CHAPTER

3

Accommodation and Shoreline Shifts

INTRODUCTION

This chapter introduces a set of core concepts relevant to sequence stratigraphy, including sediment accommodation, shoreline shifts, and the controls thereof, whose understanding is fundamental before approaching the more specialized topics related to sequence stratigraphic surfaces, systems tracts, and stratigraphic sequences. These basic concepts allow one to see why and how sequence stratigraphy works, and what the 'engine' is that unifies stratal stacking patterns across a basin into coherent models of stratigraphic architecture.

One of the key premises of sequence stratigraphy, which also served as a main incentive for its conceptual development, is that this approach allows for facies predictions from the confines of individual depositional systems to the scale of entire sedimentary basin fills (Fig. 1.2). This premise implies that depositional trends within all environments established within a sedimentary basin are synchronized to a large extent, being governed by external (allogenic) mechanisms that operate from basinal to global scales. This allogenic 'umbrella' controls regional depositional trends, and provides the basis for the definition of systems tracts and the development of sequence models of facies predictability.

Changes in depositional trends arguably represent the essence of sequence stratigraphic research (Fig. 1.3), and reflect the interplay between the space available for sediments to fill and the amount of sediment influx. The space available for sediments to fill (i.e., 'accommodation') is in turn modified by the basin-scale influence of allogenic controls, which thus provide the common thread that links the depositional trends across a sedimentary basin, from its fluvial to its marine reaches. At the limit between nonmarine and marine environments, the shoreline trajectory defines the type of depositional trend established at any given time. Shoreline trajectories are thus central to sequence stratigraphy, and their changes through time control the timing of all systems tracts and sequence stratigraphic surfaces. The effects of allogenic controls on sedimentation, the space available for sediments to fill, and shoreline trajectories and associated depositional trends are thus intricately related and form the foundation of the sequence stratigraphic approach.

ALLOGENIC CONTROLS ON SEDIMENTATION

Significance of Allogenic Controls

Sedimentation is generally controlled by a combination of autogenic and allogenic processes, which determine the distribution of depositional elements within a depositional system, as well as the larger-scale stacking patterns of depositional systems within a sedimentary basin.

Autogenic processes (e.g., self-induced avulsion in fluvial and deep-water environments) are particularly important at sub-depositional system scale, and are commonly studied using the methods of conventional sedimentology and facies analysis. Allogenic processes, on the other hand, are directly relevant to sequence stratigraphy, as they control the larger-scale architecture of the basin fill.

Allogenic controls provide the common platform that connects and synchronizes the depositional trends recorded at any given time in all environments established within a sedimentary basin, thus allowing for sequence stratigraphic models to be developed at the basin scale. This in turn is the key for the facies predictability applications of sequence stratigraphy, which are so valuable to both academic and

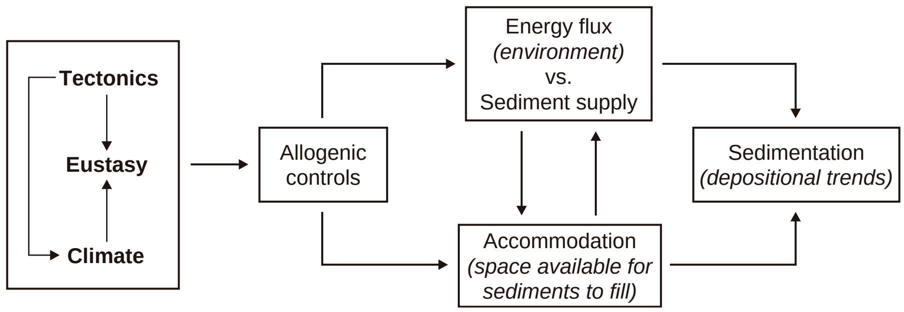


FIGURE 3.1 Allogenic controls on sedimentation, and their relationship to environmental energy flux, sediment supply, accommodation, and depositional trends (modified from Catuneanu, 2003). In any depositional environment, the balance between energy flux and sediment supply is key to the manifestation of processes of sediment accumulation or reworking. Besides tectonics, additional processes such as thermal subsidence (crustal cooling), sediment compaction, water-depth changes, isostatic, and flexural loading, also contribute to the total subsidence or uplift in the basin. Accommodation is affected by the balance between energy flux and sediment supply (i.e., increased energy 'erodes' accommodation; increased sediment supply adds to the amount of available accommodation), but it is also independently controlled by external factors such as eustasy and tectonism. At the same time, changes in accommodation controlled directly by external factors may alter the balance between energy flux and sediment supply at any location within the basin (e.g., deepening of the water as a result of sea-level rise lowers the energy flux at the seafloor). The interplay of all allogenic controls on sedimentation, as reflected by changes in accommodation and energy flux/sediment supply, ultimately determines the types of depositional trends established within the basin.

industry practitioners. The basic allogenic controls on sedimentation include the climate, tectonics, and sea-level changes, and their relationship with the environmental energy flux, sediment supply, accommodation, and depositional trends is summarized in Fig. 3.1. Tectonics is commonly equated with basin subsidence, but additional processes such as crustal cooling, crustal loading, water-depth changes and sediment compaction may also bring important contributions to the total subsidence in the basin. The dissolution and/or withdrawal of evaporites at depth have also been documented as possible subsidence mechanisms (e.g., Waldron and Rygel, 2005). Eustasy and tectonics both control directly the amount of space (accommodation) that is available for sediments to accumulate. Climate mainly affects accommodation *via* eustasy, as for example during glacio-eustatic falls and rises in sea level, but also by changing energy levels in continental to marine environments (e.g., seasonal fluvial discharge; wind regimes in eolian environments; fairweather *vs.* storm waves and currents in marine or lacustrine settings). The effect of climate is also reflected in the amount of sediment supply, by modifying the efficiency of weathering, erosion, and sediment transport processes.

It is important to note that the allogenic controls are 'external' relative to the sedimentary basin, but not necessarily independent of each other (Fig. 3.1). Eustatic fluctuations of global sea level are controlled by both tectonic and climatic mechanisms, over various time scales (Fig. 3.2). Global climate changes are primarily controlled by orbital forcing (e.g., Milankovitch cycles with periodicities of 10^4–10^5 years; Fig. 2.16), but at more local scales may also be triggered by tectonic processes such as the formation of thrust-fold belts that may act as barriers for atmospheric circulation. Tectonism is primarily driven by forces of internal Earth dynamics, which are expressed at the surface by plume or plate tectonic processes. There is increasing

Hierarchical order	Duration (My)	Cause
First order	200-400	Formation and breakup of supercontinents
Second order	10-100	Volume changes in mid-oceanic spreading centers
Third order	1-10	Regional plate kinematics
Fourth and fifth order	0.01-1	Orbital forcing

FIGURE 3.2 Tectonic and orbital controls on eustatic fluctuations (modified from Vail *et al.*, 1977, and Miall, 2000). Local or basin-scale tectonism is superimposed and independent of these global sea-level cycles, often with higher rates and magnitudes, and with a wide range of time scales.

evidence that the tectonic regimes which controlled the formation and evolution of sedimentary basins in the more distant geological past were much more erratic in terms of origin and rates than formerly inferred solely from the study of the Phanerozoic record (e.g., Eriksson *et al*., 2004; Eriksson *et al*., 2005a, b). The more recent basin-forming processes seem to be largely related to a rather stable plate tectonic regime, whereas the formation of Precambrian basins reflects a combination of competing mechanisms, including magmatic-thermal processes ('plume tectonics') and a more erratic plate tectonic regime (Eriksson and Catuneanu, 2004b). These insights offered by the Precambrian record are critical for extracting the essence of how one should categorize the stratigraphic sequences that can be observed within a sedimentary succession at different scales. This issue is discussed in more detail in the chapter dealing with the sequence stratigraphic hierarchy (Chapter 8).

Signatures of Allogenic Controls

The signature of the eustatic control on sedimentation may be recognized from (1) the tabular geometry of sedimentary sequences, suggesting that accommodation was created in equal amounts across the entire basin; (2) the synchronicity of depositional and erosional events across the entire basin, and beyond; and (3) the lack of source area rejuvenation, as it may be suggested by the absence of conglomerates along the proximal rim of the basin. The sea-level control on sedimentation has been documented in numerous case studies, with a degree of confidence that improves with decreasing stratigraphic age (e.g., Suter *et al*., 1987; Plint, 1991; Miller *et al*., 1991, 1996, 1998, 2003, 2004; Long, 1993; Locker *et al*., 1996; Stoll and Schrag, 1996; Kominz *et al*., 1998; Coniglio *et al*., 2000; Kominz and Pekar, 2001; Pekar *et al*., 2001; Posamentier, 2001; Olsson *et al*., 2002). Estimates of sea-level changes in the geological record have been obtained in recent years by backstripping, accounting for water-depth variations, sediment loading, compaction, basin subsidence and foraminiferal $\delta^{18}O$ data. Studies of the 'ice-house world' of the past 42 Ma have demonstrated a relationship between depositional sequence boundaries and global $\delta^{18}O$ increases, linking stages of sequence-boundary formation with glacio-eustatic sea-level lowerings (e.g., Miller *et al*., 1996, 1998). Even for the 'greenhouse world' of the Late Cretaceous—Early Cenozoic interval (prior to 42 Ma), backstripping studies on the New Jersey Coastal Plain, which was subject to minimal tectonic activity, indicate that sea-level fluctuations occurred with amplitudes of > 25 m on time scales of < 1 Ma (Miller *et al*., 2004). Such studies have questioned the assumption of a completely ice-free world during the Cretaceous interval, and have revamped the importance of sea-level changes on accommodation and sedimentation (e.g., Stoll and Schrag, 1996; Price, 1999; Miller *et al*., 2004).

Tectonism is a common control in any sedimentary basin, and its manifestation leads to (1) a wedge-shaped geometry of sedimentary sequences, due to differential subsidence; (2) the accumulation of coarser-grained facies along the proximal rim of the basin in relation to the rejuvenation (uplift) of the source areas; (3) variations in the maximum burial depths of the sedimentary succession across the basin, as can be determined from the study of late diagenetic minerals, fluid inclusions, vitrinite reflection, apatite fission track, etc.; (4) changes in syndepositional topographic slope gradients, as inferred from the shift in fluvial styles through time; and (5) changes in the direction of topographic tilt, as inferred from paleocurrent measurements. The role of tectonic mechanisms in the development of stratigraphic cycles and unconformities has been documented for sedimentary basins spanning virtually all stratigraphic ages, from Precambrian to Phanerozoic and present-day depositories. Early assumptions indicated that tectonic processes may operate mainly on long time scales, of > 10^6 years (e.g., Vail *et al*., 1977, 1984, 1991; Haq *et al*., 1987; Posamentier *et al*., 1988; Devlin *et al*., 1993), leaving eustasy as the likely cause of higher-frequency cyclicity, at time scales of 10^6 years or less. Advances in our understanding of tectonic processes have led to the realization that tectonically-driven cyclicity may actually develop over a much wider range of time scales, both greater than and less than 1 Ma (e.g., Cloetingh *et al*., 1985; Karner, 1986; Underhill, 1991; Peper and Cloetingh, 1992; Peper *et al*., 1992, 1995; Suppe *et al*., 1992; Karner *et al*., 1993; Eriksson *et al*., 1994; Gawthorpe *et al*., 1994, 1997; Peper, 1994; Yoshida *et al*., 1996, 1998; Catuneanu *et al*., 1997a, 2000; Catuneanu and Elango, 2001; Davies and Gibling, 2003). Therefore, the eustatic and tectonic mechanisms may compete toward the generation of any order of stratigraphic cyclicity. The challenge in this situation is to evaluate their relative importance on a case by case basis. In this light, it has been noted that the amplitudes of sea-level changes reconstructed by means of backstripping (e.g., Miller *et al*., 1991, 1996, 1998, 2003, 2004; Locker *et al*., 1996; Stoll and Schrag, 1996; Kominz *et al*., 1998; Coniglio *et al*., 2000; Kominz and Pekar, 2001; Pekar *et al*., 2001) are in many cases lower than those interpreted from seismic data (e.g., Haq *et al*., 1987), questioning the accuracy of seismic data interpretations in terms of eustatic sea-level changes (Miall, 1986, 1992, 1994, 1997; Christie-Blick *et al*., 1990; Christie-Blick and Driscoll, 1995). Field observations also indicate that the amount of erosion associated with many sequence-bounding unconformities in tectonically active basins was often greater than the inferred amplitude of eustatic fluctuations, suggesting that the basinward shifts of facies associated with stages of base-level fall are not necessarily related to changes in sea level (e.g., Christie-Blick *et al*., 1990; Christie-Blick and Driscoll, 1995). All these insights re-emphasized the importance of tectonism as a control on accommodation and sedimentation, which, in tectonically active basins, may explain the observed cyclicity at virtually any time scale.

