pl. 90) sharply overlies Quartzite 1 and consists of green, micaceous siltstones and mudstones with sharp-based 1-30cm bedded, grey, micaceous, fine sandstones which exhibit parallel lamination with primary current lineation and also show irregular wavy cross-lamination. Large horizontal burrows are common. Sharply above this facies come 1-25cm bedded sharp-based and sharp-topped "clean" sandstones with interbedded siltstones and mudstones (Pl. 91). These sandstones which are siliceous, supermature quartzarenites, have flutes and grooves, are mostly parallel laminated with cross-lamination at the tops of the beds, and show no grading. Fine-grained interbeds die out upwards and there is a gradational passage into cross-bedded sandstones. The succession is thus one of gradual upward coarsening.

On the southern side of the valley the main sandstone of Quartzite 2 is split into two parts. The lower part consists of typical cross-bedded fine to very coarse sandstones. Its topmost beds are similar to those at the top of Quartzite 1 in showing irregular bedding surfaces. These surfaces display the tops of vertical U-tubes (Diplocraterion and Arenicolites (?)) which are randomly orientated (Pl. 92).

This lower part is sharply overlain by another coarsening upward sequence which passes up into the upper part of the main sandstone which again consists of trough cross-bedded sandstones. Syringomorpha, a trace fossil of unknown origin is common at this particular horizon.

Lateral Variation

The lower finer-grained and thinner-bedded part of this unit forms a distinctive brown weathering band which makes an excellent marker horizon in all sections. Wherever it is seen it sharply overlies Quartzite 1 but although it is usually strongly bioturbated near its base it was never seen to be piped down into the underlying beds. In Sections 1, 4 and 6 this finer-grained band strongly resembles the main section but it is rather different in Section 7, a view of which is shown in Pl. 93. Symmetrical ripples are present in most sections. In Section 7 the zone of 1-30cm bedded, micaceous, very fine sandstones is about the same thickness 18m. About 2.5m above the base of the zone there is a o.6m unit of dark red, nodular, intensely bioturbated sandstone which can be traced at least two thirds of the way to section 6 and occurs progressively nearer the base of the unit in this direction. Above this bioturbated bed there are reddish-grey very fine and fine sandstones up to 1m thick as well as the usual thinner beds. These thick beds consist of large sets of steeply climbing ripples (Pl. 94, 95). Also in this zone there are several white, graded, coarse sandstones which have abundant mudflake pebbles and which, in some cases, show northward directed ripples. Such sandstones are not present in any of the other sections.

Above this zone but below the main sandstone unit of Quartzite 2 there come 11m of white, 1-5cm bedded fine to coarse sandstones with very thin mudstone partings. The sandstones are bioturbated and moderately to poorly sorted and this lithology, present also in

Section 6, represents a rather different facies to that of the more northerly sections which resemble Section 2.

The correlation of the main sandstone units of Quartzite 2 has not been completely established but it appears that a number of lensoid sand bodies are present. This can easily be seen within Breivik Valley (Fig. 65) The thinner bedded horizon which splits the main sandstone into two parts on the southern side of the valley (Section 3) is represented on the northern side (Section 2) by thinner-bedded, red sandstones. When this red band is traced down the northern side of the valley towards the NE it appears at a progressively lower position and finally disappears at the base of the sandstone. If this red band marks a sedimentary surface (i.e. is approximately isochronous) it shows that the lower part of the major sandstone at the NE end of the valley was deposited substantially later than the lower part at the head of the valley.

Another instance of lateral variation is seen between sections 7 and 6 (Fig. 66) although the detailed tracing of beds between these sections is hampered by tectonic complications.

Burrow tubes are abundant in the thin bedded sandstones of the upper part of Quartzite 2 and also occur at the tops of some medium bedded sandstones. They give the rock a honeycomb appearance in vertical section (Pl. 96, 97)(Acc. ser 19322). In rare instances these burrows have a vertically stacked arrangement. They are best referred to as varieties of *Phycodes*.

An unusual sedimentary structure occurs at the top of the thick sandstone unit in Section 7. It consists of rounded, symmetrical, straight-crested dunes with chord lengths of 35-70cm developed in a coarse sandstone (Pl. 98). The ripple heights are 5-6cm and the only internal structure seen was one instance of cross-lamination with a dip of 10° in a dune. These

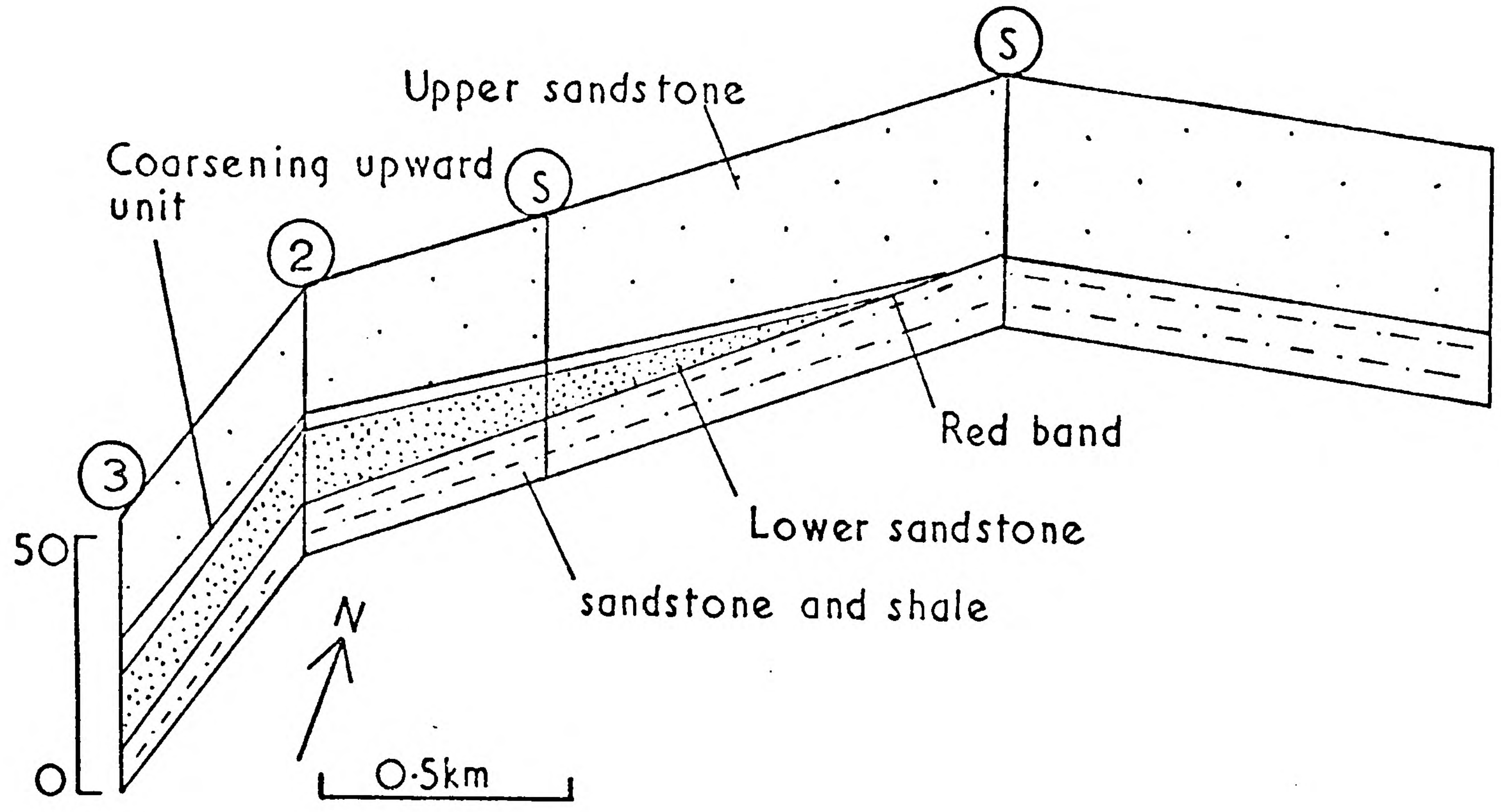


Figure 65. Lateral variation within Quartzite 2, Upper Duolbasgaissa Member, in Breivik Valley. Numbers refer to sections and S = stream.

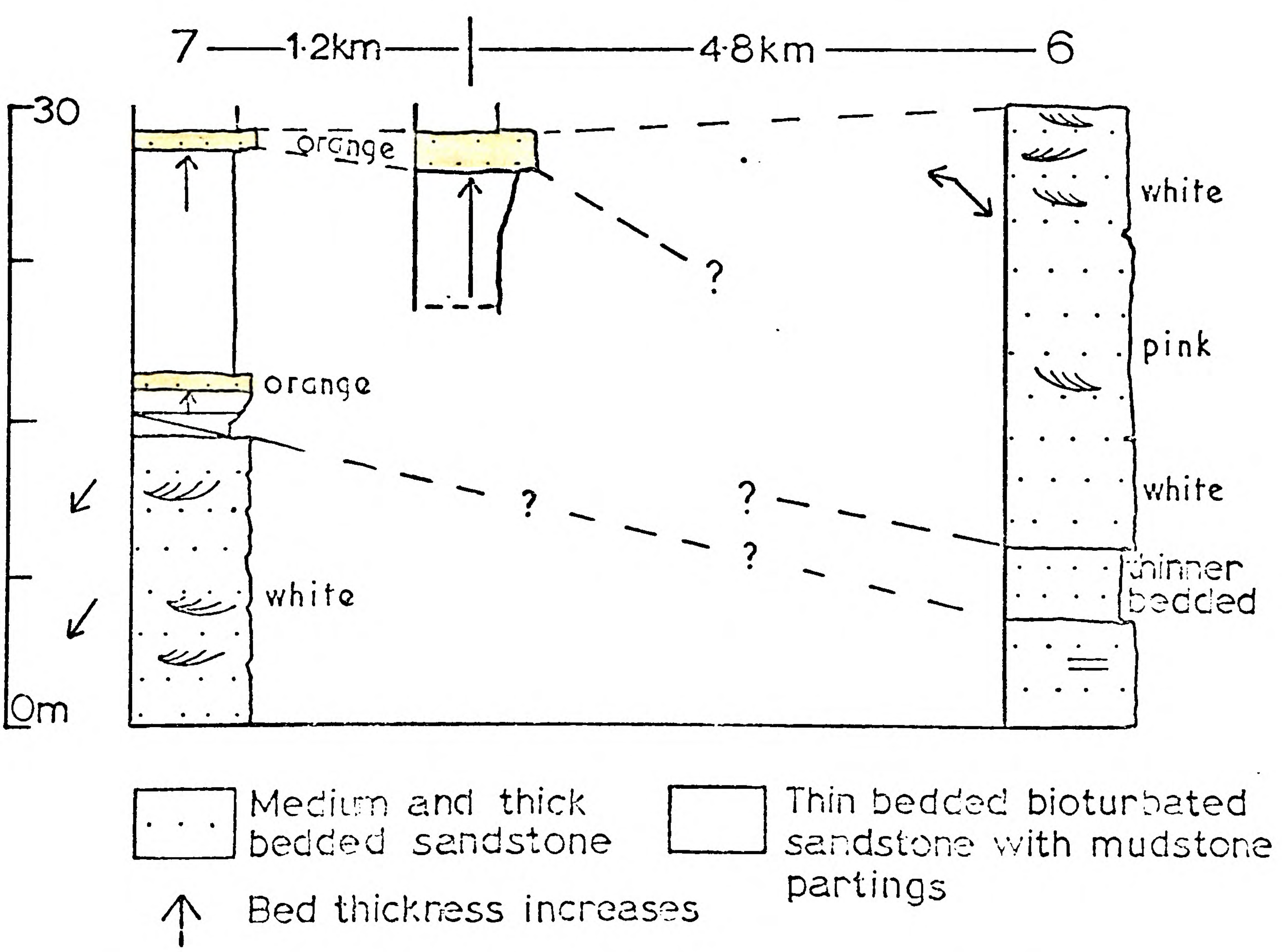


Figure 66. Lateral variation within Quartzite 2, Upper Duolbasgaissa Member, between Sections 6 and 7.

structures resemble the antidune bedding of Hand and others (1969).

The correlation line between Sections 2 and 1 marking the top of Quartzite 2 is extremely tenuous as no observations were made in the intervening area. The correlation is partly based on the presence in both sections of thin beds with abundant descication cracks just above the tie line.

Interpretation

The abrupt change to fine-grained sedimentation at the top of Quartzite 1 marks a widespread event which could have been the result of either a change in the current pattern at constant water depth or a sudden deepening of the sea with resultant decrease in current strengths. The difficulty of distinguishing between these two possibilities in tidal deposits has been discussed by Johnson and Belderson (1969). In this case the abruptness of the change suggests that an alteration in the current pattern is more likely. However, this alteration could have easily resulted, with reduced sedimentation in an increase in water depth. It is reasonable to suppose that the change was synchronous over the area and thus the top of Quartzite 1 approximates to a time plane.

The bioturbated bed near the base of Quartzite 2 in Section 7 is taken to mark a period of non-sedimentation during which extensive biogenic reworking occurred. The presence of cross-lamination above parallel lamination in the very fine sandstones of the lower part of Quartzite 2 shows that they were deposited from waning currents but in the thicker beds seen at this level in Section 7 the constant angle of ripple climb suggests deposition from rather steady currents (Allen 1970b).

Since the beds are sharp based it is evident that they were deposited from episodic currents and in water assumed to be relatively deep. However, the genesis of the currents is problematical; they could have been tidal but other origins cannot be excluded.

The white, graded, coarse sandstones which occur in this lower part of Quartzite 2 in Section 7 again represent episodic events of waning flow. The very abundant intraformational mudflakes testify to the high initial erosive capacity and the short duration of the currents. These beds were probably the result of major storms during which exceptionally strong currents (?tidal plus storm) transported material northward from an area of coarse sand accumulation. They are similar to the grits of the Lower Duolbasgaissa Member.

Accepting the relatively deep water origin of these beds the overlying sequence is a shallowing one. The overlying sharp-based and sharp-topped white sandstones of Section 2 were deposited by episodic waning currents flowing to the SSW. Taken out of context other origins for these beds might be considered (e.g. rip current deposits) but none of the alternatives is compatible with the evidence from the surrounding beds which favours offshore tidal deposition. The beds are interpreted as offshore tidal deposits, each bed again representing a single major tidal event. The area of deposition was probably downcurrent of a zone of sand waves.

It has been shown that there is rapid lateral variation within the major sandstone unit of Quartzite 2 and that it actually consists of several sandstone lenses. A tentative reconstruction of the arrangement of these lenses is shown in Fig. 67.

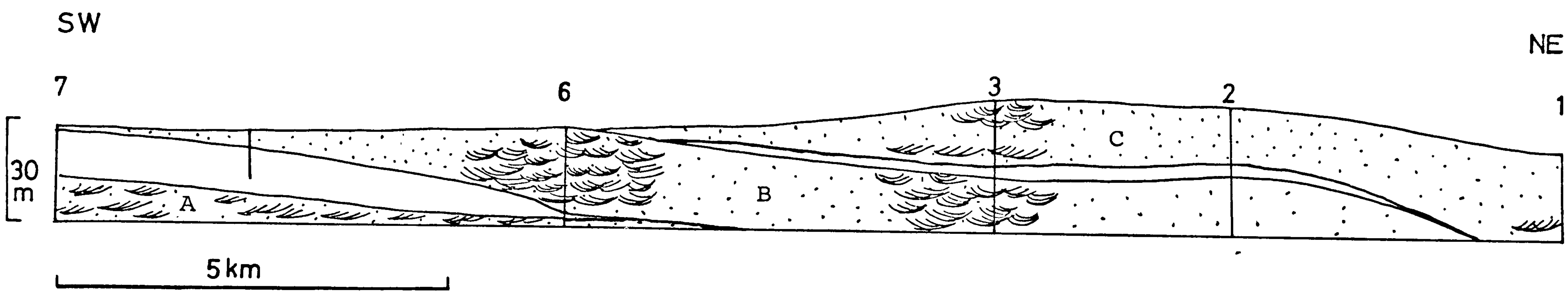


Fig. 67 . Suggested form of the sandstone bodies of Quartzite 2, Upper Duolbasgaissa Member. Cross-bedding shown only where measured. Numbers refer to measured sections.

If this reconstruction is accurate Lens A was deposited before Lens B which in turn was deposited before Lens C. Whilst Lens A was accumulating in the southwest the sea floor immediately to the northeast which was "upcurrent" of Lens A may have been a current swept region of slight deposition (sand ribbon area of Stride (1963) and Kenyon (1970)?). Indeed a series of dunes and small sand waves or a tidal current ridge might initially have been deposited in this northeastern area and then migrated southwestwards leaving the area as a sediment starved sand ribbon zone. Later the current pattern changed and the tidal sand body became fossilised to the southwest as Lens A. Alternatively it is possible that an equal thickness of sediment was deposited synchronously over the whole area but then the part in the northeast was eroded prior to deposition of Lens B. The burrowed tops of some beds (Pl. 96, 97) show that the sediment surface sometimes remained stable for sufficient time to allow biogenic reworking of the sand. If the symmetrical dune structures found at the top of Lens A in Section 7 are true antidune structures and not large wave ripples they suggest that these beds were deposited in water only a few centimetres deep (Hand 1969) and that Lens A was thus probably an emergent shoal at times.

When the current system changed through 90° Lens B formed to the northeast of Lens A which probably existed as a topographic high. However, thinner bedded sandstones representing lower energy conditions were deposited over Lens A and these beds must, at least in part, be laterally equivalent to the thick-bedded sandstones of Lens B. Again, prior to the deposition of Lens C the area to the northeast of Lens B may have been an area of non-deposition. Little information was collected from Lens C and it is uncertain whether it continues as far as Section 1.

Thus it is thought that each lens represents a tidal sand body, either a zone of dunes and small sand waves or a tidal current ridge and that each formed under different tidal current systems and at different times.

<u>Quartzite 3</u>

Main Section

The lower part of Quartzite 3 consists of 1-20cm bedded grey, rippled and cross-bedded sandstones with fine-grained partings which gradually pass up into white cross-bedded sandstones in which bed thicknesses are mostly 20-40cm. A sequence of thin-bedded sandstones, siltstones and mudstones with one 2.5m unit of purple sandstone follows and this gives way to a further coarsening upward sequence, 20m thick in which medium-bedded cross-stratified sandstones with fine-grained interbeds pass up into low angle ($\leqslant 15^{\circ}$) cross-bedded sandstones (Pl. 99). In the medium-bedded sandstones cross-bedding dips are dominantly to the NW with a minor trend to the SE. The low angle beds dip NW and in many respects resemble the beds at the top of Quartzite 1. Although the dip of the foresets is rather less the angle of dip often increases towards the top of the set reaching 15° in some cases.

Lateral variation

The correlation of the main sandstones of Quartzite 3 is fairly well established between Sections 2 and 7 but is more dubuious between Sections 1 and 2. The lower major sandstone is a distinctive unit of constant 20m thickness from Breivik Valley southwards and always occurs above a coarsening upward sequence of mainly thin-bedded sandstones which becomes thicker towards the southwest. In Section 7 a distinct bimodal palaeocurrent pattern is present, all the foreset dip directions in the upper half being diametrically opposed to those in the lower.

The upper major sandstone also continues through all sections south of Breivik Valley but the large low angle cross-beds seen in Section 3 were not seen elsewhere and the unit mostly consists of a coarsening upward sequence with medium-bedded sandstones with fine-grained partings at the top. In Section 1 a coarsening upward sequence of red trough cross-bedded sandstones seems to correlate with this horizon.

Interpretation

The ubiquitous rapid change to fine-grained sedimentation at the top of Quartzite 2 probably marks another phase of deepening and changing current pattern. Assuming that deepening did occur the overlying coarsening upward sequence is probably a shallowing one. Since this sequence is thinnest in the northeast it is likely that higher energy conditions reached the northeast before the southwest. The relatively constant nature of the lower major sandstone suggests that more uniform conditions existed over the area than during the later part of Quartzite 2 times and the relatively small scale of the cross-sets shows that only small sand waves were present. At least in Sections 6 and 7 the current system had a 120^{o}-300^{o} orientation; that is perpendicular to the line of outcrop.

The upper sandstone unit of Quartzite 3 caps another widespread possibly shallowing sequence in which both bed thickness and grain size increase upward. The large, low angle cross-beds seen at the top of this sandstone in the main section (Pl. 99) are believed to be the lowest parts of even larger cross-beds which must have been 6-8m high before their upper part was eroded. Superficially these beds resemble those at outer margins of a steeply shelving beach where wedges of sand interfinger with finer-grained offshore deposits but the increasing angle of dip on this foresets reaching 15° makes such an origin unlikely. Thus this horizon is interpreted as the remnant of a large sand wave which migrated northwestwards. Since large sand waves usually only move during times of high current velocities the fine-grained intercalations at the base of these cross-sets were probably deposited during the intervening quieter periods when the lee of the sand wave provided a sheltered environment for such deposition. Assuming a straight crest for this sand wave a minimum water depth of about 40m can be estimated (Allen 1970c, p. 78 79). However, this value may be too high if the sand wave has a strongly curved crest

Quartzite 4

Main Section

Quartzite 4 follows these beds sharply and in its lower part consists of medium-bedded dark red sandstones which are flat-bedded or have small-scale low angle cross-beds. The sandstones are mainly of very fine sand grade although occasional granule conglomerates are present. Trace fossils are abundant and include large horizontal burrows, Rusophycus, Cruziana and Skolithos. This facies passes up through thin-bedded sandstones into the final major sandstone unit of the member. This consists of the typical cross-bedded sandstone facies and is distinctive in being red in its lower half and white above (Pl. 86).

Lateral variation

The lowest part of Quartzite 4 everywhere consists predominantly of dark red sandstones which are mostly fine-grained but in Section 7 there are also poorly sorted conglomerate sandstones with particles up to 2cm in diameter (Pl. 101)(Acc. series 19324). These coarse-grained beds are composed of various varieties of silica (rose quartz, polycrystalline metamorphic quartz, chert) and intraformational mudstone and sandstone pebbles. The large grains are sub-angular but this is largely due to intense pressure solution of the abundant chert grains and most particles were initially subrounded. These beds represent the "conglomerate" of Føyn (1937, p. 106) as his section, measured above Stappogiedde, corresponds to Section 7. Although somewhat similar coarse-grained horizons occur as far north as Breivik Valley they are both less common and finer-grained in these localities.

The remainder of Quartzite 4 seems to be easily correlatable between Sections 1, 2 and 5 as was also seen to be broadly similar at outcrops between Sections 6 and 7 although no measurements were made at these latter places.

Interpretation

The very fine grained red sandstones in the lower part of Quartzite 4 mark a return to predominantly lower energy conditions. However, the poorly sorted pebbly sandstones with their abundant intraformational pebbles were probably the result of storm conditions when material from a number of areas became mixed, rapidly transported, and then deposited before the sediment could be properly sorted. The major sandstone of Quartzite 4 represents a widespread return to higher energy conditions but a lack of data prevents any detailed interpretation.

