(i.e., the rate of eustatic fall exceeds the rate of subsidence), but positive in areas where rapid subsidence prevails over the rates of sea-level fall (Jervey, 1988; Catuneanu *et al*., 1998b).

As sedimentation rates also vary along dip oriented sections, the interplay of accommodation and sedimentation results in even more complex water-depth trends characterized by different rates of change (e.g., slow *vs.* rapid deepening or shallowing), or direction of change (shallowing *vs.* deepening) between various areas in the basin (Jervey, 1988; Catuneanu *et al*., 1998b). Despite this variability in accommodation and water-depth trends within a basin at any given time, sequence stratigraphic models account for only *one* reference curve of base-level changes relative to which all systems tracts and sequence stratigraphic surfaces are defined (Fig. 1.7). This reference curve describes changes in accommodation *at the shoreline*. The interplay between sedimentation and *this* curve of base-level changes controls the transgressive and regressive shifts of the shoreline, which are referred to in the nomenclature of systems tracts (e.g., 'transgressive systems tracts', or 'regressive systems tracts'; Fig. 1.7). These issues of numerical modelling, and their consequences for the timing of specific events during the evolution of the basin, are dealt with in more detail in Chapter 7.

The success of sequence stratigraphic analyses depends on the understanding of the basic principles. Common sources of confusion are related to the concepts of (1) base-level changes *vs.* (2) water-depth changes *vs.* (3) shoreline shifts (transgressions, regressions) *vs.* (4) grading trends (fining- and coarsening-upward). Keeping these concepts separate is as important as separating data from interpretations. Water shallowing is often confused with base-level fall, and similarly, water deepening may be confused with base-level rise. Base-level changes are measured *independent of the sediment that accumulates on the seafloor* (i.e., base level relative to datum; Figs. 3.12 and 3.15), whereas water-depth changes include the sedimentation component (i.e., sea level relative to the seafloor; Fig. 3.12). For example, either water deepening or shallowing may occur during a stage of base-level rise, as a function of the balance between the rates of creation and consumption of accommodation. Grading is a characteristic of facies that can be directly observed in outcrops, core, or well logs. Describing the rocks in terms of fining- and coarsening-upward trends is always objective, and does not necessarily translate in terms of specific base-level or water-depth changes. Grading indicates a consistent change through time in *sediment supply* across the area of observation, such as the progradation of the sediment entry points associated with shoreline regression. The trend associated with this lateral shift of facies, coarsening-upward in this example, may occur during base-level rise, base-level fall, water shallowing, or water deepening at the point of observation. The correlation between grain size and marine water depth is only safely valid for nearshore areas, where changes with depth in depositional energy are more predictable, but it may be altered offshore where the balance between wave, tide, gravity, and contour currents is less predictable. In the latter situation, the sediment transport energy may fluctuate independently of water-depth changes, and hence no linear correlation between water depth and grain size can be established. Other possible confusions, between base-level changes and shoreline shifts, or between water-depth changes and shoreline shifts, are addressed in the following section of this chapter. These issues are also examined in more detail, using numerical models, in Chapter 7.

SHORELINE TRAJECTORIES

Definitions

The interplay between base-level changes and sedimentation controls the fluctuations in water depth, as well as the transgressive and regressive shifts of the shoreline (Fig. 3.15). The types of shoreline shifts are critical in a sequence stratigraphic framework, as they determine the formation of packages of strata associated with particular depositional trends and hence characterized by specific stacking patterns, known as systems tracts.

A transgression is defined as the landward migration of the shoreline. This migration triggers a corresponding landward shift of facies, as well as a deepening of the marine water *in the vicinity of the shoreline*. Transgressions result in retrogradational stacking patterns, e.g., marine facies shifting towards and overlying nonmarine facies (Fig. 3.18). Within the nonmarine side of the basin, the transgression is commonly indicated by the appearance of tidal influences in the fluvial succession, e.g., sigmoidal cross-bedding, tidal (heterolithic wavy, flaser, and lenticular) bedding, oyster beds and brackish to marine trace fossils (Shanley *et al*., 1992; Miall, 1997). Retrogradation is the diagnostic depositional trend for transgressions, and is defined as *the backward (landward) movement or retreat of a shoreline or of a coastline by wave erosion; it produces a steepening of the beach profile at the breaker line* (Bates and Jackson, 1987). As defined by Bates and Jackson (1987), the terms 'shoreline' and 'coastline' are often used synonymously, especially when referring to processes that occur over geological (Milankovitch band and

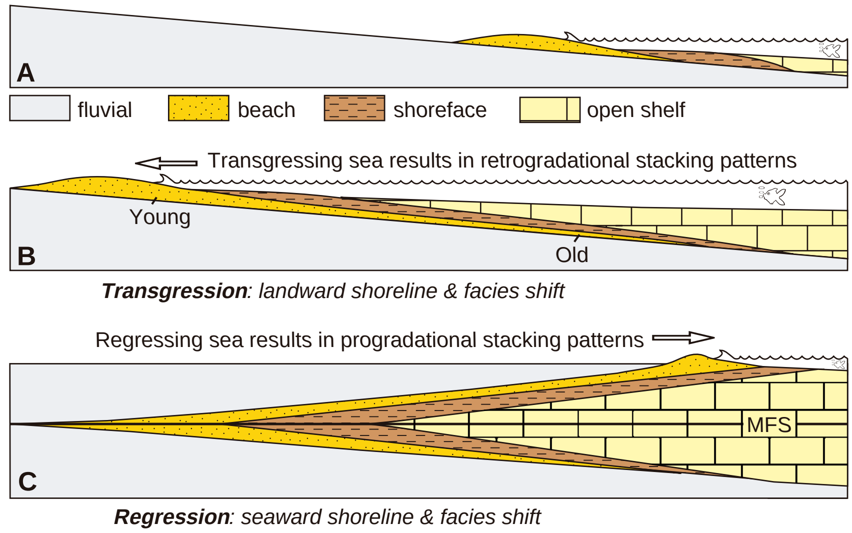


FIGURE 3.18 Transgressions and regressions. Note the retrogradation and progradation (lateral shifts) of facies, as well as the surface that separates retrogradational from overlying progradational geometries. This surface is known as the maximum flooding surface (MFS).

larger) time scales. In the solar to calendar band of time (hundreds of years and less), however, there is a tendency to regard 'coastline' as a limit fixed in position for a relatively long time, and 'shoreline' as a limit constantly moving across the intertidal area (i.e., the intersection of a plane of water with the beach, which migrates with changes of the tide or of the water level) (Bates and Jackson, 1987). In the context of this book, reference is made mainly to processes that operate over geological time scales, above the solar-band range, and therefore the terms 'shoreline' and 'coastline' are used interchangeably.

A regression is defined as the seaward migration of the shoreline. This migration triggers a corresponding seaward shift of facies, as well as a shallowing of the marine water *in the vicinity of the shoreline*. Regressions result in progradational stacking patterns, e.g., nonmarine facies shifting towards and overlying marine facies (Fig. 3.18). Progradation is the diagnostic depositional trend for regressions, and is defined as *the building forward or outward toward the sea of a shoreline or coastline (as of a beach, delta, or fan) by nearshore deposition of river-borne sediments or by continuous accumulation of beach material thrown up by waves or moved by longshore drifting* (Bates and Jackson, 1987).

The direct relationship between transgressions and regressions, on the one hand, and water deepening and shallowing, on the other hand, is only safely valid for the shallow areas adjacent to the shoreline (see *italics* in the definitions of transgressions and regressions). In offshore areas, the deepening and shallowing of the water may be out of phase relative to the coeval shoreline shifts, as subsidence and sedimentation rates vary along the dip of the basin (Catuneanu *et al*., 1998b). For example, the Mahakam delta in Indonesia (Verdier *et al*., 1980) provides a case study where the progradation (regression) of the shoreline is accompanied by a deepening of the water offshore, due to the interplay between sedimentation and higher subsidence rates. Also, the progradation of submarine fans during the rapid regression of the shoreline often occurs in deepening waters due to the high subsidence rates in the central parts of many extensional basins.