Climate changes within the 10^4–10^5 years Milankovitch band are attributed to several separate components of orbital variation, including orbital eccentricity, obliquity, and precession. Variations in orbital eccentricity, which refers to the shape (degree of stretching) of the Earth's orbit around the Sun, have major periods at around 100 and 413 ka. Changes of up to 3° in the obliquity (tilt) of the ecliptic have a major period of 41 ka. The precession of the equinoxes, which refers to the rotation (wobbling) of the Earth's axis as a spinning top, records an average period of about 21 ka (Fig. 2.16; Imbrie and Imbrie, 1979; Imbrie, 1985; Schwarzacher, 1993). In addition to Milankovitch-band processes, other astronomical forces may affect the climate over shorter time intervals, from a solar band (tens to hundreds of years range; e.g., sun-spot cycles), to a high-frequency orbital band (e.g., nutation cycles of the motion of the axis of rotation of the Earth about its mean position, with a periodicity of about 18.6 years) and a calendar band (cyclicity related to seasonal rhythms, such as freeze–thaw, varves, or fluvial discharge cycles, and other sub-seasonal effects driven by the Earth–Moon system interaction) (e.g., Fischer and Bottjer, 1991; Miall, 1997). Fluctuations in the syndepositional paleoclimate may be reconstructed by combining independent research methods such as (1) thin section petrography of the detrital framework constituents in sandstones, looking at the balance between stable and unstable grains; (2) the mineralogy of the early diagenetic constituents, assuming a short lag time between the deposition of the detrital grains and the precipitation of early diagenetic minerals; (3) the isotope geochemistry of early diagenetic cements; and (4) foraminiferal $\delta^{18}O$ data. Each of these techniques may potentially be affected by drawbacks when it comes to the unequivocal interpretation of syndepositional paleoclimates, so their use in conjunction allows for more reliable conclusions (e.g., Khidir and Catuneanu, 2003). The role of climate as a major control on sedimentation has been emphasized in numerous case studies, including Blum (1994), Tandon and Gibling (1994, 1997), Miller *et al*. (1996, 1998), Blum and Price (1998), Heckel *et al*., (1998), Miller and Eriksson (1999), Ketzer *et al*. (2003a, b) and Gibling *et al*. (2005).

Relative Importance of Allogenic Controls

The relative importance of climate, tectonism, and sea-level change on sediment accommodation is illustrated in Fig. 3.3. In marine environments, the balance between eustasy and subsidence changes according to the subsidence patterns that characterize each tectonic setting. For example, the rates of subsidence in extensional settings increase in a distal direction, and the opposite is true for foreland systems (Figs. 2.62 and 2.63). In fluvial environments, the effect of sea-level change

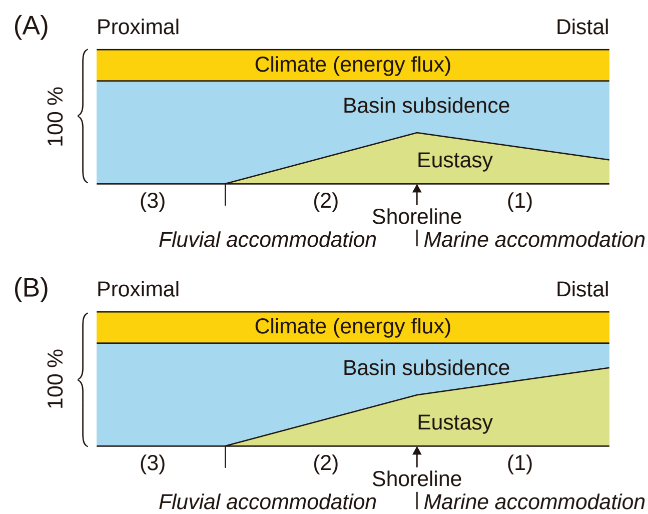


FIGURE 3.3 Relative importance of allogenic controls on accommodation in (A) extensional and (B) foreland basins. Subsidence patterns affect the balance between subsidence and eustasy in marine environments. Sedimentary basins may be subdivided into three distinct areas, based on the dominant controls on accommodation: (1) marine (or lacustrine, if eustasy is substituted with lake level) environments, where the amount of available accommodation is mainly controlled by subsidence and sea-level change; (2) downstream reaches of fluvial environments, which are still affected by sea-level change; and (3) upstream reaches of fluvial environments, unaffected by sea-level change. Note that the vertical scale suggests relative contributions of allogenic controls, and not actual amounts of accommodation. Accommodation increases in a distal direction in extensional basins, and in a proximal direction in foreland settings (Figs. 2.62 and 2.63). Variations in energy flux induced mainly (but not exclusively) by climate may affect accommodation in all environments. The boundaries that separate the relative contributions of eustasy, subsidence and climate may shift depending on local conditions. See also Fig. 3.4 for the actual processes that facilitate the climatic, subsidence and eustatic controls on fluvial and marine accommodation.

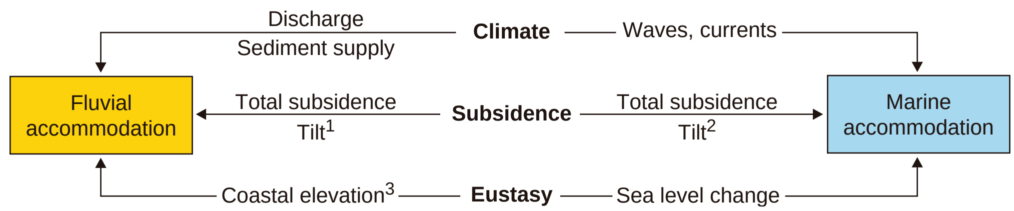


FIGURE 3.4 Processes that enable climatic, subsidence and eustatic controls on fluvial and marine accommodation. Notes: [1]—differential subsidence may modify the water velocity in fluvial systems; [2]—differential subsidence may influence the type of gravity flows that are manifest in marine/lacustrine environments; [3]—changes in coastal elevation may trigger shifts in slope gradients (and corresponding fluvial-energy flux) in the downstream reaches of fluvial systems. Ultimately, all allogenic controls modify the balance between sediment supply and energy flux in each depositional environment, leading to the manifestation of processes of erosion (negative accommodation) or sediment accumulation (positive accommodation).

diminishes in an upstream direction. Beyond the landward limit of eustatic influences, fluvial processes of aggradation or erosion are entirely controlled by climate and tectonism. Fluctuations in environmental energy flux, largely (but not exclusively) controlled by climate over various time scales, also have an impact on the amounts of accommodation that are available in each depositional environment (Figs. 3.3 and 3.4). Increases in energy flux result in losses of available accommodation, whereas decreases in energy flux allow for more sediment accumulation. Such fluctuations in environmental energy may occur from seasonal and sub-seasonal time scales (e.g., seasonal changes in mean precipitation rates and their impact on fluvial discharge, or the effect of fairweather *vs.* storm conditions on marine waves and currents) to longer-term time scales (e.g., Milankovitch cycles of glaciation and deglaciation, and their long-term effects on fluvial discharge).

The total amount of subsidence in the basin is arguably the most important control on accommodation, as the overall geometry of the basin fill ultimately reflects the pattern of basin subsidence (Figs. 2.62 and 2.63). As sea-level change commonly affects accommodation only in restricted portions of a basin (zones 1 and 2 in Fig. 3.3), subsidence also provides a common thread for the general patterns of accommodation changes across the entire basin. These overall trends are modified by fluctuations in energy flux, as explained above, and also by the superimposed effects of sea-level change. Figure 3.3 only provides a schematic illustration of these basic principles, and the boundaries that separate the relative contributions of the main allogenic controls on accommodation may shift as a function of local conditions in each sedimentary basin. These issues are discussed in more detail below, as well as in subsequent chapters of this book.

SEDIMENT SUPPLY AND ENERGY FLUX

Sediment Supply

Sediment supply is an important variable in sequence stratigraphic analyses, and it refers to the amount (or flux) and type (grain size) of sediment that is supplied from source areas to depositional areas by various transport agents, including gravity, water, and wind. The importance of sediment supply in stratigraphy, and especially on the manifestation of transgressions and regressions, was recognized at least since the eighteenth century, when Hutton attributed the migration of shorelines to the shifting balance between riverborne sediment supply and the marine processes of sediment reworking within the receiving basin (in Playfair, 1802). These early ideas have been subsequently refined in landmark publications by Lyell (1868), who related the progradation of deltas to an excess of sediment supply; Grabau (1913), who linked transgressions and regressions to the interplay of sediment supply and the 'depression' caused by subsidence within the receiving basin (precursor of what we call today 'accommodation'); and Curray (1964), who reiterated the role of sediment supply and relative sea level as the primary controls on transgressions and regressions. Following the birth of seismic and sequence stratigraphy in the 1970s and 1980s, the integration of sediment supply in modern stratigraphic analyses has become the norm (e.g., Jervey, 1988; Flemings and Jordan, 1989; Jordan and Flemings, 1991; Swift and Thorne, 1991; Thorne and Swift, 1991; Schlager, 1992, 1993; Johnson and Beaumont, 1995; Helland-Hansen and Martinsen, 1996; Catuneanu *et al.*, 1998b; Cross and Lessenger, 1999; Paola *et al.*, 1999; etc.)

Sediment supply is primarily a by-product of climate and tectonism. A wetter climate increases the amount of sediment supply, *via* increased efficiency of weathering and erosion, and so does the process of tectonic uplift *via* source area rejuvenation. The transport capacity of the transport agents may also increase under wetter climatic conditions (e.g., higher river discharge) and as a result of increased slope gradients due to tectonic tilt. In addition to the direct controls exerted by climate (e.g., *via* precipitation rates, temperature fluctuations) and source area tectonism, the substrate lithology and the vegetation cover of the sediment source areas also influence the flux and grain size of the sediment transported by rivers or wind (Blum, 1990; Einsele, 1992; Miall, 1996).

Sediment supply is critical to the stratigraphic architecture of any sedimentary basin, as it is one of the fundamental variables that determine the type of depositional trends in all fluvial to marine environments (Fig. 3.1). Once accommodation is made available by subsidence or sea-level change, the lithology, location, and stacking patterns of depositional elements are largely a function of the volume and type of sediment supply. At the same time, as a consequence of sediment accumulation, more accommodation is created as a result of isostatic sediment loading (Matthews, 1984; Schlager, 1993). The relationship between sedimentation and accommodation is thus a two-way process/response interaction, as sedimentation does not only consume accommodation made available by other mechanisms, but may also create additional space as sediment aggradation/loading proceeds. This fact is valid for all fluvial to marine environments, as isostatic sediment loading contributes to the total subsidence in the basin that is otherwise caused by tectonic, thermal, or sediment compaction processes.

Sediment Supply *vs.* Environmental Energy Flux

Variations in sediment supply may also be conducive to the manifestation of depositional processes of aggradation or erosion, but the significance of such variations is relative to the energy flux of each particular environment. In marine basins, sediment is transported by a variety of subaqueous currents, including wave-induced (longshore, rip), tidal, contour, or gravity flows, and the nature of processes at the seafloor (sediment accumulation *vs.* erosion) is dictated by the balance between the energy (transport capacity) of the current and its sediment load. A marine current that has more energy than that required to transport its sediment load (i.e., underloaded flow) erodes the seafloor, whereas a current that drops its energy below the level that is required to transport its entire sediment load (i.e., overloaded flow) results in aggradation. The same principle applies to fluvial and eolian systems, where the balance between the energy of the transport agent (water, wind) and its sediment load controls surface processes of aggradation or downcutting (Figs. 3.5 and 3.6). Even though the role of sediment supply in reducing or increasing the amount of available accommodation is not captured in Fig. 3.3, it is implied that the 'energy flux' factor stands for this dynamic energy/sediment balance, as an increase in energy flux *relative to sediment supply* leads to a loss of accommodation, and a decrease in energy flux *relative to sediment supply* results in a gain of accommodation.

Shifts in the balance between energy flux and sediment supply may be caused by each of the allogenic controls on accommodation (climate, subsidence/uplift, or sea-level change; Figs. 3.1 and 3.4), either independently or in any combination thereof. In the early days of sequence stratigraphy it was generally implied that sea-level change exerts the main control on stratigraphic architecture, and implicitly on processes of aggradation or erosion (Vail *et al*., 1977; Posamentier *et al*., 1988). In the 1990s, tectonism was emphasized as an equally important control, and the combination of eustatic and tectonic processes was invoked as the key driving force behind surface processes of deposition or sediment reworking (Hunt and Tucker, 1992; Posamentier and Allen, 1999). Climate was generally left out of sequence stratigraphic models, as it was the most difficult allogenic mechanism to quantify, but its effect on sediment aggradation or erosion was proven to be as important as the control exerted by eustasy or tectonism (Blum, 1994; Blum and Price, 1998; Gibling *et al*., 2005). Syndepositional surface processes of aggradation or erosion ultimately reflect the interplay of all three allogenic controls, whose effects may enhance or cancel each other out depending on local circumstances. The Late Cenozoic fluvial record of the U.S. Gulf Coast provides an example where climate and sea-level change promoted opposite depositional trends during stages of glaciation and interglaciation. In this case study, the climatic control on fluvial discharge outpaced the effects of sea-level change, leading to fluvial aggradation during glacial periods (driven by a drop in fluvial discharge/energy flux, in spite of the coeval glacio-eustatic fall) and fluvial erosion during interglacial stages (as a result of increased fluvial discharge due to ice melting, and despite the rise in sea level) (Blum, 1990, 1994). Similar examples of fluvial incision triggered by climate-controlled increases in discharge during times of glacial melting and global sea-level rise are also found in western Canada (Fig. 3.7).