Quartzite 5

This unit consists of thin-bedded sandstones, siltstones and mudstones capped by a thin horizon of cross-bedded sandstones. A similar sequence was seen in Sections 1, 2 and 5 but it was not established whether one continuous cross-bedded sandstone unit or a number of lenticular bodies are present.

The unit marks the last phase of high energy tidal conditions before the quieter conditions which prevailed during the lower part of the overlying Kistedal Formation.

In conclusion two points must be emphasised. Firstly, nowhere within this member is these any convincing evidence of marginal sediments such as beaches, barrier islands or lagoons. There are no examples of very low angle cross-bedding or any other features typical of beaches (Lane 1963). All the fine grained horizons which sharply overly cross-bedded sandstones contain trace fossils which are typical of fully marine conditions; so if lagoons were present they had a free circulation with the sea. Secondly, even within the framework of an offshore depositional environment for these beds the above discussion has only attempted to point out some of the most likely interpretations of the sedimentary history; the number of possible interpretations is almost limitless. Indeed since so much of what must have been deposited has not been preserved, large chapters of the sedimentary history are inevitably not recorded.

Palaeocurrent Data

Some of the data shown in Fig. 62 is rather unsatisfactory because of the low numbers of readings but the common bipolar distributions are particularly obvious. These form in two ways as is shown in Fig. 69. Type 1 is exemplified by the top of Quartzite 1 in Section 6 and Type 2 by the lower major sandstone of Quartzite 3 in Section 7. The bipolarity is mainly on a WNW-ESE axis with theWNW direction dominant. However, at other horizons there is unidirectional cross-bedding to the SW thus giving a trimodal pattern when all the data are combined (Fig. 68). Thus in contrast to the Upper Breivik Member and the Lower Duolbasgaissa Member most sediment transport was along a WNW-ESE axis and southwesterly transport was less common.

Trace Fossil Distribution

There is a great variety of trace fossils in this member and only the most prominent have been shown in Fig. ~~15~~ 61 and mentioned in the text. Three assemblages can be recognised and they are closely related to sedimentary facies and hence to environmental energy. In the siltstones and very fine sandstones there are large horizontal burrows. In the thin-bedded sandstones with fine-grained interbeds Ruspphycus predominates whilst Plagiogmus, large horizontal burrows and Rhizocorallium are occasionally seen; Phycodes is abundant in some beds and not others and its distribution is not understood. In the cross-bedded sandstones Syringomorpha and Skolithos linearis are the most common forms although neither is abundant. Rusophycus is occasionally present and near the tops of sand bodies vertical U-tubes are

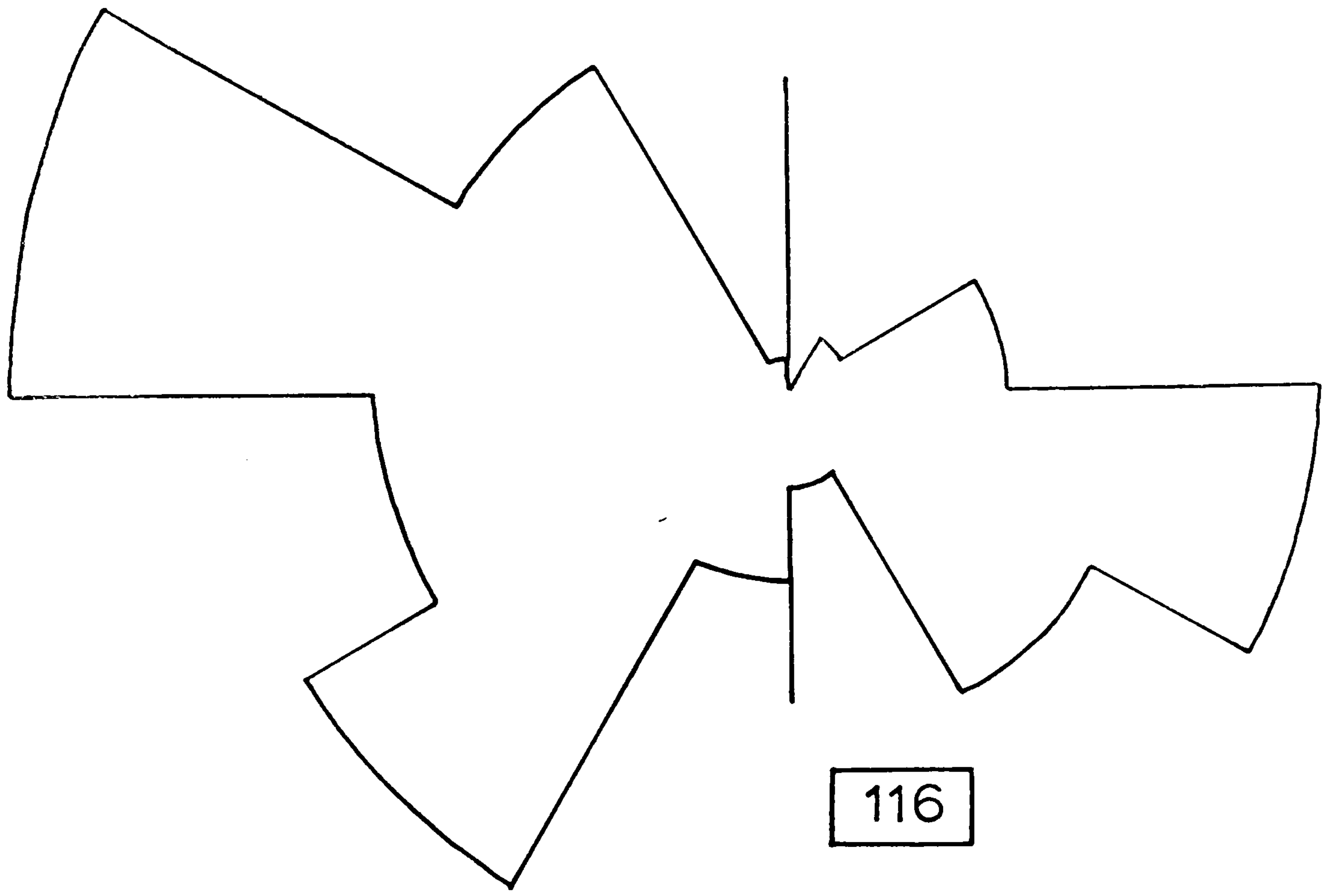


Figure 68. Total palaeocurrent data from cross-bedding in the Upper Duolbasgaissa Member.

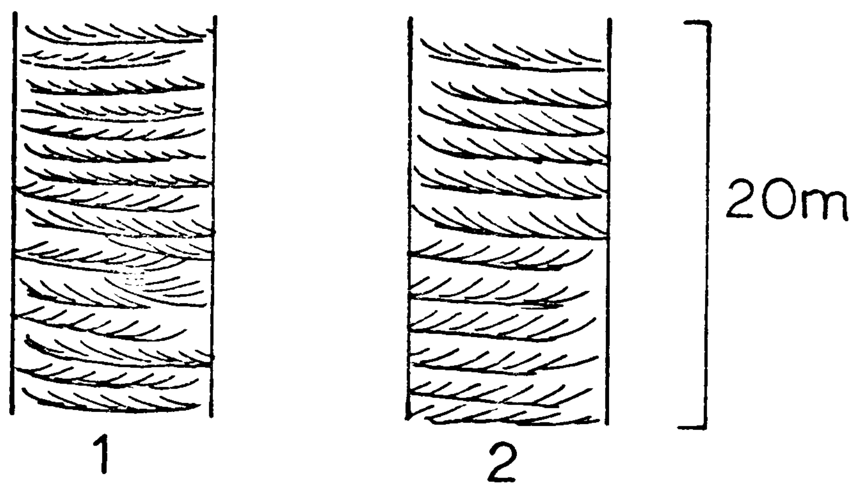


Figure 69. Two of the ways in which bipolar cross-bedding palaeocurrent patterns are developed in the Upper Duolbasgaissa Member.

common. Usually only the tops of these burrows are seen and often it is not possible to say if the burrows are Diplocraterion or Arenicolites but a few definite examples of both forms have however been seen. The first two assemblages are similar to those of the Lower Duolbasgaissa Member and fall within Seilacher's (1964, 1967) Cruziana facies whilst the last assemblage corresponds to Seilacher's Skolithos facies.

Seilacher, in a simple, depth-controlled model, attributed a littoral environment to beds of the Skolithos facies and a deeper, sub-littoral to wave base environment to beds of the Cruziana facies. However, Crimes (1970) has suggested that the trace fossil distribution is related more directly to environmental energy than to depth. Thus, whilst Seilacher's model is probably valid in situations in which environmental energy is closely inversely related to water depth, it falls down where the relationship is more complex as it probably is in most tidal seas and is inferred to be in this case. Therefore it is unwise to infer in this member that the Skolithos facies was always deposited in shallower water than the Cruziana facies even though in most cases it probably was. Also, the sedimentological evidence suggests a sub-littoral rather than littoral origin for beds of the Skolithos facies in this facies in this succession assuming that Seilacher uses "littoral" as synonomous with "inter-tidal".

Thus trace fossils do not provide an independent measure of water depth.

4, 5. SYNTHESIS OF THE UPPER BREIVIK? LOWER DUOLBASGAISSA AND UPPER DUOLBASGAISSA MEMBERS.

The succession has been explained as a shallowing sequence deposited on a tide dominated shelf. The conclusions of sections 4, 2-4, 4 and of this section are in Fig. 70. During Upper Breivik times and most of Lower Duolbasgaissa times the current system had a NE-SW orientation. However, during Upper Duolbasgaissa times a WNW-ESE system predominated and southwesterly currents were less important although still present.

Why should a tidal current system change its orientation through approximately 90° ? Johnson and Belderson (1969) suggested that alterations in basin physiography can change tidal current patterns in epicontinental seas. During shallowing the basin changed in depth and probably also in area and shape if we assume that there was regression of shorelines. Once the shallowing had continued for some time a critical time must have been reached at which the basin physiography favoured a switch of the currents through 90° . This time is marked by the base of the 20m Quartzite. A reversion to earlier conditions marked the remainder of the Lower Duolbasgaissa Member but in the Upper Duolbasgaissa Member the second orientation was favoured again.

Usually tidal currents flow either parallel or perpendicular to coastlines (Off 1963, Belderson and Stride 1966) and thus a 90° switch in direction may have been due to a change from one of these orientations to the other. It was concluded earlier that in the lower two members the sea was deeper in the southwest of the area than in the northeast and the slight southwesterly increase

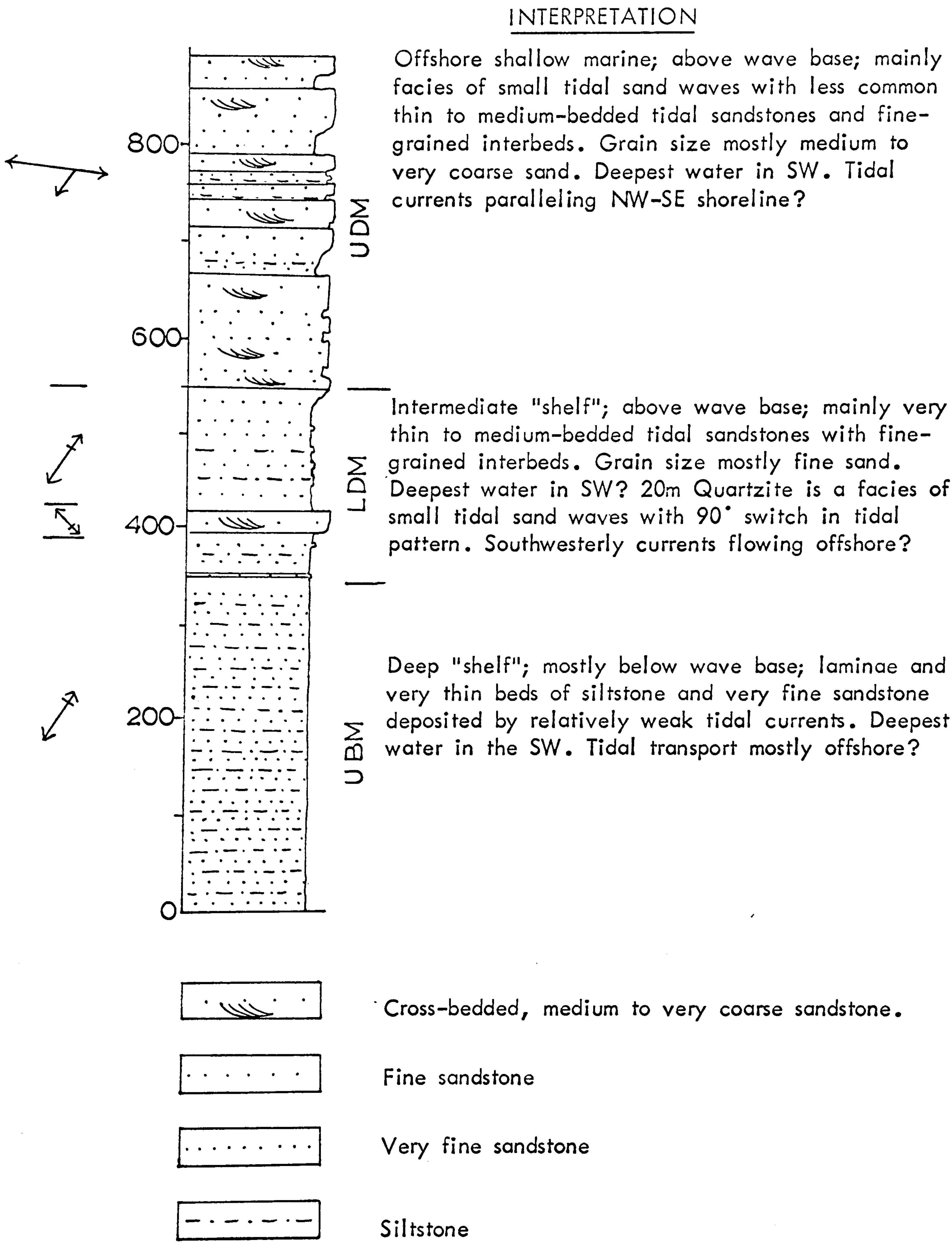


Figure 70. Interpretation of the Upper Breivik Member and Duolbasgaissa Formation.

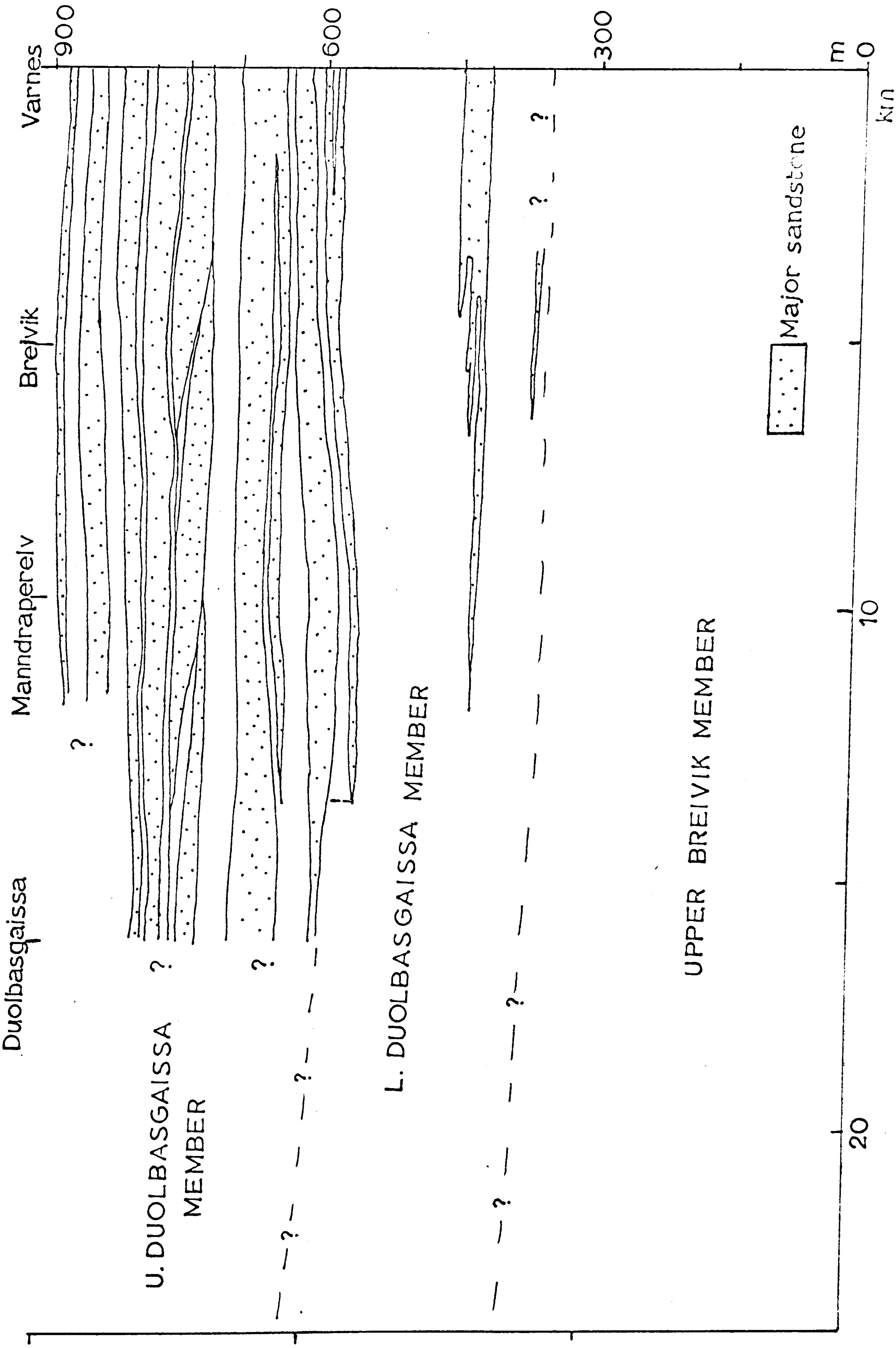
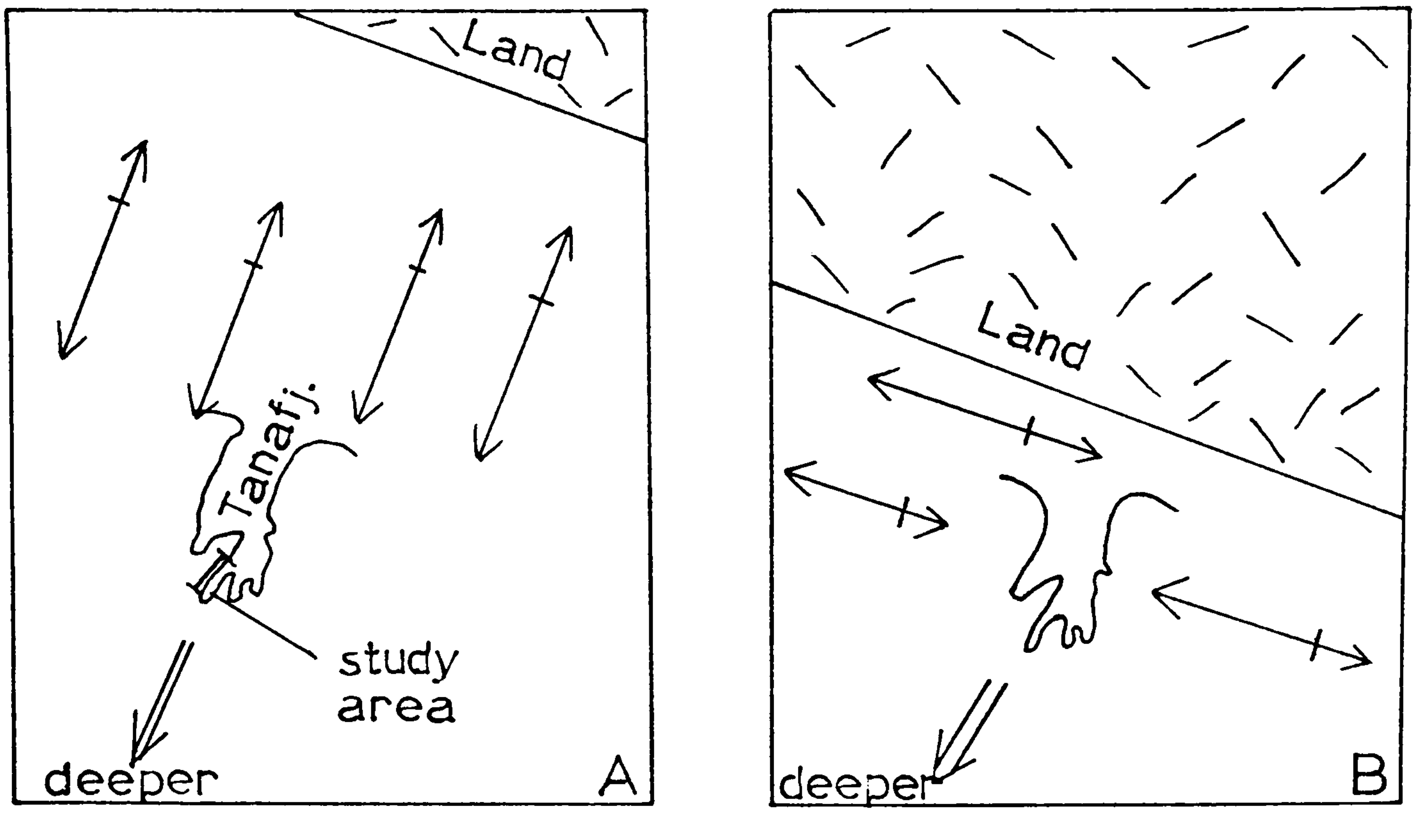


Figure 71. Cross section through the Upper Breivik Member and Duolbasgaissa Formation.

in thinner bedded sediment in the Upper Duolbasgaissa Member suggests that this situation may have existed throughout the sequence (Fig. 71). Thus it is more probably that the shoreline had a NW-SE orientation; it must have always lain to the NE of the study area and the sea deepened away from it towards the SW. Under this interpretation the NE-SW orientated system is on-shore-offshore with the offshore mode being dominant and the later WNW-ENE trend probably paralleled the shore. The model is summarised in Fig. 72.

The reasoning behind this model is inevitably biased because the study is only 2-dimensional and the study area may well be very small in comparison with the size of the basin. However, hopefully the model has some connection with the truth and in the next section it will be further developed by considering the regional geological setting of the sediments.



DOUBLE HEADED ARROWS SHOW TIDAL CURRENT PATTERN.

Fig. 72. Tentative palaeogeographies and tidal current patterns for the upper part of the succession.

