CHAPTER 6

NYBORG FORMATION

6.1 Introduction

The five members of the Nyborg Formation (section 2.3 and fig. 41) are lithologically distinct and can be readily identified in the field. The sedimentological history of the formation is conveniently divided into four phases which do not correspond exactly to the members:

1) Post-glacial transgression, member 1.
2) Basin fill, member 2, and most of member 3.
3) Transition from deep to shallow water conditions, upper part of member 3.
4) Shallow marine conditions, very top of member 3 and members 4 and 5.

This scheme is the basis for the description and interpretation of the Nyborg Formation in this chapter.

The area of study of each phase of deposition is delimited by significant erosion of the Nyborg Formation beneath the sub-Mortensnes Tillite unconformity (section 2.3, fig. 6), as well as the vagaries of present day exposure. As a result, members 1 and 2 can be studied over the whole area, member 3 occurs only around Tanafjord, the upper part of member 3 and the lower part of member 4 can be studied along the southeast coast of the Digermul Peninsula, and the entire member 4 and member 5 can be observed only at Trollfjord, where all 5 members of the Nyborg Formation are exposed.

6.2 Post-Glacial Transgressive Phase

Member 1

6.2.1 Introduction

Member 1 consists largely of purple and buff-yellow dolomitic shales, with dolomite developed locally at the base, and the

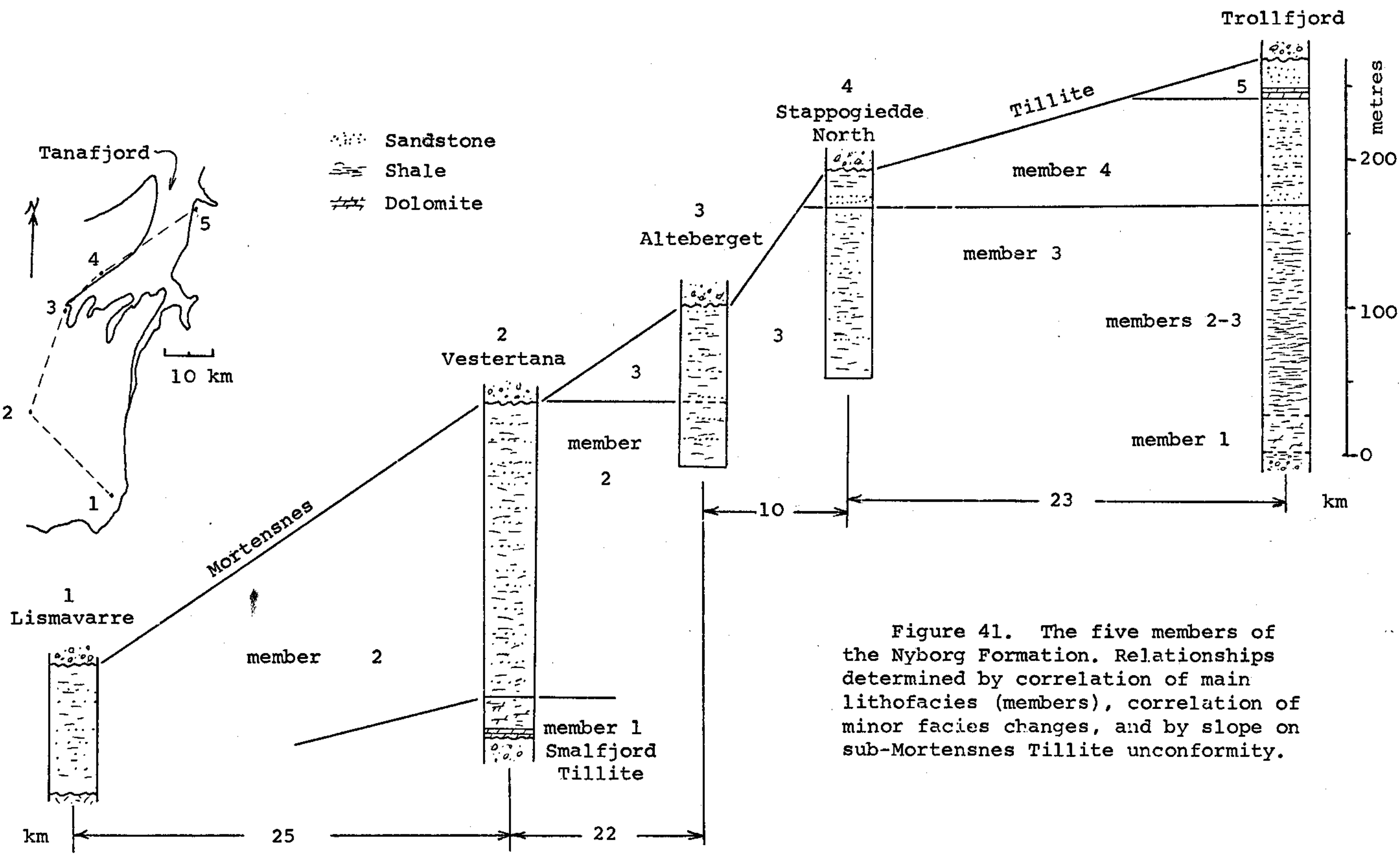

Figure 41. The five members of the Nyborg Formation. Relationships determined by correlation of main lithofacies (members), correlation of minor facies changes, and by slope on sub-Mortensnes Tillite unconformity.

proportion of dolomite decreasing gradually upwards. The member, 10-50 m thick, rests on the Smalfjord Tillite, or crystalline basement just SE of Skippagurra, (fig. 17) and is overlain by member 2 of the Nyborg Formation which is marked by the appearance of sandstone beds. The five facies in member 1 are described below, followed by an account of their occurrence in the Tanafjord, Laksefjord and Varangerfjord areas.

6.2.2 The Facies

Member 1 includes these five facies:

A) Dolomite: 0-2 m of dolomite occurs at the base of many sections. Type A1, which usually occurs at the base, is buff-weathering, and coarsely horizontally laminated. The lamination rarely forms convex structures up to 50 cm high, but usually less than 10 cm high, and about twice as wide. The laminae consist of alternating micrite (up to 5 mm thick) and of fine spar (up to 2 mm thick) with occasional lenses of coarse spar with silica-infilled cavities. The laminae of fine spar may divide laterally imparting a brecciated appearance to the rock. Polished specimens show the micrite as flat angular pebbles, rimmed with fine spar, which grades into coarse spar and silica filling the voids. It has the appearance of an intraformational conglomerate. Convex lamination is shown in Plate 41.

Type A2, usually above A1, is finely parallel-laminated dolomite, with occasional gentle scours and undulations. The lamination is caused by grain size variation. The thinnest laminae consist of angular grains of very fine quartz sand. The thicker laminae, up to about 2 mm, are composed of dolomite grains up to coarse silt size, but mostly smaller, with scattered grains of quartz silt, and mica flakes. When present, facies A2 grades up into facies B. Both dolomite facies weather buff-yellow. (see Pl.42).

B) Dolomite and Shale: This facies , usually 5-25 m thick, occasionally occurs at the base of member 1. It consists of alternating parallel laminae of purple mudstone and buff-yellow weathering dolomite micrite. The laminae may be a fraction of a millimetre to 3-4 mm thick. At only one locality, Mortensnes, (see below), ripple cross-lamination was observed. The proportion of dolomite invariably decreases upwards. The facies may grade up into facies C, or pass quickly into member 2.

C) Purple Shale: This occurs in most sections between facies B and member 2. It consists of finely parallel-laminated purple mudstone, with a large clay-sized fraction.

D) Dolomite Conglomerate: This facies was observed locally in most areas. It consists of dolomite conglomerate in lenticular, continuous or discontinuous beds up to 40 cm thick intercalated with either facies B or C. The beds are sharply, or occasionally erosively based, and may show slight grading. The conglomerate is clast-supported with the interstices filled with purple mudstone, or grey sandstone. The pebbles are faintly parallel-laminated micrite, plate shaped, up to 30 x 4 cm, usually much smaller and angular to subangular, occasionally rounded. Curved pebbles are rarely seen, but interpenetration was not. Shale flakes are rare. Pebbles are orientated subhorizontally, in which case they may be imbricated, or they may be haphazardly arranged (Pl. 44).

E) Sandstone: At two localities moderately to poorly sorted, fine grained sandstone occurs in member 1. Near Ellakløften (fig. 42), massive, fine, grey-green sandstone, 120 cm thick occurs above the basal dolomite, followed by about 5 m of shale and sandstones. Internal structures were not observed. At Smalfjord (locality 34, fig. 36) tillite of member E is overlain by 4 m of crudely bedded, drab-purple and green fine sandstone. Neither top nor basal contact was observed.

6.2.3 Facies Distribution

Tanafjord

At Rødberget (fig. 42) tillite, probably of member 5 of the Smalfjord Tillite is overlain by about 20 cm of fine

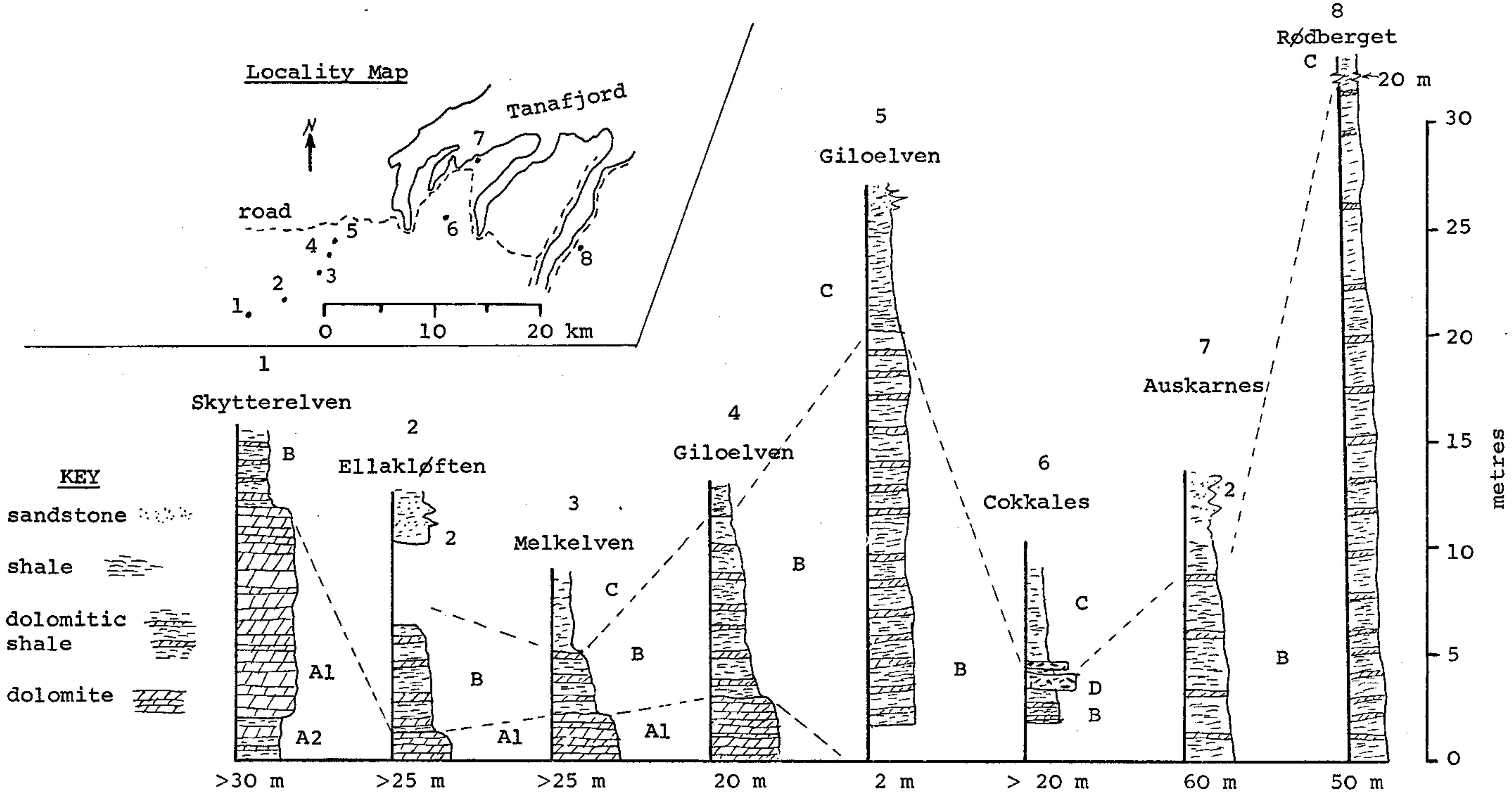

Figure 42. Distribution of facies of member 1 of the Nyborg Formation around Tanafjord. (A1, A2, B, C, D, are facies, 2 refers to member 2, Nyborg Formation. Number below refers to the thickness of the underlying Smalfjord Tillite).

sandstone with copper oxide film coating joint surfaces. After a gap comes facies B of member 1. Nearby, blocks of dolomite conglomerate outcrop on the top of the tillite, but contacts were not exposed. It is not clear whether this is part of the tillite, or facies D in the Nyborg Formation. The dolomite pebbles are white and equant, suggesting the Porsanger Dolomite rather than that typical of facies D conglomerate. However, within the lower part of facies B thin beds of dolomite conglomerate, facies D, occur. Facies B is here about 30 m thick, and is followed by about 20 m of facies C.

At Auskarnes (fig. 42) facies B occurs at the base of member 1, resting apparently gradationally above laminated tillite part of member E of the Smalfjord Tillite. Shale flakes in the basal part of the Nyborg Formation at Auskarnes suggest local erosion. Scattered crystalline clasts occur in the basal 10 cm of the Nyborg Formation. At Cokkales, (fig. 42), one thick bed of facies D occurs several metres from the base of member 1, here facies B.

West of Vestertana (localities 1-5, fig. 42), member 1 shows striking lateral changes in thickness of the component facies. While facies B thickens to the north, facies A thins out, as does the Smalfjord Tillite (see figure 5). Facies A is unusually thick at Skytterelven (fig. 42) to the west, where it sharply overlies coarse, conglomeratic sandstone, interpreted as fluvial, at the top of the Smalfjord Tillite.

At Trollfjord, the lower three members of the Nyborg Formation are not well defined. The lower 20 m is yellow and green banded, while the next 5 m is purple and yellow banded. The presence of the yellow dolomitic mudstone bands suggests inclusion in member 1, similar to facies B to the south. This dolomitic mudstone grades over a few centimetres from the parallel-laminated siltstone with outsized clasts at the top of the Smalfjord Tillite. These rocks are succeeded by about 150 m of purple and green siltstone which is attributed to members 2 and 3, although sandstone only appears in the upper part.

Laksefjord

The development of member 1 is similar to that in Tanafjord (fig. 43). At most localities south of Laksefjord, a basal bed of dolomite, usually facies A1, was observed to lie sharply above the Smalfjord Tillite (fig. 43). The basal dolomite bed of facies A1 at Aldoskaidde contains several convex mounds up to 50 cm high and 1 m wide, which may be of organic origin. Although the bed appears to have a gradational base caused by the increasing dolomite content upwards in the underlying Smalfjord Tillite, an erosive contact is suggested by the presence of both siltstone and tillite facies immediately below the dolomite. East of the Adamselv, the tillite immediately underlying the dolomite varies in lithology from place to place. At Cuobbojavrek where the Smalfjord Tillite appears to be represented locally by up to 15 m of clast free siltstone (Føyn, 1967, p.53), no dolomite occurs in the lowest part of the Nyborg Formation observed, usually separated by a small gap from the tillite.

Varangerfjord

Southeast of Skippagurra a small hill of crystalline rock just north of the main area of basement, and completely surrounded by bog is partly covered by a thin layer of tillite equivalent to the Smalfjord Tillite, and by patches of sandstone and dolomite of member 1 (fig. 17). At the base of member 1 is about 50 cm of faintly laminated fine-grained grey-green sandstone, overlain by up to 1 m of slightly tectonically deformed, pink and buff, parallel-laminated dolomite with calcite veins. It is similar to facies A2. The sediments dip up to 33°, concordant with the crystalline surface upon which they rest. Along the margin of the basement area to the south, patches of mudstone and siltstone

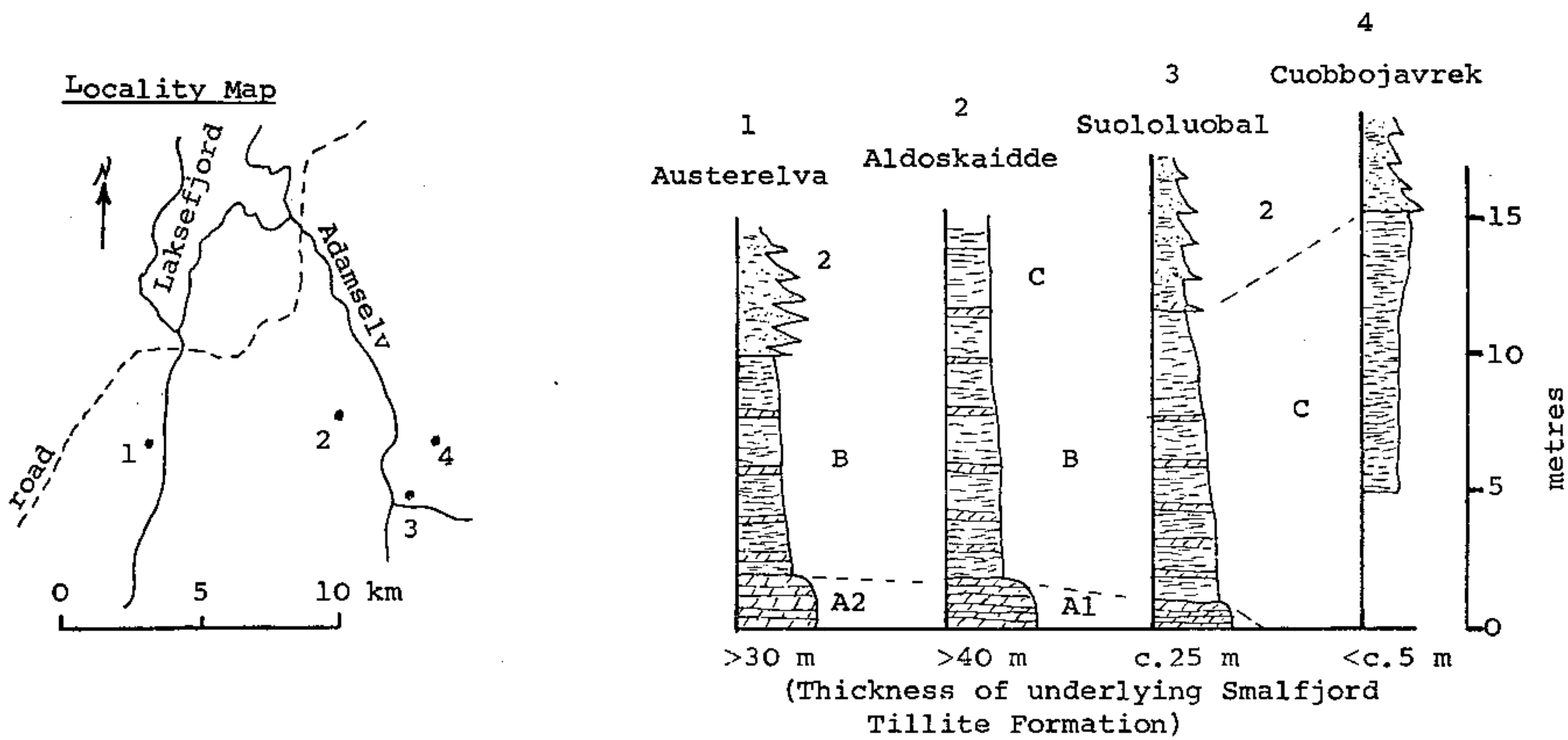

Figure 43. Distribution of facies of member 1 of the Nyborg Formation in the area south of Laksefjord. (A1, A2, B and C are facies, 2 refers to member 2, Nyborg Formation). Lithologies as in fig. 42.

occur on the crystalline rock surface and dip steeply.

In the coastal exposure 1 km west of Nesseby (fig. 17) member 1 is separated by a gap of 1 m from the Smalfjord Tillite. Member 1 consists of 1 m of facies B dolomite and shale followed by 3 m of facies B with beds of facies D dolomite conglomerate. The beds are tightly folded. These are overlain by poorly sorted sandstone and shale, part of member 2.

Above the Bergeby, in gorges 3 km from its mouth, are excellent exposures of undeformed, flat-lying members 1 and 2. In the lower 3 m of the section beds of facies D up to 60 cm thick (Pl.44) occur locally, interbedded with facies B, slightly dolomite mudstone. In adjacent exposures thin, sharp-based, occasionally graded beds of poorly sorted sandy siltstone with scattered dolomite and shale flakes (Pl.45) occur in the lower 5 m. These have been partly deformed by penecontemporaneous slurrying and faulting (Pl.46). Faults are normal, and dip to the north. Upwards, the proportion of dolomite clasts in the beds, and of dolomite in the matrix, decreases, grading into member 2.

Behind the village of Mortensnes, 1 km west of Mortensnes point, 5 m of member 1 outcrops in the hillslope; the basal and top contacts are not seen. Lenticular beds of facies D dolomite conglomerate are intercalated with facies B dolomite (Pl. 43), which here contains ripple cross-lamination. Some beds contain appreciable quantities of sand as a matrix, and in an upper layer, occasionally cross-laminated, overlying the dolomite pebbles and imparting a graded appearance to the beds. The orientation of fourteen ripple cross-sets showed that currents flowed to the west, north and east.

6.2.4 Interpretation

Facies

The most abundant facies in member 1, B and C, both consist of fine sediment suggesting deposition from suspension in quiet water, with delicate parallel-lamination and rare ripple cross-lamination at Mortensnes probably due to sporadic weak currents. Similar conditions, but with less supply of terrigenous mud, probably prevailed for the deposition of facies A2, and gentle scours attest to rare stronger current activity. The micritic texture of the dolomite in both facies A1 and A2 suggests a primary, or very early diagenetic origin for the dolomite (Folk, 1968).

