Available online at www.sciencedirect.com
SCIENCE DIRECT®
ELSEVIER
Earth-Science Reviews 62 (2003) 283–326
EARTH-SCIENCE REVIEWS
www.elsevier.com/locate/earscirev

Shelf-margin deltas: their stratigraphic significance and relation to deepwater sands

Szczepan J. Porębski[a,*], Ronald J. Steel[b]

[a] *Polish Academy of Sciences, Institute of Geological Sciences, Kraków Research Centre, Senacka 1, 31-002 Kraków, Poland*
[b] *University of Wyoming, Department of Geology and Geophysics, Laramie, WY 82071, USA*

Received 5 November 2001; accepted 12 November 2002

Abstract

Shelf-margin or shelf-edge deltas are a common constituent of Quaternary shelves, where their genetic link to falling and lowstand sea level has been well established, but they are rarely reported from older successions. Recognition of this delta type in the ancient record is nevertheless important, because (1) it helps to position the fossil shelf edge, (2) it provides insight into how sand budgets were partitioned between shelf edge, slope and basin-floor settings, and (3) it can contribute to the debate concerning three versus four systems-tract systematics and the positioning of a sequence boundary.

Shelf-margin deltas create wide (tens of kilometers), high (hundreds of meters) and steep (3–6°) clinoforms, because they build across the relatively deepwater shelf margin. Such bodies tend to form strike-elongate wedges that initially thicken basinwards, attaining a maximum thickness of 50–200 m just outboard of the shelf edge, then thin on the middle/lower slope, whereas they commonly pinch out landwards by lapping back onto shelf shales. Prior to reaching the shelf edge, small-scale (tens of meters), tangential foresets (usually $<3°$ slope) of the shelf delta downlap onto the preexisting outer-shelf surface, whereas below the shelf edge, the shelf-margin deltaic foresets become longer, steeper (3–6°) and more turbidite prone, and they downlap onto the preexisting slope of the shelf margin. The top of the shelf-edge delta package is a sharp, planar to incised surface that truncates the delta increasingly landwards, merging eventually with the basal downlap surface. The clinoform series created by this transition from outer-shelf to shelf-margin deltas tends to have a horizontal-to-downward trajectory and, eventually, in the most distal bundles, an aggrading to backstepping trajectory as relative sea level begins to rise.

The key facies association in shelf-edge deltas is the mouth bar-to-delta front association. Mouth-bar facies, landwards of the shelf edge, consist mainly of thick, clean, flat to low-angle and ripple-laminated medium to fine sands. Basinwards from the shelf edge, such sands commonly alternate with heterolithic slumped unit, creating characteristic slumped to laminated couplets. Because the delta front is superimposed on a preexisting, steep and extended shelf margin, it commonly contains (beyond the mouth bars) thick successions (up to many tens of meters) of sandy, slope turbidites. Shelf-margin deltas differ from inner shelf deltas in showing: (1) an order of magnitude higher clinoforms (hundreds instead of tens of meters), and strike-elongated, locally pod-like sand bodies commonly affected and augmented by growth faulting; (2) paleoecological evidence of abrupt shallowing (foreshortened stratigraphy); (3) turbidite-prone delta-fronts; (4) larger scale and greater abundance of slope-controlled soft-sediment deformation and (5) the general absence of a paralic 'tail' along the trailing edge of the delta front.

Shelf-margin deltas can be classified into two types. Stable shelf-margin deltas are usually tens of meters thick and are not associated with shelf-edge incision or major slope collapse/disruption features. Slope turbidites are common and occur as unconfined sheets and lobes. Such deltas form when relative sea level falls no lower than the level of the shelf platform or, after

* Corresponding author. Fax: +48-12-422-16-09.
E-mail addresses: ndporebs@cyf-kr.edu.pl (S.J. Porębski), rsteel@uwyo.edu (R.J. Steel).

doi:10.1016/S0012-8252(02)00161-7

longer sea-level falls, when rivers reestablish at the shelf edge on the rise. Unstable shelf-margin deltas are associated with large-scale slope collapse, listric growth faults and sometimes with salt diapirs. Slides are common on the upper slope, whereas imbricated sediment packages and compressional ridges affect the slope toe. Slope turbidites can be ponded within structurally controlled mini-basins on the slope. Such deltas are often related to unusually great sand influx to the shelf edge, or simply to prolonged or large fall of relative sea level below the shelf edge.

The diachronous erosional unconformity atop the shelf-to-shelf-edge deltaic wedge is viewed by some researchers as the most easily recognizable and persistent surface on the shelf and slope. This surface contrasts with the downlap erosional surface developed from the beginning of sea-level fall, which is viewed by other researchers as the key boundary surface within this complex. It should be noted that sea level can fall for a long period of time (tens of thousands of years) before deltas even reach the shelf margin and, therefore, before significant volumes of sand are delivered across the shelf break. Hence, the time of incision of the shelf edge and emplacement of deepwater sand is commonly long after the initial fall so the time of (maximum) relative lowstand of sea level may be a more practical choice for the timing of the sequence boundary.

Recognition of shelf-margin deltas and analysis of their architecture help in the prediction of presence or absence of basin-floor fans. Deltaic complexes that are aggradational to backstepping, downlap onto disrupted or complex slopes, and overlie an incised shelf and shelf edge, predict that there should be basin-floor fans present. In contrast, delta complexes that show a prolonged and preserved progradational to downstepping architecture implies the presence of turbidite accumulations on the slope but not on the basin floor.

Keywords: Shelf-margin delta; Forced regression; Depositional systems tract; Sequence boundary; Basin-floor fan

1. Introduction

There is a range of delta types that develop across different segments of the shelf during a cycle of fall and rise of relative sea level (Porębski and Steel, 2001), particularly when the fall exposes the former shelf surface. The shelf-edge or shelf-margin delta is an end-member of that spectrum that becomes prominent during the falling stage of relative sea level on the shelf, but may also result from an unusually high supply of sediment, at other stages in the cycle. There has been much interest in shelf-margin deltas, because: (1) they form at the morphologic shelf break, the boundary between major facies belts at continental margins, as well as at the margins of other types of basins; (2) they probably represent the primary increments by which shelf margins grow (Morton and Suter, 1996); (3) they are the potential staging areas for supply of sand to the deep sea (e.g., Curray and Moore, 1964; Winker and Edwards, 1983; Suter and Berryhill, 1985; Mayall et al., 1992; Morton and Suter, 1996; Sydow and Roberts, 1994), even though this potential may not be realized unless the shelf-edge deltas become incised by their own driving rivers (Steel et al., 2000); and (4) they commonly are prolific reservoirs for hydrocarbons (e.g., Mayall et al., 1992; Hart et al., 1997).

Interest in shelf-margin deltas has grown with the increased attention paid to the stratigraphic response of depositional systems to falling relative sea level (Plint, 1988; Posamentier et al., 1992; Mellere and Steel, 1995; Plint and Nummedal, 2000). By definition, a shelf-margin delta requires its fluvial distributary channels to empty near the shelf break. This involves shoreline regression across the entire shelf, and commonly implies forced regression—i.e., a regression that is forced by active fall in relative sea level and, thus, is independent of sediment supply (Posamentier et al., 1992). Although unusually high sediment flux, combined with low-energy coast conditions, may bring deltas close to the shelf break during highstands (e.g., the Balize Lobe of the present-day Mississippi Delta), many shelf-margin deltas have been proven to be forced-regressive and lowstand features (Posamentier et al., 1992; Tesson et al., 1993; Sydow and Roberts, 1994; Hernández-Molina et al., 2000; Kolla et al., 2000).

There are a number of issues regarding these deltas that remain unsolved or controversial. (1) Why have pre-Pleistocene deltas been reported from the literature so rarely? (2) To what extent is the identification of deltas at the shelf edge a predictor of coeval deepwater sand on the slope or basin floor? (3)

Because shelf-edge or perched-slope deltas result from sea-level fall to the shelf platform or below, can they also help resolve the debate around the positioning of the sequence boundary, above or below the falling-stage or early lowstand tract?

In the present work, our objectives are: (1) to review the characteristics of shelf-margin deltas, and to show that they develop as one member in a spectrum of delta types developed from sediment delivery during a cycle of relative sea-level fall and rise on the shelf; we suggest that they are common in the pre-Pleistocene record, though rarely termed 'shelf-margin deltas' in the literature; in order to help identify them in ancient successions, we provide criteria with which to distinguish this member of the delta family from inner-shelf, highstand deltas; (2) to evaluate the connection, or lack of one, between the presence of shelf-edge deltas and the transport of sand down into deepwater areas. This point is significant if deltaic facies at paleoshelf edges are to be used as a predictor of the presence or absence of deepwater sands in basinal areas.

2. Delta classification, sea-level change and shelf-margin growth

Deltas have been classified traditionally in terms of depositional process–product reaction, as emphasized in facies modelling. Fisher et al. (1969) categorized deltas as high-constructive and high-destructive systems, depending on relationships between the fluvial input and the degree of reworking by marine processes. Galloway (1974) expanded this approach in his ternary classification, based on Holocene (i.e., sea-level highstand) examples, in which delta type is defined by the relative contribution of fluvial, wave or tidal energy flux that was dominant during deposition at the seaward edge of the delta. In other classifications of deltas, an emphasis has also been put on foreset/topset geometries (Gilbert, 1890; Fisk, 1961), sediment grain size (Orton and Reading, 1993) and delivery system type (Holmes, 1965; Chough et al., 1990; Friedman et al., 1992).

In these classifications, the overall delta regime reflects the sum of the environmental controls acting

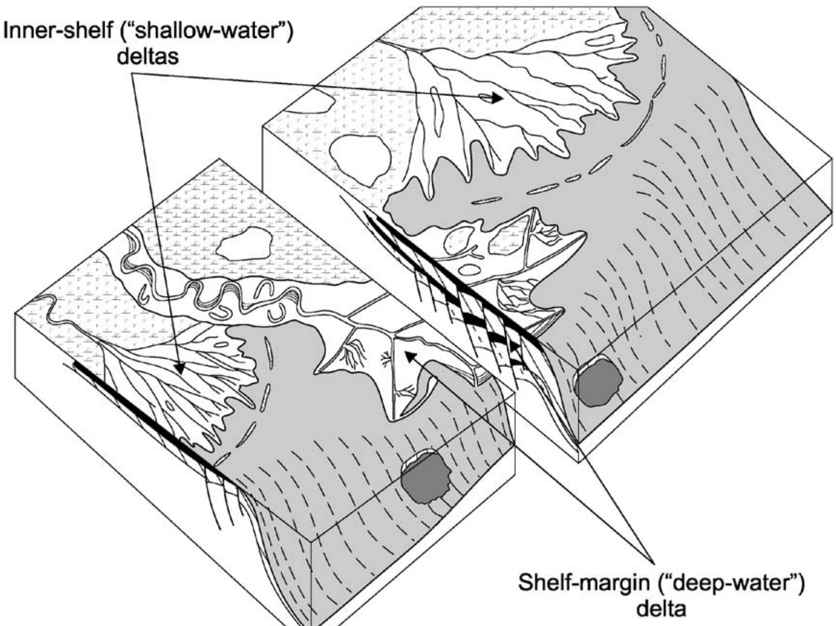


Fig. 1. Cartoonal representation of inner-shelf delta versus shelf-margin delta based on the Paleocene–Eocene Upper Wilcox Rosita delta system (after Edwards, 1981). Note the expansion of the clinoform height and the mouth-bar thickness in the shelf-margin delta. The model implies the contemporaneity of both types of deltas, as exemplified by the modern shelf off the Louisiana coast. However, it is now well proved that inner-shelf deltas tend to be predominantly highstand features while shelf margin deltas form during forced regression and early lowstand.

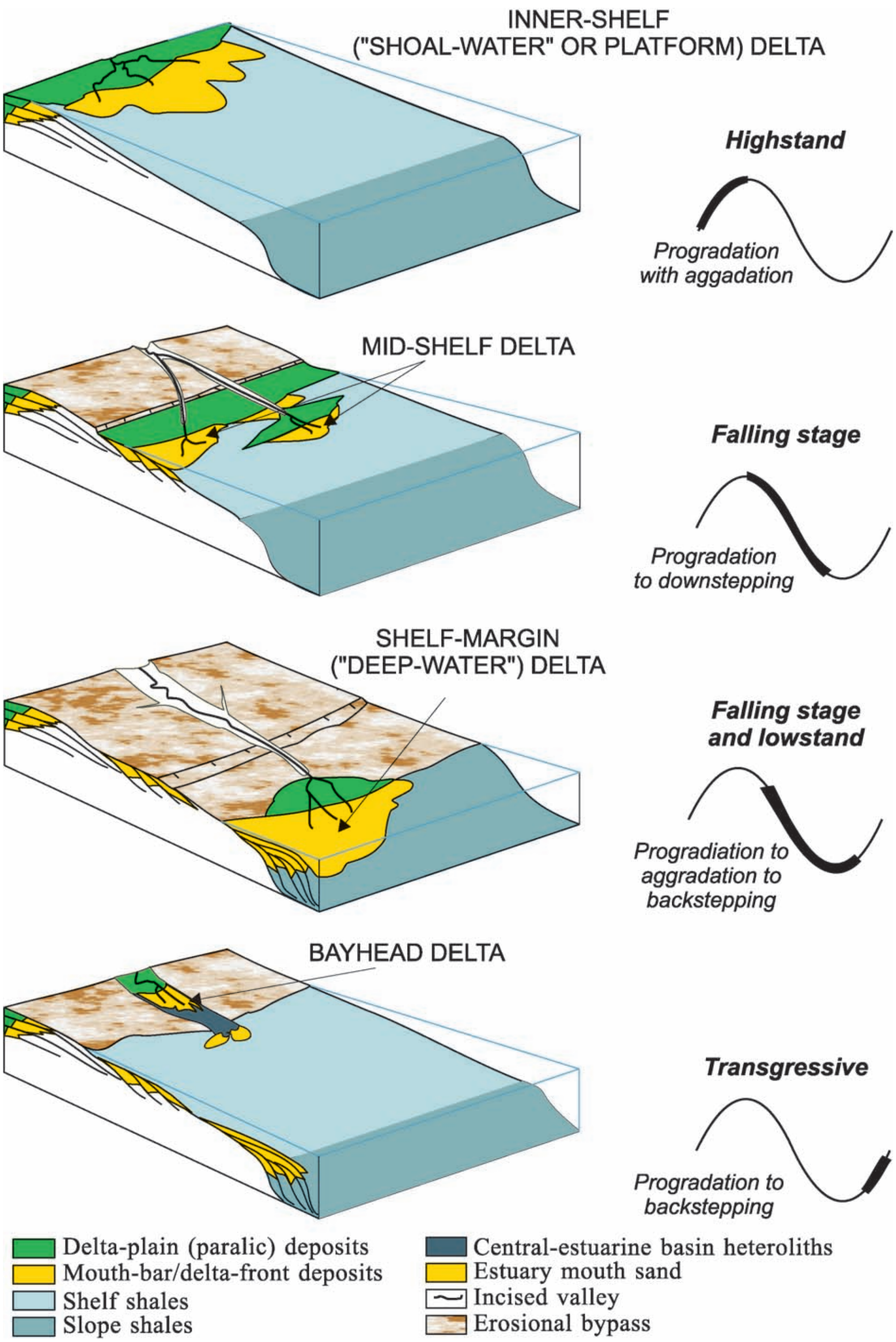


Fig. 2. Classification of shelf deltas in terms of relative sea-level change (based on Porębski and Steel, 2001).

on the deltaic system, which otherwise remains in a steady state, only incidentally affected by relative sea-level changes. More recent research has now emphasized how sea-level change (increased rate of sea-level rise) may generally change regressive deltaic shorelines into transgressive estuarine or barrier shorelines (Penland et al., 1988; Boyd et al., 1992), but there has been little attempt to show the effect of sea-level change on the deltas themselves (but see, Postma, 1995; Kolla et al., 2000; Porębski and Steel, 2001).

3. Deltas and relative sea-level change

Deltas can prograde to any location across the shelf, and some transit to the shelf margin itself, so that two principal types have been distinguished: (1) shoal-water or inner-shelf deltas; and (2) shelf-margin or shelf-edge deltas (Edwards, 1981; Suter and Berryhill, 1985; Reading and Collinson, 1996, p. 181). Although deltas formed on the middle and inner shelf are different from those sited near the shelf edge (Edwards, 1981; Winker, 1982; Elliot, 1989), the importance of this has not been well appreciated so far among facies modellers and sequence stratigraphers. On the basis of his work in the Paleocene–Eocene Rosita delta system in the Gulf of Mexico, Edwards (1981) suggested that inner-shelf deltas tend to be "horsetail" and develop a wide delta-front sheet sand, whereas those that reach the shelf break are characteristically locally thickened in response to on-growth faulting (Fig. 1). Suter and Berryhill (1985)

Table 1
Main distinctive characteristics of shelf deltas (see also Kolla et al., 2000, and references therein)

	Bayhead deltas	Inner-Shelf Deltas	Mid-Shelf Deltas	Shelf-Margin Deltas
Shape	Confined, funnel-shaped	Birdfoot, lobate to cuspate	Erosional pods common	Lobate to strike-elongate
Clinoform slope	?	Usually < 0.001*	$< 0.5°$, but can be as much as $8°$**	Up to $8°$, usually $3–6°$
Clinoform height	Several meters	Few tens of meters	Few tens of meters	Several hundreds of meters
Thickening trend	Landwards	Landwards	Seawards, or no distinct trend	Seawards, with maximum near shelf break
Dominant energy flux at delta front	Fluvial; can be tide influenced	Fluvial, wave or tide	Fluvial or wave	Fluvial or wave; tide influence may become important during late lowstand
Delta-top facies	Fluvial distributaries	Delta plain with thick distributary channel sand	Thin to absent delta plain; thick distributary channel sand below incised valley systems; erosional bypass in places	No delta plain; thin fluvial distributaries where delta perches below shelf edge. Erosional bypass common
Delta-slope facies	Heterolithic, ripple-laminated foresets	Varies depending on delta front regime	Sandstone to heterolithic foresets; rare turbidite sands	Sandstone to heterolithic foresets; slumped mouth-bar sand embedded in prodelta shales; common hyperpycnal sand turbidites. Growth faults and diapirs
Updip termination	Intertonguing with paralic deposits	Intertonguing with paralic deposits	Erosional pinchout within mid-shelf shales	Erosional pinchout within outer-shelf or upper slope shales
Downdip termination	Pinchout within central estuarine-basin heteroliths	Pinchout with inner-shelf to mid-shelf shales	Pinchout within outer-shelf shales	Pinchout within slope shales
Clinoform breakpoint trajectory	Landward rising to basinward rising	Basinward rising	Horizontal to basinward falling	Basinward falling, aggraded to landward rising
Systems tract	Transgressive	Highstand	Falling stage	Falling stage to lowstand

*See Table 6.2 in Reading and Collinson (1996).
**Posamentier and Morris (2000).

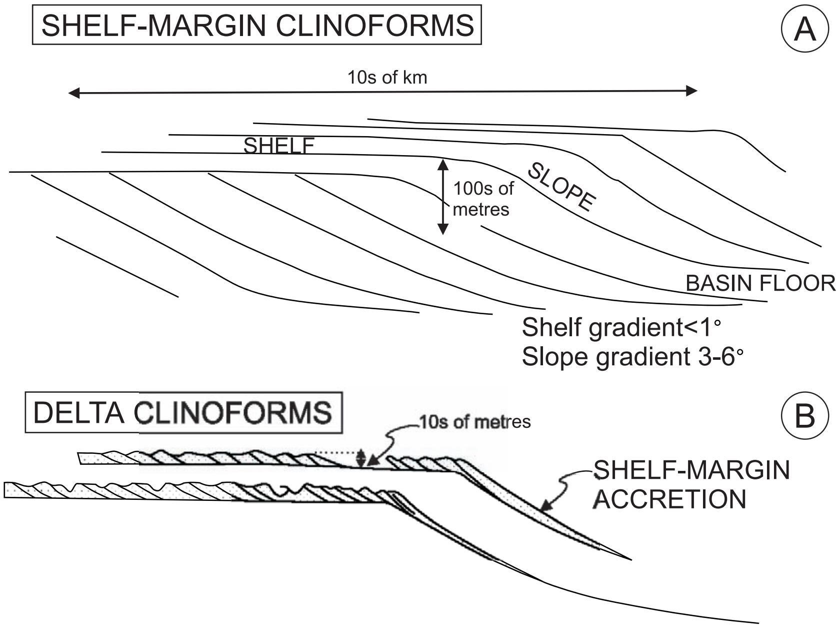


Fig. 3. Distinction between clinoforms generated by (A) shelf-margin accretion, and by (B) prograding shelf deltas. Note that the two types coincide only when deltas reach the shelf edge and drape partly down the slope.

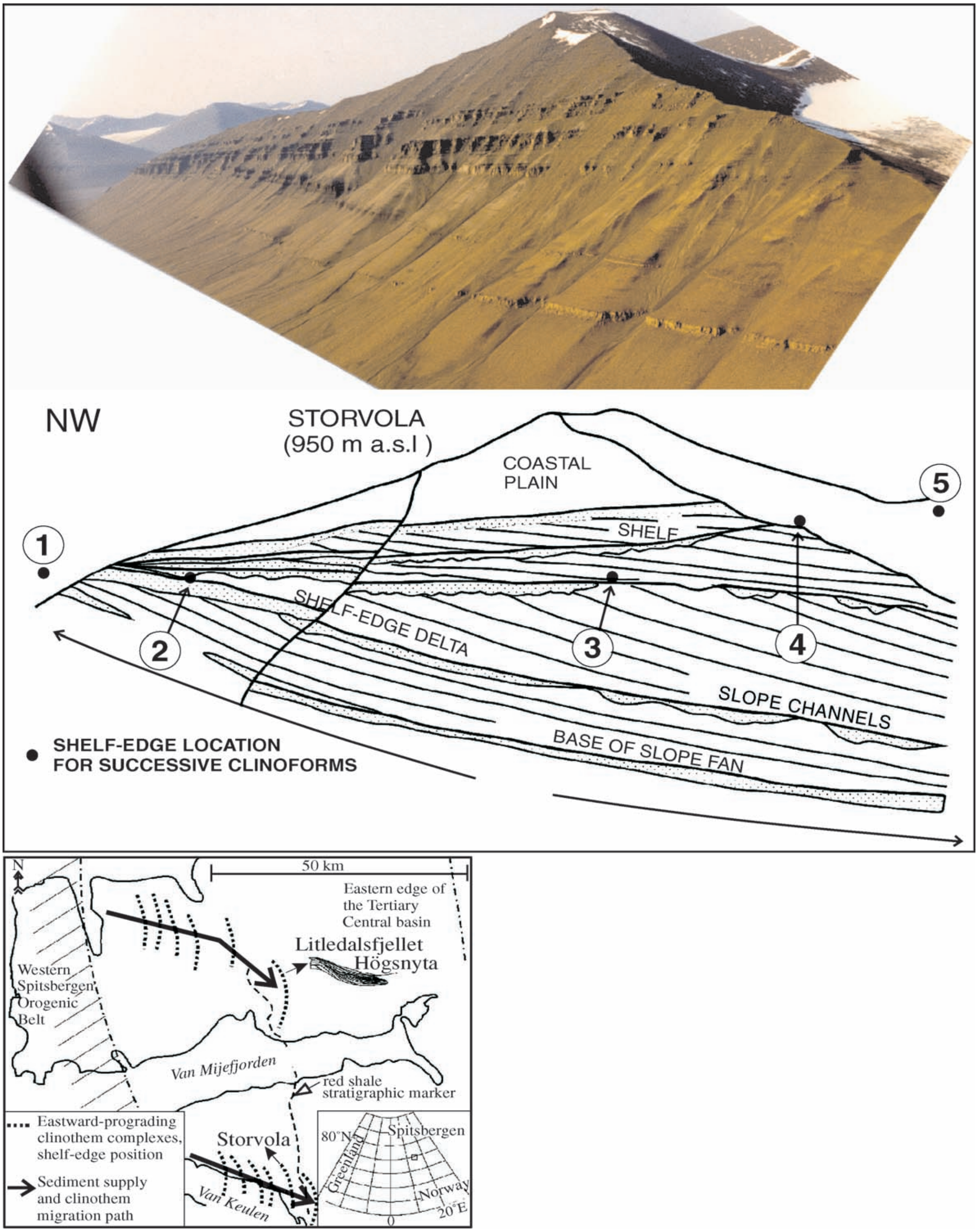
demonstrated, in the Pleistocene deposits of Gulf of Mexico, that prograding deltaic clinoforms sited on the middle shelf tend to be thin, patchily developed and show low-angle clinoforms, whereas those located near the shelf margin and on the upper slope produce thick, localized, strike-oriented wedges of steeply dipping strata, commonly bearing signs of slope failure. These differences are believed to reflect increased accommodation, when the deltaic shoreline, having transited the shelf during falling sea level, meets deeper water near the shelf break (Suter and Berryhill, 1985). At the opposite extreme, a regional sea-level rise that transforms the deltaic shoreline into an estuary will eventually result in small, dip-elongated and often tide-dominated bayhead deltas formed at the landward end of flooded valleys (Dalrymple et al., 1992; Nichol et al., 1996). Inner-shelf (platform or shoal-water) deltas, in turn, tend to be wide and backed by thick paralic facies, reflecting their aggradational-to-progradational (highstand) delta-front trajectory during formation.