A. Upper Breivik and most of Lower Duolbasgaissa times.

B. Most of Upper Duolbasgaissa time.

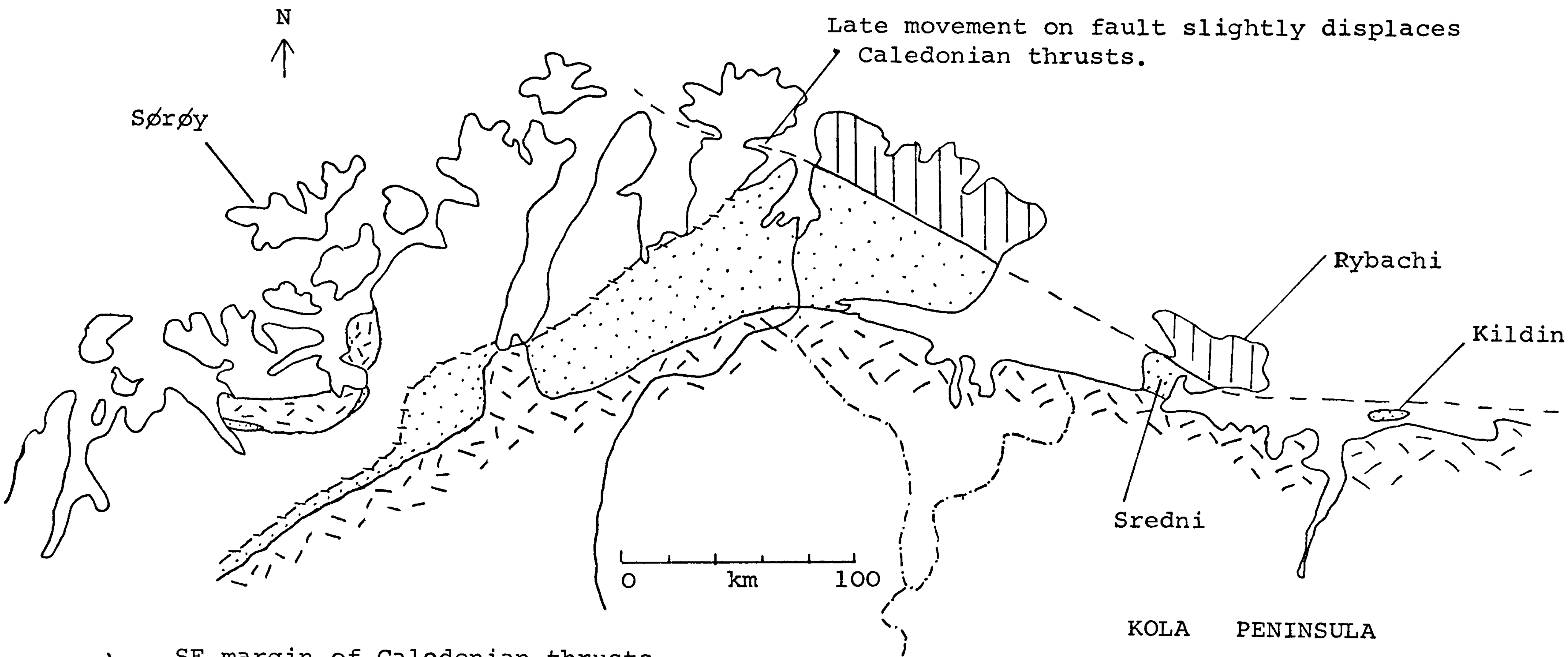
CHAPTER 5

PALAEOGEOGRAPHICAL AND PALAEOGEOLOGICAL SPECULATIONS

Regional pattern of sedimentation in Finnmark and adjacent areas

From the areas in Finnmark where autochthonous or parautochthonous latest Precambrian and Lower Cambrian rocks are now exposed the thickest accumulation is in the Tanafjord area and progressive thinning occurs to the southwest, towards Kunes and Halkkavarre. There is also limited evidence in the upper part of the Manndraperelv Member of thinning to the east of Tanafjord. Beyond Halkkavarre the Dividal Group is of almost constant thickness and can be traced along the edge of the Caledonian metamorphic belt into southern Scandinavia (Føyn 1967). The Dividal Group thus forms a thin sequence of latest Precambrian and Cambrian rocks lapping over the Baltic Shield. It is difficult to know how far this cover originally extended. Although there is no direct evidence it is probable that the Tanafjord succession thinned to a Dividal Group thickness to the south of Tanafjord where rocks of this age are no longer present but where the underlying Nyborg Formation and tillites thin and disappear.

Late Precambrian and Cambrian sediments are known on Sørøy (Fig. 73) where archaeocyathids of upper Lower Cambrian or Middle Cambrian age occur in the upper part of the Klubben Quartzite Group (Pringle and Sturt 1969). The Klubben Quartzite Group contains herringbone cross-bedding and is believed to be of shallow marine origin (D. Roberts 1968a). It is 1800m thick and could be equivalent to the Stappogiedde, Breivik, and Duolbasgaissa Formations. It is overlain by further shallow marine beds (110-135m) which pass up into a turbidite formation, the



SE margin of Caledonian thrusts.

"Older Sandstone Series", Vestertana, Digermul, and Dividal Groups and lateral equivalents.

Precambrian crystalline rocks.

Barents Sea Group, Raggo Group and probable lateral equivalents.

International frontier.

Figure 73. Outline geological map of Finnmark and part of the Kola Peninsula.

Hellefjord Schist Group. which is 700-900m thick (D. Roberts 1968b). However since these beds sit within the Caledonian metamorphic belt they have probably been moved very considerably from their site of deposition and their relationship to the nearby Dividal Group is uncertain.

Since the East Finnmark succession is thickest at the north end of the Digermul Peninsula it is reasonable to assume that it must have continued some way to the north of its present outcrop, that is north of the line of the Trollfjord-Komagelv fault. The history of this fault is poorly known but assuming that the Vestertana and Digermul Groups and the "Older Sandstone Series" were deposited to the north of it there must have been considerable uplift (4-5km?) on its northern side. This uplift occurred before the Caledonian thrusting since the thrust blocks are only slightly displaced along the line of continuation of the fault (Føyn 1969). This displacement was probably a small later movement.

To the east it is possible that rocks of equivalent age occur on Sredni and Kildin (Fig. 73) but the age and the exact succession of the rocks in these areas is in doubt (Seidlecka and Seidlecki 1967).

Palaeogeographical evidence from lateral facies changes and palaeocurrents

No truly consistent pattern emerges either from the palaeocurrent data or the lateral facies variations. At the base of the Innerelv Member a shoreline was retreating southwards across the area. In the Innerelv Member currents flowed from both the west and southeast; in the Manndraperelv Member there is reasonable evidence of a shoreline in the northeast of the area and of a possibly arcuate shelf prograding to the west. The Lower Breivik Member provides no information but in the Upper Breivik Member and Duolbasgiassa Formation the shoreline is most likely to be somewhere to the northeast. Thus apart from the base of the succession, on the two occasions when there is any evidence of a shoreline it was probably to the northeast. However, this does not necessarily imply that sediment was derived from a northeasterly land mass as will be shown below.

Another palaeogeographical feature which can be inferred is that when tidal currents were strong the sea was probably connected to a major ocean. Without such a connection tidal currents in small seas are always weak. Thus during the Breivik and Duolbasgáissa Formations there was a connection to an ocean but in the Innerelv and Manndraperelv Members this connection may have been absent or more restricted.

Petrographical evidence as to the nature of the source area

Since the sediments are largely of shallow marine origin and were deposited in a tectonically stable environment the amount of feldspar in them suggests that the source area was composed largely of igneous or high grade metamorphic rocks (see Appendix D for all details). The presence of metamorphic polycrystalline quartz shows that metamorphic rocks formed some part of the source area but the absence of any distinctive "metamorphic" heavy minerals suggests that it consisted mainly of igneous rocks. The types of feldspar suggest that these igneous rocks were acidic. The chert grains show the presence of some sedimentary or possibly low grade meta-sedimentary rocks in the source area and the mature heavy mineral suite fits a sedimentary source slightly better than an igneous one (Pettijohn 1957).

It is concluded that the source area of the sediments consisted mainly of acid igneous rocks but with subsidiary metamorphic and sedimentary (or low grade metasedimentary) rocks.

Possible location of the source area

Assuming that the source area lay somewhere to the north or east of the Tanafjord area Figure 73 shows that this region is divided geologically into two parts by the Trolljord-Komagelv fault and its probably continuation along the northern coastline of the USSR (Seidlecka and Seidlecki 1967). From geophysical work this present coastline of the USSR in the Kola area coincides with a major change in the Earth's crust (Demenitskaya and others 1968, Beliayevsky and others 1968) (Fig. 74). To the south the crust consists of a Precambrian crystalline complex the Baltic or Fenno-Scandian Shield. However this thins abruptly to the north and the crust under the Barents Sea is composed of a pile of sediments 15-20km thick sitting on a thin "basaltic" layer. A large part of these sediments are believed to be of Upper Precambrian age and the thick sequences of the Barents Sea Group and on Rybachi probably represent the upfaulted margins of this sedimentary sequence. The Trollfjord-Komagelv fault is thus the surface expression of a major tectonic feature and must have a long history dating back into the Precambrian. Thus it can be deduced that in latest Precambrian and Lower Cambrian times any land area to the north of this line would have been composed of sedimentary rocks and any to the south would have been largely crystalline.

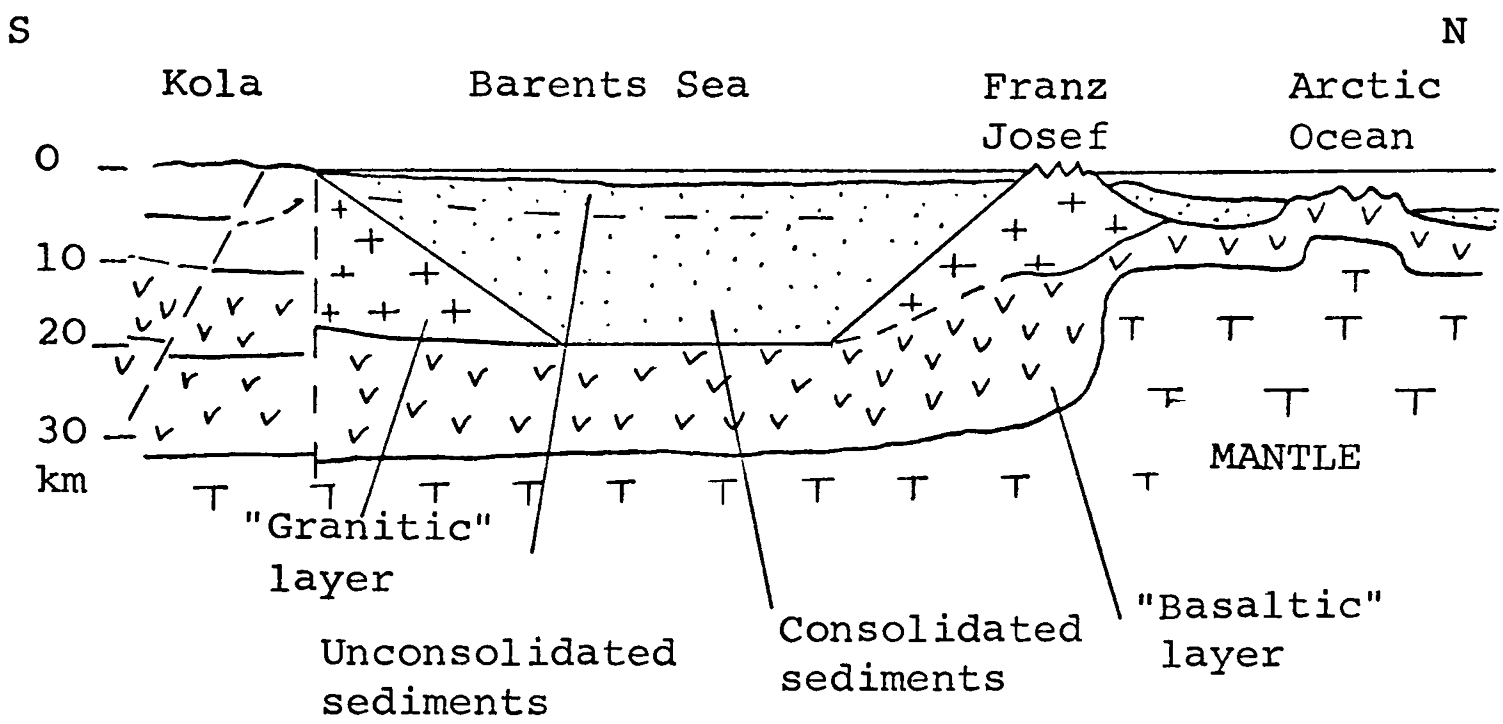
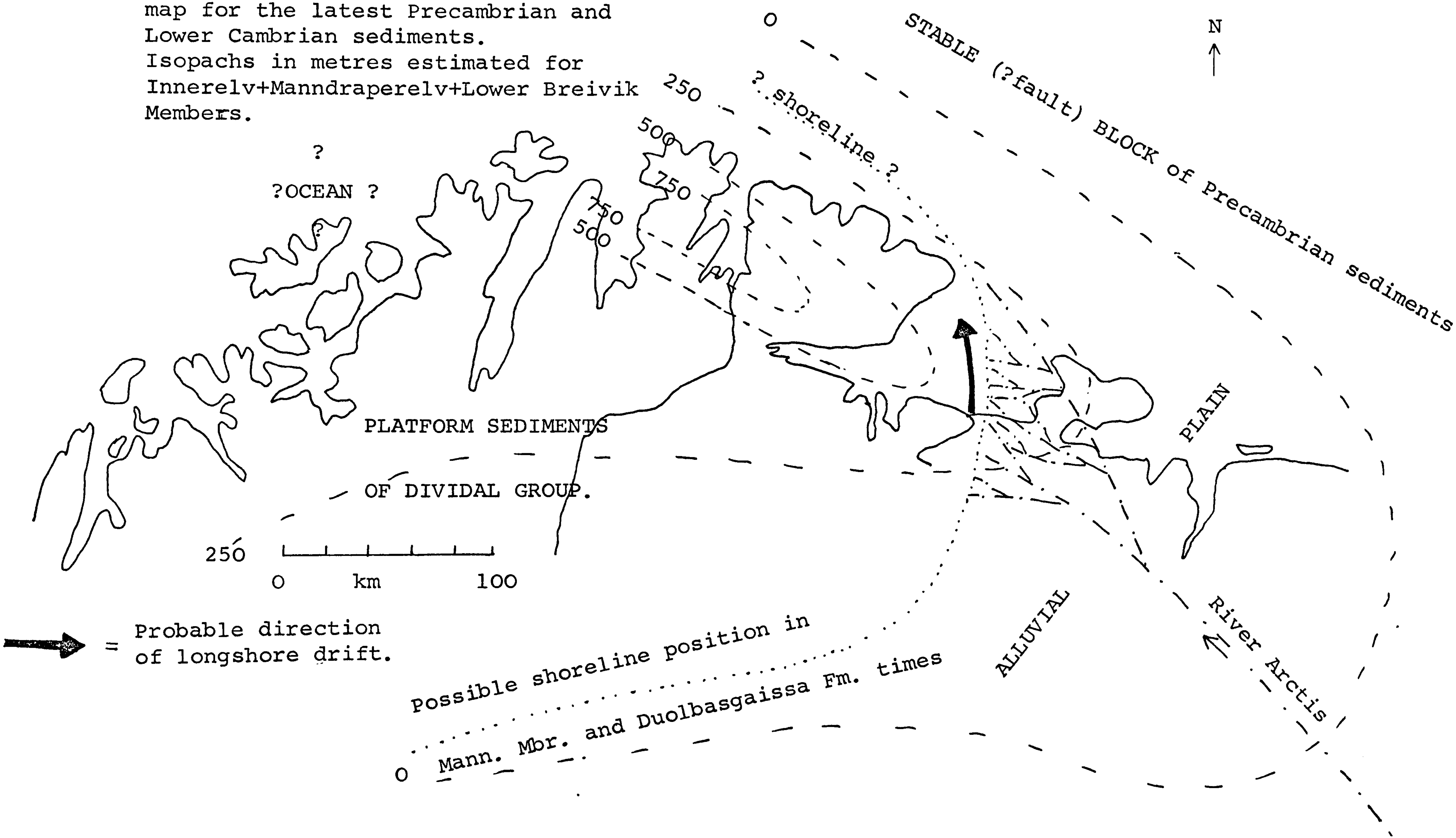


Figure 74. Simplified crustal section from the Kola Peninsula to the Arctic Ocean (after Demenitskaya and others 1968).

Figure 75. Very tentative palaeogeography and isopach map for the latest Precambrian and Lower Cambrian sediments.
Isopachs in metres estimated for Innerelv+Manndraperelv+Lower Breivik Members.



Therefore the source lands of the sediments lay to the south of the eastward extension of the Trollfjord-Komagelv fault in the region of the Kola Peninsula and Karelia. At the present day many varied Precambrian rock types are present in this area (Sederholm 1930, 1932). Granite gneisses and granites predominate but there are also patches of metasedimentary and metavolcanic complexes which are generally of younger Precambrian age and could have been more extensive in latest Precambrian and Cambrian times. Iron ore bands are common in places which could account for the wealth of opaque heavy minerals in the sediments.

Thus although there is no really convincing connecting evidence the geology of the Kola-Karelia area is compatible with its having been the source area. It is concluded that the land to the northeast provided little sediment and may have been no more than a stable non-subsiding block. A very tentative palaeogeographical reconstruction is sketched in Fig. 75.

CHAPTER 6

CONCLUSIONS OF GENERAL SIGNIFICANCE

1.

The main types of current capable of transporting sand in the shallow marine (shelf) environment are; - i) Semipermanent currents, ii) Tidal currents, iii) Wave drift currents, iv) Coastal storm surge currents, v) River generated currents. The movement of sand is often the result of the interaction of several of these currents but one usually predominates.

2.

From the known and inferred characteristics and products of these currents in modern seas the data from ancient sandstones can be evaluated to determine which type of current was most important in depositing the sandstones. The late Precambrian to Lower Cambrian succession in East Finnmark has been interpreted using this approach.

3.

This succession shows many examples of sharp based sandstones whose sole marks and internal structures suggest that they were deposited by waning currents. However, only in the case of certain beds in the Manndraperelv Member is it suggested that these beds are turbidites. In other cases these waning currents are of one of the shallow marine types outlined above.

4.

A transgressive fluvial to marine sequence with preserved inter-tidal deposits is found at the Lillevatn Member - Innerelv Member transition on the Digermul Peninsula.

5.

Coastal storm surge deposits are possibly present in the Innerelv Member.

6.

Two small scale (< 90m) sequences showing turbidite to shallow water transitions in the Manndraperelv Member differ from those described by Walker (1969) and represent a new type. The turbidity currents were generated by stirrring of sediment on the shelf rather than by a river generated process. The turbidites infilled a basin which formed by differential tectonic subsidence when subsidence greatly outweighed. sediment supply.

7.

Beach, lagoon and transgressive shelf sand facies are described in the First Coarsening Upward Sequence of the Manndraperelv Member.

8.

Two suites of offshore tidally dominated sediments are found in the Lower Breivik Member and the Upper Breivik Member plus Duolbasgaissa Formation. In both cases sediment transport was probably at a maximum when tidal currents. were combined with storms.

9.

Resonance of the basin is believed to have been particularly important in the amplification of tidal currents in the Lower Breivik Member.

10.

By comparison with the recent sediments in the seas around Great Britain the various facies of the Upper Breivik Member and Duolbasgaissa Formation are believed to have been formed at different positions along tidal current transport paths.

APPENDIX A

Nomenclature for Description of Sediments

The nomenclature for petrographic descriptions is based broadly on the proposals of Folk (1968) except that the mineralogical triangle and compositional rock names are those of McBride (1963).

Under the Folk scheme rocks are named as follows:- grain size: prominent orthochemical cements, textural maturity, notable or unusual transported constituents, main rock name.
e.g. fine sandstone: siliceous, supermature, glauconite-bearing quartz-arenite.

Because of the considerable diagenetic modification which has occurred in most of the sediments Folk's textural groups are too detailed to be applicable and a simplified system has been devised (Figure). Estimates of textural maturity were based on a visual comparison of thin sections with the sorting and rounding diagrams of Folk (1968, p. 103, 104).

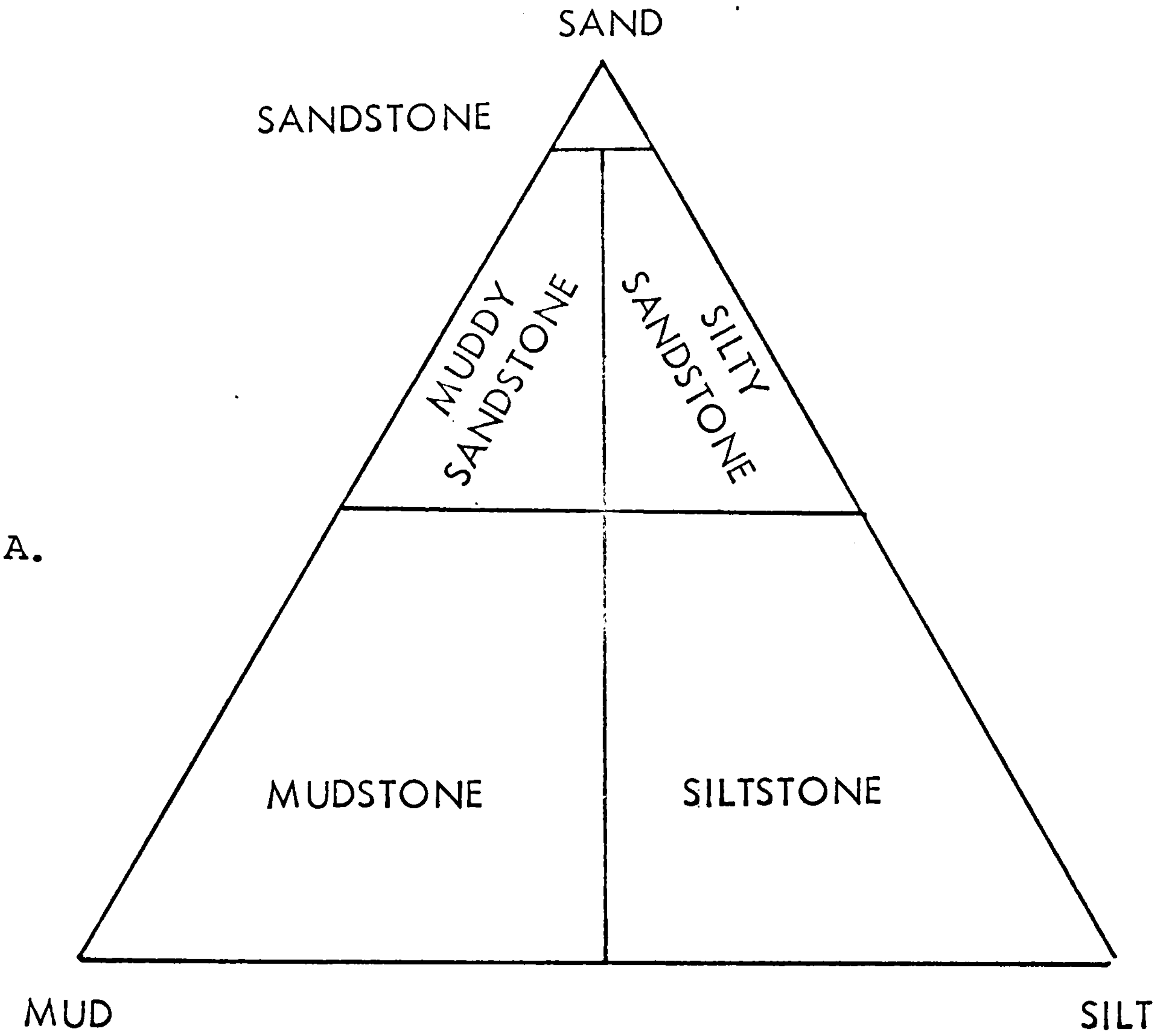
Bed thickness classification (after Ingram(1954))

0-0.3cm	Thinly laminated
0.3-1.0cm	Thickly laminated
1-3cm	Very thin-bedded
3-10cm	Thin-bedded
10-30cm	Medium bedded
30-100cm	Thick-bedded
100- cm	Very thick-bedded

Lateral persistence of beds

	length : thickness ratio
Lenticular, laterally impersistent	<50:1
Moderately laterally persistent	50:1 - 500:1
Laterally persistent	>500:1

A.