Ultimately, *all processes of aggradation or erosion are linked to the shifting balance between environmental energy*

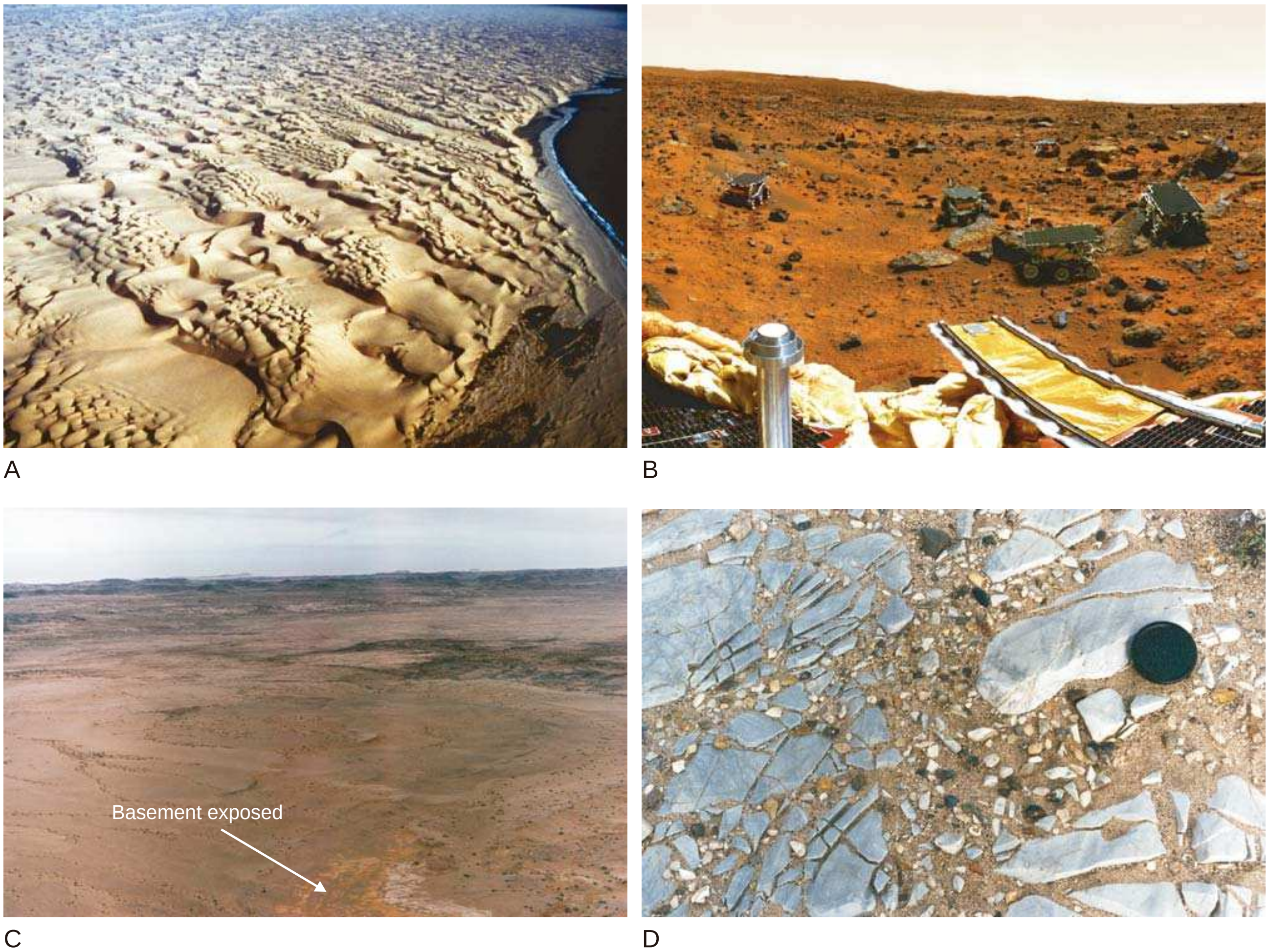


FIGURE 3.5 Surface processes that reflect the dynamic interplay of sediment supply and wind energy in eolian environments. Sediment supply exceeding the transport capacity (energy) of winds results in the accumulation of sand as sheets or dunes, depending on flow regimes. Winds stronger relative to their sediment load lead to erosion and the formation of deflation surfaces. A—sand dunes in the Namib Desert (Namibia), formed as a result of abundant sediment supply (sediment supply > wind energy; photo courtesy of Roger Swart); B—deflation surface on Mars (wind energy > sediment supply; photo courtesy of NASA); C—deflation surface in the Namib Desert, Namibia (wind energy > sediment supply); D—deflation surface in the Namib Desert, Namibia (detail showing the concentration of heavy minerals as lag deposits on top of the Precambrian dolomites basement rocks).

flux and sediment supply (i.e., aggradation occurs only where sediment supply outpaces energy flux, and erosion occurs only where energy outpaces sediment load). In turn, accommodation is closely related to the shifting balance between energy flux and sediment supply, both as a control but also as a controlled variable (see the two-way relationship indicated in Fig. 3.1). On the one hand, the balance between energy flux and sediment supply affects the amounts of available accommodation, although accommodation is also independently controlled by other factors as well (Figs. 3.1, 3.3, and 3.4). As a general rule, accommodation is inversely proportional to energy flux (i.e., an increase in energy 'erodes' accommodation) and directly proportional to sediment supply (i.e., an increase in sediment supply adds to the amount of available accommodation; Fig. 3.6). On the other hand, changes in accommodation controlled directly by allogenic mechanisms may also affect the balance between energy flux and sediment supply within the basin. For example, an increase in accommodation, such as in response to subsidence or sea-level rise tends to reduce the energy level at the seafloor, thus promoting sediment aggradation. This explains why, in virtually any situation, depositional trends may ultimately be related to shifts in the balance between energy flux

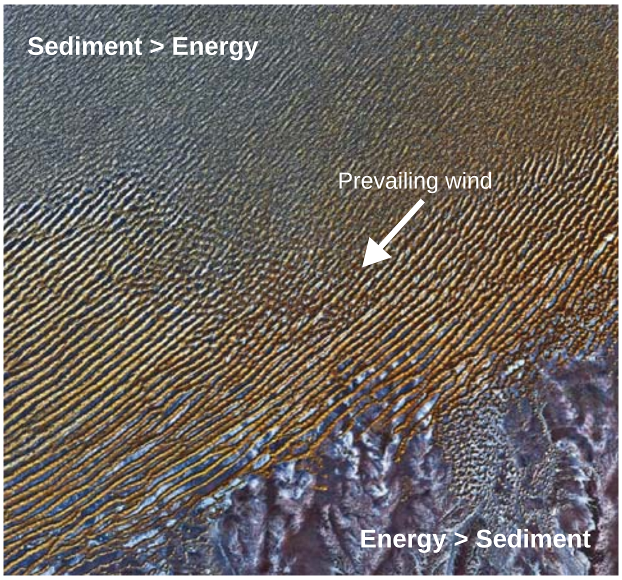


FIGURE 3.6 Satellite image of southern Arabian Peninsula showing a gradual shift in the balance between sediment supply and wind energy from the upper-left corner of the image (sediment supply dominant) to the lower-right corner of the image (wind energy dominant). Accommodation is positive where sediment supply dominates, leading to the accumulation of sand on top of the basement rocks. Accommodation is negative where energy is in excess, leading to the exposure and erosion of the basement rocks. The longitudinal dunes shown in this image are parallel to the prevailing northeasterly winds, and are the equivalent of parting lineation of the upper flow regime of subaqueous bedforms. The wind regime in this case is higher energy relative to the wind regime that generated the transversal dunes shown in Fig. 3.5A, which are the equivalent of dunes of the lower flow regime of subaqueous bedforms.

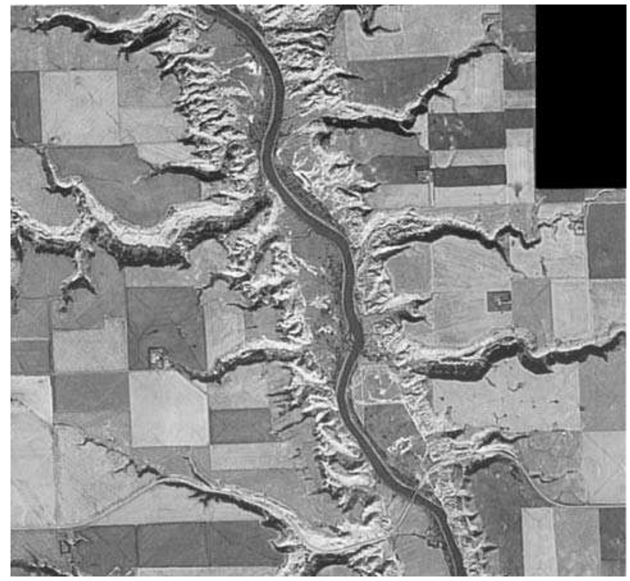


FIGURE 3.7 Aerial photograph showing the modern incised valley of the Red Deer River (Alberta). Note farm houses for scale. Tributaries are also incised, which is one of the diagnostic features of incised valleys. The incision of the Red Deer River valley was climate-controlled, and caused by the significant increase in fluvial discharge associated with the rapid glacial melting during the Late Pleistocene.

and sediment supply. The two-way process/response relationship between energy flux/sediment supply, on the one hand, and accommodation, on the other, is illustrated in Fig. 3.1.

A simple illustration of how a shifting balance between sediment supply and environmental energy flux may affect accommodation and depositional processes in a shallow-marine setting is presented in Fig. 3.8. The scenario in Fig. 3.8 assumes that sediment is supplied by a river that flows along its graded profile, to a stable coastline that is not affected by subsidence or sea-level changes. The elimination of the effects of subsidence and sea-level change on accommodation allows for a direct evaluation of the depositional processes that take place in this shallow-marine environment in response to the interplay of sediment supply and wave energy. If sediment supply and environmental energy flux are in perfect balance (case A in Fig. 3.8), all sediments will bypass this area, without erosion or aggradation, being removed by longshore drift. In this case accommodation is zero, in spite of the available water column in the marine environment, and the base level is superimposed on the seafloor—in other words, the seafloor corresponds to a graded profile. If sediment supply outpaces the capacity of the environment to remove it, sediment aggradation and progradation will occur (case B in Fig. 3.8). In this case, base level is above the seafloor and accommodation is positive. Where the energy of the environment outpaces sediment supply, erosion of the seafloor will occur (case C in Fig. 3.8). In this case, base level is below the seafloor, accommodation is negative, and coastline erosion may lead to the retrogradation of the shoreline. An important lesson from this diagram is that the amount of available *accommodation is not measured to the sea level*, but rather to a graded profile (base level) that may be in any spatial relationship with the sea level and the seafloor. The situation depicted in Fig. 3.8 is a simplification of the common reality, which is that other factors, such as subsidence and sea-level change, may also affect accommodation in parallel with (and independent of) fluctuations in energy and/or sediment supply (Fig. 3.8).

This discussion indicates that accommodation and sediment supply are not independent variables, as they are often in a process/response relationship that is modulated by environmental energy flux. Consequently, the axiom that the sequence stratigraphic architecture

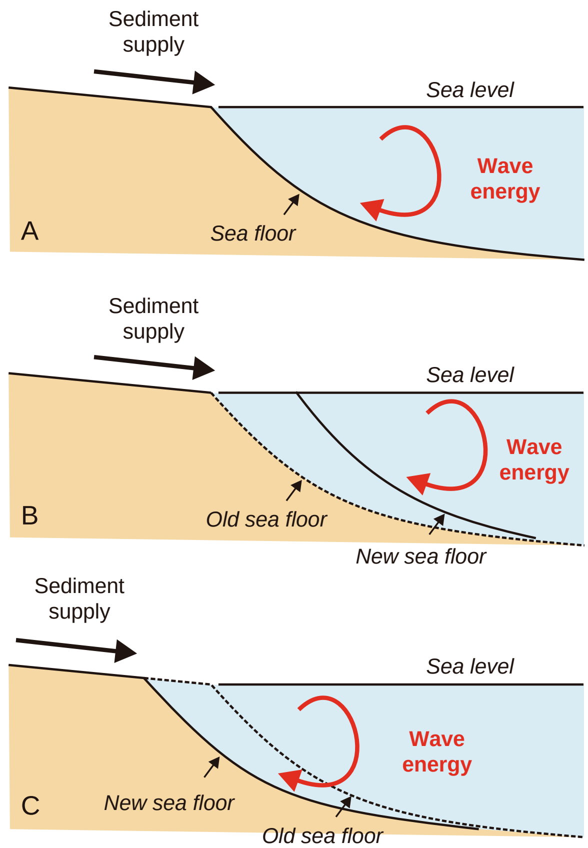


FIGURE 3.8 Relationship between energy flux, sediment supply, base level and accommodation in a shoreface environment that is not affected by subsidence or sea-level change. A—sediment supply is perfectly balanced by wave energy. In this case, all sediment bypasses the area, base level is superimposed on the seafloor, and accommodation is zero; B—sediment supply outpaces wave energy. In this case, sediment aggradation and progradation take place, base level is above the seafloor (superimposed on the sea level, as drafted in the diagram), and accommodation is positive; C—sediment supply is outpaced by wave energy. In this case, coastal and seafloor erosion take place, base level is below the seafloor, and accommodation is negative. Note that *accommodation is not measured to the sea level*, but rather to a graded profile (base level) that may be in any spatial relationship with the sea level and the seafloor. Where accommodation is not affected by subsidence or sea-level change, it is entirely controlled by fluctuations in energy flux and sediment supply. Also note that no confusion should be made between accommodation (space available for *sediments* to fill, measured from the seafloor to the base level) and water depth (space available for *water* to fill, measured from the seafloor to the sea level). For example, more volume is made available for water to fill in case C, but accommodation is negative due to the exceedingly strong wave energy.

is controlled by the interplay between the rate of change in accommodation and the rate of sediment supply (e.g., Schlager, 1993) is only valid as an approximation, since the two variables depend on each other. For this axiom to be true, the approximations being made are that accommodation is measured to the sea level, rather than to the base level (in which case accommodation becomes independent of sediment supply), and that sediment supply is proportional to the sedimentation rates. In reality, none of these approximations are entirely accurate, as discussed above and also in more detail in the following section of this chapter. One has to keep in mind the difference between *sediment supply*, which is measured as a flux, and *sedimentation* (rate), which is measured as a change in vertical distance at any location. Depending on energy flux conditions, a high sediment supply does not necessarily translate into high rates of sedimentation. While accommodation depends on sediment supply, it is measured independently of sedimentation. Therefore, the correct relationship in terms of the controls on stratigraphic architecture is portrayed by the interplay between the rate of change in accommodation and the rate of sedimentation, as both are measured, independently of each other, in units that reflect changes in vertical distance at any particular location. Further discussion on this topic is provided in the following section of this chapter.