Thus, for a given tectonic setting and sufficient sediment supply, it is the rate and direction of relative sea-level change that produces different delta types, each of them displaying a fairly distinguishable set of stratigraphic characteristics (Fig. 2, Table 1). For a given sea-level cycle, these types are likely to form a predictable evolutionary pattern in which they either form intergradational transitions, or occur as distinct entities separated in space and time by major discontinuities.

4. Shelf generation and shelf-margin accretion

Although the term shelf is commonly used to denote the fringe of a continent—the continental shelf—we suggest that the same term is fully appro-

Fig. 4. Example of a shelf-edge delta (labelled 2) draping halfway down a shelf-margin slope on Storvola, Spitsbergen. The sandstone bench below represents basin-floor fans of an older clinoform that peeled off the shelf at the left end of the mountain. The sandstone bench above has its shelf edge (labelled 3) a few kilometers to the SE. The five clinoform complexes migrated successively to the SE. The overall trajectory of the shelf edge (see successive shelf-edge locations) also reveals an aggradational component. Map shows this and other localities mentioned further in the text with respect to Eocene shelf-margin deltas in Spitsbergen.


NW
STORVOLA
(950 m a.s.l)
COASTAL
PLAIN
SHELF
SHELF-EDGE DELTA
SLOPE CHANNELS
BASE OF SLOPE FAN
1
2
3
4
5
SHELF-EDGE LOCATION
FOR SUCCESSIVE CLINOFORMS
N
50 km
Eastern edge of
the Tertiary
Central basin
Litledalsfjellet
Högsnyta
Western
Spitsbergen
Orogenic
Belt
Van Mijefjorden
red shale
stratigraphic marker
Eastward-prograding
clinothem complexes,
shelf-edge position
Sediment supply
and clinothem
migration path
Storvola
Van Keulen
80°N
Greenland
Spitsbergen
Norway
0
20°E

priate for a shallow-marine platform located around the margin of a deeper basin irrespective of the basin's tectonic setting. This is a morphological shelf. Such shelves, commonplace at the edges of marine basins, can vary in their width from a few kilometers to several hundred kilometers, and generally slope basinwards at less than 0.1° (Posamentier and Allen, 1999, their table 2). A characteristic feature of the outer edge of the shelf is a break or 'rollover' in the gradient, leading down onto a slope most commonly 2–7° steep. In our experience, this slope break occurs (and the use of the term shelf is warranted), if the maximum relief from shelf-platform edge to basin floor exceeds 150–200 m (Fig. 3A). During sea-level highstand, the shelf can support a water column of up to several hundred meters at its outer edge.

Such basin-margin platforms, facilitated by a preceding period of low sediment flux and/or high subsidence, are constructed by the long-term balance between sediment accommodation and sediment bypass, and the gradual extension or progradation of the front margin of this platform into the basin. A coastal sedimentary prism is thus constructed, and this commonly has its maximum seaward extent during falling stage and lowstand of relative sea level, when sediment supply is high and accommodation is decreasing (Posamentier and Allen, 1999). At this time, the shelf-margin delta drapes across the shelf edge, and the delta slope and the slope below the preexisting shelf break coincide, though the former is still somewhat steeper than the latter (Fig. 3B). Other than at this particular time, the smaller scale delta clinoforms should not be confused with the shelf-edge clinoforms. The platform thus constructed then becomes a shelf during subsequent rise of relative sea level, as the former subaerial platform drowns and the shore zone retreats landwards. This process repeats during successive cycles of relative sea-level fall and rise. When the deltaic or other shore zone system manages to deliver sediment as far out as the shelf edge, there occurs accretion of the shelf margin and the generation of deepwater clinoforms (Figs. 3 and 4). The clinoforms have a height approximately equal to the height of the deepwater slope, connecting the shelf edge with the basin floor, though if the water depth is great and the slope morphology complex, as it is on many passive margins, the clinoforms are muddy long before they reach the basin floor, or the clinoform toes themselves downlap long before the basin floor is reached. Because of the water depth and accommodation increase beyond the shelf edge, the progradational distance of any set of clinoforms beyond the older shelf edge is somewhat limited. In this way, shelf margins prograde mainly during sea-level stillstand and fall, whereas they aggrade mainly during relative sea-level rise. Shelf deltas, because of their powerful fluvial drive, play a prime role in this shelf aggradation and shelf-margin accretion (Fig. 4) (Morton and Suter, 1996).

5. General characteristics of shelf-margin deltas

5.1. Data base

Much of our knowledge about shelf-margin deltas derives from Quaternary continental shelves, and is primarily based on high-resolution seismic images (Table 2). This includes Costa de Nayarit, Mexico (Curray and Moore, 1964), Gulf of Mexico (Lehner, 1969; Winker, 1982; Suter and Berryhill, 1985; Morton and Price, 1987; Suter et al., 1987; Kindinger, 1988; Thomas and Anderson, 1991; Sydow et al., 1992; Sydow and Roberts, 1994; Winn et al., 1995, 1998; Morton and Suter, 1996; Hart et al., 1997; Kolla et al., 2000), off eastern United States coast (Matteucci and Hine, 1987), West Africa (McMaster et al., 1970; Pegler, 1999), and Mediterranean Sea (Aksu and Piper, 1983; Aksu et al., 1987; Farrán and Maldonado, 1990; Tesson et al., 1990, 2000; Trincardi and Field, 1991; Posamentier et al., 1992; Chiocci, 1994; Hernández-Molina et al., 2000). In contrast to this Pleistocene data, there is relatively little documentation of older shelf-margin deltas. Notable exceptions include, from passive margin settings, a Cretaceous shelf edge adjacent to the Baltimore Canyon (Poag et al., 1990), Tertiary of Gulf of Mexico (Edwards, 1981; Winker, 1982; Galloway, 1989, 1990; Mayall et al., 1992; Xue and Galloway, 1995), and Miocene clinoforms offshore New Jersey (Fulthorpe and Austin, 1998; Fulthorpe et al., 1999), from transpressional basins, Eocene of Spitsbergen (Steel et al., 2000; Plink-Björklund et al., 2001), from foreland basins, Miocene of the Carpathian Foredeep Basin in Poland (Porębski, 1999), and from intra-

Table 2
Summary characteristics of Pleistocene shelf-margin deltas

Location	Gulf of Mexico	Izmir Bay, East Mediterranean Sea	Rhône Shelf, West Mediterranean Sea	Guinea–Sierra Leone Shelf, West Africa
Delta regime	Fluvial dominated	Fluvial dominated	Wave dominated	Wave-dominated
Tectonic setting	Passive margin; basinwards downwarping due to sediment load	Active strike–slip-related extension	Passive margin; basinwards downwarping due to sediment load	Passive margin
Physiographic setting				
Shelf width	70–140 km	40–60 km	75 km	35–190 km
Shelf gradient	< 0.05°		1–3°, < 0.1° near the shelf edge	≤ 1°
Depth of shelf break	75–100 m	110–120 m	120–135 m	85–115 m
Slope gradient	0.6–1.5°, maximum 3°	Gentle	< 10°	4°
Delta geometry				
Shape	Multilobate to strike-elongate	Lobate to strike-elongate	Lobate	
Length*	25–60 km	< 60 km	40–50 km	30–40 km
Width	28–120 km	?		90 km and more
Maximum thickness	60–90 m, locally above 180 m	5–50 ms, controlled by syndepositional block faulting	50 m	30–95 to above 130 m
Clinoform				
Dip	3–6°, maximum 8°	4–6°	1.5°(?)	4–6°
Seismic expression	Oblique–tangential reflectors steepening basinwards, with toplap terminations below planar scour surface. Sigmoidal forms in the most seaward bundles. Offlap associated with downstepping ensued by backstep in youngest bundles. Common internal downlap surfaces. Progradation centres defined by bidirectional downlap	Oblique reflectors terminating updip by toplap and changing downwards into gently dipping shingled forms. Transparent facies in distal delta	Oblique–tangential reflectors showing downward decreasing amplitude, intercalated with hummocky reflection packages. Individual bundles separated by steeper erosional truncations of regional extent. Clinoform bounded at top by channeled surfaces with toplap terminations below. Overall downstepping to backstepping pattern in time	Offlapping stack of thinning-up clinoforms, each above major erosional unconformity with toplap. Terminations below and downlap above
Facies				
Upper foresets	Upward coarsening from distal to proximal delta front. Distal delta front: fauna poor, weakly burrowed thinly interbedded ripple and flat laminated fine sand and silt. Proximal delta front: 15–20 m of medium-grained sand with silt partings; weakly to non-burrowed, massive or flat laminated, rare ripple lamination, dispersed plant matter, very low fossil content	Medium-grained sand, numerous broken shells	Centimeter- to decimeter-thick beds of fine sand to silt alternating with silty clay laminae; microfauna of lower shoreface to mid-shelf aspects	

(continued on next page)

Table 2 (*continued*)

Location	Gulf of Mexico	Izmir Bay, East Mediterranean Sea	Rhône Shelf, West Mediterranean Sea	Guinea–Sierra Leone Shelf, West Africa
Facies				
Lower foreset	Mud, silt and clay, locally highly bioturbated, moderately fossiliferous; thin beds of laminated fine sand; pyritised microfaunal burrows, carbonate concretions; layered plant matter concentration	Muds an silt with sparse benthonic fauna, rare planktonic forams	Mud, silt and clay, locally highly bioturbated, moderately fossiliferous; thin beds of laminated fine sand; pyritised microfaunal burrows, carbonate concretions; layered plant matter concentration	
Soft-sediment deformation	Growth faults and associated slope basins; common contorted and sheared intervals. Extensive wedge-shaped, mounded to chaotic seismic zones interpreted as slumps and slides; slide scars filled with oblique progradational facies. Shale and salt diapirs at lower slope	Imbricated slump blocks in lower prodelta slope	Retrogressive slumping and faulting at shelf edge	
Delta top	Seaward shallowing channels, 1–5 km wide and 30–40 m deep, merging landwards into incised valleys up to 30 km wide and 60 m in relief. Channel fills of medium to coarse, locally pebbly sand; trough cross-bedding, minor parallel lamination		Channels, several hundreds of meters wide and 10–15 m deep, filled with hummocky to subparallel seismic facies, interpreted as passive infill of slump scars. U-shaped troughs, >50 m deep, interpreted as canyons grading updip into incised valleys	Incised valley systems, 15 m deep in outer shelf and up to 90 m inshore, passing seawards into cemented surface of marine erosion
Incision of delta edge	No major fluvial incisions except off Mississippi River Delta; common troughs initiated by retrogressive slope failure		Slump-related canyon heads and gullies	No major incisions
Sources	Lehner, 1969; Sangre et al., 1978; Winker, 1982; Winker and Edwards, 1983; Suter and Berryhill, 1985; Suter et al., 1987; Sydow et al., 1992; Sydow and Roberts, 1994; Morton and Suter, 1996; Hart et al., 1997; Winn et al., 1995, 1998	Aksu and Piper, 1983; Aksu et al., 1987	Tesson et al., 1990; Posamentier et al., 1992; Tesson et al., 1993, 2000	McMaster et al., 1970; Pegler, 1999

* Distance from shelf break to landward pinchout onlap.

cratonic basins, Namurian of County Clare, Ireland (Pulham, 1989; Collinson et al., 1991).

The bias towards overrepresentation of Pleistocene shelf-margin deltas in the data set reflects the fact that (1) multiple glacieustatic oscillations left a remarkably clear and fairly accessible record of long-distance shoreline shifts, across the present-day continental shelves, and that (2) high-magnitude (100–130 m), high-frequency (10's to few 100's of thousands of years) and strongly asymmetric sea-level fluctuations were particularly instrumental in producing this type of delta. However, the remarkable scarcity of shelf-margin deltas reported otherwise justifies the development of criteria that can be used to recognize shelf-margin deltas in outcrop.

It was suggested by Steel et al. (2000) that the key to recognizing older shelf-margin deltas is in the correct identification of the ancient shelf edge, i.e., the recognition of shelf-slope clinoforms that require exceptionally large outcrops (more than 10-km outcrop length is required to see shelf edge connecting with basin floor, if slope gradient is 3° and relief from shelf edge to basin floor is >500 m). In the Eocene examples of Steel et al. (2000) and Plink-Björklund et al. (2001), shelf-edge deltas are common features of the shelf margin, especially during time intervals when the shelf edge and upper slope areas were not canyonised. It is likely that many other ancient deltaic units, and particularly those showing tens to hundreds of kilometers of regressive transit across shallow shelves (e.g., some of those at the youngest levels in the Cretaceous Western Interior Basin, USA—Asquith, 1970), are also of shelf-edge type. They have simply not been referred to in these terms possibly because of an assumed ramp-type setting, or because their forced regressive or lowstand character has not been documented (a notable exception being the Panther Tongue Member of the Star Point Formation—Posamentier and Morris, 2000). Some of the Western Interior Basin's shelf-ridge sand bodies have now been reinterpreted as examples of lowstand shelf deltas, though not developed across a shelf-slope break (Mellere and Steel, 1995; Bhattacharya and Willis, 2001).

5.2. Overall geometry

Isopach maps show shelf-margin deltas as multilobate, arcuate to lunate bodies (Suter and Berryhill, 1985). However, a striking feature is that the bodies show a considerable elongation along the shelf edge, and can be 30–90 km wide in this direction (Figs. 5

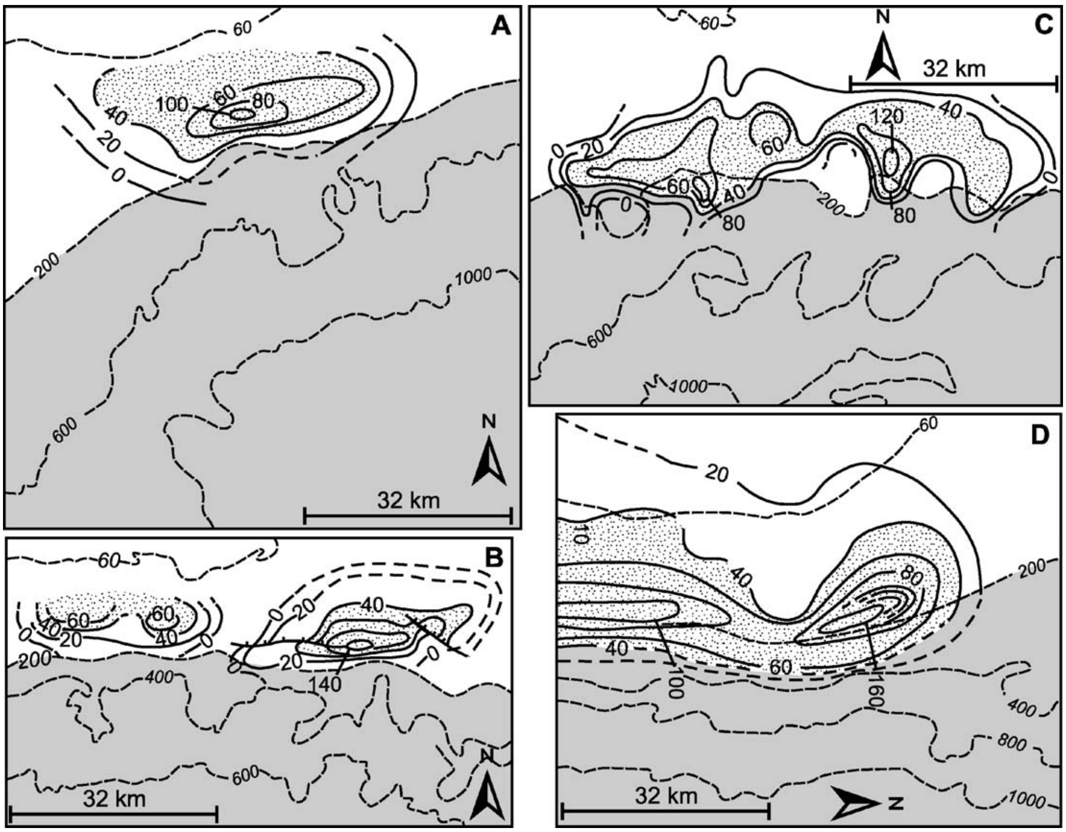


Fig. 5. Geometry and dimensions of Wisconsinan shelf-margin deltas in the northwest Mexican Gulf off the Texas coast (A and B), and the ancestral Mississippi (C) and Rio Grande (D) rivers (modified after Suter and Berryhill, 1985). Deltaic depocentres are dotted; shaded areas correspond to continental slope.

and 6). This strike-elongation is associated with the abrupt increase of accommodation outboard of the shelf edge (stalling the regression, as it were), reflecting an increased influence of tectonic and sediment load-induced subsidence rather than any particular hydraulic regime of the outer shelf. The highest subsidence rate near the shelf edge is promoted by the gravity spreading of the associated slope whose instability increases with an increased sediment supply (Winker, 1982). In dip cross-sections, a shelf-margin delta complex appears as a clinoformed wedge that thickens towards the shelf edge, attaining a maximum thickness of 50–200 m at or just below the shelf edge, and then thins gradually down onto the upper to middle slope (Figs. 7 and 8). In areas affected by growth faulting, the thickest part of the wedge tends to be located slightly landwards from the physiographic shelf break (Edwards, 1981; Morton and Suter, 1996). Farther landwards, the wedge commonly pinches out by onlap, and the pinchout zone lies some 25–60 km behind the shelf break (Aksu and Piper, 1983; Lehner, 1969; McMaster et al., 1970; Suter and Berryhill, 1985; Posamentier and Allen, 1999).

5.3. Bounding surfaces

The base of outer-shelf and shelf-edge deltaic clinoform complexes downlaps onto the preexisting shelf surface; however, consecutive reflections tend to show asymptotic lower ends rather than an abrupt downlap onto this surface (Morton and Suter, 1996; Posamentier

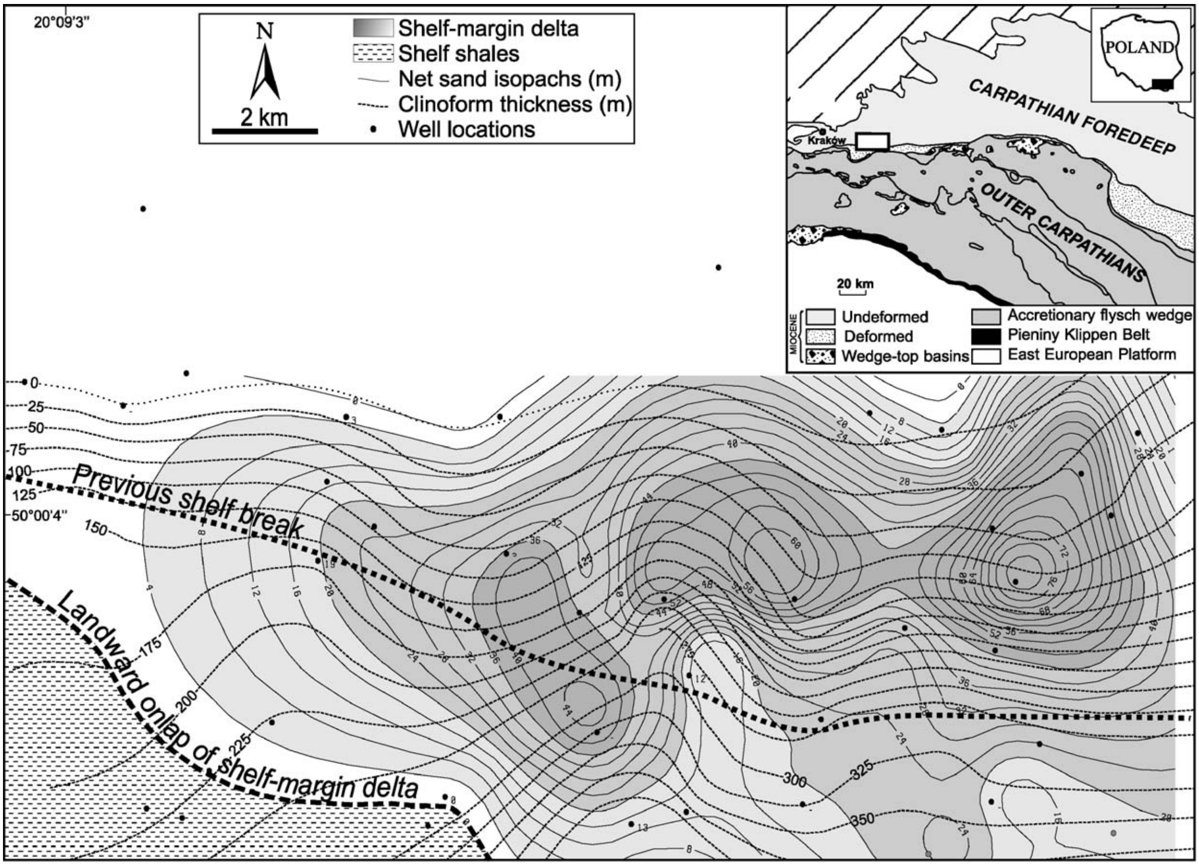


Fig. 6. Clinoform thickness and net-sand isopach map showing the subsurface geometry of Badenian shelf-margin delta that is aligned along and extends downdip (northwards) beyond the former shelf break, Carpathian foredeep, southeast Poland (Porębski et al., in press). Inset map shows the location of studied succession within the geological framework of the Carpathians.