B.

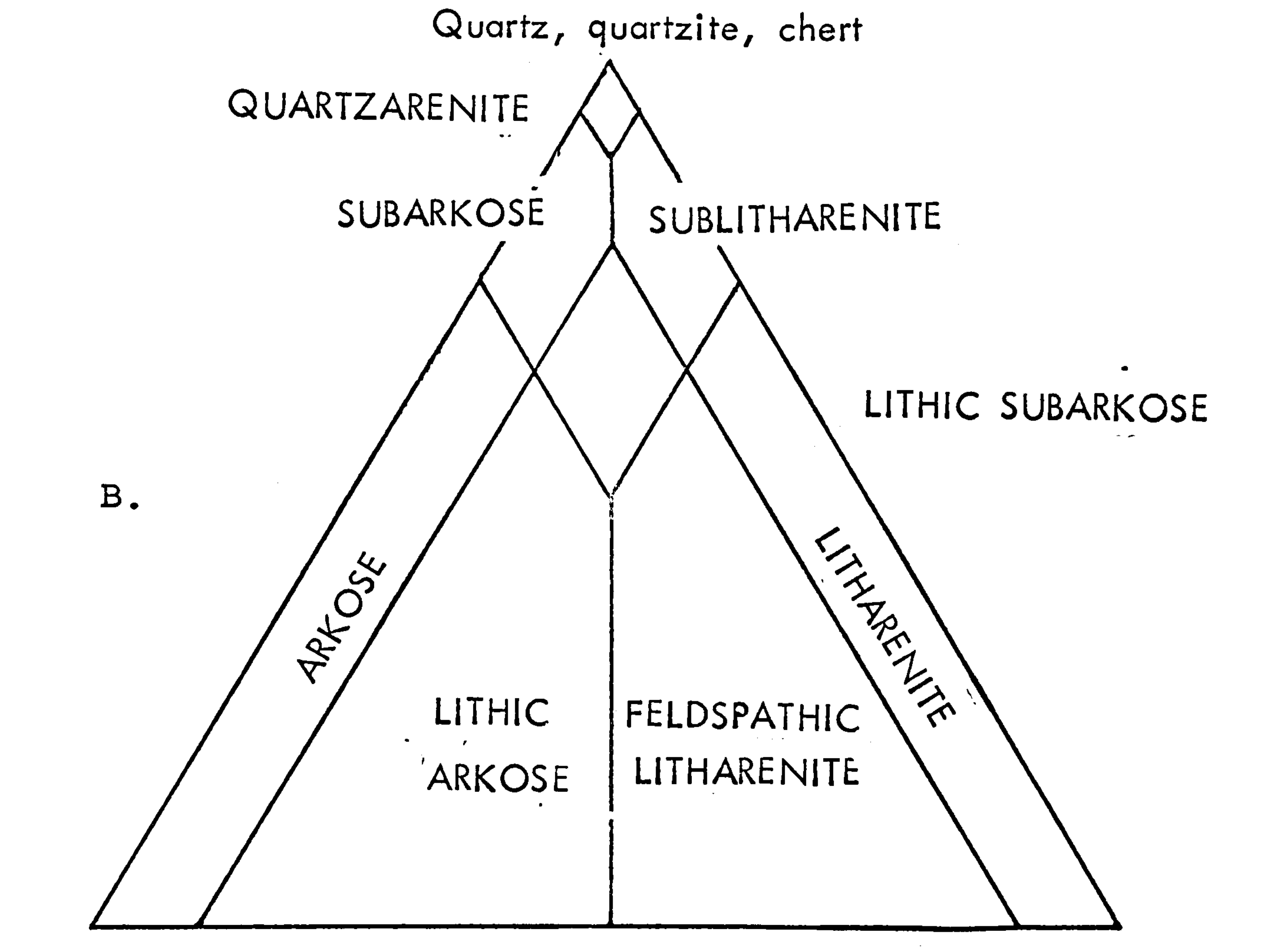


Fig. . A. Textural classification of sediments modified from Folk (1968).
B. Mineralogical classification after McBride (1963).

APPENDIX B

MAPS OF LOCALITIES

KEY TO LOCALITIES

1. "Areholmen" section; Lillevatn Member - Innerelv Member.
2. Main section, Innerelv Member.
3. Main section, Lower Sandstone, Manndraperelv Member.
4. Main section, First and Second Coarsening Upward Sequences, Manndraperelv Member.
5. Main section, Lower Breivik Member.
6. Section 0-53m, Upper Breivik Member.
7. Main section, Upper Breivik Member.
8. Main section, Lower Duolbasgaissa Member.
9. Main section, Upper Duolbasgaissa Member.
10. Larsholmen section, Innerelv Member.
11. Larsholmen section, Manndraperelv Member.
12. Innerelv section, Lower Breivik Member.
13. 20 m Quartzite sections as shown on Fig. 59. 1.
14. 2.
15. 4.
16. 5.
17. 6.
18. 7.
19. 8.
20. Lower Duolbasgaissa Member and Upper Breivik Member on south side of Hill 551
21. Upper Duolbasgaissa Member sections of Fig. 62. 1
22. 4
23. 5
24. 6
25. 7
26. Upper Duolbasgaissa Member section on south side of Hill 551.

GEOLOGICAL MAP of the southeastern part of the Digermul Peninsula (after Reading 1965). Many minor faults not shown.

APPENDIX B

MAPS OF LOCALITIES

MEASURED SECTIONS IN THE LEIRPOLLEN AREA

MAP 1

1. First Coarsening Upward Sequence; eastern tributary of Stygdalen.

2. Secend Coarsening Upward Sequence; eastern tributary of Stygdalen.

3. First Coarsening Upward Sequence; western tributary of Vaderelv with waterfall near confluence.

4. Lower Sandstone, First and Second Coarsening Upward Sequences; Vaderelv main valley.

5. Base of Lower Breivik Member section; Vaderelv main valley.

6. Position where Føyn found Platysolenites in Lower Breivik Member.

MAP 2

7. First Coarsening Upward Sequence; stream north of Lievlamfjeldet with good exposure by waterfall.

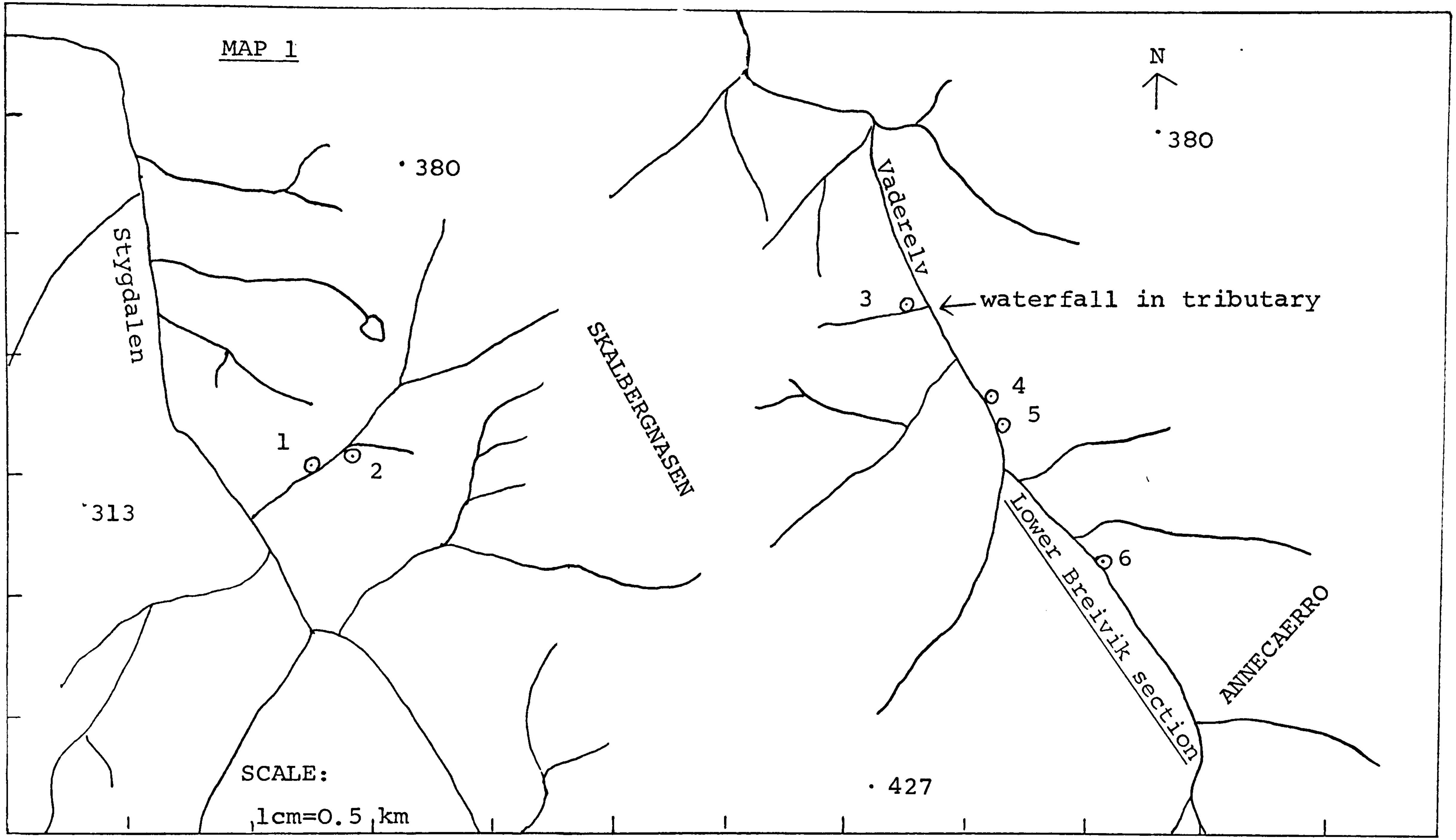
8. Upper part of First Coarsening Upward Sequence.

9. Lower part of Lower Sandstone; fault gulley at top of escarpment overlooking Tana River.

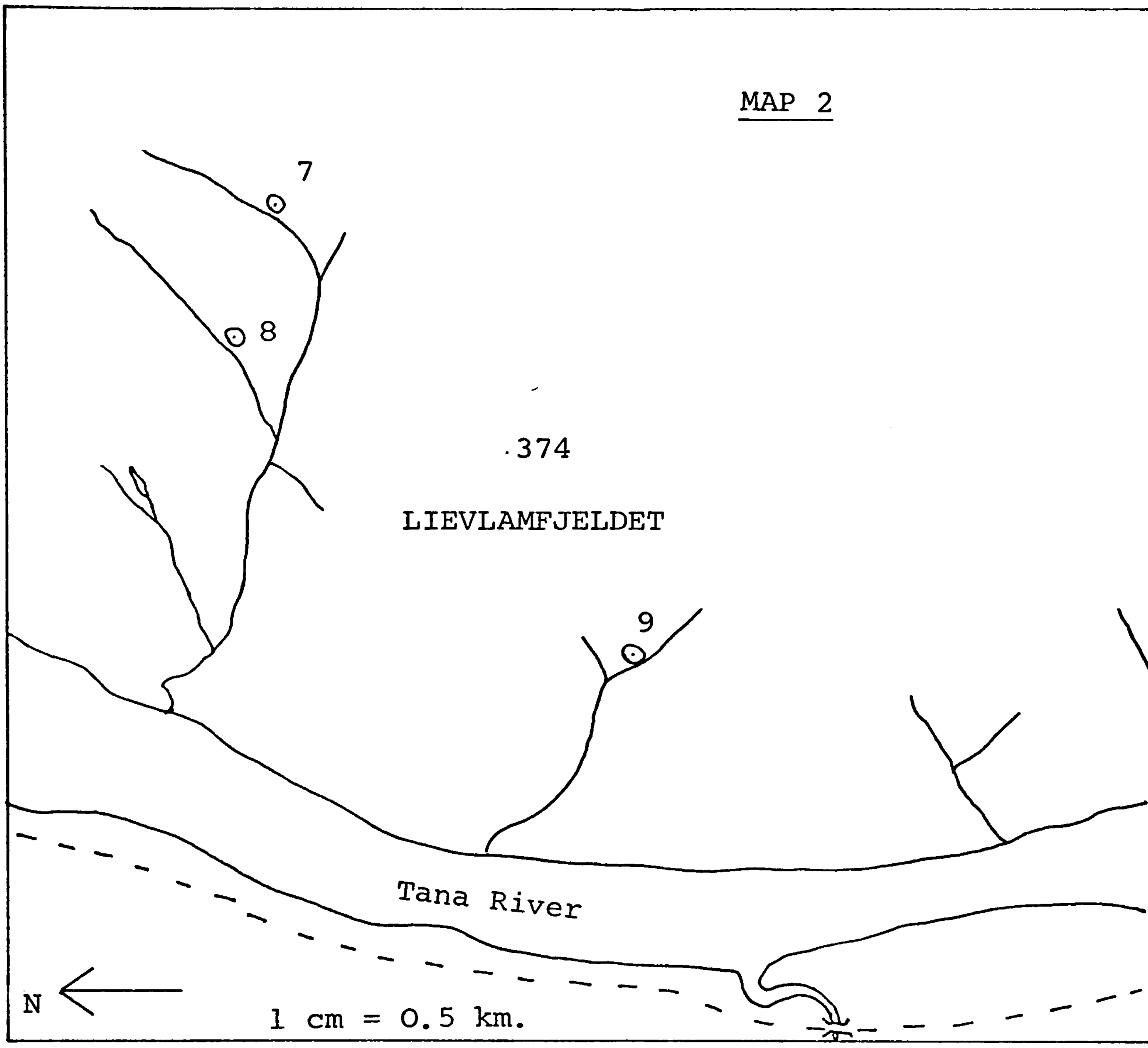
10. Innerelv Member exposed near the small stream and on steep hillside.

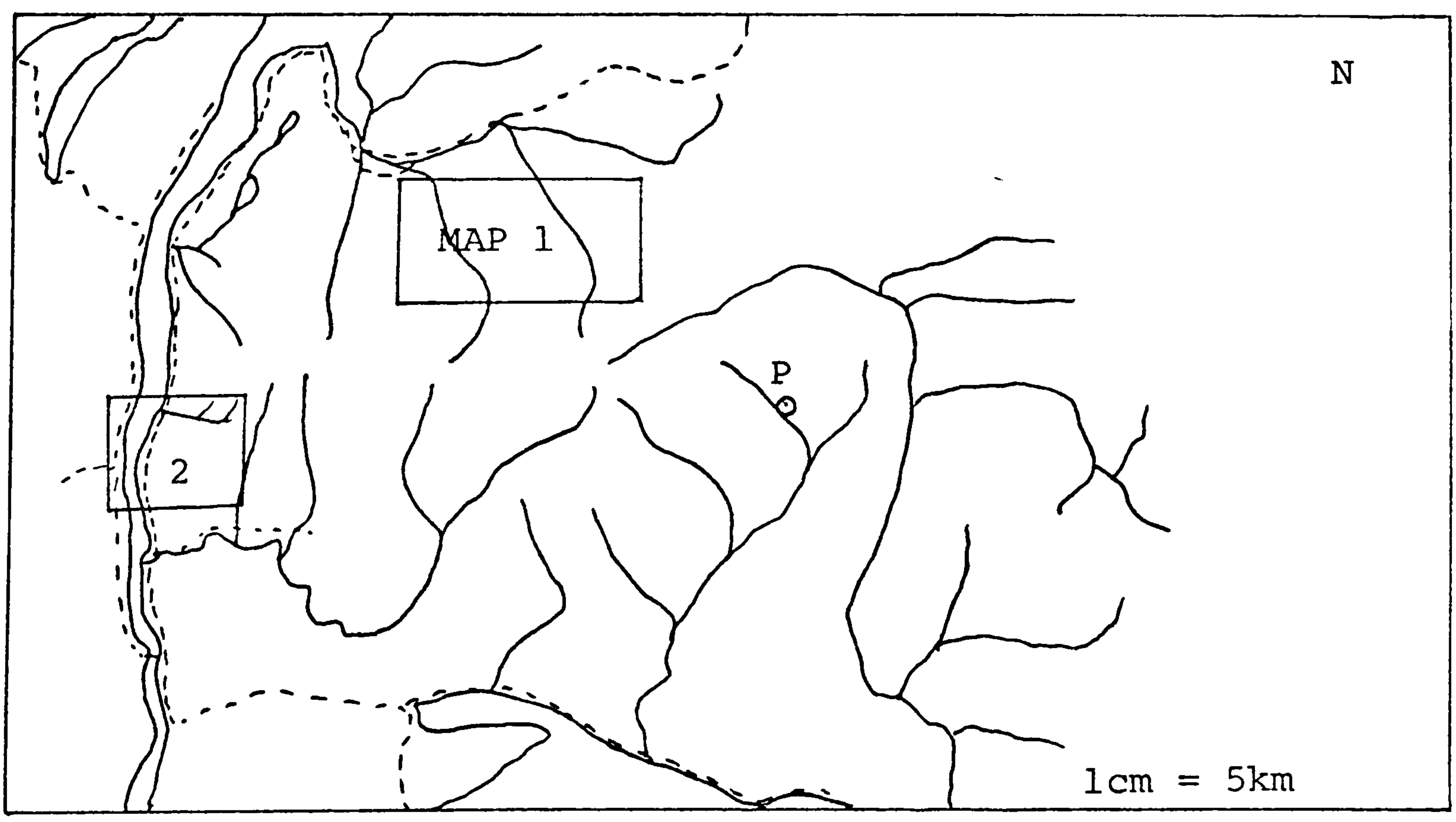
General map

P. Lower Sandstone, First and Second Coarsening Upward Sequences. Perledalen, a tributary of the Bergeby.

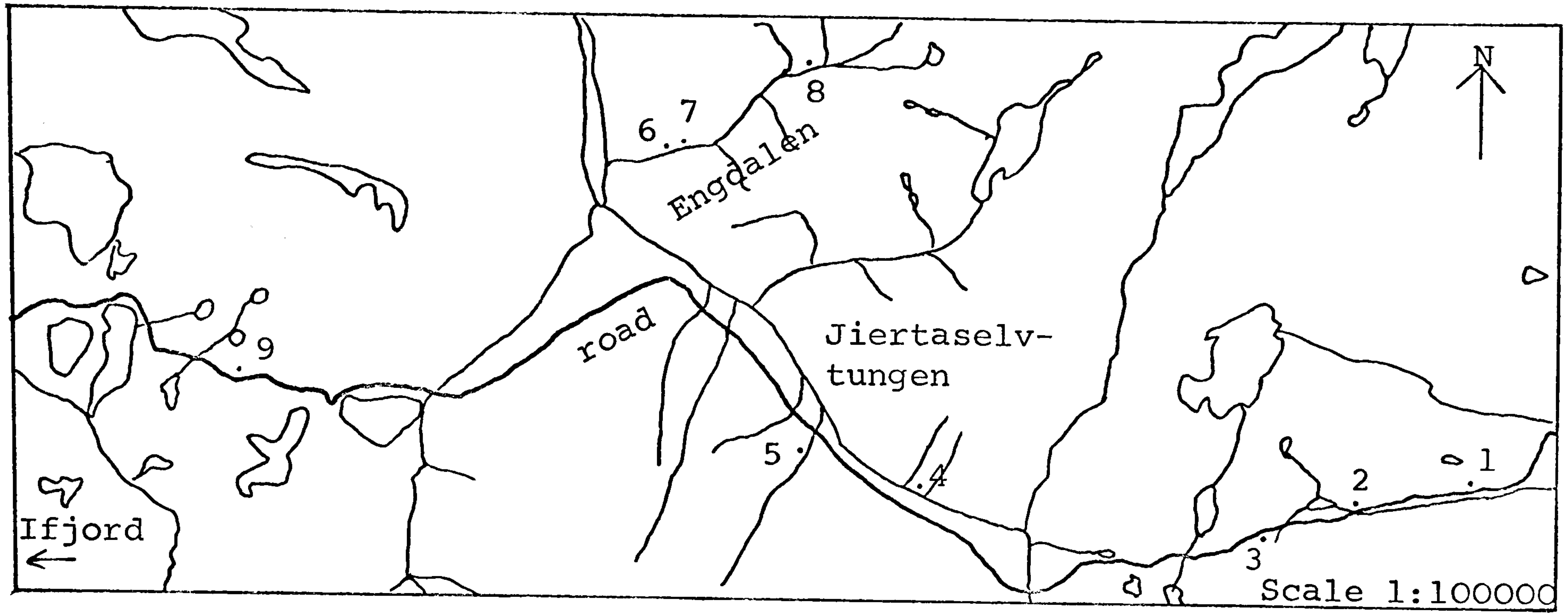


LOCALITY MAP FOR THE EASTERN PART OF THE LEIRPOLLEN AREA SHOWING LOCALITIES OF MEASURED SECTIONS.



