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

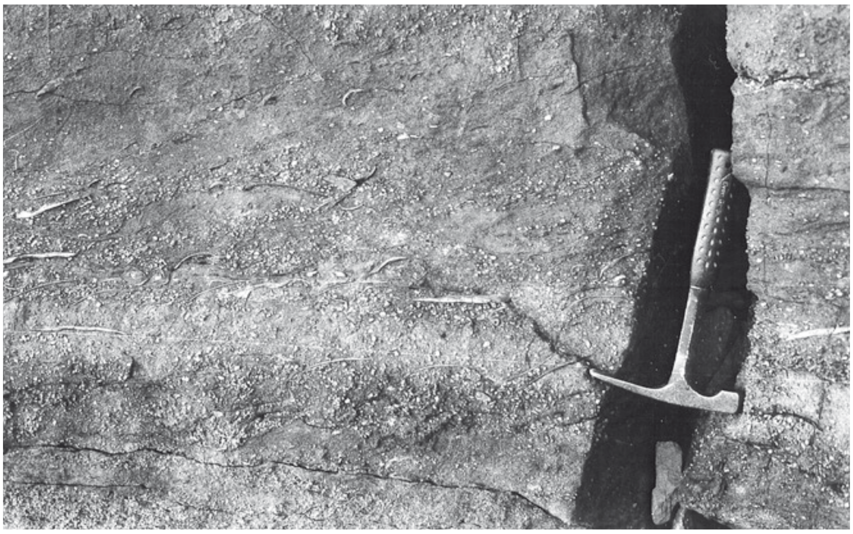
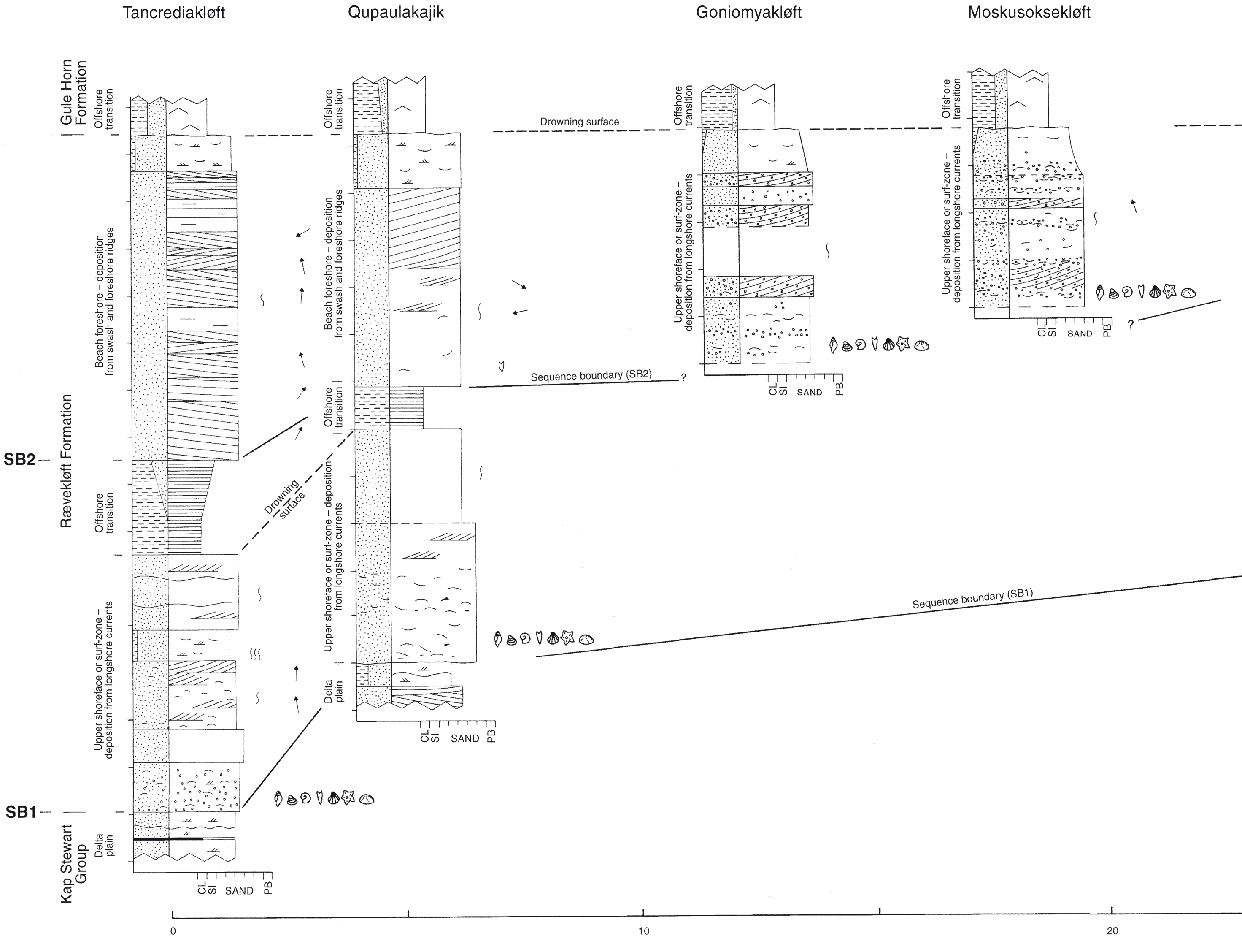


