important gold placers in the Witwatersrand Basin, has been identified as such, and is associated with a regressive surface of marine erosion (Fig. 5.43).

Facies relationships across these unconformable surfaces are critical to establish their nature and sequence stratigraphic significance (Fig. 4.9), which in turn are important to evaluate the geographic distribution of the associated placers, as well as to predict changes in grades and placer quality along dip. In the case of subaerial unconformities, the level of reworking, and implicitly the textural maturity of the lag deposit, is proportional to the amount of base-level fall and downcutting. During this process, the finer sediment fractions are removed, allowing for a concentration of coarser clasts. As the amount of fluvial incision in response to base-level fall changes along dip, commonly decreasing in an upstream direction, the best reef quality tends to be found adjacent to the paleoshoreline. Such a reef not only loses quality upstream, but it also thins until it eventually disappears beyond the area of influence of base-level changes. Depending on the distance between paleoshoreline and the proximal rim of the basin, such placers may not have a physical expression along the basin margins, and may be missed if exploration is solely based on mapping basin margin unconformities. Similarly, shallow-marine forced regressive placers that overlie regressive surfaces of marine erosion, develop only offshore relative to the paleoshoreline at the onset of base-level fall, and may be missed where exploration is based solely on mapping basin margin unconformities.

The shoreline is therefore a central element in the exploration for placer deposits, because it limits the lateral development of all placer types. The subaerial unconformity-related placers may be found only landward relative to the end-of-fall paleoshoreline, whereas the regressive surface of marine erosion can form only seaward from the onset-of-fall paleoshoreline. Consequently, a successful exploration program must include paleogeographic reconstructions for successive time steps, upon which reliable sequence stratigraphic models may be built.

LOWSTAND SYSTEMS TRACT

Definition and Stacking Patterns

The lowstand systems tract, when defined as restricted to all sedimentary deposits accumulated during the stage of early-rise normal regression (*sensu* Hunt and Tucker, 1992), is bounded by the subaerial unconformity and its marine correlative conformity at the base, and by the maximum regressive surface at the top (Figs. 4.6, 5.4, and 5.5). Where the continental shelf is still partly submerged at the onset of base-level rise, following forced regression, the basal composite boundary of the lowstand systems tract may also include the youngest portion of the regressive surface of marine erosion (Fig. 5.6; also see Fig. 4.23, and the discussion in Chapter 4). The lowstand systems tract forms during the early stage of base-level rise when the rate of rise is outpaced by the sedimentation rate (case of normal regression; Figs. 4.5 and 4.6). Consequently, depositional processes and stacking patterns are dominated by low-rate aggradation and progradation across the entire sedimentary basin. As accommodation is made available by the rising base level, this 'lowstand wedge' is generally expected to include the entire suite of depositional systems, from fluvial to coastal, shallow-marine and deep-marine (Fig. 5.44).

Lowstand deposits typically consist of the coarsest sediment fraction of both nonmarine and shallow-marine sections, i.e., the lower part of a fining-upward profile in nonmarine strata, and the upper part of an upward-coarsening profile in a shallow-marine succession (Fig. 4.6). Sediment mass balance calculations indicate, however, that the grading trends observed within shallow-marine successions do not correlate with the grading trends that characterize the age-equivalent deep-water deposits (Fig. 5.11). Thus, preferential trapping of the coarser sediment fractions within aggrading fluvial and coastal to shallow-marine systems starting at the onset of base-level rise, reduces not only the net amount of sand supplied to the deep-water environment, but also the sand/mud ratio of the sediment load transported by turbidity currents. As a result, the lowstand sediments of the basin-floor submarine fan complex are overall finer-grained relative to the underlying late forced regressive deposits (Fig. 5.5). The maximum grain size of the sediment transported by gravity flows during the lowstand normal regression is also expected to decrease with time, due to the gradual lowering in fluvial slope gradients and related competence following the onset of base-level rise (Fig. 5.11). Consequently, in contrast to the high-density turbidity currents of the late stage of forced regression (Fig. 5.27), the deep-water portion of the lowstand systems tract is dominated by low-density turbidites (Fig. 5.44). The transition from high-density to low-density turbidites at the onset of base-level rise is illustrated in Fig. 5.37. Due to their lower sediment/water ratio, the low-density turbidity currents tend to be underloaded on the continental slope (high energy relative to sediment load), where channel entrenchment rather than aggradation is often recorded (Figs. 5.44 and 5.45). Beyond the toe of the

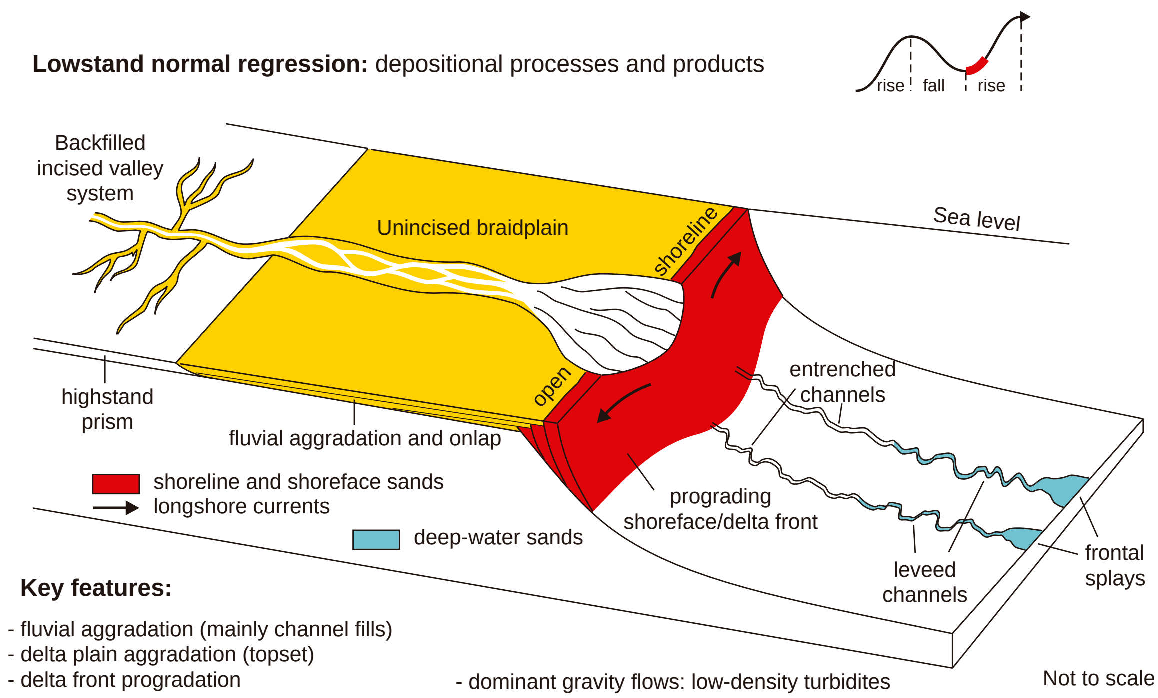


FIGURE 5.44 Depositional processes and products of the lowstand systems tract (modified from Catuneanu, 2003). In contrast to the falling-stage systems tract (Figs. 5.26 and 5.27), the sediment of this stage of early-rise normal regression is more evenly distributed between the fluvial, coastal, and deep-water systems. Sand is present in amalgamated fluvial channel fills, beach, and delta front systems, as well as in submarine fans. The 'lowstand prism' gradually expands landward *via* fluvial aggradation and onlap. Aggradation on the continental shelf in fluvial to shallow-marine environments reduces the amount of sediment supply to the deep basin, and hence the turbidity currents of this stage are dominantly of low-density type, being underloaded (entrenched) on the continental slope and aggradational only on the low-gradient basin floor where the energy of the flow drops below the threshold of balance with the sediment load. The top of all early-rise normal regressive deposits is marked by the maximum regressive surface.

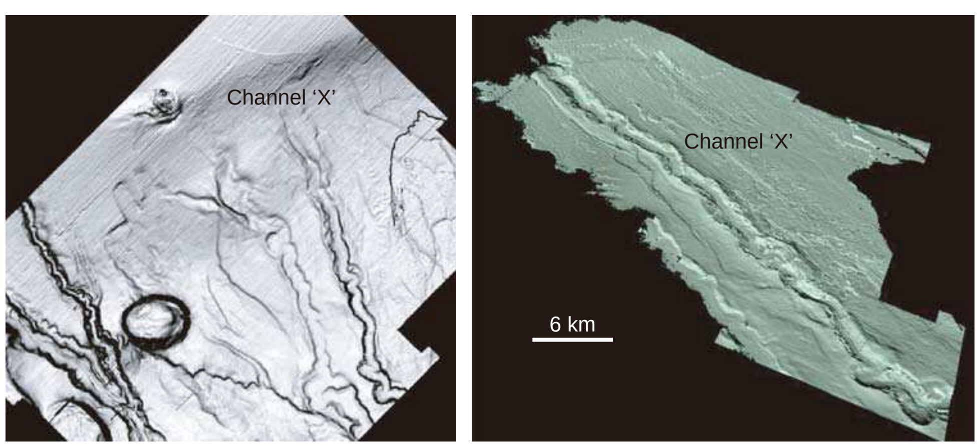


FIGURE 5.45 Entrenched turbidite channels on a continental slope (Late Pleistocene, De Soto Canyon area, Gulf of Mexico; images courtesy of H.W. Posamentier). Channel entrenchment indicates flow energy in excess of sediment load, which is most likely to occur in the case of low-density turbidity currents. Such currents commonly characterize the early stages of base-level rise (lowstand normal regression and early transgression). For scale, channel 'X' is approximately 1.8 km wide. Note that channel entrenchment on the continental slope (lowstand normal regression—early transgression) occurs after the aggradation of late forced regressive leveed channels. Consequently, relict levees may be preserved adjacent to the entrenched channels. These entrenched channels on the continental slope tend to become aggrading leveed channels on the basin floor (see text for details).

continental slope, on the basin floor, the lowstand turbidity currents may become overloaded as energy drops in response to decreasing seafloor gradients. As a result, the basin-floor setting is likely to record aggradation of leveed channels during the stage of lowstand normal regression (Posamentier and Kolla, 2003; Figs. 5.44 and 5.46–5.48). Such leveed channels are expected to develop further into the basin relative to the underlying leveed channels of the falling-stage systems tract because the muddier sediment of the low-density turbidity currents sustains the formation of levees over larger distances (compare the location of frontal splays in Figs. 5.27 and 5.44). More details on the depositional trends recorded in the deep-water environment follow below, as well as in Chapter 6.

Coastal aggradation during lowstand normal regression (i.e., relative increase in coastal elevation; Fig. 5.44) triggers a decrease in slope gradient in the downstream portion of fluvial systems (Fig. 5.6), which induces a lowering with time in fluvial energy and an overall upward decrease in grain size (Figs. 4.6 and 5.11). These decreases in fluvial energy and the grain size of the riverborne sediment that is made available at the shelf edge, explain the decrease in the maximum grain size of the sediment transported by gravity flows to the deep-water environment during the lowstand stage, as discussed above. The increase with time in the rate of base-level rise also contributes to the overall fining-upward fluvial profile, as it creates more accommodation for floodplain deposition and increases the ratio between floodplain and channel sedimentation (Fig. 5.11). Lowstand fluvial deposits typically accumulate on an uneven, immature topography, and contribute to the peneplanation of a nonmarine landscape sculptured by differential erosion during base-level fall (Fig. 5.12). Due to topographic irregularities at the stratigraphic level of the subaerial unconformity, the nonmarine portion of the lowstand systems tract may display a discontinuous geometry, with significant changes in thickness along dip and strike.

Typical examples of lowstand fluvial deposits include amalgamated channel fills (low accommodation systems) overlying subaerial unconformities (Fig. 5.49), which may accumulate within incised valleys (forming the entire valley fill or only the lower portion thereof; Figs. 5.12 and 5.25) or across areas formerly occupied by unincised, bypass falling-stage rivers (Fig. 5.44). The accumulation of lowstand fluvial sediments

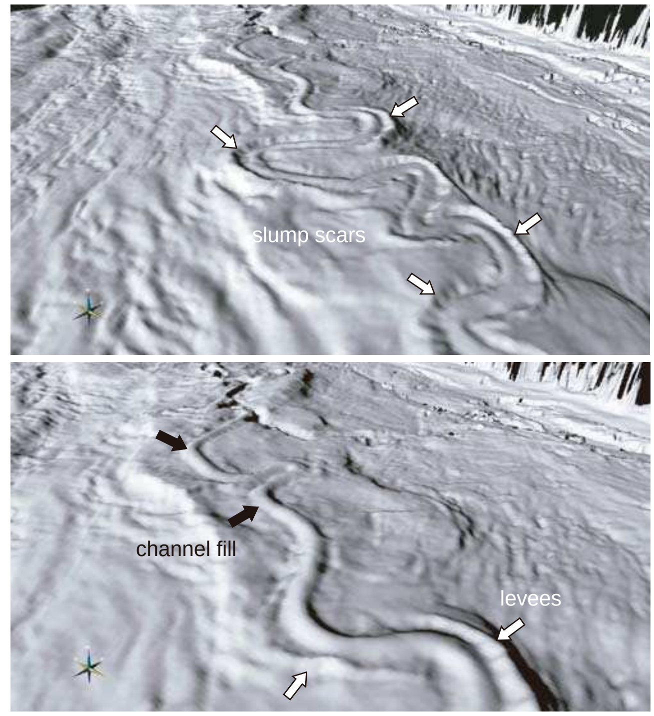


FIGURE 5.47 Depositional elements of a low-density turbidite leveed-channel system on a basin floor (details from Fig. 5.46; images courtesy of H.W. Posamentier). The raised appearance, with a convex-up top, of the channel fill is caused by postdepositional differential compaction. Levees are better developed along the outer channel bends, and their inner margins are characterized by the presence of scoop-shaped slump scars. The relief of the channel belt above the adjacent basin plain is approximately 65 m. The channel fill is approximately 625 m wide.

