A forced regressive shelf-margin wedge formed by transition-slope progradation: lowermost Cretaceous Rauk Plateau Member, Jameson Land, East Greenland

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The Middle Jurassic – lowermost Cretaceous succession of Jameson Land, East Greenland records a marine, overall regressive–transgressive–regressive cycle with regressive maxima in the Late Bajocian and Late Volgian separated by a transgressive maximum in the Kimmeridgian. Smaller-scale regressive interludes took place in the Late Callovian and Mid Oxfordian. A shelf-slope-basin physiography started to develop in the Late Callovian due to increasing rifting and a relief of several hundred metres was attained during maximum end-Jurassic regression and deposition of the Volgian Raukelv Formation. The formation consists of a forestepping stack of laterally extensive shelf-edge wedges, each several tens of metres thick, composed of coarse-grained sandstone, showing high-angle clinoform bedding and containing marine body and trace fossils. These clinoform beds are superimposed on the large-scale clinoforms of the shelf–slope–basin. The wedges onlap older shelf deposits in a landward direction and are overlain by thin transgressive sandstones or mudstones, or directly by the next coarse-grained wedge. The top wedge, comprising the Rauk Plateau Member, is of Late Volgian (i.e. earliest Cretaceous) age and is characterized by steep clinoforms truncated by internal erosional downlap surfaces. The clinoforms are simple avalanche beds, a few tens of centimetres thick, or they may be several metres thick and contain large-scale cross-bedded intrasets of probable tidal origin. The erosional events were associated with downshift of the succeeding clinoforms, recording minor sea-level fall and forced regression. The top surface of the Rauk Plateau wedge is incised by a system of minor channels leading to a large canyon-like valley. The wedge was deposited by transition-slope progradation below wave base during a period of sea-level stillstand punctuated by minor, stepwise falls. It provides an excellently exposed example of a laterally derived, coarse-grained shelf-margin wedge, showing high-angle clinoform bedding and representing an ancient counterpart to Holocene and Late Pleistocene prograding infralittoral wedges seen on seismic profiles across Mediterranean shelf edges.

Key words: Late Volgian, Cretaceous, East Greenland, Raukelv Formation, Rauk Plateau Member, shelf-margin wedge, transition-slope, forced regression.

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Outer shelf and shelf-edge sandbodies deposited during sea-level stillstand, fall and lowstand are well known from high-resolution seismic profiles across continental margins (e.g. Suter & Berryhill 1985; Tesson *et al.* 1990, 1993; Trincardi & Field 1991; Sydow & Roberts 1994; Morton & Suter 1996; Chiocci *et al.* 1997). The sequence stratigraphic position of strata deposited during sea-level fall and forced regression has been much debated (e.g. Posamentier *et al.* 1992; Hunt & Tucker 1992, 1995; Helland-Hansen & Gjelberg 1994; Kolla *et al.* 1995; Posamentier & Allen 1999; Plint & Nummedal 2000).

Documentation of outcrop examples of shelf-edge sandbodies ideally requires continuous lateral exposure of deposits from the terrestrial basin margin, across the shelf to the palaeo-shelf edge, and further over the slope to the basin. Such exposures are only rarely available and interpretation is in most cases based on the recognition of criteria indicative of forced regression, a basinally-isolated position of the

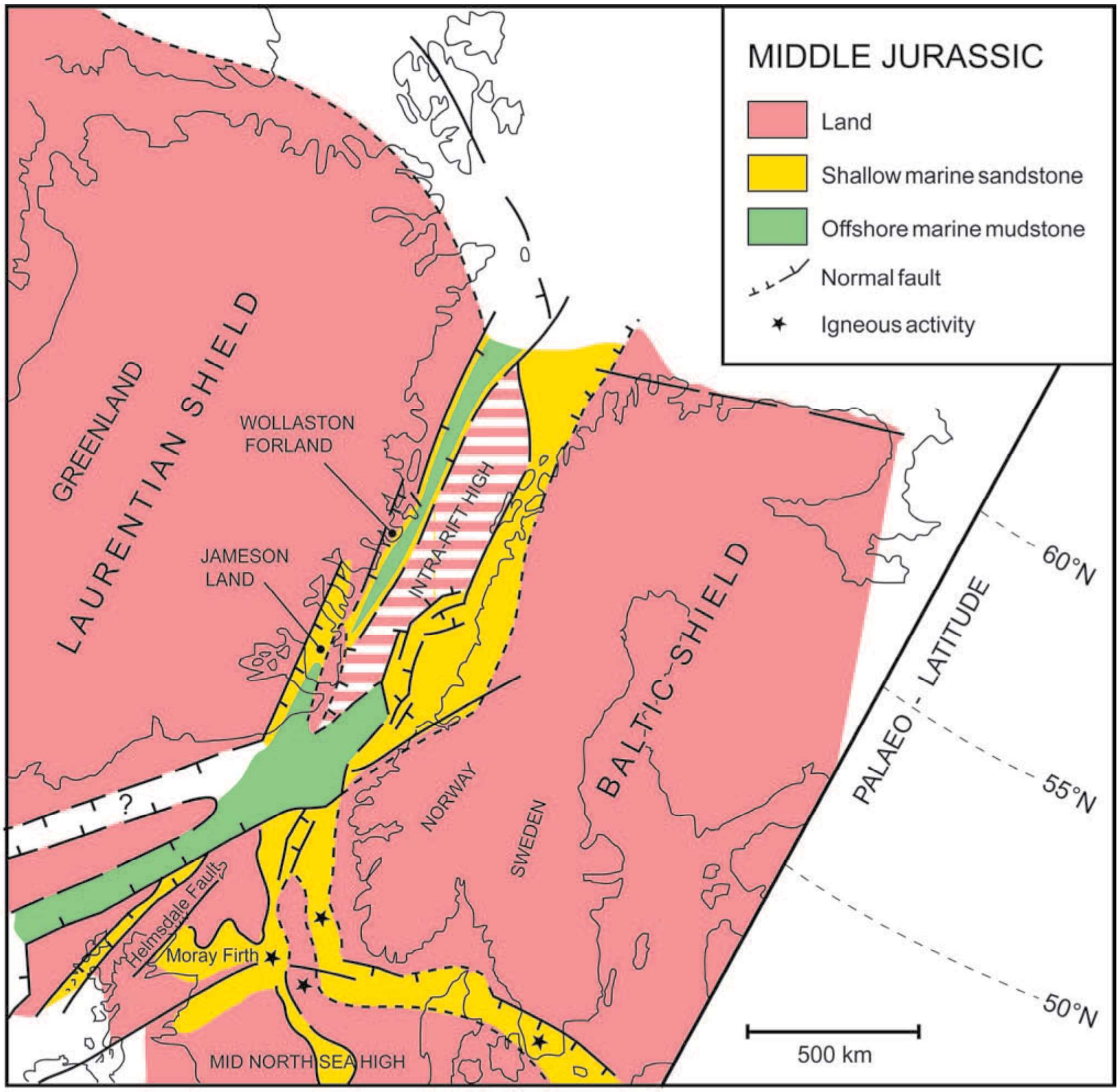


Fig. 1. The Late Jurassic rifted seaway between Greenland and Norway, continuing southward into the rifted straits of NW Europe. The positions of study area in Jameson Land, East Greenland is indicated. Based on Ziegler (1988), Doré (1992), and own data.

sandbody, and comparison with well documented recent examples.