Fig. 6. Poorly sorted medium to very coarse-grained cross-bedded sandstones with scattered granules and shell fragments (Facies association a). From the Rævekløft Formation, Harris Fjeld. See Fig. 1 for location. Hammer 32 cm long.



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-

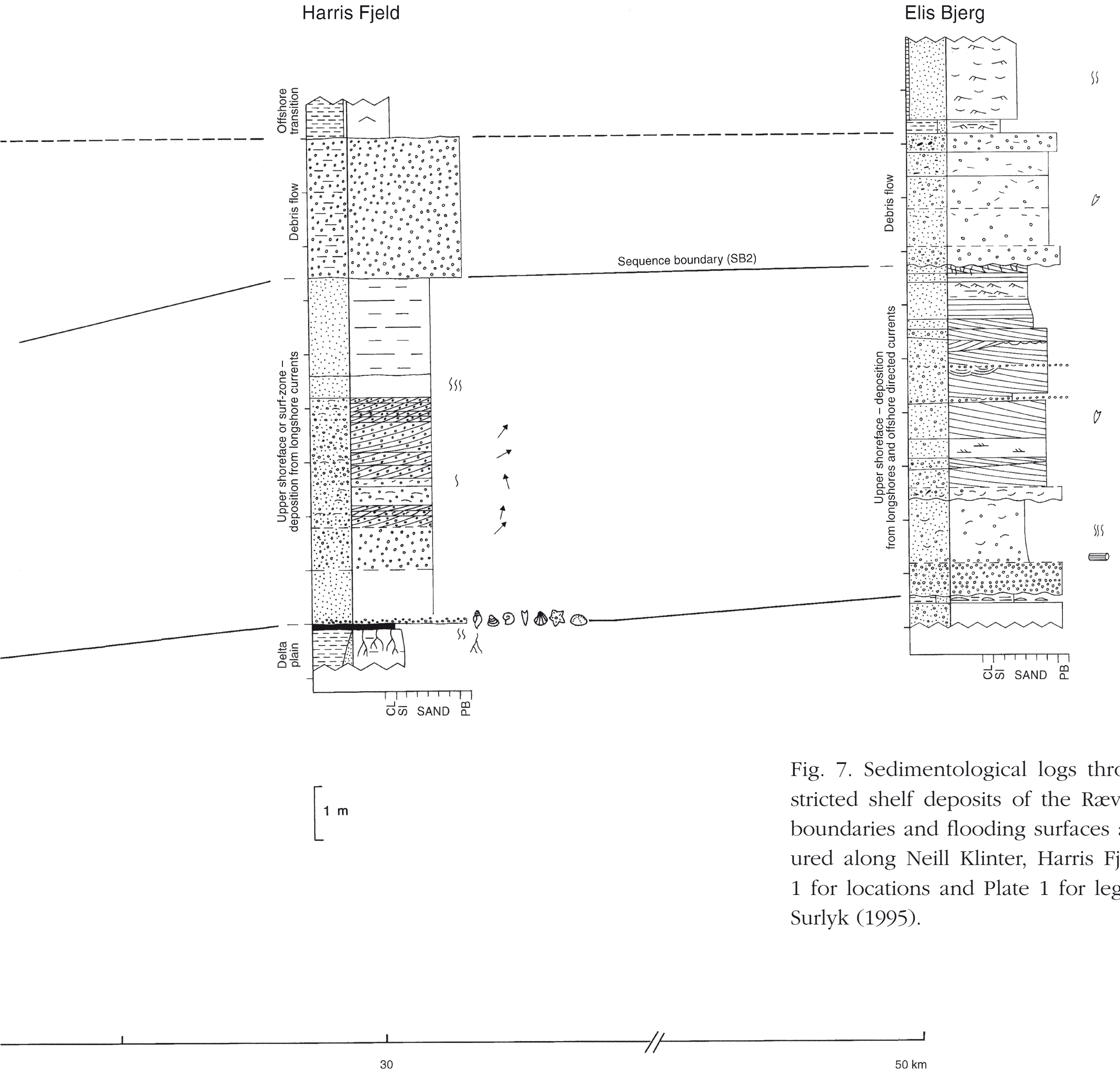


Fig. 7. Sedimentological logs through the shoreface and restricted shelf deposits of the Rævekløft Formation. Sequence boundaries and flooding surfaces are indicated. Profiles measured along Neill Klinter, Harris Fjeld and Elis Bjerg. See Fig. 1 for locations and Plate 1 for legend. Modified from Dam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