Transgressions, as well as two types of regressions may be defined as a function of the ratio between the rates of base-level changes and the sedimentation rates at the shoreline (Fig. 3.19). The top sine curve in Fig. 3.19 idealizes the cyclic rises and falls of base level through time, allowing for equal periods of time of base-level fall and rise. This symmetry is often distorted in real case studies, but the principles remain the same regardless of the shape of the reference base-level curve. During the falling leg of the base-level cycle, accommodation is reduced by external controls (primarily the interplay of subsidence and sea-level change), and the shoreline is forced to regress irrespective of the sedimentation factor. This type of regression driven by base-level fall is known as *'forced' regression* (Posamentier *et al*., 1992b). During the rising leg of the base-level cycle, accommodation is created and consumed at the same time, so the actual direction of shoreline shift depends on the interplay of these two competing forces. Sedimentation tends to dominate in the early and late stages of base-level rise, when the rates of rise are low, whereas rising base level tends to be the dominant factor around the inflexion point of the reference curve, when the rates of rise are highest.

To better understand the changes in the direction of shoreline shift that may occur during base-level rise, the bottom sine curve in Fig. 3.19 displays the *rates* of

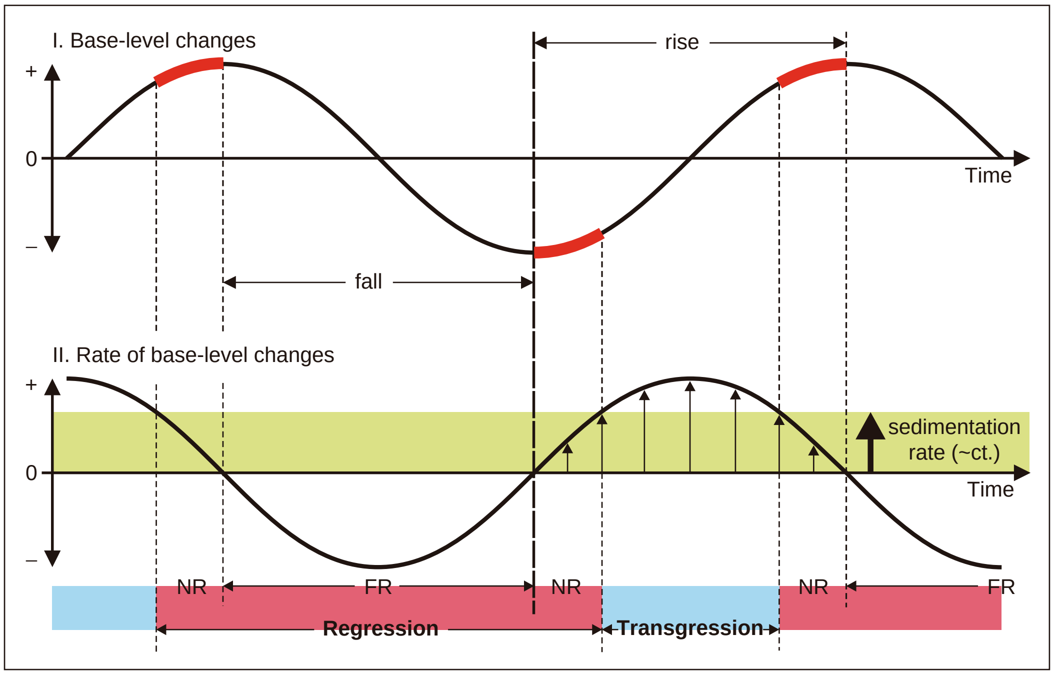


FIGURE 3.19 Concepts of transgression, normal regression, and forced regression, as defined by the interplay between base-level changes and sedimentation. The top sine curve shows the magnitude of base-level changes through time. The thicker portions on this curve indicate early and late stages of base-level rise, when the rates of base-level rise (increasing from zero and decreasing to zero, respectively) are outpaced by sedimentation rates. The sine curve below shows the rates of base-level changes. Note that the rates of base-level change are zero at the end of base-level rise and base-level fall stages (the change from rise to fall and from fall to rise requires the motion to cease). The rates of base-level change are the highest at the inflection points on the top curve. Transgressions occur when the rates of base-level rise outpace the sedimentation rates. For simplicity, the sedimentation rates are kept constant during the cycle of base-level shifts. The reference base-level curve is shown as a symmetrical sine curve for simplicity, but no inference is made that this should be the case in the geological record. In fact, asymmetrical shapes are more likely, as a function of particular circumstances in each case study (e.g., glacio-eustatic cycles are strongly asymmetrical, as ice melts quicker than it builds up), but this does not change the fundamental principles illustrated in this diagram. Abbreviations: FR—forced regression; NR—normal regression.

base-level change (first derivative of the top sine curve), which may be compared directly with the rates of sedimentation. In this diagram, sedimentation rates are assumed to be constant during a full cycle of base-level shifts, for simplicity, but other mathematical functions can be used as well to reflect more realistic fluctuations in sedimentation rates through time. What is important to emphasize is that in the early stages of base-level rise, when the rates of rise are low as increasing from zero, sedimentation rates are most likely to outpace the rates of creation of accommodation, leading to a *'normal' regression* of the shoreline, thus continuing the regressive trend of the falling leg. The timing of the end of shoreline regression is therefore not the end of base-level fall at the shoreline, but rather during the early stages of base-level rise. Once the increasing rates of base-level rise outpace the rates of sedimentation, a *transgression* of the shoreline begins (Fig. 3.19).

In the late stages of base-level rise, when the rates of rise are low as decreasing to zero, sedimentation takes over once again triggering a second *'normal' regression* of the base-level cycle. The timing of the end of shoreline transgression is therefore not the onset of base-level fall, but rather during the late stages of base-level rise (Fig. 3.19).

The discussion above implies that transgressive stages may be shorter in time (less than half of a cycle) relative to the regressive stages (normal plus forced), given a symmetrical curve of base-level changes. The actual balance between the temporal duration of transgressive and regressive stages changes with the basin, depending on the dominant allogenic controls on accommodation, as well as on sediment supply. In *foreland basins* for example, where flexural tectonics is the main control on accommodation, stages of flexural subsidence (and base-level rise) are significantly shorter in time relative to the stages of isostatic rebound (base-level fall) in the basin (Catuneanu, 2004a). In this case, a cycle of base-level shifts tends to be strongly asymmetrical, in the favour of isostatic uplift (base-level fall) and associated forced regressions. Therefore, transgressions in this tectonic setting tend to be short-lived events relative to the much longer regressive stages that intervene between transgressive events. *Extensional basins*, on the other hand, are dominated by long-term subsidence, which, combined with cyclic fluctuations in sea level, lead to asymmetrical base-level curves, this time in the favour of base-level rise (Jervey, 1988; Posamentier and Vail, 1988; Posamentier *et al*., 1988). In this case, transgressions may potentially last longer than the regressive stages, but their relative durations are ultimately controlled by the interplay of accommodation and sedimentation. Where sedimentation rates are higher than the rates of base-level rise, as recorded in many divergent continental margin settings, normal regressions become the dominant type of shoreline shift (Fig. 2.65).

As explained above, Fig. 3.19 helps to eliminate the confusion between base-level changes and shoreline shifts. A common misconception is that base-level fall equates with shoreline regression, and base-level rise signifies shoreline transgression, by neglecting the effect of sedimentation. In reality, the turnaround point from base-level fall to subsequent base-level rise in the shoreline area is temporally offset relative to the turnaround point from shoreline regression to subsequent transgression with the duration of the early rise normal regression. Similarly, the onset of shoreline regression is separated in time from the onset of base-level fall at the shoreline by the duration of late rise normal regression (Fig. 3.19).