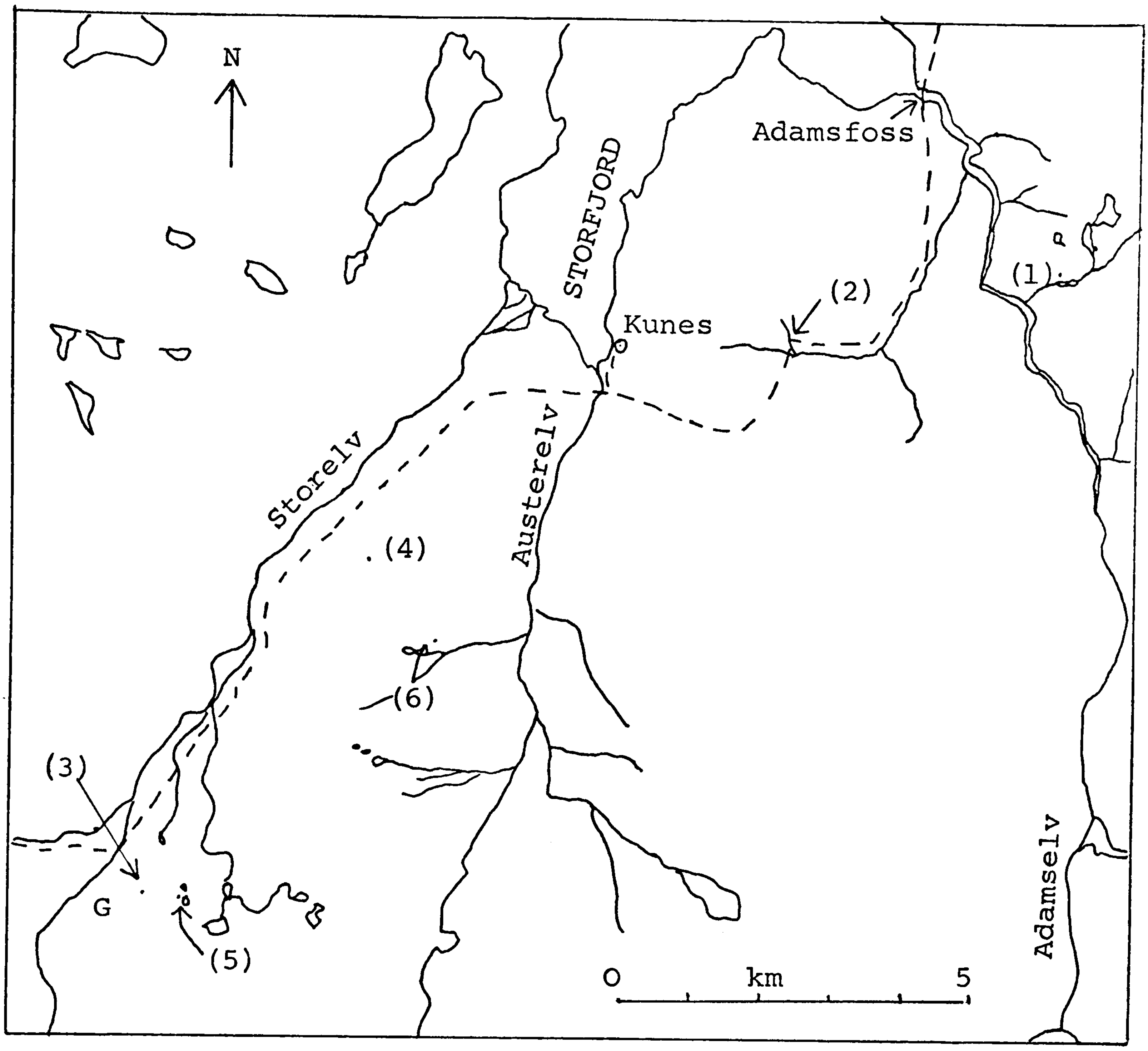


Location map for measured sections in the Leirpollen area showing the positions of detailed maps 1 and 2. P = Perle-dalen which is not on maps 1 or 2.



1. Exposure of Lillevatn Mbr - Innerelv Mbr junction (Vestertana section).
2,3. Exposures of Innerelv Mbr (Vestertana section).
4,5,6, Localities of measured sections of the Manndraperelv Member (Jiertaselvtungen).
7. Exposure of Lower Breivik Mbr.
8. Exposure of the top of the Lower Breivik Mbr.
9. Isklövervandene section (Manndraperelv Mbr).

LOCALITY MAP OF THE AREA BETWEEN IFJORD AND VESTERTANA



LOCALITY MAP FOR THE AREA AT THE HEAD OF LAKSEFJORD

(1) "Adamsfoss" locality for sections in Innerelv, Manndraperelv and Lower Breivik Members

(2) Stream section where Føyn (1967) found Platysolenites

(3) Kunes (south) section 1; in gulley on east side of Guorggaabmer (G), a metamorphic klippe.

(4) Kunes (north) section in Mannraperelv and Lower Breivik Members. Section is at top of escarpment and is clearly seen from the road.

(5) Kunes (south) section 2; section measured near two small lakes about 500m east of Locality (3)

(6) Kunes section in Innerelv Member; measured in small fault gulley just north of the stream named Fierramelva (Fieranjoga)

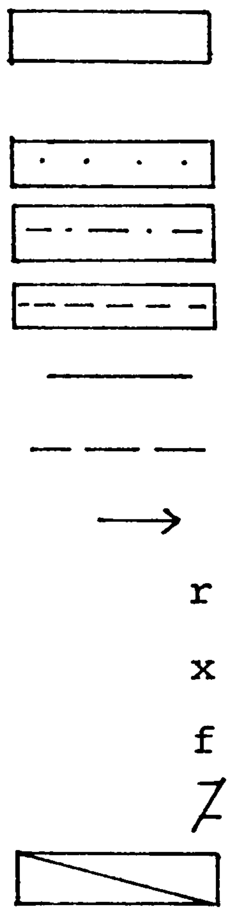
The best geological map of this area is given by Føyn (1967)

In 1970 a road was built from the main road running just south of locality (1) to give access to a hydro-electric station but as yet no bridge has been constructed across the Adamselv (or is planned ?) and so only heavy vehicles can get across. Unless this road can be used one must scramble along the east bank of the Adamselv or perhaps use a boat.

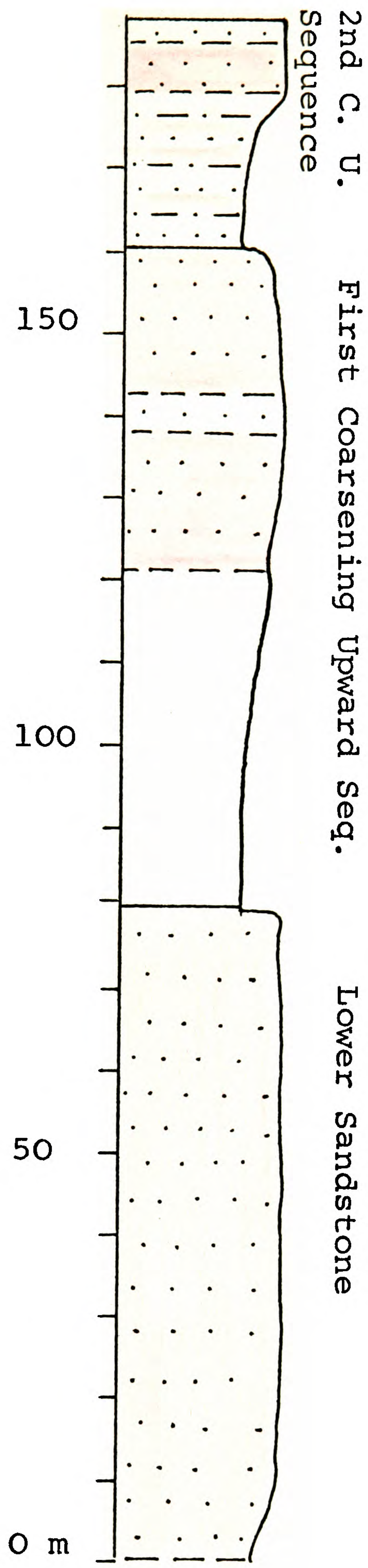
APPENDIX C

FURTHER MEASURED SECTIONS IN THE MANNDRAPERELV AND LOWER BREIVIK MEMBERS

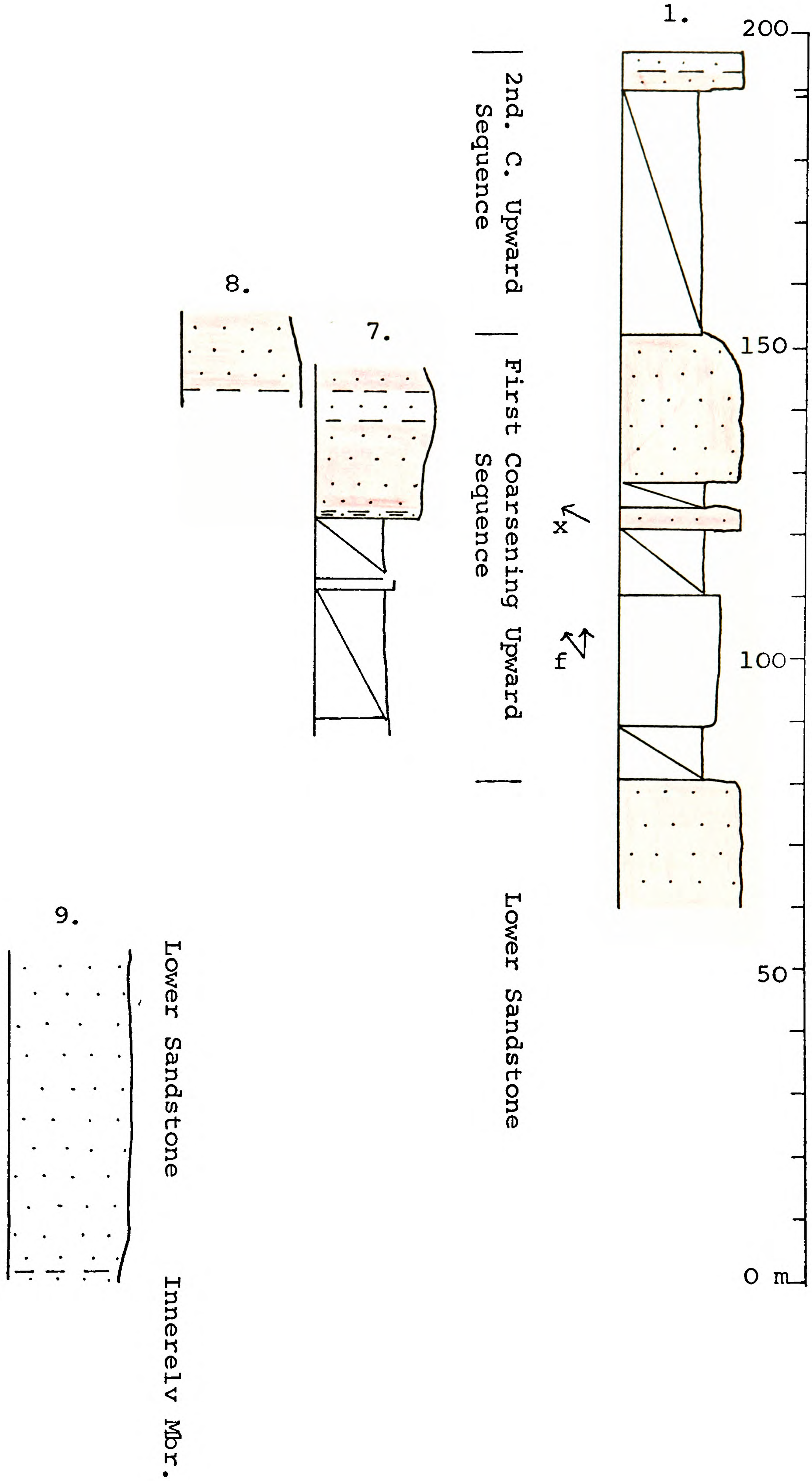
KEY for MANNDRAPERELV MEMBER sections.



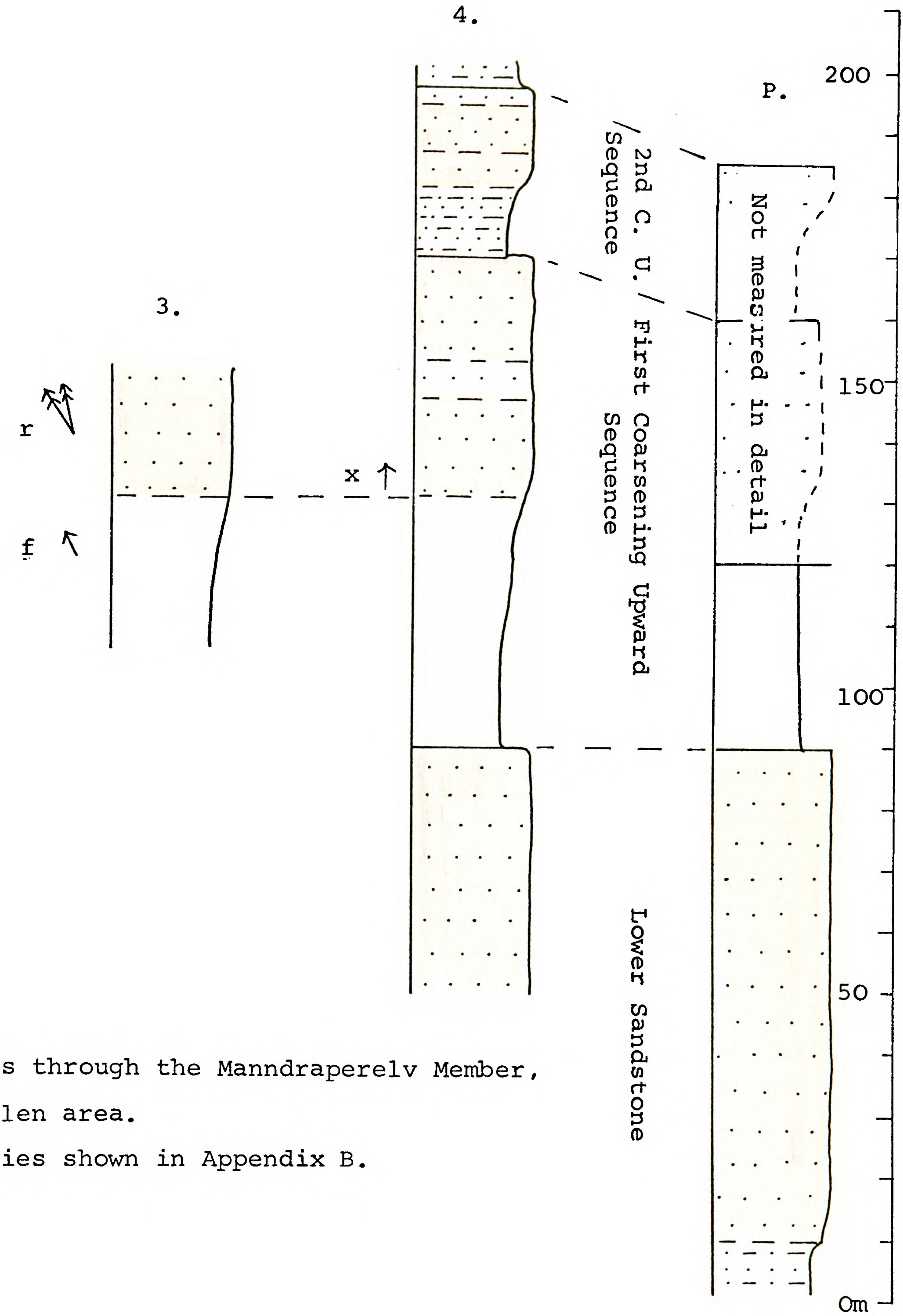
	Green mudstone and siltstone with interbedded graded sandstones.
	Sandstone
	Siltstone
	Mudstone
	Sharp contact
	Gradational contact
	Palaeocurrent direction
r	Asymmetrical ripple
x	Cross-lamination
f	Flute mark
	Fault
	No exposure



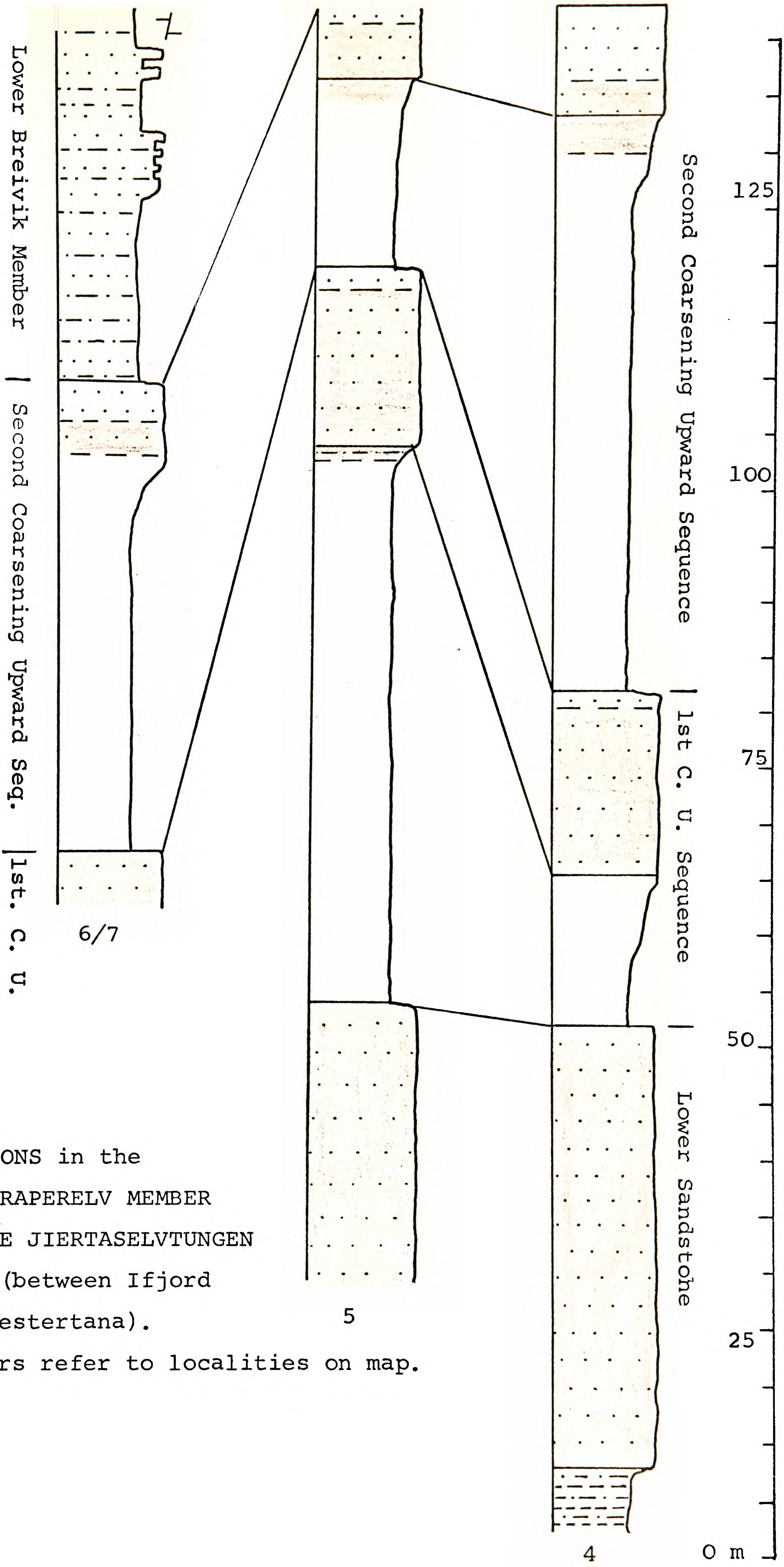
Composite section through the Manndraperelv Member in the Leirpollen area based on sections P, 1,3, 4, 7, 8, 9. Localities shown in Appendix B.



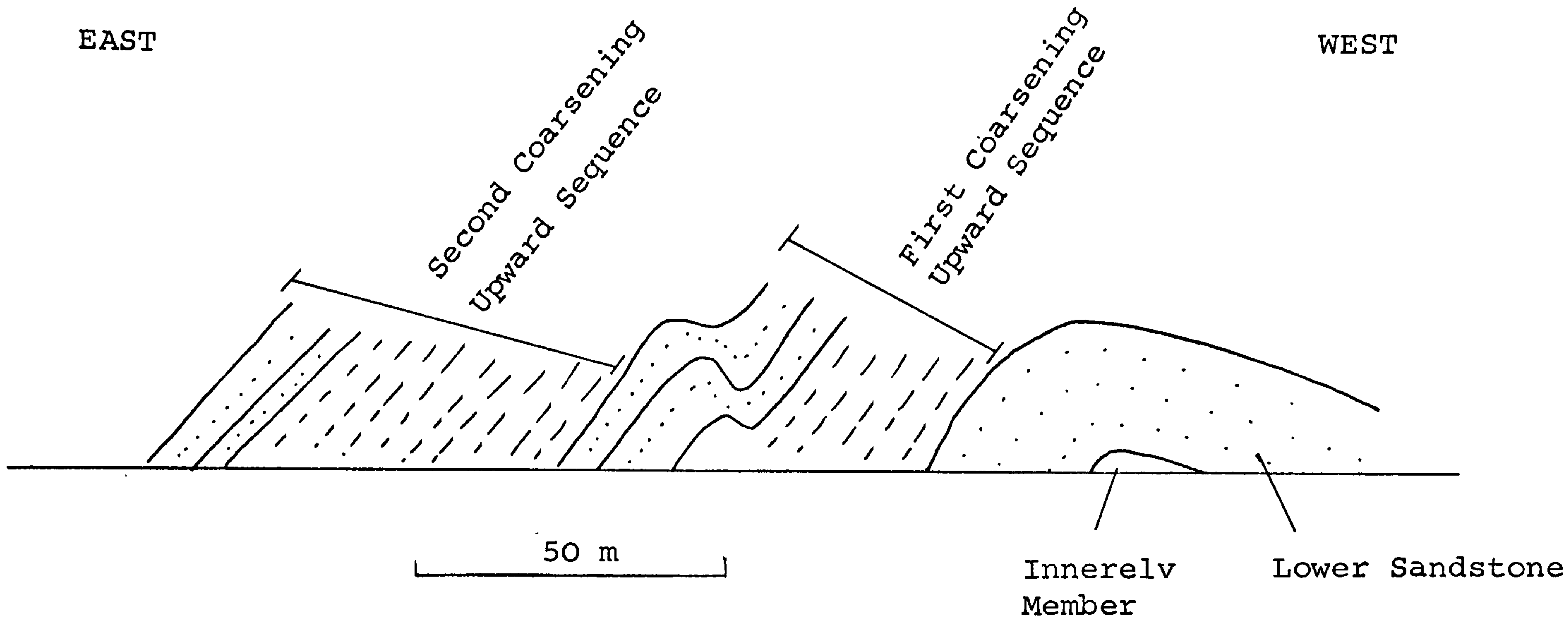
Sections through the Manndraperelv Member, Leirpollen area.



Sections through the Manndraperelv Member,
Leirpollen area.
Localities shown in Appendix B.



SECTIONS in the
MANNDRAPERELV MEMBER
IN THE JIERTASELVTUNGEN
AREA (between Ifjord
and Vestertana).
Numbers refer to localities on map.