The amounts of available marine accommodation may be modified by all three allogenic controls, whose relative importance varies with the basin (Fig. 3.3). Fluvial processes of aggradation or erosion (positive or negative fluvial accommodation, respectively) are increasingly influenced by sea-level change towards the shoreline and by climate and tectonism towards the source areas (Blum, 1990; Posamentier and James, 1993). In nonmarine regions, eustasy is therefore a more important downstream factor, whose importance diminishes in a landward direction, whereas climate and tectonism compensate this trend by becoming increasingly important upstream (Fig. 3.3). More details about the intricate process/response relationship between the allogenic controls, accommodation, and sedimentation are provided in the following sections of this chapter, as well as throughout the book.

SEDIMENT ACCOMMODATION

Definitions—Accommodation, Base Level, and Fluvial Graded Profiles

The concept of sediment 'accommodation' describes the amount of space that is available for sediments to fill, and it is measured by the distance between base level and the depositional surface (Jervey, 1988). This concept was initially applied to marine environments, as a tool to enable mathematical simulations of progradational basin-filling on divergent continental margins (Jervey, 1988). In this context, base level was equated, at first approximation, with sea level, and hence the original definition of 'accommodation' did not require further explanations of what the meaning of 'base level' may be in continental environments. It is now widely agreed that accommodation may be made available in both fluvial and marine environments by the combined effects of climate, tectonism, and sea-level change (Fig. 3.3). The expansion of the concept of accommodation into the nonmarine portion of sedimentary basins brought about further scrutiny of the concept of base level, which led to conflicting ideas and terminology (Shanley and McCabe, 1994; Fig. 3.9).

Base level (of deposition or erosion) is generally regarded as a global reference surface to which long-term continental denudation and marine aggradation tend to proceed. This surface is dynamic, moving upward and downward through time relative to the center of Earth in parallel with eustatic rises and falls in sea level. For simplicity, base level is often approximated with the sea level (Jervey, 1988; Schumm, 1993). In reality, base level is usually below sea level due to the erosional action of waves and marine currents (Fig. 3.8). This spatial relationship between sea level and base level is

Base level (Twenhofel, 1939): highest level to which a sedimentary succession can be built.

Base level (Sloss, 1962): an imaginary and dynamic equilibrium surface above which a particle cannot come to rest and below which deposition and burial is possible.

Base level (Bates and Jackson, 1987): theoretical limit or lowest level toward which erosion of the Earth's surface constantly progresses but rarely, if ever, reaches. The general or ultimate base level for the land surface is sea level.

Base level (Jervey, 1988): ... is controlled by sea level and, at first approximation, is equivalent to sea level ... although, in fact, a secondary marine profile of equilibrium is attained that reflects the marine-energy flux in any region.

Base level (Schumm, 1993): the imaginary surface to which subaerial erosion proceeds. It is effectively sea level, although rivers erode slightly below it.

Base level (Cross, 1991): a surface of equilibrium between erosion and deposition.

Base level (Cross and Lessenger, 1998): a descriptor of the interactions between processes that create and remove accommodation space and surficial processes that bring sediment or that remove sediment from that space.

Base level (Posamentier and Allen, 1999): the level that a river attains at its mouth (i.e., either sea level or lake level), and constitutes the surface to which the equilibrium profile is anchored.

There are two schools of thought regarding the concept of base level:

(1) Base level is more or less the sea level, although usually below it due to the action of waves and currents. The extension of this surface into the subsurface of continents defines the ultimate level of continental denudation. On the continents, processes of aggradation versus incision are regulated via the concept of graded (equilibrium) fluvial profile. Graded fluvial profiles meet the base level at the shoreline.

(2) The concept of base level is generalized to define the surface of balance between erosion and sedimentation within both marine and continental areas (the "stratigraphic" base level of Cross and Lessenger, 1998). In this acceptance, the concept of graded fluvial profile becomes incorporated within the concept of base level. The stratigraphic base level will thus include a continental portion (fluvial base level = graded fluvial profile) and a marine portion (marine base level ~ sea level).

The drawback of the second approach is that fluvial base-level shifts are controlled by marine base-level shifts, especially in the downstream reaches of the river system, and hence the two concepts are in a process/response relationship. This suggests that it is preferable to keep these two concepts separate as opposed to incorporating them into one "stratigraphic base level". This is the approach adopted in this book, where the fluvial base level is referred to as the fluvial graded profile, and the marine base level is simply referred to as the base level.

FIGURE 3.9 Definitions of the concept of base level.

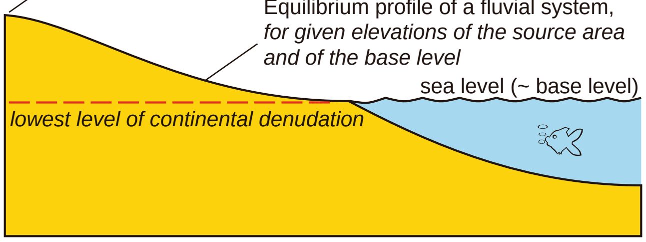


FIGURE 3.10 The concept of base level, defined as the lowest level of continental denudation (modified from Plummer and McGeary, 1996). Graded (equilibrium) fluvial profiles meet the base level at the shoreline. As the elevation of source areas changes in response to denudation or tectonic forces, graded fluvial profiles adjust accordingly. Graded profiles also respond in kind to changes in base level. See also Fig. 3.9 for alternative definitions of base level.

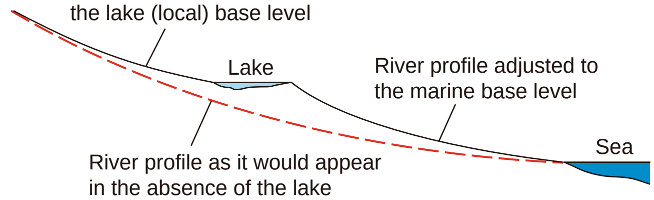


FIGURE 3.11 Marine and local base levels as illustrated by a river flowing into a lake and from the lake into the sea (modified from Press and Siever, 1986). In each river segment, the graded profile adjusts to the lowest level it can reach.

also supported by the fact that rivers meeting the sea erode below sea level (Schumm, 1993), i.e., to the base level (Fig. 3.9).

Figure 3.10 shows a marine to continental area, in which base level is approximated with sea level. The base level may be projected into the subsurface of the continents, marking the lowest level of subaerial erosion (Plummer and McGeary, 1996). The surface topography tends to adjust to base level by long-term continental denudation. Between the source areas that are subject to denudation and the marine shorelines, processes of nonmarine aggradation may still take place when the amount of sediment load exceeds the transport capacity (energy flux) of any particular transport agent (gravity-, air-, or water-flows).

Coupled with the concept of base level, fluvial equilibrium (graded) profiles are particularly important to understanding processes of sedimentation in continental areas. For any given elevations of the source area and of the body of water into which the river debouches, fluvial systems tend to develop a dynamic equilibrium in the form of a graded longitudinal profile (Miall, 1996, p. 353). This equilibrium profile is achieved when the river is able to transport its sediment load without aggradation or degradation of the channels (Leopold and Bull, 1979). Rivers that are out of equilibrium will aggrade or incise in an attempt to reach the graded profile (Butcher, 1990, p. 376). In this context, fluvial systems start adjusting to new equilibrium profiles as soon as the elevation of source areas, the level of the body of water into which the river debouches, and/or any shifts in the balance between fluvial-energy flux and sediment load that these changes may trigger, are modified due to factors such as tectonism, climate, or sea-level change. An equilibrium profile may be below or above the land surface (triggering incision or aggradation, respectively), and it merges with the base level at the marine shoreline (Fig. 3.10). In a more general sense, the base level for fluvial systems is represented by the level of any body of water into which a river debouches, including sea level, lake level, or even another river (Posamentier and Allen, 1999; Fig. 3.11). Surface processes in inland basins dominated by eolian processes may also be related to local base levels, which are represented by deflation surfaces associated with the level of the groundwater table (Kocurek, 1988).

The marine base level (~ sea level) and the fluvial graded profiles are sometimes used in conjunction to define a composite ('stratigraphic') base level, which is the surface of equilibrium between erosion and deposition within both marine and continental areas (Cross, 1991; Cross and Lessenger, 1998; Fig. 3.9). At any given location, the position of this irregular 3D surface is determined by the competing forces of sedimentation and erosion, and it may be placed either above the land surface/seafloor (where aggradation occurs), or below the land surface/seafloor (where subaerial/submarine erosion occurs).

The debate regarding the relationship between base level and the fluvial graded profile still persists in current sequence stratigraphic terminology. One school of thought argues that the term 'base level' should apply to both concepts, as the same definition can describe them both (i.e., a dynamic surface of equilibrium between deposition and erosion; Barrell, 1917; Sloss, 1962; Cross, 1991; Cross and Lessenger, 1998). A second school of thought restricts the term 'base level' to the level of the body of water into which the river debouches, where an abrupt decrease in fluvial-energy flux is recorded (Powell, 1875; Davis, 1908; Bates and Jackson, 1987; Schumm, 1993; Posamentier and Allen, 1999; Catuneanu, 2003). Terminology is trivial to some extent, but there seems to be value in keeping the concepts of graded fluvial profile and base level separate, as they are in a process/response relationship—i.e., the position in space of the fluvial graded profile is in part a function of the elevation of the base level (Fig. 3.9). This is the approach adopted in this book.

Proxies for Base Level and Accommodation

As the base level is an imaginary and dynamic 4D surface of equilibrium between deposition and erosion, largely dependent on fluctuations in environmental energy and sediment supply (Fig. 3.8), the precise quantification of accommodation at any given time and in any given location is rather difficult. For this reason, proxies may be used for an easier visualization of the available accommodation. At first approximation, sea level is a proxy for base level (Jervey, 1988; Schumm, 1993), and so the available accommodation in a marine environment may be measured as the distance between the sea level and the seafloor. Both the sea level and the seafloor may independently change their position with time relative to the center of Earth in response to various controls, and therefore the amount of available accommodation fluctuates accordingly. Sea level is one of the primary allogenic controls on sedimentation, and it is in turn controlled by climate and tectonism, as discussed in the previous sections (Fig. 3.1). The upward and downward shifts in the position of the seafloor relative to the center of Earth depend on two main parameters, namely the magnitude of total subsidence or uplift, and sedimentation. The amount of available accommodation at any given time and in any given location therefore equals the balance between how much accommodation is created (or destroyed) by factors such as tectonism and sea-level change, and how much of this space is consumed by sedimentation at the same time. The distinction between these two members of the accommodation equation (creation/destruction *vs.* consumption) is one of the key themes of sequence stratigraphy, which allows one to understand the fundamental mechanisms behind the formation of systems tracts and sequence stratigraphic surfaces.

