

DELTAS

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ABSTRACT: Deltas are discrete shoreline protuberances formed where a river enters a standing body of water and supplies sediments more rapidly than they can be redistributed by basinal processes, such as tides and waves. In that sense, all deltas are river-dominated and deltas are fundamentally regressive in nature. The morphology and facies architecture of a delta is controlled by the proportion of wave, tide and river processes; the salinity contrast between inflowing water and the standing body of water, the sediment discharge and sediment caliber, and the water depth into which the river flows. The geometry of the receiving basin (and proximity to a shelf edge) may also have an influence. The simple classification into river, wave and tide-dominated end-members must be used with caution because the number of parameters that control deltas is more numerous.

Other depositional environments, such as wave-formed shorefaces or barrier-lagoons can form significant components of larger wave-influenced deltas, but conversely smaller bayhead or lagoonal deltas can form within larger barrier-island or estuarine systems. As deltas are abandoned and transgressed they may also be transformed into another depositional systems (e.g. transgressive barrier lagoon system or estuary). Delta plains also contain distributary river channels and their associated floodplains and bays, which can be equally classified as both fluvial and deltaic environments.

Tens to hundred meter thick, sharp-based blocky sandstones within many ancient mid-continent deltas have been routinely interpreted in the rock record as distributary channels, although many of these examples are now reinterpreted as incised fluvial valleys. Distributary channels may show several orders of sizes and shapes as they bifurcate downstream around distributary mouth bars. Bifurcation is inhibited in strongly wave-influenced deltas, resulting in relatively few terminal distributary channels and mouth bars flanked by extensive wave-formed sandy barriers or strandplain deposits. In shallow-water fluvial dominated deltas, tens to hundreds of shallow, narrow and ephemeral terminal distributary channels can form intimately associated with mouth bars that form larger depositional lobes. Tides appear to stabilize distributary channels for hundred to thousands of years, inhibiting avulsion and delta switching.

As deltas prograde they form upward coarsening facies successions, as sandy mouth bars and delta front sediments build over muddy deeper water prodelta facies. Deltas display a distinct down-dip clinoform cross sectional architecture. Many large muddy deltas show separate clinoforms, the first at the active sandy delta front and the second on the muddy shelf. Along-strike facies relationships may be less predictable and depositional surfaces may dip in different directions. Overlapping delta lobes typically result in lens-shaped stratigraphic units that exhibit a mounded appearance.

All modern deltas grade up-dip from marine into non-marine environments, and Walther's Law predicts that deltas should show a marine to nonmarine transition as they prograde. However, in many low accommodation settings, topset alluvial or delta plain facies can be removed or reworked by wave or tidal erosion during transgression, resulting in top-eroded deltas. Historically, some of these top-eroded deltas have been interpreted as distal shelf deposits, not related to shoreline processes. Sequence stratigraphic concepts, however, allow facies observations to be placed within a larger context of controlling allocyclic mechanisms which allow the correct interpretation of larger delta systems of which only small remnants may be preserved.

WHAT ARE DELTAS, AND WHY ARE THEY IMPORTANT?

Much of the sediment transferred from land to sea is carried by rivers and deposited at the shoreline in the form of deltas. About 25% of the world's population lives on deltaic coastlines and wetlands (Syvitski et al., in press). Prediction of growth and decay of modern deltas is critical in areas such as Louisiana and in much of Asia where rampant dam building has caused an immense decrease in discharge of freshwater to the world's oceans, resulting in enormous stresses to these coastal ecosystems as they experience subsidence and land loss (Vörösmarty et al., 1997). Sixty percent of the world's rivers are affected by reservoirs (Syvitski et al., in press). The USA National Inventory of Dams shows 38,100 dams over 6 m high blocking the flow in the Mississippi River drainage basin alone. The muddy deltaic coastline immediately south of Bangkok, Thailand, for example, is retreating at 12 m per year as a consequence of the extreme demands made by a huge population on the precious freshwater resource supplied by the Chao Phraya River (Vongvisessomjai, 1990). Similar problems exist along the entire Indo-Gangetic delta

plain. Better prediction of deltas, the critical boundary between land and sea, is needed.

From the economic perspective, deltas have been estimated to host close to 30% of all of the world's oil, coal, and gas deposits (Tyler and Finley, 1991). Much of these resources are in areas that have been productive for many years, such as the oil and gas deposits of the Cretaceous Interior Seaway and Gulf of Mexico and the Carboniferous coal deposits of the UK and the Eastern USA. However, as production declines and global energy needs continue to grow, new and better facies models will be required to improve the extraction of oil and gas. Significant fresh-water resources also occur in delta deposits, and exploitation of these aquifers requires robust facies models for deltas.

This paper reviews deltaic facies models from the sedimentological to the regional stratigraphic perspective. It begins by presenting a brief historical overview of delta studies and then proceeds with defining what a delta is, describing the various components of deltaic depositional systems, in terms of typical sub-environments and facies successions, and then finished with the larger perspective provided by facies architectural and sequence stratigraphic studies.

Historical Background

The concept of a delta dates back to Herodotus (c. 400 BC), who recognized that the alluvial plain at the mouth of the Nile had the form of the capital Greek letter Δ (Fig. 1). The first study of ancient deltas was that of Gilbert (1885), who described Pleistocene fresh-water gravelly deltas in Lake Bonneville, Utah. Gilbert recognized a basic threefold subdivision of delta deposits into topset, foreset, and bottomset units (Fig. 2), a terminology that remains in use to this day. Barrell (1912) extended these subdivisions to the much larger scale of the Devonian Catskill wedge in the Appalachians, and provided the first explicit definition of the essential features of a delta as:

“... a deposit partly subaerial built by a river into or against a body of permanent water. The outer and lower parts are necessarily constructed below water level, but its upper and inner surface must be land maintained or reclaimed by the river building from the sea. A delta, therefore, consists of a combination of terrestrial and marine, or at least lacustrine strata, and differs from other modes of sedimentation in this respect” (Barrell, 1912, p. 381).

Barrell considered the recognition of associated nonmarine facies crucial in distinguishing ancient deltas from estuaries. This criterion is no longer required because in deltas deposited during times of falling sea level (e.g., lowstand, forced regressive, or falling-stage systems tracts), subaerial topset facies may either not be deposited or may be eroded during subsequent transgression, yielding “top-truncated” delta deposits (Plint, 1988; Posamentier et al., 1992; Hart and Long, 1996; Bhattacharya and Willis, 2001; Martinsen, 2003).

Our understanding of modern deltas developed rapidly during the last fifty years, beginning with work on the Mississippi Delta published in the 1950s and early 1960s (e.g., Shepard et al., 1960). Scruton (1960) recognized that deltas are essentially cyclic in nature and consist of a progradational, “constructive phase”

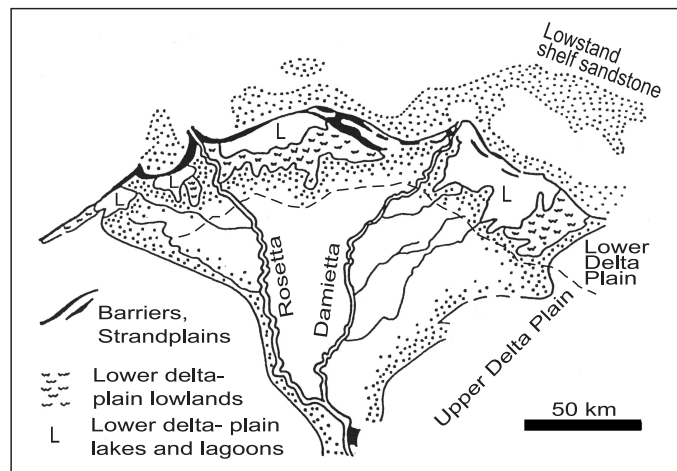


FIG. 1.—Environments and facies in the modern Nile delta. Only the Rosetta and Damietta Branches are presently active. Stipple indicates older reworked delta sands (Scheiing and Gaynor, 1991) rather than active “sand plumes” (Coleman et al., 1981). Since construction of the Aswan Dam, water and sediment discharge to the delta have decreased, and the entire delta is undergoing transgression. From Bhattacharya and Walker, (1992) based on Fisher et al. (1969) and Sestini (1989).

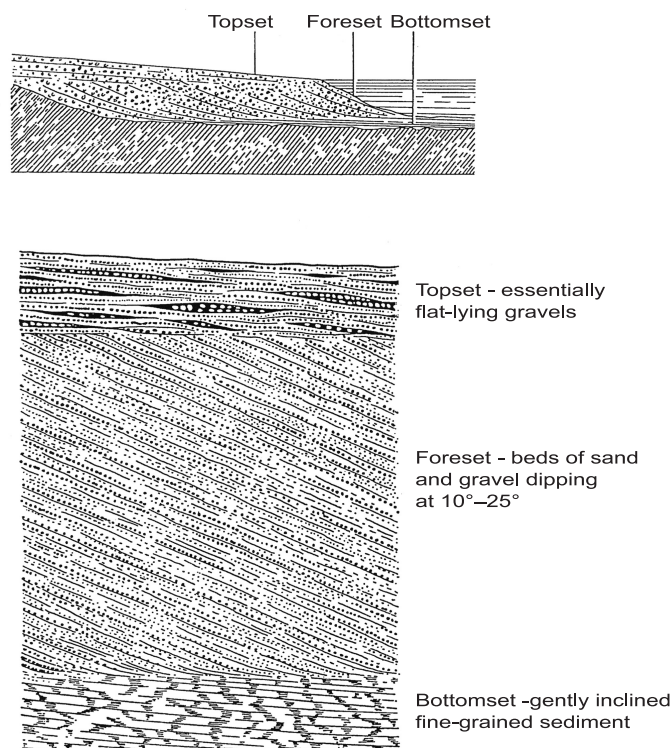


FIG. 2.—Cross-sectional facies architecture and vertical facies succession of a delta showing threefold subdivision into topset, foreset, and bottomset strata. From Elliott (1986) after Gilbert (1885) and Barrell (1912).

usually followed by a thinner retrogradational “destructive phase” coinciding with delta abandonment. He also illustrated a vertical “deltaic sequence” (Scruton, 1960) of coarsening- and sandier-upward facies related to progradation of bottomset, foreset, and topset strata (Fig. 3).

The Gulf Coast region of the U.S.A. (Florida to Texas) historically has been an important focus for research on modern and ancient deltas, primarily because of the economic importance of deltas as oil and gas reservoirs. Coleman and Wright (1975) compiled a global data base of 34 modern deltas and developed a six-fold classification based on sand distribution patterns (Fig. 4) with accompanying “typical” vertical facies profiles. One of the most widely used classification schemes is that of Galloway (1975), who subdivided deltas according to the dominant processes controlling their morphology: rivers, waves, and tides (Fig. 5). These two studies emphasized the importance of the overall shape of a sediment body in defining the type of delta, although this has recently come under some criticism (Dominguez, 1996; Rodriguez et al., 2000; Fielding et al., 2005a). A spate of research, focused on coarser-grained, high-latitude delta systems, led to an appreciation of the importance of grain size, water depth, and feeder type as controlling variables on delta type (Colella and Prior, 1990; Postma, 1990; Orton and Reading, 1993).

Improvements in seismic and side-scan sonar imaging led to the recognition of regional-scale synsedimentary deformation in the subaqueous parts of modern deltas (Coleman et al., 1983; Winker and Edwards, 1983). Similar features have now been recognized in outcrops of several ancient deltas (Nemec et al., 1988; Martinsen, 1989; Pulham, 1989; Bhattacharya and Davies, 2001; Bhattacharya and Davies, 2004; Wignall and Best, 2004).