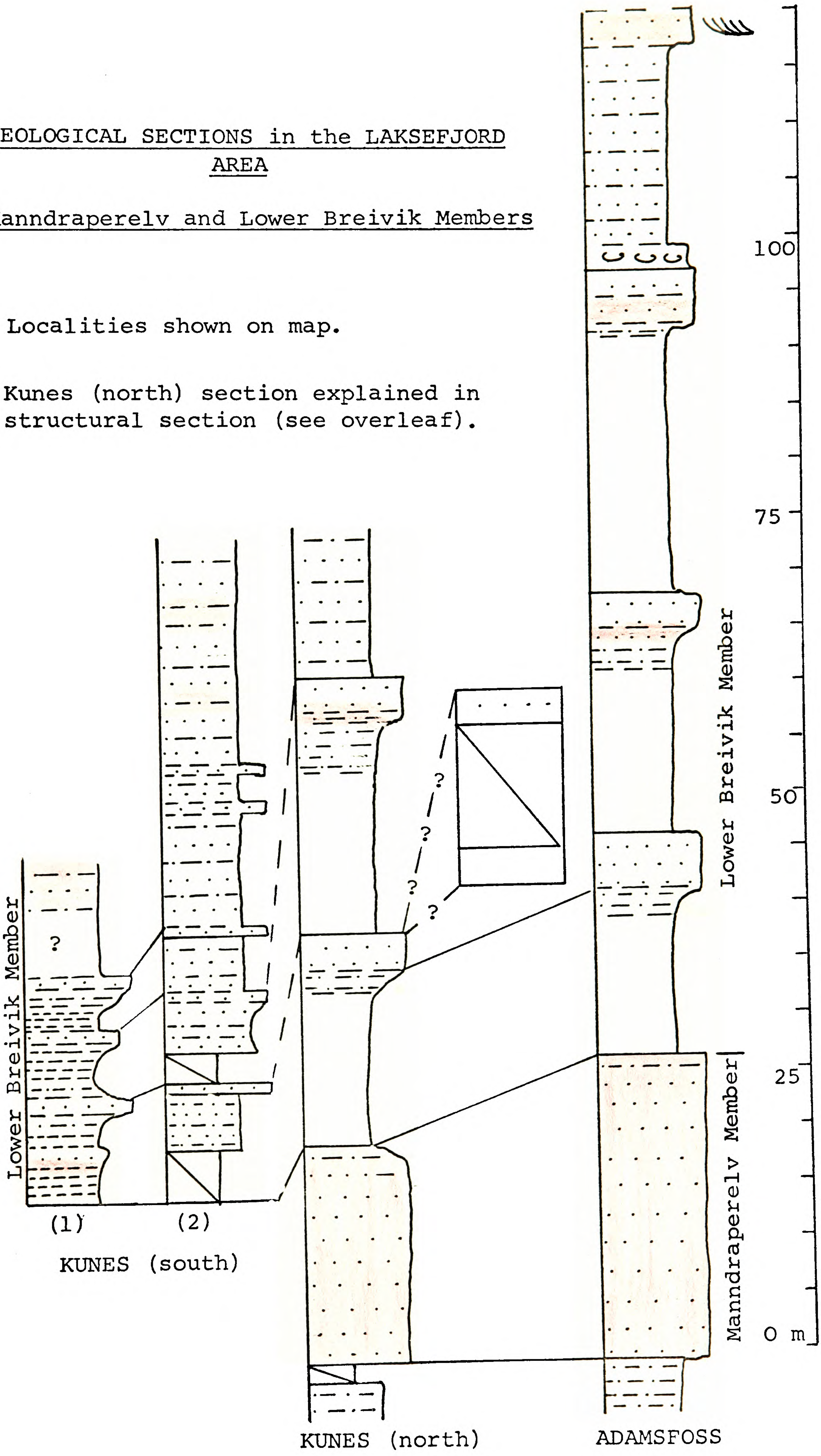
STRUCTURE IN THE STORELV VALLEY AT LOCALITY 4 (JIERTASELVTUNGEN)

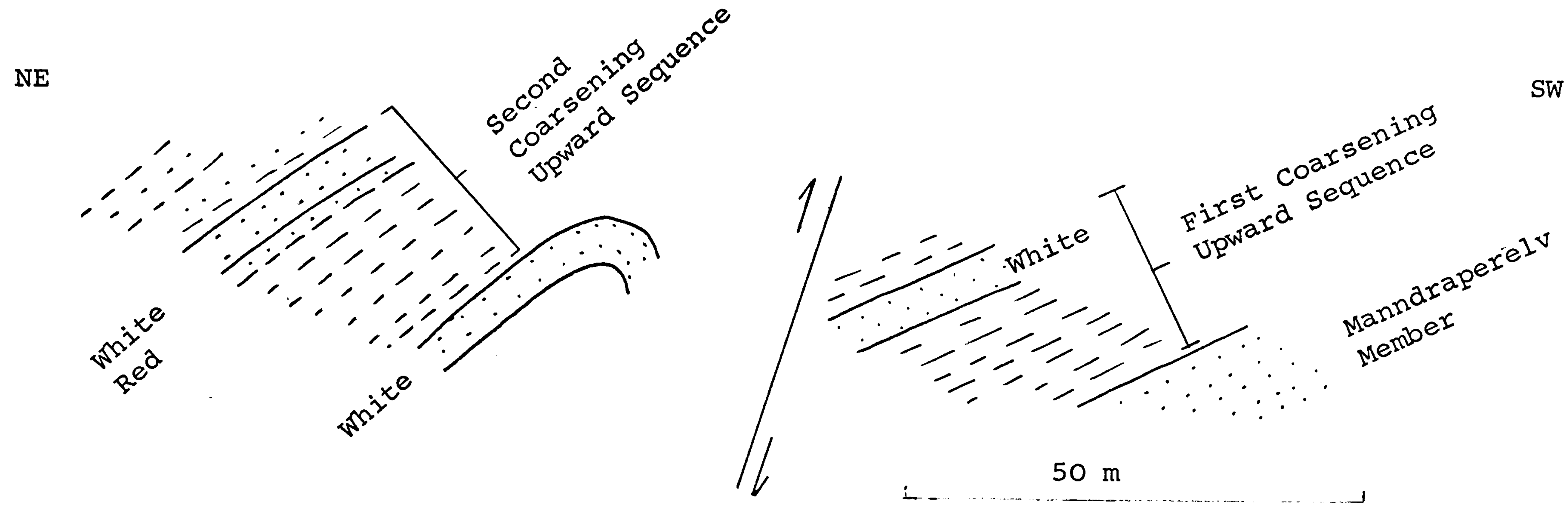
GEOLOGICAL SECTIONS in the LAKSEFJORD AREA

Manndraperelv and Lower Breivik Members

Localities shown on map.

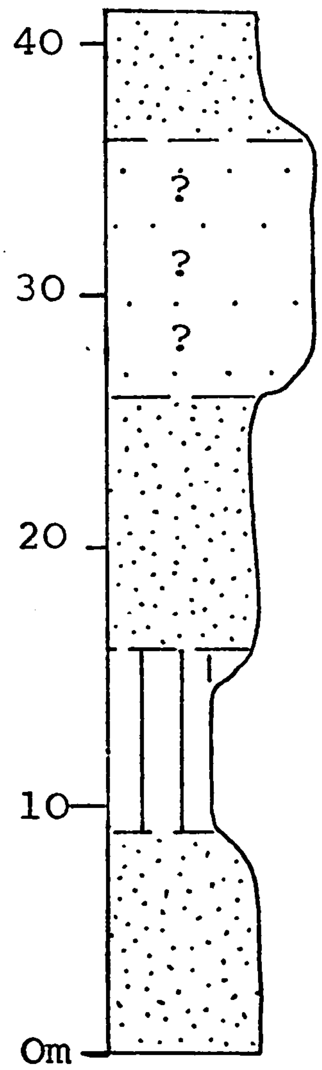
Kunes (north) section explained in structural section (see overleaf).





VIEW OF THE KUNES (north) SECTION AS SEEN FROM THE ROAD. (see map for locality)

The section shown in the text and in this appendix assumes that the two white sandstones are identical. If however the sandstone to the NE is a higher sandstone it is probably the top of the second coarsening upward sequence . This alternative would make the section more comparable with that at Adamsfoss.



TOP OF EXPOSURE SOUTH OF FAULT (stream)

Grey and purple ssts. Many intraclasts

Red sst. Beds to 50cm. Thinner bedded at base. Grouped within Facies 4 but shows many similarities to the red ssts of the Manndraperelv Member.

Gradational from Manndraperelv Mbr below. Beds rather lenticular, some up to 50cm. Thinner bedded at top. Phycodes pedum 3m above base.

KEY

	Facies 1.	Siltstone with 0.1 - 10 cm bedded sandstones
	Facies 2.	0.1 - 10 cm bedded sandstones with subordinate siltstones
	Facies 3.	10-100 cm bedded fine and very fine sandstones
	Facies 4.	10-30 cm bedded medium to coarse cross- bedded sandstones.

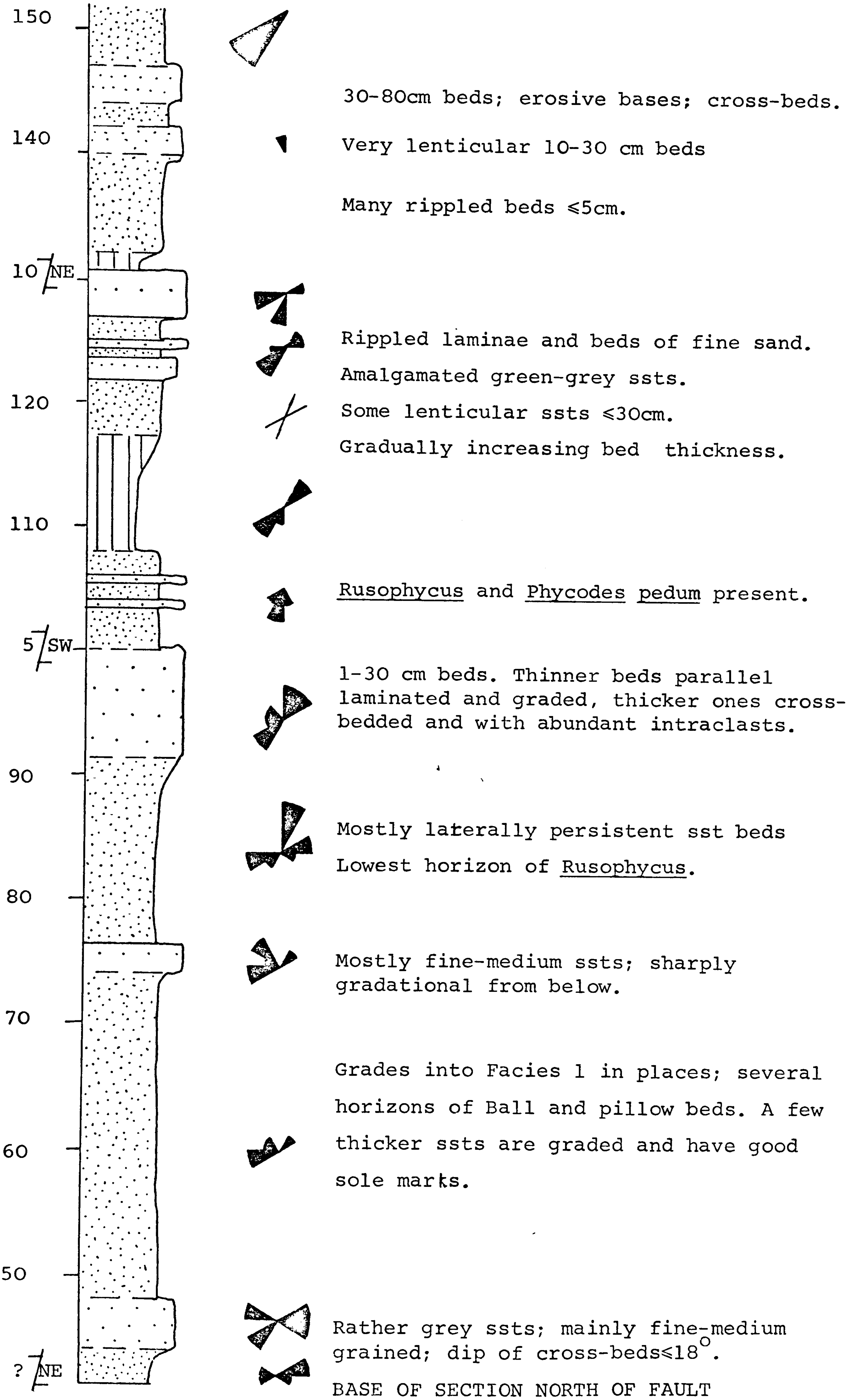
Fault: Downthrown side and amount of throw in metres

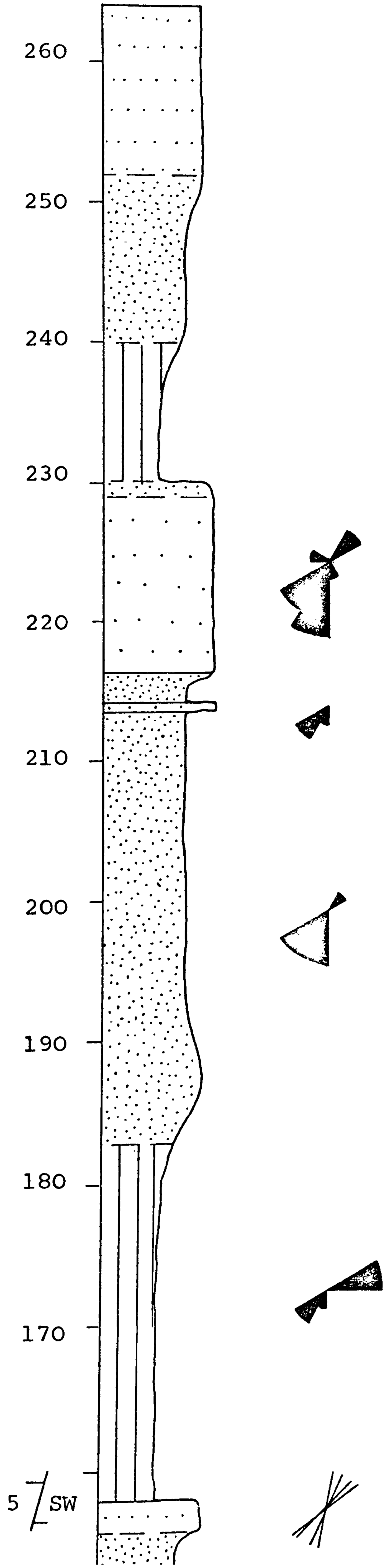
Rose diagrams show measurements on cross-bedding in Facies 4 and cross-lamination in other facies; smallest division shown equals one reading.

Axial lines shown are channel orientations.

Sst. = sandstone   Mst. = mudstone

DETAILED SECTION THROUGH THE LOWER BREIVIK MEMBER AS EXPOSED ON THE COAST SOUTH OF BREIVIK, DIGERMUL PENINSULA.


150
30-80cm beds; erosive bases; cross-beds.
140
Very lenticular 10-30 cm beds
Many rippled beds ≤5cm.
10 NE
Rippled laminae and beds of fine sand.
Amalgamated green-grey ssts.
120
Some lenticular ssts ≤30cm.
Gradually increasing bed thickness.
110
Rusophycus and Phycodes pedum present.
5 SW
1-30 cm beds. Thinner beds parallel laminated and graded, thicker ones cross-bedded and with abundant intraclasts.
90
Mostly laterally persistent sst beds
Lowest horizon of Rusophycus.
80
Mostly fine-medium ssts; sharply gradational from below.
70
Grades into Facies 1 in places; several horizons of Ball and pillow beds. A few thicker ssts are graded and have good sole marks.
60
50
Rather grey ssts; mainly fine-medium grained; dip of cross-beds≤18°.
? NE
BASE OF SECTION NORTH OF FAULT



Bioturbated at top: suggests slow sedimentation.

Distinctive white sst band with little silt; some vague cross-bedding.

Many irregular sandstones with complex internal stratification; Small sst dykes profuse at some levels.

Bioturbated beds at top of Facies 4

Rather thin-bedded ssts, lenticular, primary current lineation well seen: Some Facies 3 ssts also present.

Variable lithology on a small scale; abundant bioturbation at several levels with Phycodes particularly abundant.

Thicker ssts which are very lenticular. Washed out Phycodes on bases.

Erosion surface overlain by mst with rippled medium ssts.
Erosively based 10-100 cm bedded ssts

APPENDIX D

PETROGRAPHY

No detailed petrographic analysis was made but observations suggest that there are no significant mineralogical changes either laterally or vertically within the succession. All the sandstones are subarkoses or quartzarenites with the feldspar content varying from 0-15%. In general the coarsest sandstones have the least feldspar. Lithic fragments are absent apart from intr-formational material, which is ignored for classification purposes, and polycrystalline quartz and chert. The latter two are grouped at the quartz pole since their distribution is clearly a function of grain size rather than mineralogical maturity, their abundance increasing with increasing grain size. This relationship confiems the work of Conolly (1965). Moderate to intense pressure solution is a feature of all sandstones.

Information is given below of some features of the principle components of the sandstones.

Quartz, polycrystalline quartz, chert

Almost all quartz grains have slight or strong undulatory extinction but since the sediments have been strongly folded this may be largely a secondary affect and thus gives no indication of the original proportion of undulatory and non-undulatory grains.

Chert and polycrystalline quartz grains are a common minor constituent of sandstones of medium grade or coarser and may constitute as much as 10% of the grains in very coarse sandstone. In some polycrystalline grains the quartz crystals are elongate and show a preferred orientation suggesting a metamorphic origin. Chert grains show a variety of textures. Many consist entirely of minute equant grains (microcrystalline quartz of Folk and Weaver (1952)) and have a homogenous texture. Others show variations in grain size with circular or elliptical patches of microcrystalline quartz set in a mosaic of coarser grains. Grains of this latter type are similar to some of the silicified limestones described by Swett (1965,) e.g. Pl. 4c) and to the chert pebbles described by Wells (1947) which he believed to be silicified limestones. In this case the original rock, may have been a silicified oolitic or pelletal limestone with a spary calcite cement. However, not all the chert and aggregates of fine-grained quartz crystals are primary. Some obviously formed during diagenesis as infills of voids and possibly as a replacement of weathered feldspars.

Feldspar

Twinned plagioclase is the most obvious feldspar in thin section and often represents at least 50% of the total feldspar content. Polysynthetic twinning is more common than simple twinning. The compositions, using the Michel Levy method, are always in the range Au_{10-40}. These feldspars are often fresh but all degrees of alteration are also seen. Microcline is usually less common than twinned plagioclase and cross-hatched twinning is rarely well seen. Perthites are uncommon.

Untwinned feldspars (presumably mainly orthoclase and üntwinned plagioclase) are less easy to detect particularly in the finest sandstones but they can be at least as abundant as twinned plagioclase. Weathered feldspars are frequently altered to chlorite.

? Glauconite

A light green coloured mineral occurs in small quantities in some sandstones in the Lower Breivik Member of the Leirpollen area and in the upper part of Member IV at Halkkavarre. It is found as well rounded grains of the same size or slightly larger than the surrounding quartz grains. It has moderate relief and some grains show slight pleochroism from green to straw-coloured. Under crossed nichols it shows first order yellow or grey interference colours. It often appears to have a fine-grained granular texture.

Whilst this texture suggests that the mineral is probably glauconite the low birefringence is more typical of chamosite. However, whatever its original composition is has probably become somewhat degraded during subsequent diagenesis and tectonism.

Heavy minerals

The only common heavy minerals are zircon, tourmaline, leucoxene and hematite. Ilmenite, magnetite (?) and rutile are occasionally formed. Tourmaline occurs as both green and blue varieties. Both tourmaline and zircon show variable roundness. The heavy minerals rarely make up more than 1.5% of the grains. Leucoxene is the most common and was recognised by its white appearance in relfected light. Hematite is only common in red or white sandstones whereas leucoxene is found in all types. This distribution is probably due to diagenetic changes rather than differences in source area and is analagous to the situation described by Flemal (1969).

Others

Muscovite is especially common in very fine and fine sandstones but biotite is rare. Both are often partially altered to chlorite.

Cements

In the beds of medium sandstone and coarser quartz overgrowths are the main cementing agent with carbonate cement being of secondary importance. Since this carbonate is iron rich as is shown by staining and by its common alteration to limonite it is either siderite or ferroan dolomite. Ferroan calcite was excluded using Dickson's (1965) staining technique. Where a carbonate cement is present the quartz grains are frequently corroded. Sometimes euhedral crystals of carbonate are present and these probably grew at a late stage in the diagenetic history.

In the finer beds, particularly those with appreciable matrix, the cement is usually a mixture of fine grained carbonate and chert with a few quartz overgrowths. The matrix, which consists largely of chlorite, and the cement are intimately intergrown and surround sand grains which are always markedly corroded.

Hematite cementation is a feature of the red sandstones and siltstones of the Manndraperelv Member. In these beds the detrital grains are often markedly corroded by the cement.

APPENDIX E

COMPARISON WITH OTHER PHANEROZOIC EXAMPLES

In this thesis the main method of interpreting sedimentary environments has been to make inferences about the ancient sediments directly from our knowledge of the processes operating in modern shallow marine environments. Only occasionally have comparisons been made with other described ancient examples. This appendix is designed to point out a few ancient sequences which can, to varying extents, be compared with those described in the text.

Lillevatn Member - Innerelv Member Transition

The cross-bedded siltstones seen in the "Areholmen" section on the Digermul Peninsula can be compared with interlaminated very fine sandstones and siltstones in the Miocene of Libya which have depositional dips of up to 20° (Selley 1969, p.437). These Miocene sediments have been interpreted as point bar deposits in estuarine channels. However, these are much larger scale channels than those found in Finnmark.

Smith (1968) has interpreted fining upward sequences in the Silurian of Pennsylvania as point bar deposits of tidal creeks but in this case high angle lateral accretion deposits are apparently absent.

Innerelv Member

Ancient comparisons with this member are fully discussed in the text.

Manndraperelv Member

The part of this Member which can be compared most convincingly with other ancient examples is the beach - lagoon complex in the First Coarsening Upward Sequence. The best known ancient beach sequences are undoubtedly those of the Cretaceous of the western U.S.A. (Lane 1960, Masters 1963, Weimer 1966, Campbell 1971).

Lower Breivik Member

I know of no ancient sediments which can be compared closely with this member. The most similar are probably the southern facies of the Ffestiniog Stage in the Cambrian of Wales (Crimes 1970) and the Meadfoot beds of S.W.England (Richter 1967).

Upper Breivik Member and Duolbasgaissa Formation

In general terms the beds of this part of the succession can be compared with the Ffestiniog Stage (Crimes 1970). The Upper Breivik and Lower Duolbasgaissa Members are closest to the south Wales, Harlech Dome and St Tudwal's Is. exposures and the Upper Duolbasgaissa Member is most similar to the N. Caernarvonshire exposures. Earlier workers had suggested a turbidite origin for some of these deposits but Crimes argues convincingly that these are shallow marine sediments.

The cross-bedded sandstones of the Upper Duolbasgaissa Member are also comparable with a number of deposits to which a tidal origin has been assigned. For example, the Upper Devonian of S.W. Ireland (Kuijpers1971), the Cambrian Eriboll Quartzite of Scotland (Swett et al. 1971) and the Lower Greensand of Leighton Buzzard (A.Bentley, unpublished B.Sc. thesis, Oxford).

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Trace fossils

Edited by
T.P. Crimes
and
J.C. Harper

NOTE

The sedimentological interpretation of the Upper Duolbasgaissa Member given here differs from that in the main text. The interpretation given here was based on preliminary studies and was later found to be wrong. However, the new interpretation of the Upper Duolbasgaissa Member as an offshore tidal deposit makes no difference to the main conclusions of the paper.

Trace fossils from the late Precambrian and Lower Cambrian of Finnmark, Norway.