The micritic texture of the intraclasts in facies A1 conglomerate was probably also due to deposition in quiet, occasionally agitated water, but the development of the conglomerate argues for strong currents. The fact that the conglomerate could form indicates that the upper surface of the micrite became consolidated soon after deposition. The formation of the conglomerate could be due to subaerial shrinkage, or erosion by currents. Subaerial shrinkage would also be consistent with the hardening of the dolomite soon after deposition, as is observed in modern intertidal and supratidal environments where evaporation is an important process. Ancient examples of desiccation cracks are usually filled with spar (Matter, 1967). The effect of strong currents in a shallow marine, offshore environment would probably be to form the shallow scours noted in facies A2. Whether or not the spar layers are associated with polygonal cracks could not be determined as no bedding plane surfaces in facies A1 were observed. Thus, facies A1 is tentatively assigned to a tidal (?intertidal or supratidal) environment. If the mound structures are organic (?algal) in origin, they support this interpretation. However, this must remain uncertain because of the absence of typical stromatolite structures and algal filaments.

Facies D, dolomite conglomerate, is not an <u>in situ</u> deposit formed by alternating strong and weak currents, in the same environment. The facies is developed in distinct beds, with the intervening sediment often lacking in dolomite. The large size of the clasts suggests that strong currents transported them. The sandy layer at the top of some beds may reflect the waning of the current. Clast-supported beds, the most common type, reflect bedload transport of pebbles, or the small proportion of fine-grained material in the current. Poorly sorted, matrix-supported beds may have been deposited entirely from suspension.

The lenticularity and discontinuous nature of the beds, and occasional matrix-supported beds suggests transport in high density, turbulent currents, which deposited their load rapidly, in a small area. The disrupted fabrics of some of the beds suggest that onee deposition began, the stationary pebbles exerted a strong drag on pebbles still in motion, and that where the current was sufficiently strong, the bed of pebbles may have flowed in response to the drag. Currents which deposited facies D may have been propogated by storms (Ball, 1971). Perhaps these currents were storm generated turbidity currents (Hayes, 1967).

The origin of facies E, sandstone, is uncertain due to the absence of sedimentary structures, and the difficulty of observing its relation to other facies.

Distribution of Facies

The lenticular form of the facies in member 1, and the discontinuous aspect of facies A dolomite may be a result of deposition on a surface with appreciable relief. The dolomite clasts in facies D were almost certainly derived from neighbouring areas where dolomite (facies A) was being deposited. It is reasonable to suspect that the dolomite originally formed in topographically high areas, and then subsequently resedimented to form the conglomerate (facies D) in lower areas. Consistent with this is the absence of shale flakes from facies A dolomite. These considerations strengthen the interpretation of dolomite facies A as inter- or supratidal deposits, and the more shaly facies B and C as relatively quiet (possibly deeper) water sediments.

Lateral variations in thickness of the Smalfjord Tillite, and the facies in member 1 add support to this interpretation. Facies A is present where the Smalfjord Tillite is well developed, but absent where the tillite is unusually thin or absent. Also, facies B and C are thickest where the Smalfjord Tillite and facies A are thinnest, but thin where the tillite and facies A are prominent. These relationships suggest that facies A developed over topographically high areas, where the tillite was thickest, and that facies B and C developed in topographically low areas where the tillite was thinnest (fig. 44). Also, facies D, dolomite conglomerate is developed in facies B and C where dolomite facies A is not present, suggesting that facies D was deposited in relatively deeper water. It appears that the relief of the upper surface of member 1, was less than that of its lower surface (fig. 44).

The vertical sequence of facies A to B to C thus indicates a successive increase in water depth, correspondong to a transgression. If this interpretation is correct, it implies that subaerial conditions were developed at the end of the Smalfjord Tillite glaciation over a considerable part of the study area.

Palaeoflow and Palaeoslope

Current directions obtained from cross-lamination and pebble imbrication in facies B and D at Mortensnes indicate currents flowed to the east, west and north. The orientation of soft-sediment faulting in facies B and D along the Bergeby indicate a northward dipping slope.

6.2.5 Conclusions

Member 1 was deposited during a transgression over an irregular surface. Dolomite was deposited on the highs in the intertidal or supratidal zone, while mud accumulated in the lows

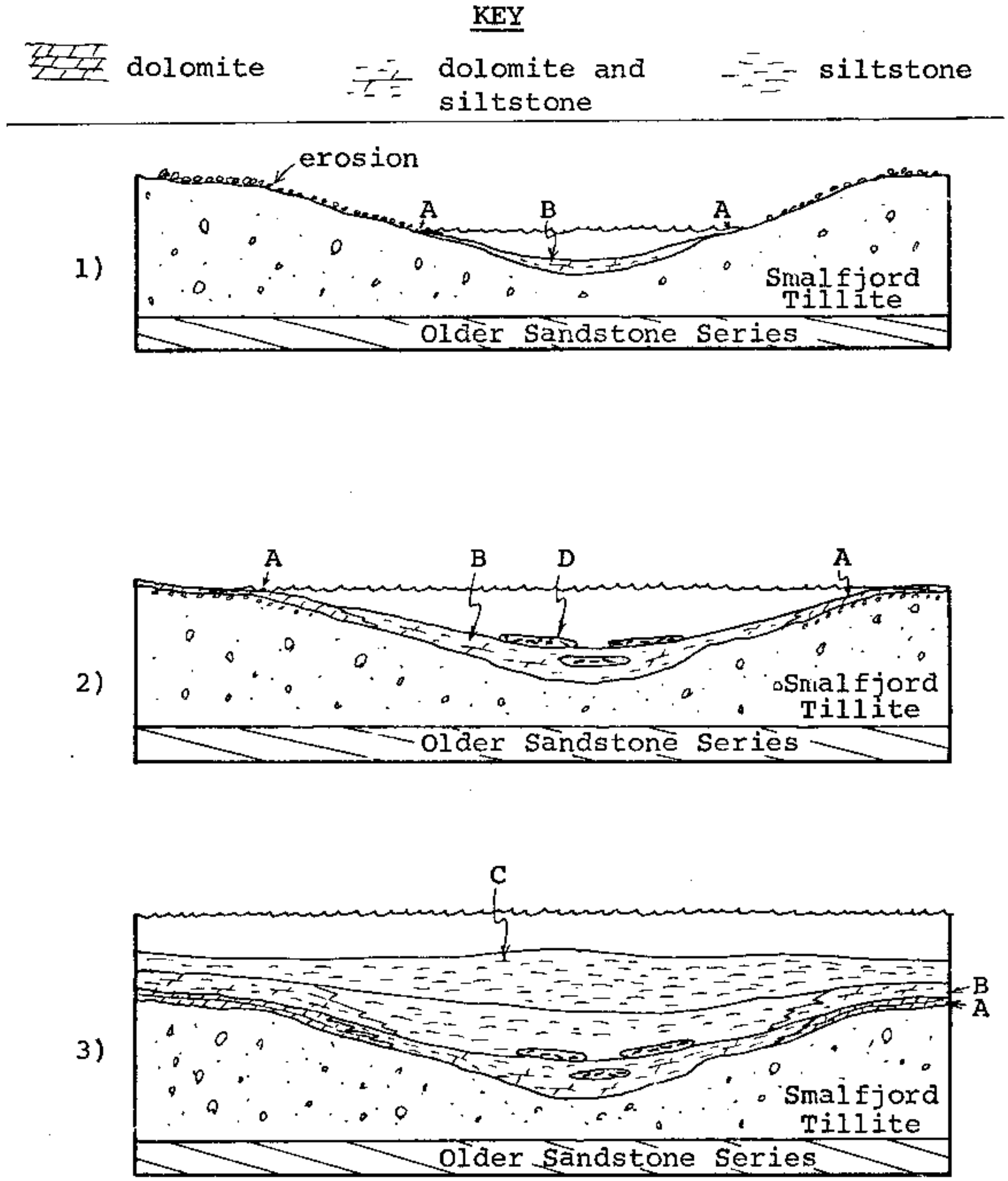

Figure 44. A model for the deposition of the facies in member 1 of the Nyborg Formation. 1) Facies A is deposited in shallow water, while facies B in deeper water. 2) Facies A is deposited on the 'highs', facies B in the 'deeps', and storms transport dolomite intraclasts into the deep environment. 3) Facies C is deposited over all areas. Steps 1-3 form a transgressive sequence, upon which member 2 was later deposited.

in quiet, deeper water. Occasionally dolomite interclasts were washed into the deep areas, possibly by storm generated density currents. As the transgression proceeded and the deposition of mud replaced that of dolomite as the water deepened over the entire area.

The relief of the area was lower at the end than at the commencement of deposition of member 1.

Basin Fill

6.3 Members 2 and 3

6.3.1 Introduction, Member 2

Most of the outcrop of the Nyborg Formation is of member 2. Its thickness cannot be accurately measured because it is almost always highly folded (Pl. 1). Its probable thickness is about 150-200 m. South of Vestertana it is eroded into by the Mortensnes Tillite, and east of Mortensnes it is cut out completely (figs. 6 and 41).

Member 2 consists of red-brown to grey sandstones intercalated with purple shale. In general the sandstones are usually fine to very fine grained, poorly to moderately sorted, in parallel-sided beds, 5 - 100 cm thick. There is an overall decrease in grain size from Varangerfjord to Tanafjord (Føyn, 1937). The sandstone to shale ratio is highly variable, but ranges from about 1/3 to 2/3. The high percentage of garnet in certain beds was pointed out by Føyn (1937).

6.3.2 Description, Member 2

Tanafjord

South of Vestertanafjord large areas are underlain by member 2 of the Nyborg Formation (fig. 2). Sandstone beds are mostly 5 - 30 cm thick and can be traced laterally up to 100 m, the extent of the exposure, without wedging out or changing thickness. Beds tend to occur in bundles, in which they are also thicker, and the proportion of shale is lower (Pl. 47). Bundles alternate with horizons that have fewer, and thinner sandstone beds. Such bundles and intervening zones tend to be about 10-30 m thick.

Sedimentary structures are not generally visible in inland exposures. In loose blocks and occasional fresh (glacially scoured, or cliff sections) exposures, internal structures can be seen clearly. Beds appear to have a sharp or erosive base, a sharp top, are poorly graded and often show the Bouma (1962) sequence T a-c* in the thicker, coarser beds, and the sequence Tc in the thinner beds. Flutes and shale flakes are rarely observed.

Along the coast east of Sjursjok (fig. 2), two types of sandstone beds can be distinguished near the base of member 2. One type, similar to typical intermediate turbidites, consists of sharp or erosive based, thin to medium graded beds with well developed T a-d and T b-d sequences. Most beds are composed largely of fine and very fine sand, but coarse and very coarse sand is present at the base of some beds. Two sets of ripples indicated currents to the north. The second type includes mainly thin beds of fine and very fine sandstone with gradational basal and top contacts. These beds are largely cross-laminated internally, and grading is not apparent. Intercalated with both types of sandstone beds is parallel-laminated silty mudstone.

In a small quarry along the coast north of Sjursjok two lithologies are observed in the upper part of member 2. The lower lithology consists of thin (5-10 cm) ripple and ripple-drift cross-laminated fine and very fine sandstones intercalated with ripple cross-laminated sandy siltstone. The ripples are about 10-15 cm in length, and about 2-3 cm high. The base and

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* Following the notation introduced by Bouma (1962) the letters a - e, when referred to turbidites, have these meanings: a lower massive, occasionally graded division; b lower parallel-laminated division; c ripple cross-laminated division; d upper parallel-laminated division; and e pelitic mudstone division.

top of the beds appear gradational. Above, the second lithology consists of thin (less than 5 cm), very fine sandstones, ripple laminated and parallel-laminated, intercalated with parallel-laminated silty mudstone. The ripples in the sandstone are mostly 10-20 cm in length, and 2-3 cm high. A few ripples, sometimes occurring as rows of isolated lenses in the siltstone are 2-3 cm in length, 2-3 mm high. On bedding plane surfaces these ripples are straight-crested and asymmetrical. Most beds in the second lithology have sharp or erosive bases, some with small flutes, and grading. Current directions appear to have been consistently to the southwest. Where a continuous section through the upper 40 m of member 2 was observed at Alteberget (fig. 2) alternations between these two lithologies are noted (Pl. 48). Green colouration appears gradually without any other notable changes, marking the transition into member 3.

At Rødberget member 2 is exposed in a section with numerous gaps, and some folding. The thickness is estimated to be about 50 m. Sandstones are absent in the upper 20 m; purple siltstone grades up into the grey-green siltstone of member 3.

At Trollfjord, member 1 (distinguished by the presence of dolomite) is followed by 100 m of alternating purple and green siltstone lacking sandstone beds. It is thus not possible to separate members 2 and 3 at this section. The total thickness between members 1 and 4 is about 150 m.

Laksefjord

Careful study of member 2 was not made south of Laksefjord. At Austerelva (fig. 43) about 20 m of member 2 are present between member 1 and Mortensnes Tillite. It consists of thin, fine and very fine sandstones, mainly parallel-laminated and cross-laminated, intercalated with purple mudstone. Some beds are well graded, sharp based and pass upward from cross-lamination into parallel-lamination. Yellow bands are present in the lower part, while green bands are present in the upper part. Ripple cross-lamination indicates that currents flowed to the northeast.

East of Aldoskaidde member 2 reaches at least 50 m in thickness. It is mainly thin to medium bedded fine sandstones.

Varangerfjord

The lower part of member 2 is exposed along the coast 1 km west of Nesseby, and along the Bergeby River in a gorge about 3 km from the mouth (fig. 17). West of Nesseby, 2 m of purple and green, poorly sorted sandy siltstone beds, intercalated with purple shale occur above member 1, 4 m above the Smalfjord Tillite. The beds are well graded with dolomite clasts at the base. Above are 2 m of graded and inversely graded beds with granules of quartz and feldspar, and numerous tabular dolomite clasts (Pl. 49). The beds are strongly erosive. Over about 10 m the grain size decreases to medium sandstone, bed thickness is about 20-40 cm, dolomite clasts die out, and mudstone intercalations appear. Beds are well graded, and highly erosive. Above, the grain size becomes finer, bed thickness continues to decrease, thickness of mudstones increases, and bases are less erosive.

Along the Bergeby, poorly sorted sandy siltstones occur within 3-4 m of the Smalfjord Tillite, above member 1. Beds are thin, graded, sharp based, with red shale flakes, and dolomite pebbles which die out rapidly up the section. Granule beds were not observed here. Beds are generally thin and fine-grained, and siltstone intercalations compose at least 50% of the thickness.

Between Varangerbotn and Skipagurra, along the basement, member 2 consists of thin to thick bedded poorly sorted sandstones, with thin mudstone intercalations. The sandstone locally contains granules and coarse sand grains scattered within the basal part of the bed, and mudflakes higher in the bed. Contacts are usually sharp and erosive; amalgamation of sandstones occurs. Away from the basement area, and probably higher stratigraphically member 2 consists of thin to medium bedded, rarely thick bedded sandstones, mostly fine to very fine grained. Contacts with the mudstone are sharp, and grading was not observed.

At Hamarnes, (fig. 17) beneath the Mortensnes Tillite, member 2 is again coarser. The level is probably about 50 m above member 1. Sandstones are fine to medium grained and contain well developed Bouma sequences (Pls. 50 and 51) with excellent fluted bases (see Bjørlykke, 1967) indicating current flow to 280° and 290°. Similar beds are exposed in the roadcut just to the northwest. However, they are eroded into by up to 3 m of fine to very coarse sandstone (Pl. 52), which is in turn eroded into by the Mortensnes Tillite. A similar coarse sandstone facies was seen southeast of Skipagurra. The sandstones are thin to medium bedded, with little intercalated mudstone. The erosion surface into the underlying facies is steep, overhanging at the east end of the outcrop, but horizontal over the rest of it. The overhang trends approximately N-S. Internally sandstones show grading and inverse grading, and occasional parallel-lamination, cross-bedding, and soft sediment deformation, mainly loading and ball-and-pillow structure. Cross-bedding azimuths recorded are 340°, 310°, 330°, 340°, 330°, 345°, and one ripple cross-lamination to 300°.

At Tan~~y~~ Bru (Bridge) 4 km north of Skipagurra good exposures a of member 2 occur in the east bank of the Tana River. The section consists of thin to medium lenticular sandstones, rarely thick, graded and inversely graded with coarse and very coarse sand near the base of the bed, rare shale flakes and few mudstone intercalations. Amalgamation is common (Pl. 53). Most beds are internally massive, in some cases with a parallel-laminated division (b) above. Cross-bedding was seen at the base of one bed (Pl. 54). The set is 10 cm thick, and the foresets dip 30° to the north. Foresets are planar, and gently concave at the base. Grain size is coarse sand. The top of the set is eroded into, and slightly deformed. The sandstones here contain about 10% feldspar, 20% rock fragments (mainly polycrystalline quartz), about 7% garnet, and 10-15% matrix. Finer-grained beds tend to contain a large proportion of biotite flakes altered to haematite.

6.3.3 Interpretation, Member 2

The monotonous intercalation of laterally continuous sandstones in mudstone in a sequence over a hundred metres thick is immediately suggestive of deposition by turbidity currents. In certain inland exposures and coastal outcrops, the lateral continuity and certain sedimentary structures: erosive, occasionally fluted base, grading, shale flakes, and various kinds of Bouma sequences indicate that the sandstones are turbidites. This is consistent with the scarcity of evidence for current activity in the associated mudstones. Poor inland exposure and unsuitable weathering, and the overall fine grain size of most of the sandstones may account for the relative scarcity of grading and of a gradational upper contact. These sandstones do not have a typical turbidite appearance (as noted by Bjørlykke, 1967), but probably are turbidites.

However, some sandstones have gradational contacts, are ungraded, and are intercalated with siltstones which show signs of current activity. This is the case at Sjursjok in parts of member 2. This facies was described by Reading and Walker (1966, pp.189 and 193, fig. 7) as their facies C-2, which they attributed to occasional current activity in quiet water. The present author is in agreement with their interpretation.

Straight-crested ripples indicate oscillitory water motion, caused by waves. As such ripples have been reported at depths of 2,000 m, they need not indicate shallow depths. The length of the ripples corresponds to the displacement of sediment during the passage of the waves (Harms, 1969). In view of the rarity of wave-formed ripples, the water depth must have been below wave base for most waves. Considering that current velocities on the order of 10 cm/sec are capable of transporting coarse silt and fine sand (Allen, 1970), deposition of member 2 in a depth of 100 m is not unreasonable as velocities of about 30 cm/sec occur on one day a year, at depths of 40 m in protected shallow shelf areas (Draper, 1967).

In the upper part of the sequence at Nesseby, graded sandstones intercalated with mudstone are probably turbidites (intermediate or distal), while the coarse beds below, with scoured bases and no mudstone intercalations appear to be very proximal turbidites or "fluxoturbidites". Such deposits may have formed in submarine fan channels, or at the channel-open fan junction. The exposure does not allow the geometry of the facies to be determined; according to Stanley (1969) fluxoturbidites deposited in submarine channels should have a shoestring geometry. On the other hand, migration of such channels, if it occurred, would form a sheet-like deposit.

The transition from members 1 to 2 at this locality is marked by a change in clast composition from dolomite (facies D, dolomite conglomerate, derived from facies A dolomite), to terrigenous quartz, feldspar and crystalline rock fragments.

The coarse sandstones eroding into intercalated sandstone and mudstone near Hamarnes is also atypical of turbidites. However, they appear to have some of the features of "fluxoturbidites" as compiled by Walker (1970, p.235). When compared with the adjacent turbidites, the coarse sandstones are seen to consist largely of the coarsest components of the available material. In light of the erosive base of the coarse sandstone unit it is suggested that they are a submarine fan channel deposit. The only characteristic of fluxoturbidites which Stanley (1969, pp. DJS-9-7) felt to be diagnostic, a shoestring geometry, cannot be tested in this case either. Nevertheless, the sandstone has those properties originally defined by Kuenen (1958) as diagnostic of fluxoturbidites; "coarse grain, thick bedding, poor development of grading, sole markings, and of shales" (in Stanley, 1969, p. DJS-9-6), especially in comparison with the adjacent turbidites.

A facies similar in some respects is seen at Tana Bru which may represent proximal fan deposits. The lenticularity of the beds, coarse grain size, and scarcity of shale interbeds are all typical features of proximal turbidites, while the presence of grading suggests an intermediate position between proximal and distal (Walker, 1967). Cross-stratification has been described from turbidite deposits of coarse grain-size (see Allen, 1970, for references), and have been discussed theoretically by Allen (1970), who concluded that the structure is compatible with the turbidity current model and will form if sufficient coarse material is entrained by the flow in the appropriate hydraulic conditions.

Palaeocurrents

Palaeocurrent directions determined from flutes, and dune and ripple cross-stratification indicate that currents flowed mainly to the north in the Tanafjord area, and to the west and northwest in the Varangerfjord area. An important departure from this trend are the southwest currents shown by cross-lamination in the upper part of member 2 at Sjursjok. It is apparent that turbidity currents flowed to the north and northwest, while other types of currents flowed to the southwest. The nature of these currents is uncertain.

The rarity of wave formed ripples, the fine grain size, and the restriction of significant current activity to the turbidite sandstones suggest that deposition was in greater depths than at which most waves were capable of agitating the bottom.

6.3.4 Member 3, Description

Due to lack of exposure and the overlying unconformity (see Chapter 2), member 3 is nowhere exposed in its entirety. Near Alteberget (fig. 2) 75 m are present above member 2, and along the southeast coast of the Digermul Peninsula (fig. 5) 125 m are present below member 4. The two sections cannot be correlated.