Correct identification of shelf-edge sandbodies is of considerable economic importance (Posamentier *et al*. 1992). They are typically laterally extensive, oriented parallel to the coastline, basinally-isolated, and wedge out in a landwards direction where they onlap older shelf deposits. They commonly consist of clean sand overlying outer shelf and upper slope mud and overlain by fine-grained transgressive deposits. Most modern, Holocene and ancient examples of lowstand shelf-edge wedges are from passive continental margins of the Mediterranean Sea and the Gulf of Mexico.

Gravitational failure of shelf-edge sandbodies forms the source of massive sands transported by sliding, slumping and concentrated gravity flows to the slope, base-of-slope and proximal basin where they interfinger with muds. The Upper Jurassic Hareelv and Olympen Formations of East Greenland

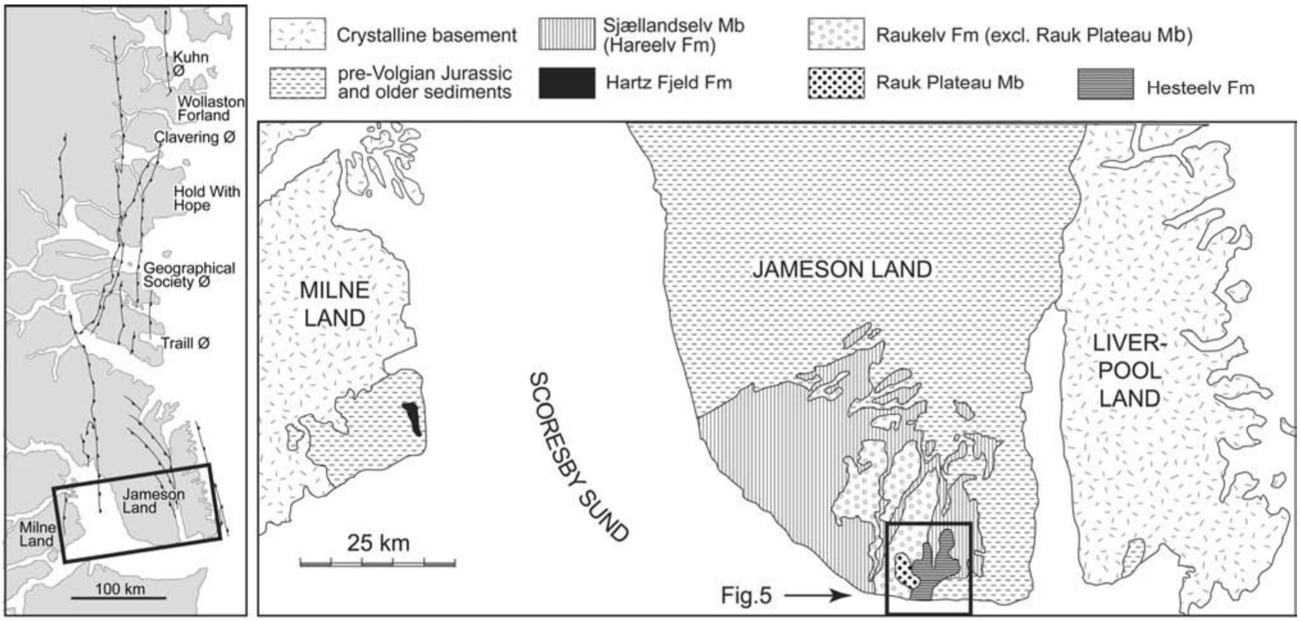


Fig. 2. Present-day distribution of the Raukelv Formation and the correlative lower Hennigryggen Member (Hartz Fjeld Formation) in southern Jameson Land and Milne Land, respectively. Distribution of the Rauk Plateau Member is indicated; note its position in the axial part of the Jameson Land Basin. The Hesteelv Formation constitutes the fill of a valley incised in the most distal shelf-edge part of the Rauk Plateau Member and the adjacent slope.

provide excellent examples of such interbedded gravity-flow sandstones and dark slope and basinal mudstones (Surlyk 1987; Surlyk & Noe-Nygaard 2001, 2003; Larsen & Surlyk, 2003; Bruhn & Surlyk 2004; Surlyk, Gjelberg & Noe-Nygaard *in press*).

The Upper Volgian (i.e. lowermost Cretaceous) Rauk Plateau Member is the youngest in a series of lowstand shelf-margin sandstone wedges of the Volgian Raukelv Formation exposed in southernmost Jameson Land (Surlyk & Noe-Nygaard 1991). Broadly correlative proximal coastal and shelf sandstones are located 80 km to the west in Milne Land where they are represented by the lower Hennigryggen Member of the Hartz Fjeld Formation (Figs 1, 2) (Birkelund *et al.* 1984; Surlyk *et al.* 1993). The Rauk Plateau Member is 27 m thick at the palaeo-shelf edge and wedges out completely over a few kilometres in an up-dip, landward direction and thus occupies a distal, basinally-isolated position. It is coarser grained, poorer sorted and lithologically more immature than the underlying wedges.

The aim of this study is to describe the coarse-grained, high-angle clinoform bedded shelf-margin wedge of the Rauk Plateau Member, to interpret the depositional mechanisms and sequence stratigraphy in terms of relative sea-level fluctuations during deposition. The member is a good ancient analogue to Holocene and Late Pleistocene prograding infralittoral wedges seen on seismic profiles across modern continental margins, notably in the Mediterranean Sea and the Gulf of Mexico.

Geological setting

In Jurassic – Early Cretaceous times, a rifted seaway existed between Greenland and Norway, and the north-south trending Jurassic rift basin of East Greenland was located along its western margin. The exposed onshore part of the East Greenland basin is about 600 km long, reaching a maximum width of 200 km in Jameson Land at its southern end (Figs 1, 2).

Jurassic rifting in East Greenland was initiated in the Late Bajocian, increased during the Late Oxfordian – Kimmeridgian, climaxed in the Volgian (latest Jurassic – earliest Cretaceous), and gradually faded out in the Ryazanian – Early Valanginian (Surlyk & Noe-Nygaard 2000; Surlyk 2003). The rift event was characterized by block faulting and tilting in areas north of Jameson Land, whereas the wide Jameson Land Basin behaved as a structural entity, characterized by more gentle westwards tilting and differential subsidence. The eastern and western margins of the basin were controlled by deep-seated master faults but the position of the coastlines shifted across the fault zones in response to changes in sediment influx and sea level. In Volgian tim, the Jameson Land

