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Stratigraphy of the Neill Klinter Group; a Lower – lower Middle Jurassic tidal embayment succession, Jameson Land, East Greenland

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Cover
Exposure of the Neill Klinter Group in the type area along Hurry Inlet in the south-eastern part of Jameson Land. Sandstones and mudstones of the Neill Klinter Group overlie the scree-covered Kap Stewart Group. Fine-grained units in the Neill Klinter Group are intruded by Tertiary sills. The view is from Astartekløft towards the north. The airport at Constable Pynt is seen on the second delta. Exposed section *c.* 200 m thick.

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Abstract

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The sediments of the Neill Klinter Group of Jameson Land, East Greenland were deposited in a wide, shallow, wave, storm and tidally-influenced marine embayment situated at the western margin of the Jurassic seaway between Greenland and Norway. The group is formally defined and a new lithostratigraphic scheme is erected on the basis of recent sedimentological, biostratigraphic and sequence stratigraphic studies. The Neill Klinter Formation is changed to group rank; the Rævekløft, Gule Horn and Ostreaelv Members are revised and raised in status to formations, and the Sortehat Formation is redefined and included in the group. The Gule Horn Formation is divided into two new members and the Ostreaelv Formation into seven new members. The Neill Klinter Group is up to 450 m thick.

Sandy and muddy material was transported into the embayment from source areas to the east, west and north. A number of sub-environments are represented in the succession, including: restricted and bioturbated offshore, storm-dominated offshore transition zone, wave and storm-dominated shoreface, storm-dominated sandy shoal, subtidal sand sheet, ebb-tidal delta, tidal channel, wave and storm-dominated lagoon, and ephemeral stream delta.

The Neill Klinter Group consists of seven sequences and is characterised by a near absence of parasequences, interpreted as reflecting high influx rates of sand into the land-locked embayment. Continuous filling of accommodation space and erosion resulted in amalgamation of sedimentary packages, and poor development of facies cyclicity.

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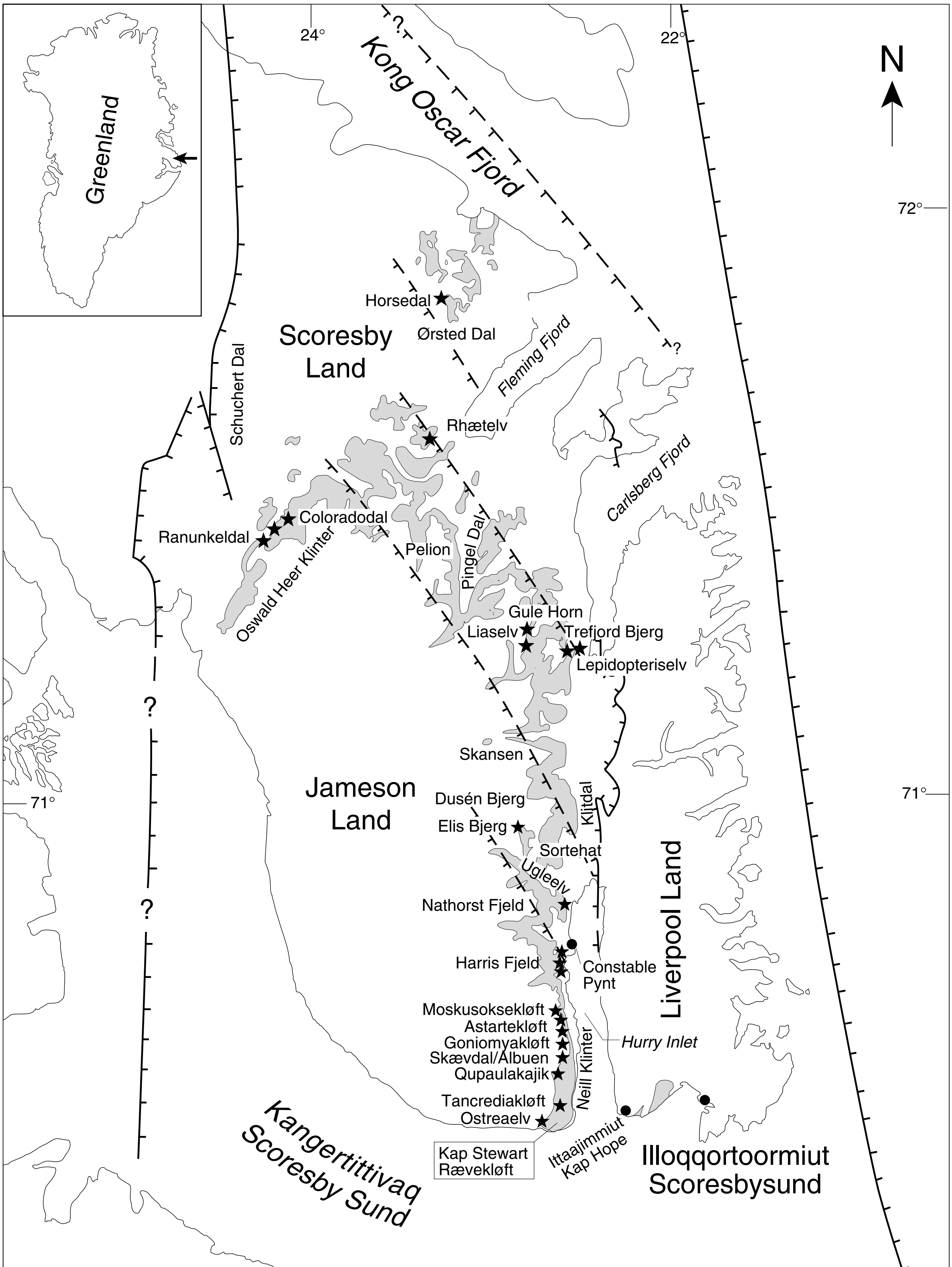

Fig. 1. Map of the Jameson Land region showing outcrops of the Neill Klinter Group and location of measured sections described in this work. Simplified after maps of the Geological Survey of Greenland. Modified from Dam & Surlyk (1995).

Previous investigations

Lower Jurassic deposits in the Scoresby Sund region were discovered in 1822 by William Scoresby Jr. at Kap Stewart, the south-easternmost point of Jameson Land (Fig. 1). Since then the Lower Jurassic of Jameson Land, especially the exposures in the cliffs of Neill Klinter along the west coast of Hurry Inlet, has been the subject of several studies. The early exploration history was documented by Rosenkrantz (1934), Koch (1939) and Donovan (1957).

Den østgrønlandske Expedition led by V. Ryder in 1891–92 reinvestigated the classical locality of W. Scoresby Jr. at Kap Stewart and along the Neill Klinter (Fig. 1) and was the first of a series of expeditions to carry out research in this area. The rich Rhaetian – Lower Jurassic floras of the Kap Stewart Formation (Hartz, 1896) were discovered during Ryder's expedition. Marine faunas of the formation overlying the plant-bearing deposits were described and figured by Lundgren (1895), who dated them to the Callovian, although the material included an ammonite fragment which was later determined as belonging to the Pliensbachian Jamesoni Zone.

A French expedition led by J. B. Charcot visited Scoresby Sund in 1925, and collected fossils from the Lower Jurassic marine strata at Kap Stewart (Fig. 1). The results were summarised by Haug (1926) who recognised the Pliensbachian age of the fauna which had previously been assigned to the Callovian by Lundgren (1895). Haug (1926) listed 27 species, but the collection was not described or figured in any detail.

Systematic geological investigations in East Greenland were initiated during the first expedition of Lauge Koch in 1926–27. T. M. Harris described the Rhaetian – Lower Jurassic floras from the coastal sections of Hurry Inlet in a series of monographs (Harris, 1926, 1931a, b, 1932a, b, 1935, 1937) and the Lower Jurassic deposits of Jameson Land were for the first time studied in detail by A. Rosenkrantz who erected the Neills Cliff Formation (Rosenkrantz, 1929) and sampled in the south-eastern part of the basin. Extensive collections were made of the invertebrate faunas, numbering more than 140 Pliensbachian species and 60 Toarcian species. The faunas were listed in subsequent stratigraphic publications together with a preliminary study of the geology of the Lower Jurassic in south-eastern Jameson Land and Liverpool Land (Rosenkrantz, 1929, 1934, 1942). Illustrations and descriptions of the crustaceans *Glyphea rosenkrantzi* Van Straelen and *Glyphea* sp., and the bivalve *Velata hartzi* Rosenkrantz were given by van Straelen (1929) and Rosenkrantz (1956). Doyle (1991) described Lower Jurassic belemnites from Jameson Land and Liverpool Land collected by Rosenkrantz, and discussed their stratigraphical and biogeographical significance. The remaining part of the fauna has never been described or figured.

The next expedition, also planned and led by Lauge Koch, was the Danish Two-Year Expedition which started in 1936. During the expedition geological mapping and stratigraphical investigations were undertaken of the coastal area of East Greenland from Scoresby Sund, at latitude 70°N, to Kuhn Ø at 75°N. No special attention was paid to the Lower Jurassic succession during this expedition, but a brief account of the geology of Jameson Land was given by Stauber (1940).

Following the expeditions of Lauge Koch no major attention was paid to the Lower Jurassic of Jameson Land until a 5-year mapping programme was initiated in 1968 by the Geological Survey of Greenland and the University of Copenhagen. Surlyk *et al.* (1973) established a lithostratigraphic scheme for the uppermost Triassic – Lower Cretaceous succession of Jameson Land and Scoresby Land. It included the Lower Jurassic Neill Klinter Formation, which was divided into the Rævekløft, Gule Horn and Ostreaelv Members. The Sortehat Member was included in the Middle Jurassic Vardekløft Formation. Upper Triassic – Jurassic trace fossils were described by Bromley & Asgaard (1972), and a palynological study of the Jurassic formations of Surlyk *et al.* (1973) in south-eastern Jameson Land was undertaken by Lund & Pedersen (1985).

The first sedimentological study of the Neill Klinter Formation in south-eastern Jameson Land was presented by Sykes (1974) who recognised a series of stacked coarsening-upward units, 8–40 m thick, which he interpreted as regressive offshore-estuarine successions. Sykes (1974) recognised shelf deposits which passed up-wards into cross-bedded sandstones representing fields of migrating dunes and mega-ripples which had accumulated in low-relief estuarine channel extensions and shoals. A WNW–ESE trending coastline was suggested, with strong bi-directional tidal currents operating normal to it. A brief lithological description of the Neill

Klinter Formation of Surlyk *et al.* (1973) in the Kap Hope area on Liverpool Land was given by Birkenmajer (1976; Fig. 1).

The palaeogeographic setting of the Lower Jurassic Neill Klinter Formation and the Sortehat Member (*sensu* Surlyk *et al.*, 1973) and their position within the sedimentary evolution of the Jameson Land Basin was described by Surlyk (1977a, 1990b) and Surlyk *et al.* (1981), and a low-resolution sequence stratigraphic interpretation was presented by Surlyk (1990a, 1991). The diverse trace fossil assemblages of the Neill Klinter Formation and their palaeoenvironmental significance were described by Dam (1990a, b), and a detailed sequence stratigraphic interpretation of the Neill Klinter Formation and a correlation with the time-equivalent Lower Jurassic succession on the mid-Norwegian shelf was undertaken by Dam & Surlyk (1995). Dam & Surlyk (1995) also included a new preliminary lithostratigraphy for the Neill Klinter Formation. However, this lithostratigraphy did not respect all the recommendations from the North American Commission on Stratigraphic Nomenclature (1983) or the International Stratigraphic Guide (Salvador, 1994), and some revisions have been made in the present paper. A micropalaeontological study of the Neill Klinter Formation has recently been completed (Koppelhus & Dam, in press; Koppelhus & Hansen, in press), and the organic geochemistry of the Sortehat Formation has been studied by Krabbe *et al.* (1994).

Since the work of Sykes (1974) no detailed sedimentological studies of the Neill Klinter Formation (*sensu* Surlyk *et al.*, 1973) have been undertaken until the results of recent field work presented in Dam (1991), Dam & Surlyk (1995) and in this study.

Geological setting

The Late Palaeozoic – Mesozoic Jameson Land Basin is located in the present-day land areas of Jameson Land and Scoresby Land, at the southern end of the East Greenland rift system (Figs 1, 2). This system is part of a larger rift complex located between Greenland and Norway before the opening of the North Atlantic Ocean (e.g. Ziegler, 1988; Fig. 2). The Jameson Land Basin is bounded to the east and west by major N–S trending faults, and to the north, by a number of reactivated Devonian NW–SE cross-faults (Surlyk, 1977a, 1978, 1990; Dam *et al.*, 1995). The southern boundary is unknown, but the basin probably extended south of Scoresby Sund (Fig. 1) where it is buried beneath Tertiary plateau basalts. The basin was initiated in the Devonian by extensional collapse of the over-thickened crust of the Caledonian mountain belt. The Devonian phase was probably associated with strike-slip or oblique-slip deformation resulting in the development of NW–SE trending transverse faults in the north-eastern part of the Jameson Land Basin (Fig. 1). This regime changed during Late Carboniferous – Early Permian times to a more orthogonal extensional stress field, resulting in development of basin margin half grabens (Surlyk *et al.*, 1984, 1986; Surlyk, 1990b; Larsen & Marcussen, 1992). The period of extensional tectonism was followed by Late Permian through Cretaceous subsidence governed by cooling and thermal contraction, interrupted by episodes of rifting and faulting in the Triassic, Middle Jurassic and Late Jurassic – Early Cretaceous (Surlyk, 1977a, b, 1990b; Clemmensen, 1980a; Surlyk *et al.*, 1981, 1986; Larsen & Marcussen, 1992).

Triassic – early Middle Jurassic sedimentation in the Jameson Land Basin was influenced by tectonic subsidence, climate, drainage pattern and eustatic sea-level changes. This time interval records a long term change from a warm arid to a more temperate humid climate. Early Triassic continental red beds with subordinate carbonates and evaporites, indicating an arid climate, were gradually succeeded during Middle to Late Triassic times by playa mudstones, lacustrine carbonates and flood plain mudstones and sandstones indicative of a more humid climate (Clemmensen 1978a, b, 1979, 1980a, b; Bromley & Asgaard, 1979). By latest Triassic time the climate had become temperate and humid as indicated by the disappearance of red beds, evaporites and carbonates, and by a change to drab and dark coloured sediments and the incoming of rootlet horizons, thin coal beds and perennial lacustrine deposits of the Kap Stewart Formation (Fig. 3; Dam, 1991; Dam & Surlyk, 1992, 1993; Surlyk *et al.,* 1993). This long term

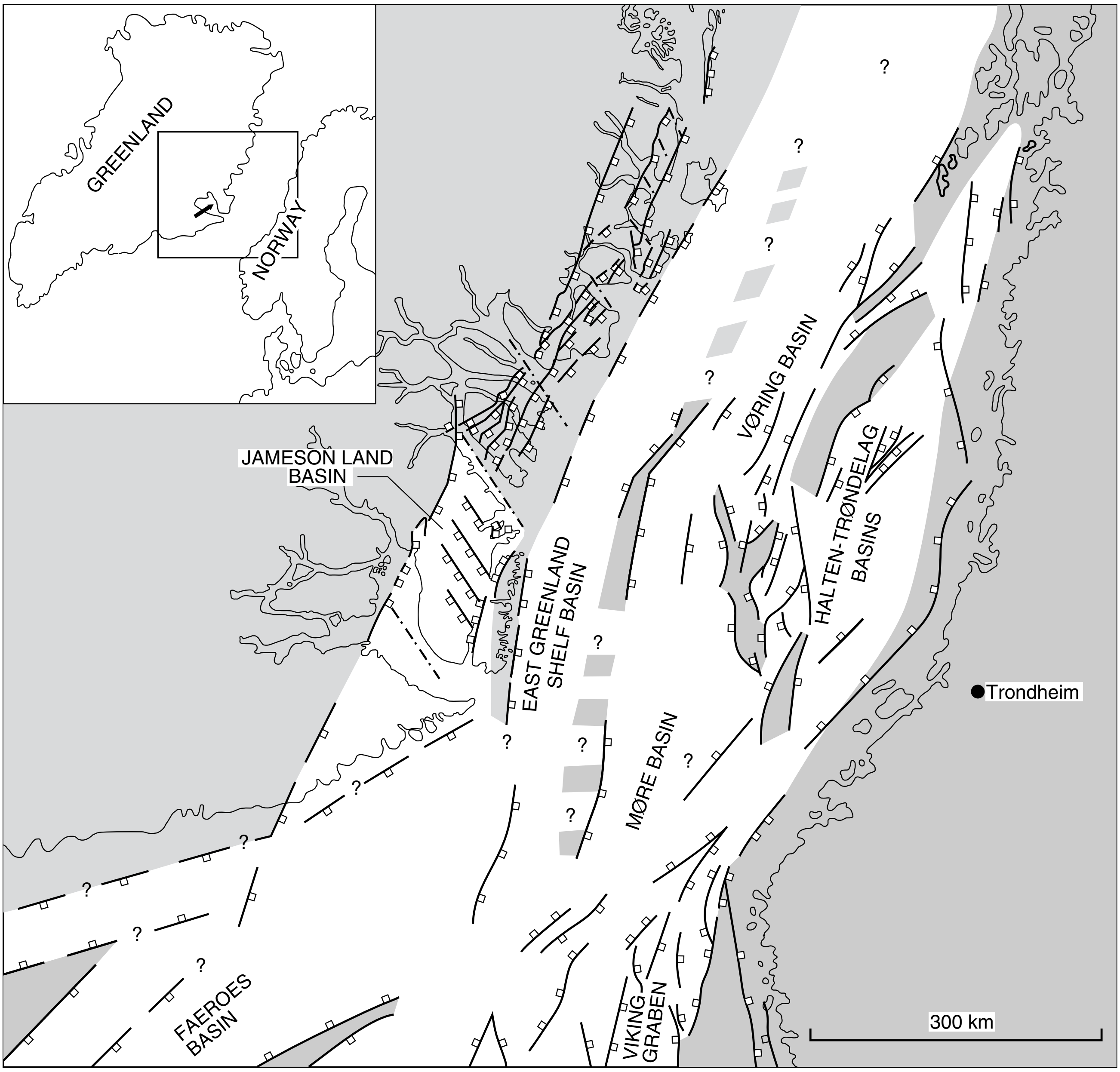

Fig. 2. Tectonic map of the North Atlantic region in Early Jurassic times showing reconstructed position of Greenland and Norway. The tone indicates position of major highs. From Dam & Surlyk (1995).

climatic change was probably caused by a gradual drift of the Triassic continent from a subtropical near-equatorial position in Middle Triassic times to a temperate position in Rhaetian – Early Jurassic times (Smith & Briden, 1977; Clemmensen, 1980a). This was accompanied by a long-term eustatic sea-level rise during the Early Jurassic and the Rhaetian – Sinemurian lacustrine complex turned into a shallow marine embayment marking the first fully marine inundation of the basin since the Late Permian – Early Triassic. The embayment was restricted to the same depositional basin as the lacustrine complex.

The shallow marine, Lower – lower Middle Jurassic Neill Klinter Group occurs in Jameson Land and Scoresby Land, and a small fault-bounded outlier is exposed in southern Liverpool Land (Fig. 1). In exposures the group varies in thickness from 300 m along the eastern basin margin to *c.* 450 m in the central part of the basin. The group and most of its formations and members show an overall sheet geometry, although the thicknesses

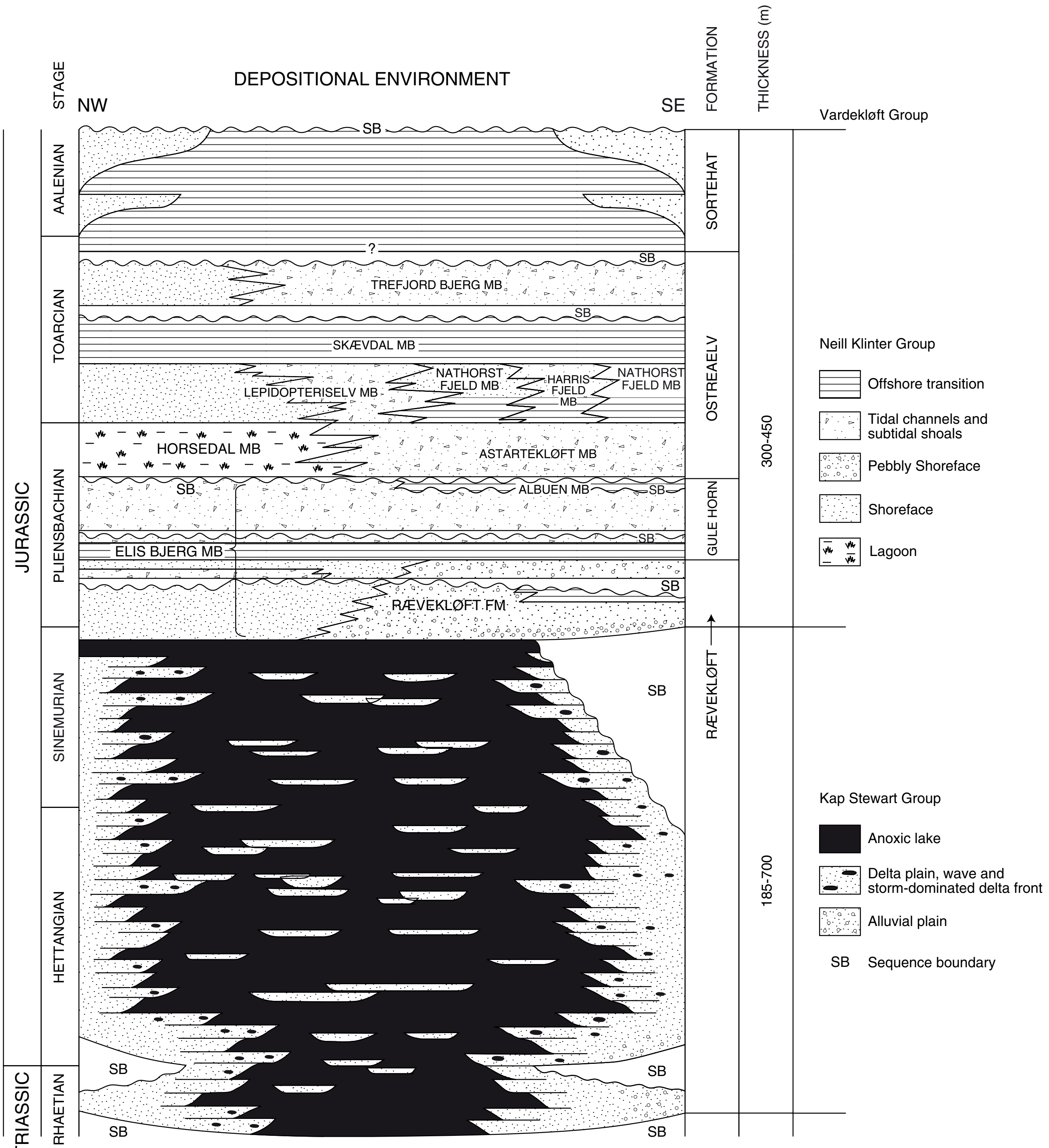

Fig. 3. Stratigraphy and depositional environments of the Upper Triassic – lower Middle Jurassic succession of the Jameson Land Basin. Modified from Dam & Surlyk (1995).

are greatest in the basin centre and decrease towards the margins. The boundary between the Kap Stewart Group (Surlyk *et al.*, in press) and Neill Klinter Group is developed as an erosional unconformity along the south-eastern basin margin, representing a major hiatus corresponding to the Sinemurian Stage (Fig. 3).The unconformity passes basinwards into a correlative conformity and the contact between the lacustrine mudstones of the Kap Stewart Group (Surlyk *et al.*, in press) and the shallow marine sandstones of the Neill Klinter Group is gradational (Fig. 4; Dam, 1991; Dam & Surlyk 1992, 1993). The Neill Klinter Group is overlain with a sharp contact by the Bajocian – Callovian Vardekløft Group (Fig. 3; Surlyk *et al.*, 1973; Surlyk 1990a, b, 1991; Engkilde & Surlyk, in press).

The sediments of the Neill Klinter Group were deposited in a shallow marine embayment during a period of relative tectonic quiescence and the facies pattern was mainly controlled by relative sea-level fluctuations, sediment influx and basinal currents, and differential subsidence along reactivated deep-seated Devonian faults.

Sykes (1974) recognised that the Gule Horn Member (Elis Bjerg Member in this paper) is made up of a series of stacked coarsening-upward successions, 8–40 m thick, in the south-eastern part of the basin. He proposed a depositional model similar to the recent Heligoland Bight of the southern North Sea and interpreted the coarsening-upward units as formed by progradation of estuaries across a shallow shelf. This simplified interpretation is not supported by the present study. Estuarine deposition takes place during transgression, and the whole succession shows very large variations.

The Jurassic – lowermost Cretaceous succession of East Greenland was interpreted within a large-scale low-resolution sequence stratigraphic framework by Surlyk (1990a, 1991). The deposits of the Lower Jurassic Neill Klinter Formation were considered as part of a transgressive systems tract passing into a thin highstand systems tract. The boundary between the two systems tracts was located within the lower, more fine-grained part of the Sortehat Formation.

A high-resolution sequence stratigraphic study of the Neill Klinter Formation and sequence stratigraphic correlation with the equivalent formations on the mid-Norwegian shelf was presented by Dam & Surlyk (1995). The Pliensbachian – Toarcian succession in both regions consists of six correlative sequences.

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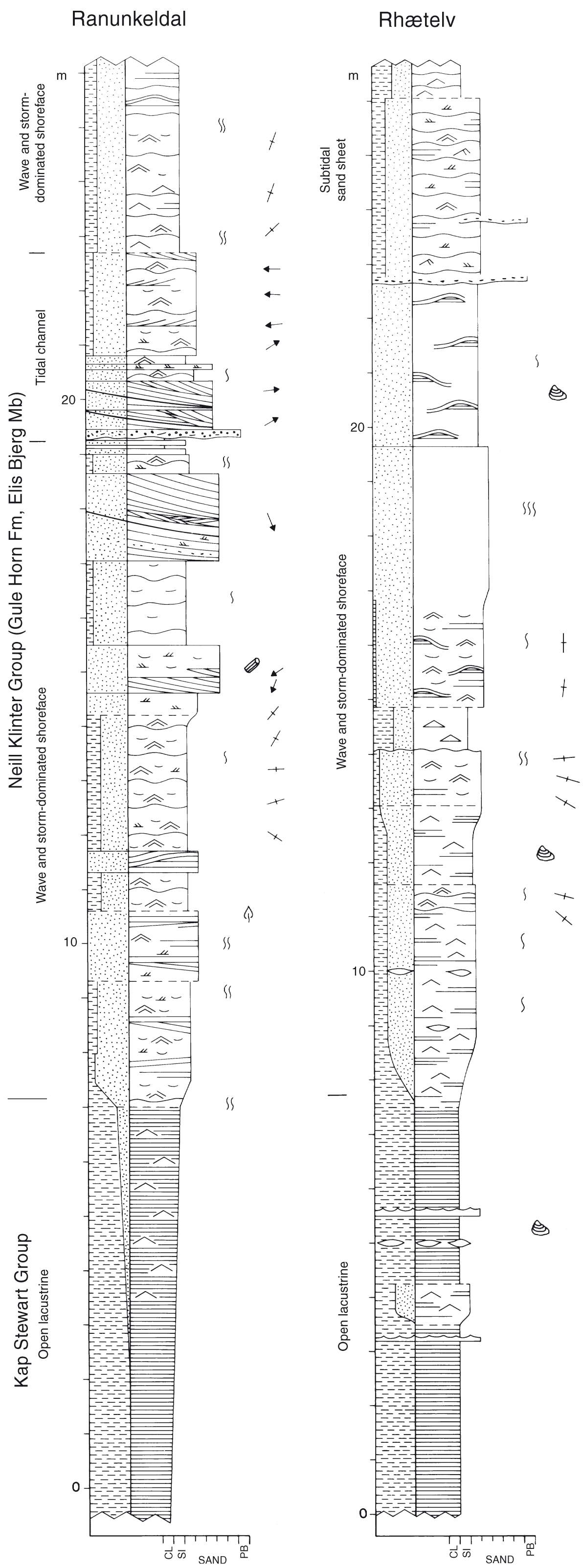


Fig. 4. Sedimentological logs from the north-western (Ranunkeldal) and northern (Rhætelv) parts of the basin showing the gradual transition from the fresh water lacustrine mudstones of the Kap Stewart Group to wave and storm-dominated upper shoreface (Facies association e) and subtidal sand sheet deposits (Facies association f) of the Elis Bjerg Member of the Gule Horn Formation. See Fig. 1 for locations and Plate 1 for legend.

Lithostratigraphy

The lithostratigraphic scheme of Surlyk *et al.* (1973) was the result of a regional 1:100 000 mapping programme of Jameson Land and Scoresby Land. Recent sedimentological and sequence stratigraphic studies allow subdivision on a much finer scale. The possibility of tracing very thin stratigraphic slices over several tens of kilometres necessitates a more elaborate terminology.

In the following section the Neill Klinter Formation of Surlyk *et al.* (1973) changes rank to group; the Rævekløft, Gule Horn and Ostreaelv Members undergo minor revisions and change rank to formations. The Gule Horn Formation is divided into two new members and the Ostreaelv Formation into seven new members (Fig. 3). The Sortehat Formation is redefined and included in the Neill Klinter Group. The new scheme is part of a major revision of the Jurassic – lowermost Cretaceous lithostratigraphy in East Greenland currently taking place (Surlyk *et al.* in press) in which the overlying Vardekløft Formation and the underlying Kap Stewart Formation change their status to group. These names are used in this bulletin (Fig. 3).

Neill Klinter Group

new group

History. The Neill Klinter Group includes the Neill Klinter Formation (*sensu* Surlyk *et al.*, 1973) and the Sortehat Member of the Vardekløft Formation (*sensu* Surlyk *et al.*,1973). The Neill Klinter Formation was recognised as a stratigraphic unit and named the Neills Cliff Formation by Rosenkrantz (1929) in the south-eastern part of the basin. Surlyk *et al.* (1973) redefined and amended the name and extended the formation to the northern and north-western parts of the basin (Fig. 1). They established the boundaries of the formation and subdivided it into the Rævekløft Member (base), Gule Horn Member and Ostreaelv Member (top). Detailed summaries of earlier investigations of the strata composing the Neill Klinter Formation were given by Rosenkrantz (1934) and Donovan (1957), and in the section on previous investigations above. Strata now forming the Sortehat Formation were initially defined as the basal member of the Middle Jurassic Vardekløft Formation (Surlyk *et al.*, 1973) and more recently excluded from that formation and first established as a separate member of the Neill Klinter Formation and later as a separate formation (Surlyk, 1990a, b, 1991). Detailed sedimentological and stratigraphical work have shown that the Sortehat Member of Surlyk *et al.* (1973) is genetically linked to the Neill Klinter Group, and is therefore included in the group as a formation.

Name. From Neill Klinter, the cliff section on the west side of Hurry Inlet in the south-eastern part of Jameson Land (Fig. 1).

Type area and reference localities. The type area for the group is the cliffs of Neill Klinter that are composed mainly of sediments of this group. Well-exposed reference sections occur at Harris Fjeld, Nathorst Fjeld, on the northern slope of Elis Bjerg, Lepidopteriselv, Liaselv, Rhætelv, on the eastern side of Horsedal and in Ranunkeldal (Fig. 1).

Thickness. The group is about 300 m thick at Neill Klinter, 360 m at Lepidopteriselv, approximately 450 m in Ørkenbjergene (Rhætelv), and 270 m at Horsedal (the thickness of the Sortehat Formation is not included at this locality).

Lithology, facies associations and depositional environments. A detailed lithological description of the strata that now make up the group in the type area was given by Rosenkrantz (1934). The formation commences with fossiliferous arkosic sandstones of the Rævekløft Formation. These are overlain by heterolithic micaceous quartz sandstones, mudstones, clean sandstones and sandy mudstones. Thin conglomerates are intercalated at several levels. The sediments show a great variety of primary sedimentary structures and trace fossils (Surlyk *et al.*,1973; Dam, 1990a, b). The group is topped by a thick succession of silty mudstones. Detailed descriptions of facies, facies associations and interpretation of the depositional environments of the group are presented under the individual formations.

Boundaries. The lower boundary coincides with a basin margin unconformity at the base of the Rævekløft Formation in the Hurry Inlet area (Fig. 5). Arkosic sandstones and mudstones of the underlying Kap Stewart Group are sharply overlain by coarse-grained pebbly

Fig. 5. Major basin margin unconformity (SB1) (arrowed) between the Hettangian delta plain deposits of the Kap Stewart Group and the Pliensbachian (Jamesoni Zone) shoreface deposits of the Rævekløft Formation. Notice the thin coal horizon and rootlet beds just below the unconformity. Harris Fjeld. See Fig. 1 for location. Backpack for scale. From Dam & Surlyk (1995).

sandstones with a rich marine fauna of the Rævekløft Formation (Rosenkrantz, 1934; Surlyk *et al.*, 1973; Dam, 1990b). The formation is absent from the northern, western and central parts of the basin, where the Gule Horn Formation rests directly on the Kap Stewart Group. In these areas black organic-rich paper shales of the uppermost Kap Stewart Group, give way to well-sorted fossiliferous sandstones of the Gule Horn Formation (Fig. 4; Surlyk *et al.*, 1973; Dam & Christiansen, 1990; Dam, 1991; Dam & Surlyk, 1993).

The upper boundary is placed at a sharp unconformity between the muddy siltstones of the Sortehat Formation and the sandstones of the Vardekløft Group (Surlyk *et al.*, 1973; Surlyk, 1990b; Engkilde, 1994; Engkilde & Surlyk, in press).

Distribution. The group crops out along the whole length of Neill Klinter and further north along the west side of Klitdal and Carlsberg Fjord. It forms plateau areas west of the head of Fleming Fjord and is exposed in the valleys east of Schuchert Dal. Further outcrops occur in the central part of Scoresby Land and in the southern part of Liverpool Land (Fig. 1; Surlyk *et al.*, 1973).

Geological age. The Neill Klinter Group has an Early– early Middle Jurassic age. The base of the group (Ræve-kløft Formation) contains a Lower Pliensbachian fauna of the Jamesoni and Davoei Zones, and the upper part of the group (Skævdal and Trefjord Bjerg Members of the Ostreaelv Formation) yields a Toarcian fauna and flora (Rosenkrantz, 1934; Donovan, 1957; Callomon, 1961, personal communication, 1993; Doyle, 1991; Koppelhus & Dam, in press) and the uppermost Sortehat Formation contains an Aalenian – Early Bajocian microflora (Underhill & Partington, 1994; Koppelhus & Hansen, in press).

Subdivisions. The Neill Klinter Group is subdivided from below into the Rævekløft, Gule Horn, Ostreaelv and Sortehat Formations (Fig. 3).

Rævekløft Formation

new formation

History. This formation corresponds to the Rævekløft Member of Surlyk *et al.* (1973).

Name. The formation is named after the ravine Rævekløft close to the south-easternmost tip of Jameson Land (Fig. 1).