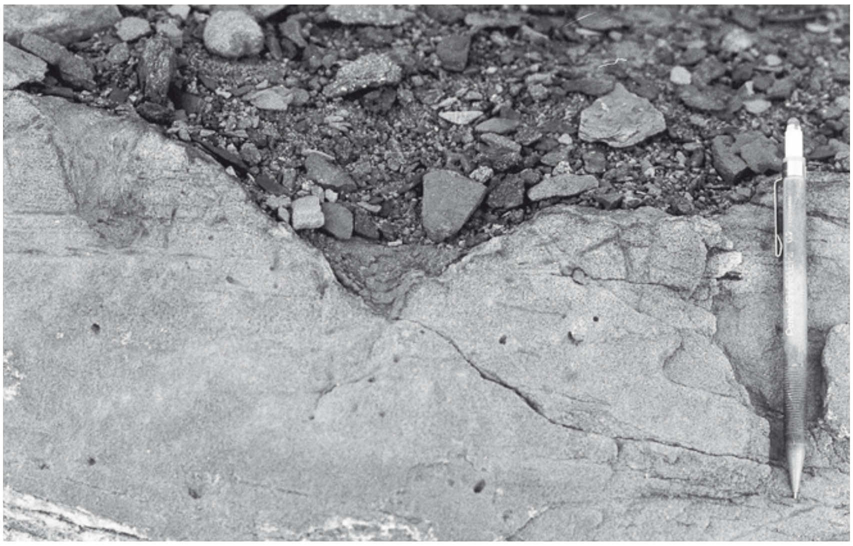


Fig. 8. Bivalve burrows from a level in the Rævekløft Formation that can be traced laterally for several hundreds of metres at the north slope of Elis Bjerg. See Figs 1 and 7 for location. Pen 14 cm long.

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 southeasternmost 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

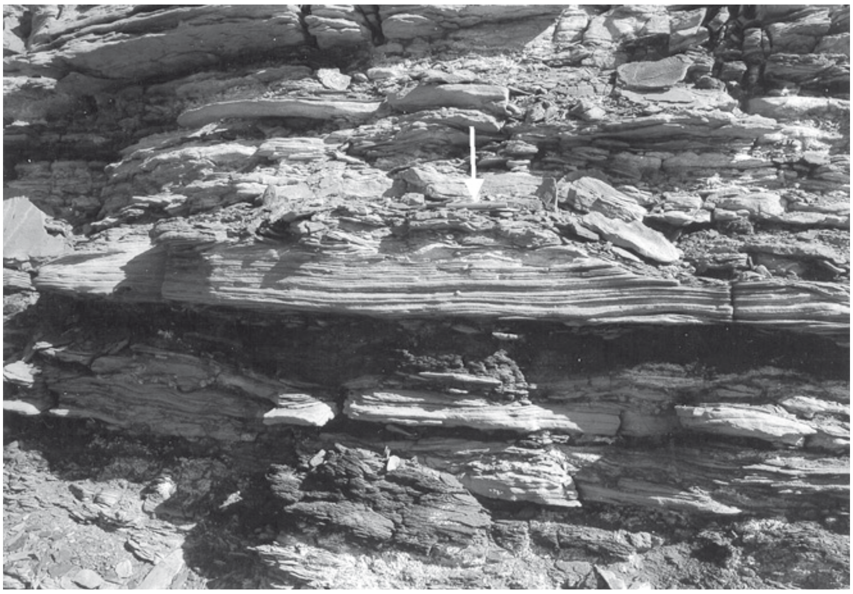


Fig. 9. Laterally persistent storm sandstones (tempestites) interbedded with homogeneous mudstones of the storm-dominated lower shoreface association (Facies association d). Lower part of the Elis Bjerg Member, Ranunkeldal. See Figs 1, 10 and 11 for location. Pen 14 cm long (arrow).