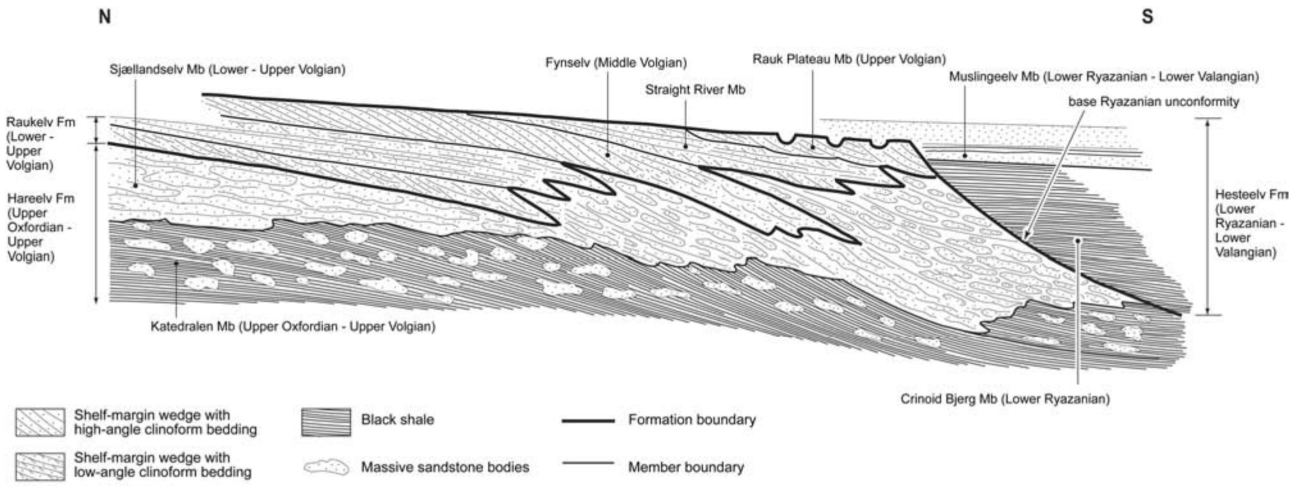


Fig. 3. Stratigraphic N–S section through the Upper Jurassic – lowermost Cretaceous succession of the Jameson Land area. Note the distal, shelf-edge position of the Rauk Plateau Member. Stratigraphic nomenclature based on Surlyk (2003).

Basin received a high input of very coarse-grained, immature, pebbly sand from the north and west forming the marine Raukelv Formation, the top part of which forms the subject of this study (Surlyk & Noe-Nygaard 1991; Surlyk 2003).

Basin physiography

The north–south trending axis of the Jameson Land Basin was essentially horizontal during the tectonically relatively quiescent Early Jurassic time interval. Onset of rifting in the Late Bajocian resulted in a complete reorganisation of drainage patterns and an axial southerly transport system was developed. A shelf-slope-basin topography with a southward dipping slope was formed in the Late Callovian and the greatest physiographic differentiation was reached in the Volgian (Larsen & Surlyk 2003; Surlyk 2003; Bruhn & Surlyk 2004; Surlyk *et al.* in press).

The shorelines followed the north–south orientated basin margins and changed to an east–west direction across the rift axis. The shorelines and associated shelf edges thus had an elongate horseshoe geometry opening towards the south. The western basin margin fault zone is today concealed by the inner Scoresby Sund fjord, but a Jurassic – lowermost Cretaceous outlier occurs in Milne Land to the west of the zone (Fig. 2). In Callovian–Oxfordian times the northern shelf edge was situated in central Jameson land and the western shelf edge in the Milne Land area. A relatively deep-water basin was located over the basin axis in southern Jameson Land.

The shelf edge prograded towards the east and south and reached the basin axial region in southernmost Jameson Land in the Late Volgian. The rate of Late Jurassic – earliest Cretaceous long term shelf progradation towards the south was of the order of 15 km/myr. Progradation was punctuated, however, by a number of flooding events during which the shoreline shifted northward and westward for tens of kilometres.

Basinal water depth in the intracratonic Jameson Land Basin was much smaller than at the foot of a passive continental margin and probably never exceeded a few hundred metres. The slope inclinations were at times as high as about 20°, especially in periods of rapid progradation when the coastline reached the shelf edge and caused oversteepening and failure of the upper slope (Surlyk & Noe-Nygaard 2001, 2003). The water depth at the base-of-slope was as much as about 130 m even during the latest stage of basin fill, increasing to several hundred metres further offshore to the south. It is important to distinguish the large-scale shelf-slope-basin clinoforms from the thinner clinoform-bedded shelf-margin wedges. The former are well exposed in southernmost Jameson Land (Surlyk & Noe-Nygaard 2001, fig. 4). The latter are almost an order of magnitude smaller and were formed by progradation of marine reworked deltas across the shelf to the shelf-slope break.

Upper Jurassic – lowermost Cretaceous stratigraphy

The Jurassic – lowermost Cretaceous lithostratigraphic nomenclature of East Greenland erected by Surlyk *et al.* (1973) is under revision and the revised framework was presented in Surlyk (1993). The new scheme is used for the deposits described here and is briefly commented upon because several of the units are interpreted quite differently in the older literature. The Upper Jurassic – lowermost Cretaceous of Jameson Land is referred to the Olympen, Hareelv and Raukelv Formations. The Lower – Middle Oxfordian Olympen Formation consists of shelf-margin wedge sandstones and associated slope mudstones and massive gravity flow sandstones (Larsen & Surlyk 2003; Bruhn & Surlyk 2004). The redefined Upper Oxfordian – Volgian Hareelv Formation (Katedralen, Sjællandselv and Salix Dal Members) is composed of slope and basinal mudstones and massive sandstones (Surlyk 1987; Surlyk & Noe-Nygaard 2001, 2003; Surlyk *et al. in press*), whereas the Volgian Raukelv Formation comprises marine shelf-margin wedge sandstones (Fig. 3). The Raukelv Formation was originally defined as composed of (from below) the Sjællandselv, Salix Dal and Fynselv Members (Surlyk *et al.* 1973). The Sjællandselv and Salix Dal Members now form the upper part of the Hareelv Formation. The Raukelv Formation is redefined and now comprises six members; the top member which forms the subject of this study is termed the Rauk Plateau Member (Fig. 3).

The Raukelv Formation as redefined was deposited over 7 Ma during an interval of overall eustatic lowstand at the end of the Jurassic Period (Hallam 1988, 1992; Haq *et al.* 1988; Jacquin *et al.* 1998; Surlyk 1990, 2003; Sahagian *et al.* 1996). The formation consists of four high-angle and two low-angle clinoform-bedded, coarse-grained, pebbly sandstone wedges with maximum thicknesses between 15 and 50 m (Fig. 3). They are interbedded with a few thinner clinoform sandstone beds and cross-bedded transgressive sheet sandstones. The main regressive-transgressive packages may thus represent time intervals of about 1 Ma each. Regional mapping of the clinoform beds forms the basis for subdivision of the formation into six members, the uppermost of which is the Upper Volgian Rauk Plateau Member (Figs 3, 4). The member has been traced over an area of 12 km^2 and overlies a southward thickening low-angle clinoform-bedded wedge, which is up to 45 m thick, and is composed of stacked tidal cross-beds showing palaeocurrents to the south.

Fig. 4. High-angle clinoform-bedded coarse-grained sandstone of the Rauk Plateau Member, southernmost Jameson Land. View towards the south-east.