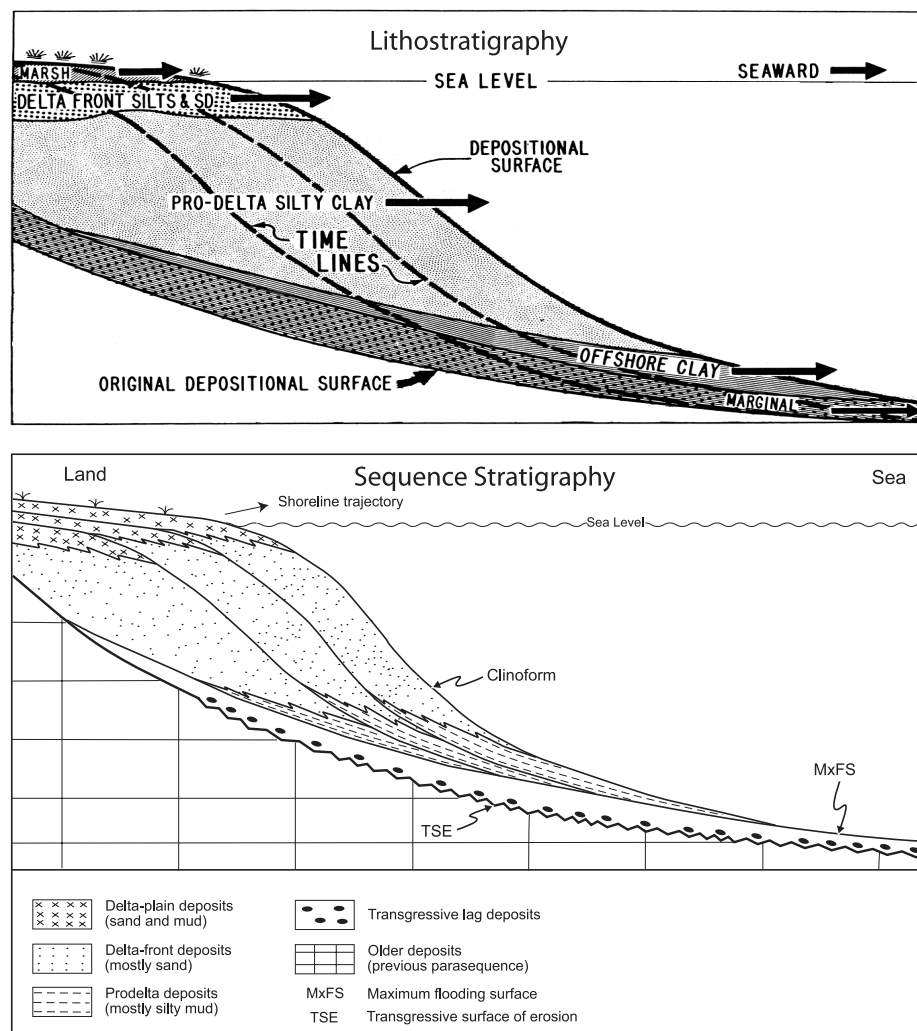


FIG. 3.—Early example of a delta clinoform, showing topset, foreset, and bottomset strata (Scruton, 1960). **A)** Lithostratigraphic representation shows facies boundaries as undulating but apparently sharp. Arrows indicate direction of progradation. Most modern delta studies still show facies contacts in this manner. **B)** Correct representation of facies boundaries versus timelines. Bed boundaries are more likely to follow the time lines (From Gani and Bhattacharya, 2005).

These are also critical for creation of traps in many shelf and offshore deltas such as the Gulf of Mexico and Nigeria (e.g., Evamy et al., 1978; Berg and Avery, 1995).

A large body of research, begun with advent of seismic and sequence stratigraphy, emphasizes the evolution of modern and Quaternary deltas in the context of relative sea-level changes (Dominguez et al., 1987; Boyd et al., 1989; Williams and Roberts, 1989; Carbonel and Moyes, 1987; Hart and Long, 1996; Hori et al., 2002; Ta et al., 2002; Sydow and Roberts, 1994; various papers in Sidi et al., 2003; various papers in Anderson and Fillon, 2004, and the application of these concepts to ancient deltas (Galloway, 1989a, 1989b; Bhattacharya and Walker, 1991; Martinsen, 1993; Tesson et al., 1993; Bhattacharya, 1994; Gardner, 1995; Garrison and van den Bergh, 1997; Plint, 2000; Bhattacharya and Willis, 2001; Garrison and van den Bergh, 2004).

Several recent studies have documented examples of tide-influenced deltas, which until recently have been the least well documented in the ancient record (e.g., Maguregui and Tyler,

1991; Jennette and Jones, 1995; Mellere and Steel, 1995, 1996; Dalrymple, 1999; Willis et al., 1999; Willis and Gabel, 2001; Bhattacharya and Willis, 2001; Ta et al., 2002; Ta et al., 2005; Hori et al., 2002; Davies et al., 2003; Allison et al., 2003; Dalrymple et al., 2003; Lambiase et al., 2003; White et al., 2004; Willis, 2005). There has also been an increasing focus on studies of muddy prodeltaic shelves linked to Modern highstand deltas, such as the Po, Ganges–Brahmaputra, Amazon, and Orinoco (Nittrouer et al., 1986; Kuehl et al., 1997; Correggiari et al., 2001; Liu et al., 2002; Warne et al., 2002; Cattaneo et al., 2003; Correggiari et al., 2005; Neill and Allison, 2005).

Definitions

Deltas have been defined as “discrete shoreline protuberances formed where rivers enter oceans, semi-enclosed seas, lakes or lagoons and supply sediments more rapidly than they can be redistributed by basinal processes” (Elliott, 1986, p. 113).

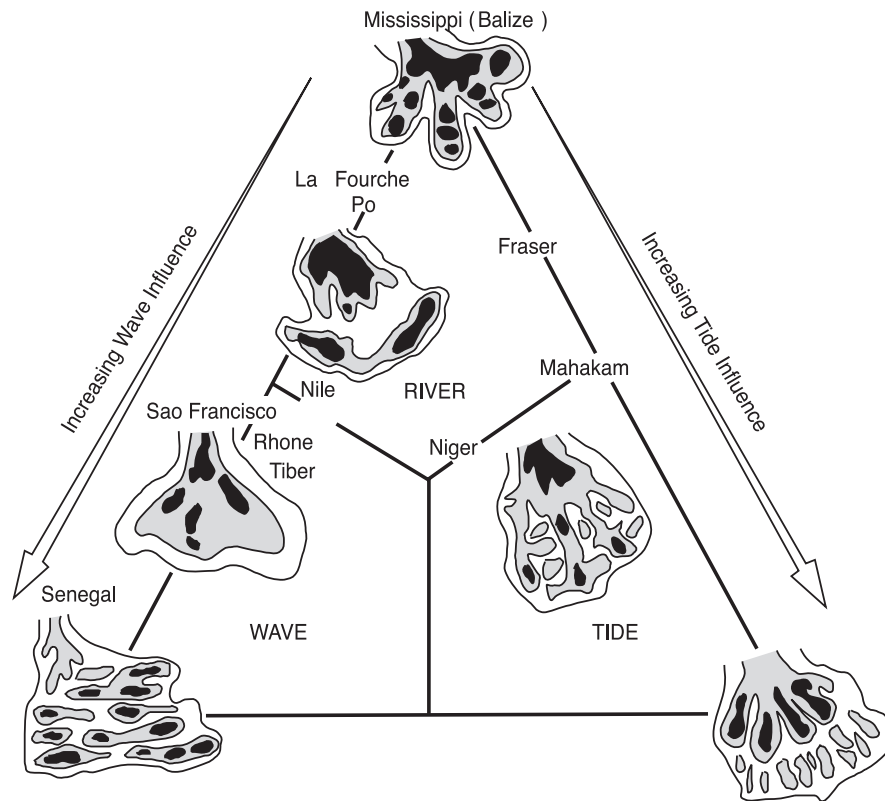


FIG. 4.—Sandbody geometries of the six delta types of Coleman and Wright (1975) plotted on the river-, wave-, and tide-dominated tripartite classification of Galloway (1975), from Bhattacharya and Walker (1992). Note that all sand bodies narrow and thicken towards a point (fluvial) source. Also note similarity of tide-dominated isolith to river-dominated end member.

By this definition, all deltas are to some degree river-influenced. Deltas are therefore fundamentally regressive in nature (Dalrymple, 1999).

The term delta has also been applied to many ancient facies successions or clastic wedges that show a marine to nonmarine transition, or which contain a marine–fluvial or lacustrine–fluvial interface (Alexander, 1989), following the early definition of Barrell (1912). Although a shoreline must be crossed in such a transition or interface, the identification of the shoreline as specifically deltaic usually requires good three-dimensional control of facies patterns. This may consist of maps of lithofacies distributions showing a thickening and narrowing of the clastic succession toward the point of fluvial input, and the required seaward protuberance of the shoreline (Fig. 4).

Distinguishing Deltas from Other Depositional Systems

Much of the sediment in a delta is derived directly from the river that feeds it, in contrast with estuaries in which sediment is derived both from the marine and fluvial realm (Dalrymple, 1999). Estuaries have also been defined as fundamentally transgressive depositional systems, in contrast to deltas which are regressive (Dalrymple et al., 1992).

In barrier-island systems, sediment is supplied alongshore (Reinson, 1992). The terms ebb-tidal delta and flood-tidal delta have also been applied to sediment accumulations that form around tidal inlet channels in barrier-lagoon depositional systems (Reinson, 1992). Barrier islands may form components of

larger wave-influenced delta systems (Bhattacharya and Giosan, 2003). In particular, a river can act as a groyne, or hydraulic barrier, that traps sediment carried in the longshore drift system (Fig. 6; e.g., Dominguez, 1996; Rodriguez et al., 2000; Bhattacharya and Giosan, 2003). Barrier-islands can also form during the transgression of a delta, such as the Chandeleur islands in the Gulf of Mexico, which are the remnants of a now abandoned delta lobe of the Mississippi delta (Fig. 7; Boyd and Penland, 1988). Where basinal processes redistribute sediment to the point that the fluvial source and delta morphology can no longer be recognized, more general environmental terms such as paralic, strandplain, or coastal plain may be more preferable (Alexander, 1989).

Deltas occur at a wide variety of scales ranging from continental-scale depositional systems, such as the modern Mississippi delta (Fig. 8) with an area of about 28,500 km², to components of other depositional systems such as bayhead deltas within estuarine or lagoonal systems. Many continental-scale deltas, such as the Danube in Romania (Fig. 9) and the Mississippi, may contain smaller-scale crevasse deltas within larger-scale lobes, resulting in a complex and hierarchical facies architecture.

RIVER-MOUTH PROCESSES

A delta forms when a river of sediment-laden freshwater enters a standing body of water, loses its competence to carry sediment, and deposits it. The theory of jets has been widely applied to explain the dynamics of how river plumes interact

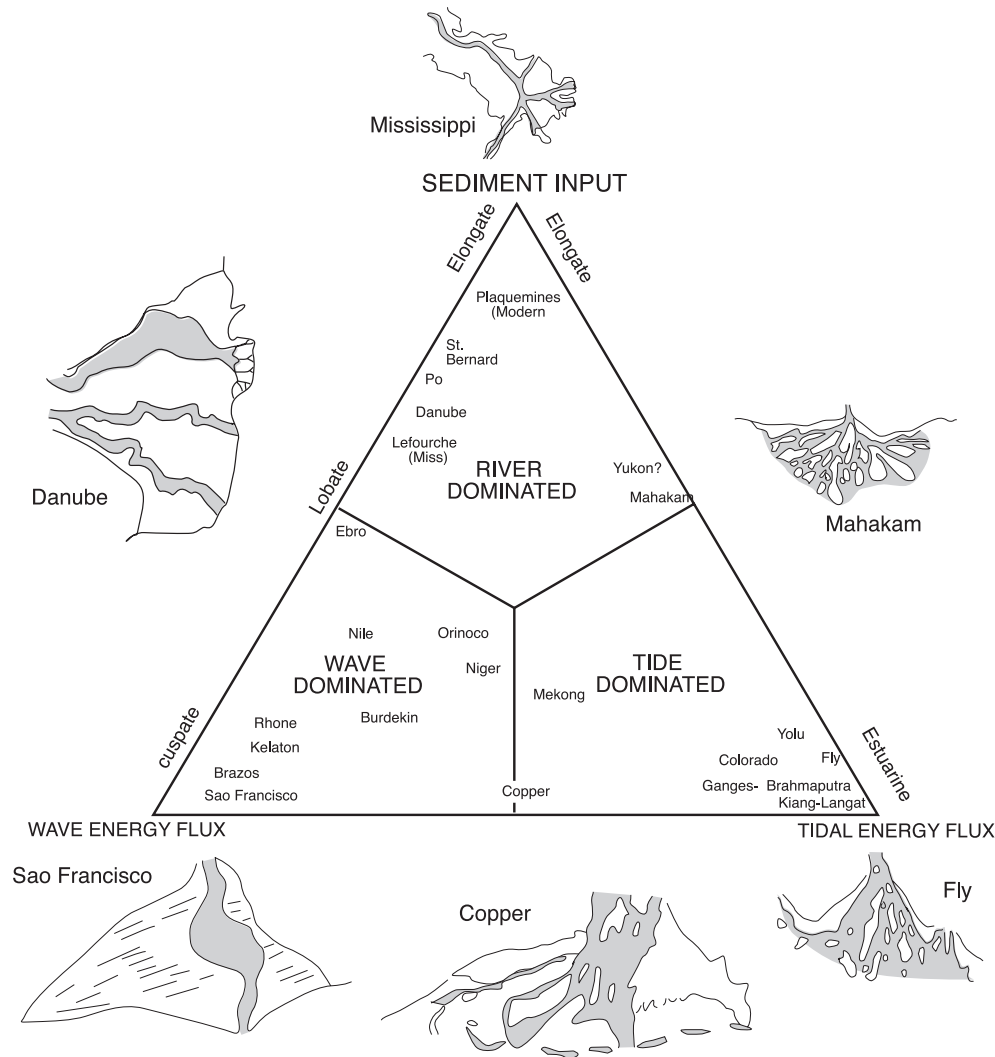


FIG. 5.—Tripartite classification of deltas, into river-, wave-, and tide- dominated end members (Galloway, 1975). Tide-dominated end-members are noted as being “estuarine”. This prompted Walker (1992) to abandon the concept of a tide-influenced delta. Also note that the São Francisco and Brazos deltas are considered as type examples of wave-dominated end-members.

with the body of water that they flow into (e.g., Bates, 1953; Wright, 1977; Orton and Reading, 1993; Nemec, 1995). The internal facies distribution and external morphology of a deltaic deposit depends upon (1) whether the river outflow is more dense (hyperpycnal), equally dense (homopycnal), or less dense (hypopycnal) than the standing body of water, (2) the interaction of the river plume with marine processes, which can include waves, tides, storms, and ocean currents, and biogenic reworking (Fig. 10), (3) the physical position of the delta in the basin, such as the shelf edge, and (4) the degree to which river-derived sediments are reworked by marine processes.