Type and reference localities. The thickest development of the formation occurs at Rævekløft, at Kap Stewart and in the southern part of Liverpool Land, between Kap Hope and Scoresbysund (Fig. 1; Surlyk *et al.*, 1973). The section at Rævekløft, designated the type section, was first described by Rosenkrantz (1934, p. 38, pl. 9), and in more detail by Surlyk *et al.* (1973). Well-exposed reference sections occur at Tancrediakløft, Qupaulakajik, at Harris Fjeld, and on the northern slopes of Elis Bjerg (Fig. 1).

Thickness. The formation is 15 m thick at the type locality and Tancrediakløft, and more than 20 m at Kap Hope. It thins to 9 m on Harris Fjeld and on the northern slope of Elis Bjerg, and disappears entirely

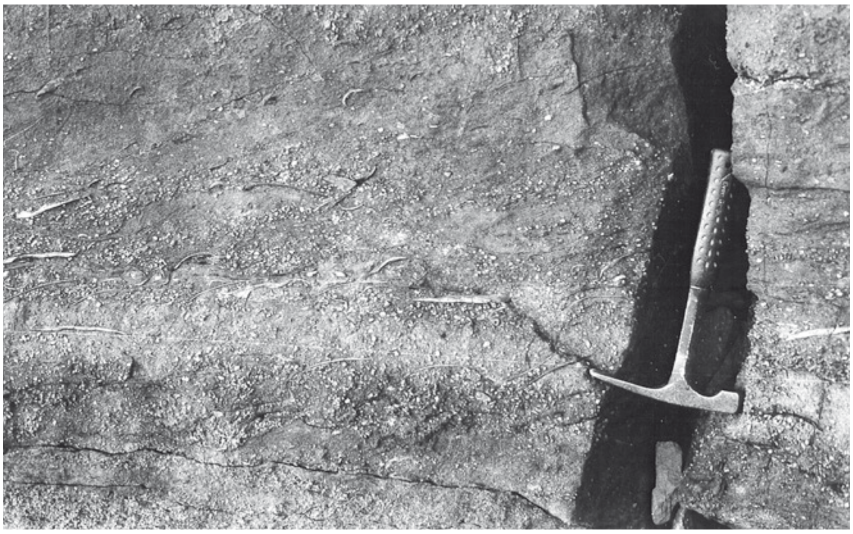
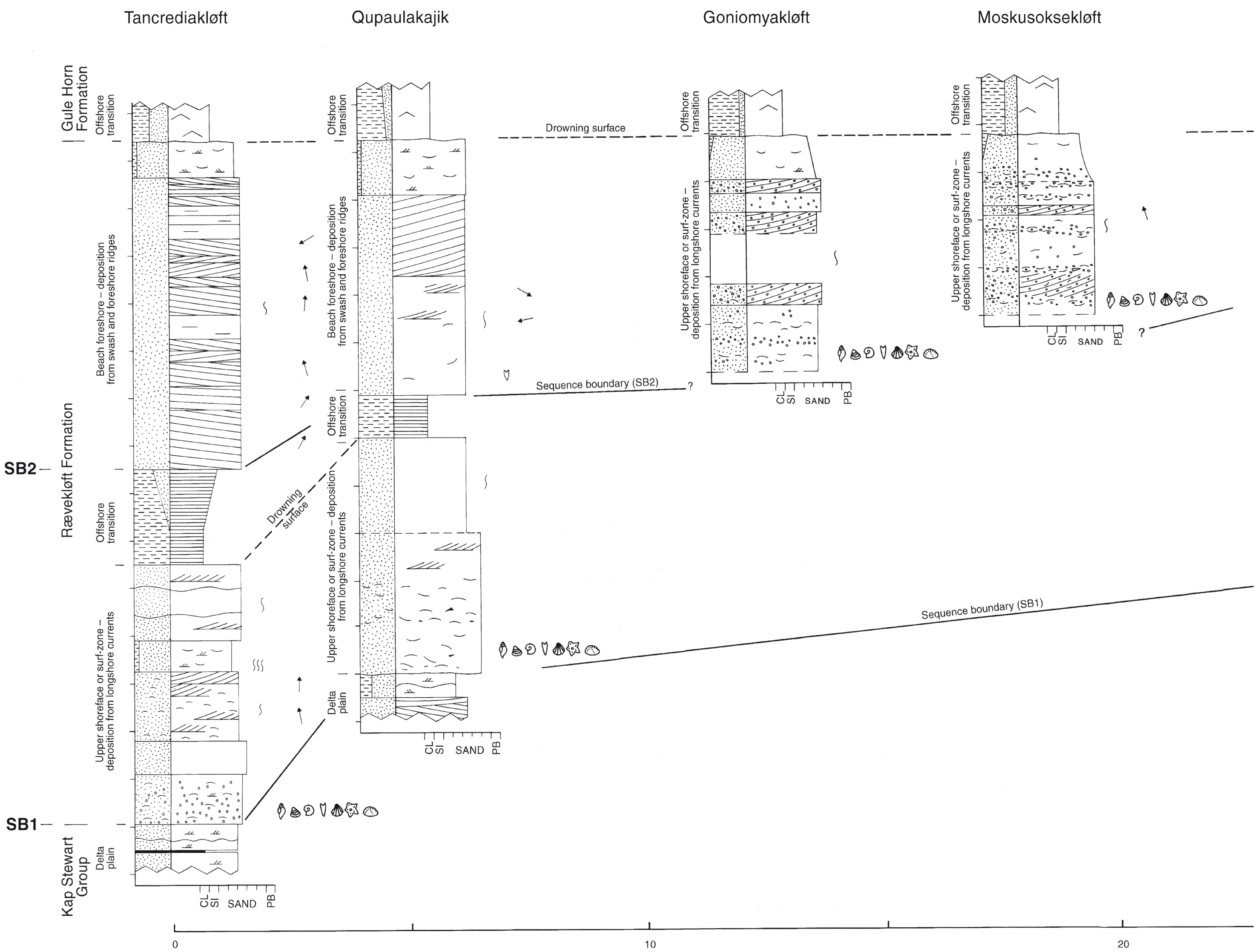


Fig. 6. Poorly sorted medium to very coarse-grained cross-bedded sandstones with scattered granules and shell fragments (Facies association a). From the Rævekløft Formation, Harris Fjeld. See Fig. 1 for location. Hammer 32 cm long.



north of Dusén Bjerg (Rosenkrantz, 1934, 1942; Surlyk *et al.*, 1973).

Lithology. The formation is mainly composed of cliff-forming units of fossiliferous, pebbly, medium to coarse-grained sandstone hardened by concretionary cement, and pebble conglomerates with a sandy and muddy matrix. The sandstones are grey but weather to a reddish-brown colour (Surlyk *et al.*, 1973).

Facies associations and depositional environments. Three facies associations are recognised in the Rævekløft Formation (a–c).

a. Shoreface association. This association constitutes most of the formation. It consists of massive, planar and trough cross-bedded, poorly sorted, medium to very coarse-grained sandstones, and parallel laminated, low-angle cross-bedded and cross-laminated, fine to coarse-grained well-sorted sandstones (Figs 5–7). The massive sandstones occur in beds, 0.25–10 m thick. The cross-bedded sets occur as single sets or more commonly as cosets composed of 2–8 sets, each 0.1–0.7 m thick. Foresets are angular or tangential and dip 10–27°. Pebbles, logs and transported brachiopods, gastropods, cephalopods, bivalves, echinoids and crinoids commonly occur along the bottomsets and scattered along the foresets (Fig. 6). The foresets dip unimodally towards the NNE, parallel to the eastern basin margin.

The medium to very coarse-grained sandstones are characterised by elements of the *Diplocraterion* ichnocoenosis (Dam, 1990b). It includes common, up to 50 cm long, slender *Diplocraterion parallelum*, show-

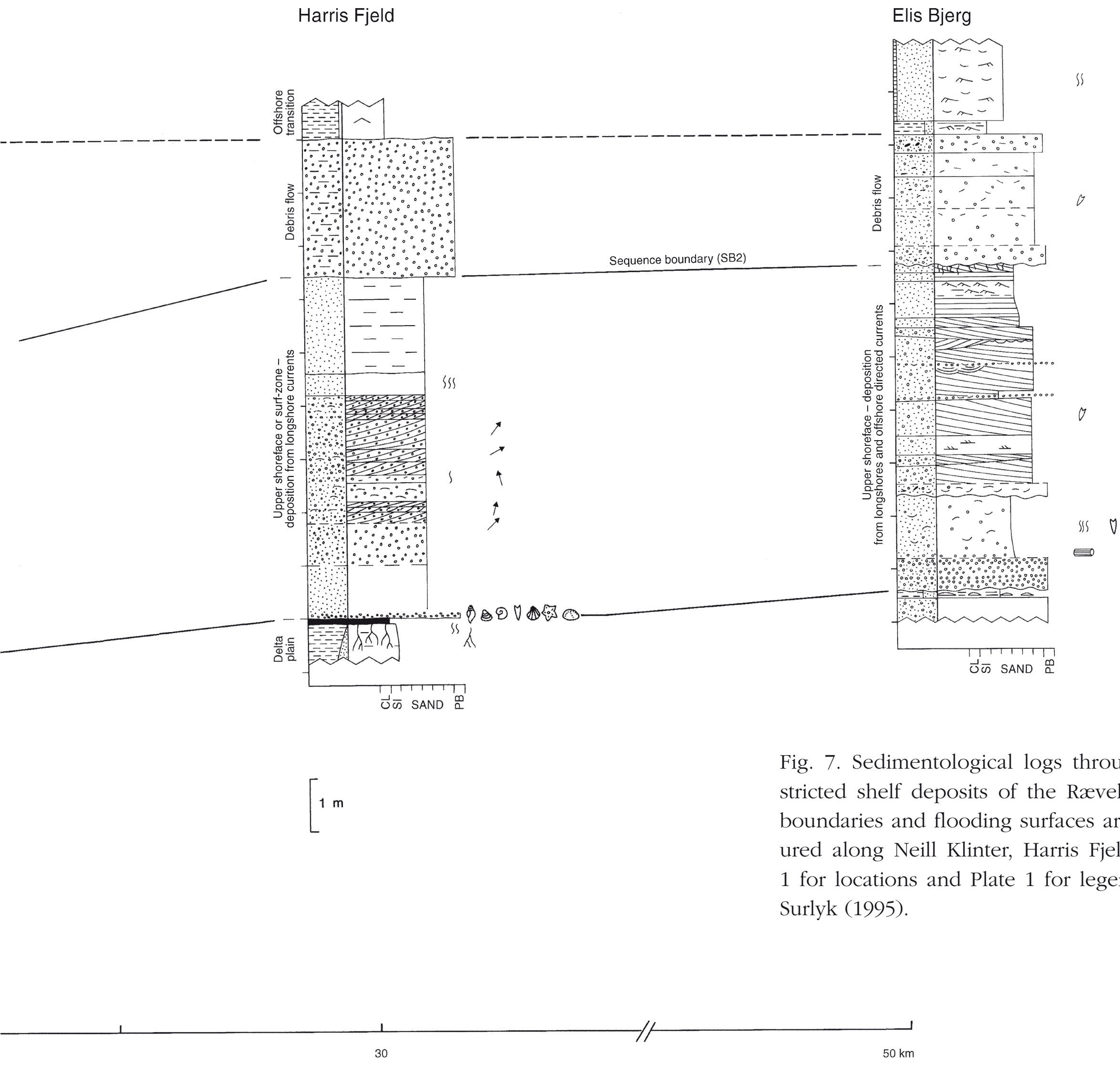


Fig. 7. Sedimentological logs through the shoreface and restricted shelf deposits of the Rævekløft Formation. Sequence boundaries and flooding surfaces are indicated. Profiles measured along Neill Klinter, Harris Fjeld and Elis Bjerg. See Fig. 1 for locations and Plate 1 for legend. Modified from Dam & Surlyk (1995).

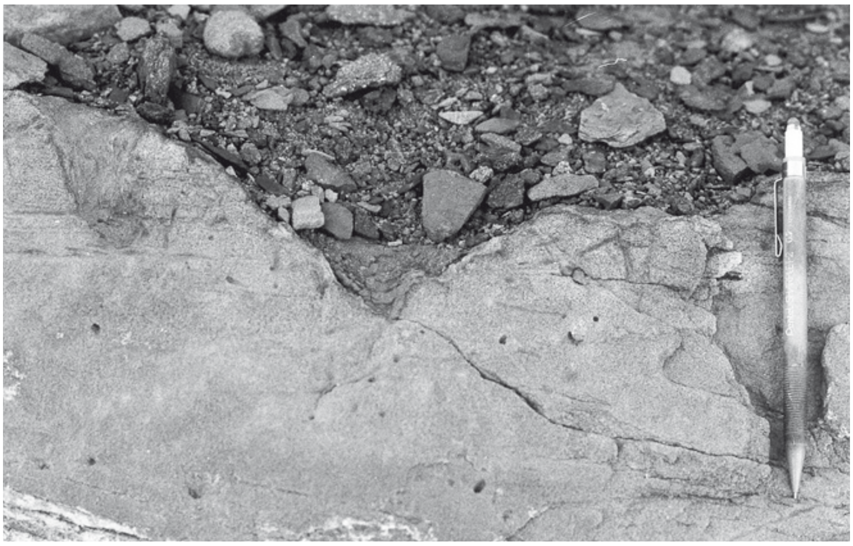


Fig. 8. Bivalve burrows from a level in the Rævekløft Formation that can be traced laterally for several hundreds of metres at the north slope of Elis Bjerg. See Figs 1 and 7 for location. Pen 14 cm long.

ing both protrusive and retrusive spreiten, and rare *Rhizocorallium irregulare* and *Ophiomorpha nodosa*.

The well-sorted, micaceous fine to coarse-grained sandstones occur in sheet-like beds, up to 6 m thick, at Qupaulakajik and on the northern slope of Elis Bjerg (Fig. 1). In the ravine between Dusén Bjerg and Elis Bjerg the sandstones are arranged in sharp-based, fining-upward units, 30–70 cm thick. They show indistinct parallel lamination, parallel and climbing ripple cross-lamination, and low-angle cross-bedding. The low-angle cross-bedded sets are wedge-shaped, 0.15–1.7 m thick, and occur in cosets, up to 6 m thick. The foreset dip angles vary from 5° to 22°, and the palaeocurrent directions are generally westwards towards the centre of the basin. Transported belemnites and bivalves occur occasionally in the sandstones. On the northern slope of Elis Bjerg a level with bivalve burrows can be followed laterally for several hundreds of metres (Fig. 8).

Interpretation. The poorly sorted cross-bedded medium to very coarse-grained sandstones are interpreted as representing fields of small-scale 2-D and 3-D dunes or shoreface ridges on the upper shoreface migrating in a longshore direction towards NNE. The lack of bedding in the massive sandstones is probably due to the poor sorting of the sediment which makes internal structures indistinct. The *Diplocraterion* ichnocoenosis reflects a high-energy, well-aerated intertidal to shallow subtidal environment (Fürsich, 1975). The relatively long burrows of *Diplocraterion parallelum* and the presence of both protrusive and retrusive spreiten, indicate that the burrow acted as a protective shelter against repeated erosion and deposition in the high-energy and unstable upper shoreface.

The well-sorted, fine to coarse-grained sandstones arranged in sharp-based, fining-upward units suggest episodic deposition of sands from waning flows. The internal arrangement and the offshore directed palaeocurrent directions suggest deposition from storm-generated flows reminiscent of low-density turbidity currents in a shoreface environment.

The low-angle cross-bedded fine-grained sandstones present at Qupaulakajik probably represent migrating foreshore ridges. The indistinct parallel laminated sandstones were probably formed in the swash zone.

b. Debris flow association. This association consists of a single, sharply based, conglomeratic bed occurring at the top of the formation at Harris Fjeld and on the northern slope of Elis Bjerg (Figs 1, 7). It is 2–3 m thick, and has an along-strike lateral extent exceeding 20 km. It consists of well-rounded quartzite pebbles up to 4 cm in maximum diameter occurring scattered in a muddy sandstone matrix. Transported bivalves and belemnites occur occasionally at Elis Bjerg, but are absent at Harris Fjeld, where the bed is massive. In the ravine between Elis Bjerg and Dusén Bjerg the coarser clasts occur in thin, indistinct laminae.

Interpretation. The mixed lithology and the general absence of an organised fabric or grading suggest deposition from a high viscosity debris flow, in which clasts were transported as a result of matrix strength (e.g. Lowe, 1982). However, the occasional occurrence of coarser clasts at distinct levels at Elis Bjerg, indicates local shearing or pulsating surging during deposition. The general lack of marine indicators at Harris Fjeld suggests that the source area for the debris was terrestrial. The debris flow deposits probably represent a catastrophic flooding episode on a flood plain and shoreface, possibly triggered by minor tectonic activity in the sediment source area.

c. Offshore transition association. This association consists of a single coarsening-upward bed, 1–2 m thick, of alternating parallel laminated mudstones and fine-grained sandstone streaks showing incipient wave ripples. It contains marine palynomorphs (Koppelhus & Dam, in press). The bed is only present in the south-easternmost part of the basin at Qupaulakajik and Tancrediakløft, just beneath the foreshore sandstones of the uppermost part of the formation (Fig. 7).

Interpretation. The presence of sandstone streaks showing incipient wave-generated structures and marine palynomorphs, suggests that the sediments were deposited from suspension in the offshore transition zone with restricted wave activity.

Fossils. Fossils are restricted to certain levels separated by largely barren intervals. Rosenkrantz (1934) identified a lower division (the lower sandstone bed in Tancrediakløft and Qupaulakajik; Fig. 7), with a diverse dominantly European fauna of 150 species, dominated by bivalves and including gastropods, cephalopods, echinoids and crinoids, and an upper division (the upper sandstone bed in Tancrediakløft and Qupau-lakajik; Fig. 7) containing approximately 20 molluscan species. Trace fossils include common *Diplocraterion parallelum* and rare *Rhizocorallium irregulare*, *Ophiomorpha nodosa* and bivalve burrows (Dam, 1990a, b).

Boundaries. The formation is separated from the underlying delta plain deposits of the Kap Stewart Group by a basin margin unconformity and is commonly initiated by a thin transgressive lag conglomerate of fragmented body fossils and quartzite pebbles (Surlyk, 1990a, b; Dam, 1991). The upper boundary occurs at the sharp lithological change from the sandstones and conglomerates of the Rævekløft Formation to the shaly unfossiliferous base of the Elis Bjerg Member (Surlyk *et al.*, 1973).

Distribution. The formation is exposed along the length of Neill Klinter and as far north as Dusén Bjerg and in the southern part of Liverpool Land. Surlyk *et al.* (1973) included lenses of arkosic sandstones containing similar fossils, occurring to the west in Schuchert Dal, in the Rævekløft Formation. The present study shows that these beds extend throughout the northern and western parts of the basin, but contain a very different set of facies associations and are better included in the Gule Horn Formation.

Geological age. Rosenkrantz (1934) considered the lower division of the Rævekløft Formation as belonging to the Lower Pliensbachian Jamesoni Zone. He dated the upper division to the Ibex Zone, but Callomon (1961, p. 261) identified the ammonite *Androgynoceras* representing the *maculatum* group, indicating the lowermost Davoei Zone. All the belemnites recovered from the formation by Rosenkrantz were apparently collected from the lower Jamesoni Zone division. They indicate that the Jamesoni Zone division of Rosenkrantz (1934) includes the Lower Pliensbachian Jamesoni to at least the Ibex Zone and possibly the lower Davoei Zone (Doyle, 1991). The combined fossil evidence thus indicates a Jamesoni to early Davoei Zone age for the formation.

The Rævekløft Formation is probably time equivalent to parts of the Gule Horn Formation throughout the central, northern and western parts of the basin.

Gule Horn Formation

new formation

History. This formation corresponds to the Gule Horn Member of Surlyk *et al.* (1973).

Name. The formation is named after the mountain Gule Horn west of Carlsberg Fjord, Jameson Land (Fig. 1).

Type and reference localities. Surlyk *et al.* (1973) chose the exposures at the mountain Gule Horn as the type locality because of the thick development of the unit. Well-exposed reference sections occur along the length of Neill Klinter, at Harris Fjeld, Nathorst Fjeld, Elis Bjerg, Lepidopteriselv, Liaselv and Ranunkeldal, and Horsedal (Fig. 1, Plate 2).

Thickness. The formation is 75–100 m thick along Neill Klinter, 100–105 m at Harris Fjeld, Lepidopteriselv and Liaselv, 110 m at Horsedal, and 185 m at Rhætelv.

Lithology. The lithology of this formation is very characteristic and consists of heterolithic thin-bedded micaceous sandstones and mudstones and cross-bedded sandstones with thin mudstone drapes on the foresets. Thin-bedded sandstones show wave ripple and lingoid current ripple marks on parting planes. Thin mud chip

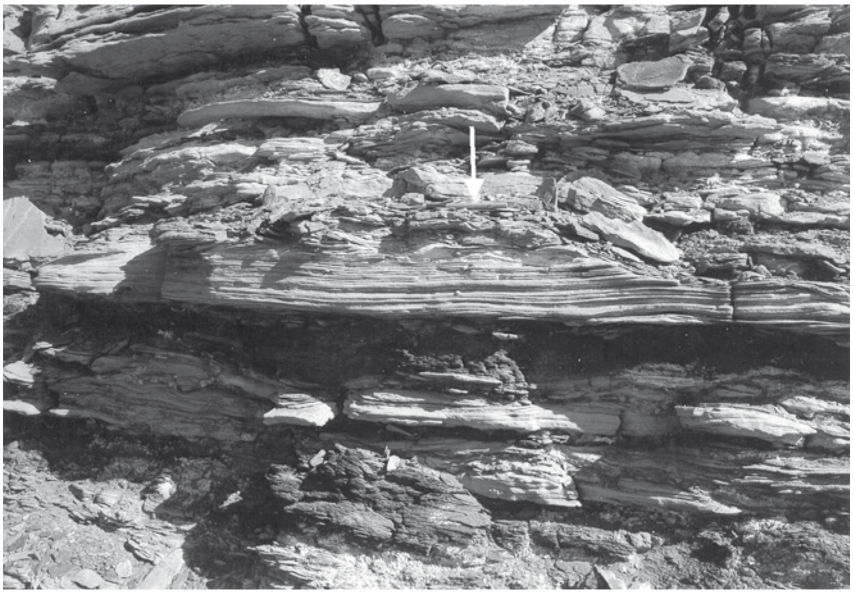


Fig. 9. Laterally persistent storm sandstones (tempestites) interbedded with homogeneous mudstones of the storm-dominated lower shoreface association (Facies association d). Lower part of the Elis Bjerg Member, Ranunkeldal. See Figs 1, 10 and 11 for location. Pen 14 cm long (arrow).

conglomerates occur at several levels. Detailed descriptions of facies, facies associations and interpretation of depositional environments of the formation are presented under individual members.

Fossils. The formation is generally unfossiliferous, but bivalves do occur in the lowermost part of the formation in the Rhætelv and Ranunkeldal sections, and a few fragmented bivalves and belemnites have been found in thin-bedded conglomerates and massive pebbly muddy sandstone beds along Neill Klinter. The formation contains a diverse assemblage of trace fossils, including *Ancorichnus ancorichnus, Arenicolites* isp., *Asteriacites lumbricalis, Bergaueria* isp., *Cochlichnus anguineus, Cruziana* isp., *Curvolithos multiplex, Diplocraterion parallelum, Gyrochorte comosa, Gyrophyllites kwassicensis, Helminthopsis magna, Jamesonichnites heinbergi, Lockeia amygdaloides, Monocraterion tentaculatum, Ophiomorpha nodosa, Palaeophycus alternatus, Phoebichnus trochoides, Phycodes auduni, Phycodes bromleyi, Phycosiphon* isp., *Planolites beverleyensis, Rhizocorallium irregulare, Taenidium serpentinum, Teichichnus* isp., *Thalassinoides* isp. and unnamed trackways.

Boundaries. Along Neill Klinter the lower boundary occurs at a sharp lithological change from the fossiliferous sandstones and conglomerates of the Rævekløft Formation to the unfossiliferous mudstones at the base of the Gule Horn Formation (Fig. 7; Plate 2; Surlyk *eet al.*, 1973). In the northern, western and central parts of the basin the black organic-rich paper shales of the Kap Stewart Group, give way to well-sorted fossiliferous sandstones of the Gule Horn Formation (Fig. 4; Surlyk *et al.,* 1973; Dam & Christiansen, 1990; Dam, 1991; Dam & Surlyk, 1993).

Along Neill Klinter the upper boundary is sharp and placed between the interbedded sandstone and mudstone of the Albuen Member and the overlying cross-bedded sandstones of the Astartekløft Member. In the northern and central parts of the basin the upper boundary is placed where cross-laminated and cross-bedded sandstones of the Elis Bjerg Member give way to wave ripple cross-laminated, parallel laminated and hummocky cross-stratified sandstones, arranged in small coarsening-upward successions, of the Horsedal Member.

Distribution. Same as for the group.

Geological age. Body fossils are very rare in the Elis Bjerg and Albuen Members, and none are age diagnostic. Belemnites and ammonites from the upper part of the group (Nathorst Fjeld, Lepidopteriselv and Skævdal Members) suggest that the oldest strata of these members belong to the Commune Subzone (the oldest subzone of the Lower Toarcian Bifrons Zone) or even to the lowermost Toarcian Tenuicostatum Zone (Semicelatum Subzone) (Doyle, 1991; J. H. Callomon, personal communication, 1993), suggesting that the Gule Horn Formation has a Late Pliensbachian age in the south-eastern part of the basin, where the member overlies the Rævekløft Formation. Dinoflagellate cysts suggest that the Gule Horn Formation has a Late Pliens-

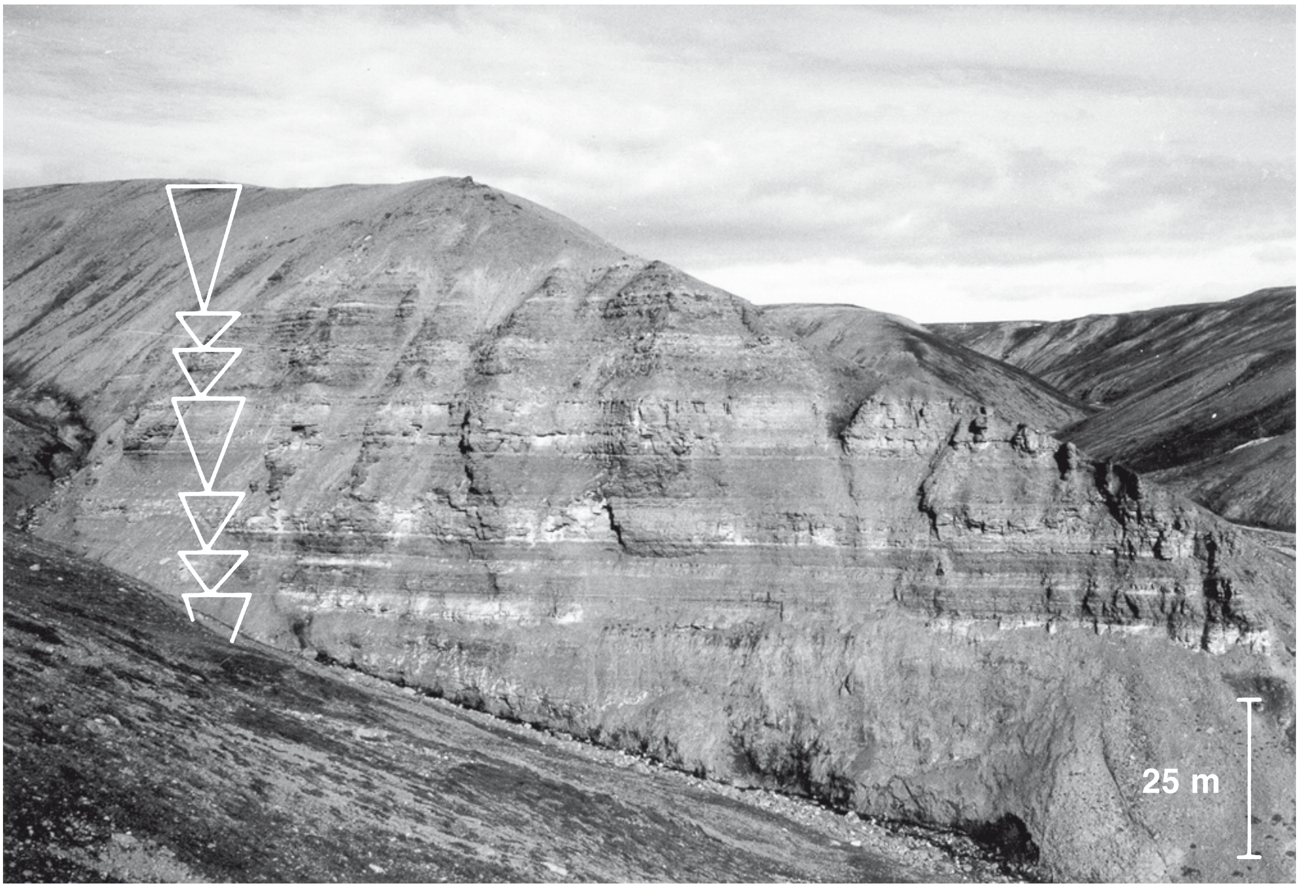


Fig. 10. Stacked coarsening-upward successions formed by progradation of wave and storm-dominated shorefaces (Facies associations d and e) into a storm-dominated offshore shelf. Lower part of the Elis Bjerg Member, Ranunkeldal. See Fig. 1 for location. From Dam & Surlyk (1995).

bachian (probably Margaritatus Zone age) to Early Toarcian age along Neill Klinter (Koppelhus & Dam, in press). Where the Gule Horn Formation shows a gradual transition to the underlying Kap Stewart Group (Fig. 4), the latest part of the Kap Stewart Group suggests an uppermost Sinemurian to earliest Pliensbachian age and the Gule Horn Formation an Early to Late Pliensbachian age (Koppelhus & Dam, in press).

Subdivisions. The Gule Horn Formation is subdivided into the Elis Bjerg Member (base) and Albuen Member (top) (Fig. 3).

Elis Bjerg Member

new member

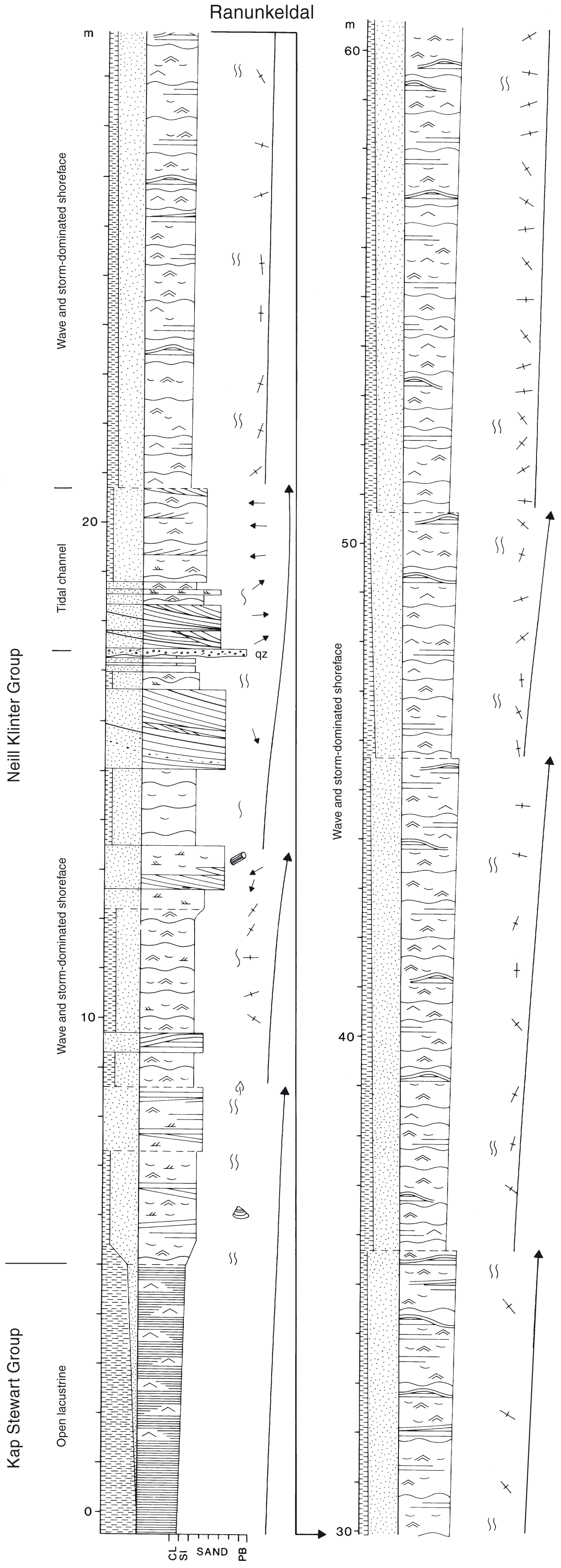
History. The strata composing this member were included in the lower part of the Gule Horn Member of Surlyk *et al.* (1973).

Name. The member is named after the mountain Elis Bjerg, north-west of Hurry Inlet, eastern Jameson Land (Fig. 1).

Type and reference localities. One of the most complete, well-exposed and easily accessible sections of the member occurs at the northern slope of Elis Bjerg which is designated the type section. Well-exposed reference sections occur along the length of Neill Klinter, at Harris Fjeld, Nathorst Fjeld, Lepidopteriselv, Ranunkeldal and Horsedal (Fig. 1, Plate 2).

Thickness. The member is 90–95 m at Elis Bjerg, Harris Fjeld, Lepidopteriselv and Liaselv, 75 m along Neill Klinter, 110 m at Horsedal and 185 m at Rhætelv.

Lithology. The lithology of the member is very characteristic and consists of heterolithic thin-bedded micaceous sandstones and mudstones and cross-bedded sandstones with thin mudstone drapes on the foresets.



Thin-bedded sandstones show wave ripple and linguoid current ripple marks on parting planes. Thin mud chip conglomerates occur at several levels.

Facies associations and depositional environments. Six facies associations are recognised in the Elis Bjerg Member (c–h) (Plate 2).

c. Offshore transition association. This association consists of isolated beds of alternating parallel laminated mudstones and fine-grained sandstone streaks, 1–3 m thick, and a single laterally extensive bed, up to 6 m thick, at the base of the Elis Bjerg Member. This bed has been followed laterally for 78 km in the south-eastern part of the basin (Plate 2). All the isolated beds occur at the base of small coarsening-upward successions. The sandstone streaks show incipient wave ripples. The mudstones contain marine palynomorphs (Koppelhus & Dam, in press).

