examples have been documented in the rock record as well (Plint, 1988, 1991, 1996; Posamentier *et al*., 1992b; Ainsworth, 1994; Plint and Nummedal, 2000; Posamentier and Morris, 2000; Fig. 4.28).

The regressive surface of marine erosion is one of the most prominent sequence stratigraphic surfaces, with a strong physical expression in the rock record due to the contrast in facies across the scoured contact, even though both the underlying and overlying deposits are coarsening-upward, as being part of a regressive succession (Figs. 4.9, 4.28, 4.29, and 4.31). The process of wave scouring during forced regression leads to the exhumation of semi-lithified marine sediments, resulting in the formation of firmgrounds colonized by the *Glossifungites* ichnofacies tracemakers (MacEachern *et al*., 1992; Chaplin, 1996; Buatois *et al*., 2002). Such firmgrounds separate deposits with contrasting ichnofabrics, largely due to the abrupt shift in environmental conditions that prevailed during the deposition of the juxtaposed facies across the contact. Both MacEachern *et al*. (1992) and Buatois *et al*. (2002) provide case studies where the regressive surface of marine erosion, marked by the *Glossifungites* ichnofacies, separates finer-grained shelf deposits with *Cruziana* ichnofacies from overlying shoreface sands with a *Skolithos* assemblage. The basinward extent of the forced regressive *Glossifungites* firmground is limited to the area affected by fairweather wave erosion, beyond which the stratigraphic hiatus collapses, being replaced by the correlative conformity *sensu* Hunt and Tucker (1992) (Fig. 4.24). Synonymous terms for the regressive surface of marine erosion include the *regressive ravinement surface* (Galloway, 2001) and the *regressive wave ravinement* (Galloway, 2004).

Maximum Regressive Surface

The maximum regressive surface (Catuneanu, 1996; Helland-Hansen and Martinsen, 1996) is defined relative to the transgressive-regressive curve, marking the change from shoreline regression to subsequent transgression (Fig. 4.6). Therefore, this surface separates prograding strata below from retrograding strata above (Fig. 4.32). The change from progradational to retrogradational stacking patterns takes place *during* the base-level rise at the shoreline, when the increasing rates of base-level rise start outpacing the sedimentation rates (Fig. 4.5). As a result, the end-of-regression surface forms within an aggrading succession, sitting on top of lowstand normal regressive strata, and being onlapped by transgressive 'healing phase' deposits (Figs. 4.9 and 4.32). As the youngest clinoform associated with shoreline regression, the maximum regressive surface downlaps the pre-existing seafloor in a basinward direction, and drapes the preceding regressive clinoforms. Hence, the underlying lowstand normal regressive strata do not terminate against the maximum regressive surface (Fig. 4.9).

The maximum regressive surface is generally conformable (Fig. 4.9), although the possibility of seafloor scouring associated with the change in the direction of shoreline shift at the onset of transgression, which triggers a change in the balance between sediment load and the energy of subaqueous currents, is not excluded (Loutit *et al*., 1988; Galloway, 1989). The maximum regressive surface may also be scoured in the transition zone between coastal and fluvial environments, in relation to the backstepping of the higher energy intertidal swash zone (transgressive beach) over the fluvial overbank deposits of the lowstand (normal regressive) systems tract (Catuneanu *et al*., in press; Fig. 4.33). Where conformable, the maximum regressive surface is not associated with any substrate-controlled ichnofacies (Fig. 4.9). Where the transgressive marine facies are missing, the marine portion of the maximum regressive surface is replaced by the maximum flooding surface, and this composite unconformity may be preserved as a firmground or even hardground, depending on the amounts of erosion and/or synsedimentary lithification, colonized by the *Glossifungites* and *Trypanites* ichnofacies, respectively (Pemberton and MacEachern, 1995; Savrda, 1995). As this unconformity forms basinward relative to the shoreline position at the end of regression, within a fully marine environment, no xylic substrates (woodgrounds: the *Teredolites* ichnofacies) are expected to be associated with it.

The end of shoreline regression event (Fig. 4.7) marks a change in sedimentation regimes, as reflected by the balance between sediment supply and environmental energy, in all depositional systems within the sedimentary basin, both landward and seaward relative to the shoreline. As a result, the maximum regressive surface may develop as a discrete stratigraphic contact across much of the sedimentary basin, from marine to coastal and fluvial environments (Figs. 4.9, 4.32, and 4.34). The preservation potential of the end-of-regression surface is highest in the deep- to shallow-marine environments, where it tends to be onlapped by aggrading transgressive strata, and is lower in coastal to fluvial settings, where it may be subject to wave scouring during subsequent shoreline transgression (Fig. 3.21). Landward from the end-of-regression shoreline, the preservation of the maximum regressive surface depends on the balance between the rates of aggradation in the transgressive coastal to fluvial environments and the rates of subsequent transgressive wave-ravinement erosion in the upper shoreface. There are cases where this transgressive wave scouring may remove not only the

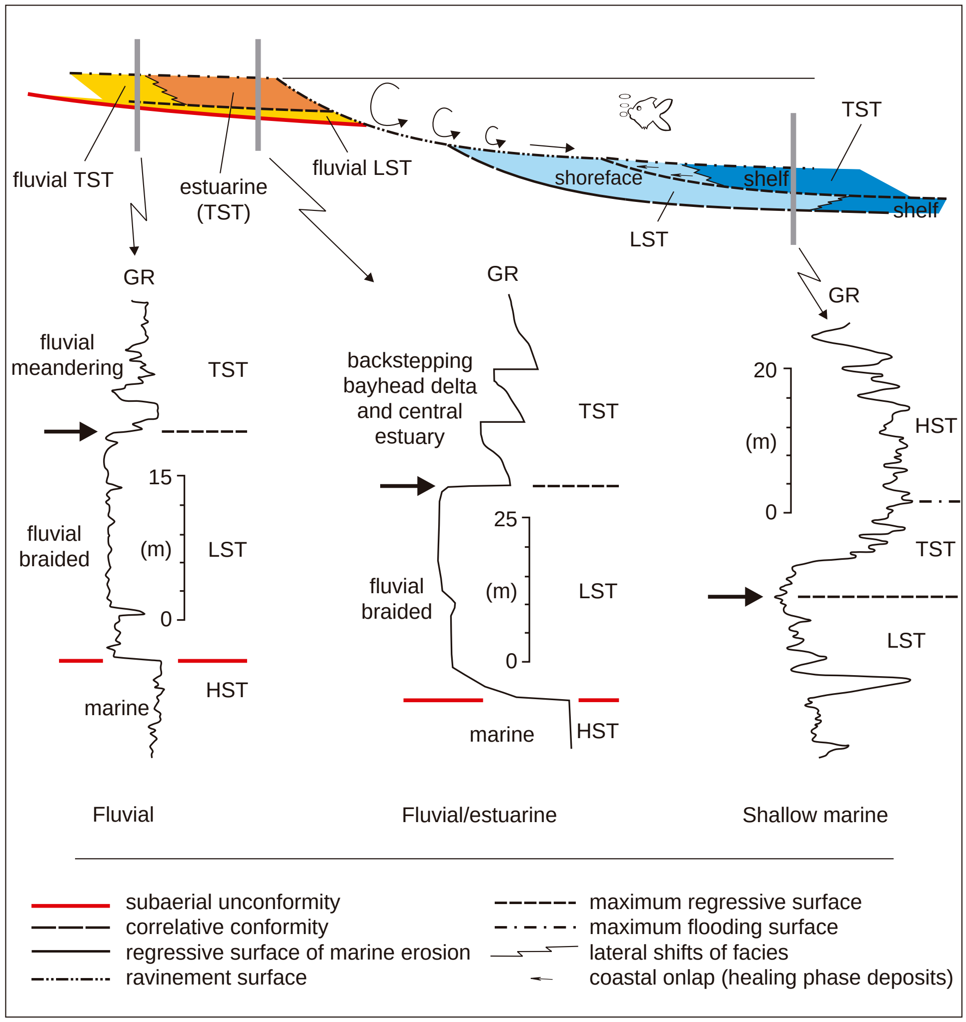


FIGURE 4.32 Well-log expression of the maximum regressive surface (arrows; modified from Catuneanu, 2002, 2003). See Fig. 4.9 for a summary of diagnostic features of the maximum regressive surface. Log examples from Kerr *et al*. (1999) (left and center) and Embry and Catuneanu (2001) (right). Abbreviations: GR—gamma ray; LST—lowstand systems tract; TST—transgressive systems tract; HST—highstand systems tract.

transgressive coastal to fluvial deposits, but also all underlying coastal to fluvial lowstand normal regressive deposits as well. In such cases, the transgressive wave scour, the maximum regressive surface and the subaerial unconformity are all amalgamated in one unconformable contact (Embry, 1995). In a more general scenario, however, the preservation of coastal to fluvial lowstand normal regressive deposits in the rock record depends on the duration of normal regression and the rates of sediment aggradation in coastal to fluvial environments prior to the transgressive wave scouring. Prolonged stages of lowstand normal regression may result in the formation of relatively thick topsets of aggrading and prograding coastal to fluvial strata, which drape the subaerial unconformity and are preserved from subsequent transgressive wave-ravinement erosion (Fig. 4.34). In such cases, the maximum regressive surface has the potential of being mappable across much of the sedimentary basin, within both marine and fluvial successions (Fig. 4.34).

In deep-marine deposits, the maximum regressive surface is most difficult to identify within the facies succession of the submarine fan complex on the basin floor, because the end-of-regression event occurs *during* a stage of waning down in the amount of terrigenous sediment that is delivered to the deep-water environment. For this reason, no physical criteria for outcrop, core, or well-log analysis have been developed to map the maximum regressive surface within the gravity-flow deposits that accumulate on the basin floor. More detailed discussions on the nature of gravity-flow deposits that accumulate in the deep-water environment during the various stages of the base-level cycle are provided in Chapters 5 and 6 of this book. On continental slopes, the maximum regressive surface is the *youngest prograding clinoform* which is onlapped by the overlying transgressive 'healing phase' deposits (Fig. 4.34). Where afforded by high resolution seismic data, the extension of this youngest prograding slope clinoform into the deeper portions of the basin may

FIGURE 4.33 Outcrop photograph of a maximum regressive surface (yellow arrow) at the contact between fluvial normal regressive strata (facies A) and the overlying backstepping beach deposits (facies B) (Bahariya Formation, Lower Cenomanian, Bahariya Oasis, Western Desert, Egypt). The maximum regressive surface is scoured by high-energy swash currents during the earliest stage of shoreline transgression. Underlying the maximum regressive surface, the fluvial strata correlate with a prograding and aggrading delta, and are part of the lowstand systems tract. The backstepping beach is the only preserved portion of the transgressive systems tract. The beach deposits are truncated at the top by a subaerial unconformity (red arrow, base of incised valley), and are overlain by coarse fluvial channel fills (facies C, part of a younger lowstand systems tract; Catuneanu *et al.*, in press). Note the landward shift of facies recorded across the maximum regressive surface, in contrast with the basinward shift of facies associated with the subaerial unconformity.

provide a clue of where to trace the maximum regressive surface within the basin-floor succession.

In shallow-marine systems, the maximum regressive surface is relatively easy to recognize at the top of coarsening-upward (prograding) deposits (Figs. 4.35–4.37). Depending on the rates of subsequent transgression, as well as on the location within the basin, the maximum regressive surface may or may not be associated with a sand/shale lithological contrast. Cases A, B, C, and D in Fig. 4.35 provide examples of maximum regressive surfaces that correspond to a *sand/shale contact*, suggesting rapid transgression and/or an abrupt cut-off of sediment supply as the transgression was initiated. Under these conditions, the sediment is trapped within the retrograding shoreline systems at the onset of transgression, leading to sediment starvation offshore and hence an abrupt facies change at the maximum regressive surface (Loutit *et al.*, 1988). Where the transgression is slower and/or the sediment supply is high and continues to be delivered offshore, the peak of coarsest sediment may occur *within the sand*, and the sand/shale contact is above the maximum regressive surface, within the overlying transgressive succession (Fig. 4.35E). Farther offshore relative to the paleoshoreline, into lower shoreface and shelf systems, the maximum regressive surface occurs within silty–shaly successions, marking the peak of coarsest sediment (end of progradation; Figs. 4.36 and 4.37). In such settings, the position of the maximum regressive surface is often evident from the breaks in slope gradients that can be observed in outcrops (Figs. 4.36 and 4.37). The end-of-progradation event (top of coarsening-upward trend) does not necessarily correspond to the peak of shallowest water depth, especially in offshore areas. The peak of shallowest water is usually recorded within the underlying regressive (lowstand) deposits, while the maximum regressive surface

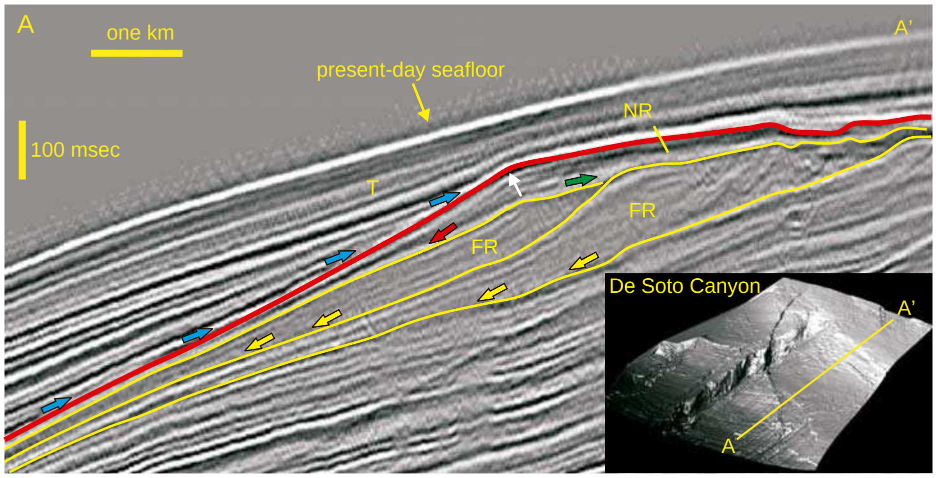
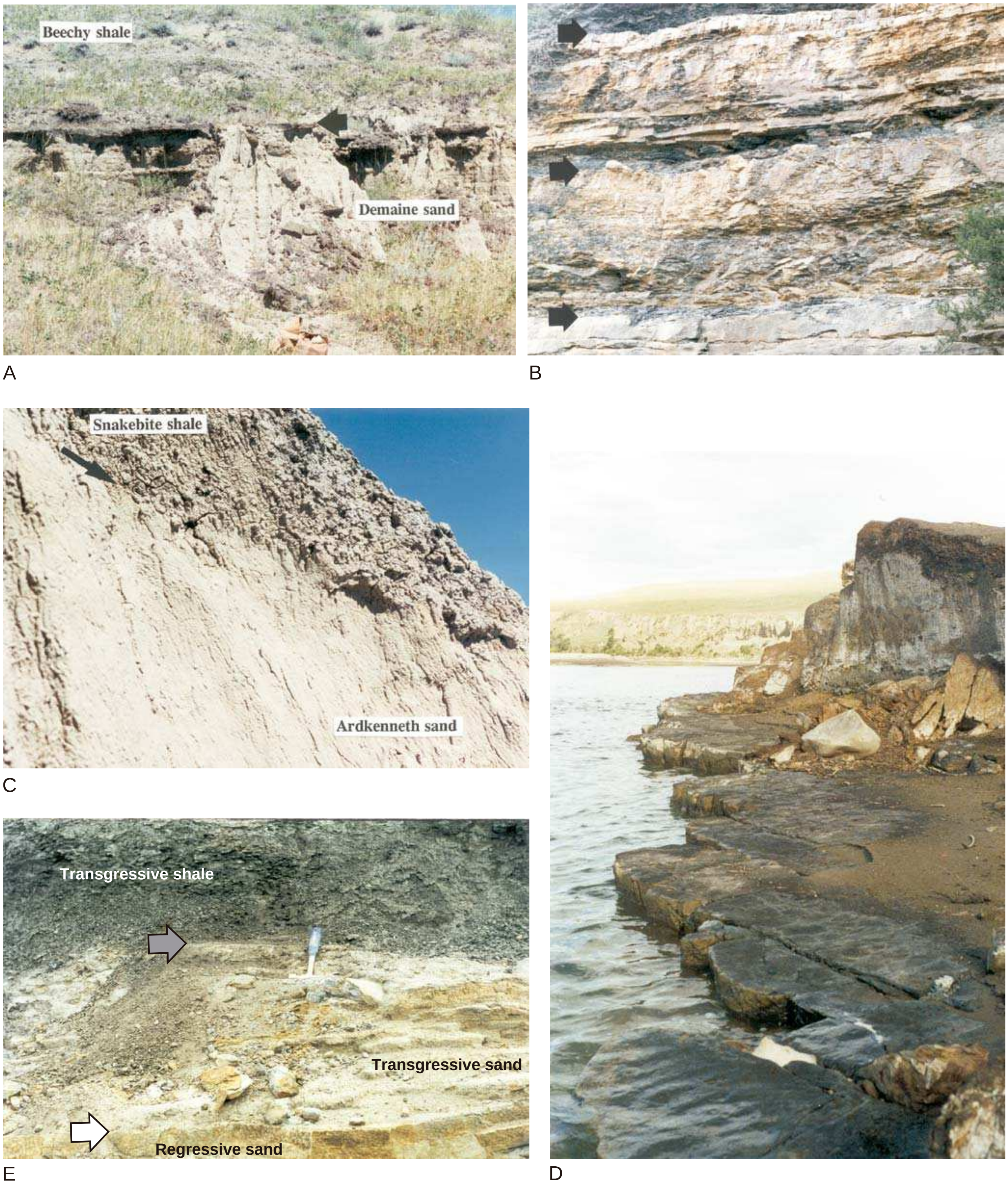