Historically, most marine deltas have been assumed to be hypopycnal, but many rivers experience dramatic changes in discharge as a function of seasonal climate change or as a result of major floods associated with storms. As a consequence, many rivers can alternate from hypopycnal to hyperpycnal conditions, even in fully marine settings (Nemec, 1995; Mulder and Syvitski, 1995; Parsons et al., 2001). Many river plumes may show both

hyperpycnal and hypopycnal plumes at the same time (Nemec, 1995; Kineke et al., 2000). Homopycnal conditions are the least common, because only small density differences are required for a flow to become either hypopycnal or hyperpycnal.

Much of the active sand deposition occurs in a **distributary-mouth bar** (also referred to as a stream-mouth bar or a middle-ground bar). Mouth bars are a fundamental architectural element in modern deltas that can coalesce to form complex bar assemblages that in turn build regional-scale depositional lobes. Mouth bars scale broadly to the width of flow, although flow widths in distributary channels can vary both spatially and temporally. Individual bars can be on the order of several kilometers long in relatively large rivers like the modern Atchafalaya delta (Fig. 11; Van Heerden and Roberts, 1988; Tye, 2003). The size and shape of a mouth bar also depends on the angle of dispersion of a plume, the flow conditions (hyperpycnal, hypopycnal, or homopycnal), and the forces that act on the river plume (buoyant, inertial, and frictional forces and basal processes).

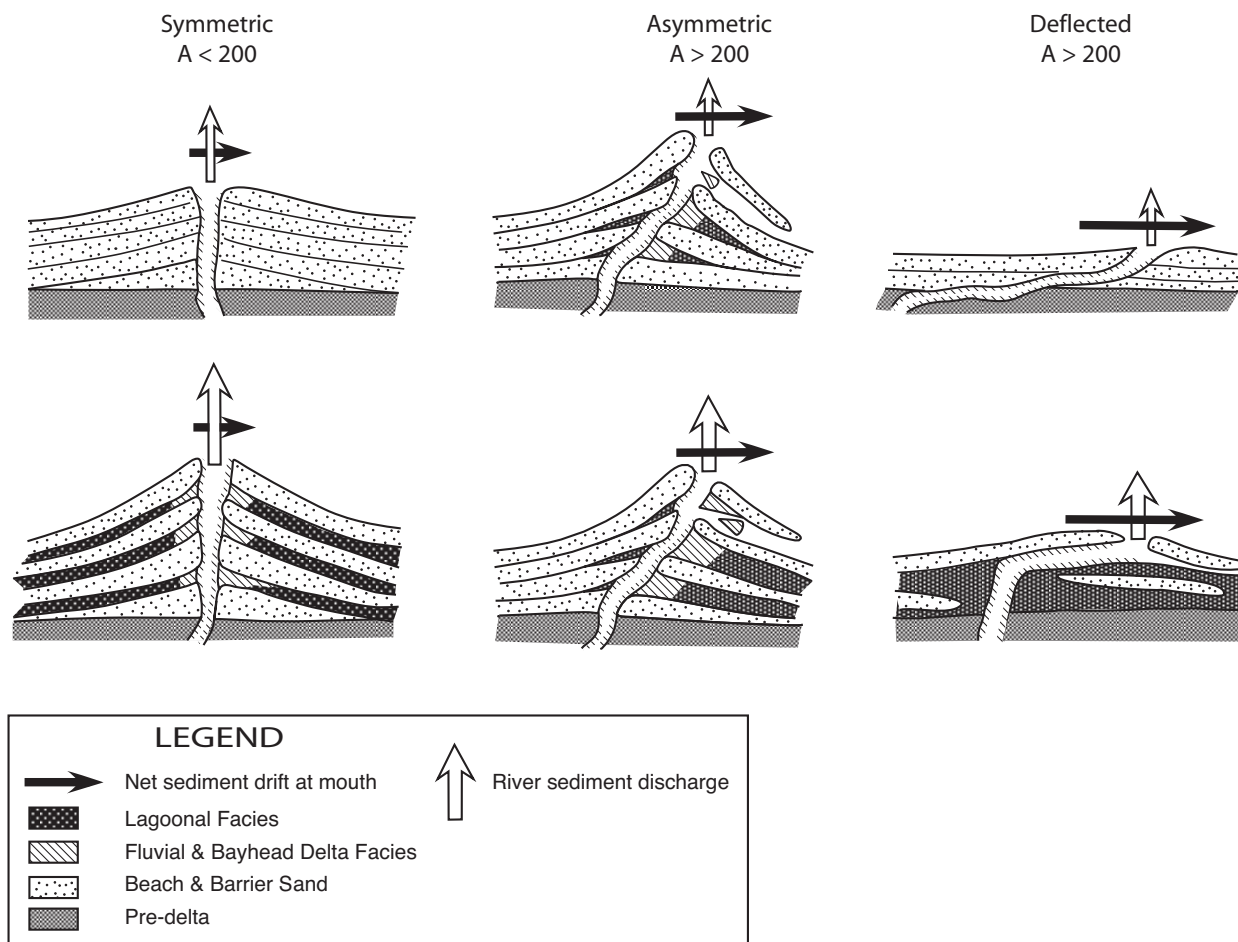


FIG. 6.—Morphology of wave-influenced deltas. Top row represents lower fluvial discharge compared to bottom row. River plume acts as a groyne that traps sediment updrift (after Bhattacharya and Giosan, 2003). Asymmetry index represents the ratio of fluvial sediment discharge to alongshore sediment transport rate.

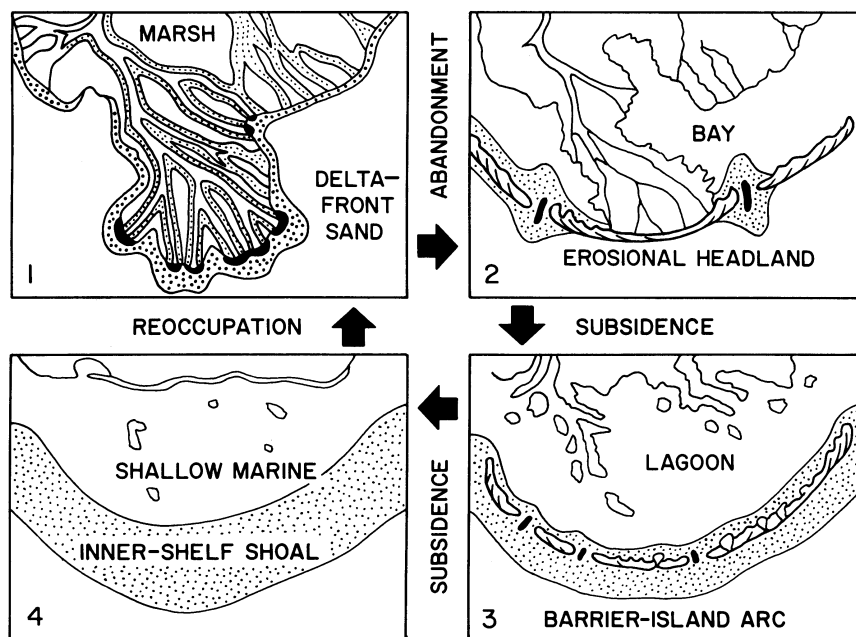


FIG. 7.—Evolution of Mississippi delta lobes from progradation to abandonment (from Boyd et al., 1989). Delta goes through an initial cycle of progradation, during which it shows a river-dominated character. As it is abandoned, it forms into a barrier-lagoon system. The barrier is ultimately drowned to form a relict shelf shoal.

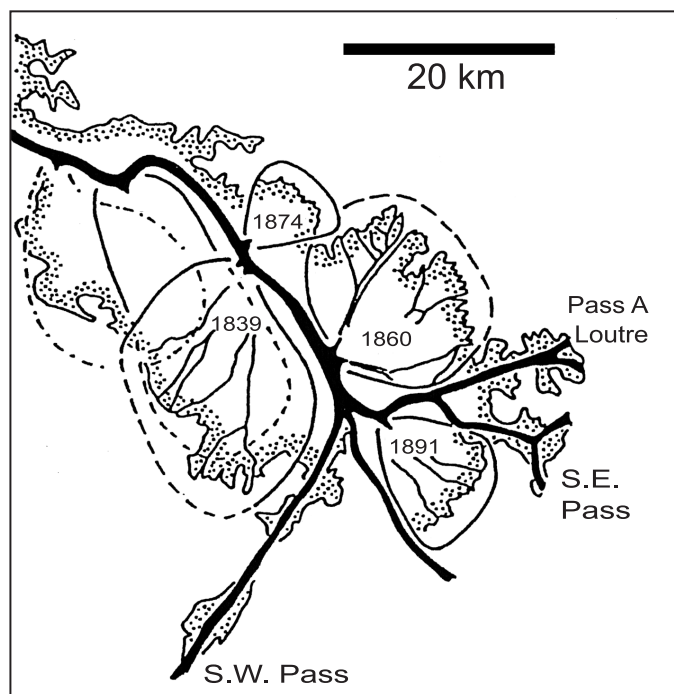


FIG. 8.—Infilling of interdistributary bays by historically dated crevasse “subdeltas” in the modern Mississippi birdfoot delta. Note the large variation in scale of deltas and distributary channels. At least three orders of branching can be discerned (from Bhattacharya and Walker, 1992; simplified from Coleman and Gagliano, 1964).

Tye (2004) compiled data on the dimensions of modern mouth bars and other sandy elements in the Atchafalaya delta in the Gulf of Mexico and the Colville, Kuparuk, and Sagavinortik rivers in Alaska. His data showed that bar widths ranged from 100 m to 3 km and bar lengths ranged from 140 m up to a maximum of nearly 7 km. Modal mouth-bar widths are between 120 to 410 m, and modal lengths are between 250 m to 610 m. A compilation by Reynolds (1999) of ancient mouth-bar sand bodies shows considerably larger dimensions. His study showed that mouth-bar sand bodies range from 1.1 km to 14 km wide with lengths of between 2.6 km to 9.6 km. Average sand-body widths are about 3 km, and average lengths are about 6 km. Reynolds suggests that mouth-bar sandstones are typically twice as long as they are wide. Clearly, the average values of ancient sand bodies (Reynolds, 1999) versus their modern geomorphic counterparts (Tye, 2003) illustrate that ancient examples represent the migration and growth of modern bars, and hence give larger dimensions.

Because bars are bedload features, they induce an enormous amount of form friction, in excess of that associated with grain and bedform roughness. This form friction significantly lowers bed shear stress and causes channel discharge to decrease as well as causing a change in the direction of flow around the bar (i.e., bifurcation of the channel). As bifurcation continues, the system may become unstable, initiating an autocyclic upstream avulsion of the feeding distributary.

Mouth bars can accrete downstream, laterally, and upstream (e.g., Van Heerden and Roberts, 1988; Corbeanu et al., 2004; Olariu et al., 2005; Olariu and Bhattacharya, 2006). Downstream accretion is an important process by which deltas grow and prograde. Upstream accretion of sand, caused by the form fric-

tion described earlier, may initiate growth faults (Bhattacharya and Davies, 2001, 2004) and may be important in causing avulsions as distributary channels become choked with sediment and switch course.

Waves smooth out and elongate mouth bars in a shore-parallel direction (Fig. 12; Wright, 1977; Fielding et al., 2005b). The ability of waves to extend a bar downdrift depends on the ratio of flood frequency to longshore-drift transport capacity. In deltas with high-wave-energy regimes or very infrequent floods (e.g., centennial floods), mouth bars may be extended for many kilometers or more alongshore. Tides commonly dissect the bar, or elongate it in a shore-normal direction. Tides may also cause distributary channels, and in turn the associated bars, to be stable for centuries, resulting in length-to-width ratios of up to 10 (Reynolds, 1999).

Hypopycnal (Buoyancy- and Friction-Dominated)

Where a river enters salt water, the density of the fresh river water plus suspended sediment load may be less than that of the sea water, causing hypopycnal flow (Fig. 10). Suspended muds are carried out into the receiving basin as a buoyant plume, resulting in lower depositional slopes.

