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Shelf-margin clinoforms and prediction of deepwater sands

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ABSTRACT

Early Eocene successions from Spitsbergen and offshore Ireland, showing well-developed shelf-margin clinoforms and a variety of deepwater sands, are used to develop models to predict the presence or absence of turbidite sands in clinoform strata without significant slope disturbance/ponding by salt or mud diapers. The studied clinoforms formed in front of narrow to moderate width (10–60 km) shelves and have slopes, 2–4°, that are typical of accreting shelf margins. The clinoforms are evaluated in terms of both shelf-transiting sediment-delivery systems and the resultant partitioning of the sand and mud budget along their different segments. Although this sediment-budget partitioning is controlled by sediment type and flux, shelf width and gradient, process regime on the shelf and relative sea-level behaviour, the most tell-tale or predictive signs in the stratigraphic record appear to be (1) sediment-delivery system type, (2) degree of shelf-edge channelling and (3) character of shelf-edge trajectory through time. The clinoform data sets from the Porcupine Basin (wells and 3-D seismic) and from the Central Basin on Spitsbergen (outcrops) suggest that river-dominated deltas are the most efficient delivery systems for dispersing sand into deep water beyond the shelf-slope break. In addition, low-angle or flat, channelled shelf-edge trajectories associate with co-eval deepwater slope and basin-floor sands, whereas rising trajectories tend to associate with muddy slopes and basin floors. Characteristic features of the shelf-edge, slope and basin-floor segments of clinoforms for these trajectory types are documented. Seismic lines along the slope to basin-floor transects tend to show apparent up-dip sandstone pinchouts, but most of these are likely to be simply sidelap features. Dip lines aligned along the axes of sandy fairways show that stratigraphic traps are unlikely, unless slope channels become mud-filled or are structurally partitioned. Another feature that is prominent in the data sets examined is the lack of slope onlap. During the relative rise of sea level back up to the shelf, the clinoform slopes are generally mud-prone and they are characteristically aggradational.

INTRODUCTION

Deepwater sands, generated largely by a variety of sediment-gravity flows, accumulate in a range of slope and basin-floor environments. They are particularly common on the deepwater shelf margins of basins. They can sometimes be seen in clear relationship with a delivery system from an adjacent shelf edge, and in such cases, the nature of the linkage between the shallow-water and deep-water areas can be determined by critical examination of the different segments (shelf, slope and basin floor) of the shelf-margin system. Although the term 'clinoform' can be used for sigmoidal-shaped surfaces on a wide range of spatial scales, we use it here to refer to the sigmoidal shape of the surface profile (time line) of accetionary shelf margins. These clinoforms have an amplitude of 100–1000s of metres, in contrast to delta clinoforms that are normally less than 100-m high. The term 'clinoform' is therefore used here in a broader way than originally intended by Rich (1951). Rich (1951) used the term only for the 'slope' part of the sigmoidal feature whereas we are using it to include the more flat-lying, adjacent shelf and basin-floor segments of the feature. The term clinothem should be used when referring specifically to the rock body and its lithological detail (Rich, 1951). The 'topset' of the clinoform thus represents the shallow-water morphological shelf-platform; the 'slope' of the clinoform is the shelf margin that grades down into deep water, whereas, the 'bottomset' represents the deepwater toe-of-slope to basin-floor areas. In deepwater basins where water depth exceeds approximately 1 km or so, the clinoform slopes can be irregular and ponded because of salt or mud movement (e.g. Prather *et al*., 1998). We follow Swift & Thorne (1991) and Steel & Olsen (2002) in arguing that the term 'shelf' should not be restricted to the sedimentary, accretionary platform that forms at the margins of continents, but that it be used for the analogous shallow-water platforms that

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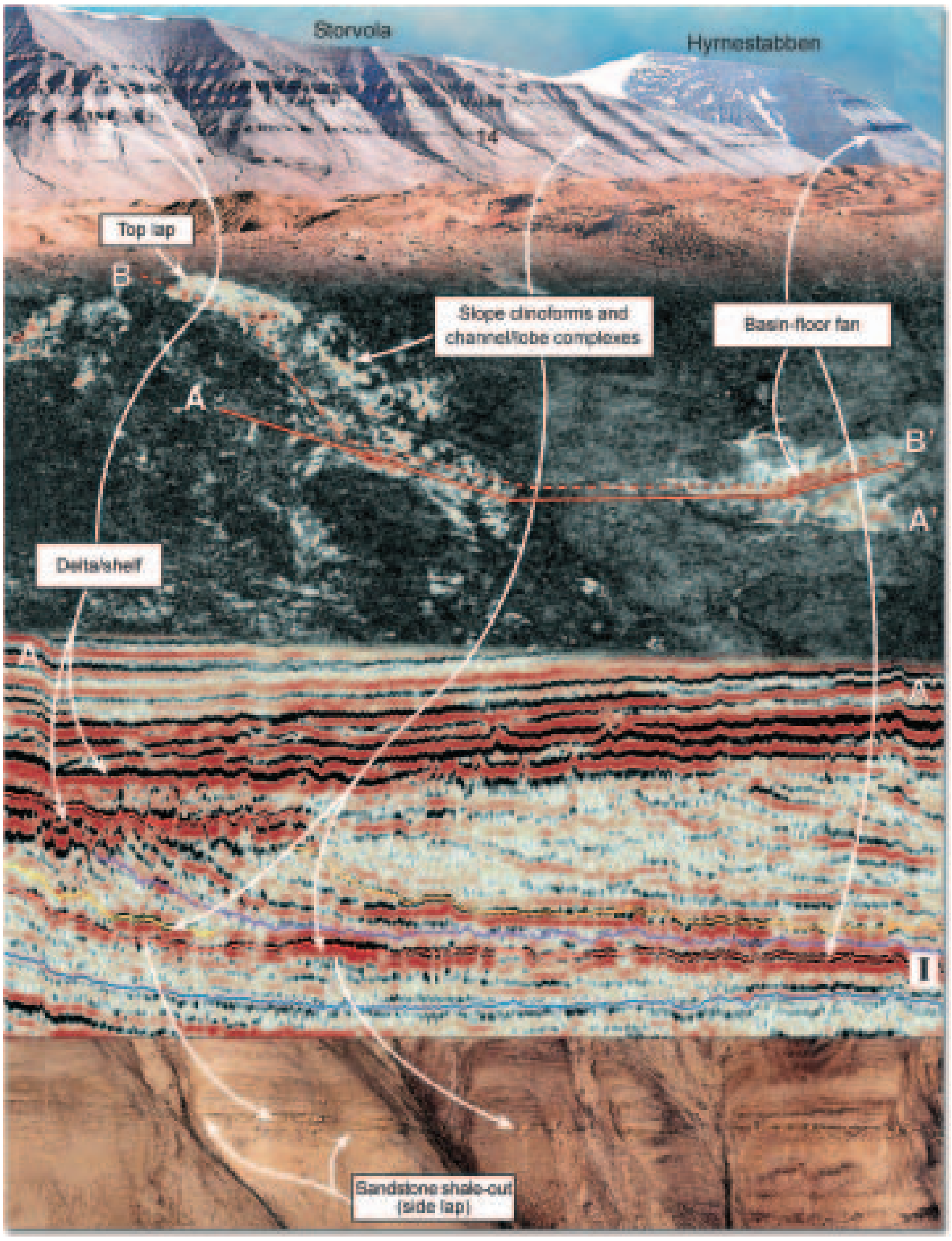

Fig. 1. Comparison of clinoforms and sandstone geometries of Porcupine Basin (illustrated with a flattened 3-D seismic line and amplitude map) and 2-D seismic-scale outcrops from Spitsbergen. Clinoform I (red horizon) on the seismic is compared with the mountainside Clinoform 14, dipping from left to right on Storvola. Several single-cycled high-amplitude reflectors pinching out at toe of slope are compared with sandstone bodies (slope channels) in shales at the base of the Storvola clinoform. On the horizon attribute map along clinoform I, the basin-floor reaches of the clinoform show a fan-like feature analogous to the Hyrnestabben fan seen at the end of the Storvola clinoform. The attribute map also shows a channel–lobe complex on the slope of the clinoform and indicates that the high-amplitude terminations on the seismic line A–A′ are side-lapping and not slope-onlapping. This is verified on transect B–B′ that shows a high-amplitude clinoform from basin floor to the shelf edge, thus increasing the risk for hydrocarbon trapping in this clinoformed slope setting (see Fig. 10a). 2-D sandstone geometries and facies can thus be investigated on the mountainside and these details can then be used to calibrate seismic geometries, and finally give 3-D geomorphology. Integration of outcrop studies and seismic interpretation/mapping thus gives the most robust geological model. Seismic: Black amplitudes = soft acoustic kick. Amplitude map: Warm colours = strong soft amplitudes. Seismic and mountain outcrop are at approximately the same scale; outcrop view length is about 15 km. From Johannessen *et al.*, 2000.

actively build out from many types of basin margin, in response to there being significant water depth (usually >200 m) in front of the shelf, irrespective of the tectonic setting.

We describe and illustrate a variety of deepwater sandstones that occur within clinoform strata, where the context/water depth of individual sandstone packages is seen clearly from its position/height on the clinoform. We have chosen to compare two clinoform successions of Early Eocene age, one from the Porcupine Basin west of Ireland, using wells and a 3-D seismic data set, the other from the Central Tertiary Basin of Spitsbergen, using very large-scale mountainside outcrops (Fig. 1). Both data sets provide high-resolution insight, allowing the examination of details within clinoform strata on a variety of time scales. Our main aim is to suggest and test a geologically based predictive model for the likely presence or absence of deepwater sands on clinoforms. The work is highly relevant to current scientific interest: Firstly, it analyses how the sediment budget of rivers is eventually partitioned out across shelf, slope and deepwater areas (source-to-sink) during cycles of relative sea-level change. Secondly, it considers the extent to which deepwater sands may be derived from primary river floods rather than purely from slumping and recycling of sediment. The predictive model that is developed focusses heavily on shelf and shelf-edge data, particularly shelf facies, outer-shelf morphology and shelf-edge trajectory data, for the same clinoforms. As regards the applied relevance of this research, we also discuss the degree to which sands are likely to be connected or disconnected along the slope segment of clinoforms, how slope sands are likely to link with the sands of other segments on the same clinoform and if these sandstones are likely to provide the potential for updip stratigraphic pinchouts along clinoform systems (Fig. 1). All of these objectives are achieved by (1) systematic examination of the external character, the internal stacking geometry and the deposits of clinoforms along their basin-floor, slope and shelf-edge reaches in the two data sets, (2) determination of how clinoforms change their character during the course of a sediment supply/accommodation cycle, at differing time scales. The clinoforms we analyse are broadly similar to those described, though from other perspectives, by Winker & Edwards (1983), Christie-Blick & Driscoll (1995), Fulthorpe & Austin (1998), Pirmez *et al*. (1998), Steckler *et al*. (1999), Driscoll & Karner (1999), Porebski & Steel (2003), Sydow *et al*. (2003), Saller *et al*. (2004) or Hadler-Jacobsen *et al*. (2005).

SEQUENCE STRATIGRAPHY OF CLINOFORM SUCCESSIONS

Figure 2 shows the basic clinoform concept. The shelf reaches of the clinoform are almost flat, dipping basinwards with a very low gradient (commonly $<0.1^{\circ}$). This segment of the clinoform is the platform across which deltas or other shoreline types regress and transgress during shelf-transit cycles. The shelf platforms considered here are narrow to moderately wide (10–60 km).

At times of highstand of relative sea level, the shoreline tends to be located on the inner shelf, whereas, at times of lowstand, the shoreline has the potential to prograde out to the outer shelf. If the shoreline is able to regress entirely to the edge of the pre-existing shelf platform, sand can be delivered out onto and beyond the shelf edge, and the shelf-slope system (shelf margin) thus has the potential to accrete basinwards. The basinward and landward movements of the sand-delivery system, usually deltas, thus describe regressive-to-transgressive shelf-transit cycles (see also Cross & Lessenger, 1998; Steel & Olsen, 2002) or systems tracts (Embry, 1993). These cycles can be classified into two basic types (Fig. 2a); one where the regressive-to-transgressive turnaround of the delivery system happens at or before the shelf edge (therefore little or no sands delivered over the shelf edge), and the other where regression continues to and beyond the shelf edge with delivery of significant volumes of sand into deepwater areas. These two types of shelf-transit cycle are significantly different from each other because of shelf width, sediment supply or sea level change, and the consequent presence or absence of an associated deepwater sand-prone package (Fig. 2a). The rationale for these two types of cycle probably also partly underpinned the designation of Types 2 and 1 stratigraphic sequences of some previous authors (e.g. Posamentier & Vail, 1988).

For shelves with a width of those considered here, the regressive shelf-transit time (time taken by a regressing delta or other shoreline to reach the shelf edge area) is probably less than 100k years, as judged from the modelling of modern rivers and their deltas (Burgess & Hovius 1998; Muto & Steel, 2002). This transit time is controlled not only by sea-level behaviour, sediment flux and shelf width, but also by shelf gradient and delta-front gradient (Muto & Steel, 2002). Thus, because the strata deposited from the regressive and transgressive shelf transit (eventually with intervening lowstand segment) is the basic building block of a stratigraphic sequence, and shelf-transit time will filter or constrain the duration of such sequences, the fundamental time scale of clinoform sequences is likely to be a few hundred thousand years or less.

Clinoform morphology

It is important not to confuse shelf-margin clinoforms with deltas. The former have heights of hundreds of metres, whereas deltas are generally tens of metres high (Fig. 2c). However, sediment delivery by deltas is the main mechanism by which shelf-margins accrete, and when a delta has regressed as far as the shelf edge, the delta front happens to coincide with the upper slope of the shelf margin. The shelf segment of clinoforms represents the low-gradient shelf platform across which the sediment-delivery system regresses and transgresses, as discussed above (Fig. 2). The slope segment of clinoforms extends from the shelf-slope break down to the deepwater base-of-slope, generally with an average gradient of less than 3° (and rarely exceeding 6°). The length of the slope segment of clinoforms depends on the water depth as well as the gradient, but in the two cases considered here, the un-decompacted water depths vary from 250 to 400 m, giving slope lengths of 4–12 km. This deepwater slope is generally muddy during times of restricted sediment supply, i.e. either when relative sea level is 'high' and shorelines are

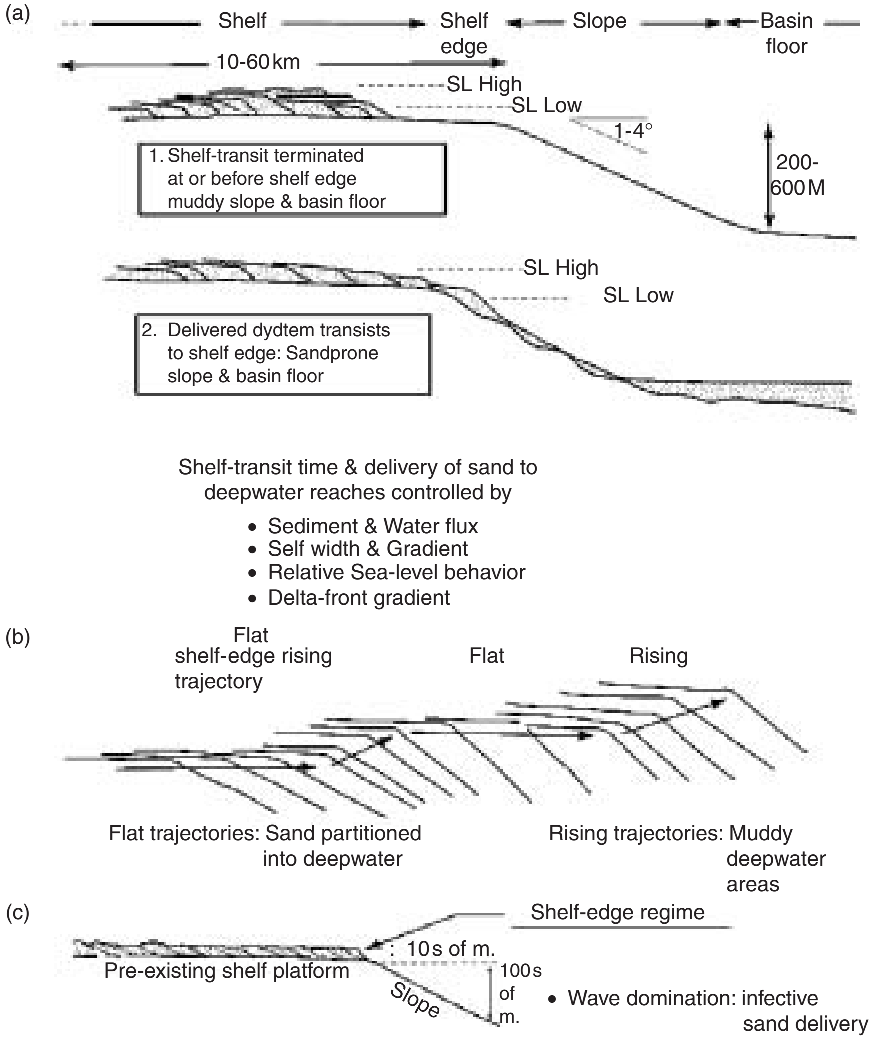

Fig. 2. The basic clinoform concept used in this work. (a) Deltas transiting the shelf may or may not reach the shelf-edge area. In the former case, sand may be delivered to the slope and/or basin-floor, whereas in the latter case the deepwater areas are likely to accumulate mainly mud. Factors controlling shelf-transit time are listed. (b) Accretion of the shelf edge through time produces flattish or rising shelf-edge trajectories. Sand is preferentially partitioned into deep water during flattish shelf-edge accretion, as argued further below. (c) Scale difference between clinoforms produced by deltas (10s of m) and by shelf-margin accretion (100s of m). However, notice that when the deltas reach the shelf edge, the two clinoform types coincide, and the shelf margin is accreting almost solely by the addition of deltaic sediment. The clinoform amplitude at this time reflects the water depth in the basin at relative lowstand of sea level. The process regime at the shelf edge is also an important control on sand delivery into deep water.

back on the inner-middle shelf, or simply when sediment flux through the entire system is low. However, the slope can be sand-prone when it is channelized or when it supports sandy, shelf-edge-attached aprons, usually when sediment flux from the shelf edge is high, i.e. commonly when shorelines are at or near the shelf edge. Although the slope is more likely to be sand-prone when sea level is low, any relationship between relative sea level and sand–mud distribution on the slope would be expected to disappear as shelf width is reduced, so that the slope can then be sand-prone at any time. However, we have been surprised, in the present study, by the degree to which shelves with high sediment supply and with a width of less than 20 km still seem to maintain relatively mud-prone margins when the sea level was high (shelf-edge trajectory was rising), and sand-prone margins when the sea level was low (trajectory was flat, see definitions below). High sediment-supply shorelines that regress across the shelf during rising relative sea level (e.g. Burgess & Hovius, 1998) still tend to produce muddy slopes because they deposit and store much of their sediment budget on the aggrading coastal plain and shelf during their regressive transit (Fig. 2b). The basin-floor segment of the studied clinoforms is relatively flat. The basin floor tends to be muddy during the time interval

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when the sediment-delivery system is located farther landwards of the outer shelf, but can be sand-prone when the sediment-delivery system is sited at or near the shelf edge. However, again, this need not be the case when shelf width is initially minimal or supply is very high.