N. L. Banks

A conformable succession of dominantly shallow marine sediments of late Precambrian and Lower Cambrian age is present in the Tanafjord area of Finnmark, northern Norway. In this succession trace fossils first occur a short distance above the late Precambrian (Varangian) tillites. Their abundance and diversity increases rapidly in latest Precambrian and early Cambrian sediments. The absence of biogenic activity in older sediments is not due to the lack of suitable sedimentary facies. The incoming and diversification of trace fossils reflects principally the development of the Phyla Annelida, Arthropoda and Mollusca. The rate of early metazoan evolution and the factors which controlled it are briefly discussed.

1. Introduction

What is the earliest evidence of undoubted metazoan life? How rapidly did the Metazoa evolve in late Precambrian and Cambrian times? In investigating these problems, trace fossils as well as body fossils may give some indications as to the state of development of certain phyla. Some trace fossils are sufficiently distinctive to allow reasonable inferences to be made as to the phylum of the animal concerned (Glaessner 1969). With the appearance of metamerically segmented worms (i.e. annelids), sustained burrowing became possible for the first time in metazoan history (Clark 1964). Thus the first occurrence of well-developed burrow systems may reflect the initial development of the Phylum Annelida. From studies of the ecology of modern animals and the functional morphology of fossils, it seems likely that most biogenic structures in Palaeozioc sediments were produced by members of the Phyla Annelida, Arthropoda and Mollusca. It is universally agreed by zoologists that these phyla are closely linked in their origins, and probably all evolved at about the same time. Thus the development of trace fossils in late Precambrian and early Cambrian times probably more accurately reflects the evolution of these phyla than of metazoans as a whole.

Trace fossils from late Precambrian and Cambrian strata have been described and discussed by many authors. Seilacher (1956) described material from Pakistan and the U.S.A. and concluded that trace fossils were very rare in Precambrian rocks and showed an explosive differentiation at the beginning of the Cambrian. Glaessner (1969) studied trace fossils from late Precambrian and Cambrian successions in southern and central Australia and supported Seilacher's thesis of a rapid differentiation of soft bodied benthonic life at the base of the Cambrian. However, he also showed that several distinctive forms occur in late Precambrian rocks.

The purpose of this paper is to describe the incoming and development of trace fossils in a conformable succession of late Precambrian and Lower Cambrian sediments in the Tanafjord area of Finnmark, northern Norway, and to attempt to

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distinguish between those changes in biogenic activity which are the result of variations in the sedimentary environment, and those which may be reflections of the evolution of animal life at that time.

2. Geological background

The geology of the Tanafjord area of Finnmark (Fig. 1) was first elucidated by Føyn (1937) following the observations of Holtedahl (1918; 1931). The most complete section of the latest Precambrian and younger rocks is found on the Digermul Peninsula where Reading (1965) established the presence of a conformable succession of sedimentary rocks, 3000 m thick, extending from late Precambrian (Varangian, Eocambrian) tillites through the Cambrian and into the Tremadoc (Fig. 2). The succession can be divided into two parts:

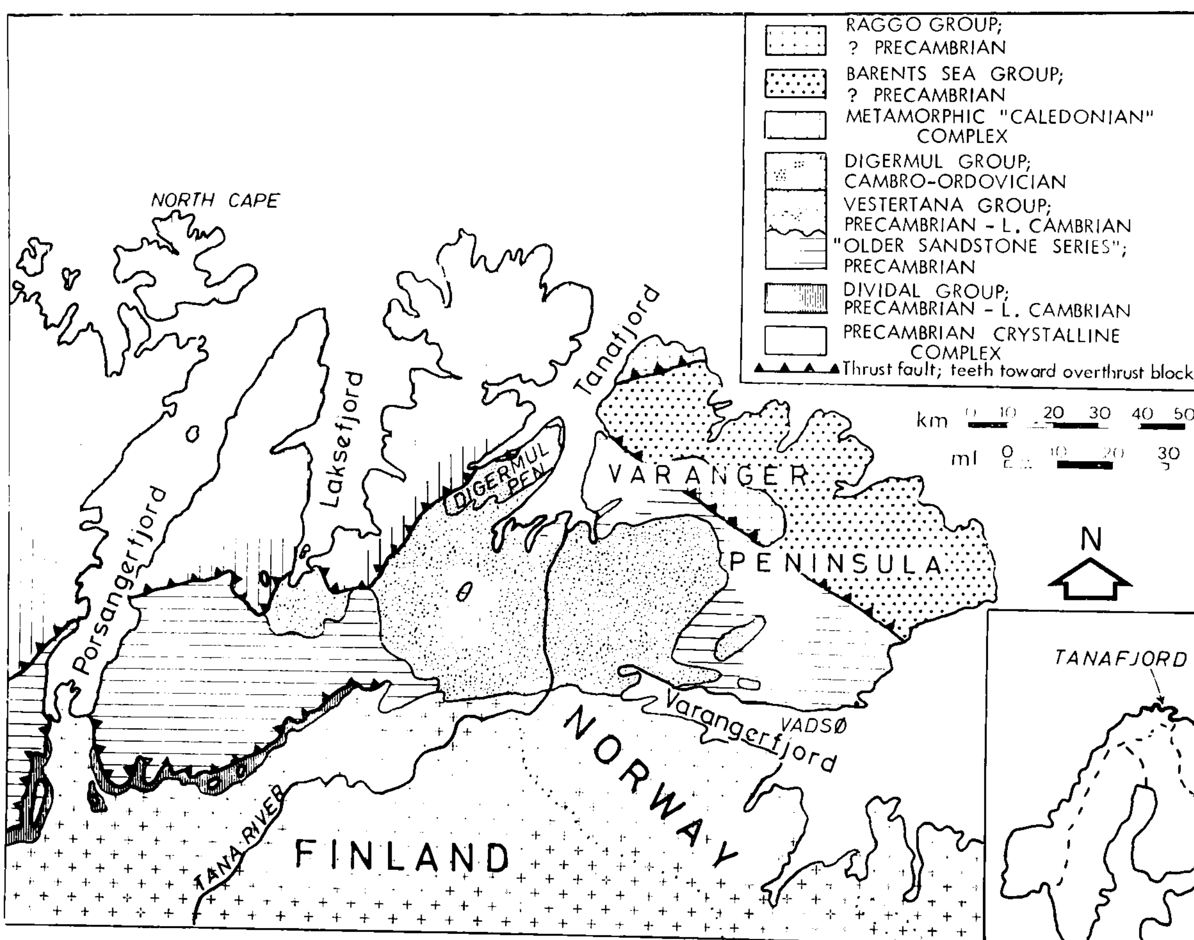


Fig. 1. Geological Map of East and Mid Finnmark. After Føyn (1937; 1967), Reading (1965) and Siedlecka and Siedlecki (1967).

2. Digermul Group: 1500 m—Duolbasgaissa, Kistedal and Berlogaissa Formations. Lower Cambrian—Tremadoc.

1. Vestertana Group: 1550 m—Lower Tillite, Nyborg, Upper Tillite, Stappogiedde and Breivik Formations. Precambrian-Lower Cambrian.

Reading considered that the majority of the sediments were deposited in a shallow marine basin, although occasional deepening allowed sedimentation by turbidity currents to occur in a non-agitated environment.

Together with the sediments of the unconformably underlying "Older Sandstone Series" this succession forms a northward thickening wedge of dominantly clastic miogeosynclinal sediments. This wedge is bounded to the south by the Fenno-scandian basement, to the northwest by an overthrust "Caledonian" metamorphic complex and to the northeast it is separated by a tectonic discontinuity from the sediments of the Barents Sea Group (Siedlecka and Siedlecki 1967). Part of the wedge can be correlated with the Dividal Group (Føyn 1967), a relatively condensed succession of late Precambrian and Cambrian sediments whose outcrop can be traced from Finnmark, through northern Sweden into southern Scandinavia.

Palaeontological studies on the Digermul Peninsula by Henningsmoen (in Reading 1965) revealed good Middle Cambrian and younger faunas although only fragments of *Holmia* sp. were found below the Middle Cambrian. I have now found several specimens of *Platysolenites antiquissimus* Eichwald at a horizon about 150 m above the base of the Breivik Formation (Fig. 2). This fossil, recently interpreted by Hamar (1967) as a serpulid worm tube, is well known from other localities in Finnmark (Føyn 1967), from southern Scandinavia and the Baltic regions. It is considered indicative of a very low horizon in the Lower Cambrian which is thus at least 950 m thick in this succession.

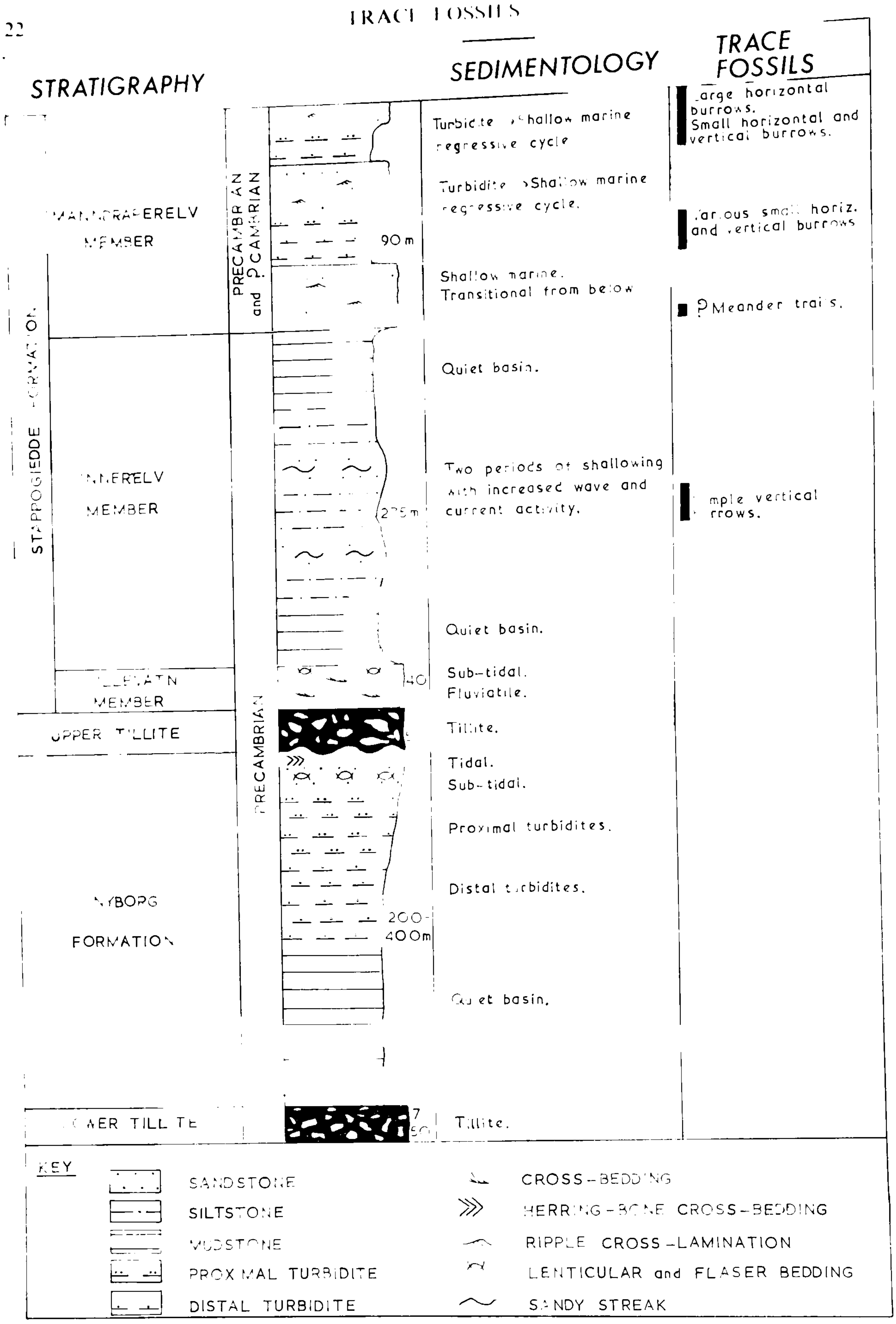
3. Depositional history

Only a brief description and discussion of the sedimentology need be given here but a more detailed account has been given by Banks *et al.* (in press). The main sedimentological features of the succession are summarised in Figure 2.

The sedimentation of the two tillites and the immediately adjacent sediments was described in detail by Reading and Walker (1966). The upper part of the Nyborg Formation is a well developed regressive sequence passing from distal turbidites through proximal turbidites into a wide variety of shallow marine sediments. The latter include herring-bone cross-stratified sandstones of probable tidal origin. The unconformity at the base of the Upper Tillite is considered to be due to glacial scouring and it is probable that no significant time break is represented. Above the Upper Tillite the transition from the Lillevatn Member to the Innerelv Member was interpreted by Reading and Walker (1966) as a transgressive sequence, passing from a fluviatile facies through a sub-tidal facies into a basinal facies with mudstone deposition.

Two periods of shallowing occurred within the Innerelv Member. They are indicated by transitions from mudstone through thin bedded fine sandstones and siltstones into medium to thick bedded lenticular sandstones; the latter often sit within large low-angle scours which also contain irregularly bedded, rippled, fine sandstones and siltstones resembling the sandy streak facies of de Raaf *et al.* (1965). The scours may have been produced by exceptional wave activity associated with storms.

Quiet basin mudstones occur again at the top of the Innerelv Member and pass gradually upward into red siltstones and sandstones which mark the base of the Manndraperelv Member. This member includes a lower unit consisting predominantly of fine red sandstones, deposited in a fairly active offshore environment, which is overlain by two regressive cycles 40–90 m thick. These cycles pass up from thin turbidites with interbedded mudstones into more proximal turbidites and



Figs.2a (above) and 2b (opposite). Sedimentological interpretation and trace fossils of the Vestertana Group and the lower part of the Digermul Group. Partly after Reading and Walker (1966).

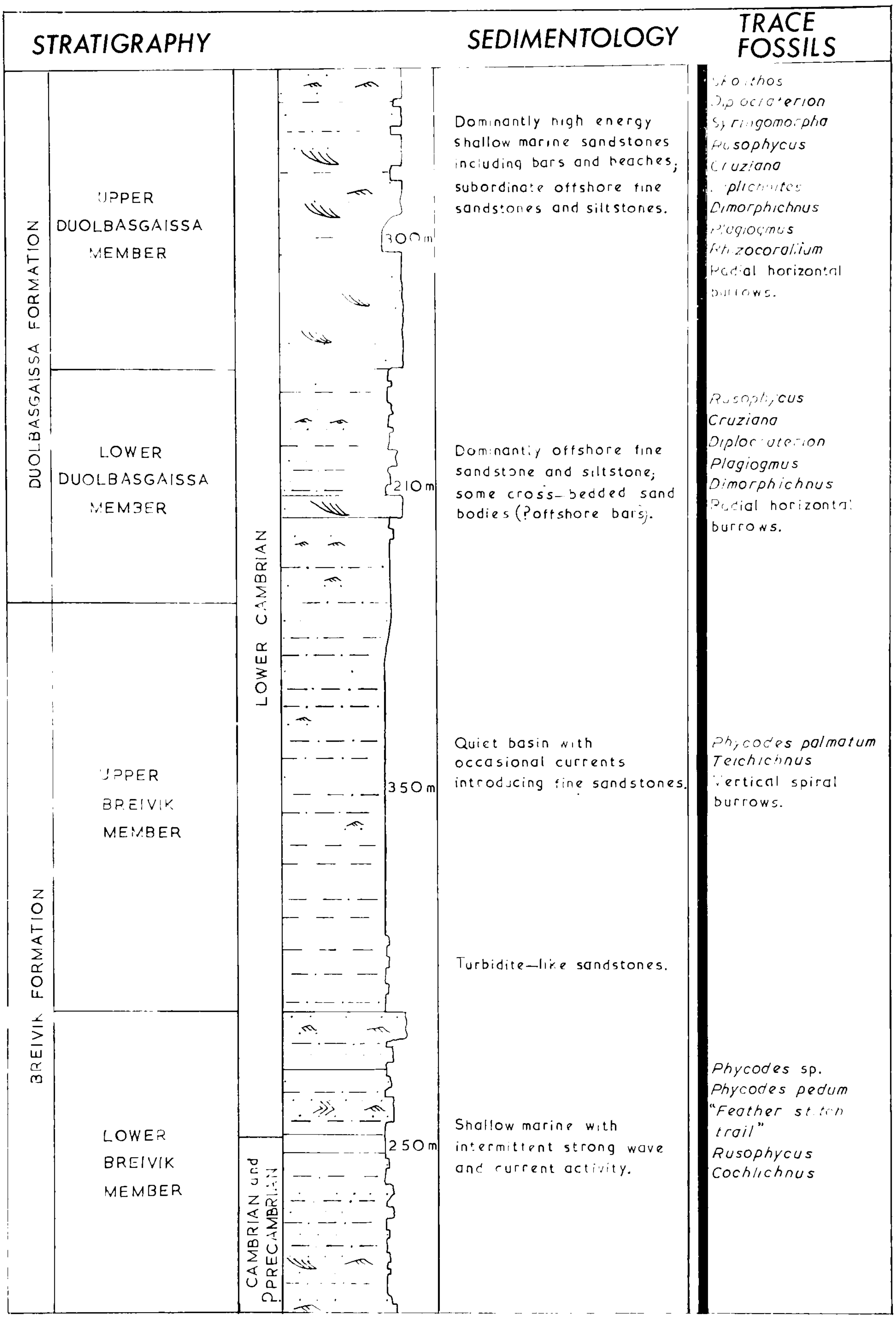

STRATIGRAPHY
SEDIMENTOLOGY
TRACE FOSSILS
DUOLBASGAISSA FORMATION
UPPER DUOLBASGAISSA MEMBER
LOWER DUOLBASGAISSA MEMBER
BREIVIK FORMATION
UPPER BREIVIK MEMBER
LOWER BREIVIK MEMBER
LOWER CAMBRIAN
CAMBRIAN and PRECAMBRIAN
300 m
210 m
350 m
250 m
Dominantly high energy shallow marine sandstones including bars and beaches; subordinate offshore fine sandstones and siltstones.
Dominantly offshore fine sandstone and siltstone; some cross-bedded sand bodies (?offshore bars).
Quiet basin with occasional currents introducing fine sandstones.
Turbidite-like sandstones.
Shallow marine with intermittent strong wave and current activity.
Rusophycus
Cruziana
Dimorphichnus
Rhizocorallium
Radial horizontal burrows.
Rusophycus
Cruziana
Diplocraterion
Plagiogmus
Dimorphichnus
Radial horizontal burrows.
Phycodes palmatum
Teichichnus
Vertical spiral burrows.
Phycodes sp.
Phycodes pedum
"Feather stitch trail"
Rusophycus
Cochlichnus

Fig. 2b.

then into shallow marine sandstones. The majority of the upper sandstones are similar to those of the lower unit, but in one section lagoonal and swash zone facies are also present. All the evidence from sedimentary structures, vertical and lateral facies relationships and petrography suggests that the sediments of the upper parts of the cycles represent true marine rather than deltaic conditions.

The transition upward from the Manndraperelv Member (Stappogiedde Formation) into the Breivik Formation is lithologically gradational. The Lower Breivik Member, a series of rapidly alternating sandstones, siltstones and mudstones, is sedimentologically complex, but in general appears to have been deposited in a shallow marine environment, mostly above wave base. Large fluctuations in current strength are indicated by the interbedding of cross-stratified clean sandstones with siltstones and mudstones. Lenticular and flaser bedding (Reineck 1967) occur and some horizons are extensively bioturbated.

The transition into the Upper Breivik Member is abrupt. The lower part of the Upper Breivik Member consists of siltstones with occasional sharp-based fine sandstones which often fill channels, and were probably deposited by turbidity currents. This facies passes upward into silts with many thin rippled sandstones which are interpreted as having been deposited in a gently agitated "shelf" environment.

The sediments of the Duolbasgaissa Formation were deposited in a nearer shore environment. Offshore thin bedded fine sandstones and siltstones interfinger with coarser sand bodies deposited under higher energy conditions. These sand bodies consist mainly of cross-bedded and flat-bedded orthoquartzites which were deposited as a series of bars and beach/barrier sands. The Lower Member consists almost entirely of the offshore facies whilst in the Upper Member thick sand bodies predominate.

4. Trace fossils

Trace fossils have been recorded from the Digermul Peninsula by Strand (1935), Føyn (1937) and Reading (1965). Reading considered that trace fossils first occurred in the upper part of the Nyborg Formation but the specimen on which this statement was based is now believed to be of inorganic origin (Reading personal communication). Further study of the Nyborg Formation has failed, as yet, to provide any evidence of animal life (M. B. Edwards personal communication and my own observations). Similarly, studies in the "Older Sandstone Series", a series of shallow marine and fluviatile sediments lying with slight regional unconformity

Plate 1

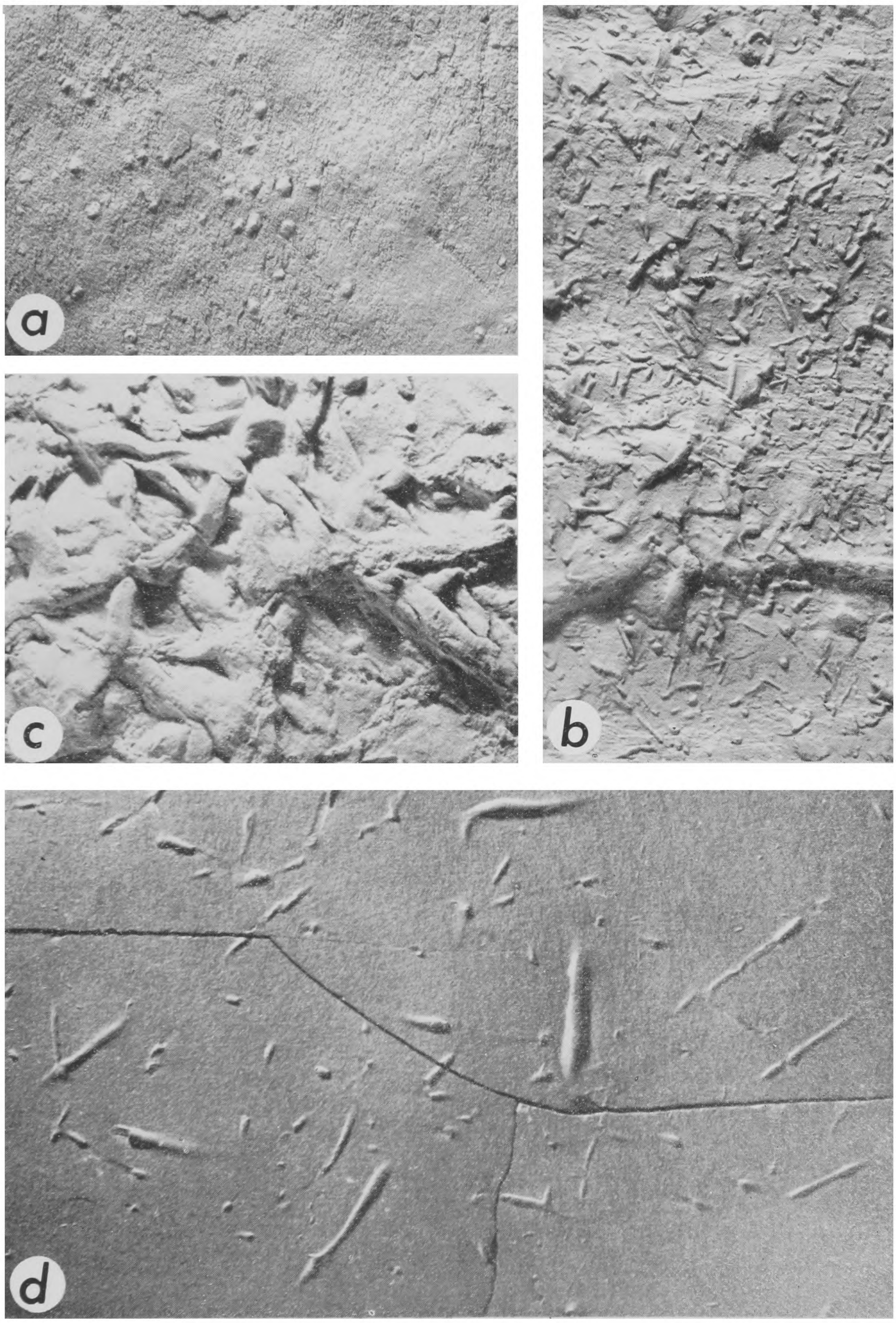
a Hypichnial casts; passively filled simple vertical burrows. 150 m above base of Innerelv Member (× 1).

b Hypichnial and exichnial casts showing various burrowing patterns. Turbidite facies, 75 m above base of Manndraperelv Member (× 0·9).

c Simple horizontal burrows; hypichnial and exichnial casts. Shallow marine facies, 180 m above base of Manndraperelv Member (× 1·2).

d Hypichnial and exichnial casts showing horizontal burrows with occasional branching. Turbidite facies, 75 m above base of Manndraperelv Member (× ·075).