Figure 3.12 helps to define some of the basic concepts involved in the accommodation equation, such as eustasy (sea level relative to the center of Earth), relative sea level (sea level relative to a datum that is independent of sedimentation), and water depth (sea level relative to the seafloor). A change in relative sea level is a proxy for how much accommodation was created or lost during a period of time, independent of sedimentation, whereas water depth is a proxy for how much accommodation is still available after the effect of sedimentation is

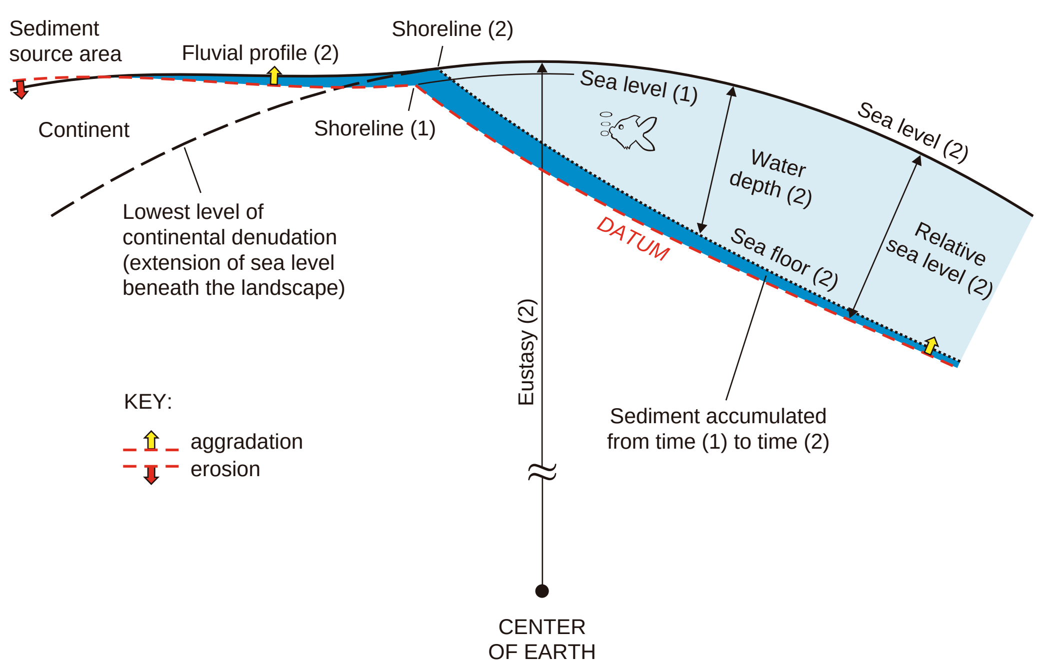


FIGURE 3.12 Eustasy, relative sea level, and water depth as a function of sea level, seafloor, and datum reference surfaces (modified from Posamentier *et al*., 1988). The datum is a subsurface reference horizon that monitors the amount of total subsidence or uplift relative to the center of Earth. In this diagram, the datum corresponds to the ground surface (subaerial and subaqueous) at time (1). Sedimentation (from time 1 to time 2 in this diagram) buries the datum, which, at any particular location, may be visualized as a G.P.S. that monitors changes in elevation through time (i.e., distance relative to the center of Earth).

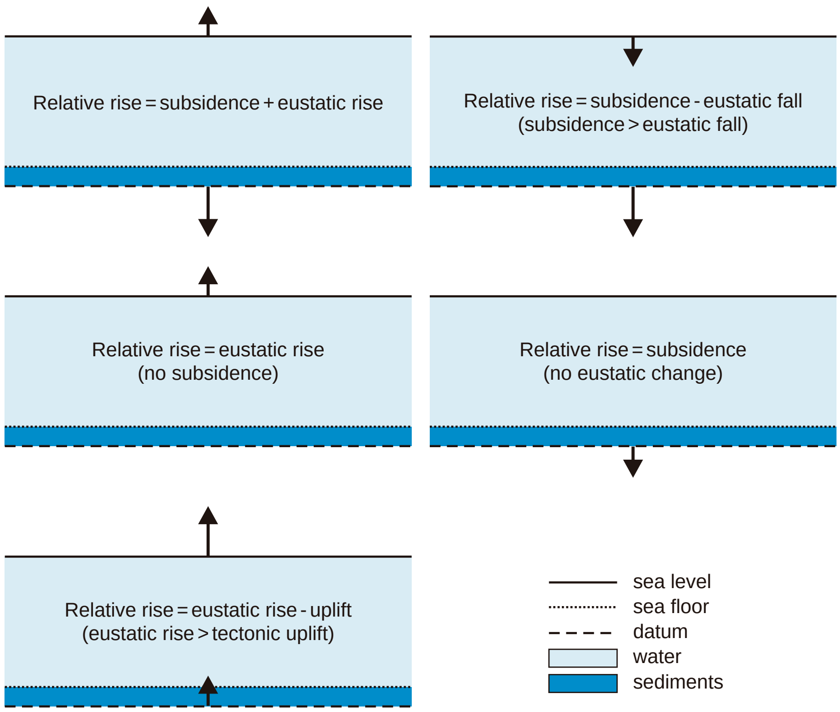


FIGURE 3.13 Scenarios of relative sea-level rise. If base level is equated with sea level for simplicity (by neglecting the energy of waves and currents), then relative sea-level rise becomes synonymous with base-level rise. Note that the newly created accommodation may be consumed by sedimentation at any rates, resulting in the shallowing or deepening of the water. The length of the arrows is proportional to the rates of vertical tectonics and eustatic changes.

also taken into account. The datum in Fig. 3.12 monitors the total amount of subsidence or uplift (including the effects of sediment loading and compaction) recorded in any location within the basin relative to the center of Earth. This datum reference horizon is taken as close to the seafloor as possible in order to capture the entire subsidence component related to sediment compaction, but its actual position is not as important as the change in the distance between itself and the sea level. This is because we are more interested in the *changes* in relative sea level (i.e., changes in the distance between the datum and the sea level), which reflect how much accommodation is created or lost during a period of time, rather than the actual *amount* of relative sea level (i.e., the actual distance between the datum and the sea level) at any given time. Different scenarios for rises and falls in relative sea level are illustrated in Figs. 3.13 and 3.14.

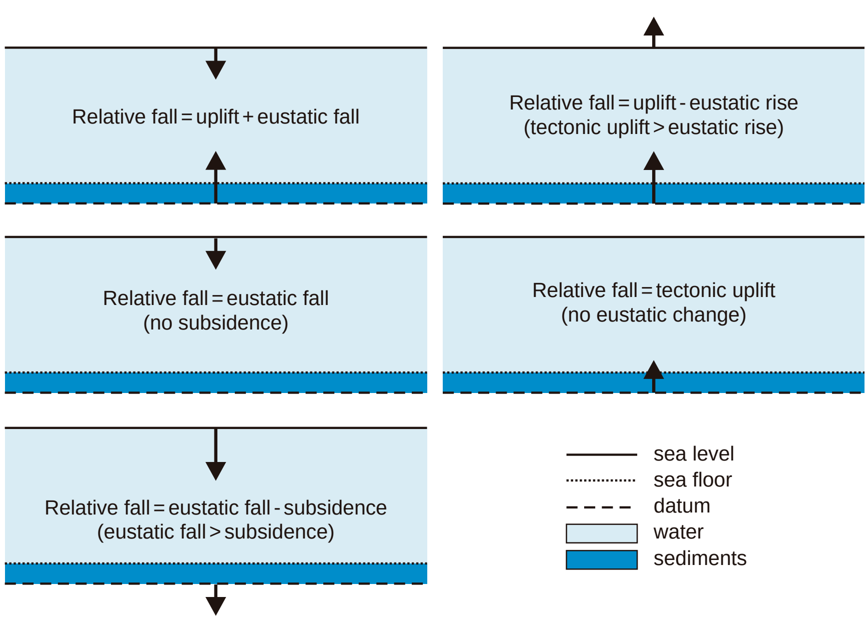


FIGURE 3.14 Scenarios of relative sea-level fall. If base level is equated with sea level for simplicity (by neglecting the energy of waves and currents), then relative sea-level fall becomes synonymous with base-level fall. Falling base level results in loss of available accommodation, and almost invariably in the shallowing of the water. The length of the arrows is proportional to the rates of vertical tectonics and eustatic changes.

The separation between relative sea-level changes and sedimentation is a fundamental approach in sequence stratigraphy, which allows for the comparison between their rates as independent variables. The balance between these rates (creation/destruction of accommodation *vs.* consumption of accommodation) controls the direction and type of shoreline shifts, and implicitly the timing of all sequence stratigraphic surfaces and systems tracts. This approach is therefore key to a proper understanding of sequence stratigraphic principles. Failure to do so may result in confusions between relative sea-level changes, water-depth changes, and the directions of shoreline shift. Simple calculations show that the relative sea level may rise even during stages of sea-level fall, if the rates of subsidence are high enough (Fig. 3.13). For example, if the sea level falls at a rate of 5 m/1000 years but the subsidence rate is 9 m/1000 years, the relative sea level rises with 4 m/1000 years, which means that accommodation is created at a rate of 4 m/1000 years. If the sedimentation rate in that particular location is 3 m/1000 years, it means that accommodation is created faster than it is consumed, and hence the water is deepening, in this case at a rate of 1 m/1000 years. If the location in this example is placed in the vicinity of the shoreline, then the increase in water depth is likely to be associated with a shoreline transgression. As shown by numerical modeling, the correlation between water-depth changes and the direction of shoreline shift (i.e., water shallowing = regression, and water deepening = transgression) is only truly valid for shallow-marine areas, and it may be distorted offshore (Catuneanu *et al*., 1998b).

Changes in Accommodation

The above discussion on the controls on accommodation is based on the assumption that sea level is a proxy for base level. This is true at first approximation, but in reality base level is commonly *below* the sea level, due to the energy flux brought about by waves and currents (Fig. 3.8). As noted by Schumm (1993), this is also supported by the fact that at their mouths, rivers erode slightly below the sea level. The actual distance between base level and sea level depends on environmental energy, as for example the base level is lowered during storms relative to its position during fairweather. Such energy fluctuations usually take place at seasonal to sub-seasonal time scales, at a frequency that is higher than most highest-frequency cycles investigated by sequence stratigraphy. Longer-term shifts in base level, at scales relevant to sequence stratigraphy, are generally controlled by the interplay of eustasy and total subsidence. In other words, the proxies used in the above discussion (i.e., sea level for base level, and relative sea-level changes for changes in accommodation) are acceptable in a sequence stratigraphic analysis. The most complete scenario that illustrates the interplay of the controls on accommodation and shoreline shifts in a marine environment is presented in Fig. 3.15.

Similar to the way relative sea-level changes are measured, base-level fluctuations *relative to the datum* define the concept of base-level changes. As base level is not exactly coincident with sea level, due to the energy flux of waves and currents, the concepts of relative sea-level changes and base-level changes are not identical although they follow each other closely (Fig. 3.15). A rise in base level (increasing vertical distance between base level and the datum) creates accommodation. Sedimentation during base-level rise results in the consumption of the available accommodation at lower or higher rates relative to the rates at which accommodation is being created. The former situation implies water deepening, whereas the latter implies water shallowing. At any given time, the amount of accommodation that is still available for sediments to accumulate is measured by the vertical distance between the seafloor and the base level. Similarly, a fall in base level (decreasing vertical distance between base level and the datum) destroys accommodation. Almost invariably, such stages result in water shallowing in that particular location, irrespective of the depositional processes.

The contrast between the rates of change in accommodation and the sedimentation rates in locations placed in the vicinity of the shoreline allows one to understand why the shoreline may shift either landward or seaward during times of relative sea-level (base-level) rise. Accommodation outpacing sedimentation generates transgression (i.e., accommodation is created faster than it is consumed by sedimentation), whereas an overwhelming sediment supply may result in shoreline regression (i.e., accommodation is consumed more rapidly than it is being created). In either situation, the river mouth moves accordingly, landward or seaward, connecting the continuously adjusting fluvial profile to the shifting base level. In the case of a delta that progrades during a stage of base-level rise, for example, the newly created space is not sufficient to accommodate the entire amount of sediment brought by the river, and as a result the river mouth shifts seaward. This shift triggers a change in depositional regimes from prodelta and delta front environments, where sedimentation is limited to the space between the seafloor and the base level, to delta plain and alluvial plain environments (landward relative to the shoreline), where depositional trends (aggradation, bypass, or erosion) are governed by the

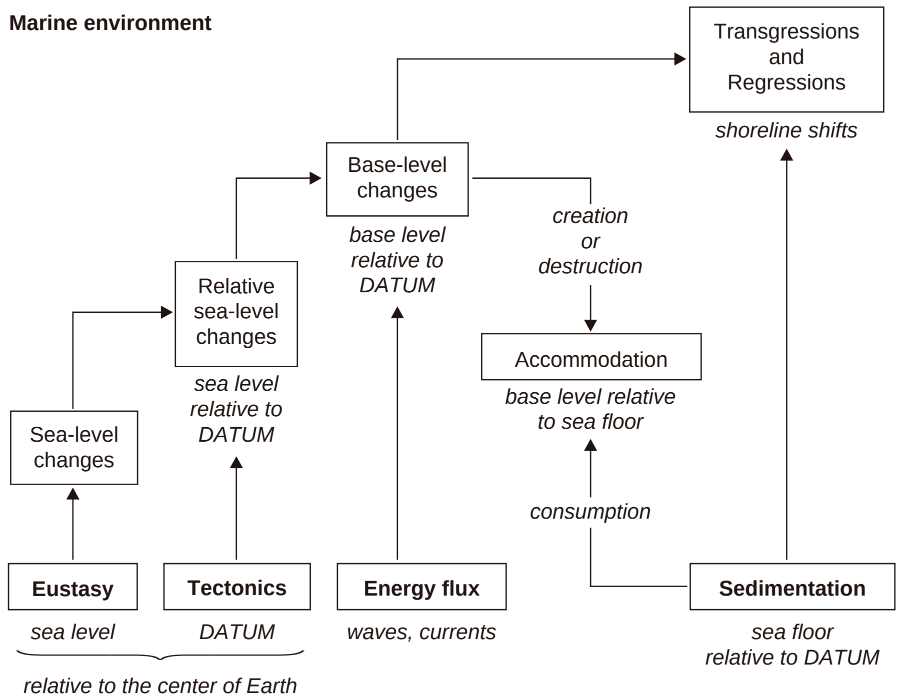


FIGURE 3.15 Controls on accommodation and shoreline shifts in a marine environment (modified from Catuneanu, 2003). This diagram also applies to lacustrine environments by substituting sea level with lake level. See Fig. 3.12 for the definition of the *DATUM*. The energy flux lowers the base level *via* the effects of waves, wave-generated currents, tidal currents, contour currents, or gravity flows. Short-term climatic changes (seasonal to sub-seasonal time scales) are accounted for under energy flux, whereas the longer-term climatic changes (e.g., Milankovitch type) are built into eustasy. The 'energy flux' box stands for the dynamic balance between environmental energy and sediment supply, as an increase in energy *relative to sediment supply* leads to base-level fall (loss of accommodation), and a decrease in energy *relative to sediment supply* leads to base-level rise (gain of accommodation). Note the difference between 'sediment supply' (load moved by a transport agent) and 'sedimentation' (amount of vertical aggradation). For example, depending on energy flux conditions, high sediment supply does not necessarily result in high sedimentation rates. Base-level changes depend on sediment supply, but are measured independently of sedimentation. In contrast, relative sea-level changes are independent of both sediment supply and sedimentation. This flow chart is valid for zone 1 in Fig. 3.3.

relative position between the fluvial graded profile and the actual fluvial profile.