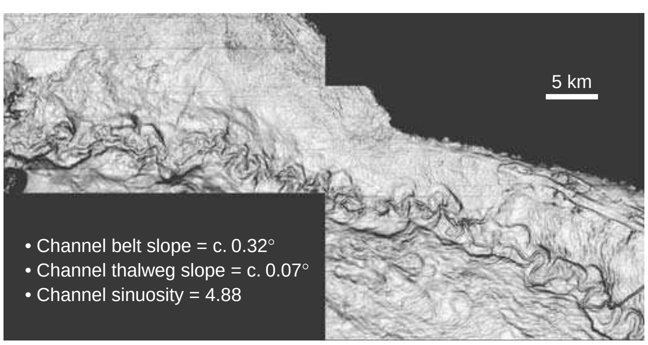


FIGURE 5.46 Aggrading turbidite leveed channel on a basin floor (Late Pleistocene, De Soto Canyon area, Gulf of Mexico; modified from Posamentier, 2003; image courtesy of H.W. Posamentier). Well developed basin-floor leveed channels are typical of low-density turbidity flows, whose mud content is high enough to sustain the formation of levees over large distances. Such basin-floor leveed channels are commonly age-equivalent with entrenched channels on the continental slope (Fig. 5.45). The change in the character of syndepositional processes from erosional on the slope to aggradational on the basin floor relates to changes in slope gradients and associated energy of the flow. Low-density turbidity currents tend to be underloaded on the continental slope (insufficient sediment load relative to energy), but they become overloaded on the basin floor where the energy of the flow decreases significantly. This is in contrast with the high-density turbidity currents, whose larger sediment load allows them to aggrade even on the steeper-gradient continental slope (Figs. 5.39–5.42).

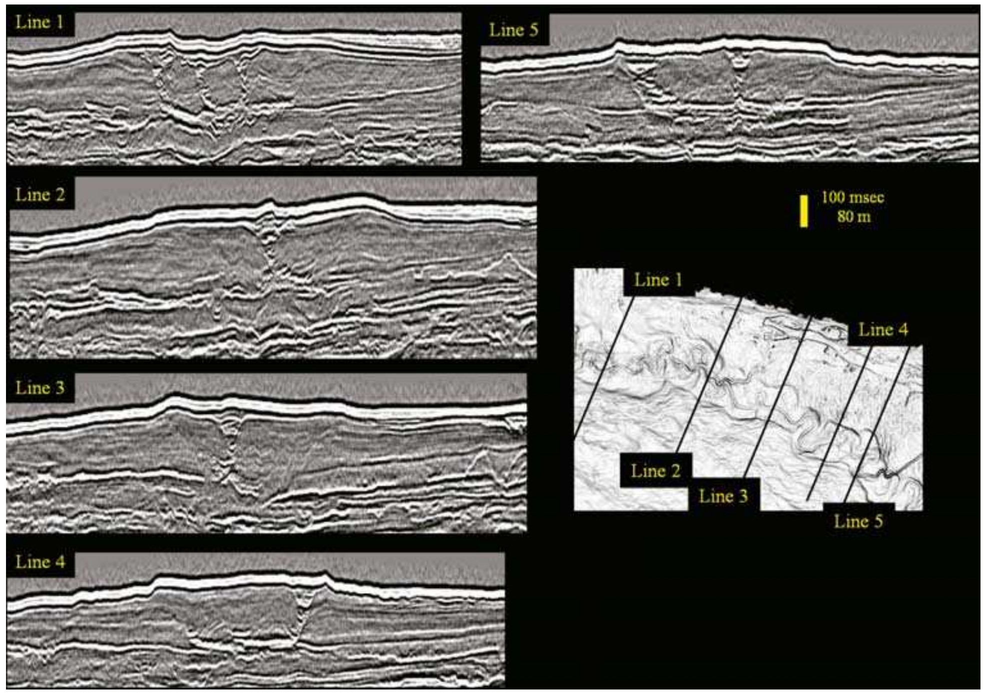


FIGURE 5.48 Transverse sections through the leveed channel in Fig. 5.46 (images courtesy of H.W. Posamentier). The 2D seismic lines indicate channel aggradation, as well as lateral migration with time. Note that the sandy channel fill is characterized by higher amplitude seismic reflections relative to the surrounding finer-grained facies of the overbank environment. Levees are also built by finer-grained material relative to the channel fill. For scale, the channel fill is approximately 625 m wide.

begins within topographic lows, and it is commonly assumed that incised valleys are at least in part filled with such deposits (e.g., models developed by Shanley and McCabe, 1991, 1993, 1994; Wright and Marriott, 1993; Gibling and Bird, 1994). There are also cases, however, where the lowstand fluvial deposits are missing from the stratigraphic architecture of incised-valley fills, due to either nondeposition or erosion during subsequent transgression. In such cases, the fluvially-cut surface at the base of the incised valley is modified into a transgressive surface of erosion, and the incised valley may be entirely filled by transgressive systems tract deposits (e.g., Dalrymple *et al.*, 1992; Ainsworth and Walker, 1994).

Within the lowstand successions of amalgamated channel fills, paleosols may be present, reflecting syndepositional conditions of limited accommodation on floodplains (Fig. 2.17). Such paleosols are often 'wet' and immature, as being formed during stages of base-level rise, and may be associated with poorly developed, if any, coal seams (Fig. 2.17). The poor development of coal seams within the lowstand systems tract is explained by the low rates of creation of accommodation coupled with the generally high influx of clastic sediment.

The inclusion of fluvial deposits under analysis within any conventional systems tract (e.g., 'lowstand' in this case; Fig. 5.49) implies a particular type of *shoreline shift* during the accumulation of fluvial facies (e.g., lowstand normal regression in this case). The documentation of a process/response relationship between contemporaneous fluvial and marine environments is therefore important to justify the usage of the standard (lowstand – transgressive – highstand) systems tract nomenclature. Figure 5.50 provides examples of sedimentary structures that document the manifestation of marine influences on fluvial processes at the time of sedimentation. In the absence of such evidence, where no relationship can be established between fluvial processes and any shifts of a coeval shoreline, the concepts of low- and high-accommodation systems tracts may provide a more realistic approach to describing fluvial deposits in a sequence stratigraphic framework. This is generally the case in overfilled basins or in areas of sedimentary basins that are beyond the influence of marine base-level changes (e.g., zone 3 in Fig. 3.3). The low- and high-accommodation systems tracts are discussed in more detailed in a subsequent section of this chapter.

Lowstand fluvial strata are commonly depicted as the product of sedimentation within high-energy rivers of braided type, due to the steepening of the landscape gradient that is generally expected during stages of base-level fall. While this may often be the case (e.g., Figs. 4.32 and 4.38), one should not exclude the possibility of meandering lowstand systems, which are particularly prone to develop within incised meander belts (Figs. 5.12 and 5.19). Lowstand rivers of

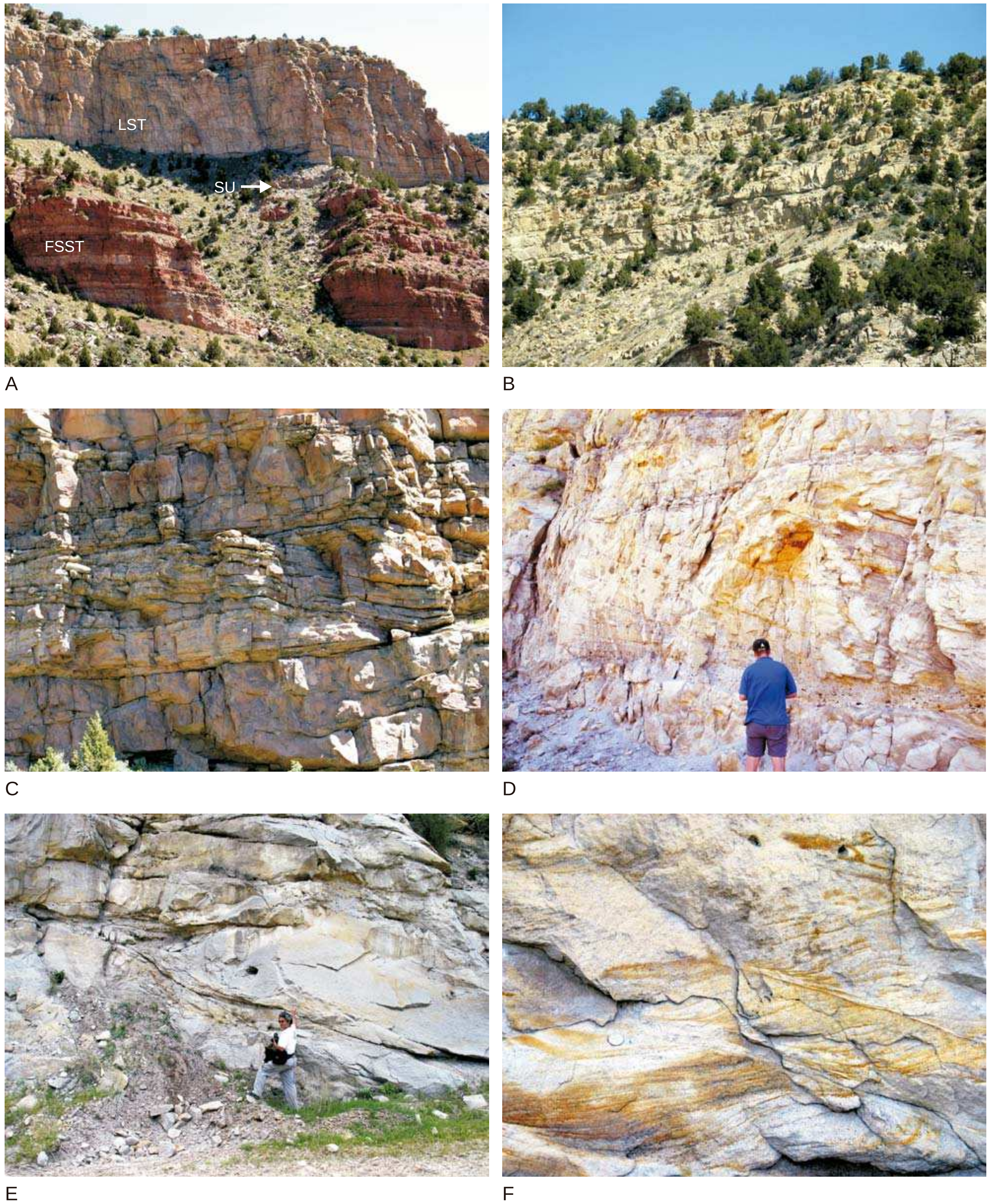
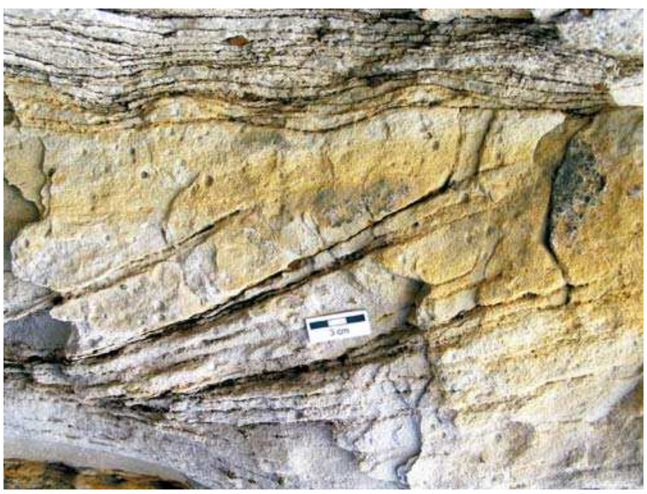


FIGURE 5.49 Outcrop expression of lowstand fluvial systems (Castlegate Formation, Utah). A—amalgamated channel fills (upper part of the photograph, lighter color: Castlegate Formation) separated from the underlying forced regressive shoreface deposits (lower part of the photograph, reddish color: Blackhawk Formation) by the subaerial unconformity; B, C, D, E—amalgamated channel fills of braided fluvial systems; F—climbing dunes indicating high sediment supply in the high-energy braided streams. Abbreviations: FSST—falling-stage systems tract; LST—lowstand systems tract; SU—subaerial unconformity.

A



B

FIGURE 5.50 Tidal influences in lowstand fluvial systems, indicating proximity to the coeval shoreline (Castlegate Formation, Utah). A—mud drapes associated with cross-bedding; B—sigmoidal bedding, mud drapes, and worm burrows. Evidence of marine influences suggests that fluvial processes are at least in part controlled by marine base-level changes; therefore, the fluvial deposits under analysis accumulated in zone 2 in Fig. 3.3. The process/response relationship between contemporaneous fluvial and marine environments justifies the usage of the standard systems tract nomenclature ('lowstand' in this case), which makes direct reference to syndepositional shoreline shifts. If fluvial systems accumulate beyond the reach of marine influences (e.g., in overfilled basins, or within zone 3 in Fig. 3.3), the usage of the lowstand—transgressive—highstand systems tract nomenclature becomes redundant, and the concepts of low- *vs.* high-accommodation systems tracts are more appropriate (see discussion below on low- and high-accommodation systems tracts).

braided type are most likely to establish themselves in areas formerly occupied by bypass falling-stage systems, where the channel pattern is not constrained by incised valleys that tend to preserve the plan-view morphology of low-energy highstand rivers. In such cases, where the flow is not constrained by an erosional landscape, rivers may easily adjust their channel pattern (e.g., from meandering to braided) to reflect the change in energy levels from highstand to lowstand conditions. In contrast, incised valleys tend to retain the meander pattern inherited from highstand rivers, and impose that pattern on the younger and higher energy lowstand systems (Fig. 5.20; see also discussion in the section dealing with the falling-stage systems tract).