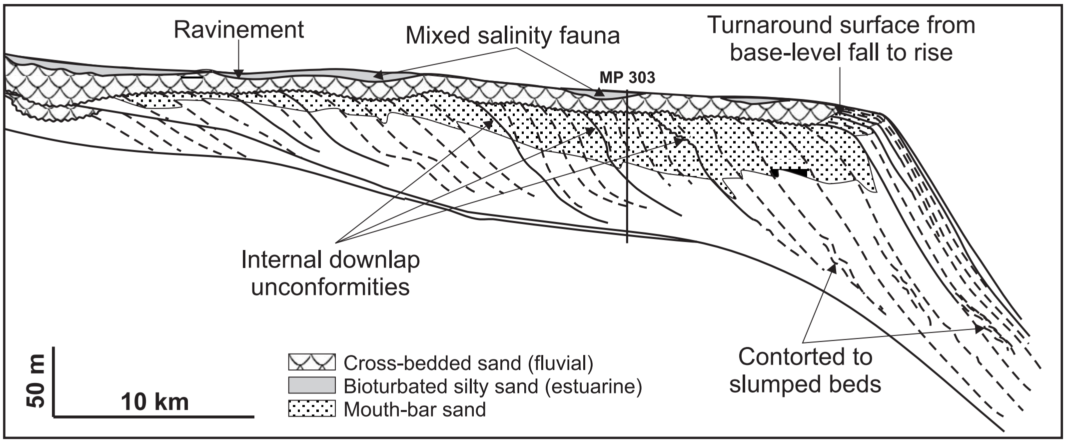


Fig. 7. Dip cross-section through Late Pleistocene shelf-margin delta formed during falling stage to early sea-level rise. Note the seaward steepening and thickening of outer-shelf delta clinoform complex and the thickening of mouth-bar sands. Part of the fluvial sands is time equivalent to the youngest, backstepping clinoform bundle. Internal downlap unconformities, where of regional persistence, are believed to record the stepwise nature of sea-level fall (modified from Sydow and Roberts, 1994). See Fig. 12A, for core log from the MP303 boring.

et al., 1992; Tesson et al., 2000). At the landward reaches of the shelf-edge delta complex, this basal surface invariably truncates the underlying succession, emphasized by a toplap truncation of older shelf, shelf-edge or slope reflections (Sydow and Roberts, 1994) (Fig. 9). Basinwards of the shelf margin the same

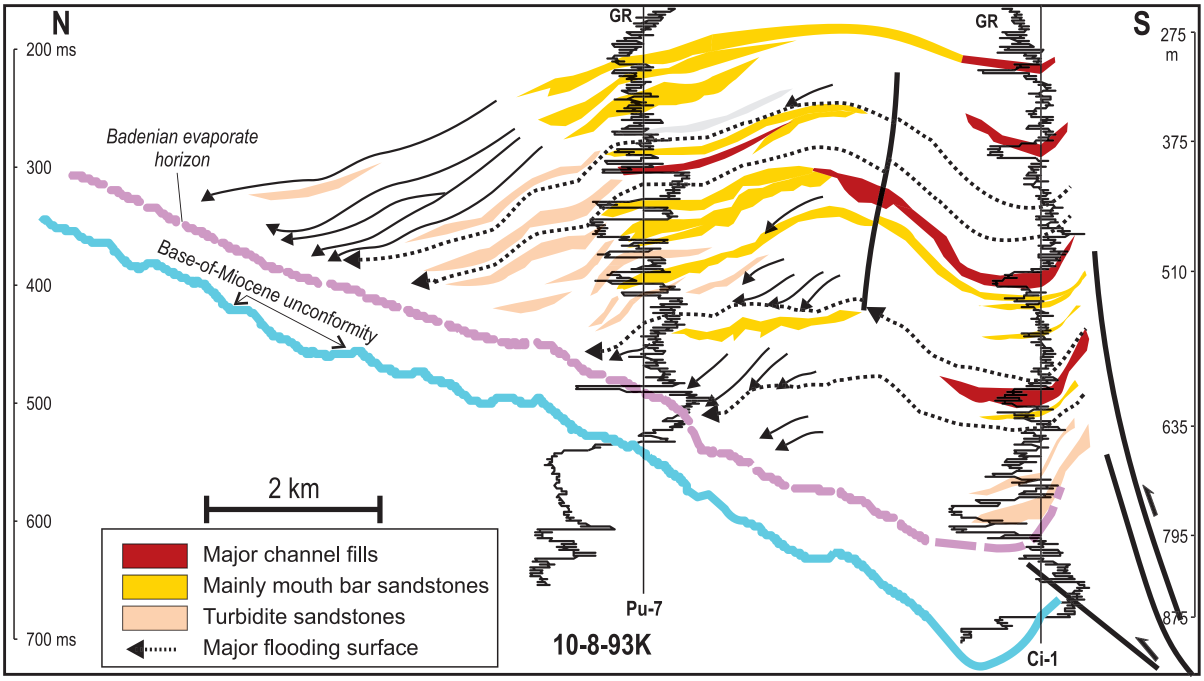


Fig. 8. Interpreted line drawing from seismic dip-section through Badenian shelf margin in the Carpathian foredeep (southeast Poland), showing the distribution of sand bodies (Porębski et al., in press). Note that the sand bodies thin and pinch out landwards (to the south), and form the thickest accumulation near the shelf edge which is constructed mainly of mouth-bar and slope turbidite sandstones. See Fig. 19B, for location of section.

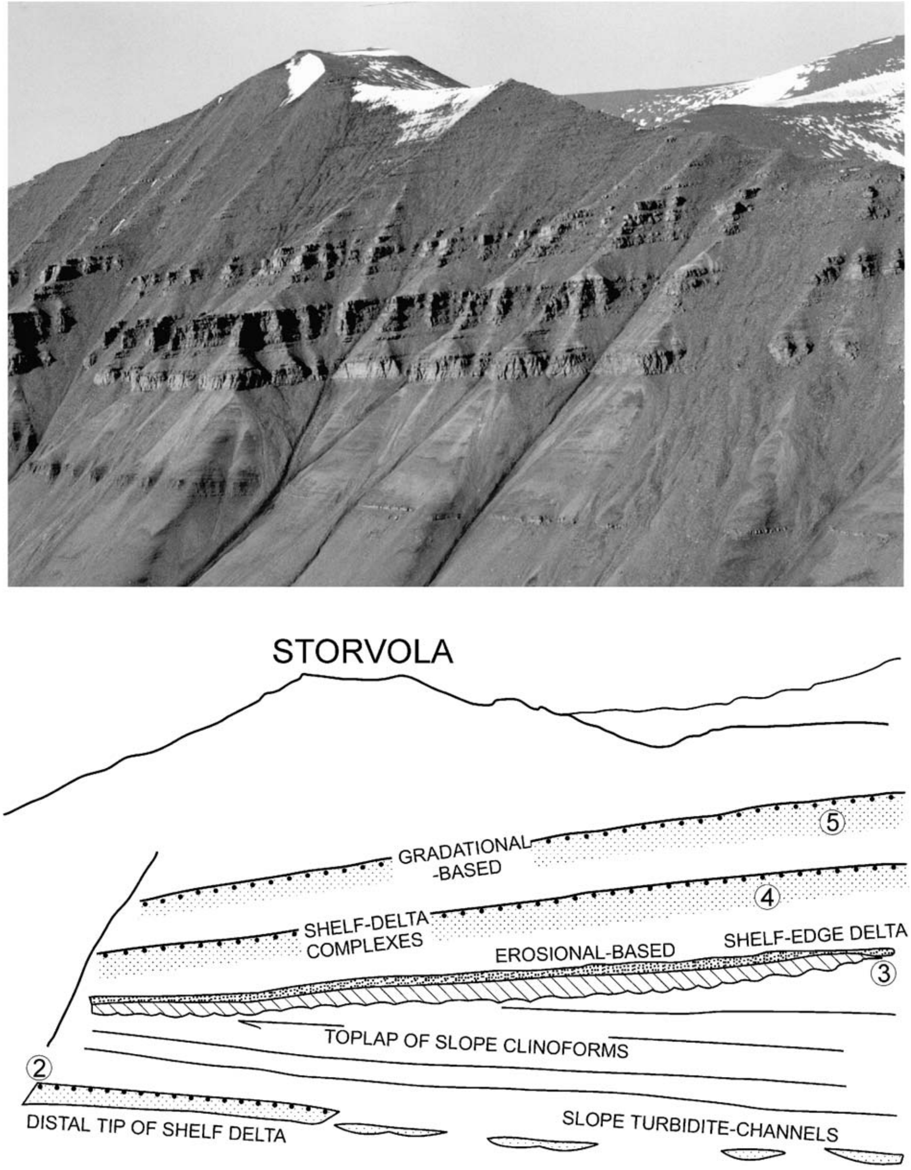


Fig. 9. A series of four clinoform complexes (2–5) on Storvola, Spitsbergen (see also Fig. 4). Clinoform complex 3 (see Fig. 4) represents a shelf-margin delta that has a sharp and incised base. It truncates toplapping, older slope clinoforms (shaly), because of its development during falling relative sea level. Note that clinoform complexes 4 and 5 are gradationally based, showing an upward coarsening. They occupy an inner to mid-shelf (highstand [?]) position. At the level of clinoform 2, the distal tip of a late lowstand, shelf-margin delta can be seen to downlap onto lower slope turbidite channels.

downlap surface persists, but the downlapping seismic reflections eventually flatten and approach the gradient of the slope, and the surface eventually becomes conformable into the basin. The upper surface of the shelf-margin delta complex tends to be a high-amplitude reflection. It displays a planar to incised geometry along the shelf reaches of the complex (Fig. 10), truncating the internal clinoforms of the delta increas-

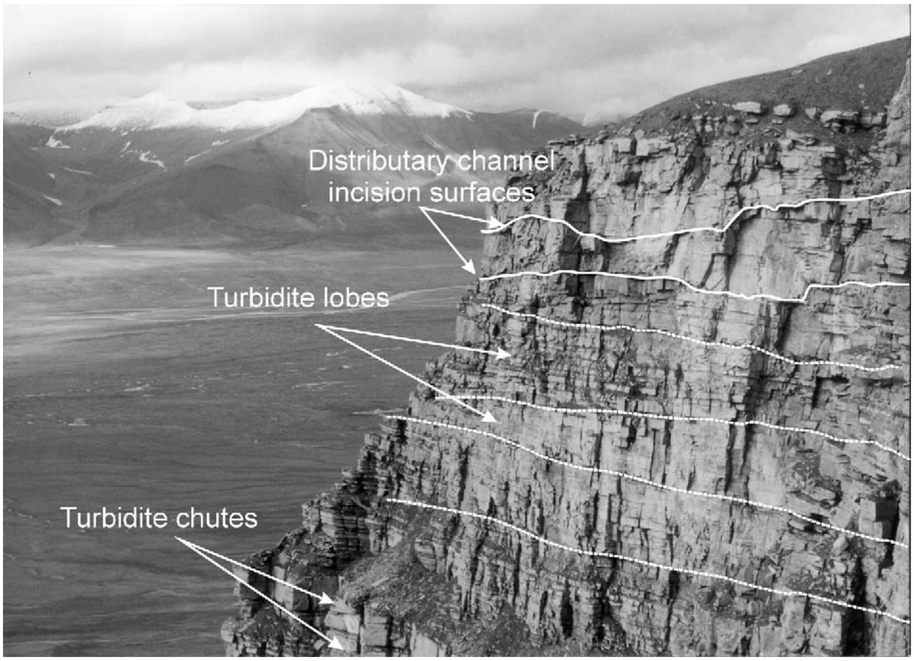


Fig. 10. Distributary channel truncation down into top of 60-m-thick shelf-edge delta succession. Note the slight discordances between successive delta-front lobes. Except for the distributary channel fill, all sand beds are turbidites; Litledalsfjellet, Spitsbergen.

ingly landwards, and eventually amalgamating with the basal downlap surface described above. Along the shelf-edge to upper slope reaches of the system, this upper surface becomes conformable with the slope-elongated clinoforms (Morton and Suter, 1996). Stratigraphically, this surface bears signs of the erosion associated with the basinward propagation of a fluvial distributary channel feeder, and its subaqueous-equivalent erosion surface, possibly induced by hyperpycnal flows (Plink-Björklund et al., 2001), although this evidence can be removed by a subsequent transgressive ravinement. This ravinement surface is likely to amalgamate or erode through the regressive erosion surface at the top of the delta in the region of the shelf edge. Farther landwards, it is likely to diverge from the regression surface, separated from it by estuarine or coastal deposits, if sediment supply is still high during transgression.

5.4. Clinoform bundling and slopes

Clinoform geometry tends to change basinwards as the deltas prograde from the outer shelf, across the shelf edge and onto the upper-slope reaches of the system (Fig. 7). While the deltas are prograding on the shelf, the visible, small-scale clinoforms (low-angle oblique forms) correspond with the accreting delta front, and the slope angles are consistent with this. On reaching the shelf edge, however, the front of the delta then progrades onto a preexisting slope that can be as steep as 3–5°. The visible clinoforms at and just below the shelf edge thus become steep (4–7°) reflections with long tangential ends and eventually sigmoidal forms with strongly asymptotic toes on the slope (Kolla et al., 2000; see also examples in Posamentier and Allen, 1999). The seaward steepening of the clinoform slope tends to be associated with the basinward segregation of sand (Sydow and Roberts, 1994) that attains maximum net values at the newly formed shelf break (Fig. 7). This pattern of steepening of deltaic clinoform slopes at the shelf margin, accompanied by a sand body thickness change at this point (from 20–30 up to 70 m), has also been described from Eocene (Plink-Björklund et al., 2001) and Miocene (Porębski et al., in press) shelf edges (Fig. 8).

The progradational trajectory of the entire shelf-margin delta complex is commonly interrupted by internal downlap surfaces (Fig. 7) that partition the complex into individual, stacked and laterally offset clinoform bundles (Sydow and Roberts, 1994; Tesson et al., 2000). Such breaks presumably reflect either slight changes in progradational direction or in the process regime affecting the deltas. In their distal reaches, such internal discontinuities can be associated with a high-amplitude reflection that can be traced in cores to nodular or layered carbonate concentrates (Sydow and Roberts, 1994). Strike sections through a single clinoform bundle may show a bidirectional downlap indicative of the core of a prograding delta lobe (Sydow et al., 1992).

Clinoform offlap breaks that record successive positions of the prograding shelf margin, tend to be preserved in the most basinward bundles. Near the termination of the shelf-margin delta trajectory, the offlap break tends to shift downwards and basinwards because of continued relative sea-level fall (Fig. 11). The offlap break tends to acquire an upward aggrading and eventually backstepping trajectory within the youngest bundles, as relative sea-level rise produces a prograding lowstand wedge and eventually a transgressive systems tract (Sydow and Roberts, 1994; Kolla et al., 2000; Plink-Björklund et al., 2001). Increasingly, transgressive systems tract deposits overlying lowstand deltas are being correctly identified in terms of tidally dominated, shelf-margin estuaries (Rabineau et al., 1998; Berné et al., 1998; Schellpeper, 2000) and/or shelf ridges (Posamentier, 2002).

5.5. Facies spectrum in shelf-margin deltas

In contrast to the well-documented geometry of shelf-margin deltas from seismic data, there is a marked deficiency of detailed facies information available for these systems. The situation has been improved recently by lithofacies, biostratigraphic and paleobathymetric data from a series deep borings that penetrated the Late Pleistocene Lagniappe Delta (Kindinger, 1988), east of the present Mississippi Delta (Sydow and Roberts, 1994; Winn et al., 1995, 1998; Scott et al., 1998; Kolla et al., 2000), and from spectacularly exposed shelf-margin deltas in the Early Eocene foreland basin of Spitsbergen (Steel et al., 2000; Plink-Björklund et al., 2001), as well as in the Namurian of County Clare, Ireland (Pulham, 1989; Collinson et al., 1991). The data from the Lagniappe Delta document mud-prone to thin-bedded heterolithic,

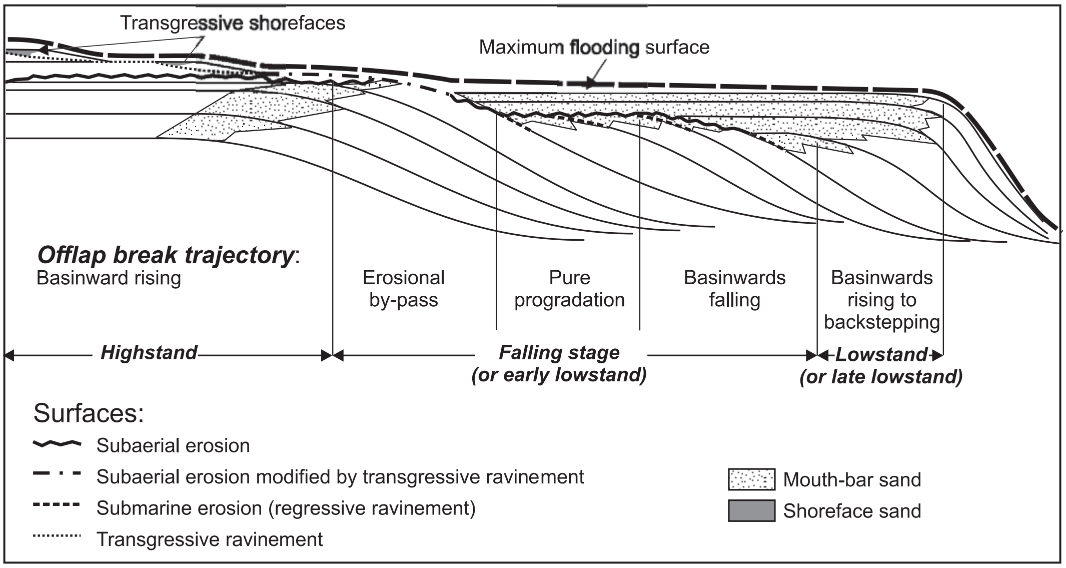


Fig. 11. Concept of offlap break trajectory (partly based on Helland-Hansen and Martinsen, 1996) and architectures of shelf-margin deltas.

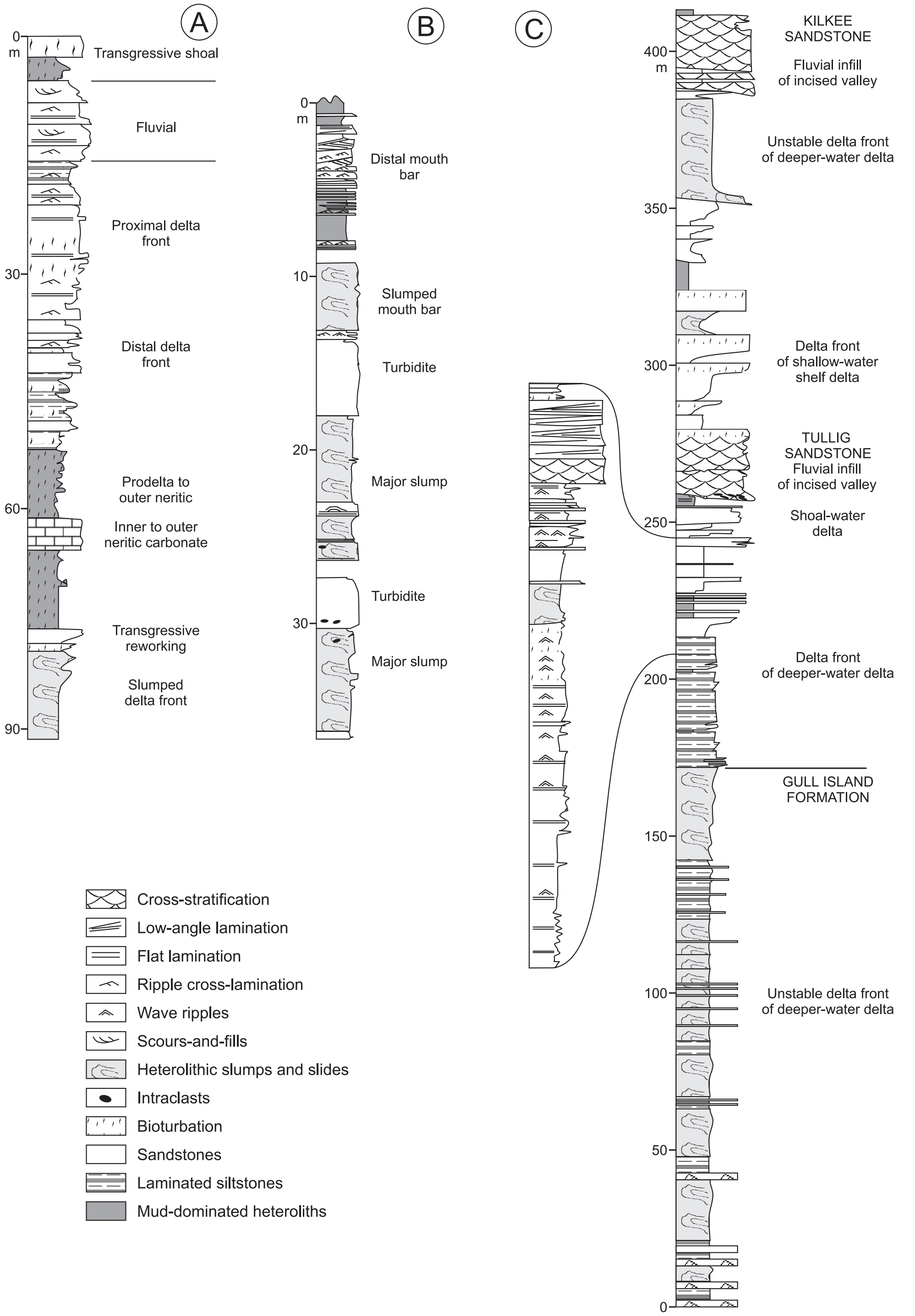
locally slumped character of the shelf-edge delta front. Mayall et al. (1992) documented alternations of heterolithic slumps with clean, thick-bedded sands in the Pliocene of the Mississippi Canyon area. On the other hand, evidence from Eocene and Miocene shelf-margin deltas in foreland basins of Spitsbergen and Poland show that similar mud-prone delta-front sets can alternate with those dominated by thick-bedded turbidite sandstones (Steel et al., 2000; Plink-Björklund et al., 2001; Porębski, 1999; Porębski et al., in press). Xue and Galloway (1995) also mapped several thick sand-rich pods beyond the shelf margin in the Middle Paleocene Wilcox subgroup, offshore Texas, and ascribed their origin to slump lobes or turbidite resedimentation of shelf-edge mouth-bar facies. All the above examples share a characteristic feature, i.e., slope deformation and collapse, whereas differences in the sand percentage on the delta-front deposits reflect mainly variations in the calibre and sediment volume delivered to the slope. Although the tectonic setting of the host basin strongly influences sediment supply and large-scale stacking patterns, it probably has relatively little influence on the internal character of individual shelf-edge deltas. Deltas form and generally are able to regressively transit the entire shelf width in less than 100,000 years (Burgess and Hovius, 1998; Muto and Steel, in press)—a time scale within which they will be only moderately affected by tectonics.

If the above examples from Gulf of Mexico, Spitsbergen, Carpathians and Ireland seem few, on which to base our knowledge on shelf-edge delta facies, it is because it is largely these literature examples that have been examined and documented in a shelf-edge position.

5.6. Facies landwards of the shelf edge

Along the shelf traverse, the upper parts of the deltaic succession (steepest part of delta clinoforms) are dominated by sandy mouth bars (Mayall et al., 1992; Hart et al., 1997; Plink-Björklund et al., 2001) or proximal delta-front deposits (Fig. 12A) (Sydow and Roberts, 1994; Winn et al., 1995, 1998). Mouth-bar sands are fine to medium grained, often clean, well sorted and bedded on a decimeter to meter scale. They form up to 20-m-thick units that show a "blocky" to slightly serrated, coarsening-upward gamma-ray well-log response. Beds are non- to weakly bioturbated and poorly to non-fossiliferous, although isolated mollusc shells, shell debris and dispersed plant matter can locally be present. Parallel laminated sandstone beds intercalate with current-rippled and "structureless" sandstone beds, and with siltstones. The sandy mouth-bar unit grades basinwards and downwards into a more heterolithic succession attributable to delta-front deposition (Fig. 12A). It comprises non- to weakly burrowed, silty sand and sandy silt, locally rich in plant matter, which form a host for thin beds/bedsets of well-sorted, plane-parallel laminated and graded-laminated fine- and very fine-grained sandstone. The fossil content is low, though can be higher than in the mouth-bar/proximal delta-front sand (Sydow and Roberts, 1994). Long tangential to parallel toeset units correspond to distal delta front and prodeltaic sediments which are chiefly silty clay and clay in grade. These sediments are moderately fossiliferous and locally strongly bioturbated. Fossils are of neritic to bathyal affinity. Thin beds of parallel laminated sand and thick, unburrowed, structureless mud beds are commonly interpreted as delta-front turbidites (Morton and Suter, 1996; Winn et al., 1998).