FIGURE 4.34 Maximum regressive surface (red line) on a dip-oriented, 2D seismic transect (location shown on the 3D illuminated surface) (De Soto Canyon area, Gulf of Mexico; image courtesy of H.W. Posamentier). This surface tops all fluvial to deep-marine strata that accumulate during lowstand normal regression. The maximum regressive surface may onlap the subaerial unconformity in a landward direction (fluvial onlap), and is onlapped by transgressive facies in the deep-water environment (marine onlap; blue arrows). The white arrow indicates the shoreline trajectory during lowstand normal regression. It is inferred that the normal regressive facies are marine seaward from the white arrow (downlapping the underlying forced regressive deposits; red arrow), and nonmarine in the opposite direction (onlapping the subaerial unconformity; green arrow—fluvial onlap). In a marine environment, the maximum regressive surface is the *youngest clinoform associated with shoreline regression*. 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. Abbreviations: FR—forced regressive deposits; NR—normal regressive deposits; T—transgressive deposits.

forms within deepening water—see discussion in Chapter 7. For this reason it is preferable to describe the trends in terms of *observed* grading (coarsening- *vs.* fining-upward) as opposed to *inferred* bathymetric changes (shallowing- *vs.* deepening-upward).

In coastal settings, the maximum regressive surface underlies the earliest estuarine deposits (Fig. 4.6). The contact between estuarine and underlying fluvial facies diverges from the maximum regressive surface beyond the initial length of the estuary at the onset of

FIGURE 4.35 Outcrop examples of maximum regressive surfaces in proximal shallow-water settings. A—maximum regressive surface (arrow) in a conformable marine succession. The top of the prograding (coarsening-upward) shoreface is marked by a concretionary layer of siderite-cemented sandstone, indicating the preferential fluid migration pathway during diagenesis. In this example, the onset of transgression is accompanied by an abrupt cut-off of sediment supply to the marine environment. Sediment trapping within the retrograding shoreline systems results in sediment starvation on the shelf (Loutit *et al*., 1988) (contact between Demaine and Beechy members, Bearpaw Formation, Late Campanian, Saskatchewan, Western Canada Sedimentary Basin); B—high-frequency maximum regressive surfaces (arrows) in a conformable deltaic succession. Maximum regressive surfaces are marked by concretionary layers (coarsest sand, prone to preferential precipitation of diagenetic cements), and are overlain by thin transgressive shales (Late Permian Waterford Formation, Ecca Group, southern Karoo Basin); C—maximum regressive surface (arrow) in a conformable marine succession, at the top of coarsening-upward prograding shoreface sands. The sharp lithological contrast across this surface indicates rapid transgression and/or a cut-off of sediment supply as the transgression is initiated (contact between Ardkenneth and Snakebite members, Bearpaw Formation, Late Campanian, Saskatchewan, Western Canada Sedimentary Basin); D—maximum regressive surface (top of coarsening-upward prograding shoreface sands) exposed by the subaerial erosion of the overlying (and more recessive) transgressive shales (top of the Kipp Member, Bearpaw Formation, Late Campanian, Oldman River, Alberta, Western Canada Sedimentary Basin); E—maximum regressive surface (white arrow) in a conformable marine succession, at the top of coarsening-upward prograding shoreface sands. Note that in this case the transition to the overlying transgressive facies is more subtle, and the facies contact between sand and shale (*flooding surface*, grey arrow) is above the maximum regressive surface (top of the Ryegrass Member, Bearpaw Formation, Late Campanian, Alberta, Western Canada Sedimentary Basin).


Beechy shale
Demaine sand
A
B
Snakebite shale
Ardkenneth sand
C
Transgressive shale
Transgressive sand
Regressive sand
E
D

A B

FIGURE 4.36 Outcrop examples of maximum regressive surfaces in distal shallow-water settings (arrows). A—maximum regressive surface in a conformable lower shoreface to shelf succession (top of the Magrath Member, Bearpaw Formation, Late Campanian, St. Mary River, Alberta, Western Canada Sedimentary Basin); B—maximum regressive surface in a conformable shelf succession (Beechy Member, Bearpaw Formation, Late Campanian, Saskatchewan, Western Canada Sedimentary Basin). In both cases, the slope breaks indicate textural changes across the maximum regressive surfaces, from coarsening-upward (below) to fining-upward (above).

transgression, becoming progressively younger in an upstream direction (a within-trend facies contact that forms *during* shoreline transgression; Figs. 4.6 and 4.38). Therefore, where dealing with fluvial to estuarine successions it is important to differentiate between the *stratigraphically lowest* surface that defines the base of estuarine facies, which is the low-diachroneity maximum regressive surface, and the highly diachronous facies contact that becomes younger landward with the rate of shoreline transgression. The distinction between these two types of contacts may be made on the basis of juxtaposed facies: the maximum regressive surface separates fluvial from overlying central estuary facies, whereas the within-trend (transgressive) facies contact separates fluvial from overlying bayhead deltas (in a wave-dominated estuarine

A B

FIGURE 4.37 Maximum regressive surface (arrows) in a conformable succession of prodelta facies (Campanian Panther Tongue Formation, Utah). The break in slope gradients indicates textural changes across the surface, from coarsening-upward (below) to fining-upward (above). Photograph B: detail from A.

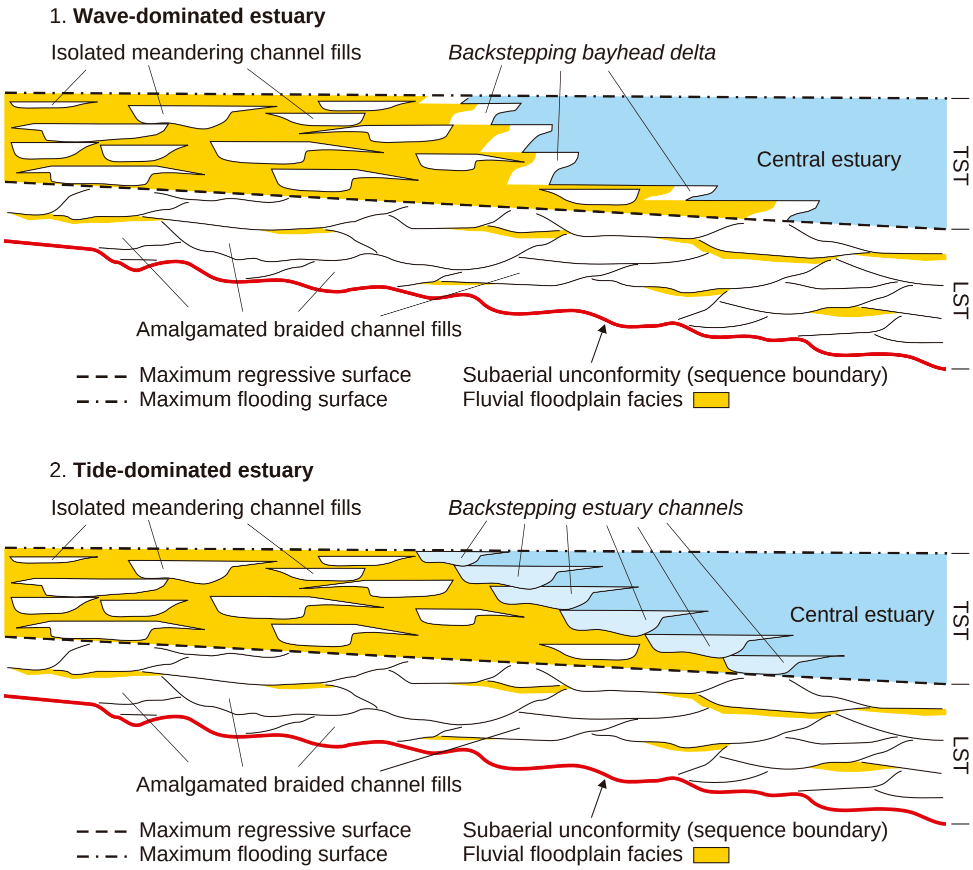


FIGURE 4.38 Dip-oriented stratigraphic cross-sections through fluvial to estuarine successions in wave- and tide-dominated settings (modified from Kerr *et al*., 1999). The lowstand systems tract (LST) is composed of amalgamated braided channel-fill facies resting on a sequence boundary with substantial erosional relief. The transgressive systems tract (TST) is composed of meandering fluvial deposits (isolated ribbons encased in well-developed floodplain facies) and correlative estuarine facies towards the coastline. The maximum regressive surface may be traced at the base of central estuary facies, and at the contact between braided and meandering systems farther inland. Beyond the landward limit of the estuary at the onset of transgression, the facies contact between estuarine and fluvial facies becomes highly diachronous (a within-trend facies contact, within the TST), and may be traced at the base of backstepping bayhead deltas (in wave-dominated settings) or at the base of backstepping estuary channels (in tide-dominated settings).

setting) or estuary channels (in a tide-dominated estuarine setting) (Fig. 4.38).

The extension of the maximum regressive surface into the fluvial part of the basin is much more difficult to pinpoint, but at a regional scale it is argued to correspond with an abrupt decrease in fluvial energy, i.e., a change from amalgamated braided channel fills to overlying meandering systems (Kerr *et al*., 1999; Ye and Kerr, 2000; Fig. 4.38). This shift in fluvial styles across the maximum regressive surface is suggested by the grain size threshold in Fig. 4.6, and is attributed to the formation of the low energy estuarine system at the beginning of transgression, which would induce a lowering in fluvial energy upstream. The link between the formation of estuaries and the coeval lowering in fluvial energy upstream is provided by the increased rates of coastal aggradation at the onset of transgression, which result in a decrease in the slope gradient of the fluvial graded profile and a corresponding change in fluvial energy levels, fluvial styles, and sediment load. Notwithstanding these general principles, much work is still needed to properly document the physical attributes of the nonmarine portion of maximum regressive surfaces. There is increasing evidence that the commonly inferred 'braided' nature of the lowstand fluvial systems (Kerr *et al*., 1999; Ye and Kerr, 2000; Figs. 4.32 and 4.38), even though valid in many cases, may not be representative as a generalization. Lowstand fluvial systems of meandering type have also been documented (e.g., Miall, 2000; Posamentier, 2001; see also the discussion in Chapter 5 regarding the nature of lowstand fluvial deposits), especially within incised valleys, and in such cases the identification of the nonmarine portion of the maximum regressive surface may require more in-depth studies than the simple observation of fluvial styles. Where the maximum regressive surface develops within a succession of meandering stream deposits (lowstand normal regressive below and transgressive above), the stratigraphically lowest sedimentary structures, fossils and trace fossils associated with tidal influences may provide the evidence for the onset of transgression. In this case, well-log and seismic data are not sufficient for unequivocal interpretations, and core or outcrop studies need to be performed for detailed facies analyses.

Following the general trend of fluvial onlap recorded by the underlying lowstand normal regressive deposits, which form a wedge that gradually expands and becomes thinner upstream, the nonmarine portion of the maximum regressive surface may also onlap the subaerial unconformity. The location of the landward termination of the maximum regressive surface depends on basin physiography (landscape gradients), duration of lowstand normal regression, and the rates of fluvial aggradation during lowstand normal regression.

The maximum regressive surface is also known as the *transgressive surface* (Posamentier and Vail, 1988), *top of lowstand surface* (Vail *et al*., 1991), *initial transgressive surface* (Nummedal *et al*., 1993), *conformable transgressive surface* (Embry, 1995), and *maximum progradation surface* (Emery and Myers, 1996). The maximum regressive surface has a low diachroneity along dip that reflects the rates of sediment transport (Catuneanu, 2002; Fig. 4.9). The diachroneity rates may substantially increase along strike, due to the variability in the rates of subsidence and sedimentation (Catuneanu *et al*., 1998b). More details about the temporal attributes of this, as well as all other stratigraphic surfaces, are provided in Chapter 7.

Maximum Flooding Surface

The maximum flooding surface (Frazier, 1974; Posamentier *et al*., 1988; Van Wagoner *et al*., 1988; Galloway, 1989) is also defined relative to the transgressive–regressive curve, marking the end of shoreline transgression (Figs. 4.5 and 4.6). Hence, this surface separates retrograding strata below from prograding (highstand normal regressive) strata above (Figs. 4.9 and 4.39). The presence of prograding strata above

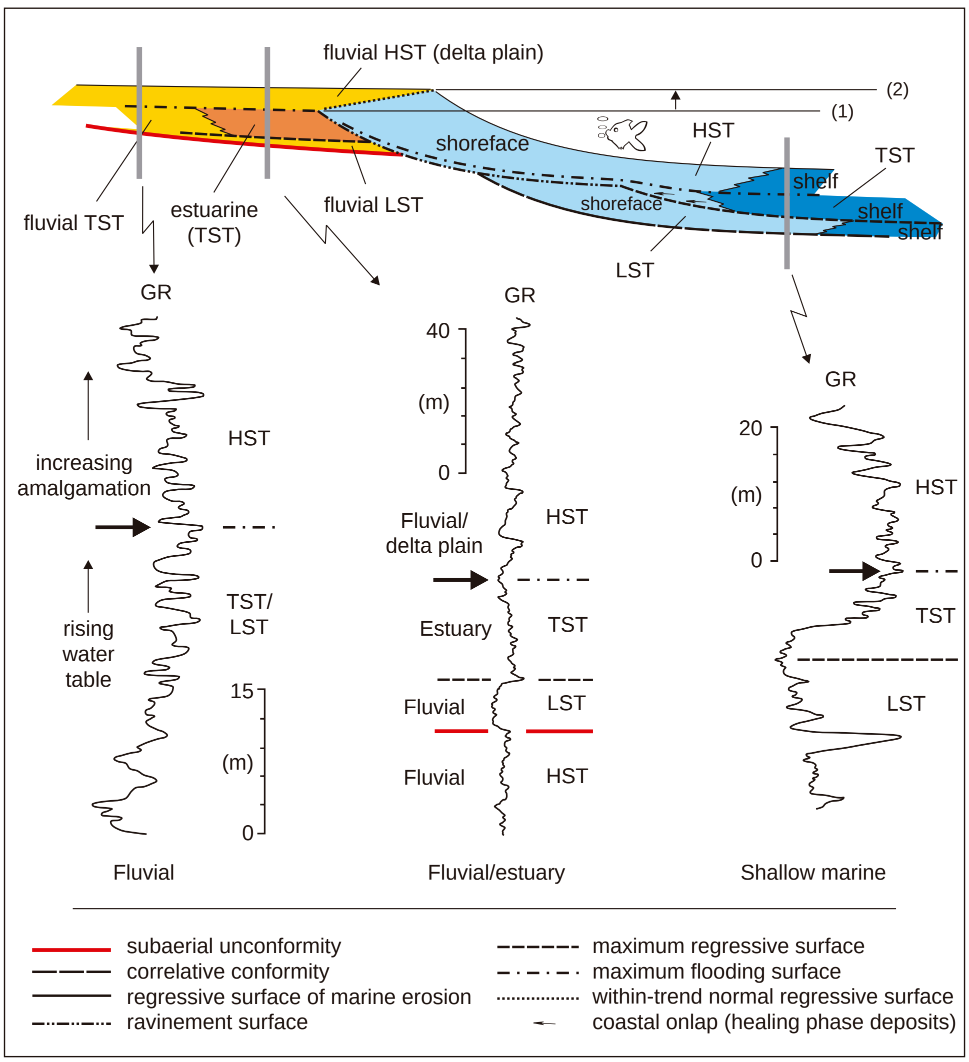


FIGURE 4.39 Well-log expression of the maximum flooding surface (arrows; modified from Catuneanu, 2002, 2003). See Fig. 4.9 for a summary of diagnostic features of the maximum flooding surface. Log examples from the Wapiti Formation, Western Canada Sedimentary Basin (left and center), and Embry and Catuneanu (2001) (right). Abbreviations: GR—gamma ray; LST—lowstand systems tract; TST—transgressive systems tract; HST—highstand systems tract.