Following the stage of base-level fall, when most of the shelf becomes subaerially exposed, the lowstand systems tract may include shelf-edge deltas with diagnostic topset geometries (Fig. 5.4). Triggered by the rise in base level at the shoreline, the aggradation of lowstand fluvial strata starts from the delta plain area and gradually extends upstream by onlapping the subaerial unconformity (Figs. 5.4 and 5.5). This trend of fluvial onlap widens the stratigraphic hiatus that is associated with the subaerial unconformity in a landward direction, as the age of the overlying fluvial strata is increasingly younger upstream (Fig. 5.5). This scenario is valid under the assumption that rivers flow along graded profiles upstream of the point of fluvial onlap, which allows for a simple modelling that only takes into account the downstream controls (i.e., base-level changes) on fluvial processes. Under this assumption, the lowstand fluvial deposits that accumulate due to base-level rise form a wedge-shaped prism that thins upstream towards the youngest point of fluvial onlap (Fig. 5.4). The distance along dip that is subject to fluvial onlap depends on several factors, including the duration of the lowstand stage, the amount of sediment supply and the rates of coastal aggradation, and the topographic gradients of the land surface. A flat topography (e.g., in a low-gradient shelf-type setting) coupled with high sediment supply are conducive to fluvial aggradation over a large area, whereas a steep topography (e.g., in a high-gradient ramp setting, such as a continental slope or a fault-bounded basin margin) coupled with low sediment supply restrict the size of the area that is subject to fluvial aggradation (Blum and Tornqvist, 2000). In the latter case, the subaerial unconformity may be overlain directly by transgressive fluvial strata over much of its extent (Embry, 1995; Dalrymple, 1999). Data from the Gulf Coastal Plain of the U.S.A. indicate that the landward limit of fluvial onlap correlates to the amount of sediment supply, and is inversely proportional to the gradient of the onlapped floodplain surface. This distance may vary significantly, from approximately 40 km in the case of the steep-gradient, low-sediment-supply Nueces River to at least 300–400 km for the low-gradient, high-sediment-supply Mississippi River (Blum and Tornqvist, 2000).

The preservation potential of coastal and adjacent lowstand fluvial strata may be low due to subsequent transgressive ravinement erosion (Fig. 5.6). Figure 4.53 illustrates an example where no lowstand fluvial strata are preserved on top of a former subaerial unconformity, which was subsequently reworked by a transgressive tidal-ravinement surface. Where central estuary facies are preserved, they may act as a buffer that protects the underlying fluvial strata from subsequent transgressive scouring. In such cases, fluvial lowstand deposits are likely to be preserved between the subaerial unconformity and the earliest estuarine strata whose timing marks the onset of transgression (Fig. 4.32). The contact between lowstand fluvial and the overlying estuarine facies is the maximum regressive surface (Fig. 5.6). This stratigraphic contact tends to be sharp, because of the rapid development of the estuarine system as soon as the shoreline starts its landward shift (Figs. 4.32 and 4.39). This sequence stratigraphic surface should not be confused with the facies contact between transgressive fluvial facies and the overlying estuarine strata, which develops *within* the transgressive systems tract (Fig. 5.6). The latter facies shift is gradational, with significant interfingering between tidally-influenced fluvial and estuarine deposits, and is highly diachronous with the rate of shoreline transgression. The differentiation between lowstand fluvial and transgressive fluvial facies based solely on well logs may be difficult, unless a more regional understanding of the study area is achieved. Such a distinction is greatly facilitated where core and/or outcrop data are available to allow the observation of sedimentary structures associated with tidal influences, which are much more abundant within the downstream reaches of transgressive fluvial systems.

Seaward from the coastline, the shoreface deposits of the lowstand systems tract are generally gradationally based, excepting for the earliest lowstand shoreface strata which may be sharp-based, as they overlie the regressive surface of marine erosion (Fig. 5.6). Beyond the fairweather wave base, the extent of shelf facies may be limited due to the potential proximity of the shoreline to the shelf edge at the end of forced regression (Fig. 5.4). In this case, the subtidal facies may pass directly into deep-water slope facies, which consist primarily of gravity-flow deposits (Figs. 4.15 and 5.4).

In contrast to the trends discussed in the case of highstand normal regression, the seaward shift of the lowstand shoreline *decelerates* during the lowstand stage because the rates of base-level rise increase with time, from zero until the turnaround from regression to transgression is achieved (Fig. 3.19). As a result, increasingly more sediment is required to fill the newly created accommodation at the shoreline, and so the lowstand normal regressive deltaic lobes ('parasequences') become thicker with time and in an offshore direction (Fig. 5.13). If a portion of the continental shelf is still submerged at the end of forced regression, and as accommodation is limited during early lowstand, the oldest coastal to subtidal sandstones of the lowstand systems tract tend to have a wider geographic distribution across the shelf due to rapid autocyclic shifting imposed upon deltaic lobes (Fig. 5.13). Such shallow-marine reservoirs are expected to have a better connectivity relative to their late lowstand counterparts, which display a more pronounced aggradational component (i.e., vertical rather than lateral stacking). In parallel to these trends of change in the thickness and stacking patterns of the deltaic lobes that accumulate during lowstand normal regression, the average grain size of successive delta lobes is also expected to decrease in a seaward direction in response to the lowering with time in fluvial energy during base-level rise (Fig. 5.13). The latter trend of change in average grain size along dip parallels the one observed in the case of highstand deltas, and is the opposite of what characterizes forced regressive deltas (Fig. 5.13). Irrespective of these trends of change in grain size from older to younger deltaic lobes along dip, vertical profiles in any particular location show invariably a reverse grading (coarsening-upward) due to the progradation of delta front facies over finer prodelta sediments (Fig. 5.11).

Economic Potential

Petroleum Plays

Rising base level during the lowstand normal regression provides accommodation across the entire basin, from fluvial to marine environments. Sediment budget observations indicate a concentration of the coarsest riverborne sediment within fluvial and coastal depositional systems, which arguably form the best reservoirs, with the highest sand/mud ratio, of the lowstand systems tract. The trapping of sand within aggrading fluvial to shallow-marine systems following the onset of base-level rise results in a net decrease in the volume of sediment available for deep-water gravity flows, and also in a lowering of the sand/mud ratio in submarine fans (Figs. 4.18 and 5.11). Shelf-edge deltas and correlative strandplains continue to prograde the upper slope, with the development of a topset in response to coastal aggradation (Figs. 3.22 and 5.4). Increased elevation at the shoreline triggers fluvial aggradation, starting from the shoreline and gradually expanding upstream (fluvial onlap), which explains the wedging out of lowstand fluvial reservoirs toward the basin margins (Figs. 5.4 and 5.44).

The petroleum plays of the lowstand systems tract are therefore diverse in terms of origin and syndepositional processes, ranging from fluvial to coastal, shallow- and deep-marine systems. In this way, the entire dip profile of the basin offers exploration opportunities within this systems tract. A key for recognizing the 'lowstand wedge' on seismic lines is the presence of a topset (rather than offlap and truncation, typical of forced regressions) associated with the shelf-edge deltas and correlative strandplains (Figs. 3.22 and 5.44). Landward from the shelf edge, fluvial reservoirs are mainly represented by amalgamated channel fills, due to the low amounts of accommodation available during this stage. These are the best fluvial reservoirs of the entire base-level cycle, with the highest sand/mud ratio. The shelf-edge deltas that prograde the upper continental slope, and their open shoreline beach and shoreface correlative systems, also trap a significant amount of sand and may form good reservoirs that are laterally extensive along strike (Fig. 5.44). These normal regressive shelf-edge reservoirs are often topped by high amplitude reflectors on seismic lines, marking the change in acoustic impedance from the transgressive and highstand shales above to the underlying lowstand sand-rich reservoirs (Fig. 3.22). Beyond the shelf-edge deltas and their correlative strandplains, riverborne sediment continues to be delivered to the deeper basin but in decreasing amounts and with a decreasing sand/mud ratio in response to the increasing rates of base-level rise at the shoreline (Fig. 5.11). As more and more sand is trapped in aggrading fluvial and coastal systems, the submarine fan receives less and less sand relative to the pelagic fallout, which generates the overall fining-upward trend noted for the slope fans in Fig. 4.18 above the correlative conformity. This fining-upward trend is also shown in Figs. 5.5 and 5.11.

The significance of the lowstand systems tract, and the various portions thereof, in terms of sediment budget, potential petroleum reservoirs, and petroleum source and seal rocks is summarized in Fig. 5.14. As suggested in this table, the lowstand systems tract tends to be the most balanced among all systems tracts in terms of the relatively even distribution of reservoirs across the basin. The lowstand fluvial deposits (amalgamated channel fills), where preserved from subsequent transgressive scouring, form the best reservoirs of the entire fluvial portion of a stratigraphic sequence. Equally good reservoirs may form in coastal, shallow-water and deep-water environments during the lowstand normal regression of the shoreline (Fig. 5.14). The change in the type of dominant gravity flows that manifest during the lowstand time, from high-density turbidity currents at the end of base-level fall/onset of base-level rise to low-density turbidity currents, has important consequences for the lithology, morphology and location of deep-water reservoirs within the basin, as discussed above (Figs. 5.11, 5.27, 5.37, and 5.44). These issues are explored further in Chapter 6.

The main risks for the exploration of lowstand reservoirs relate to charge, seals, and source rocks, especially toward the basin margins. Even within a shelf setting, however, fluvial, coastal, and shallow-water lowstand reservoirs may be sealed by overlying transgressive shales, which may be fluvial, estuarine or shallow-water in nature (Fig. 5.14). The risks of exploration of lowstand reservoirs decrease toward the deep portion of the basin, as lowstand turbidites, which travel farther into the basin relative to the falling-stage gravity flows, stand a good chance of being in direct contact with transgressive/highstand source and seal facies both below and above.

Coal Resources

The lowstand systems tract is defined by high sediment supply in an overall low accommodation setting, and therefore environmental conditions are generally unfavourable for peat accumulation (Fig. 5.15). The architecture of lowstand fluvial systems is commonly described by amalgamated channel fills, partly because of the lack of sufficient accommodation, and partly due to the tendency of lowstand rivers to be of higher energy following the steepening of the topographic profile as a result of tilt and/or differential erosion during the stage of base-level fall (e.g., Catuneanu, 2004a; Figs. 4.32 and 4.38). The limited amount of fluvial accommodation affects particularly the overbank environment, which may also be subject to scouring by laterally shifting fluvial channels, and therefore no significant coal deposits are generally associated with the lowstand systems tract. As the rates of base-level rise increase with time during the lowstand stage, gradually more accommodation becomes available to the overbank environment, and so chances of peat accumulation and subsequent coal development tend to improve toward the top of the lowstand systems tract (Fig. 5.15).

Placer Deposits

No unconformities form *during* the lowstand normal regression, but the lowstand systems tract is closely associated with all three types of unconformity-related placer deposits. The subaerial unconformity and the youngest portion of the regressive surface of marine erosion are found at the base of the lowstand systems tract, whereas the oldest portion of the transgressive ravinement surface commonly truncates the top of lowstand deposits (Fig. 5.6). The first two unconformity-related placer types are described in more detail in the section dealing with the falling-stage systems tract; the placers associated with transgressive ravinement surfaces are discussed in the following section.

Following the end of base-level fall, the accumulation of coarse sediments may continue in amalgamated channels during the early stage of lowstand normal regression. These lowstand deposits are particularly prone to 'reef' facies development in the case of gravel-bed fluvial systems. Such 'depositional' reefs (as opposed to unconformity-related reefs) involve only limited reworking of the underlying sediments, and so they may be of economic significance especially where the mineralization is the result of precipitation from hydrothermal fluids. Where conditions are favorable for the aggradation of coarse-grained clasts following the onset of base-level rise, the depositional lowstand reefs add to the thickness of the underlying placers represented by lag deposits associated with subaerial unconformities or regressive surfaces of marine erosion.

TRANSGRESSIVE SYSTEMS TRACT

Definition and Stacking Patterns

The transgressive systems tract is bounded by the maximum regressive surface at the base, and by the maximum flooding surface at the top. This systems tract forms during the stage of base-level rise when the rates of rise outpace the sedimentation rates at the shoreline (Figs. 3.19, 4.5, and 4.6). It can be recognized from the diagnostic retrogradational stacking patterns, which result in overall fining-upward profiles within both marine and nonmarine successions (Figs. 4.6, 5.5, and 5.11). As the rates of creation of accommodation are highest during shoreline transgression (Fig. 4.5), the transgressive systems tract is commonly expected to include the entire range of depositional systems along the dip of a sedimentary basin, from fluvial to coastal, shallow-marine and deep-marine (Fig. 5.4).

The transgressive fluvial and coastal deposits may potentially be thick, due to the high sedimentation rates stimulated by the available accommodation, although exceptions do occur under particular circumstances (Fig. 3.20; see also discussion below). The trapping of large amounts of terrigenous sediment within aggrading fluvial and coastal systems during transgression results in a cut-off of sediment supply to the marine environment (Loutit *et al*., 1988). As a consequence, transgressive shallow-marine deposits accumulate primarily in areas adjacent to the shoreline, with correlative condensed sections or even unconformities in the more distal portions of the shelf (Galloway, 1989). Triggered by the lack of sediment supply and a regime of hydraulic instability during rapid base-level rise, the shelf edge region is generally subject to non-deposition and/or sediment reworking during transgression (Fig. 5.4). As a result, the transgressive systems tract tends to be composed of two distinct wedges separated by an area of non-deposition around the shelf edge, one on the continental shelf consisting of fluvial to shallow-marine deposits, and one in the deep-water setting consisting of gravity-flow deposits and pelagic sediments (Figs. 4.41 and 5.4). Both these wedges shift toward the basin margin during transgression, following the general retrogradational trend, by onlapping the landscape and the seascape, respectively, in a landward direction (Fig. 5.5). The gradual expansion of the transgressive depozone in a continental shelf-type setting is associated with *fluvial onlap* (leading edge of the transgressive wedge). Within the deep-water portion of the basin, the transgressive deposits are often seen onlapping the continental slope, forming a transgressive slope apron associated with *marine onlap* (Galloway, 1989; Figs. 4.2, 4.3, 4.41, 5.4, and 5.5). In addition to the fluvial and marine onlaps that characterize the leading edges of the two transgressive wedges, *coastal onlap* is also an important type of stratal termination, diagnostic for transgression, forming within the continental shelf-based transgressive wedge by the shift of shoreface facies on top of the landward-expanding wave-ravinement surface (Figs. 4.2, 4.3, 5.4, and 5.5).