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-

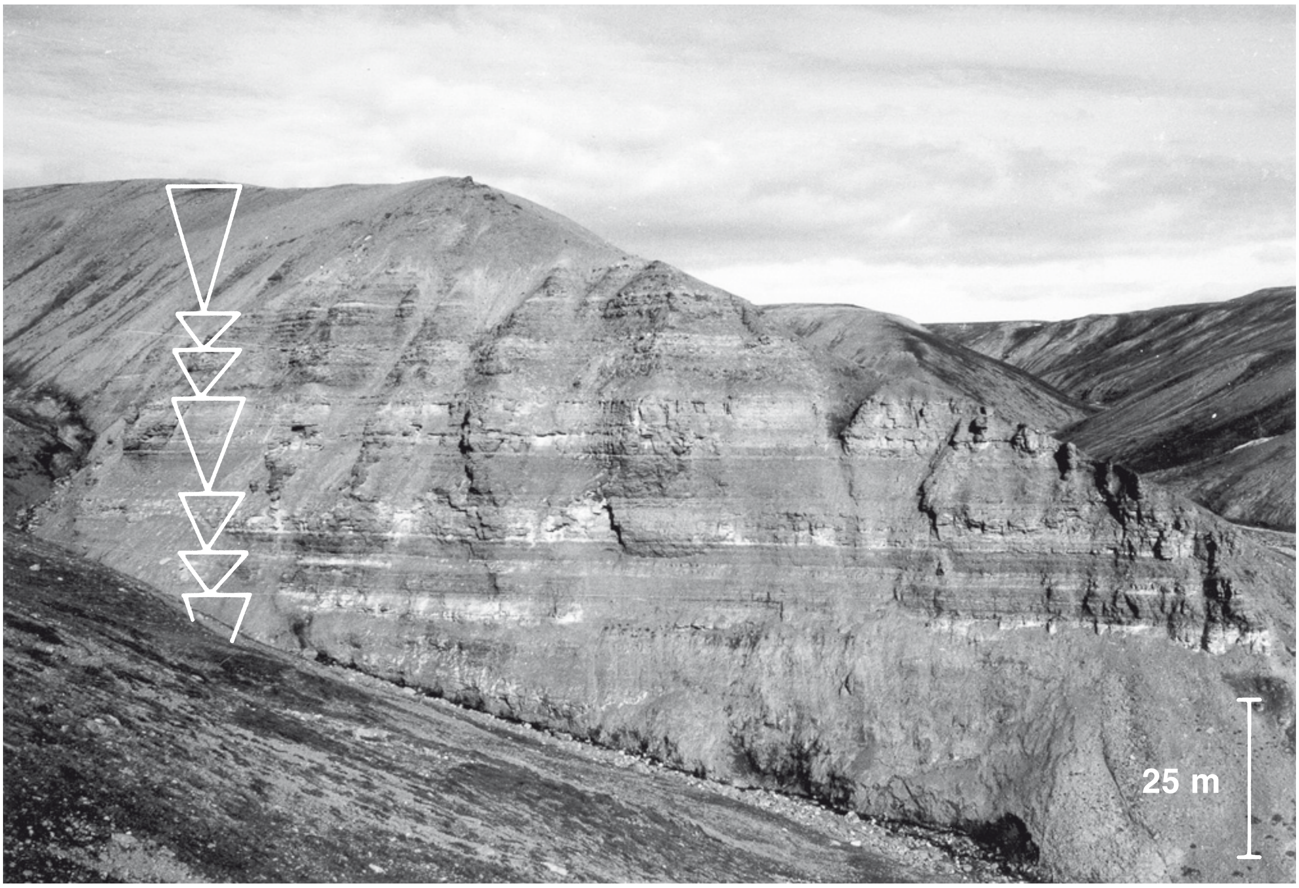


Fig. 10. Stacked coarsening-upward successions formed by progradation of wave and storm-dominated shorefaces (Facies associations d and e) into a storm-dominated offshore shelf. Lower part of the Elis Bjerg Member, Ranunkeldal. See Fig. 1 for location. From Dam & Surlyk (1995).