Interpretation. The presence of sandstone streaks showing incipient wave-generated structures and marine palynomorphs, suggests that the sediments were deposited from suspension in the offshore transition zone with restricted wave activity.

d. Storm-dominated lower shoreface association. This association constitutes most of the member in the north-western part of the basin along Schuchert Dal and at Rhætelv. It consists of alternating mudstones and well-sorted fine-grained sandstones. The sandstones range from millimetre-thick streaks showing incipient wave ripple lamination (Facies M_1 of De Raaf *et al.*, 1977), to thin and thick-bedded (5–150 cm) laterally persistent parallel-sided beds (Fig. 9). The sandstones are sharp-based and may be overlain by thin lag conglomerates. Internally the bedded sandstones show parallel lamination, hummocky and swaley cross-stratification. Laterally the hummocks show pinch and swell with an amplitude of up to 65 cm and hummock heights are less than 10 cm. Wave ripple cross-lamination and mud-draped wave ripple formsets are common at the top of the beds. In the Ranunkeldal section the laminae and beds are arranged in thickening and coarsening-upward successions, up to 15 m thick (Figs 10, 11). The storm-dominated lower shoreface association contains

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Fig. 11. Sedimentological log through stacked coarsening-upward successions formed by progradation of wave and storm-dominated shorefaces into a storm-dominated offshore shelf. Lower part of the Elis Bjerg Member, Ranunkeldal. See Fig. 1 for location and Plate 1 for legend.

Fig. 12. Cross-bedded medium to coarse-grained sandstones of Facies association e that form a sheet in the upper 20–25 m of the Ranunkeldal section and constitute a yellow-orange weathering plateau on top of the Oswald Heer Klinter along the north-western basin margin. See Fig. 1 for location. Hammer 32 cm long.

the *Planolites* ichnocoenosis with common *Planolites beverleyensis, Taenidium serpentinum, Gyrochorte comosa, Helminthopsis magna*, rare *Phycosiphon* isp., *Phoebichnus trochoides* and *Gyrophyllites kwassicensis* (Dam, 1990b).

Interpretation. The mudstones were deposited during fair-weather periods. The laminated and bedded sandstones were deposited as tempestites during periodic storms at water depths below fair-weather wave base but above storm-wave base. The thickening and coarsening-upward cycles in Ranunkeldal showing upward thickening of storm-generated layers are shallowing-upward cycles with distal overlain by proximal tempestites (cf. Hamblin & Walker, 1979; Aigner & Reineck, 1982; Dott & Bourgeois, 1982; Aigner, 1985; Duke, 1985). The sands were probably derived from coastal erosion and transported by downwelling flows during storms (cf. Snedden *et al.*, 1988).

The trace fossils were mainly produced by infaunal organisms combining the activity of deposit-feeding and locomotion (endostratal pascichnia burrows) and deposit-feeding organisms that systematically mined the sediment for food at one particular place (fodinchnia burrows) (Dam, 1990a, b). The low diversity suggests a relatively low oxygen content within the sediment (cf. Ekdale & Mason, 1988). However, the dominance of pascichnia over stationary fodinichnia suggests that the interstitial environment, supporting production of pascichnia, must have been characterised by at least some oxygen to allow respiration. A dysaerobic bottom environment is accordingly suggested for this association (Dam, 1990b).

e. Wave and storm-dominated upper shoreface association. This association consists of fine-grained or medium to coarse-grained sandstones, overlying the storm-dominated lower shoreface deposits of Facies association d in the north-western and central parts of the basin.

The fine-grained sandstones form sheet-like units, up to 11 m thick, locally overlain by tidal channel deposits. The sandstones are low-angle cross-bedded, parallel laminated, hummocky cross-stratified or massive (Figs 4, 11). The massive beds show a high degree of bioturbation (up to 100%) and are characterised by the *Taenidium* ichnocoenosis, dominated by *Taenidium serpentinum* with occasional *Gyrochorte comosa* and *Curvolithos multiplex* (Dam, 1990b).

The medium to coarse-grained sandstones are cross-bedded, and occur in the upper 20–25 m of the Ranunkeldal section and form a yellow-orange plateau on top of Oswald Heer Klinter along the north-western part of the basin (Fig. 1). Cross-lamination and bedding show undulatory lower boundaries, opposed unidirectional cross-bedded lenses, offshooting and draping laminations, symmetrical ripple form sets, features characteristic of wave action (Fig. 12; cf. De Raaf *et al.*, 1977). Foresets show asymmetrical bipolar dip directions toward NE and SW. Sets are 0.1–0.25 m thick. The cross-bedded sandstones are closely associated with wave ripple cross-laminated sandstones showing symmetrical ripples on parting planes. Ripple crestline orientations show a rather large scatter but are dominated by ESE–WNW directions. The sandstones are commonly burrowed by *Gyrochorte comosa* and *Planolites beverleyensis*.

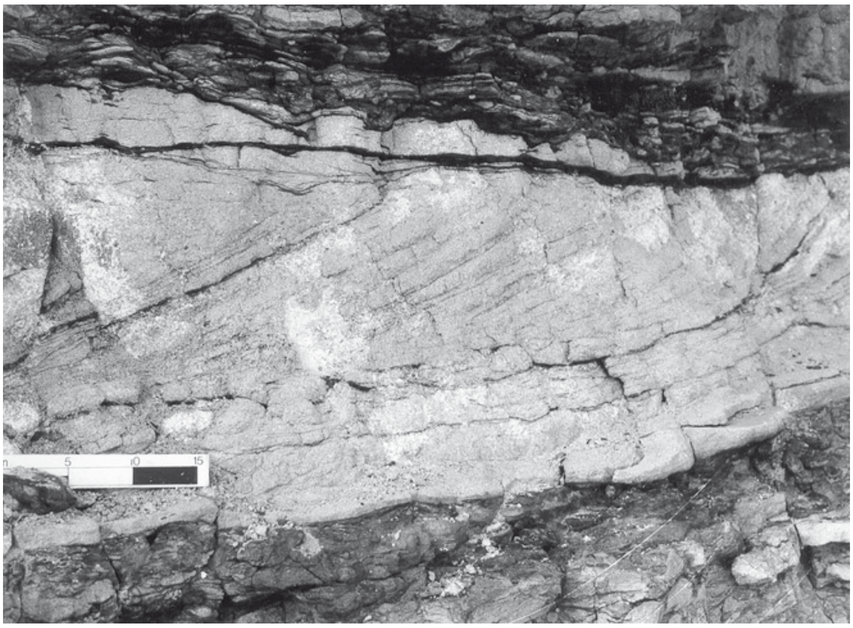

Fig. 13. Cross-bedded sandstone with mudstone drapes and heterolithic bedding of the subtidal sand sheet association (f). The heterolithic bed is intensively burrowed by *Diplocraterion parallelum*. Elis Bjerg Member, Astartekløft. See Fig. 1 for location.

Interpretation. Most internal structures of the shoreface association show features diagnostic of wave and storm action (cf. De Raaf *et al.*, 1977). The low-angle cross-bedded and parallel laminated fine-grained sandstones represent migration of swash and foreshore ridges on the beach foreshore. Hummocky cross-stratification formed during storms by aggradation or translation in a combined or oscillatory flow regime (Dott & Bourgeois, 1982; Swift *et al.*, 1983; Allen, 1985; Surlyk & Noe-Nygaard, 1986).

The *Taenidium* ichnocoenosis is dominated by burrows of deposit-feeders occurring in isolated beds indicating periods of low energy. Pascichnia-dominated ichnocoenoses like this occur where bottom and interstitial waters are dysaerobic (cf. Ekdale & Mason, 1988).

The cross-bedded, medium to coarse-grained sandstones represent fields of symmetrical dunes on the shoreface, similar to those recorded from modern shoreface and offshore transition settings (cf. Newton & Werner, 1972; Hunter *et al.*, 1982, 1988; Cacchione *et al.*, 1984; Leckie, 1988). The general asymmetrical bipolar dip directions toward NE and SW, roughly paral-

Fig. 14. Cross-bedded cosets of the subtidal sand sheet association (f). Lower part of the Elis Bjerg Member, Harris Fjeld. See Fig. 1 for location. Person for scale. From Dam & Surlyk (1995).

Fig. 15. Heterolithic deposits of the subtidal sand sheet association (f), arranged in a coarsening-upward succession. Elis Bjerg Member, Astartekløft. See Fig. 1 for location. Hammer 28 cm long.

lel to the western basin margin, suggest that the dunes formed from oscillatory or oscillatory-dominant coast-parallel flows.

f. Subtidal sand sheet association. This association is characteristic of the member in the south-eastern, eastern, central and northern parts of the basin. It consists of sandstone beds and alternating laminated and thinly bedded sandstones and mudstones. Thinly bedded siderite pebble conglomerates occur at several levels.

The sandstones occur in planar and trough cross-bedded medium to very coarse-grained beds, 0.1–5.5 m thick, with scattered plant fragments and mudstone intraclasts (Figs 13, 14). Thin mudstone drapes, single or couplets, commonly occur on the foresets (cf. Visser, 1980). They flatten out along the toesets and extend as horizontal layers for several metres, merging with the lower set boundaries. Mudstone drapes may extend to the top of the foresets or they may be truncated on the mid-foreset slope by reactivation surfaces, commonly veneered by mudstone intraclasts (cf. Boersma, 1969; Visser, 1980). Groups of foresets may display repetitive lateral thickening and thinning like the bundled up-building of Visser (1980). The cross-bedded sandstones may be capped by ripple cross-laminated sandstones. Combined palaeocurrent data show bimodal to bipolar directions toward N–NE and S–SW indicating reversing currents. Heterolithic beds separating cross-bedded sets and individual foresets may be densely burrowed by *Diplocraterion parallelum* (Fig. 13).

The heterolithic deposits are arranged in both coarsening and thickening-upward (Fig. 15), and fining and thinning-upward successions. Bedding planes display lunate ripple trains and less commonly wave ripple formsets. Sediment transport directions in succeeding sand laminae are commonly opposite (Fig. 16). Combined palaeocurrent data of this facies association also show bimodal to bipolar directions towards N–NE and S–SW, suggesting reversing currents.

The heterolithic beds are characterised by the *Cochlichnus* ichnocoenosis containing the most diverse trace fossil assemblage of the Neill Klinter Group, including abundant *Arenicolites* isp. 3 and *Cochlichnus anguineus*, common *Ancorichnus ancorichnus, Asteriacites lumbricalis, Bergaueria* isp., *Jamesonichnites heinbergi,*

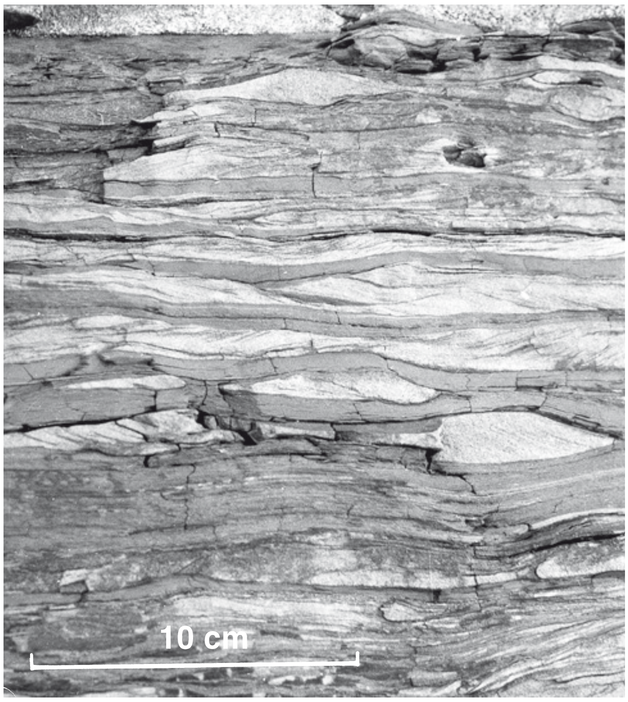

Fig. 16. Wavy bedding of the subtidal sand sheet association (f). Notice that sediment transport directions in succeeding sandstone laminae are commonly opposite, suggesting that the thinly interbedded sandstone and mudstone can be regarded as tidal bundles. Elis Bjerg Member, Astartekløft. See Fig. 1 for location.

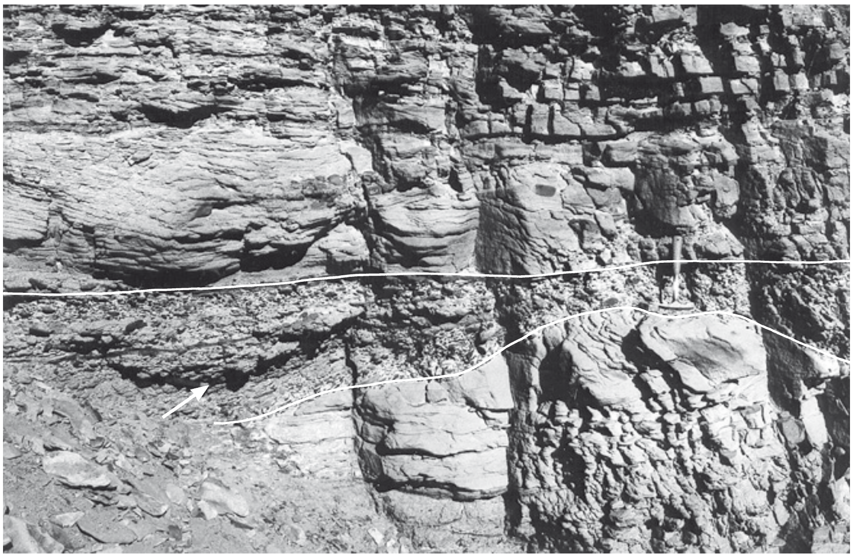

Fig. 17. Intraformational siderite clast conglomerate of the subtidal sand sheet association (f) (arrow). Elis Bjerg Member, Nathorst Fjeld. See Fig. 1 for location. Hammer 28 cm long.

Diplocraterion parallelum, Gyrochorte comosa, Palaeophycus alternatus, Phoebichnus trochoides, Phycodes auduni, Phycodes bromleyi, Planolites beverleyensis, Taenidium serpentinum, Thalassinoides isp. and rare *Arenicolites* isp. 1, *Cruziana* isp., *Curvolithos multiplex, Helminthopsis magna, Lockeia amygdaloides, Ophiomorpha nodosa, Phycosiphon* isp., *Rhizocorallium irregulare, Teichichnus* isp. and type 2 trackways (Dam, 1990a, b).

Continuous and discontinuous thinly bedded intraformational mudstone clast conglomerates and sandstones showing large-scale scour and fill structures are interbedded with the sandstones and heterolithic deposits (Fig. 17). The large-scale scours occur as local channel-like depressions, up to 3 m deep, along laterally continuous erosion surfaces. Intraformational mudstone conglomerates are present in the deeper parts of the scours. The lateral extent of the discontinuous conglomerates is generally less than 10 m, and only rarely can they be followed for more than a few tens of metres.

Interpretation. The presence of mudstone couplets, reactivation surfaces veneered by mudstone intraclasts and foreset bundles indicate that the cross-bedded sandstones were deposited from migrating small and medium-scale subtidal dunes (Types I–IV of Allen, 1980) in a strong subtidal current regime (e.g. Boersma, 1969; Visser, 1980). The abundance of *Diplocraterion parallelum* suggests a high-energy, shallow subtidal to intertidal environment (Fürsich, 1975). The opposite transport directions in succeeding sandstone laminae suggest that they are tidal bundles showing reversal of currents of equal strength separated by slack-water stages (e.g. Reineck *et al.*, 1968; Tessier & Gigot, 1989; Dreyer, 1992). The bundled upbuilding of the heterolithic beds and the lack of indications of subaerial exposure indicate subtidal conditions. The trace fossil assemblage of the heterolithic beds reflects a wide variety of feeding habits and behavioural categories, indicating a well-aerated medium to low-energy shallow marine environment, with abundant food supply for both suspension and deposit-feeders, and mobile carnivores (Dam, 1990b). Although cyclical thickness variations of bundles occur in both dunes and heterolithic deposits, attempts to determine neap-spring-neap cyclical bundle successions were unsuccessful. This is due to large random variations in bundle thickness of the tidal sandwaves and interbedding with wave and storm-reworked sandstones in the heterolithic deposits. Random variations in tidal bundle thickness is expected in shelf settings which are affected by storms (cf. Yang & Nio, 1985).

Most likely the large-scale erosion surfaces of this association were formed by winnowing or erosion by offshore directed storm surges, which only left a thin lag conglomerate in the incised depressions. The discontinuous, thinly bedded conglomerates probably represent tidal scouring or deflation of more local origin.

g. Tidal channel association. This association is characteristic of the member in the south-eastern and northern parts of the basin. It consists of fining and thinning-upward successions, up to 9 m thick (Fig. 18). The successions are bounded below by a laterally continuous erosion surface, commonly with small erosional pockets up to 1 m wide and 25 cm deep. The surface is commonly draped by a laterally continuous lag conglomerate composed of plant remains and sideritic mudstone intraclasts. The conglomerate is overlain by coarse-grained sandstones with distinct trough and planar cross-bedded cosets, with set heights decreasing upwards from 2 m to less than 0.5 m. Thin, single or double mudstone drapes are common on the foresets. Foreset dip directions show a bimodal to bipolar distribution with a dominance towards WNW, indicating flow towards the basin centre. Generally, the upper parts of the successions show an upward increase in number of intercalated mudstone layers, and the cross-bedded sandstones grade into flaser, wavy and lenticular-bedded heteroliths. The fining-upward units form tabular, laterally extensive bodies that can be followed for more than 11 km along strike. The heterolithic deposits are commonly extensively burrowed by *Diplocraterion parallelum*, rare *Arenicolites* isp. and *Monocraterion tentaculatum*.

Other fining-upward successions, up to 16 m thick, show unimodal, basinwards-directed palaeocurrent directions towards WNW and absence of mudstone couplets, tidal bundles and bioturbation. However, cross-bedded sets may still be separated by thin mudstone layers.

Interpretation. The distinct fining-upward units were formed by lateral migration of ebb-dominated subtidal channels. The base of the units was formed by erosion along the active channel thalweg. The small erosional pockets are interpreted as small scour troughs along the thalweg. The succeeding cross-bedded cosets were formed by 2-D and 3-D small-scale dunes migrating on the channel floor. The upper parts of the successions were deposited along the inner bend of the channel. The presence of tidal structures and the absence of any indications of subaerial exposures suggest shallow subtidal deposition. The trace fossil assemblage indicates a high-energy, shallow subtidal to intertidal environment (cf. Fürsich, 1975). Similar fining-upward successions formed by subtidal channel fills have been described by Rizzini (1975), Goldring *et al.* (1978) and Yang & Nio (1989).

The fining-upward units without clear tidal characters are interpreted as deposited in proximal supra- or intertidal channels. Tidal influence is suggested by the presence of mudstone layers separating individual sets. They were probably deposited during slack-water stages.

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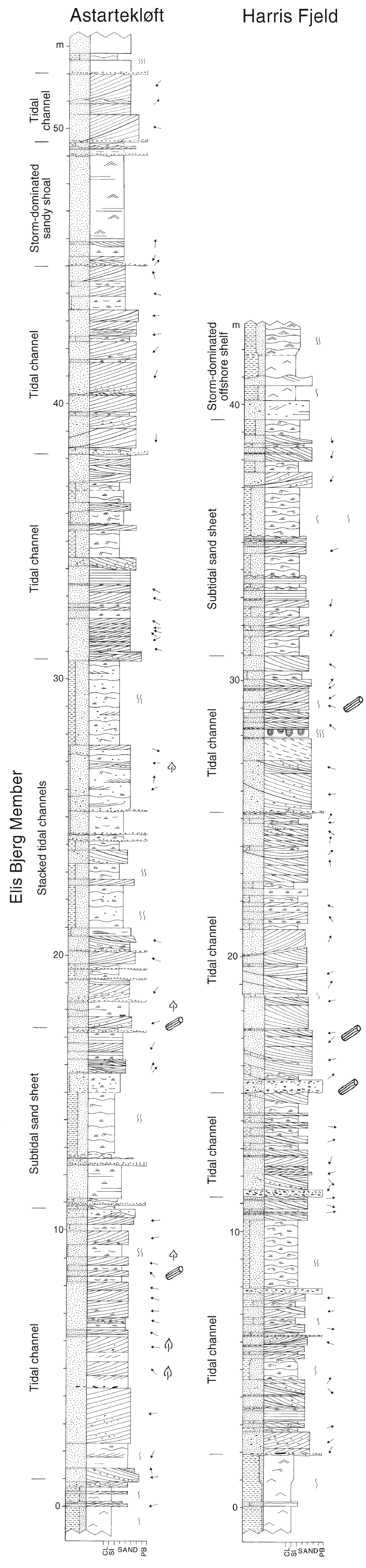


Fig. 18. Sedimentological logs through stacked tidal channel, subtidal sand sheet, and storm-dominated sandy shoal deposits. Elis Bjerg Member, Astartekløft and Harris Fjeld. See Fig. 1 for location and Plate 1 for legend.

Fig. 19. Amalgamated storm sandstones of the storm-dominated sandy shoal association (h), Elis Bjerg Member, Harris Fjeld. See Fig. 1 for location. Each bed marks a major storm event.

h. Storm-dominated sandy shoal association. This association is characteristic of the Elis Bjerg Member in the south-eastern, eastern and northern parts of the basin. It consists of single or amalgamated sandstone beds closely associated with subtidal sand sheet and tidal channel deposits (Figs 18, 19). The association occurs in laterally continuous successions that have been followed along strike of the basin for more than 25 km. Each bed is 0.15–1.55 m thick and forms amalgamated bed-sets, up to 5.5 m thick. The bases of the sandstone beds are sharp and may be overlain by thin lag conglomerates. Internally the bedded sandstones are parallel laminated. Fine-grained beds may show hummocky cross-stratification. Parting planes show parting lineation. Current ripple cross-lamination and lunate ripple form sets showing unidirectional foreset dip directions are common at the top of the beds. Wave ripple form sets are less common.

Interpretation. The sharp bases of sandstone beds and the dominance of parallel lamination suggest episodic deposition from waning flows characterised by combined flow-processes or intense bed shear due to storm-wave activity. Hummocky cross-stratification is formed by aggradation or translation in a combined or oscillatory flow regime (e.g. Dott & Bourgeois, 1982; Swift *et al.*, 1983; Allen, 1985; Surlyk & Noe-Nygaard, 1986). The cross-laminated sandstones were formed during decelerating flow, and the unidirectional palaeocurrent directions indicate deposition by currents with a unidirectional component. The interbedding with tidal channel and subtidal sand sheet deposits suggests that the succession represents storm-dominated sandy shoals.

Fossils. The member is generally unfossiliferous, but bivalves do occur in the lowermost part of the member in the Rhætelv and Ranunkeldal sections, and a few fragmented bivalves and belemnites have been found in thin-bedded conglomerates along Neill Klinter. The member contains a diverse assemblage of trace fossils of the *Cochlichnus* ichnocoenosis (Dam, 1990a, b), including abundant *Arenicolites* isp., *Cochlichnus anguineus, Diplocraterion parallelum, Taenidium serpentinum*, common *Ancorichnus ancorichnus, Asteriacites lumbricalis, Bergaueria* isp., *Curvolithos multiplex, Gyrochorte comosa, Jamesonichnites heinbergi, Palaeophycus alternatus, Phoebichnus trochoides, Phycodes auduni, Phycodes bromleyi, Planolites beverleyensis, Thalassinoides* isp. and rare *Cruziana* isp., *Gyrophyllites kwassicensis, Helminthopsis magna, Lockeia amygdaloides, Monocraterion tentaculatum, Ophiomorpha nodosa, Phycosiphon* isp., *Rhizocorallium irregulare, Teichichnus* isp. and unnamed trackways.

Boundaries. The fine-grained basal part of the Elis Bjerg Member rests with a sharp boundary on the sandstones and conglomerates forming the top of the Rævekløft Formation (Fig. 7; Plate 2). In the northern, western and central parts of the basin, the Elis Bjerg Member rests directly on the Kap Stewart Group (Fig. 4). In these areas the boundary is emphasised by black paper shales at the top of the Kap Stewart Group, grading into well-sorted fossiliferous sandstones, rich in marine trace fossils, of the Elis Bjerg Member.

Along Neill Klinter the upper boundary is marked by the appearance of alternating mudstones and well-sorted fine-grained sandstones associated with coarse-grained pebbly sheet sandstones and massive beds of the Albuen Member. In the northern part of the basin the upper boundary is placed at the occurrence of the small coarsening-upward successions, rich in thin coal seams and rootlet beds, of the Horsedal Member.

Distribution. Same as the formation.

Geological age. Body fossils are almost completely lacking and none are age diagnostic. Belemnites and ammonites from the Nathorst, Lepidopteriselv and Skævdal Members suggest that the oldest strata of these members belong to the Commune Subzone (the oldest subzone of the Lower Toarcian Bifrons Zone) or even to the lowermost Toarcian Tenuicostatum Zone (Semicelatum Subzone) (Doyle, 1991; J. H. Callomon, personal communication, 1993), suggesting that the Elis Bjerg Member has a Late Pliensbachian age in the south-eastern part of the basin, where the member overlies the Rævekløft Formation. Dinoflagellate cysts suggest that the Elis Bjerg Member has a Late Pliensbachian (probably Margaritatus Zone age) to Early Toarcian age along Neill Klinter (Koppelhus & Dam, in press). Where the Elis Bjerg Member shows a gradual transition to the underlying Kap Stewart Group, palynomorphs in the uppermost part of the Kap Stewart Group suggest a latest Sinemurian to earliest Pliensbachian age and the Elis Bjerg Member an Early to Late Pliensbachian age (Koppelhus & Dam, in press).

Albuen Member

new member

History. Strata composing this member were previously included in the upper part of the Gule Horn Member of Surlyk *et al.* (1973).

Name. The member is named after a kink on the western coastline (albuen is the Danish word for 'the elbow') of Hurry Inlet, south-east Jameson Land, above which the member is well-exposed (Fig. 1).

Type and reference localities. One of the most complete and well-exposed sections of the member occurs at Albuen. This is designated the type section. Other well-exposed reference sections occur at Qupaulakajik, in Goniomyakløft, Astartekløft, Moskusoksekløft, and at Harris Fjeld (Figs 1, 20).

Thickness. The member is 22 m thick at the type locality, and more than 26 m at Qupaulakajik. Further north the member thins to 7 m at Harris Fjeld and to less than 1 m at Nathorst Fjeld. The member wedges out north of Dusén Bjerg.

Lithology. The member is characterised by alternating mudstones and well-sorted, fine-grained sandstones. Coarse-grained, pebbly beds moulded into symmetrical ripples, and massive beds composed of muddy sandstones with scattered quartzite pebbles, granitic boulders, ooids, bivalves and fragmental belemnites, are commonly interbedded with the alternating mudstones and fine-grained sandstones.

Facies associations and depositional environments. Two facies associations are recognised in the member (i and j; Fig. 20).

i. Storm-dominated lower shoreface association. This association constitutes most of the Albuen Member and consists of heteroliths composed of alternating mudstones and well-sorted very fine to fine-grained sandstones. The sandstones range from millimetre-thick streaks showing incipient wave ripple lamination (similar to Facies M_1 of De Raaf *et al.*, 1977), to thin, laterally persistent, parallel-sided beds, less than 30 cm thick. The bases of the sandstones are sharp and may be draped by thin lag conglomerates. Internally the bedded sandstones show parallel lamination, hummocky and swaley cross-stratification. The wavelength of the hummocks are up to 65 cm with heights less than 10 cm. Wave ripple cross-lamination and mud-draped wave ripple formsets are common at the top of the beds. Occasionally, the laminae and beds are arranged in thickening-upward units, less than 15 m thick, grading into massive fine to medium-grained sandstones (Fig. 20). The massive sandstones show a high degree of bioturbation, with *Taenidium serpentinum* completely obliterating the primary physical structures.

Coarse-grained pebbly sandstone sheets, moulded into large symmetrical ripples, are occasionally interbedded with the heterolithic deposits (Fig. 21). The coarse-grained ripples occur in sharply based laterally persistent beds. Ripples are up to 30 cm high and wave lengths reach 175 cm. Crest lines are straight to sinuous, commonly showing bifurcations. Stratification within the ripples is only rarely visible.

The shelf association of the member contains elements of the *Planolites* ichnocoenosis including common *Planolites beverleyensis, Taenidium serpentinum* and *Gyrochorte comosa* (Dam, 1990b).

Interpretation. The mudstones were deposited in a lower shoreface environment during fair-weather periods. The laminated and bedded sandstones were deposited as tempestites during periodic storms at water depths below fair-weather wave base but above storm-wave base. The coarsening-upward units are shallow-

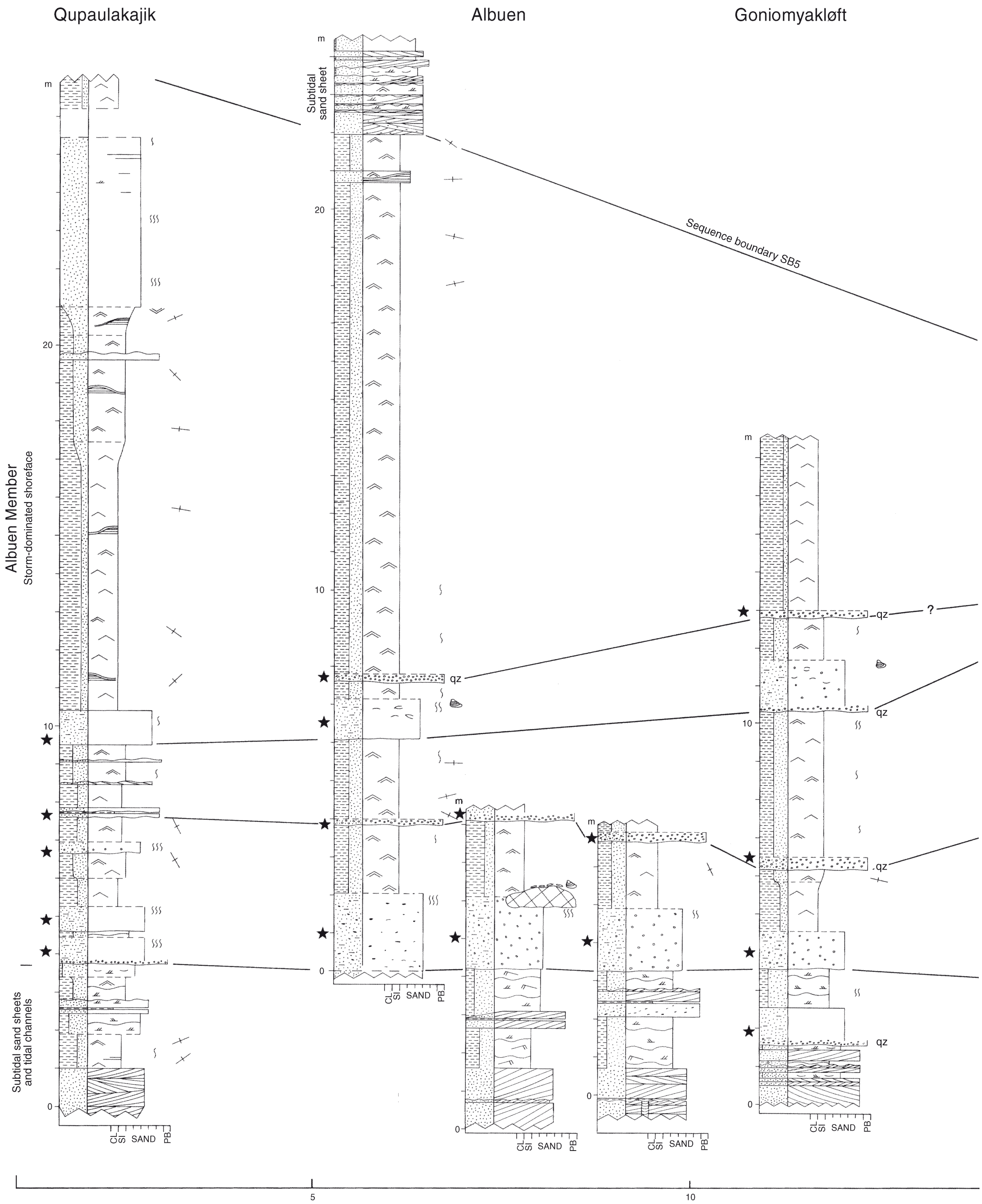
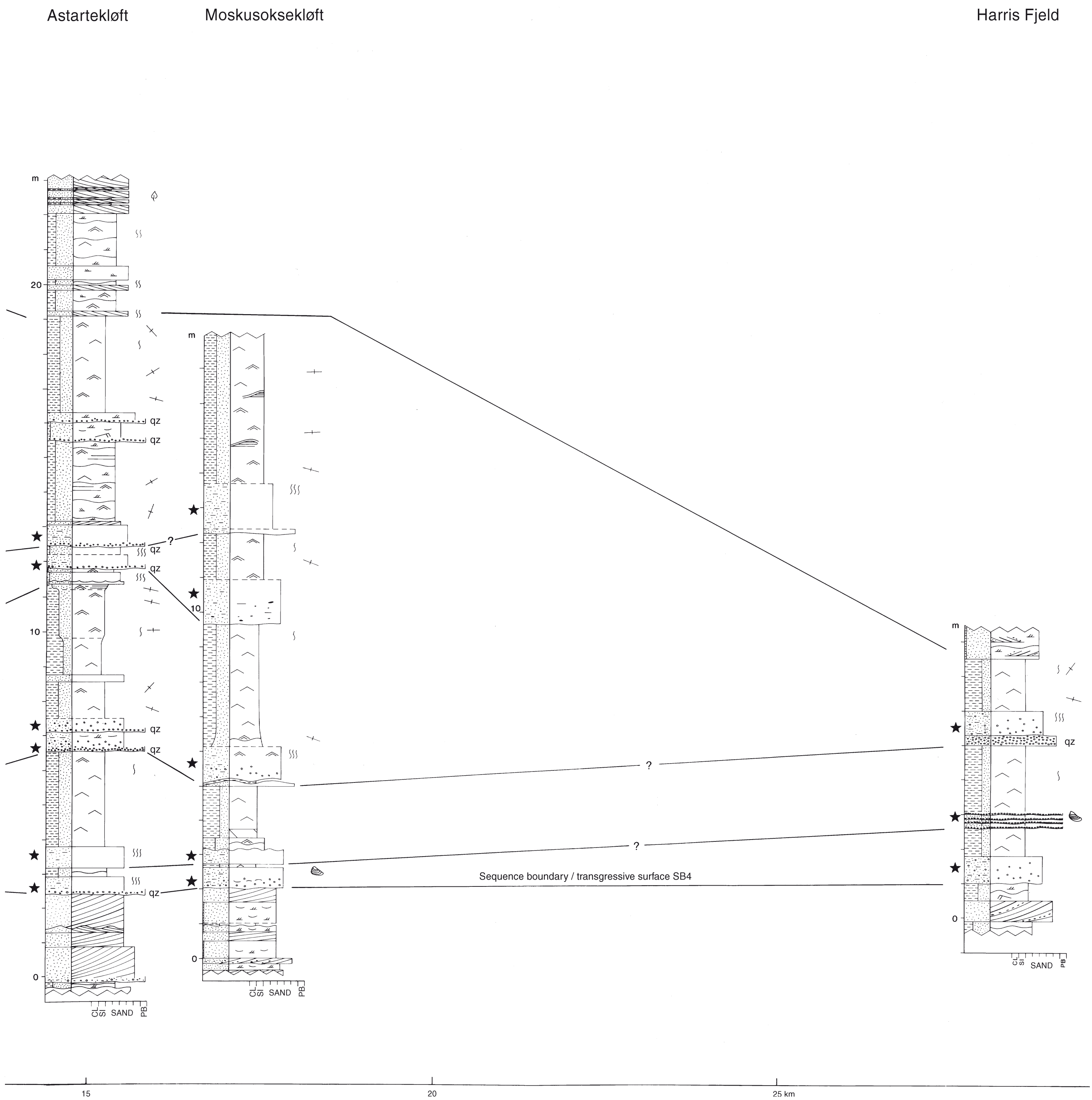


Fig. 20. Sedimentological logs through the debris flow and storm-dominated lower shoreface deposits of the Albuen Member. Asterisks mark the submarine debris flow deposits. Measured along Neill Klinter. See Fig. 1 for locations and Plate 1 for legend.