Shelf-edge trajectory

Shelf-edge trajectory (Steel & Olsen, 2002; Sydow *et al.*, 2003), the pathway taken by the shelf edge during the development of a series of accreting clinoforms (Fig. 2b), can vary in its inclination or gradient in much the same way as that described for simple shoreline trajectories (Helland-Hansen & Martinsen, 1996). Flat or even slightly downward-directed shelf-edge trajectories, created by the accretion of successive clinoforms onto the shelf margin, suggest a stable to slightly falling relative sea level through time. Such flat or low-angle trajectories are consistent with optimal delivery of sediment across the shelf, eventually allowing fluvial channel systems to have direct access to the deepwater slope beyond the shelf edge. Provided this sediment supply is channelled or focused as it crosses the slope, large volumes of sand can reach the toe of slope and basin floor (Fig. 2b). On the other hand, when the shelf-edge trajectory rises through time, reflecting an overall rising relative sea level, there is a tendency for much less or even no sand to be delivered into deepwater areas. The implication of this type of trajectory is that a much greater percentage of the sediment budget is stored on the contemporary shelf and coastal plain, with little being partitioned into the deepwater areas. These conclusions about the relationship between shelf-edge trajectory and sediment storage/bypass were noted by Steel & Olsen (2002) and are a major theme of the present paper and the data sets discussed below.

Evaluation of shelf-edge trajectory along clinoformed successions is therefore a useful first exercise on either seismic or large-outcrop data sets; the trajectory gives a clear signal as to whether a large part of the sediment budget, in a given time interval, is likely to have been retained (deposited) in the upstream reaches of the system, or likely to have been partitioned into the deepwater reaches of the basin. Large sediment-supply systems can of course show both a rising shelf-edge trajectory and a delivery of sand into deepwater (e.g. Sydow *et al.*, 2003), but the delivery duration and the proportion of the budget going into deepwater in such cases are likely to be severely limited, unless the shelf is narrow. For this highstand sediment-delivery scenario to have some chance of success (deepwater sand emplacement), shelf width, coastal-plain gradient, delta-front slope and rate of rise of relative sea level should all be at a minimum (Muto & Steel, 2002).

Facies within the three segments of the clinothems

Sedimentary facies within shelf-slope-basin clinothems, both for those with sand-prone and for those with mud-prone deepwater segments, are illustrated in Fig. 3. The shelf segment of the clinothem has an older part that is regressive/lowstand, usually consisting of river-dominated deltaic or wave-dominated shorezone facies and an overlying younger part that is transgressive and contains estuarine or barrier/lagoonal deposits with an overlying cap of upward-deepening, offshore deposits. The slope segment can be mud-prone or sand-prone, and where it is sandy, channelized or sheet-like turbidites dominate (e.g. Plink-Bjorklund *et al.*, 2001). Sometimes, here also an early sand-prone set of channels and a later back-stepping, muddy channel-levee system can be seen. The basin-floor segment of the clinothem can also be either sandy or muddy. When sandy, the basin-floor deposits are turbidites, and these tend to construct elongate finger-shaped to fan-shaped bodies that build out on the basin floor (e.g. Crabaugh & Steel, 2004).

An appreciable amount of time is required for the construction of the clinothems shown in Fig. 3. Clinothem building requires (1) the regressive shelf transit of the sediment-supply system, (2) the delivery and accumulation of sediment gravity flows on the slope and basin-floor areas, (3) a retreat/re-establishment of the sandy depocentre back up on the shelf edge, and then (4) a transgressive transit of the shoreline back across the shelf. This represents a succession of events during a cycle of varying accommodation to sediment supply (A/S cycle, see Cross & Lessenger, 1998), effectively during a falling-to-rising relative sea-level cycle. This cycle is the basic building block of sequence stratigraphy and is likely to have a time duration of a few 100 kyr at most, as argued above. On the basis of this time duration (Steel & Olsen, 2002), and on the basis of the likely restricted areal extent of the cycle (Embry, 1995), it can be referred to as a fourth-order stratigraphic cycle. The clinothem constructed as above also contains a number of key sequence stratigraphic surfaces and intervening stratal intervals showing specific stacking patterns, or systems tracts.

Stratigraphic surfaces and systems tracts within the clinothems

Erosion and sequence boundary

The basic stratigraphic sequence that constitutes the clinothem is picked most easily within the outermost shelf or slope succession (at least with outcrops), because here the sandstone sheets or tongues stand out clearly within a muddy succession. Sequences are less easy to pick within the more proximal reaches of the sand-prone shelf succession, but the boundaries of sequences are usually picked here on the basis of erosion surfaces that occur at the top of the regressive part of clinothems (Fig. 3). These erosion surfaces are created by subaerial distributary channels cutting into the top of the regressing delta or shorezone as it transits the shelf. The erosion is especially marked and widespread when the system is prograding in a setting with low accommodation-to-sediment supply ratio (low A/S), i.e. during prolonged stable, or slightly falling relative sea level, as recognized in the outcrop or seismic section by a flattish trajectory of shelf-margin accretion (Fig.

2a,b). This sequence boundary continues to be generated whereas relative sea level continues to fall, but is no longer generated as soon as sea level begins to rise. This sequence boundary may still be marked by an erosion surface on the slope, but eventually becomes a conformable surface on the basin-floor (Fig. 3). The basin-floor fan is partly the product of erosion and sediment by-pass along the shelf (though also a product of continued normal river discharge at the shelf edge) and is thus broadly co-eval with the sequence boundary on the shelf.

Intra-lowstand flooding surface

After the development and abandonment of the basin-floor fan system, there is usually a still significant delivery of sediment from the shelf and accumulation of sediment onto the deepwater slope, despite the rise of the base level at this time. Mud-prone channel-levee systems typically develop above and slopewards of the top surface of the fan. The earlier disrupted and channelled topography on the slope becomes gradually smoothed and blanketed by mud, and, in the Spitsbergen outcrops of the present study, there is a thick (10s of m) mud-prone section that accumulated back to the shelf edge and onto the outermost shelf. This mud-prone interval, referred to as the intra-lowstand flooding interval (Plink-Bjorklund and Steel, in press) (Fig. 3), demonstrates that the clinothem was aggrading in a fairly uniform manner at this time (we can document no slope onlap). With eventual rise of sea level above the shelf platform, deltas tend to re-establish at the

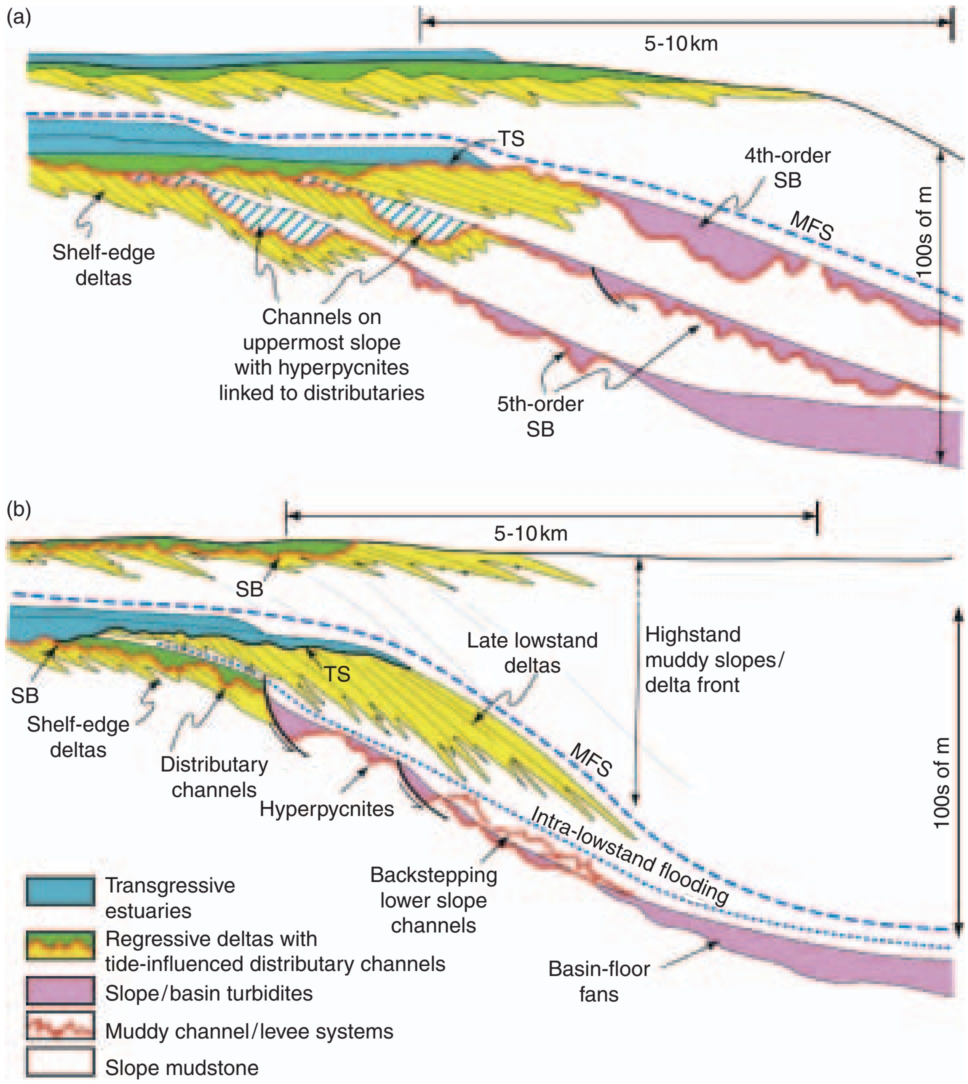

Fig. 3. Different styles of clinoform development during (a) persistent, flat-trajectory progradation and (b) flat to slightly rising-trajectory progradation. Depositional environments of fourth-order clinoforms with (lower parts of (a) and (b)) and without (upper (a) and (b)) deepwater sands. (a) The lower clinoform develops sand on both slope and basin-floor areas, and splits into a series of fifth-order slope and basin-floor segments. (b) The lower clinoform has a lowstand complex of (1) an initial phase of fluvial channelling at the shelf edge, growth faulting and channelling on the slope, and basin-floor fan growth, (2) a mid phase of muddy channel-levee systems on the lower slope and (3) a late phase of shelf-edge deltas that downlap onto the earlier segments. The entire complex is then overlain by mudstones. In both scenarios, the deltas become estuaries or barrier systems during transgression and are eventually capped by a maximum flooding surface.

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shelf edge and produce a strike-extensive, regressive shelf-edge delta wedge that downlaps onto the intra-lowstand flooding surface (Fig. 3b). This deltaic wedge can prograde considerably beyond the earlier shelf edge in the same clinoform and produces a sediment wedge that lies largely below the older shelf edge. These late deltas (Kolla, 1993) have been referred to as the late lowstand wedge (Posamentier & Vail, 1988), or as the late prograding complex (Vail, 1987). They represent the last stage of the lowstand interval and immediately precede shelf transgression. These deltas are less prominent or may not exist where prolonged flat-trajectory progradation occurs through a series of clinothems (Fig. 3a).

Transgressive surface

The transgressive surface (TS in Fig. 3) is picked on the shelf where the first marine deposits overlie non-marine deposits, or where brackish-water, muddy back-barrier deposits have flooded back across the underlying regressive shorezone deposits. Above the initial transgressive surface both tidal and wave ravinement surfaces can occur as the transgression continues. Tidal ravinement is important when tidal channels migrate laterally and landwards within an estuary, and this is followed in time by a wave ravinement surface as the wave-dominated, outer coastline also retreats landwards. In a landward direction, the transgressive surface eventually can become coincident with the subaerial sequence boundary. In a basinward direction, and especially towards the shelf edge, there tends to be some significant separation between this surface and the sequence boundary, so that there is a stratal package on the shelf margin between the two surfaces. This package results from aggradation on the margin although sea level is still relatively low, but during relative sea level rise, prior to the beginning of transgression back across the shelf (Fig. 3b).

Maximum flooding surface

Transgression during continued rising of relative sea level causes the depocentre to migrate landwards back across the shelf, and blankets the underlying shelf and slope with mudstones. The stratigraphic level at which there is maximum marine incursion landwards, producing the deepest water facies across the entire clinoform, is referred to as the maximum flooding surface (MFS in Fig. 3).

Systems tracts

The entire transit of the sand-delivery system from inner shelf to the deepwater depocentre could be seen as a single, regressive or basinward-stepping tract, and the return transit from deep to shallow water as a single transgressive or landward-stepping tract (e.g. Embry & Johannessen, 1992; Embry, 2002). This simple two-fold systems tract division would be quite appropriate where there was no fall of relative sea level, and where highstand shelf-edge deltas reached the shelf edge. However, because the regressive transit sometimes involves a significant fall of relative sea level, and there can be a significant regressive wedge of sediment between the sequence boundary and the transgressive surface, many researchers prefer to further subdivide the regressive strata into two (highstand and lowstand) or three (highstand, falling-stage and lowstand) systems tracts where the data allows this.

THE STUDIED CLINOFORMED SUCCESSIONS

Central Tertiary Basin, Spitsbergen

The Eocene study succession on Spitsbergen developed within a large north-to-south trending foreland or transpressional basin, the Central Tertiary Basin of Spitsbergen (Kellogg, 1975; Steel & Worsley, 1984; Steel *et al.*, 1985; Helland-Hansen, 1990) (Fig. 4a). This basin formed during the late Palaeocene and early Eocene, contemporaneous with the development of the West Spitsbergen Orogenic Belt, along the western coast of Spitsbergen (Harland, 1969; Lowell, 1972; Craddock *et al.*, 1985). The larger-scale setting was one of transpressional plate movement as the Eurasian and Greenland plates slid past each other, and as a new oceanic crust was created in the Norwegian-Greenland Sea (Myhre *et al.*, 1982; Eldholm *et al.*, 1984; Spencer *et al.*, 1984). The basin has generally been considered to be a Foreland Basin with some long-term oblique-slip motion, although Blythe and Kleinspehn (1998) have suggested that it may be a piggy-back basin.

During the early Eocene, the Central Basin filled (>2 km) west to east with marine and non-marine clastic sediment. Basin infilling appears to have been asymmetric, generated by a series of more than 20, eastward-sloping clinoforms with amplitudes of 180–350 m that record the eastward migration of the depocentre (Helland-Hansen, 1992; Nyberg *et al.*, 1995; Steel & Olsen, 2002). The clinoforms, reflecting a basinward accretion of the Eocene shelf margin, highlight timelines through the coastal-plain, marine-shelf, slope and basin-floor stratigraphy. The illustrated clinoform transect is located on Van Keulenfjorden (Fig. 4b).

Porcupine Basin, offshore W Ireland

The Porcupine Basin (Fig. 4c) is one of a series of Atlantic Borderland basins that developed within a tectonostratigraphic framework related to the protracted opening history of the North Atlantic Ocean (Johnston *et al.*, 2001). The basin trends north–south and contains up to 5 km of Mesozoic and 4 km of Tertiary succession (Shannon *et al.*, 1993). It is likely that the Tertiary sediment infill of the Porcupine Basin reflects ridge-push effects associated with spreading-ridge adjustments in the North Atlantic. The ridge push probably produced an uplift of non-faulted basin margins to provide sediment-source areas, such as the Irish hinterlands in the east, the Porcupine High in the north and west, as well as rapid basin subsidence (Shannon

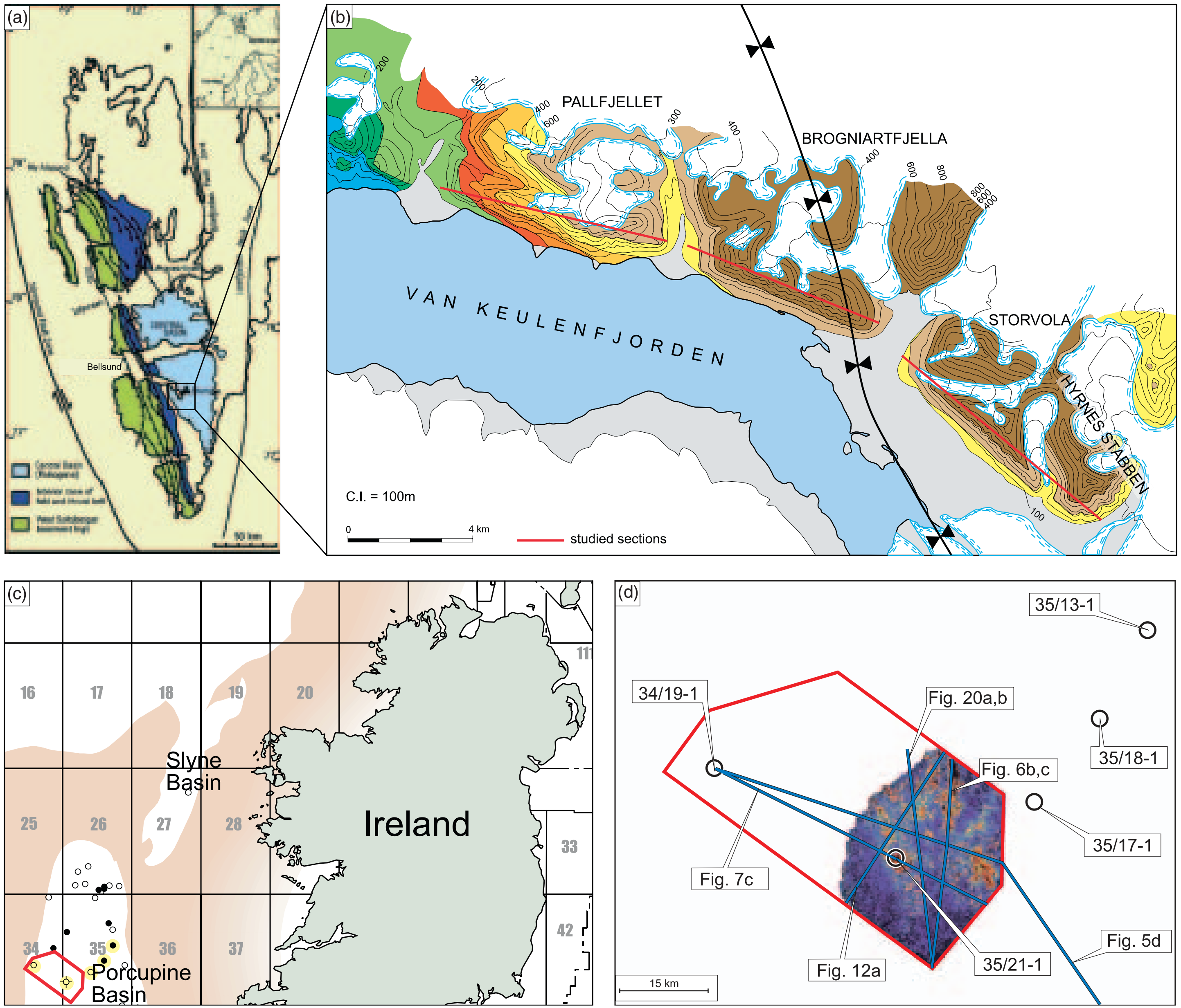


Fig. 4. (a) The island of Spitsbergen, showing the Central Tertiary Basin flanked on its west by the West Spitsbergen Orogenic Belt. Van Keulenfjorden branches to the southeast from Bellsund. (b) The Lower Eocene study transect on northern side of Van Keulenfjorden, along Pallfjellet, Brogniartfjellet, Storvola and Hyrnestabben mountainsides (courtesy T. Olsen). (c) Porcupine Basin, west of Ireland, is a NS Jurassic rift with a cretaceous and tertiary post-rift infill history. Thirty-five wells, most with a Jurassic target, have been drilled in the basin. (d) 2-D seismic coverage and the outline of the 3-D area. The NW part of the 3-D area is dominated by shelf deposits whereas the amplitude map shows the slope and basin floor of one of the studied sequences (3-D survey is Chevron-operated 5/95 and Statoil-operated 8/95). The Sarsfield well 35/21-1 is the only well to targeted the Eocene clinoforms, but with a negative result (Conroy *et al.*, 2003). Wells 35/17-1 and 35/18-1 penetrated these clinoforms, but lateral to the high seismic amplitudes, and thus without sandstones. Well 34/19-1 is in the 3-D area but is too proximal to penetrate the clinoforms.

et al., 1993). During the early Eocene, the Porcupine Basin was infilled north to south, along the old Jurassic failed-rift arm. This involved a major deltaic clastic wedge that built a series of southward-sloping shelf margin clinoforms. Four Palaeogene seismic sequences, arranged as a series of forward stepping to back stepping units, have been documented. Each sequence comprises a 25–75 km wide belt of deltaic to shelf strata, leading down the clinoforms to basin-floor mounded deposits that are interpreted as submarine-fan complexes (Shannon, 1992; Shannon *et al*., 1993).