The fluvial portion of the transgressive systems tract commonly shows evidence of tidal influences (Shanley *et al*., 1992; Shanley and McCabe, 1993), and is characterized by an overall fining-upward vertical profile (Fig. 4.6). This overall grading trend reflects both an upward decrease in maximum grain size caused by a decline with time in the competence of the rivers, and also a lowering of the sand/mud ratio (channel *vs.* overbank sedimentation) in response to accelerating base-level rise following the lowstand normal regression (Fig. 5.11). The latter feature of the vertical profile also translates into an upward decrease in the degree of amalgamation of transgressive channel-fill sandstones, which are often described as isolated ribbons engulfed within floodplain fines (Shanley and McCabe, 1993; Wright and Marriott, 1993). The decline with time in the energy of transgressive fluvial systems parallels a corresponding decrease in topographic gradients, which in turn is triggered by coastal aggradation coupled with the general pattern of fluvial sedimentation during base-level rise. As in the case of lowstand and highstand normal regressions, the sedimentation of fluvial deposits during transgression *in response to base-level rise* starts from the downstream reaches of rivers, where the fluvial succession is thickest, gradually expanding upstream (Fig. 5.5). This pattern of fluvial onlap explains the wedge-shaped geometry of the fluvial transgressive package, which thins landward from the shoreline, leading to the observed decrease in topographic gradients and fluvial energy during transgression (Fig. 5.4). Following this style of fluvial sedimentation established at the onset of base-level rise, the transgressive fluvial deposits often extend farther toward the basin margins relative to the underlying lowstand fluvial strata, by onlapping the subaerial unconformity (see fluvial onlap in Figs. 5.4 and 5.5). Such predictable trends could, however, be altered if fluvial processes are influenced by controls other than base-level changes, notably by climate and/or source area tectonism. As accommodation is generated rapidly during transgression, and the water table rises in parallel with the base level, the fluvial portion of the transgressive systems tract often includes well developed coal seams (Fig. 4.42).

The transgressive fluvial deposits may form a significant portion of incised-valley fills, or may aggrade in the interfluve areas of former incised valleys. Where incised valleys are inherited from previous stages of base-level fall and are not entirely filled by lowstand deposits, their downstream portions are commonly converted into estuaries at the onset of transgression (Dalrymple *et al*., 1994). In such cases, the lowstand fluvial deposits that overlie the subaerial unconformity may be scoured, or partly reworked, by estuarine channels and tidal-ravinement surfaces (Rahmani, 1988; Allen and Posamentier, 1993; Ainsworth and Walker, 1994; Breyer, 1995; Rossetti, 1998; Cotter and Driese, 1998). Where not reworked by the tidal-ravinement surface, the contact between lowstand fluvial and the earliest (stratigraphically lowest) overlying estuarine facies is represented by the maximum regressive surface. In this setting, the maximum regressive surface is relatively easy to map in outcrop or core, at the abrupt change from coarse fluvial sand and gravel (lowstand deposits) to the overlying estuarine facies comprising finer-grained and more varied lithologies with abundant tidal structures such as clay drapes and flasers (see Allen and Posamentier, 1993, for the case study of the Holocene Gironde incised valley in southwestern France; Fig. 4.52). This contrast between lowstand fluvial and overlying transgressive estuarine facies may also be strong enough to be seen in well logs, at the contact between 'clean' and blocky sand and the younger, more interbedded and finer-grained lithologies (Fig. 4.32).

In coastal settings, the transgressive systems tract may include backstepping foreshore (beach) deposits, diagnostic estuarine facies (particularly in the case of smaller rivers), and even proper deltas in the case of large rivers (Figs. 5.51 and 5.52). The formation and preservation of transgressive coastal deposits depends on the rates of base-level rise, sediment supply, the wind regime and the amount of associated wave-ravinement erosion, and the topographic gradients at the shoreline. Coastal aggradation is favoured by high rates of base-level rise, weak transgressive ravinement erosion, and shallow topographic gradients (e.g., in low-gradient shelf-type settings; Fig. 5.6). Steeper topographic gradients (e.g., in high-gradient ramp settings) tend to induce coastal erosion in relation to a combination of factors including higher fluvial energy, wave ravinement, and slope instability (Fig. 3.20). This may explain the common lack of estuarine facies in fault-bounded basins, but also in areas characterized by extreme wind energy and associated strong wave-ravinement erosion (Leckie, 1994).

In the case of erosional coastlines, where transgressive coastal facies are not preserved in the rock record, transgressive fluvial deposits are likely to be missing as well (Fig. 3.20). In this case, the coastal to nonmarine portion of the transgressive systems tract is replaced by a subaerial unconformity with an associated hiatus that is age-equivalent with the marine transgressive deposits. A modern analog is represented by the incised estuaries and fluvial systems of the Canterbury Plains, New Zealand, where the transgressive coastline is dominated by erosional processes (Figs. 3.24–3.26).

In the case of aggrading coastlines (Figs. 3.20 and 5.6), both coastal and fluvial deposits have a high preservation potential. The character of the coastline may change along strike from transgressive to normal regressive as a function of the shifting balance between the rates of base-level rise and the rates of sedimentation in *open shoreline* settings (Fig. 5.52). As such, prograding strandplains are typical of normal regressive coastlines, whereas backstepping beaches define transgressive coastlines (Fig. 5.52). The boundary between coeval transgressive and normal regressive coastlines in Fig. 5.52 may either be constrained by spatial variations in sedimentation rates or by strike variability in subsidence rates, or both. The mechanisms controlling the change in depositional trends along a coastline have been investigated by Wehr (1993), who noted that 'spatial variations in sedimentation rates … might locally shift the onset of progradation to an earlier time and delay the onset of retrogradation.' These issues were further tackled by Martinsen and Helland-Hansen (1995), Helland-Hansen and Martinsen (1996) and Catuneanu *et al*. (1998b), who summarized the various types of shoreline trajectories that may develop in response to the strike variability in subsidence and sedimentation.

As depicted in Fig. 5.52, the defining element that is common among all types of transgressive coastlines is

River mouth environments		Conditions: River mouth	Conditions: Open shoreline
Deltas	Prograding deltas *(large rivers, high sediment supply)*	Sedimentation > Base-level rise	Base-level rise > Sedimentation
Deltas / Estuaries	Retrograding ("bayhead") deltas/ Incomplete (drowned) estuaries *(smaller rivers, unincised channels)*	Base-level rise > Sedimentation Drowning of river <	Base-level rise > Sedimentation Transgression of open shoreline
Estuaries	Complete estuaries *(smaller rivers, incised valleys)*	Base-level rise > Sedimentation Drowning of river >	Base-level rise > Sedimentation Transgression of open shoreline

FIGURE 5.51 River-mouth environments of the transgressive systems tract. Estuaries are commonly regarded as river-mouth environments diagnostic for transgression, as being associated with a retrogradational shift of facies and forming only during the landward shift of the shoreline. While this is true, a wider array of river-mouth environments, ranging from estuaries to proper deltas, may be part of a larger-scale transgressive systems tract as a function of the balance between the rates of sedimentation and the rates of base-level rise at the river mouth. The two end members of this array include the estuaries, usually in the case of smaller rivers (the rates of base-level rise outpace the sedimentation rates at the river mouth), and the prograding (proper) deltas in the case of large rivers (the rates of sedimentation outpace the rates of base-level rise at the river mouth). Note that in the latter case, deltas may only be considered as part of a transgressive systems tract where the adjacent open shoreline is transgressive; otherwise, we deal with normal regressive (lowstand or highstand) deltas (Fig. 5.52). Between proper (prograding) deltas and fully developed estuaries, retrograding ('bayhead') deltas may also develop where estuaries are partly drowned by the rapid transgression of the adjacent open shorelines. This situation is prone to occur in the case of unincised channels, where the transgression of the open shoreline tends to be faster than the rate of drowning of the river (the river's sediment supply is higher than the amounts of sediment available for the construction of backstepping beaches). Such bayhead deltas represent only one of the several sub-environments of a typical (complete, or fully developed) estuary, and hence are marked in the diagram as 'incomplete' estuaries. Complete estuaries tend to form at the mouth of incised valleys, which facilitate the drowning of rivers at rates that are higher relative to the rates of transgression of the adjacent open shorelines.

the retrogradational character of open shorelines. Within this overall transgressive setting, river-mouth environments may show a range of depositional trends, from prograding deltas to retrograding and fully developed estuaries, as a function of the balance between accommodation and riverborne sediment supply. At one end of the spectrum, large rivers with high sediment load may prograde into the basin in spite of the transgression recorded by the adjacent open shorelines (case B in Fig. 5.52; Figs. 5.53 and 5.54). In such cases, the rates of base-level rise are higher than the rates of aggradation of backstepping beaches, but lower than the sedimentation rates at the river mouth. At the other end of the spectrum, smaller rivers that do not supply enough sediment to fill the entire accommodation created by base-level rise are converted into estuaries characterized by a retrogradational shift of facies. In such cases, the rates of base-level rise outpace the rates of aggradation both within the estuaries and along the adjacent open shorelines. Within this context of retrogradational river-mouth environments, two situations may be envisaged (Figs. 5.51 and 5.52). Firstly, where the transgression of the open shoreline lags (is slower than) the transgression of the river, a fully developed estuary will form, which is commonly the case with incised valleys (Dalrymple *et al.*, 1994; case D in Fig. 5.52). Secondly, where the transgression of the open shoreline is faster than the transgression of the river, which is generally the case with unincised channels, the estuary is drowned (i.e., incompletely developed) and represented only by its retrograding bayhead delta subenvironment (Fig. 5.51; case C in Fig. 5.52). The formation of bayhead deltas is favoured in wave-dominated estuaries, and it is unlikely in tide-dominated settings (Figs. 4.46 and 4.47; Allen, 1991; Reinson, 1992; Dalrymple *et al.*, 1992; Allen

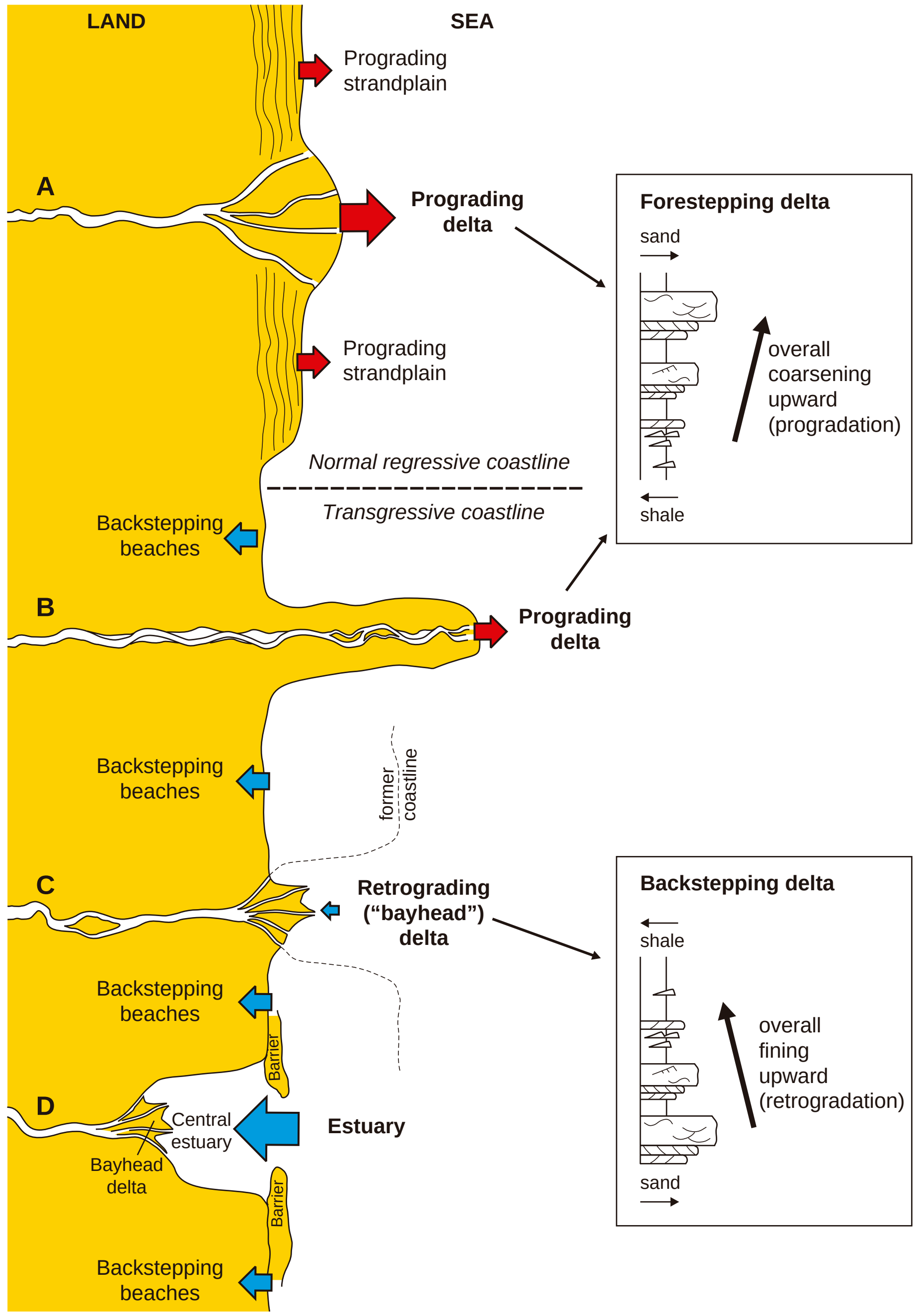


FIGURE 5.52 Types of coastlines that may develop during base-level rise. Not to scale. The progradational or retrogradational character of the shoreline in both river-mouth and open shoreline settings is dictated by the balance between sedimentation rates and the rates of base-level rise. Normal regressive coastlines (lowstand, highstand) are defined by progradation of both river-mouth and adjacent open shoreline settings. Transgressive coastlines are defined by retrogradation of the open shoreline, while the river mouth could be progradational ('proper' deltas) or retrogradational (estuaries, or portions thereof/bayhead deltas) (Fig. 5.51). Note that the definition of deltas and estuaries is based on stratigraphic criteria (progradational *vs.* retrogradational depositional trends, respectively), irrespective of the mechanisms of sediment redistribution adjacent to the coastline (i.e., river-, wave- or tide-dominated settings). A—normal regressive delta (for an example of a prograding and aggrading strandplain in an open shoreline setting, see Fig. 3.22); B— 'proper' (progradational/forestepping) delta in a transgressive setting (see Fig. 5.53 for a modern analogue); C—retrogradational ('bayhead') delta in a transgressive setting (see Fig. 5.51 for the conditions that may lead to the formation of retrograding/backstepping deltas); D—fully developed estuary.

and Posamentier, 1993; Zaitlin *et al.*, 1994; Shanmugam *et al.*, 2000).