Astartekløft
Moskusoksekløft
Harris Fjeld
m
20
10
0
qz
?
Sequence boundary / transgressive surface SB4
CL
SI
SAND
PB
15
20
25 km

Fig. 21. Coarse-grained symmetrical ripple formset of the storm-dominated lower shoreface association (i) of the Albuen Member. Astartekløft. See Fig. 1 for location. Hammer 32 cm long.

ing-upward cycles with lower shoreface deposits overlain by upper shoreface sandstones. The sands were probably derived from coastal erosion and transported by downwelling flows during storms (cf. Snedden *et al.*, 1988).

The coarse-grained ripples are similar to those present in the underlying Rhaetian – Sinemurian Kap Stewart Group (Clemmensen, 1976; Dam, 1991; Dam & Surlyk, 1992, 1993). The symmetry of the ripple forms and the association with other wave and storm-produced facies indicate that the coarse-grained ripples are wave ripples, formed by large amplitude, long period storm waves, like those recorded from modern shelf settings at water depths up to 100 m (Newton & Werner, 1972; Hunter *et al.*, 1982; Cacchione *et al.*, 1984; Leckie, 1988). However, similar ripples have also been recorded from bays with a fetch of only a few tens of kilometres, at water depths of 2 to 20 m (e.g. Masuda & Makino, 1987; Hunter *et al.*, 1988).

The trace fossils of the heterolithic deposits were mainly produced by infaunal organisms combining the activity of deposit-feeding and locomotion (endostratal pascichnia burrows) (Dam, 1990a, b). The low diversity suggests a limitation of oxygen supply within the sediment (cf. Ekdale & Mason, 1988). However, the dominance of pascichnia suggests that the interstitial environment, supporting production of pascichnia, must have contained at least some oxygen to allow

Fig. 22. Submarine debris flow deposits (Facies association j) interbedded with storm-dominated lower shoreface deposits (Facies association i). Albuen Member, Albuen. See Fig. 1 for location.

Fig. 23. Granitic boulder sitting on top of submarine debris flow deposits (Facies association j). Encrusting oysters occur on upper surface of the boulder (arrow). Albuen Member, Albuen. See Fig. 1 for location. Hammer 32 cm long.

respiration. Hence a dysaerobic bottom environment is suggested by this ichnocoenosis (Dam, 1990b). The low diversity may also result from a more stressed, brackish water environment, as inferred by the palynomorphs (Koppelhus & Dam, in press).

j. Submarine debris flow association. This association consists of massive pebbly muddy sandstone beds that are interbedded with storm-dominated lower shoreface deposits (Facies association i) (Figs 20, 22). The beds have a sharp lower boundary, are up to 2 m thick, and can be followed laterally for more than 30 km along strike (Fig. 20). The beds contain quartzite pebbles, and rare granitic boulders with a diameter up to 1.5 m, fragmental belemnites and bivalves, including *Protocardia* sp., *Camptonectes* sp. and *Isocyprina* sp., and ooids scattered in a matrix of muddy sandstones. The upper boundary of the beds is blurred due to strong bioturbation by *Arenicolites* isp. 1 (Dam, 1990b). The upper surfaces of some of the boulders at the top of the beds are encrusted by oysters (Fig. 23).

Interpretation. The mixed lithology and the absence of an organised fabric or grading of the massive beds suggest deposition from high viscosity debris flows in which clasts were supported by matrix strength. The clast composition suggests a source area characterised by gravel beaches or deltas. The ooids may have formed in tidal bars or within tidal deltas and indicate that strong bottom currents existed in the sediment source area. The debris flows may have been triggered by major storms, tectonic activity or catastrophic floods in the sediment source area.

The debris flow deposits represent major environmental changes in substrate conditions on the storm-dominated shelf. The Planolites ichnocoenosis of the heterolithic deposits is characteristic of dysaerobic bottom environments (see Facies association d), whereas the presence of the *Arenicolites* isp. 1 and oysters encrusting boulders on top of the marine debris flow deposits suggest a well-aerated environment. The *Arenicolites* isp. 1 ichnocoenosis includes traces very similar to those made by modern and ancient opportunistic suspension-feeding tube-dwelling polychaetes, colonising marine habitats after major environmental changes. Thus, the presence of the *Arenicolites* ichnocoenosis suggests that during debris flow deposition the general dysaerobic conditions of the bottom substrate of the Albuen Member shelf were interupted by short periods of well-aerated conditions in the bottom water, with no net sedimentation and abundant food supply in suspension (Dam, 1990b).

Fossils. Body fossils include fragmented belemnites and bivalves such as *Protocardia* sp., *Camptonectes* sp. and *Isocyprina* sp. Trace fossils include common *Arenicolites* isp., *Planolites beverleyensis*, *Taenidium serpentinum*, and rare *Gyrochorte comosa*.

Boundaries. The muddy deposits of the member rest on the cross-bedded sandstones and heterolithic deposits of the Elis Bjerg Member with a sharp boundary (Fig. 20). The upper boundary is non-erosional along Neill Klinter, but erosional at Nathorst Fjeld and Elis Bjerg. It is placed where the muddy deposits are overlain by cross-bedded sandstones of the Astartekløft Member.

Distribution. The member occurs along the length of Neill Klinter, at Harris Fjeld and as far north as Dusén Bjerg, where it wedges out.

Geological age. Body fossils are generally rare and none are age diagnostic. Palynomorphs suggest a Late Pliensbachian age (Koppelhus & Dam, in press).

Ostreaelv Formation

new formation

History. This formation corresponds to the Ostreaelv Member of Surlyk *et al* (1973).

Name. The formation is named after the river Ostreaelv, in the south-eastern part of Jameson Land, where the uppermost part of the formation is well-exposed (Fig. 1; Surlyk *et al.*, 1973).

Type locality and reference localities. The upper part of the formation is well-exposed at Ostreaelv (Surlyk *et al*, 1973). This is designated the type section. Well-exposed reference sections occur along Neill Klinter at Albuen, at Harris Fjeld and Nathorst Fjeld and at Lepidopteriselv and Horsedal (Fig. 1).

Thickness. The formation is less than 90 m at the type locality, 125 m at Albuen, 122 m at Harris Fjeld, more than 112 m at Nathorst Fjeld, 157 m at Liaselv and more than 155 m at Horsedal.

Lithology. The Ostreaelv Formation has a very variable lithological composition. One of the most distinctive characteristics of the formation is the appearance of body fossils that are very rare in the underlying Gule Horn Formation. At the type section, along Neill Klinter, and at Nathorst Fjeld the formation is dominated by fine to very coarse-grained sandstones. Many sandstones are cross-bedded and rhythmical clay drapes are common on the foresets. Concretionary cement has hardened some beds. At Lepidopteriselv the formation consists of medium to very coarse-grained, concretionary, cross-bedded sandstones, occasionally with logs, interbedded by bioturbated sandy mudstones. From Lepidopteriselv westwards towards Liaselv and southwards the cross-bedded sandstones pass laterally into medium-grained bioturbated sandstones. At Horsedal the formation is initiated with a unit of thin coarsening-upward successions composed of sand-streaked mudstones grading upward into fine to medium-grained sandstones. Coal seams and rootlet beds commonly occur on top of the coarsening-upward successions. This unit is overlain by a unit composed of cross-bedded, cross-laminated, parallel-laminated and hummocky cross-stratified fine to medium-grained sandstones alternating with bioturbated sandstones. At this locality thin conglomerates, composed of discoidal quartzite pebbles, occur at several levels.

Fossils. Body fossils occur at several horizons, most commonly in those hardened by concretionary cement. Rosenkrantz (1934) identified a diverse, dominantly European fauna of 69 species, dominated by bivalves and including brachiopods, crinoids and cephalopods. Fish scale and rootlet horizons have also been found in the Horsedal Member. Trace fossils are common and include *Arenicolites* isp., *Cochlichnus anguineus, Cruziana* isp., *Curvolithos multiplex, Diplocraterion habichi, Diplocraterion paralellum, Gyrochorte comosa, Gyrophyllites kwassicensis, Helminthopsis magna, Lockeia amygdaloides, Nereites* isp., *Ophiomorpha nodosa, Palaeophycus* isp., *Parahaentzschelinia surlyki, Phoebichnus trochoides, Phycodes bromleyi, Planolites beverleyensis, Scolicia* isp., *Taenidium serpentinum, Teichichnus* isp., *Thalassinoides* isp., *Rhizocorallium irregulare,* and *Rhizocorallium* isp.

Boundaries. Along Neill Klinter the lower boundary is sharp and placed between the alternating sandstones and mudstones of the Albuen Member and the overlying cross-bedded sandstones of the Astartekløft Member. In the northern and central parts of the basin the lower boundary is placed where cross-laminated and cross-bedded sandstones of the Elis Bjerg Member give way to wave ripple cross-laminated, parallel laminated and hummocky cross-stratified sandstones arranged in small coarsening-upward successions of the Horsedal Member.

Throughout the exposed part of the basin the upper boundary of the formation, between the sandstones of the Trefjord Bjerg Member and the dark mudstones of the Sortehat Formation, is flat, very sharp and in places pebble strewn.

Distribution. Same as for the group.

Geological age. Palynomorphs, belemnites and ammonites suggest a latest Pliensbachian – earliest Aalenian age of the formation (Doyle, 1991; J. H. Callomon, personal communication, 1993; Koppelhus & Dam, in press).

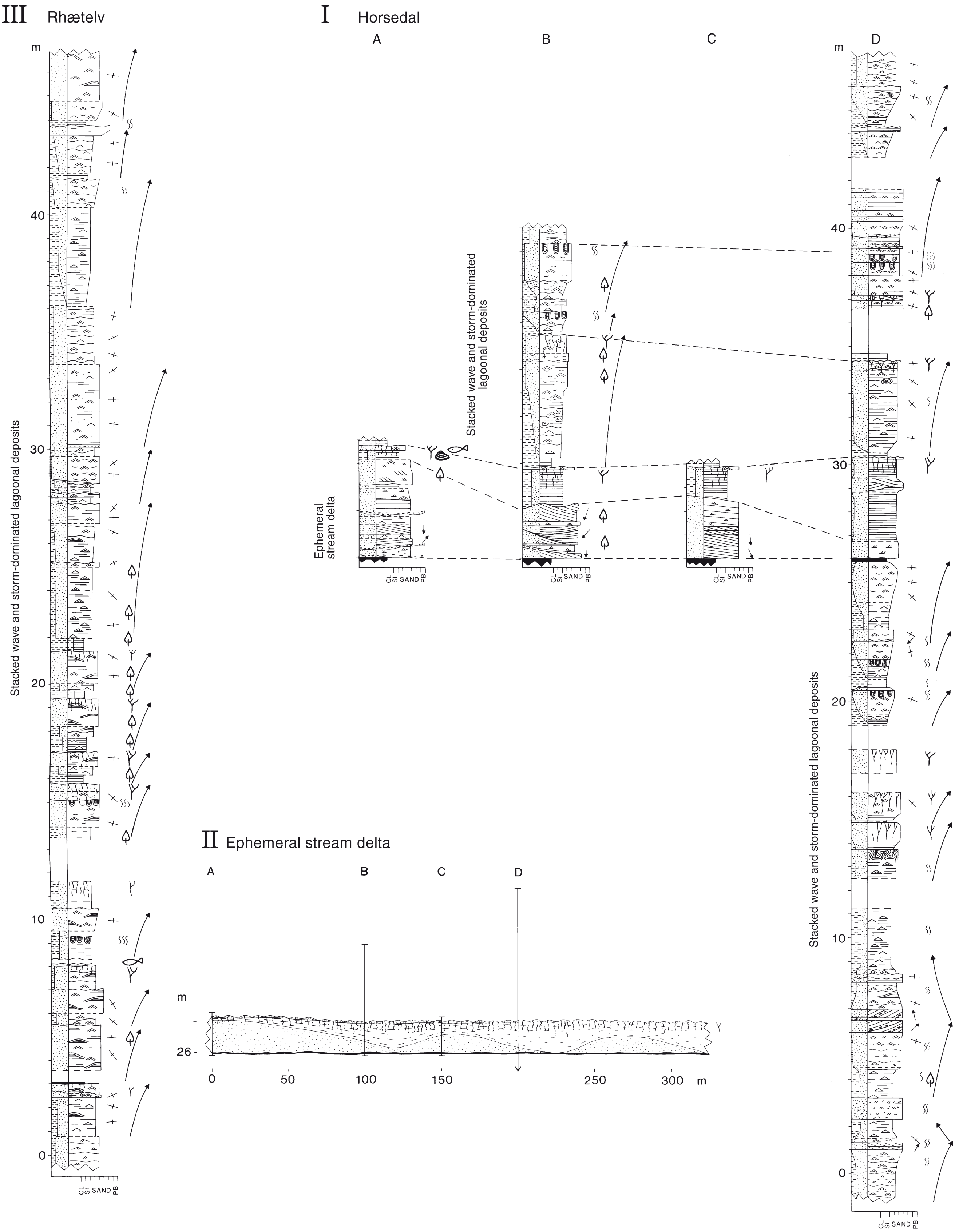


Fig. 24. Sedimentological logs through stacked wave and storm-dominated coarsening-upward lagoonal (Facies association k) and ephemeral stream delta (Facies association l) deposits of the Horsedal Member. I = section measured in Horsedal; II = lateral profile showing the morphology of the ephemeral stream delta deposits in Horsedal. Vertical lines show location of vertical profiles from I; III = section measured in Rhætelv, *c.* 25 km south of Horsedal. See Fig. 1 for locations and Plate 1 for legend.

Fig. 25. Stacked, coarsening-upward sheet sandstones formed by repeated progradation of the shoreface into a wave and storm-dominated lagoon (Facies association k). Horsedal Member, Horsedal. See Fig. 1 for location. Encircled person for scale.

Subdivisions. The formation is subdivided into seven members. They are the Horsedal, Astartekløft, Lepidopteriselv, Harris Fjeld, Nathorst Fjeld, Skævdal and Trefjord Bjerg Members (Fig. 3).

Horsedal Member

new member

History. The sediments of this member have not previously been recognised as a stratigraphic unit. It has not been possible to decide whether the sediments were included in the Ostreaelv or Gule Horn Members (*sensu* Surlyk *et al.*, 1973) in previous studies.

Name. The member is named after the valley Horsedal north of Ørsted Dal, in Scoresby Land (Fig. 1).

Type and reference localities. Two nearly complete sections occur in Horsedal and at Rhætelv; the former is chosen as type section and the latter as reference section (Figs 1, 24).

Thickness. The member is 58 m thick at the type locality and 50–55 m at Rhætelv and Liaselv.

Lithology. The member is dominated by thin coarsening-upward successions, 1–6 m thick, composed of sand-streaked mudstones grading upward into fine to medium-grained sandstones. Coal seams and rootlet beds commonly occur on top of the coarsening-upward units. Parting planes often show wave ripples.

Facies associations and depositional environments. Two facies associations are recognised in the Horsedal Member (k and l).

k. Wave and storm-dominated lagoonal association. This association is characteristic of the member and consists of small coarsening-upward units, 1–6 m thick (Figs 24, 25). The lower part of the units consists of streaked mudstones. The streaks are composed of fine-grained silty sandstones showing incipient wave ripple lamination (cf. Facies M_1 of De Raaf *et al.*, 1977). The lower streaked mudstones grade upward into wavy and flaser-bedded heterolithic deposits and fine to medium-grained sandstones. The wavy and flaser-bedded deposits show a large variety of structures typical of wave action (cf. Facies M_2 and S_1 of De Raaf *et al.*, 1977). A maximum depositional water depth for these sediments of approximately 10 m is estimated on the basis of wave ripple features following the method of Diem (1985). The fine to medium-grained sandstones are low-angle cross-bedded, parallel laminated or massive. Beds of single and amalgamated hummocky and swaley cross-stratified, fine-grained sandstones occur within the coarsening-upward successions. Root-like structures occasionally occur in distinct wave-worked beds on top of the coarsening-upward units. They are vertical, sand-filled and have a thin coaly lining. They commonly show upward dichotomous branching, and thin bituminous coal seams may overlie the beds.

The *Diplocraterion parallelum* ichnocoenosis characteristic of this association is generally monospecific, including only *Diplocraterion parallelum*, but *Arenicolites* isp., *Cochlichnus anguineus, Helminthopsis magna, Lockeia amygdaloides* and *Planolites beverleyensis* may occur (Dam, 1990b). *D. parallelum* is restricted to certain levels, which are strongly bioturbated (up to 100%), and can be followed for several hundreds of metres.

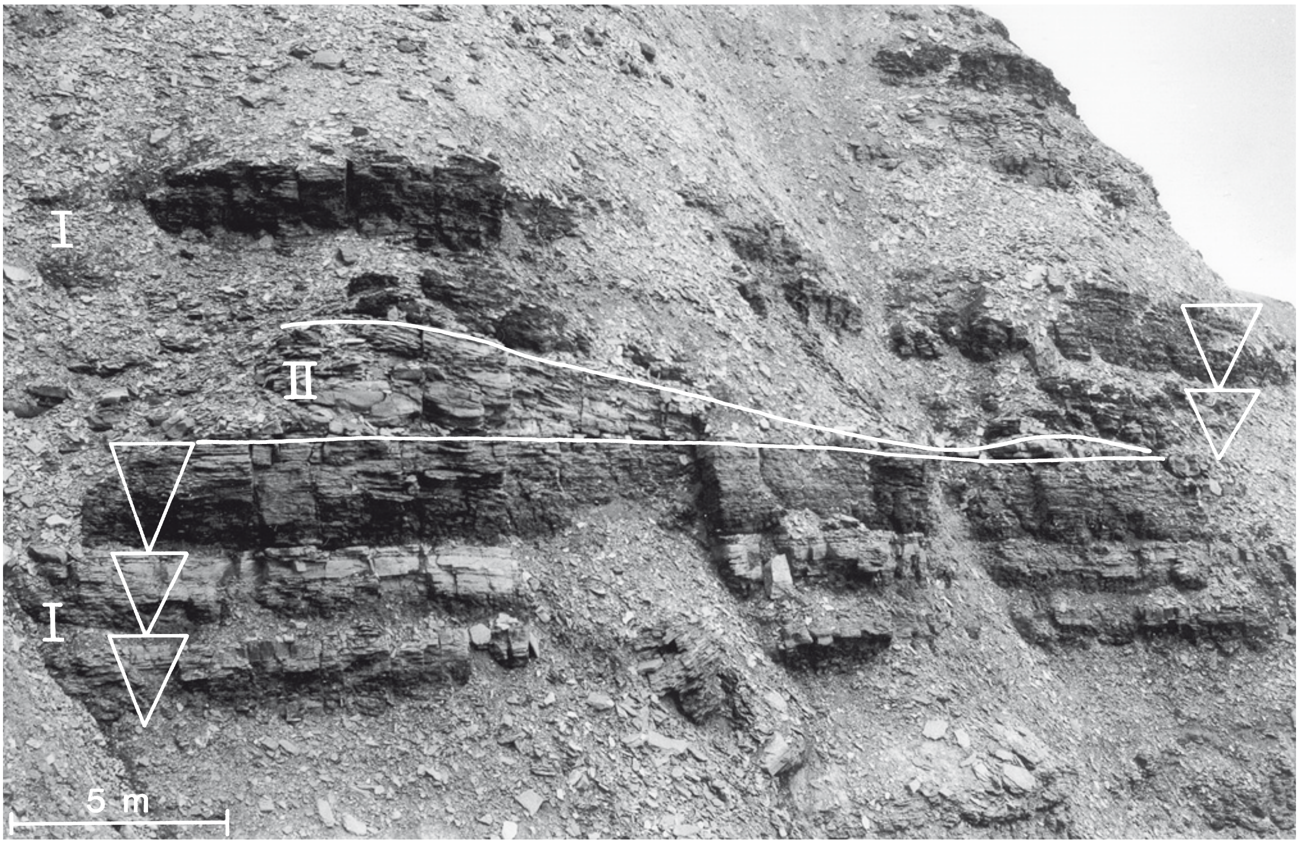


Fig. 26. Ephemeral stream delta lobe sandstones associated with wave and storm-dominated lagoonal deposits. I = stacked wave and storm-dominated lagoonal deposits (Facies association k); II = ephemeral stream delta deposits (Facies association l). Horsedal Member, Horsedal. See Fig. 1 for location.

Leaves, bivalves and concretions with fish remains commonly occur.

Interpretation. The physical structures show features formed by wave and storm action (cf. De Raaf *et al.*, 1977). The internal structures of the thin, fine to medium-grained coarsening-upward units (facies M_2 to S_1 of De Raaf *et al.*, 1977) reflect an increase in wave energy and a shallowing-upward tendency. The low-angle cross-bedded and parallel laminated fine-grained sandstones formed by migration of swash and foreshore ridges on the beach foreshore. The coarsening-upward units represent wave-dominated beaches or bay-head delta systems prograding into an extensive lagoonal environment. A lagoonal environment is supported by the palynomorphs which indicate a freshwater depositional environment (Dam & Koppelhus, in press).

The root-like structures and coal seam couplets represent autochthonous plant remains, formed when the swamp covered the lagoonal delta fronts during delta abandonment. The dichotomous upward-branching of some of the root-like fossils suggests that their upper parts are fossil stems in growth position. The sandy fills indicate that the plants had hollow stems and roots (aquatic plants) and that they grew during sedimentation. Similar plant fossils have been observed in the underlying Kap Stewart Group (Dam, 1991; Dam & Surlyk, 1993), in the Hettangian–Sinemurian Sose Bugt Member of Bornholm, Denmark (Surlyk *et al.*, 1995), and in the Wealden of SE England (Allen, 1981).

l. Ephemeral stream delta association. This association consists of sharply based lenses, with a mound-shaped upper surface, composed of poorly sorted medium to very coarse-grained sandstones. The lenses occur at a single stratigraphic level in Horsedal, associated with wave and storm-dominated lagoonal shoreface deposits (Facies association k; Figs 24, 26). In 2-D sections, the lenses are convex-upward, regularly spaced, and almost isolated from one another. They are up to 4 m thick and thin laterally to less than 25 cm. The distance between the crests of neighbouring lenses range from 75 m to more than 200 m. The sandstones are predominantly cross-bedded, parallel or cross-laminated and form a single fining-upward succession. The lenses are draped by thin form-concordant cross-laminated beds. In one section the basal scoured surface is succeeded by three distinct fining-upward units, 30–150 cm thick, in erosional contact with each other (Fig. 24, IA). Foreset dip directions are generally towards the south, perpendicular to the 2-D sections. The boundary to the overlying mudstones is sharp but non-erosional.

Interpretation. The geometry and poor sorting of the sandstone bodies, and the internal facies arrangement, suggest deposition from unidirectional erosive flows during a single depositional event. The vertical and lateral facies variation within the lenses, and the presence of small erosionally-based fining-upward units, within a major fining-upward succession, reflect highly variable and pulsating hydrodynamic conditions. The sharp, non-erosional contact to the overlying mudstones and the form-concordant bedding suggest that the surface of the 'lenses' represents the original morphology. The palaeocurrent directions indicate a flow direction towards the south, perpendicular to the northern WNW–ESE orientated basin margin, suggesting that the lenses represent cross-sections of small subaqueous lobes that prograded from land into a protected lagoonal environment. The lobes were probably formed during a major flooding event. The flood flow apparently behaved like a plane-jet, scouring the bottom before deposition of the subaqueous ephemeral delta lobes. The form-concordant bedding formed from migrating ripple trains on top of the lobe surface during waning flow.

Fossils. Plant fossils and fish scales are common in concretionary layers. Trace fossils commonly include *Diplocraterion parallelum*, and rare *Arenicolites* isp., *Cochlichnus anguineus, Helminthopsis magna, Lockeia amygdaloides* and *Planolites beverleyensis*.

Boundaries. The lower boundary is placed where cross-laminated and cross-bedded sandstones of the Elis Bjerg Member give way to wave ripple cross-laminated, parallel laminated and hummocky cross-stratified sandstones arranged in small coarsening-upward successions. The Horsedal Member is erosionally overlain by cross-bedded, hummocky cross-stratified and bioturbated sandstones of the Lepidopteriselv Member.

Distribution. The member occurs in Scoresby Land, and in the northern part of Jameson Land where it can be traced as far south as Liaselv. It has not been recorded south of this locality (Fig. 1).

Geological age. None of the recovered body fossils are age diagnostic. Belemnites and ammonites from the overlying Nathorst Fjeld, Lepidopteriselv and Skævdal Members, suggest a Commune Subzone age (the oldest subzone of the Lower Toarcian Bifrons Zone) or even a lowermost Toarcian Tenuicostatum Zone age (Semicelatum Subzone) (Doyle, 1991; J. H. Callomon, personal communication, 1993), suggesting that the Horsedal Member is of Late Pliensbachian age. The member is probably time-equivalent to the Astartekløft Member.

Astartekløft Member

new member

History. Strata included in this member were first described by Rosenkrantz (1934, p. 83, pl. 12, 430–468 m) at Nathorst Fjeld. They form the lower third of the Ostreaelv Member of Surlyk *et al.* (1973) along Neill Klinter, Nathorst Fjeld and Elis Bjerg.

Name. The member is named after the ravine Astartekløft in Neill Klinter in the south-eastern part of the basin (Fig. 1).

Type and reference localities. One of the most well-exposed and complete developments of the member occurs in Astartekløft and is designated the type locality. Well-exposed reference sections occur at Nathorst Fjeld, where the member forms an almost vertical cliff section, Harris Fjeld and along Neill Klinter at Moskusoksekløft, Goniomyakløft, Albuen and Qupaulakajik (Figs 1, 27, 28).

Thickness. The member is 25 m thick at the type section, 18–43 m along Neill Klinter, and 29 m at Nathorst Fjeld.

Lithology. Along Neill Klinter the member consists of cross-laminated and cross-bedded sandstones alternating with thin mudstone beds, similar to the Elis Bjerg Member. Plant fragments are common, and logs and belemnites occur occasionally. At Nathorst Fjeld the member consists of medium to very coarse-grained cross-bedded sandstones, which pass upward into medium-grained bioturbated sandstones with bivalves and crinoids.

Facies associations and depositional environments. Along Neill Klinter the member includes four facies associations (f, g, h and m). They are very similar to the analogous facies associations of the Elis Bjerg Member and are designated as such. However, they are described separately in this section. The tidal channel succession which occurs at Nathorst Fjeld has a very different character from the tidal channel successions otherwise recognised in the Gule Horn Formation and Astartekløft Member and is therefore treated separately.

f. Subtidal sand sheet association. This association is characteristic of the member all along Neill Klinter in the south-eastern part of the basin. It consists of sandstone beds and alternating laminated and thinly bedded mudstones and sandstones arranged in a single coarsening-upward or several smaller coarsening or fining-upward successions (Fig. 27).

The sandstones occur in planar and trough cross-bedded medium to very coarse-grained beds with scattered plant fragments and logs. Each set is 10–145 cm thick and has tangential foresets. Thin single mudstone drapes and mudstone clasts commonly occur along foresets. They flatten out along toesets and extend downward into heterolithic beds. Reactivation surfaces are common and occasionally they are capped by cross-laminated sandstones showing a foreset dip direction in the opposite direction to the master bedding. The cross-bedded sets are generally separated by thin heterolithic beds, 10–35 cm thick. Bedding planes may show lunate current ripples and occasionally wave ripples. Cone-in-cone structures are common in the heterolithic beds. The sandstones are characterised by elements of the *Diplocraterion* ichnocoenosis and the heterolithic beds include elements of the *Cochlichnus* ichnocoenosis of Dam (1990b). Trace fossils include common *Curvolithos multiplex, Diplocraterion parallelum, Gyrochorte comosa, Ophiomorpha nodosa, Phoebichnus trochoides, Planolites beverleyensis, Taenidium serpentinum, Teichichnus* isp. and *Rhizocorallium* isp.

Combined palaeocurrent data show bipolar directions toward SSW and NNE indicating reversing currents. The dominant direction is towards SSW.

Interpretation. The presence of mudstone drapes, reactivation surfaces capped by cross-laminated sandstones, bipolar palaeocurrent directions, and lack of any indications of subaerial exposure indicate that the sandstones were deposited as migrating small and medium-scale subtidal dunes in a strong subtidal current regime (e.g. Boersma, 1969, Visser, 1980). The presence of elements of the *Diplocraterion* ichnocoenosis suggests a high-energy, shallow subtidal to intertidal environment (Fürsich, 1975).

The presence of heterolithic beds between the cross- bedded sandstones suggests that these represent inter-dune areas. The trace fossil assemblage of the heterolithic beds reflects a wide variety of feeding habits and behavioural categories, indicating a well-aerated medium to low-energy shallow marine environment, with abundant food supply for both suspension and deposit-feeders, and mobile carnivores (Dam 1990b).

g. Tidal channel association. This association is characteristic of the member at Albuen (Fig. 27). It consists of fining and thinning-upward successions, up to 8.5 m thick. The successions are bounded by a lower erosion surface, commonly draped by a lag conglomerate composed of mudstone and sandstone intraclasts. The conglomerate is overlain by coarse-grained sandstones with distinct trough and planar cross-bedded cosets, with set heights decreasing upwards from 90 cm to 20 cm. The sets may be separated by thin mudstone layers. Scattered mudstone clasts are common in the sandstones. Foreset dip directions are unimodal towards west, in a basinwards direction. The upper parts of the successions show an upward increase in number of intercalated mudstone layers, and the cross-bedded sandstones grade into bioturbated flaser, wavy and lenticular-bedded heteroliths. The heterolithic deposits are commonly extensively burrowed by *Arenicolites* isp., *Diplocraterion parallelum* and *Rhizocorallium* isp., elements of the *Diplocraterion* ichnocoenosis of Dam (1990b).

Interpretation. The distinct fining-upward units were formed by lateral migration of channels. The base of the units was formed by erosion along the active channel thalweg. The succeeding cross-bedded cosets were formed by 2-D and 3-D small-scale dunes migrating on the channel floor. The upper heterolithic parts of the successions were deposited along the inner bend of the channel. The lack of any clear tidal characters are interpreted as reflecting deposition in intertidal channels. Some tidal influence is, however, suggested by the presence of mudstone layers separating individual sets. They were probably deposited during slack-water stages. The trace fossil assemblage indicates a high-energy, shallow subtidal to intertidal environment (cf. Fürsich, 1975).

h. Storm-dominated sandy shoal association. This association is characteristic of the Astartekløft Member at Albuen (Fig. 27). It consists of amalgamated fine to medium-grained sharply-based sandstone beds closely associated with tidal channel deposits. Each bed is 15–110 cm thick and forms an amalgamated bed-set, 4.8 m

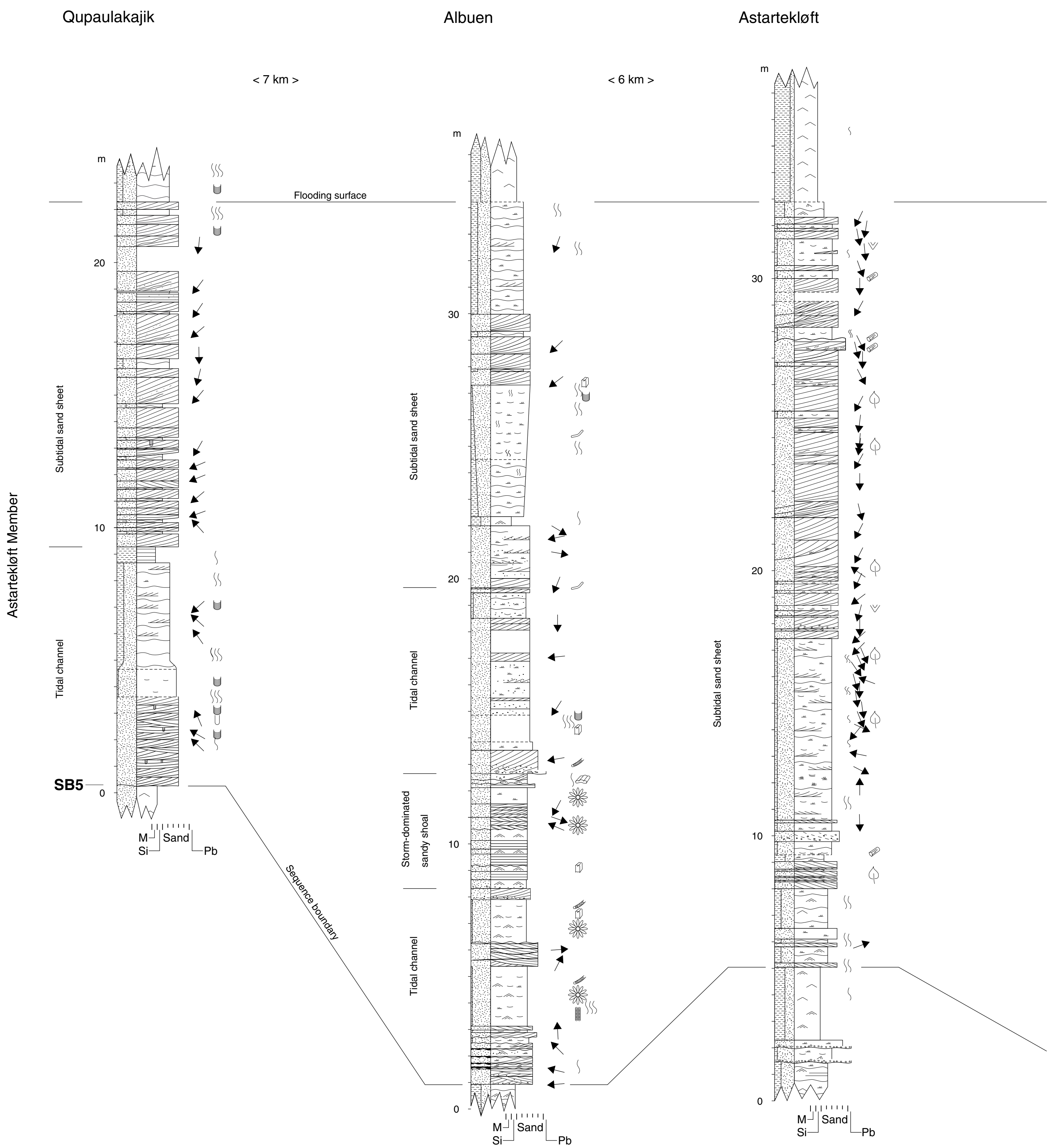
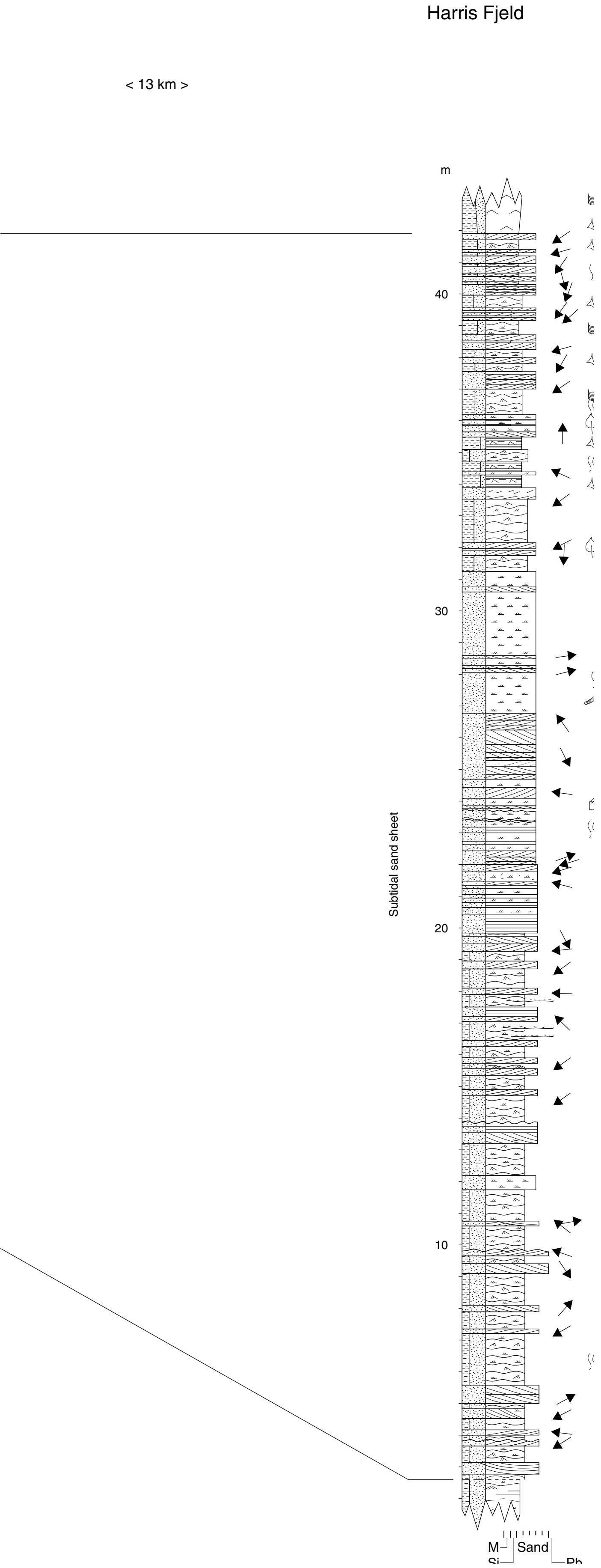


Fig. 27. Sedimentological log through the tidal channel, subtidal sand sheet and storm-dominated sandy shoal deposits of the Astartekløft Member. See Fig. 1 for locations and Plate 1 for legend. Modified from Dam & Surlyk (1995).



thick. Internally the sandstones are parallel-laminated, grading upward into wave ripple cross-lamination. Some beds may show cross-bedding with scattered mudstone clasts. The sandstones are bioturbated weakly to heavily by *Gyrochorte comosa, Ophiomorpha nodosa, Phoebichnus trochoides, Planolites beverleyensis, Taenidium serpentinum*, and elements of the *Planolites* and *Taenidium* ichnocoenoses (Dam, 1990b).