The vertical facies succession described above apparently does not differ substantially from that found in an inner-shelf, river-dominated delta. However, the overall upward-coarsening trend in shelf-margin deltas is associated with paleoecological evidence of an extremely rapid shallowing. The close proximity of shallow- and deep-marine biotic indicators is common in shelf-margin deltaic successions (Winker, 1982). One of best example of this is provided by the VK774 well drilled through the Pleistocene Lagniappe Delta (eastern Gulf of Mexico), where 20–30 m of sediments records a transition from upper bathyal to middle neritic depths, i.e., 100–200 m of decreasing water depth (Winn et al., 1998; see also Kolla et al., 2000). This phenomenon, referred to as foreshortened stratigraphy (Posamentier and Morris, 2000, p. 36), together with the general absence of delta-plain facies appear to be one of the main distinctive features of shelf-margin deltas. The absence of delta-plain facies and the foreshortening of the vertical succession result from relative sea-level fall and decreasing accommodation during delta progradation on the outer shelf.


A
Transgressive shoal
Fluvial
Proximal delta front
Distal delta front
Prodelta to outer neritic
Inner to outer neritic carbonate
Transgressive reworking
Slumped delta front
B
Distal mouth bar
Slumped mouth bar
Turbidite
Major slump
Turbidite
Major slump
C
KILKEE SANDSTONE
Fluvial infill of incised valley
Unstable delta front of deeper-water delta
Delta front of shallow-water shelf delta
TULLIG SANDSTONE
Fluvial infill of incised valley
Shoal-water delta
Delta front of deeper-water delta
GULL ISLAND FORMATION
Unstable delta front of deeper-water delta
Cross-stratification
Low-angle lamination
Flat lamination
Ripple cross-lamination
Wave ripples
Scours-and-fills
Heterolithic slumps and slides
Intraclasts
Bioturbation
Sandstones
Laminated siltstones
Mud-dominated heteroliths

5.7. Facies basinwards of the shelf edge

As a delta reaches the shelf edge and begins to prograde onto the upper slope, there is a significant change in facies development, due to the increased gradient of the substratum that promotes mass-gravity deformation and flow (Mayall et al., 1992; Winn et al., 1998). Over half of heterolithic delta-front units in the inferred shelf-margin deltas in the Namurian of County Clare, Ireland, reveal evidence of growth faulting, slumping and sliding (Pulham, 1989), and such phenomena are even more pronounced in the underlying turbiditic slope deposits of the Gull Island Formation (Fig. 12C) (Martinsen and Bakken, 1990; Collinson et al., 1991).

In the Spitsbergen examples, which provides a considerable insight into the 3D facies geometry of a fossil deltaic shelf margin (from outer shelf to basin plain—Figs. 4 and 9), the sand-prone mouth bars consist of both thick, massive sandstones, scours filled with single, cross-stratified sets and flat to low-angle laminated sandstones. In addition, however, units of soft-sediment deformed sandy mudstones/siltstones now abound (Fig. 13). If falling sea level causes the mouth-bar system to perch at successively lower levels on the upper slope, a series of erosional terraces can develop that cut back onto the shelf edge (Plink-Björklund and Steel, 2002). The delta-front facies basinwards of the shelf edge tends to be heterolithic as above, but clean, plane-parallel to ripple-laminated, or massive sand beds now spectacularly alternate vertically with slumped, sandy mudstone units, up to several meters thick (Fig. 14). Mayall et al. (1992) provided detailed core descriptions of such heterolithic slumps, 6–15 m thick, each capped by 3–5 m of clean, structureless to parallel-laminated sand from the upper slope of a Pliocene shelf-margin delta in the Mississippi Canyon Block 109 (Fig. 12B). The occurrence of such repeated, heterolithic slump/clean sandstone couplets appears as one of the clearest criteria for identifying an outcrop location as being below the shelf edge (the steeper slope angle may not be obvious close up on the outcrop).

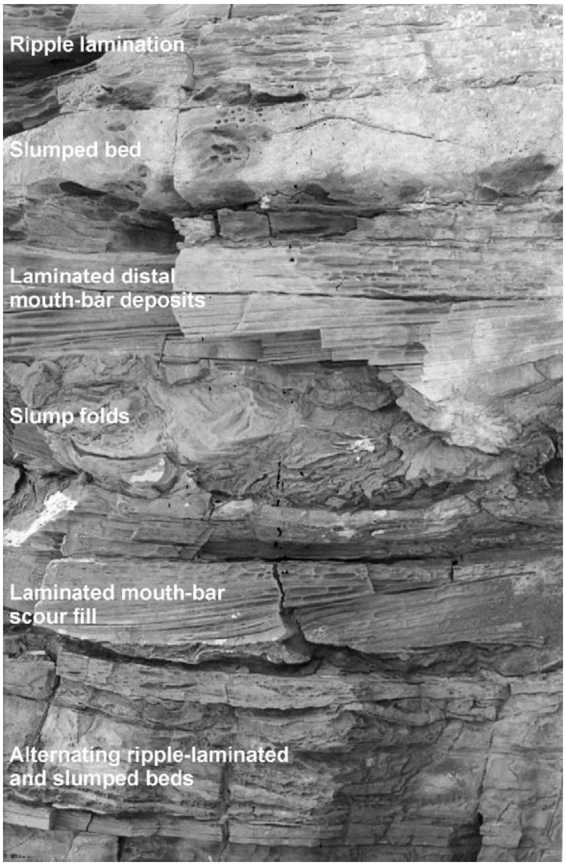


Fig. 13. Alternations of slumped beds with scour-based, single sets of cross strata, occurring in the distal mouth-bar succession just basinwards of a shelf edge. Slumped bed in centre is 25 cm thick; Storvola, Spitsbergen.

The clean, sharp-based sandstone beds described above that are structureless or flat laminated and ripple capped are more clearly identified as turbidites (Fig. 15) than the analogous delta-front beds on inner-shelf deltas. In the Spitsbergen examples, such beds can be seen to stack vertically to form upward-thickening

Fig. 12. Examples of vertical facies successions in shelf-margin deltas. (A) Core log from borehole MP303 through the Pleistocene Lagniappe Delta, Gulf of Mexico (modified from Sydow and Roberts, 1994, and Kolla et al., 2000, with interpretation after Sydow and Roberts, 1994). The well is located landwards of the former shelf break (comp. Fig. 7). (B) Core log from well MC109-3(OH), Middle Pliocene, Mississippi Canyon. Block 109, showing repetitions of slump/sand-turbidite couplets (modified from Mayall et al., 1992). (C) Composite section for the lower Central Clare Group, Ireland (modified from Elliot and Davies, 1996, and based partly on Pulham, 1989; Martinsen, 1989; Collinson et al., 1991).

Fig. 14. Mouth-bar succession (20 m thick), showing multiple slumped beds. The slump in centre of succession is 3 m thick, and originated from a slump scar that can be seen halfway along the cliff. The slumped beds are mud-rich sandstones, whereas the intervening distal mouth-bar beds are clean, laminated sandstones; Storvola, Spitsbergen.

Fig. 15. Thin-bedded, ungraded sandy turbidites typical of the slope association when shelf-edge deltas drape across the shelf margin; Storvola, Spitsbergen.

slope lobes, up to 20 m thick, and are frequently cut by small turbidite-filled channels or chutes that extend down the slope for several kilometers (Mellere et al., in press). Such chute- or channel-topped lobes are clearly progradational on the slope, but tend to systematically thin downslope and interfinger gradually with thicker units of rippled siltstones and mudstones. The existence of similar slope channels, or chutes, on delta-slope lobes have been inferred from well-log responses in the sand-prone delta-front deposits in the middle Wilcox subgroup shelf-edge deltas in Texas (Xue and Galloway, 1995).

On the Spitsbergen Eocene deltaic shelf-margin successions, sandy turbidites are extremely abundant (Fig. 16) , but it is not always clear if the turbidites derive from slumps or from sediment-laden currents directly output from the delta (hyperpycnal flows). However, the decelerating/waning nature of these delta-front flows, inferred from their thinning and termination on the slope, suggest relatively low-density, non-ignitive flows. Such sandy turbidites make up the bulk of sand-prone clinothems in which they are connected updip to mouth-bar sandstones.

Thick, massive to crudely laminated sand beds dissected by upflow-dipping shear planes and showing convoluted to locally wave-rippled tops are common constituents in the delta-front/slope deposits off Badenian shelf-margin deltas in the Carpathian Foredeep of Poland (Fig. 17). On well logs, such sands are commonly identifiable as "blocky" units 5–30 m thick that, together with intervening heterolithic intervals, form sand-prone accumulations up to 120 m thick (Fig. 18). The latter are parallel to the shelf break and extend for 14–30 km along strike and 2–4 km down on the slope (2–4°) (Fig. 19). On seismic sections, the landward ends of thicker sand bodies appear either to pinch out within slope heteroliths in a shingled fashion, or onlap onto an unconformity that is traceable further up either into thin mouth-bar deposits or directly below transgressive shelf shales. Where there is poor seismic control and/or only discontinuous core data, the repeated occurrence of "blocky" sands pinching out updip within shales of outer neritic to bathyal affinities strongly implies proximity to a fossil shelf edge.

5.8. Soft-sediment deformation

Common features of clinoform seismic images, particularly at or below the shelf margin, are discon-

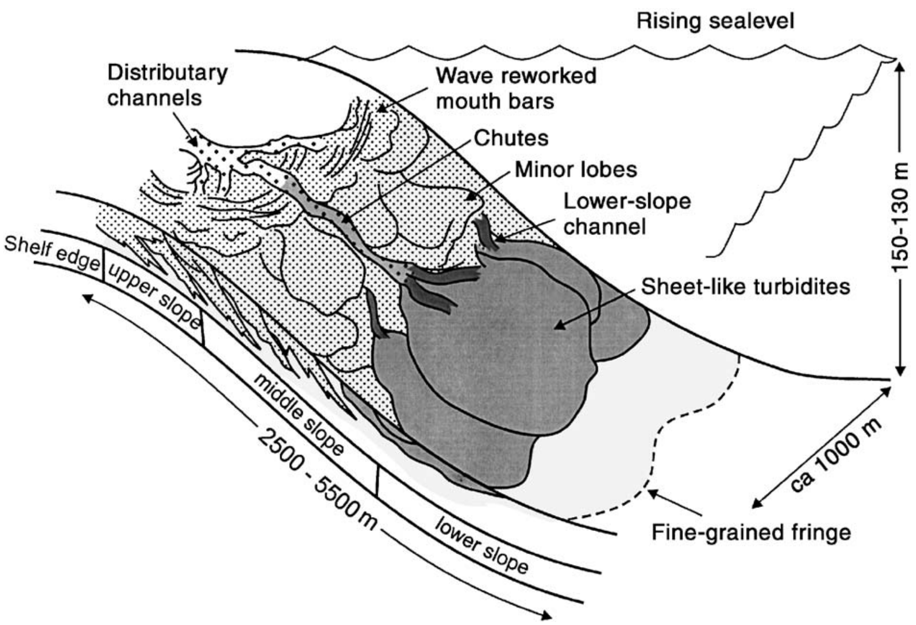


Fig. 16. Schematic drawing of the turbiditic slope association commonly developed when shelf deltas reach the shelf margin (modified from Plink-Björklund et al., 2001).

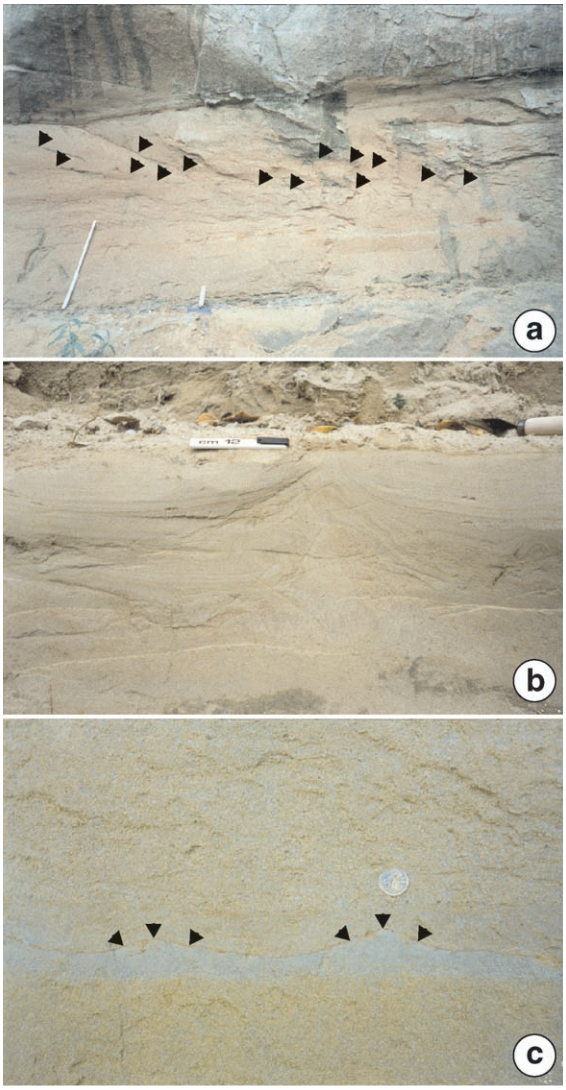


Fig. 17. Features of thick-bedded, massive sandstones deposited within a perched mouth bar. Zabawa, Carpathian foredeep (modified from Porębski and Oszczypko, 1999). (a) Listric shear planes (arrowed) verging downflow (to the left) and merging downwards into the intrabed, flat decollement surface. (b) Water-escape convolutions near top of massive sand bed. (c) Symmetrical ripple forms (arrowed) in a centimeter-thick siltstone parting at the contact of thick, massive sandstone beds.

tinuous, discordant to wedge-shaped units of chaotic, mounded and transparent seismic facies, interpreted as the product of slope-gravity collapse (e.g., Lehner, 1969; Sangree et al., 1978). The combination of mud-rich upper slope and high sediment influx makes shelf-margin deltas particularly susceptible to gravity-driven soft-sediment deformation on a variety of scales. Extensional strain dominates the shelf edge and uppermost slope, and this commonly results in regional down-to-basin, listric growth faults that are subparallel to and centred at the shelf break, whereas compression lower on the slope generates imbricated folds and thrusts, piled up slides and diapiric intrusions (Winker, 1982; Galloway, 1989). Sediment-load subsidence, when acting together with growth faulting, expands delta thickness significantly and results in the predominant strike alignment of shelf-edge depocentres irrespective of the delta-front regime (e.g., Edwards, 1980; Xue and Galloway, 1995). Evidence of large-scale slumping and sliding is widespread, and includes both contorted and imbricated sediment packets seen in cores, as well as chaotic to mounded seismic facies best interpreted as failure deposits. Such deposits are particularly common in base-of-slope areas, but they can also be present high on the slope (Mayall et al., 1992; Martinsen and Bakken, 1990; Collinson et al., 1991).

Magnificent examples of shelf-edge rotational slides and related scars filled with sandy debrites and high-density turbidites were reported from an Early Cretaceous shelf-margin in eastern Spitsbergen (Nemec et al., 1988). These same Cretaceous shelf-edge collapse features are later healed over by a 60–80-m-thick series of deltas that reestablished at the shelf edge when sea level reattained that position (Fig. 20). It seems likely that a mixture of rotational slides and cohesive slumps overlain by remobilised sand that filled the post-collapse hummocky relief can produce chaotic seismic facies, whereas more transparent zones intercalated with parallel reflections can be interpreted in terms of thicker, sandier intervals (cf., Morton and Price, 1987).

As exemplified by parts of the Tertiary succession of the northern Gulf of Mexico (Edwards, 1980, 1981; Winker, 1982; Galloway, 1989; Morton, 1993) and of the Namurian of County Clare (Collinson et al., 1991), large-scale slope-gravity spreading and gliding are almost certainly triggered and facilitated by a large sediment supply (and accompanying sediment-load subsidence) to the shelf margin by deltas. This deformation, particularly when enhanced by salt withdrawal, can result in the delta-fed slopes having a complex topography, with semicircular to strike-elon-

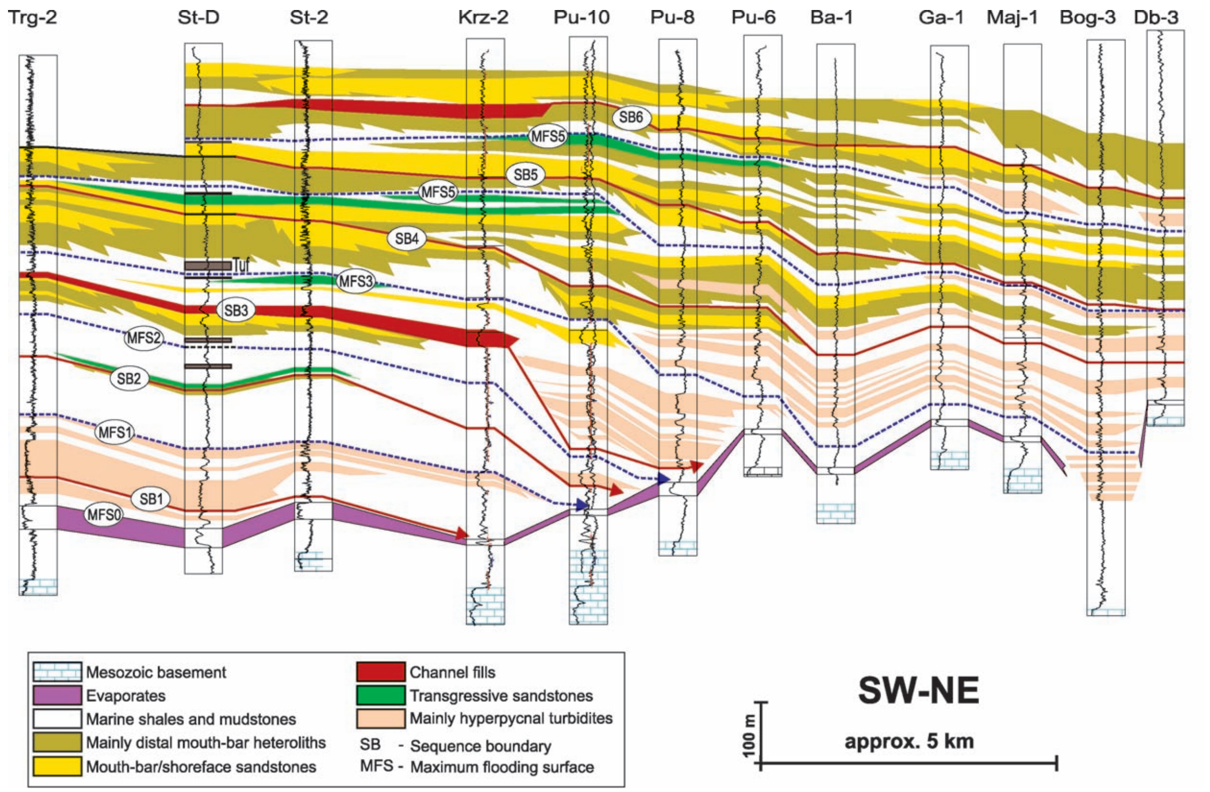


Fig. 18. Dip correlation through Badenian deltaic shelf margin in the Carpathian foredeep (modified from Porębski, 1999). See Fig. 19B, for location of section.

gated elliptical slope basins that may or not be connected by dip-trending gullies and channels in a ‘‘fill-and-spill’’ style (Prather et al., 1998). These perched basins have onlapping fills containing sand-rich turbidites, whereas the linking channels may show a local levee development (Badalini et al., 1999; Prather et al., 1998). Obviously, collapsing of a shelf-margin delta into a series of rotated and tilted blocks, as demonstrated for example in the Ubit Field, offshore Nigeria (Clayton et al., 1998), will create a disrupted slope topography that would promote the trapping of delta-derived, resedimented sands within irregular ‘basins’ relatively high on the slope itself. This would contrast with a predominantly rilled to gullied slope topography, as shown by Miocene clinoforms off New Jersey (Fulthorphe and Austin, 1998; Fulthorphe et al., 1999) and the Pleistocene, wave-dominated shelf margin off the Rhône Delta (Tesson et al., 1990, 2000). As observed in modern delta-fed slopes, gullies or chutes tend to form on the slope immediately off major distributaries, apparently in some relationship to enhanced sediment flux (Coleman and Prior, 1988). The gullies are believed to be initiated either as slide/slump scars (Prior and Coleman, 1978), or as erosional rills carved by sediment-gravity flows on an oversteepened slope (Pratson and Coakley, 1996) that, once formed, can propagate upslope and may intersect the shelf edge. Gullies are modified by erosion imposed by sediment-gravity flows and may possibly evolve into canyons (Pratson and Coakley, 1996). This type of disrupted slope topography is likely to signify sediment by-pass on the slope, causing sand transfer to deepwater areas beyond the base of slope. Consequently, slope facies can be expected to be rich in cohesive slumps, sandy debris-flow deposits and high-density turbidites, most of them derived directly or indirectly (slumped) from mouth-bar deposits (cf., Mayall et al., 1992). Such

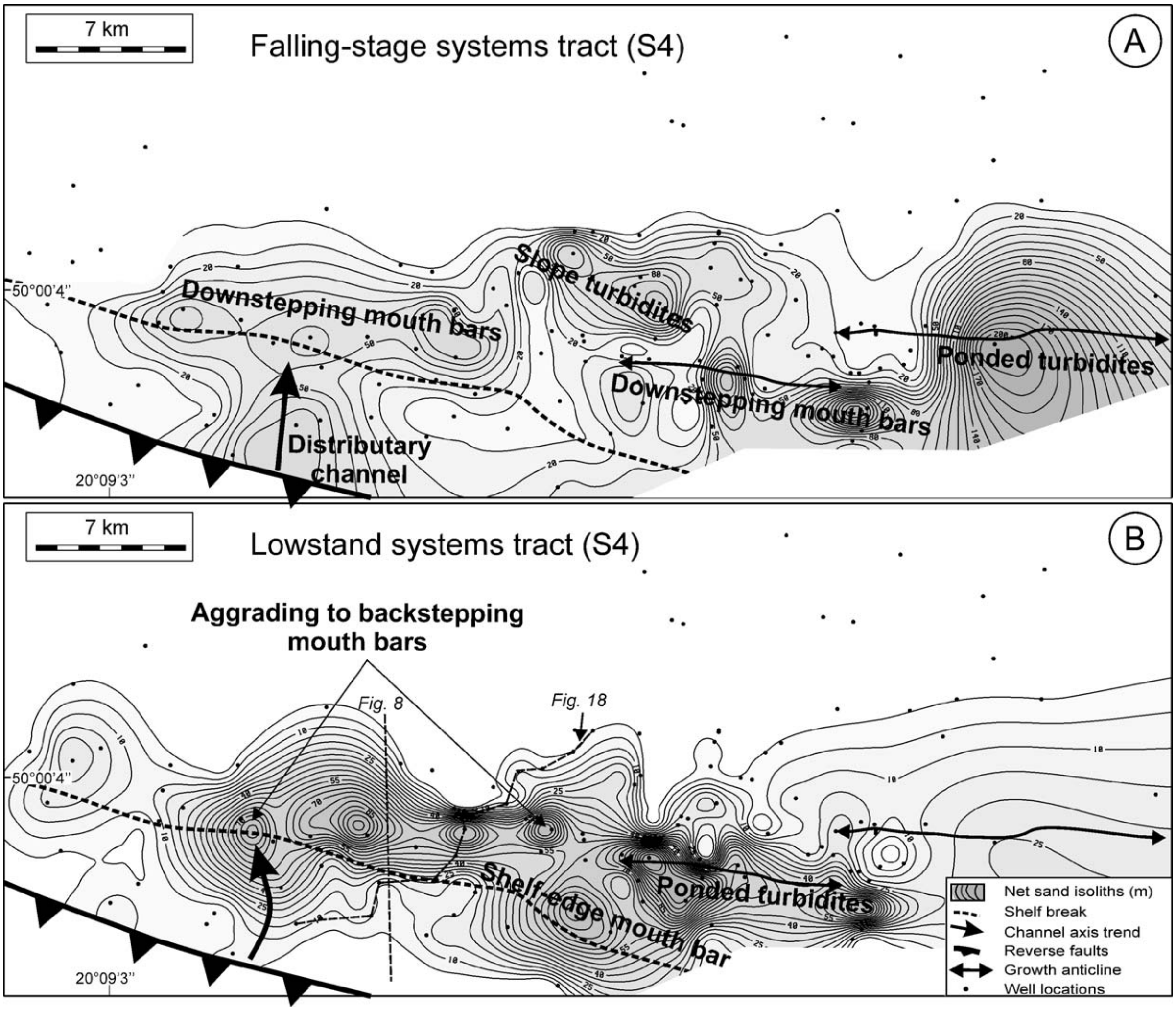


Fig. 19. Palaeogeographic interpretation of the falling-stage and lowstand systems tracts in Badenian deltaic shelf margin in the subsurface of the Carpathian foredeep (modified from Porębski et al., in press).

gullied slopes may reflect either a mature stage in slope evolution, due to diminished sediment flux, or development of a more wave-dominated regime along the shelf edge.