The fluvial graded profile is the conceptual equivalent of the marine base level in the nonmarine realm, as it describes the imaginary and dynamic surface of equilibrium between deposition and erosion in the fluvial environment. In this context, the amount of *fluvial accommodation* is defined as the space between the graded profile and the actual fluvial profile (Posamentier and Allen, 1999). If we compare this definition with the concept of *marine accommodation*, discussed above, the graded profile is the equivalent of the base level, and the actual fluvial profile is the counterpart of the seafloor in the marine environment. If we follow this comparison even farther, we notice that the sea level, which is used as a proxy for base level, does not have an equivalent in the fluvial realm, which makes the visualization of fluvial accommodation rather difficult as there is no physical proxy for the fluvial graded profile. The only observable surface is the actual fluvial landscape, whose position relative to an independent datum changes in response to surface processes of aggradation or erosion (Fig. 3.12). In turn, these surface processes are triggered by an attempt of the river to reach its graded profile.

The graded profile is 'anchored' to the base level at the river mouth, and as the base level rises and falls, this anchoring point moves either landward or seaward, or up or down, triggering an in-kind response of the graded profile (Posamentier and Allen, 1999). Therefore, base-level changes exert an important control on graded profiles, and implicitly on fluvial accommodation, especially in the downstream reaches of the fluvial system (Shanley and McCabe, 1994; Fig. 3.16). The position of graded profiles also depends on fluctuations in energy flux, which are mainly attributed to the effects of climate on a river's transport capacity (Blum and Valastro, 1989; Blum, 1990; Fig. 3.16). Such energy fluctuations may

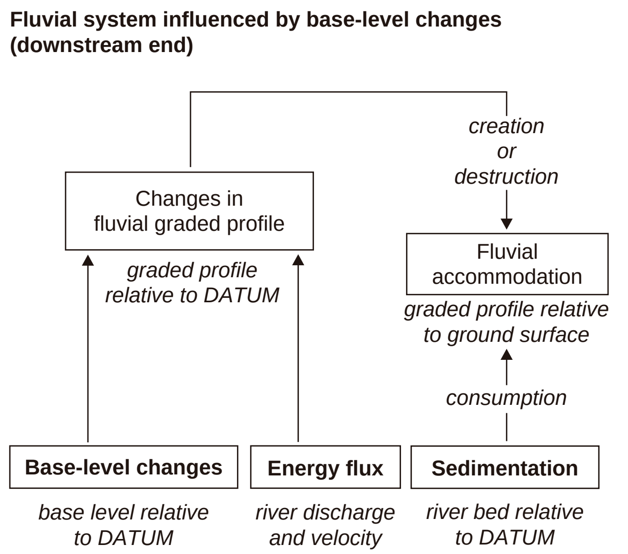


FIGURE 3.16 Controls on fluvial accommodation in the downstream reaches of a fluvial system. See Fig. 3.12 for the definition of the *DATUM*. The energy flux is mainly controlled by short to longer term climatic changes (especially the discharge component), but also by tectonic tilt. This flow chart is valid for zone 2 in Fig. 3.3.

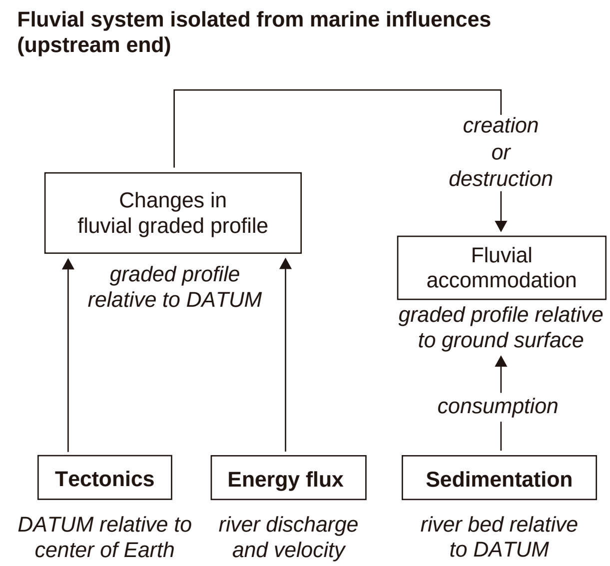


FIGURE 3.17 Controls on fluvial accommodation in the upstream reaches of a fluvial system. See Fig. 3.12 for the definition of the *DATUM*. The energy flux is mainly controlled by short- to longer-term climatic changes (especially the discharge component), but also by tectonic tilt. This flow chart is valid for zone 3 in Fig. 3.3.

be recorded over different time scales, from seasonal climatic changes that may occur with a frequency higher than the highest-frequency cycles studied by sequence stratigraphy, to Milankovitch-scale orbital forcing.

The effect of base-level changes on fluvial processes (aggradation *vs.* erosion) is only 'felt' by rivers within a limited distance upstream relative to the river mouth, which is usually in a range of less than 200 km (Miall, 1997). Beyond the landward limit of base-level influences, rivers respond primarily to a combination of tectonic and climatic controls (Fig. 3.17). Tectonism dictates the overall geometry of fluvial sequences, as the creation of fluvial accommodation follows the patterns of regional subsidence. For example, the rates of subsidence induced by flexural loading in a foreland basin increase in a proximal direction, toward the center of loading, whereas the rates of thermal and mechanical subsidence in an extensional basin increase in a distal direction. Superimposed on these general trends, the climatic control on runoff and discharge also affect the position of graded profiles, as discussed above (Fig. 3.17).

The role of climate as a control on accommodation is always difficult to quantify, as it operates *via* other variables such as eustasy and environmental energy flux. In a marine environment, the short-term climatic changes (seasonal to sub-seasonal time scale) translate into fluctuations in energy flux, whereas the longer-term changes are accounted for under eustasy (Fig. 3.15). In the case of fluvial environments, both short- and longer-term climatic changes are reflected in the fluctuations in energy flux, as there is no physical proxy for the graded profile that could be related to the longer term climate shifts (Figs. 3.16 and 3.17). Climate is also relevant to the 'sedimentation' box in all cases (Figs. 3.15–3.17), as the amount of sediment supply transferred from source areas to the sedimentary basin depends on the efficiency of weathering and sediment transport processes, both partly dependent on climate.

Changes in accommodation, in conjunction with the rates of sedimentation, represent a key control on depositional trends, which are in turn reflected by specific shoreline shifts (e.g., progradation is associated with shoreline regression, and retrogradation relates to shoreline transgression). Quantitative modeling of the interplay between subsidence, sea-level change and sedimentation shows that even though the shoreline may only shift in one direction along a dip oriented profile at any given time, accommodation may change with different rates, and even in opposite directions, along the same cross-sectional profile (Jervey, 1988; Catuneanu *et al.*, 1998b). This coeval change in the rates and sign of accommodation shifts is caused by differential subsidence, which is usually the norm in any sedimentary basin. The higher the contrasts in the rates of differential subsidence between various areas in the basin, the more pronounced the difference between the amounts of available accommodation will be. For example, during a stage of sea-level fall, accommodation may be negative in slowly subsiding areas (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

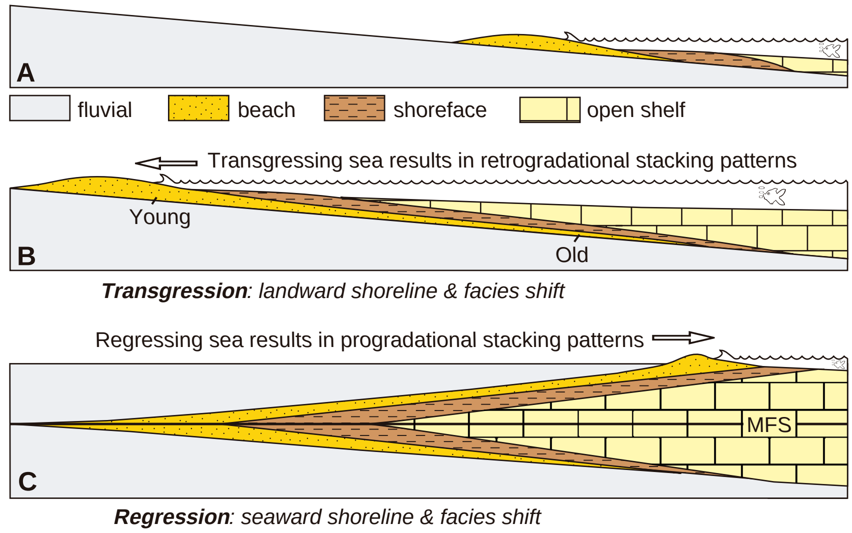


FIGURE 3.18 Transgressions and regressions. Note the retrogradation and progradation (lateral shifts) of facies, as well as the surface that separates retrogradational from overlying progradational geometries. This surface is known as the maximum flooding surface (MFS).

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

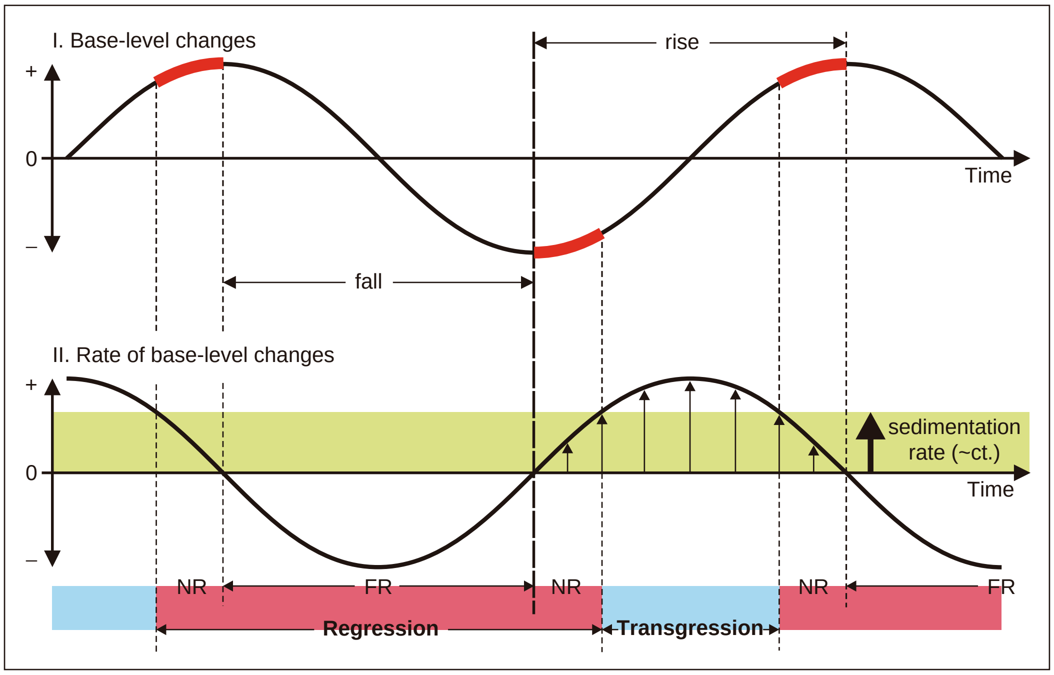


FIGURE 3.19 Concepts of transgression, normal regression, and forced regression, as defined by the interplay between base-level changes and sedimentation. The top sine curve shows the magnitude of base-level changes through time. The thicker portions on this curve indicate early and late stages of base-level 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