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

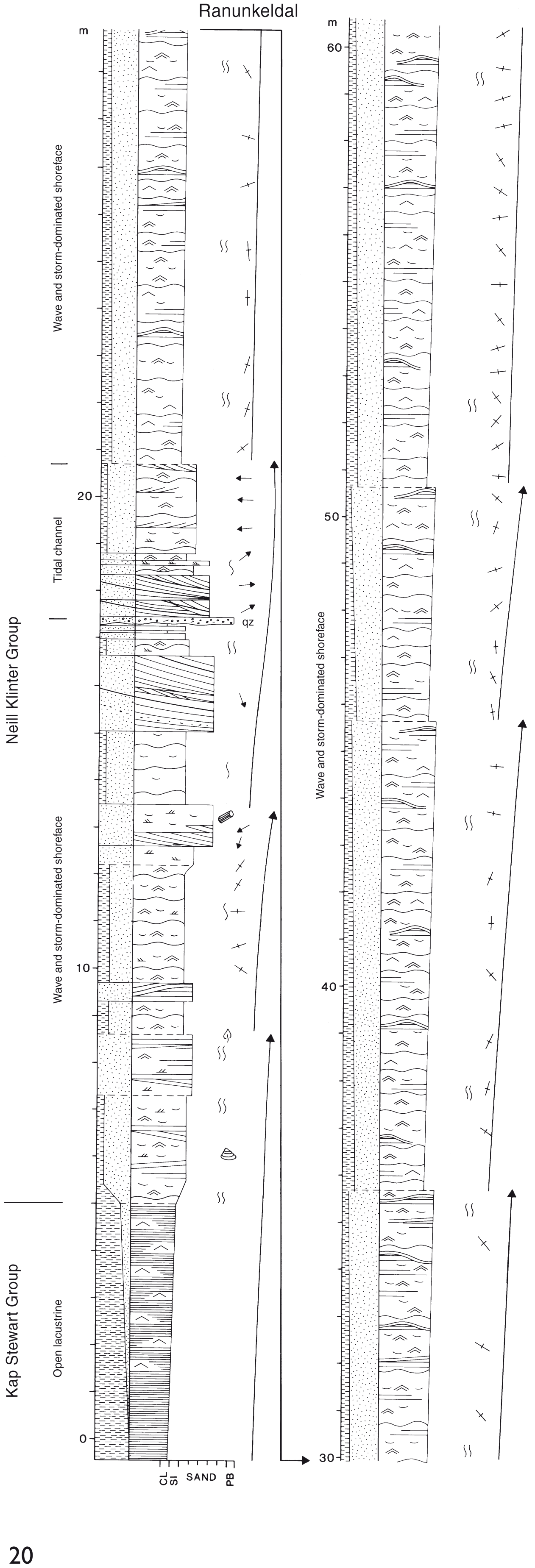


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Fig. 11. Sedimentological log through stacked coarsening-upward successions formed by progradation of wave and storm-dominated shorefaces into a storm-dominated offshore shelf. Lower part of the Elis Bjerg Member, Ranunkeldal. See Fig. 1 for location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

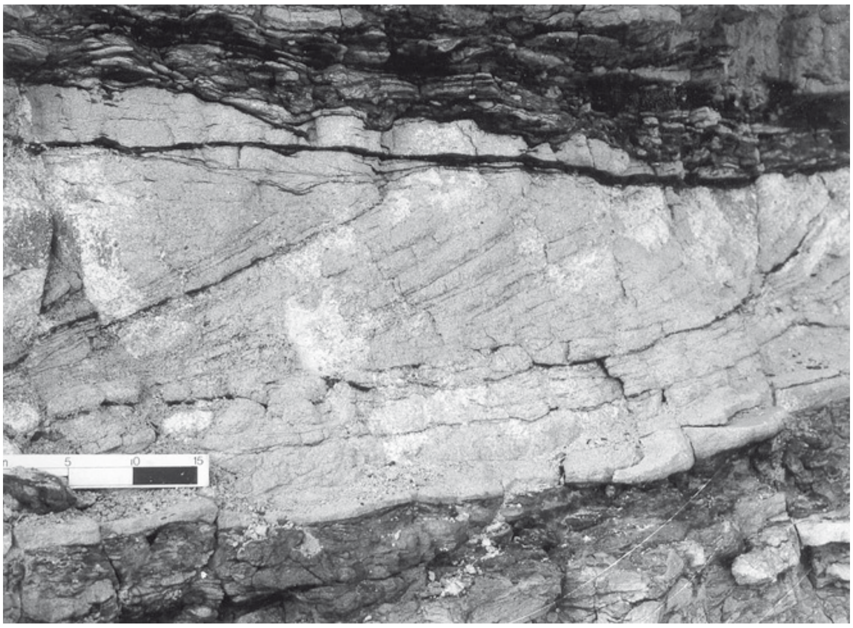

Fig. 13. Cross-bedded sandstone with mudstone drapes and heterolithic bedding of the subtidal sand sheet association (f). The heterolithic bed is intensively burrowed by *Diplocraterion parallelum*. Elis Bjerg Member, Astartekløft. See Fig. 1 for location.