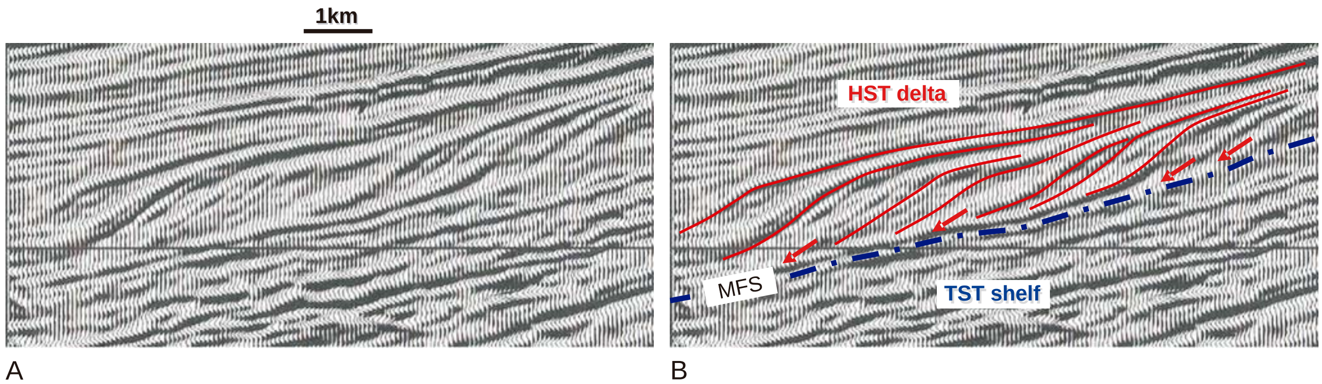


FIGURE 4.40 Seismic expression of a maximum flooding surface in a coastal to shallow-marine setting (A—uninterpreted seismic line; B—interpreted seismic line; modified from Brown *et al*., 1995). The maximum flooding surface overlies transgressive shelf facies, and is downlapped by a highstand (normal regressive) delta. For this reason, the maximum flooding surface is also known as a 'downlap surface'.

identifies the maximum flooding surface as a *downlap surface* on seismic data (Fig. 4.40). The change from retrogradational to overlying progradational stacking patterns takes place *during* base-level rise at the shoreline, when sedimentation rates start to outpace the rates of base-level rise (Fig. 4.5). The maximum flooding surface is generally conformable, excepting for the outer shelf and upper slope regions where the lack of sediment supply coupled with instability caused by rapid increase in water depth may leave the seafloor exposed to erosional processes (Galloway, 1989; Fig. 4.41). The maximum flooding surface is also known as the *maximum transgressive surface* (Helland-Hansen and Martinsen, 1996) or *final transgressive surface* (Nummedal *et al*., 1993). The maximum flooding surface has a low diachroneity along dip that reflects the rates of sediment transport (Catuneanu, 2002; Fig. 4.9). As in the case of the maximum regressive surface, the diachroneity rates may substantially increase along strike due to the variability in subsidence and sedimentation rates (Catuneanu *et al*., 1998b).

Maximum flooding surfaces are arguably the easiest stratigraphic markers to use for the subdivision of stratigraphic successions, especially in marine to coastal plain settings, because they lie at the heart of areally extensive condensed sections which form when the shoreline reaches maximum landward positions (Galloway, 1989; Posamentier and Allen, 1999). Such condensed sections are relatively easy to identify and correlate on any type of data, as they consist dominantly of fine-grained, hemipelagic to pelagic deposits accumulated during times when minimal terrigenous sediment is delivered to the shelf and deeper-water environments. Condensed sections are typically marked by relatively transparent zones on seismic lines, due to their lithological homogeneity. They also tend to exhibit a high gamma-ray response caused by their common association with increased concentrations of organic matter and radioactive elements. One must note, however, that the generally inferred correlation between condensed sections and organic-rich sediments is subject to exceptions, as the deposition and

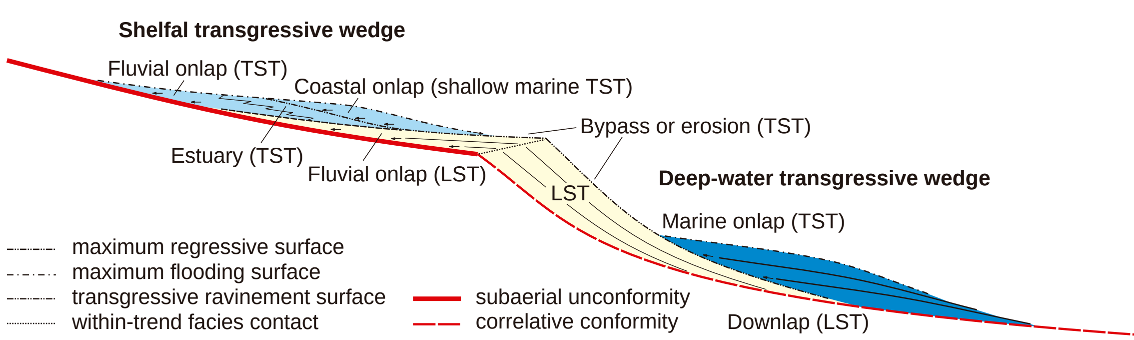


FIGURE 4.41 Stratigraphic expression of transgressive strata. Note that the transgressive systems tract may consist of two distinct wedges, one on the continental shelf and one in the deep-water environment, separated by an area of sediment bypass or erosion around the shelf edge.

preservation of organic matter may merely reflect stages of restricted bottom-water circulation, diminished terrigenous sediment supply, and/or the accumulation of carbonaceous mudstones in paralic environments, which may not necessarily correspond to times of maximum shoreline transgression (Posamentier and Allen, 1999). At the same time, condensed sections associated with stages of maximum flooding may contain glauconite and/or siderite, or other carbonates or biochemical precipitates (Fig. 4.42) which may exhibit a wide range of log motifs (Posamentier and Allen, 1999). For these reasons, well-log data must be integrated with any other available data sets, as well as with the observation of the regional stratal stacking patterns, for more reliable interpretations. As a general principle, '… the identification of a condensed section and a maximum flooding surface should be based on the identification of a convergence of time horizons rather than degree of radioactivity. Converging well-log correlation markers, converging seismic reflections, or converging strata can indicate convergence of time horizons.' (Posamentier and Allen, 1999).

Maximum flooding surfaces have a high preservation potential, being overlain by aggrading and prograding highstand normal regressive deposits, and can be identified in all depositional environments of a sedimentary basin, seaward and landward from the shoreline, on the basis of stratal stacking patterns (Fig. 4.9). The broad areal extent, as well as its consistent association with fine-grained, low energy systems across the basin, makes the 'maximum flooding' a surface that is, in many instances, easier to identify than the subaerial unconformity, and potentially more useful as a stratigraphic marker for basin-wide correlations. The basin-wide extent of the transgressive tract may, however, be hampered by the absence of transgressive deposits in the area around the shelf edge. For this reason, the transgressive systems tract usually comprises two distinct wedges, one on the continental shelf consisting of fluvial to shallow-marine facies, and one in the deep-water environment (Fig. 4.41). Each of these transgressive wedges is topped by a conformable maximum flooding surface which onlaps the fluvial landscape or the continental slope in a landward direction, and downlaps the shallow or deep-marine seafloor in a basinward direction (Figs. 4.9 and 4.41). The downlap type of stratal terminations may, however, be only apparent in a transgressive context, as potentially marking the base of a sedimentary unit at its erosional rather than depositional limit (Fig. 4.2), which is why depositional downlap is commonly restricted to regressive deposits (Fig. 4.3). Where transgression is accompanied by sediment aggradation, the transgressive strata do not terminate against the maximum flooding surface, which rather drapes the underlying deposits. This principle is valid for all conformable stratigraphic surfaces listed in Fig. 4.9, meaning that sedimentary strata do not terminate against a younger conformable surface. Where the transgressive facies are absent, the maximum flooding surface truncates the underlying regressive deposits (Fig. 4.9).

FIGURE 4.42 Coal seam (1 m thick) in a coastal setting, overlain by a 50 cm thick limestone bed (photograph courtesy of M.R. Gibling; Pennsylvanian Sydney Mines Formation, Sydney Basin, Nova Scotia). The coal lies within the transgressive systems tract. The limestone bed (arrow) marks a maximum flooding level with restricted detrital supply, and it is overlain by the highstand systems tract.

In a marine succession, the maximum flooding surface is placed at the top of fining-upward (transgressive) deposits. This trend is generally valid in both deep-water settings, where the maximum flooding surface marks the top of waning-down gravity-flow deposits (base of highstand pelagics—see the following chapters for more details), as well as in shallow-water environments. Seaward from the shoreline, on the shelf, the transgressive deposits may be reduced to a condensed section, or may even be missing. In the latter situation, the maximum flooding surface is superimposed on and reworks the maximum regressive surface. Figure 4.43 provides an example where the transgressive deposits are present, and hence the succession is conformable. In this case, the maximum flooding surface corresponds to the peak of finest sediment, marking the top of a fining-upward (transgressive) succession. This surface is not easy to pinpoint in outcrop or core, as it is not associated with a lithological contrast, and it requires thin section textural analysis for unequivocal identification. However, such a conformable maximum flooding surface is easier to recognize on well logs, which are more sensitive in recording changes in grain size. Under restricted detrital supply conditions,

FIGURE 4.43 Maximum flooding surface in a conformable shallow-marine succession, at the base of the transition zone between shelf and overlying shoreface facies. The succession is younging to the left. The vertical dashed line marks the peak of finest sediment (top of retrograding succession). This conformity is difficult to pinpoint in the field because of the lack of lithological contrast, and requires thin section textural analysis for accurate identification. Transition between the Sherrard and Demaine members (Bearpaw Formation, Late Campanian), Saskatchewan, Western Canada Sedimentary Basin.

the maximum flooding level may also be marked by condensed sections of carbonate facies (Fig. 4.42). Where the transgressive deposits are missing, the maximum flooding surface is scoured and replaces the maximum regressive surface. In this case, the maximum flooding surface is associated with a lithological contrast and separates two coarsening-upward successions (Fig. 4.44).

Where transgressive deposits are present and the succession is conformable, the top of fining-upward retrograding marine facies does not necessarily correspond to the peak of deepest water, especially in offshore areas. The peak of deepest water is usually recorded within the overlying regressive (highstand) deposits (see Chapter 7 for more details). This is why, as in the case of the maximum regressive surface, grading terms that reflect *observations* (coarsening- *vs*. fining-upward) are preferred over bathymetric terms that reflect *inferred* changes in water depth (shallowing- *vs*. deepening-upward).

The ichnological signature of maximum flooding surfaces in a marine succession is highly variable, depending on the dominant synsedimentary process (i.e., sediment aggradation *vs*. bypass, erosion and/or

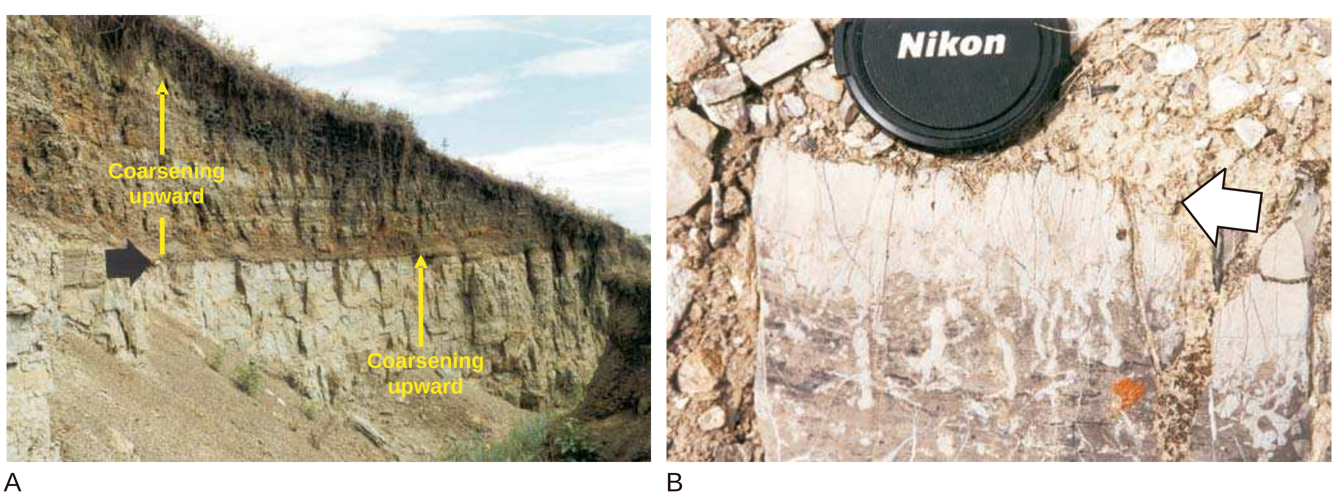


FIGURE 4.44 Outcrop examples of maximum flooding surfaces (scoured) that rework the underlying maximum regressive surfaces. The transgressive facies are missing. A—Young Creek Member (Bearpaw Formation, Early Maastrichtian), Castor area, Alberta, Western Canada Sedimentary Basin; B—firmground associated with the *Glossifungites* ichnofacies, formed as a result of prolonged sediment starvation (Mississippian Shunda Formation, Talbot Lake area, Jasper National Park).

lithification) that affects the seafloor during the maximum transgression of the shoreline. Due to the decrease in sediment supply to the marine environment during shoreline transgression, the maximum flooding surface is often associated with firmgrounds or hardgrounds, as a function of degree of seafloor cementation (Fig. 4.44B), although softgrounds may also form where sedimentation rates are high enough to maintain an unconsolidated seafloor (Fig. 4.43) (Pemberton and MacEachern, 1995; Savrda, 1995; Ghibaudo *et al*., 1996). Ghibaudo *et al*. (1996) provide a case study where the maximum flooding surface is represented by a firmground with burrows infilled with glauconitic sandstone. This stratigraphic contact ('omission' surface) is interpreted to correspond to a period of very low sedimentation rates or nondeposition, where the lack of clastic input allowed for glauconite formation and concentration, intense seafloor burrowing and increased cohesiveness of the substrate (Ghibaudo *et al*., 1996). In this example, the formation of the firmground was accompanied by a decrease in the water's oxygen levels at the seafloor, as evidenced by the preservation of plant debris as well as by the abundance of *Phycosiphon incertum* and *Planolites* traces (Ghibaudo *et al*., 1996). The landward shift of facies during transgression is also confirmed by the change in softground ichnofacies across the firmground, from *Cruziana* below to *Zoophycos* above (Ghibaudo *et al*., 1996). The latter ichnofacies is consistent with an oxygen-deprived setting (Pemberton and MacEachern, 1995; Ghibaudo *et al*., 1996), although the association between maximum flooding surfaces and oxygen-deficient ichnocoenoses is not necessarily a valid generalization, especially in the proximal regions of shallow-marine environments where the water may be well oxygenated during times of maximum shoreline transgression (Savrda, 1995). At the opposite end of the spectrum, Siggerud and Steel (1999) provide a case study where the maximum flooding surface formed during a time of continuous seafloor aggradation, which did not allow for the formation of firmgrounds or hardgrounds. In this case, the position of the maximum flooding surface is inferred on the basis of changes in ichnofabrics, corresponding to the point of highest bioturbation index. The increased level of bioturbation at the maximum flooding surface softground, which is not necessarily accompanied by any abrupt changes in ichnofacies across the conformable stratigraphic contact, correlates with the amount of sediment supply delivered to the marine environment (and the corresponding rates of seafloor aggradation), which is lowest during the time of maximum shoreline transgression. This example is relevant to all conformable shallow-marine successions, where *sediment supply* (as opposed to inferred changes in water depth) is the main switch that controls the observed grading patterns, sedimentation rates, and associated levels of bioturbation. Besides softgrounds, firmgrounds and hardgrounds, maximum flooding surfaces may also be represented by woodgrounds especially in coastal regions where marine flooding results in the inundation of forested coastal plains (Savrda, 1995). Such woodgrounds are common at all flooding surfaces that form during shoreline transgression, and are preserved *within* the transgressive systems tract, so it is only the *youngest* woodground of any transgressive succession that indicates the position of the maximum flooding surface. It can be concluded that all substrate-controlled ichnofacies may, under different circumstances, be associated with maximum flooding surfaces (Fig. 4.9), although softgrounds characterized by increased bioturbation indexes and changes in ichnofabrics in conformable marine successions should not be ruled out (Savrda, 1995; Siggerud and Steel, 1999).