The succession of transgressive and regressive shoreline shifts illustrated in Fig. 3.19 represents the most complete scenario of stratigraphic cyclicity, where one forced regression, two normal regressions and one transgression manifest during a full cycle of base-level changes. In practice, simplified versions of stratigraphic cyclicity may also be encountered, such as: (1) repetitive successions of transgressive and normal regressive facies, where *continuous base-level rise* in the basin outpaces and is outpaced by sedimentation in a cyclic manner; and (2) repetitive successions of forced and normal regressions, where the *high sediment input* consistently outpaces the rates of base-level rise (hence, no transgressions). The stratal geometries associated with these basic types of shoreline shifts are presented below.

Transgressions

Transgressions occur when accommodation is created more rapidly than it is consumed by sedimentation, i.e., when the rates of base-level rise outpace the sedimentation rates at the shoreline (Fig. 3.19). This results in a retrogradation (landward shift) of facies. The main processes that take place in the transition zone between nonmarine and marine environments during transgression are summarized in Fig. 3.20. These processes involve both sediment reworking and aggradation, depending on the balance between environmental energy flux and sediment supply in each location along the dip-oriented profile. The key for understanding these processes is the fact that the shoreline trajectory involves a combination of landward and upward shifts, which implies that the concave-up, wave-carved shoreface profile gradually migrates landward on top of fluvial or coastal facies. Assuming that the gradient of the nonmarine landscape is shallower than the relatively steeper upper shoreface profile, which is the case in most coastal regions, the landward translation of the shoreline triggers active wave scouring in the upper shoreface, in an attempt to carve a steeper profile that is in equilibrium with the wave-energy flux. This scour surface continues to form and expand in a landward direction for as long as the shoreline transgresses, and it is one of the sequence stratigraphic surfaces, diagnostic for transgression.

The scour surface cut by waves during the shoreline transgression (wave-ravinement surface) is onlapped by the aggrading and retrograding lower shoreface and shelf deposits (Fig. 3.20). The combination of wave scouring in the upper shoreface and deposition in the lower shoreface is required to preserve the concave-up shoreface profile that is in equilibrium with the wave energy during transgression (Bruun, 1962; Dominguez and Wanless, 1991). The onlapping deposits that accumulate in the lower shoreface and shelf environments 'heal' the bathymetric profile of the seafloor which, following shoreline transgression, has a gradient that is too steep relative to the new, lower energy conditions. These onlapping shallow-marine sediments form a transgressive wedge known as 'healing-phase' deposits (Posamentier and Allen, 1993; Fig. 3.20). The patterns of sediment redistribution as a result of wave-ravinement erosion in the upper shoreface during transgression are illustrated in Fig. 3.21. Note that the sediment eroded in the upper shoreface is transported both in landward and seaward directions. The portion of the sediment carried towards the coast may form backstepping beaches or estuary-mouth complexes, whereas the sediment carried offshore generates healing-phase wedges. Healing-phase deposits are relatively easy to recognize on seismic lines, as they form a package of convex-up reflections that onlap the last (youngest) regressive clinoform (Fig. 3.22).

The rise in base level at the shoreline promotes coastal aggradation in estuarine (river-mouth) or

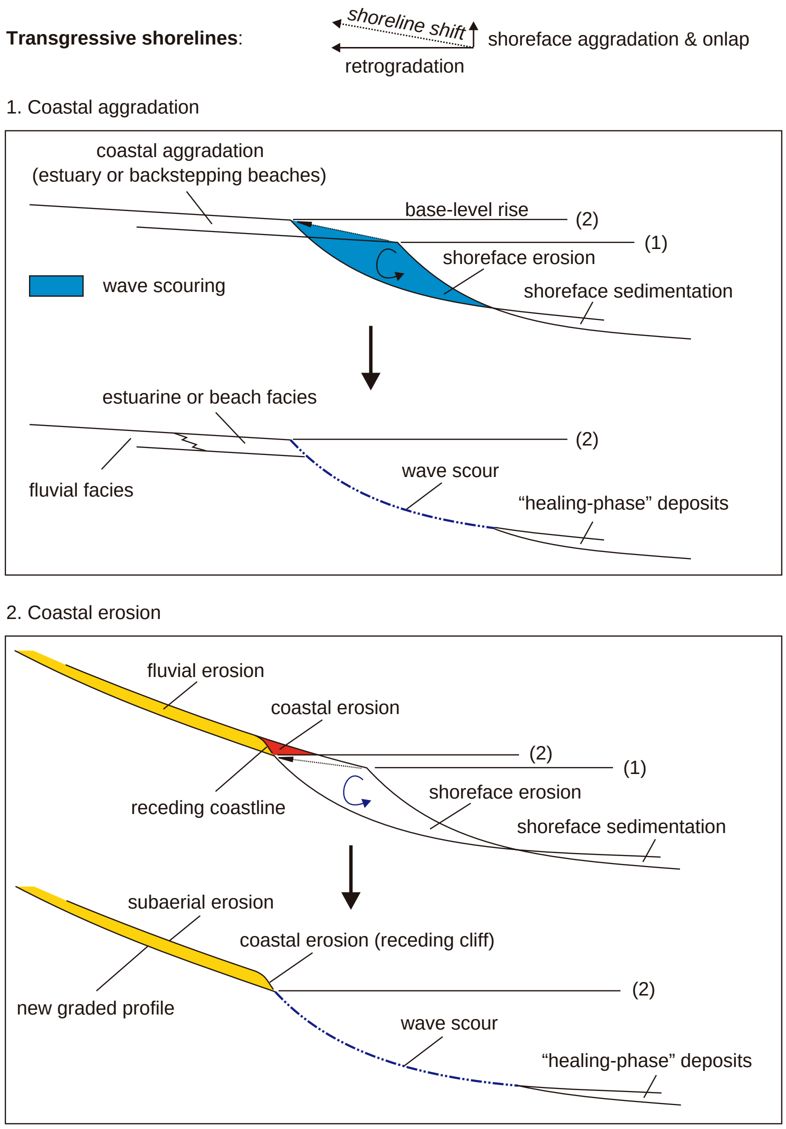


FIGURE 3.20 Shoreline trajectory in transgressive settings (from Catuneanu, 2003). Transgressions are driven by base-level rise, where the rates of base-level rise outpace the sedimentation rates in the shoreline area. The balance between the opposing trends of aggradation (in front of the shoreline) and wave scouring (behind the shoreline) determines the type of transgressive coastline. Irrespective of the overall nature of coastal processes (aggradation *vs.* erosion), the scour cut by waves in the upper shoreface is onlapped by transgressive lower shoreface and shelf ('healing phase') deposits. Low-gradient coastal plains are prone to coastal aggradation, whereas steeper coastal plains are prone to coastal erosion. In both cases, the gradients may be shallower than the average shoreface profile (approximately 0.3°).

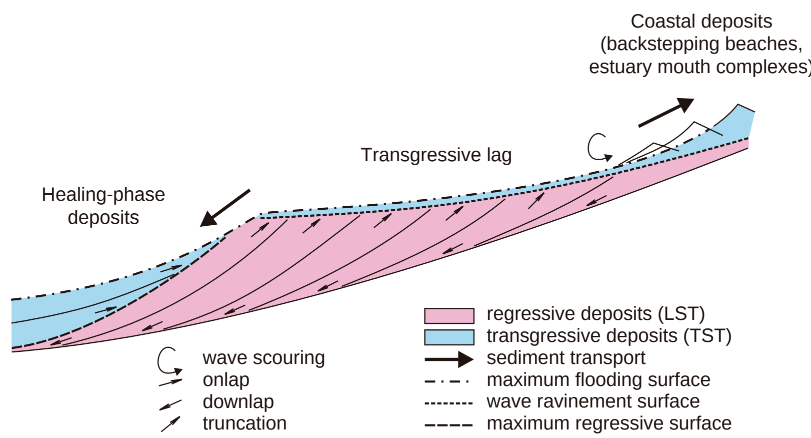


FIGURE 3.21 Patterns of sediment redistribution during shoreline transgression (modified from Posamentier and Allen, 1993, and Willis and Wittenberg, 2000). Some sediment is carried landward as backstepping beaches (open shorelines) or backstepping estuary-mouth complexes (river-mouth settings), while the coarser fraction typically mantles the ravinement surface as a transgressive lag. The transgressive coastal deposits may or may not be preserved as a function of the balance between the rates of coastal aggradation and the rates of wave-ravinement erosion. In addition, some sediment is transported seaward of the last clinoform of the underlying progradational deposits (LST) and forms a wedge-shaped deposit referred to as the healing-phase unit. Abbreviations: LST—lowstand systems tract; TST—transgressive systems tract. The definition of sequence stratigraphic surfaces follows in Chapter 4.