Interpretation. The sharp bases of sandstone beds and the dominance of parallel lamination suggest episodic deposition from waning flows characterised by combined flow-processes or intense bed shear due to storm-wave activity. The cross-laminated tops were formed during decelerating flow. The interbedding with tidal channel deposits suggests that the successions represent storm-dominated shoals. The *Planolites* and *Taenidium* ichnocoenoses are very characteristic of deposits of storm-dominated environments in the Neill Klinter Group, and indicate that between storm events, periods of low-energy and oxygen-limitation prevailed in the substratum.

m. Tidal channel association. This association is characteristic of the Astartekløft Member at Nathorst Fjeld. It consists dominantly of planar and elongate trough cross-bedded medium to very coarse-grained sandstone (Fig. 28), and less commonly of bioturbated, parallel and cross-laminated sandstones. Set thickness varies from 0.2 to 2.9 m. Foresets are tangential and dip angles range from 17 to 30°. Logs and plant fragments occur in the basal part of the sets. The cross-bedded sets form a coset, 29 m thick, with thinning upward set thicknesses (Fig. 28). The lower surface of the cosets may be erosional. The foresets dip unimodally towards the west.

In the upper part of the succession the sandstones are commonly burrowed by *Curvolithos multiplex, Diplocraterion habichi, Gyrochorte comosa, Ophiomorpha nodosa*, and *Taenidium serpentinum*, and fragments of the crinoid *Pentacrinus* are common.

Interpretation. The cross-bedded sandstones were deposited by extensive fields of migrating small and medium-scale 2-D and 3-D dunes. The lower erosional boundary of the succession, the general offshore directed currents, the upward thinning of set thickness and the body and trace fossil fauna suggest deposition in the subtidal parts of a tidal channel.

Boundaries. The boundary between the muddy deposits of the underlying Albuen Member and the cross-bedded sandstone of the Astartekløft Member is erosional along Hurry Inlet (Fig. 27). At Nathorst Fjeld

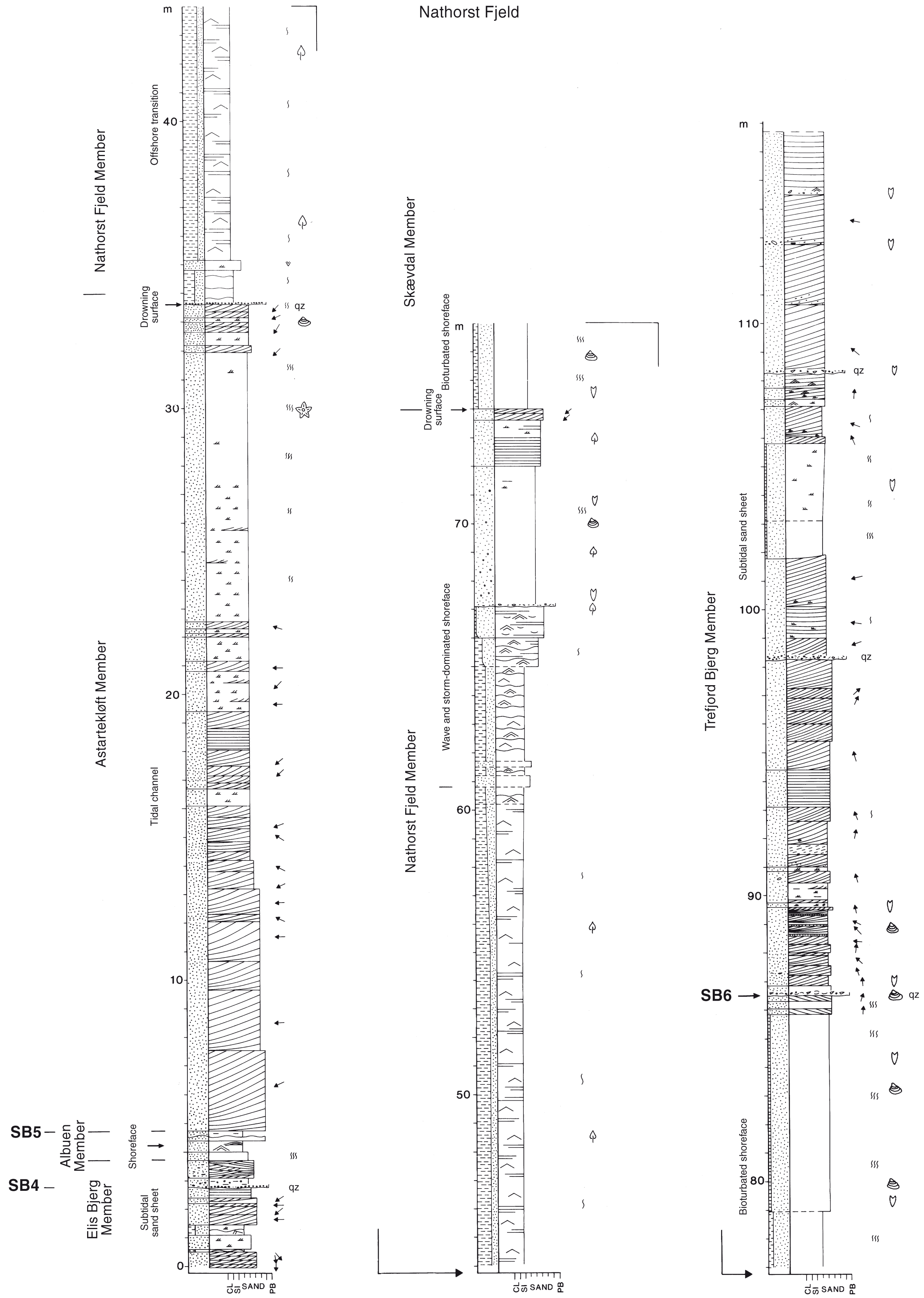


Fig. 28. Sedimentological log through the Astartekløft, Nathorst Fjeld, Skævdal and Trefjord Bjerg Members at Nathorst Fjeld. See Fig. 1 for location and Plate 1 for legend.

Fig. 29. Thin lag conglomerate separating tidal channel sandstones of the Astartekløft Member from offshore transition deposits of the overlying Nathorst Fjeld Member at Nathorst Fjeld. See Fig. 1 for location. The conglomerate consists of well-sorted, well-rounded quartzitic pebbles with a maximum diameter of 3 cm. The sandstone just below the conglomerate is strongly burrowed by *Diplocraterion parallelum*.

and Elis Bjerg the lower boundary is sharp and erosional (Fig. 28). The upper boundary between the sandstones of the Astartekløft Member and the silty mudstones of the laterally correlative Nathorst Fjeld and Harris Fjeld Members is sharp and at Nathorst Fjeld a conglomerate of extraformational pebbles, 3 cm thick, separates the members (Figs 28, 29).

Distribution. The member occurs along the length of Neill Klinter and as far north as Dusén Bjerg.

Fossils. The member has only yielded a few belemnites, oysters and crinoids. Trace fossils include common *Arenicolites* isp. 2 (Dam, 1990a) and rare *Curvolithos multiplex, Diplocraterion habichi, Gyrochorte comosa, Ophiomorpha nodosa*, and *Taenidium serpen-*

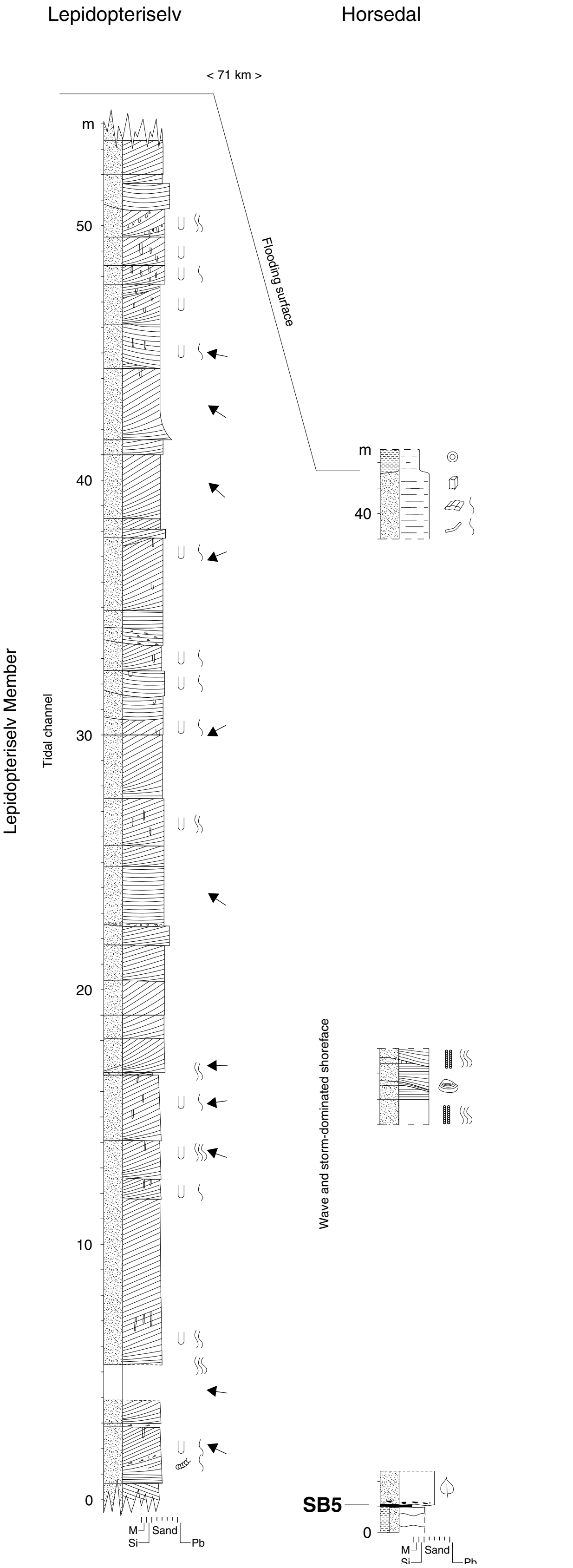


Fig. 30. Sedimentological logs through the Lepidopteriselv Member at Lepidopteriselv and Horsedal. See Fig. 1 for locations and Plate 1 for legend.

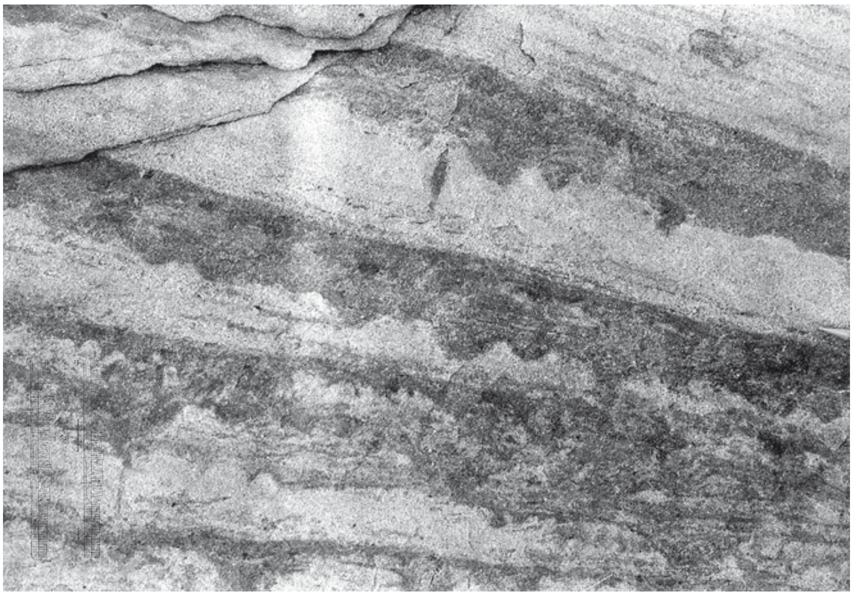


Fig. 31. Resting traces on foresets of cross-bedded sandstones of the tidal channel association (m). Lepidopteriselv Member, Lepidopteriselv. See Fig. 1 for location.

tinum. Resting traces and polychaete burrows also occur.

Geological age. None of the body fossils are age diagnostic. Dinoflagellate cysts from the base of the member suggest an early Toarcian age (Koppelhus & Dam, in press).

Lepidopteriselv Member

new member

History. North of Nathorst Fjeld the Lepidopteriselv Member constitutes the lower third of the Ostreaelv Member of Surlyk *et al.* (1973).

Name. The member is named after the river Lepidopteriselv in north-eastern Jameson Land (Fig. 1).

Type and reference localities. One of the most complete developments of the member is well-exposed at Lepidopteriselv, where it forms an almost vertical cliff section. This is designated the type section. Well-exposed reference sections occur at Liaselv, in Horsedal and Ranunkeldal (Figs 1, 30).

Thickness. The member is 60 m thick at Lepidopteriselv and 48 m at Horsedal.

Lithology. At the type section the Lepidopteriselv Member is cliff-forming and consists of medium to coarse-grained, cross-bedded, glauconitic sandstones. At Elis Bjerg cross-bedded sandstones pass laterally into medium-grained bioturbated sandstones with bivalves, belemnites, ammonites and crinoids. At Liaselv and in Horsedal the cross-bedded sandstones alternate with fossiliferous, clean, bioturbated, cross-bedded and hummocky cross-stratified fine to medium-grained sandstones.

Facies associations and depositional environments. The member includes three facies associations (m, n an d o).

m. Tidal channel association. This association is characteristic of the Lepidopteriselv Member at Lepidopteriselv, and is very similar to the Nathorst Fjeld Member at Nathorst Fjeld and the Trefjord Bjerg Member at Lepidopteriselv. It dominantly consists of a planar and elongate trough cross-bedded medium to coarse-grained glauconitic sandstone unit (Fig. 30), but cross-laminated beds also occur. In the Lepidopteriselv area the thickness of the member varies from 55 to 80 m. The lower surface of the sandstone unit is not exposed, but from the thickness variations of the member in the area it seems that the sandstones are separated from the heterolithic deposits of the Elis Bjerg Member below by a major erosional surface. The upper boundary to the bioturbated muddy sandstones of the Skævdal Member is flat and non-erosional. Set thickness of the cross-bedded sets varies from 0.35 to 6.5 m. Foresets are tangential and dip angles range from 21 to 31°. The foresets dip unimodally towards the west. Thin mudstone drapes may occur in couplets on the foresets, and enclose a thin sandstone layer displaying cross-lamination with a foreset orientation opposite to that of the cross-bedding. Groups of foresets may form bundle sequences (cf. Visser, 1980).

The cross-bedded sandstones are commonly burrowed by *Arenicolites* isp. 2. (Dam, 1990a). Resting traces may occur along foresets of the cross-bedded sandstones, and may have, together with polychaete burrows, completely mottled the bottomsets (Fig. 31). *Arenicolites* isp. 2 lacks a burrow lining and shows a great morphological variability (Dam, 1990a).

A single conglomeratic bed, 2 m thick, occurs on the northern slope of Elis Bjerg. It consists of subrounded quartz pebbles with a diameter of 5 to 30 mm, in a matrix of coarse to very coarse-grained sand. The conglomerate is weakly parallel laminated and cross-bedded.

Interpretation. The cross-bedded sandstones were deposited by extensive fields of migrating small and medium-scale 2-D and 3-D dunes. The presence of mudstone couplets and foreset bundles indicates that deposition took place in a subtidal environment with an ordinate and a subordinate current (Visser, 1980). Foreset dip directions towards the west indicate a basinward flow direction. The lower erosional surface of the cross-bedded sandstone unit and the general offshore directed flow directions suggest deposition in the subtidal parts of a tidal channel. The *Arenicolites* isp. 2 ichnocoenosis represents a low diversity assemblage of organisms that adapted themselves to well-aerated high-energy environments of great physical instability. The lack of burrow linings, great morphological variability and scattered distribution suggest that *Arenicolites* isp. 2 must have acted as shelters for only a short period of time and not as permanent domiciles, indicating a high sediment influx (Dam, 1990a, b).

n. Wave and storm-dominated shoreface association. This association constitutes the Lepidopteriselv Member in the northern part of the basin. It consists of single or amalgamated beds of cross-bedded, parallel laminated, hummocky and swaley cross-stratified very fine to fine-grained sandstones (Fig. 30). The beds have flat, sharp and erosional bases. Abundant transported bivalves and crinoids are common in the laminae, and include *Astarte* sp., *Corbicellopsis unioides, Modiolus (Strimodiolus) elongatus, Pleuromya uralensis* and *Pentacrinus* sp. The sandstones are commonly burrowed by *Ophiomorpha nodosa*. Bioturbated very fine to fine-grained sandstones, 0.1–1 m thick, occur interbedded with the parallel laminated, hummocky and swaley cross-stratified sandstones.

Interpretation. The sharp bases of individual sandstone beds and the dominance of parallel lamination suggest episodic deposition from waning flows characterised by combined flow-processes or intense bed shear due to storm-wave activity. Hummocky cross-stratification is formed by aggradation or translation in a combined or oscillatory flow regime (Dott & Bourgeois, 1982; Swift *et al.*, 1983; Allen, 1985; Surlyk & Noe-Nygaard, 1986). Amalgamation of beds is caused by scouring and erosion above storm-wave base with the removal of the fine-grained, bioturbated caps (Dott & Bourgeois, 1982; Leithold & Bourgeois, 1984). The interbedding with heavily bioturbated, wave ripple laminated sandstones suggest deposition at or just below fair-weather wave-base and above storm wave-base. The transported bivalves include both shallow and deep burrowing forms, the latter indicating that deep erosion frequently occurred on the shoreface during storms (cf. Fürsich, 1982, 1984).

The *Ophiomorpha* ichnocoenosis of the storm-deposited sandstones indicates moderate to high sedimentation rates and a low rate of reworking as shown by the complex clay-ball-lined walls (Heinberg & Birkelund, 1984). The occurrence of only *Ophiomorpha* in amalgamated sandstones indicates physically unstable conditions (i.e. high storm frequency) favouring opportunistic behaviour (Rhoads *et al.*, 1985; Vossler & Pemberton, 1989).

o. Bioturbated shoreface association. This association is characteristic of the Lepidopteriselv Member at Elis Bjerg, Liaselv and in Horsedal. The association consists of bioturbated fine to medium-grained sandstones, locally containing phosphate nodules. It may form thoroughly bioturbated successions, more than 10 m thick, laterally intercalated with tidal channel and storm-dominated inner shelf deposits (Facies associations m and n). Primary physical structures include wave ripples, hummocky cross-stratification, cross-bedding and cross-lamination.

The bioturbated deposits are characterised by the *Curvolithos* ichnocoenosis dominated by *Curvolithos multiplex*, but also include common *Thalassinoides* isp., *Ophiomorpha nodosa, Rhizocorallium irregulare, Gyrochorte comosa, Planolites beverleyensis, Arenicolites* isp. 1 (Dam, 1990a), *Diplocraterion parallelum* and *Palaeophycus* isp. and rare *Cruziana* isp. and *Taenidium serpentinum* (Dam, 1990a, b). Belemnites and bivalves are common.

Interpretation. The high degree of bioturbation and the general scarcity of preserved physical structures indicate a relatively slow sedimentation rate, little physical reworking and abundant food supply. The *Curvolithos* ichnocoenosis reflects a diverse fauna of infaunal and epifaunal suspension and deposit-feeding organisms, as well as carnivores. The trophically diverse fauna lived in a well-aerated environment (Dam, 1990b). Phosphatic concretions probably formed within the anoxic zone just below the sediment-seawater interface during low sedimentation rates (cf. Cook, 1976; O'Brien, 1990).

Boundaries. At the type locality and at Horsedal the sandstones of the Lepidopteriselv Member rest with a sharp, erosional lower boundary on the heterolithic deposits of the Astartekløft and Horsedal Members, respectively. The upper boundary is placed at the sharp lithological change from sandstones of the Lepidopteriselv Member to the muddy bioturbated sandstones of the Skævdal Member.

Fossils. Fossils are common in the member and include bivalves, crinoids, belemnites and ammonites. The ammonites were found at Elis Bjerg and Horsedal and include *Dactylioceras semicelatum* (Simpson) (also including *D. groenlandicum* Rosenkrantz 1934) and *Hildaites* sp. aff. *H. murleyi* (Moxon) (J. H. Callomon, personal communication, 1993). Those collected in Horsedal were found loose.

Distribution. The member occurs north of Nathorst Fjeld and in Ranunkeldal in the north-western part of the basin.

Geological age. The ammonite *Dactylioceras semicelatum* indicates an Early Toarcian, Tenuicostatum Zone, Semicelatum Subzone age of the Lepidopteriselv Member (J. H. Callomon, personal communication, 1993).

Nathorst Fjeld Member

new member

History. The strata composing this member were first described by Rosenkrantz (1934, p. 83, pl. 12). Along Neill Klinter and at Nathorst Fjeld the member represents the middle part of the Ostreaelv Member of Surlyk *et al.* (1973).

Name. The member is named after the mountain Nathorst Fjeld in the south-eastern part of Jameson Land (Fig. 1).

Type and reference localities. The member is well-exposed at Nathorst Fjeld which is designated the type section. Reference sections occur at Moskusoksekløft, Astartekløft and Albuen (Figs 1, 32).

Thickness. The member is 30 m thick at the type section, 35 m at Moskusoksekløft, 37 m in Astartekløft, and 29 m at Albuen.

Lithology. The member forms a single coarsening-upward unit consisting of alternating silty mudstones and thin laminae of very fine to fine-grained sandstones, grading upward into fine to coarse-grained sandstones. The sandstones are trough and planar cross-bedded, wave ripple cross-laminated and cross-bedded, hummocky cross-stratified or bioturbated. Highly disintegrated plant remains and cone-in-cone structures are common in the silty mudstones and very fine to fine-grained sandstones. Thin conglomerates with mud chips, body fossils, quartzite pebbles and plant remains occur at several levels in the fine to coarse-grained sandstones.

Facies associations and depositional environments. The member comprises two facies associations (p and q).

p. Offshore transition association. This association is characteristic of the lower part of the member, and consists of silty mudstones, forming a homogeneous unit up to 30 m thick, with very fine and fine-grained micaceous sandstone streaks (Fig. 32). The streaks show pinch and swell structures, incipient lenses and parallel lamination grading into wave ripple cross-lamination (cf. Facies M_1 of De Raaf *et al.*, 1977). The streaked mudstones locally contain comminuted plant debris, early diagenetic phosphatic concretions and calcareous concretionary layers with cone-in-cone structures. The mudstones are commonly burrowed by *Curvolithos multiplex* and *Planolites beverleyensis*.

Interpretation. The streaked mudstones were deposited from suspension in an offshore transition environment with restricted wave activity. The phosphatic concretions probably formed within the anoxic zone just below the sediment-seawater interface during low sedimentation rates (cf. Cook, 1976; O'Brien *et al.*, 1990).

q. Shoreface association. This association consists of a coarsening-upward succession, 13–24 m thick, present in the upper part of the member (Fig. 32). It has a lateral extent of several tens of kilometres. The lower part of the succession is lenticular and wavy bedded.

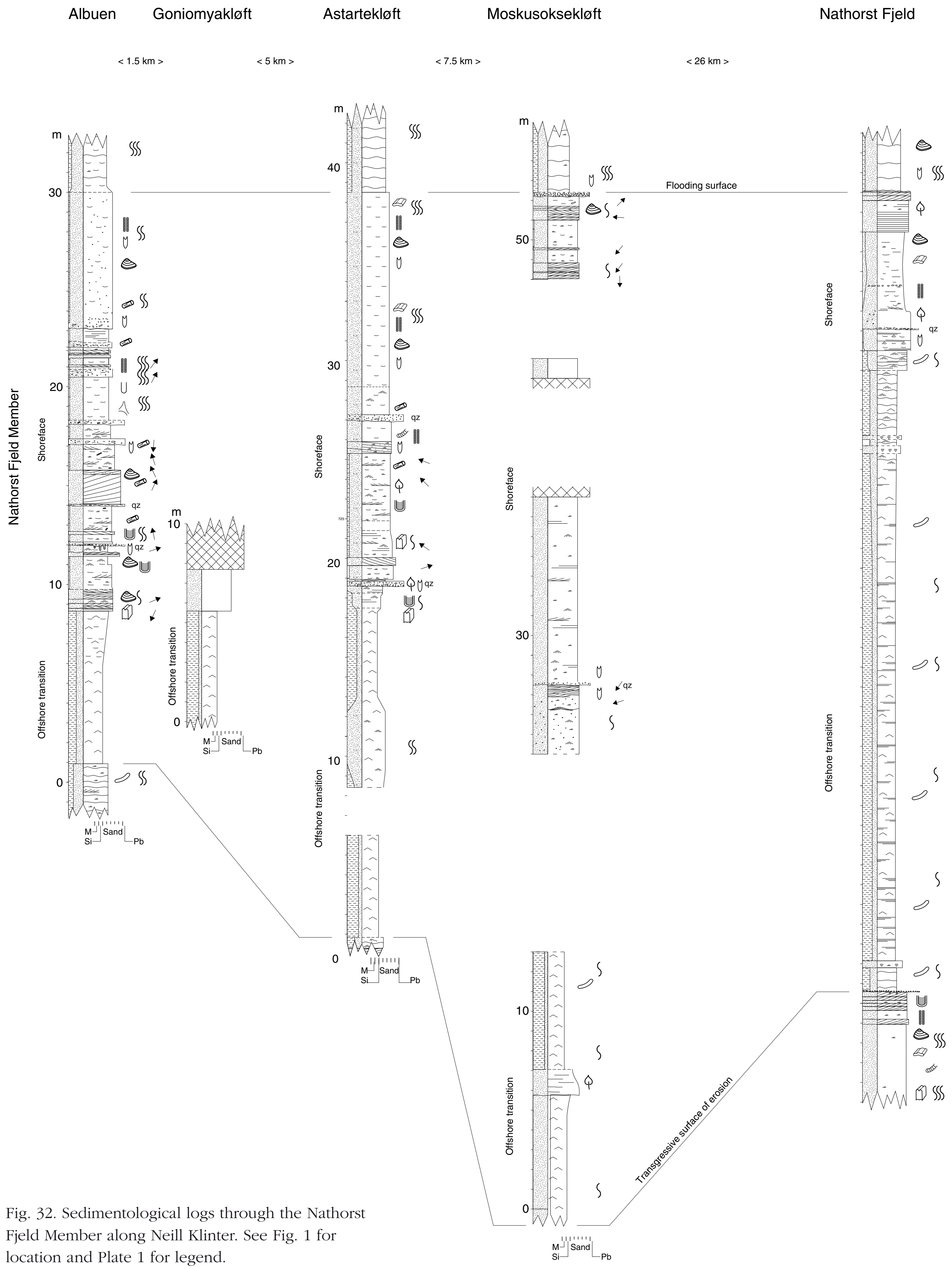


Fig. 32. Sedimentological logs through the Nathorst Fjeld Member along Neill Klinter. See Fig. 1 for location and Plate 1 for legend.

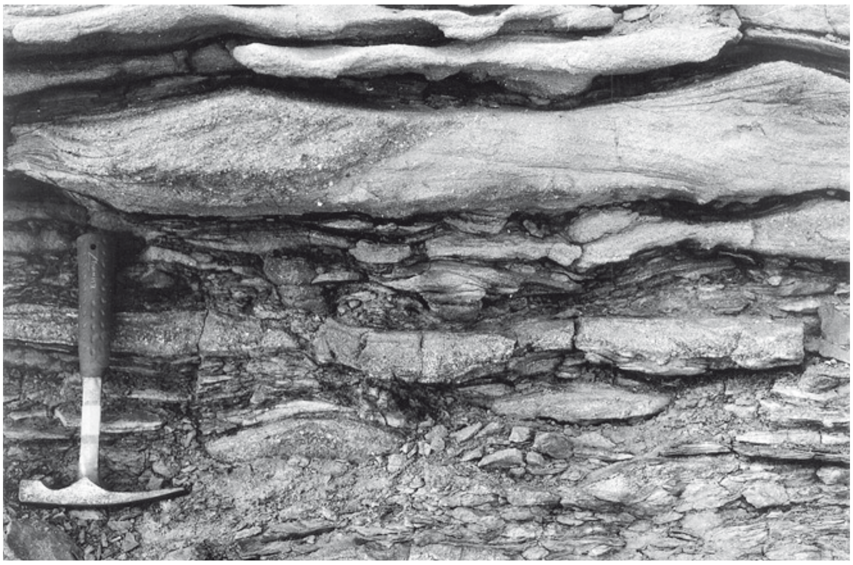


Fig. 33. Coarse-grained symmetrical ripples showing features characteristic of wave-formed ripples of the shoreface association (q) of the Nathorst Fjeld Member. Nathorst Fjeld. See Fig. 1 for location. Hammer 32 cm long.

It is erosionally overlain by poorly-sorted medium to very coarse-grained sandstones comparable to the sandstones of the Rævekløft Formation, and very fine to fine-grained sandstones. Well-rounded quartzitic pebbles, logs, plant remains and transported assemblages of bivalves and belemnites, commonly occur scattered in the medium to very coarse-grained sandstones or form thin laterally extensive conglomerates, less than 20 cm thick. The sandstones are planar and trough cross-bedded, parallel laminated or massive. The cross-bedded sets occur singly or more commonly in cosets. Set thickness varies from 0.1 to 1.5 m. The foreset dip directions are generally towards NNE and SW, but show a large scatter. Cross-bedded medium to coarse-grained sandstones also occur, and show features characteristic of wave action, such as irregular lower boundaries, opposed unidirectional cross-bedded lenses, offshooting and draping laminations, symmetrical ripple form sets (Fig. 33), similar to those of the wave and storm-dominated shoreface deposits of the Elis Bjerg Member (Facies association e). Wave lengths and amplitudes of the symmetrical ripples are 50–60 cm and 6–10 cm, respectively, and dominant crest line orientations are E–W and N–S.

The massive sandstones are strongly bioturbated and occur in beds 0.25–5.5 m thick. The sandstones contain elements of the *Curvolithos* ichnocoenosis including abundant *Curvolithos multiplex, Ophiomorpha nodosa, Taenidum serpentinum, Thalassinoides* isp., common *Arenicolites* isp., *Diplocraterion parallelum*, and rare *Rhizocorallium irregulare* (Dam, 1990b). The top of the sandstones are commonly burrowed by *Skolithos* isp. and *Arenicolites* isp.

The very fine to fine-grained sandstones show parallel lamination and hummocky cross-stratification. The upper part of the sandstones may be glauconitised and a thin lag conglomerate occasionally occurs on top of the succession.

Interpretation. The coarsening-upward succession reflects an increase in energy with time and is interpreted as a shallowing-upward succession. The cross-bedded sandstones are interpreted as representing fields of small-scale 2-D and 3-D dunes migrating longshore towards NNE and SSW on the upper shoreface. The apparent bimodality of the palaeocurrents is the only suggestion of tidal activity and no other tidal features occur. It is thus more likely that the bimodality represents local reversals of the coast-parallel current system. The cross-bedded sandstones showing features characteristic of wave action is interpreted as representing fields of large symmetrical ripples on the shoreface similar to those recorded from modern shoreface and offshore transition settings formed by oscillatory or oscillatory-dominant flows (cf. Newton & Werner, 1972; Cacchione *et al.*, 1984; Hunter *et al.*, 1988; Leckie, 1988). Hummocky cross-stratification formed during storms by aggradation or translation in a combined or oscillatory flow regime (Dott & Bourgeois, 1982; Swift *et al.*, 1983; Allen, 1985; Surlyk & Noe-Nygaard, 1986). The thin conglomerates interbedded with the sandstones were probably transported during storms or winnowing during periods of non-deposition.