Stratigraphic setting and age of study intervals

The stratigraphy of the Central Tertiary Foreland Basin on Spitsbergen is outlined in Fig. 5a, b and the relevant successions here are the Aspelintoppen and Battfjellet Formations, and the Gilsonryggen Member. These Lower Eocene stratigraphic units can be seen clearly in Fig. 5a, where the lower third of the mountainside is shale-prone (deepwater deposits of Gilsonryggen Member), the middle reaches are sand-prone (shorezone and shelf deposits of Battfjellet Formation), and the upper third is mixed sandstones and shales (coastal-plain and estuarine deposits of Aspelintoppen Formation).

Previous biostratigraphic studies have dated the study interval (i.e. from the base of the Hollendardalen Formation upwards into the Aspelintoppen Formation) to range in age from latest Palaeocene to Middle Eocene, with the Palaeocene–Eocene boundary occurring in the basal part of the Gilsonryggen shales (Manum & Throndsen, 1986). More recent dating by the present researchers (J. Powell, pers. comm., 2002) suggests that the study interval is entirely within the early Eocene. This assignment of the study interval to Early Eocene (6 myr), provides an averaged age of 200–300 000 years for the individual fourth-order clinoforms. The observed stacking of such sequences, in aggradational to progradational sequence sets, produces larger-scale cyclic successions, as discussed below.

The Palaeocene to Eocene strata of the Porcupine Basin are shown in Fig. 5c, where the environment interpretation is based on log character and lithology description. The succession is divided into three sandstone units with the informal names:

1. Lower Nummulitic Sandstone, interpreted as mainly delta-front sandstones, with a thin, overlying delta-plain succession.
2. The Coaly Sandstone, interpreted as a delta-front to delta-plain/floodplain succession.
3. Upper Nummulitic Sandstone, interpreted as inner-shelf sandstones.

In the proximal reaches of the system, these sandstones together make up a major complex of shelf-margin accretion and back stepping. The entire succession is subdivided into three transgressive-to-regressive units, where the maximum progradation of the complex is interpreted to be close to the top of the Coaly Sandstone unit, within its thick floodplain succession. The Upper Nummulitic Sandstone reflects a clear retreat of the delta system, implying a major relative sea-level rise at this time (Fig. 5c). This paper deals with only the Early Eocene Coaly Sandstone Unit (approximately 4–5 myr duration), that can be subdivided into three sequences (2a–c), as well as a series of higher order sequences.. The Coaly Sandstone unit has a variation in thickness from 380 m in its proximal position (well 34/19-1) to more than 1000 m in its distal position (well 35/21-1). During the development of the Coaly Sandstone unit, deltas transited across a flat-lying shelf platform (seismic topset relectors), delivered sands to the clinoform slopes, and down onto the flat-lying basin floor areas (Fig. 5d, e). These accreting clinoforms have un-decompacted heights of 250–350 m.

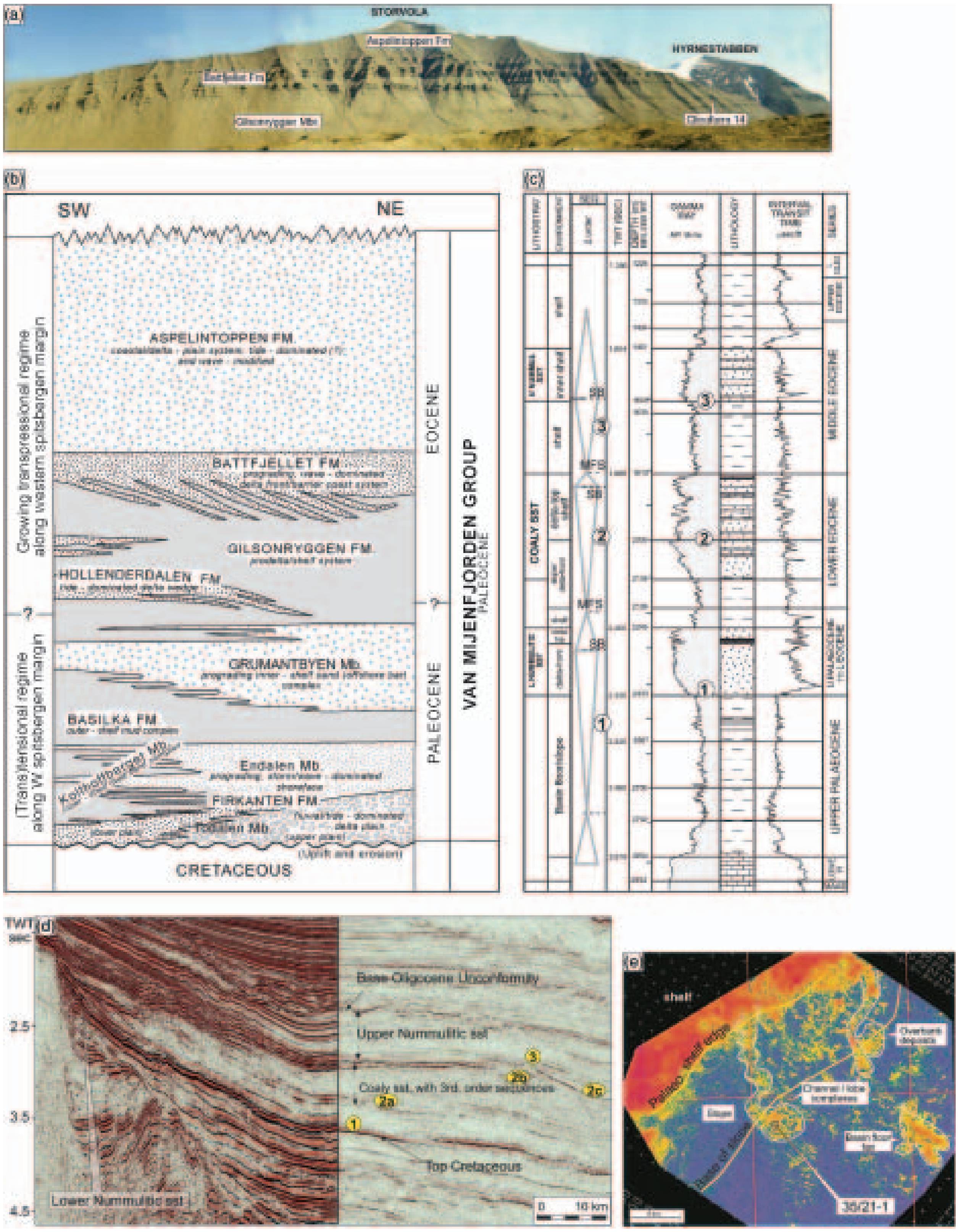
TIME SCALES, SHELF-MARGIN TRAJECTORY AND CORRELATION

In each basin the clinoforms are developed in cycles or sequences at different spatial and time scales. As these time scales are approximately 1–2 myr, few 100 and <100 kyr, and the sequence boundaries show the spatial characteristics outlined by Embry (1995), we refer to them as third-order, fourth-order and fifth-order, respectively. We consider the fourth-order clinoforms to be the basic building blocks of the successions. They stand out very clearly as regressive–transgressive successions with thick shale packages above and below, at least in their outer shelf and slope segments (Figs 3, 5a and 6a). The fourth-order clinoforms also have the most significant transgressive shale penetration back along their shelf (topset) reaches. fifth-order units, by contrast, are less extensive and more closely bundled within their fourth-order host and become amalgamated along the shelf reaches of the system (Fig. 3a). The fourth-order clinoforms then group into thicker and longer time-scale units (third order), identified by significant changes in the trajectory of shelf-margin accretion. The largest-scale trend, which might be referred to as second order, is that which is seen through the entire succession in both study areas. Although each of the four sequence orders or time trends is unlikely to have been caused by a discrete or unique control, we nevertheless attempt below to assign possible causes by at least matching the time scale of the product with a process that operates at a similar time scale.

Basin infilling: 4–6 myr time scale (second order)

The studied successions are some 1.5 km thick and of about 6 myr duration from Spitsbergen and some 1 km thick and of about 4–5 myr duration from Porcupine Basin. Each succession shows a long series of clinoforms (10s of km of basinward accretion) that migrate across the basin

(Fig. 6a–c). Each clinoform series describes an irregular but overall rising trajectory of shelf-margin growth and basin infilling, implying a general or overall relative sea level rise. This long-term trend during basin infilling was probably generated largely by tectonic and compactional subsidence. In the Eocene Central Basin of Spitsbergen, the large-scale, systematic progradation of the shelf margin is seen in Fig. 6a. In Porcupine Basin, two of the large-scale clinoformed units (sequences 1 and 2 in Fig. 5c,d) document major phases of deltaic progradation into the



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basin, whereas, the Upper Nummilitic Sandstone/sequence 3, with lower amplitude seismic reflectors, is interpreted as back stepping (Fig. 5c, d) with respect to the older units.

The very long-term basin-infill trend is one of overall relative sea level rise, allowing a km-scale stacking of strata. At this time scale and thickness, it is likely to have been a tectonic control, possibly also reflecting major depocentre shifting. In Spitsbergen, basin infilling at this scale involved giant lateral accretion (clinoform migration west to east) as deepwater locations gradually became shoreline and then subaerial coastal plains. In the Porcupine Basin, Shannon et al. (1993) postulated changing differential subsidence across basin margins.

Shelf-margin growth cycles: 1–2 myr time scale (third order)

Despite the overall tendency of the shelf margin to show stratigraphic rise through time in both areas (Fig. 6), there is a clear variation along the growth transect, from sets of clinoforms that have a flat growth trajectory to sets that are aggrading at a low angle (see large arrows in Fig. 6a, c). Three stratigraphic units can be distinguished within the basinal succession in each area on the basis of these trends and bounding transgressive flooding intervals. Such units represent changes in the style of shelf-margin growth at third-order time scale, as the basin fills.

Spitsbergen

In the Spitsbergen succession, the individual third-order units have an aggradational-to-progradational style (see units numbered 1–3 in Fig. 6e). Below the first sandstones in the study area on Spitsbergen is a 200 m thick shale succession that is interpreted as the aggradational component of the first third-order sequence (Fig. 6e). Above this thick shale package, there are a series of four basin-floor fans (labelled 1–4 in Fig. 6a) that represent the progradational upper part of this first third-order sequence. Clinoforms 5–14 (Fig. 6a) show a similar overall trend of aggradation followed by flat progradation, and represent the second third-order sequence (Fig. 6e). The turbidite-filled slope channels just below the shelf break in clinoforms 8–10 strongly suggest the presence of deepwater basin-floor fans (fans 8–10 cannot be proven because of outcrop limitation). Clinoforms 11–14 are well exposed and 11, 12 and 14 have basin-floor fans. Clinoforms 15–18 (Fig. 6a) have a more aggradational style, as seen from the stacking pattern in their shelf to shelf-edge reaches. The chance of development of basin-floor fans in this series is likely to be much reduced because the aggradational growth implies that much of the sediment budget was stored within the shelf and coastal plain segments of the clinoforms. There are clear signs of more abrupt basinward shift in sequences 19 and 20, possibly implying a lower angle of growth for the shelf-edge trajectory in this late part of the third third-order unit (Fig. 6e).

Porcupine basin

We deal here only with the Coaly Sandstone Unit (sequences 2a–c, in Fig. 6c), which shows a large-scale growth style analogous to the interval containing Clinoforms 8–18 in the Spitsbergen Central Basin (Fig. 6a). The upper, high-amplitude package with parallel reflectors in the seismic data (Fig. 6b, c) is interpreted as deltaic shorelines on the shelf segment of Porcupine clinoforms. Below this 'tramline' package, there are a set of clinoform slope segments that gradually flatten as they approach the basin floor. The three third-order sequences of the Coaly Sandstone are separated by levels of marine flooding that penetrate far back across the shelf (2a–c in Fig. 6c). Based on the presence of flooding between each tram-line reflector, each of the third-order packages can be further sub-divided into higher-order units. Sequence 2a consists of two fourth-order clinothems; Sequence 2b comprises three fourth-order clinothems and Sequence 2c is difficult to subdivide (Fig. 6c).The strong parallel reflectors terminate at the shelf edge (interpreted in terms of maximum progradation of the shorezone, near the shelf-edge break) and the pathway of successive clinoforms at this location defines the trajectory or style of growth of the shelf margin. Sequence 2a clearly developed during an interval of low accommodation, as it has a flat growth trajectory. The points used to reconstruct the trajectory are closely spaced, and the tram lines are separated by only a very thin shoreline/coastal plain succession. Sequences 2b and 2c have less closely spaced tram lines (implying aggradation), producing a somewhat more steeply rising, shelf-edge trajectory. This change through 2a–c gives the overall stratigraphic 'rise' seen on the larger scale of the entire succession.

Fig. 5. (a) Storvola mountainside showing clinoform 14 crossing through the Batfjellet and Gilsonryggen lithostratigraphic units. (b) Tertiary succession in the Central Basin (from Steel *et al.*, 1985). The clinoforms described span the Aspellintoppen, Battfjellet and Gilsonryggen units. (c) Well 35/13-1 in Porcupine Basin is a key stratigraphic well because of its shelf-edge position (location, Fig. 4d). Three large-scale (second-order) transgressive–regressive sequences occur in this well, but only sequence 2 is treated in this paper. (d) Regional Porcupine seismic line (location, Fig. 4d) showing three second-order Palaeocene–Eocene sequences (1–3) prograding across the top cretaceous surface. Strong parallel reflectors represent deltaic/shoreface sandstones, slope strata are generally poorly imaged and basin floor has weak parallel reflectors. Occasional high amplitudes on slope and basin-floor might indicate sandstones. Maximum progradation of the Eocene system is well imaged at the top of the Coaly sandstone sequence (2c); weaker parallel reflectors of the back-stepping Upper Nummulitic unit (3) lie above. (e) RMS total volume amplitude map of sequence 2a, showing sandstone distribution below the shelf edge from two clinoforms (Fig. 10). Two main channel–lobe complexes are imaged (as well as some slope gullies) between the shelf break and base of slope. Warm colours = strong soft amplitudes.

Cause of third-order sequences

Third-order sequences consist of an aggradational-progradational stack of smaller clinoforms on a thickness scale of 100s of metres, and time scale of a few my. These scales also strongly suggest tectonic control. In Spitsbergen, the units architecturally resemble those seen in other Foreland Basins (Heller *et al*., 1998) and those generated by Foreland Basin numerical models (Flemmings & Jordan, 1989) in response to periods of thrust activity followed by isostatic uplift of the source area. Thrust loading initially causes subsidence to outpace sediment influx, with resultant vertical aggradation of basin-margin shoreline cycles. Reduced rates of thrust loading and eventual uplift caused sediment supply to greatly outpace accommodation so that there was strong shoreline progradation of clastic sedimentary wedges much farther into the basin. The third-order stacking patterns in Porcupine Basin are also likely to reflect long-term accommodation/sediment flux changes in the basin, driven by tectonics. Changing subsidence rates across basin margins has been suggested by Shannon *et al*. (1993), and possible mantle plume control has been put forward by White and Lovell (1997), and by Jones *et al*. (2001).

Correlation of sequences: shelf to deepwater segments of clinoforms

There is significant expansion of thickness and lithology change from shelfal, shallow-water segments of clinoforms to deeper water, basinal segments of the same clinoforms. This change, and its importance for correct correlation is illustrated from seismic/well data and outcrop examples in Fig. 7.

The outcrop correlation from shelf and shelf-edge to deepwater slope and basin floor (western to eastern Storvola) in Clinoforms 12–18 (Fig. 7a), over a distance of less than 10 km, illustrates the importance of (a) predicting the approximate slope gradient for correct well correlation), and (b) being aware of the likely drastic lithology change from shelf to slope and basin-floor segments of the clinoforms. The Storvola correlation diagram also emphasizes the importance of correctly tracking (a) regressive versus transgressive components of the shelf segment of individual clinoforms, and (b) the potential complexity of the slope and basin-floor segments of clinoforms, with their range of early and late turbidite accumulations (Fig. 7a).

An analogous correlation diagram and seismic section, illustrating other exploration considerations, is shown from Porcupine Basin (Fig. 7b, c). The most striking difference between the two correlated wells within the 3-D-seismic area is the lack of slope succession in well 34/19-1, in contrast to the thick, shale-prone slope lithologies in well 35/21-1. This contrast is because of location change from shallow-water reaches of the clinoforms to deeper-water basinal reaches of the same clinoform packages. Sequence 1 (Fig. 7b) is not penetrated in the basinal 35/21-1 well (Sarsfield), but the seismic line (Fig. 7c) suggests a sandstone pinchout before the well position. Based on the regional seismic line (Fig. 6c), Sequence 2 (Coaly Sandstone, Fig. 7b) can be divided into three progradational packages (2a–c). These third-order units are tied and correlated across the 3-D area. The maximum flooding surface between sequences 1 and 2a is not penetrated in well 35/21-1 but the seismic tie line suggests it to be only some tens of metres below TD. The shoreface section in sequence 2a in well 34/19-1 is, based on biostratigraphy and the seismic tie, correlated to the base of slope/basin floor section in the 35/21-1 well, and includes a 26-m thick, fine-grained turbiditic sandbody interpreted as a deep-water fan lobe. The prograding shoreface deposits in the lower part of sequence 2b in well 34/19-1 expand greatly into a thick mud-dominated slope succession with only one thin sandy slope channel, in well 35/21-1. The lower part of the overlying coastal plain section passes laterally into shoreface deposits in 35/21-1, whereas the upper levels of 2b and sequence 2c are interpreted as coastal plain deposits across both wells.