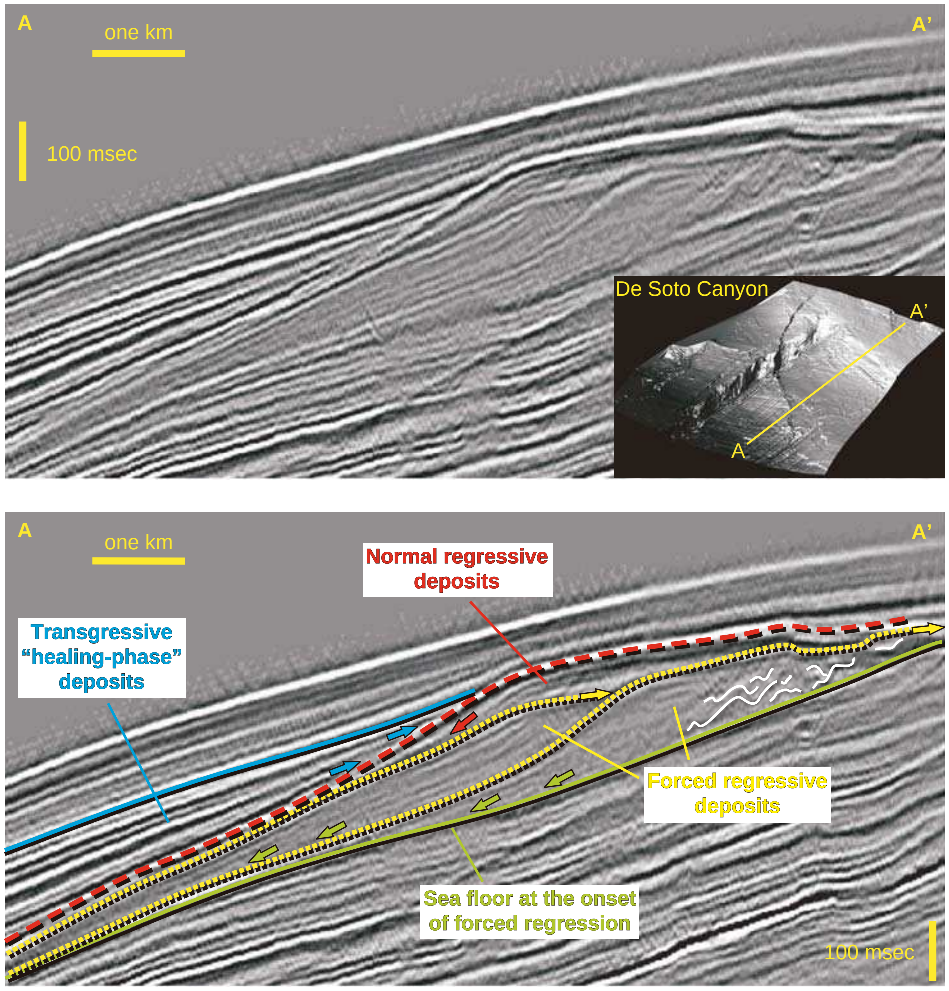


FIGURE 3.22 Shelf-edge and healing-phase deposits in the De Soto Canyon area of the Gulf of Mexico (uninterpreted and interpreted seismic lines, whose location is shown on the 3D illuminated surface) (modified from Posamentier, 2004a; images courtesy of H.W. Posamentier). Note that prograding clinoforms tend to be concave-up, in contrast with the convex-up reflections of the transgressive healing-phase wedge. The white wavy lines indicate possible slumping on the continental slope during forced regression. Regressive deposits (both normal and forced) downlap the seafloor (green and red arrows), whereas the transgressive deposits onlap the youngest prograding clinoform (blue arrows). Forced regressive deposits are associated with offlap (yellow arrows), whereas normal regressive deposits include an aggrading topset. These three genetic types of strata (forced regressive, normal regressive, and transgressive) are *independent* of the sequence stratigraphic model of choice, and their recognition is more important than the nomenclature of systems tracts or even the position of sequence boundaries, which are *model-dependent* (Fig. 1.7). For this reason, shoreline shifts, and their associated sediment dispersal systems, form the conceptual core of sequence stratigraphy as they control the formation and timing of all systems tracts and stratigraphic surfaces irrespective of the model of choice. Note that the (lowstand) normal regressive deposits shown on the 2D seismic transect include a prograding and aggrading strandplain in an open shoreline setting rather than a shelf edge delta, which is small and restricted to the channel area captured on the 3D illuminated surface. The distribution of sediment from the river mouth (shelf edge delta) to the open shoreline setting is attributed to longshore currents. For scale, the channel on the 3D illuminated surface is approximately 1.8 km wide, and 275 m deep at shelf edge. The illuminated surface is taken at the base of forced regressive deposits.

beach (open shoreline) environments. However, the tendency of coastal aggradation is counteracted by the wave scouring in the upper shoreface, as the latter gradually shifts in a landward direction. The balance between these two opposing forces, of sedimentation *vs.* erosion, determines the overall type of transgressive coastline (Fig. 3.20). Coastlines dominated by aggradation lead to the preservation of estuarine or backstepping beach facies in the rock record (Fig. 3.23). Coastlines dominated by erosion are associated with

A

B

FIGURE 3.23 Estuarine facies preserved in the rock record, showing tidally-influenced inclined heterolithic strata (Dinosaur Park Formation, Belly River Group, Alberta). A—estuary-channel point bar (the section is approximately 4 m thick); B—amalgamated estuary channel fills (the section is approximately 6 m thick). The preservation of estuary facies in the rock record indicates coastal aggradation during shoreline transgression, which means that the rates of aggradation in the estuary outpaced the rates of wave scouring in the upper shoreface. This scenario is conducive to the preservation of underlying lowstand normal regressive fluvial deposits, which are protected from transgressive wave scouring by the estuary strata.

A

B

FIGURE 3.24 Coastal erosion in a transgressive open shoreline setting (Canterbury Plains, New Zealand). A—wave-ravinement erosion outpaces coastal aggradation in spite of rising base level. As a result, a receding cliff forms instead of backstepping beaches. Beyond the cliff face, the coastal plain is subject to subaerial erosion. B—note the gravel beach, indicating high energy upper shoreface-shoreline systems. The gravel is supplied by (1) coastal erosion of the wave-cut cliff, which consists of gravel-rich Quaternary deposits, and (2) rivers (Fig. 3.25). In this open shoreline setting, the riverborne gravel is redistributed along the coastline by strong longshore currents.

unconformities in the nonmarine part of the basin, whose stratigraphic hiatuses are age-equivalent with the transgressive marine facies. Regardless of the overall nature of coastal processes, the wave-ravinement surface is onlapped by transgressive shallow-marine ('healing-phase') deposits, which provides a clue for understanding the transgressive nature of some subaerial unconformities.

A modern example of an erosional transgressive coastline is represented by the shore of the Canterbury Plains in the Southern Island of New Zealand (Leckie, 1994). In this wave-dominated setting, the rates of wave erosion outpace the rates of coastal aggradation in both open shoreline and river-mouth settings. As a result, estuaries are incised into the coastal plain, and the open shorelines are marked by receding cliffs (Figs. 3.24–3.26). The extreme wave energy that leads to overall coastal erosion is caused by oceanic swell originating as far

FIGURE 3.25 Shallow gravel-bed braided system, supplying coarse-grained sediment to the Canterbury Plains shoreline. From the sediment entry points (river mouths), the gravel is reworked and redistributed along the open coastline by strong longshore currents. Southern Alps, New Zealand.