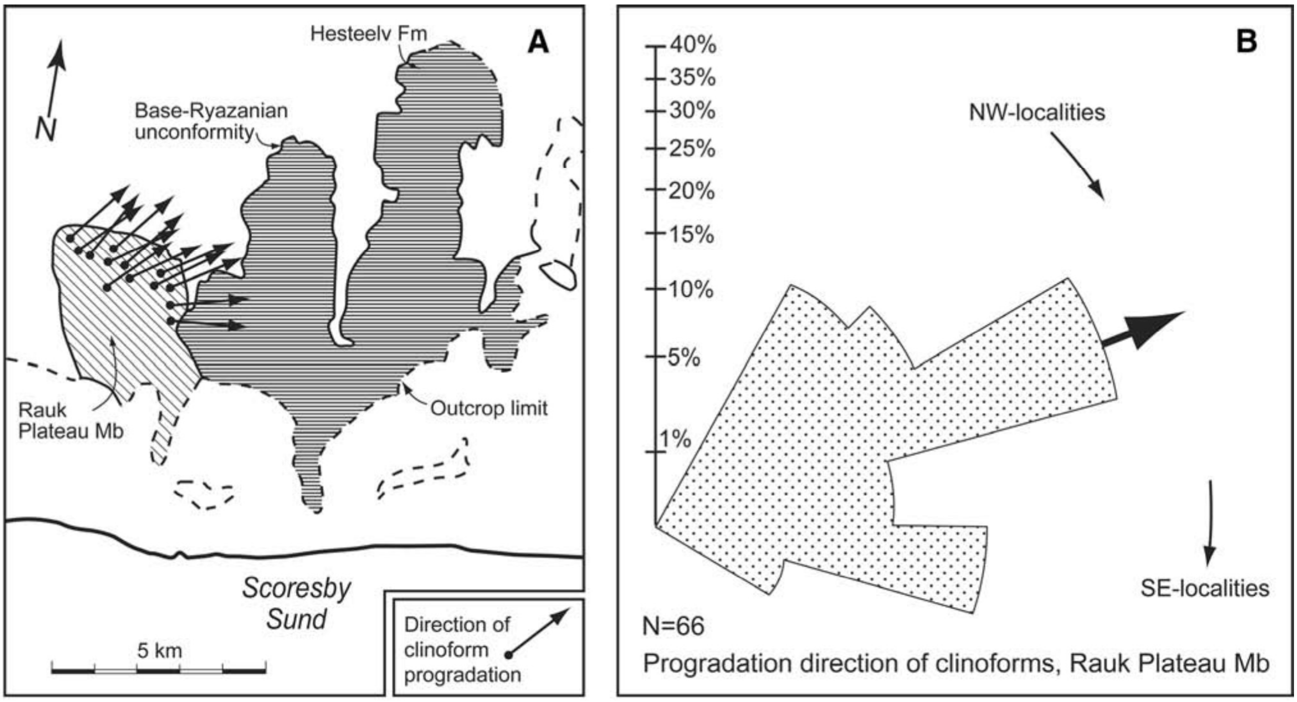


Fig. 5. (A) Clinoform progradation directions in the Rauk Plateau Member. (B) Equal area rose for all measurements of clinoform progradation directions. There is a slight change in progradation direction with time from north-east to east, going from northern over north-eastern to eastern outcrops.

Sedimentology

The Rauk Plateau Member is high-angle clinoform-bedded and prograded from the western basin margin towards east-northeast and thus represents a roughly shore-parallel lateral infill of the N–S trending rift basin (Figs 1, 2). The shelf edge which is marked by the most distal, south-eastern margin of the member is situated almost 10 km basinward of the previous shelf edge of the Middle Volgian Fynselv Member.

The member has a wedge-shaped geometry, is up to 27 m thick, and forms the top bed in most of the area of exposure. It wedges out or is cut out by erosion associated with the base-Ryazanian unconformity 3 km northwest of the northernmost outcrop margin where Ryazanian strata rest directly on older members of the Raukelv Formation (Fig. 2). The Rauk Plateau Member is composed of poorly sorted, coarse-grained, pebbly quartz sandstone with some feldspar and glauconite. The quartz grains are highly angular and seem to be first cycle, whereas the quartzite pebbles are very well rounded, flat and ellipsoidal in outline, and up to 10 cm long. The top surface which coincides with the base-Ryazanian unconformity is draped by a red-stained pebble lag with an average maximum pebble size of 5 cm (N=24), showing north–south and east–west long axis orientations.

The clinoforms of the member dip up to 25° and the progradation direction is towards east-northeast, averaging 067° (Figs 4–6). The clinoforms show tangential downlap with long asymptotic toes (Figs 4, 5), whereas their tops are truncated by the base-Ryazanian unconformity (Fig. 3). There are no indications of an original sigmoidal configuration of the clinoforms. The toesets show roughly the same grain sizes as the higher parts of the clinoforms and there are thus no marked vertical changes in grain size through the clinoform bed except for the pebble lag on the top surface.

The clinoforms may form bundles, in most cases some metres thick, separated by iron-stained, strongly burrowed surfaces, representing pause planes (Fig. 4). In some cases the pause planes are closely spaced with a distance of only 30 cm. The clinoforms are simple avalanche sets in the thinner bundles but in most cases they show large-scale trough cross-bedded intrasets, generally 0.5–1.5 m thick, but sets up to 3 m thick are not uncommon (Figs 4–10). The thicker trough cross-beds have gently curved bases, and tangential or more commonly sigmoidal foresets with preserved form sets and brinkpoints (Fig. 10A). The foresets commonly show a bundled upbuilding with the bundles separated by convex-upwards unidirectional reactivation surfaces (cf. de Mowbray & Visser 1984), but the coarse grain size and complete absence of clayey or silty laminae hamper unequivocal demonstration of tidal periodicity (Figs 8, 9). The cross-bedding is highly similar to that described from the Eocene Vlierzele Sands of Belgium which are interpreted as representing longitudinal tidal sand ridge deposits (Houthuys & Gullentops 1988). The cross-beds show a systematic upwards change in palaeocurrent direction from broadly south at the base of the clinoforms, over southeast in the middle, to east at the top (i.e. from about 180° to about 100°) (Figs 7, 8). They were formed by large dunes that migrated obliquely down the clinoforms and gradually changed orientation to a direction roughly parallel to the clinoforms and to the basin axis at the foot of the clinoforms. The dunes are interpreted as governed by tidal current regime and were almost totally built up by the dominant tidal current. However, bidirectional 'herring-bone' type cross-beds also occur with palaeocurrent directions towards the north-northeast and south-southwest.

The currents forming the cross beds most likely had a strong tidal component as suggested by the sigmoidal foresets, reactivation surfaces, bundlewise upbuilding, and occasional bidirectional palaeocurrent directions. The underlying clinoform beds of the Raukelv Formation (e.g. the Fynselv and Straight River Members, Fig. 3) show similar large-scale, commonly sigmoidal cross-bedding with a distinct tidal influence with dominant southern and subordinate northern current directions (Surlyk & Noe-Nygaard 1991).

Large, transported plant fragments, including branches, twigs and leaves occur abundantly in the clinoform beds especially along the burrowed pause planes. Trace fossils occur scattered and include vertical *Ophiomorpha nodosa*, *Planolites* and *Diplocrateri-*

Fig. 6. Coarse-grained sandstones of the Rauk Plateau Member showing high-angle clinoform bedding. Erosion surface (dashed) truncates the lower part of the clinoforms to the right (arrows) and gradually becomes conformable with the upper part of the clinoforms. The downlap surface of the older clinoform package to the right is situated at the base of the section (DLP), whereas the downlap surface of the younger clinoform package to the left is situated several metres lower. The upper boundary of the unit is an erosion surface which coincides with the regional base-Ryazanian unconformity. Person for scale (in box). Vertical arrow shows position of section shown in close-up in Figures 7–8.