The *Curvolithos* ichnocoenosis reflects prolonged periods of well-aerated conditions, intermediate energy and slow deposition (Dam, 1990b).

Fossils. Several thin, fossiliferous levels occur in the upper part of the member and contain bivalves, brachiopods, ammonites, belemnites, crinoids and vertebrates. Fossils collected at Nathorst Fjeld were listed by Rosenkrantz (1934, p. 82–84, 494–511 m) and include the belemnite *Megateuthis* sp. and the ammonite *Dactylioceras* sp. Trace fossils include common occurrences of *Arenicolites* isp., *Curvolithos multiplex, Diplocraterion parallelum, Gyrochorte comosa, Ophiomorpha nodosa, Planolites beverleyensis, Rhizocorallium irregulare, Taenidium serpentinum* and *Thalassinoides* isp.

Boundaries. The lower boundary is sharp and is marked by the change from the sandstones of the Astartekløft Member to the alternating mudstones and thinly laminated sandstones of the Nathorst Fjeld Member. At Nathorst Fjeld the boundary is marked by a very thin pebble conglomerate (Fig. 29). The upper boundary is also sharp and is marked by a change from the sandstones of the Nathorst Fjeld Member to the bioturbated sandy mudstones of the Skævdal Member.

Distribution. The member occurs along the length of Neill Klinter and at Nathorst Fjeld.

Geological age. At his locality 2 at Nathorst Fjeld, Rosenkrantz (1934) collected a specimen of the belemnite *Parapassolotheuthis polita* (Simpson, 1866) at an altitude of 494 m, and of '*Parabrachybelus*' *subaduncatus* (Voltz, 1930) at 509 m (Doyle, 1991).

Although there is a discrepancy in altitude the lithological descriptions of Rosenkrantz (1934), suggest that the level with *P. polita* belongs to the upper part of the Nathorst Fjeld Member and the level with *'P.' subaduncatus* to the overlying Skævdal Member. The two species have restricted ranges and are not known to be widespread in Europe. *P. polita* is only recorded from the lower Toarcian latest Falciferum Zone or lowermost Bifrons Zone (Commune Subzone) in Britain, while *'P.' subaduncatus*, so far recorded only from mainland Europe, has a range probably restricted to the uppermost Toarcian Levesquei Zone (Doyle, 1991).

The ammonite *Dactylioceras semicelatum* (Simpson) has been collected at the base of the Skævdal Member at Nathorst Fjeld (C. Bjerrum, personal communication, 1996) and in the Lepidopteriselv Member on top of Elis Bjerg indicating an Early Toarcian Tenuicostatum Zone, Semicelatum Subzone age (J. H. Callomon, personal communication, 1993). Sequence stratigraphically the Nathorst Fjeld Member is correlated with the Lepidopteriselv Member. This discrepancy in the age of the two members may indicate that *P. polita* cannot be used stratigraphically, or that the *D. semicelatum* at the base of the Skævdal Member is reworked. Palynomorphs suggest an Early Toarcian age of the Nathorst Fjeld Member (Koppelhus & Dam, in press).

Harris Fjeld Member

new member

History. Strata referred to the Harris Fjeld Member were included in the Ostreaelv Member by Surlyk *et al*. (1973). Deposits from the unit were figured in Sykes (1974, pl. 4, fig. 1).

Name. The member is named after the mountain Harris Fjeld at Hurry Inlet, south-eastern Jameson Land (Fig. 1).

Type locality. The member is only known from Harris Fjeld (Fig. 1).

Thickness. The member is 34–40 m thick.

Lithology. The Harris Fjeld Member consists of tabular, heterolithic or sandy clinoform beds, up to 18 m thick, that laterally pass into silty shales and cross-bedded sandstones.

Facies associations and depositional environments. The member is mainly made up of the ebb-tidal delta Facies association (r). However, minor occurrences of subtidal sand sheet (Facies association f) and offshore transition deposits (Facies association p) also occur.

r. Ebb-tidal delta association. This association includes the largest sedimentary structures identified in the Neill Klinter Group, and is characteristic of the Harris Fjeld Member. The association consists of tabular, low-angle clinoform beds, 6–21 m thick (Figs 34–37). The clinoforms are tangential and dip 5–15° basinwards towards west. The beds can be followed for 2 km along strike and pass laterally into offshore transition mudstones or shoreface sandstones of the Nathorst Fjeld Member (Facies associations p and q). The clinoform beds gradually overlie silty shales or cross-bedded sandstones of the subtidal sand sheet association (Facies association f). The upper boundary is sharp and erosional and is succeeded by restricted shelf mudstones of the Skævdal Member (Figs 34–37).

The clinoform beds are either heterolithic (Fig. 35)

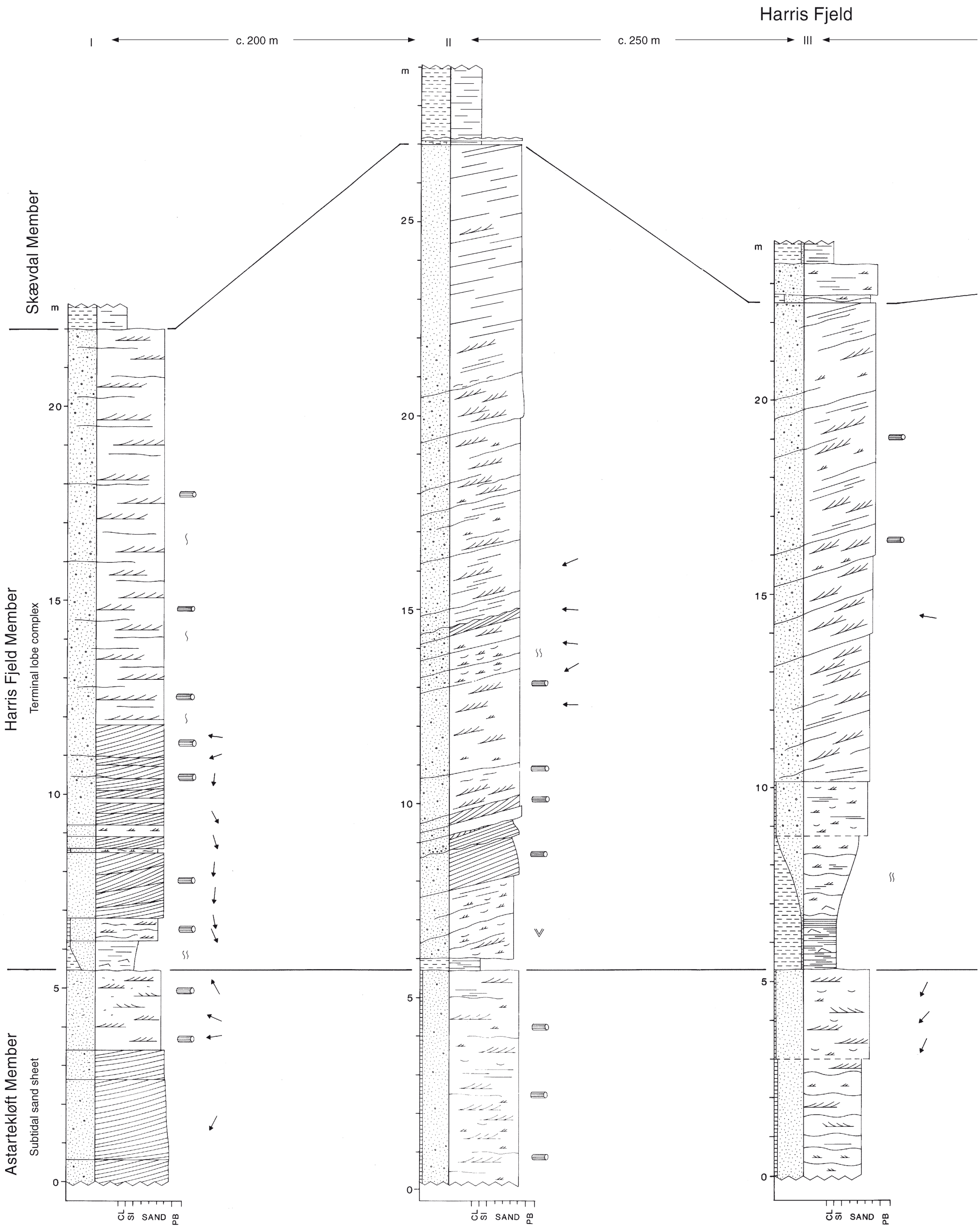
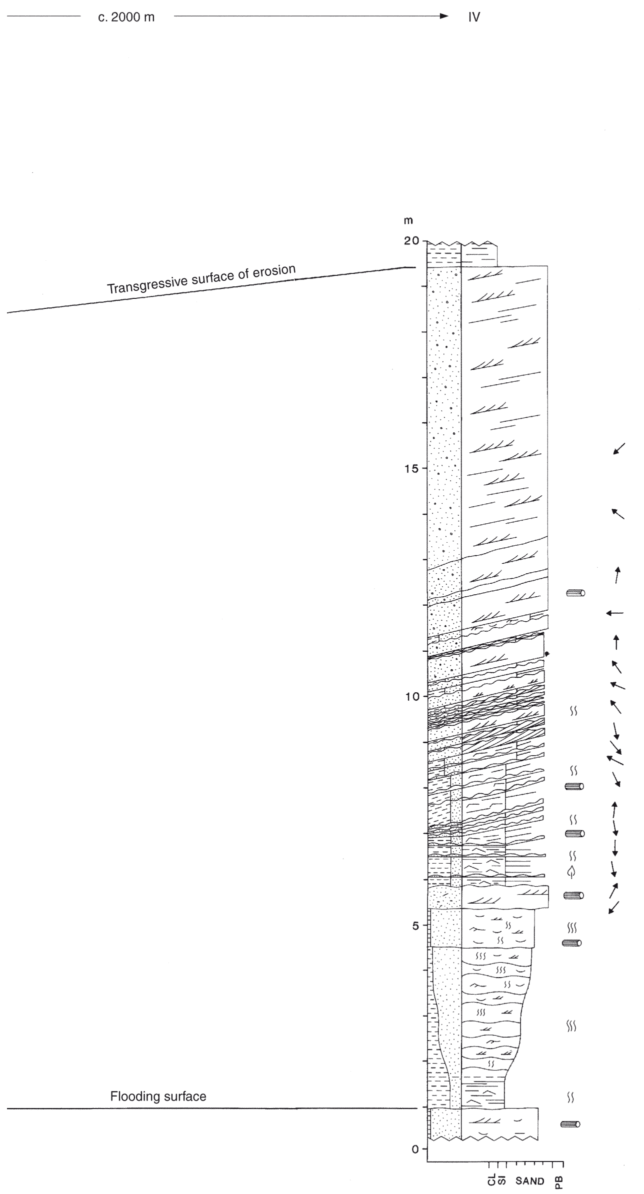


Fig. 34. Sedimentological logs through the Harris Fjeld Member at Harris Fjeld. See Fig. 1 for location and Plate 1 for legend.



or sandy throughout (Figs 36, 37). The heterolithic beds consist of alternating mudstone layers and planar and trough cross-bedded sandstone intrasets. The intrasets are 5–35 cm thick. Mudstone drapes and mudstone intraclasts are common on the foresets of the intrasets. The thickness of the intrasets decreases down-dip, whereas the thickness of mudstone layers increases, resulting in the overall coarsening-upward pattern seen in vertical sections. The foreset dip direction of the intrasets in the upper part of the heterolithic sets is parallel to the dip direction of the clinoforms. Along the toe of the clinoforms the dip directions of the intrasets are perpendicular to the dip direction of the clinoforms. Trace fossils are rare and only occur along the toe of the clinoforms, but include elements of the *Cochlichnus* ichnocoenosis (Dam, 1990b).

The sandy clinoform beds consist of alternations of coarser and finer structureless beds, 5–25 cm thick. Continuous mudstone laminae and intrasets commonly occur. Simple avalanche foresets showing inverse grading have not been recorded. A lateral transition from the sandy to the heterolithic clinoform beds occurs from south to north at Harris Fjeld (Fig. 34).

Interpretation. The limited lateral extent, the association with tidal channel deposits and the basinward dip directions suggest that the clinoforms were formed by progradation of ebb-tidal deltas onto the shelf in front of tidal channels. The structures are very similar to those documented from the Lower Eocene Roda Sandstone in the southern Pyrenees, and those of the present-day tidal and ebb-delta environment of the Eastern Scheldt mesotidal system, south-west Netherlands (Yang & Nio, 1989; Crumeyrolle *et al.*, 1993). The limited extent of the sets in 2-D strike sections indicates that the lobes were 2–10 km wide. The systematic change in palaeocurrent directions of the small 2-D and 3-D dunes and ripples, which form the intrasets, shows that they migrated with an increasingly oblique angle down the surface of the ebb-tidal delta. The association with tidal channel and subtidal sand sheet deposits suggest that the intrasets formed from ebb-tidal currents. The mudstones separating the intrasets may be the result of tidal slack-water periods.

The sharp upper erosional surface of the clinoform beds, overlain by marine mudstones, was formed by shoreface erosion during an ensuing transgression.

Fossils. A few fragmented belemnites and bivalves occur. Rare trace fossils include *Diplocraterion parallelum, Planolites beverleyensis* and *Teichichnus* isp. (Dam, 1990b).

Boundaries. The upper boundary is sharp and erosional and is overlain by mudstones of the Skævdal Member (Figs 34–37). The lower boundary is more diffuse, but is placed at the base of the first thin coarsening-upward succession grading into the clinoform beds.

Distribution. The member only occurs at Harris Fjeld.

Geological age. The member is considered to be time-equivalent to the Nathorst Fjeld and Lepidopteriselv Members, suggesting an Early Toarcian age.

Fig. 35. Heterolithic clinoform bed (Facies association r) of an ebb-tidal delta of the Harris Fjeld Member. The complex is composed of two smaller and one larger coarsening-upward successions. The larger succession shows low-angle clinoform bedding with dips less than 15°. Along the clinoforms, mudstone layers alternate with intrasets of cross-bedded sandstones with mudstone drapes and mudstone clasts. TSE - Transgressive surface of erosion. Harris Fjeld. See Fig. 1 for location. From Dam & Surlyk (1995).

Fig. 36. Sandy clinoform bed (Facies association r) of an ebb-tide delta of the Harris Fjeld Member at Harris Fjeld. See Fig. 1 for location. Set is 21 m thick. TSE - Transgressive surface of erosion.

Fig. 37. Sandy clinoforms below an erosional transgressive surface (TSE) of the Harris Fjeld Member. Note the upper erosional boundary of the ebb-tidal delta (Facies association r), succeeded by mudstones. Harris Fjeld. See Fig. 1 for location.

Skævdal Member

new member

History. Strata now referred to the Skævdal Member were previously included in the Ostreaelv Member of Surlyk *et al.* (1973).

Name. The member is named after the ravine Skævdal in Neill Klinter, in the south-eastern part of Jameson Land (Fig. 1).

Type and reference localities. A well-exposed section in the cliff at Skævdal is designated the type section. Well-exposed reference sections occur at Nathorst Fjeld and Lepidopteriselv (Figs 1, 38).

Thickness. The member is 31 m thick at the type section, 4 m at Harris Fjeld, and 12 m at Nathorst Fjeld.

Lithology. The member consists of bioturbated and fossiliferous sandy mudstones and muddy very fine to fine-grained sandstones, locally containing phosphate nodules (Figs 39, 40). Primary physical structures have not been observed in the muddy sandstones. Wave ripples, hummocky cross-stratification, cross-bedding and cross-lamination are locally present in the fine to medium-grained sandstones. Ooids occur scattered in the mudstones at Horsedal.

Facies association and depositional environment. The member is entirely made up of bioturbated shoreface deposits (Facies association o).

o. Bioturbated shoreface association. This association is characteristic of the Skævdal Member. The association consists of bioturbated sandy mudstones and muddy sandstones, locally containing phosphate nodules. It forms a thoroughly bioturbated basin-wide succession, up to 31 m thick. The bioturbated deposits are characterised by the *Curvolithos* ichnocoenosis dominated by *Curvolithos multiplex*, but also include common *Thalassinoides* isp., *Ophiomorpha nodosa, Rhizo-*

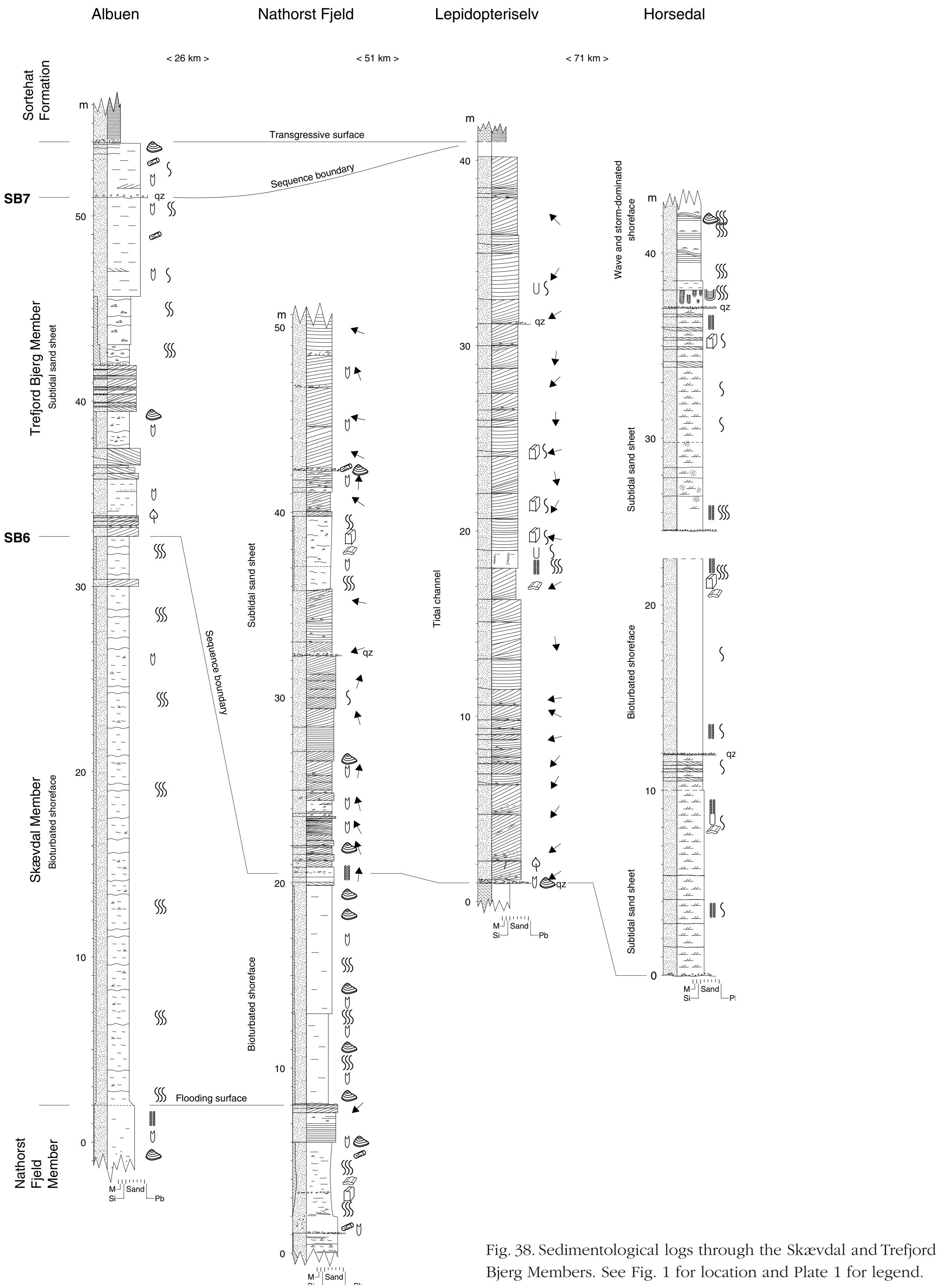


Fig. 38. Sedimentological logs through the Skævdal and Trefjord Bjerg Members. See Fig. 1 for location and Plate 1 for legend.

corallium irregulare, Gyrochorte comosa, Planolites beverleyensis, Arenicolites isp. 1 (Dam, 1990a), *Diplocraterion parallelum, Palaeophycus* isp. and rare *Cruziana* isp. and *Taenidium serpentinum* (Dam, 1990a, b). Belemnites and bivalves are common.

Along Neill Klinter, vertical alternations in mud content in the heavily bioturbated muddy sandstones of the Skævdal Member suggest that the sediments were originally deposited as heteroliths (Figs 39, 40).

Interpretation. The high degree of bioturbation and the lack of physical structures indicate a relatively slow sedimentation rate, little physical reworking and abundant food supply. The *Curvolithos* ichnocoenosis reflects a diverse fauna of infaunal and epifaunal suspension and deposit-feeding organisms, as well as carnivores. The trophically diverse fauna lived in a well-aerated environment (Dam, 1990b). Phosphatic concretions probably formed within the anoxic zone just below the sediment-seawater interface during low sedimentation rates (cf. Cook, 1976; O'Brien *et al.*, 1990).

Fossils. Fossils occur at several levels and comprise belemnites (including *'Parabrachybelus' subaduncatus*), bivalves, brachiopods and rare ammonites. Trace fossils include abundant *Curvolithos multiplex*, common *Thalassinoides* isp., *Ophiomorpha nodosa, Rhizocorallium irregulare, Gyrochorte comosa, Planolites beverleyensis, Arenicolites* isp., *Diplocraterion parallelum, Palaeophycus* isp., *Phoebichnus trochoides* and rare *Cruziana* isp. and *Taenidium serpentinum*.

Fig. 39. Bioturbated heterolithic deposits (Facies association o) of the Skævdal Member at Albuen, showing a slight coarsening-upward trend. The original heterolithic nature of the deposits can still be recognized in spite of the complete bioturbation of the sediments. See Fig. 1 for location.

Fig. 40. Detail of the bioturbated sediments (Facies association o) of the Skævdal Member at Albuen. See Fig. 1 for location. Pen 14 cm long.

Fig. 41. Erosional unconformity separating bioturbated sandy shoreface mudstones (Facies association o) of the Skævdal Member from overlying cross-bedded tidal channel sandstones (Facies association m) of the Trefjord Bjerg Member. Lepidopteriselv. See Fig. 1 for location. Persons for scale. From Dam & Surlyk (1995).

Fig. 42. Erosional unconformity separating bioturbated sandy shoreface deposits (Facies association o) of the Skævdal Member from the overlying cross-bedded sandstones (Facies association m) of the Trefjord Bjerg Member. Notice the upward thinning of cross-bedded sets. SB6 – Sequence boundary 6. Lepidopteriselv. See Fig. 1 for location.

Boundaries. The lower boundary is sharp and in places erosional and placed where the muddy sandstones and sandy mudstones of the Skævdal Member overlie the sandstones of the Nathorst Fjeld, Harris Fjeld and Lepidopteriselv Members. The upper boundary to the sandstones of the Trefjord Bjerg Member is sharp and erosional at Lepidopteriselv (Figs 38, 41, 42) and Horsedal and sharp, but non-erosional along Neill Klinter.

Distribution. The distribution of the Skævdal Member is the same as for the Neill Klinter Group.

Geological age. The presence of the belemnite *'Parabrachybelus' subaduncatus* suggests a latest Toarcian Levesquei Zone age for the Skævdal Member, whereas the ammonite *D. semicelatum* found at the base of the member (C. Bjerrum, personal communication, 1996) indicates an Early Toarcian Tenuicostatum Zone, Semicelatum Subzone age. This suggests that either the ammonite is reworked or the Skævdal Member has a very long age range (see discussion earlier). Palynomorphs suggest a Late Toarcian – Early Aalenian age (Koppelhus & Dam, in press).

Fig. 43. Bored siderite clasts from the basal conglomerate of the Trefjord Bjerg Member at Lepidopteriselv. See Fig. 1 for location. Scale 1 cm on rock (encircled).

Trefjord Bjerg Member

new member

History. Strata now referred to the Trefjord Bjerg Member include the uppermost part of the Ostreaelv Member of Surlyk *et al.* (1973).

Name. The member is named after Trefjord Bjerg, west of the head of Carlsberg Fjord (Fig. 1).

Type and reference localities. One of the most well-exposed and complete developments of the member occurs at Lepidopteriselv, along the southern slope of Trefjord Bjerg; this is designated the type section. Well-exposed reference sections occur along Neill Klinter at Albuen, at Harris Fjeld and Nathorst Fjeld and at Ostreaelv and Horsedal.

Thickness. The member is *c.* 40 m thick at the type section, 30 m at Ostreaelv, 21 m at Albuen, 45 m at Harris Fjeld, and at least 43 m at Horsedal.

Lithology. At Ostreaelv, along Neill Klinter, and at Nathorst Fjeld the member is dominated by fine to very coarse-grained sandstones. Many sandstones are cross-bedded and rhythmical clay drapes are common on the foresets. Concretionary cement has hardened some beds. At the type section the member consists of medium to very coarse-grained, concretionary, cross-bedded sandstones, occasionally with logs, in places with interbedded bioturbated sandstones. From Lepidopteriselv towards west to Liaselv and towards south the cross-bedded sandstones pass laterally into medium-grained bioturbated sandstones. At Horsedal the member is composed of cross-bedded, cross-laminated, parallel-laminated and hummocky cross-stratified fine to medium-grained sandstones alternating with bioturbated sandstones. At this locality thin conglomerates, composed of discoidal quartzite pebbles, occur at several levels.

Facies associations and depositional environments. The member is made up of three facies associations (m, n and o).

m. Tidal channel association. This association is characteristic of the member at the type locality, where it is cliff-forming. It consists dominantly of planar and elongate trough cross-bedded medium to very coarse-grained glauconitic sandstones (Figs 38, 41, 42), and less commonly of parallel and cross-laminated sandstones. Thin mudstone drapes may occur in couplets on the foresets (cf. Visser, 1980), and flatten out along

Fig. 44. Interbedded bioturbated (I) and parallel-laminated and hummocky cross-stratified (II) very fine to fine-grained sandstones of the wave and storm-dominated shoreface association (n). Trefjord Bjerg Member, Horsedal. See Fig. 1 for location. Hammer 32 cm long.

the toesets, where they extend as horizontal layers for several metres. Mudstone drapes may extend to the top of the foresets or pinch out on the mid-foreset slope due to truncation by reactivation surfaces. The coupled mudstone drapes enclose a thin sand layer displaying cross-lamination with a foreset orientation opposite to that of the cross-bedding. Groups of foresets may display a repetitive lateral thickening and thin- ning, similar to the bundle sequences of Visser (1980). Set thickness varies from 0.2 to 3.5 m with 0.5 to 2 m being the most common. The troughs commonly exceed 15 m in width. Foresets are tangential and dip angles range from 14° to 31°. Logs, pebbles, and transported assemblages of belemnites, bivalves and occasional crinoids occur in the basal part of the sets. The cross-bedded sets form a coset, 40 m thick, with a general thinning upward in set thickness (Figs 38, 42). The lower surface of the coset is erosional and overlain by a thin lag conglomerate of belemnites, bivalves, well-rounded quartzitic pebbles less than 2 cm in diameter, and bored siderite clasts (Fig. 43). The foresets dip unimodally towards SW. Sets are in places separated by bioturbated sandstones. The bioturbated sandstones include *Arenicolites* isp. 2, *Curvolithos multiplex, Diplocraterion habichi, Gyrochorte comosa, Ophiomorpha nodosa* and *Rhizocorallium* isp. (Dam, 1990a). The top of the sandstone unit at several localities at Lepi-dopteriselv is heavily bioturbated by *Diplocraterion habichi,* and loose blocks also suggest *Phycodes bromleyi*.

Interpretation. The cross-bedded sandstones were deposited by extensive fields of migrating small and medium-scale 2-D and 3-D dunes. The presence of mudstone couplets and foreset bundles indicates that deposition took place in a subtidal environment with ordinate and subordinate currents (Visser, 1980). The lower erosional boundary of the cross-bedded cosets, the general offshore directed currents and the upward thinning of set thickness suggest deposition in the ebb-dominated subtidal part of a tidal channel. The trace fossil assemblage represents a low diversity assemblage of organisms that adapted themselves to well-aerated high-energy environments of great physical instability. The lack of burrow linings, great morphological vari- ability and scattered distribution suggest that *Arenico-lites* isp. 2 must have acted as shelters for only a short period of time and not as permanent domiciles, indicating a high sediment influx (Dam, 1990a, b). The complex clay-ball-lined walls of *Ophiomorpha nodosa* indicate moderate to high sediment influx and a low rate of reworking (Heinberg & Birkelund, 1984).

n. Wave and storm-dominated shoreface association. This association constitutes the Trefjord Bjerg Member in the northern part of the basin. It consists of single or amalgamated beds of parallel laminated, hummocky and swaley cross-stratified, very fine to fine-grained sandstones. The beds have flat, sharp and erosional bases, occasionally succeeded by thin lag conglomerates consisting of well-sorted, well-rounded, oval quartzitic pebbles and cobbles. Parallel lamination with parting lineation is commonly the main, or only, sedimentary structure (Fig. 44). Abundant transported bivalves are common in the laminae, and include *Astarte* sp., *Corbicellopsis unioides, Modiolus (Strimo-diolus) elongatus and Pleuromya uralensis.* Wave ripple cross-laminated sandstones and symmetrical ripples usually cap the beds. A maximum depositional depth for the wave ripple laminated beds of approximately 13 m is estimated on the basis of wave ripple features, using the method of Diem (1985). Bioturbated very fine to fine-grained sandstones, 0.1–1 m thick, occur interbedded with the parallel-laminated, hummocky and swaley cross-stratified sandstones. The beds are thoroughly bioturbated and characterised by the *Rhizocorallium* and *Phoebichnus* ichnocoenoses of Dam (1990b). The *Rhizocorallium* ichnocoenosis consists primarily of *Rhizocorallium irregulare, Gyrochorte comosa, Parahaentzschelinia surlyki, Curvolithos multiplex, Nereites* isp., and rare *Scolicia* isp., *Gyrophyllites kwassicensis* and trackways. The *Phoebichnus* ichnocoenosis is monospecific, only including the complex burrow *Phoebichnus trochoides*, occurring in very large numbers at certain levels.

Interpretation. The sharp bases of individual sandstone beds and the dominance of parallel lamintion suggest episodic deposition from waning flows characterised by combined flow-processes or intense bed shear due to storm-wave activity. Hummocky cross-stratification is formed by aggradation or translation in a combined or oscillatory flow regime (Dott & Bourgeois, 1982; Swift *et al.*, 1983; Allen, 1985; Surlyk & Noe-Nygaard, 1986). Amalgamation of beds is caused by scouring and erosion above storm-wave base with the removal of the fine-grained, bioturbated caps (Dott & Bourgeois, 1982; Leithold & Bourgeois, 1984). The interbedding with heavily bioturbated, wave ripple laminated sandstones suggest deposition at or just below fair-weather wave-base and above storm wave-base. The transported bivalves include both shallow and deep burrowing forms (cf. Fürsich, 1982, 1984). The latter indicating that deep erosion frequently occurred on the shoreface during storms.

The *Rhizocorallium* ichnocoenosis represents the activity of both deposit-feeding organisms and carnivores (Dam, 1990a, b). The ichnocoenosis is present in horizons that represent well-aerated periods of relative low-energy conditions. The *Phoebichnus* ichnocoenosis reflects a monospecific fauna of exclusively deposit-feeding organisms and is indicative of very quiet conditions with deposition of organic-rich material (Heinberg & Birkelund, 1984). This type of stationary fodinichnial ichnocoenosis probably exemplifies an oxygen-limited environment that was exploited thoroughly by a population of opportunistic organisms (cf. Ekdale & Mason, 1988).

o. Bioturbated shoreface association. This association consists of bioturbated fine to medium-grained sandstones, locally containing phosphate nodules. It forms thoroughly bioturbated successions, up to 55 m thick, laterally intercalated with tidal channel and wave and storm-dominated shoreface deposits (Facies associations m and n). Primary physical structures are rare and include wave ripples, hummocky cross-stratification, cross-bedding and cross-lamination. The bioturbated deposits are characterised by the *Curvolithos* ichnocoenosis dominated by *Curvolithos multiplex*, but also include common *Thalassinoides* isp., *Ophiomorpha nodosa, Rhizocorallium irregulare, Gyrochorte comosa, Planolites beverleyensis, Arenicolites* isp. 1, *Diplocraterion parallelum, Palaeophycus* isp., rare *Cruziana* isp. and *Taenidium serpentinum* (Dam, 1990a, b). Belemnites and bivalves are common.

Interpretation. The high degree of bioturbation and the general scarcity of preserved physical structures indicate a relatively low sedimentation rate, little physical reworking and abundant food supply. The *Curvolithos* ichnocoenosis reflects a diverse fauna of infaunal and epifaunal suspension and deposit-feeding organisms, as well as carnivores. The trophically diverse fauna lived in a well-aerated environment (Dam, 1990b).

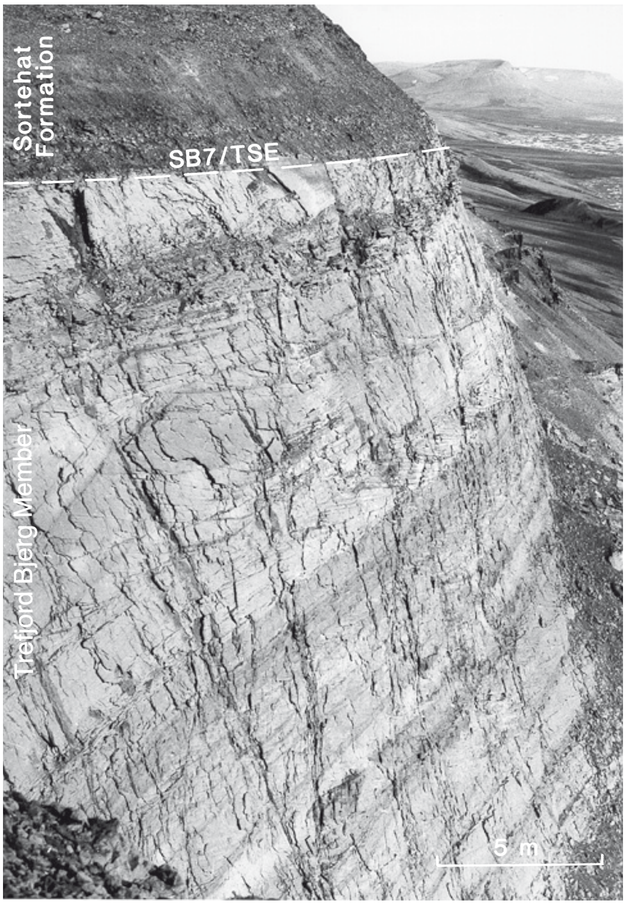


Fig. 45. Cross-bedded sandstones of the Trefjord Bjerg Member overlain by black mudstones of the Sortehat Formation. The boundary is interpreted as a coalesced sequence boundary and transgressive surface. SB7 - Sequence boundary 7, TSE - Transgressive surface of erosion. North of Astartekløft. See Fig. 1 for location. From Dam & Surlyk (1995).