FIGURE 3.26 Coastal erosion in a transgressive river-mouth setting (A—panoramic view and B—close up). Wave-ravinement erosion outpaces coastal aggradation in spite of rising base level. As a result, the estuary is incised, about 20 m into the coastal plain. The width of the incised estuary is about 1 km. Ashburton River, Canterbury Plains, New Zealand.

away as 2000 km. The wave-cut cliffs, which may be up to 25 m high, recede at a rate of approximately 1 m per year. Coastal erosion lowers the fluvial graded profile below the topographic profile (Fig. 3.20), causing the rivers to incise 1.5–4.2 mm per year in the vicinity of the coastline. The amount of incision gradually decreases inland from the coast, until it becomes minimal 8–15 km upstream (Leckie, 1994).

Forced Regressions

Forced regressions occur during stages of base-level fall, when the shoreline is forced to regress by the falling base level irrespective of sediment supply (Fig. 3.19). A variety of processes may accompany the forced regression of the shoreline in the transition zone between marine and nonmarine environments, including erosion, aggradation, or a combination of both. These processes affect both fluvial and marine environments, and the manifestation of one over the other (erosion *vs.* aggradation) in any region depends on the relative position between the energy flux equilibrium profile (fluvial graded profile or base level) and the ground surface (subaerial or subaqueous).

In shallow-marine settings, equilibrium profiles are generally concave-up and reflect the energy flux of fairweather waves. These profiles are dynamic, being sensitive to any changes in marine-energy flux that may occur during storms or due to the activity of marine currents. The dominant processes that manifest during forced regression in a shallow-marine environment are therefore a function of the relative position between the wave equilibrium profile and the seafloor. Low-gradient seafloors are more susceptible to wave erosion during a fall in base level, whereas steeper seafloors (with a gradient higher than the gradient of the wave equilibrium profile) are less affected by the wave-energy flux, being rather prone to aggradation (Fig. 3.27). Seafloor gradients in coastal regions are in turn controlled by the basin physiography, as well as by the dominant process of sediment distribution in the subtidal areas adjacent to the coastline.

In wave-dominated coastal settings, such as open shorelines or wave-dominated deltas, the preservation of the concave-up seafloor profile that is in equilibrium with the wave energy requires coeval deposition and erosion in the upper and lower parts of the subtidal area, respectively (Bruun, 1962; Plint, 1988; Dominguez and Wanless, 1991; Fig. 3.27). As the shoreline shifts basinward, the upper subtidal forced regressive deposits downlap the scour generated in the lower subtidal zone (Fig. 3.27). At the same time, the subaerially exposed area is commonly subject to sediment starvation, pedogenesis, or fluvial and wind degradation. The amount of nonmarine downcutting is generally proportional to the magnitude of base-level fall, but it also depends on the changes in slope gradients of the ground surface exposed by the fall in base level (see Posamentier, 2001, for a discussion of incised *vs.* unincised fluvial bypass systems).

In the case of river-dominated deltas, the angle of repose of delta front clinoforms is generally steeper than the gradient required to balance the energy of the waves, so there is no reason for wave scouring in the lower delta front area (Fig. 3.27). Therefore, the marine scour surface that forms in shallow-marine wave-dominated settings during forced regression is missing from the stratal architecture of forced regressive river-dominated deltas. In the former case, a vertical profile through the shallow-marine forced regressive succession shows an abrupt shift of facies from offshore muds to upper subtidal sands (Figs. 3.28 and 3.29), whereas this facies shift is gradational in the latter situation (Fig. 3.30).

Landward relative to the shoreline, processes of fluvial erosion or aggradation reflect changes in fluvial-energy flux that are in part controlled by the contrast between the gradients of the fluvial and seafloor profiles at the onset of forced regression. As the shoreline regresses and the seafloor becomes subaerially exposed, steeper seafloor gradients (relative to the fluvial profile at the onset of forced regression) lead to increased fluvial-energy flux and incision, whereas shallower seafloor gradients trigger a decrease in fluvial-energy flux and sediment aggradation (cases A and C in Fig. 3.31, respectively). Both processes of fluvial incision or aggradation propagate gradually from the shoreline upstream through a series of landward-migrating knickpoints (Figs. 3.31 and 3.32). Each knickpoint represents an abrupt shift in slope gradients along the fluvial profile at a particular time, and it is the change in fluvial-energy flux induced by such shifts in slope gradients that triggers aggradation or fluvial incision. A downstream increase in valley slope is prone to fluvial incision (case A in Fig. 3.31; Fig. 3.32), whereas a downstream decrease in valley slope promotes fluvial aggradation (case C in Fig. 3.31) (Pitman and Golovchenko, 1988; Butcher, 1990; Posamentier and Allen, 1999). The fluvial response to such changes in valley slope is in fact much more complex than depicted in Fig. 3.31, as rivers may internally adjust their flow parameters (e.g., the degree of channel sinuosity) in order to adapt to changing topographic gradients without aggradation or incision (Schumm, 1993).

The diagrams in Fig. 3.27 illustrate a scenario where the gradient of the seafloor in the subtidal zone is steeper than the gradient of the downstream fluvial profile,

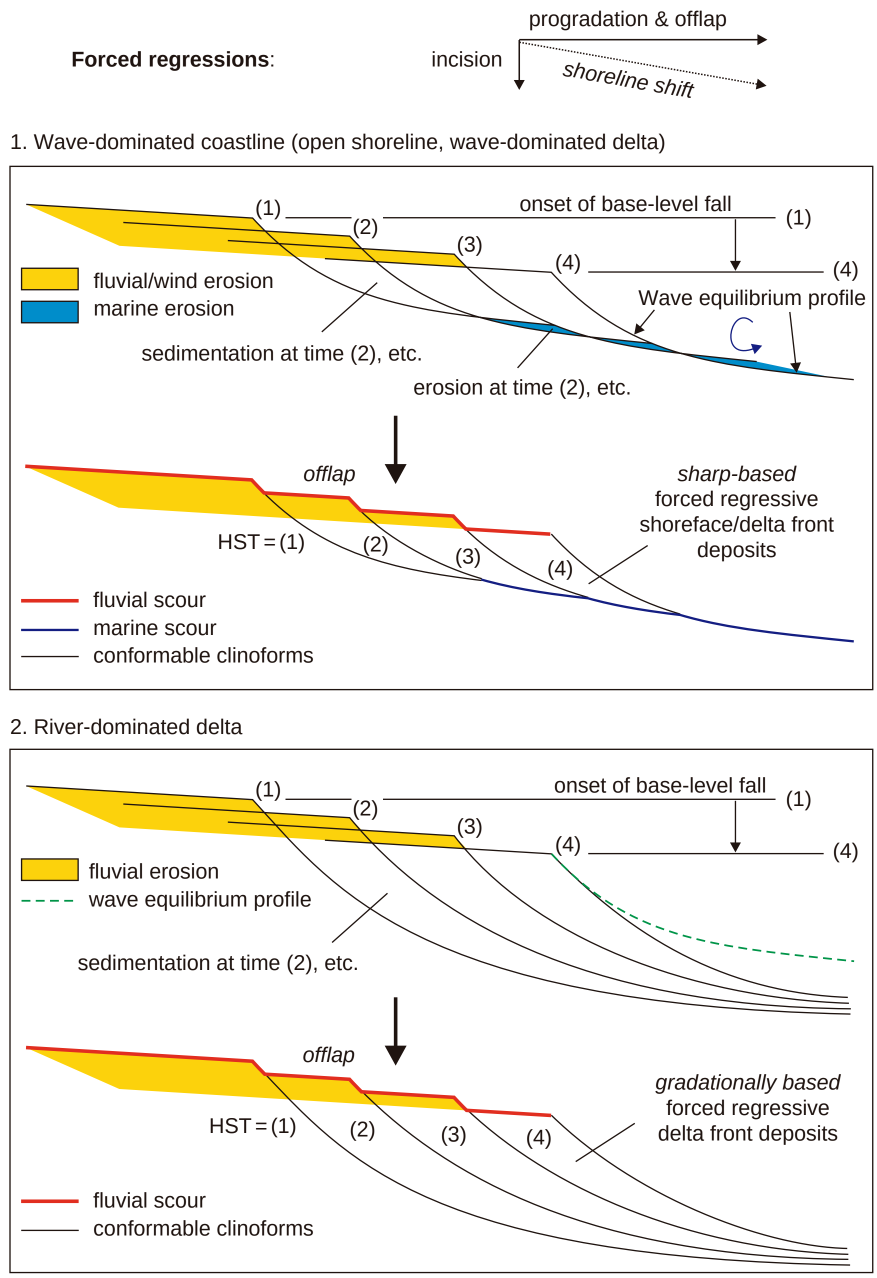


FIGURE 3.27 Shoreline trajectory in forced regressive settings (modified from Catuneanu, 2003). Forced regressions are driven by base-level fall, irrespective of sediment supply, and the rates of progradation are generally high. Wave-dominated subtidal settings are characterized by low gradients of the seafloor, which is subject to wave scouring in order to preserve a profile that is in equilibrium with the wave energy. River-dominated deltas generally have delta front clinoforms that are steeper than the wave equilibrium profile, and therefore no wave scouring takes place during forced regression. HST—highstand systems tract.