Fig. 7. Oblique view of section shown in Figs 6 and 7. Close-up of the section arrowed in Figure 6. Note the truncation surface which is seen in both dip and strike view. Note also the cross-bedded intrasets of the clinoforms which shows palaeocurrent directions to the right (south), almost perpendicular to the clinoform progradation direction and the dip of the truncation surface.

on habichi, a metre-long, vertical U-shaped burrow which occurs in great densities, extending downward from the top surface of the clinoform bed (177/m^2 and 218/m^2 counted on two bedding plane examples).

The regular succession of conformable clinoforms is commonly truncated by erosion surfaces which are conformable with the upper part of the clinoforms and become progressively more discordant towards their base (Figs 6–8, 11A). The truncation surfaces show roughly the same dip directions as the clinoforms. Thus, in one example the dip directions of a truncated clinoform and the truncation surface are both towards the east-northeast (68º). The intrasets in the clinoform below this truncation surface show palaeocurrent directions changing from south (185º) at the base, over southeast (160º, 148º, 124º) to east-southeast (107º) at the top of the clinoform.

The number of truncation surfaces and the horizontal distance between them is not known but several surfaces have been recognized in a 1 km long lateral profile. The base of the member shows downwards shifts of the order of a few metres across each truncation surface and the lower boundary is thus a composite downstepping erosion surface (Figs 6, 11A).

A network of tributary channels is incised in the top surface of the member, terminating distally in an up to 125 m deep and 12 km wide canyon-like valley cut into the distal south-eastern margin of the member. The channels are mainly 5–10 m deep, commonly sinusoidal with sharp bends, and their margins dip up to 24º (Fig. 12A). The channels do not show any signs of lateral migration and thus occupied a stable position. The channel floors are iron-stained, pebble strewn and densely burrowed by *D. habichi* (Figs 12B, 13). The staining causes an intense bright red weathering colour which eases identification of channels and adjacent interfluve areas. The axis of the wide incised valley plunges towards the south and the valley is filled with an onlapping coarsening-upward succession, up to 125 m thick, of dark marine mudstones and sandstones forming the Ryazanian Hesteelv Formation (Figs 2, 3) (Surlyk *et al.*, 1993).

Environmental interpretation of the Rauk Plateau Member

The high-angle clinoform-bedded wedge of the Rauk Plateau Member is highly similar to a Gilbert-type delta in terms of the steep angle of the clinoforms, coarse grain-size, angular grains, poor sorting, and high content of terrestrial plant remains. This interpretation is contradicted, however, by the basinally-isolated position, the marine trace fossil assemblage, the commonly sigmoidal large-scale cross-bedded nature of the intrasets, the overall longshore southward palaeocurrent direction of intrasets at the base of the clinoforms, and the complete absence of features indicative of emergence or shallow water such as wave- and storm-induced structures. Progradation is thus thought to have taken place below wave base in the shoreface-transition zone of Pomar & Tropeano (2001). The Fynselv Member (Fig. 3), which is laterally very extensive and equally clinoform-bedded shows no signs of wave or storm action, contains a rich marine fauna, has sigmoidal clinoforms, and a sheet-like geometry parallel to the shoreline and was formed by seawards progradation towards the east. This indicates that the Fynselv Member does not represent a Gilbert-type delta but was formed by shoreface-transition zone progradation below wave base (Surlyk & Noe-Nygaard 1991).

The very coarse-grained probably fluvially-derived sand and gravel was thus not deposited directly at the mouth of the river but was reworked by strong, probably tidally-enhanced, mainly southwards flowing, coast-parallel marine currents to be deposited in relatively deep water as a shelf-margin wedge completely detached from the river mouth. Modern steep-slope, marine-reworked deltas are few in number but good examples which show great similarity with the high-angle clinoform-bedded members of the Raukelv Formation have been described from Canadian fjords by Hart & Long (1996). The presence of a narrow U-shaped gulf open to the south seems to have had an amplifying effect on tidal currents throughout Mid and Late Jurassic time intervals characterized by low sea level (Surlyk 2003; Surlyk & Noe-Nygaard 1991; Surlyk *et al.* 1993; Engkilde & Surlyk 2003).

The internal truncation surfaces and associated downshifts of the basal downlap surface reflect erosion of the lower part and toe of the clinoforms by enhancement of the southwards flowing marine currents following minor relative sea-level falls of a few metres. During renewed progradation the truncation surface was downlapped by the succeeding clinoforms which built outwards and downwards with respect to the eroded clinoforms. There is no evidence of fluvial erosion during progradation and the upper surface of the clinoform bed first became emergent and subject to fluvial incision after the system had prograded to the shelf edge during stepwise relative sea-level fall.

Fig. 8. Close-up of strike section of the clinoforms shown in Figure 6 where the position of the section is indicated by a vertical arrow. Large-scale cross-bedded intrasets of the clinoforms show palaeocurrents almost perpendicular to the clinoform progradation direction. Note the gradual deflection of the palaeocurrent direction upwards through the clinoform set. The foresets of the lower cross beds dip to the right out of the picture (south), those in the middle to the right parallel to the plane of the picture (southeast), those in the upper part to the right obliquely into the picture, and those at the top into the picture (east) almost at a right-angles to the lowest cross beds. Note bundled upbuilding of foresets in several beds and convex-upwards reactivation surfaces.

The incised valley and associated channels were formed after deposition of the Rauk Plateau Member by fluvial erosion and shelf margin collapse, following a major base-level fall in the latest Volgian,

Fig. 9. Bundled upbuilding and reactivation surfaces of large-scale planar cross-bedded intrasets in the clinoforms seen in Fig. 8.