Phosphatic concretions probably formed within the anoxic zone just below the sediment-seawater interface during low sedimentation rates (cf. Cook, 1976; O'Brien *et al.*, 1990).

Fossils. Fossils are common at several levels and include bivalves, belemnites and ammonites. Trace fossils include abundant *Curvolithos multiplex, Ophiomorpha nodosa, Phoebichnus trochoides, Taenidum serpentinum, Thalassinoides* isp., common *Arenicolites* isp., *Diplocraterion habichi, Diplocraterion parallelum, Rhizocorallium irregulare, Thalassinoides* isp., *Gyrochorte comosa, Nereites* isp. *Parahaentzschelinia surlyki, Planolites beverleyensis* and rare *Cruziana* isp., *Gyrophyllites kwassicensis, Scolicia* isp., and trackways.

Boundaries. At Ostreaelv, Trefjord Bjerg and Horsedal, the coarse-grained sandstones of the Trefjord Bjerg Member rest with a sharp, erosional boundary on the muddy sandstones of the Skævdal Member. The boundary is overlain by a lag of bored siderite clasts, extraformational pebbles, logs and belemnites (Figs 41–43). Along Hurry Inlet there is a sharp non-erosional contact to the underlying Skævdal Member. The upper boundary is placed at the sharp change from the sandstones of the Trefjord Bjerg Member to the overlying dark silty mudstones of the Sortehat Formation (Fig. 45).

Distribution. The distribution of the Trefjord Bjerg Member is the same as for the Neill Klinter Group.

Geological age. None of the fossils found in the Trefjord Bjerg Member are age diagnostic. The presence of the belemnite *'Parabrachybelus' subaduncatus* in the underlying Skævdal Member, suggests an age no older than the uppermost Toarcian Levesquei Zone. Palynomorphs suggest a Late Toarcian – earliest Aalenian age (Koppelhus & Dam, in press).

Sortehat Formation

redefined

History. The strata composing the Sortehat Formation were originally defined as the basal member of the Middle Jurassic Vardekløft Formation (now Vardekløft Group) by Surlyk *et al.* (1973). They were recently excluded from this formation and first established informally as a member of the Neill Klinter Formation, later as a separate formation (Surlyk, 1990a, b, 1991). It is here included in the Neill Klinter Group. A detailed description is presented in Koppelhus & Hansen (in press).

Name. From the mountain Sortehat by Ugleelv at the head of Hurry Inlet (Fig. 1).

Type and reference localities. One of the most complete and well-exposed sections occurs at Sortehat which is designated the type section. Well-exposed reference sections occur at Albuen, Goniomyakløft, Trefjord Bjerg, Liaselv, Pelion, Coloradodal and Pingel Dal (Fig. 1).

Thickness. The thickness varies between 60 and 120 m (Surlyk *et al.*, 1973). At the type locality it is *c.* 100 m thick (Krabbe *et al.*, 1994).

Lithology. The member is very uniform throughout the exposed parts of the basin. The lower part consists of dark-grey to black mudstone with concretionary layers and nodules of claret-coloured ironstone. In the upper part the mudstones become more silty, the ironstone decreases or disappears and lenses or layers of fine sandstone, partly concretionary, appear in the succession. The upper part may contain layers of calcareous concretions. Fossils include bivalves, mainly oyster fragments, and belemnites which commonly occur in the top of the formation. Plant fragments are common. Part of the mudstones and fine sandstones are burrowed by *Curvolithos multiplex* and *Planolites beverleyensis* (Surlyk *et al.*, 1973; Krabbe *et al.*, 1994).

Facies association and depositional environment. One facies association is recognised in the Sortehat Formation (s).

s. Restricted embayment association. This association consists of silty mudstone with very fine and fine-grained micaceous sandstone streaks. The sandstone streaks show pinch and swell structures, incipient lenses and parallel lamination grading into wave ripple cross-lamination. The streaked mudstones locally contain highly disintegrated plant debris and wood fragments, early diagenetic phosphate concretions and calcareous concretionary layers. Fossils include bivalves, mainly oyster fragments, and belemnites which commonly occur in the top of the formation. At Albuen a large number of inarticulate brachiopods occur in the basal part of the formation. The mudstones form one or two coarsening-upward successions, and are commonly burrowed by *Curvolithos multiplex* and *Planolites beverleyensis* (Surlyk *et al.*, 1973; Krabbe *et al.*, 1994).

Interpretation. The mudstones were deposited in an offshore environment, probably in a relatively deep embayment with restricted wave energy. The very fine and fine-grained sandstones were deposited as tempestites during periodic storms in water depths below fair-weather wave base but above storm-wave base. The coarsening-upward successions are shallowing-upward cycles with offshore deposits overlain by lower shoreface deposits. The sands were probably derived from coastal erosion and transported by downwelling flows during storms (cf. Snedden *et al.*, 1988). Both macrofossils, organic geochemistry (Krabbe *et al.*, 1994) and palynomorphs (Koppelhus & Hansen, in press) suggest an increase of salinity in the depositional environment from the base to the top of the formation.

Fossils. Fossils include bivalves, mainly oyster fragments, and belemnites which commonly occur in the top of the formation. The sediments are commonly burrowed by *Curvolithos multiplex* and *Planolites beverleyensis.*

Boundaries. The boundary between the Sortehat and Ostreaelv Formations is one of the most conspicuous lithological boundaries in the Mesozoic succession of East Greenland. Throughout the exposed part of the basin the contact between the sandstones of the Neill Klinter Group and the dark mudstones of the Sortehat Formation is flat and very sharp and in places pebble strewn. The Sortehat Formation is overlain by the sandy Vardekløft Group with a sharp, unconformable contact (Surlyk *et al.*, 1973; Engkilde & Surlyk, in press).

Distribution. Same as the group.

Geological age. No age diagnostic macrofossils have been found. Palynomorphs suggest an Aalenian – Early Bajocian age of the formation (Underhill & Partington, 1994; Koppelhus & Hansen, in press).

Palaeogeographic evolution

The shallow marine sediments of the Pliensbachian – Early Bajocian Neill Klinter Group were deposited in a more than 200 km long and 100 km wide fault-bounded embayment. Deposition was influenced by tidal and storm wave currents.

The embayment was restricted to the same depositional basin as the underlying Rhaetian–Sinemurian lake complex of the Kap Stewart Group (Surlyk *et al.*, 1981; Surlyk, 1990b; Dam, 1991; Dam & Surlyk, 1990, 1993). In the central, northern, and north-western parts of the basin the change from lacustrine to shallow marine conditions resulted in a succession in which lacustrine mudstones, with clear non-marine indicators, coarsen-upward into fine-grained sandstones with marine body and trace fossils, associated with a marked increase of the ratio of total sulphur to total organic carbon (Dam & Christiansen, 1990; Dam & Surlyk, 1993). This coarsening-upward succession is remarkable in that it reflects a shallowing-upward tendency at the same time as it marks a change from freshwater to marine conditions. This suggests that the lacustrine basin and probably much of the rift complex between Greenland and Norway was low-lying and that the basin was separated from the nearby sea by shallow barriers. The barriers were flooded during the Early Jurassic eustatic rise, resulting in increased sediment supply and changed circulation patterns within the basin that could give rise to the coarsening-upward succession. The marine link was probably established through a south-eastern entrance along the present-day Scoresby Sund or a southwards extension of the basin. However, this cannot be clearly determined on the basis of the present-day exposures. The Early Jurassic marine inundation has been recorded throughout the North Atlantic rift complex and coincides with a long term Late Triassic – Early Jurassic eustatic sea-level rise (e.g. Haq *et al.*, 1987; Hallam, 1988; Surlyk, 1990a).

A striking feature of the Lower – lower Middle Jurassic Neill Klinter Group is the marked sheet-like geometry of both members and formations on several scales along the exposed basin margins (Fig. 46). This indicates that the Lower – lower Middle Jurassic sediments in the Jameson Land Basin were deposited during a period of relative tectonic quiescence, and facies patterns were mainly controlled by relative sea-level fluctuations, sedimentary influx and marine dispersal systems. However, reflection seismic data (Christiansen *et al.,* 1991) clearly indicate that the Upper Triassic – lower Middle Jurassic Kap Stewart and Neill Klinter Groups show a general thickness increase towards the central parts of the basin, with additional minor depocentres in the eastern part of the basin. The latter are aligned parallel to deep Devonian NW–SE trending transverse basement faults (Dam *et al.,* 1995). The trend of the faults is also clearly visible on Landsat images as major topographic lineaments and are linked to present-day topographic lineaments along which gentle large-scale flexures are commonly developed. These flexures are related to minor Mesozoic and younger deformations. The deep faults also exerted an important control on facies distribution throughout Rhaetian – Pliensbachian times without actually truncating the Mesozoic succession (Dam *et al.,* 1995). The Upper Triassic – lower Middle Jurassic succession thus demonstrates evidence of subtle fault control on subsidence, as seen from isopach maps. However, possible direct field evidence of Early Jurassic tectonic activity along the bounding faults includes the presence of extensive debris flow deposits associated with the storm-dominated offshore deposits of the Albuen Member and the establishment of an apparently structurally controlled lagoon (Horsedal Member) in the northern part of the basin during latest Pliensbachian times.

The changing palaeogeography of the basin through Early – early Middle Jurassic times is described in the following section. Three palaeogeographical reconstructions have been made based on lithostratigraphic, palaeoenvironmental and tectonic evidence.

Latest Sinemurian – Early Pliensbachian marine inundation and establishment of a shallow marine embayment (Rævekløft Formation and Elis Bjerg Member)

The lower 74 m of the Elis Bjerg Member in the north-western part of the basin (Ranunkeldal), are well-exposed and comprise a succession of storm-dominated offshore transition zone, wave and storm-dominated shoreface and tidal channel deposits (Figs 10, 11). The sediments can be divided into six well-defined coarsening-upward units, 5–35 m thick, formed

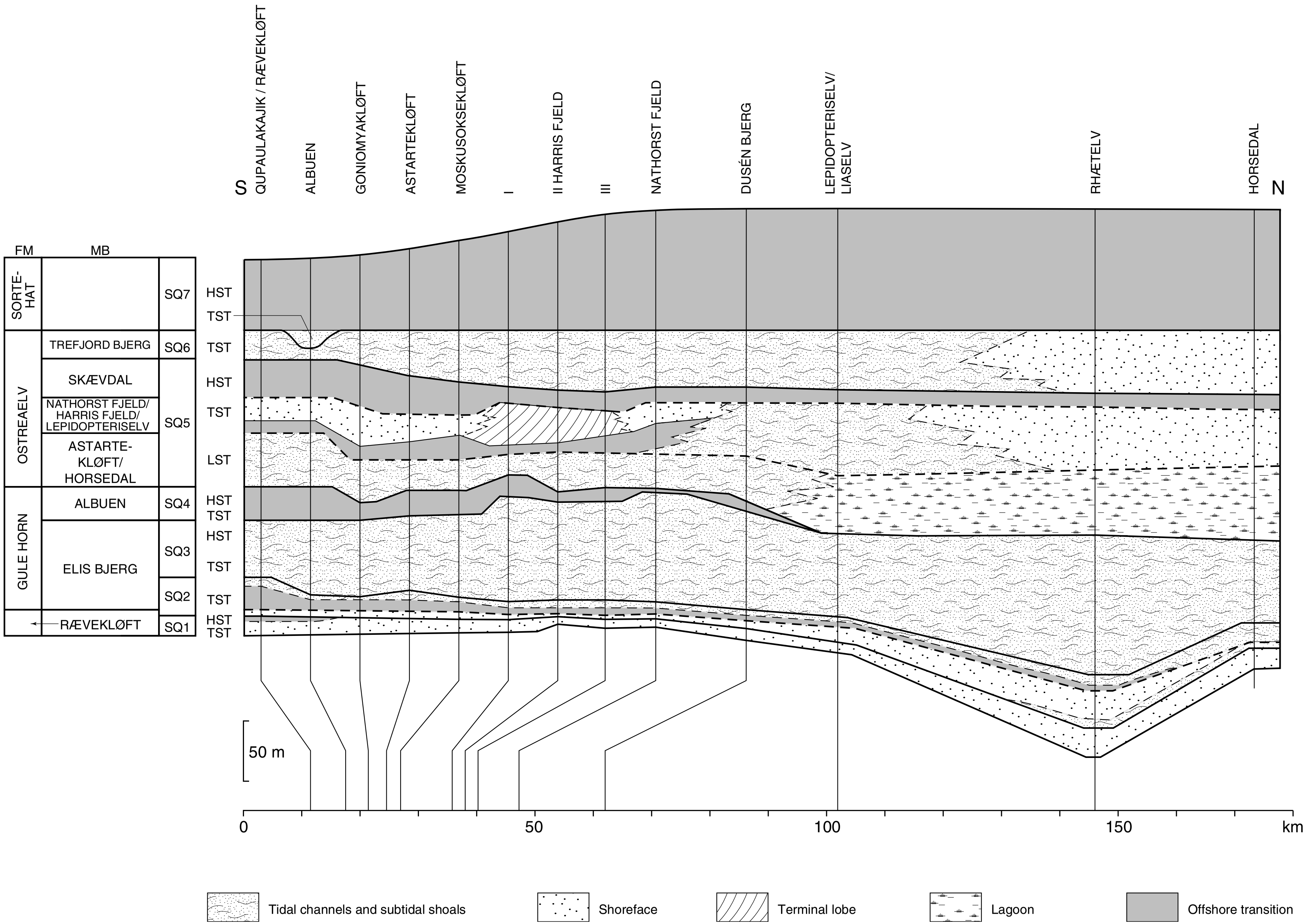


Fig. 46. Sequence stratigraphic and depositional environment correlation panel of the Lower – lower Middle Jurassic Neill Klinter Group, East Greenland. Position of measured sections are marked. Bold-faced line = sequence boundary; Stippled bold-faced line = major flooding surface.

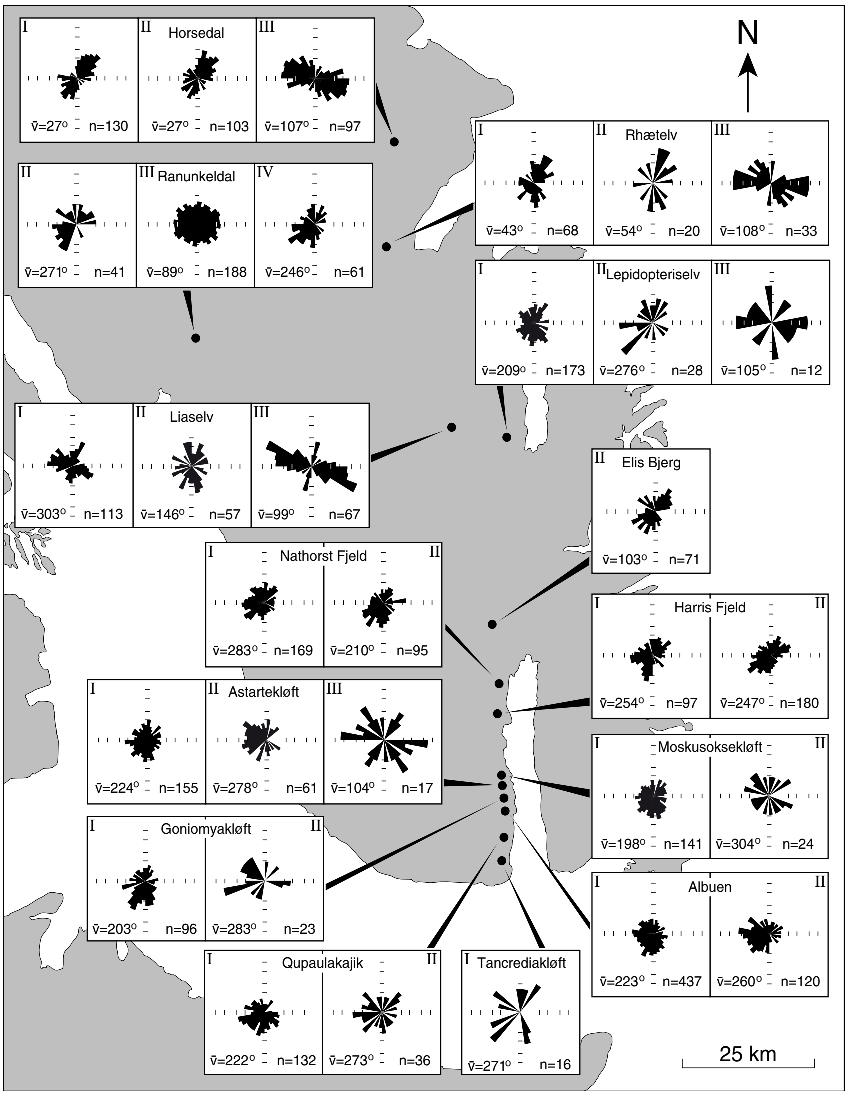


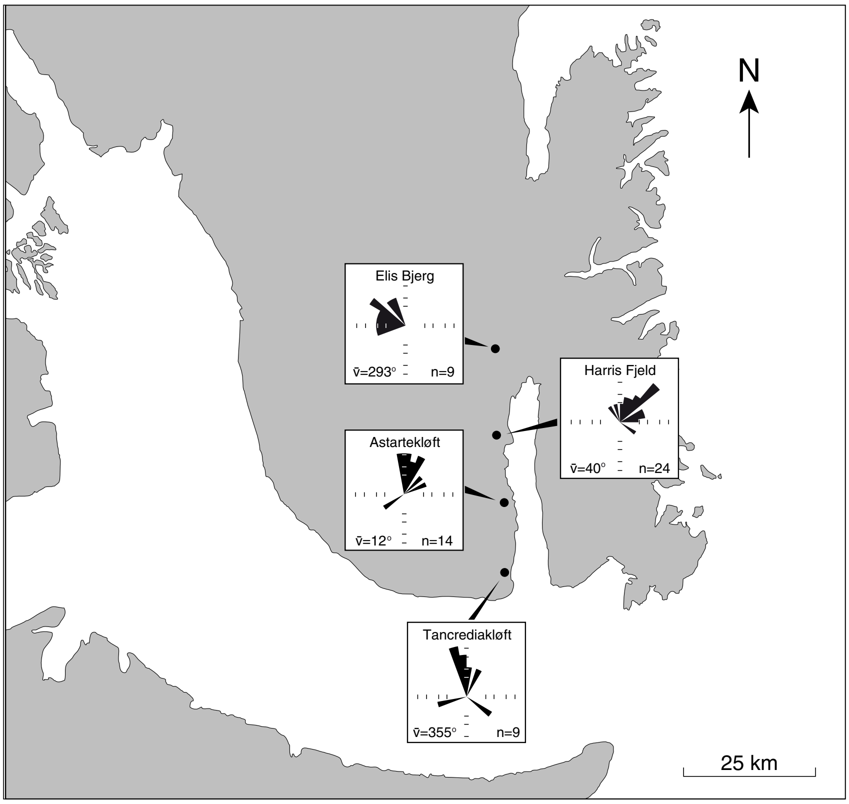
Fig. 47. Palaeocurrent and wave ripple data from the Elis Bjerg Member.
I = foreset azimuths of cross-lamination;
II = foreset azimuths of cross-bedding;
III = crestline orientation of wave ripples;
IV = foreset azimuths of wave-ripple cross-bedding.
Rose diagrams are shown as true area plots. Marks at 5, 10, 20 and 30 % frequency.
v = vector mean;
n̄ = number of measurements.

by progradation of wave and storm-dominated shoreface (Facies association e) across a storm-dominated lower shoreface transitional environment (Facies association d). Tidal channel deposits may top the successions (Facies association g). Crest line orientations of wave ripples in the shoreface and storm-dominated lower shoreface transitional deposits show a rather large scatter with a preferred NW–SE orientation, perpendicular to the western basin margin (Fig. 47). Palaeocurrent data from the cross-bedded shoreface sandstones indicate asymmetrical bipolar NE and SW current directions, roughly parallel to the western basin margin (Fig. 47). No tidal features are present and the close association with wave ripples suggest that the dominant mode of transportation was by wind-driven coast-parallel currents.

Along the south-eastern part of the basin the succession is initiated with the coarse-grained foreshore and shoreface deposits of the Lower Pliensbachian Rævekløft Formation (Facies association a and b) (Figs 6, 7). At the south-easternmost localities (Tancrediakløft and Qupaulakajik) the foreshore and shoreface sandstones are separated by a unit of offshore transition mudstones (Facies association c) (Fig. 7). The mudstones represent a flooding event coincident with an apparent hiatus corresponding to the Ibex Zone between the Jamesoni and Davoei Zones. Palaeocurrent data from the cross-bedded shoreface sandstones show unimodal directions towards N–NE, parallel to the eastern basin margin, suggesting deposition from coast-parallel currents (Fig. 48).

In the south-eastern part of the basin the Elis Bjerg Member is 73–95 m thick and shows a general increase in thickness towards the north (Plate 2). The member is initiated with an offshore transition mudstone unit, 5–13 m thick (Facies association c) which extends along

Fig. 48. Palaeocurrent data (foreset azimuths) from the Rævekløft Formation. Rose diagrams are shown as true area plots. Marks at 5, 10, 20 and 30% frequency; v̄ = vector mean; n = number of measurements.



the Neill Klinter and at least as far north as Elis Bjerg. The mudstones pass upward into closely interbedded deposits of extensive subtidal sand sheets, storm-dominated sandy shoals and relatively deep, actively migrating tidal channels (Facies associations f, g and h). One of the most common sedimentological features is the along-strike lateral transition between the sandy subtidal sand sheet, tidal channel, and storm-dominated shoal associations. The coarsening-upward part of the sandstone bodies reflects progradation of tidal channels where lower shoreface and offshore transition zone deposits pass upwards into subtidal dune field and tidal channel sandstones. The sandstone bodies commonly consist of several smaller cycles of limited lateral extent (1–2 km) which are probably due to autocyclic shift in depocentres. The Elis Bjerg Member appears to thicken along strike towards Harris Fjeld (Plate 2). The thickness increase is associated with an increased amount of tidal channel deposits and a major tidal channel complex is suggested to have been situated in the present-day Harris Fjeld area.

Combined palaeocurrent data from the tidally-influenced facies show two bipolar populations, suggesting reversing currents approximately perpendicular and parallel to the eastern basin margin, respectively (Fig. 47). The N–S to NE–SW population, parallel to the basin margin, is associated with subtidal sand sheet deposits, whereas the E–SE to W–NW population is associated with tidal channel deposits. The latter population is strongly asymmetric with dominantly offshore directed palaeocurrent directions, suggesting an eastern source area covering parts of the present-day Liverpool Land area and bedform migration in response to offshore (ebb) flows.

In the north-eastern (Lepidopteriselv, Liaselv) and northern (Horsedal) parts of the basin, the Elis Bjerg Member is 95–100 m thick. It consists of 5–6 poorly developed coarsening-upward or coarsening-fining upward successions, 4–44 m thick. The coarsening-upward successions were formed by progradation of a subtidal dune field across a ripple field or storm-dominated lower shoreface. Tidal channel deposits only top the Elis Bjerg Member in the Horsedal section. The member thickens towards the central parts of the basin and is 185 m thick at Rhætelv, but the nature of the exposure does not allow study of number and nature of cyclic successions.

Wave ripple data from the north-eastern (Lepidopteriselv and Liaselv), northern (Horsedal) and central (Rhætelv) parts of the basin show dominant ESE–WNW and ENE–WSW crestline orientations, parallel to the northern basin margin and perpendicular to the east-

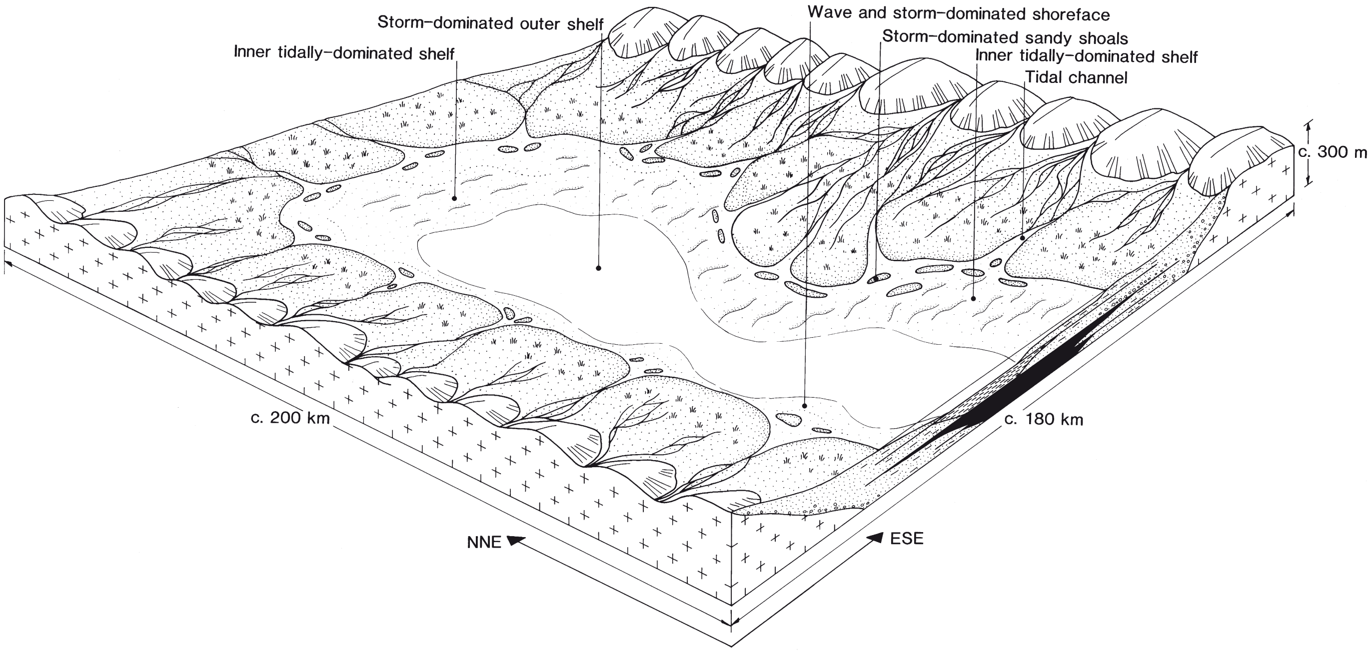


Fig. 49. Palaeogeographical reconstruction of the Jameson Land Basin during Early Pliensbachian times.

ern and western basin margins (Fig. 47). Palaeocurrent data from the cross-bedded and cross-laminated subtidal sandstone sheets show generally bipolar NNE–SSW directions, parallel to the eastern and western basin margins (Fig. 47).

An Early Pliensbachian palaeogeographic reconstruction of the Jameson Land Basin is shown in Fig. 49. Material was transported into the shallow marine embayment from northern, eastern and western land areas. Along the eastern and northern parts of the embayment sediment was evenly distributed via ebb-dominated tidal channels into extensive subtidal dune fields driven mainly by tide-enhanced currents. An important source area appears to have been present east of the present-day Harris Fjeld area. The dune fields passed basinwards into restricted or wave and storm-dominated shelf environments.

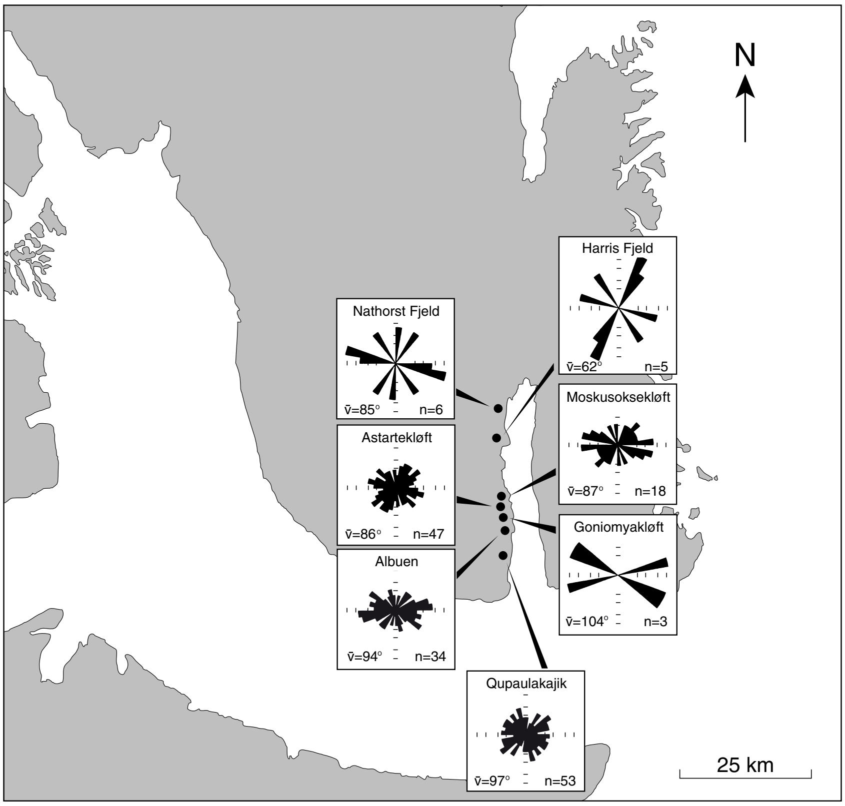
Along the western part of the basin the dominant mode of sand transport appears to have been in northwards and southwards migrating dune fields mainly driven by wind-induced coast-parallel currents. The dune fields passed basinwards into a storm-dominated offshore shelf with deposition of sheet sands from downwelling flows during storms.

Late Pliensbachian sea-level rise and establishment of a storm-dominated offshore shelf (Albuen Member)

Following the deposition of the Elis Bjerg Member, a major landward shift of facies occurred, and in the south-eastern part of the basin a wedge of storm-dominated lower shoreface sediments (Facies association i), associated with submarine debris flow deposits (Facies association j), were deposited in a shallow shelf environment (Fig. 20). The wedge is 25 m thick at Qupaulakajik, thins to less than 2 m at Nathorst Fjeld and disappears north of this location. The lower boundary of the wedge is a prominent flooding surface and the wedge appears to onlap the top of the Elis Bjerg Member. At Nathorst Fjeld the Albuen Member is erosionally overlain by the Astartekløft Member.

At Nathorst Fjeld the crestline of wave ripples from the offshore sandstones of the Albuen Member shows preferred E–W orientations, perpendicular to the eastern basin margin (Fig. 50). The debris flow deposits associated with the storm-dominated lower shoreface deposits may have been triggered by either major storms, tectonic activity or catastrophic floods in the sediment source area. As prominent submarine debris

Fig. 50. Palaeowave ripple data from the Albuen Member. Rose diagrams are shown as true area plots. Marks at 5, 10, 20 and 30% frequency.
v̄ = vector mean;
n = number of measurements.



flow deposits only occur in this interval where they are relatively common, it is suggested that they reflect tectonic activity along the margin of the eastern borderland.

Establishment of a latest Pliensbachian lagoon controlled by deep Devonian faults (Astartekløft and Horsedal Members)

A major basinward shift in facies took place in the latest Pliensbachian and comprises two spatially separated successions of subtidal channels and subtidal shoals (Astartekløft Member), and wave and storm-dominated lagoonal deposits (Horsedal Member).

At Nathorst Fjeld the base of the Astartekløft Member is marked by a major erosional unconformity, and the Albuen Member is almost completely eroded away. At this locality the Astartekløft Member consists of a single fining-upward tidal channel succession (Facies association m) overlain by a thin coarsening-upward succession (Fig. 28). The palaeocurrent data from this locality indicate flows towards the west suggesting bedform migration in response to offshore directed (ebb) currents, from an eastern source area covering the present-day Liverpool area (Fig. 51).

Further south along Neill Klinter, the unconformity between the Albuen and Astartekløft Members disappears and passes into a correlative conformity, and the shoreface deposits of the Albuen Member pass into subtidal sand sheet deposits arranged in an overall coarsening-upward succession (Facies association f) (Fig. 27). The combined palaeocurrent data from the Astartekløft Member at Neill Klinter show two bipolar populations, suggesting reversing currents approximately perpendicular and parallel to the eastern basin margin, respectively (Fig. 51).

The Horsedal Member has been traced for more than 75 km in the northern and north-eastern parts of the basin and shows an extremely uniform thickness (ranging between 50 and 60 m) and facies development, very different from the contemporaneous Astarte-kløft Member. The sediments are arranged in stacked coarsening-upward units, 1–6 m thick, produced by wave and storm-dominated shoreface progradation into a protected lagoon (Facies association k) that probably covered more than 8000 km^2 (Figs 24, 25, 52). A single ephemeral stream delta succession is interbedded with the shoreface deposits (Facies association l; Figs 24, 26). Wave ripple data indicate an E–W to ESE–

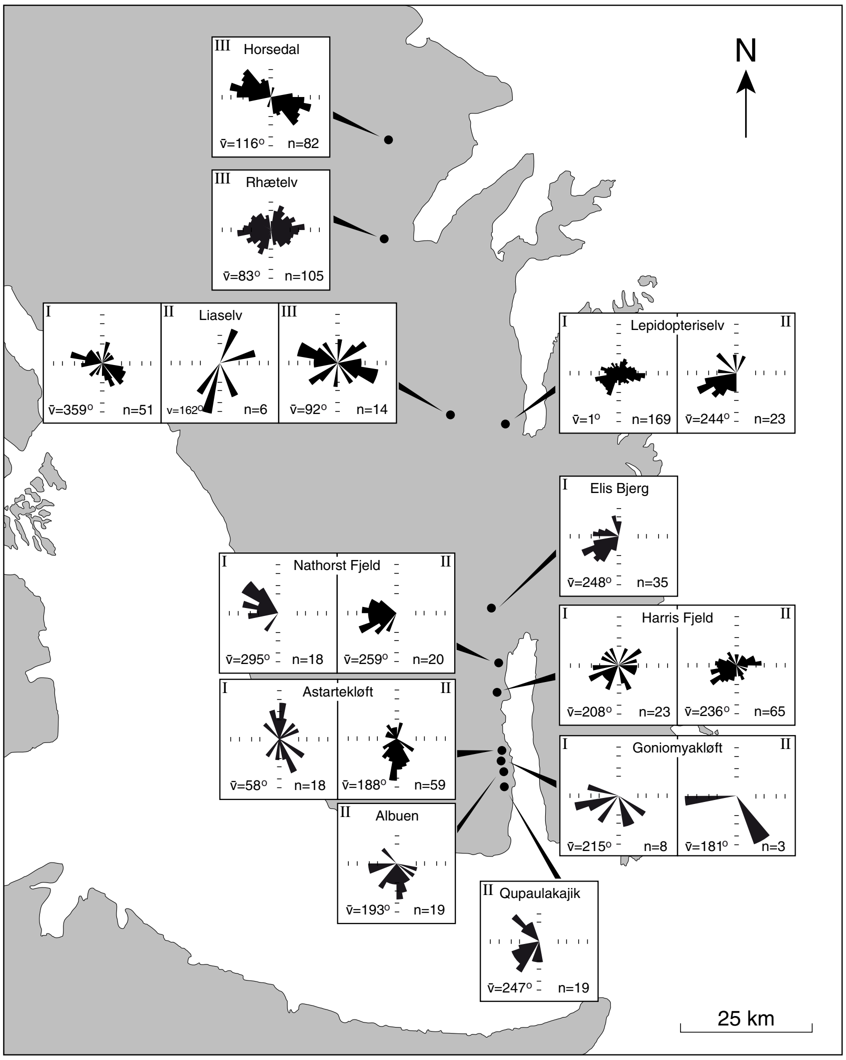


Fig. 51. Palaeocurrent and wave ripple data from the Astartekløft and Horsedal Members.
I = foreset azimuths of current cross-lamination;
II = foreset azimuths of current cross-bedding;
III = crestline orientation of wave ripples.
Rose diagrams are shown as true area plots. Marks at 5, 10, 20 and 30% frequency.
v = vector mean;
n = number of measurements.