In summary, although there is practically no lower limit for the slope inclination that can induce gravity collapse (e.g., Field et al., 1982), shelf-margin delta slopes appear to be subject to deformation, slumping and collapse on a much greater scale than are the slopes of inner-shelf deltas. The slope oversteepening created by anomalously great sediment flux to the shelf margin promotes mass wasting and efficient sand transport to the base-of-slope setting. Once the deltaic wedge attains a critical weight, it will start to deform internally by growth faulting and the resultant slope basins may become efficient sediment traps.

6. Fluvial feeder to the shelf-margin deltas

With good coverage of seismic data, shelf-margin deltaic clinoforms commonly can be seen to merge landwards into a buried network of fluvial channels (Fig. 21). This has been particularly well documented

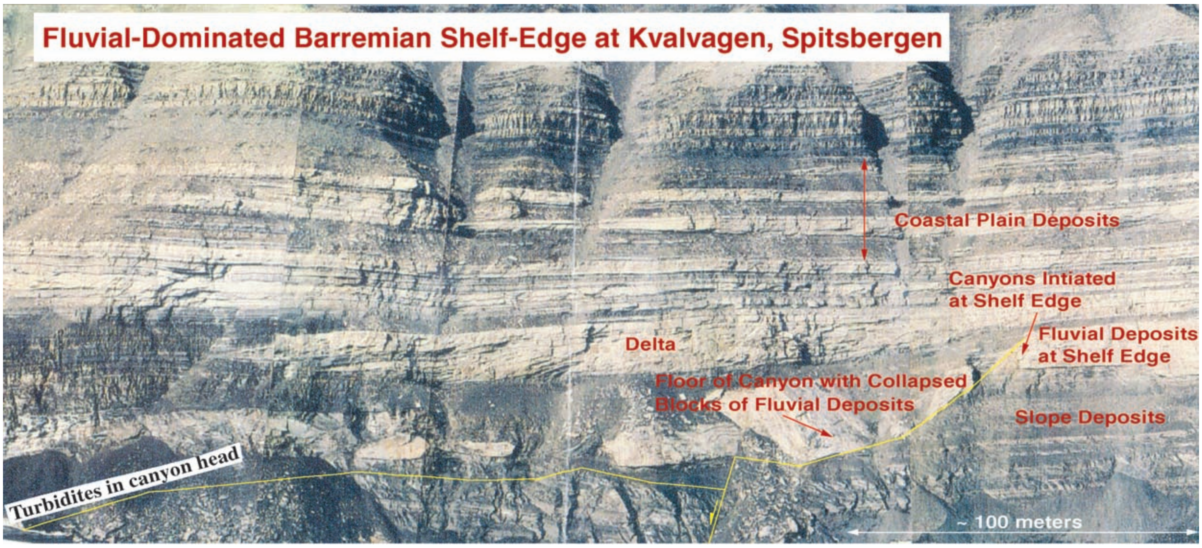


Fig. 20. Features of a falling-stage, Barremian shelf-edge collapse and upper canyon (Kvalvagen, Spitsbergen). Fluvial blocks 100 m long have fallen down into canyon head. Turbidite beds are present in canyon head at left end. Shelf-edge deltas are healing across earlier collapse area, after sea level again rose above the shelf edge (modified from Nemec et al., 1988; Steel et al., 2000).

in the Quaternary shelf of the Gulf of Mexico (Suter and Berryhill, 1965; Morton and Price, 1987; Kindinger, 1988; Suter et al., 1987). Although most of these channel systems have been interpreted as incised valleys, it is probably more correct to assume that channels that merge basinwards into thick shelf-margin deltas are the deltaic distributary channels. The formation of incised valleys on the shelf requires the shelf to be fully emerged, i.e., accompanied by the relative sea-level fall below the former shelf edge (Steel et al., 2000; Posamentier, 2001, p. 1789). Such conditions would promote incision and by-pass of the shelf margin rather than its accretion through deltaic deposition.

Near the shelf edge, the feeder or distributary channels can be hundreds of meters to several kilometers wide and 30–40 m deep (Suter and Berryhill, 1965; see also examples in Posamentier and Allen, 1999). Channels tend to show composite infills (Fig. 22), and the infill-lithology varies from clean, cross-bedded medium-grained, occasionally pebbly sand (Sydow and Roberts, 1994) to lithologies that are heterolithic and sometimes show upward-fining grain-size trends (Morton and Suter, 1996). The bases of channels incise the substrate and can in places be seen to truncate the underlying clinoforms. The latter is a signal of some deepening incision of the distributary channels, possibly during relative fall of sea level. However, the filling in some channels is thought to have been concomitant with the deposition of the most distal, aggrading clinothems (Fig. 7), usually after the initial rise of sea level (Morton and Suter, 1996; Sydow and Roberts, 1994), although there is a growing evidence that fluvial deposits can be formed and preserved during falling stage, and can in fact be more common than assumed so far (Blum, 1990; Posamentier, 2001). The top of the channelised strata is a flat, landward-rising truncation surface that extends to inter-channels areas and is invariably interpreted as a transgressive ravinement surface.

Farther landwards, the erosional relief on top of the clinoform sets increases and distributary channels merge into a network of incised fluvial valleys that can be up to 20 km wide and 60 m in relief, as in the southwestern Louisiana shelf (Suter et al., 1987). Suter and Berryhill (1985) emphasized that such an association is one of the main features that allow shelf-margin deltas to be distinguished from other types of shelf-edge progradation. In the absence of evidence of a connecting fluvial drainage system, clinoforms at the shelf-margin have sometimes been interpreted in terms of either prograding shelf-edge shorefaces (Hovland and Dukefoss, 1981; Trincardi

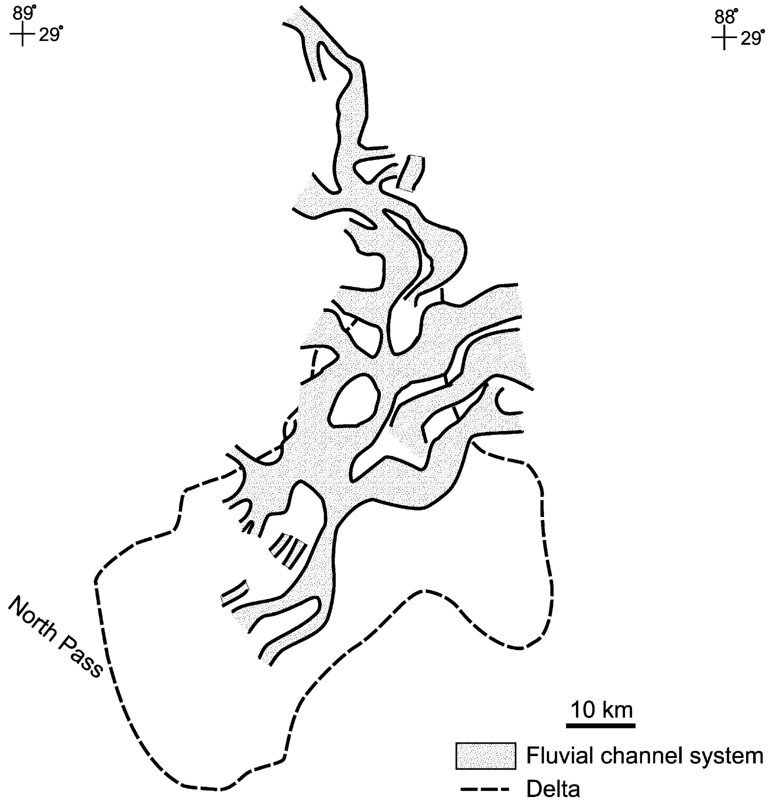


Fig. 21. Geometry of fluvial channel system associated with Late Pleistocene Lagniappe Delta, outer Mississippi–Alabama shelf (modified after Kindinger, 1988).

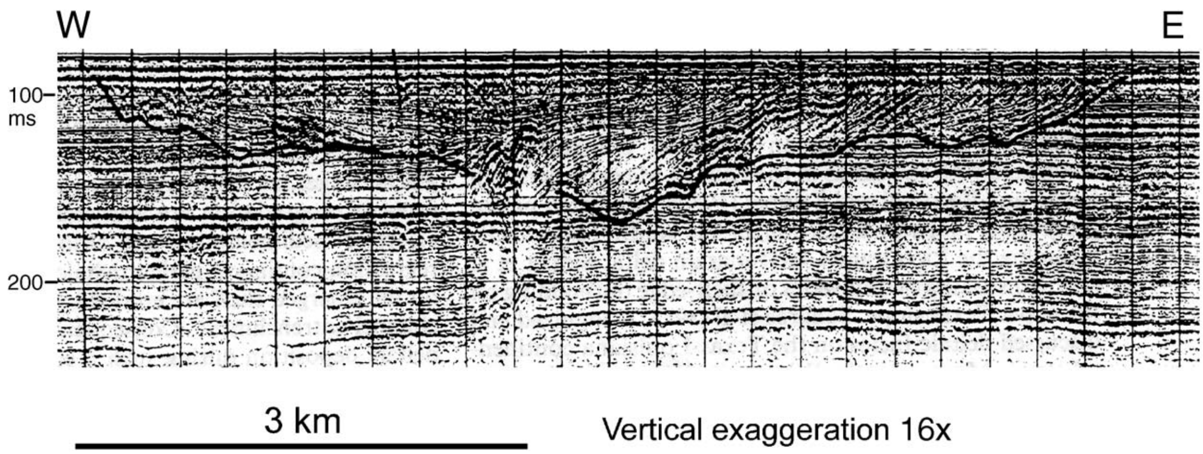


Fig. 22. Transverse seismic section through Late Pleistocene fluvial channel in outer Louisiana shelf, showing compound infill (after Suter and Berryhill, 1985).

and Field, 1991; Tesson et al., 1993; Berné et al., 1998; Trincardi and Correggiarri, 2000). However, where a fluvial feeding system is absent, care should be taken in making a non-deltaic interpretation. One of the central and inherent aspects of shelf-edge delta progradation is precisely the ability of the delta to cannibalise and remove evidence of its updip coastal plain during falling relative sea level. Moreover, subsequent transgression can certainly remove this evidence (e.g., 6–15 m depth of Holocene ravinement on the Gulf Coast—Rodriguez et al., 2001, p. 850) causing even greater exaggeration of a 'detachment' of the delta (cf., Ainsworth and Pattison, 1994). In addition, it is relatively difficult to transport large quantities of medium- or coarse-grained sand out to the shelf margin without fluvial feeders, except in cases of very narrow (< 20–30 km) shelves.

In other cases, the absence of fluvial feeders turns out to be merely apparent, due to the subtlety of shelf-drainage features coupled with the low resolution of seismic data. This is exemplified by Miocene clinoforms on the New Jersey shelf (Fulthorpe and Austin, 1998). Only when these clinoforms were analysed on images with a 5-m vertical resolution was the presence of distributary channels (100–400 m wide, 30 m deep and 10–15 km in spacing) revealed (Fulthorpe et al., 1999). Channel-like erosional forms having fills characterized by both concave-up and chaotic reflections have been observed at the top of prograding clinothem units of the western Rhône shelf (Tesson et al., 2000). These features were interpreted as shelf-edge slide/slump scars because of their in-filling style and their apparent restricted extent inboard of the clinoform offlap breaks (Tesson et al., 2000). However, distributary-channel origin was invoked for similar styles of infill in the Upper Pleistocene shelf-margin delta in the Eugene Island, offshore Louisiana (Hart et al., 1997).

There is also another way in which some of the fluvial feeder infills have been reinterpreted recently. Although the upper, channelised erosion surface of shelf-edge deltas is indeed likely to have been the feeder conduits for the delta-front growth (and this is seen especially in the youngest clinoforms of the set, as noted above), the in-filling of the irregular relief on this surface commonly happens later during transgression. During transgression, the earlier deltas become transformed into estuaries or embayments that overlie the axis of the previous distributary system (cf., Boyd et al., 1992). Shelf-edge delta/shelf-edge estuary couplets are now being recognised as the normal components of many "shelf" sand tongues, and this is confirmed by the strong tidal signatures and landward-directed paleocurrents now identified in the strata filling some of the so-called distributary channels (Schellpeper, 2000). Pleistocene examples of shelf-edge deltas transforming to shelf-edge estuaries during turnaround and transgression have been described by Berné et al. (1998). In all these cases, the erosive or incised tops of the regressive tongues or wedges testify to the former presence of fluvial feeders, even though the deposits in the upper part of the tongues are of transgressive origin.

7. Generation of shelf-margin deltas

7.1. Importance of sediment supply

Deltaic progradation out across a shelf to its margin can be caused by either (1) anomalously great siliciclastic sediment flux with fairly stable relative sea level (though there may be a problem with great water depth on the outer shelf) or (2) normal sediment flux with forced regression during relative sea-level fall. For example, Paleocene regional uplift within the continental interior of North America resulted in the generation of delta-driven shelf-margin accretion onto the inherited Cretaceous shelf edge in the Gulf of Mexico (Edwards, 1981; Galloway, 1989, 1990). Climatic cooling combined with uplift of the Appalachians caused a tenfold increase in clastic sediment input to the New Jersey shelf margin during the Miocene (Poag and Sevon, 1989). This resulted in delta-driven clinoforms offlaping onto the former, shallow-dipping carbonate ramp (Fulthorpe and Austin, 1998; Olsson et al., 2002). Although controlled by extrabasinal factors, great clastic influx and the resulting shelf-margin accretion need to be neither synchronous basinwide (Winker, 1982; Galloway, 1989, 1990) nor dependent on eustatic sea-level change. The latter is exemplified by the Balize Lobe of the modern Mississippi Delta, which during the Holocene transgression and subsequent highstand has prograded over a low-energy, storm-dominated shelf to its present location within 30 km of the shelf edge (Frazier, 1967). However, the

Mississippi River, with its continental-wide drainage and former access to continental ice sheets, is a river in a class of its own.

7.2. Evidence of sea-level fall during shelf-margin growth

In contrast, Pleistocene shelf-margin accretion on the Gulf Coast, though also delta-driven, appears to have been controlled by eustatic changes. There is a general acceptance that shelf-edge deltas commonly prograded during falling (though not necessarily much below the shelf edge) and lowstand of sea level. Supportive arguments include (see also Posamentier and Morris, 2000)

- landward delta termination by onlap-pinchout, so that the delta is separated by a zone of sediment by-pass from the highstand shoreline (Morton and Suter, 1996; Tesson et al., 2000);
- rapid upward-shoaling trend (Sydow and Roberts, 1994; Winn et al., 1998; Kolla et al., 2000) leading to foreshortened stratigraphy (Posamentier and Morris, 2000);
- seaward descending elevation and trajectory of sigmoidal clinoform tops (Curray and Moore, 1964) (that changes commonly to an ascending pattern in the youngest clinoform bundles);
- erosional truncation of clinoforms at their landward side (Winn et al., 1998), including the cutting of valleys on the shelf;
- chronostratigraphic evidence, based on oxygen-isotope ratios, showing that this upper unconformity surface at its final stage corresponds to the maximum sea-level lowstand (Kolla et al., 2000; Scott et al., 1998).

7.3. Shelf-margin delta growth during fall-to-rise cycle of relative sea level

A shelf-margin delta system comprises a clinoform set whose upper bounding unconformity represents the trace of either the fluvial feeder, or a combined fluvial-erosion/transgressive ravinement surface. As modelled by Sydow and Roberts (1994), eustatic sea-level fall forced both the fluvial feeder and the clinoforms across the shelf, with erosion and deposition prevailing at the trailing edge and leading edge of the system, respec-

Fig. 23. Edge of a small valley incision (12 m deep) cut into the outer shelf-edge delta deposits of an Eocene shelf-margin clinoform; Storvola, Spitsbergen.

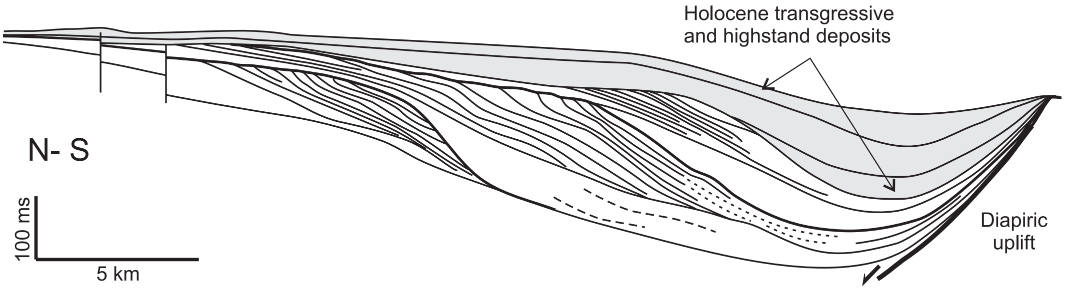


Fig. 24. Minisparker seismic section through late Wisconsinan Mississippi delta, showing the progradation and aggradation of deltaic shelf edge due to relative sea-level rise generated by local, diapir-related subsidence (after Suter and Berryhill, 1985).

tively. This process results in cannibalisation and incision of feeder channels into the former proximal clinoform tops (Fig. 23) and leads to an increased segregation of sand within more distal bundles (Fig. 7). The latter is consistent with the commonly observed seaward steepening of the upper foreset segments of the clinoforms (Lehner, 1969). The sigmoidal shape of the youngest descending clinoforms indicates that the shoreline did not drop below the shelf break (cf., Trincardi and Field, 1991). The latest stages of accretion at the leading edge of the system show an active up-building, with well-preserved sigmoidal clinoforms, indicative of early rise of sea level. Internal downlap surfaces within the clinoform body reflect either (1) lobe abandonment followed by reactivated delta growth (Sydow et al., 1992) or, (2) when laterally persistent, short periods of stillstand or small rises in relative sea level, superposed on an overall falling trend (Posamentier et al., 1992; Tesson et al., 2000). The aggradational stacking pattern of the youngest deltaic clinoforms, commonly taken as evidence of sea-level turnaround (fall to rise—Fig. 7), may in some cases reflect only a local rise in relative sea level. This is exemplified by the rising clinoforms break-point trajectory in the late Pleistocene Mississippi Delta (Fig. 24), due to sediment loading and subsidence behind a growing diapir (Suter and Berryhill, 1985).

8. Criteria for distinguishing shelf-margin deltas from other shelf deltas

There is a rich literature on the geometry, facies character and variability of inner mid-shelf deltas (Morgan, 1970; Galloway, 1974; Elliot, 1986; Coleman, 1988; Whateley and Pickering, 1989; Friedman et al., 1992; Orton and Reading, 1993; Reading and Collinson, 1996). With the advent of sequence stratigraphy, shelf-margin deltas became an important member of the delta family, and some of the early researchers (e.g., Edwards 1981; Suter and Berryhill, 1985) made a point of drawing some key distinctions between shelf-margin and inner-shelf delta types. Since that time, however, the fact these different delta types formed under differing sea-level regimes, and the consequence of this for their geometry, internal architecture and facies has been largely ignored. We summarize below the key differences.

The main distinctive criteria of shelf-margin deltas, in contrast to inner-shelf deltas include (Table 1)

- the very large scale of the clinoforms, that can be an order of magnitude greater in amplitude than that of mid- and inner-shelf deltas;
- the resultant thick, strike-elongated isopach image of shelf-margin deltas is commonly augmented by growth faulting;
- sigmoidal dip cross-sectional shape, with the thickest part located near the offlap break of a preexisting shelf margin;
- the landward pinchout by onlap onto shelf shales and the basinward pinchout by downlap within hemipelagic shales;
- evidence of foreshortened stratigraphy;
- the characteristic turbidite-prone nature of the delta-front/prodelta segment;
- the abundance and a large scale of slope-controlled soft-sediment deformation;
- the absence of a paralic “tail” along delta’s trailing edge.

Shelf-margin deltas show by far the largest clinoforms among all shelf delta types, because they build out into deep water beyond the shelf break. As soon as a delta reaches the former shelf edge, the initially thin-bedded, heterolithic delta front will become increasingly turbiditic in character because the fluvial feeder will empty directly onto a longer (minimum 2–4 km in length, if water depth exceeds 200 m) and steeper slope, promoting hyperpycnal discharges. Although some mid-shelf, wave-dominated deltas can build up steep (2–4°, maximum 8°) turbidite-rich slopes (see, Panther Tongue example in Posamentier and Morris, 2000), the resultant clinothems are an order of magnitude smaller (up to 15 m in the above example) and turbidite beds are few and thin, because of shallow water depth and the consequent short slope length.

In inner-shelf deltas, the combination of high-suspended-load yield with low slopes promotes the development of heterolithic, thin-bedded delta-front deposits in which, depending on delta-front regime, wave-modified or purely traction to suspension-fall-out structures dominate (Elliot, 1986; Friedman et al., 1992; numerous papers in Morgan, 1970, and in Whateley and Pickering, 1989). Such conditions promote also chiefly vertical soft-sediment deformation, such as mud diapirs and localized bar front slumps (Fisk, 1961; Coleman and Prior, 1988) rather than large-scale slope wasting. Invariably, inner-shelf deltas develop thick delta-plain deposits (e.g., Elliot, 1986; Coleman, 1988; Penland et al., 1988), because they are formed during conditions of slightly rising relative sea level (highstand systems tract). This would contrast with falling-stage shelf-margin deltas that do not have a paralic tail along their trailing edge, because such a tail either barely develops during forced regression due to the decreasing accommodation behind the delta front, or is entirely removed during subsequent transgressive ravinement.

9. Types of shelf-margin delta

As seen from the preceding review, two types of shelf-margin delta can be distinguished on the basis of the style and scale of internal gravity-induced deformation.

9.1. Sequence-stratigraphic context

During the development of sand-rich lowstand complexes beyond and below the shelf platform, shelf-edge deltas basically develop at two distinct times: during the fall of relative sea level that forces deltas to migrate out to the shelf edge, and during the subsequent rise of relative sea level back to a level above the shelf platform (provided sea level dropped below this platform to allow the development of a full-scale lowstand complex).