In coastal settings, the maximum flooding surface is placed at the top of the youngest estuarine facies, marking the turnaround point to subsequent delta plain sedimentation (Figs. 4.6, 4.38, and 4.39). Landward from the coastline, criteria for the recognition of the maximum flooding surface in the fluvial portion of the basin have been provided by Shanley *et al*. (1992), mainly based on the presence of tidal influences in fluvial sandstones. Sedimentary and biogenic structures that may suggest a tidal influence in fluvial strata include sigmoidal bedding, paired mud/silt drapes, wavy and lenticular bedding, shrinkage cracks, multiple reactivation surfaces, inclined heterolithic strata, complex compound cross-beds, bidirectional cross-beds, and trace fossils including *Teredolites*, *Arenicolites*, and *Skolithos* (Shanley *et al*., 1992). Tidal influences in fluvial strata generally extend for tens of kilometers inland from the coeval shoreline (Shanley *et al*., 1992), although, depending on river discharge and tidal range, such influences, including tidal-current reversals, may occur as far as 130 km (Allen and Posamentier, 1993) or even over 200 km inland from the river mouth (Miall, 1997). Farther upstream, the maximum flooding surface corresponds to the highest level of the water table relative to the land surface (Fig. 4.39), which, given a low sediment input and the right climatic conditions, may offer good conditions for peat accumulation at the basin scale. As a result, the position of the maximum flooding surface may be indicated by regionally extensive coal seams (Hamilton and Tadros, 1994; Tibert and Gibling, 1999). Given its association with high water table conditions, the maximum flooding surface is likely included within floodplain and/or lacustrine sediments, and it

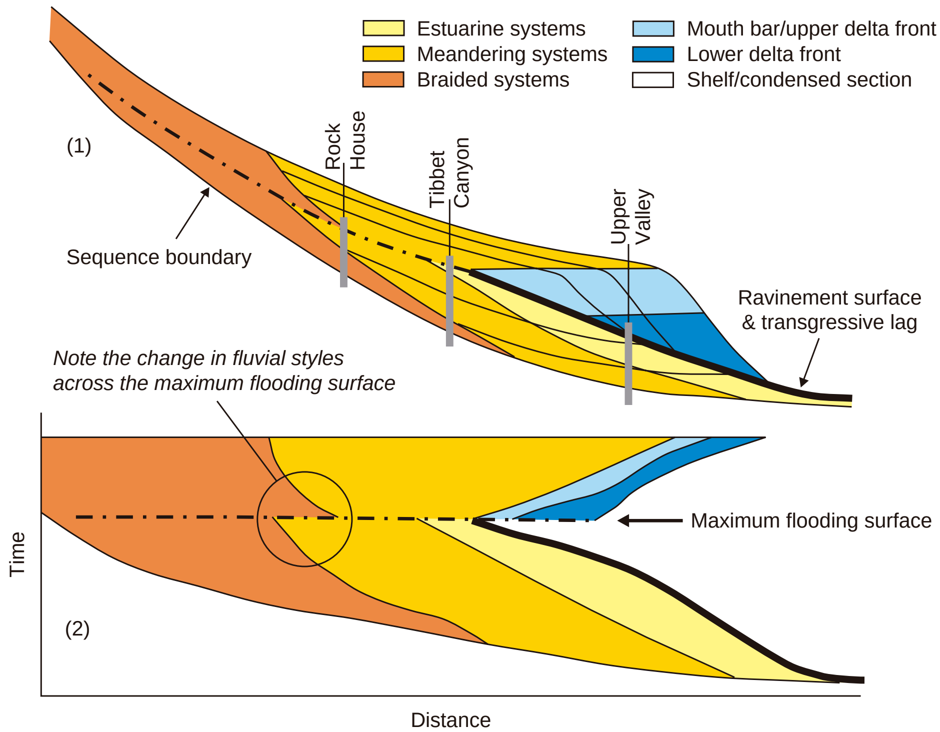


FIGURE 4.45 Dip-oriented stratigraphic cross-section (1) and chronostratigraphic (Wheeler) diagram (2) of a fluvial to shallow-marine depositional sequence (modified from Shanley *et al.*, 1992). This case study, based on the Upper Cretaceous succession in southern Utah, suggests that the end of the estuary life time (end of transgression) is accompanied by an abrupt shift in fluvial styles upstream, which provides a criterion for the recognition of the nonmarine portion of the maximum flooding surface. The landward shift through time of the boundary between braided and meandering stream facies explains the fining-upward trend within the fluvial part of each systems tract. This trend is also suggested in Fig. 4.6 (including the threshold of facies shift across the maximum flooding surface), and has been observed in other case studies as well (e.g., Catuneanu and Elango, 2001).

may be prograded by crevasse deltas and lake deltas as the balance between accommodation and sedimentation shifts again in the favor of the latter.

The position of the maximum flooding surface in fully fluvial successions may also be indicated by an abrupt increase in fluvial energy, from meandering to overlying braided fluvial systems, as the end of the estuary life time triggers a rapid seaward shift of the river mouth (Shanley *et al.*, 1992; Fig. 4.45). This change in fluvial styles across the maximum flooding surface is suggested by a grain size threshold in Fig. 4.6. It should be noted, however, that this scenario only reflects the particular circumstances of a case study, and it may not be adequate as a generalization. Depending on the patterns of differential subsidence and sediment supply of each basin, other changes in fluvial styles may also be envisaged across maximum flooding surfaces. As a general principle, continuous coastal aggradation during transgression and subsequent highstand normal regression contributes towards a gradual *decrease* in the gradient of fluvial graded profiles. As a result, a lowering with time in fluvial energy should be expected, unless the effects of tectonism and differential subsidence overprint this trend. Irrespective of the actual change in fluvial energy levels and corresponding fluvial styles across the maximum flooding surface, which should therefore be studied on a case-by-case basis, the highstand fluvial deposits overlying a maximum flooding surface record an abrupt decline in tidal structures, as well as a gradual increase in the degree of channel amalgamation as the amount of available accommodation decreases towards the end of base-level rise (Fig. 4.39; Wright and Marriott, 1993; Shanley and McCabe, 1993; Emery and Myers, 1996). Most of the current models of fluvial sequence stratigraphy acknowledge these changes in sedimentary structures and the ratio between fluvial architectural elements, without accounting for a shift in fluvial styles across the maximum flooding surface.

Transgressive Ravinement Surfaces

Transgressive ravinement surfaces are scours cut by tides and/or waves during the landward shift of the shoreline. In the majority of cases, the two types of transgressive ravinement surfaces (i.e., tide- and wave-generated) are superimposed and onlapped by the transgressive shoreface (i.e., coastal onlap). Such amalgamated transgressive scours form commonly in open shoreline settings, and, where all retrograding facies are preserved, separate backstepping (transgressive) beach deposits below from transgressive shoreface strata above. Depending on the amount of ravinement scouring during transgression, the beach and underlying fluvial transgressive facies may not be preserved, and in this case the transgressive ravinement surface may truncate older, normal regressive (lowstand or even highstand) strata. For this reason, the facies that may be found below a transgressive ravinement surface are variable, from fluvial to coastal or shallow-marine, whereas the facies above are always shallow-marine (Fig. 4.9).

In transgressive river-mouth settings, either wave- or tide-dominated, the two types of transgressive ravinement surfaces may be preserved as distinct scoured contacts separated by the sandy deposits of the estuary-mouth complex (Figs. 4.46 and 4.47). In such cases, the tidal and wave scouring during shoreline transgression take place at the same time but in different areas, within the estuary and the upper shoreface, respectively (Figs. 4.46 and 4.47). As a result, the tidal-ravinement surface is placed at the contact between central estuary muds (or older variable facies where central estuary sediments are not preserved) below, and the estuary-mouth complex above (Fig. 4.9). The age-equivalent wave-ravinement surface is placed at the contact between the estuary-mouth complex below, and the transgressive shallow-marine deposits above. This scenario is based on the assumption that the rates of aggradation of the estuary-mouth complex are higher than the rates of subsequent wave-ravinement erosion, because otherwise the wave-ravinement surface would rework the tidal-ravinement surface, and the two contacts would be superimposed. Where the estuary-mouth complex is preserved in the rock record, the wave-ravinement surface is always intercepted in vertical profiles at a higher stratigraphic level than the tidal-ravinement surface, due to the retrogradational shift of facies during transgression (e.g., see Allen and Posamentier, 1993, for a case study).

The transgressive ravinement surfaces provide the most favorable conditions for the formation of substrate-controlled ichnofacies, as they are omission surfaces that are always scoured and overlain by marginal-marine to shallow-marine facies. Depending on the amount of tidal and/or wave scouring, as well as on the nature of facies that are subject to erosion, the transgressive ravinement surfaces may be marked by firmgrounds (*Glossifungites* ichnofacies; Figs. 2.25

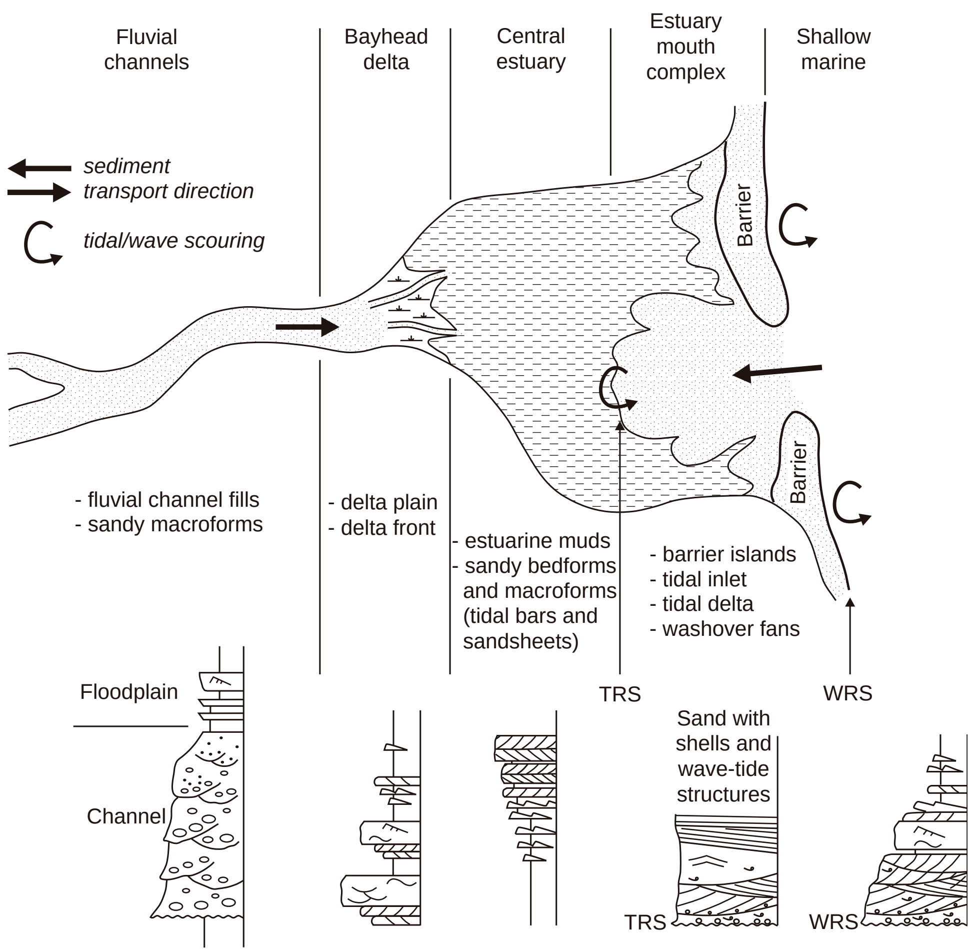


FIGURE 4.46 Tidal- and wave-ravinement surfaces in a wave-dominated estuarine setting (modified from Dalrymple *et al*., 1992; Reinson, 1992; Zaitlin *et al*., 1994; Shanmugam *et al*., 2000). As facies retrograde during transgression, both tidal- and wave-ravinement surfaces expand in a landward direction, truncating central estuary and estuary-mouth complex facies, respectively. Abbreviations: TRS—tidal-ravinement surface; WRS—wave-ravinement surface.

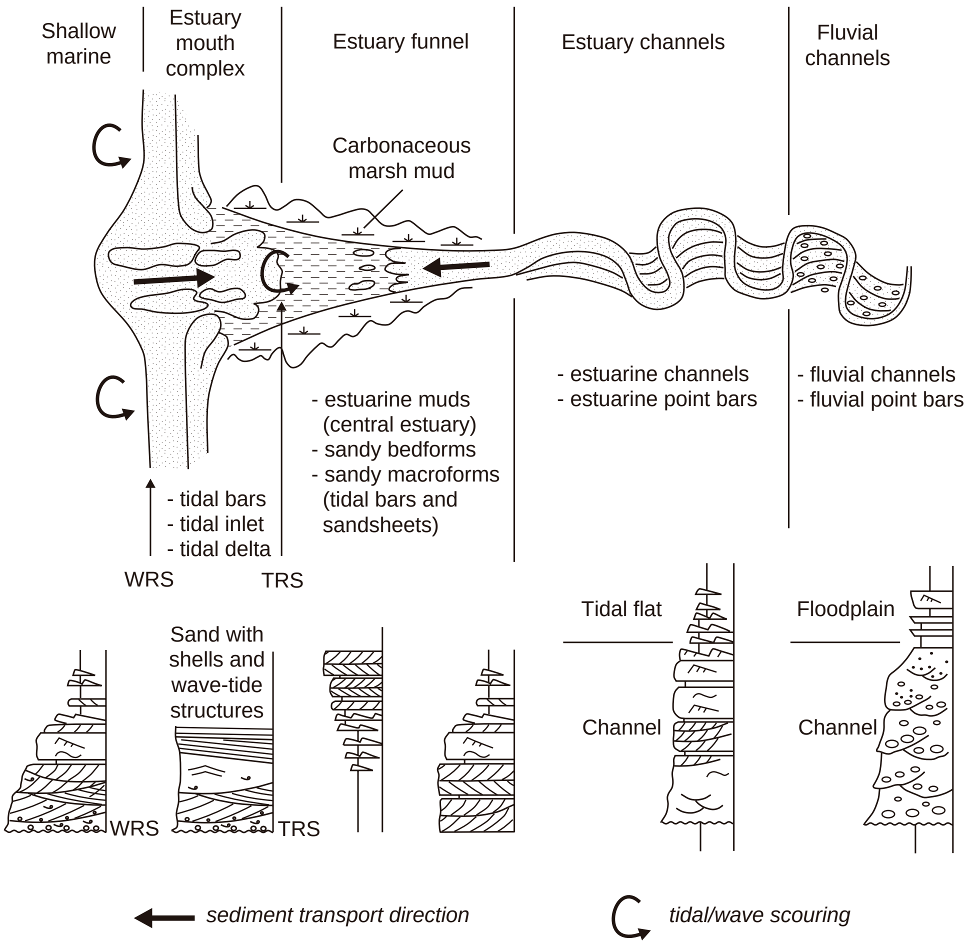


FIGURE 4.47 Tidal- and wave-ravinement surfaces in a tide-dominated estuarine setting (modified from Allen, 1991; Dalrymple *et al.*, 1992; Allen and Posamentier, 1993; Shanmugam *et al.*, 2000). As facies retrograde during transgression, both tidal- and wave-ravinement surfaces expand in a landward direction, truncating central estuary and estuary-mouth complex facies, respectively. Abbreviations: TRS—tidal-ravinement surface; WRS—wave-ravinement surface.

and 2.26), hardgrounds (*Trypanites* ichnofacies; Fig. 2.27), or woodgrounds (*Teredolites* ichnofacies; Fig. 2.28). The colonization of these substrates takes place within a relatively short interval of time, depending on the rates of shoreline transgression, either during or immediately after the ravinement surface is cut (MacEachern *et al.*, 1992). Following the formation of substrate-controlled ichnofacies, transgressive ravinement surfaces are gradually onlapped by the landward-shifting marginal-marine to shallow-marine facies (i.e., coastal onlap: Figs. 4.2 and 4.9). Numerous case studies documenting the ichnology of transgressive ravinement surfaces have been published from both modern settings and ancient successions (e.g., MacEachern *et al.*, 1992, 1999; Taylor and Gawthorpe, 1993; Pemberton and MacEachern, 1995; Ghibaudo *et al.*, 1996; Krawinkel and Seyfried, 1996; Pemberton *et al.*, 2001; Gingras *et al.*, 2004).