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,*

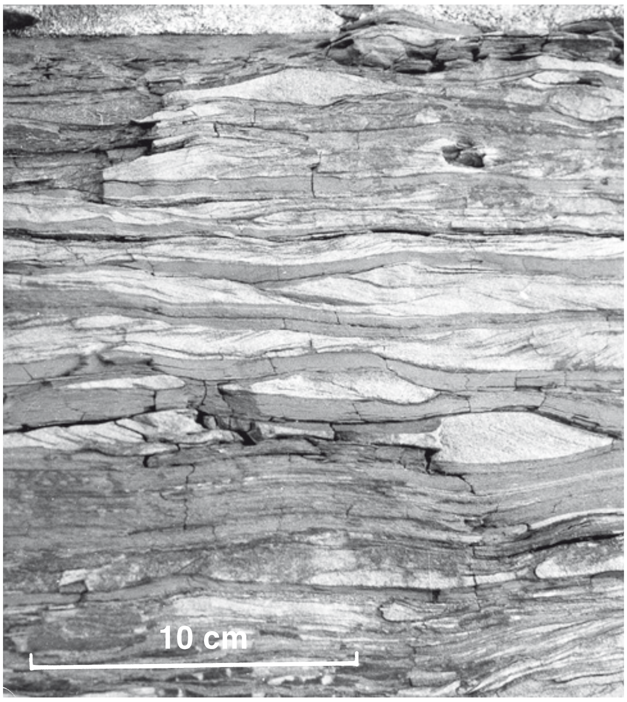

Fig. 16. Wavy bedding of the subtidal sand sheet association (f). Notice that sediment transport directions in succeeding sandstone laminae are commonly opposite, suggesting that the thinly interbedded sandstone and mudstone can be regarded as tidal bundles. Elis Bjerg Member, Astartekløft. See Fig.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

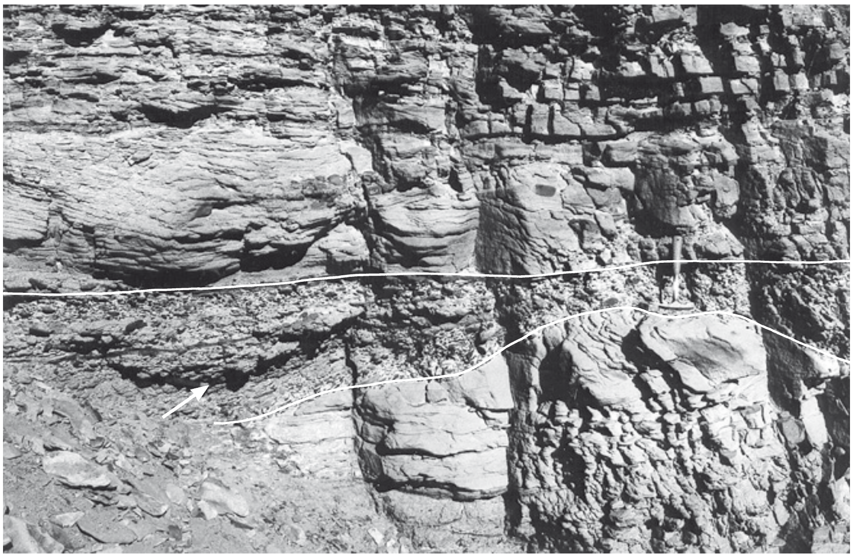

Fig. 17. Intraformational siderite clast conglomerate of the subtidal sand sheet association (f) (arrow). Elis Bjerg Member, Nathorst Fjeld. See Fig. 1 for location. Hammer 28 cm long.

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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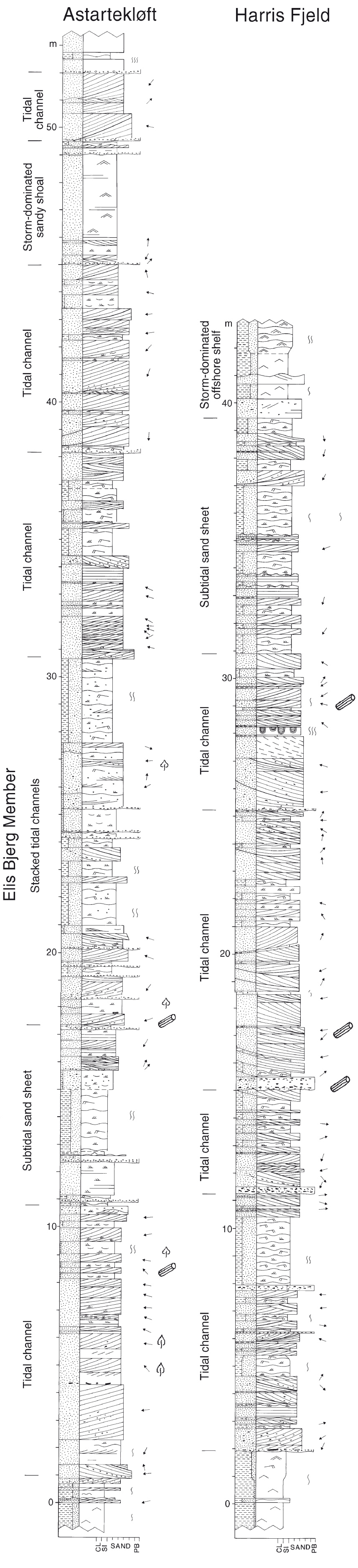