The deltas that develop during fall of relative sea level (with variable potential for subsequent preservation) would be variously classified as deposits within the early lowstand systems tract (Posamentier and Vail, 1988), within the late highstand systems tract (Van Wagoner, 1995) and within the forced regressive or falling-stage systems tract (Hunt and Tucker, 1992, 1995; Helland-Hansen and Gjelberg, 1994; Plint and Nummedal, 2000). Shelf-margin deltas developed at this time in the cycle have the potential to be severely or entirely cannibalised by valley incision on the shelf and canyon incision on the upper slope, as sea level continues to fall. However, it is important to be aware that they do develop at this time because significant remnants of these deltas can be preserved on the outer shelf and upper slope between and below zones of incision.

The deltas that reestablish at the shelf margin during rise of relative sea level, after a period of lowstand below the shelf platform, have been referred to as the lowstand wedge prograding complex (Vail, 1987), as the late lowstand wedge (Posamentier and Vail, 1988), the late lowstand systems tract (Normark et al., 1993) and the late lowstand (Kolla, 1993). The nature and timing of this late accretion has been well illustrated by Posamentier et al. (1991) and is shown in Fig. 25.

Although we have emphasized the role of changing relative sea level in the generation of these deltas, high supply of sediment can be additionally critical for their creation, as noted below, where we outline the characteristics of the two main classes of shelf-margin deltas.

9.2. Unstable shelf-margin deltas

Unstable shelf-margin deltas (Edwards, 1980) are characterized by great sediment thicknesses (up to

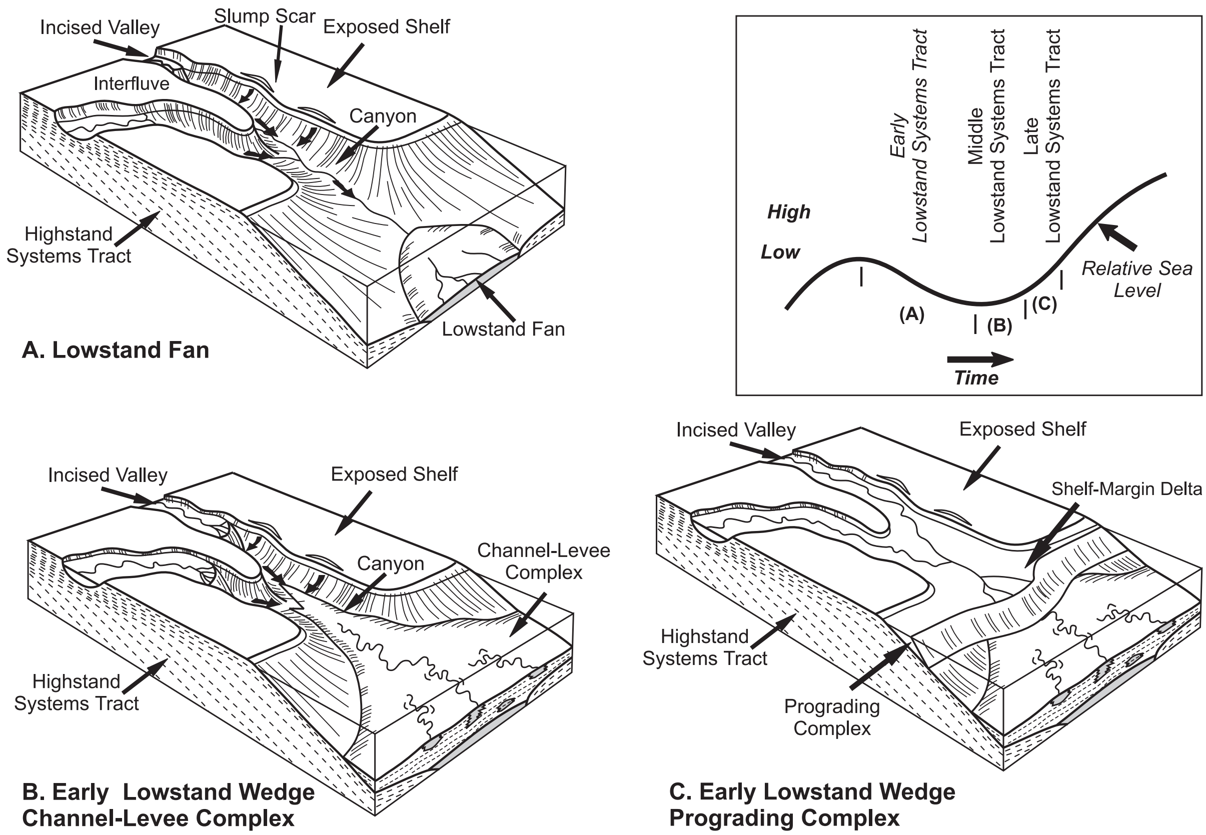


Fig. 25. Block diagrams of the lowstand systems tracts (modified from Posamentier et al., 1991; Normark et al., 1993). Note that the deposition of basin-floor fan is promoted when incised valley becomes linked to canyon on the slope during relative sea-level fall (A), while shelf-margin delta is the dominant component of the prograding complex that is formed during early rise (C).

several hundreds of meters) and associated listric growth faults and rollovers along the shelf edge (Fig. 26A). Rotational slides are common on the upper slope, whereas compressional ridges affect the slope toe (Winker, 1982). Where the slope is long, the above irregularities tend to pond sand turbidites and heterolithic facies on the slope. Such deltas develop due to a combination of great accommodation space available in the deep water beyond the shelf platform, with great sediment flux to the shelf edge, either due to tectonically active source areas or anomalously big rivers. High sediment supply and relative sea-level fall below the shelf edge promote development of canyons which once connected with shelf valleys, will cannibalise the former deltaic shelf margin and bypass much of the sediment directly to the basin floor to build basin-floor fans. The formation of unstable shelf-margin deltas appears to be related commonly to very high sediment supply, as well as to relative fall of sea level, and can be facilitated by the flow of evaporates in the underlying slope sediment that results in additional accommodation. Tertiary deltas of the Gulf Coast belong to this type (Edwards, 1980, 1981; Winker, 1982; Winker and Edwards, 1983; Xue and Galloway, 1995; Prather et al., 1998).

9.3. Stable shelf-margin deltas

These are typified by smooth shelf-margin morphologic profile and little to no growth faulting along the shelf edge (Fig. 23B). Large-scale sliding/slumping is not common, although it may locally be voluminous (e.g., Lehner, 1969), and it is essentially limited to collapsing mouth-bars. The slope tends to

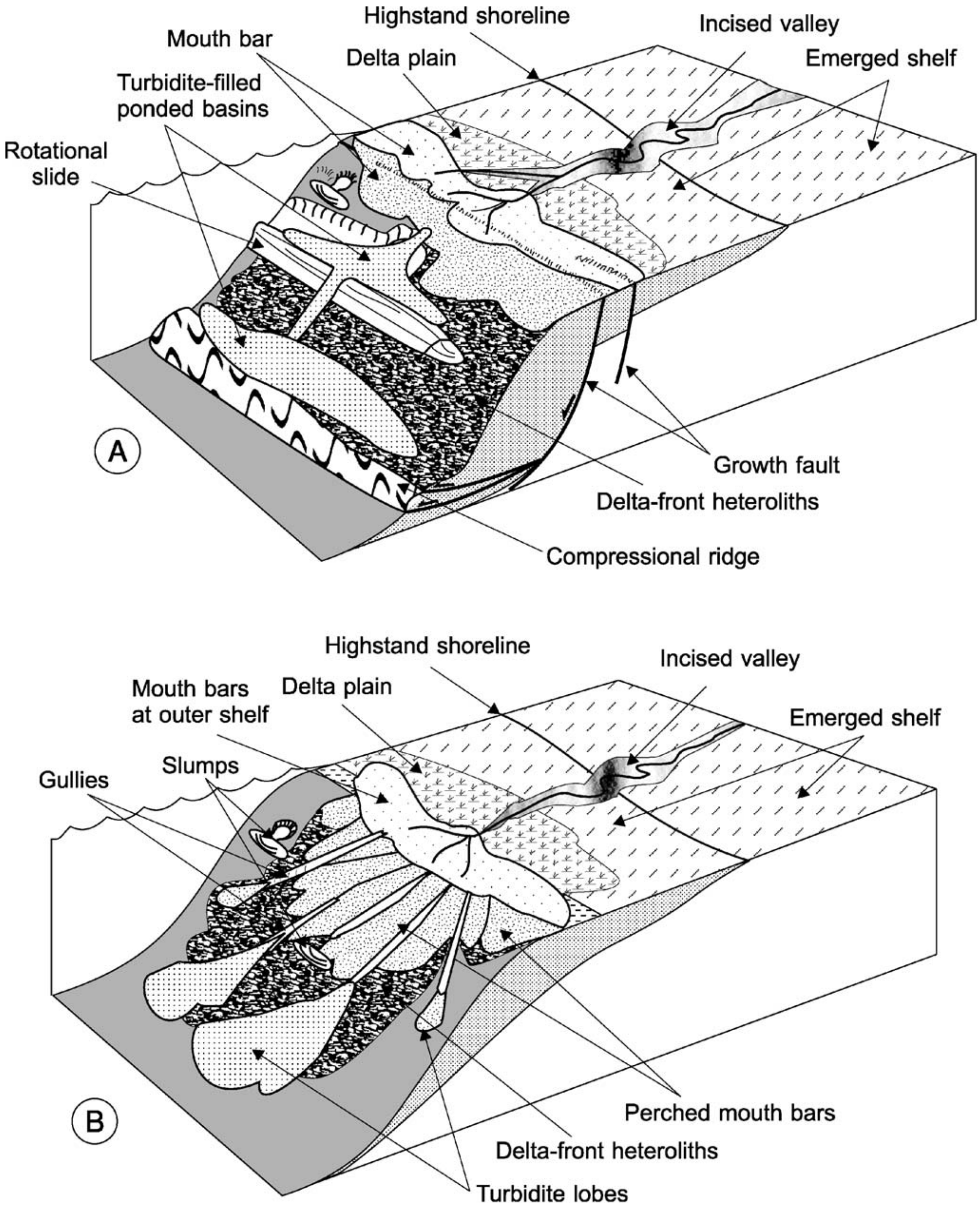


Fig. 26. Cartoonal representation of unstable (A) and stable (B) shelf-margin deltas. Because of an association with sea-level fall below the shelf edge, both types of shelf-margin deltas, during their main regressive transits, tend to have their delta plain and upper delta front removed by erosion.

be smooth, rilled to gullied with neither major canyons cut in it nor any incised valleys breaching the shelf edge. The delta-front/slope facies vary from mud to sand prone. In the latter case, it can contain thicker turbidite sands of hyperpycnal origin (they fail to ignite on the slope, and therefore die on the slope). These sands tend to form strike-oriented to lobate bodies of limited lateral extent that can reach down to the middle or even the lower slope. Stable shelf-margin deltas appear to be associated with (1) small to modest rivers that reestablish at the shelf edge after sea level rises back to this level, just prior to transgression. Such deltas can also form with (2) modest relative sea-level falls to the level of the shelf platform, but not below. Many Quaternary shelf-margin deltas belong to this category.

10. Implications for sequence stratigraphy

10.1. Sea-level drop and shelf-edge incision

Although submarine canyons form in a variety of ways (Pratson et al., 1994), they seem to share one feature, i.e., an association with high sediment supply to the shelf edge and upper slope, whether or not they are visibly backed by fluvial valleys (May et al., 1983). Such conditions promote slope loading, slope deformation and headward erosion by retrogressive mass wasting and turbidity currents, and appear to be responsible for the initiation of most canyons located off major deltas (Coleman et al., 1983). With relative sea-level fall below the shelf edge, such canyons are able to capture river inflow and become connected to incised valleys on the shelf (Figs. 23 and 25), as well as to a coeval basin-floor fan (Posamentier and Vail, 1988). Well-documented examples of submarine canyons that are directly connected to fluvial erosional systems at their shelf side include the Rhône and Nile canyons that were generated during a spectacular sea-level drop associated with the Messinian salinity crisis. Examples of similar linkage between canyons and fluvial systems have been shown from the Cretaceous and Eocene by Nemec et al. (1988) (Fig. 20), Steel et al. (2000) and by Mellere et al. (in press). A similar origin has also been proposed for the Zaire Canyon, whose head lies well within the Zaire River estuary (Droz et al., 1996). As implied by classic sequence stratigraphic models, establishment of such a direct fluvial-valley/canyon connection enhances the sediment transfer directly to the basin floor (see discussion in Posamentier and Allen, 1999). This incision at the shelf edge tends to cannibalise, partly or completely, the earlier formed falling-stage deltas, and precludes the generation of new deltas until sea level reestablishes above the shelf edge during later rise.

As shown by Quaternary examples, falling sea level generates a prograding deltaic-shelf margin which is fed by distributary channel discharge plus the products of erosion of the increasingly emergent inner shelf. The latter is overridden by subaerial incision, cut progressively basinwards by the fluvial feeder (Fig. 27). In this way, the shelf margin progrades through the welding of successive deltaic clinoforms, and is apparently never overtaken by the zone of fluvial erosion, except where sea level falls significantly below the shelf edge, and fluvial systems and slope canyons link directly with each other. Talling (1998) predicted that during sea-

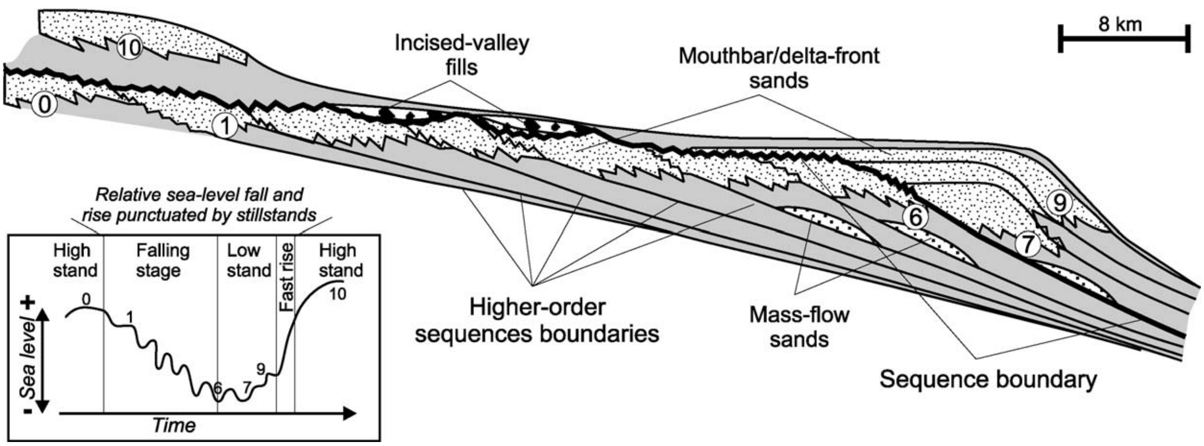


Fig. 27. Stratigraphic architecture generated at shelf edge during prolonged, stepped fall followed by rapid rise in the sea level (isotopic stages 5 to 1), as exemplified by Late Pleistocene Lagniappe Delta, Gulf of Mexico (after Kolla et al., 2000, slightly modified). The stepwise fall resulted in a series of downstepping shelf-edge deltaic increments, each based by minor unconformity attributed to a high-order sequence boundary. The pronounced unconformity atop of forced regressive deltaic wedge is taken as the master sequence boundary.

level falls not extending beyond the shelf break, incisions will be restricted to the inner shelf, with a maximum relief of 20–70 m centred at the former highstand coastal shoreline, decreasing both land- and seawards in relief (see also Posamentier and Allen, 1999). This seems to be the case for most of the incised valley feeders of Pleistocene shelf-margin deltas. In fact, evidence of dissection of topographic deltaic-shelf edge by incised valleys is at least equivocal at this time. Farrán and Maldonado (1990) recognized several canyons incised into Quaternary shelf-margin deltaic units off the Ebro Delta, but interpreted them as submarine incisions. Tesson et al. (2000, p. 137) described prominent U-shaped troughs extending 5–10 km landwards of the shelf break and showing sagging reflections. They interpreted this type of trough as "canyon grading updip in some instances into an incised valley", but did not elaborate further on this subject. Weimer (1990) noted that 10–15 Pleistocene canyons on the Louisiana slope had updip connections to incised valleys. However, except for the ancestral Mississippi River-canyon system, no such a connection was proved in the northern Gulf of Mexico (Suter et al., 1987; Sydow and Roberts, 1994; Morton and Suter, 1996). Posamentier (2001) noted the scarcity of incised valley systems in the Pleistocene of offshore Java, and concluded that their formation would be limited only to those very short periods of the maximum lowstand when sea-level fall exceeded 110–120 m.

Rate of sea-level fall can be classified as rapid or slow, when it is greater or less, respectively, than the rate of subsidence at the shelf edge (Posamentier and Vail, 1988). Rapid fall below the shelf edge results in dissection of the shelf by fluvial erosion. However, the effects of Pleistocene glacieustatic sea-level falls demonstrate, at least on passive margin shelves, that even large magnitude (120–150 m) falls that are associated with high-frequency, asymmetrical sea-level changes should not a priori be expected to result in fluvial dissection of the shelf break (McMaster et al., 1970; Trincardi and Field, 1991; Morton and Suter, 1996). Such falls are apparently not fast enough, probably because they are interrupted by stillstands and small rises, so that during much of the period of overall sea-level fall, the shelf was never fully exposed (Posamentier, 2001).

Sydow and Roberts (1994) noted that the water depth of the present-day shelf break on passive margins (75–135 m, Table 2) is close to the sea-level minimum estimates (120 m) for the last glacial, thus, promoting shelf-edge accretion rather than dissection during sea-level fall. During this glacial period, sea level was falling at ca. 1.2 m/ka, whereas Mississippi Delta lobes are presently subsiding locally at a rate of up to 10 m/ka (Nummedal, 1983). These figures highlight the importance of subsidence in deltaic shelf-margin settings (Winker, 1982; Galloway, 1989; Tesson et al., 1993), which is perhaps the most important single factor preventing the extension of incised valleys through to the deltaic shelf edge.

Morton and Suter (1996) suggested that the formation of linked incised valley-submarine canyon systems could be facilitated by the presence of a river mouth near the shelf margin when sea level begins to fall, i.e., at the peak of highstand. This would create particularly steep shelf-to-slope gradients so that fall of sea level below the shelf break would effectively cause the entrenchment of the fluvial system and cannibalisation of former deltas at the shelf margin. Burgess and Hovius (1998) have calculated the time necessary for 24 modern deltas to arrive at the shelf edge in present-day highstand (stillstand) conditions. This required transit times that varied between 8.5 and 116.5 ka. However, these calculations did not take into account the effect of rising relative sea level that is an integral part of the highstand systems tract. Using a 2-m/ka eustatic rise that is typical for the last 7 ka (Fairbanks, 1989), these minimum times would have to be increased by a factor of 1.09–1.69 (1.30 on average) and, in some cases, the deltas would never reach the shelf edge (see Muto and Steel, in press). This necessary transit time greatly exceeds the duration of fourth-order glacieustatic sea-level highstand intervals (for example, ca. 18 ka for the last 120-ka sea-level oscillation), and so, not many of the 24 modern deltas would be able to deliver sand to the shelf edge or to the adjacent deepwater slope during even slightly rising relative sea level. We should add that the necessary transit time above probably also exceeds the regressive transit time of fourth-order regressions during non-glacial periods.

Pleistocene glacieustatic sea-level cycles are strongly asymmetrical, because they reflect periods of slow cooling and rapid warming (Shackleton, 1987). This results in prolonged conditions of stepped sea-level fall and very short periods of sea-level rise, as also reflected in a characteristic asymmetry of forced regressive deposits versus transgressive and highstand deposits (e.g., Chiocci, 1994; Hernández-Molina et al., 2000). To the extent that the Pleistocene sea-level curve can be used as analogue to glacieustatic oscillations in the more distant past, it seems very unlikely that glacieustatic control alone is able to bring highstand deltas consistently to the shelf margin, even during third-order cycles (cf., Haq et al., 1987). Other limiting factors include an increasing tendency for rivers to switch in an autocyclic manner as they lengthen, and the inherent tendency of prograding deltas, during even modest rates of rising sea level, to be subject to auto-retreat (Muto and Steel, 1992, 2000).

10.2. Systems tract assignment

There is a general consensus that Pleistocene deltas prograded across the shelf and located at the shelf-margin during relative sea-level fall. However, the systems tract assignment of such forced regressive deposits and the stratigraphic position of the associated sequence boundary are still a matter of controversy. Such deltas have been assigned to early lowstand (Posamentier and Vail, 1988), stable highstand (e.g., Torres et al., 1995), highstand (Sydow and Roberts, 1994), late highstand (Van Wagoner, 1995), early lowstand (Van Wagoner et al., 1990; Posamentier and Allen, 1993; Posamentier and Morris, 2000), detached lowstand (Ainsworth and Pattison, 1994) and to the forced regressive systems tract (Hunt and Tucker, 1995). Clearly, this reflects conflicting tendencies, on the one hand, to describe stratigraphic architecture in terms of classic sequence stratigraphy, and on the other, to modify and extend existing terminology in the light of new understanding of stratigraphic patterns (cf., Wilson, 1998; Friedman and Sanders, 2000). This debate is not merely a matter of nomenclature; it has consequences for facies predictions.

Shelf-margin deltas that form during relative sea-level fall are, by definition, forced regressive deposits, and are directly or indirectly the feeders to sharp-based shoreline sandstones. Such deltas, when occurring below the shelf platform to perch on the slope, have also been used to provide evidence in the debate regarding the three- versus four-systems-tract sequence terminology (Hunt and Tucker, 1992, 1995; Kolla et al., 1995). However, it should be noted that shelf deltas also form during rising relative sea level, as when they reestablish at late lowstand, as sea level rises to back above the shelf margin. This delta type has been designated as the 'late prograding complex' of the classic lowstand systems tract (Posamentier et al., 1991). There are significant differences in architectural style between these two cases of shelf-edge delta. The one shows stepped fall during progradation, in contrast to the other that shows aggradation during progradation (Figs. 11 and 27).

The erosional unconformity along the top of a shelf-margin deltaic body begins to form already during initial relative sea-level fall (Sydow and Roberts, 1994). Initially, fluvial channels incise vertically, but their lateral migration increases when the rate of fall slows (Hart and Long, 1996). Thus, the unconformity continues to be generated throughout the period of stepped fall to the point of lowest relative sea level (Suter and Berryhill, 1985), and it then passes into a conformable surface at the turnaround to relative sea-level rise (Hart and Long, 1996; Kolla et al., 2000). Below the initial shelf break, this unconformity becomes an onlap surface that separates the downstepping forced regressive deposits formed during the sea-level fall from the aggrading-to-backstepping deltas formed during the early rise in relative sea level. The initial forced regressive surfaces are unlikely to be of basinwide extent, whereas the erosion surface that continues to be generated across the tops of the delta all during the fall to the minimum sea-level position is likely to be widespread. This mitigates against including the forced regressive deposits within the lowstand systems tract.