In all exposures seen, the lower part of member 3 represents a continuation of member 2, but with a grey-green rather than reddish-brown colour. North of Sjursjok the change in colour is gradational, the two colours alternating for up to about 20 m. Member 3 sandstones are mainly thin, occasionally medium bedded, fine to very fine grained, and often occur in bundles. In most of the area beds have a planar, sharp base, but in the section at Sjursjok, a fluted base is very common. Three beds show flutes which indicate that the current flowed to the north. Well exposed sections north of Sjursjok show many ripple cross-lamination current directions to the west, northwest and north.

At eastern localities, Rødberget, Leirpollen and Lavvonjargga, (fig. 63), member 3 consists of grey-green, laminated, flaggy siltstones, with occasional flat straight-crested ripple marks on parting planes.

Where the upper part of the member is observed, at Stappogiedde and Trollfjord, it coarsens upward, over about 50 m into the overlying member 4, dominated by purple sandstones and associated lithologies (see section 6.5).

6.3.5 Member 3, Interpretation

The similar sedimentary features of member 3 to member 2 over most of the area suggests that member 3 is also largely a turbidite deposit. The deep water origin is supported by the upward transition into a shallow marine environment (see below, section 6.5). However, at eastern localities, the coarser grain size, siltstone, and the presence of ripple marks suggest that these were deposited in shallower water. As with member 2, the predominating direction of transport was to the north.

No changes in sedimentary features were observed concommitant with the change in colour for which there is no apparent cause.

6.3.6 Discussion and Conclusions

Member 2 and the lower part of member 3 were deposited largely by turbidites, but also in part by other marine currents. Within the turbidite group of facies, fluxoturbidites representing ancient fan channels, and proximal turbidites representing proximal fan deposits are abundant in the south of the area, particularly in the lower parts of the section. To the north, around Vestertanafjord, turbidites are intermediate to distal with shale interbeds ubiquitous throughout the section.

To the east, around Leirpollen, the upper part of member 2, and the part of member 3 preserved beneath the sub-Mortensnes Tillite unconformity lack turbidite sandstones. The straight-crested ripples and coarser grain size suggest shallower conditions of sedimentation than around Vestertanafjord.

Palaeocurrent directions obtained from flutes, cross-bedding and cross-lamination indicate that turbidite currents flowed to the NW in the Varangerfjord area, and to the north in the Tanafjord area. Currents other than turbidity currents flowed to the SW, as seen at Sjursjok in the upper part of member 2. The lateral facies changes, palaeocurrents, and the overall decrease in grain size to the north indicate that member 2 and the lower part of 3 were deposited by submarine fan channels along the southern margin of the area, which fed submarine fans (probably coalescent into an apron as suggested by the uniform palaeocurrents) building out to the north. This, in addition to the shallower conditions in the north and east, may explain why turbidites are lacking from member 2 and the lower part of member 3 at Trollfjord. While the palaeoslope had a pronounced dip to the north in the southern part of the area, it may have been negligible or dispersed slightly to the south in the northern part of the area.

The rapid sedimentation indicated by the immature composition and texture of the sandstone, and the turbidite mode of deposition apparently succeeded in filling the basin; the upper part of member 3 described in the following section, records the gradual changeover to shallow marine processes which controlled the deposition of members 4 and 5.

6.4 Regressive Phase

Upper Part of Member 3

6.4.1 Introduction

The upper part of member 3 at Stappogiedde North, Digermul Peninsula (fig. 45) is a 55 m thick transition between the lower, deep water part of member 3 and the shallow water, subtidal deposits of member 4. In this account of the regression, two groups of facies are distinguished: background and sandstone beds. Background sediments are finer than the sandstone beds, and are composed largely of clay and silt grade material. As set forth in the preceding sections on members 2 and 3, the sandstones are interpreted as turbidites, while the shales largely reflect the "normal" conditions at the site of deposition. A study of each of these two groups of facies independently leads to an understanding of how the regression took place, and by combining them, a unique pattern of sedimentation can be seen.

The sequence was described briefly by Reading and Walker (1966) who also interpreted it as a regressive sequence, with turbidity current deposition in quiet water giving way upwards to mudstone deposition in shallow, agitated conditions.

Brief observations were also made at other localities on the Digermul Peninsula and at Trollfjord (fig. 5).

6.4.2 Background Sediment

Facies

Background sediment is subdivided into three facies which are readily identified in the field. These are:

Facies 1: parallel-laminated mudstone (Pl.55), composed of alternating fine laminae of silt and clay grade sediment. The facies is typically cleaved and was probably deposited from suspension, in the absence of wave and current activity.

Facies 2: rippled siltstone and very fine sandstone (Pls.55, 56, 57). Ripples occur as cross-laminated lenses, or very thin lenticular beds surrounded by mudstone. In cross section the ripples are symmetrical to asymmetrical, with rounded tops, and are < 1 cm high, and < 15 cm in length, averaging about 7 cm (fig. 46). The ripples are often in sharp contact with the surrounding sediment, and tend to be of the same, or slightly coarser average grain size than the sediment they are immersed in. The few bedding plane exposures seen in the lower part of the section, show straight-crested ripple marks while a wide variety of ripple marks were seen near the top (see below). Cross-lamination dips predominantly to the north, but also to the south. This facies was probably formed by wave and current ripples (Harms, 1969) (see below).

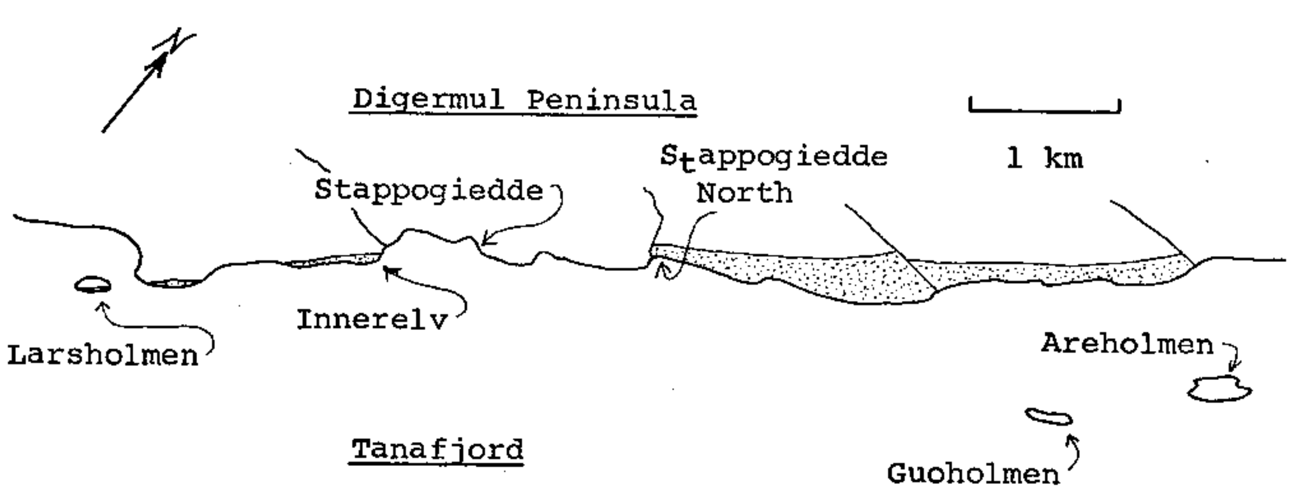


Figure 45. Localities in the Nyborg Formation, dotted, southeast coast, Digermul Peninsula, Tanafjord

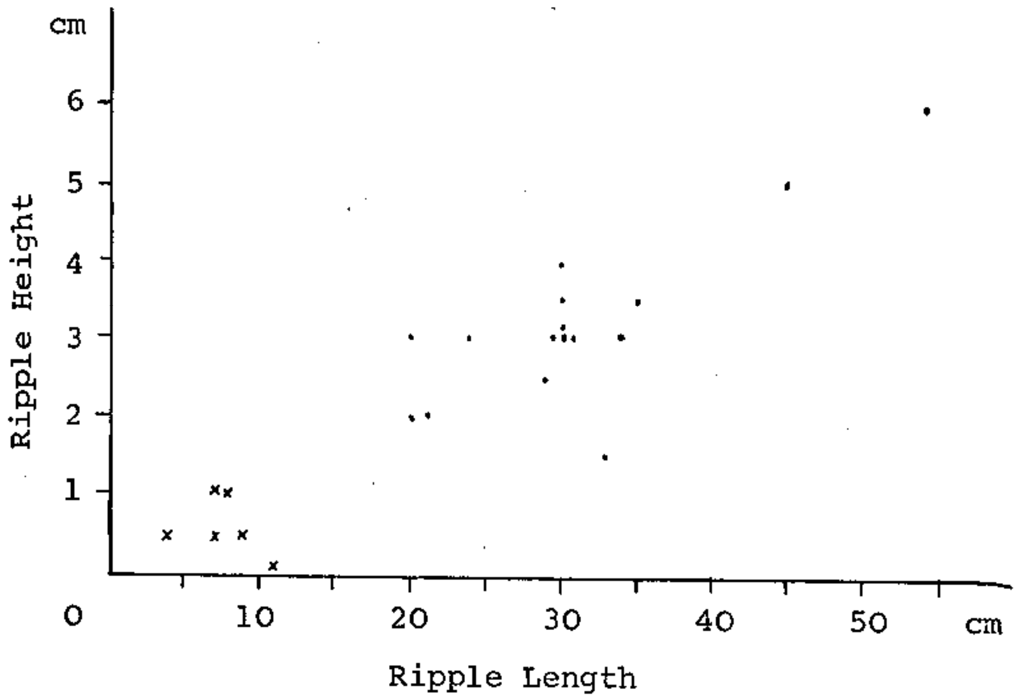


Figure 46. Measurements of height and length of ripples in the background sediment (×) and in facies c of the sandstone beds (·).

Facies 3: parallel-laminated siltstone and very fine sandstone (Pl.58). The sediment occurs in parallel sided or slightly lenticular laminae up to 5 mm thick. The facies is typically uncleaved. Each lamination may be internally finely laminated, or appear massive; grading was never seen in this facies. Deposition was probably largely from suspension, possibly in the presence of weak currents. This is suggested by the fact that the facies grades into facies 1, and that the laminae often have the appearance of a drape. Reineck (1963) has shown that finely laminated sand in the shallow parts of the German North Sea can be attributed to the action of waves placing sand into suspension.

Vertical Distribution of Background Facies

Based on the type and proportion of facies present, the sequence has been divided into four units (fig. 47):

Unit A: lacks ripples, consists almost entirely of facies 1.

Unit B: consists largely of facies 1, but facies 2, ripple cross-lamination is uniformly present (Pl.55). The contact between units A and B is about 55 m below the base of member 4.

Unit C: consists of a mixture of facies 1, 2 and 3 (Pl.56) in varying proportions. Facies 3 becomes important in the lower part, and is dominant at the top, while facies 1 decreases rapidly up section (fig. 47). Facies 2 increases slightly upwards. The boundary between units B and C is placed at about 35 m below the base of member 4.

Unit D: consists mainly of facies 3 (Pls. 57 and 58). Facies 2 is also present, but facies 1 consists only of thin mudstone partings between laminae of facies 3, the two facies noticeably different in grain size (Pl.58). The boundary between units C and D is about 15 m below the base of member 4.

A wide variety of ripple marks were observed in units 3 and 4 (Pl.59). These include straight and sinuous-crested forms, with linguoid ripples uncommon. Most ripples are asymmetrical in profile, and have rounded crests. Wave lengths are mostly 5-10 cm, and heights are generally < 3 mm. The axes are strongly aligned E-W (fig. 48A) and are asymmetrical mainly to the north, but also to the south. The presence of straight crests, and asymmetry indicates a combined flow origin for most ripples (Harms, 1969), but linguoid ripples formed under current dominated flows, while symmetrical ripples formed under wave dominated flows. The bimodal palaeocurrent directions suggest the activity of tidal currents. Similar trends are observed in the same part of the section south of Innerelv (fig. 48B).

Significance of the Vertical Distribution of Background Facies

Two rather distinct processes seem to have been important in the deposition of the background sediment. These are:

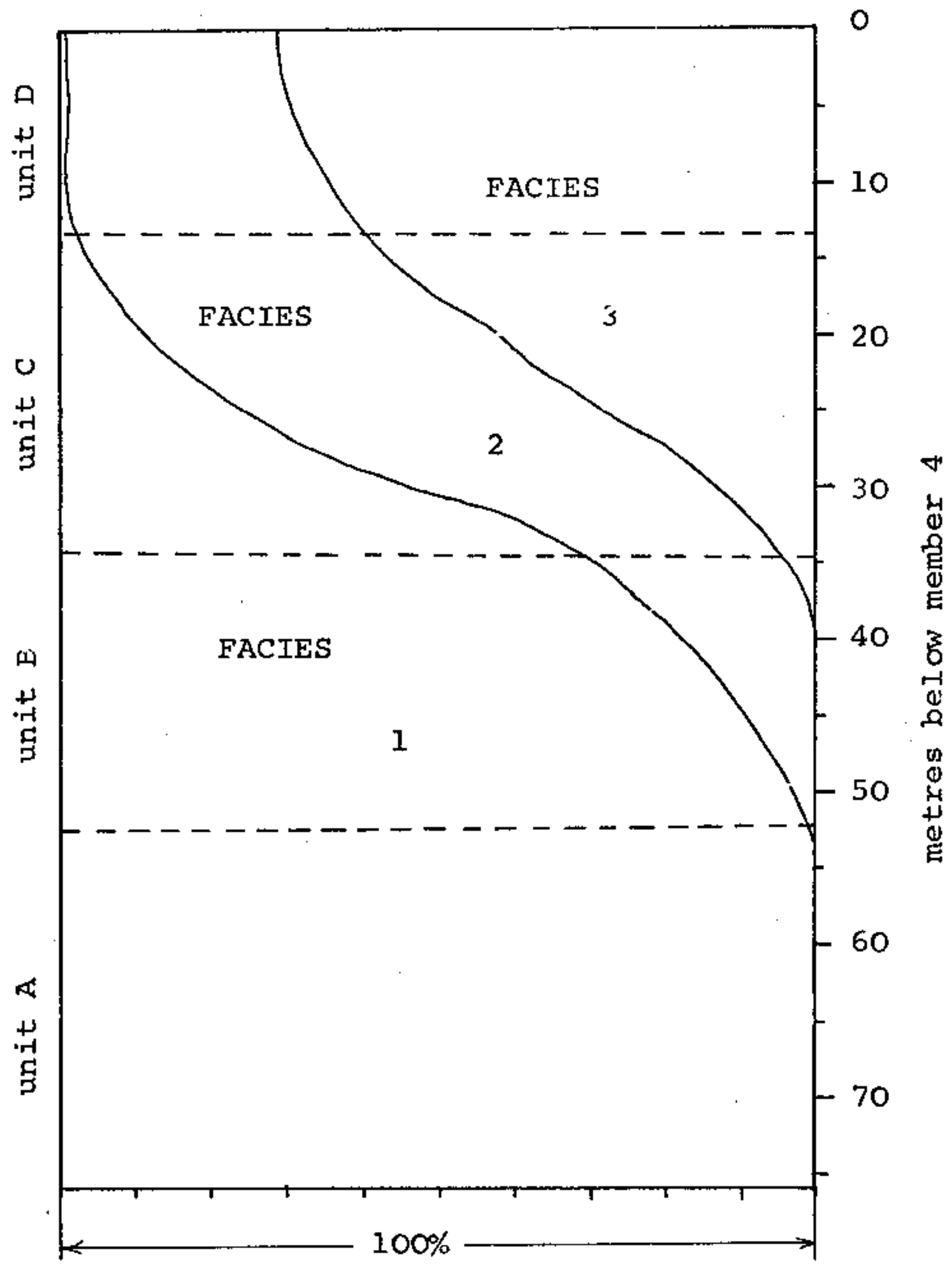


Figure 47. Diagrammatic portrayal of the subdivision of member 3 background sediment into facies and units. Facies 1 is parallel-laminated mudstone, facies 2 rippled silt and sand, and facies 3 parallel-laminated silt and sand. There is an overall increase in grain size and wave and current agitation upwards.

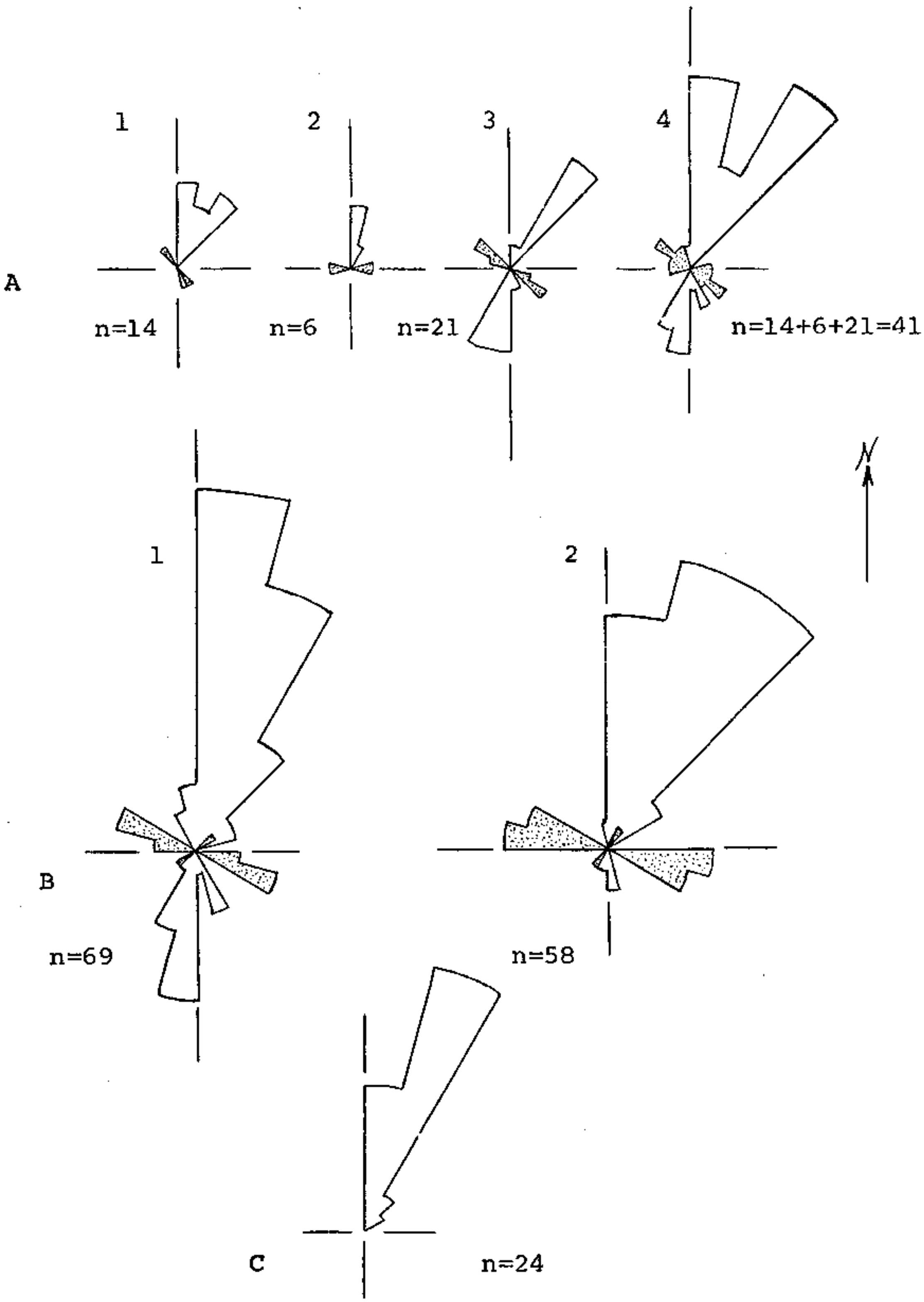


Figure 48. Palaeocurrents, upper part of member 3, Nyborg Formation, Digermul Peninsula. A: ripple marks in background and on reworked sandstones at Stappogiedde North. 1) about 20 m below member 4, 2) about 16 m below member 4, 3) top 8 m of member 3. B: ripple marks in top 20 m of member 3; 1) south of Innerelv, 2) north of Larsholmen. Plain area: combined flow and current ripples (asymmetrical). Dotted area: axes of symmetrical ripple marks. C: direction of sole marks,

fallout of fine sediment from suspension (facies 1 and 3) and rippling of the bottom sediment (facies 2). The similarity in grain size between the ripples and the adjacent non-rippled background sediment suggests that ripples formed from the reworking of the bottom sediment, and that their formation was not associated with the influx of new sediment. This is visible particularly in unit D (Pl.60). The ripples in facies 2 form a population distinct from the ripples in the sandstone facies (described below) in terms of length and height (fig. 76). From the overall increase in importance of facies 2 upwards (fig. 47), it appears that wave and current agitation increased up the section. Because of the contrast in dimensions, ripples in the background can be distinguished from those in the sandstone beds (this is important in unit D where the grain size of the two facies groups is very similar).

The deposition from suspension of both facies 1 and 3 suggests that they were deposited by the same event which had different effects in different places. One lamina of facies 3 may have been deposited in a shallow environment, while a corresponding lamina of facies 1 may have formed in a deeper environment. Thus a lamina of one would be expected to pass laterally into a lamina of the other. Such a deposit could form by the offshore movement of a mass of water with suspended fine sediment: first the coarse fraction, and then progressively finer sediment would be deposited as turbulence waned, possibly related to decreasing wave agitation.

According to these interpretations, wave activity was responsible mainly for modification of the bottom into a ripple form, and for maintaining fine sediment in suspension. The dominant factor in sediment transport would have been the offshore movement of sediment-laden masses of water. Such large-scale movements of water may be related to storm activity, see section 6.5).

Studies cited by Potter and Pettijohn (1963) indicate that ripple crests in ancient and recent sediments tend to parallel the coastline. This suggests that the ancient coastline for member 3 was aligned WNW-ESE.