which is the case in the majority of coastal regions. Other situations may, however, occur as well, as illustrated in Fig. 3.31. These three possible scenarios may explain why rivers do not always incise during stages of base-level fall, as commonly inferred in the sequence stratigraphic literature (case A in Fig. 3.31; Fig. 3.32), but they may also bypass (case B in Fig. 3.31) or even aggrade (case C in Fig. 3.31) during the forced regression of the shoreline. As discussed earlier in this chapter, however, changes in base level controlled by tectonism and sea-level change, which are accounted for in Fig. 3.31, may be overprinted by the effect of climate change to the extent that processes of fluvial incision or aggradation may proceed in a fashion that is opposite to what is normally expected from relative sea-level changes (Blum, 1990, 1994). All these aspects of fluvial sedimentation are detailed more in Chapter 4 (discussion on subaerial unconformities), Chapter 5 (discussion of the falling-stage systems tract) and Chapter 6 (discussion of fluvial processes in a sequence stratigraphic framework).

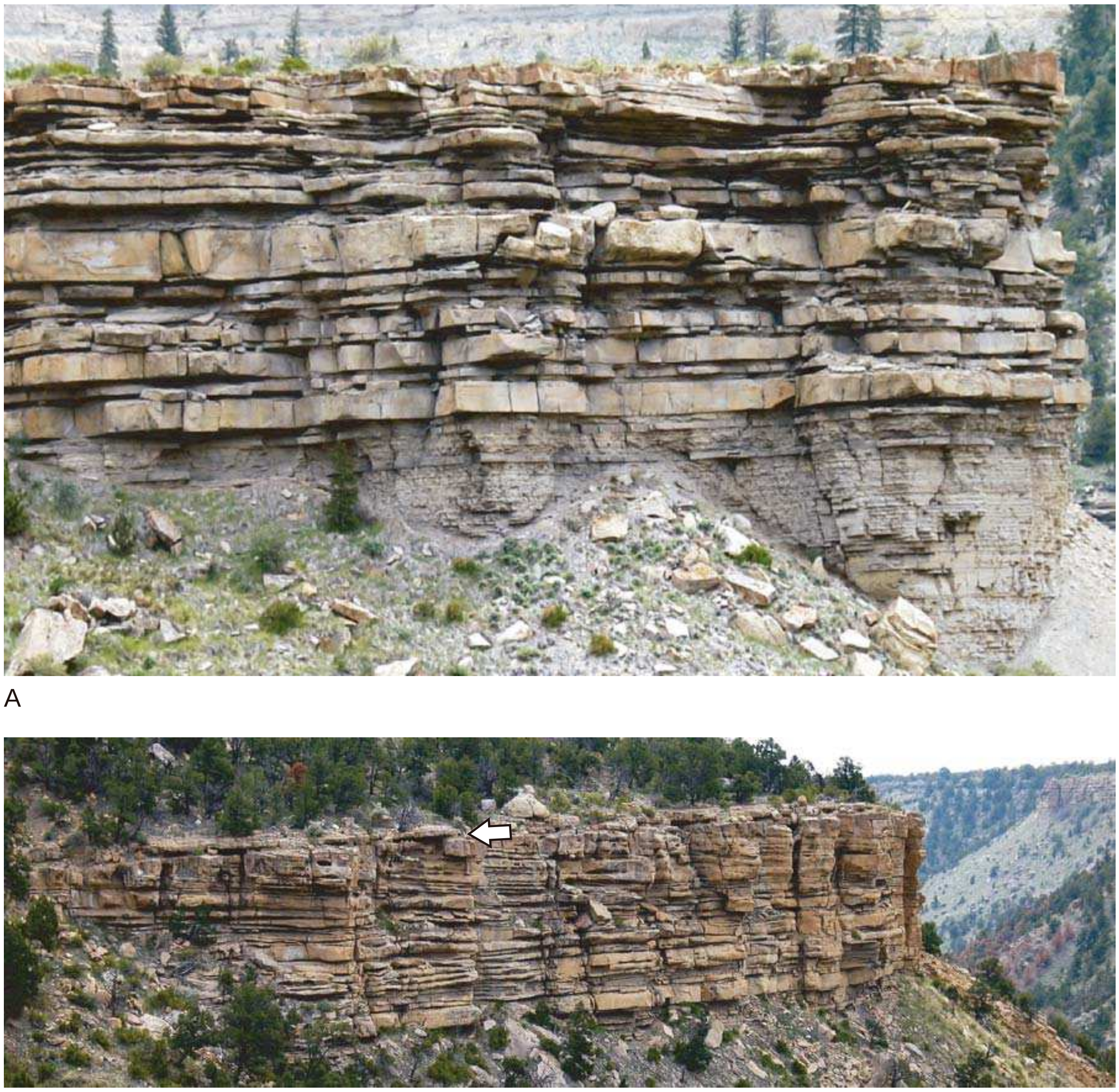
Stages of forced regression are generally characterized by a significant increase of sediment supply to the deep-water depositional systems. This is due to (1) a lack

FIGURE 3.28 Forced regressive, wave-dominated shoreface sands (with A—swaley cross-stratification–Fig. 3.29) abruptly overlying inner shelf interbedded sands and muds (B). The upper shoreface sands (A) are 'sharp-based' due to wave scouring in the lower shoreface during base-level fall. The exposed section below the wave scour is approximately 2 m thick. Blackhawk Formation, Utah.

of accommodation in the fluvial to shallow-marine environments, and therefore the terrigenous sediment tends to bypass these settings and be delivered to the deep-water environment; and (2) additional sediment may be supplied by erosional processes in the fluvial and lower shoreface environments.

The stratal architecture of shallow-marine forced regressive deposits is a function of sediment supply, rates of base-level fall, and gradient of the seafloor (Ainsworth and Pattison, 1994; Posamentier and Morris, 2000). The interplay of these variables controls the character of the forced regressive prograding lobes,

FIGURE 3.29 Swaley cross-stratification in wave-dominated, upper shoreface sandstones. Blackhawk Formation, Utah.