The overall vertical profiles of prograding (forestepping) and retrograding (backstepping) deltas of the transgressive systems tract are presented in Fig. 5.52. Note that in both cases, the overall coarsening- and fining-upward profiles, respectively, consist of higher-frequency coarsening-upward successions reflecting short-term progradation of riverborne sediments in each of the two types of transgressive river-mouth settings. In the longer term, however, the overall trend reflects the dominant direction of facies shift. This overall vertical profile is punctuated by high-frequency flooding events which terminate the deposition of each individual short-term coarsening-upward succession.

The accumulation of shallow-marine facies in a transgressive setting is governed by a set of first principles, including: sediment supply to the shallow-marine environment is limited during shoreline transgression, as most riverborne sediment is trapped

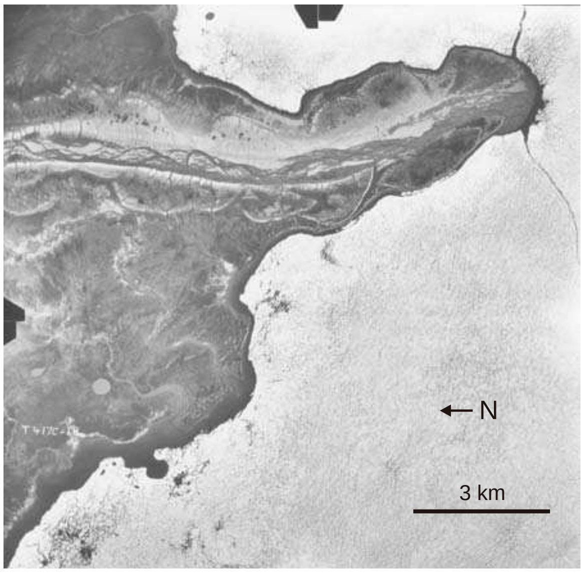


FIGURE 5.53 Aerial photograph showing a river-dominated, prograding delta in an overall transgressive setting (case B in Fig. 5.52; Chads Point, western Sabine Peninsula, Melville Island, Canadian Archipelago; photograph courtesy of J. England). The high sediment supply of the river causes the delta to prograde in spite of the transgression recorded by the adjacent open shorelines. The rate of post-glacial base-level rise is higher than the rate of aggradation of backstepping beaches along the open shoreline, but it is lower than the sedimentation rates at the river mouth.

within rapidly aggrading fluvial and coastal systems (*caveat*: see the case of coastal erosion in Fig. 3.20); an additional source of sediment for the shallow-marine environment is provided by processes of wave erosion in the upper shoreface during transgression; these sediments are transported landward during fairweather to form backstepping beaches or estuary-mouth complexes, and are dispersed seaward on the shelf by storm surges and tidal currents to form sheet-, ridge-, or wedge-shaped deposits; and sedimentation in a transgressive marine environment tends to 'heal' the seascape profile by smoothing out the breaks in seafloor gradients. The latter process leads to the formation of '*healing-phase wedges*' in the low areas of the seafloor in an attempt to re-establish a graded seafloor profile (Posamentier and Allen, 1993). Healing-phase wedges may form in various settings of the transgressive marine environment, each involving similar depositional processes but different scales of observation, from shoreface (wedge thickness in a range of meters; Fig. 3.20), to shelf (wedge thickness up to tens of meters, filling the low area outboard of the youngest regressive clinoform; Figs. 3.21 and 5.55) and even the deep-water environment (wedge thickness up to hundreds of meters, smoothing out the difference in gradients between the continental slope and the basin floor; Figs. 3.22, 5.56, and 5.57). In addition to healing-phase wedges, transgressive shallow-marine

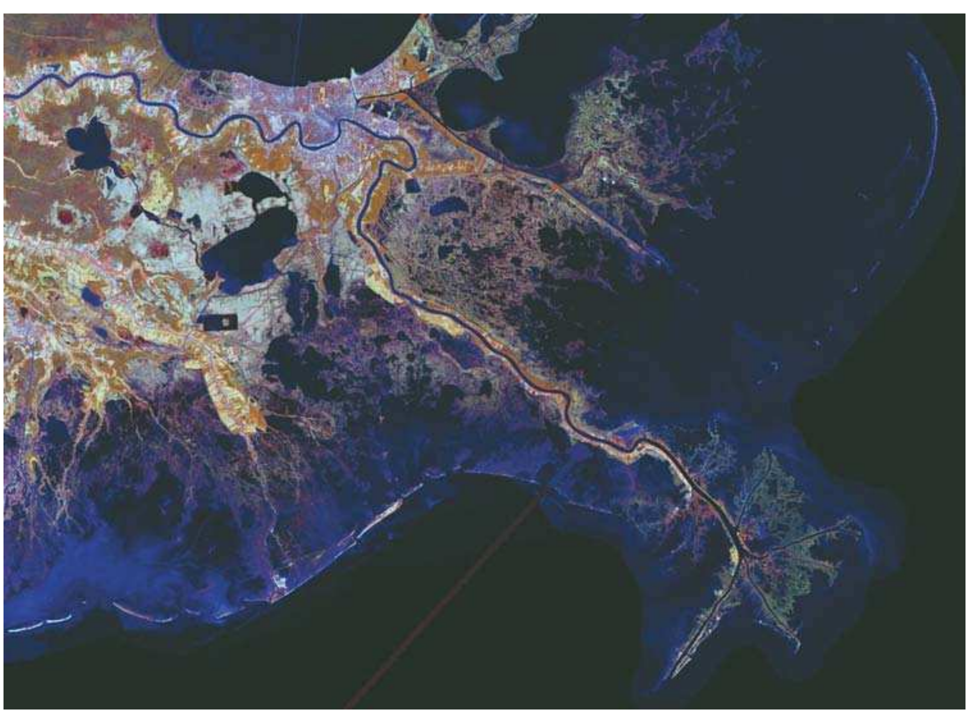


FIGURE 5.54 Satellite image showing a river-dominated, prograding delta in an overall transgressive setting (case B in Fig. 5.52; Mississippi delta, Louisiana; image released by the U.S. Geological Survey National Wetlands Research Center). The high sediment supply of the river causes the delta to prograde in spite of the transgression recorded by the adjacent open shorelines.

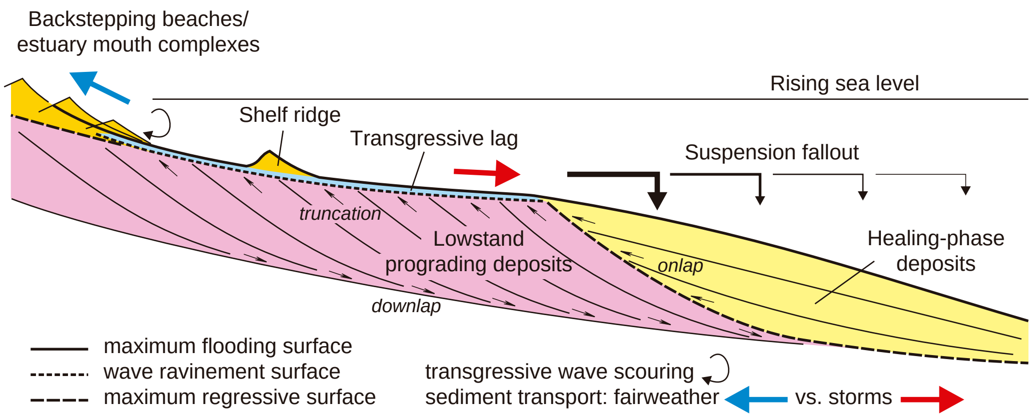


FIGURE 5.55 Coastal to shallow-marine deposits of the transgressive systems tract (not to scale; modified from Posamentier and Allen, 1993, 1999). At the scale of the continental shelf, discernable transgressive deposits include backstepping beaches (open shoreline settings) and estuary-mouth complexes (retrograding river-mouth settings), transgressive lag deposits overlying the wave-ravinement surface that forms in the upper shoreface, sand sheets and/or sand ridges that form within inner shelf environments in relation to storm surges and tidal currents, and more distal healing-phase wedges that fill the low areas of the seafloor outboard of the last regressive clinoform. At a smaller scale, healing-phase wedges may also form in the transgressive lower shoreface environment, smoothing out the slope break created by wave scouring in the upper shoreface (not shown in this diagram; Fig. 3.20). Note that the wave-ravinement surface invariably removes the seaward termination of the nonmarine portion of the maximum regressive surface. Assuming that the rates of fluvial and coastal aggradation during transgression are higher than the amount of transgressive wave scouring in the upper shoreface, the maximum regressive surface and the wave-ravinement surface diverge in a landward direction. The healing-phase wedge shown in this diagram is primarily composed of relatively fine-grained sediment accumulated from suspension. As the fallout rate decreases with distance in a seaward direction, the overall geometry of bedding surfaces changes from concave-up (shape that is inherited from the youngest regressive clinoform) to flat and eventually convex-up.

deposits may also include *transgressive lags* (Figs. 3.30B and 4.50), and *shelf-sand deposits* with a sheet-like or ridge-like geometry (Fig. 5.55). Under restricted detrital supply conditions, the shallow-marine portion of the transgressive systems tract may also be represented by *carbonate condensed sections* (Fig. 4.42). The overall thickness of the shallow-water portion of the transgressive systems tract decreases toward the shelf edge, where transgressive deposits are commonly missing (Galloway, 1989; Figs. 4.41 and 5.4).

The process of wave scouring in the upper shoreface in response to the landward translation of the shoreface profile during shoreline transgression is a key to understand the stratigraphy and the sediment budget of the shallow-marine portion of the transgressive systems tract. The balance between the processes of erosion and sedimentation that are caused by this shift of the shoreface profile (see the lever point between shoreface erosion and shoreface sedimentation in Fig. 3.20) was first pointed out by Bruun (1962), and subsequently incorporated into the sequence stratigraphic theory by Dominguez and Wanless (1991), Posamentier and Chamberlain (1993), and others. The seaward transport of the sediments generated by wave erosion in the upper shoreface onto the shelf is primarily attributed to storm surges (dispersive sediment transport, resulting in the formation of *shelf sand sheets*) and tidal currents (more channelized style of sediment transport, resulting in the formation of *shelf sand ridges*) (Curray, 1964; Swift, 1968, 1976; Swift and Field, 1981; Belknap and Kraft, 1981; Demarest and Kraft, 1987; Kraft *et al*., 1987; Rine *et al*., 1991; Snedden *et al*., 1994; Snedden and Kreisa, 1995; Abbott, 1998; Snedden and Dalrymple, 1999; Posamentier, 2002). During the process of sediment transport, the coarsest clasts produced by wave scouring or otherwise available within the shoreface environment are left behind as a *transgressive lag* on top of the wave-ravinement surface (Swift, 1976; Figs. 3.30B, 4.50, and 4.51). The finer sediment, which includes most of the sand-size fraction of clasts, is transported farther onto the shelf to form sheet-like and ridge-like *shelf-sand deposits*. *Shelf sand sheets* have been documented in numerous studies (e.g., Swift, 1968; Swift and Field, 1981; Belknap and Kraft, 1981; Demarest and Kraft, 1987; Kraft *et al*., 1987; Masterson and Paris, 1987; Masterson and Eggert, 1992; Helland-Hansen *et al*., 1992; Eschard *et al*., 1993; Abbott, 1998), and are known to form relatively thin (1–3 m thick) but

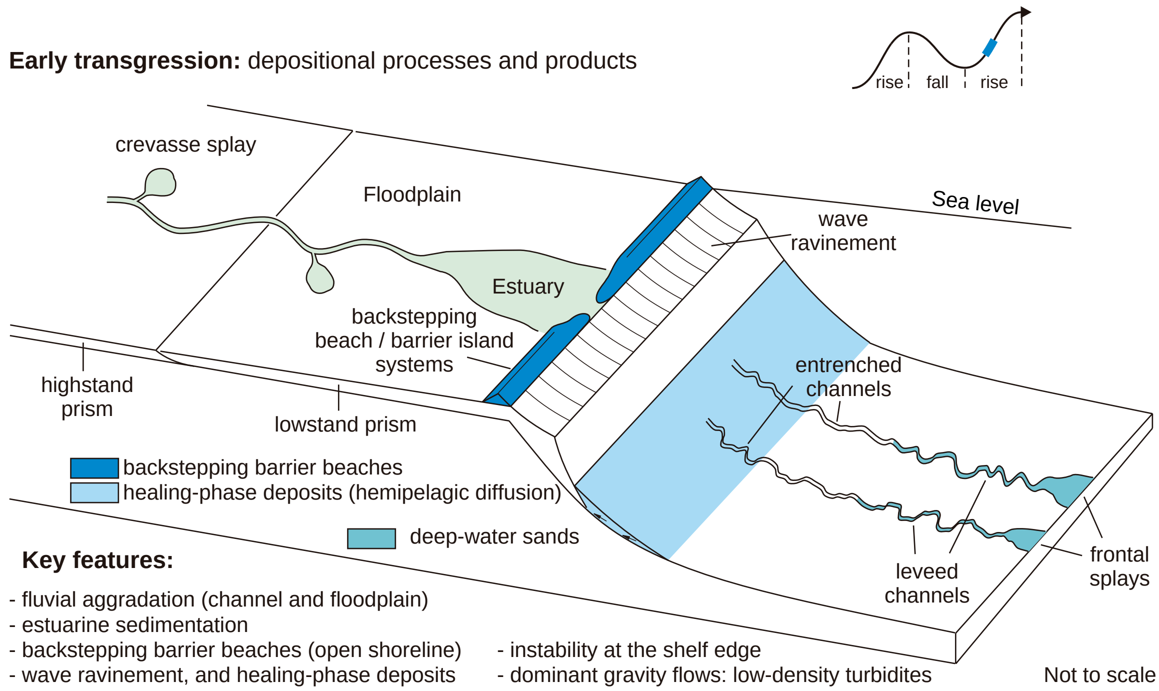


FIGURE 5.56 Depositional processes and products of the *early* transgressive systems tract (modified from Catuneanu, 2003). Rapid rates of base-level rise trigger a retrogradational shift of facies on the continental shelf, where most of the riverborn sediment is now trapped in fluvial, coastal and shallow-marine systems. Wave-ravinement processes erode the underlying normal regressive shelf-edge deltas and open shoreline systems, continuing to supply sand for the deep-water turbidity flows. These turbidity flows tend to be of low-density type, similar to the ones of the lowstand systems tract (Fig. 5.44). Such low-density turbidity currents are underloaded on the steep continental slope (flow energy > sediment load, which causes entrenchment), but become overloaded/aggradational on the low-gradient basin floor (sediment load > flow energy). The lowstand and early transgressive low-density turbidity flows travel farther into the basin relative to the high-density late falling-stage flows because the higher proportion of mud sustains the construction of levees over larger distances. Healing-phase wedges are typical for the transgressive systems tract, and infill, or heal over, the bathymetric profile established at the end of regression. Estuaries are diagnostic for transgression, but retrograding or even prograding deltas may also form in river-mouth settings during the transgression of the open shoreline, primarily as a function of degree of channel incision and sediment supply (Figs. 5.51 and 5.52; see text for details).