Wave-Ravinement Surface

The wave-ravinement surface is a scour cut by waves in the upper shoreface during shoreline transgression, in an attempt to maintain the shoreface profile that is in balance with the wave energy (Bruun, 1962; Swift *et al.*, 1972; Swift, 1975; Dominguez and Wanless, 1991; the 'wave scour' in Fig. 3.20). This erosion may remove as much as 10–20 m of substrate (Demarest and Kraft, 1987; Abbott, 1998), as a function of the wind regime and related wave energy in each particular coastal region. Under exceptional circumstances, in coastal settings characterized by extreme wave energy, the thickness of material being removed by ravinement scouring may reach 40 m, as documented along the Canterbury Plains of New Zealand (Leckie, 1994). At the opposite end of the spectrum, the amount of erosion associated with transgressive wave scouring may be negligible where the transgressed surface is indurated by various pedogenic processes (Fig. 4.14). The wave-ravinement surface is onlapped during the retrogradational shift of facies by transgressive (fining-upward) shoreface deposits (coastal onlap), and it may overlie any type of depositional system (fluvial, coastal, or marine). The wave-ravinement surface is highly diachronous, with the rate of shoreline transgression (Fig. 4.9).

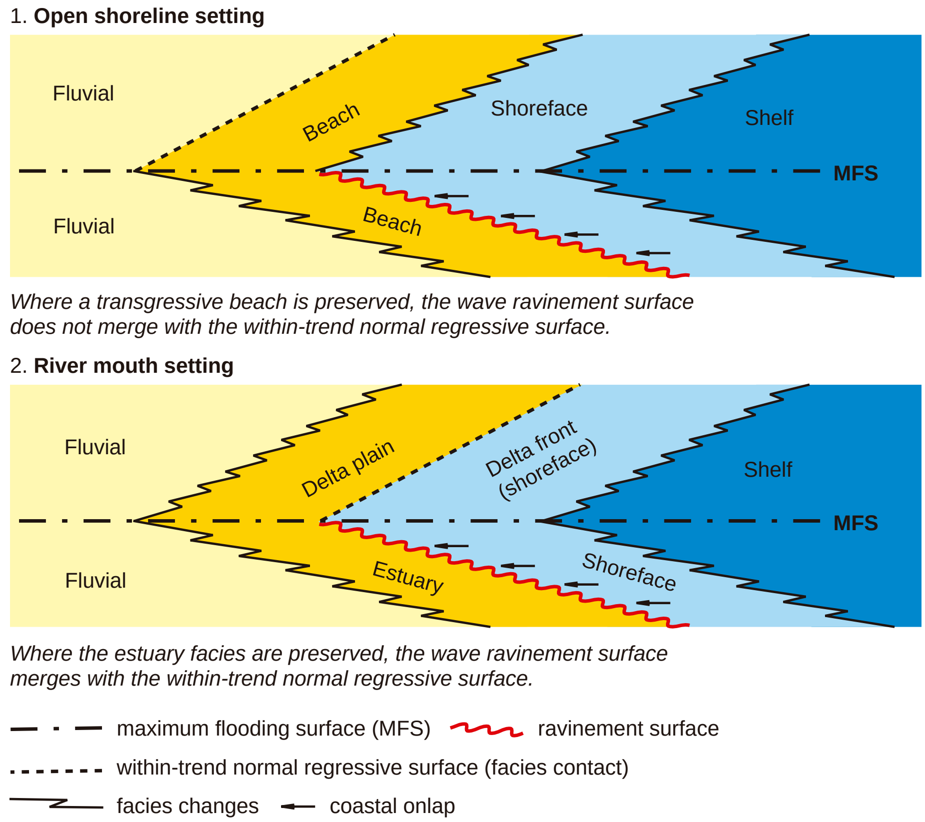


FIGURE 4.48 Architecture of facies and stratigraphic surfaces at the point of maximum shoreline transgression (from Catuneanu, 2002). The position of the within-trend normal regressive surface varies with the type of coastline, between open shoreline and river-mouth settings. The wave-ravinement surface always sits at the base of transgressive shoreface facies. The maximum flooding surface separates retrograding from overlying prograding geometries. Within the transgressive systems tract, the facies contact between shoreface sands and the overlying shelf shales defines the within-trend flooding surface.

In a vertical profile that preserves the entire succession of facies, the wave-ravinement surface separates coastal strata below (backstepping foreshore and backshore facies in an open shoreline setting, or estuarine facies in a river-mouth setting) from shoreface and shelf deposits above (Figs. 4.6, 4.48, and 4.49). Where the transgressive coastal and fluvial deposits are not preserved, the wave-ravinement surface may rework the underlying lowstand normal regressive strata and even the subaerial unconformity (Embry, 1995; Fig. 4.49). In the latter case, the wave-ravinement surface becomes part of the sequence boundary. The chances for a wave-ravinement surface to replace the underlying subaerial unconformity depend on the balance between the thickness of the lowstand normal regressive strata and the amount of subsequent wave-ravinement erosion, and are highest in the case of short stages of lowstand normal regression and/or low rates of aggradation during the lowstand normal regression. Where stages of lowstand normal regression result in the deposition of thick (> 20 m) fluvial to coastal deposits, the subaerial unconformity is preserved as such in the rock record (Fig. 4.34).

In stratigraphic sections located immediately landward from the shoreline position at the onset of transgression, it is common for the wave-ravinement surface to rework the maximum regressive surface and the underlying lowstand beach, coastal plain or delta plain strata, and therefore to be found within a fully shallow-marine succession (Fig. 4.49). In such cases, the distinction between a wave-ravinement surface and the marine portion of a maximum regressive surface (scoured and conformable contacts, respectively, both separating coarsening-upward strata below from fining-upward strata above; Figs. 4.9, 4.32, and 4.49), solely from the study of well logs, may be difficult (compare the well logs in Figs. 4.32 and 4.49). Under these circumstances, additional information (e.g., core material) is required for the unequivocal identification of the wave-ravinement surface, in order to document the scoured nature of this stratigraphic contact (Fig. 2.26). Owing to their mode of formation, wave ravinement surfaces are commonly marked by the concentration of transgressive lag deposits, which can be best observed in outcrop or core (Fig. 4.50). Where developed within fully marine successions, wave-ravinement surfaces are commonly demarcated by firmgrounds (*Glossifungites* ichnofacies; MacEachern *et al*., 1992) or hardgrounds (*Trypanites* ichnofacies; e.g., case study by Krawinkel and Seyfried, 1996,

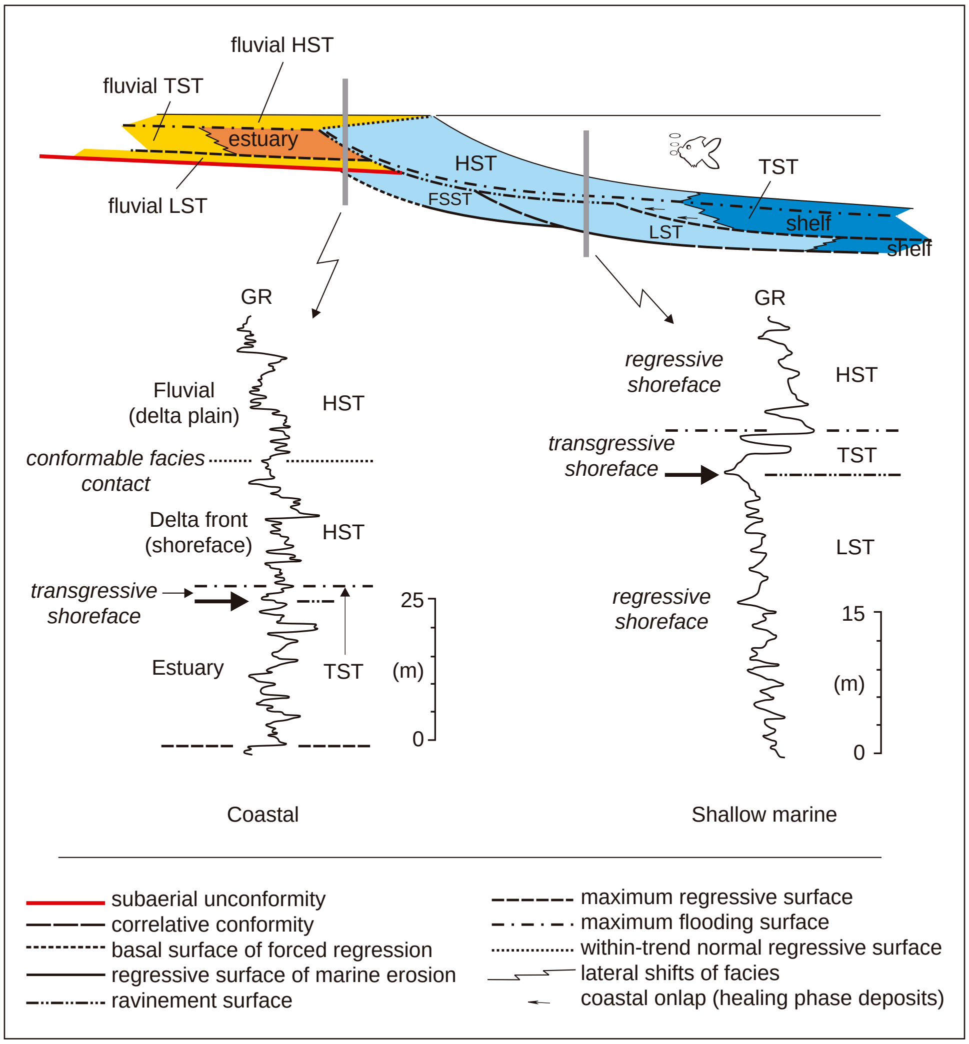


FIGURE 4.49 Well-log expression of the transgressive wave-ravinement surface (arrows; modified from Catuneanu, 2003). See Fig. 4.9 for a summary of diagnostic features of the transgressive wave-ravinement surface. Note that in fully shallow-marine successions, the transgressive wave-ravinement surface may replace the maximum regressive surface, if the log is located landward relative to the shoreline position at the onset of transgression. Log examples from the Bearpaw Formation (left) and Embry and Catuneanu (2001) (right). Abbreviations: GR—gamma ray; LST—lowstand systems tract; TST—transgressive systems tract; FSST—falling-stage systems tract; HST—highstand systems tract.

where the wave-ravinement surface is a wave-cut platform with *Gastrochaenolites* borings and a thin veneer of transgressive lag, cut into regressive shoreface deposits and overlain by transgressive shoreface facies). In stratigraphic sections located farther inland relative to the shoreline position at the onset of transgression, the chances of preservation of nonmarine deposits beneath a wave-ravinement surface are higher, and as a result such transgressive scours are commonly cut into rooted nonmarine facies capped by firmgrounds (*Glossifungites* ichnofacies) or woodgrounds (*Teredolites* ichnofacies) (MacEachern *et al.*, 1992; Pemberton *et al.*, 2001). The presence of coal beds within the nonmarine succession that is subject to transgressive wave scouring may limit the amount of downcutting, due to the more resilient nature of coal, and as a result many wave-ravinement surfaces are found directly on top of xylic substrates (Fig. 4.51).

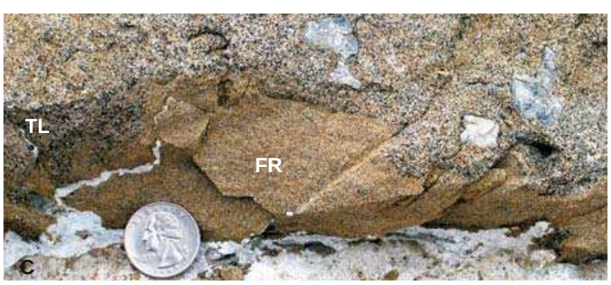
The term ‘wave-ravinement surface’ was introduced by Swift (1975); synonymous terms include the *transgressive surface of erosion* (Posamentier and Vail, 1988), *shoreface ravinement* (Embry, 1995) and *transgressive ravinement surface* (Galloway, 2001). Figure 4.51 provides a field example of a ravinement surface that separates coal-bearing fluvial floodplain strata from the overlying transgressive shoreface facies. In this example, no coastal deposits are preserved following the wave-ravinement erosion, and the fluvial deposits are transgressive (fluvial transgressive facies in Fig. 4.6). As a result, this particular wave-ravinement surface develops *within* a transgressive systems tract, and it is not part of a systems tract or sequence boundary.

Tidal-Ravinement Surface

The tidal-ravinement surface is a scour cut by tidal currents in coastal environments during shoreline

A

B



C

FIGURE 4.50 Outcrop examples of transgressive lag deposits associated with wave-ravinement surfaces. A—plan view of a wave-ravinement surface, showing the presence of transgressive lag deposits (plant debris in this case). In this example, the wave-ravinement surface is at the top of coarsening-upward shoreface deposits (a 'parasequence'), and it is overlain by transgressive shales (not shown). This type of sharp lithological contact, from sands below to shales above, qualifies the wave-ravinement surface as a 'flooding surface' (Late Cretaceous Blackhawk Formation, Utah); B, C—wave-ravinement surface associated with transgressive lag deposits ('TL'—coarse sandstone with shell fragments). In this example, the wave-ravinement surface is at the top of forced regressive delta front deposits ('FR'), is overlain by transgressive marine shales (not shown), and reworks the subaerial unconformity. This wave-ravinement surface also fits the definition of a 'flooding surface,' and, in this case, is part of the sequence boundary (Campanian Panther Tongue Formation, Gentle Wash Canyon, Utah).

FIGURE 4.51 Wave-ravinement surface separating transgressive shoreface facies with Oyster coquina from the underlying coal-bearing fluvial facies. The latter are interpreted as transgressive (Hamblin, 1997), hence this portion of the ravinement surface develops within a transgressive systems tract and it is not a systems tract or sequence boundary. Contact between the Dinosaur Park Formation (Belly River Group) and the Bearpaw Formation, southern Alberta, Western Canada Sedimentary Basin.

transgression. Depending on the nature of coastal deposits that are subject to scouring, as well as the magnitude of tidal erosion, tidal-ravinement surfaces may be demarcated by firmgrounds (Fig. 2.25), hardgrounds (Fig. 2.27) or woodgrounds (Fig. 2.28). The formation of such scour surfaces may be observed along present-day transgressive coastlines (Figs. 2.25B, 2.27B, and 2.28), or in the rock record where the fill of tidal channels is preserved from subsequent transgressive wave-ravinement erosion (Figs. 2.25A and 2.27A). The process of tidal reworking of the underlying transgressive or normal regressive (lowstand or even highstand) deposits is equally important in open shoreline and river-mouth settings, although the type of coastline is a critical factor that controls the preservation of the tidal-ravinement surface as a distinct stratigraphic contact in the stratigraphic record. In open shoreline settings, tidal reworking in the intertidal to coastal plain areas is followed by wave erosion in the upper shoreface, as the shoreline shifts in a landward direction during transgression. For this reason, the tidal-ravinement surface is generally replaced, shortly after formation, by the landward-expanding wave-ravinement surface. This is why the wave-ravinement surface is commonly the only type of transgressive ravinement scour that is referred to in the majority of studies.