6.4.3 Turbidite Sandstone Beds

In the 72 m of section measured along the coast north of Stappogiedde North (fig. 45), 394 beds were measured and described. To warrant inclusion as a bed, a limit of a minimum thickness of 1 cm was arbitrarily set. The beds consist of poorly to moderately sorted silty very fine sandstone, with fine sand and medium sand present at the top of the sequence. The grains are mainly angular to sub-rounded quartz with some feldspar, muscovite, haematite and leucoxene. Tourmaline and zircon are rare. The matrix is a fine-grained groundmass of carbonate and micas. Heavy minerals occur scattered about, rather than in laminae.

Almost the entire number of beds is composed of the following four sandstone facies: massive (Bouma division a (1962)), parallel-laminated (b), ripple cross-laminated (c), and thin (1-5 cm) sandstones which grade up from massive sandstone to siltstone (a->e), (Walker, 1965). Soft-sediment deformation is infrequent.

Sole marks formed by current scour were observed on the bases of 30 beds, and current directions could be taken from 24 of these. Flutes were observed under 23 beds, longitudinal furrows were observed once. These scour marks indicate the turbidity currents flowed consistantly ENE (fig. 48C). This was probably the dip direction of the palaeoslope.

<u>Facies</u>

Massive Sandstone (a)

Massive sandstone forms the lower part, or the entirety of many sharp-based, parallel-sided sandstone beds (Pl.61). It may pass up into parallel-laminated or cross-laminated sandstone. The division is usually 4-10 cm thick, but the maximum observed was 32 cm. Because it underlies the parallel-laminated division, it is thought to have originated in the upper flow regime (Walker, 1965).

Parallel-Laminated Sandstone (b)

This division occurs in both parallel-sided and lenticular beds, mainly as T_{a-b}, and T_{b-c} sequences respectively (Pls. 61, 62). It occurs rarely over a fluted base, most often over a sharp base, and is usually 6-10 cm thick. Primary current lineation was not observed, but internal parting surfaces are rare due to the jointing and lack of fissility. The facies was probably deposited in the plane bed flow of the lower part of the upper flow regime.

Ripple Cross-Laminated Sandstone (c)

This division occurs more frequently at the base, than within sandstone beds (Pl.62). Such beds have most of the flutes that were observed in the sequence. The division tends to be about 2-4 cm thick, the height of the ripples present.

Most beds formed largely of division c consist of a series of laterally spaced-out ripples which due to the variation in crest and trough heights, imparts a pinch-and-swell appearance to the bed (Pl.63) (pinch-and-swell has been used in a different sense by Walker (1969)). The base of such beds is often undulatory, apparently due to subsequent compaction, and possibly, to a small extent, due also to the scouring of flutes. The upper surface of the beds is mostly sharp, but may be gradational where concordant to the internal lamination on the lee side of a ripple, or erosive where discordant on the stoss side of a ripple (Pl.62).

Ripple length is 20-54 cm, averaging about 30 cm and heights of 2-4 cm prevail, the ripples being distinct in size from those in the background sediment (fig. 46). In the only bedding plane exposure observed, crests are continuous and slightly sinuous.

While most "swells" are composed of one ripple with cross-lamination dipping about 10° to the north, and slightly concave upwards, there are in some cases minor erosion surfaces separating ripple cosets within the bed. Where these internal scours are developed, the overlying lamination may dip gently to the south, presumably in the upcurrent direction. Ripple-drift cross-lamination was seen once.

Rippled sandstone was obviously formed by the migration of ripples, attributed to the lower part of the lower flow regime. During deposition, most ripples migrated downcurrent with erosion on the stoss side and deposition on the lee side. The rounded crests and gentle dip of the foreset lamination indicate that a well-defined separation eddy did not form in the lee of the ripples. The high rates of deposition from suspension ascribed by Jopling and Walker (1968) to very rounded (sinusoidal) ripples associated with ripple-drift cross-lamination is not consistent with the absence of ripple-drift in the present sequence. Current ripples with rounded crests have been produced in a flume by the addition of fine sediment (clay) to the load (Simons et al., 1963). The rounded form was attributed by these workers to the increased cohesion of the bed.

The present author is inclined to believe that important changes in ripple morphology may occur when the sediment is composed largely of the finer spectrum of material capable of being carried as bed load (coarse silt). In such a situation the distinction between transport as bed load versus suspended load becomes indistinct, and the validity of Allen's analysis of ripple-drift, which considers bedload transport and deposition only, ignoring suspended load, may be open to question. The virtual absence of ripple-drift, as mentioned above, suggests that the ripples were reworking the bed, and did not form while the rate of deposition was high, a condition which is required for the development of ripple-drift cross-lamination.

A particularly difficult case to account for is the occurrence of apparently upstream (southward) dipping laminae in one bed (Pl.64). The current direction is extrapolated from the adjacent beds which contain flutes all which indicate a current direction to the north. The lamination occurs directly above a thin layer of massive sandstone. If the laminations are backsets, further investigations into the behaviour of silt and very fine sand, with varying concentrations of suspended sediment in the flow, may explain this occurrence.

Massive Sandstone Grading to Siltstone (a->e)

This facies consists of thin (mainly 1-5 cm) parallel-sided beds with massive very fine sandstone at the base grading up into massive siltstone (Pl.65). Sole marks were observed on the base of two beds; in other cases the base appeared sharp. The siltstone at the top of the bed appears to grade into the overlying background sediment.

The beds may have formed rapidly, largely from suspension, with insufficient time for stable bed forms to develop (Wal~~ton~~ker, 1967).

Other Features

In addition to the four main facies, other structures, forming only a minor part of the sequence, were observed. Most prominent is soft-sediment deformation which is shown by convoluted lamination and ball-and-pillow structure (Pl.66). These have been included as "slumps" in figure 49*. It is uncertain whether or not true slumps occur in the sequence.

In the upper part of the sequence the effects of reworking become more important as erosion of the beds is obvious. A few beds have been cut by deep scours (Pl.67) which suggests much more catastrophic erosion than that by wave agitation (Pl.60).

Turbidite Facies in the Vertical Sequence

Discussion

The increasing grain size and agitation upwards in the background sediment has been interpreted as reflecting progressive shallowing of the sea. Although the shallowing could have occurred by vertical sedimentation alone, the presence of turbidites suggests that a slope was present. The slope may not have been necessary for the propogation of the turbidity currents, but probably would have been required for their maintenance (continuous supply of energy).

It thus seems likely that the transition from deep to shallow water conditions was associated with the forward progradation of a submarine slope across the area. It is uncertain how important subsidence and vertical movements of sea-level were, compared to the rate of sedimentation. Considering the distance of 55 m betweeen the base of member 4, which must have been near sea-level, and the lowest occurrence of oscillation ripple marks, there is no apparent reason to infer an important change in relative sea-level during deposition of the sequence. It is no surprise that Reading and Walker (1966), who determined a maximum thickness of 9 m for the regressive, shallowing sequence invoked a rapid drop in sea-level (p.193). These authors evidently considered the dying out of turbidite sandstones near the top of the sequence as indicative of shallowing, rather than the appearance of agitation and upward coarsening of the background sediment, first noted 55 m below member 4.

Two assumptions, stated below, allow the vertical sequence deposited by 394 (recorded) turbidity currents in the same place over a long period of time, to be viewed as a large area of the deposit of one "average" turbidity current formed during a relatively short period of time. These are:

1) During the forward progradation of the slope, in the time interval studied, environmental conditions which could influence the development of the turbidity currents did not change significantly, and

2) the 394 beds represent different depositional stages of the same "average" turbidity current; only one kind of turbidity current was responsible for the deposition of all of the beds.

Thus, beds at the top of the sequence provide a view of the currents during their early stage in shallow water, and further down the sequence, the change in conditions as the currents flowed downslope into deeper, quieter water can be observed.

Some Aspects of Turbidity Currents and Their Deposits

A turbidity current in an overall depositional phase loses power because of the decrease in density compared to the surrounding fluid of constant density (mixing with the fluid also contributes to dilution but this may be offset by erosion; rates of erosion and deposition may be low where transportation is by autosuspension, see Middleton, 1966C, 1969). The loss in power is reflected by the downstream lateral decrease in maximum grain size, and the horizontal succession of sedimentary structures indicating decreasing flow regime. Even if a current is increasing in velocity down a steep slope, deposition may ensue by the passage of the current over a given point; the turbidity current is most powerful in the region of the head, and the power decreases rapidly into the body and the tail of the current (Middleton, 1967).

Aside from the effect of a decrease in density of the current with time, the slope over which the current is flowing can vary with distance. A sufficiently steep slope can offset the decay of power so that maximum flow power generated by the current in the head region will increase with time. As a result, increased erosion may take place beneath the head, and a higher flow regime will develop in the depositing part of the current.

The vertical sequence of sedimentary structures at one point indicates deposition by a waning current; the locus of deposition is first within the head region, and then in the body and tail regions. This effect also may cause graded bedding. The sedimentary structure preserved at the base of the bed is an indication of the bed form on the bottom when the current began to deposit its load. For a constant grain size, the basal bed form is a guide to the relative maximum power developed in the turbidity current at the time when deposition began (Allen, 1970). Beds that begin with facies b or c are referred to as "base-absent" sequences (Walker, 1965).

Some turbidites lack the overlying facies that should ideally be deposited by a gradually waning current. These are termed "top-absent" sequences (Walker, 1965), and their mode of origin is uncertain. Walker suggested that a stabilized, cohesive bed deposited rapidly from the traction carpet would not be reworked by the current, so that excess turbulence would develop and the current would transport away the remaining sediment in the overlying flow. Walker supposed that this "rejuvenated" current would be capable of reworking a cohesionless bed, but not of depositing sediment. (The present author does not understand Walker's concept. He refers to a current that cannot and then can rework the bed, and a bottom which is first cohesive, and later cohesionless). Middleton (1969) has suggested that a large turbidity current may set up a large-scale circulation of the surrounding water, so that after the main body of the current passed by, the "entrained flow" would rework the top of the freshly-deposited turbidite.

In the present sequence, the nearly constant grain size, mainly fine and very fine sand, and coarse silt, allows us to use the bed forms as an accurate indication of the changes in flow power of the depositing turbidity current.

Vertical Distribution of Sandstone Bed Facies: Description and Interpretation

Relative Slope

The relative slope over which the turbidity currents flowed may be determined by analysing the distribution of facies occurring at the base of the turbidites. This distribution (figs. 49 and 50) shows a three-part pattern:

1) From the top of member 3 down to about 25 m (distances are measured down from the base of member 4) a and b beds (in this and the following two sections, underlined facies refers to the facies which occurs at the base of the bed only, and is not a description of the bed) are slowly decreasing in importance, while c and a->e beds are increasing.

2) From about 25 m to 47 m there are a number of striking changes. First, c drops rapidly, then b increases sharply peaking at 34, and then dropping off sharply, c peaks again at 44, and a peaks at 47. Except for the lower c peak, there is an overall trend from c to b to a beds.

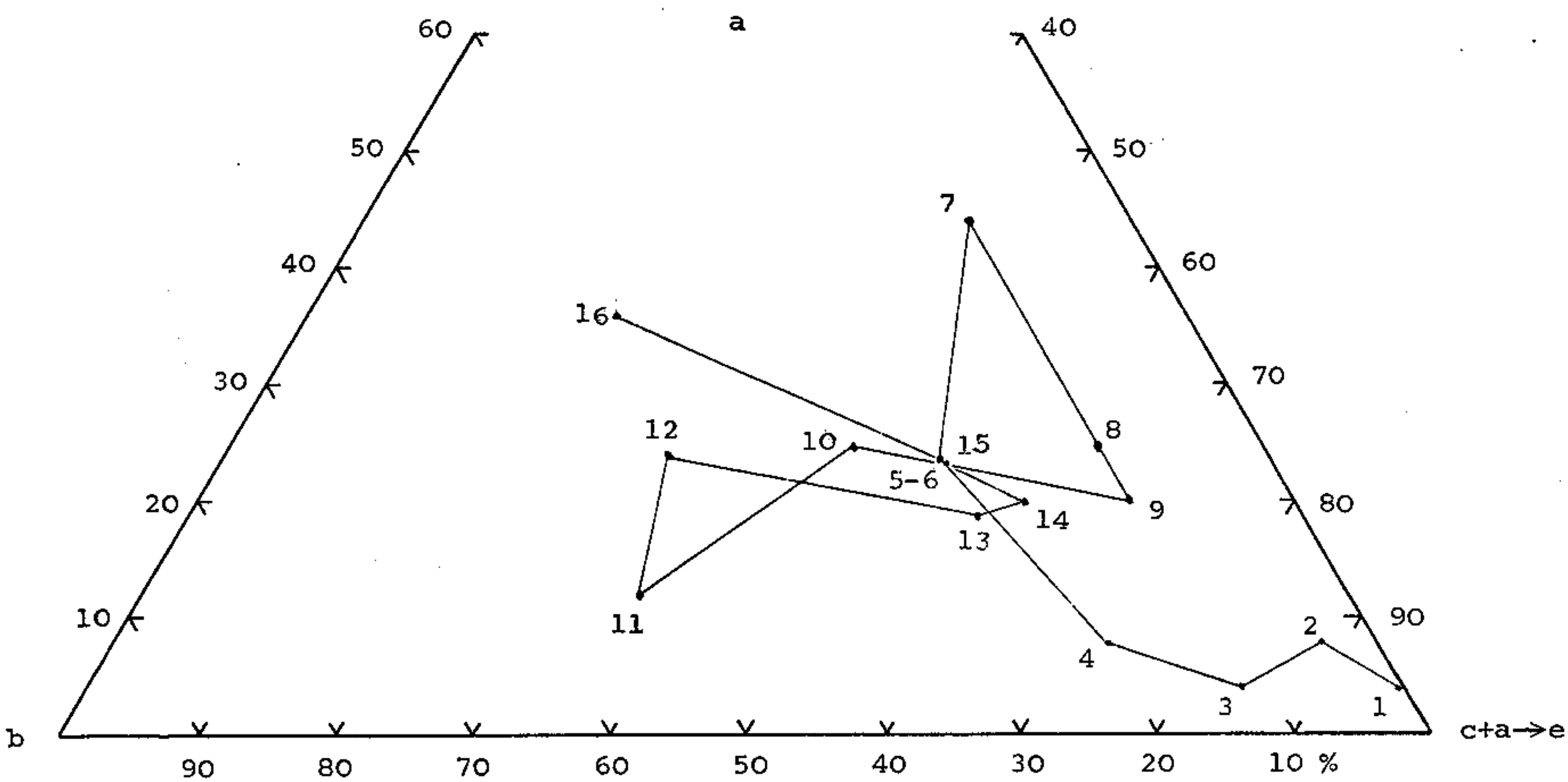


Figure 50. The changing proportion of a-based, b-based, and c-based + a→e sequences in turbidite sandstones of the upper part of member 3, Stappogiedde North, Digermul Peninsula. Numbers refer to 25 bed intervals, from the base upwards. Deceleration of the turbidity currents occurs in 16-14, on the shelf, in 7-1, on the lower part of the slope and the basin. On the greater part of the slope, 14-7, trends are irregular.

3) From 47 m down to 72 m a decreases, almost dying out altogether, b has a gentle peak between 50 m and 60 m, and both c and a->e increase.

In the uppermost part, the current was slowing down, depositing sediment in successively lower flow regime conditions. However, in the central part, there is a change to high flow power, and the current was apparently increasing in velocity. Finally, in the lower part, the current again lost power, and was slowing down. These relative changes indicate that the central part had a steeper gradient than the upper and lower part. It thus seems appropriate to label the upper part the shelf, the central part the slope, and the lower part the basin, across which environments the current was respectively decelerating, accelerating, and decelerating again. The fact that the sequence is only about 70 m thick shows that there is no comparison in scale with the major continental features.

The Shelf-Slope Boundary

The sharp drop in c beds and the sharp rise in b beds indicates that the break in slope occurred between 25 m and 34 m, most likely between 25 m and 30 m, as acceleration was evidently rapid.

The Slope-Basin Boundary

The sharp drop in a beds between 47 m and 52 m suggests that the gradient was levelling off. This is accompanied with increases in b and c beds. However, the most steady trends are present below 55 m, and this value is taken as the junction between the slope and basin.

Proximality Index

The proximality index (P_1, Walker, 1967) was partly intended by Walker to indicate the relative distance a turbidity current had travelled. In the present example it appears that slope rather than distance travelled has a more profound influence on P_1 (fig. 49). P_1 decreases uniformly in the shelf and basin parts, but is very irregular in the slope area. The uniform decrease of P_1 below 55 m suggests that the currents were steadily losing power in this interval, i.e. travelling further over a uniformly low gradient.

Sandstone Bed Thickness

Bed thickness varies irregularly in the upper half of the sequence (fig. 49). Towards the lower part of the slope, 41 m to 52 m it increases to a peak of 9.5 cm, and then falls off rapidly to 3.3 cm at the bottom of the section. It appears that the greatest amount of deposition occurred around the decrease in gradient. As the slope levelled off, the ability of the current to transport its load decreased. Consequently rapid deposition ensued, and the bed deposited downcurrent thins appreciably.

Proximality and Bed Thickness

There is a linear relationship between P_1 and bed thickness (fig. 51). This is consistent with the findings of Walker (1967, fig. 5, with regard to values below 70%). Walker suggested that thicker beds are related to higher flow regimes. This is probably correct for most turbidite sequences in a flysch setting in which deposition of turbidites occurred on a relatively flat, nearly horizontal basin floor, the currents being fed by steeply sloping submarine canyons. In the present example, bed thickness and P_1 each appear to be independently related to the gradient over which the turbidity current flowed.

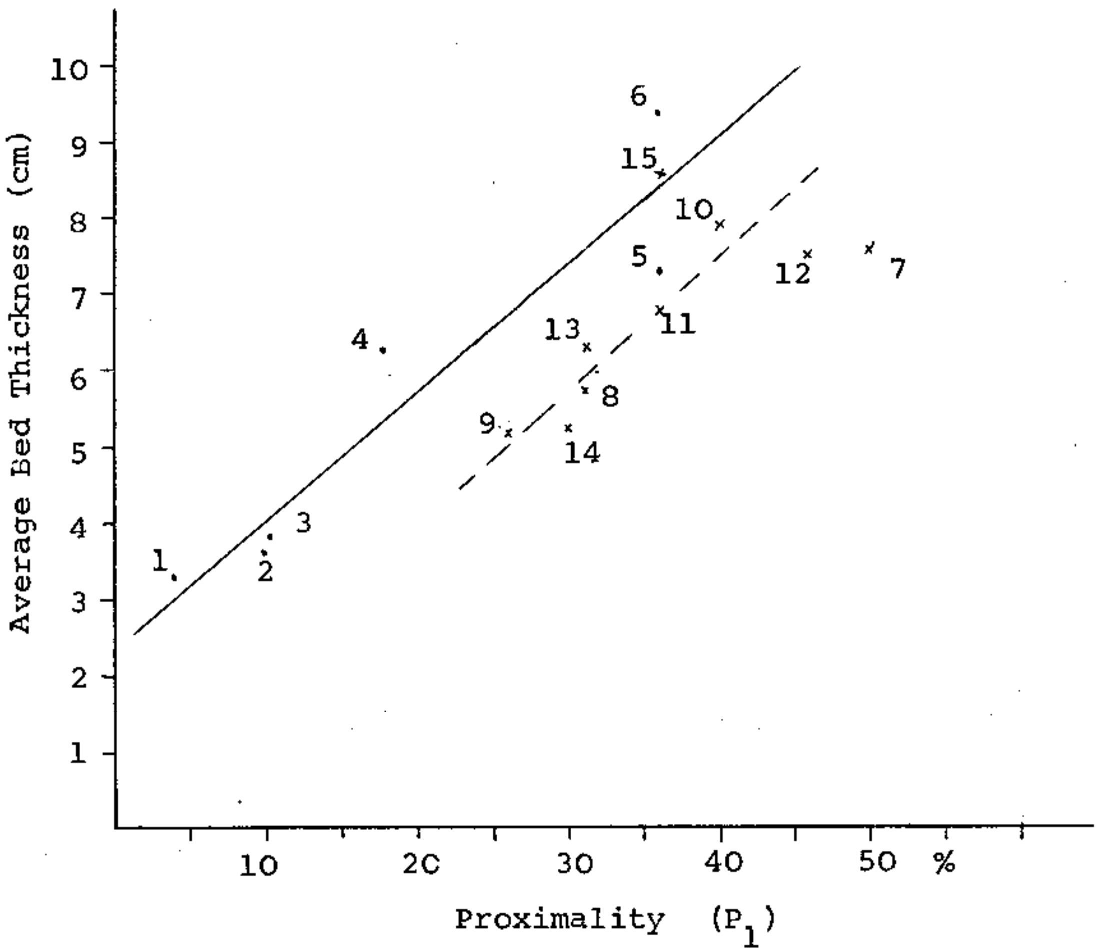


Figure 51. Proximality (P_1) and bed thickness of turbidites in upper part of member 3, Nyborg Formation, Stappogiedde North, Digermul Peninsula. Numbers refer to 25 bed intervals, from the base of the section, upwards. Lowest 6 points show steadily increasing bed thickness and proximality, and group around solid line. Points 7-15 show irregular trends in proximality and bed thickness, but plot around the dashed line. The partition of values may reflect deposition from currents that are accelerating and decelerating, rather than uniformly decelerating.

Background Sediment Thickness

The background sediment tends to be relatively thick (fig. 49) in the shelf area, probably due to the high rate of reworking of the sandstone beds. The value falls off sharply at the shelf-slope boundary. This may be due to the increased rate of deposition of sandstone beds, a decreasing extent of reworking of sandstone beds, a fall off in rate of deposition of background, or erosion beneath turbidity currents. The thickness of background sediment increases down the slope, possibly due to the increasing depth, through which wave agitation diminished allowing more fine sediment to settle out of suspension. The coarser texture of the background in the shelf environment suggests that the fines were being transported away, presumably into deeper, quieter water.