A

B

FIGURE 3.30 Forced regressive, river-dominated deltaic succession (Panther Tongue, Utah). A—conformable shift of facies from prodelta to the overlying delta front deposits. The delta front sands are 'gradationally based', as no wave scouring took place during the progradation of the delta; B—relatively steep delta front clinoforms (dipping to the right in the photograph, at an angle of 5–15°). As the clinoforms are steeper than the wave equilibrium profile (approximately 0.3°), no wave scouring took place during the progradation of the delta. The delta front succession is topped by a transgressive lag (sandstone layer—see arrow), which in turn is overlain by transgressive shale. Hence, no delta plain deposits are present.

which may be attached *vs.* detached, stepped-topped *vs.* smooth-topped, and spread over short or long distances (Fig. 3.33). Criteria for the recognition of shallow-marine forced regressive deposits in outcrop, core, well logs and seismic data are also provided by Posamentier and Morris (2000). Perhaps the most important defining signature of coastal to shallow-marine forced regressive deposits is their offlapping (seaward downstepping) character, which is caused by the fall in relative sea level (Fig. 3.27). This stratal stacking pattern may be observed on seismic lines (Fig. 3.22), and it is particularly significant for the exploration of age-equivalent deep-water

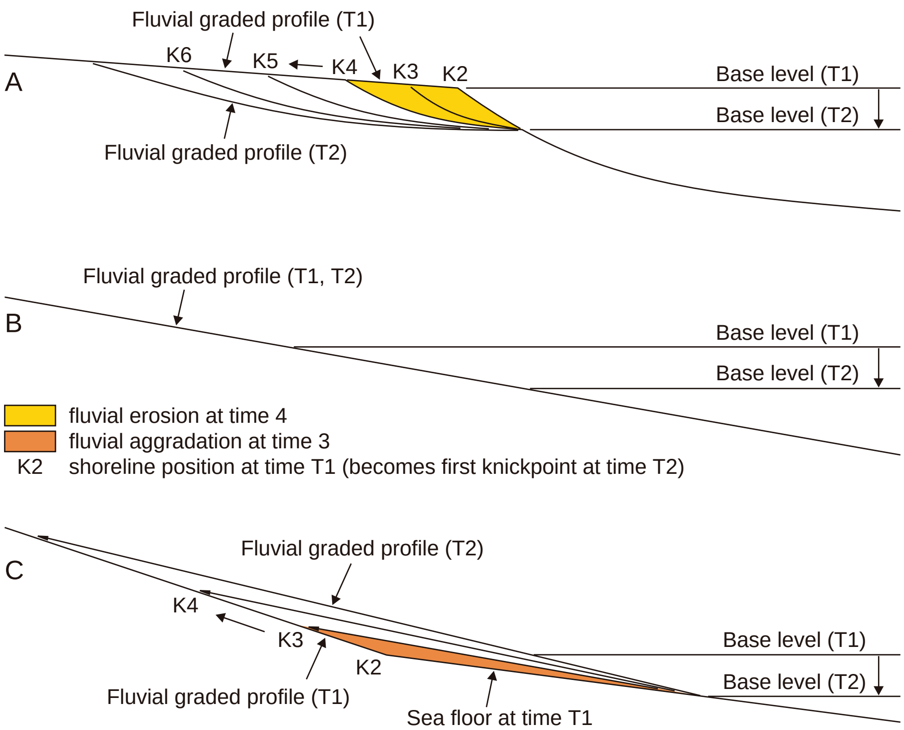


FIGURE 3.31 Fluvial responses to base-level fall, as a function of the contrast in slope gradients between the fluvial and the seafloor profiles at the onset of forced regression (modified from Summerfield, 1985; Pitman and Golovchenko, 1988; Butcher, 1990; Schumm, 1993; Posamentier and Allen, 1999; Blum and Tornqvist, 2000). A—fluvial incision; B—fluvial bypass; C—fluvial aggradation. Knickpoints (K) mark abrupt changes in the gradient of fluvial profiles. A downstream increase in slope gradient (and corresponding fluvial-energy flux) is prone to fluvial erosion (case A). A downstream decrease in slope gradient (and corresponding fluvial-energy flux) is prone to fluvial aggradation (case C). Knickpoints migrate upstream with time, resulting in a landward expansion of the subaerial unconformity (case A) or in a backfill of the landscape to the level of the new graded profile, accompanied by fluvial onlap of the old graded profile (case C). Case A is most likely, case C is least likely. Case B may describe the forced regression across a continental shelf, where minor fluvial incision (or aggradation) may still occur below the seismic resolution.

reservoirs (more details on this topic are presented in Chapters 5 and 6). Offlapping forced regressive deposits may also be observed in modern environments, such as for example in areas that are currently subject to post-glacial isostatic rebound at a rate that exceeds the present day rate of sea-level rise (Fig. 3.34).

Normal Regressions

Normal regressions occur during early and late stages of base-level rise, when sedimentation rates outpace the low rates of base-level rise at the shoreline (Fig. 3.19). In this case, the newly created accommodation is totally

FIGURE 3.32 Upstream-migrating fluvial knickpoint (arrow) along a small-scale, actively incising 'valley'. Note the decrease in the elevation of the 'coastal plain' as a result of base-level fall. The older coastal plain, which existed during the early stage of incision, is now preserved as a stranded terrace.

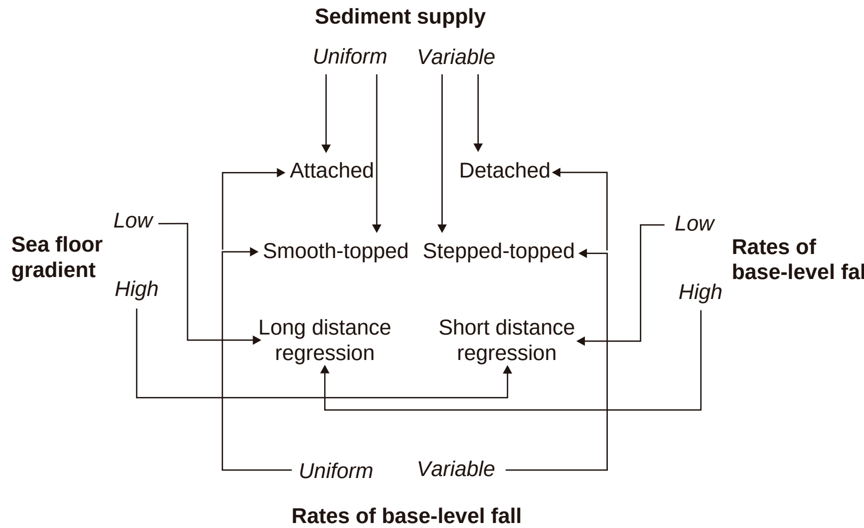


FIGURE 3.33 Stratal architecture of shallow-marine forced regressive deposits, as a function of sediment supply, rates of base-level fall and gradient of the seafloor. The interplay of these variables may result in a variety of possibilities, with the prograding forced regressive lobes being attached or detached, stepped-topped or smooth-topped, and spread over short or long distances (see Posamentier and Morris, 2000, for a more detailed discussion).

consumed by sedimentation, aggradation is accompanied by sediment bypass (the surplus of sediment for which no accommodation is available), and a progradation of facies occurs (Fig. 3.35). Such seaward shifts of facies result in the formation of conformable successions, which consist typically of coarsening-upward shallow-marine deposits topped by coastal to fluvial facies (Fig. 3.36). Normal regressive successions may develop in both river-mouth (deltaic) and open coastline settings. In the former case, the vertical profile records a shift from prodelta, to delta front and delta plain facies (Fig. 3.36), whereas in the latter setting the change is from shelf to shoreface and overlying beach and alluvial facies (Figs. 3.37 and 3.38).

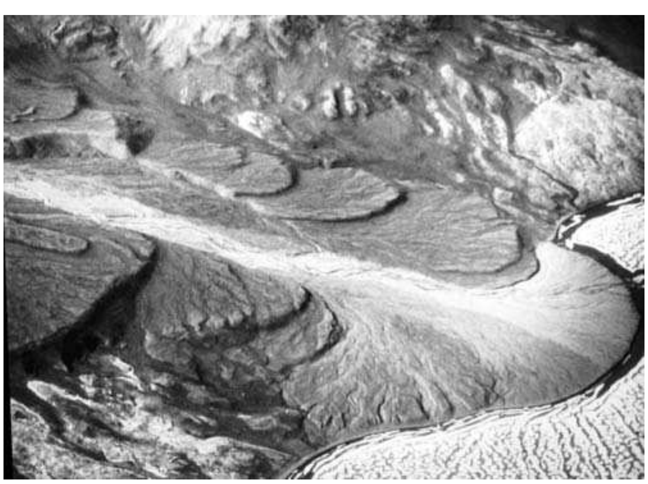


FIGURE 3.34 Modern forced regressive delta showing offlapping stratal stacking patterns (photo courtesy of J. England). In this case, the fall in base level is triggered by post-glacial isostatic rebound in the Canadian arctics, at a rate that exceeds the rate of present day sea-level rise.