Hypopycnal mud plumes may be deflected along the shelf by waves, ocean gyres and other oceanic circulation currents. In cases where this mud is trapped within the littoral zone, it may form a hyperconcentrated fluid mud that accretes to the shoreline. Winnowing of this mud may cause shells or sands to form thin beach deposits that armor the underlying mud and allow the downdrift muddy coastline to prograde, forming a **chenier plain** (Rine and Ginsburg, 1985; Augustinius, 1989; Penland and Suter, 1989; Draut et al., 2005). These are common on the downdrift margins of muddy delta systems, such as the Chenier plain of the Louisiana coast, which lies downdrift of the mighty Mississippi outflow (Penland and Suter, 1989; Draut et al., 2005). The Camau peninsula is a largely muddy accumulation that forms the downdrift wing of the Mekong delta (Ta et al., 2002). Nearly 50% of mud trapped in the modern Orinoco delta in Venezuela is actually derived from the Amazon (Rine and Ginsburg, 1985; Warne et al., 2002). The Amazon muds are thus carried along the shelf for a distance of over 1000 km. Mud from the modern Po delta has also been tracked for several hundred kilometers to the south along the Adriatic coast (Fig. 13; Cattaneo et al., 2003).

River bedload typically stops moving at the point of flow expansion, forming the mouth bar, whereas suspended load muds continue to be transported basinward. Hypopycnal deltas are thus characterized by a distinct separation of the friction-inducing sandy bedload from the buoyant suspended muddy load. Depending on plume stability, muddy plumes may produce subaqueous distributary channels with well developed levees that may cause the mouth bar and channel to form elongate bar-finger sands (Fisk, 1961).

At low stage, the more dense sea water can intrude many kilometers upstream into the river, forming a salt wedge, as is seen in many modern deltas, such as the Po in Italy (Nelson, 1970). Salt wedges can bring marine or brackish fauna into the distributary channels, as observed in delta-plain distributary channels in the Cretaceous Ferron Sandstone in Utah, U.S.A. (e.g., Corbeanu et al., 2004). During low discharge, muds in the overlying buoyant plume may flocculate and be deposited through suspension, settling as extensive bar drapes in the river and mouth-bar areas. These bar drapes can form fluid-mud layers that may be flushed out onto the shelf during subsequent floods. Settling of sediment within a hypopycnal plume may result in unstable fingers of sediment collapsing through the water column and becoming

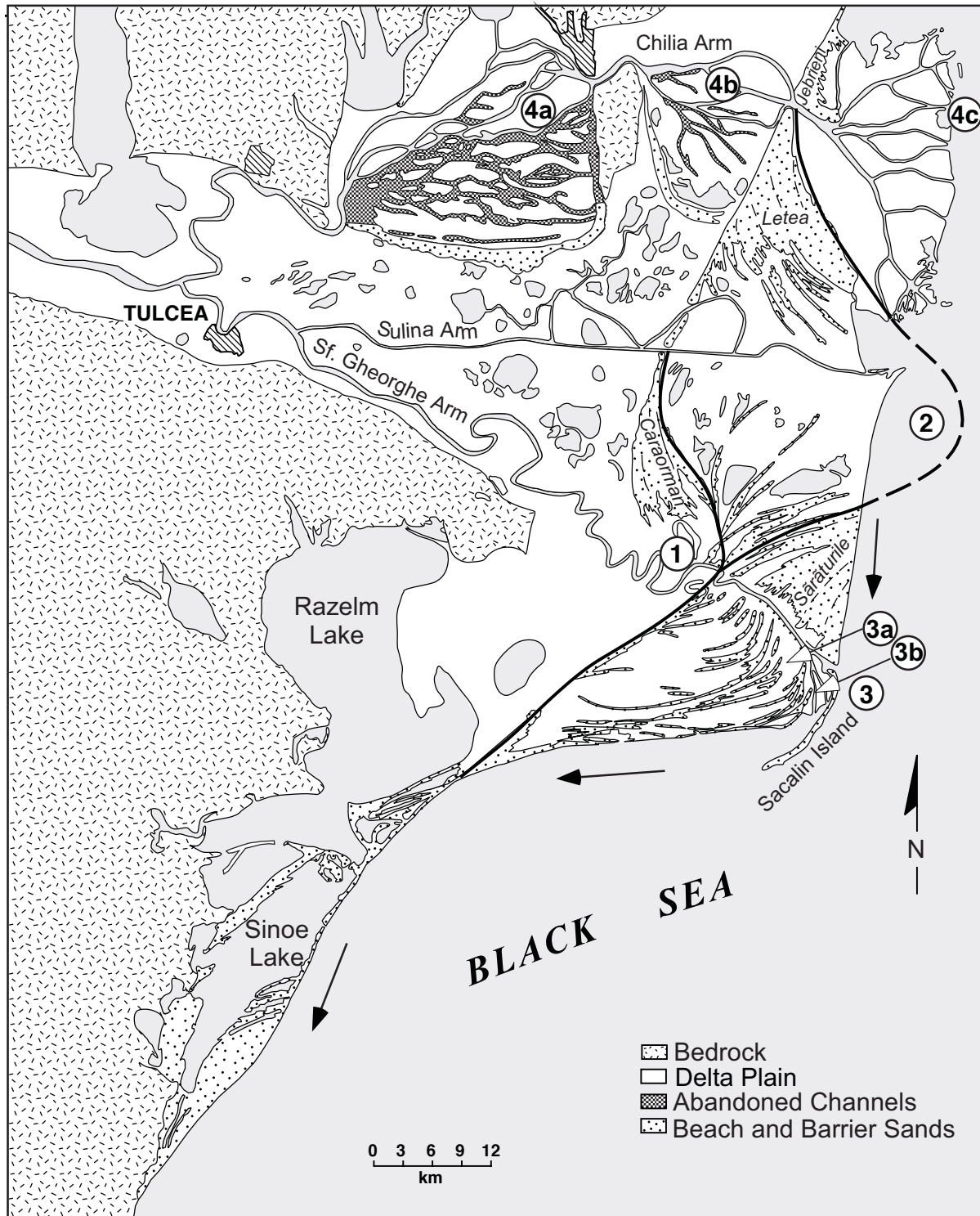


FIG. 9.—History of the Danube delta plain. The highest-discharge, northern branch feeds a highly river-dominated delta lobe 4c, comprising numerous bifurcating distributary channels with only minor wave reworking. The southernmost branch feeds the distinctly asymmetric wave-influenced lobe 3. The updrift side of lobe 3 comprises amalgamated beach ridges of the Saraturile Formation whereas the downdrift side comprises river-dominated bay-head deltas (3b) building behind a wave-formed barrier island. The asymmetry is preserved in the older lobes 1 and 2. The central lobe 2 is largely inactive and is presently being destroyed. Sands from lobe 2 are carried south by longshore drift to accumulate in the vicinity of lobe 3. Successive ages and outlines of lobes are: 1, 9000–7300 yr BP; 2, 7300–2500 yr BP; 3 and 4, 2900 yr BP–present (based on radiocarbon dates of Panin et al., 1983). Figure is based on map prepared by Gastescu (1992).

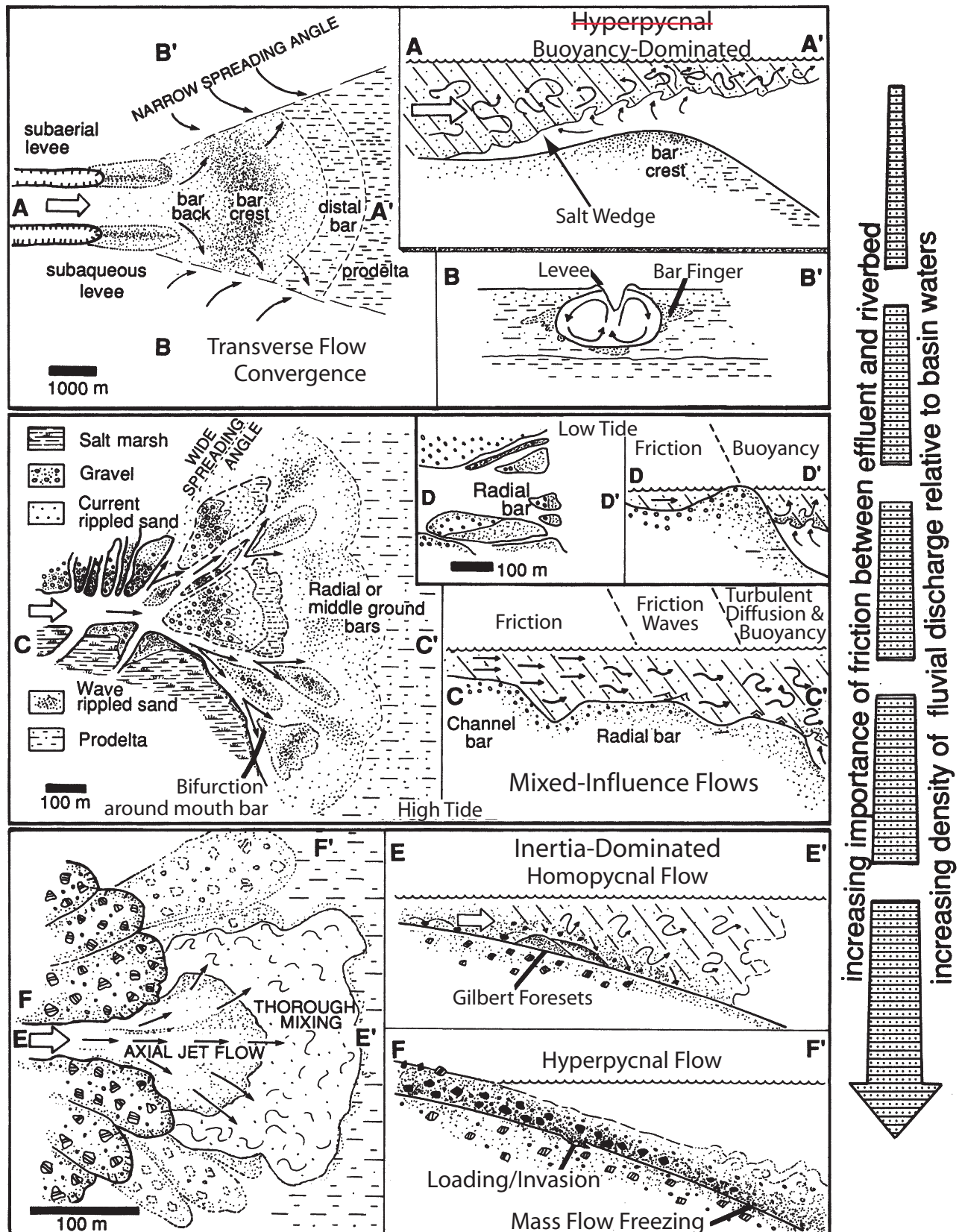


FIG. 10.—Examples of mouth-bar processes in river-dominated deltas (from Reading and Collinson, 1996, after Orton and Reading, 1993) incorporating ideas of Bates, 1953, Wright (1977) and others. See text for discussion.