The chances of preservation of the tidal-ravinement surface as a distinct stratigraphic contact are enhanced

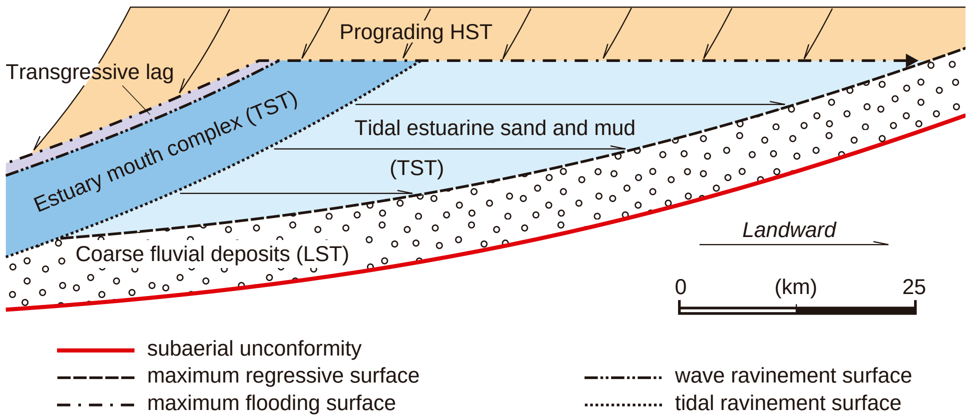


FIGURE 4.52 Stratigraphic model of an incised-valley fill, based on the Gironde estuary (modified from Allen and Posamentier, 1993). Note the spatial relationship between sequence stratigraphic surfaces, as well as their relation with the various facies of the incised-valley fill. This case study provides a most complete scenario, where all systems tracts that form during base-level rise are represented in the rock record of the valley fill.

in transgressive river-mouth settings, where the rates of aggradation of the estuary-mouth complex outpace the rates of subsequent wave-ravinement erosion (Figs. 4.46 and 4.47). In such settings, the two transgressive ravinement surfaces are separated by the sandy deposits of the estuary-mouth complex (Fig. 4.52). In a most complete scenario, where most estuarine facies are preserved, the tidal-ravinement surface occurs at the contact between central estuary muds below and estuary-mouth sands above (Allen and Posamentier, 1993; Fig. 4.52). The preservation of the underlying central estuary muds depends on the balance between the rates of aggradation in the central estuary and the rates of subsequent tidal erosion, as all subenvironments shift in a landward direction. In turn, these two opposing forces, of sedimentation *vs.* erosion, are a function of several variables, including sediment supply, available accommodation, and tidal range. Higher sediment supply contributes towards increased rates of aggradation, whereas a higher tidal range increases the magnitude of tidal scouring, counteracting the effect of sedimentation.

The documentation of tidal-ravinement surfaces is most common in case studies involving incised-valley fills, as the bulk of such deposits is generally tidally influenced and estuarine in origin. The Gironde estuary in France provides a classic example of a mixed tide- and wave-influenced coastal setting, where the fill of the incised valley preserves a full succession of lowstand fluvial, transgressive estuarine, and highstand deltaic sedimentary facies (Fig. 4.52; for core photographs, see Fig. 6 of Allen and Posamentier, 1993). This case study provides a good example of a tidal-ravinement surface at the contact between central estuary and overlying estuary-mouth facies (Allen and Posamentier, 1993). In coastal settings characterized by rapid transgression following the onset of base-level rise, high tidal range, and/or reduced accommodation, the lowstand fluvial deposits, as well as the low energy central estuarine facies may not be preserved in the rock record. In such cases the tidal-ravinement surface reworks the subaerial unconformity, and the underlying highstand facies may range from fluvial to shallow-marine (Figs. 4.9, 4.53, and 4.54). Irrespective of the nature of underlying facies, the preservation of a tidal-ravinement surface as such requires the presence of estuary-mouth complex deposits on top (Fig. 4.9). As with the wave-ravinement surface, the tidal-ravinement surface is highly diachronous, with a rate that matches the rate of shoreline transgression.

WITHIN-TREND FACIES CONTACTS

In addition to the seven sequence stratigraphic surfaces described above, facies contacts associated with a strong physical expression may also be recognized *within* the various systems tracts. Such lithological discontinuities may be caused by shifts in depositional environments accompanied by corresponding changes in environmental energy and sediment supply during transgressions or regressions, and are surfaces of lithostratigraphy or allostratigraphy. They are not proper sequence stratigraphic surfaces as they do not serve as systems tract boundaries. In a sequence stratigraphic approach, within-trend facies contacts need to be dealt with only after the framework of sequence stratigraphic surfaces has been constructed. A discussion of the most prominent types of within-trend facies contacts follows below.

Within-trend Normal Regressive Surface

The within-trend normal regressive surface is a conformable facies contact that develops *during* normal regressions at the top of prominent shoreline

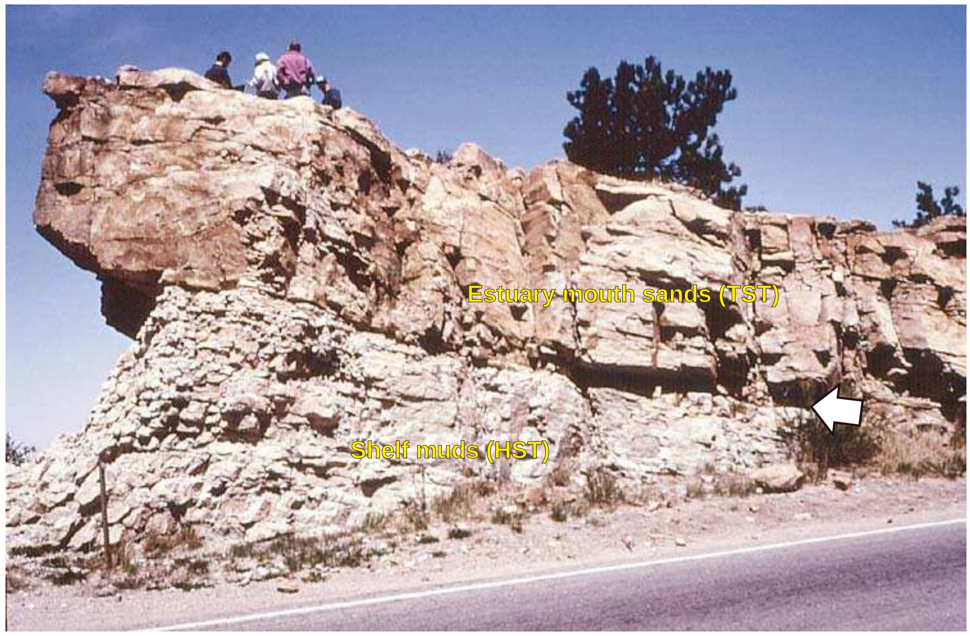


FIGURE 4.53 Incised-valley fill within the Muddy Formation (Ft. Collins, Colorado), showing a *tidal-ravinement surface* (arrow) at the contact between highstand shelf deposits below (Ft. Collins Member) and a transgressive estuary-mouth complex above (Horsetooth Member) (photograph courtesy of H.W. Posamentier). The tidal-ravinement surface reworks the subaerial unconformity, thus becoming part of the sequence boundary. In this example, neither lowstand nor central estuary (transgressive) deposits are preserved following the tidal-ravinement scouring. The transgressive *wave-ravinement surface* is expected to rework the top of the estuary-mouth complex.

sands (Figs. 4.9 and 4.55). The formation of this facies contact therefore requires coeval progradation and aggradation, which bring lower energy supratidal sediments on top of higher energy subtidal to intertidal facies. The underlying prominent coarser deposits may

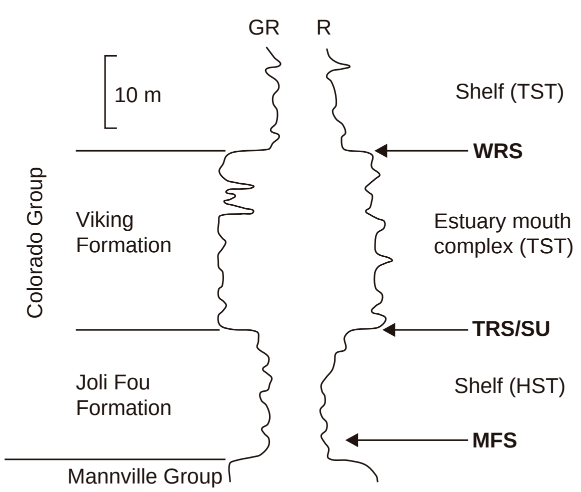


FIGURE 4.54 Well-log expression of a tidal-ravinement surface at the contact between highstand shelf deposits below and transgressive estuary-mouth sands above (Colorado Group, Crystal Field, Alberta). The estuary-mouth complex forms the fill of an incised valley, and is capped by a wave-ravinement surface. The tidal-ravinement surface reworks the subaerial unconformity, thus becoming part of the sequence boundary. Abbreviations: GR – gamma ray log; R – resistivity log; TST – transgressive systems tract; HST – highstand systems tract; WRS – wave ravinement surface; TRS – tidal ravinement surface; SU – subaerial unconformity; MFS – maximum flooding surface.

be represented by beach sands in an open shoreline setting, or by delta front sands in a river-mouth setting (Fig. 4.48), and are usually overlain by alluvial deposits dominated by floodplain fines. Due to its formation during a stage of coastal aggradation, the within-trend normal regressive surface is not demarcated by any substrate-controlled ichnofacies (Fig. 4.9). Instead, this facies contact may be associated with intertidal softground ichnofacies such as *Psilonichnus* or *Skolithos* (Fig. 2.21). This surface has a strong physical expression (i.e., an abrupt facies shift from sand to overlying mud; Fig. 4.55), which makes it easy to identify in outcrop and subsurface, and has the potential to form over large distances, depending on the duration and rates of normal regression. In spite of its prominent physical characteristics and possible regional extent, the within-trend normal regressive surface has little value for chronostratigraphic correlations as it is highly diachronous, with the rate of shoreline normal regression (Fig. 4.9).

It is important to note that the mere contrast in lithologies (mud over sand) is not sufficient for the proper identification of this facies contact as a within-trend normal regressive surface, as other facies contacts, such as some flooding surfaces for example, may also exhibit a similar juxtaposition of facies. Therefore, in addition to the observation of lithologies, other key attributes of the underlying and overlying deposits need to be explored, including depositional trends, bathymetric contrasts, and the direction of syndepositional shoreline shift. For example, even though within-trend normal regressive surfaces and flooding surfaces

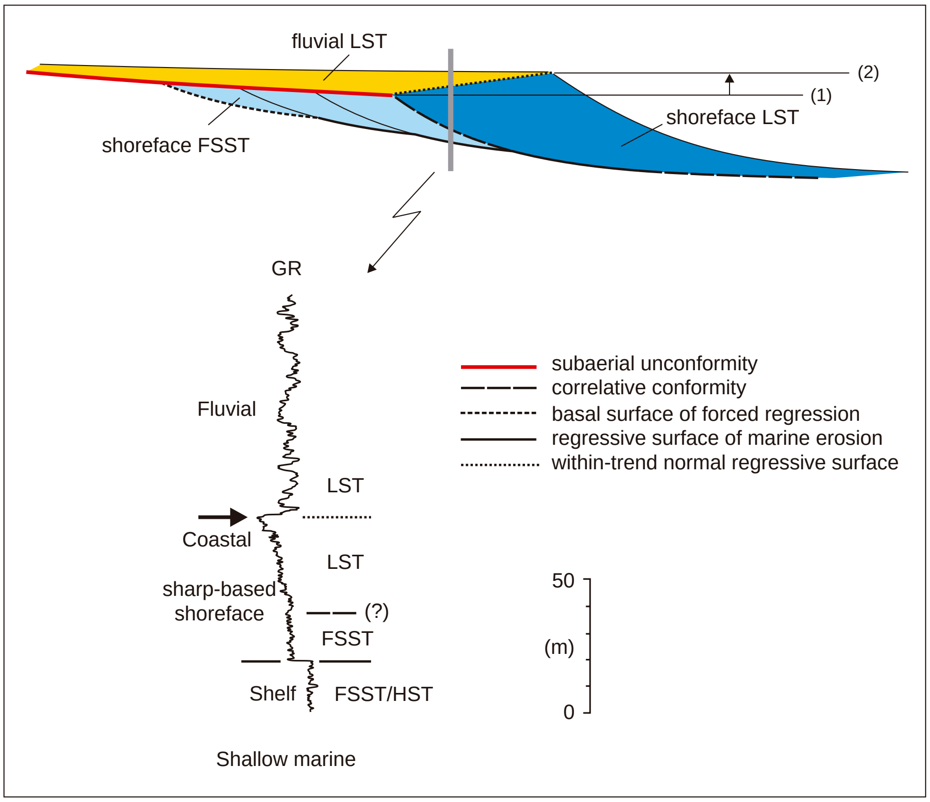


FIGURE 4.55 Well-log expression of the within-trend normal regressive surface (arrow). See Fig. 4.9 for a summary of diagnostic features of the within-trend normal regressive surface. Log example from the Lea Park Formation, Western Canada Sedimentary Basin. Note that such conformable facies contacts may occur within either lowstand or highstand systems tracts, where preserved from subsequent transgressive ravinement or subaerial erosion, respectively. Abbreviations: GR—gamma ray log; LST—lowstand systems tract; HST—highstand systems tract; FSST—falling-stage systems tract.

may display similar lithological signatures, the former are generated during *progradation* and water *shallowing* in the nearshore area, whereas the latter form during shoreline *transgression* and reflect water *deepening* in the coastal region. As explained in Chapter 3, and further detailed in Chapter 7, the association between regression and water shallowing, as well as between transgression and water deepening, is safely valid only for the shallow-water environment in the vicinity of the shoreline.

Even where a seaward shift of facies across a sand-to-overlying mud contact is documented, ruling out the interpretation of the contact as a flooding surface, the identification of a within-trend normal regressive surface solely based on well logs may be difficult, due to the possible confusion with the subaerial unconformity (e.g., compare the well-log expression of the two surfaces in Figs. 4.13, 4.29, and 4.55). For unequivocal identification, additional evidence from core or nearby outcrops is required to document the nature (scoured *vs.* conformable) of the stratigraphic contact under investigation (Fig. 4.9). In contrast to the subaerial unconformity, which truncates the underlying deposits and is also associated with offlap and fluvial onlap, the within-trend normal regressive surface is part of a conformable succession where no stratal terminations are recorded in relation to the adjacent, older and younger strata (Fig. 4.9).

Within-trend normal regressive surfaces may form during both lowstand and highstand normal regressions. In the case of highstand normal regressions, the within-trend normal regressive surface may or may not connect with the landward termination of the transgressive wave-ravinement surface, depending on the type of coastal setting (Fig. 4.48). In the case of lowstand normal regressions, the within-trend normal regressive surface connects with the basinward termination of the subaerial unconformity (Fig. 4.55). Field examples of within-trend normal regressive surfaces are provided in Figs. 3.36 and 4.56. The preservation potential of within-trend normal regressive surfaces may be hampered by subsequent transgressive ravinement erosion, in the case of lowstand systems tracts, or by subaerial erosion in the case of highstand systems tracts. Even where preserved from such larger-scale erosional processes, the within-trend normal regressive surface may be scoured locally by distributary channels in coastal plain or delta plain environments (Fig. 4.56B).