Fig. 18. Sedimentological logs through stacked tidal channel, subtidal sand sheet, and storm-dominated sandy shoal deposits. Elis Bjerg Member, Astartekløft and Harris Fjeld. See Fig. 1 for location and Plate 1 for legend.

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-

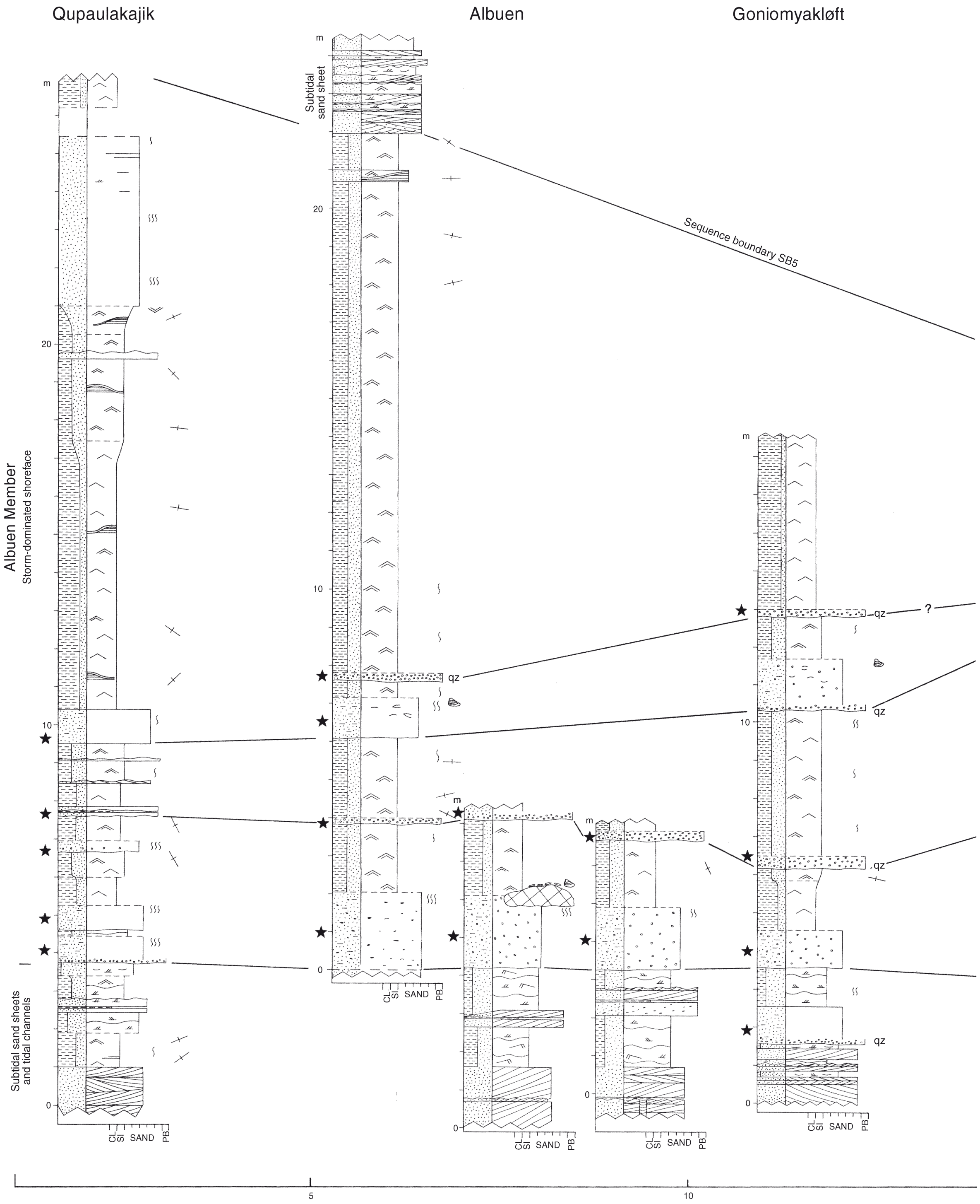

Fig. 20. Sedimentological logs through the debris flow and storm-dominated lower shoreface deposits of the Albuen Member. Asterisks mark the submarine debris flow deposits. Measured along Neill Klinter. See Fig. 1 for locations and Plate 1 for legend.