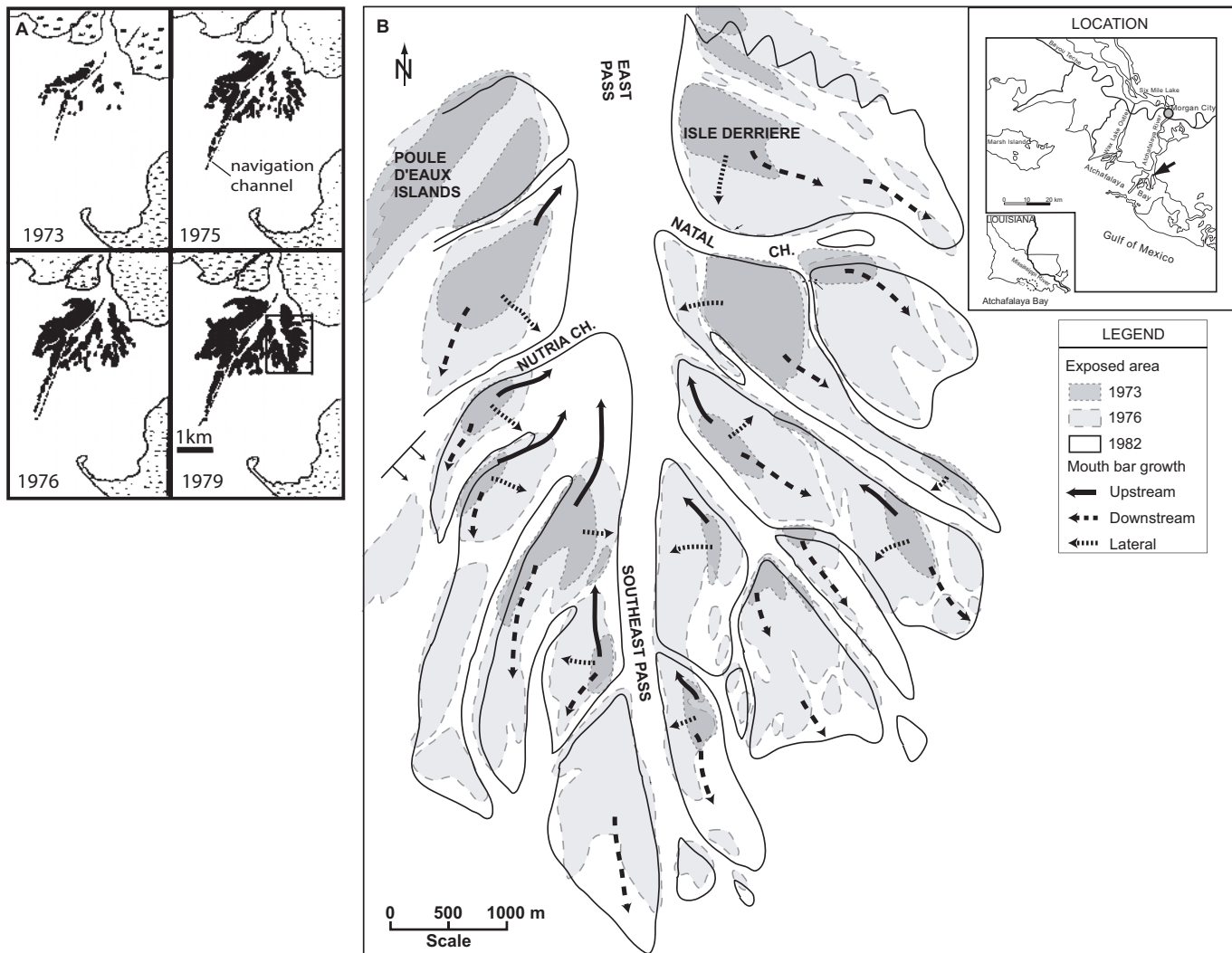


FIG. 11.—Development of a shallow water delta in Atchafalaya Bay, Mississippi Delta, U.S.A. **A)** River-dominated lobe forms by the coalescing of distributary-mouth bars (black), suggesting friction dominance. **B)** As the delta grows, the mouth bars accrete upstream and downstream (compare 1976 and 1982 shorelines). Note that there are numerous orders of distributary channels, culminating in small terminal distributary channels. Also note the scale of the mouth bars, which are on the order of several hundred meters wide and one to several kilometers in length (from Olariu and Bhattacharya, 2006; after Van Heerden and Roberts, 1988).

concentrated enough to produce a hypervycnal flow (Nemec, 1995; Parsons et al., 2001).

Hypervycnal (Inertia-Dominated)

In freshwater lakes, sediment concentrations less than 1 kg/m^3 produce hypervycnal conditions whereas sediment concentrations greater than the density caused by dissolved salt in seawater (about 35 to 45 kg/m^3) may be required to generate hypervycnal flows in marine settings (Mulder and Syvitski, 1995; Parsons et al., 2001). These flow conditions dominate where sediment-laden streams enter freshwater lakes, as occurs in many alpine or periglacial environments (Eyles and Eyles, 1992). Many marine settings, however, are also hypervycnal (e.g., Wright et al., 1988; Mulder and Syvitski, 1995; Plink-Björklund and Steel, 2004) and hypervycnal conditions have

been shown to occur in marine setting at sediment concentrations of $1\text{--}5 \text{ kg/m}^3$ (Parsons et al., 2001). Such low-concentration hypervycnal flows may occur where marine water is colder than fresh river outflow, or where the shallow marine setting is brackish, such as occurs at many delta fronts. Hypervycnal flows may cause sediment to bypass the shoreline or mouth bar and be deposited on the offshore shelf as density underflows. Because the momentum of the hypervycnal flow exceeds the ability of the standing body of water to stop the motion or lift the plume by buoyant forces, hypervycnal deltas have been referred to as “inertia-dominated” (Bates, 1953). Hypervycnal flows can be important in feeding deep-water systems, especially during times of low sea level or in areas with narrow shelves where the river may be delivering sediment directly into deep water. The resulting deposit of a hypervycnal flow is either a fluid mud, or a silty or sandy graded bed (i.e., turbidite;

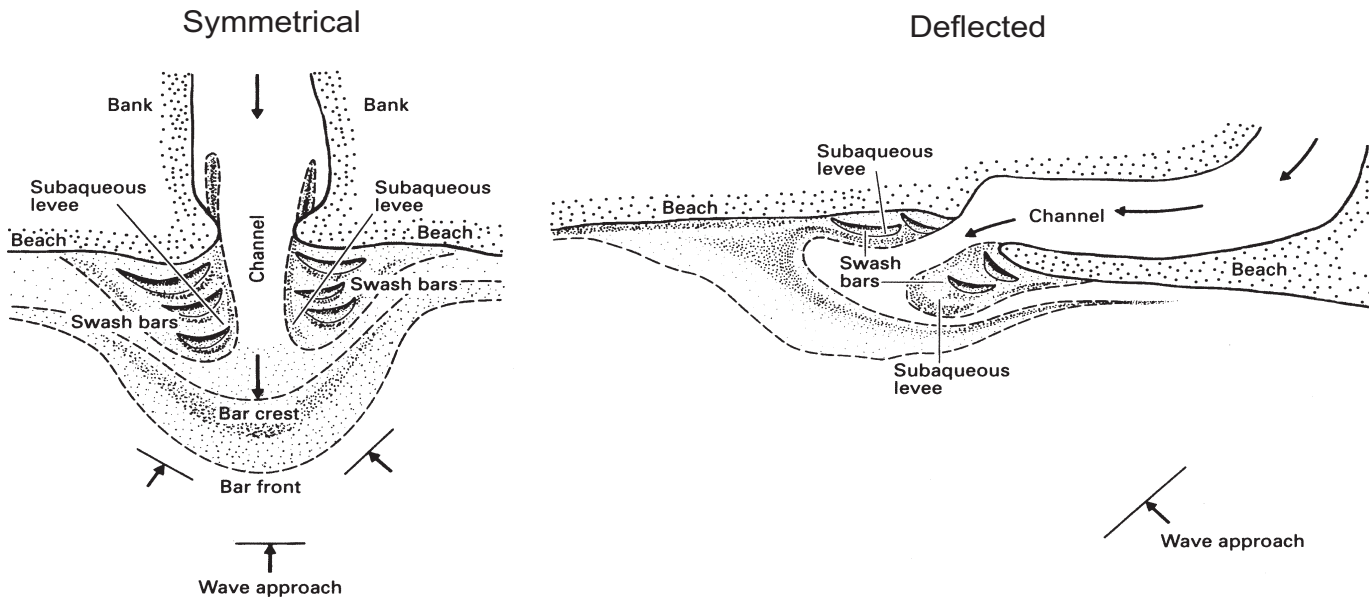


FIG. 12.—**A**) Symmetrical mouth bars, versus **B**) deflected mouth bars. As a result of oblique wave approach (compare with Figure 6). From Reading and Collinson (1996), after Wright (1977).

Fig. 14A, B, C). Sands may occur as thinner, wedge-shaped sheets, or fining-upward shallow undulating channel deposits (Olariu et al., 2005; Gani and Bhattacharya, 2005; Plink-Björklund and Steel, 2005; Plink-Björklund and Steel, 2004). Hyperpycnal turbidites typically show more complex internal geometry than surge-type turbidites (Mulder and Alexander, 2001; Plink-Björklund and Steel, 2004). The sustained flows associated with hyperpycnal flows may result in thick, massive beds that typically show inverse grading at the base (associated with increasing flood discharge) followed by normal grading as the flood wanes. Alternation of structureless to parallel-laminated sand-

stones may also indicate more sustained flows (Plink-Björklund and Steel, 2004).

For sandy systems there has been significant debate about how sandy delta front turbidites form. One hypothesis is that delta-front turbidites are fed by true hyperpycnal flows directly from the proximal delta front caused by rapid sedimentation (e.g., Mulder et al., 1996). The alternate hypothesis is that sandy sediment is first “stored” in a proximal mouth bar, which builds up to a threshold slope and then becomes unstable. Floods, storms, or earthquakes may trigger a delta-front sediment gravity flow. Both processes have been documented for the Modern

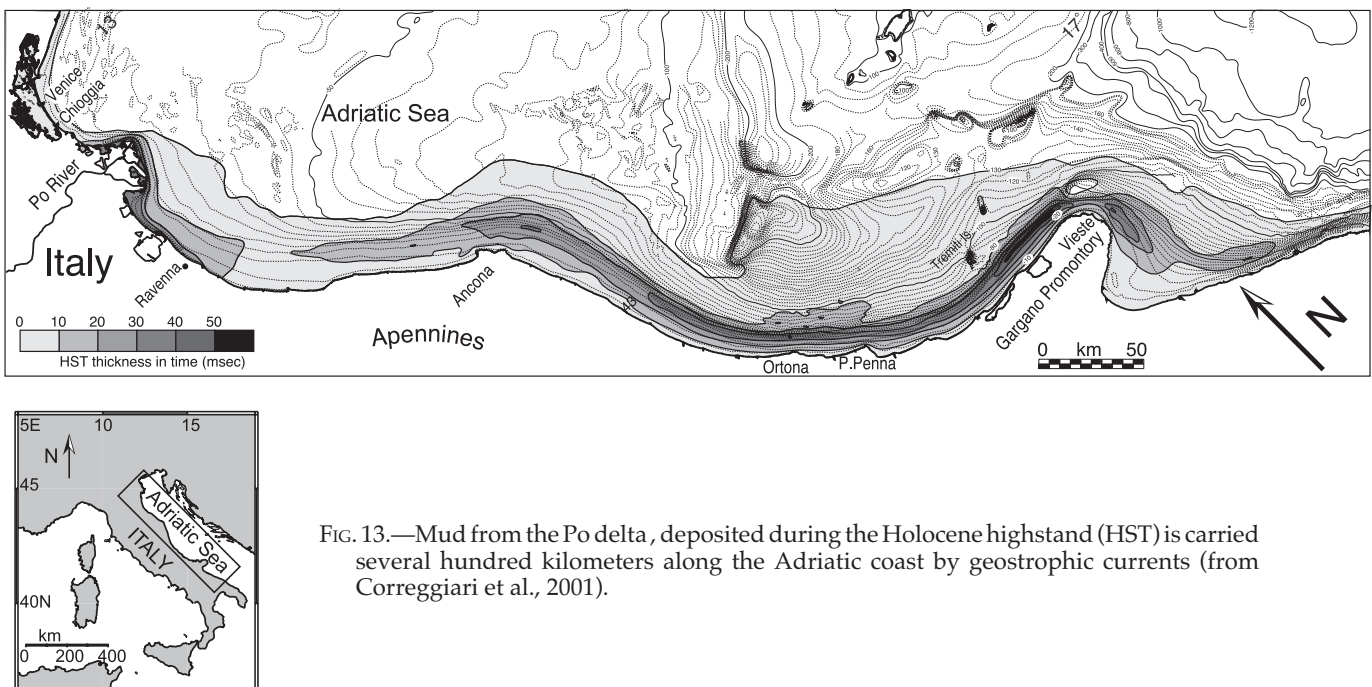


FIG. 13.—Mud from the Po delta, deposited during the Holocene highstand (HST) is carried several hundred kilometers along the Adriatic coast by geostrophic currents (from Correggiari et al., 2001).