10.3. Sequence boundary

Exxonian depositional sequences are bounded by unconformities that are formed during relative-sea-level falls (Vail et al., 1984). A stepped fall results in a series of downstepping wedges, each based by a regressive surface of marine erosion, and cumulatively forming a basal downlap surface. The high-frequency erosion surfaces at the base of each downstepping wedge record basinwards facies shift and represent higher order sequence boundaries (Posamentier et al., 1992). Such boundaries merge upwards and landwards into a major erosional unconformity that reflects subaerial incision often modified subsequently by transgressive ravinement. A debate has now centred as to whether the master sequence boundary should be placed at (1) the first downlap surface separating normal (highstand) regression from forced regressive deposits (Posamentier et al., 1992; Tesson et al., 2000; Posamentier and Morris, 2000), or

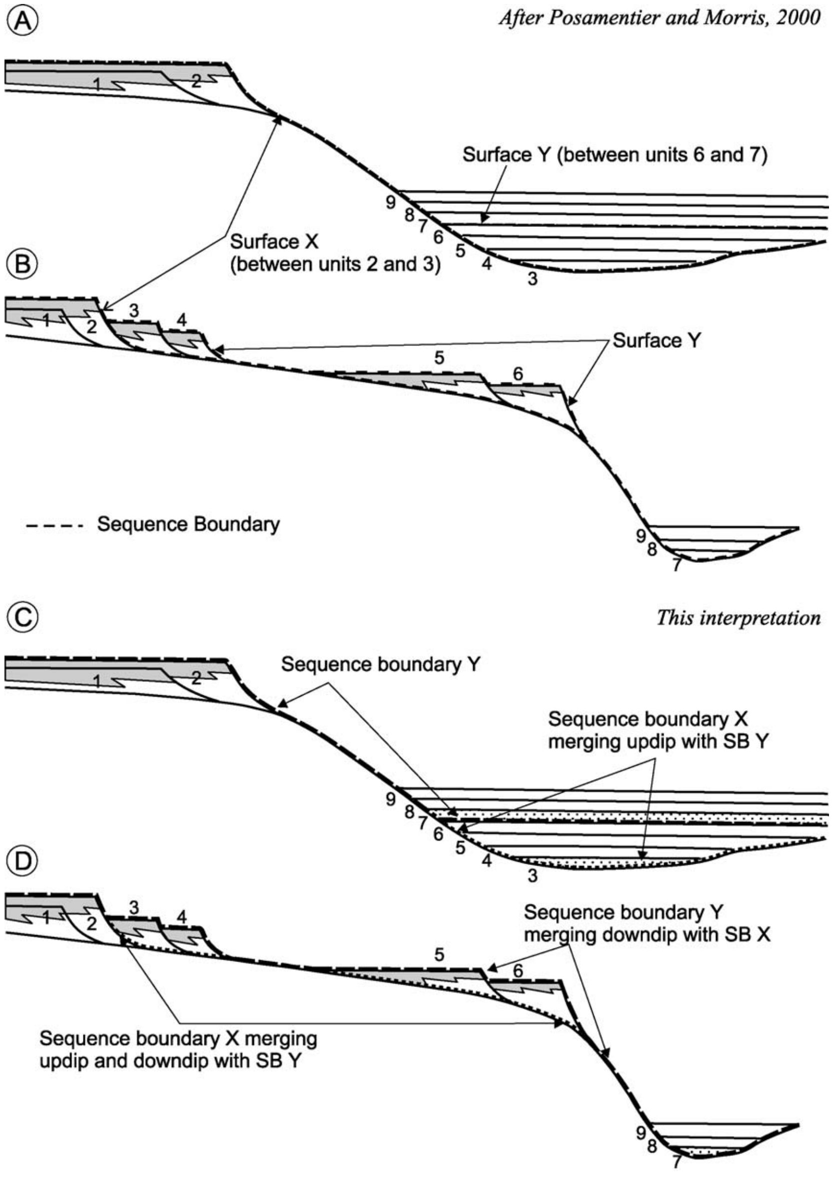


Fig. 28. (A and B) Stratigraphic architecture of shelf margin through different physiographic settings in a basin affected by the same relative sea-level fall, showing different placement of potential master sequence boundary (modified from Posamentier and Morris, 2000). In (A), highstand shoreline reached the shelf edge and mass wasting to the deep-water area both began already at time 3, while in (B), this had not started until time 7. Posamentier and Morris (2000) argue that isochronous though largely cryptic surface X is a better candidate for the sequence boundary than surface Y, because the latter gives a problematic correlation between the locations (A) and (B). (C and D) Alternative interpretation of the same sections, envisaging two sequence boundaries. Sequence boundary Y, though diachronous and polygenetic, is well recognizable across the entire basin, because it corresponds at its final stage to the maximum sea-level fall that coincides at both locations with enhanced mass-flow deposition (dotted).

at (2) the erosional unconformity that is perpetuated across the top of the forced regressive wedge (Hunt and Tucker, 1992, 1995; Helland-Hansen and Gjelberg, 1994; Plint and Nummedal, 2000).

An increasing use of the second choice has been based on the fact that this surface commonly forms the most prominent and easily recognizable discontinuity on the shelf. This approach has recently been criticized by Posamentier and Morris (2000; see also Posamentier and Allen, 1999), who point out that this surface is diachronous and its location will vary depending on variations in local physiography and on the extent of database available (data window); hence, paleogeographic maps based on such an assignment will be meaningless. However, it should be stressed that this is a weakness of all prominent discontinuities, as most are diachronous. Posamentier and Morris (2000) (Fig. 28A and B) propose to locate the master sequence boundary at the base of the first forced regressive wedge and its correlative conformity into the basin. They argue that such a choice will not only provide the necessary isochroneity to the sequence boundary but it will also help resolve stratigraphy within a basin in which the same relative sea-level fall affected the highstand shoreline that is near the shelf edge along one part of the coast, but far inboard of the edge in another part.

The main problem with the sequence boundary so defined is that for much of its extent it remains cryptic and difficult to identify. Moreover, although on cartoon drawings and conventional seismic sections, such a surface may appear as a downlap unconformity (Vail et al., 1977), it is not necessarily so when traced downdip. This boundary continues downdip as a correlative conformity that, however, can be difficult to pinpoint within a wide zone of ascending downlapping contacts. Furthermore, Posamentier and Morris (2000) omitted to indicate on their drawing (Fig. 28A and B) that in both coastal segments surface Y, atop the forced regressive wedge, forms another sequence boundary—one that actually corresponds to the maximum sea-level drop within the portrayed stratal architecture (Fig. 28C and D). Thus, we conclude that this architecture shows two sequence boundaries: one, located between wedges 2 and 3, running atop the highstand and below the first forced regressive wedge, corresponds to an initial sea-level fall that affected only a small portion of the shelf, albeit producing a basin-floor fan offshore location B. The other surface (Y) that is located between wedges 6 and 7, running atop the forced regressive body, corresponds to the final fall (the sum of falls between wedges 2–3 and 6–7) that affected the entire shelf area and can be associated with basin-floor fans across the entire basin plain (Fig. 28C and D).

Nonetheless, we feel that choices (1) versus (2) should not be set up as alternatives. In some situations, particularly in ramp settings where a seaward part of the forced regressive deposits is detached from their proximal part, the master sequence boundary should occur below the detached segment as the latter may actually correspond to lowstand systems tract developed during initial sea-level rise (e.g., Mellere and Steel, 1995). In the presence of a shelf break, the upper unconformity appears to be a more correct choice. In all cases, the available data window will affect any choice; this cannot be remedied by putting the sequence boundary at more or less cryptic surfaces.

11. Concluding remarks: architecture of shelf-margin deltas as predictor for slope and basin-floor turbidites

It is clear from the preceding discussion that shelf-margin deltas record essentially periods when the sediment is stored at the shelf margin or when sediment contributes to the accretion of the shelf margin, rather than necessarily periods when sand is delivered into the basin floor. This implies that shelf-margin deltas and basin-floor fans do not generally or automatically form temporally linked entities, though a genetic link between shelf-edge deltas and deepwater systems nonetheless is very likely.

In dip section, some shelf-margin delta complexes reveal a bipartite stratal architecture, with a lower and older part composed of progradational and downstepping increments, and an upper and younger part with an aggradational to backstepping stacking pattern (Fig. 27). There may or may not be significant erosion between these two parts. This pattern records deposition during relative sea-level fall and subsequent early rise, with the downdip length of each segment reflecting the duration of forced regression (or early lowstand) and late lowstand, respectively. At least three stratigraphic scenarios portraying either fully developed, base-missing, or top-missing patterns can be envisaged (Fig. 29).

1. The presence of a marked aggrading-to-backstepping shelf-margin deltaic wedge, overlying a widespread erosional surface on the shelf and downlapping onto an irregular or disrupted slope implies conditions of shelf-margin accretion that were preceded by a period of significant sea-level fall below the shelf edge, shelf-edge incision and shelf-edge delta cannibalisation (Fig. 29A). This situation is the one portrayed in the classical Exxonian depositional sequence model associated with Type 1 sequence boundary (Posamentier and Vail, 1988). This scenario predicts the downdip

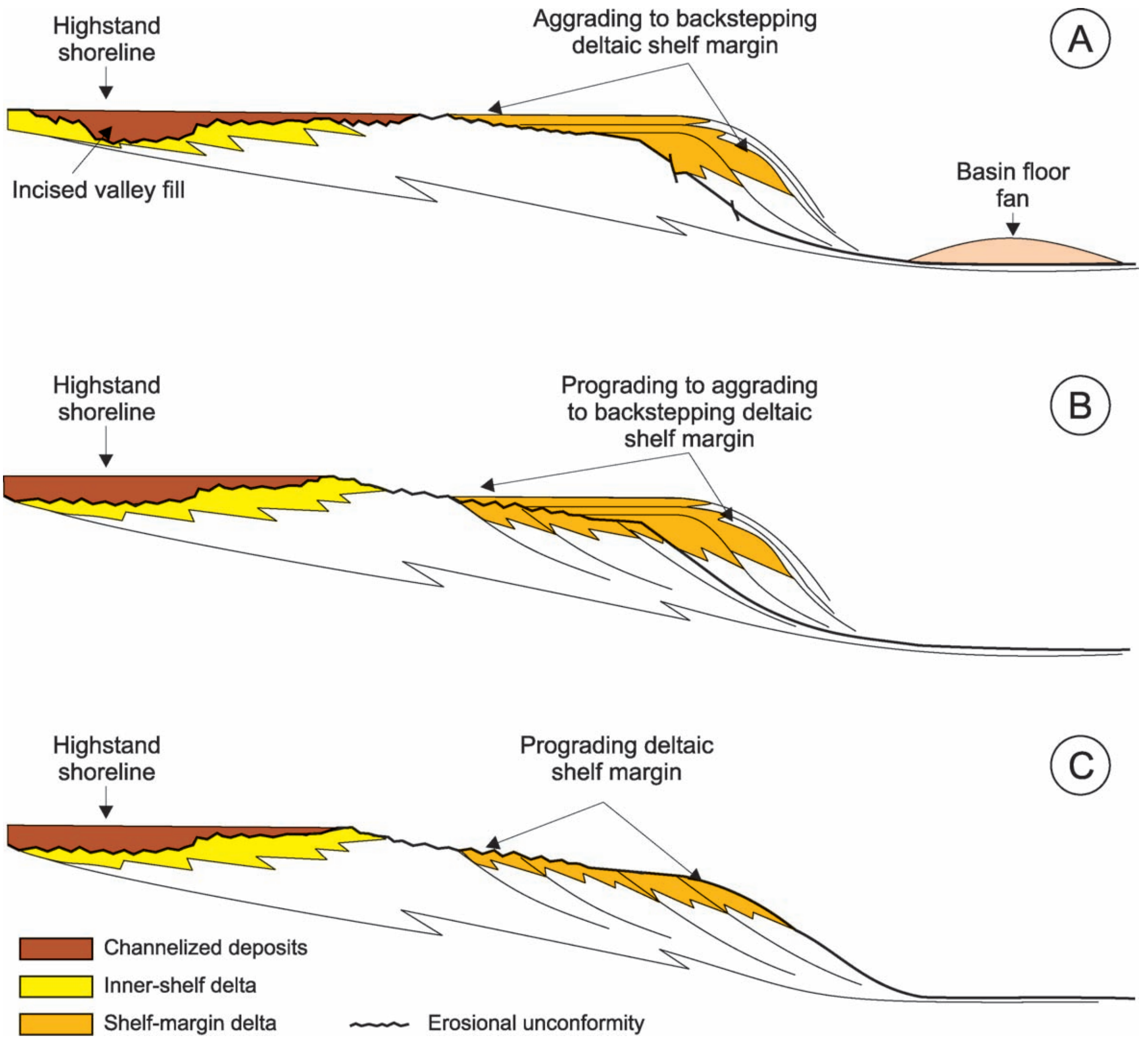


Fig. 29. Stratigraphic architectures of deltaic shelf margin as potential predictors for the presence or absence of basin-floor fans. (A) Aggrading-to-backstepping complex overlying a growth faulted/disrupted slope and unconformity on the shelf (falling-stage delta deposits cannibalised during sea-level fall below the shelf edge). This scenario broadly corresponds to Exxon's Type 1 sequence. (B) Prograding-to-aggrading-to-backstepping shelf-margin delta complex, without internal unconformity (Exxon's Type 2 sequence). (C) Prograding shelf-margin delta complex overlain by transgressive shales—this scenario may reflect either reduced sediment supply during progradation, or relative sea-level fall turning abruptly to rise.

presence of basin floor-fans, as documented in the Eocene deltaic shelf-margin of Spitsbergen (Steel et al., 2000; Plink-Björklund et al., 2001).

2. Fully developed shelf-margin delta complexes signify optimum shelf-margin accretion, and consist of a fully preserved, falling stage regressive segment, an overlying aggrading segment and a retrogressive cap (Fig. 29B). There will be an unconformity above the falling-stage deposits on the shelf, but no unconformity on the slope. In such cases, it is likely that the relative sea level fell as the deltas prograded to the shelf edge, but the fall did not persist below the shelf edge. The complex continued to accrete during sea-level rise and transgression. It is unlikely that there was significant shelf-edge incision, and no basin-floor fans are predicted. This type of deltaic complex resembles the Exxon Type 2 sequence (Posamentier and Vail, 1988). Posamentier (2001) suggested that even for large-magnitude (>100 m) sea-level oscillations, it is mainly the duration of fall relative to the duration of rise that determines whether or not valley incision will extend across the entire shelf. His data shows that the longer the fall takes to gradually expose the shelf, the smaller is the chance for dissection of the shelf edge and, if this happened, it would be limited to a very short period of maximum lowstand. We might then predict that the likelihood of getting a basin-floor fan would increase for relatively small lengths of the progradational-to-downstepping segment and the fan, if developed at all, should be looked for on a surface correlative to the turnaround surface from the base level fall to rise and not on a downlap surface marking the initial sea-level drop.
3. The presence of a shelf-margin delta complex with only a progradational-to-downstepping segment overlain by transgressive shales, and no evidence of shelf-edge incision by the delta's distributaries implies conditions of prolonged forced regression, but without sea-level fall below the newly formed shelf edge. The rapid turnaround to capping transgressive shales suggests decreased sediment supply and rapid retrogradation of the deltaic system, without any significant intervening aggradational phase. In such a case, the deltaic shelf-margin accreted by the broad dispersal of deltaic mouth bars. However, the lack of incision and lack of consequent focusing of sediment delivery through a valley-canyon system, predicts that little or no sand escaped to the basin floor (Fig. 29C).

The above architectural features at the shelf margin could be used as an initial predictor of the likely presence or absence of time-equivalent basin-floor fans. If, in addition, valley incision on the shelf and/or canyon incision on the slope can be imaged, or if some of the other features of unstable shelf-margins that are described above can be documented, then our ability to predict deepwater sand on the basin floor is even more likely.

Acknowledgements

RJS thanks the WOLF-consortium members (Norsk Hydro, Conoco, Phillips, BP and PDVSA) for continued and enthusiastic support for the shelf-margin research work in Spitsbergen. SJP thanks Institute of Energy Research and University of Wyoming for support during Spring 2000 when much of this review was undertaken. The study was also supported by a State Committee of Scientific Research (KBN) Grant T12B05218 to SJP, and by Landmark Graphics via the Landmark University Grant Program (grant to Institute of Geological Sciences, Polish Academy of Sciences). The final manuscript benefited greatly from critical and insightful review by Henry W. Posamentier, Mike Mayall and Gerald M. Friedman.

References

Ainsworth, R.B., Pattison, S.A.J., 1994. Where have all the lowstands gone? Evidence for attached lowstand systems tracts in the Western Interior of North America. Geology 22, 415–418.
Aksu, A.E., Piper, D.J.W., 1983. Progradation of late Quaternary Gediz delta, Turkey. Mar. Geol. 54, 1–25.
Aksu, A.E., Piper, D.J.W., Konuk, T., 1987. Late Quaternary tectonic and sedimentary history of outer Izmir and Candarli Bays, western Turkey. Mar. Geol. 76, 89–104.
Asquith, D., 1970. Depositional topography and major marine environments, Late Cretaceous, Wyoming. AAPG Bull. 63, 324–340.
Badalini, G., Kneller, B., Winker, C.D., 1999. Late Pleistocene Trinity–Brazos turbidite system: depositional processes and architectures in a ponded mini-basin system, Gulf of Mexico continental slope. AAPG Bull. 83, 1297–1298.

Berné, S., Lericolais, G., Marsset, T., Bourillet, J.F., de Batist, M., 1998. Erosional offshore sand ridges and lowstand shorefaces: examples from tide- and wave-dominated environments of France. J. Sediment. Res. 68, 540–555.

Bhattacharya, J.P., Willis, B.J., 2001. Lowstand deltas in the Frontier Formation, Powder River basin, Wyoming: implications for sequence stratigraphic models. AAPG Bull. 85, 261–294.

Blum, M.D., 1990. Climatic and eustatic controls on Gulf coastal plain fluvial sedimentation: an example from the Quaternary rivers, Texas. In: Weimer, P., Posamentier, H.W. (Eds.), Siliciclastic Sequence Stratigraphy—Recent Developments and Applications. AAPG Mem., vol. 58, pp. 259–283.

Boyd, R., Dalrymple, R.W., Zaitlin, B.A., 1992. Classification of clastic coastal depositional environments. Sediment. Geol. 80, 139–150.

Burgess, P.M., Hovius, N., 1998. Rates of delta progradation during highstands: consequences for timing of deposition in deep-marine systems. J. Geol. Soc. London 155, 217–222.

Chiocci, F.L., 1994. Very high-resolution seismics as a tool for sequence stratigraphy applied to outcrop scale—examples from the eastern Tyrrhenian margin Holocene/Pleistocene deposits. AAPG Bull. 78, 378–395.

Chough, S.K., Hwang, I.G., Choe, M.Y., 1990. The Miocene Domusan fan-delta, southeast Korea: a composite fan-delta system in back-arc margin. J. Sediment. Petrol. 60, 445–455.

Clayton, C.A., Anis, M., Cooley, T.W., Troyer, W.H., Wallace, J.P., Cohen, M.F., Fadase, A.O., Pebdani, F.N., Odior, E.G., Ekworomadu, E.O., Nwaeri, E.U., Membere, A.H., 1998. Integrated reservoir management, Ubit field, offshore Nigeria. AAPG Bull. 82, 1901.

Coleman, J.M., 1988. Dynamic changes and processes in the Mississippi River delta. Geol. Soc. Am. Bull. 100, 999–1015.

Coleman, J.M., Prior, D.B., 1988. Mass wasting on continental margins. Annu. Rev. Earth Planet. Sci. 16, 101–119.

Coleman, J.M., Prior, D.B., Lindsay, J.F., 1983. Deltaic influences on shelf edge instability processes. In: Stanley, D.J., Moore, G.T. (Eds.), The Shelfbreak, Critical Interface on Continental Margins. SEPM Spec. Publ., vol. 33, pp. 121–137.

Collinson, J.D., Martinsen, O., Bakken, B., Kloster, A., 1991. Early fill of the Western Irish Namurian Basin: a complex relationship between turbidites and deltas. Basin Res. 3, 223–242.

Curray, J.R., Moore, D.G., 1964. Pleistocene deltaic progradation of continental terrace, Costa de Nayarit, Mexico. In: Van Straaten, L.M.J.U. (Ed.), Deltaic and Shallow Marine Deposits. Elsevier, Amsterdam, pp. 193–214.

Dalrymple, R.W., Zaitlin, B.A., Boyd, R., 1992. Estuarine facies models: conceptual basin and stratigraphic implications. J. Sediment. Petrol. 62, 1130–1146.

Droz, L., Rigaut, F., Cochonat, P., Tofani, R., 1996. Morphology and recent evolution of the Zaire turbidite system (Gulf of Guinea). Geol. Soc. Am. Bull. 108, 253–269.

Edwards, M.B., 1980. The Live Oak Delta complex: an unstable, shelf/edge delta in the deep Wilcox trend of south Texas. In: Meyes, S.D., Van, W.W., Bebout, D.G., Nault, M. (Eds.), Transactions of the 30th Annual Meeting of the Gulf Association of Geological Societies. AAPG Regional Meeting and the 27th Ann. Meeting of the Gulf Coast Section of SEPM., vol. 30, Gulf Coast Ass. Geol. Societies, New Orleans, pp. 71–73.

Edwards, M.B., 1981. Upper Wilcox Rosita delta system of South Texas. Growth-faulted shelf-edge deltas. AAPG Bull. 65, 54–73.

Elliot, T., 1986. Deltas. In: Reading, H.G. (Ed.), Sedimentary Environments and Facies. Blackwell, Oxford, pp. 113–154.

Elliot, T., 1989. Deltaic systems and their contribution to an understanding of basin-fill successions. In: Whateley, M.K.G., Pickering, K.T. (Eds.), Deltas: Sites and Traps for Fossil Fuels. Geol. Soc. Spec. Publ., vol. 41, pp. 3–10.

Elliot, T., Davies, S., 1996. High Resolution Sequence Stratigraphy of an Upper Carboniferous Basin-Fill Succession, County Clare, Western Ireland. Department of Earth Sciences, University of Liverpool. Guidebook.

Fairbanks, R.G., 1989. A 17,000-year glacio-eustatic sea level record: influence of glacial melting rates on the Younger Dryas event and deep ocean circulation. Nature 342, 637–642.

Farrán, M., Maldonado, A., 1990. The Ebro continental shelf: quaternary seismic stratigraphy and growth patterns. Mar. Geol. 95, 289–312.

Field, M.E., Gardner, J.V., Jennings, A.E., Edwards, B.D., 1982. Earthquake induced sediment failures on a 0.25° slope, Klamath River delta, California. Geology 10, 542–546.

Fisher, W.L., Brown, L.F., Scott, A.J., McGowen, J.H., 1969. Delta Systems in the Exploration for Oil and Gas. Bureau of Economic Geology, University of Texas, Austin, 78 pp.

Fisk, H.N., 1961. Bar-finger sands of the Mississippi Delta. In: Peterson, J.A., Osmond, J.C. (Eds.), Geometry of Sandstone Bodies—A Symposium. Am. Ass. Petrol. Geol., Tulsa, OK, pp. 29–52.

Frazier, D.E., 1967. Recent deltaic deposits of the Mississippi River: their development and chronology. Gulf Coast Ass. Geol. Soc. Trans. 31, 287–315.

Friedman, G.M., Sanders, J.E., 2000. Comments about the relationships between new ideas and geologic terms in stratigraphy and sequence stratigraphy with suggested modifications. AAPG Bull. 84, 1274–1280.

Friedman, G.M., Sanders, J.E., Kopaska-Merkel, D.C., 1992. Principles of Sedimentary Deposits—Stratigraphy and Sedimentology. Macmillan, New York, 717 pp.

Fulthorpe, C.S., Austin, J.A., 1998. Anatomy of rapid margin progradation: three-dimensional geometries of Miocene clinoforms, New Jersey margin. AAPG Bull. 82, 251–273.

Fulthorpe, C.S., Austin, J.A., Mountain, G.S., 1999. Buried fluvial channels off New Jersey: did sea-level lowstands expose the entire shelf during the Miocene? Geology 27, 203–206.