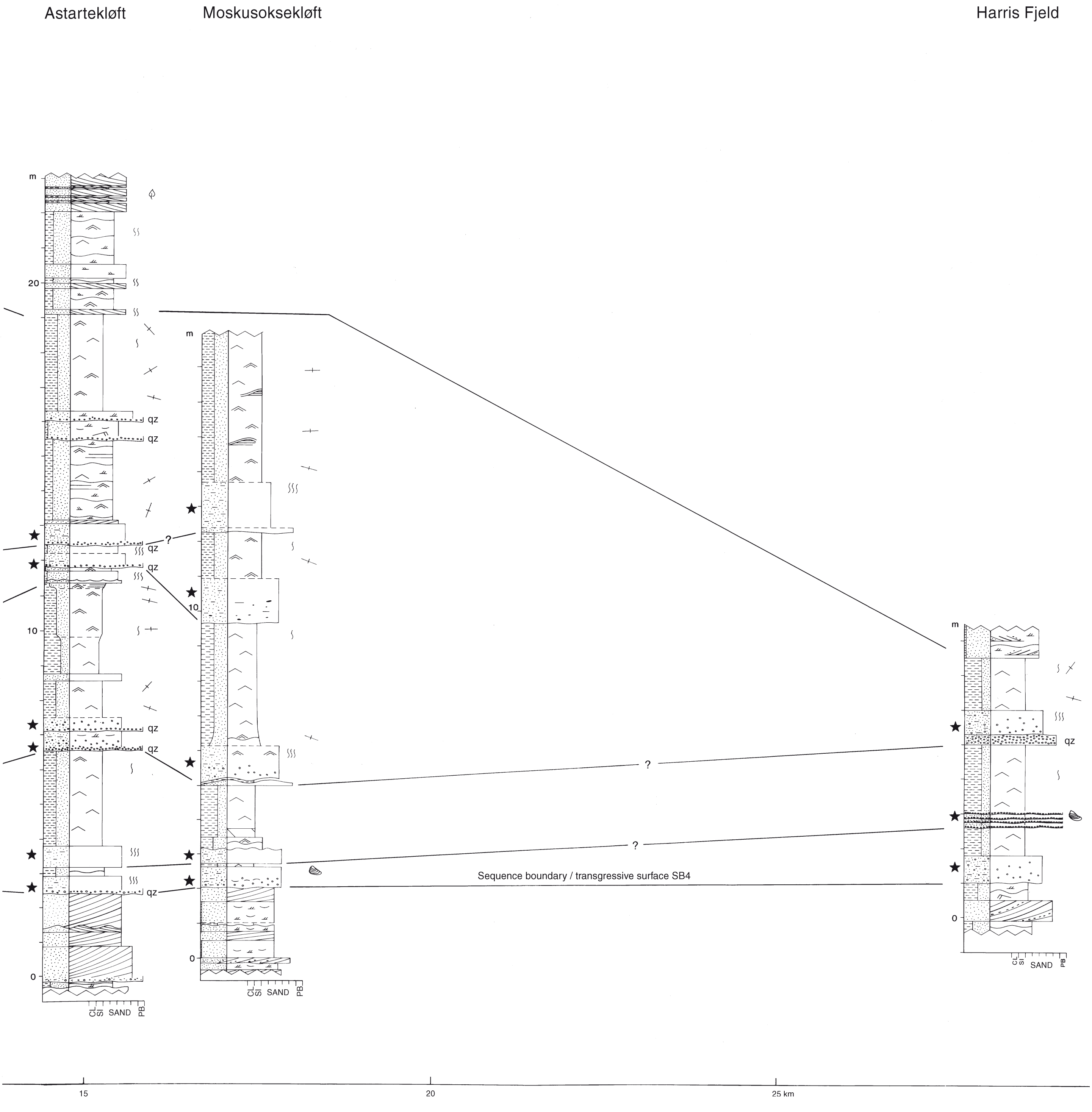
Astartekløft
Moskusoksekløft
Harris Fjeld
m
20
10
0
qz
?
Sequence boundary / transgressive surface SB4
CL
SI
SAND
PB
15
20
25 km

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).