Besides within-trend normal regressive surfaces, as defined above in coastal settings, other, but less prominent facies contacts may be identified as well within normal regressive systems tracts. Notably, within the shallow-water environment, the facies contact between prodelta (deltaic bottomset) and the overlying delta

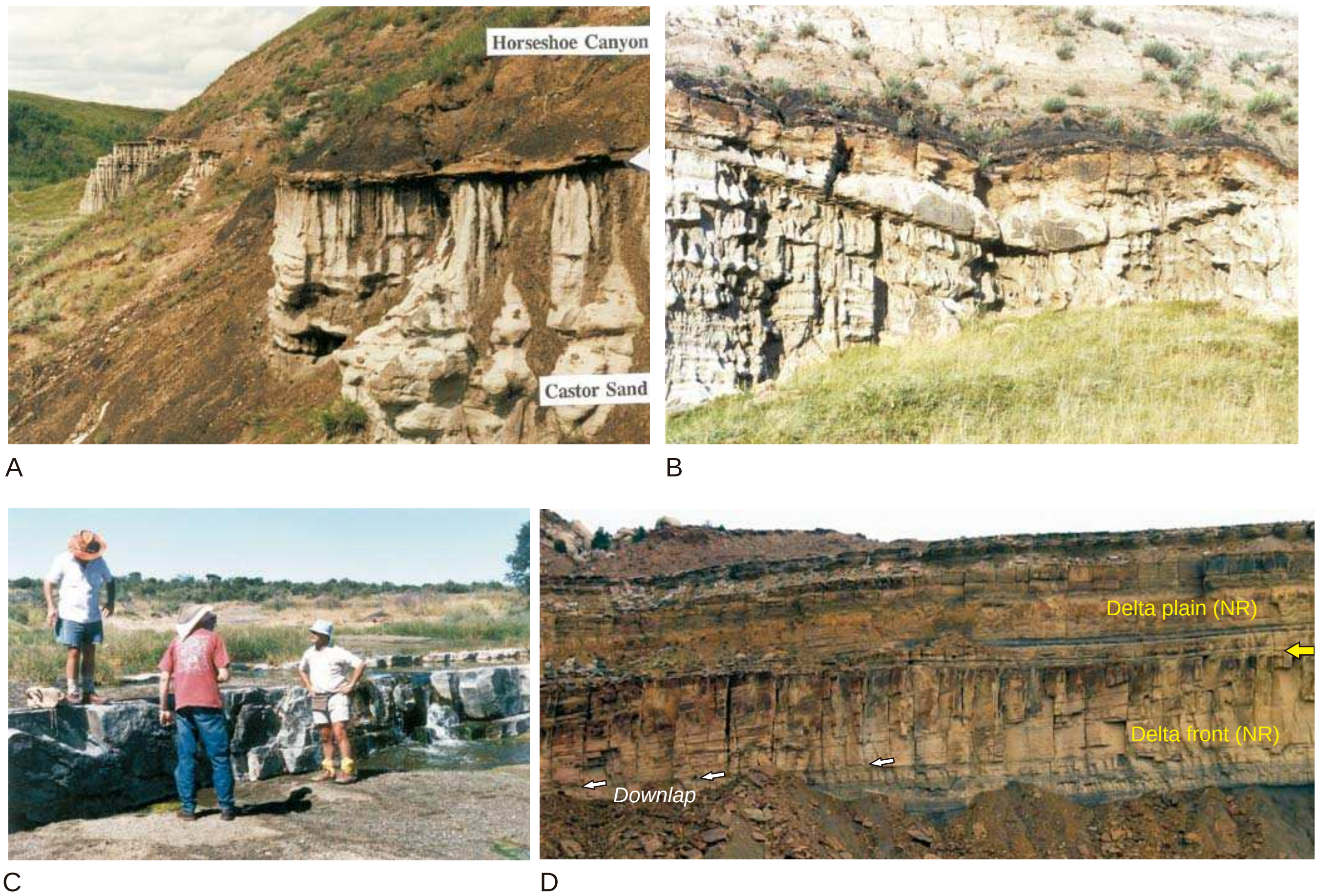


FIGURE 4.56 Outcrop examples of within-trend normal regressive surfaces. A—within-trend normal regressive surface separating beach sands from the overlying coal-bearing fluvial strata. This facies contact is conformable, mappable over a relatively large area, but is highly diachronous, with the rate of shoreline regression. Contact between the uppermost regressive shoreline sands of the Bearpaw Formation and the overlying Horseshoe Canyon Formation (Early Maastrichtian, Castor area, Western Canada Sedimentary Basin); B—distributary channel that scours locally the within-trend normal regressive surface in photograph A; C—within trend normal regressive surface (top of prograding strandplain) exposed by the erosion of the overlying fluvial floodplain deposits (contact between the Ecca and Beaufort groups, Late Permian, Karoo Basin); D—within-trend normal regressive surface (larger arrow) at the conformable facies contact between delta front (deltaic foreset) and the overlying coal-bearing delta plain deposits (deltaic topset). The photograph shows the river-dominated, normal regressive Ferron delta prograding from right to left (Late Cretaceous, Utah). Abbreviation: NR—normal regressive.

front (deltaic foreset) in river-mouth settings, or between shelf facies and the overlying prograding shoreface in open shoreline settings, may be identified as a mappable surface (sharp contact) in some cases, although in general the transition between these depositional environments tends to be gradational (Fig. 3.36). A possible reason for this gradual transition, as opposed to an abrupt and mappable facies contact, is that normal regressions are generally *slow*, hence there is sufficient time for wave-driven sediment mixing between the subtidal and the deeper-water environments. This makes it difficult, in most cases, to pinpoint a single surface as the base of delta front or subtidal facies in a normal regressive systems tract. This situation is often in contrast to what is expected in the case of forced regressions, as explained in the following section of this chapter.

The within-trend normal regressive surface is a lithologic discontinuity that may be used in lithostratigraphic and allostratigraphic analyses, but it is not part of a systems tract boundary or of a sequence boundary. For this reason, the within-trend normal regressive surface is not a proper sequence stratigraphic surface (Fig. 4.8). It may, however, be used to fill in the internal facies details of sequences and systems tracts once the main sequence stratigraphic framework is outlined by mapping and correlating the sequence stratigraphic surfaces.

Within-trend Forced Regressive Surface

The within-trend forced regressive surface is a conformable facies contact that develops *during* forced regressions at the base of prograding delta front facies of *river-dominated deltas* (Figs. 4.9 and 4.57). This type of within-trend facies contact does not develop in wave-dominated settings, either river-mouth or open shorelines, because in such settings the regressive surface of marine erosion forms instead (Fig. 4.23). It is also noteworthy that the within-trend normal regressive surface does not have an equivalent in a forced regressive coastal setting, where delta plain and fluvial

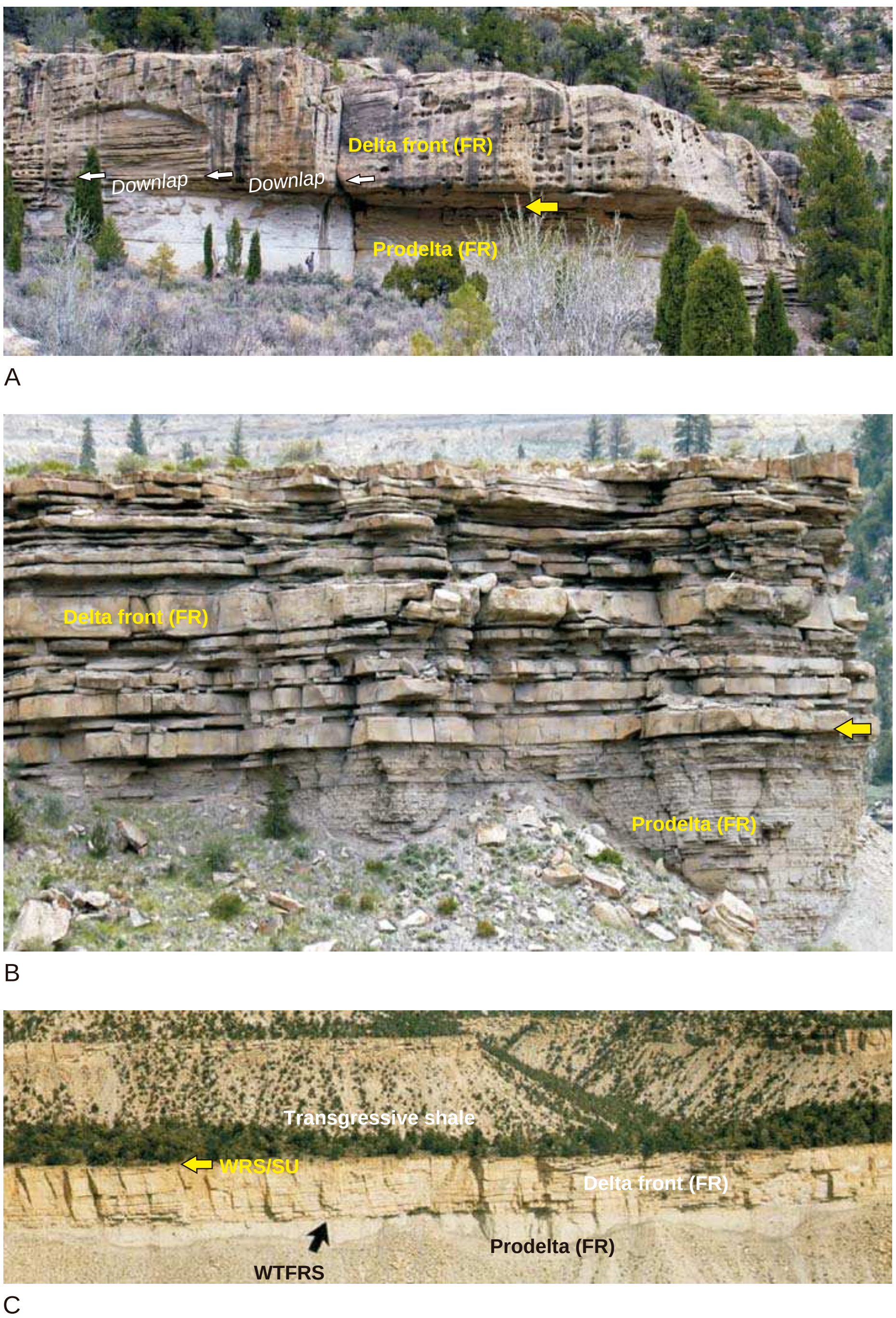


FIGURE 4.57 Outcrop examples of within-trend forced regressive surfaces at the conformable facies contact between prodelta (deltaic bottomset) and the overlying coarser-grained delta front deposits (deltaic foreset) (river-dominated, forced regressive Panther Tongue delta, Late Cretaceous, Utah). For scale, note person in image A. The delta front succession may reach up to 20 m in thickness. The within-trend forced regressive surface forms only in *river-dominated* deltaic settings, and it is highly diachronous, younging basinward with the rate of forced regression. A—steep delta front clinoforms (approximately 27°) associated with grain flow deposits (sand avalanches in a Gilbert-type delta); B—finer-grained delta front deposits (relative to A) associated with lower-angle clinoforms (approximately 10°) and the manifestation of turbidity flows; C—panoramic view showing that the forced regressive deltaic succession is truncated at the top by a composite unconformity that represents a transgressive wave-ravinement surface reworking a subaerial unconformity. Abbreviations: FR—forced regressive; WTFRS—within-trend forced regressive surface; WRS—transgressive wave-ravinement surface; SU—subaerial unconformity.

deposits are missing, being replaced in the rock record by the subaerial unconformity (Figs. 4.20—wave-dominated setting, and 4.26—river-dominated setting). As with any within-trend facies contact, the within-trend forced regressive surface is characterized by high diachroneity, becoming younger in a basinward direction with the rate of shoreline's forced regression (Fig. 4.9).

The conformable facies contact between prodelta and overlying delta front facies of forced regressive river-dominated deltas tends to be sharper than the corresponding facies contact in normal regressive settings, because forced regressions are relatively *fast*, and hence there is less time for mixing between delta front and prodelta sediments. As a result, the within-trend forced regressive surface tends to be prominent (sharp lithological contact), and therefore relatively easy to map in outcrop and subsurface (Figs. 4.57 and 4.58). Although the within-trend forced regressive surface appears, from a distance, to be a unique facies contact between prodelta and delta front facies (Fig. 4.57), detailed analyses from a closer range reveal that the change from prodelta to the overlying delta front facies takes place within a relatively narrow zone of facies transition; as such, no single lithological contact can be picked unequivocally as the within-trend forced regressive surface, which, in reality, amalgamates a few meters thick transitional interval (e.g., about 4–5 m on the well log in Fig. 4.58). For this reason and in spite of the relatively sharp lithological contrast that defines the within-trend forced regressive surface at a larger scale (Fig. 4.57), the forced regressive delta front deposits in a river-dominated setting are still 'gradationally based', rather than 'sharp-based' (Fig. 3.27), because the succession is conformable (i.e., no regressive surface of marine erosion is present) and the change from prodelta to delta front facies is gradational even though the transition takes places rapidly, within a relatively narrow interval (compare the log in Fig. 4.58, which shows gradationally based delta front deposits in a conformable succession, with the logs in Fig. 4.29, which show a much sharper, and unconformable, facies contact at the base of the forced regressive shoreface or delta front deposits that prograde during forced regression in a wave-dominated setting). Also, in contrast to the sharp-based delta front or shoreface deposits that accumulate in wave-dominated settings, the gradationally based delta front succession that overlies the within-trend forced regressive surface is potentially thicker than the depth of the fairweather wave base (assuming preservation from subsequent subaerial and transgressive ravinement erosion), because the toe of the delta front clinoforms that prograde in a river-dominated

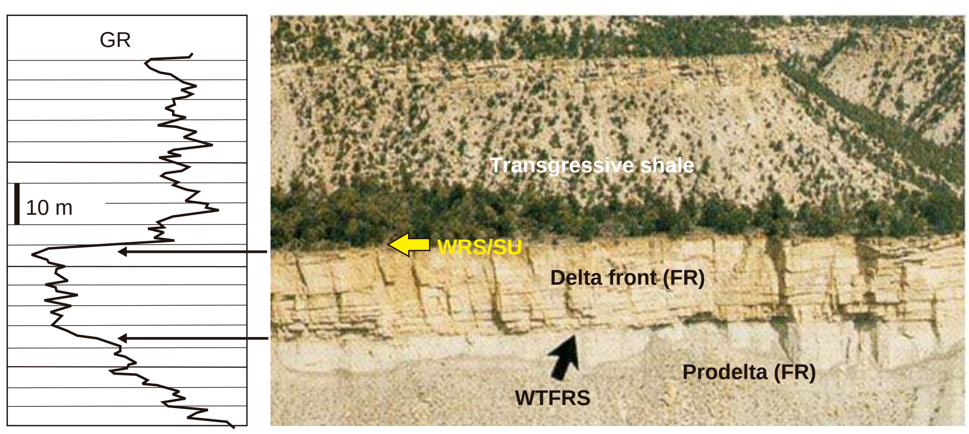


FIGURE 4.58 Well-log expression of a within-trend forced regressive surface (modified from images provided by H.W. Posamentier). The outcrop photograph shows the river-dominated, forced regressive Panther Tongue delta (image C in Fig. 4.57). Note that the deltaic succession, including the transition from prodelta to delta front facies, is conformable. The delta front interval (about 20 m in this example) is likely thicker than the depth of the fairweather wave base, because the toe of the delta front clinoforms that prograde in a river-dominated setting may reach depths greater than the fairweather wave base. Note that, from a distance, the within-trend forced regressive surface looks like a unique and well-defined facies contact (see also additional outcrop examples in Fig. 4.57). From close range, however, no single surface can be picked unequivocally as a unique lithological contact between prodelta and delta front facies. In reality, the within-trend forced regressive surface corresponds to a narrow zone of facies transition that may reach a few meters in thickness. As such, the delta front facies of river-dominated forced-regressive deltas are gradationally based (see also Fig. 3.27, and compare this well log with the logs provided in Fig. 4.29). Abbreviations: GR—gamma ray log; FR—forced regressive; WTFRS—within-trend forced regressive surface; WRS—transgressive wave-ravinement surface; SU—subaerial unconformity.

setting may reach depths greater than the fairweather wave base (Figs. 3.27 and 4.58).

The within-trend forced regressive surface may be used as a proxy for the basal surface of forced regression (seafloor at the onset of base-level fall—the conformable portion of Posamentier and Allen's, 1999, sequence boundary), even though the latter is known to be placed below, within the underlying finer-grained facies (Posamentier and Allen, 1999). This approximation is permitted by (1) the high rates of forced regression, coupled with (2) the low rates of sedimentation on the continental shelf in front of the prograding delta front. These two conditions imply that the within-trend forced regressive surface (above) and the basal surface of forced regression (below) are relatively close spatially (with and without a physical expression, respectively), although, due to the time required by the shoreline to regress, the two surfaces diverge in a basinward direction.

Within-trend Flooding Surface

The flooding surface is defined as 'a surface separating younger from older strata across which there is evidence of an abrupt increase in water depth. This deepening is commonly accompanied by minor submarine erosion or nondeposition' (Van Wagoner, 1995). Even though widely used in sequence stratigraphic work, the term 'flooding surface' is one of the most controversial concepts in sequence stratigraphy, as it allows for multiple meanings. The ambiguous nature of the above definition was discussed by Posamentier and Allen (1999) who emphasized that it is not clear whether the flooding surface forms merely as a result of increasing water depth in a marine (or lacustrine) environment, or actual flooding of a previously emergent landscape. What is clear is that flooding surfaces, commonly marked by abrupt facies shifts from sand to overlying mud in shallow-water settings, form invariably during shoreline transgression, and are topped by marine (or lacustrine) strata. The nature of the underlying deposits is however contentious, as they can vary from fluvial to coastal and shallow-water (Fig. 4.9).