Flat vs. rising shelf-margin trajectory in the data sets

As discussed above, some growth segments showing a flat shelf-edge trajectory contrast with others showing rising trajectory, in both basins. An outcrop example of the former scenario, where the topset segments of successive clinoforms are highly amalgamated and the migrating shelf-edge has a flat trajectory, is shown in Fig. 8a–c. This location involves Clinoforms 6–12 on Brogniartfjellet (Fig. 6a). The flat trajectory predicts that these clinoforms should lead down into well-developed fans on the basin floor, because the amalgamation and trajectory style imply significant sediment bypass and partitioning of sand into the downslope deepwater areas. Clinoforms 6–9 do show sand-prone slope channels, but lack of outcrop prevents verification of basin-floor fans, except for Clinoforms 11 and 12 that do have outcropping fans. An outcrop example of a rising shelf-edge trajectory is shown in Fig. 8d, where Storvola Clinoforms 14–18 (Fig. 6a) have topset segments that are separated by marine or brackish-water shales (in contrast to topsets being amalgamated as in Fig. 8a, b). This scenario predicts that smaller volumes of sand were delivered across the shelf margin into deepwater areas. This appears to hold true for Clinoforms 15–18, which have few sandy slope channels and no basin-floor fans. Another example of rising clinoform trajectories with minimal associated deepwater sands occurs on the mountain Litledalsfjellet, as described by Deibert *et al*. (2003).

From Porcupine Basin, seismic examples show similar features. Sequence 2a (Fig. 6b, c) has a flat progradational style (comparable with that on Brogniarfjellet, Fig. 8a–c), whereas sequences 2b and c show a rising style of shelf margin, comparable with that discussed above for Storvola (Fig. 8d). The volume attribute map for sequence 2a below the shelf edge (Fig. 8e), shows several high-amplitude areas interpreted as channel–lobe complexes and a basin floor fan, suggesting that this flat trajectory style has a significant tendency to supply sand down the slope and onto the

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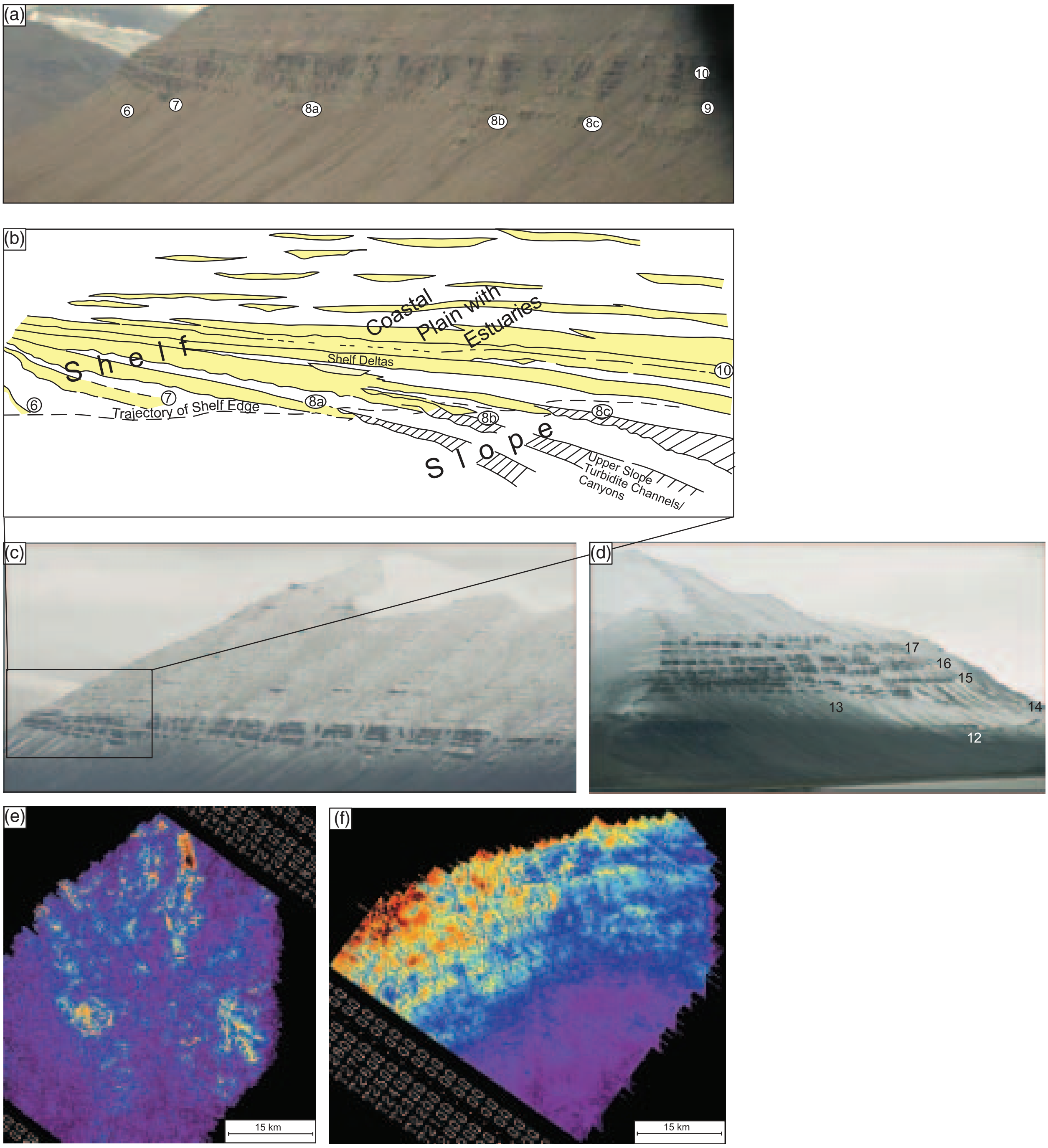

Fig. 8. Photo (a) and interpreted sketch (b) of a clinoform series from Brogniartfjellet showing a flattish shelf-edge trajectory during growth of the shelf margin. Photo (c) is a Brogniartfjellet overview. The shelf segments of successive clinoforms are sandy and amalgamated, implying low accommodation and vigorous sediment delivery out across the shelf edge and down into the deepwater reaches of the system. (d) Contrasting clinoform series on Storvola where there is aggradation of topsets (with some shales between) and a rising shelf-edge trajectory. This style of shelf-margin growth implies much sediment storage on the shelf and coastal plain of the system, and relatively little sand delivery into deepwater. (b and e) RMS total volume amplitude maps from sequence 2a (e) showing high-amplitude geomorphic features on slope and basin floor during times of flat-trajectory progradation. Sequence 2b (f) shows no such geomorphic features (only gradual decrease of amplitude down the slope) during times of rising-trajectory progradation, suggesting little or no sand delivery to the slope and basin floor at this time (see Fig. 6b, c).

basin floor. However, the volume attribute map for sequence 2b (Fig. 8f) shows no geomorphic features on the slope and basin floor. Only a gradual decrease of evenly distributed high-amplitudes down the slope is seen, and this is interpreted in terms of a clinoform where stratigraphic rise caused a relatively large proportion of the sediment budget to be stored on the coastal plain and shelf, with little being delivered across the shelf edge. There is thus a clear tendency that rising shelf-edge trajectories associate with clinoforms that deliver only small proportions of their sand budget down to slope channels and basin floor fans, whereas flat or falling trajectories allow a much greater proportion to pass the shelf edge. Note, however, that this interpretation is limited by the extent of the 3-D seismic data (strike extent of 25 km), and that these conclusions do not exclude significant strike variability.

Clinoform building blocks: Hundreds of 1000s of years time scale (fourth order)

Within the third-order units described above, there are individual clinoforms that have distinctive shelf, slope and basin-floor segments. These are the basic building blocks of the succession, and, when studied from proximal to distal reaches, document how the available sediment budget was partitioned from coastal plain (source) to deepwater (sink). Details of these fourth-order clinoforms are discussed below, where the characteristics of their three separate segments are separately treated.

Individual fourth- or fifth-order clastic wedges of clinoforms, with thickness scale of a few 10s of metres and time scale of up to several 100 kyr, are likely to have been much more influenced by climate (sediment supply) and eustatic sea-level changes because these controls operate within the Milankovitch range of time scales. These time scales are also consistent with numerically modelled delta shelf-transit times, and so the clinoforms on this scale are controlled directly by sediment supply, shelf width and gradient, delta-front gradient and relative sea level behaviour (Muto & Steel, 2002).

Tens of 1000s years time scale (fifth order)

In some areas, particularly where shelf-edge growth trajectory is near horizontal or falling, there are higher frequency units that 'peel off' from the fourth-order clinoforms, as separate sandy tongues that head down a channelized slope. In general, the transgressive tracts of such high-order clinoforms are much less extensive (landwards) than the parent fourth-order transgressive tracts, i.e. the fifth-order units tend to amalgamate with each other at their landward ends, forming a fourth-order topset or shelf segment. These are also discussed further below.

FEATURES OF INDIVIDUAL CLINOFORMS (FOURTH & FIFTH ORDERS)

Clinoform scale and dimensions

A typical example of a fourth-order clinoform from Spitsbergen is shown in Figs 5a and 7a. The undecompacted amplitude of this clinoform, from shelf-slope break to basin floor, is about 300 m. The Storvola clinoform has a slope (Fig. 7a) of some 3.5°, an apparent slope length of nearly 5 km, and the thickness of its slope deposits is 60–70 m. Its main basin-floor fan (on Hyrnestabben, Fig. 9) is up to 55 m thick and outcrops for 3 km beyond the base of slope before it is eroded by a present-day glacier. The regressive shelf segment of this Storvola clinoform (including the by-pass erosion surface on the updip alluvial/coastal plain on Brogniartfjellet) has a minimum exposed length of 12–13 km, and there is an overlying transgressive segment (well represented by estuarine deposits back on the proximal areas) of about the same extent. The total length of the clinoform from points of maximum transgression to maximum regression on the basin floor is more than 20 km. Two sedimentary logs through this clinoform are shown in Fig. 7a. Examples of fifth-order clinoforms that 'peel off' from a fourth-order clinoform can be seen on Brogniartfjellet (Figs 8a, b and 14a).

Typical examples of fourth (and fifth)-order clinoforms from Porcupine Basin are shown within third-order sequence 2a in Figs 6c and 10a, d. The undecompacted amplitude of these clinoforms, from shelf-slope break to basin floor, is about 250–300 m. They have an apparent slope length of 6–7 km, which gives a slope angle of some 2–3°. Some of these fourth-order clinoforms are built up of smaller fifth-order clinoforms, seen especially well on the slope. These fifth-order clinoforms are a single seismic cycle thick, and can have a thickness of up to 30 m. Seismic correlation and tie to a proximal well shows that some fourth-order sequences have a thick (20 m) marine shales below a stacked (60-m thick) coarsening-upward shoreline section, which, in turn, is capped by a thin (2–3 m) marine shale. The well position is close to the proximal reaches of the marine transgression, giving total apparent length of the clinoform from maximum regression to maximum transgression, of some 25 km (Fig. 7b, c).

Basin-floor fan segment of clinoforms

Details of the orange (55-m thick) and yellow (up to 25-m thick) fans associated with Van Keulenfjorden clinoform 14 (Fig. 6a) are shown in Fig. 9a–d. The fans prograded away from the viewer and into the plane of the photo because of longitudinal skewing of the fans at the base of the slope. There appears to be a compensational thickness relationship between the orange and yellow fans. These fans consist mainly of thick-bedded ($>$0.5-m thick), ungraded sandstones and thin-bedded ($<$0.5-m thick), ripple-laminated sandstones. The facies and vertical stacking tendency of thin-to-thick bedded turbidites is shown in Fig. 9d. As shown in Fig. 9c, the thick-bedded facies can locally pass laterally into the thin-bedded facies, although there is a general tendency for the thicker and thinner beds to dominate in the upper and lower levels of fans, respectively. No deep ($>$ 1.5 m) channel features are observed on these fans and they are thus interpreted as fans or lobes with a shallow braided channel system. Details of the facies, geometry and skewing of some of these fans are discussed in Crabaugh & Steel (2004). Figure 11 illustrates a progressive basinward stacking of a series of basin-floor fans (associated with sequences 1–4 in Fig. 6a) on Palfjellet

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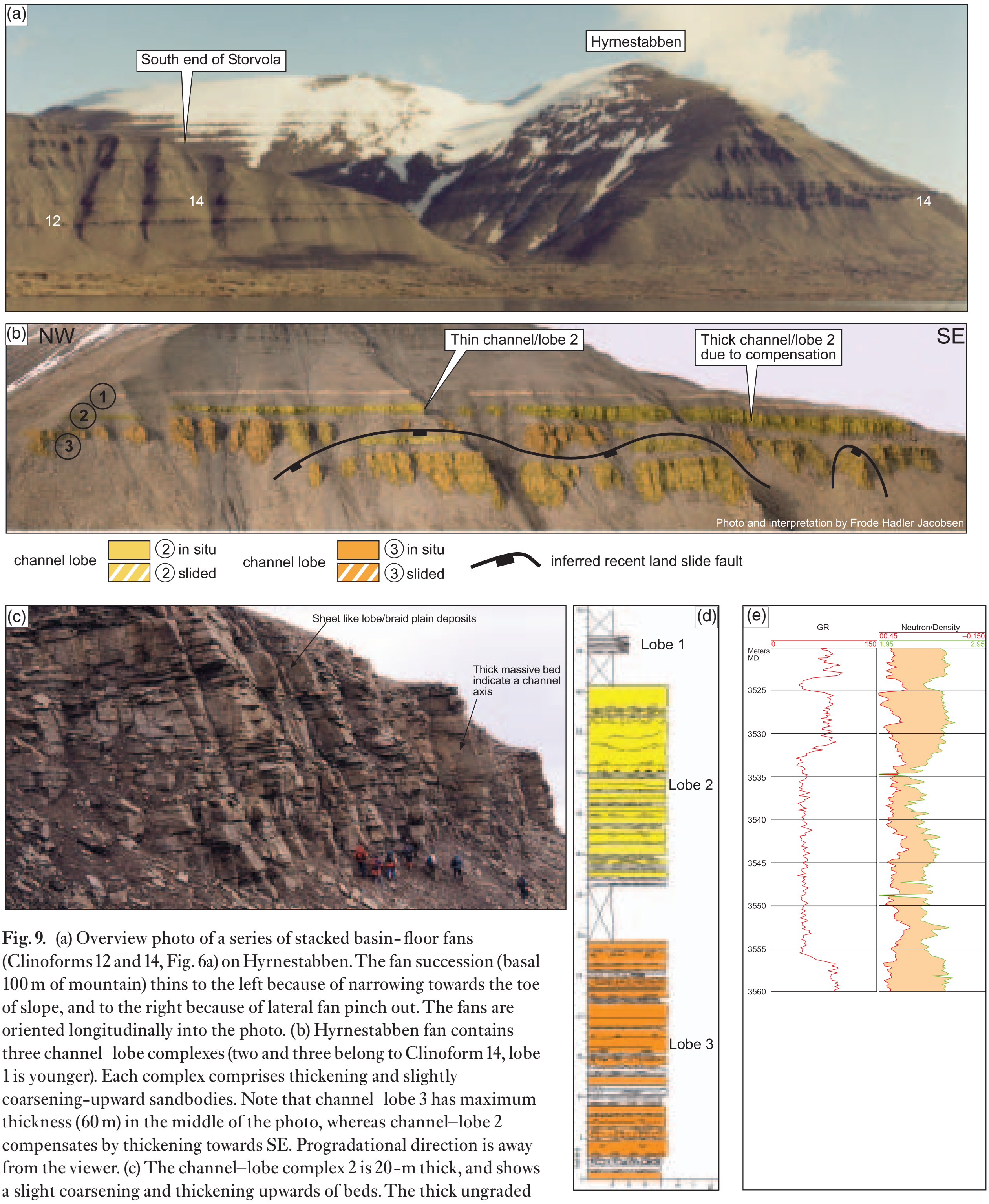
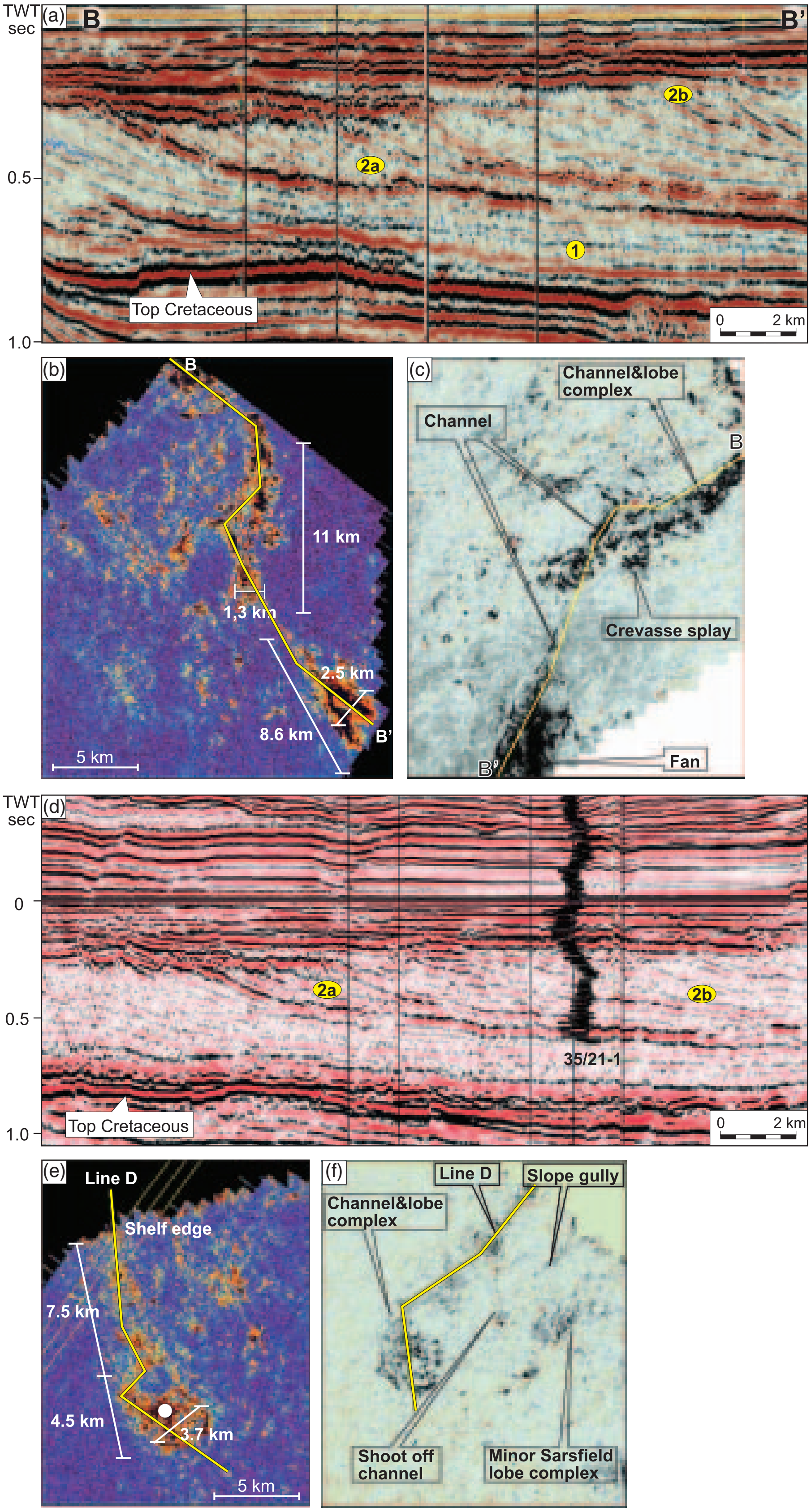