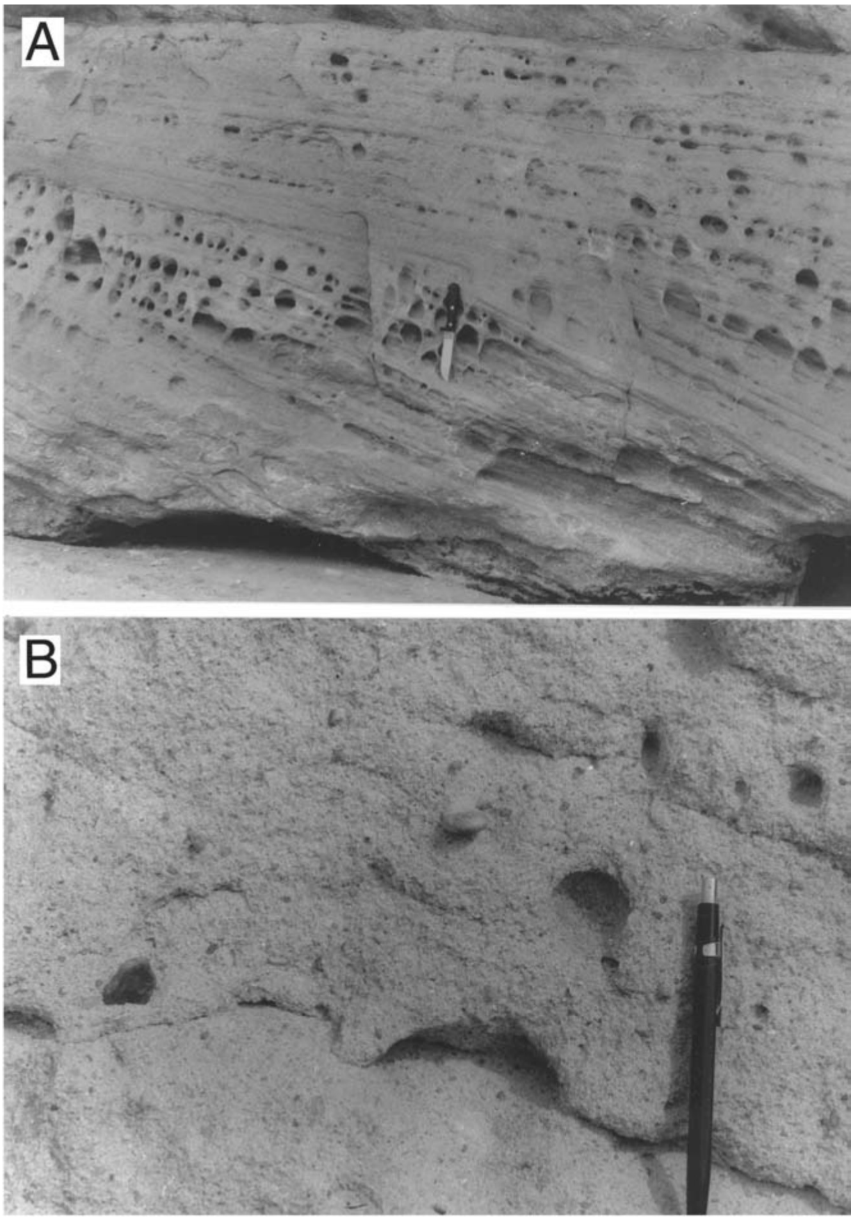
when the top of the member became emergent. The purely erosional, incised nature of the channels indicates that base-level fall was rapid, creating a system of permanent, non-migrating channels (cf. Posamentier & Allen 1999; Hart & Long 1996).

Sequence stratigraphy

The Rauk Plateau Member was deposited during the culmination of the long-term eustatic sea-level fall during the latest Jurassic (Hallam 1988, 1992; Haq *et al.* 1988; Surlyk 1990, 2003; Sahagian *et al.* 1996; Jacquin *et al.* 1998). Maximum lowstand in East Greenland occurred in the latest Volgian, and culminated with the formation of the base-Ryazanian unconformity, which forms the upper boundary of the member. Most of the depositional Jameson Land basin was subaerially exposed during this lowstand and the sea became restricted to a narrow U-shaped gulf located over the basin axis and open to the south. The gulf was only about 30 km wide at this time, and west of the gulf the 80 km wide shelf became subaerially exposed. The submerged part of the basin thus attained the physiography of an estuary. This is a significant feature of marine rift basins which is commonly overlooked. Estuaries are characteristically – and according to many definitions – formed during sea-level rise by transgressive drowning of incised river valleys (see review by Perillo 1995). However, similar narrow, funnel-shaped gulfs commonly form in marine rift basins during maximum sea-level fall. This development of similar physiographies under opposite movements of sea level is important because dominance of certain oceanographic processes is commonly linked to different segments of a sea-level cycle. Several authors have thus suggested that development of strong tidal systems is mainly associated with early sea-level rise when incised river valleys are flooded and estuaries formed (e.g. Dalrymple *et al.* 1992; Mellere & Steel 1995; Mellere 1996).

The dip directions of the high-angle clinoforms indicate that the shelf-margin wedge of the Rauk Plateau Member prograded towards the east-northeast away from the western basin margin (Fig. 5). The clinoform slopes and probably also the top of the unit – by analogy with all the underlying clinoform beds – were covered by large subaqueous dunes that migrated towards the south, driven by strong, probably tidally enhanced, coast-parallel bottom currents.

Fig. 10. (A) Cross-bed with sigmoidal foresets forming an intraset in the lower part of the clinoform bed of the Rauk Plateau Member at the same locality as Figure 6. Note the internal convex-up reactivation surface. Knife for scale, 20 cm long. (B) Close-up view of foreset, showing its coarse-grained, granule-rich nature. Pencil for scale; part shown on picture 12 cm long.



The currents were deflected towards eastern directions down the clinoform slopes and back again towards the south-east and south at their base.

The iron-stained, burrowed pause planes separating clinoform bundles were formed during periodic stillstands in progradation. The more widely spaced erosional downlap surfaces marking downward shifts of several metres of clinoform packages are interpreted to represent minor sea-level falls during progradation of the whole member.

The channel network on the top of the member is interpreted to have been incised by small rivers during a phase of rapid sea-level fall, leading to emergence. The valley at the distal margin of shelf-margin wedge was formed during maximum sea-level fall and was probably strongly augmented by slumping of valley walls at the shelf-slope break and upper slope. Further deepening of the valley and its walls took place by marine transgressive current and wave erosion during subsequent sea-level rise. The eroded top surface of the member and the bottom of the incised valley form a major sequence boundary.

The canyon-like valley was filled with a fully marine coarsening-upward mudstone–sandstone succession of the Hesteelv Formation, constituting a lowstand prograding wedge (Fig. 3). Marine sedimentation was resumed on the former shelf when the valley margins, tributary channels and interfluve areas eventually became flooded. The coincident sequence boundary and flooding surface of the interfluve area