Ratio of Bed/Background Thickness

This ratio (fig. 49) shows a marked downward decrease between 52 m and 57 m, the slope-basin boundary. Interestingly, there is no fall off in the ratio along the basin floor. The reasons for these trends are obscure.

Facies Sequences

Eight types of facies sequence are distinguished from the section of 394 sandstone beds (fig. 49). In the shelf part of the section, at the top a, b and bc sequences are the most abundant, and c is infrequent, while at the shelf break, about 25 m, c has increased to a major peak, and the other sequences have fallen off. Down the upper part of the slope, b and then a sequences have peaks followed by c and a->e sequences. Below, in the lower part of the slope, ab peaks, and ac, abc, and bc sequences have rather broad peaks extending for about 20 m. Centering on the slope-basin boundary is a broad b peak, and in the basin, c and a->e sequences increase rapidly towards the bottom of the section.

The changes in the shelf part of the sequence are consistent with the waning of the flow and the shallow gradient (although reworking has apparently eradicated much of the primary deposits). In the upper part of the slope where acceleration is occurring the kinds of sequences deposited are mainly top-absent, particularly a and b. These contrast with the ab, ac, abc, and bc sequences which occur in the lower part of the slope and the top of the basin. Over the lower part of the slope the basal facies of the beds gradually changes from a, to b, to c indicating waning of the flow as discussed earlier. It is possible, in the light of these observations, that the peak of a->e sequences at about 41 m may largely be a misinterpretation, in the field, of a beds. Both facies are massive internally, the difference being largely the presence or absence of grading at the top of the bed.

It thus appears that top-absent sequences formed when the current was accelerating, and top-present sequences formed when the current was decelerating. The latter case is normal for most flysch turbidite sequences, as mentioned earlier.

This relationship of top-absent and top-present sequences to changes in velocity (flow power) may perhaps be explained by the hydraulic jump. This has been mentioned by Middleton (1969), and the effect of the hydraulic jump has been considered in detail by Komar (1971). Where supercritical flow (Froude number, $Fr > 1$) passes downstream into subcritical flow ($Fr < 1$), the thickness of the flow increases, and the velocity decreases rapidly at a junction called the hydraulic jump. Presumably, a hydraulic jump would also exist where subcritical flow was passing downstream into supercritical flow. Unfortunately there is no close relationship between bed form and Froude number (Simons et al., 1965; Harms and Fahnestock, 1965) so that it is impossible to determine the nature of the flow from the sedimentary structures present. However, the increase in velocity associated with a higher slope would cause a thinning of the current, which would be emphasized in the presence of a hydraulic jump. It seems reasonable to expect that the sequence formed where the current is accelerating would be different from that where the current is decelerating. Sufficient acceleration might offset the decrease in velocity which takes place during the passage of a turbidity current over a point, and prevent the formation of the waning flow sequence of bedforms. Similarly, with a decreasing gradient, the thickness of the current would increase, and waning would occur gradually as the current slowed. A complete waning flow sequence might be expected. Rapid deceleration might form the a->e sequences, but the data do not unequivocally support this.

6.4.4 Synthesis

Having established the relative gradients of the prograding marine environments, these can be related to the background sediment. In the upper 15 m of the shelf (which extends down to 25 m) the background is unit D, with wave agitation, reworking of the bottom and deposition of thick parallel sandstone laminations, possibly related to tidal current activity (fig. 54). Down the slope, passing through units C and B, wave agitation decreases, and the background is much finer. The most rapid decrease in grain size in the background occurs between about 15 and 30 m corresponding to the outer shelf and upper slope environments. It is possible that the silt and mud deposition in this area was an important agent in maintaining the gradient of the slope. The transition from unit B to A occurs approximately at the slope-basin boundary where wave activity is very slight. It is possible that deposition of fines from suspension also helped to maintain the lower part of the slope.

6.4.5 Discussion

We pause here to consider the validity of the second assumption underlying the interpretations presented above. Perhaps most important is the problem of the genesis of the turbidity currents. An origin by slumping within the slope area could be considered for those sandstone beds situated in the central and lower parts of the sequence. However, the near absence of any signs of lateral mass movement of sediment, such as slump scars, rotated bedding, and strongly lenticular massive beds, suggests that this origin is very unlikely. Alternatively, the turbidity currents may have originated from the shallow water part of the sequence either by the introduction of sediment from rivers in flood, or by the downslope movement of sediment-laden coastal waters. In terms of the environmental context, the top of member 3 is overlain by shallow water tidal sediments. However, the immaturity of the sediment in both members 3 and 4 indicate that these marine environments were being supplied by a nearby, active fluvial source. Deep channels, not observed in the present sequence, have been observed in a deltaic context (Walker, 1966a). According to Walker turbidity currents would be capable of eroding channels if they were fast, under-laden with sediment, and have no traction carpet. Walker (1966b) described diverging palaeocurrents at the basin margin, a pattern which suggests a point source (discrete feeder channels) in the adjacent slope, causing the progradation of fans. In the present case, the tightly grouped palaeocurrents in the basinal turbidities (fig. 48C) indicates an apron rather than fan topography, with a line rather than a point source of turbidity currents, and with little divergence of flow directions. In his papers, Walker does not explain convincingly why the currents form discrete channels, rather than flow as sheets down the slope. Strong erosion is not a sufficient condition. If a point source for the introduction of turbidity currents into the basin (i.e. a delta distributary) was a necessary prerequisite for the submarine channels, then the absence of channels in the present sequence would suggest that turbidity currents did not form by the direct introduction of sediment from rivers, or their distributaries.

The activity of storms in shallow coastal waters is capable of suspending large amounts of material which can be transported offshore and converted into a turbidity current during the storm surge ebb (Hayes, 1967). In the offshore area the flow is not channelised, but has a sheet-like flow pattern, with sediment transport predominantly offshore. (The nature of the shallow marine conditions at the top of member 3 will be touched on in a later section).

It thus appears likely that one type of turbidity current was responsible for the deposition of most of the sandstone beds in the upper part of member 3. Obviously, the density, size and extent of individual currents would vary according to the specific conditions in which they were formed.

6.4.6 Conclusions

The upper part of member 3 consists of sandstone beds deposited by turbidity currents, and background sediment deposited by normal, in situ processes. Vertical changes in the background sediment reveal a gradual shallowing up of the sequence which can be traced 55 m below the shallow marine, tidal sediments of member 4. The background changes from a mudstone at the base to a mixture of mud and very fine sand at the top, and ripple marks and cross-lamination increase gradually upwards. The changes are attributed to the progradation of an inclined depositional surface, rather than to vertical sedimentation alone.

The turbidite sandstones do not show a uniform trend in the section. Rather, they indicate a gentle relative slope in the upper part, a steep relative slope in the middle part, and again a gentle slope in the lower part of the section. These are interpreted respectively as shelf, slope and basin environments. While passing from the shelf to the slope, the turbidity currents accelerated depositing sediment with structures indicative of progressively higher flow regime, at the base of top-absent sequences. While crossing onto the lower gradients of the lower slope and basin, the currents decelerated, depositing sediment with structures indicative of a progressively lower flow regime at the base of top-present, waning flow turbidites (fig. 52).

Changes in the thickness of the turbidity current flow which result from a changing gradient, and which may be accompanied by a hydraulic jump are possibly the determining factor in whether or not the bed will be top-absent. Thinning of the flow and failure of the current to wane slowly enough, may prevent the normal turbidite sequence from being deposited.

Ripple marks and flutes combine to show that the palaeoslope was to the NNE or NE, and the coastline was approximately perpendicular to this.

6.4.7 Other Localities

Digermul Peninsula

The thickness of the zone at the top of member 3 which contains few sandstone beds and consists largely of unit D background sediment appears to thicken to the north (fig. 53) (Reading and Walker, 1966, facies C-4, fig. 2). However, the many observations made at sections between Larsholmen and Areholmen (fig. 45) indicates that the proportion but not the character of beds changes rapidly from one exposure to the next.

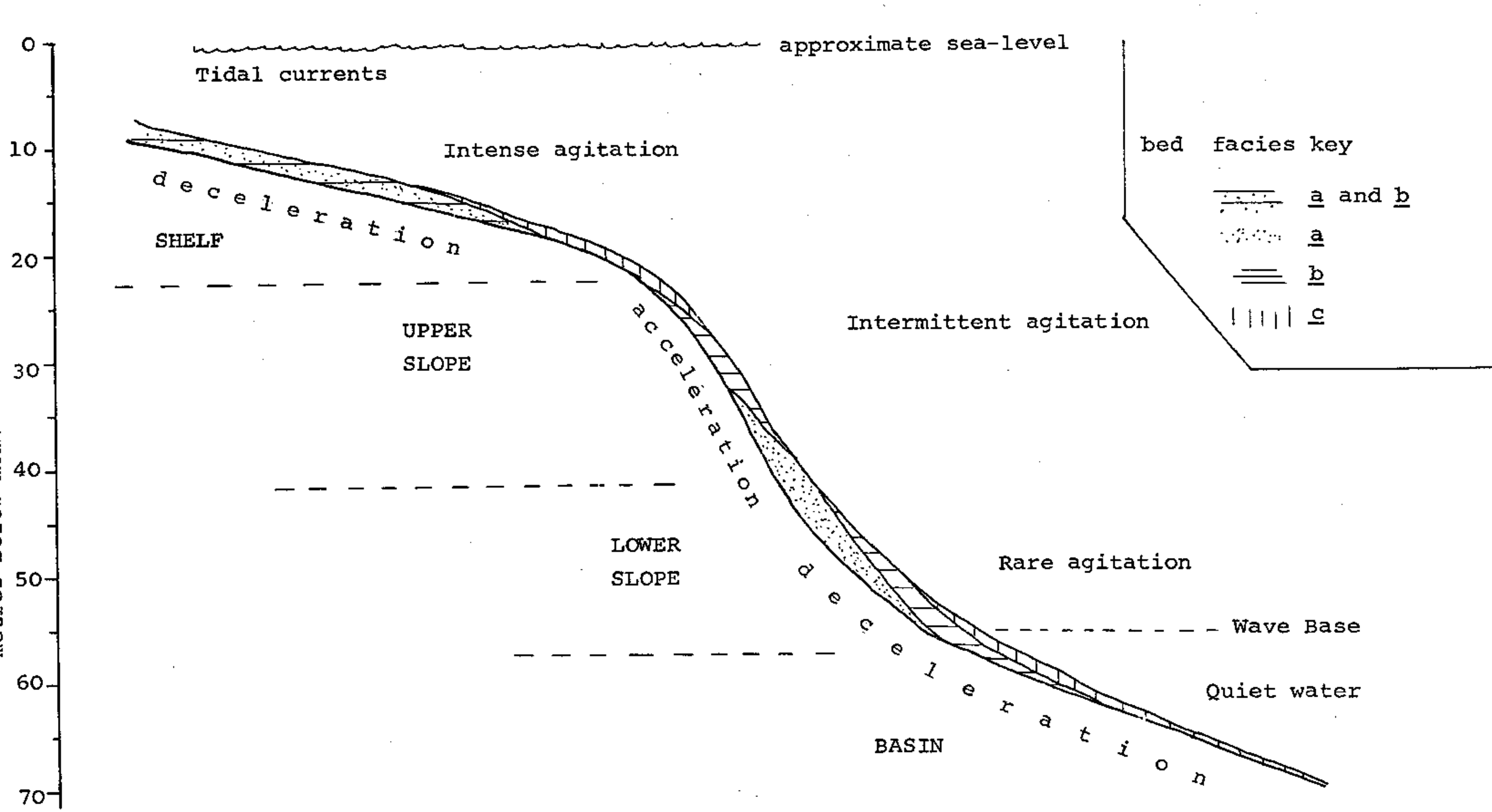


Figure 52. Conditions of sedimentation of upper part of member 3, Nyborg Formation. Sandstone facies are the deposit of a "typical" turbidity current. Tidal currents and wave agitation were important in reworking and deposition of background sediment.

The most important bed type is thick beds composed of massive and parallel-laminated sandstone, facies a and b. As the only complete coastal exposure occurs at Stappogiedde North, a proper test of lateral variation cannot be made. Observations made at the other localities do not counter the interpretation of the main section.

Trollfjord

The section at Trollfjord is in some ways quite different from that at Stappogiedde North. Because of discontinuous exposure only 40 m of section at the top of members 2-3 can be examined. These show agitation at the base, mainly in the form of current ripples. The agitation increases up the section, and lenses of sandstone are common 15 m below member 4. Unlike at Stappogiedde, this part of the section is reddish, rather than grey-green. Ten metres below member 4 the colour changes to grey-green and thin and medium bedded sandstone become rather abundant. Some of these closely resemble the sandstone beds at the top of member 3 at Stappogiedde North, but others appear to have gradational bases. Convoluted lamination is abundant. Marking the contact with member 4 is a 2 m thick massive grey-green sandstone with large ball-and-pillow structures. Thus, the section at Trollfjord also shows a broad shallowing upward trend, and a sharp contact with member 4.

Figure 49

Explanation

Aside from slumps and flutes which are shown individually, data is represented as the average of 25-bed non-overlapping groups. The average is plotted along the thickness midpoint for each group. In the 394 beds, 15 25-bed groups are present, with the 16th group containing 19 beds. Charts that refer to frequency are of the number of occurrences of a feature out of 25. Thus, these can be used as a percentage as well. Background and bed thickness refer to the sum of these values in each 25-bed group. For proximality, P_1 (Walker, 1967, $a + \frac{1}{2}b - a\rightarrow e$) is used, where each facies refers to the number of times it occurs at the base of a bed.

For basal facies frequency, the following applies:

a ———

b – – – –

c —·—·—·—

a->e ————

Background
Slumps
Flutes
25 Bed Interval
Background Thickness
Bed Thickness
Ratio Bd / Bgd
Average Bed Thickness
Basal Facies Frequency
Proximality
Frequency of Facies Sequences
Environment
unit D
unit C
unit B
unit A
increasing agitation
metres below member 4
a
ab
ac
abc
b
bc
c
a→e
SHELF
UPPER SLOPE
LOWER SLOPE
BASIN

6.5 Shallow Marine Sedimentation

Members 4 and 5

6.5.1 Introduction

Members 4 and 5 reflect the shallow marine conditions which prevailed after the filling of the basin by members 2 and 3. Member 4 is characterised by purple sandstones and interbedded shale while member 5 contains white sandstone with dolomite at the base (fig. 53). Although member 4 forms a relatively continuous strip of outcrop between Stappogiedde North and Areholmen (fig. 45), and is discontinuously exposed in coastal outcrops between Larsholmen and Innerelv, the most detailed investigation was of the excellently exposed section along the coast at Stappogiedde North. Observations of all the other exposures aided in the understanding of the member.

Lack of time did not permit a careful investigation of the Nyborg Formation at Trollfjord, but the observations made there show the critical importance of the section in interpreting members 4 and 5.

In addition to the standard procedure of detailed section measuring, an intensive study of the palaeocurrents was carried out in member 4. Due to the virtually infinite number of structures available to measure, an arbitrary limit of 50 in a particular unit at an outcrop was set. The bedding dip of 25^{o}-30^{o} does not seriously affect ripple mark orientations, but all cross-bedding orientations were recorded by measuring the orientation of cross-bed foresets and rotating the bedding to horizontal by stereographic rotation.

In the procedure used, the sediments of member 4 at Stappogiedde North were divided into facies, and then grouped together into kindreds, and facies associations.

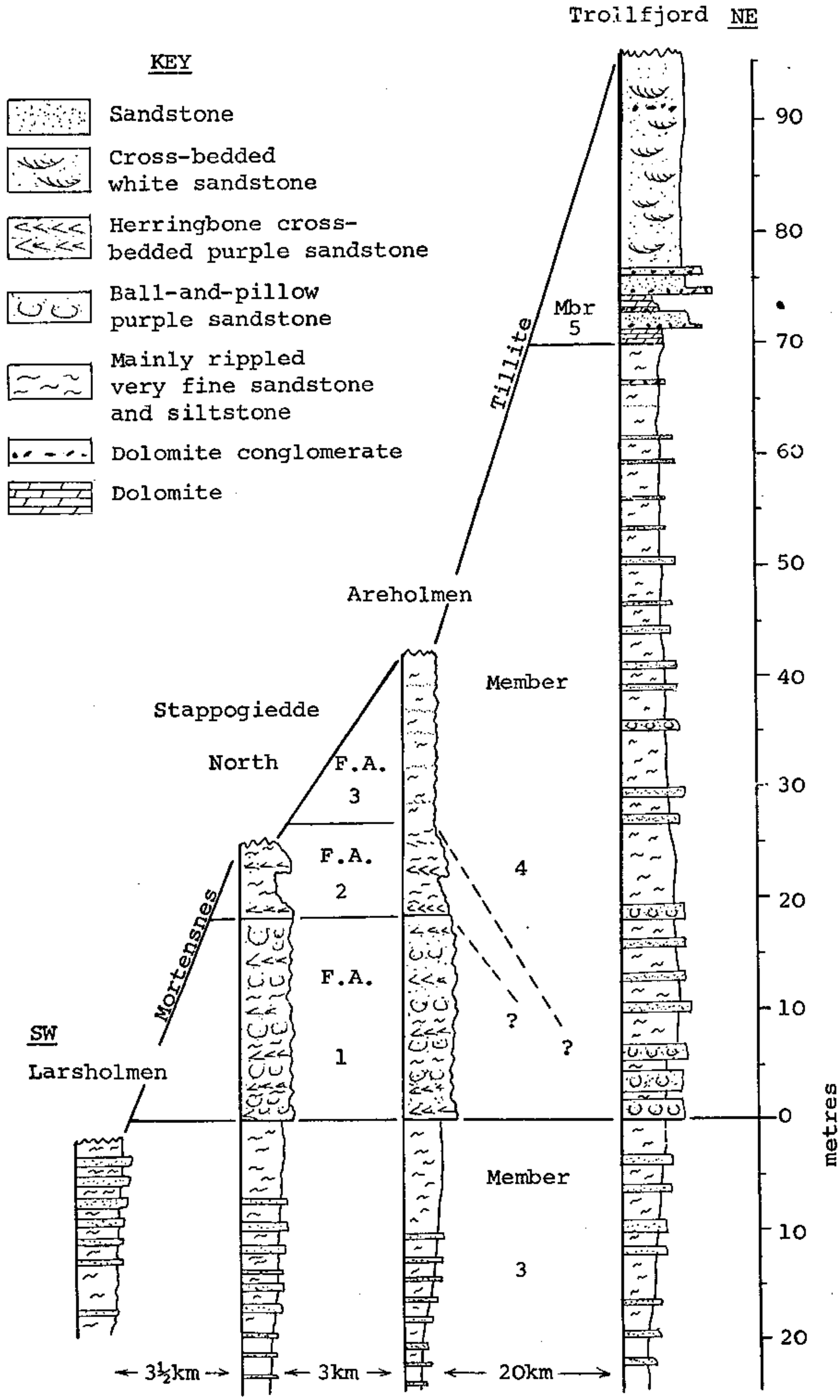


Figure 53. Stratigraphy and sedimentological outline of the top of member 3, and members 4 & 5, Nyborg Formation, Digermul Peninsula and Trollfjord.

6.5.2 Facies

Introduction

The extremely varied sedimentary rocks in member 4 were ruthlessly grouped into eight facies. These are rather easily distinguished on the basis of sedimentary structure, colour and grain size. The principle sedimentary structures are cross-bedding, parallel lamination, flaser bedding, graded bedding, massive beds, and those formed by soft-sediment deformation. Sandstone generally occurs as either purple or grey to grey-green. Grain size falls into three groups: medium to coarse sandstone, very fine to fine sandstone, and silty mudstone. Generally, the coarser sandstones are purple, and the finer are grey to grey-green.

Facies A

Cross-Bedded Sandstone

This facies consists of sets of cross-bedded purple sandstone mainly 5-15 cm thick varying from beds one set thick to cosets up to 2 m thick, of many interlocking sets.

Sets are laterally continuous, or lenticular due to erosion by overlying sets. Cross-bedding is planar to somewhat concave with the lower part of the foreset merging tangentially into the base of the set (Pl.68). Foresets dip between 8° and 34°, though 20° is average, and the foresets of successive sets often dip in opposing directions, imparting a "herringbone" appearance. In cross-section, perpendicular to the current flow, sets appear parallel-laminated, either horizontal or slightly inclined, or filling shallow troughs. Festoon trough cross-bedding was not observed. The sets are interpreted as having formed by the migration of sinuous, or straight-crested dunes in the upper part of the lower flow regime. Penecontemporaneous ball-and-pillow structure (Potter and

Pettijohn, 1963) is common in this facies (Pl.69). One example of overturned cross-bedding was noted.

Thin sections show moderately sorted medium to coarse grained sandstone. The grains are subangular to subrounded quartz (including undulatory and polycrystalline), plagioclase and microcline with accessory chert, magnetite and zircon. Feldspars show all stages of decomposition. Grains are coated with haematite and both haematite and quartz are cementing agents. The matrix, largely mica, gives the rock a greywacke like appearance.

Five hundred foreset orientation measurements show that most cosets have a bimodal current distribution with the modes approximately 180° apart (fig. 54*). The relative size of each of the modes in a coset is highly variable. A bimodal, bipolar current distribution is strong evidence for a tidal environment. As these modes are parallel to the sole mark directions in member 3, currents appear to have been predominantly onshore-offshore, rather than longshore.

Several sets near the base of one coset preserve the modified form of the dunes, and shed further light on their size, shape and mode of formation (Pl.70). The sets are about 15 cm high, 2 m long, and average 3-4 m crest-to-crest length (fig. 55). The foresets are truncated at the top of the set on both stoss and lee slopes. Subsequent deposition of cross-stratification in the up-current trough seems to have been related to the rounding of the dune. Apparently a reversal in current direction caused erosion of the lee side of the dune, and deposition on the stoss side. *(Follows page 176).