Fig. 9. (a) Overview photo of a series of stacked basin-floor fans (Clinoforms 12 and 14, Fig. 6a) on Hyrnestabben. The fan succession (basal 100 m of mountain) thins to the left because of narrowing towards the toe of slope, and to the right because of lateral fan pinch out. The fans are oriented longitudinally into the photo. (b) Hyrnestabben fan contains three channel–lobe complexes (two and three belong to Clinoform 14, lobe 1 is younger). Each complex comprises thickening and slightly coarsening-upward sandbodies. Note that channel–lobe 3 has maximum thickness (60 m) in the middle of the photo, whereas channel–lobe 2 compensates by thickening towards SE. Progradational direction is away from the viewer. (c) The channel–lobe complex 2 is 20-m thick, and shows a slight coarsening and thickening upwards of beds. The thick ungraded beds (upper fine- to medium grained) with sheet-like and lenticular geometries are interpreted as transitional between shallow channels and lobes, and are interpreted as relatively proximal on an unconfined basin-floor fan. (d) Sedimentary log in turbidite beds of the orange and yellow fan lobes on Hyrnestabben. The thick beds are structureless and ungraded, whereas the thin beds are ripple laminated (from Crabaugh & Steel, 2004). (e) Log from Porcupine Basin (well 35/21-1, Fig. 7b, c), showing deposits as a distal lobe at toe of slope. GR suggests massive sandstones, whereas the neutron/density log shows two slightly upward-coarsening to fining motifs. High spikes on the density log (green) may be organic-rich layers.

at the western end of the Van Keulenfjorden transect. These fans, each separated by a few 10s of metres of shales, can be seen to pinch out progressively farther basinwards through time (Figs 6a and 11). The uppermost fan (fan 4), thinner than the others, can be seen to cap the basin-floor succession in Fig. 11. The greater degree of internal defor-



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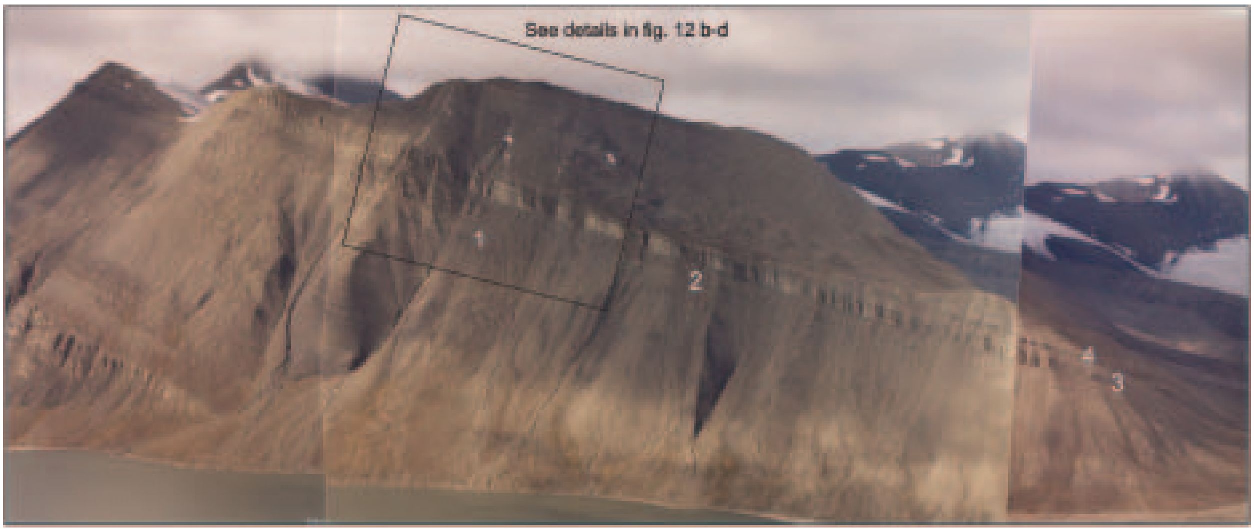

Fig. 11. Stacked basin-floor fans on Pallfjellet (sequences 1–4, Fig. 6a). Highlighted section is Fig. 12b. Below the fan bodies there is a 200-m thick slope-to-basin floor shale. Sandstones in the lower left are shallow-marine deposits of Grumantbyen Formation (see Fig. 5b).

mation and erosive channelling seen in this fan suggests that it represents a landward-thinning lower-slope apron, whereas the underlying three fans are on the basin floor. It is likely that the fans identified here belong to different clinoforms, because of the thick shale package between each. Each of the intervening shale packages is likely to expand greatly in their thickness slopewards, if it is correct that they represent the shingled toes of different clinoforms.

In Porcupine Basin, an arbitrary dip-directed seismic line down a clinoform in sequence 2a (Fig. 10a) shows an almost continuous high-amplitude reflector from the shelf edge and down into the basin floor. The amplitude maps (Fig. 10b, c) along this clinoform shows several basin-floor features. A 9-km long and 2.5-km wide well-developed, fan-shaped body with two distinct branches or lobes are observed some 10-km basinwards of the toe-of-slope area. This single-cycle seismic event, some 30–40-m thick, is interpreted as a basin floor fan. The channel feeder to the fan narrows and thins in the updip direction, and is masked by a slightly younger channel/lobe complex, but is again seen at the toe of slope before it turns east and merges into the common slope channel (Fig. 10a). Where the channel feeder is narrow, it also has weaker seismic amplitude (Fig. 10a). It is believed that this transitional area from basin floor to toe of slope reaches of the system comprises continuous sand, but with a thickness close to seismic resolution.

The high-amplitude feature at the toe of slope (Fig. 10b, c) is 1–2-km wide, extends some 11 km down the slope, lacks internal channels and has no fan geometries like the fan described above. Amplitude terminations are associated with rounded ridges (pressure ridges associated with toe trusts), which could suggest that this feature is a complex of debris flows. However, observation of a crevasse splay amplitude feature (Figs 5e and 10 b, c) suggest a more long-lived active sand supply. This high-amplitude feature at the toe of slope is thus interpreted as a channel–lobe complex, and may be similar to the Hyrnestabben fans on Spitsbergen.

An arbitrary dip-directed line, also in sequence 2a but along a slightly younger clinoform than described above, shows high seismic amplitudes from the shelf edge and to some distance beyond the toe of slope. No clear basin-floor fan is found within this clinoform (Fig. 10d–f). The toe of slope amplitude feature is some 5-km long and 3.5-km wide. The upper slope part of this feature is some 7-km long and 1–2-km wide. Amplitude termination consists again of rounded ridges, but chute channels off-shooting from the slope segment of the channel suggest an active sand supply system (Fig. 10f). The fact that the high-am-

Fig. 10. (a) Arbitrary flattened depositional-dip line across the high-volume amplitude map of sequence 2a (Fig. 1, line B–B^1 and 5(e), right part). This construction shows that the clinoform at lowest part of sequence 2a is a shelf-attached, continuous high-amplitude reflector along the entire slope to basin-floor profile, suggesting continuous sandstone. Sequences 2a and b on the shelf are built up of several flat-lying high-amplitude reflectors (fourth-order sequences). From each seismic reflectors, individual slope clinoforms can be seen to peel off, suggesting that slope clinoforms have fifth-order status. (b) RMS attribute map generated along the continuous high-amplitude clinoform on seismic line a. (c) Horizon slice map generated along the same clinoform as in (b). (d) Seismic line is a constructed dip section of the Sarsfield clinoform and includes well 35/21-1 which penetrated a 26 m-thick, toe-of-slope sandstone, interpreted as a turbidite lobe. The clinoform (middle part of sequence 2a) shows high amplitudes from basin floor up to the shelf reflectors. Along the high amplitudes there are some dim spots on the slope, interpreted as faults that were critical to the trap integrity in the pre-drill analysis (Conroy *et al.*, 2003). (e) RMS attribute map generated along the continuous high-amplitude seismic reflector (clinoform) on seismic line d. (f) Horizon slice map generated along the same clinoform as in (e).

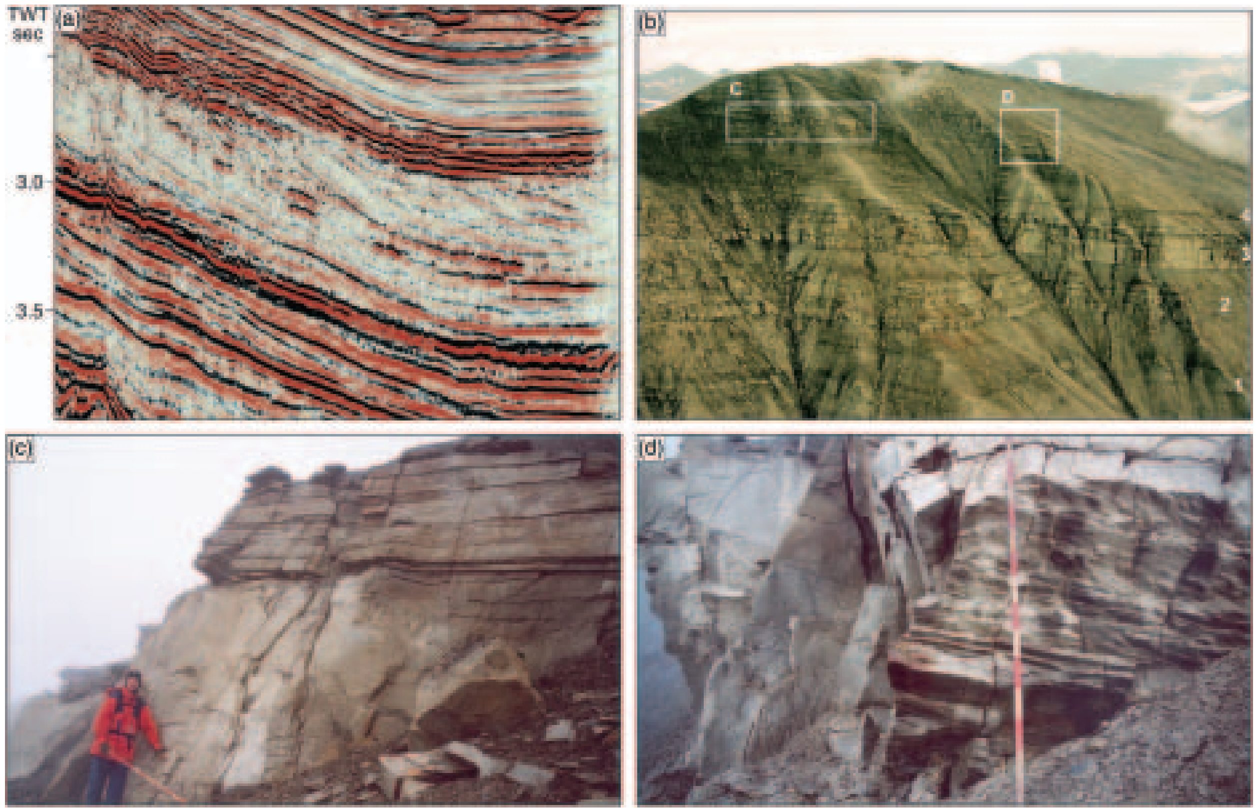

Fig. 12. (a) Random seismic line across the Porcupine 3D data set (location Fig. 4d). Scattered lenses of high-amplitude reflectors in the section are interpreted as sandy slope channels, gullies or debrites. Analogous lens-shaped sandstone geometries on the slope occur on Pallfjellet (b), Storvola (Fig. 1) and Areniusfjellet (Fig. 13) (3-D examples of such apparently isolated sandstone bodies are shown on seismic section and amplitude maps of Figs 1, 5e, 8 and 10). (b) Pallfjellet (highlight in Fig. 11) showing stacked, 10–30-m thick, basin-floor turbidite fans (1–4) separated by basin-floor shales (5–20-m thick), overlain by 150-m thick slope mudstones. Two slope channels (insets (c) and (d)) occur within the slope section. Pallfjellet is capped by shallow-marine sandstones. (d) Slope channel C (from (b)) that is massive, coarse grained in lower part, and 8.5-m deep. (d) Turbidite-filled slope channel D (in (d)) with grooves and flutes along channel side.

plitude links directly up into the shelf again supports an active sand supply system and not a slump-related debris flow. This feature is thus interpreted as a channel–lobe complex, the Sarsfield channel–lobe complex. The Sarsfield amplitude feature was part of Statoil's stratigraphic-prospect portfolio and was drilled (well 35/21-1) with a negative result in 2000 (Conroy *et al.*, 2003).The well penetrated the lower slope part of the amplitude (Figs 5e, 7c and 10d, e) and a 26-m thick, fine-grained sandstone body was documented. The gamma ray log through this body shows a box-form or blocky pattern, with a sharp base and top (Figs 7b and 9e), a log pattern expected more from a slope channel or a debris flow rather than from a fan lobe. However, the density log shows two faint upward-coarsening packages suggesting more organized deposition than would be expected on a debris flow (Fig. 9e). The current interpretation is that the exploration well penetrated the distal part of the Sarsfield channel–lobe complex.

Slope segment of clinoforms

The slope segments of clinoforms in the Spitsbergen data set are generally mud-prone, although at certain times, parts of the slope can become sand-prone. Medium to coarse-grained sandy channels that feed basin-floor fans usually form on the slope at an early stage of the fall-to-rise cycle. Examples of such early channels are shown on Arenuiusfjellet's lower slope (Fig. 13), Pallfjellet's middle slope (Figs 11 and 12b–d), and on Brogniartfjellets upper slope (Figs 14 and 15b, c). Somewhat finer-grained and smaller channels with associated muddy levees form on the lower slope at the latest stages of fan development or right after fan abandonment. These mid-stage channel systems sometimes appear to backstep towards the slope with time, and examples of these are shown on Storvola and in Porcupine Basin (Fig. 1). At the latest stages of the lowstand accumulations, there sometimes occur broad sandy aprons dominated by sheet-like turbidites from late-stage shelf-edge deltas on the upper-middle slope. It should be emphasised that such sand-prone slopes are fairly rare, occurring only when a sand supply system reaches the shelf edge and sand delivery is effective across the shelf margin. The late-stage sandy aprons (late prograding wedges in Exxon terminology) that extend down from the shelf edge are the result of delta re-establishment on the outer shelf as the relative sea level rose back up to a

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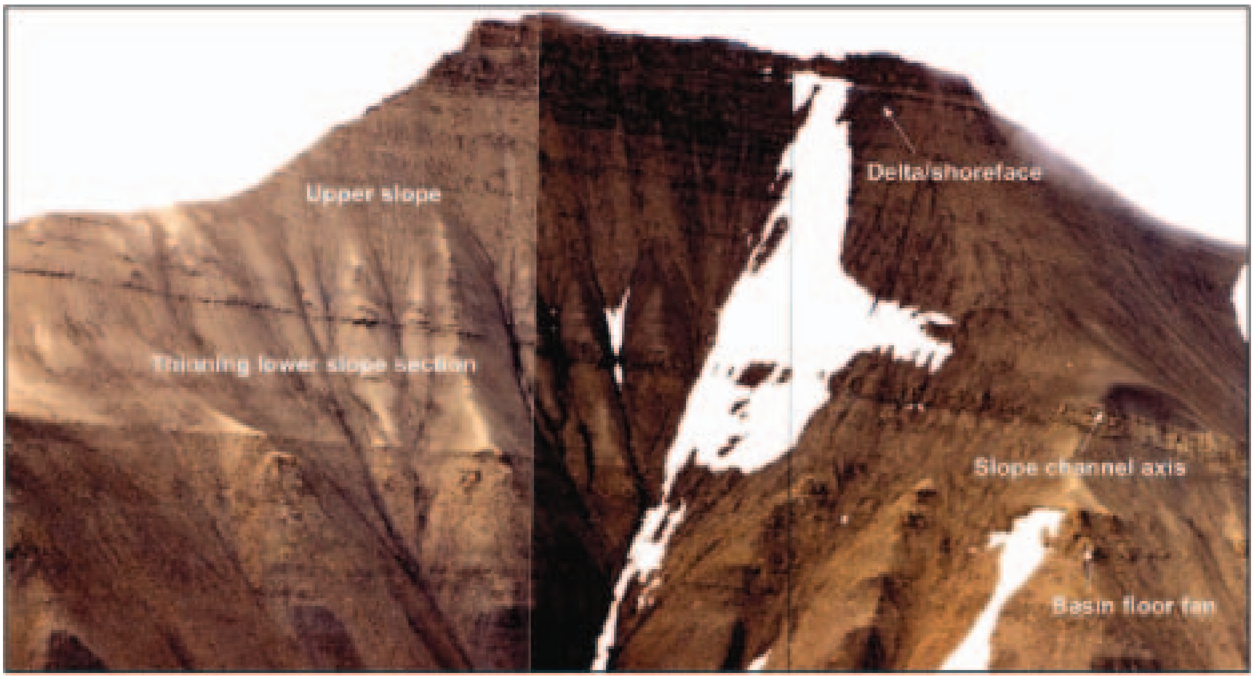

Fig. 13. Obliquely cut slope channel on middle right Areniusfjellet, thinning upslope on the clinoform. Note that the shale section between basin-floor fans (lower part of the succession) and the slope channel, thins downslope, showing that the slope channel occurs on the clinoform slope. Younger prograding shelf can be seen in the uppermost part of the mountain.

shelf position (Exxon Type 1 system), or are the result of deltas simply perching on the shelf edge during the entire lowstand interval (shelf-margin tract; Exxon Type 2 system). Examples of the latter are detailed by Mellere *et al.* (2002) and by Plink-Bjorklund *et al.* (2001), whereas examples of the former are discussed by Steel *et al.* (2001, 2003). The mid-stage channel-levee systems were termed slope fans by Vail (1987), early lowstand wedges or channel-levee systems by Posamentier *et al.* (1991) and Type ll turbidite systems by Mutti (1985).

A slightly oblique seismic line to depositonal dip (Fig. 12a) across the Porcupine data set shows a shale-prone, low- amplitude slope to basin floor succession (see also the 500-m thick, shaly slope section penetrated in well 35/21-1 (Fig. 7b). Scattered and clusterd lenses of high-amplitude reflection in the section (Fig. 12a) are interpreted as sand-prone slope channels, gullies or debris flows. They appear to be randomly distributed and disconnected down the slope, but amplitude maps along each clinoform show more continuous zones or areas that express high-amplitude geological features such as channels, lobes and fans surrounded by mud-prone low-amplitude areas. The typical width of such slope channels is 1–2 km. They are single-storey ($<$30-m thick) and continuous down through the entire slope (Fig. 10).