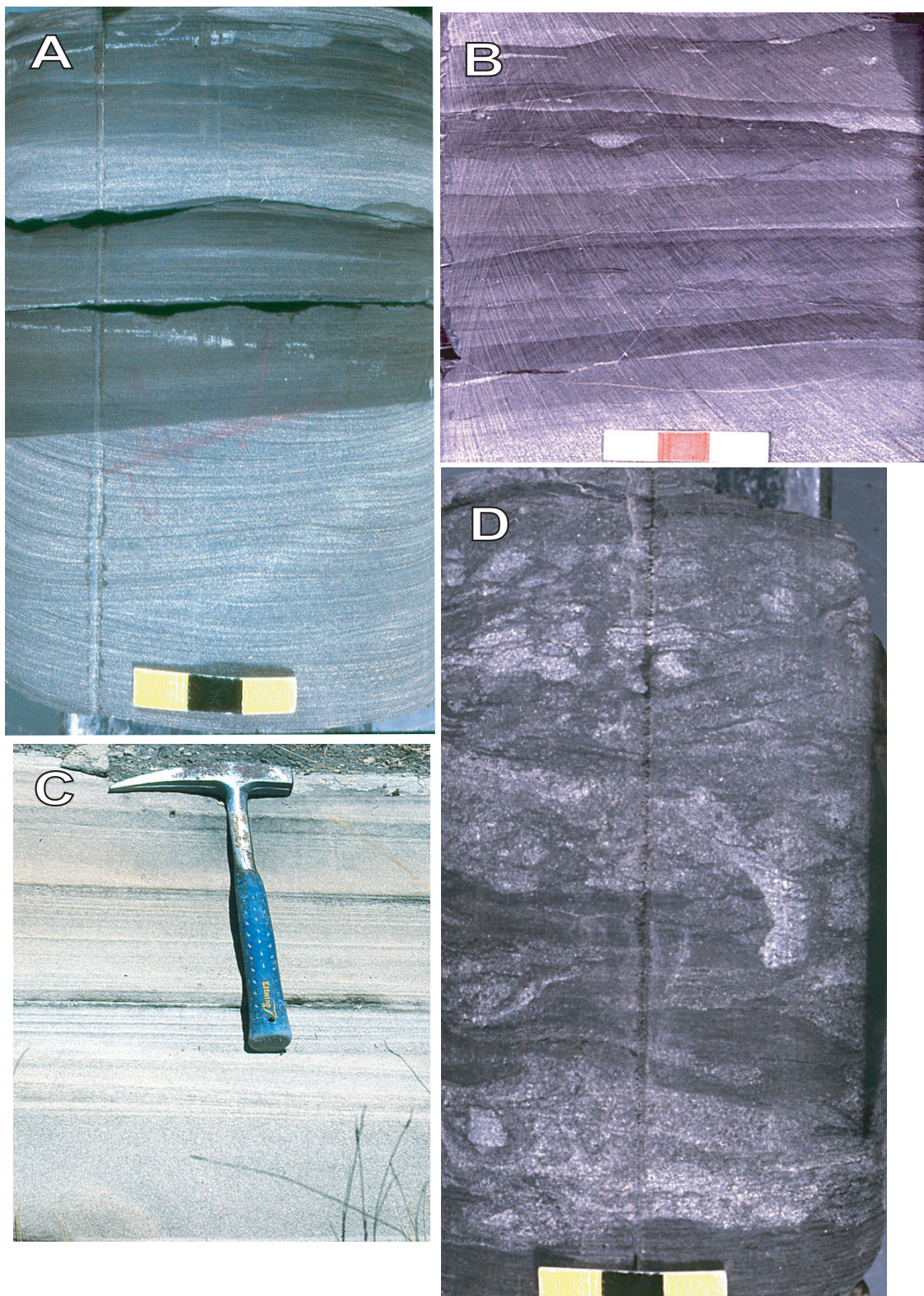


FIG. 14.—**A)** Aggrading wave-rippled sandstones interbedded with normally graded siltstones and claystones, Cretaceous Dunvegan Formation, Alberta, Canada. **B)** Interbedded normally graded very fine-grained sandstones and siltstones with lightly burrowed mudstones. Prodelta mudstones of the Cretaceous Dunvegan Formation, Alberta, Canada. **C)** Normally graded to flat-stratified sandstones of the Cretaceous Panther Tongue sandstone, Utah, U.S.A. A–C are interpreted as delta-front sediment-gravity-flow deposits. **D)** Pervasively bioturbated, non-deltaic sandy mudstone, Cretaceous Dunvegan Formation, Alberta. Except for hammer in Part C, scale bar is 3 cm.

Sepik river mouth in Papua New Guinea (Kineke et al., 2000). In the second scenario, all that is required is rapid sedimentation of the sandy load of the river, which could occur in either hypopycnal or homopycnal settings. In the second scenario, mud may never be deposited from a hyperpycnal flow, and the deposits thus comprise mud deposited from hypopycnal flows that shows less "fluvial" influence (e.g., more normal marine biota), interbedded with rapidly deposited sands with burrowed tops (MacEachern et al., 2005). Muds deposited from suspension accumulate at rates an order of magnitude slower than hyperpycnal muds and consequently show much higher degrees of bioturbation (Fig. 14D; Allison et al., 2000).

Rivers that frequently experience hyperpycnal conditions are typically small "dirty" systems that drain high-relief, tectonically active terrains, such as the Eel River, which feeds the Northern California coast (Mulder and Syvitski, 1995; Syvitski and Morehead, 1999). However, these systems are usually not hyperpycnal throughout the year. Most sediment discharge occurs during rare, large-magnitude floods. 90% of the yearly Eel River discharge, for example, occurred in just a few days of flooding. The rest of the year, the Eel is hypopycnal and carries very little sediment (Syvitski and Morehead, 1999). During low-discharge periods, sediments deposited during major floods may be significantly reworked by waves and tides.

A "pulsed" depositional history characterizes many deltas, such as the Brazos, in the Texas Gulf Coast, the Danube, in the Black Sea, the Senegal in Africa, and the Burdekin delta in Australia (Bhattacharya and Giosan, 2003; Fielding et al., 2005a). Growth of these deltas is confined to very short periods of major flood activity associated with storms. Depending on flood frequency, which can be seasonal or centennial, flood-borne sediment can be completely reworked over time. The Brazos and Burdekin deltas have recently been redefined as "flood-dominated", rather than wave-dominated systems (Rodriguez et al., 2000; Fielding et al., 2005a, 2005b). Storms at the downstream end of a delta system may do little to increase sediment discharge, even though they may cause flooding, whereas storms in the hinterland may be far more important in terms of increasing sediment discharge. In continental-scale deltas, such as the Nile, the Amazon or the Mississippi, there may be no obvious link between coastal storms and hinterland storms, which may be completely out of phase. Rivers associated with continental drainages, in excess of 10^6 km², have been suggested to rarely, if ever, go hyperpycnal (Mulder and Syvitski, 1995), although if marine waters are already brackish, or where marine water is cold, even large rivers may go hyperpycnal frequently (Parsons et al., 2001; Plink-Björklund and Steel, 2004).

Homopycnal (Friction-Dominated)

In homopycnal settings there may be a greater degree of mixing between the river and standing body of water. These situations are common in fresh-water deltas and can also occur in marine settings where the amount of bed load is high. In shallow water, friction at the bed causes rapid deceleration and development of a mouth bar that causes the associated distributary channel to bifurcate, and settings of this kind have been referred to as "friction-dominated" (Wright, 1977). However, friction is important in both hypopycnal and hyperpycnal deltas. "Friction-dominated" mouth bars are more fan shaped than buoyancy-dominated mouth bars, and may be dominated by traction current features such as climbing ripples and cross bedding (Fig. 10). Deltas of this kind are characterized by close-to-angle-of-repose foreset beds, such as seen in the gravelly freshwater deltas originally described by Gilbert (1885). There are several good

examples of marine steep-fronted Gilbert type deltas, such as the Modern Alta delta in Norway (Corner et al., 1990). Numerous ancient examples are given in Colella and Prior (1990), and more recently published examples from Europe include Burns et al. (1997), Ulicny (2001), and Soria et al. (2003).

DELTA ENVIRONMENTS

Deltas comprise three main geomorphic environment of deposition (Fig. 15): the subaerial delta plain (where river processes dominate), the delta front (the coarser-grained area where river and basinal processes interact), and the prodelta (primarily muddy). These three environments roughly coincide with the topset, foreset, and bottomset strata of early workers, although the boundaries overlap and specific definitions of the delta front are not widely agreed on.

Delta Plain

The delta plain is defined by the presence of distributary channels. It includes a wide variety of nonmarine to brackish, paralic to wetland sub-environments including swamps, marshes, tidal flats, lagoons, and interdistributary bays. Although readily distinguished in most modern delta environments (e.g., Ta et al., 2005; Fielding et al., 2005a), the distinction of these various shallow-water, brackish wetland environments is not routinely attempted in ancient settings (but see McCabe, this volume).

The landward limit of modern delta plains is typically taken at the point in the alluvial realm where trunk streams become unconfined and distributive (typically immediately downstream of the alluvial valley). In many cases, this is the nodal avulsion point on an alluvial plain. In modern settings, the delta plain can be subdivided into a lower delta plain, marked by tidal incursion of sea water, and a more landward upper delta plain, in which major distributary channels still occur but in which there is no incursion of marine water (Fig. 15; Coleman and Prior, 1982).

The demarcation between these areas is referred to as the bay line (Posamentier et al., 1988). Rivers may experience tidal modulation of flow far upstream of the actual marine incursion, depending on the ratio of tidal forces to river discharge. The landward limit of incursion depends on slope and discharge. In ancient settings, the bay line may be indicated by the landward limit of marine or brackish-tolerant fossils or trace fossils, although tidally influenced cross stratification may occur farther upstream of any measurable brackish influence. The seaward limit of the lower subaerial delta plain is defined at either the high-tide shoreline (Elliott, 1986) or the low-tide shoreline, which includes the foreshore (e.g., Coleman and Prior, 1982).

The upper delta plain is a fluvial environment, although in rare cases it may be indirectly tide influenced. Lakes lack tides, and consequently the distinction between the upper and lower delta plain cannot be made in lacustrine deltas. Steeply sloping fan deltas, adjacent to scarps, have very limited delta plains.

Delta Front

The **delta front** is defined as the shoreline and adjacent dipping sea bed (Elliott, 1986). It is defined as the area dominated by coarser sediment (sand or gravel) that includes subaqueous topset and foreset beds. However, many studies of modern delta systems do not include the foreshore and shoreface environments within the delta front but rather treat this as a separate intertidal to subtidal "delta platform" environment (Fig. 16; e.g., Coleman and Prior, 1982; Ta et al., 2002; Roberts

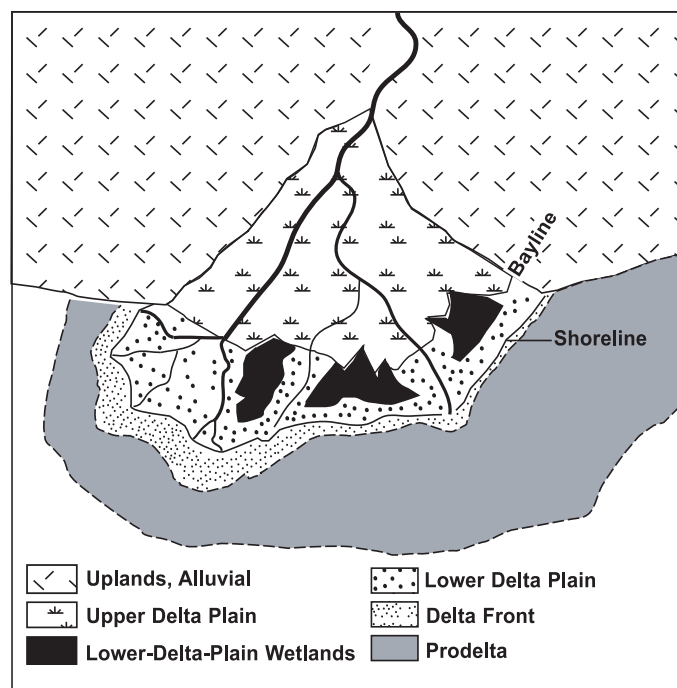


FIG. 15.—Major areal subdivisions of a delta. The upper delta plain is essentially nonmarine and characterized by distributive river systems.

and Sydow, 2003). The width of this subtidal platform can be up to several kilometers where tidal range is high (e.g., Corner et al., 1990; Hori et al., 2002; Ta et al., 2002; Allison et al., 2003; Roberts and Sydow, 2003).

River-dominated delta fronts typically consist of a complex association of terminal distributary channels and mouth bars that coalesce to form bar assemblages and depositional lobes (e.g., Van Heerden and Roberts, 1988). In hypopycnal river-dominated settings, especially those with minimal tides or waves, the posi-

tions of distributary channels may be fixed for long periods, forming elongate bar fingers, such as in the deeper-water mud-dominated Mississippi “birdfoot” delta (Figs. 8, 17; Fisk, 1961). However, the elongation of the modern Mississippi “birdfoot” delta is somewhat artificial, because it has been maintained for many decades by the U.S. Army Corps of Engineers. By contrast, in siltier or sandier systems deposited in shallower water, or not stabilized by human interference, distributaries switch more rapidly and coalesce to form more lobate deltas, as in the Lafourche (Fig. 17) and Atchafalaya (Fig. 11) deltas (Olariu and Bhattacharya, 2006).

The seaward-dipping slope associated with the distal margin of a distributary-mouth bar is also sometimes referred to as the **distal delta front** and can form a relatively continuous sandy fringe in front of the active zone of mouth bars. Internally, the distal bar is physically built by rapidly decelerating “frontal splays”. In high-slope delta fronts, these can be expressed as normally or inversely graded beds deposited from waning turbidites or grain flows (Porebski and Steel, 2003; Plink-Bjorklund and Steel, 2005; Olariu et al., 2005). In coarser deltas, these deposits produce the classical foreset geometries that define Gilbert deltas (e.g., Soria et al., 2003). Several researchers (e.g., Coleman and Prior, 1982) reserve the term delta front to refer *only* to this distal bar environment. The term delta front has also been applied to refer to mid-shelf muddy clinoform strata, seaward of any significant sand deposits (Fig. 16; Roberts and Sydow, 2003).