potentially continuous (over several hundred square kilometres) blankets of upward-fining sandy deposits. *Shelf ridges* are also sand prone, usually consisting of 5–10 m thick and regionally extensive upward-fining successions of well-sorted, cross-bedded to bioturbated fine- to coarse-grained sediment (Snedden *et al.*, 1994). A case study from the Miocene section of offshore northwest Java shelf provides high-resolution seismic images calibrated with well logs and core that reveal some of the geomorphological characteristics of these self ridge deposits (Posamentier, 2002). The features are described as large-scale, up to 17 m thick elongated bodies ranging from 0.3 to 2.0 km wide and more than 20 km long. They are asymmetric, thicker along the sharp leading edge, gradually thinning toward a more irregular trailing edge (Figs. 5.58–5.62). Smaller-scale sand waves are generally observed on top of shelf ridges, oriented oblique to the long axes of the ridges and also to the direction of ridge migration (Posamentier, 2002). Shelf sand ridges overlie transgressive wave-ravinement surfaces abandoned on the shelf following the retreat of the shoreline, and tend to be oriented parallel to the axes of structural embayments that may channelize the energy of tidal currents. Shelf ribbons, which are the smaller version of shelf ridges, may also concentrate sand in a transgressive shelf setting at scales of less than 5 m thick and less than 100 m wide (Posamentier, 2002). Both types of transgressive shelf-sand deposits (sheet-like and ridge-like) may form excellent regional reservoirs encased in shelf fine-grained seal facies. The formation of such shelf-sand deposits is favored during transgression, when

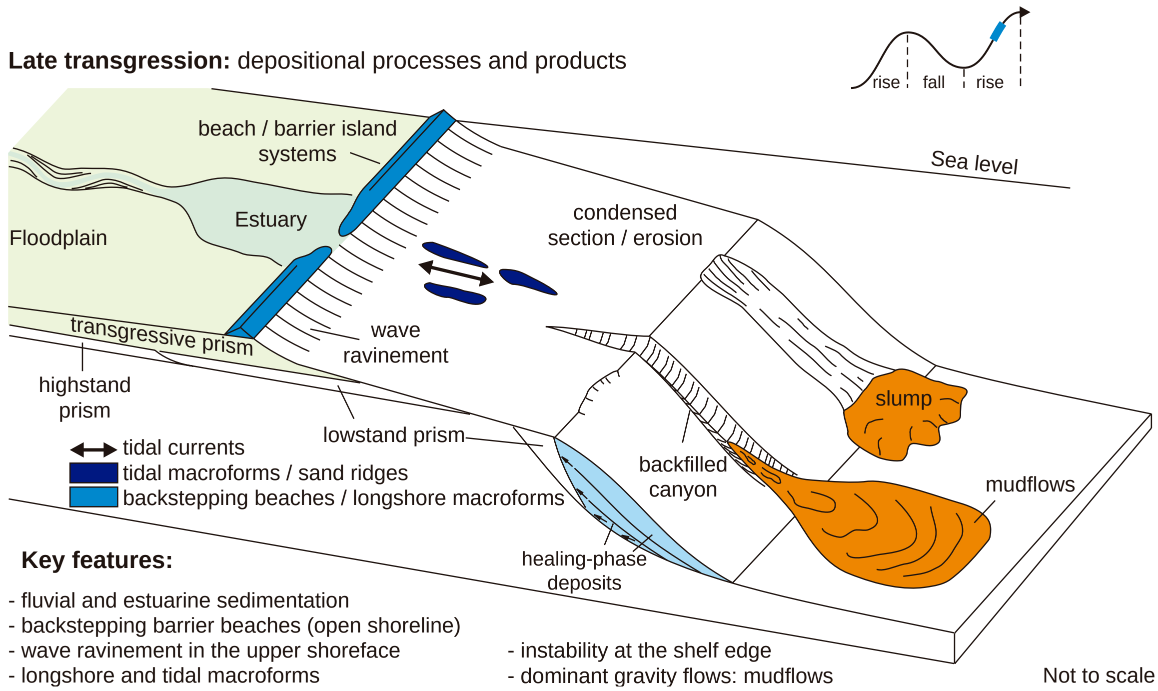


FIGURE 5.57 Depositional processes and products of the *late* transgressive systems tract (modified from Catuneanu, 2003). Most of the terrigenous sediment is trapped in the fluvial to shallow-marine transgressive prism, which includes fluvial, estuarine, deltaic, open shoreline, and lower shoreface deposits. Additional sand is incorporated within shelf macroforms (sheets, ridges, ribbons) generated by storm surges and tidal currents. Such shelf-sand deposits are generally associated with the transgressive systems tract, as the best conditions to accumulate and the highest preservation potential are offered to shelf macroforms that form during shoreline transgression (Posamentier, 2002). As base level rises rapidly during transgression, hydraulic instability at the shelf edge generates mudflows in the deep-water environment. The top of all transgressive deposits is marked by the maximum flooding surface. Where the transgressive deposits are missing (e.g., in the outer shelf-upper slope areas subject to nondeposition or erosion), the maximum flooding surface reworks the maximum regressive surface. The river-mouth settings may become estuaries (shown in the diagram) or deltas, depending on the balance between accommodation and sedimentation (Figs. 5.51 and 5.52; see text for details).

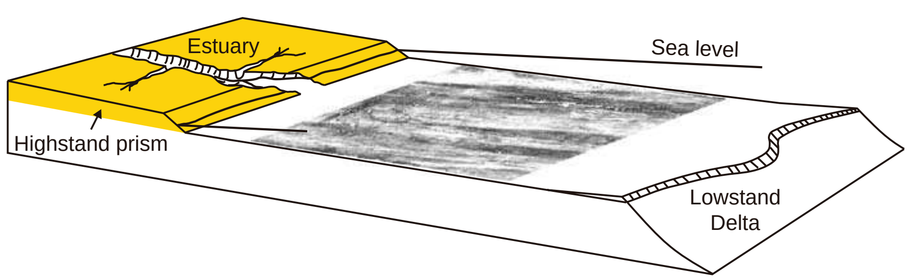


FIGURE 5.58 Reflection amplitude extraction map showing Miocene shelf ridges, offshore northwest Java (not to scale; modified from Posamentier, 2002; seismic image courtesy of H.W. Posamentier). These ridges (white features on the map, corresponding to high negative amplitudes) are several hundred meters wide and several kilometers long, and are observed along a horizon slice approximately 775 m subsea. The formation of such shelf-sand deposits is favored during transgression, when a significant portion of the continental shelf is submerged. Subsequent aggradation during the highstand base-level rise and normal regression confers on these ridges a high preservation potential in the rock record. This is why shelf-sand deposits are now recognized as a significant shallow-water component of the transgressive systems tract (Posamentier, 2002). No other systems tract offers such favorable conditions for the formation and preservation of significant sand-prone shelf macroforms.

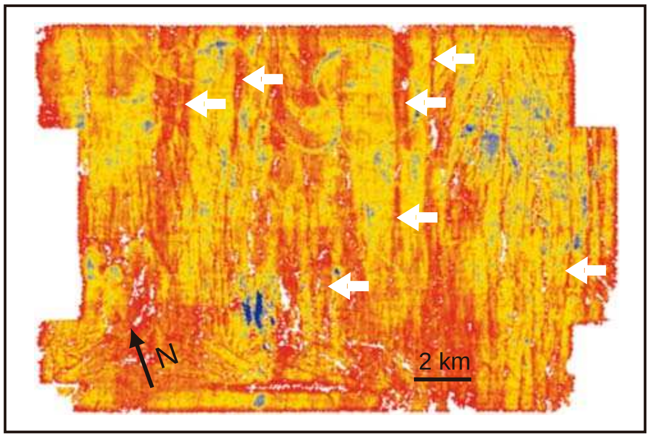


FIGURE 5.59 Reflection amplitude extraction map showing Miocene shelf ridges, offshore northwest Java (modified from Posamentier, 2002; image courtesy of H.W. Posamentier). These ridges (shown as red bands on the map; see arrows) are several hundred meters wide and several kilometers long, and are observed along a horizon slice approximately 810 m subsea. The formation of such shelf-sand deposits is favored during transgression, when a significant portion of the continental shelf is submerged. Subsequent aggradation during the highstand base-level rise and normal regression confers on these ridges a high preservation potential in the rock record. This is why shelf-sand deposits are now recognized as a significant shallow-water component of the transgressive systems tract (Posamentier, 2002). No other systems tract offers such favorable conditions for the formation and preservation of significant sand-prone shelf macroforms.

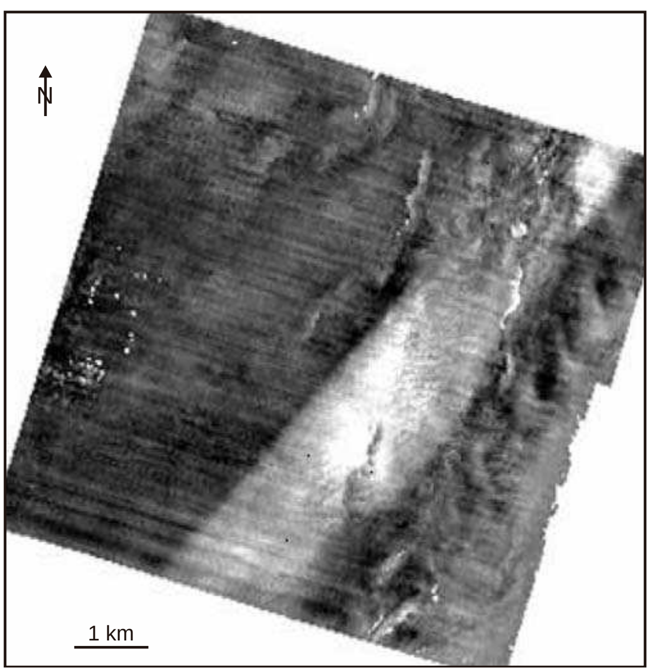


FIGURE 5.60 Reflection amplitude extraction map showing a close-up of a Miocene shelf ridge, offshore northwest Java, from a horizon slice approximately 720 subsea (modified from Posamentier, 2002; image courtesy of H.W. Posamentier). The sharply defined northwestern edge of the ridge (white feature on the map, corresponding to high negative amplitudes) is interpreted as the leading edge of the macroform. See Fig. 5.61 for a contrast between leading and trailing edges.

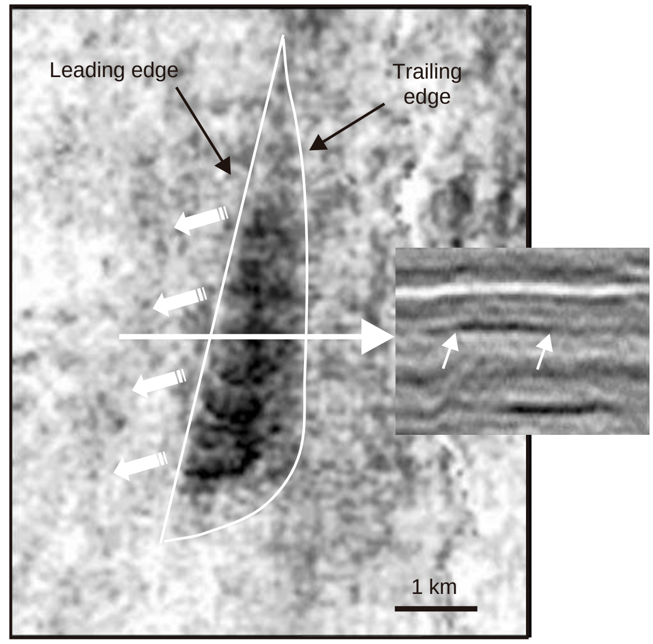


FIGURE 5.61 Morphology of a Miocene shelf ridge, offshore northwest Java, as seen on an amplitude extraction map from a horizon slice located approximately 775 m subsea (modified from Posamentier, 2002; images courtesy of H.W. Posamentier). Note the cross-sectional expression of the shelf ridge on the 2D seismic line. The shelf ridge has an asymmetrical shape in plan view, with a straight and well-defined leading edge, and a more irregular trailing edge. The direction of ridge migration is indicated by the wide arrows. Due to limitations imposed by vertical seismic resolution, the shape of the shelf ridge in cross sectional view is difficult to assess on the 2D seismic line, although the width of the sandy macroform can be estimated from the amplitude anomaly (the two small arrows indicate the edges of the macroform). The shape of shelf ridges in cross sectional view may be assessed significantly better using well logs (Fig. 5.62).