Within the 3-D data set in the Porcupine Basin there are channelized sands on the slope only in sequence 2a that has a flattish shelf-edge trajectory (Figs 6b, c and 8e). This observation is similar to that described as scenario 3 below, from the flat-trajectory, shelf-edge progradation on Spitsbergen. In contrast, an amplitude map from the slope segment of a third-order clinoform set with rising trajectory shows evenly distributed amplitudes along the slope, probably representing sheet sands (similar to scenario two described from Spitsbergen shelf-edges, Fig. 8e, f, compare maps). The slopes of clinoforms are generally mud-prone (Fig. 7), but two fifth-order clinoforms are sand-prone along certain reaches (Fig. 10). High seismic amplitudes along these clinoforms show several small gullies and two more continuous channel features, supplying sand to toe of slope lobes and a well-developed basin-floor fan. The channel features are straight to slightly sinuous and are typically 1–2-km wide. Comparison of full and far offset seismic data also confirms that slope segments generally are mud-prone (Fig. 21).

Shelf-edge segment

Shelf-edge scenarios

The shelf-edge segments of clinoforms in the Spitsbergen transect show three different conditions, each of which leads to a predictable sand-partitioning scenario in the deepwater beyond the shelf edge. In some cases, the shelf edge region is simply shale-prone (the sand-delivery system retreated transgressively back across the shelf before it neared the shelf margin), and this scenario leads predictably to little sand being delivered into the deepwater area. In a second scenario, the sandy deltas do reach the shelf margin, but there is no significant river-channel downcutting at the shelf edge and therefore no direct linkage between the delta distributaries and slope channels or gullies. In this scenario, large sand volumes can be delivered as delta-front, turbidite sheets onto the upper-middle slope (up to 60-m thick on uppermost slope), but this sand is dispersed as unconfined flows, and little of the sand reaches the basin floor because of a lack of major channels on the upper-slope apron (Mellere *et al*., 2002). Nevertheless, the strike extent of such shelf-edge-attached aprons or wedges can be several 10s of kilometre, implying that significant volumes of sand have been delivered beyond the shelf edge. In the third scenario, where there is more significant channelling and more continuous clinoform peel-off from a prograding shelf edge with a flat-trajectory, the delta-distributory system seems to have been more focused, so that the river-derived sediment gained direct access to the slope and linked into existing or newly created slope channels. A good example of this scenario can be seen on the shelf-edge outcrops of Brogniartfjellet (Fig. 14). In the break-of-slope reaches of some fifth-order clinoforms, there are interpreted steep-fronted, subaqueous mouth bars (12–15-m high foresets, Fig. 14b, c) that appear to feed sand directly into upper-slope channels. In other cases, river/upper slope channels up to 10m deep, and filled with what we interpret as hyperpycnal flows, cut down into older delta-front sands, to feed directly onto the slope (Fig. 14a–d). Such sand supply that was focused within slope channels produced turbidites that eroded and incorporated sediment on the slope and so were able to reach the base of slope as high-density flows. These flows constructed sand-prone fans on the basin floor.

Porcupine Basin fourth-order clinoforms that show flat progradational trajectories display frequent fluvial channels in their outer-shelf/shelf-edge segments, whereas the similar segments of clinoforms that have a rising shelf-edge trajectory have more smooth reflector morphology and apparently contain far fewer distributary channels (Fig. 6b, c, compare the flat-lying segments of sequences 2a and b/c). However, even where clinoform topsets are well channelized and show a flat shelf-edge growth pattern, the channels do not erode far down or entirely through the underlying high-amplitude strata, and the channel system is thus not deeply incised. Nevertheless, such channels systems are often connected with high amplitudes at the shelf break, or high on the slope, strongly suggesting that these fluvial channels have a large potential to supply sandstone down the slope and out into the basin floor (Fig. 15a). Upper-slope areas lateral to the fluvial channels, or within third-order clinoform sets that show a rising shelf-edge trajectory, are seismically more transparent and are therefore believed to have less potential to deliver sand down into deepwater areas.

Relationship between shelf-edge and slope reflectors

One of the most common conclusions drawn about clinoforms, in the literature (e.g. Steel & Olsen, 2002), is that (a) when they prograde with a rising trajectory then the relative sea level is rising and there is complete interfingering between the shelf-edge delta system and the slope system, whereas (b) when they prograde with a flat or downward trajectory, the apparent toplap relationship between shelf and slope reflects continuous and deep erosion between shelf-edge and slope depositional systems. We agree with the former but only cautiously with the latter. Although there are likely to be cases where toplap erosion does occur between shelf and slope (by rapid sea level fall below the shelf edge), our seismic (showing apparent toplap) and outcrop (suggesting erosion from a distance) transects show rather that deep erosion is localized and that the relationship between slope and shelf edge is still transitional, but with a very compressed or foreshortened upward coarsening between the two. We suggest that identification of the interfingering is difficult because it is commonly below the resolution of the seismic data and masked by the high-amplitude contrast between sandy shelf edge and shaly slope. At places where the upper slope is sand-prone, the interfingering link between the shelf and slope systems is well imaged (see shelf-attached clinoforms in Figs 10, 15a and 18). This is also illustrated in Fig. 6b, in the left-hand part of sequence 2a (at 2.9 s), where small high-amplitude dipping reflectors (sandy gullies) merge and interfinger with the flat high-amplitude reflectors of the shelf.

In the Spitsbergen outcrops, there is a clear architectural difference between the deltaic units involved in shelf edges that were prograding with 'rise' (they are thick, have parasequences or higher-frequency sequences and and are upwards coarsening) (e.g. Fig. 16a, Clinoforms 16 and 17) and deltaic units within flatly prograding shelf margins (they are thinner, coarser, show few higher-frequencey sequences and abruptly coarsen upwards) (e.g. Fig. 16a, Clinoform 15). In both cases there is an inter-fingering between shelf and slope systems, albeit foreshortened in the latter case.

Shelf segment

One aspect of the clinoform analysis both from the Central Basin and Porcupine Basin that differs most from many of the previously published analyses is the degree of aggradation that has been documented on the shelf during each fourth-order sequence. This deposition occurred both during regressive and transgressive shelf transits of the sand-delivery system (Fig. 6). Transgressive tracts are particularly thick, often making up more than half the thickness of the shelf succession in any sequence. These transgressive deposits are commonly thick estuar-

Fig. 15. (a) seismic line, slightly oblique to depositional dip, showing clinoforms with a flat-progradational trajectory, splitting up into fifth-order clinoforms at the shelf break. Similar scale and type of clinoform splitting/divergence is seen from Brogniartfjellet (Fig. 14a, b). (b) Clinoform 8 on Brogniartfjellet, showing an analogous clinoform 'peel- off' as in (a). The lower sandstone bench (an upper- slope channel) diverges from the horizontal shelf-edge strata above. Note landward thinning of shelf muds to the left, between slope sandstone bench and overlying shoreline/delta benches (c) Detail of the upper-slope channel on Clinoform 8 showing an offlapping or side-lapping, large-scale lateral accretion of hyperpycnites (interpretation courtesy D. Mellere).

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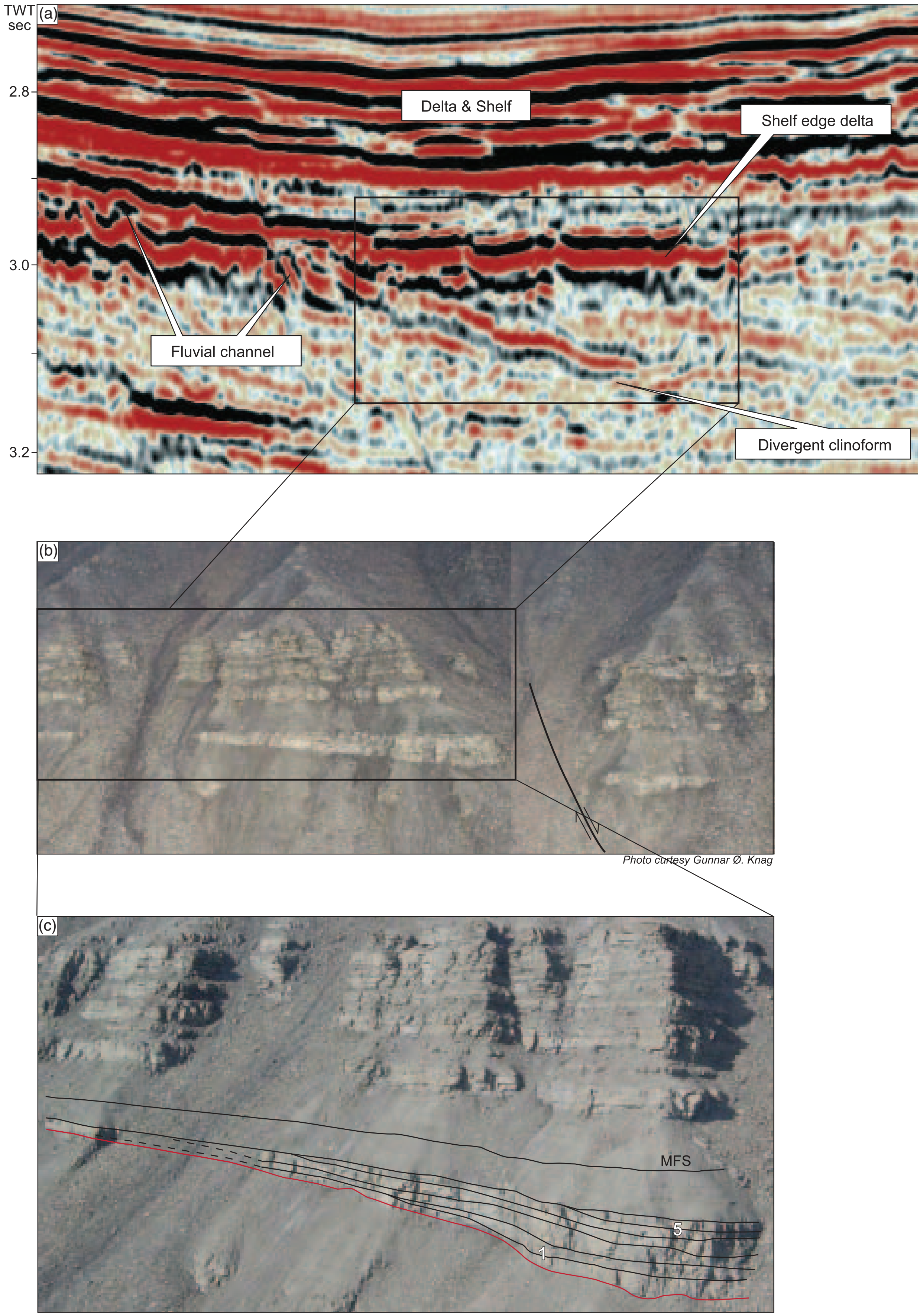
TWT sec
(a)
2.8
3.0
3.2
Delta & Shelf
Shelf edge delta
Fluvial channel
Divergent clinoform
(b)
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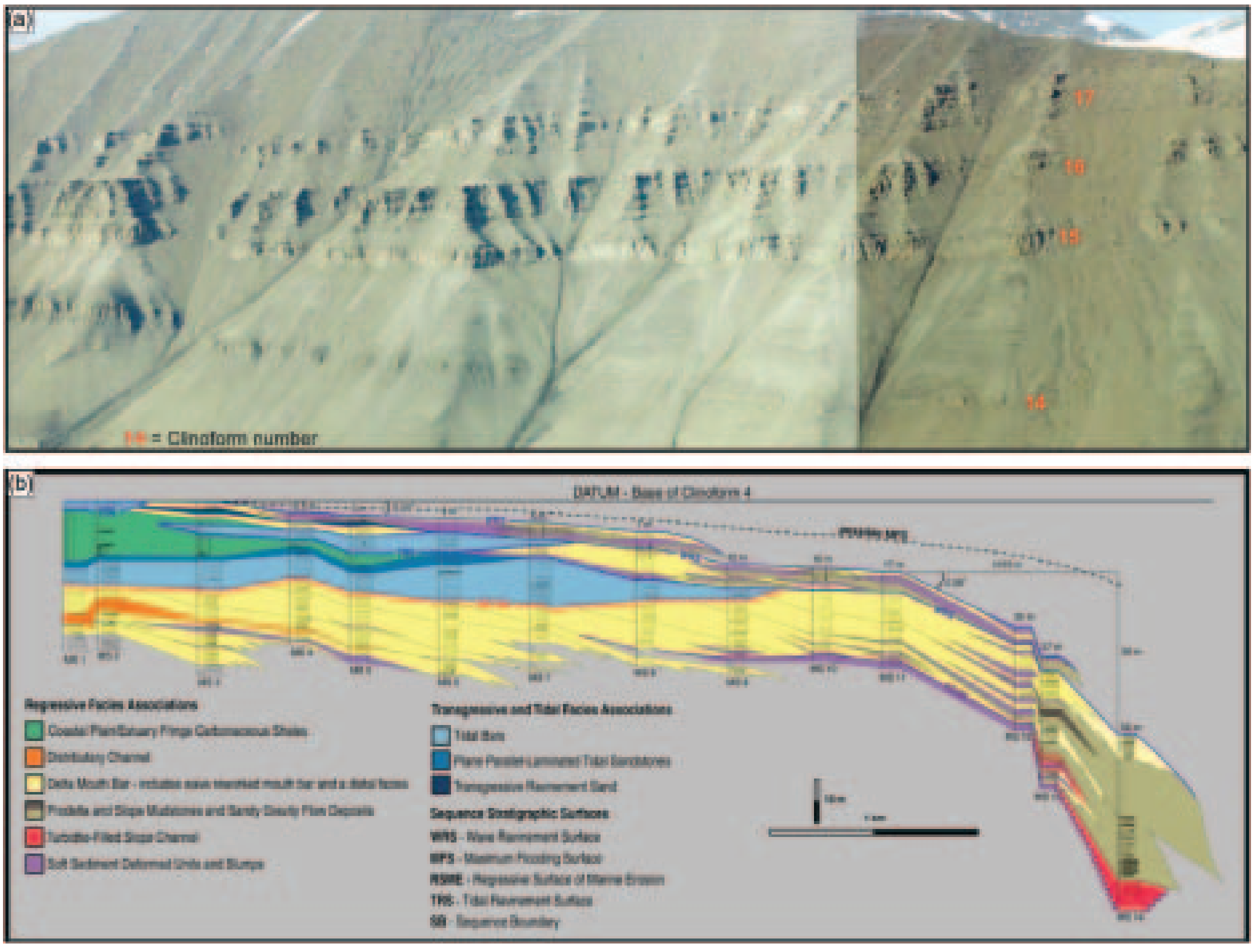

Fig. 16. (a) Helicopter shot of shelf segments of clinoforms (15–17) on Storvola that show quite different architectural styles. Clinoform 15 is relatively sharp-based, thin and accumulated with a flat to falling shoreline trajectory, whereas Clinoforms 16 and 17 are gradually coarsening upwards, have many higher order sequnces, are thicker and accumulated with a slightly rising shoreline trajectory. (b) Architectural diagram of Clinoform 15, showing how only the lower half of the sandstone body accumulated regressively from shelf deltas. The upper half is a transgressive estuarine system that was sand-prone at its mouth and muddier landwards (from M. Schellpeper, in Steel *et al.*, 2003).

ine sands sited at the outer shelf reaches of clinoforms (Fig. 16b), which become lagoonal mudstones farther back on the shelf. The transgressive tract then dominates the coastal-plain succession farther landwards, where tidally influenced deposits overlie fluvial sediments in most sequences.

The regressive part of the shelf segment of any sequence is commonly deltaic (e.g. Mellere *et al.*, 2002), but may also be wave-dominated strandplain. Where the regressive segment occurs within a set of sequences showing a shelf-edge trajectory that is flat or slightly falling, the thickness of the regressive deposits tends to be modest (10–30 m) and its base is fairly abrupt (Fig 16a, Clinoform 15). In contrast, where the regressive tract occurs within a set of sequences that have a 'rising' shelf-edge trajectory, the regressive deposits of any sequence tend to be thicker (30–50 m), more clearly sub-divided into upward-coarsening units (parasequences or higher-order sequences) up to 10m thick, and tend to be more gradationally based (Fig. 16a, Clinoforms 16 and 17).

The shelf segments of clinoforms in Porcupine Basin comprise parallel, continous high-amplitude seismic reflectors, with occasional channel features (Figs 6 and 15). Log patterns observed on these clinoform topsets suggest coarsening-upward shorezone strata, 'box-form' and slightly fining-upward stream-channel units, as well as muddy strata and coal beds, representing delta-front and delta-plain coastal environments (Fig. 7b).

3-D CLINOFORM DEVELOPMENT AND EXPLORATION SIGNIFICANCE

Conventional slope-onlap vs. clinofom aggradation models

Conventional sequence stratigraphic models for deep-water, lowstand clastic deposition and deepwater exploration (e. g. Posamentier & Vail 1988) usually portray sandy detached basin-floor fans that show some onlap back towards the muddy slope, thus ensuring a degree of slope-

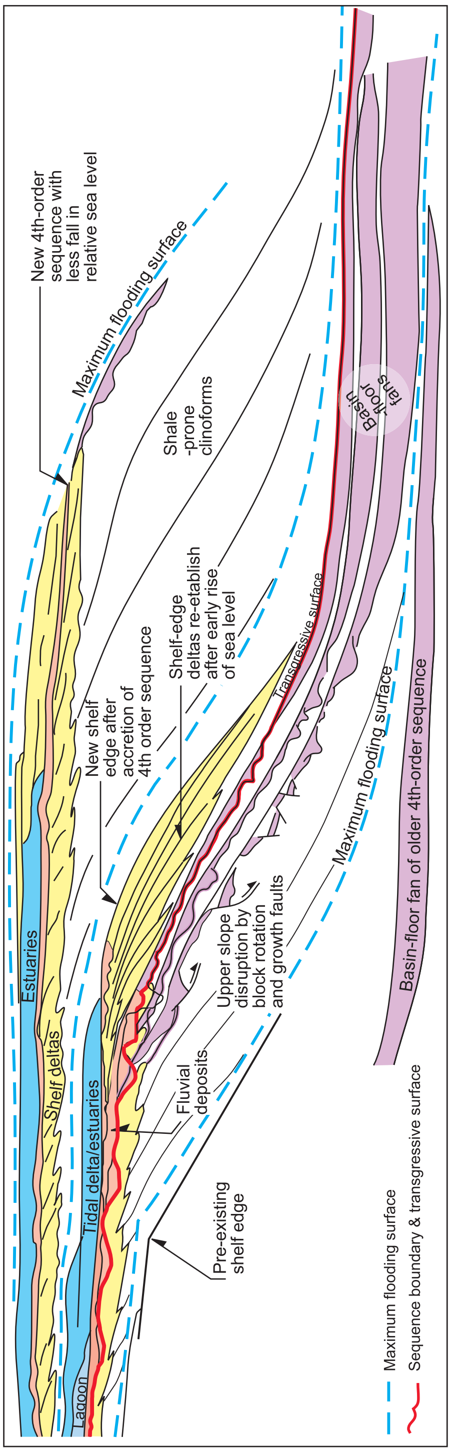

Fig. 17. (a) Conceptual diagram for sand delivery to deep water emphasizing that (a) there can be multiple fans developed from a single fourth-order clinoform, and (b) the common continuity (albeit spatially irregular) of sand fairways observed in the Spitsbergen and Porcupine clinoforms.

ward trapping. These fans are overlain by more muddy channel-levee systems, that generally are located slopewards of the fans, before being downlapped by the late lowstand prograding wedge, a muddy wedge of distal, shelf-edge delta deposits (Posamentier *et al.*, 1991).