The dip angle of the prograding clinoforms (Fig. 3.35) depends on the dominant controls on sediment distribution in the subtidal area, as well as on sediment supply. In the case of wave-dominated open coastlines, or wave-dominated deltas, the angle of repose is very low, averaging 0.3° (mean gradient of the wave equilibrium profile). This angle is steeper in the case of river-dominated deltas, ranging from less than a degree (where rivers bring a significant amount of fine-grained suspension load, and the sediment transport in the delta front environment is primarily attributed to low-density turbidity flows) to approximately 30° (Gilbert-type deltas, where the riverborne sediment is dominantly sandy and its transport within the delta front environment is largely linked to the manifestation of grain flows). In either case, the creation of accommodation in the coastal and adjacent fluvial and shallow-marine regions is prone to aggradation along the entire nearshore profile, and hence no significant fluvial or wave scouring are expected to be associated with this type of shoreline shift (Fig. 3.35). As a result, normal regressive shoreface or delta front deposits are *gradationally based* (Fig. 3.36), in contrast with the forced regressive shoreface or wave-dominated delta front facies which are *sharp-based* (Figs. 3.27 and 3.28).

The process of coastal aggradation, in response to rising base level, also confers another important diagnostic feature that separates normal regressive from forced regressive deposits (Figs. 3.27 and 3.35). As accommodation is positive in the coastal region, a *topset* of intertidal to supratidal deposits (delta plain in river-mouth settings, Fig. 3.36; or beach/strandplain sediments in open shoreline settings, Fig. 3.38) accumulates and progrades on top of the shallow-marine delta front/shoreface facies (Fig. 3.35). Such a topset is absent in the case of forced regressions, where the subtidal facies

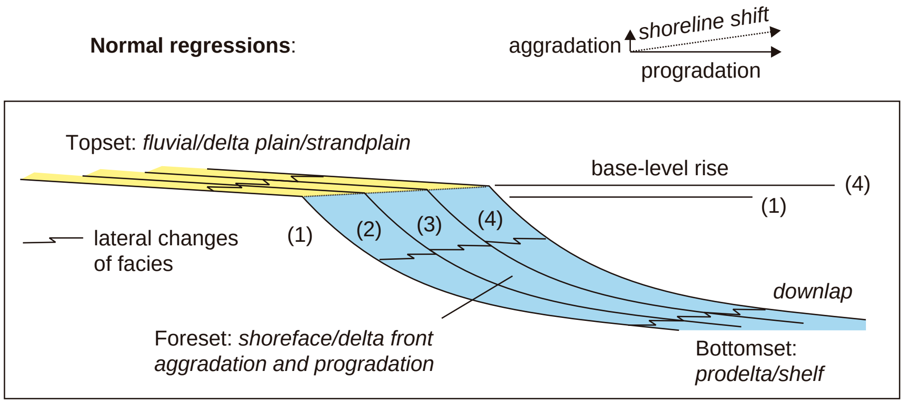


FIGURE 3.35 Shoreline trajectory in normal regressive settings, defined by a combination of progradation and aggradation in fluvial to shallow-marine systems. Normal regressions are driven by sediment supply, where the rates of base-level rise at the shoreline are outpaced by sedimentation rates. Normal regressions occur during early and late stages of base-level rise, when the rates of creation of accommodation are low (Fig. 3.19). Progradation rates are generally low. Normal regressions are prone to aggradation in fluvial, coastal (delta plains in river-mouth settings, or strandplains along open shorelines), and marine environments.

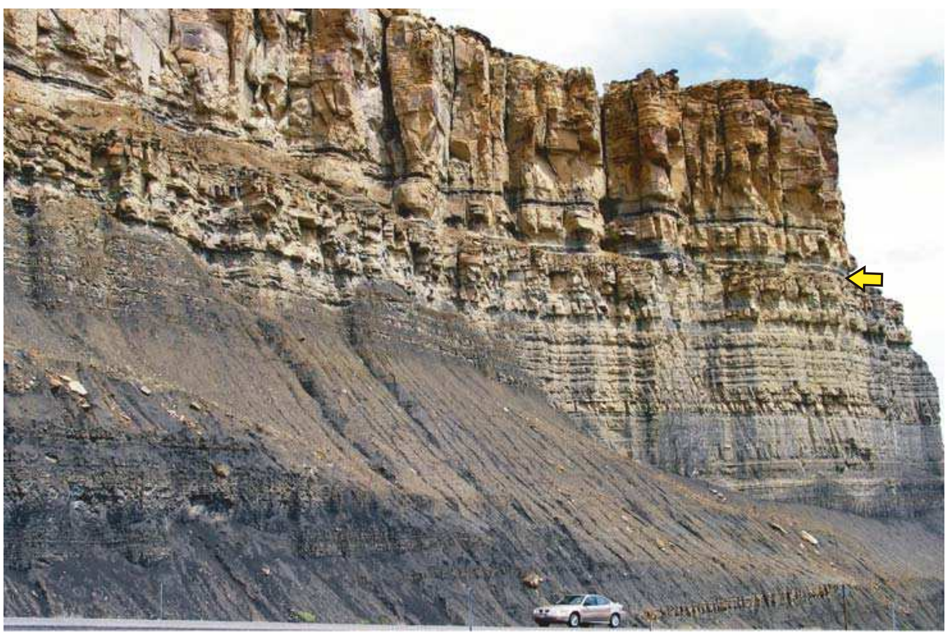


FIGURE 3.36 Normal regressive deltaic succession (river-dominated delta), showing a conformable transition from shallow-marine muds and sands (shelf, prodelta, delta front) to coastal and fluvial deposits (Ferron Sandstone, Utah). The arrow points at the conformable facies contact between delta front sands and the overlying coal-bearing delta plain and fluvial facies. This facies contact marks the base of the deltaic topset (Fig. 3.35).

FIGURE 3.37 Aggrading upper shoreface sandstones in a wave-dominated open coastline setting. These wave ripple-marked strata are interpreted as part of a late rise (highstand) normal regressive systems tract (Rubidge *et al*., 2000). Waterford Formation (Late Permian), Ecca Group, Karoo Basin.

FIGURE 3.38 Aggrading beach deposits in a normal regressive setting. The sands are massive, with low-angle stratification, typical of foreshore open-shoreline systems. The beach sands overlie coarsening-upward shelf to shoreface deposits (in subsurface in this particular location), and are overlain by fluvial floodplain facies. The latter contact is sharp but conformable. Uppermost Bearpaw Formation sands (Early Maastrichtian), Castor area, Western Canada Sedimentary Basin.

offlap and are truncated by processes of subaerial erosion (Fig. 3.27). The thickness of topset successions varies with the case study, depending on the duration of normal regression, the rates of coastal aggradation, and available sediment supply. The topset may be identified in core or outcrop based on facies analysis, but its recognition on seismic lines as a distinct unit may or may not be possible, depending on seismic resolution relative to the unit's thickness (Fig. 3.22).

The surface that separates the topset package from the underlying subtidal deposits is always represented by a conformable (and diachronous, with the rate of shoreline regression) facies contact (dotted line in Fig. 3.35; Fig. 3.36). The upper boundary of the topset unit may also be conformable, where no subsequent erosion reworks it (e.g., in the case of early rise 'lowstand' normal regressions, where the topset is overlain by transgressive fluvial and/or estuarine strata), but often it is scoured by subaerial erosion (e.g., late rise 'highstand' topsets truncated by subaerial unconformities) or transgressive reworking (e.g., early rise 'lowstand' topsets truncated by tidal- or wave-ravinement surfaces). The preservation potential of topset packages is higher in the case of early rise ('lowstand') normal regressive deposits, as the creation of accommodation continues following the maximum regression of the shoreline, and lower in the case of late rise ('highstand') normal regressive successions which are followed by stages of base-level fall and potential subaerial erosion.