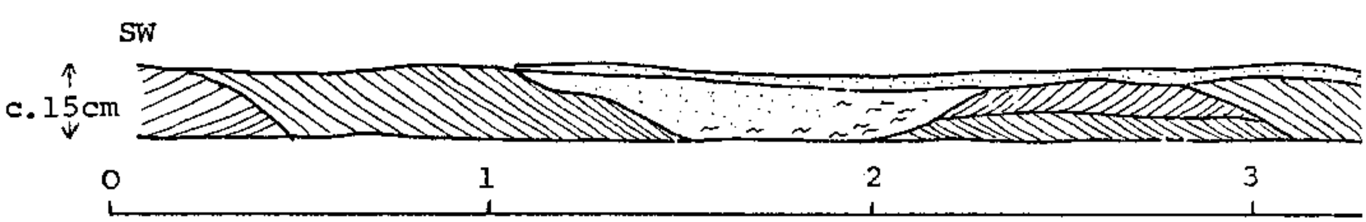


Figure 55. Single bed containing several cross-sets at the ba

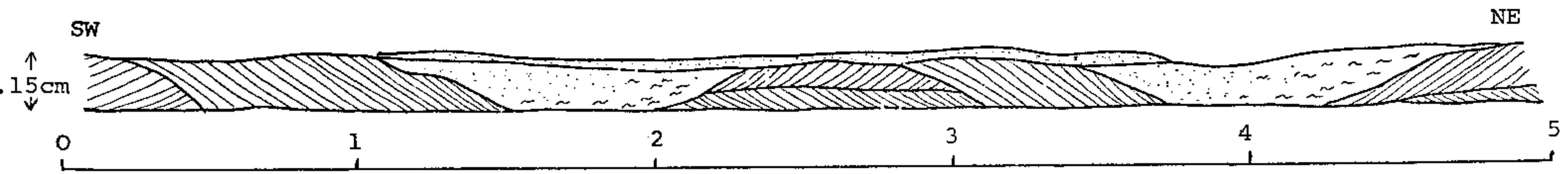


Figure 55. Single bed containing several cross-sets at the base of unit K, facies association 1, member 4, Nyborg Formation, Stappogiedde North, Digermul Peninsula. Currents flowed north, south, and north again each time modifying earlier structures. After a final period of erosion, the troughs were filled with massive and flaser bedded very fine sandstone. (see Plate 70).

Rounded dunes have been recorded from present day fluvial (Collinson, 1970) and intertidal (Klein, 1970) environments. Currents in the fluvial environment are mainly undirectional (Allen, 1966b). Small erosion surfaces in the lamination of some of the foresets (lee side) may be reactivation surfaces (Collinson, 1970) which formed by dune modification between periods of forward dune migration, i.e., during current reversals.

Straight-crested ripples occur on planes between cross-sets, and foreset surfaces may occasionally appear ripple modified. This indicates the alternation of wave and weak currents, with strong currents associated with dune migration. Dune modification probably took place during low or high tide under low velocity flow, while dune migration probably occurred during the ebb and flood tides when peak velocities were reached (e.g. Klein, 1970).

Thick cosets are very continuous laterally; several could be traced up to 3 km, the length of the outcrop, but unexposed stretches, across which the units were correlated may conceal channel margins. One such channel margin is described below (section 6.5.3).

Facies B

Parallel-Laminated Sandstone, Fine to Medium Grained

Parallel-laminated purple sandstone, fine to medium grained and moderately sorted occurs as units up to 40 cm thick. The lamination is horizontal or slightly undulating, and is laterally persistent (Pls. 68, 69). Interrupting the lamination are rare ripple horizons (Pl.68) and small, shallow scours (Pl.71). Primary current lineation was not observed.

In thin section, the lamination consists of alternating medium sand and fine sand layers up to 6 mm thick, but mainly 1-3 mm thick, with haematite grains concentrated in the coarser layers. Boundaries between adjacent laminae are both sharp and gradual. The facies usually has sharp or erosive contacts with the adjacent facies.

Parallel-laminated fine to medium grained sand can form by plane bed flow in the lower part of the upper flow regime (Harms and Fahnestock, 1965). The structure has been described from numerous sedimentary environments: beach (Thompson, 1937), fluvial channels (Harms, MacKenzie and McCubbin, 1963), tidal estuary and inlet (Terwindt, 1971; Reineck and Singh, 1967) and offshore marine (Reineck, 1963). Pseudoplanar stratification has been described from the Platte River (Smith, 1971) where its formation has been attributed to the migration of low amplitude sand waves in very shallow water. Such an origin for the present sediments can be ruled out because of the fluctuating water level associated with a tidal environment.

The absence of low angle inclined sets of parallel lamination, and the moderate sorting in the present deposits appears to rule out a beach origin (e.g. Campbell, 1971). In a description of tidal deposits of Devonian age in SW Ireland, De Raaf and Boersma (1971) noted the absence of primary current lineation in the parallel-laminated sandstones and suggested that this was due to rapidly slackening currents causing high rates of fall-out from suspension. However, the scarcity of bedding plane exposures in the present rocks might also explain why primary current lineation was not observed.

Viewing the facies on its own, only the hydrodynamic flow conditions under which it formed can be interpreted with certainty.

Facies C

Parallel-laminated sandstone, very fine grained

This facies occurs in units up to 30 cm thick. The lamination is delicate and continuous, and although mud is absent, the facies often grades into the flaser bedded facies

(Pl.72).

Very fine grained parallel-laminated sandstone is also present in some graded beds observed. These are considered as a distinct facies below.

The sediment was probably deposited under conditions of low flow power in the lower part of the upper flow regime (see Allen, 1970).

Facies D

Flaser bedded very fine grained grey sandstone

Flaser bedded sandstone occurs in units up to 115 cm thick, though typically less than 40 cm. The coarser component of the facies is very fine sandstone and coarse silt arranged into laminae and lenses up to several mm thick. These are separated by continuous and discontinuous laminae (flasers) of mud or silty mud usually no thicker than 1 mm. According to the classification of Reineck and Wunderlich (1968) this facies may be termed bifurcated wavy flaser bedding (Pl.72).

A few bedding planes show the ripple morphology to be mainly straight or sinuous crested, mainly asymmetrical ripples. These features, plus the rounded profiles suggest that waves and currents simultaneously shaped the bed, forming combined flow ripples, very similar to those at the top of member 3 (Pl.59).

Most workers have interpreted flaser bedding as tidal in origin, the contrast in grain size reflecting ebb and flood currents versus slack water conditions. Recently, it has been suggested that marine mud layers are more likely formed over a long period of time, determined by suspended sediment concentration, and wave activity, possibly related to storms (McCave, 1970). According to McCave, the origin of mud on tidal flats may be due possibly to tidal rhythm, but more likely to larger scale fluctuations mentioned above.

Although it is not possible to be certain about the exact mode of deposition of the mud and sand, it is likely that flaser bedding, characterised by the strong separation of grain sizes is most often developed in a shallow, tidally influenced environment as this is where it has been most often described to date. The absence of structures indicating subaerial exposure suggest that this facies is subtidal in origin.

Facies E

Graded Beds

A variety of graded beds occur in member 4. All are less than 20 cm thick and comprise a sandy base and a silty, muddy top (fig. 56). The sandstone base includes very fine to medium grained parallel-laminated sandstone, which grades up, sometimes through a zone of ripple cross-lamination, into silty mudstone. The sandstone base may be lenticular (Pl.73), erosive, and contain mud flakes. A few beds grade upwards from massive sandstone into massive silty mudstone. The massive silty mudstone at the top of the bed is usually grey-brown, and strongly cleaved. Lenses of coarser sediment, such as sand, were not observed in the mudstone.

In thin section, the medium sandstone parts of graded beds consists of quartz, including strained, and polycrystalline and feldspar including plagioclase and micrcline. Of the two slides examined, one has dolomite grains and a dolomite cement, while the other has no dolomite grains and a silica cement. Both slides have haematite coatings on the grains, and a distinctive green mineral with a high relief and low birefringence. The green mineral occurs in grains mostly sand sized, or smaller, and often well rounded. The chemical composition of four grains of the mineral, determined from one thin section by electron microprobe analysis (Appendix) shows

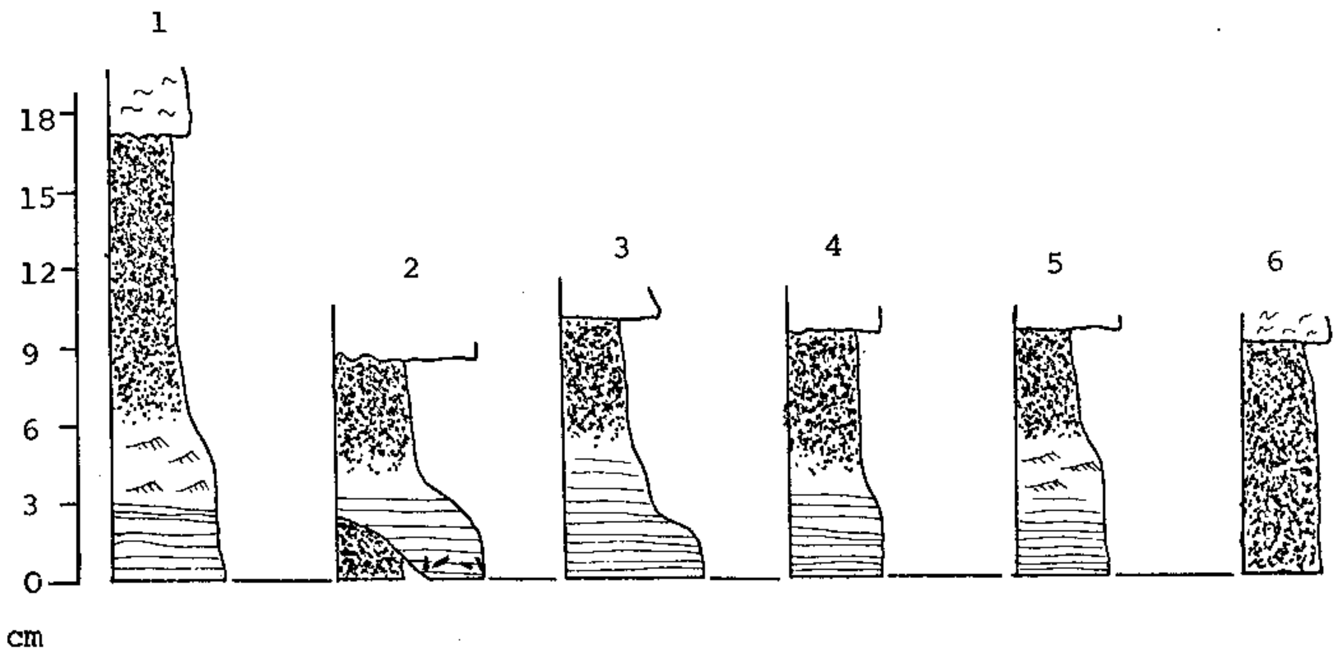


Figure 56. Facies E, graded beds, in member 4 of the Nyborg Formation, Stappogiedde North, Digermul Peninsula. Beds 1,2 & 3 are from facies association 2, beds 4,5 & 6 are from facies association 3. The bases of beds 1,4 & 5 are fine and very fine sandstone, bases of beds 2 & 3 are medium sandstone. Shaded area is silty mudstone.

a composition similar to that of glauconite (James, 1966) especially with regard to the distinctive high K_2O value. These grains are thus probably glauconite pellets.

The sandstone-mudstone unit appears to be a waning flow deposit formed during a single event. At first, the current may have scoured its bed, entraining some of the bottom as mud flakes, and then during deceleration deposited first sand in bed forms indicative of decreasing flow power (Allen, 1970) and then silty mud which fell rapidly out of suspension. It is likely that the thick mud units, massive and not laminated, were due to the high mud concentration of the current, rather than to a relatively long period of deposition from weak, low concentration currents.

Facies F

Silty mudstone

Beds of cleaved, massive silty mudstone less than 40 cm thick occur in sharp contact with other facies. Rarely, very faint parallel lamination is observed. Some beds appear to be very continuous laterally, traceable over several kilometres; most however cannot be traced from one outcrop to the next.

The base of one bed (Pl.74) shows linear ridges interpreted as groove marks, which attest to the strength of the depositing current.

As with the graded beds, the silty mudstone beds are interpreted as having been deposited by a waning current during one relatively quick event. The grain size was most likely controlled by the sediment content of the current.

Facies G

Deformed Sandstone

Deformed sandstone occurs as local zones of deformation in cosets of purple cross-bedded sandstone, or other facies, and as discrete beds made up entirely of deformed masses of sandstone with intervening spaces filled with massive sandstone.

Four kinds of deformation were observed. (1) Small load balls, pseudonodules, occur in fine grained massive sandstones up to 30 cm thick. These have an ovoid shape in cross section, and formed by vertical quasifluid movement within the bed (see Potter and Pettijohn, 1963). (2) Convoluted lamination, formed by vertical movements, occurs in fine sandstone. (3) On a larger scale is ball-and-pillow structure (Potter and Pettijohn, 1963). This structure often occurs as zones within facies A, cross-bedded medium grained purple sandstone cosets and as beds made up entirely of the structure, and in facies B, medium grained parallel-laminated sandstone. In the former case zones of deformation may change laterally or vertically into undeformed sandstone (Pls.68, 69). Such zones may be small, less than 1 m across, or they may make up most of the coset. Horizontal cross-sections through the structures were not observed, but vertical cross-sections show circular to considerably flattened balls of sandstone occasionally surrounded by massive sandstone (Pl.76a). The balls show lateral symmetry in these sections, the plane of symmetry being perpendicular rather than oblique to the bedding.

The absence of lateral asymmetry is interpreted to indicate the lack of lateral movement in the deformation of the bed. Vertical movement predominated, possibly caused by dewatering, or an unstable density stratification of depositional origin.

The tops of ball-and-pillow beds are often eroded into by massive sandstone. In contrast, some graded beds have ball-and-pillow at the base, which grades up into the overlying sediment. (4) Slumped bedding is characterised by masses of sandstone which are chaotically bent, squeezed, and sometimes broken (Pls.71, 75, 76b, c, d). Lateral movement probably played an important role in the formation of beds with this structure.

Facies H

Massive Sandstone

Massive fine to medium grained, poorly sorted sandstone occurs in beds up to 20 cm thick. Most of the beds rest erosively upon deformed sandstone. Massive sandstone that contains small load balls, that surrounds pillow structure, or that occurs with slumped beds probably originated by churning of originally stratified sediment. Erosively based massive sandstones without attendant deformation structures may have been deposited rapidly from suspension by strong, turbulent currents.

6.5.3 Facies Kindreds

Introduction

The term kindred is introduced here to indicate facies which have a marked tendency to occur together in the succession and which are interpreted as having formed by processes which are likely to occur together in space and time. Thus the term fluvial "channel" facies might be the equivalent of the kindred facies: cross-bedded, flat-bedded, and ripple-laminated sandstone. The term is intermediate in scale between facies and facies association.

AB Kindreds

Thick Beds of Cross Bedded and Parallel-Laminated Medium Sandstones

Within the close relationship between thick cosets of facies A and facies B (Table 18), there is a significant tendency for facies A to pass up into facies B, but not the other way round. These two-part sequences are termed AB kindreds.

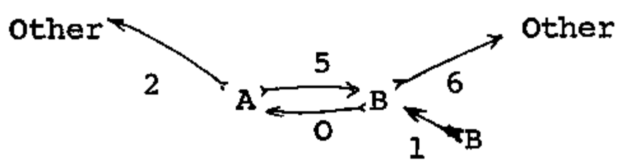


Table 18. Transitions involving facies A and B.

Point bar accretion in a laterally meandering stream is a well known mechanism for forming a sequence of this type (Visher, 1965; Allen, 1965). Recent work by Allen (1970) has shown the quantitative relationship of parameters such as channel geometry, water surface slope, and flow velocity in controlling sedimentation within an alluvial channel. Because Allen's model is based on unidirectional flow alone, his findings cannot quantitatively be applied to these tidal sediments. However the importance of unidirectional ebb and flood currents suggests that there may be overall similarities between fluvial and tidal channel deposits.

Several other lines of evidence suggest a channel origin. Figure 57 shows an exposure in which one AB kindred is scouring into another. Along the steep margin of the scour is a breccia of sandstone fragments which are apparently locally derived. The occurrence of intraformational conglomerates in fluvial and tidal channels is well known (Kelling, 1968; Terwindt, 1971).

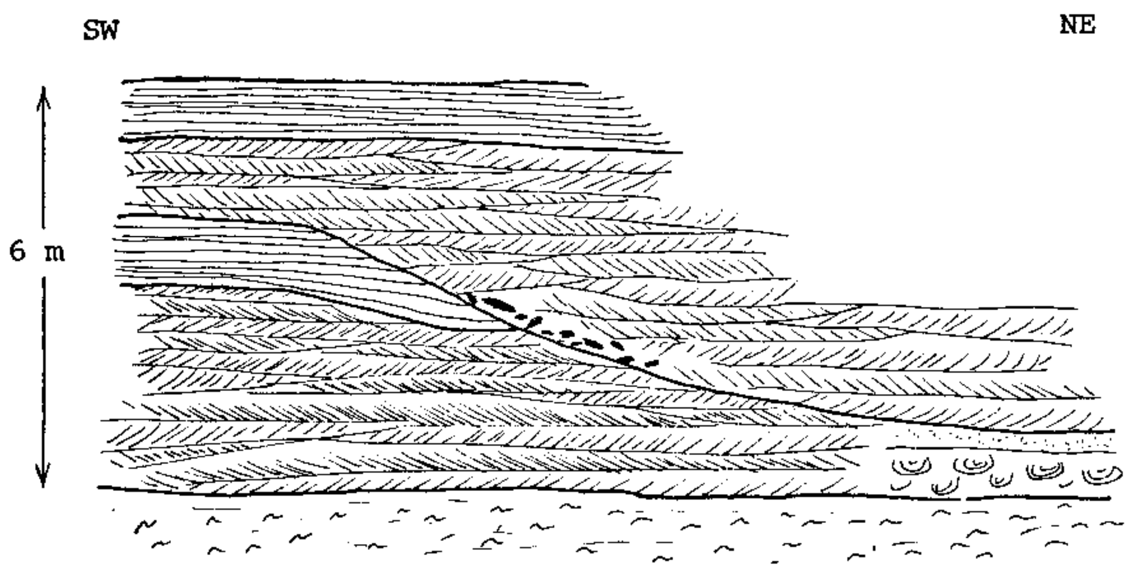


Figure 57. Channel margin at the base of an AB kindred in facies association 1, member 4, Nyborg Formation, between Larsholmen and Innerelv (locality L-17, fig. 69). The underlying silty-sandstone is the top of member 3. No vertical exaggeration.

Vertically from facies A upwards to facies B there is a slight decrease in mean grain size which is an attribute of meandering river and tidal channel point bar deposits (Visher, 1965; Oomkens and Terwindt, 1960).

Palaeocurrent observations from shallow marine subtidal sediments of the North Sea (Reineck, 1963) indicate that cross-sets having bimodal current and offshore tend to be associated with tidal channels or inlets. Cross bedding modes (fig. 54) in AB units trend parallel to the palaeoslope as indicated by flute marks in member 3, and perpendicular to the coastline as indicated by ripple marks at the top of member 3 (fig. 48). In rivers, cross bedding is thought to roughly parallel the channel in which they formed (Allen, 1966a). This suggests that channels in which the AB units may have formed were orientated perpendicular to the coastline and are in this respect similar to tidal channel deposits in the North Sea described by Reineck (1963).

This direction, NE-SW, corresponds to the orientation of the outcrop under study. This may explain why these units are laterally continuous for distances up to 3 km; the section exposes the channel deposits longitudinally rather than transversely.

Two additional ways of forming the AB units are: 1) a migrating subtidal sand bar and 2) a beach building out over an offshore dune field. Complex subtidal sand bars described by Reineck (1963) generally have palaeocurrents whose direction is related to the environment in which the bars exist. Offshore bars tend to have cross-beds with bipolar orientations trending oblique to, or parallel to the shoreline, while bars in estuaries and other large channels tend to trend perpendicular to the coastline. Thus, the palaeocurrent evidence suggests that although bars in channels may have been an important part of channel processes and important in the deposition of AB kindreds, it is unlikely that offshore bars were the environment of deposition. In addition, parallel-laminated sands seem to be more abundant in the shallow zones of coastal channels and inlets rather than on offshore bars which seem to be composed almost entirely of cross-bedded sands. However, the erosive base of many of the AB kindreds suggests that a migrating sand bar was not important as an environment. It is also difficult to see how a migrating sand bar could deposit the AB sequence.

Considering the second alternative, a beach building out over an offshore dunefield, a sequence attributed to such an event has been described from the Pleistocene of Oregon (Clifton et al., 1971). The sequence is 4 m high and the orientations in the dunes (numbering only a dozen) are highly varied. This irregular trend in shoreface cross-stratification was also described from the Cretaceous of New Mexico (Campbell, 1971). Here, the foreshore stratification dipped gently offshore, and built out over a coarsening upward sequence which has quiet water marine mudstones at the base. All of these features are not comparable with the AB kindred or its setting in the sequence, and consequently the regressive beach model can be ruled out.

A migrating channel appears to be the likeliest environment of deposition for the AB kindred.

CD Kindreds

Parallel-Laminated and Flaser Bedded Very Fine Sandstone

In certain parts of the section, facies C and D occur in close association, occasionally alternating from one to the other. Facies C may also be isolated in beds, but this occurrence is not considered here.

Whether the very fine sand bottom is shaped into a plane bed, or a rippled surface depends on the relative importance of unidirectional currents and wave activity respectively. The factors controlling the deposition of mud are somewhat more controversial. While it has been argued that a high suspended mud concentration and a low wave effectiveness cause mud deposition (McCave, 1970), it has generally been held that the marked changes in current strength may form sand and clay alternations, with the sand, either laminae or lenses, being deposited during high velocities, and the mud being deposited during slack water (Reineck, 1967).