The Porcupine and Spitsbergen data discussed herein provide an alternative model that we think may be widely applicable to clinoformed successions in areas with narrow to medium-width shelves in front of the shelf margin. In this setting, the sediment supply/accommodation interaction is such that significant shelf-margin accretion occurs although the basin is being entirely infilled (sediment over-filled basins of Hadler-Jacobsen *et al.*, 2004). In this scenario the deepwater basin-floor and slope deposits are fed directly or indirectly, mainly but not exclusively when relative sea level is low, from coeval deltas at the shelf edge, out onto the slope clinoforms. The critical aspect of this model is that there is little or no slope onlap generated during the sea level rise back to the shelf edge, but instead there is aggradation along much of the length of the clinoform. The degree to which deepwater sandstones on the basin floor or slope are detached or attached to the shelf edge system will be discussed in the following section.

Porcupine: sand continuity along true depositional-dip lines

A random seismic line across the Porcupine 3-D data set shows clinoform development with discontinuous, high-amplitude reflectors, representing apparently discontinuous sandstones within an overall shaly slope to basin-floor setting (Fig. 12a). However, if the seismic line is selected carefully (based on an amplititude map) so as to link together features such as fans, lobes and channels, then the high amplitudes show continuity along the clinoform (Figs 1 and 10). For example, line A–A′ on the lower-slope reaches of the clinoform (Fig. 1) shows high-amplitude reflectors that apparently pinch-out updip on this seismic line. This gives an impression that the basin-floor fan is detached from any sands higher on the slope. In the conventional slope-onlap model, this scenario could predict a stratigraphic hydrocarbon trap in the updip areas. However, if the seismic line had been chosen along transect B–B′ (Fig. 1), a more continuous high amplitude would have been seen along the clinoform from basin floor to shelf edge, indicating that the fan was attached updip to good slope sands (Fig. 10a). Thus, random seismic lines along slope to basin-floor transects are likely to show apparent updip sandstone pinchouts, and most of these are likely to be simply sidelap features. Dip lines aligned down the axes of continous sandy fairways will show that stratigraphic traps are unlikely. In general, in clinoformed successions similar to those discussed here, we would caution against slope-trap interpretations based on two dimensional seismic lines and 3-D seismic diplines that are not tracking along specific fairway systems. Comparison of three parallel lines with a dip line aligned down the sand

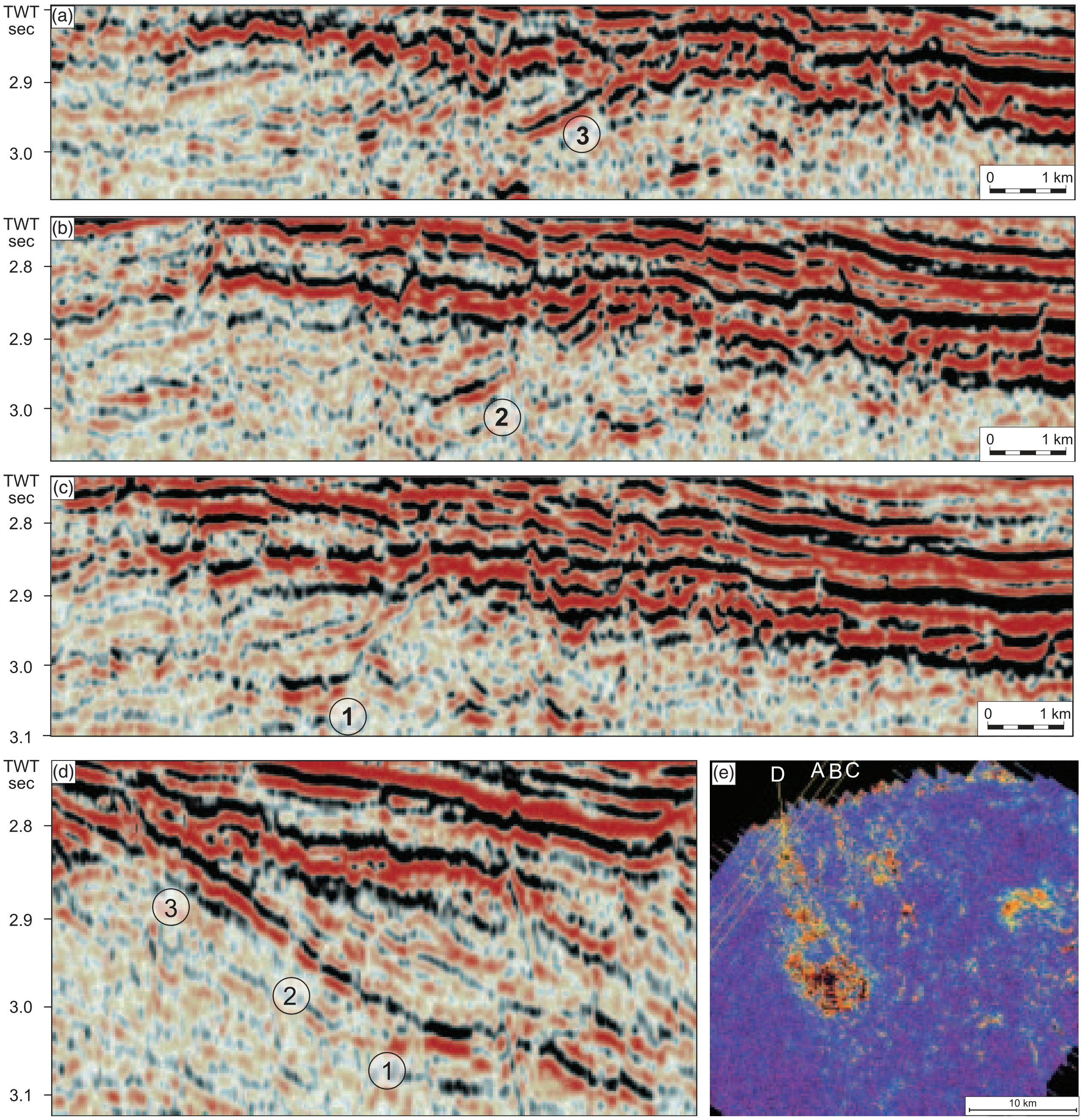

Fig. 18. Three proximal, parallel seismic lines (slightly oblique to depositional dip) showing relatively steeply-dipping clinoforms. The three circled high-amplitude reflectors on lines A–C, represent apparently isolated channel features encased in muddy slope deposits. Line C is most distal with a totally isolated channel in strike view. A is proximal and shows that the channel feature is more elongated and attaches directly up onto the flat-lying shelf-edge delta. Seismic line D is a dip line linking the high amplitudes (slope channels) in A and C together, giving a sigmoidal, shelf-edge-attached, continous high amplitude in the upper slope area. E: RMS amplitude map generated along high amplitude in D and shows location map for seismic lines (Fig. 10d–f).

fairway axis (Fig. 18) shows how critical it is to select true dip lines to confirm whether a disappearance of high amplitudes is caused by sidelap (no trapping) or by true updip pinchout (trapping). In most cases of slope or shelf-edge trapping there has to be an additional tectonic overprint; for example, thrusts faults or four-way closures related to shale or salt diapirs or extensional faults/syn-sedimentary slump faults in the upper-slope.

Spitsbergen: sand-prone channels on upper, middle and lower slope

Data from the Svalbard examples strongly suggest similar conclusions as regards common continuity of sandstones along the slope to basin-floor segments of clinoforms (Fig. 17). Basin-floor fans are quite common, and are found or predicted in half of all clinoforms along the Van Keulenfjorden transect (Fig. 6a), though not necessarily so frequently

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along other transects in the same succession. Apparent toe-of-slope sandstone pinchout (updip) is observed on Storvola (Clinoform 14, Figs 1 and 6a) and Areniusfjellet (Fig. 13). Paleocurrents readings indicate that these sandstones have their depositional dip at a slight angle to the section of the mountainside, and that the sandstone pinchout is likely to be a sidelap effect, very similar to high-amplitude terminations in the seismic data from Porcupine Basin (Fig. 1). Well developed channels are found in the middle reaches of the slope segment of Clinoform 5 on Palfjellet (Figs 6a, 11 and 12b) and good examples of upper-slope, turbidite-filled channels being fed directly by sandy delta distributaries are seen in Clinoforms 7–9 on Brogniartfjellet (Figs 14 and 15). The Svalbard 2-D outcrop data set shows that sandstones occur on all reaches of the clinoform slopes. Using occasional 3-D information from the outcrops and information from the 3-D data set from Porcupine, we conclude that slope-to-basin floor clinoforms develop with continuous sandstone fairways at some locations (albeit narrow in places) on the slope (Fig. 17). This view is consistent with the 'aggradational' style of build-up style of these clinoforms, as discussed below.

Spitsbergen and Porcupine: aggradation on the clinoform rather than slope onlap

In addition to continuity of sand along slope channels/fairways, there is evidence in both data sets of significant muddy-slope aggradation, both during and after basin-floor fan growth. This aspect is clearly at odds with conventional slope-onlap models (see also Saller *et al.*, 2004), and suggests rather that there was some sediment aggradation on the clinoform through much of the fourth-order cycle. Although aggradation was most marked during base-level rise, we have also documented this during the base-level fall half cycle (see also Blum & Tornquist, 2000). Some of the early slope channels that fed the basin-floor fan were clearly filled, were abandoned and were buried in shales whereas other slope channels became active and continued to feed the basin floor. This sometimes resulted in several basin-floor fans developing within the early phase of a single fourth-order cycle (Fig. 9a, b). The early sandy slope fairways accumulated within an aggrading shale-prone slope environment that resulted in a 30–50-m-thick package on the slope segment of the clino-

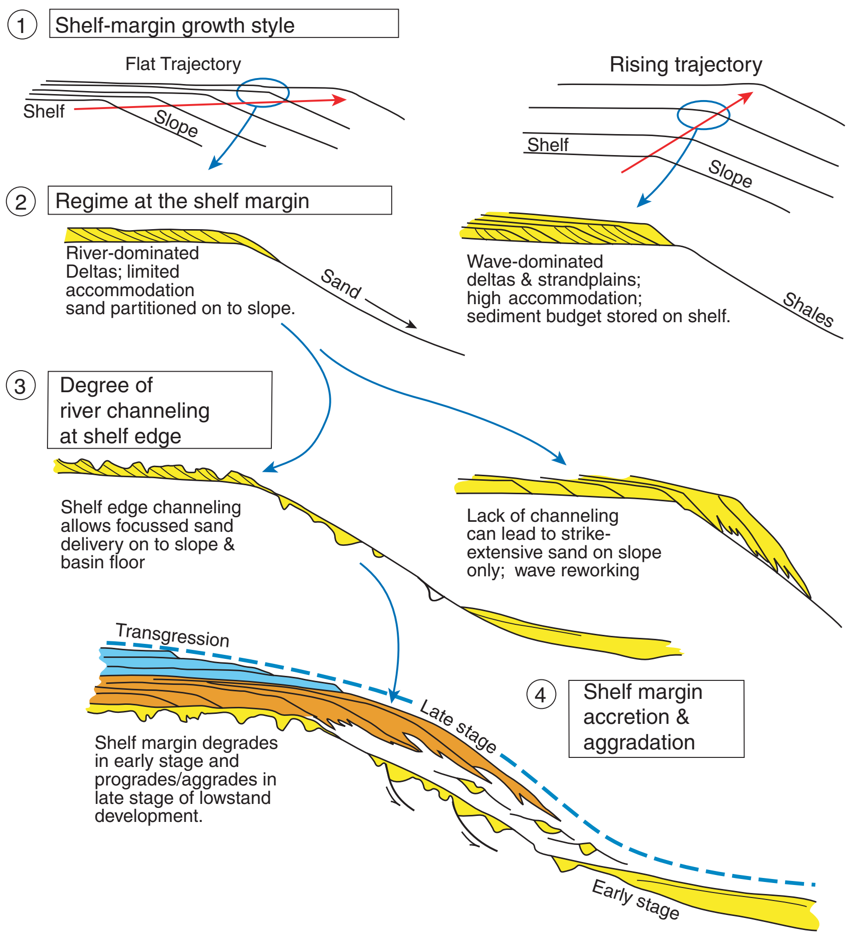

Fig. 19. Summary diagram outlining the importance of (1) shelf-edge growth style, (2) sedimentary regime on the outer shelf, and (3) degree of river channelling at the shelf edge for prediction of presence or absence of deepwater sands on the clinoform, and the eventual partitioning of this sand between slope and basin-floor areas.

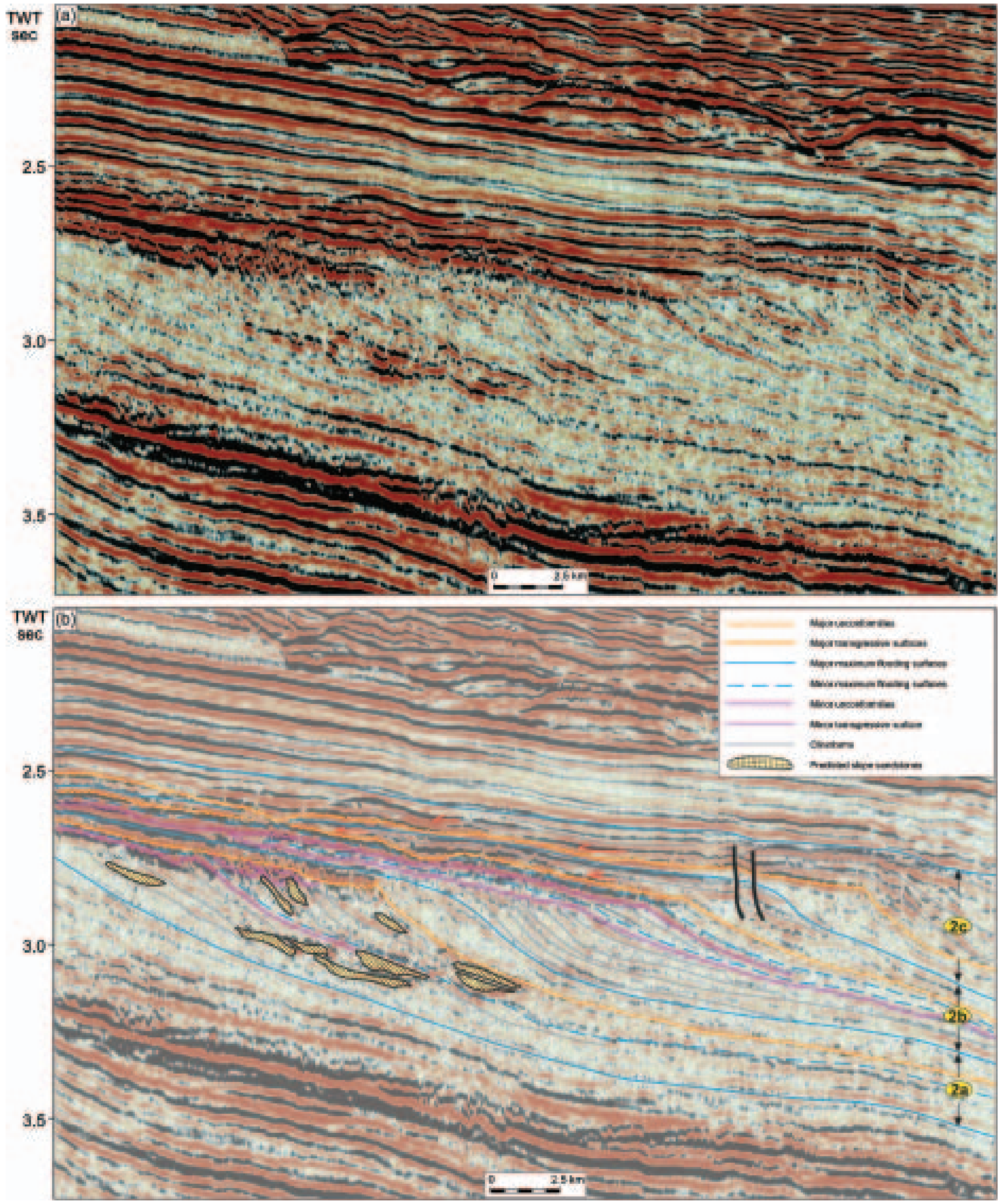

Fig. 20. (a and b) Un-interpreted (a) and interpreted (b) seismic lines (slightly oblique to depositional dip). Sequence 2a shows shelf-edge progradation with a flat trajectory, and single and clustered high amplitudes (channel–lobe sandstones) on the slope area, suggesting sand on the clinoform slope. Sequence 2b and c were deposited with a rising progradational trajectory, show lower amplitudes in the slope area, thus suggesting lack of sandstones on the slope. Scattered and clustered lenses of high-amplitude reflectors are interpreted as sandy slope channels, gullies or debrites location (Fig. 4d).

form. This package accumulated before a downlapping wedge of muddy deltaic sediment, which, in turn, was overlain by the maximum flooding interval of the fourth-order sequence (Fig. 3b).

DEEPWATER SANDSTONE PREDICTION

This study strongly suggests that mapping the trajectory patterns of shelf-edge clinoform sets is a powerful tool for prediction of sandstones below the shelf edge (Fig. 19).

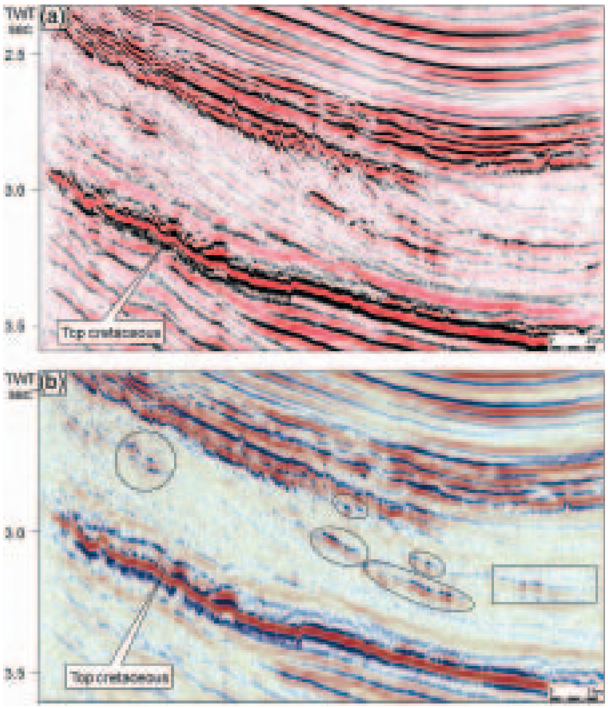

Fig. 21. (a) Near offset seismic line with several high amplitudes on the slope (b). Same line as in (a), but displayed as far offset. High amplitudes indicate sandstones (circled). Another area (marked with a rectangle) has lost amplitude from near to far offset, which gives an indication of a lack of good sandstone.