Delta Front Versus Shoreface

In wave-influenced deltas, mouth bars may be reworked into a distinct shoreface, which can be considered as part the delta front (Barrell, 1912). The **shoreface** is the seaward-dipping equilibrium surface that forms in response to the asymmetry of shoaling fair-weather waves (Barrell, 1912; Bruun, 1962; Swift, 1968). The shape and extent of the shoreface depends on the interaction of sediment supply and wave energy expended at the coast (Walker and Plint, 1992). Non-oceanic settings, or coasts with wide shelves, typically lack swell waves, for example, and have a correspondingly diminished shoreface in which fair-weather waves affect only sediments deposited in a few meters or

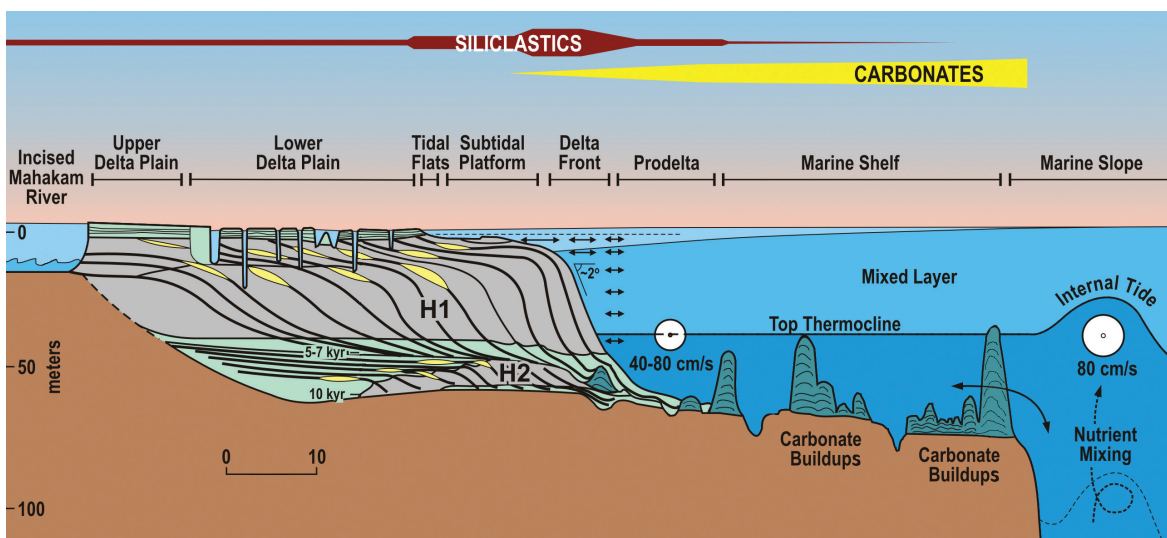


FIG. 16.—Morphometric subdivisions of the Mahakam delta, Kalimantan, Indonesia. Note that muddy subaqueous foreset is referred to as the “delta front” (modified after Roberts and Sydow, 2003).

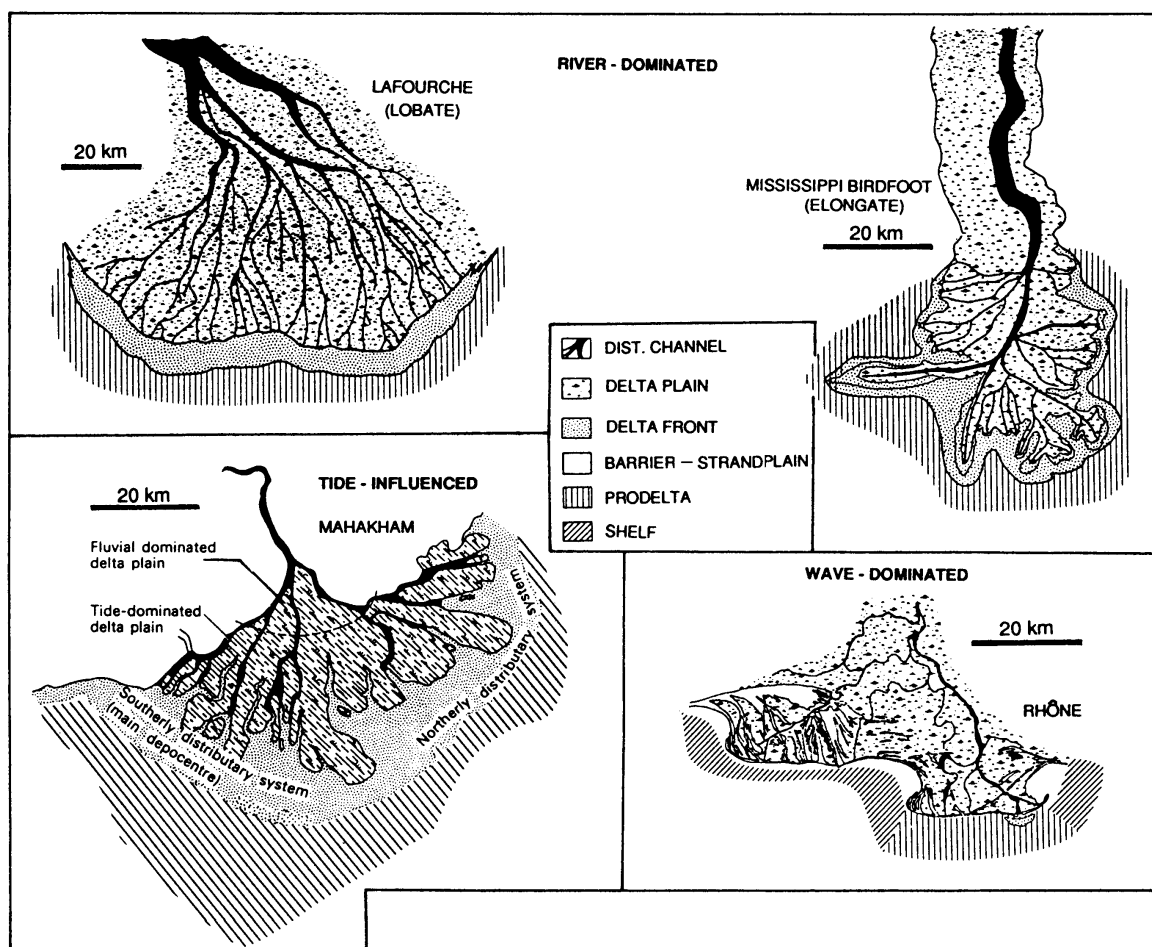


FIG. 17.—Representative modern examples of river-dominated, wave-dominated, and tide-influenced deltas. Modified from Fisher et al. (1969). River-dominated deltas are classified into lobate (shoal-water), and elongate (deep-water or birdfoot deltas). In the Mahakham example (after Allen et al., 1979), delta-front deposits comprise sandy siltstones and mudstones. Figure from Bhattacharya and Walker (1992).

less. This may nevertheless impart a smooth-fronted appearance to the delta, but sediments deposited below the effects of fair-weather waves will record the original depositional processes. In mesotidal or macrotidal settings, tidal process may mask the effects of waves. Many smooth-fronted modern deltas, such as the Brazos, Burdekin, Baram/Trusan, and Mekong, while showing the effects of shallow-wave reworking, show a dominance of river-flood or tidal facies in the underlying sediments (Rodriguez et al., 2000; Lambiase et al., 2003; Ta et al., 2002, 2005; Fielding et al., 2005a, 2005b).

In wave-modified deltas, the shoreface may form part of the delta front. Shorefaces can also form in the absence of a river. The shoreface can also be entirely erosional, especially during transgression, where sediment supply may be minimal (e.g., Bruun, 1962; Swift, 1968; Nummedal and Swift, 1987; Kraft et al., 1987). This erosion forms a **ravinement surface** that commonly removes 5–10 m of the topset portions of a delta. These ravinement surfaces form profoundly significant bounding discontinuities that are the key to identifying and mapping ancient top-truncated delta deposits (e.g., Weise, 1980; Walker and Plint, 1992; Hart and Long, 1996; Posamentier and Allen, 1999; Bhattacharya and Willis, 2001; Martinsen, 2003).

Prodelta

The **prodelta** has historically been interpreted as the area where fine mud and silt settle slowly out of suspension. Prodelta deposits may be more or less burrowed, depending on sedimentation rates. Prodelta muds may merge seaward with fine-grained hemipelagic and commonly calcareous sediment of the basin floor. The preservation of silty or sandy lamination is commonly taken to mark the influence of the river, as opposed to total bioturbation of the basin-floor sediments in areas away from the active river (Fig. 14D; Allison et al., 2000; Neill and Allison, 2005; MacEachern et al., 2005). Where the sediments are rhythmically laminated, a tidal influence may be inferred (Smith et al., 1990). Because of the abundant suspended sediment, certain types of vertical filter feeders and other organisms that produce open vertical burrows of the *Skolithos* ichnofacies tend to be suppressed (e.g., Moslow and Pemberton, 1988; Gingras et al., 1998; MacEachern et al., 2005).

The term prodelta and shelf have been presented historically as mutually exclusive environments (e.g., Walker, 1984), which may be a serious error. Many of the world's muddy shelves, such as the Adriatic (Fig. 13), Black Sea, Amazon, Bay of Bengal, Papua

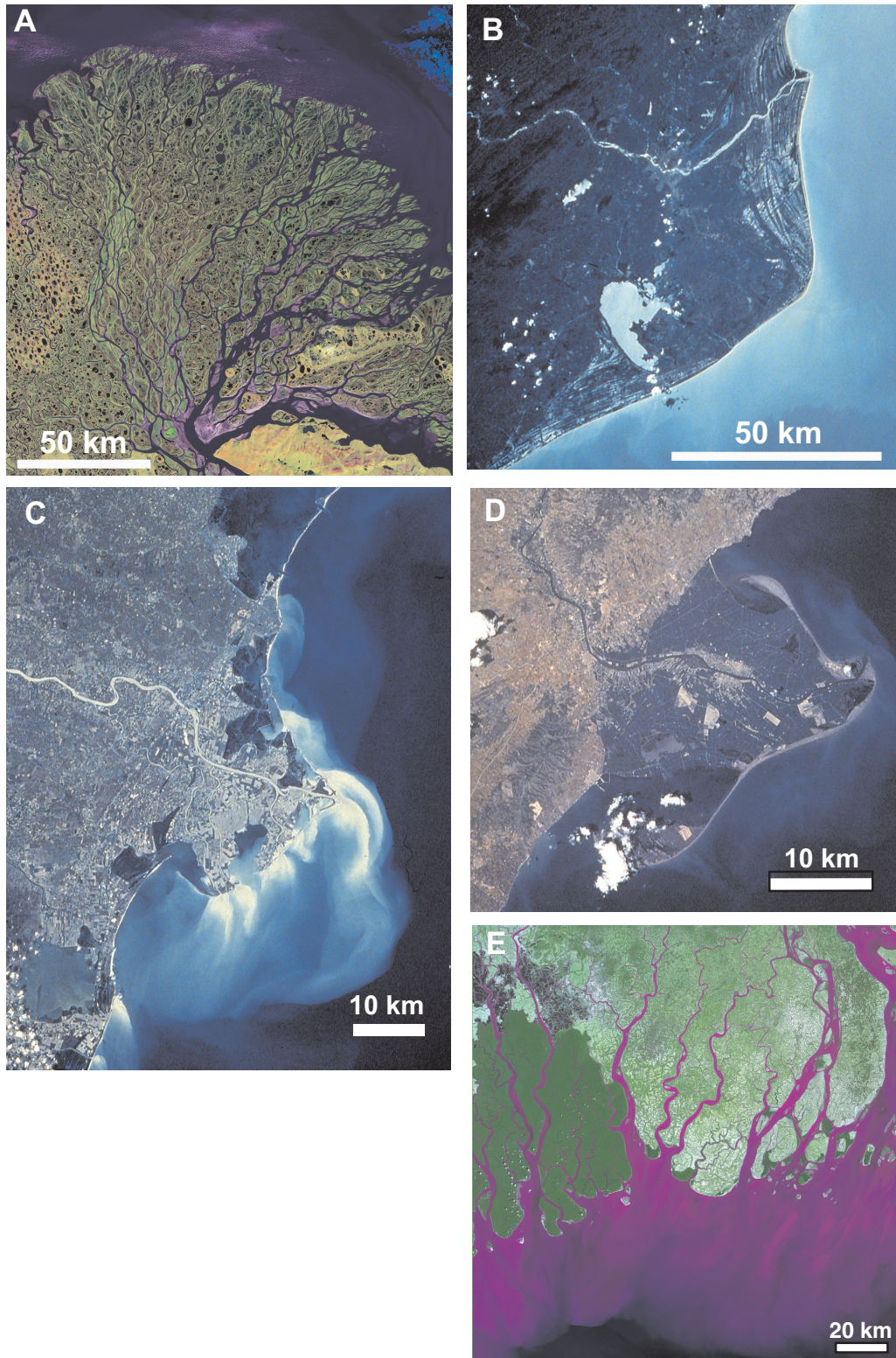


FIG. 18.—Comparison of distributary-channel branching patterns in a river-dominated versus wave-dominated deltaic coastline. **A)** River-dominated Lena River delta (Russian Arctic) shows numerous orders of branching with many tens of terminal distributary channels. **B)** Wave-dominated coastline associated with the Paraiba do Sul, Brazilian coast. **C)** Po delta, Italy. **D)** Ebro delta, Spain. Bifurcation is inhibited in wave-dominated deltas because the river is unable to prograde into the basin as rapidly. This effectively allows the river to maintain its grade, which in turn inhibits avulsion. **E)** Tide-dominated Ganges–Brahmaputra delta shows highly elongate channels. Photos courtesy of NASA.