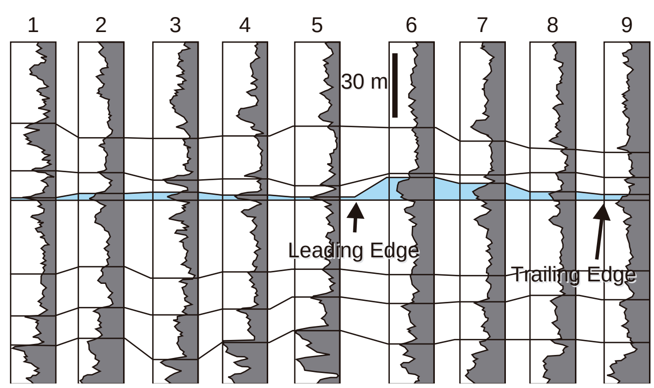


FIGURE 5.62 Well-log cross-section of correlation showing the morphology of a Miocene shelf ridge, and adjacent sand-sheet deposits, located approximately 850 m subsea, offshore northwest Java (modified from Posamentier, 2002; well logs courtesy of H.W. Posamentier). The length of the cross-section is approximately 6 km. Note the asymmetrical shape of the shelf ridge, with a thicker and better defined leading side, and a tapering trailing side. The integration of 3D seismic and well-log data (e.g., Fig. 5.61 and the well logs presented in this Figure) allows for a full 3D reconstruction of the shelf ridge morphology.

a significant portion of the continental shelf is submerged. Subsequent aggradation during the highstand base-level rise and normal regression confers on these macroforms a high preservation potential in the rock record. This is why shelf-sand deposits are now recognized as a significant shallow-water component of the transgressive systems tract (Posamentier, 2002). No other systems tract offers such favourable conditions for the formation and preservation of significant sand-prone shelf macroforms.

In addition to transgressive lags and shelf-sand macroforms, onlapping healing-phase wedges also form an integral part of, and are diagnostic for the transgressive systems tract (Figs. 3.20, 3.21, 3.22, 5.6, and 5.55–5.57; see also diagrams in Dominguez and Wanless, 1991, and Posamentier and Chamberlain, 1993, based on earlier work by Bruun, 1962). A common feature of all types of healing-phase wedges that may form in different areas of the marine environment and at different scales, is that they fill bathymetric lows in an attempt to re-establish a graded seafloor profile (Posamentier and Allen, 1993; Figs. 3.20–3.22, 5.6, and 5.55–5.57). These healing-phase depozones are invariably asymmetrical, with steeper slope gradients on the landward side, as they inherit the shape of shoreface or delta front profiles in shallow-water settings (Figs. 3.20, 3.21, 5.6, and 5.55), or of the continental slope in deep-water settings (Figs. 3.22, 5.56, and 5.57). The asymmetrical shape of these depozones confers on the healing-phase deposits a wedge-shaped geometry, as they onlap the proximal side of the bathymetric low and taper gradually in a distal direction (Fig. 5.55). Healing-phase wedges may form in lower shoreface, shelf and deep-water environments, each setting providing different amounts of accommodation and hence being associated with different spatial scales. Small-scale *lower shoreface* healing-phase wedges that fill seascape irregularities carved by waves during transgression (e.g., Fig. 3.20) overlie and onlap the wave-ravinement surface (and its associated transgressive lag) and may be overlain by shelf-sand deposits. Medium-scale *shelf* healing-phase wedges overlie and onlap the maximum regressive surface (the youngest prograding clinoform of the lowstand shoreface/delta; Fig. 5.55), and may also be overlain by shelf-sand deposits. Finally, large-scale *deep-water* healing-phase wedges tend to smooth out the difference in slope gradients between the continental slope and the basin floor, and onlap the maximum regressive surface on the continental slope (Figs. 5.56 and 5.57). Note that only healing-phase wedges that fill bathymetric lows created *during* transgression may overlie and onlap the wave-ravinement surface; healing-phase wedges that fill *existing* bathymetric lows at the onset of transgression develop basinward relative to the distal termination of the wave-ravinement surface, and hence they overlie and onlap the maximum regressive surface instead.

Irrespective of their location within the basin, on the continental shelf or in the deep-water setting, all healing-phase wedges share common features regarding the processes involved in their formation and the resulting stratal geometry. In the early stage of transgression, when the shoreline is closer to the bathymetric low area that is being infilled, sediment supply is higher and depositional processes are dominated by a combination of gravity flows and suspension sedimentation. The resulting lower portion of the healing-phase wedge is relatively coarse-grained, and may include a significant amount of sand. As transgression proceeds and the shoreline becomes remote relative to the bathymetric low area, sediment supply diminishes and the accumulation of the healing-phase wedge continues primarily from suspension fallout. This upper portion of the healing-phase wedge is relatively fine-grained, being composed mainly of silt and mud. The typical vertical profile of a fully-developed healing-phase wedge is therefore fining-upward, showing an increase in the concentration of sand beds towards the base in relation to the activity of non-channelized hyperpycnal flows. Up section, the balance between hyperpycnal and hypopycnal flow deposits changes in the favour of the latter as the supply of sand is gradually cut off. Given the nature of processes that contribute to the supply of sediment to the healing-phase depozones, which involve wave action to a large extent, sediment sources may be considered linear and the transport of sediment is primarily by diffusion rather than being channelized. As sediment is supplied from the coastline and is moved basinward to the accumulation area, sedimentation rates within the healing-phase depozone decrease accordingly in a distal direction (Fig. 5.55). As a result, the proximal side of the healing-phase wedge grows thicker with time relative to the distal portion, and the clinoform geometry changes accordingly from concave-up towards the base (mimicking the shape of the youngest regressive clinoform) to flat and eventually convex-up towards the top (Posamentier and Allen, 1993, 1999; Figs. 5.55–5.57). In the process of infilling the bathymetric low areas, these healing-phase clinoforms onlap the steeper, landward side of the seascape (Figs. 3.22 and 5.55–5.57). Where developed in a deep-water setting, onlapping the continental slope, such healing-phase wedges correspond to the 'transgressive slope aprons' of Galloway (1989) (Figs. 3.22, 5.56, and 5.57).

In addition to transgressive slope aprons (large-scale healing-phase wedges that onlap the continental slope and form from linear sediment sources), the deep-water portion of the transgressive systems tract may also include submarine fans associated with more localized sediment sources and involving a channelized style of sediment transport. The nature of gravity flows that lead to the formation of such submarine fans is known to change during transgression, from early-transgressive low-density turbidity currents to late-transgressive cohesive debris flows (mudflows) (Posamentier and Kolla, 2003; Figs. 5.11, 5.37, 5.56, and 5.57).

The low-density turbidity currents of the early stage of transgression are similar to the ones of the underlying lowstand systems tract (Figs. 5.45, 5.46, 5.47, and 5.48; also, compare Figs. 5.44 and 5.56), which makes the recognition of the maximum regressive surface in a conformable succession of deep-water turbidites most difficult (Fig. 5.63). The trend of decrease with time in the amount of sand delivered to the deep-water environment, initiated at the onset of base-level rise by the trapping of riverborne sediment in aggrading lowstand normal regressive systems on the continental shelf, continues during transgression. This trend is illustrated by the fining-upward profile of the lowstand – transgressive portion of the basin-floor submarine fan deposits in Fig. 5.63, and is explained primarily by a combination of two different factors. Firstly, the rates of base-level rise increase from the lowstand normal regression to the subsequent early transgression, which means that increasingly more riverborne sediment is trapped within aggrading fluvial to shallow-marine systems. In turn, this leads to a decrease in the amount of riverborne sediment that is made available to the deep-water environment. Secondly, the landward shift recorded by the shoreline during transgression increases the distance between the sediment entry points (river mouths) and the shelf edge, again reducing the chance of the riverborne sediment being delivered to the deep-water environment. In addition to these two factors, the gradual decrease in energy recorded by fluvial systems in relation to the denudation of source areas coupled with coastal aggradation may also explain, although to a lesser extent, the decrease with time in the amount of riverborne sand delivered to the deep-water environment during base-level rise.

During early transgression (Fig. 5.56), the shoreline is still close to the shelf edge and therefore sand can still be delivered to the deep-water environment by low-density turbidity currents. Such low-density turbidity currents are underloaded on the steep continental slope (flow energy > sediment load, which causes entrenchment; Fig. 5.45), but become overloaded/aggradational on the low-gradient basin floor (sediment load > flow energy; Figs. 5.46–5.48). The lowstand and early transgressive low-density turbidity flows travel further into the basin relative to the high-density late falling-stage flows because the higher proportion of mud sustains the construction of levees over larger distances. During late transgression (Fig. 5.57), the sediment entry points into the marine basin are far from the shelf edge, and hence no riverborne sand is made available to the staging area for the deep-water gravity flows. The vast majority of this riverborne sediment is now trapped in the fluvial to shallow-marine transgressive prism, which includes fluvial, estuarine, deltaic, open shoreline, lower shoreface, and shelf-sand deposits. As base level rises rapidly during transgression, hydraulic instability at the shelf edge results in the erosion of outer shelf – upper slope fine-grained sediments, generating mudflows in the deep-water environment (Figs. 5.57 and 5.63). These mudflow deposits are similar to the ones of the early falling-stage systems tract (Figs. 5.33–5.36), and complete the fining-upward profile illustrated in Fig. 5.63 for the lowstand – transgressive portion of the basin-floor submarine fan complex.

The preservation potential of the transgressive deposits is generally high due to the fact that the subsequent highstand normal regression leads to sediment aggradation across the entire basin (Fig. 5.6). Generally speaking, the transgressive systems tract has the best preservation potential among all systems tracts, from the basin margin to the basin center. By comparison, the falling-stage deposits in continental shelf-type settings are strongly affected by subaerial erosion during base-level fall; the downstream fluvial portion of the lowstand systems tract is commonly affected by wave-ravinement erosion during transgression; and the fluvial to shallow-marine portion of the highstand systems tract is subject to subaerial erosion during the subsequent fall in base level.

Economic Potential

Petroleum Plays

The petroleum plays of the early transgression have a bimodal distribution, some being related to the continental shelf-based transgressive wedge, and others being part of the deep-water wedge (Figs. 4.41 and 5.56). On the continental shelf, likely close to the shelf edge, the best reservoirs are concentrated along the coastline, being represented by backstepping beaches (open shoreline settings), estuary-mouth complexes, retrograding bayhead deltas or even prograding deltas

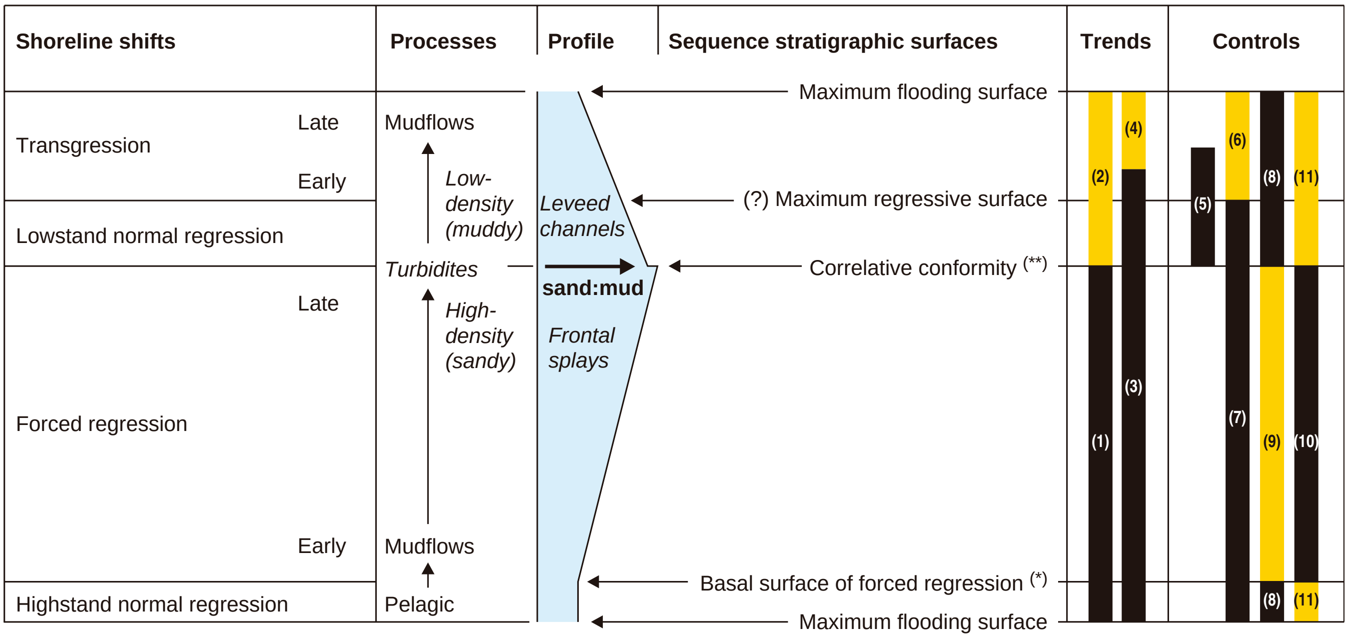


FIGURE 5.63 Composite vertical profile of a basin-floor submarine fan complex that forms during a full cycle of base-level changes, showing overall grading trends and the inferred position of the four low-diachroneity (event-significant) sequence stratigraphic surfaces (modified from Catuneanu, 2003, with additional information from Posamentier and Kolla, 2003). Key: (*) correlative conformity of Posamentier and Allen (1999); (**) *sensu* Hunt and Tucker (1992); (1) coarsening-upward; (2) fining-upward; (3) progradation of the submarine fan complex; (4) retrogradation of the submarine fan complex; (5) accelerating base-level rise (increasingly more terrigenous sediment is trapped within fluvial to shallow-marine systems, and correspondingly less sand is available for the deep-water setting—hence the fining-upward profile); (6) shoreline transgression (retrogradation of sediment entry points into the marine basin—hence the fining-upward profile); (7) shoreline regression (progradation of sediment entry points into the marine basin—hence the coarsening-upward profile); (8) decrease with time in fluvial gradients and energy during base-level rise (caused by denudation of source areas coupled with coastal aggradation—hence the fining-upward profile);(9) increase with time in fluvial gradients and energy during base-level fall (caused by differential fluvial incision or differential tectonism—hence the coarsening-upward profile); (10) no fractionation of the riverborne sediment during base-level fall: all grain-size classes are delivered to the deep-water environment; (11) fractionation of the riverborne sediment: the coarser sediment fractions are preferentially trapped on the continental shelf, leaving only the finer sediment fractions available for deep-water gravity flows (hence the sharp decrease in sand/mud ratio across the correlative conformity). Note that different controls that operate at the same time may tend to generate opposite trends (e.g., controls (5) and (8) promote a fining-upward profile, whereas control (7) promotes a coarsening-upward profile), and it is their interplay that determines the actual trend in the rock record. In this case, the onset of base-level rise, with its subsequent accelerating rates (control (5)), has the most profound influence on the balance of sediment budget across the basin, and hence it triggers the change from high- to low-density turbidites across the correlative conformity. The change from high- to low-density turbidites at the onset of base-level rise means not only a decrease in the volume of terrigenous sediment made available to the deep-water environment, affecting the *sediment/water ratio* of gravity flows, but also a decrease in the *sand/mud ratio* in these flows as the coarser fractions of the riverborne sediment are trapped first in the aggrading fluvial to coastal systems. Also note that autocyclic shifts through time in the *locus* of deposition of the different fan lobes may result in the different portions of this composite profile being found in different locations within the submarine fan complex.