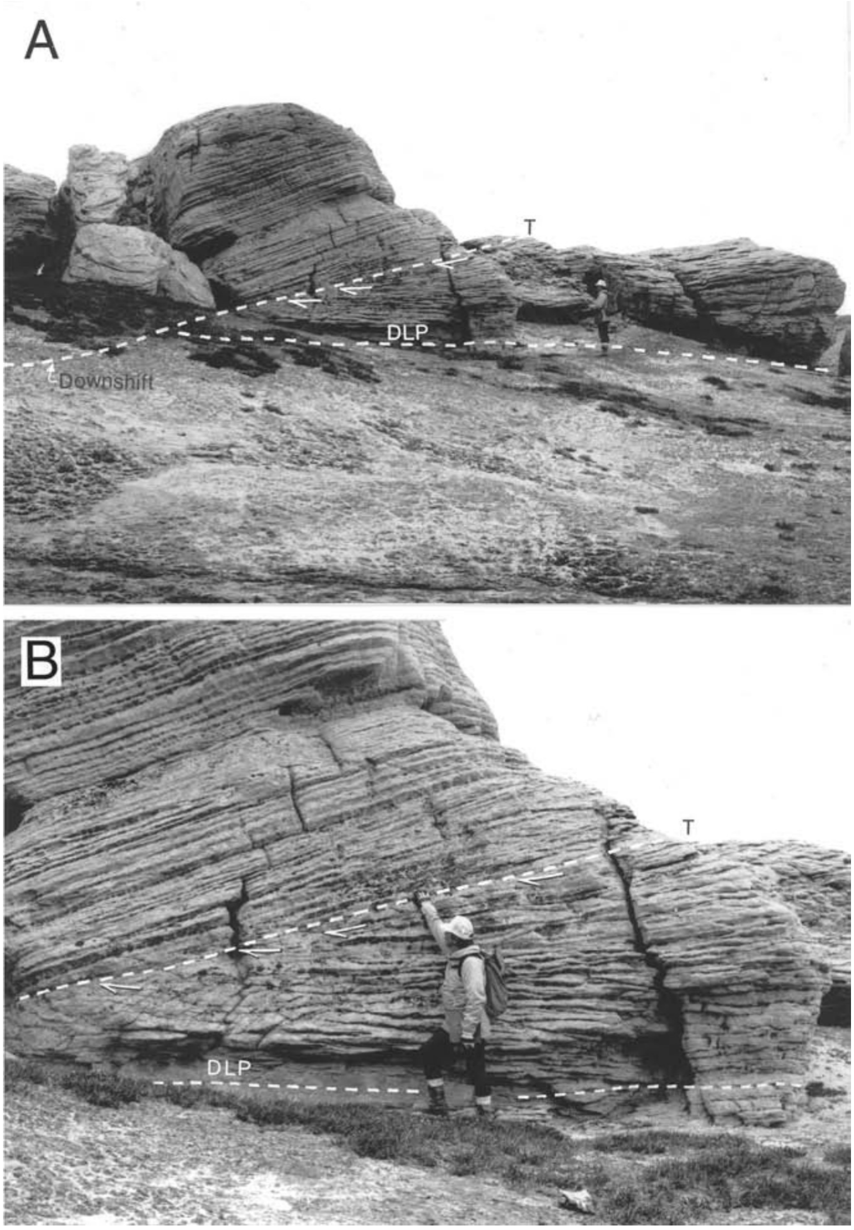


Fig. 11. (A) Clinoform truncation surface (T; arrows) cutting the toesets of the clinoforms to the right. Downlap surface (DLP) of the older clinoform bed to the right at the level of the person's feet. The downlap surface of the younger clinoform to the left is situated several metres lower, as marked by the oblique dashed line. (B) Detail of section in A, showing the basal downlap surface (DLP) of the older clinoform bed and the truncation surface (T) cutting the toesets of this unit (at person's hand). The progradation direction of the clinoforms is towards east-northeast (70°), whereas the trough cross-bedded intrasets show palaeocurrents towards southeast (135° and 144°).

is overlain by coarse-grained sandstones of the same lithology as the Rauk Plateau Member, making identification of the incised channel system rather difficult. However, they are separated by the prominent pebble lag and the overlying sandstones are structureless and represent slope and base-of-slope deposits formed during renewed progradation after the important flooding event that drowned the Rauk Plateau Member wedge.

The Rauk Plateau Member shelf-margin wedge was thus formed during late highstand and succeeding falling stage and is bounded below by a downstepping basal surface of forced regression (terminology after Hunt & Tucker 1992). The upper boundary of the member is a coincident subaerial erosion surface and marine transgressive surface of erosion, and represents a major, regional sequence boundary. The member is thus interpreted as being composed of a highstand and a forced regressive or falling stage systems tract bounded below by the composite, downstepping regressive surface of marine erosion, and above by a sequence boundary. The lowstand systems tracts of this and the preceding sequences are represented by thick, massive density flow sandstones of the Sjællandselv Member (Hareelv Formation) exposed a few kilometres to the east of the distal margin of the Rauk Plateau Member (Fig. 3).

The internal truncation surfaces separate lozenge-

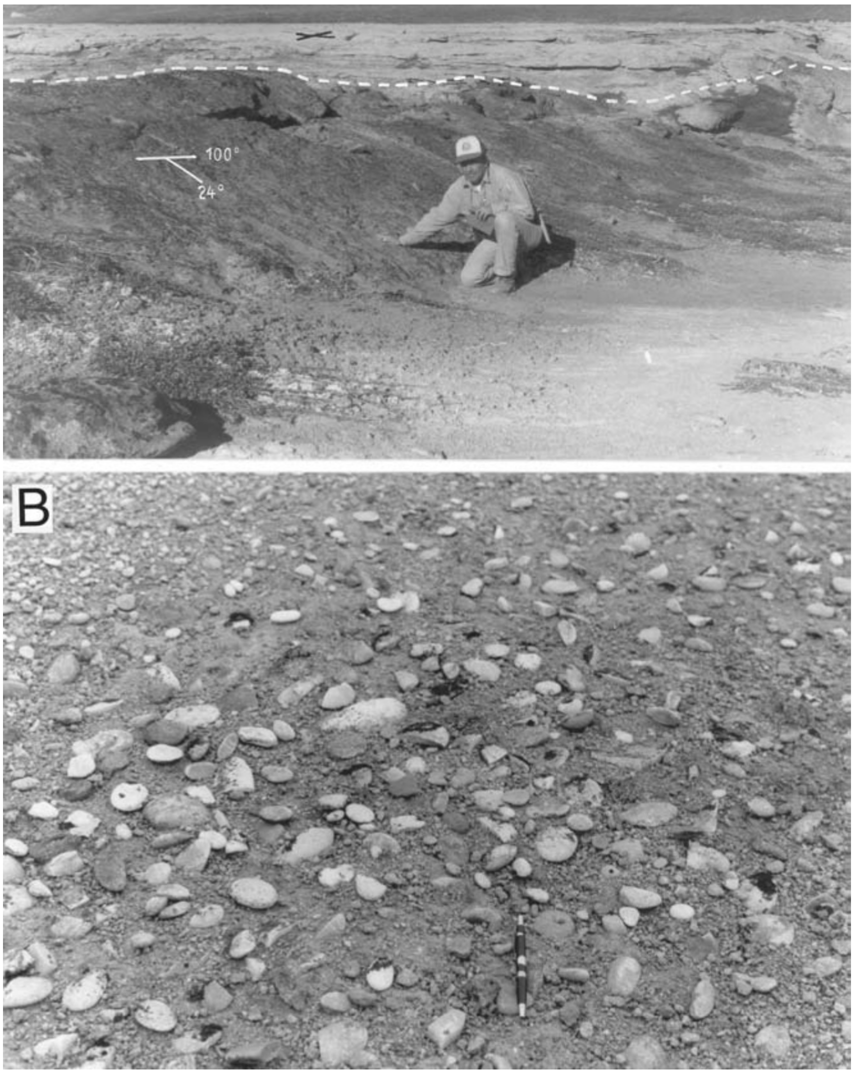


Fig. 12. (A) Margin of fluvial channel incised in the top of the Rauk Plateau Member, forming the base-Ryazanian unconformity together with the adjacent interfluve. The channel is part of a system of channels leading to a major valley incised in the margin of the member and the adjacent upper slope. Strike and dip of channel margin are 100° and 24°, respectively. The channel floor is covered with a pebble lag, is iron-stained and weathers bright-red. Logs and other plant fragments are common in the lag. The channels are all erosional and incised and are not associated with lateral accretionary bedding. (B) Quartzite pebble lag on the erosional sequence boundary surface at the top of the Rauk Plateau Member. Note the well rounded ellipsoidal outline of most pebbles. Pencil (lower centre) for scale, 14 cm long.

shaped falling stage systems tracts of higher order (terminology after Plint & Nummedal 2000) (Figs 6, 7, 11A). The toes of the truncation surfaces form the complex basal surface of forced regression of the lower order sequence represented by the whole member. Deposition of the higher-order units was initiated by onset of short-term sea-level fall causing erosion and downshift, whereas the younger part of the units was deposited during stable lowstand. The higher-order units thus include combined falling stage and lowstand systems tracts which are not separated by a marked internal boundary (Fig. 14).