In the present sequence, the occurrence of the mud largely as thin drapes, rather than thick internally laminated layers suggests to the present author that the tidal hypothesis is more applicable. This is consistent with the occurrence of herringbone cross-bedding, which is certainly tidal in origin in facies A. It is probable that the factors which McCave has stressed are more important in the offshore environment, particularly around the sand/mud boundary (McCave, 1971).

Although not formally part of member 4, the upper part of member 3, unit D can be considered in the present context. Of main interest is the alternations of sand and mud layers, the sand occurring as parallel-sided, occasionally finely laminated layers, averaging about 2-3 mm thick, separated by thin mud layers. This distinctive sediment has been termed "tidal bedding" (see Reineck, 1967) and unlike lenticular and flaser bedding, it has not been described from other than tidal environments to the author's knowledge. According to Reineck (1967) a sand lamina is deposited during the high velocity flow, but the exact mode of deposition is not specified. From the internal parallel laminated, traction deposition in the plane bed flow of the upper flow regime may be a possibility. This explanation does not appear to resolve the problem of why graded units do not form, except that changes in velocity may be too sudden (see De Raaf and Boersma, 1971).

The dual controls of type of energy input (wave versus current) and continuity of input (continuous versus alternating high and low energy level) can be portrayed diagramatically (Table 19). Under conditions of continuous activity, tidal current influence is weak, and mud-free deposits form. Where tidal influence is considerable, changing current velocity

	Energy Source	
	Wave Dominant	Current Dominant
Continuous Activity	Rippled Sandstone	Rippled/Parallel-Laminated Sandstone
Alternating Quiet and Activity	Flaser Bedding	"tidal" bedding

Table 19. Effect of energy source (wave or current) in shaping the bottom, and of the distribution of energy in time (continuous, or alternating with quiet) in depositing mud and sand in adjacent layers. This is meant to apply to a shallow tidal environment only.

and water depth will cause the intermittent deposition of mud and sand. Needless to say, this scheme only applies in a relative fashion to the present deposits, which are believed to have formed in a shallow, tidal environment. Furthermore the description of "wave" activity must also embrace wave drift, which in shallow water may be an important "current" in aid of sediment transport. The formation of ripples in the flaser facies may be thought of as a wave "imprint" on a unidirectional current rather than due only to the oscillatory motion of the water.

As for the overall environment of deposition of this kindred, the presence of only one example of low angle inclined bedding may be taken to indicate that sedimentation occurred vertically upon a horizontal surface. Deposition on point bars of laterally migrating, meandering tidal channels (Reineck, 1967) appears to be ruled out. Also, as described earlier, the absence of emergence criteria appears to rule out an intertidal origin. Possible environments include subtidal shoals, in an estuarine or lagoonal context, or subtidal nearshore deposits.

GH Kindreds

Beds of Slumped Sandstone and of Massive Sandstone

The vertical relationships between facies G and H (Table 20) supports the field observation that facies G slumped bed is often succeeded by a bed of facies H, massive sandstone.

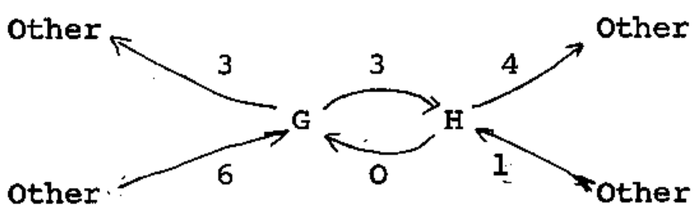


Table 20. Vertical transitions between slumped facies G beds and massive sandstones, facies H, and other facies.

A similar relationship is seen in the upper part of member 3, within the top 10 m. In general the kindred may be described as follows:

1) there is an erosive base upon which rests slumped sandstone, usually of facies B, parallel-laminated purple sandstone. The slumped sandstone is lenticular (Pls.75, 76b).

2) erosively above the slumped sandstone facies G is a massive sandstone bed facies H, laterally more extensive than the slumped bed (fig. 58).

From the abrupt, erosive contact between the two facies it appears that they were deposited as two separate primary depositional events. This can be contrasted with unit M (fig. 54) which consists of ball-and-pillow structure in the lower part, and massive sandstone above. The contact between

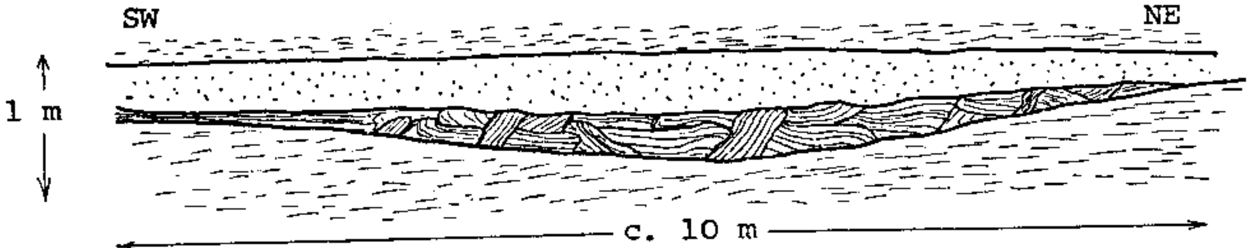


Figure 58. Slumped, parallel-laminated purple sandstone, facies G in a shallow channel, overlain erosively by massive sandstone, facies H. About 2 m below top of member 3, Nyborg Formation, adjacent lithology is rippled very fine sandstone and siltstone. Locality L-16 (fig. 69) between Larsholmen and Innerelv.

the two parts is sometimes diffuse as well as sharp, and there are blocks of the deformed sediment floating in the massive sandstone. This suggests that the massive sandstone formed by the churning of the laminated sandstone which, as it flowed upwards past the pillows, carried along fragments of sandstone with it.

Considering the occurrence of the GH kindred in the top of member 3, just the presence of parallel-laminated purple sandstone in that context implies currents of extraordinary strength. From the erosive base of the kindred, and the lenticular form, deposition within a channel seems likely. It is possible that the facies G and H are two parts of a channel-fill sequence related to a specific event. The slumping of facies B, parallel-laminated purple sandstone to form facies G slumped sandstone may be triggered off by the high bed shear stress associated with the deposition of facies H, massive sandstone. Just what the event was it is impossible to be certain of, but one possibility is the storm surge flood and ebb currents which are capable of considerable channel erosion and redeposition of sediment along a coastal area during a major storm (Hayes, 1967).

Other Facies Relationships

Facies E and facies F, beds with silty mudstone tend not to occur together in the section, but each facies occur in groups in separate parts of the section (fig. 54). These facies were thus not deposited at the same place at the same time.

Both facies E and F were deposited rapidly by currents containing a great deal of silt and clay. It is possible that these facies were deposited in different places simultaneously, and conversely, at the same place as the environments shifted laterally. This possibility is discussed below.

6.5.4 Facies Associations

Introduction

The facies and kindreds just discussed can be related on a larger scale in order to piece together the varied processes occurring over a long period of time within a major environment. The associations are divided according to the changing distribution of facies and kindreds in the vertical section (fig. 54).

Facies Association 1

Subtidal Shoals and Tidal Distributary Channels

This association includes the basal 18 m of member 4 at Stappogiedde North and consists of alternating AB, CD and GH kindreds (fig. 54). A transition matrix (Table 21a) of the sequence can be converted into a transition probability matrix (Table 21b) by taking the product of the proportions of each kindred involved in the transition to the total number of that kindred in the sequence. The probability matrix shows that the modal sequence is of the form AB->GH->CD->AB, with the sequence AB->GH->AB of secondary importance.

a)

	AB	CD	GH
AB	1	2	5
CD	3	0	1
GH	3	3	0

b)

	AB	CD	GH
AB	.018	.100	.52
CD	.107	0	.042
GH	.071	.100	0

Table 21. Vertical transitions of kindreds in facies association 1, member 4, Nyborg Formation, Stappogiedde North. a) frequency of transitions, b) probability of transitions.

Placing the AB kindred at the base of the sequences is reasonable if the formation of the AB->GH->CD sequence was related to the lateral migration of a channel in which coarser, higher energy facies (A and B) occur below finer, lower energy facies (C and D). In detail, it appears that while facies A and B accumulated within a channel (see 6.5.3, kindreds AB), facies C and D were deposited adjacent to the channel. From its position in the sequence, it would appear that facies G and H were preserved most frequently at the margin of the channel.

The topographic differentiation of grain size and sedimentary structures was apparently controlled by the distribution of wave and current activity. Tidal currents were strongest in the deep water of the channels. The higher bed shear stress there caused the formation of dunes. Towards the margin of the channels in shallower water, the plane bed formed. Along the channel, but outside of it, tidal currents were relatively weak and wave agitation, which is limited in intensity by depth, was most effective (fig. 59). A similar relationship between grain size and topography has been noted in the outer part of the Gironde estuary, where tidal currents and wave activity are thought to be the controlling processes (Allen, 1971).

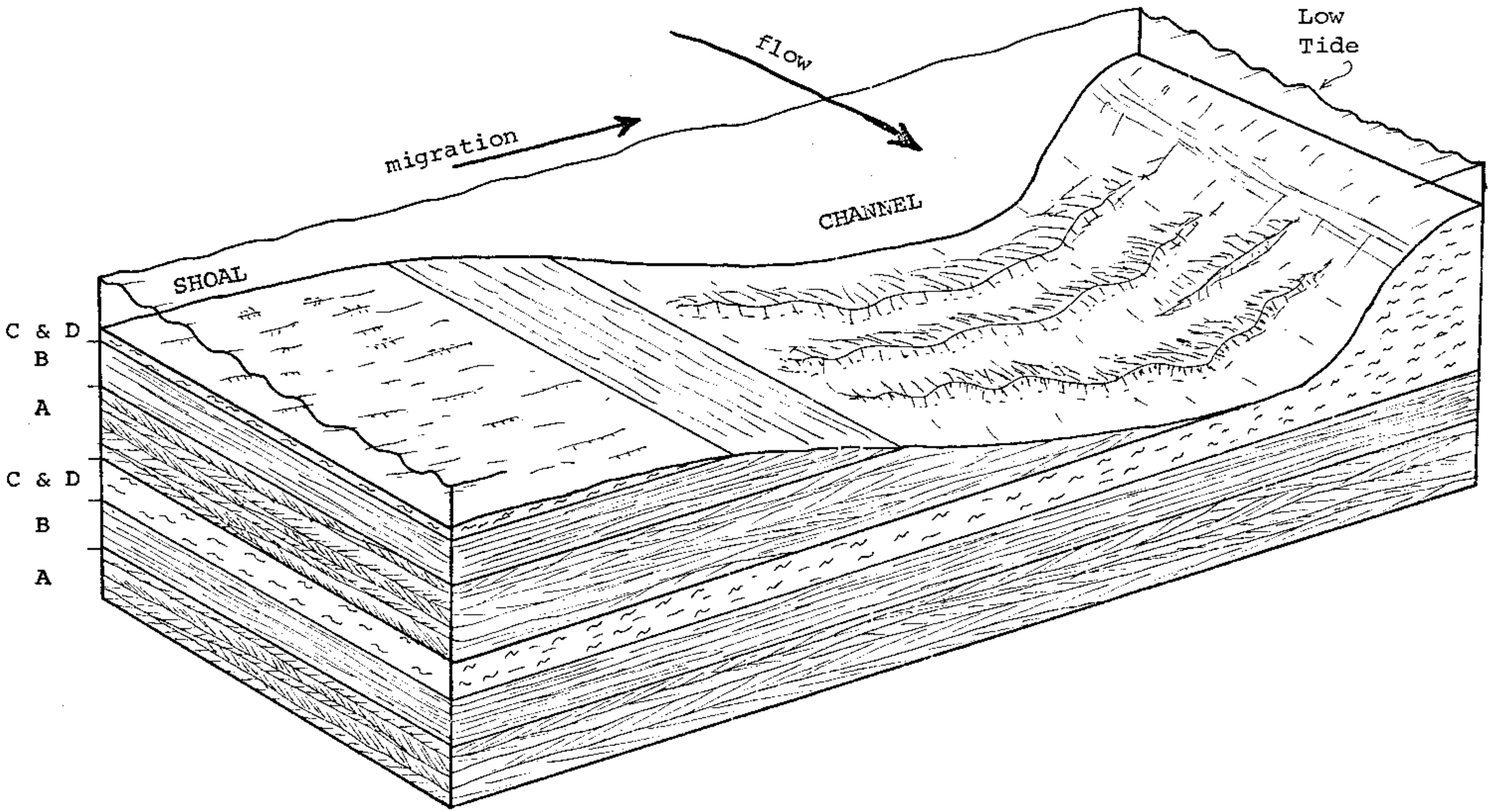


Figure 59. Channel and shoal model for deposition of facies association 1, member 4, Nyborg Formation, Stappogiedde North. Ebb and flood currents cause dune migration in the channel, flat bed at the channel margin. Waves and weak currents cause ripple migration on the shoals. Lateral migration of channels builds up a sequence of facies A, B, and C&D.

Unfortunately, the bed forms are not described in this report. However, the absence from member 4 of intertidal mudflats and channels, and beaches and barriers, as described in earlier sections, strengthens the interpretation of the environment as a subtidal area with a channel and shoal morphology.

It was suggested in an earlier section that kindred GH, deformed and massive sandstone, may have formed by rapid sedimentation, possibly during a storm. This style of deposition would contrast with that of kindreds AB and CD which formed by continued action of one process in one place over a long period of time, current or wave activity respectively. The presence of GH kindreds between AB and CD may be due to the restriction of the processes that formed it to the channel margin, or to subsequent erosion of storm deposits by later normal, "fair-weather" processes. Where the depth of reworking by normal tidal currents is considerable, i.e., where dunes are migrating, there would be little chance of the preservation of storm deposits. In the shoal areas, where the depth of reworking is probably less than a few centimetres as indicated by the absence of significant scours, it appears that GH was deposited only rarely, perhaps because the flow of storm currents was largely confined to the already developed subtidal channels.

The immature composition and texture of the purple sandstone and its occurrence at the top of a strongly regressive sequence suggests that the source of the sediment was more or less directly from a river, and not due to reworking or winnowing of offshore sediments. Thus the present channels are different from many present day tidal channels which, although they may be associated with outbuilding, formed in response to the Holocene transgression, and as a result were and still are supplied largely by offshore-derived sand. Regardless of this distinction, it is possible that with a fluvially dominated supply the only major difference would be petrographic, and not in the depositional structures of the tidally influenced sediments.

In summary, this facies association was laid down as migrating subtidal channels (kindred AB) and shoals (kindred CD). Channel deposition was dominated by normal tidal currents, while shoal deposition was dominated by wave activity. Occasionally, storm deposits (kindred GH) were preserved on the margins of the channel where the depth of reworking by normal currents was low. The supply of sediment was directly from a river, suggesting that the subtidal channels were distributaries.

Facies Association 2

Distributary - Lagoon Transition

This facies association, 7.5 m thick, consists of relatively thick CD kindreds separated by beds of facies A or AB kindreds, and by groups of facies E and facies F. These two latter facies first appear, and facies E is best developed, in this association.

Crossing from association 1 to 2: 1) AB kindreds decline sharply in thickness and frequency indicating the reduced importance of channel sedimentation. In addition these sandstones can be seen to wedge out over several hundred metres. Facies A and B also occur here as thin beds. 2) CD kindreds show a marked increase indicating the widespread existence of shoal areas, or a decrease in the rate of erosion by channels. 3) Groups of beds of facies F and, to a lesser extent, facies E occur intercalated with CD kindreds and their development above purple medium-grained sandstone (facies A and B) suggests that they occur within a fining upwards sequence (units N, O, P, and Q, R, S, fig. 54), and thus possibly in relation to a channel. 4) Facies G and H are rare in the Stappogiedde North section but occur at the same horizon at other localities as lenticular beds of deformed, (usually ball-and-pillow) purple sandstones. These are not overlain by massive sandstone as in facies association 1. These beds are laterally continuous for less than a few hundred metres, and may be up to 50 cm thick. No preferred relation to other facies was noted.

CD kindreds are again interpreted as "uniform" sedimentation deposits, but beds of facies A, B, E and F appear to be due to relatively short-lived events. The origin of these beds is further discussed below.

Facies Association 3

Lagoon

This association is characterized by the alternation of CD kindreds, facies F, and beds of facies A, B and E. Thick units of facies A and B are absent, facies E is scarcer than in the transitional association, and while beds of deformed (internally loaded and convoluted) sandstone are present, massive sandstone in distinct beds is absent (fig. 54).

The association is 12 m thick, 1 km north of Stappogiedde, but about 17 m are present at Areholmen (fig. 53). It is correlated with 70 m of similar rocks which comprise member 4 at Trollfjord (fig. 53). On the basis of the interpretation of the section at Trollfjord, this association seems to have been deposited in a protected shallow marine area, a lagoon which was finally overridden by a landward migrating barrier bar, and offshore marine sediments after being filled in by sediment both from land and from barrier washover.

The distinctive feature of this association is the haphazard intercalation of facies of contrasting grain size, sorting and colour. A transition matrix indicates that the sequence of facies is near to a random one in which successive beds are not related to each other preferentially. This suggests that successive, contrasting, genetically unrelated processes formed the sequence.

6.5.5 Trollfjord

Introduction

At Trollfjord, about 20 km NNE of Stappogiedde (fig. 5) the highest parts of the Nyborg Formation are exposed. Member 4 is 70 m thick and consists entirely of sediments similar to facies association 3 at Stappogiedde (fig. 53). The colour fluctuates between purple and grey-green. Member 5, 25 m thick, consists almost entirely of sandstone with two dolomite beds at the base. This member is observed only at Trollfjord.

Member 4

Member 4 consists of alternating beds highly contrasting in grain size, sorting and colour. The background sediment is largely laminated siltstone and flaser bedded sandstone. At the base are three thick beds of grey-green ball-and-pillow sandstones up to 2 m thick. Two other similar beds are found higher in the member. Interbedded at frequent intervals are thin to medium bedded, poorly sorted feldspathic sandstones, both grey-green and purple in colour. In addition, beds having the appearance of turbidites, showing Bouma's (1962) divisions, a and b occur in groups.

Upwards in the section, purple colour decreases and is replaced by grey-green colour, and then grey-brown near the top. About 15 m from the top of the member fine-grained white sandstones appear in thin, continuous beds. They are much better sorted and have a lower feldspar content than the surrounding sandstones which are drab coloured and poorly sorted. Ten metres from the top, ball-and-pillow and convoluted sandstones increase markedly in abundance and 5 m from the top, dolomite pebbles, small and rounded, appear at the bases of some graded sandstones.

Member 5

The base of member 5 is marked by a 60 cm parallel and irregularly laminated grey-brown dolomite bed which sharply overlies the top of member 4. Above a 1 m sandstone with abundant oscillation ripple marks is overlain by a second massive and parallel-laminated dolomite bed with dolomite pebbles. In thin section, the dolomite is finely laminated with quartz silt and sand grains, and thin platy fragments of dolomite scattered in a fine-grained dolomite matrix. Above, a sandstone similar to that between the dolomite beds is overlain by a dolomite pebble layer which marks the top of the basal part of member 5. The sandstone in the basal part is cross-bedded in part, contains many oscillation ripples, and consists of very well sorted coarse sand. Where the original grain boundaries are not destroyed by pressure solution they appear to be very well rounded. Feldspar is almost absent, composing less than 1% of the rock. The sandstone is a supermature quartzarenite (Folk, 1968).

Overlying the dolomitic basal part of member 5 is about 20 m of trough cross-bedded white sandstones with oscillation ripples, well rounded shale flakes in scours, and a few dolomite pebbles in layers. Parallel lamination and massive beds also occur. Ripple crests trend to 10°, 40°, 55° and 90°. No preferred orientation was noted. The sandstone consists of moderately sorted, rounded, fine to medium sand with a considerable amount of feldspar, including microcline and plagioclase. Accessory minerals include zircon and tourmaline. The sandstone is a mature subarkose. It is the highest unit observed in the Nyborg Formation, and is unconformably overlain by the Mortensnes Tillite.

Interpretation and Discussion

Considering the high feldspar content of most of the sandstones in the Nyborg Formation, the quartzarenite at the base of member 5 can be interpreted only as a beach deposit. In this context the adjacent dolomite beds may be coastal deposits, perhaps formed on inter- or supratidal flats. The platy dolomite fragments may be eroded algal mats. Similar associations in lithology are known in present day shorelines (e.g. Ball, 1967).

The gradual change from poorly sorted, immature sandstone up into well sorted mature sandstone through member 4 indicates the increasing marine influence. As the sandstones occur interbedded with laminated siltstone and flaser bedded very fine sandstone, these beds were introduced into a quiet environment during periods of exceptionally strong currents. This quiet environment, clearly not open marine, may have been a bay or lagoon. The sources of the sandstone beds in member 4 are varied. The poorly sorted immature sandstones were land (river) derived, and are similar to the medium sandstone which occur in facies association 1 at the base of member 4 at Stappogiedde North. In contrast, the well sorted sandstones, and the dolomite pebbles at the base of some of the sandstones were derived from marine areas. These may be resedimented equivalents of the shoreline deposits at the base of member 5.

A sequence of lagoonal deposits followed by beach deposits can be attributed to the landward migration of a barrier-lagoon couple (Fischer, 1961; Dillon, 1970). Applying this model to the present sequence, member 4 would represent lagoonal deposits, and member 5, barrier and offshore deposits. A principal mechanism by which a barrier migrates landward is by the transport of barrier sediment into the lagoon during storms by currents and large waves (Hayes, 1967), and by the progradation of barrier washover fans into the lagoon (Ball, 1967). Thus, the lagoon is naturally filled with sediment derived from the barrier. A rise in sea-level will ensure that some of the lagoon deposits will be preserved. Both the increased importance of marine processes, the landward migration of a barrier and the preservation of the lagoon deposits, suggest that a relative rise in sea-level occurred during the deposition of members 4 and 5.