(river-mouth settings). The formation of estuary-mouth complexes depends on the degree of estuary development. For example, a fully established estuary in a wave-dominated setting will have all its subenvironments represented, including the bayhead delta, the central estuary and the estuary-mouth complex (Fig. 5.52). This is commonly the case with transgressions that flood rivers that flow within incised valleys, where the drowning of the river is faster than the transgression of the adjacent open shoreline (Fig. 5.51). In such settings, the preservation of estuary-mouth complexes (e.g., Fig. 4.53) also indicates that the rates of wave-ravinement erosion are less than the rates of sedimentation within the estuary. The same condition applies for the preservation of backstepping beaches and barrier island systems; otherwise, in coastal settings characterized by extreme wave energy, the preservation of coastal reservoirs is unlikely (Leckie, 1994; Figs. 3.24 and 3.26). An incomplete (drowned) estuary, where the flooding of the river is outpaced by the transgression of the adjacent open shoreline, is only represented by the backstepping bayhead delta, without the establishment of the central estuary and estuary-mouth complex subenvironments (Fig. 5.52). This situation is prone to occur in the case of relatively small rivers (low sediment supply) that flow within unincised channels (Fig. 5.51). This discussion is meant to reveal the complexity that may be encountered in the 'real world,' where a seemingly endless number of possibilities may be envisaged depending on local circumstances. This argues, once again, that sequence stratigraphic modeling needs to be performed on a case-by-case basis, as rigid and 'universal' templates are bound to be misleading when applied indiscriminately to all case studies.

Landward from the shoreline, the potential for petroleum exploration of the transgressive systems tract is generally moderate to poor because of the extensive development of fine-grained floodplain facies in response to the rapid rates of base-level rise. Fluvial reservoirs are represented by isolated channel fills, levees, and crevasse splay deposits engulfed within floodplain fines. Seaward from the shoreline, onlapping healing-phase deposits trap part of the sand supplied by wave-ravinement processes, whereas the surplus of sediment continues to feed the deep-water submarine fans *via* low-density turbidity flows, for as long as the shoreline is still close to the shelf edge (Fig. 5.56). These early transgressive turbidites are commonly expected on the low-gradient basin floor, forming the fill of leveed channels and also building relatively small, distal frontal splays. No such reservoirs are expected on the steeper-gradient continental slope, where channels tend to be entrenched (erosion > sedimentation) due to the underloaded nature (energy flux/transport capacity > sediment load on a steep seascape) of the low-density turbidity flows. As the diluted turbidity currents of the early stage of transgression are increasingly dominated by finer-grained sediment (Fig. 5.63), which sustains the formation of levees, they tend to travel farthest into the basin relative to all other gravity flows that are recorded during a full cycle of base-level changes. These early transgressive turbidity currents are similar in nature to the flows of the previous lowstand stage, but are expected to be of even lower density because of the accelerating base-level rise that allows for more riverborne sediment to be trapped within aggrading fluvial to shallow-marine systems on the continental shelf.

During late transgression, the decreasing rates of base-level rise at the shoreline still outpace sedimentation rates (Fig. 5.57). A significant portion of the shelf is now submerged, and the combined activity of storm surges and tidal currents may lead to the accumulation of shelf-sand deposits, including ridges oriented normal to the shoreline (Posamentier, 2002). Elsewhere, especially in areas closer to the shelf edge, the shelf is generally subject to sediment starvation and condensed sections are likely to form (Loutit *et al*., 1988). Rapid increases in water depth lead to shelf edge instability, which results in the manifestation of gravity flows (Galloway, 1989). Such flows are mud-rich, involving fine-grained outer shelf sediments that accumulated far from the sediment entry points. As such, the sand/mud ratio of the gravity-flow deposits accumulated in the deep-water environment during rising base level (lowstand normal regression to transgression) records an overall decrease, from turbidites to mudflows (Figs. 5.44, 5.56, 5.57, and 5.63). This fining-upward trend of the rising stage in the deep-water basin is completed by the accumulation of pelagic/hemipelagic sediments of the highstand (late rise) normal regression at the top of the submarine fan complex (Figs. 5.7 and 5.63).

The petroleum plays of the late transgression are concentrated in the fluvial to shallow-marine depositional systems (the transgressive wedge that develops on the continental shelf; Fig. 4.41). The characterization of fluvial, coastal, and lower shoreface reservoirs is similar to what was described above in the case of early transgression. These reservoirs may include sandy fluvial architectural elements (channel fills, levees, crevasse splays) engulfed within floodplain fines; backstepping beaches, bayhead deltas and estuary-mouth complexes; prograding deltas; and healing-phase deposits in the lower shoreface. The new addition to the types of petroleum plays that characterize late transgression stages is represented by the shelf-sand deposits referred to above (Fig. 5.57; Posamentier, 2002). Transgressive shelf macroforms can be recognized both on modern shelves and in the stratigraphic record. As noted by Posamentier (2002), these macroforms 'are thought to have formed as a result of erosion and subsequent reworking of sand-prone deltaic and/or coastal plain deposits by shelf tidal currents... These transgressive systems tract deposits have significant exploration potential because they are commonly sand prone and tend to be encased in shelf mudstone seal facies.' During late transgression, the terrigenous sediment entry points are too far from the shelf edge to make any significant contribution to the deep-water gravity flows, so no more sand is fed to the submarine fans. Mudflows are, however, still active due to the general instability around the shelf edge, both in the outer shelf and upper slope areas (Fig. 5.57).

The main risks associated with the exploration of the transgressive systems tract rest with the identification of reservoirs, even though such facies may be found within all depositional systems that accumulate during the shoreline transgression (Fig. 5.14). The best transgressive reservoirs are commonly related to coastal settings (estuarine, deltaic, and beach sands), although their preservation in the rock record requires a number of conditions to be fulfilled, including a relatively weak wave-ravinement erosion during transgression. Such conditions need to be assessed on a case-by-case basis in the process of sequence stratigraphic analysis. In addition to coastal facies, shelf-sand deposits and deep-water turbidites may also make good prospects for petroleum exploration. The main contribution, however, of the transgressive systems tract to the development of petroleum systems within a sedimentary basin is the accumulation of source rocks and seal facies, within most transgressive depositional environments (Fig. 5.14). Transgressive shallow-marine shales, for example, usually form regionally extensive covers across continental shelves, which may serve as reference units for stratigraphic correlation that can be easily identified on 2D seismic lines, based on their 'transparent' seismic facies (Fig. 5.64).

Coal Resources

The transgressive systems tract is arguably the best portion of a stratigraphic sequence for coal exploration. The time of end-of-shoreline transgression marks the peak for peat accumulation and subsequent coal development because the water table is at its highest level relative to the landscape profile, following a time characterized by a high accommodation to sediment supply ratio during the transgression of the shoreline (Fig. 5.15). This balance between accommodation and sedimentation, tipped in favor of the former during transgression, represents a fundamental prerequisite that optimizes environmental conditions for significant accumulations of peat deposits. However, the condition that accommodation > sedimentation is necessary but not sufficient, as vegetation growth also depends on climatic constraints. Assuming that all favorable conditions are fulfilled, the best developed coal seams are expected to overlap with the maximum flooding surface (Hamilton and Tadros, 1994; Fig. 5.15). The timing of the maximum flooding surface is also relatively late in the stage of base-level rise (Fig. 4.7), which means that denudated source areas now supply

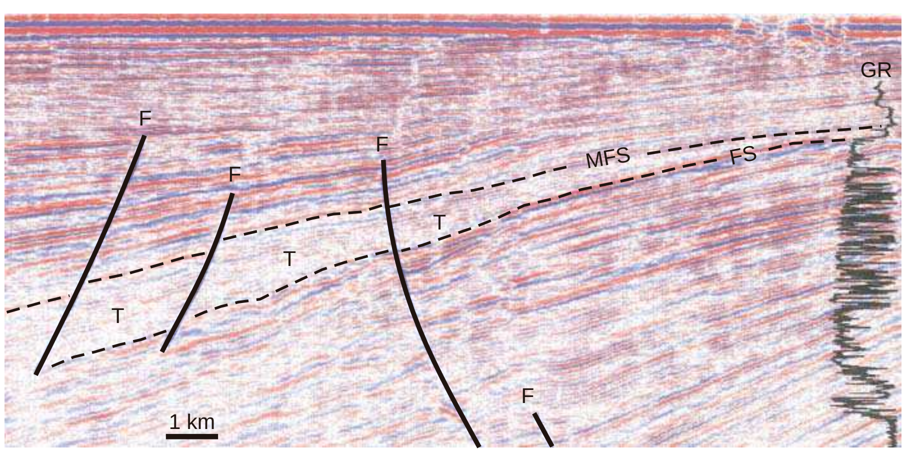


FIGURE 5.64 Seismic line showing a Pliocene to recent succession accumulated within the tectonic setting of a continental shelf (image courtesy of PEMEX). The seismic facies are calibrated with a gamma ray log. Note the regionally extensive transgressive shale that can be mapped on the seismic line as a 'transparent' facies. This transgressive shale forms a stratigraphic marker that can be used for regional correlation, and it is bounded by a flooding surface at the base and by a maximum flooding surface at the top. The transgressive shale accumulated within an outer shelf environment (below the storm wave base), following an episode of abrupt flooding that can be recognized across the basin. The underlying facies (below the flooding surface) accumulated mainly above the storm wave base, in inner shelf to beach environments. The maximum flooding surface is overlain by regressive (highstand) deposits. Abbreviations: T—transgressive shale; F—faults; FS—flooding surface; MFS—maximum flooding surface.

less sediment than in the earlier stages of base-level rise (such as during the lowstand normal regression). The scenario described in this section fits the view of standard sequence stratigraphic models, which predict coastal and fluvial aggradation during stages of shoreline transgression. One has to be aware, however, that exceptions do occur, such as in the situation described in case 2 in Fig. 3.20 (see Chapter 3 for a detailed discussion). In such cases, where coastal erosion prevails in spite of the rising base level, the nonmarine environment may also be dominated by erosional processes or sediment bypass, leading to the formation of subaerial unconformities (Leckie, 1994).

Placer Deposits

Transgressive ravinement surfaces, which are the product of wave or tidal scouring and reworking in near-shore environments during shoreline transgression, may be associated with lag deposits that have the potential of forming economically-significant placers. The G.V. Bosch and Stilfontein reefs of the Witwatersrand Basin are examples of such transgressive placers (Catuneanu and Biddulph, 2001; Fig. 5.43). The geographic distribution of transgressive placers is strictly controlled by the location of paleoshorelines, and, along dip-oriented transects, it is restricted to the area that is limited by the shoreline trajectories at the onset and end of transgressive stages. Once again, as in case of the other two unconformity-related placer types (subaerial unconformities and regressive surfaces of marine erosion – see section on the falling-stage systems tract), the paleoshoreline is a central element in the exploration for placer deposits because it limits the lateral extent of the transgressive reefs. Depending on where the maximum transgressive shoreline is located in relation to the basin margins, transgressive placers may be missed if exploration is solely based on the mapping of basin-margin unconformities.

REGRESSIVE SYSTEMS TRACT

Definition and Stacking Patterns

The regressive systems tract includes all strata that accumulate during shoreline regression, i.e., the entire succession of undifferentiated highstand, falling-stage, and lowstand deposits (Fig. 5.65). As such, this systems tract is defined by progradational stacking patterns across the basin. The concept of regressive systems tract was introduced in the sequence stratigraphic literature by Embry and Johannessen (1992), as part of their transgressive-regressive sequence model (Figs. 1.6 and 1.7), and it was subsequently refined in follow-up publications by Embry (1993, 1995).

The amalgamation of all regressive deposits into one undifferentiated systems tract is particularly feasible where the available data base is insufficient to observe stratal terminations (e.g., offlap) and stacking patterns, and thus to separate between the different genetic types of regressive deposits. In such instances, the use of the regressive systems tract over individual lowstand, falling-stage, and highstand systems tracts is preferable, due to the difficulty in the recognition of some of the surfaces that separate the lowstand, falling-stage, and highstand facies (notably, the correlative conformity and the conformable portions of the basal surface of forced regression; Embry, 1995). The identification of conformable sequence stratigraphic surfaces that serve as systems tract boundaries is virtually impossible in individual boreholes, where only well-log and core data are available. For example, if we only had well logs (2) and (5) in Fig. 5.65, it would be impossible to estimate where the basal surface of forced regression and the correlative conformity, respectively, are placed within the conformable and coarsening-upward succession of prograding shallow-marine strata. Knowledge, however, of the regional architecture and stacking patterns of this succession, as afforded by seismic data for instance, helps to infer where these conformable surfaces are placed along the cross sectional profile. Such additional insights into the stratigraphic architecture of the studied succession allow one to map the basal surface of forced regression as the oldest clinoform associated with offlap, and the correlative conformity as the youngest clinoform associated with offlap (Fig. 5.65). The application of these criteria may, however, be limited by a number of factors, including the degree of preservation of offlapping stacking patterns in the rock record, as discussed in Chapter 4.

The regressive systems tract, as defined by Embry (1995), is bounded at the base by the maximum flooding surface within both marine and nonmarine portions of the basin. At the top, the regressive systems tract is bounded by the maximum regressive surface in a marine succession, and by the subaerial unconformity in nonmarine strata. The latter portion of the systems tract boundary is taken by definition (Embry, 1995), even though there is a possibility that lowstand fluvial strata (still regressive) may be present above the subaerial unconformity. In this practice, all fluvial strata directly overlying the subaerial unconformity are assigned to the transgressive systems tract (Embry, 1995). A drawback of this approach in delineating the upper boundary of the regressive systems tract, which coincides with the boundary of the T–R (transgressive–regressive) sequence, consists in the fact that the