At a semantic level, the usage of the word 'flooding' as a generic term that fits all the above scenarios of facies juxtaposition was challenged by Posamentier and Allen (1999) who proposed that 'flooding' should be restricted to situations where water overflows onto land that is normally dry. This definition is consistent with the common meaning of the word 'flooding', and implies subaerial exposure of the section below, prior to inundation. Following this rationale, and in order to avoid semantic confusions, Posamentier and Allen (1999) suggest replacing the term 'flooding surface' as defined by Van Wagoner (1995) with the more generic term 'drowning surface' to indicate a stratigraphic contact across which an abrupt water deepening is recorded. In this terminology, flooding surfaces become a special case of drowning surfaces, where shallow-water facies overlie nonmarine deposits. A practical problem with this approach is that evidence for subaerial exposure prior to the marine (or lacustrine) flooding is required in order to identify a stratigraphic contact as a 'flooding surface' *sensu* Posamentier and Allen (1999). Such evidence, however, may or may not be preserved in the rock record, depending on the intensity of transgressive ravinement erosion which may remove paleosols, root traces, or any other proof of subaerial exposure prior to flooding. On practical grounds, therefore, the more generic 'drowning surface' (or flooding surface *sensu* Van Wagoner, 1995) is easier to work with in terms of designating facies contacts generated by shoreline transgression, irrespective of the nature of the underlying deposits. In spite of the terminological arguments discussed by Posamentier and Allen (1999), the generic term of 'flooding surface' as defined by Van Wagoner (1995) is still the one that is most commonly used in current sequence stratigraphic work. Part of the reason is that the 'flooding surface' is heavily entrenched in the literature, despite the possible misleading connotation associated with the meaning of the word 'flooding'. In addition to this, the term 'drowning' was already coined as part of the 'drowning unconformity' concept, which is widely used in the context of carbonate sequence stratigraphy (Schlager, 1989).

Flooding surfaces are best observed in coastal to nearshore shallow-marine settings, where evidence of water deepening based on facies relationships is unequivocal (Fig. 4.59). Typical flooding surfaces may cap regressive successions (i.e., deltaic lobes in river-mouth settings or beach/shoreface deposits in open shoreline settings; Figs. 4.35B and 4.59), or transgressive sands (Figs. 4.35E and 4.60). In the former case, the transgressive deposits are typically absent or very thin, and the flooding surface may represent the only evidence of transgression in addition to the occasional transgressive lags (Kamola and Van Wagoner, 1995). Flooding surfaces have correlative surfaces in the coastal plain and shelf environments (Kamola and Van Wagoner, 1995), and possibly beyond, into the alluvial plain and deep-water settings, respectively. However, the identification of such correlative surfaces in nonmarine or deep-water deposits, unless based on uniquely correlatable strata such as volcanic ash beds, serves little purpose and may only be a source of confusion (Posamentier and Allen, 1999).

The definition provided by Van Wagoner (1995) is general enough to allow different types of stratigraphic

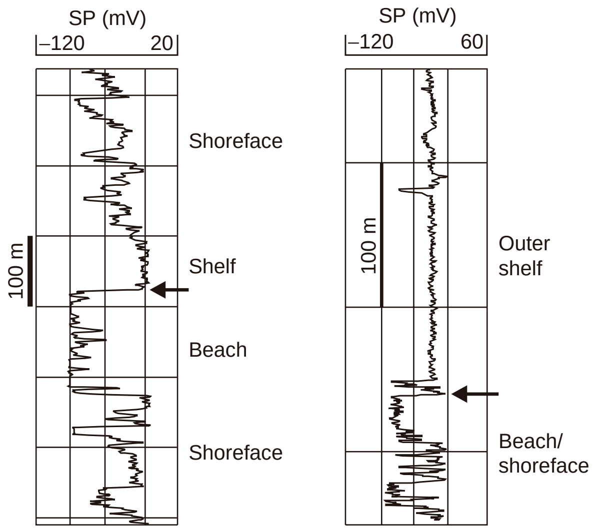


FIGURE 4.59 Flooding surfaces (arrows) at the contact between normal regressive shoreface and beach sands, and the overlying shelf mudstones. In these examples, the flooding surfaces are most likely represented by transgressive wave-ravinement surfaces. Above the flooding surfaces, the transgressive deposits may be very thin.

contacts to be candidates for flooding surfaces. The *transgressive ravinement surface* is often considered a 'flooding surface' (Posamentier and Allen, 1999: 'an overflowing of water onto land that is normally dry') (Fig. 3.30), but other surfaces that form in fully marine successions satisfy the definition of a flooding surface as well: the *maximum regressive surface*, where there is an abrupt cut-off of sediment supply at the onset of transgression (cases A, B, C, and D in Fig. 4.35); the *maximum flooding surface*, where the transgressive strata are missing and the maximum flooding surface reworks the maximum regressive surface (Fig. 4.44); or a *within-trend facies contact*, where the sand/shale contact occurs within the transgressive succession (Figs. 4.35E and 4.61). As the transgressive ravinement, maximum regressive, and maximum flooding surfaces are already defined in an unequivocal manner, the within-trend type of flooding surface is the only new surface left to be considered (Figs. 4.35E, 4.60, and 4.61). This within-trend facies contact, separating transgressive sands from the overlying transgressive

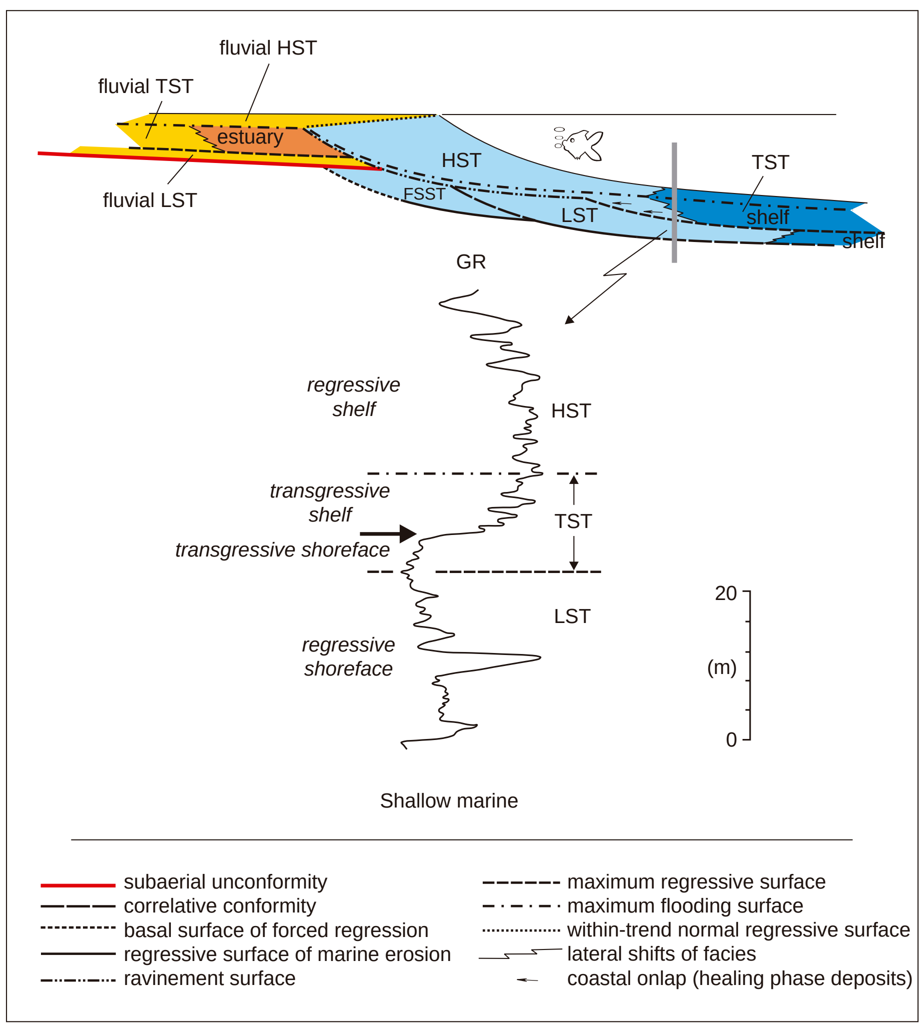


FIGURE 4.60 Well-log expression of the within-trend flooding surface (arrow; modified from Catuneanu, 2003). See Fig. 4.9 for a summary of diagnostic features of the within-trend flooding surface. Log example from Embry and Catuneanu (2001). Abbreviations: GR—gamma ray; LST—lowstand systems tract; TST—transgressive systems tract; HST—highstand systems tract; FSST—falling-stage systems tract.

FIGURE 4.61 Within-trend flooding surface (arrow in image A) at the contact between transgressive shoreface deposits (Bad Heart Formation, Coniacian) and the overlying transgressive outer shelf shales (Puskwaskau Formation, Santonian) (photographs courtesy of Andrew Mumpy). This flooding surface corresponds to an episode of abrupt water deepening within the marine basin, which led to sediment starvation and the development of a firmground on the seafloor. The substrate immediately underlying the flooding surface is burrowed, and indurated by subaqueous seafloor cementation. The stage of nondeposition required by the formation of this firmground ('omission' surface) provided a proper environment for the formation of substrate-controlled ichnofacies. No lag deposits, or other evidence of scouring, are associated with this flooding surface. Images A—D show the indurated nature of the firmground (approximately the top 20 cm of the sediment underlying the flooding surface); images E and F show the fabric of the substrate-controlled ichnofacies.

shales, is not in a position to serve as a systems tract or sequence boundary, which is why it is not a surface of sequence stratigraphy. Similar to the within-trend normal regressive and forced regressive surfaces, the within-trend flooding surface may, however, be used to resolve the internal facies architecture of a systems tract (transgressive systems tract in this case) once the sequence stratigraphic framework is established.

As the flooding surface may change its meaning depending on case study, from a within-trend facies contact to an actual sequence stratigraphic surface, its defining features, associated stratal terminations and temporal attributes may vary significantly (Fig. 4.9). For this reason, the diagnostic features listed in Fig. 4.9 for the flooding surface cover a spectrum wide enough to allow for all possible scenarios. For example, where the transgressive facies are missing, the flooding surface may 'borrow' the characteristics of a maximum flooding surface, truncating the strata below, being downlapped by the strata above, and separating two normal regressive successions (Figs. 4.9 and 4.44). Similarly, a transgressive wave-ravinement surface may also qualify as a flooding surface (Fig. 4.57C), displaying, in this case, a high diachroneity, variable underlying facies, and onlapping shallow-marine deposits on top (Fig. 4.9). When possessing the significance of a maximum regressive or maximum flooding surface, the flooding surface itself may onlap and downlap the pre-existing landscape and seascape in a landward and seaward direction, respectively (Fig. 4.9). In a most general sense, therefore, the flooding surface may, in terms of field attributes, fit the profile of several different types of stratigraphic contacts depending on circumstances. The common thread, however, is the fact that *flooding surfaces are always overlain by marine/lacustrine shales*, either transgressive (e.g., Figs. 4.35, 4.57C, 4.60, and 4.61) or regressive (e.g., Fig. 4.44A), accumulated in a deeper-water environment relative to the underlying facies. Some flooding surfaces may be *conformable*, where sedimentation is continuous during their formation. This is likely the case where flooding surfaces are represented by maximum regressive surfaces (cases A, B, C, and D in Fig. 4.35), or by non-omission (i.e., with no substrate-controlled ichnofacies associated with them) within-trend facies contacts (e.g., the conformable surface indicated by the grey arrow in Fig. 4.35E). Often, however, flooding surfaces are represented by 'omission' contacts, associated with a stratigraphic hiatus caused by a lack of sediment supply, sediment bypass or erosion, and as a result they are potentially demarcated by substrate-controlled ichnofacies (Figs. 4.9 and 4.61). The actual type of substrate that marks a flooding surface may vary with the location within the basin, with firmgrounds and hardgrounds forming in fully marine environments (e.g., Fig. 4.61), and all types of substrate-controlled ichnofacies (firmgrounds, hardgrounds, and woodgrounds) possibly occurring where the underlying facies are coastal or nonmarine. Such *unconformable* flooding surfaces are typically represented by maximum flooding surfaces and transgressive ravinement surfaces (e.g., Figs. 4.44 and 4.57C), but also by within-trend facies contacts that are associated with significant stages of water deepening and sediment starvation of the seafloor during transgression (e.g., Fig. 4.61).

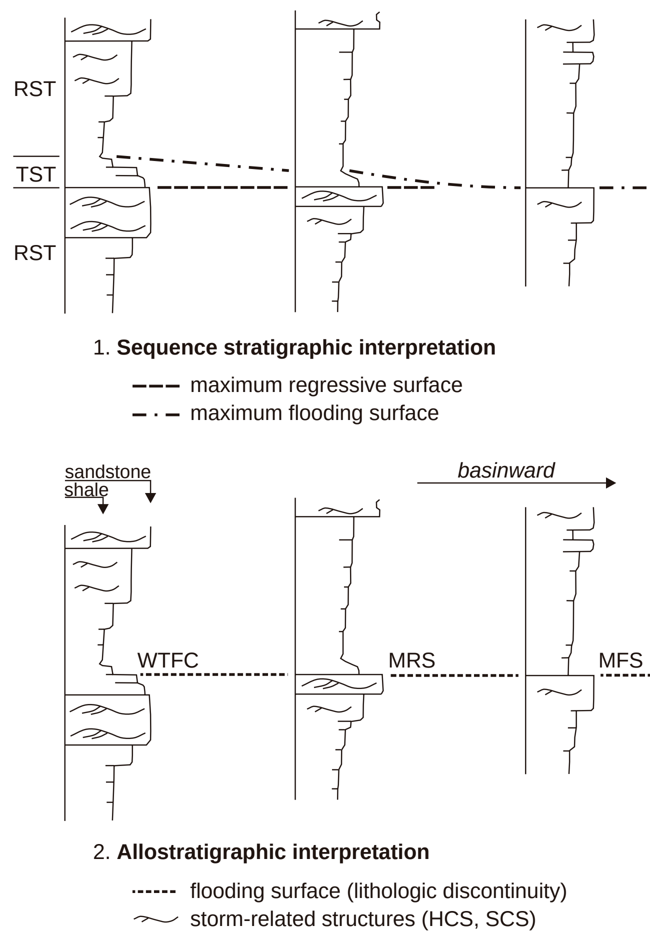


FIGURE 4.62 Shallow-marine (shoreface to shelf) succession of sands and shales interpreted in sequence stratigraphic and allostratigraphic terms. The thickness shown is about 12 m. Note that the transgressive facies thin basinward, to the point where the maximum flooding surface reworks the maximum regressive surface. The flooding surface is placed at the strongest lithological contrast. The example is from the Cardium Formation, Western Canada Sedimentary Basin. Abbreviations: WTFC—within-trend facies contact; MRS—maximum regressive surface; MFS—maximum flooding surface; TST—transgressive systems tract; RST—regressive systems tract; HCS—hummocky cross-stratification; SCS—swaley cross-stratification.

The unconformable flooding surfaces may or may not be associated with erosion of the seafloor. Where scoured, flooding surfaces are commonly overlain by a thin veneer of lag deposits, including coarse sand, granules or rip-up clasts, indicating that variable amounts of erosion have taken place in the process of their formation (Pemberton *et al*., 2001). The amount of erosion varies with the type of flooding surface, being higher in the case of transgressive ravinement surfaces and maximum flooding surfaces, and minimal (if any) in the case of maximum regressive surfaces and within-trend flooding surfaces. As a rule of thumb, the higher the amount of erosion, the greater the chance for the formation of well developed transgressive lags and substrate-controlled ichnofacies, although the latter may also form in relation to stages of sediment starvation, in the absence of any discernable scouring (Fig. 4.61). Irrespective of the stratigraphic significance of the flooding surface, the shift to deeper-water facies across the contact usually triggers an increase in faunal abundance and ichnodiversity following the flooding event (Pemberton *et al*., 2001), as well as a sharp increase in the bioturbation index (Siggerud and Steel, 1999). This change in ichnofabric across the flooding surface is accompanied by an increase in water load, which may contribute to further compaction that will enhance the firmness of the substrate, and hence generate substrate-controlled ichnofacies (Snedden, 1991).

Due to its generic nature, the flooding surface is thus too general, or vague, as a concept to pinpoint the exact type of stratigraphic contact under analysis. The usage of more specific terms, or surface types, is therefore preferred whenever sufficient data are available for the unequivocal identification of the actual type of stratigraphic contact. In a generic sense, as a lithological contact with or without sequence stratigraphic significance, the flooding surface is more appropriate for allostratigraphic studies. For sequence stratigraphic work, however, the vague nature of flooding surfaces hampers the communication of precise genetic meanings, and hence the usage of sequence stratigraphic surfaces, whenever possible, is recommended. Figure 4.62 illustrates the conceptual difference between the approaches used for sequence stratigraphic *vs*. allostratigraphic correlations. The main lithological discontinuity (the sand/shale contact, i.e., the flooding surface) is the surface of choice for allostratigraphic correlations. This surface not only transgresses time, but also changes in significance along dip, from a within-trend facies contact, to a maximum regressive surface, and finally to a maximum flooding surface (Fig. 4.62). This allostratigraphic approach is descriptive, as opposed to the sequence stratigraphic interpretation that provides a genetic framework for the rock record under analysis.