With the limited amount of exposure, one can only speculate on the specific environments represented in member 5. The presence of dolomite suggests quiet conditions, and these would most likely occur along the landward side of the barrier. The absence of low-angle cross-stratification in the sandstones in the lower part of member 5 suggests that these did not form in the swash zone of the beach. High angle cross-stratification is reported to occur in bars in the lower parts of certain beaches and in tidal channels associated with beaches (Hoyt, 1967). The mineralogical immaturity and finer grain size of the sandstones in the upper part of member 5 suggests that these formed offshore of the quartzarenites. The marked change in lithology between the lower and upper parts of member 5 may be a disconformity produced by the landward migration of the barrier as erosion, associated with the shore face cut into upper shore face and foreshore deposits with the beach barrier and lagoonal deposits preserved below (Fischer, 1961). Thus, the upper sandstones formed offshore of the erosional zone (nearshore and possibly lower shore face), where net deposition occurred (e.g. Swift et al., 1971).

In summary, member 4 at Trollfjord formed by the filling of a lagoon, first from the landward side by terrigenous sediment, and then from the seaward side by sediment from a barrier. The development of the lagoon and the landward migration of the barrier over the lagoon indicate a relative rise in sea-level.

6.5.6 Discussion

Relationship Between Stappogiedde North and Trollfjord

The interpretatation of member 5 as, in part, a shoreline deposit, showed member 4 (at Trollfjord) to be a lagoonal deposit, thereby suggesting the transgressive nature of the sequence from the base of member 4 to member 5.

How did the events at Trollfjord relate to those recorded at Stappogiedde North? Four observations are particularly relevant here:

1) At Trollfjord, the transgression began with the establishing of a lagoon, marked by the base of member 4. Because of the similarity of facies association 3 at Stappogiedde North to member 4 at Trollfjord it is likely that both formed in similar environments.

2) At Stappogiedde, the presence of tidal channel deposits rather than fluvial channel deposits above a strongly regressive sequence is surprising. The dominance of tidal processes in the mouths of present day flooded rivers (estuaries) is due to the Holocene rise in sea-level.

3) Although a channel facies was observed at the base of member 4 at Stappogiedde North, none was observed at Trollfjord, suggesting that channel environments adjacent to the sediment source, probably a river, did not reach as far north as Trollfjord.

4) Pellets of glauconite occur in sharp based, graded sandstones, facies E, of facies association 2 at Stappogiedde North. The abundance of glauconite in certain transgressive, shallow marine deposits is well known (Burst, 1958). The mineral has also been described from an estuarine environment, where it occurs as layers and lenses, as opposed to the pelletoid marine glauconite (Porrenga, 1968).

Combining these observations we see that, at the time represented by the base of member 4, a channel and shoal environment marginal to a delta was established around Stappogiedde, while somewhere to the north of Trollfjord barriers began to develop cutting off the area between there and Stappogiedde from the open sea. In this area lagoonal sediments were deposited directly over shallow marine deposits. The formation of the barriers was probably a direct result of a relative rise in sea-level. A second consequence of the relative rise in sea-level was the flooding of the tidal distributaries (facies association 1) in the south so that lagoonal conditions (facies association 3) extended further southwards. This is represented by the sequence of facies associations 1, 2 and 3 at Stappogiedde North (fig. 60). The occurrence of glauconite is consistent with overall transgressive conditions, but its exact mode of origin is uncertain.

Cause of Transgression

The cause of the change from strongly regressive, to transgressive conditions may come under either or both of two categories:

1) Decrease in sediment supply caused by a) delta switch, b) change in climate, c) depletion of sediment available for transport.

2) A eustatic rise in sea-level, or an increase in the rate of subsidence.

The presence of terrigenous sandstones high into member 4 at Trollfjord appears to rule out the first group of causes. If the sediment supply from land had been cut off, then we would not expect to find this sediment appearing in the lagoon. Also, the influx of terrigenous sand into member 5 is suggested by the high amount of feldspar in the sandstone at the top of the member. Considering the grain size, the sand could not have been derived from the underlying lagoonal deposits

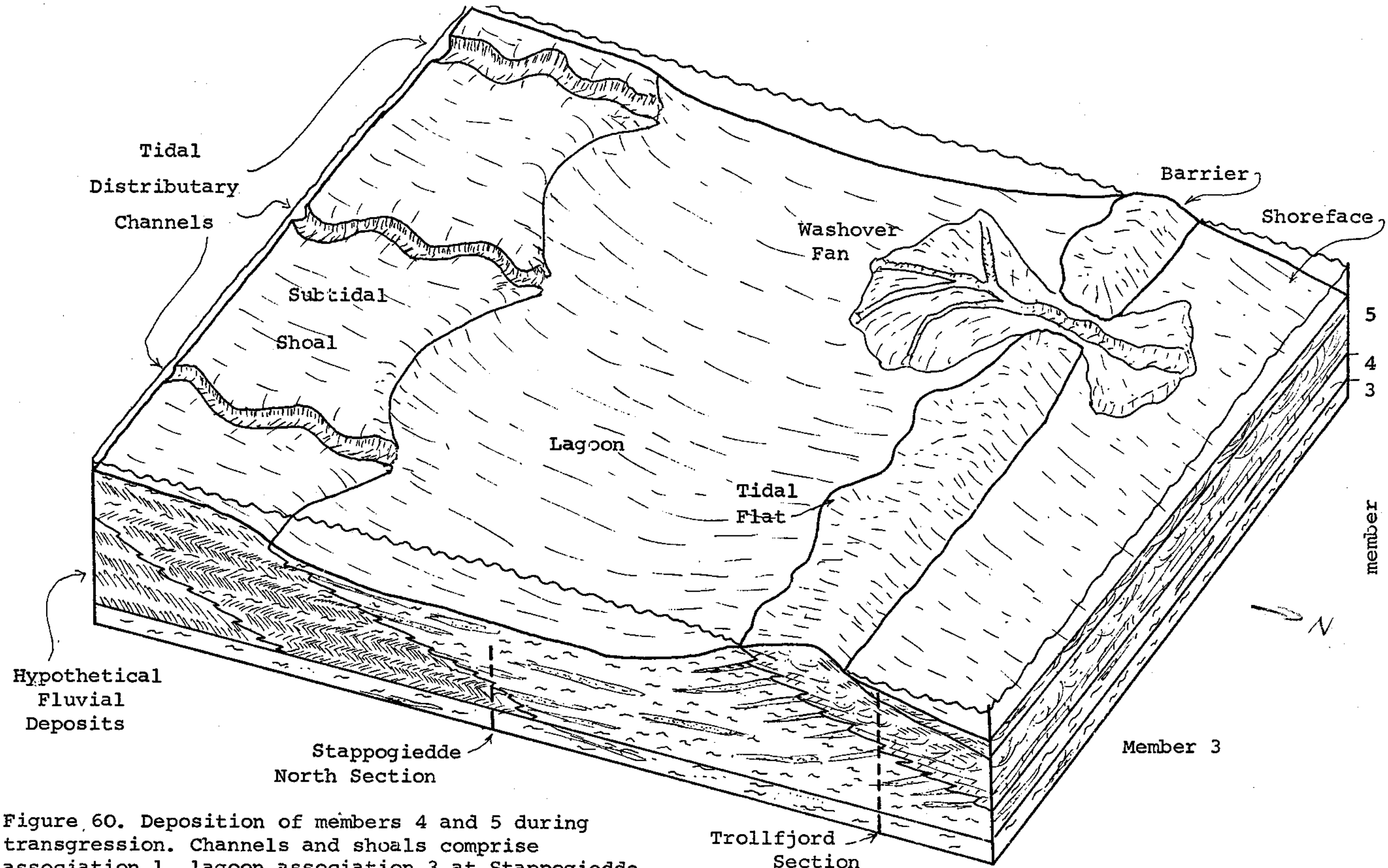


Figure 60. Deposition of members 4 and 5 during transgression. Channels and shoals comprise association 1, lagoon, association 3 at Stappogiedde North; lagoon, member 4, barrier and shoreface member 5 at Trollfjord.

into which it was eroding. However, the fact that terrigenous sediment continued to be supplied to the lagoon does not argue for or against a change in the rate of influx. It appears that without additional evidence the cause of the change to transgressive conditions cannot be determined. The thick accumulation of lagoonal sediments preserved indicates only that subsidence was taking place, and to some extent, a relative rise in sea-level must have occurred.

Normal and Catastrophic Sedimentation

We may define normal sedimentation as that which occurs by a frequently recurring, low energy process. During one event relatively little sediment is deposited, but over many events a substantial proportion of the sequence may be deposited. In contrast, catastrophic sedimentation occurs relatively rarely, but because high energies are involved, a large amount of sediment may be eroded or deposited. According to these definitions thick deposits of facies A (cross-bedded purple sandstone), B (parallel-laminated medium sandstone), C (parallel-laminated very fine sandstone), and D (flaser bedded very fine sandstone) are the products of normal sedimentation. For example facies A in an AB kindred is formed by the back and forth migration of dunes within a tidal channel. The coset is built up by a frequently recurring, relatively low energy event. Facies E (graded beds) and F (massive silty mudstone) were deposited by catastrophic processes. These were apparently high energy events, compared to the events characteristic of a given environment. For example the deposition of a purple poorly sorted sandstone, or a white well sorted sandstone in the lagoon involved transport from either a river, or the barrier respectively, under conditions of much higher energy than the gentle wave agitation and weak currents typical of the lagoon floor. The energy required to transport these beds was probably

6.5.7 Conclusions

The uppermost part of member 3, and members 4 and 5 were deposited in shallow marine environments following the regressive phase of members 2 and 3. The sections at Stappogiedde North and Trollfjord together show that members 4 and 5 formed during a transgression.

At Stappogiedde North three facies associations are recognized. Facies association 1 at the base consists largely of herringbone cross-stratification with a strong bimodal bipolar palaeocurrent pattern. The facies are well-ordered and appear to have been deposited by migrating tidal channels and intervening subtidal shoal areas. Facies association 2 forms a transition to association 3 which contains a mixture of coarse and fine facies and is interpreted as a lagoon deposit on the basis of observations at Trollfjord.

At Trollfjord member 4 consists of a normal, lagoonal, background sediment, siltstone and flaser bedded sandstone, with intercalations of purple, river-derived sandstone in the lower part, and well sorted mature marine-derived sandstone at the top of the member. These marine-derived sandstones and the dolomite pebbles which appear in several beds are believed to have been transported from a barrier during storm activity. Member 5 sharply overlies member 4 with dolomite and quartzarenites which are believed to be remnants of the barrier. Sharply above are well sorted subarkoses which may represent offshore deposits.

As the transgression began, barriers were established north of Trollfjord, so that an area including Trollfjord was transformed from open marine, to restricted lagoonal conditions. At the same time, tidal distributaries at Stappogiedde North were inundated by the rising sea-level, and covered by deposits formed in the southward migrating lagoon.

Both normal (fair-weather) and catastrophic (storm) processes can be recognized in the shallow marine deposits.

6.6 Conclusions

Each member of the Nyborg Formation provides important data for the determination of palaeogeography and palaeoslope. The formation contains within no evidence of glacial activity, and it seems probable that deglaciation was, for the present area, complete at the time.

6.6.1 Member 1, Transgression

Facies and facies distribution in the Vestertana area show that a transgression is represented by basal dolomite and shale. Transgression was on a surface with sufficient relief to influence the sedimentation of dolomite, and the mass movement of dolomite conglomerate. At Trollfjord, dolomite is a minor constituent of the member. This and the gradational contact with the underlying marine laminated tillite of the Smalfjord Tillite suggest that the area was not emergent during the transgression, and thus may have been topographically lower than the Vestertana area.

Palaeocurrent and slope indicators, soft-sediment faults, north of Varangerfjord suggest that the palaeoslope in that area was to the north during the deposition of member 1.

Thus a regional slope to the north, from Varangerfjord to Trollfjord existed during the time of deposition of member 1 (fig. 61).

6.6.2 Members 2 and 3, Basin Fill

Most of members 2 and 3 was deposited by turbidity currents in relatively deep, quiet water. Sole marks show NW currents around Varangerfjord, and north and NE currents around Tanafjord. Non-density currents, of uncertain origin flowed to the SW for a while around Vestertana.

Submarine fan channel deposits around Varangerfjord pass

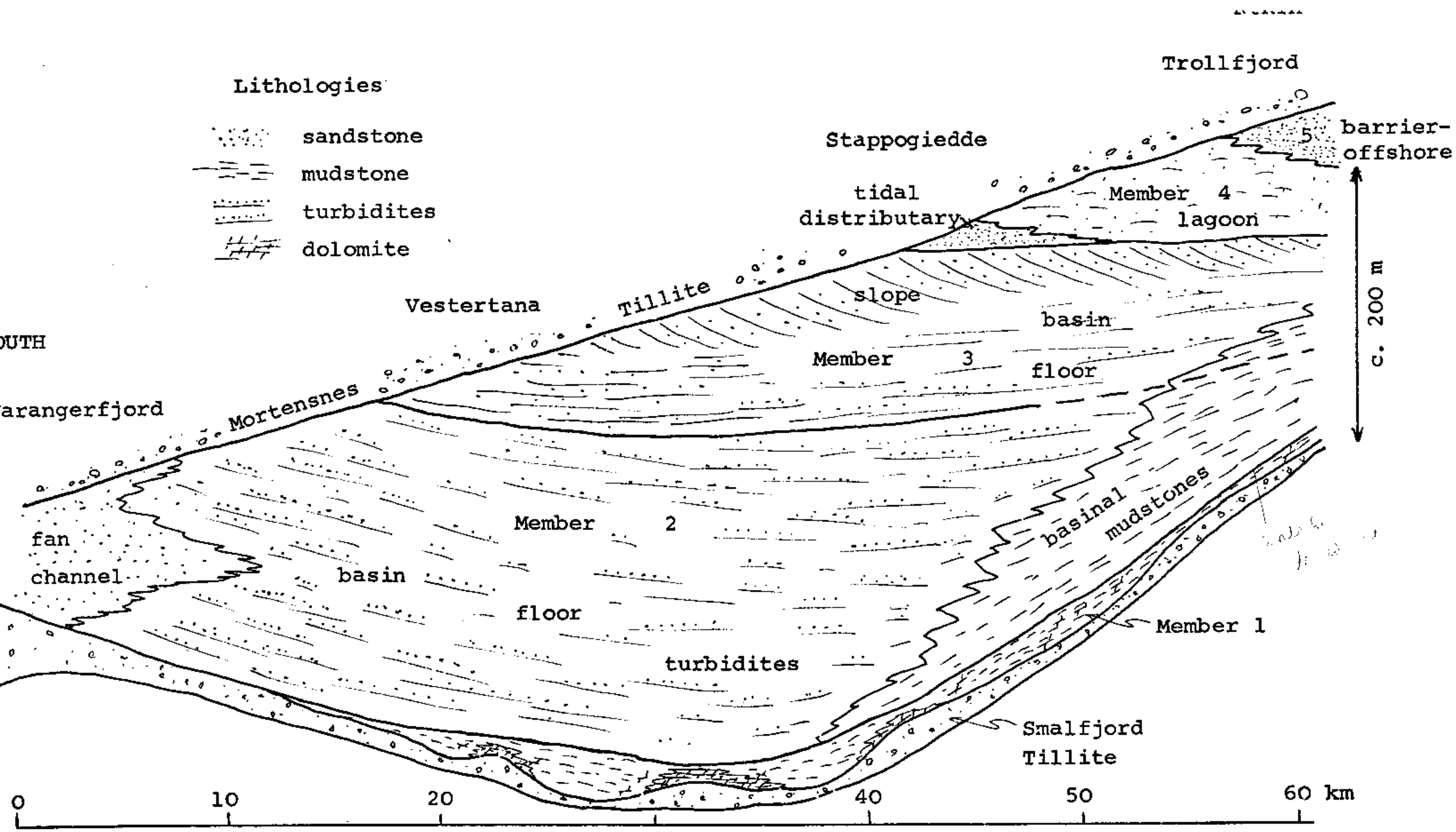


Figure 61. Simplified, partly hypothetical reconstruction of sediments and sedimentary environments in the Nyborg Formation. The source area was persistently to the south, and maximum subsidence was in the Vestertana area.

into fan deposits to the north. The diminished thickness of turbidites along the east side of Tanafjord, and the coarser grain size of the background sediment suggests that this area was relatively shallow at times.

6.6.3 Regressive Phase

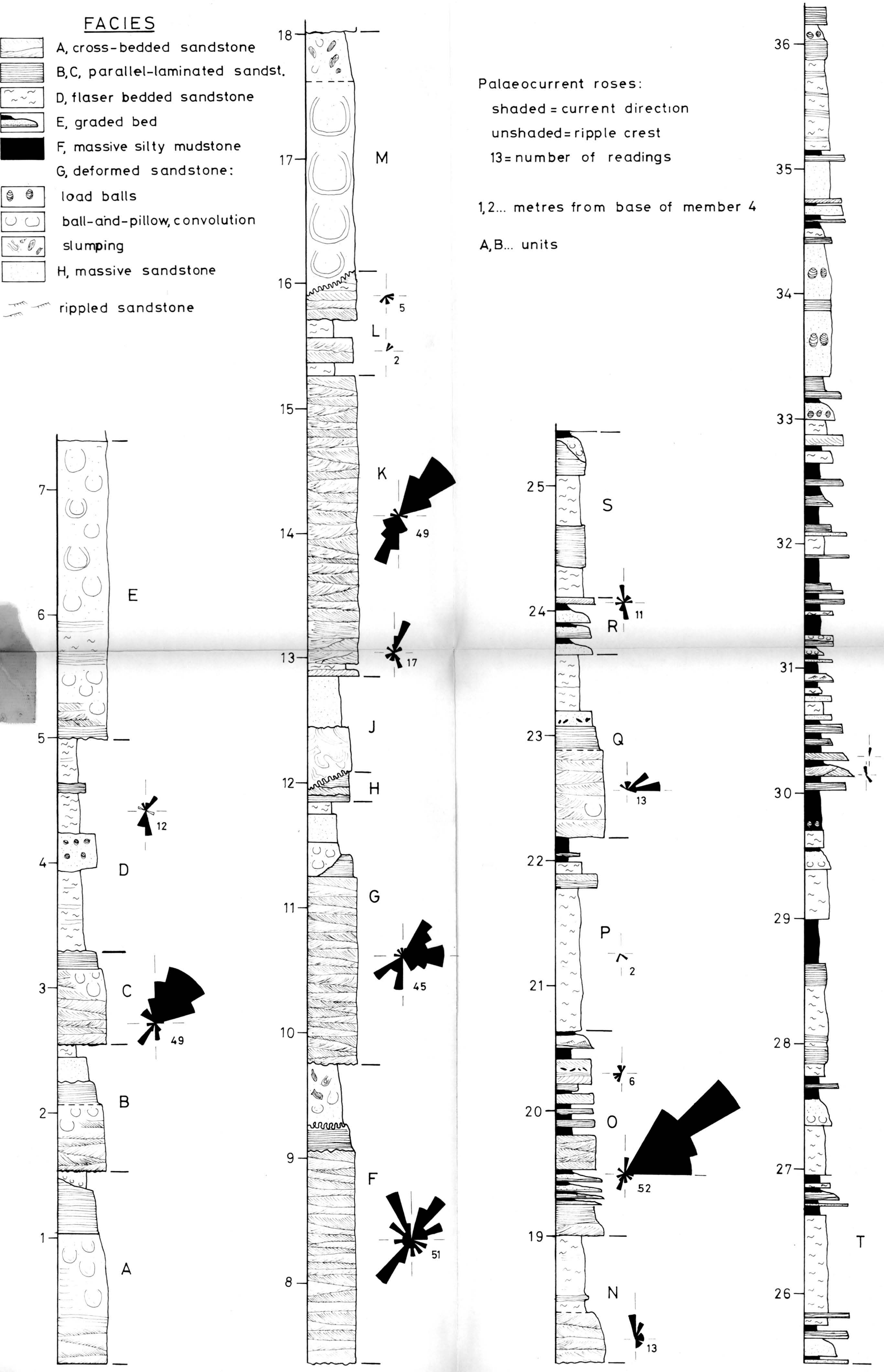
Sole marks of turbidites in the upper part of member 3 on the Digermul Peninsula show that a submarine slope, associated with the change from quiet water to shallow marine agitated environments prograded to the NNE. Unfortunately, the sequence has been eroded away beneath the Mortensnes Tillite to the south.

6.6.4 Shallow Marine Sedimentation

Palaeocurrents indicate that the coastline was orientated approximately WNW-ESE. A composite sequence of tidal channel distributaries, overlain by lagoonal deposits, followed by barrier and nearshore deposits indicate transgressive conditions. This seems to contradict the suggestion of Reading and Walker (1966, p.208) that the regressive sequence at the top of member 3 was caused by a eustatic drop in sea-level during the onset of the Mortensnes Tillite glaciation.

Figure 54

Member 4, Digermul Peninsula.
Facies association 1 (units A to M) and facies association 2 (units N to S) were measured at Stappogiedde North. Facies association 3 (T) was measured about 1 km north of Stappogiedde North, an inland exposure.
In all palaeocurrent róses north is towards the top. Units were devised in the field to aid in the lateral tracing of lithologies.


FACIES
A, cross-bedded sandstone
B,C, parallel-laminated sandst.
D, flaser bedded sandstone
E, graded bed
F, massive silty mudstone
G, deformed sandstone:
load balls
ball-and-pillow, convolution
slumping
H, massive sandstone
rippled sandstone
Palaeocurrent roses:
shaded = current direction
unshaded = ripple crest
13 = number of readings
1, 2... metres from base of member 4
A, B... units
A
B
C
D
E
F
G
H
J
K
L
M
N
O
P
Q
R
S
T