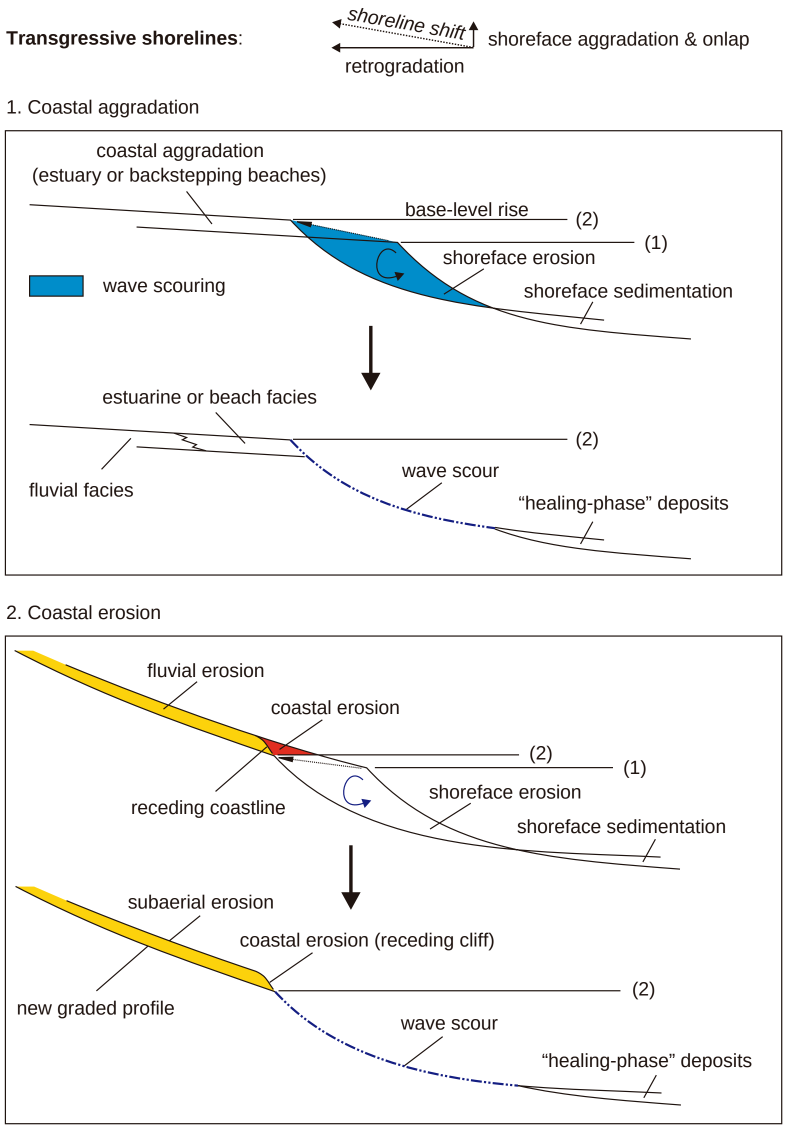


FIGURE 3.20 Shoreline trajectory in transgressive settings (from Catuneanu, 2003). Transgressions are driven by base-level rise, where the rates of base-level rise outpace the sedimentation rates in the shoreline area. The balance between the opposing trends of aggradation (in front of the shoreline) 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°).

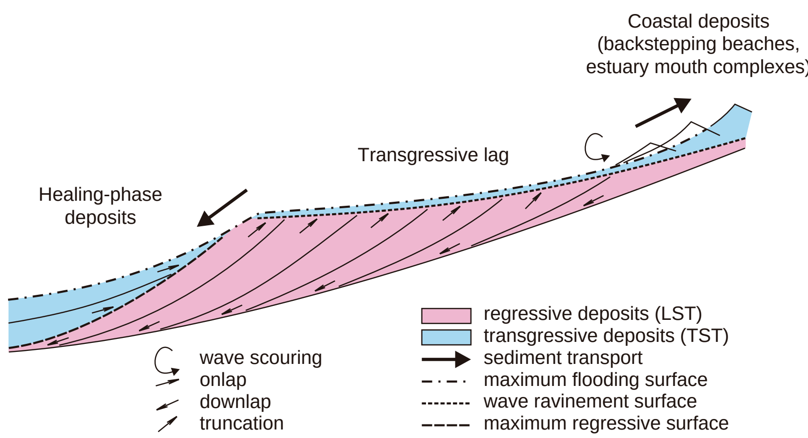


FIGURE 3.21 Patterns of sediment redistribution during shoreline transgression (modified from Posamentier and Allen, 1993, and Willis and Wittenberg, 2000). Some sediment is carried landward as backstepping beaches (open shorelines) or backstepping estuary-mouth complexes (river-mouth settings), while 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.

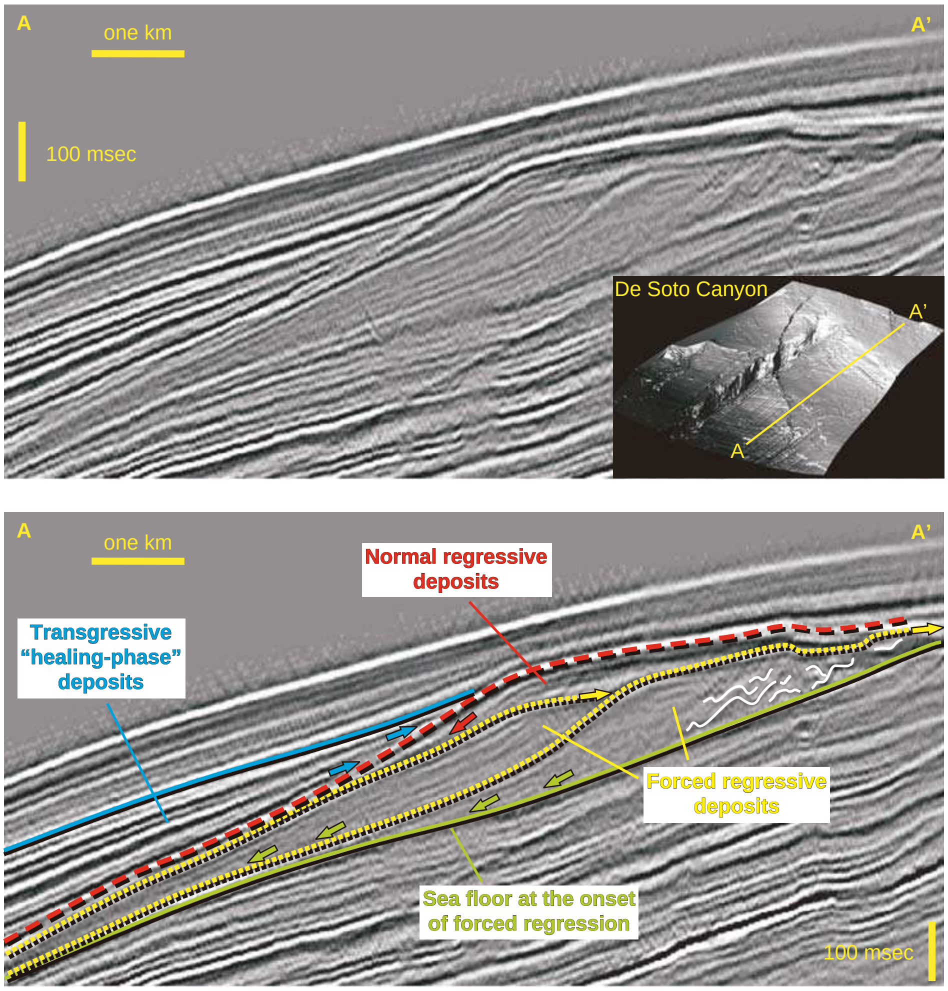
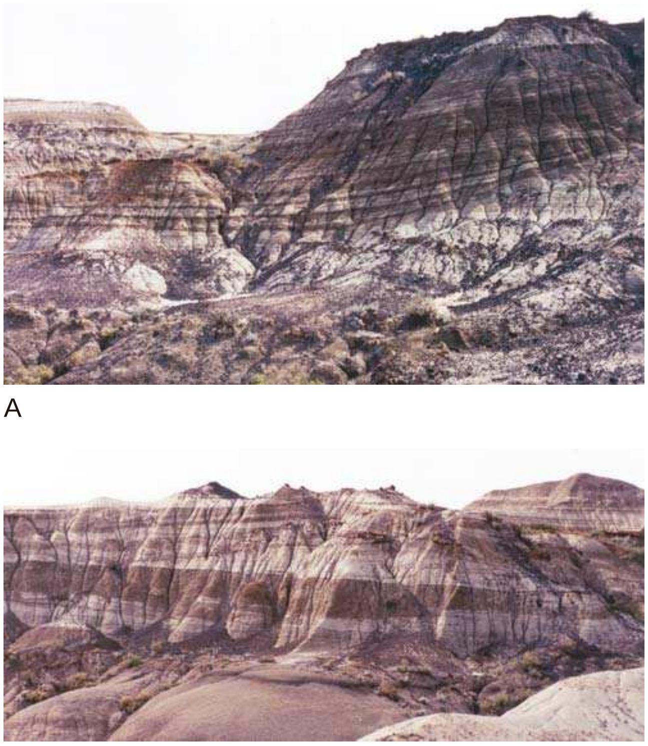


FIGURE 3.22 Shelf-edge and healing-phase deposits in the De Soto Canyon area of the Gulf of Mexico (uninterpreted and interpreted seismic lines, whose location is shown on the 3D illuminated surface) (modified from Posamentier, 2004a; images courtesy of H.W. Posamentier). Note that prograding clinoforms 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.

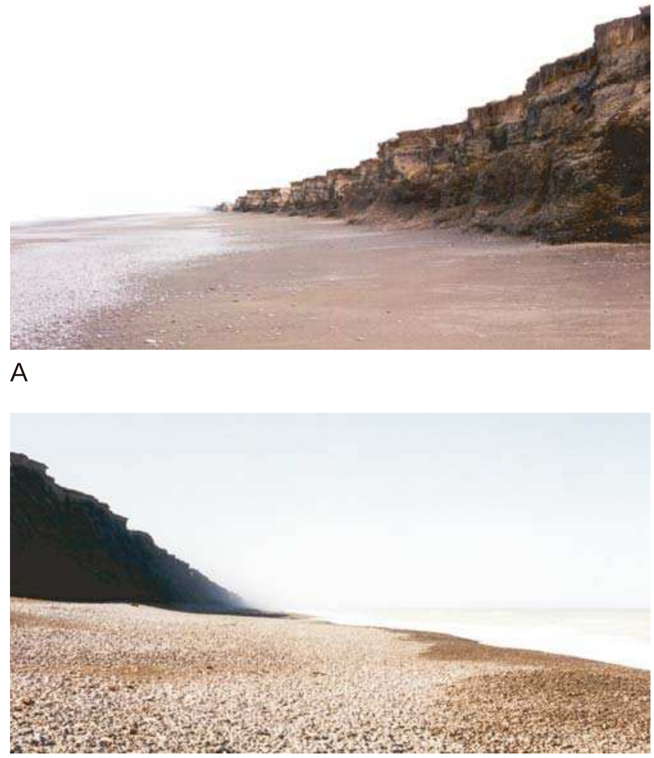
beach (open shoreline) environments. However, the tendency of coastal aggradation is counteracted by the wave scouring in the upper shoreface, as the latter gradually shifts in a landward direction. The balance between these two opposing forces, of sedimentation *vs.* erosion, determines the overall 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,