The Holocene lowstand deltas described by Hart & Long (1996) from Canada are excellent analogues for the Rauk Plateau Member. The Natashquan delta lobe formed entirely subaqueously and was shifted about 10 km away from the river mouth by longshore currents. The clinoforms of the lobe show similar internal erosion surfaces with downlap as described here from the Rauk Plateau Member (Hart & Long 1996; their fig. 3) and are likewise interpreted to record falling sea level. The top of the Canadian deltas was incised by the fluvial channels during rapid sea-level fall and the channels remained stable without showing lateral migration, similar to the channels incised in the top of the Rauk Plateau Member (Fig. 12A).

Pomar & Tropeano (2001) described large-scale cross beds of the upper Pliocene – lower Pleistocene Calcarenite di Gravina from southern Italy, showing great similarity to the clinoform beds of the Raukelv Formation, including the Rauk Plateau Member (Surlyk & Noe-Nygaard 1991). The cross beds were interpreted to represent avalanches of sediment swept out onto a depositional slope below wave base. They were termed transition-slope deposits as the term shoreface was restricted to cover only the more proximal facies dominated by wave traction. The Calcaren-

Fig. 13. Bedding plane view showing densely spaced *Diplocraterion habichi* burrows. Note that the burrows occur in pairs, representing cross sections of the metre-long U-burrows.

ite di Gravina transition-slope cross beds also show internal downlap surfaces formed during sea-level fall, similar to the Canadian deltas of Hart & Long (1996) and the Rauk Plateau Member. They are coarse grained and form laterally extensive bodies which are parallel with the shoreline and prograded seaward with large-scale foresets or clinoforms dipping up to 35°. The beds showing internal downlap surfaces were interpreted as simple sequences, comprising highstand and falling stage systems tracts separated by the downlap surface (Pomar & Tropeano 2001).

Conclusions

The Upper Volgian Rauk Plateau Member represents a marine-reworked, high-angle clinoform-bedded and coarse-grained shelf-margin wedge. It was formed by transverse, eastwards progradation to the shelf edge during a long-term sea-level fall which started in latest Jurassic, Early Volgian time and culminated in earliest Cretaceous, Late Volgian time. Failure of the shelf edge resulted in shedding of highly concentrated sandy density flows down the slope and into the proximal basin forming thick massive sandstones of the Sjællandselv Member.

The Rauk Plateau Member is characterized by steep clinoforms showing internal, probably tidally-influenced large-scale cross bedding. The clinoforms are truncated by internal erosion surfaces that are conformable with the upper part of the clinoforms and truncate the lower part. The erosional events were associated with downstepping and downlap of the base of the succeeding clinoforms, and are interpreted to record minor sea-level fall and associated forced regression. The top surface of the member is incised by a system of minor fluvial channels leading to a large canyon-like valley.

The member is interpreted to comprise a highstand and falling stage systems tract bounded below by a composite regressive surface of marine erosion, and above by a sequence boundary. The lowstand systems tract is represented by thick, massive mass flow sandstones deposited at the base-of-slope and belonging to the Sjællandselv Member. The falling stage systems tract consists of forestepping and downstepping higher order units bounded below by a forced regressive surface of erosion formed during short-

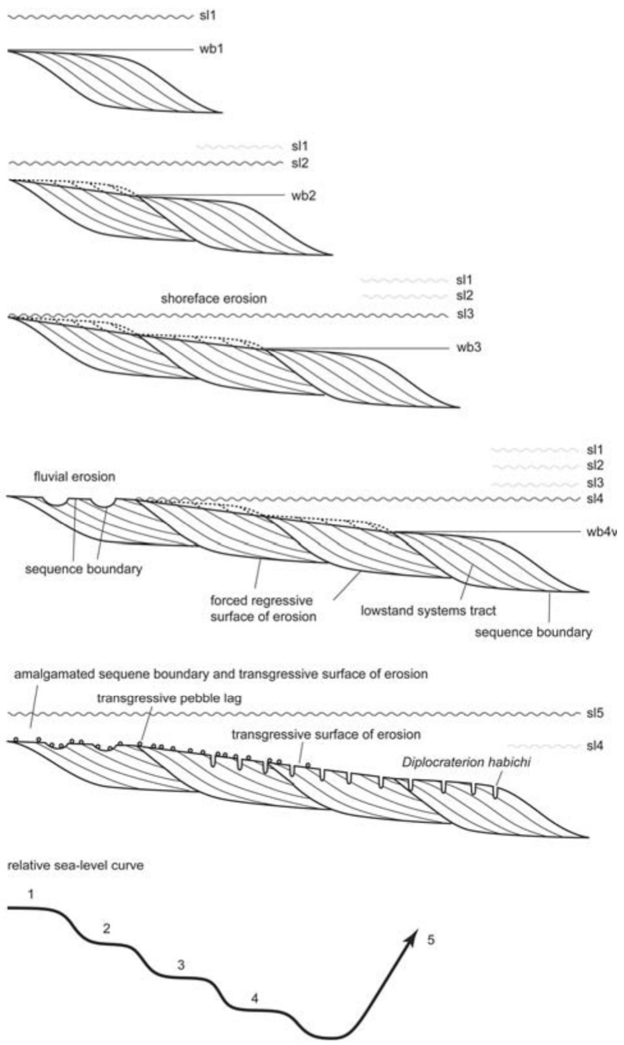


Fig. 14. Model for the formation of the erosion surfaces, truncating the high-angle clinoform bedding based on sections in Figures 6 and 11. Progradation of the shelf-margin wedge during sea-level stillstand was interrupted by a minor sea-level fall, causing erosion of the clinoform toes and downstepping of the succeeding clinoforms. The erosion surface is a forced regressive surface of marine erosion. Successive falls resulted in stepwise downstepping of the base of the member which thus represents a composite surface of regressive marine erosion. Sl 1–5 indicate successive sea levels and wb1–5 the corresponding wave bases. The corresponding relative sea-level curve shows a longer-term fall punctuated by four falls followed by a rise, during which the whole succession is truncated at the top during formation of a transgressive surface of erosion. Numbers 1–4 indicate phases of relative sea-level stillstand and progradation.

term sea-level fall and lowstand. The high-order unit thus includes combined falling stage and lowstand systems tracts which are not separated by any marked internal boundary.

The Rauk Plateau Member provides an excellently exposed example of a laterally derived, coarse-grained shelf-margin wedge, showing high-angle clinoform bedding and represents an ancient counterpart to Holocene and Late Pleistocene prograding infralittoral wedges commonly seen on seismic profiles across shelf edges in the Mediterranean Sea and the Gulf of Mexico.

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