Determination of facies/processes of the sediment-delivery system at the shelf margin and identification of significant channels at the shelf edge or on the uppermost slope are important additional predictive parameters (Fig. 19). These indicators can be the first tell-tale signs of the presence/absence of deepwater sands farther downslope.

Consider a seismic line, slightly oblique to depositional dip, indicating a cluster of high amplitudes (interpreted as channel–lobe complexes) in the lower-slope segment of clinoforms (e.g. sequence 2a in Fig. 20a, b). Sequence 2a shows shelf-margin progradation with a flat trajectory style; strong, horizontal seismic reflectors are closely spaced, indicating amalgamation of clinoform topsets and suggesting a low accommodation setting during shoreline transits and delivery of sand to the outer shelf and shelf edge. All of these features are consistent with a prediction of significant sand delivery into deepwater areas of the clinoforms, and this in turn is consistent with the features seen on the amplitude maps of these horizons (Fig. 8e). Consider also sequence 2b that, in contrast, is associated with a rising shelf-edge trajectory where the clinoform topsets are high-amplitude, horizontal reflectors (shoreline sandstones) separated by weaker amplitudes, interpreted as transgressive shales across the shelf (Fig. 20a, b). This higher-angle growth trajectory, and less amalgamated shelf succession suggests a higher accommodation setting, making it likely that a higher proportion of the sediment budget was stored on the shelf and coastal plain, with correspondingly less sediment available for delivery into deepwater. The amplitude map of this sequence shows no channel features on the slope (Fig. 8f), and confirms that searching for sand here is a high risk activity. Sequence 2c (Fig. 20a, b) represents the level of maximum progradation of the entire delta system in this part of Porcupine Basin, but there is incomplete seismic-data coverage at the end of the 3-D data set. However, the shelf-edge shows a rising trajectory style near its termination, with several small isolated (probably also fault-related) high amplitudes on the slope.

These features are interpreted as slope gullies and possibly debris flows created because of muddy slope instability at the aggrading shelf break (see also Galloway, 1989). We would predict a lack of fan development on the basin floor of 2c.

Most seismic interpretation is done on full-stacked seismic. However, based on seismic high amplitudes, there is a tendency to over-interpret sandstone presence, thickness and net to gross. However, AVO analysis in shallow sections gives a clear brightening of the best sandstones (compare near and far offset data in Fig. 21).

CONCLUSIONS

1. Seismic and outcrop examples of Eocene shelf-margin clinoforms in Porcupine Basin and in Spitsbergen together allow an unusually detailed examination of relationships between shelf-edge trajectory and differing architecture/linkage along individual shelf edge-slope-basin floor clinoforms. The analysis allows prediction of the likely presence (or absence) of sand that has bypassed the shelf break, to produce deepwater turbidite accumulations.
2. The presence of river-dominated shelf deltas, a channelized shelf-edge and slope morphology, and a shelf edge that shows a persistently flat trajectory through time together indicate that significant volumes of sand have been delivered from the shelf edge out onto the slope. These conditions are commonly accompanied by accumulation of basin-floor fans.
3. The presence of wave-dominated shorelines, or of river-dominated deltas without significant channelling or downcutting at the shelf edge, together with a persistently rising shelf-edge trajectory, is more likely to indicate a lack of focused sediment bypass and no fan accumulation on the basin floor, though some sands can accumulate on the upper slope as shelf-edge attached sheets under these conditions.
4. Persistently flat or falling shelf-edge trajectories show extensive development of sandy, early slope channels but no development of late lowstand wedges on the upper slope immediately before shelf transgression. In contrast, flat to slightly rising trajectories develop both early sandy slope channels and late heterolithic prograding wedges on the upper slope, before shelf transgres-

sion. In both of these cases, basin-floor fans can develop at an early stage in the clinoform's development.

5. During progradation with flat trajectory style, channel continuity from toe of slope to shelf edge can be seen, so that mud-filled channels or shelf-edge structure is needed to trap hydrocarbons. Little slope onlap has been observed because there is notable slope aggradation.
6. During the regressive stage of clinoform generation, the sediment budget is preferentially partitioned into the shelf and deepwater segments, whereas, during the transgressive stage, the sediment budget is partitioned mainly into the coastal plain and shelf segments. With a flat or downward-directed shelf-edge trajectory during the regressive stage, a relatively high proportion of the budget at this time is partitioned into deepwater areas, with thinner accumulations on the shelf, whereas with a rising shelf-edge trajectory during the regressive stage much more of the budget is stored in the thicker coastal plain and shelf units, with correspondingly less sediment available for eventual deepwater deposition.

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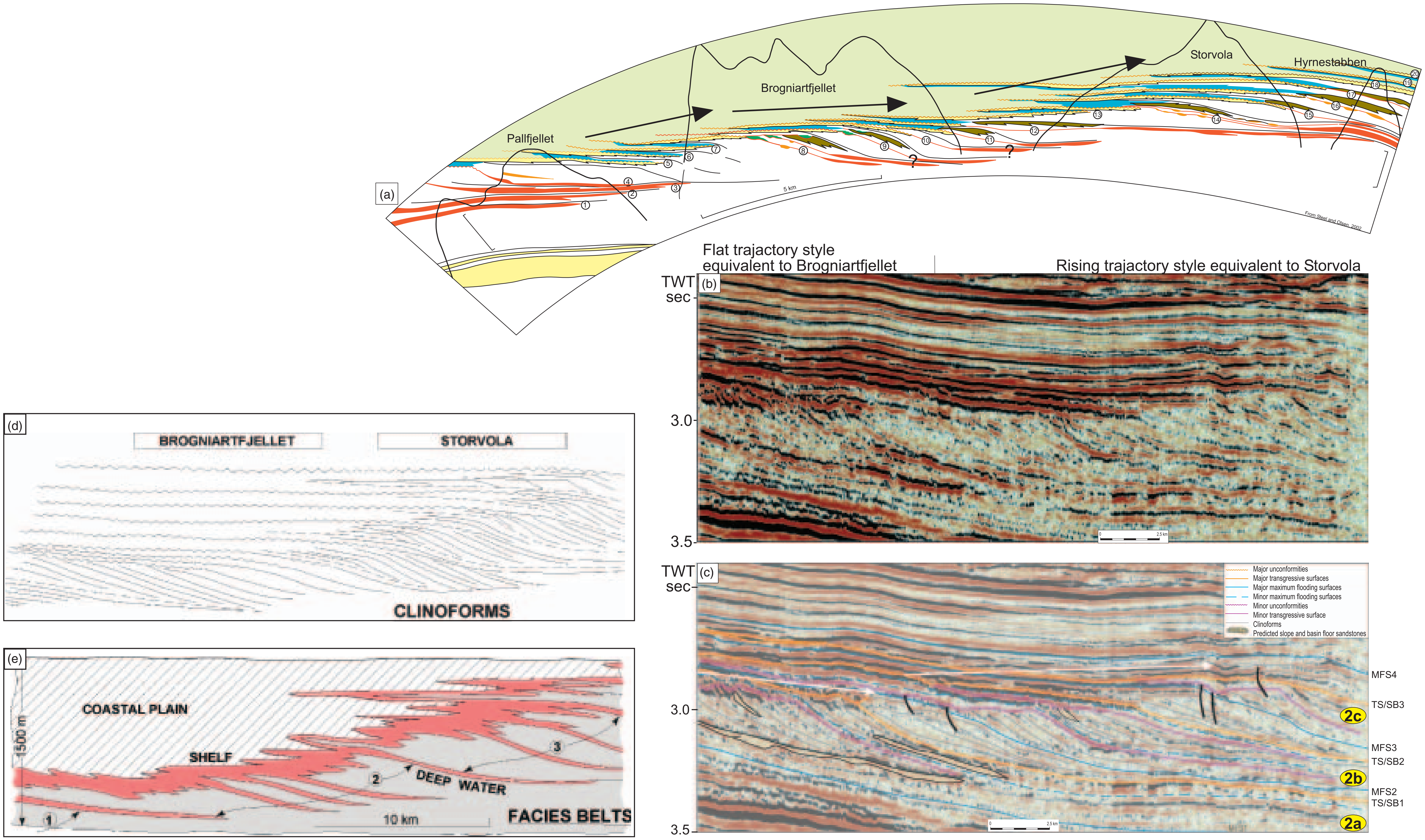

Fig. 6. (a) Van Keulenfjorden clinoform transect. The large arrows show long-term, third-order shelf-margin accretion trends, containing fourth-order clinoforms (labelled 1–20). Green = coastal plain, yellow/blue = shelf, brown/white = slope and white/red = basin floor (diagram modified from Steel & Olsen, 2002). (b and c) Porcupine Basin un-interpreted (b) and interpreted (c) oblique 3-D-dip-line at about same scale as Van Keulenfjorden transect. Third -order sequences 2a–c are bracketed by major flooding surfaces (see also Fig. 7b, c) and shelf-edge maxima occur at the termination of individual high-amplitude shelf reflectors. fourth-order sequences occur between minor maximum flooding surfaces. See text discussion on trajectories. (d and e) Schematic clinoform transect along Van Keulenfjorden, Spitsbergen showing both clinoform geometries (d) and main facies (e). Numbers 1–3 in (e) denote the three third-order sequences in the succession.

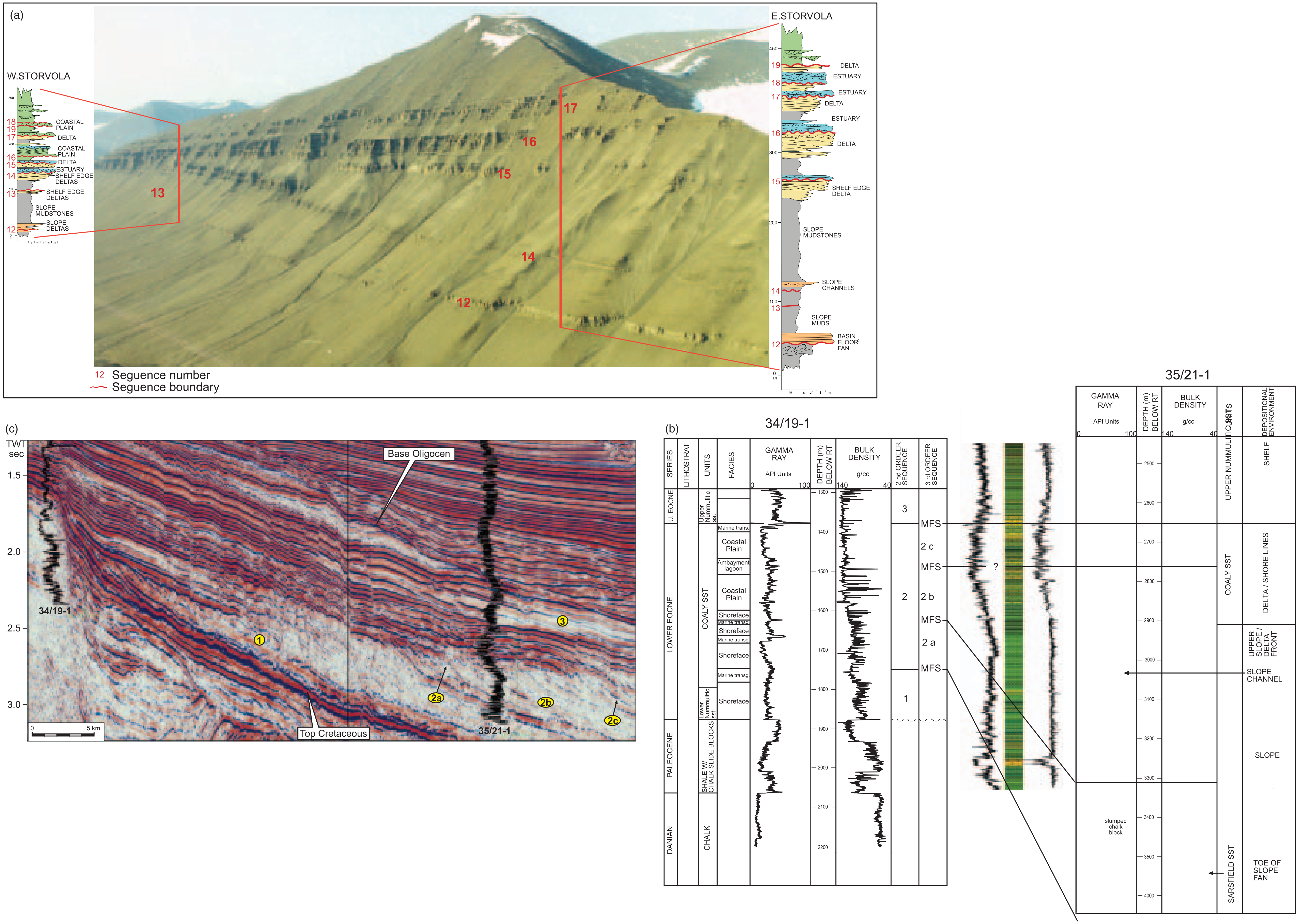

Fig. 7. (a) Contrasting sedimentary logs from the western (shelf) and eastern (deeper water) ends of the mountain Storvola in the context of clinoform development. Note the importance of slope gradient in picking the correct correlation. The basinal log is much more shale-prone than the shelf log. Note the estuarine character of the transgressive and deltaic character of the regressive segments of the shelf reaches of sequences. (b) Porcupine Basin log correlation from shallow-water (well 34/19-1) to basinal area (well 35/21-1), with corresponding dramatic thickening of succession with change from shallow to deep facies in Sequences 2a and b. Sequence 2c has a uniform thickness in the two wells, indicating basin fill and continous delta progradation across older shelf platform. (c) Seismic line showing the location of the proximal and distal wells in (b) (location, Fig. 4d).

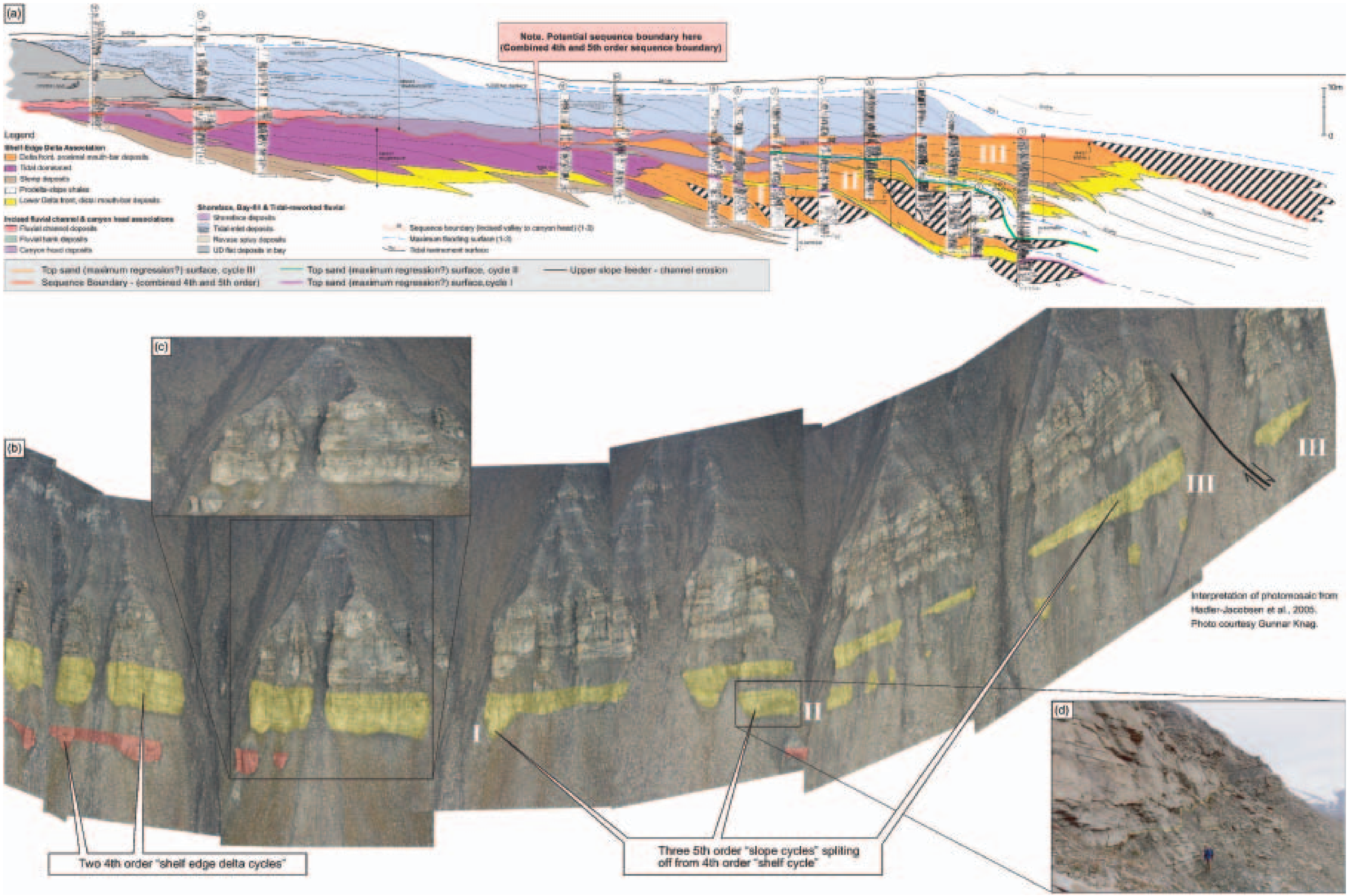

Fig. 14. (a) Interpretive reconstruction of the shelf and shelf-edge reaches of Clinoform 8 on Brogniartfjellet (Fig. 6a). The lower half of the diagram (*ca.* 20 m) is regressive whereas the upper part is backstepping and transgressive. In the regressive part, there are prograding (to the right) shelf-edge deltas that become cut by a series of channels on the shelf edge that then connect with upper-slope channels. There is a complex transition from the distributary channels on the outer shelf to turbidite-filled channels on the upper slope. The entire regressive tract is fourth-order, and the channel-dominated segments that peel off onto the slope are fifth-order. Data profiles and much of diagram are from Mellere *et al.* (2003), but shelf edge area interpretation is modified. (b) An interpreted, slightly oblique photomosaic (basis for (a) above) showing shelf-edge progradation of Clinoform 8, eastern part of Brogniartfjellet., The photo also shows how the fourth-order deltaic sequence at the shelf edge splits up into three fifth-order clinoforms. (c) Detail of what may be 12–15-m high, steep-fronted, subaqueuos mouth-bar foresets on the shelf-edge area. The foresets toplap against a multistory estuarine channel fill (see also (a) and (b)), and the intervening boundary is interpreted as a fourth order sequence boundary. (d) Example of the subaqueous reaches of river channels (> 9-m deep) cutting into older delta-front sands at the shelf edge. The channel is filled with fine and medium-grained, erosively based well laminated to current-rippled sandstone beds, possibly originating from river-generated hyperpycnal underflows.