All specimens from Digermul Peninsula.


a
c
b
d

below the Lower Tillite, have also failed to reveal the presence of trace fossils (W. P. Geddes personal communication).

The general sequence of trace fossils is summarised in Figure 2. Brief descriptions are given below of some of the early forms and more general comments are made about some of the well known later types. The classification of Martinsson (1965) is used to describe the mode of preservation of the specimens.

Innerelv Member. The first undoubted trace fossils occur 140–150 m above the base of the member in a facies of interlaminated siltstone and mudstone. In the field they appear as circular protuberances (hypichnial casts) on the bases of siltstone laminae (Pl. 1a). The diameter of the protuberances is 1-2 m. Occasionally they occur in pairs but usually there is no pattern to their distribution. No sign of burrowing can be seen in the overlying siltstone and the structures are interpreted as being passive infillings of vertical burrows.

Manndraperelv Member. Trace fossils occur intermittently throughout this member. In general they are very rare in the shallow water sandstones but common in the turbidites, where several types are present. This distribution is probably a function of the higher preservation potential of trace fossils in the latter environment. However, it may also reflect the original distribution of the infaunal population.

In the turbidite facies trace fossils are found on the undersurfaces of sandstones and siltstones. They occur most frequently on the bases of thin beds (<5 cm) although they are occasionally seen associated with thicker beds. The most common forms are:

(i) Cylindroidal burrow tubes, approximately 1 mm in diameter, infilled with fine sand or silt and often densely covering undersurfaces of beds and developed as a number of different forms (Pl. 1b). These forms include regularly sinuous horizontal burrows, loosely coiled (spiral) horizontal burrows (cf. *Helicolithus*), short horizontal burrows with Y-shaped branching, and a variety of other patterns, including vertical non-septate U-tubes. The polychaete worm *Notomastus* is known to produce rather similar burrows with short branches and spiral structures in the present day sediments of the North Sea (Reineck *et al.* 1967).

(ii) Dominantly horizontal burrows, circular or elliptical in cross-section, 2–3 mm in diameter, straight or slightly curving, often branching every 2–4 cm. The branches usually diverge at angles of 60–90°. Other more irregularly branching forms also occur. They are found as exichnial sand-filled casts lying closely below a sandstone bed or as hypichnial casts (Pl. 1d). Most of the trace fossils in this facies were probably produced by worms.

The undersurfaces of many sandstone beds show delicate groove marks produced by tools carried by the current which deposited the sand. These marks often consist of sets of fine sub-parallel striations. These are usually straight and continuous, but occasionally are gently curved, discontinuous, and occur in sets

Plate 2

a *Phycodes pedum* Seilacher, Lower Breivik Member (× 1·3).

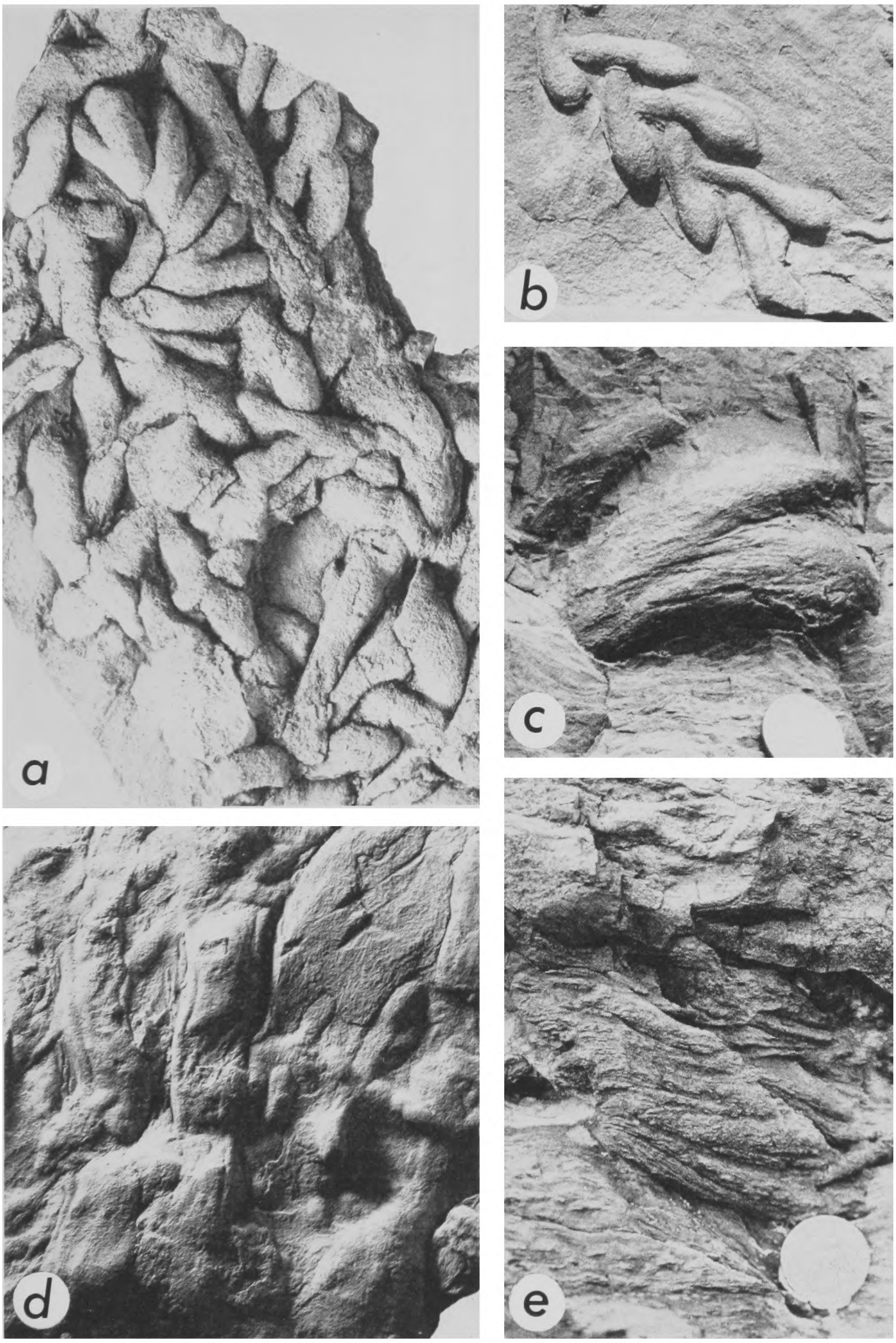
b "*Feather-stitch trail*", Lower Breivik Member (× 1).

c Vertical spiral burrow, Upper Breivik Member (× 0·55).

d Horizontal burrows with lateral grooves; hypichnial casts. Shallow marine facies, 180 m above base of Manndraperelv Member (× 1).

e cf. *Teichichnus*, Upper Breivik Member (× 0·6).

All specimens from Digermul Peninsula.


a
b
c
d
e

resembling in part the trace fossil *Dimorphichnus* Seilacher. Seilacher (1955) interpreted these marks as the tracks of a sideways-moving trilobite. It seems possible that many of these delicate markings could have been produced by the hard exoskeletons of arthropods during transport. The tool could either have been an animal dead or alive, or exoskeletons previously discarded by ecdysis. Similar structures have been described by Seilacher (1960) and Martinsson (1965). If this is so, it provides indirect evidence for the existence of arthropods at this time, although no hard parts were preserved.

In the shallow marine facies, some poorly preserved trace fossils occur a few metres above the base of the Member. They resemble structures described by Glaessner (1969 fig. 5C, D) from the Pound Quartzite of South Australia which he interpreted as meander trails.

Several types of trace fossil appear in the uppermost part of the Member. These include:

(i) Horizontal burrows 5–10 mm wide and up to 10 mm deep with well-developed lateral grooves. They are found as hypichnial or exichnial sand filled casts (Pl. 2d). The lateral grooves suggest that the animal which produced the structure may have been an arthropod.

(ii) Horizontal burrows 3–5 mm in width, without ornamentation, randomly curved and frequently crossing, found as exichnial or hypichnial sand filled casts (Pl. 1c).

(iii) Small vertical U-tubes with the width of the U *c.* 2 cm (*?Diplocraterion*) and several irregular types of horizontal and vertical burrows.

Lower Breivik Member. Here biogenic activity has reached an intensity at which marked modification of the sedimentary lamination is often visible. Many horizontal burrows are present, the majority being simple sand-filled tubes which lack any distinctive character. The trace fossil assemblage is dominated by *Phycodes*. Most specimens can be referred to *Phycodes pedum* Seilacher or are closely related to this species (Pl. 2a). *Phycodes pedum* first occurs in the lowest 3 m of the member and is persistent throughout it. The diameter of the burrow tubes varies from 2–20 mm. Seilacher (1955) considered that this species might serve as a Cambrian index fossil of world-wide extent. Its wide distribution is further confirmed by its presence in Finnmark and also in Australia (Glaessner). Its first occurrence in Finnmark, 140–150 m below the level of *Platysolenites*, suggests that the organisms responsible had developed in very earliest Cambrian, or perhaps even latest Precambrian times. *P. pedum* is now known to range up into the Lower Ordovician (Seilacher, personal communication). Probably closely related to *Phycodes pedum* is the so-called "*Feather-stitch trail*" (Pl. 2b); this, like *Phycodes*, is best interpreted as a feeding burrow.

Rusophycus is first seen about 70 m above the base of the member. *Rusophycus* is used here in general discussion to refer to paired sets of claw impressions which can be interpreted as resting marks (presumably of trilobites in most cases). *Cruziana* is used for more continuous markings, usually interpreted as the result of furrowing through the sediment.

Plate 3

a *Plagiogmus* sp., Lower Duolbasgaissa Member (× 1).

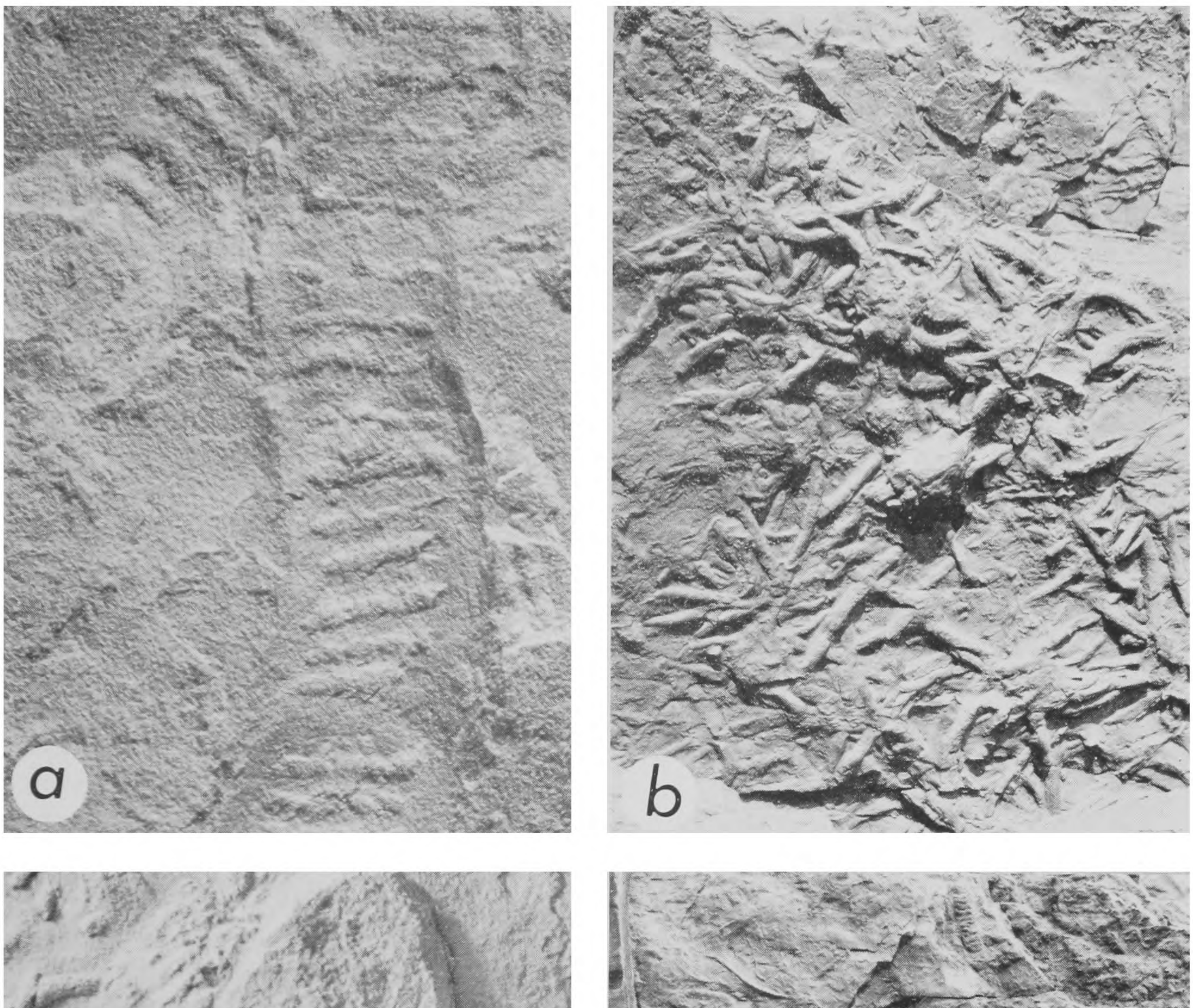
b Radial horizontal burrow patterns; exichnial casts, Lower Duolbasgaissa Member (× 0·09).

c *Rusophycus* sp., Lower Duolbasgaissa Member (× 1·25).

d Large tracks cf. *Dimorphichnus* sp., Lower Duolbasgaissa Member (× 0·07).

All specimens from Digermul Peninsula.

PLATE 3



Some *Rusophycus* are only very shallowly dug paired sets of scratch marks (cf. Cloud and Nelson 1966 fig. 1g); others are much more deeply excavated bilobate pits. The top surfaces of sandstone beds show a number of epichnial grooves. These include regular sinuous trails (*Cochlichnus*) and bilobate trails with lateral lobe-like elevations similar to the trails produced by recent crustaceans (Nathorst 1881).

Upper Breivik Member. In the lower part vertically-stacked horizontal burrows which compare with *Teichichnus* are common (Pl. 2e) and are associated with various types of vertical spiral burrows. Some of these spiral burrows are irregular (Pl. 2c) whilst others maintain a constant whorl diameter and compare with the Mesozoic-Tertiary form *Gyrolithes* except that they are much smaller (tube diameter usually less than 5 mm). In the upper parts of the Member numerous small horizontal burrows occur associated with occasional specimens of *Phycodes palmatum* (Hall), a species described in detail by Seilacher (1955).

Duolbasgaissa Formation. Three intergradational trace fossil assemblages can be recognised in this Formation. The finest grained offshore siltstone and sandstone facies is dominated by large horizontal burrows (Pl. 3b). These are sometimes found in the form of radiating tunnels from a central chamber; scratch marks on the burrow walls suggest that they were probably produced by arthropods, although worms cannot be ruled out. The radiating burrow pattern has some similarity to that of *Volkichnium* figured by Pfeiffer (1965) but the Finnmark specimens are larger and laterally more extensive.

When the beds become sandier this form, though still present, becomes less common and *Rusophycus* (Pl. 3c), *Cruziana*, cf. *Dimorphichnus* (Pl. 3d), *Diplichnites*, *Plagiogmus* (Pl. 3a) and small *Diplocraterion* appear.

In the highest energy facies consisting of flat- and cross-bedded orthoquartzites, the assemblage consists of *Skolithos*, large *Diplocraterion* (width of U *c*. 12 cm), *Syringomorpha* and *Rusophycus*. Thus with increasing current activity there is a passage through the *Cruziana* facies of Seilacher (1967) into his *Skolithos* facies.

Several different forms of *Rusophycus* and *Cruziana* occur within this Formation and these are distinct from those found above and below this level. In the Upper Member some very large forms are present which resemble *C. dispar* Linnarsson. Cloud and Nelson (1966) suggested that the organisms responsible for these structures were more prone to a sedentary life in the Cambrian than in younger rocks. It is certainly true that *Rusophycus* is much more common than *Cruziana* in the Lower Cambrian of Finnmark, but in the overlying Kistedal Formation (Middle–Upper Cambrian) the reverse relationship is true.

5. Discussion

Seilacher (1964; 1967) has shown that trace fossils can be useful as environmental indicators. A clear relationship between sedimentary facies and the morphology of the trace fossils occurring in those facies is displayed in the upper part of the Finnmark succession. Such a relationship also demonstrates the wide ecological diversity already attained by animals in the Lower Cambrian.

It is thus important to consider whether changes in the environment can account, in whole or in part, for the increase in trace fossil activity which has been described. From the earlier discussion of the depositional history this seems to be unlikely. A turbidite facies in the inter-tillite Nyborg Formation has no trace fossils, whilst a similar facies in the younger Manndraperelv Member has abundant trails and burrows. Broadly similar, shallow marine facies occur in parts of the Nyborg

Formation, the Manndraperelv Member and the Lower Breivik Member. In the Nyborg Formation biogenic activity is absent, in the Manndraperelv Member it is present but rare, and in the Lower Breivik Member it is abundant. It is possible that further study may eventually lead to the discovery of trace fossils in the Nyborg Formation, but the overall pattern of their development in the Finnmark succession is abundantly clear.

From the evidence of this succession the following tentative suggestions are made:

(i) Before and during the time of the late Precambrian glaciation in Finnmark, animals capable of producing recognisable trace fossils did not exist, or were at a very early stage of development.

(ii) They first appeared soon after the cessation of glacial sedimentation and developed rapidly, so that by the end of the Lower Cambrian a level of biogenic activity had been reached which is comparable with that existing in many younger sediments of similar facies.

To see whether the suggestions made above are typical on a world-wide scale it is important to compare the evidence from as many areas as possible. Such a review is beyond the scope of this paper, but a few points can be mentioned. Firstly, at other localities in Finnmark, where the equivalent successions are often of somewhat different facies to that on the Digermul Peninsula, the same general trend in trace fossil evolution is seen. Secondly, in the Adelaide Series of South Australia the development of trace fossils is similar to that in Finnmark. The earliest occurs a short distance above the upper of two tillites; several forms are found in the Pound Quartzite associated with the Ediacara fauna while the Lower Cambrian contains a rich assemblage (Glaessner 1969)

It is not known whether the widespread late Precambrian glacial sediments are broadly time-equivalents. It would be of considerable interest to know the absolute ages of the tillites from the point of view of estimating the rate of early metazoan evolution. On the basis of isotopic dating presented by Dunn *et al.* (1966) the Australian tillites may be about 680–750 m.y. old. Banks *et al.* (1969) consider that the glaciation in Finnmark probably ended 15–43 m.y. before the beginning of the Cambrian (all dates recalculated where necessary using a Rb^{87} half-life of $5{\cdot}0 \times 10^{10}$ m.y. to enable comparison). The base of the Cambrian is poorly defined geochronologically but by recalculating Cowie's (1964) figure a date of 606 m.y. is obtained. Thus, in Australia the record of undoubted metazoan life apparently extends back about 100 m.y. before the Cambrian whilst in Finnmark this period is less than about 40 m.y. However such estimates must, at present, be regarded as tentative.

Many supposed trace fossils have been described from much older sediments. The great majority of these can, however, now be attributed to inorganic processes (Cloud 1968b; Glaessner 1969) but there remains a small number of specimens whose origins are problematical. Nevertheless, it is unlikely that any true trace fossils exist in rocks which are significantly older than about 700 m.y. The idea that Precambrian metazoan life is restricted to a relatively short period prior to the Cambrian appears to be coming increasingly accepted (Cloud 1968b; McAlester 1968).

If it is true that there was an extensive late Precambrian glaciation it seems possible that it may have had some effect on the development of life. Harland and Wilson (1956) and Rudwick (1964) considered that there was a causal relationship between the amelioration of climate at the end of the glaciation and the basal Cambrian transgression; the former provided a trigger for the rapid evolution of life at the beginning of the Cambrian. Whilst the effect of an ameliorating climate cannot be discounted there is little evidence to link a possible eustatic rise in sea level in early Cambrian times with the glacial episodes preserved in the stratigraphical record. On the other hand, the presence of distinct faunal provinces in the Cambrian may reflect the existence of high temperature gradients between the poles and the equator at that time. Such a situation, would tend to contribute towards a high rate of phyletic evolution because of the high provinciality and relatively small species populations within each province (Valentine 1968).

Whatever the effects of climatic factors on evolution I do not consider it likely that they were fundamental in controlling the development of life. A more plausible theory is that which relates major evolutionary advance to the gradual build up of atmospheric oxygen (Nursall 1959; Berkner and Marshall 1964; 1965; Cloud 1968a, b).

6. Conclusions

(i) Trace fossils first appear a short distance above the tillites and increase rapidly in abundance and diversity in latest Precambrian and Lower Cambrian times.

(ii) The sequence of trace fossil assemblages is partly controlled by changes of sedimentary facies but these are not responsible for the overall increase in trace fossils.

(iii) The incoming of trace fossils reflects the development of annelids, arthropods and molluscs rather than of metazoans as a whole.

(iv) Trace fossils are unlikely to occur in any rocks which are significantly older than 700 m.y.

(v) The build up of atmospheric oxygen is considered to be the fundamental factor controlling the development of life. Climatic factors were probably of only minor importance.

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