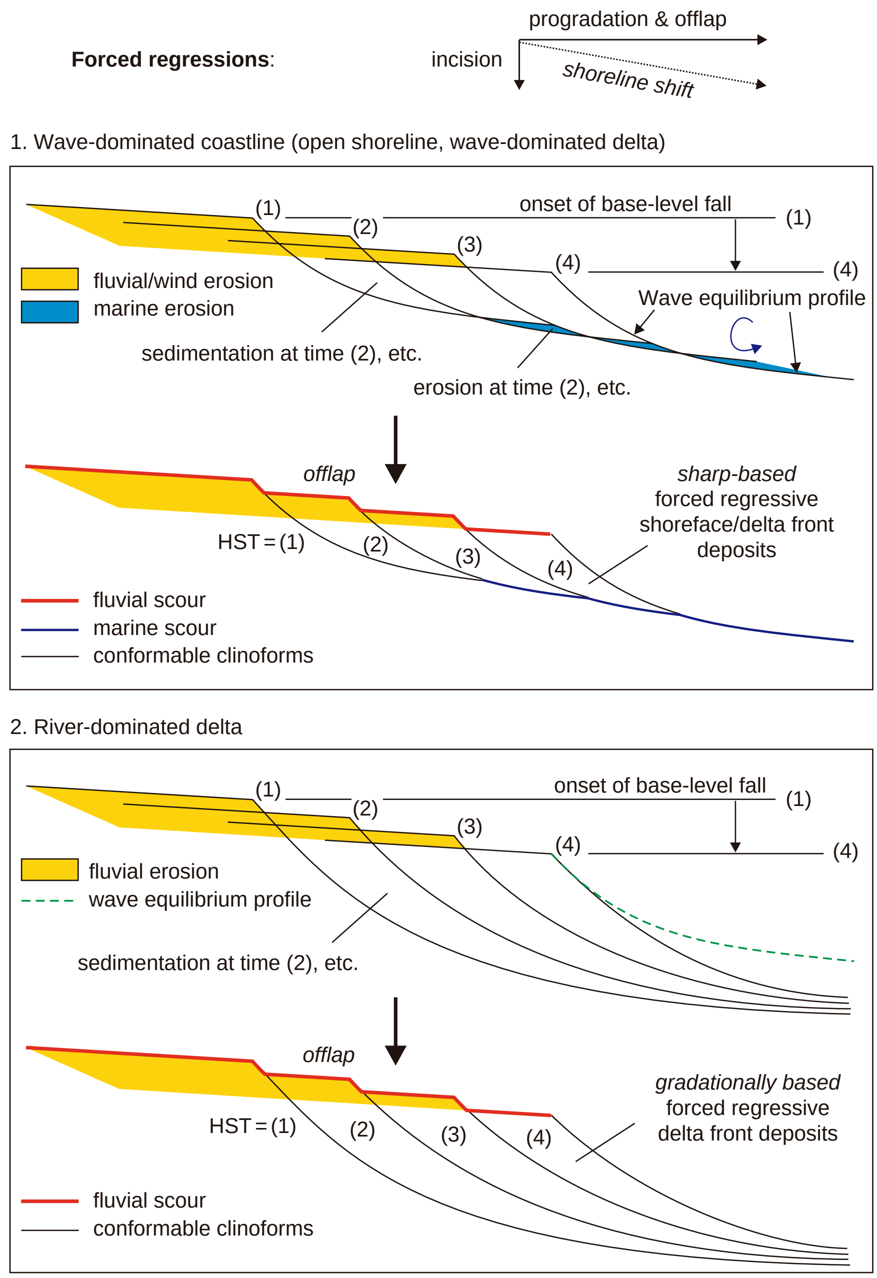


FIGURE 3.27 Shoreline trajectory in forced regressive settings (modified from Catuneanu, 2003). Forced regressions are driven by base-level fall, irrespective of sediment supply, and the rates of progradation are generally high. Wave-dominated subtidal settings are characterized by low gradients of the 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

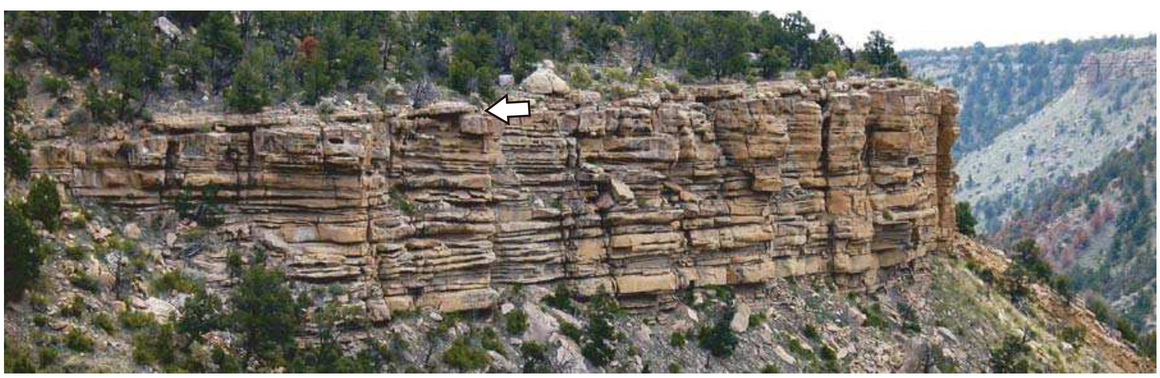
FIGURE 3.28 Forced regressive, wave-dominated shoreface sands (with A—swaley cross-stratification–Fig. 3.29) abruptly overlying inner shelf interbedded sands and muds (B). The upper shoreface sands (A) are 'sharp-based' due to wave scouring in the lower shoreface during base-level fall. The exposed section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

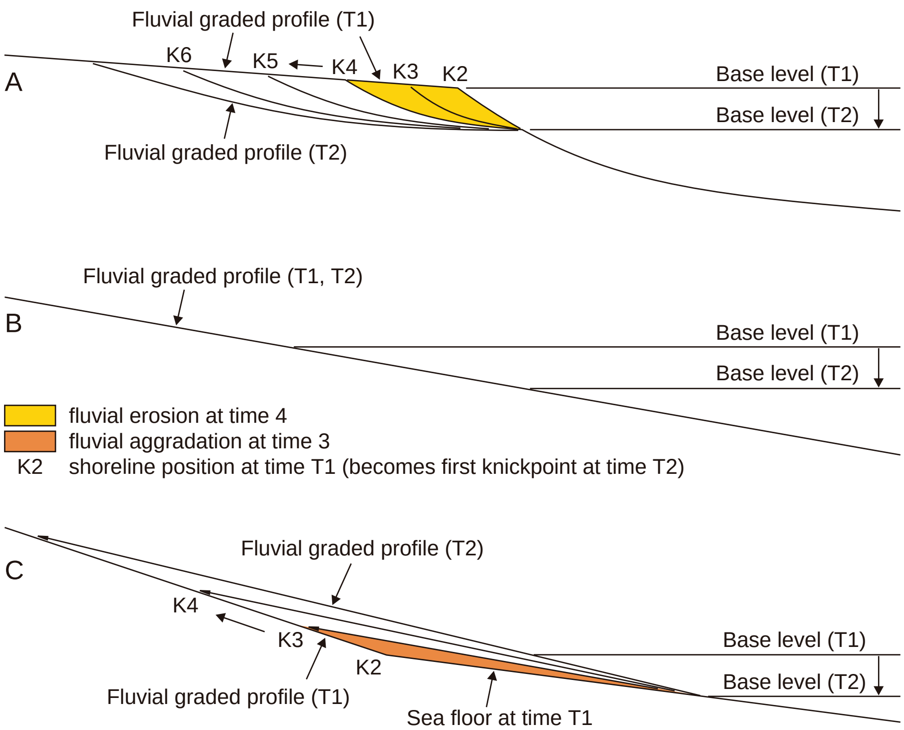


FIGURE 3.31 Fluvial responses to base-level fall, as a function of the contrast in slope gradients between the fluvial and the seafloor profiles at the onset of forced regression (modified from Summerfield, 1985; Pitman and Golovchenko, 1988; Butcher, 1990; Schumm, 1993; Posamentier and Allen, 1999; 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.

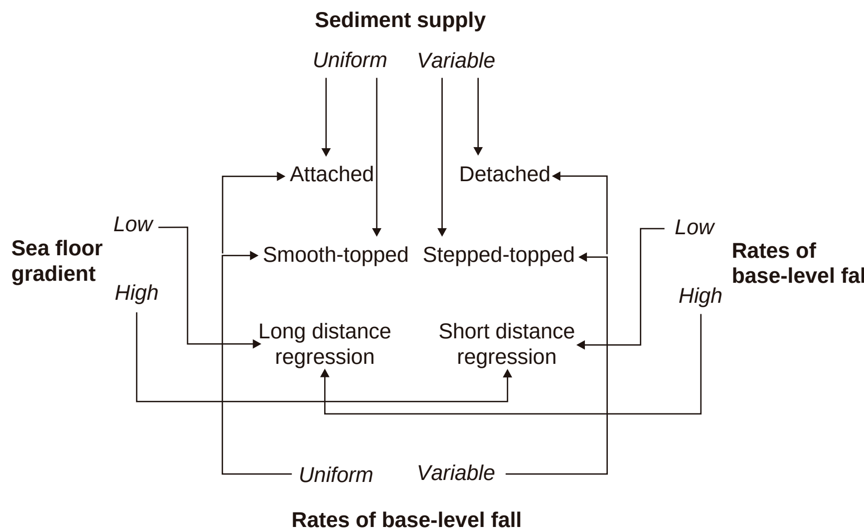


FIGURE 3.33 Stratal architecture of shallow-marine forced regressive deposits, as a function of sediment supply, rates of base-level fall and gradient of the seafloor. The interplay of these variables may result in a variety of possibilities, with the prograding forced regressive lobes being attached 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).

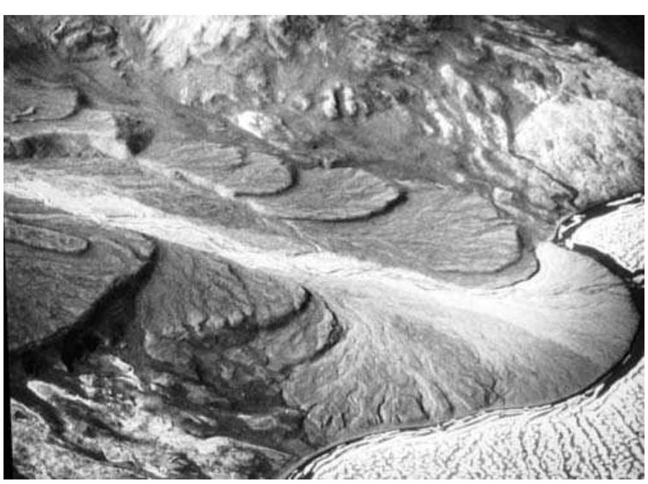


FIGURE 3.34 Modern forced regressive delta showing offlapping stratal stacking patterns (photo courtesy of J. England). In this case, the fall in base level is triggered by post-glacial isostatic rebound in the Canadian arctics, at a rate that exceeds the rate of present day sea-level rise.

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

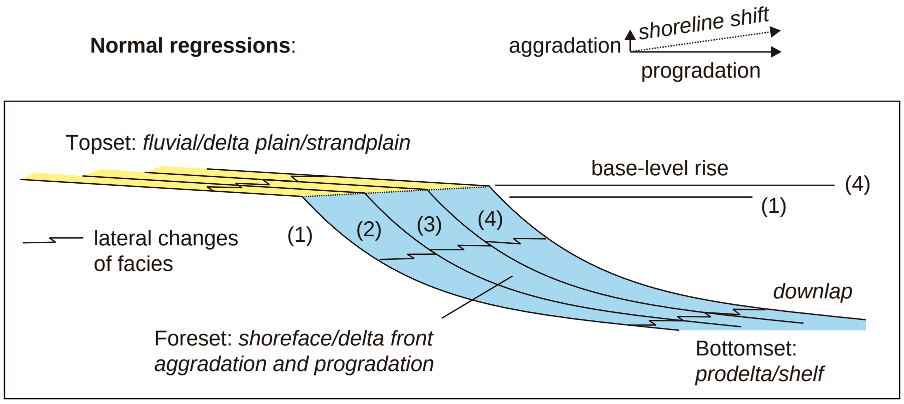


FIGURE 3.35 Shoreline trajectory in normal regressive settings, defined by a combination of progradation and aggradation in fluvial to shallow-marine systems. Normal regressions are driven by sediment supply, where the rates of base-level rise at the shoreline are outpaced by sedimentation rates. Normal 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.

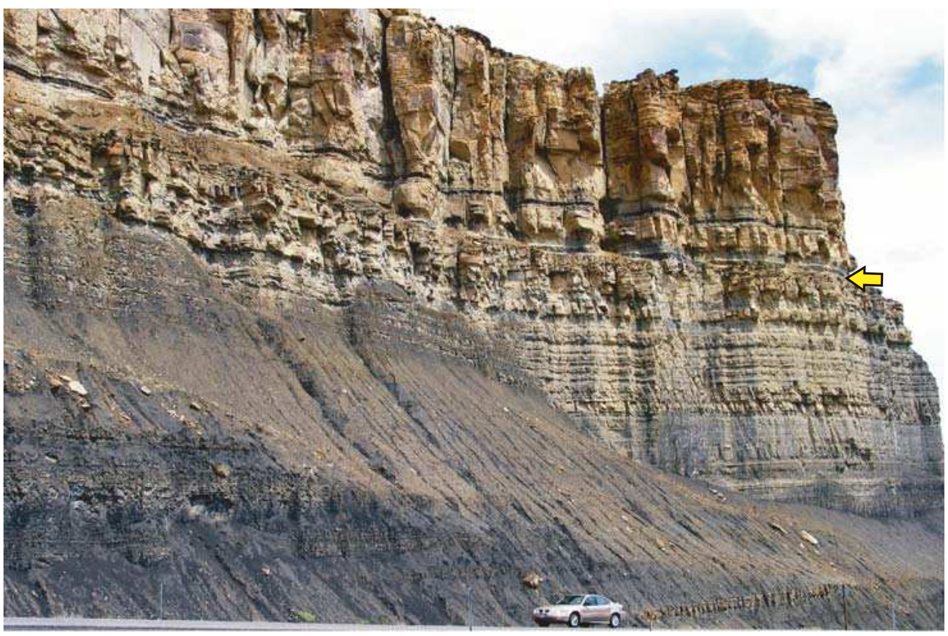


FIGURE 3.36 Normal regressive deltaic succession (river-dominated delta), showing a conformable transition from shallow-marine muds and sands (shelf, prodelta, delta front) to coastal and fluvial deposits (Ferron Sandstone, Utah). The arrow points at the conformable facies contact between delta front 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.