WNW trending shoreline, parallel to the northern basin margin (Fig. 51). No major vertical facies changes occur and the successions form an aggradational unit.

Widespread lagoonal deposition in the northern part of the basin took place contemporaneously with tidal channel and subtidal shoal deposition, without lagoonal influence in the south-eastern part of the basin. A barrier complex must thus have been situated in the present-day Dusén Bjerg – Skansen area during Late Pliensbachian times (Fig. 52). The nature and depositional strike of the barrier complex is not known, but crestline orientation of wave ripples in the lagoonal deposits suggest that the barrier had a NW–SE orientation (Fig. 52). The barrier was apparently situated above a major Devonian transverse fault, suggesting structural control of the position and formation of the barrier. A major tidal channel, situated in the Nathorst Fjeld area, was the source of a southward prograding subtidal shoal along the present-day Neill Klinter. A Late Pliensbachian palaeogeographic reconstruction of the Jameson Land Basin is shown in Fig. 52.

Early Toarcian sea-level rise and establishment of a shelf (Nathorst Fjeld, Harris Fjeld and Lepidopteriselv Members)

In the Early Toarcian the tidal channels, subtidal sand sheets and lagoonal environments were drowned and an offshore transition zone environment was established in the southern part of the basin (Facies association p). In the Harris Fjeld area the drowning was followed by progradation of ebb-tidal deltas towards the west (Facies association r), suggesting that a major tidal channel complex was still situated in this area (Figs 34–36). Further south, along Neill Klinter, and in the Nathorst Fjeld area the offshore transition deposits

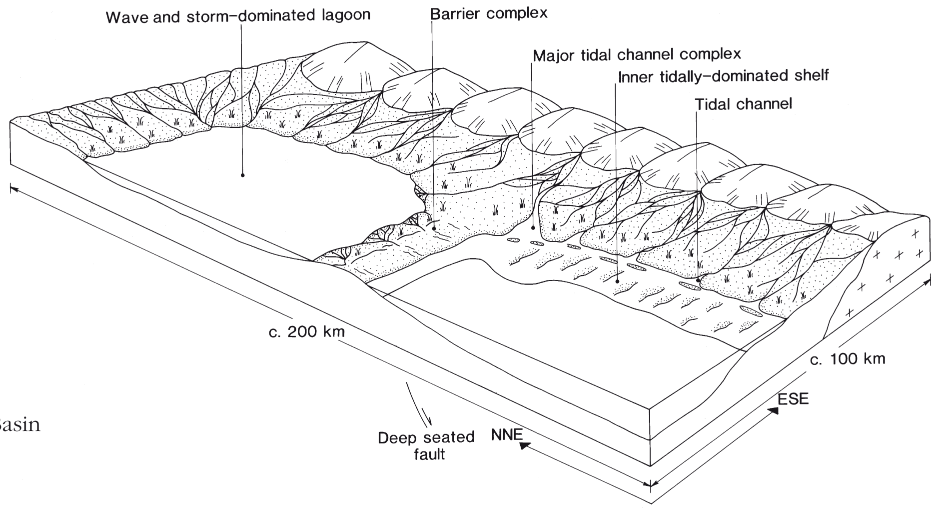


Fig. 52. Palaeogeographical reconstruction of the Jameson Land Basin during Late Pliensbachian times.

pass upward into a sheet of shoreface sandstones (Facies association q) which reflects the same progradational event as recorded by the ebb-tidal delta succession in the Harris Fjeld area (Fig. 32). Foreset dip directions of the cross-bedded shoreface sandstones are bipolar towards N and S, indicating currents roughly parallel to the eastern basin margin (Fig. 53). Wave ripple crestlines in the Nathorst Fjeld Member are oriented either N–S or E–W, parallel and perpendicular to the basin margin, respectively (Fig. 53).

At Lepidopteriselv a large tidal channel complex was established in the Early Toarcian (Facies association m). The channel deposits consist of stacked dune field sandstones (Fig. 30). Palaeocurrent data show a preferred orientation towards the west, indicating deposition from dominantly offshore directed currents (Fig. 53).

Further north, the Lepidopteriselv Member is separated from the overlying lagoonal deposits of the Horsedal Member by a transgressive surface of erosion. In this area a large shallow wave and storm-dominated

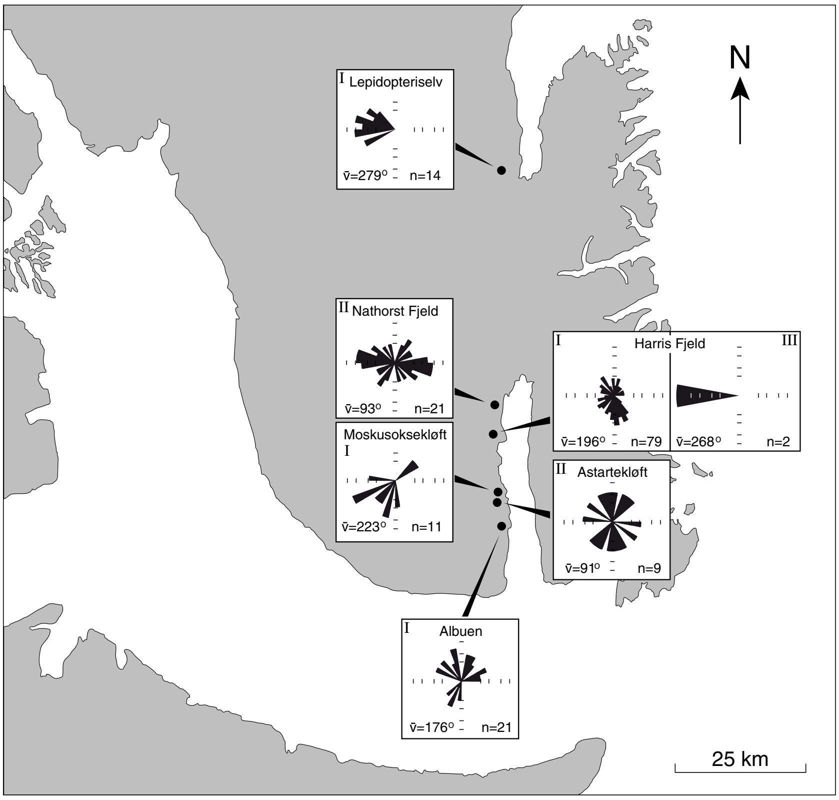


Fig. 53. Palaeocurrent and wave ripple data from the Nathorst Fjeld, Harris Fjeld and Lepidopteriselv Members.
I = foreset azimuths of cross-bedding and cross-lamination;
II = crestline orientation of wave ripples;
III = foreset azimuths of giant-scale cross-bedding.
Rose diagrams are shown as true area plots. Marks at 5, 10, 20 and 30% frequency;
v̄ = vector mean;
n = number of measurements.

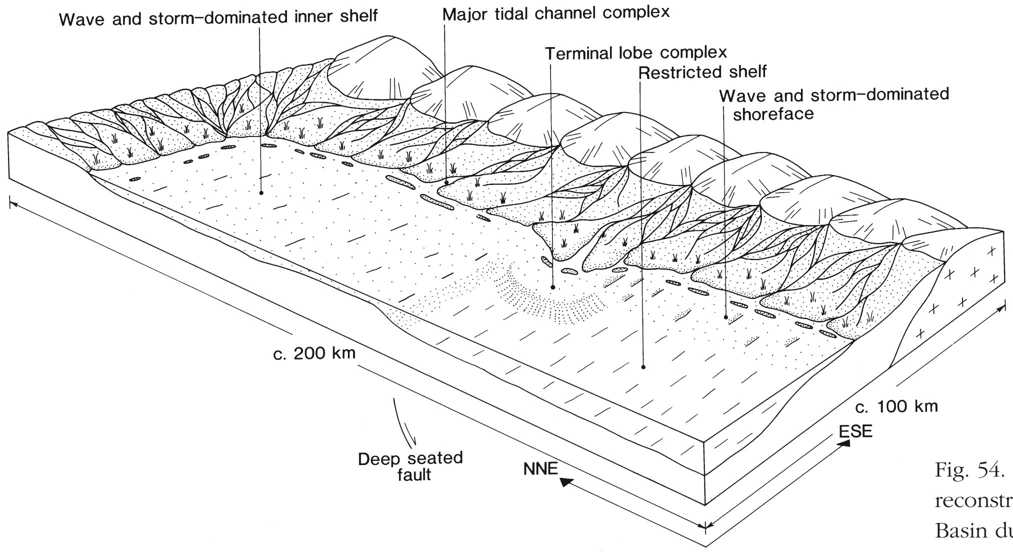


Fig. 54. Palaeogeographical reconstruction of the Jameson Land Basin during Early Toarcian times.

shelf was established that later developed into an interfingering dune field and bioturbated shallow marine environment (Facies associations n and o).

An Early Toarcian palaeogeographic reconstruction of the Jameson Land Basin is shown in Fig. 54.

Toarcian bioturbated shelf and tidal channel deposits (Skævdal and Trefjord Bjerg Members)

A major landward shift in facies took place in the Late Toarcian, and a uniform basin-wide sheet of bioturbated sandy mudstones and muddy fine-grained sandstones of the Skævdal Member was deposited (Facies association o; Figs 39, 46).

This landward shift in facies was followed by a basinward shift in facies throughout the basin represented by sheets of tidal channel, interbedded dune field and bioturbated shoreface sandstones (Facies associations m, n and o) of the Trefjord Bjerg Member. At Lepidopteriselv, the Trefjord Bjerg Member has a sharp lower contact to the muddy sandstones, and a lag conglomerate containing belemnites, bivalves, quartzite pebbles and bored siderite clasts occurs at the boundary. The presence of tidal channel deposits composed of stacked dune field sandstones, arranged in a thinning-upward succession, indicates that a major tidal channel complex still existed in this area (Figs 38, 41, 42). Palaeocurrent data from the cross-bedded sandstones show a preferred orientation towards the west, indicating deposition from dominantly offshore directed currents (Fig. 55).

West of Lepidopteriselv the cross-bedded sandstones pass into increasingly bioturbated sandstones, and at Liaselv, 8 km west of Lepidopteriselv, the sandstones are bioturbated throughout. In the south-eastern and northern parts of the basin the Trefjord Bjerg Member consists of interbedded dune field and bioturbated shoreface deposits (Fig. 38). In the south-eastern part of the basin foreset dip directions of the dune field deposits are generally towards SW, NW and N–NE indicating deposition from coast-parallel and offshore directed currents (Fig. 55). In the northern part of the basin (Horsedal) the foreset dip directions are towards SW (Fig. 55). In Horsedal the dune field sandstones pass into alternating bioturbated and wave and storm-dominated shoreface sandstones.

The upper surface between the dune field and bioturbated shoreface deposits of the Trefjord Bjerg Member and the mudstones of the Sortehat Formation is sharp throughout the basin. *Diplocraterion parallelum* and *Diplocraterion habichi* may in places penetrate down from the formation boundary and lag gravels, containing well-sorted, well-rounded quartzite pebbles with a maximum diameter of 1 cm are common at the boundary or occur dispersed in the lower 20 cm of the mudstones of the Sortehat Formation. In the Skævdal area the mudstones just above the Trefjord Bjerg Member (basal Sortehat Formation) contain abundant inarticulate brachiopods. In Horsedal, in the northern part of the basin, the interbedded dune field and bioturbated shoreface deposits are overlain by a lag conglomerate, containing flat, well-sorted, well-rounded quartzite pebbles with a maximum diameter of 8 cm. The sandstones just above the conglomerate are densely burrowed by *Diplocraterion parallelum* and are suc-

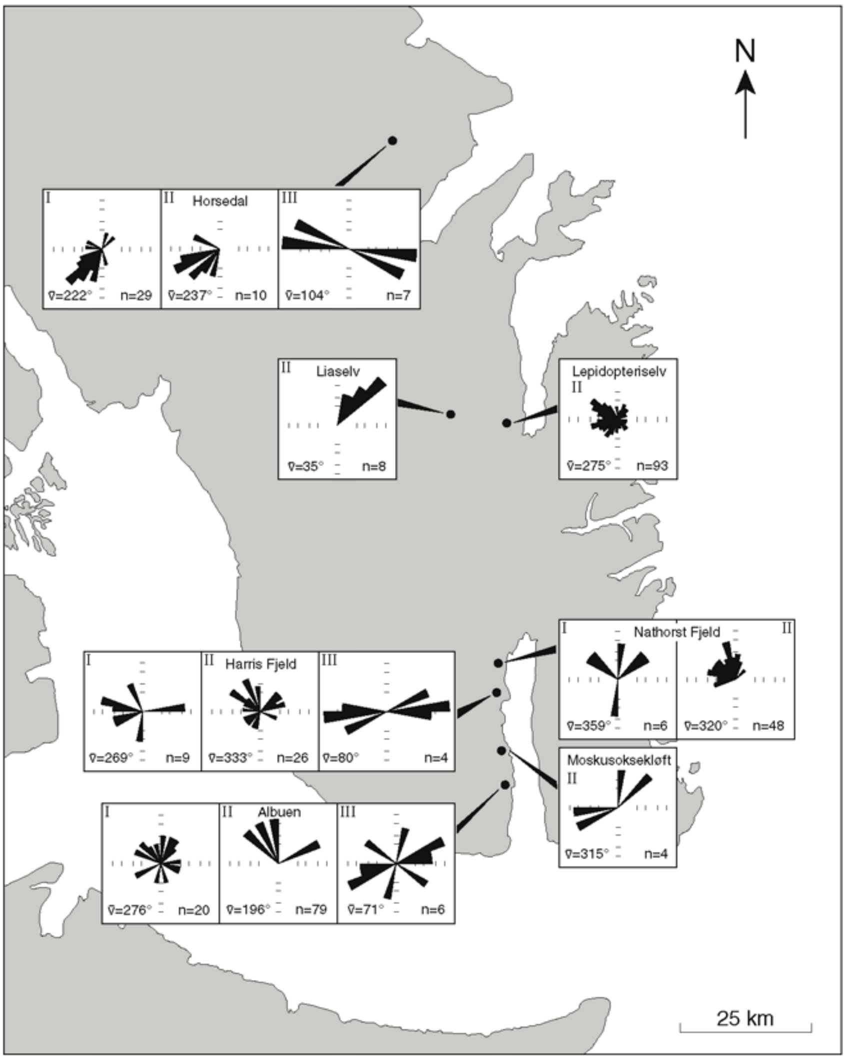


Fig. 55. Palaeocurrent and wave ripple data from the Trefjord Bjerg Member.
I = foreset azimuths of cross-lamination;
II = foreset azimuths of cross-bedding;
III = crestline orientation of wave ripples.
Rose diagrams are shown as true area plots. Marks at 5, 10, 20 and 30% frequency;
∇ = vector mean;
n = number of measurements.

ceeded by wave and storm-dominated shoreface deposits (Facies association n). Wave ripple crestlines trend WNW–ESE parallel to the northern basin margin (Fig. 55). The boundary between the Trefjord Bjerg Member and Sortehat Formation is not exposed in this area.

The dominant mode of sand transport during deposition of the Trefjord Bjerg Member appears to have been associated with extensive southwards, westwards and northwards-migrating dune fields initiated by wind-driven and tide-enhanced coast-parallel and offshore directed currents. The dune fields passed basinwards into bioturbated shoreface environments. A major tidal channel complex in the Lepidopteriselv area was the source of a large part of the sand delivered to the basin in the Late Toarcian.

Aalenian sea-level rise and establishment of a restricted embayment (Sortehat Formation)

Following deposition of the sandstones of the Trefjord Bjerg Member, a major landward shift of facies occurred throughout the basin and a thick succession of offshore mudstones and lower shoreface siltstones and very fine to fine-grained sandstones were deposited (Facies association s), arranged in one or two coarsening-upward successions (Krabbe *et al.*, 1994). A detailed sedimentological and palynological study of the formation is presented by Koppelhus & Hansen (in press).

Sequence stratigraphy

The sequence stratigraphic interpretation of the Neill Klinter Group is based on the identification and interpretation of key surfaces, including unconformities, flooding and drowning surfaces, surfaces or levels rich in siderite clasts or conglomerates composed of extraformational pebbles, and surfaces across which there are major landward or seaward shifts in facies (Dam & Surlyk, 1995).

In this way seven regional unconformities, and six regional flooding or drowning surfaces are recognised (Fig. 46). Correct interpretation of systems tracts is difficult due to the near absence of parasequences, and thus of parasequence sets showing well-defined stacking patterns. As a tentative approach it is suggested that aggradational units topped by a sharp drowning surface, draped by a lag conglomerate formed by shoreface erosion, represent lowstand systems tracts, whereas aggradational to slightly retrogradational, strongly bioturbated units represent transgressive systems tracts. We assume that extensive fine-grained units, formed during periods of regional starvation of coarse clastic material in the basin, represent transgressive systems tracts.

The sequences and key surfaces are described below in ascending stratigraphic order, and a tentative systems tract interpretation is presented. The sequences are referred to from below as SQ1, SQ2 etc., whereas sequence boundaries are abbreviated SB1, SB2 etc.

Sequence 1 (SQ1)

SQ1 consists of the lower part of the Lower Pliensbachian Rævekløft Formation (Figs 3, 6, 7).

Transgressive systems tract. In the south-eastern part of the basin the Rævekløft Formation is separated from the underlying delta-plain deposits of the Rhaetian – Sinemurian Kap Stewart Group by a major basin-margin unconformity that conceals a hiatus that in this area corresponds to all of the Sinemurian Stage (Harris, 1935; Pedersen & Lund, 1980; Figs 5, 7). Basinwards it passes into a correlative conformity, and most or all of the Sinemurian is probably represented in the basinal areas and belongs to the Kap Stewart Group, as indicated by recent palynostratigraphic investigations (E. Koppelhus, unpublished manuscript, 1996). In these areas the boundary between the Kap Stewart and Neill Klinter Groups is marked by a change from lacustrine to shallow marine deposits. The change is indicated by the appearance of marine body, micro- and trace fossils, and by a sudden increase in the ratio of total sulphur to total organic carbon (Fig. 4; Surlyk *et al.*, 1973; Dam & Surlyk, 1993; Koppelhus & Dam, in press). The basin-margin unconformity is commonly overlain by a thin conglomerate of fragmented bivalve shells and extraformational clasts (Fig. 5). At Harris Fjeld the delta plain deposits of the Kap Stewart Group are strongly burrowed by *Diplocraterion parallelum* just below the unconformity. The trace fossil assemblage makes up an omission suite and suggests minimal scouring and reworking of the delta plain deposits of the Kap Stewart Group during the Early Pliensbachian transgression (Dam, 1990b). This suggests that the Sinemurian sediments must have been eroded prior to the transgression. The unconformity and its correlative conformity constitutes the basal sequence boundary (SB1) of sequence 1 of the Neill Klinter Group.

The lower cross-bedded sandstone bed of the Rævekløft Formation is up to 7.5 m thick and was deposited in the upper shoreface and foreshore by longshore currents (Fig. 7). It contains a rich invertebrate fauna of European aspect composed of about 140 species of bivalves, gastropods, cephalopods and echinoderms.

Highstand systems tract. The transgressive systems tract is topped by a sharp, erosional drowning surface overlain by offshore transition mudstones in the south-eastern part of the basin at Tancrediakløft and Qupaulakajik (Fig. 7). The surface is interpreted as a transgressive surface of erosion formed by marine shoreface erosion during transgression. In Tancrediakløft and Qupaulakajik the highstand systems tract is represented by a small coarsening-upward mudstone unit, 2 m thick. Further north this highstand systems tract is not preserved.

Sequence 2 (SQ2)

SQ2 consists of the upper part of the Rævekløft Formation and the lowermost part of the Elis Bjerg Member and coincides with palynomorph assemblage 1 of Koppelhus & Dam (in press) at Albuen (Figs 3, 7, 46; Plate 2).

Transgressive systems tract. In the south-eastern part of the basin the lower sequence boundary (SB2) is defined by a sharp, erosional contact between the offshore mudstones of SQ1, and shoreface and foreshore pebbly sandstones of SQ2 (Fig. 7). The boundary represents a major seaward shift in facies, and is interpreted as a sequence boundary. SB2 can be followed northwards from the south-eastern basin margin over at least 50 km and forms a sharp boundary between the lower and upper Rævekløft Formation. At Elis Bjerg shoreface deposits of SQ1 are strongly burrowed by bivalves just below the unconformity (Fig. 8). Further north and towards the basin centre the unconformity passes into the correlative conformity situated about 25 m above the base of the Elis Bjerg Member, which here forms the lowest part of the Neill Klinter Group.

From Harris Fjeld and northwards SB2 is overlain by a pebbly sandstone of debris flow origin (Fig. 7). It is 2–3 m thick, and has a lateral extent exceeding 20 km. It is characterised by well-rounded pebbles scattered in a matrix of muddy sandstones. Rare transported bivalves and belemnites occur at Elis Bjerg, but are absent at Harris Fjeld (Dam, 1991). The shoreface sandstones of the upper part of the Rævekløft Formation is topped by a major drowning surface, suggesting that the sandstones of the Rævekløft Formation belong to a transgressive systems tract.

Highstand systems tract. The drowning surface at the top of the Rævekløft Formation forms the base of the Elis Bjerg Member of the Gule Horn Formation (Fig. 7; Plate 2). It is overlain by offshore transition zone mudstones showing a slight coarsening-upward trend culminating in a sharply-based tidally dominated shoreface sandstone unit, which may represent a minor forced regression. It is not known if this sharp boundary only has local significance, or if it is of more regional importance. The lower part of the Elis Bjerg Member, where it overlies the Rævekløft Formation, is interpreted as a highstand systems tract. The sharp boundary between the two units is interpreted as a transgressive surface of erosion reflecting a major landward shift in facies.

Sequence 3 (SQ3)

SQ3 includes the main part of the Elis Bjerg Member (Figs 3, 46; Plate 2) and coincides with palynomorph assemblage zones 2 and 3 of Koppelhus & Dam (in press).

Transgressive and highstand systems tracts. Sequence 2 is topped by an erosional surface draped by a quartzite pebble and belemnite conglomerate, which forms the boundary between two distinct facies packages (Plate 2). The size of quartzite pebbles exceeds those of the underlying Elis Bjerg Member, and the formation of the conglomerate appears not to be related to simple transgressive shoreface erosion. The conglomerate may represent a reworked and winnowed fluvial lag, and the erosional surface is accordingly interpreted as an unconformity, forming the third sequence boundary (SB3).

Along the eastern basin margin the overlying part of the Elis Bjerg Member is very difficult to interpret within a sequence stratigraphic framework. The deposits consists of amalgamated tidal channel and subtidal sandsheet facies packages (Plate 2). Well-defined facies cycles are not developed and parasequences can only rarely be recognised. The uppermost part of the Elis Bjerg Member is, however, characterised by the incoming of major tidal channel complexes, forming the best potential hydrocarbon reservoirs in the member (Plate 2). Although they are concentrated in the upper part of the member they can only be lithostratigraphically delineated. The dominant depositional motif seems to be aggradational to progradational. This probably represents moderate rate of creation of new accommodation space and high sediment influx in a flat-bottomed basin. Correlation of siderite clast conglomerates (Fig. 17; Plate 2) gives some clues to the presence of cryptic parasequences and thus to possible relative sea-level fluctuations. Conglomerates are rare to absent in the lowermost *c.* 20 m of SQ3, but become very common in the higher parts of the sequence. Many of the conglomerates form marker beds which can be followed for 10–20 km (Plate 2). The presence of siderite clast conglomerates shows that penecontemporaneous siderite cementation had taken place in the upper sediment layers. Early siderite cementation preferably occurs in marginal marine, brackish and terrestrial environments (e.g. Mozley & Wersin, 1992), and a high content of siderite clasts is thus interpreted as representing shallow water deposition followed by emergence and erosion.

The upward progression from siderite-poor to siderite-rich units in the south-eastern part of the basin, may thus indicate a trend from rare subaerial exposure to common exposure, and a decrease in rate of generation of new accommodation space.

In the north-western part of the basin in Ranunkeldal, the lower part of the Elis Bjerg Member is made up of storm-dominated shoreface deposits. The deposits consist of alternating mudstones and well-sorted fine-grained storm-generated sandstones. The sandstone laminae and beds are arranged in thickening and shallowing-upward units, up to 15 m thick, separated by flooding surfaces, representing well-developed parasequences (Figs 10, 11). This trend and the aggradational to progradational stacking pattern in both the south-eastern and north-western part of the basin suggests that the succession represents a transgressive systems tract overlain by a highstand systems tract.

Sequence 4 (SQ4)

SQ4 consists of the Albuen Member (Fig. 46) and coincides with palynomorph assemblage 4 of Koppelhus & Dam (in press). The Elis Bjerg Member is topped by a major erosional flooding surface (SB4) which forms the boundary to the overlying Albuen Member. SB4 is interpreted as a coalesced sequence boundary and transgressive surface, suggesting the presence of lowstand deposits in more basinal areas. The Albuen Member is about 26 m thick along the south-eastern basin margin; it gradually wedges out towards the north, and is nearly completely removed by later erosion at Nathorst Fjeld in connection with formation of SB5 (Figs 20, 46).

Transgressive systems tract. In detail the lower boundary shows a complex development. The basal few metres of the Albuen Member still have a high content of marine sandstone beds in most sections, and a clear fining-upward trend can be recognised before mudstone becomes the dominant facies (Fig. 20). The fining-upward unit is interpreted as representing the transgressive systems tract. The maximum flooding surface thus occurs a few metres above the transgressive surface in the most fine-grained part of the Albuen Member.

Highstand systems tract. The fining-upward transgressive systems tract is overlain by heteroliths of alternating mudstones and well-sorted, fine-grained storm-generated sandstones deposited in the offshore transition zone. Debris flow deposits, up to 2 m thick, are interbedded with the heteroliths and can be followed laterally for more than 30 km along strike (Fig. 20). The heteroliths form a slightly upwards coarsening unit that is interpreted as a highstand systems tract (Figs 20, 46).

Sequence 5 (SQ5)

SQ5 consists of the Astartekløft, Horsedal, Nathorst Fjeld, Harris Fjeld, Lepidopteriselv and Skævdal Members (Figs 3, 46) and coincides with palynomorph assemblage zone 5 and the lower half of zone 6 of Koppelhus & Dam (in press).

Lowstand systems tract. Along the Hurry Inlet the base of SQ5 is marked by an important erosional unconformity. It truncates the underlying mudstones of the Albuen Member which are almost completely cut out at Nathorst Fjeld (Fig. 46). Further north, the degree of truncation decreases and the unconformity passes into a correlative conformity, separating the Elis Bjerg Member and the overlying Horsedal Member (Figs 3, 46).

Along the south-eastern basin margin the unconformity is marked by fine-grained marine heteroliths overlain by subtidal sandsheet and tidal channel deposits of the Astartekløft Member, corresponding to an important and abrupt seaward shift in facies. In central and northern Jameson Land the lower part of SQ5 consists of aggradationally stacked lagoonal parasequences of the Horsedal Member (Figs 24, 25). The parasequences represent more proximal facies than the underlying tidal sandstones of the Elis Bjerg Member, and thus represent an important seaward shift in facies. The unconformity and its correlative conformity is interpreted as a sequence boundary (SB5) because it is erosional, of basin-wide nature, and marks a basinward shift in facies. The abrupt seaward shift in facies, the aggradational parasequence stacking pattern and the flooding surface at the top suggest that the Astartekløft and Horsedal Members represent a thickly developed lowstand systems tract.

Transgressive systems tract. The Astartekløft Member is overlain by an important flooding surface, in places veneered by a thin conglomerate of well-sorted, well-rounded quartzite pebbles, interpreted as a marine transgressive surface of erosion (Fig. 29). The overlying

Nathorst Fjeld Member consists of a single coarsening-upward unit composed of offshore transition zone mudstones grading upward into shoreface sandstones (Fig. 32). The Nathorst Fjeld Member interfingers with seaward prograding low-angle clinoform bedded ebb tidal delta deposits of the Harris Fjeld Member (Figs 34-37).

Further north the offshore mudstones, shoreface sandstones, and ebb-tidal delta heteroliths and sandstones of the Nathorst Fjeld and Harris Fjeld Members pass laterally into cross-bedded tidal channel sandstones of the Lepidopteriselv Member (Fig. 30). The transgressive surface of erosion separating the Astartekløft and Nathorst Fjeld Members, passes northwards into a flooding surface separating the north-eastern lagoonal deposits of the Horsedal Member from the overlying tidal shoal deposits of the Lepidopteriselv Member. Thus to the south this surface forms the boundary between coarse-grained shoreface and tidal channel deposits below and fine-grained offshore deposits above, and represents a landward shift in facies. Towards the north the lithological succession is reversed, relatively fine-grained lagoonal deposits being overlain by much coarser shoreface sandstones, but the boundary still marks a landward shift in facies.

The complex unit composed of the Nathorst Fjeld, Harris Fjeld, and Lepidopteriselv Members is interpreted as representing the transgressive systems tract.

Highstand systems tract. The transgressive systems tract is topped by a second drowning surface overlain by strongly bioturbated shoreface heteroliths of the Skævdal Member showing a slight coarsening-upward trend in several sections (Figs 38, 39). This surface may represent the maximum flooding surface of SQ5, and the basin-wide coarsing-upward shoreface deposits of the Skævdal Member possibly constitutes a thin highstand systems tract.

Sequence 6 (SQ6)

SQ6 consists of the Trefjord Bjerg Member and coincides with the upper part of palynomorph assemblage zone 6 at Astartekløft of Koppelhus & Dam (in press).

Transgressive systems tract. The Skævdal Member forming the top of SQ5 is truncated by a prominent basin-wide erosional unconformity draped by a lag conglomerate (SB6; Figs 41–43). The conglomerate pebbles consist of skeletal fragments, mudstone clasts, and bored siderite clasts. The unconformity marks a basin-wide seaward shift in facies and is interpreted as a sequence boundary (SB6). It is overlain by subtidal cross-bedded sandstones deposited in extensive sandwave fields in a tidal channel complex, and fossiliferous, bioturbated, and wave and storm-dominated shoreface sandstones of the Trefjord Bjerg Member.

Sequence 7 (SQ7)

Sequence 7 consists of the uppermost 3 m of the Trefjord Bjerg Member at Albuen and all of the Sortehat Formation and coincides with palynomorph assemblage zones 7, 8 and 9 of Koppelhus & Dam (in press) and Koppelhus & Hansen (in press).

Transgressive systems tract. The boundary between the Trefjord Bjerg Member and the Sortehat Formation is one of the most prominent lithological boundaries in the Mesozoic succession of East Greenland, and it seems to represent a major landward shift in facies caused by a significant rise in relative sea-level. However, the palynological study of Koppelhus & Dam (in press) shows that the *Botryococcus* assemblage, characteristic of the lower part of the Sortehat Formation, is also present in the uppermost 3 m of the Trefjord Bjerg Member at Albuen, above a thin conglomerate. This suggests that the thin conglomerate at Albuen marks a sequence boundary and that the sandstones above represent a thin transgressive systems tract (Fig. 46).

Highstand systems tract. At other localities the sandstones of the Trefjord Bjerg Member is topped by a locally pebble strewn drowning surface, forming the boundary between the Trefjord Bjerg Member and the black mudstones of the Sortehat Formation (Fig. 45). The boundary has a basin-wide distribution and the conglomerate that locally tops the Trefjord Bjerg Member contains pebble sizes larger than any recorded in the sandstones below. It has the field appearance of a major marine flooding surface formed by marine shoreface erosion during transgression. It is likely that the pebble conglomerate, that in places drape the flooding surface, represents a bypass zone or a reworked lag of a thin lowstand systems tract which had bypassed the Trefjord Bjerg Member. This would explain the very sharp, laterally extensive nature of the surface and the remarkable shift from fully to restricted marine conditions, as suggested by the macrofossil content (Surlyk *et al.*, 1973; Surlyk 1990b, 1991), the organic geochemistry

(Krabbe *et al.*, 1994) and the palynological assemblages (Koppelhus & Dam, in press; Koppelhus & Hansen, in press), accompanying what appears to be a major deepening and transgression. If this interpretation is correct, the Ostreaelv-Sortehat boundary, except for the Albuen section, represents a coalesced sequence boundary and top of a lowstand surface (= transgressive surface). The maximum flooding surface is probably situated in the very dark and organic-rich mudstones just above the formation boundary with the main part of the Sortehat Formation interpreted as a highstand systems tract.

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Legend to sedimentary logs

Plate I

Lithology

Mudstone

Sandstone

Pebbly sandstone

Coal

Volcanic intrusive

Concretion

Siderised rip-up mudstone clasts/conglomerate

qz Quartzite clasts/conglomerate

Bed contacts

Sharp/erosive or irregular

Sharp/planar

Gradational

Sedimentary features

Parallel lamination

Lenticular bedding

Wavy bedding

Flaser bedding

Planar cross-bedding

Trough cross-bedding

Cross-bedding with pebbles along foresets

Structureless

Structureless (with quartzite pebbles)

Slumping

Cross-lamination

Incipient wave ripple lamination

Wave ripple cross-lamination

Hummocky and swaley cross-stratification

Coarse-grained ripples

Biota

Rootlets

Plant fragments

Drifted plant stems/logs

Bivalves

Gastropods

Ammonites

Belemnites

Brachiopods

Crinoids

Echinoderms

Fish

Undifferentiated shell fragments

Trace fossils

Weak
Moderate
Intense

} bioturbation

Arenicolites isp.

Curvolithos multiplex

Diplocraterion habichi

Diplocraterion parallelum

Gyrochorte comosa

Ophiomorpha nodosa

Phoebichnus trochoides

Planolites beverleyensis

Taenidium serpentinum

Thalassinoides isp.

Unidentified burrow

Miscellaneous features

Current direction from cross-bedding and cross-lamination

Crestline orientation of wave ripples

Cone-in-cone structures

Ooids

Basement boulder

The accompanying legend is applicable to all sedimentological logs in this paper. Due to different drawing techniques the signatures may not entirely match the individual figures.

Plate 2 (in pocket). Sedimentological logs through the Elis Bjerg Member showing interbedded offshore transition, tidal channel, subtidal sand sheet and storm-dominated sandy shoal deposits. Sequence boundaries and correlatable siderite clast conglomerates are indicated. Logs are measured along a south–north trending profile line from the south-eastern to the northern parts of the basin. See Plate 1 for legend. Modified from Dam & Surlyk (1995).