Galloway, W.E., 1974. Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. In: Brousard, M.L. (Ed.), Deltas, Models for Exploration Houston Geol. Soc., SEPM Foundation, Houston, TX, pp. 87–98.

Galloway, W.E., 1989. Genetic stratigraphic sequences in Basin Analysis II: application to northwest Gulf of Mexico Cenozoic Basin. AAPG Bull. 73, 143–154.

Galloway, W.E., 1990. Paleogene depositional episodes, genetic stratigraphic sequences, sediment accumulation rates NW Gulf of Mexico Basin. In: Armentrout, J.M., Perkins, B.F. (Eds.), Sequence Stratigraphy as an Exploration Tool: Concepts and Practices in the Gulf Coast. Soc. Econ. Paleontol. Mineral., Gulf Coast Section. 11th Annual Research Conference Program and Abstracts, pp. 165–176.

Gilbert, G.K., 1990. Lake Bonneville. U.S. Geol. Surv. Monogr. 1, 438 pp.

Haq, B.U., Hardenbol, J., Vail, P.R., 1987. Chronology of fluctuating sea levels since the Triassic. Science 235, 1156–1167.

Hart, B.S., Long, B.F., 1996. Forced regressions and lowstand deltas: Holocene Canadian examples. J. Sediment. Res. 66, 820–829.

Hart, B.S., Sibley, D.M., Flemings, P.B., 1997. Seismic stratigraphy, facies architecture, and reservoir character of a Pleistocene shelf-margin delta complex, Eugene Island Block 330 Field, offshore Louisiana. AAPG Bull. 81, 380–397.

Helland-Hansen, W., Gjelberg, J.G., 1994. Conceptual basis and variability in sequence stratigraphy: a different perspective. Sediment. Geol. 92, 31–52.

Helland-Hansen, W., Martinsen, O.J., 1996. Shoreline trajectories and sequences: description of variable depositional-dip scenarios. J. Sediment. Res. 66, 670–688.

Hernández-Molina, F.J., Somoza, L., Lobo, F., 2000. Seismic stratigraphy of the Gulf of Cádiz continental shelf: a model for Late Quaternary very high-resolution sequence stratigraphy and response to sea-level fall. In: Hunt, D., Gawthorpe, R.L. (Eds.), Sedimentary Responses to Forced Regressions. Geol. Soc. London, Spec. Publ., vol. 172, pp. 329–362.

Holmes, A., 1965. Principles of Pysical Geology. Thomas Nelson, London, 1288 pp.

Hovland, M., Dukefoss, K.M., 1981. A submerged beach between Norway and Ekofisk in the North Sea. Mar. Geol. 43, M19–M28.

Hunt, D., Tucker, M.E., 1992. Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall. Sediment. Geol. 81, 1–9.

Hunt, D., Tucker, M.E., 1995. Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall-reply. Sediment. Geol. 95, 147–160.

Kindinger, J.L., 1988. Seismic stratigraphy of the Mississippi–Alabama shelf and upper continental slope. Mar. Geol. 83, 79–94.

Kolla, V., 1993. Lowstand deep-water siliciclastic depositional systems: characteristics and terminologies in sequence stratigraphy and sedimentology. BCREDP 17, 67–78.

Kolla, V., Posamentier, H.W., Eichenseer, H., 1995. Stranded parasequences and the forced regressive wedge systems tract: deposition during base-level fall—discussion. Sediment. Geol. 95, 139–145.

Kolla, V., Biondi, P., Long, B., Fillon, R., 2000. Sequence stratigraphy and architecture of the Late Pleistocene Lagniappe delta complex, northeast Gulf of Mexico. In: Hunt, D., Gawthorpe, R.L. (Eds.), Sedimentary Responses to Forced Regressions. Geol. Soc. London, Spec. Publ., vol. 172, pp. 291–327.

Lehner, P., 1969. Salt Tectonics and Pleistocene stratigraphy on continental slope of northern Gulf of Mexico. AAPG Bull. 53, 2431–2479.

Martinsen, O., 1989. Styles of soft sediment deformation on a Namurian (Carboniferous) delta slope, western Ireland Namurian basin, Ireland. In: Whateley, M.K.G., Pickering, K.T. (Eds.), Deltas: Traps for Fossil Fuels. Geol. Soc. Spec. Publ., vol. 41, pp. 167–177.

Martinsen, O., Bakken, B., 1990. Extensional and compressional zones in slumps and slides—examples from the Namurian of County Clare, Ireland. J. Geol. Soc. London 146, 153–164.

Matteucci, T.D., Hine, A.C., 1987. Evolution of the Cape Fear Terrace: a complex interaction between the Gulf Stream and a paleo-shelf edge delta. Mar. Geol. 82, 185–205.

May, J.A., Warme, J.E., Slater, R.A., 1983. Role of submarine canyons on shelfbreak erosion and sedimentation. In: Wilgus, C.K., Hastings, C.K., Kendall, S.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea Level Changes: an Integrated Approach. SEPM Spec. Publ., vol. 42, pp. 315–332.

Mayall, M.J., Yeilding, C.A., Oldroyd, J.D., Pulham, A.J., Sakurai, S., 1992. Facies in a shelf-edge delta—an example from the subsurface of the Gulf of Mexico, middle Pliocene, Mississippi Canyon, Block 109. AAPG Bull. 76, 435–448.

McMaster, R.L., de Boer, J., Ashraf, A., 1970. Magnetic and seismic reflection studies on continental shelf off Portuguese Guinea, Guinea, and Sierra Leone, West Africa. AAPG Bull. 54, 158–167.

Mellere, D., Steel, R.J., 1995. Variability of lowstand wedges and their distinction from forced regressive wedges in the Mesaverde Group, southeast Wyoming. Geology 23, 803–806.

Mellere, D., Plink-Björklund, P., Steel, R., 2002. Anatomy of shelf-edge deltas on a prograding Eocene shelf margin, Spitsbergen. Sedimentology 49, 1101–1206.

Morgan, J.M. (Ed.), 1970. Deltaic Sedimentation, Modern and Ancient. SEPM Spec. Publ., vol. 15. 312 pp.

Morton, R.A., 1993. Attributes and origins of ancient submarine slides and filled embayments: examples from the Gulf Coast Basin. AAPG Bull. 77, 1064–1081.

Morton, R.A., Price, W.A., 1987. Late Quaternary sea-level fluctuations and sedimentary phases of the Texas coastal Plain and shelf: late Quaternary sea-level fluctuations and depositional sequences, Southwest Louisiana continental shelf. In: Nummedal, D., Pilkey, O.H., Howard, J.D. (Eds.), Sea-Level Fluctuation and Coastal Evolution. SEPM Spec. Publ., vol. 41, pp. 181–198.

Morton, R.A., Suter, J.R., 1996. Sequence stratigraphy and composition of Late Quaternary shelf-margin deltas, northern Gulf of Mexico. AAPG Bull. 80, 505–530.

Muto, T., Steel, R.J., 1992. Retreat of the front in a prograding delta. Geology 20, 697–970.

Muto, T., Steel, R.J., 2000. The accommodation concept in sequence stratigraphy: some dimensional problems and possible redefinition. Sediment. Geol. 130, 1–10.

Muto, T., Steel, R.J., 2002. In defence of shelf-edge delta development during falling and lowstand of relative sea level. J. Geol. (in press).

Nemec, W., Steel, R.J., Gjelberg, J., Collinson, J.D., Prestholm, E., Oxnevad, I.E., 1988. Anatomy of collapsed and re-established delta front in Lower Cretaceous of eastern Spitsbergen: gravitational sliding and sedimentation processes. AAPG Bull. 72, 454–476.

Nichol, S.L., Boyd, R., Penland, S., 1996. Sequence stratigraphy of a coastal-plain incised valley estuary: Lake Calcasieu, Louisiana. J. Sediment. Res. 66, 847–857.

Normark, W.R., Posamentier, H., Mutti, E., 1993. Turbidite systems: state of the art and future. Rev. Geophys. 31, 91–116.

Nummedal, D., 1983. Rates and frequencies of sea-level changes: a review with an application to predict futures sea level in Louisiana. Gulf Coast Ass. Geol. Soc. Trans. 33, 361–366.

Olsson, R.K., Miller, K.G., Browning, J.V., Wright, J.D., Cramer, B.S., 2002. Sequence stratigraphy and sea-level change across the Cretaceous–Tertiary boundary on the New Jersey passive margin. In: Koeberl, C., MacLeod, K.G. (Eds.), Catastrophic Events and Mass Extinctions: Impacts and Beyond. Geol. Soc. Am. Spec. Pap., vol. 356, pp. 97–108.

Orton, G.J., Reading, H.G., 1993. Variability of deltaic processes in terms of sediment supply, with particular emphasis on grain size. Sedimentology 40, 475–512.

Pegler, E.A., 1999. Mid- to late Quaternary environments and stratigraphy of the southern Sierra Leone shelf, West Africa. J. Geol. Soc. London 156, 977–990.

Penland, S., Boyd, R., Sutter, J.R., 1988. Transgressive depositional systems of the Mississippi delta plain: a model for barrier shoreline- and shelf sand development. J. Sediment. Petrol. 58, 932–949.

Plink-Björklund, P., Steel, R.J., 2002. Sea-level fall below the shelf edge, without basin-floor fans. Geology 30, 115–118.

Plink-Björklund, P., Mellere, D., Steel, R.J., 2001. Turbidite variability and architecture of sand-prone, deepwater slopes: Eocene clinoforms in the Central Basin, Spitsbergen. J. Sediment. Res. 71, 897–914.

Plint, A.G., 1988. Sharp-based shoreface sequences and "offshore bars" in the Cardium Formation of Alberta; their relationships to relative changes in sea level. In: Wilgus, C.K., Hastings, C.K., Kendall, S.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea Level Changes: an Integrated Approach. SEPM Spec. Publ., vol. 42, pp. 357–370.

Plint, A.G., Nummedal, D., 2000. The falling stage systems tract: recognition and importance in sequence stratigraphic analysis. In: Hunt, D., Gawthorpe, R.L. (Eds.), Sedimentary Responses to Forced Regressions. Geol. Soc. London, Spec. Publ., vol. 172, pp. 1–17.

Poag, C.W., Sevon, W.D., 1989. A record of Appalachian denudation in postrift Mesocoic and Cenozoic sedimentary deposits in the U.S. middle Atlantic continental margin. Geomorphology 2, 119–157.

Poag, C.W., Swift, B.A., Schlee, J.S., Ball, M.M., Sheetz, L.L., 1990. Early Cretaceous shelf-edge deltas of the Baltimore Canyon Trough: principal sources for sediment gravity deposits of the northern Hatteras Basin. Geology 18, 149–152.

Porębski, S.J., 1999. Depositional setting of a supra-evaporite succession (Upper Badenian) in the Kraków-Brzesko area, Carpathian Foredeep Basin [English Summary]. Prace Państwowego Instytutu Geologicznego 168, 97–118.

Porębski, S.J., Oszczypko, N., 1999. Lithofacies and origin of the Bogucice sands (Upper Badenian) [English Summary]. Prace Państwowego Instytutu Geologicznego 168, 57–82.

Porębski, S.J., Steel, R.J., 2001. Delta types and sea level cycle. 2001 AAPG Annual Convention June 3–6, 2001, Denver, Colorado. Official program, vol. 10, A160.

Porębski, S.J., Pietsch, K., Hodiak, R., Steel, R.J., 2002. Origin and sequential development of Upper Badenian–Sarmatian clinoforms in the Carpathian Foredeep Basin, SE Poland. Geol. Carpathica (in press).

Posamentier, H.W., 2001. Lowstand alluvial bypass systems: incised vs. unincised. AAPG Bull. 85, 1771–1793.

Posamentier, H.W., 2002. Ancient shelf ridges—a potentially significant component of the transgressive systems tract: case study from offshore northwest Java. AAPG Bull. 86, 75–106.

Posamentier, H.W., Allen, G.P., 1993. Variability of sequence stratigraphic model: effects of local basin factors. Sediment. Geol. 86, 91–109.

Posamentier, H.W., Allen, G.P., 1999. Siliciclastic sequence stratigraphy: concepts and applications. SEPM Concepts in Sedimentology and Paleontology, vol. 9, 210 pp.

Posamentier, H.W., Morris, W.R., 2000. Aspects of the stratal architecture of forced regressive deposits. In: Hunt, D., Gawthorpe, R.L. (Eds.), Sedimentary Responses to Forced Regressions. Geol. Soc. London, Spec. Publ., vol. 172, pp. 19–46.

Posamentier, H.W., Vail, P.R., 1988. Eustatic controls on clastic deposition: II. Sequence and systems tract models. In: Wilgus, C.K., Hastings, C.K., Kendall, S.G.St.C., Posamentier, H.W., Ross, C.A., Van Wagoner, J.C. (Eds.), Sea Level Changes: an Integrated Approach. SEPM Spec. Publ., vol. 42, pp. 125–154.

Posamentier, H.W., Erskine, R.D., Mitchum Jr., R.M., 1991. Models for submarine-fan deposition within sequence stratigraphic framework. In: Weimer, P., Link, M.H. (Eds.), Seismic Facies and Sedimentary Processes of Submarine Fans and Turbidite Systems. Springer, New York, pp. 127–136.

Posamentier, H.W., Allen, G.P., James, D.P., Tesson, M., 1992. Forced regressions in a sequence stratigraphic framework: concepts, examples, and exploration significance. AAPG Bull. 76, 1687–1709.

Postma, G., 1995. Causes of architectural variation in deltas. In: Oti, M.N., Postma, G. (Eds.), Geology of Deltas. A.A. Balkema, Rotterdam, pp. 3–16.

Prather, B.E., Booth, J.R., Steffens, G.S., Craig, P.A., 1998. Classification, lithologic calibration, and stratigraphic succession of seismic facies of intraslope basins, deep-water Gulf of Mexico. AAPG Bull. 82, 701–728.

Pratson, L.F., Coakley, B.J., 1996. A model for the headward erosion of submarine canyons induced by downslope eroding sediment flows. Geol. Soc. Am. Bull. 108, 225–234.

Pratson, L.F., Ryan, W.B.F., Mountain, G.S., Twichell, D.C., 1994. Submarine canyon initiation by downslope-eroding sediment flows: evidence in late Cenozoic strata on the New Jersey slope. Geol. Soc. Am. Bull. 106, 395–412.

Prior, D.B., Coleman, J.M., 1978. Disintegrating retrogressive landslides on very low-angle subaqueous slopes, Mississippi delta. Mar. Geotechnics 3, 37–60.

Pulham, A.J., 1989. Controls on interval structure and architecture of sandstone bodies within Upper Carboniferous fluvial-dominated deltas, County Clare, western Ireland. In: Whateley, M.K.G., Pickering, K.T. (Eds.), Deltas: Traps for Fossil Fuels. Geol. Soc. Spec. Publ., vol. 41, pp. 179–203.

Rabineau, M., Berné, S., Ledrezen, É., Lericolais, G., Marsset, T., Rotunno, M., 1998. 3D architecture of lowstand and transgressive Quaternary sand bodies on the outer shelf of the Gulf of Lion, France. Mar. Pet. Geol. 15, 439–452.

Reading, H., Collinson, J.D., 1996. Clastic coasts. In: Reading, H.G. (Ed.), Sedimentary Environments: Processes, Facies and Stratigraphy. Blackwell, Oxford, pp. 154–231.

Rodriguez, A.B., Fassell, M.L., Anderson, J.B., 2001. Variations in shoreface progradation and ravinement along the Texas coast, Gulf of Mexico. Sedimentology 48, 837–853.

Sangree, J.B., Waylett, D.C., Frazier, D.E., Amery, G.B., Fennessy, W.J., 1978. Recognition of continental slope seismic facies, offshore Texas–Louisiana. In: Bouma, A.H., Moore, G.T., Coleman, J.M. (Eds.), Framework, Facies, and Oil Trapping Characteristics of the Upper Continental Margins. AAPG Stud. Geol., vol. 7, pp. 87–116.

Schellpeper, M., 2000. Basin analysis of the Central Tertiary Basin and sequence stratigraphy of an Eocene clinoform in the Battfjellet Formation, Spitsbergen. MS thesis, University of Wyoming, 63 pp.

Scott, R.W., Combes, J.M., Nissen, S.E., 1998. High-precision chronostratigraphy of a Late Pleistocene shelf-edge delta, Louisiana. J. Sediment. Res. 68, 596–602.

Shackleton, N.J., 1987. Oxygen istotopes, ice volume and sea level. Quat. Sci. Rev. 6, 183–190.

Steel, R.J., Crabaugh, J., Schellpeper, M., Mellere, D., Plink-Björklund, P., Deibert, J., Loeseth, T., 2000. Deltas versus rivers on the shelf edge: their relative contributions to the growth of shelf margins and basin-floor fans (Barremian and Eocene, Spitsbergen). Proc. GCSSEPM Foundation 20th Ann. Res. Conf., Deepwater Reservoirs of the World, Houston, Dec. 2000., pp. 981–1000.

Suter, J.R., Berryhill Jr., H.L., 1985. Late Quaternary shelf-margin deltas, northwest Gulf of Mexico. AAPG Bull. 69, 77–91.

Suter, J.R., Berryhill Jr., H.L., Penland, S., 1987. Late Quaternary sea-level fluctuations and depositional sequences, southwest Louisiana continental shelf: late Quaternary sea-level fluctuations and depositional sequences, Southwest Louisiana continental shelf. In: Nummedal, D., Pilkey, O.H., Howard, J.D. (Eds.), Sea-Level Fluctuation and Coastal Evolution. SEPM Spec. Publ., vol. 41, pp. 199–219.

Sydow, J., Roberts, H.H., 1994. Stratigraphic framework of a Late Pleistocene shelf-edge delta, northeast Gulf of Mexico. AAPG Bull. 78, 1276–1312.

Sydow, J., Roberts, H.H., Bouma, A.H., Winn, R., 1992. Constructional subcomponents of a shelf-edge delta, northeast Gulf of Mexico. In: Schmitz, D.W., Swann, C.T., Walton, K. (Eds.), 42nd Ann. Meeting and Gulf Coast Section SEPM, 39th Ann. Meeting. Gulf Coast Ass. Geol. Soc. Trans., vol. 42, Gulf Coast Ass. Geol. Societies, New Orleans, pp. 717–726.

Talling, P.J., 1998. How and where do incised valleys form if sea level remains above the shelf edge? Geology 26, 87–90.

Tesson, M., Gensous, B., Allen, G.P., Ravenne, Ch., 1990. Late Quaternary deltaic lowstand wedges on the Rhône continental shelf, France. Mar. Geol. 91, 325–332.

Tesson, M., Allen, G.P., Ravenne, C., 1993. Late Pleistocene shelf-perched lowstand wedges on the Rhône continental shelf. In: Posamentier, H.W., Summerhayes, C.P., Haq, B.U., Allen, G.P. (Eds.), Sequence Stratigraphy and Facies Associations. Int. Assoc. Sedimentol., Spec. Publ., vol. 18, pp. 183–196.

Tesson, M., Posamentier, H.W., Gensous, B., 2000. Stratigraphic organization of Late Pleistocene deposits of the western part of the Golfe du Lion Shelf (Langedoc Shelf), western Mediterranean Sea, using high-resolution seismic and core data. AAPG Bull. 84, 119–150.

Thomas, M.A., Anderson, J.B., 1991. Late Pleistocene sequences stratigraphy of the Texas continental shelf relationships to O curves. GCS Soc. Econ. Paleont. Mineral. Foundation Ann. Res. Conference. Society of Economic Paleontologists and Mineralogists Foundation, Houston, TX, pp. 265–270.

Torres, J., Savoye, B., Cochnat, P., 1995. The effects of Late Quaternary sea-level changes on the Rhône slope sedimentation (north-western Mediterranean), as indicated by seismic stratigraphy. J. Sediment. Res. 65, 368–397.

Trincardi, F., Correggiari, A., 2000. Quaternary forced regression deposits in the Adriatic basin and the record of composite sea-level cycles. In: Hunt, D., Gawthorpe, R.L. (Eds.), Sedimentary Responses to Forced Regressions. Geol. Soc. London, Spec. Publ., vol. 172, pp. 245–269.

Trincardi, F., Field, M.E., 1991. Geometry, lateral variation, and preservation of downlapping regressive shelf deposits: eastern Tyrrhenian Sea margin, Italy. J. Sediment. Petrol. 61, 775–790.

Vail, P.R., 1987. Seismic stratigraphy interpretation procedure. In: Bally, B. (Ed.), Seismic Stratigraphic Atlas. AAPG Stud. Geol., vol. 27, pp. 1–10.

Vail, P.R., Todd, R.G., Sangree, J.B., 1977. Seismic stratigraphy and global changes of sea level, part. 5: chronostratigraphic significance of seismic reflectors. In: Payton, C.E. (Ed.), Seismic Stratigraphy—Application to Hydrocarbon Exploration. AAPG Mem., vol. 26, pp. 99–116.

Vail, P.R., Hardenbol, J., Todd, R.G., 1984. Jurassic unconformities, chronostratigraphy and seal-level changes from seismic stratigraphy and biostratigraphy. AAPG Mem. 36, 129–144.

Van Wagoner, J.C., 1995. Sequence stratigraphy and marine to non-marine facies architecture of foreland basin strata, Book Cliffs, Utah, U.S.A. In: Van Wagoner, J.C., Bertram, G.T. (Eds.), Sequence Stratigraphy of Foreland Basin Deposits. AAPG Mem., vol. 64, pp. 137–223.

Van Wagoner, J.C., Mitchum, R.M., Campion, K.M., Rahmani, V.D., 1990. Silicilastic Sequence Stratigraphy in Well Logs, Cores, and Outcrops. AAPG Methods Expl. Ser. 7, 155 pp.

Weimer, P., 1990. Sequence stratigraphy, facies geometrics, and depositional history of the Mississippi Fan, Gulf of Mexico. AAPG Bull. 74, 425–453.

Whateley, M.K.G., Pickering, K.T. (Eds.), 1989. Deltas: Sites and Traps for Fossil Fuels. Geol. Soc. Spec. Publ., vol. 41. 360 pp.

Wilson, R.C.L., 1998. Sequence stratigraphy: a revolution without a cause? In: Blundell, D.J., Scott, A.C. (Eds.), Lyell: The Past is the Key to the Present. Geol. Soc. London, Spec. Publ., vol. 143, pp. 303–314.

Winker, C.D., 1982. Cenozoic shelf margins, northwestern Gulf of Mexico. In: Morad, M.A., Kilgore, J.E., Cameron Jr., D.K. (Eds.), Trans. Gulf Coast Ass. Geol. Societies, 32nd Ann. Meeting 32. Gulf Coast Ass. Geol. Societies, New Orleans, pp. 427–448.

Winker, C.D., Edwards, M.B., 1983. Unstable progradational clastic shelf margins. In: Stanley, D.J., Moore, G.T. (Eds.), The Shelfbreak, Critical Interface on Continental Margins. SEPM Spec. Publ., vol. 33, pp. 139–157.

Winn Jr., R.D., Roberts, H.H., Kohl, B., Fillon, R.H., Bouma, A.H., Constans, R.E., 1995. Latest Quaternary deposition on the outer shelf, northern Gulf of Mexico: facies and sequence stratigraphy from Main Pass Block 303 shallow core. Geol. Soc. Am. Bull. 107, 851–866.

Winn Jr., R.D., Roberts, H.H., Kohl, B., Fillon, J.A., Crux, J.A., Bouma, A.H., Spero, H.W., 1998. Upper Quaternary strata of the upper continental slope, northeast Gulf of Mexico: sequence stratigraphic model for a terrigeneous shelf edge. J. Sediment. Res. 68, 579–595.

Xue, L., Galloway, W.L., 1995. High-resolution depositional framework of the Paleocene middle Wilcox strata, Texas coastal plain. AAPG Bull. 79, 205–230.