New Guinea, Gulf of Mexico, and others, are now being interpreted as the subaqueous extension of deltas (e.g., Nittrouer et al., 1986; Kuehl et al., 1997; Michels et al., 1998; Liu et al., 2002; Cattaneo et al., 2003; Roberts and Sydow, 2003; Kuehl et al., 2005; Neill and Allison, 2005). Studies of modern muddy shelves show that much of the muddy sediment deposited by suspension out of buoyant river plumes ultimately concentrates at the seabed, forming a fluid-mud layer that may be kept in suspension by waves (e.g., Kineke et al., 1996) or moved by storms (Allison et al., 2000; Draut et al., 2005). Mud may also be introduced directly onto the seafloor by hyperpycnal flows (Mulder and Syvitski, 1995). Because sedimentation rates of fluid muds are so much higher than by suspension settling, they are probably far more important in the construction of the shelf than has historically been realized (Neill and Allison, 2005). Fluid muds may be characterized by centimeter- to decimeter-thick beds which show a lack of lamination and bioturbation and which are interbedded with graded siltstone or sandstone beds (Kuehl et al., 1986; Allison et al., 2000; MacEachern et al., 2005).

Distributary Channels

Distributary channels may show a wide range of sizes and shapes in different position on the delta (Fig. 18; Olariu and Bhattacharya, submitted). There is therefore no such thing as “a distributary channel” in many deltas. Typically, a trunk fluvial system first avulses at the point where the river becomes unconfined forming a nodal avulsion point (i.e., Nelson, 1970; Mackey and Bridge, 1995). Delta-plain channels tend to be few in number and are separated by wide areas of interdistributary bays, swamps, marshes, or lakes on the delta plain, although these interdistributary areas can be replaced by channel deposits depending on the avulsion frequency and rate of channel migration (Bristow and Best, 1993; Mackey and Bridge, 1995; Holbrook, 1996). In an ancient setting, upper-delta-plain channels may be very difficult to distinguish from fluvial channels, especially if there is no tidal influence.

Distributary channels can show several orders of branching, readily measured in modern systems but difficult to determine in ancient examples. The smallest-scale channels are referred to as “terminal distributary channels” and are intimately associated with mouth bars that form at the distal delta plain and proximal delta front (Olariu and Bhattacharya, 2006). Terminal distributary channels can extend several kilometers offshore forming, channelized to scoured facies within the delta front (Olariu and Bhattacharya, 2005).

Distributary-channel bifurcation occurs at a point where the channel can no longer cut directly through the distributary-mouth bar, forcing it to split into two smaller channels flanking the bar crest. Channel-bifurcation frequency and branching patterns are strongly dependent on slope, river discharge, water depth, and the interaction of the river plume with marine processes (Fig. 18). Multiple bifurcations are favored in low-gradient, high-discharge, river-dominated deltas, where friction is the dominant process controlling sediment dispersal and deposition (Welder, 1959; Wright, 1977). Nodal avulsion of trunk streams and distributary crevassing are common processes in river-dominated deltas because hydraulic gradients decrease as rivers and distributaries extend their courses. Friction, caused by mouth bars and plume dispersion, also reduces the discharge.

In wave-modified deltas, much of the sediment delivered to the shoreline is carried away from the river mouth by longshore transport (Figs. 6, 18). Thus, relative to river-dominated deltas, the progradation rate of wave-influenced deltas is slowed. This allows rivers feeding wave-influenced coasts to maintain a

higher slope, which inhibits avulsion. As a consequence, wave-influenced deltas typically have only a few active distributary channels (Figs. 18B, C, D) whereas river-dominated deltas can have tens to hundreds of active terminal distributary channels (Fig. 18A).

Many tidally influenced deltas show distributary channels that are stable for hundreds to thousands of years, such as in the Mekong delta (Ta et al., 2002; Ta et al., 2005) and Ganges-Brahmaputra (Kuehl et al., 2005). This results in the development of elongate bars and islands that can be tens of kilometers in length and a few kilometers wide (Fig. 18E). Increased channel stability results in far more elongate sand bodies, with higher length-to-width ratios than are typically found in fluvially-dominated delta fronts (ratio of 10 versus 2, respectively; Reynolds, 1999).

In systems with many orders of branching, younger and shorter-lived distributary channels, lower on the delta plain, tend to be straighter because of lower slopes and lower discharge, whereas longer-lived channels on the upper delta plain can show complex and highly sinuous or braided channel patterns because of higher slopes. In wave- or tide-dominated deltas, rivers may retain their trunk character all the way to the shoreline. This is why the Ganges-Brahmaputra and São Francisco rivers are braided at the river mouth.

Terminal distributary channels are difficult to recognize in subsurface because they tend to be shallow. Successive higher-order distributary channels should become thinner and narrower downstream, as has been documented in ancient channels within the Devonian “Catskill” deltaic wedge of the Appalachian basin, U.S.A. (Bridge, 2000). Thus, one way to determine if an ancient fluvial system is distributive is to see if the widths and depths of channels become smaller in more distal reaches of a clastic wedge.

Distributary Channels Versus Incised Valleys and Overapplication of the Mississippi Analogue.—

Many ancient examples of so-called river-dominated deltas exhibit thick channelized deposits overlying marine prodelta shales and have been interpreted as distributary channels cutting into their associated delta fronts (Fig. 19; e.g., Busch, 1971, 1974; Cleaves and Broussard, 1980; Rasmussen et al., 1985) despite the fact that this is rare in many modern deltas. Some of these sandstones are over 30 m thick and cut out delta-front sandstones that are only 10 m thick. Many of these deeply incised channels are now recognized as valley fills rather than distributary channels (Willis, 1997; Bowen and Weimer, 2003; Bhattacharya and Tye, 2004).

The earlier interpretations of these features as distributary channels stemmed largely from comparison with the deep and stable distributary channels of the modern Mississippi birdfoot delta. The fact that deep Mississippi distributary channels erode into underlying prodelta muds has been cited as the main reason why distributary channels, in general, do not migrate laterally (Coleman and Prior, 1982). However, the Mississippi is a continental-scale system, and the modern channel has been kept in place through the dredging efforts of the U.S. Army Corps of Engineers. It is an inappropriate analogy with which to interpret many shallower-water, mid-continent delta systems, such as developed in the Pennsylvanian and Cretaceous systems of North America, which drained considerably smaller areas (Bhattacharya and Tye, 2004).

Regional Controls on Delta Morphology

Many other factors may also influence the delta form, apart from the nature of the fluvial input and the reworking of the

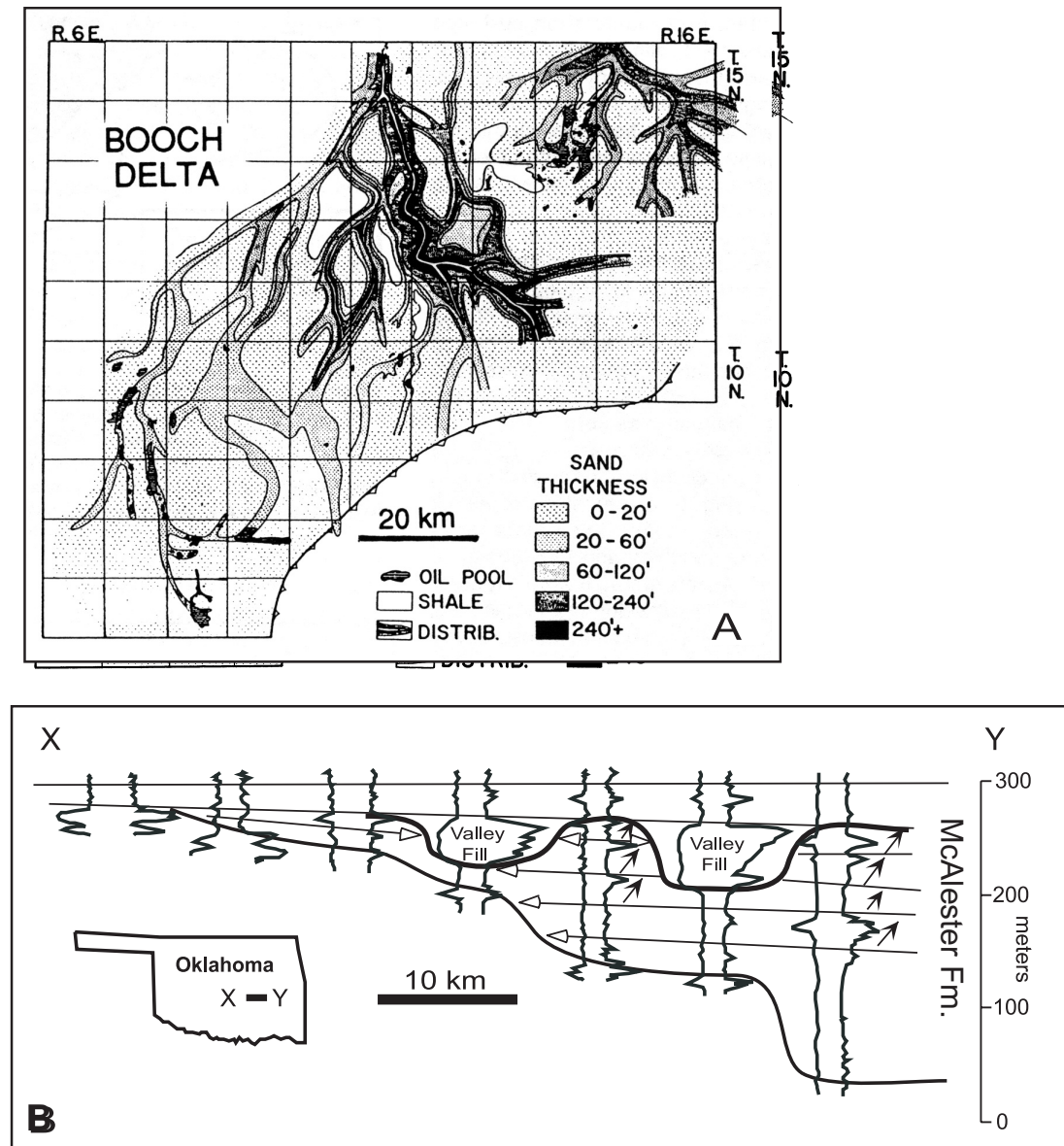


FIG. 19.—Ancient examples of dendritic “shoestring” sandstones of the Pennsylvanian Booch sandstone (Oklahoma). **A**) Map view suggests river-dominated, elongate deltas. This interpretation was heavily biased by modern Mississippi birdfoot delta (after Busch, 1971). **B**) Well-log cross section suggests over-thick valleys. Many of these systems should probably be reinterpreted as incised valleys.

deposits by waves and tides. Coleman and Wright (1975) emphasize the nature of the receiving basin, the nature of the drainage basin, the tectonic setting, and climate. In addition, relative sea-level changes influence the extent of delta growth and destruction. These factors are not all independent. Sediment type and rate of supply, for example, are a function of the size, relief, climate, and underlying geology in the drainage basin. Relief may be dependent on the tectonics of the drainage basin. Wave or tide energy may be a function of eustasy, shelf slope, and size and shape of the receiving basin, and wave energy is also related to climate (e.g., wind direction and strength).

For example, tidal range is typically enhanced within coastal embayments. Many of the world’s largest deltas are tide-domi-

nated, because they lie within major straits or continental reentrants, such as the tide-influenced Ganges–Brahmaputra delta (Fig. 18E), which lies at the head of the Bay of Bengal. The tropical deltas of southeast Asia lie within a part of the world that has very low wave heights but high tides, which also contributes to the generally tide-dominated nature of deltas there (Nummedal et al., 2003). Sediment type and rate of supply were also influenced by the absence of land plants in pre-Devonian rocks, resulting in higher sedimentation rates and a greater proportion of fan deltas (Stow, 1986; MacNaughton et al., 1997). Deltas formed against scarps or faults typically form as fan deltas with little to no delta plain.

The focus of geologists on studying sandstone has resulted in the erroneous notion that many rivers carry primarily sand. Most

ivers carry between 85% to 95% mud (Schumm, 1972), chiefly in suspension. Mud-free rivers are rare in nature, and most modern mud-free rivers owe their lack of suspended material to the fact that the suspended load is deposited in dams, far upstream of the river mouth. There are very few studies of Modern systems that document how mud is partitioned between the delta plain (i.e., floodplain) versus the prodelta shelf environment and what are the key controlling factors on sediment partitioning (e.g., Goodbred and Kuehl, 1999). In contrast, sequence stratigraphic studies of ancient systems have demonstrated that sediment is partitioned over geological time as a consequence of changes in accommodation, sediment supply, and sea-level change (e.g., Jervey, 1988; Posamentier et al., 1988; Helland-Hansen and Gjelberg, 1994; Posamentier and Allen, 1999).

CLASSIFICATION OF DELTAS

The commonly used tripartite classification of deltas (Fig. 5; Galloway, 1975) is based on the idea that the ratio of fluvial, wave, and tidal processes results in different and identifiable plan-view morphology of resulting deposits as well as characteristic internal facies successions. Unfortunately, there has been a natural tendency for workers to force-fit their particular example into one of the end-member categories (e.g., Bhattacharya and Walker, 1992; Dominguez, 1996), despite the fact that most deltas are likely to be mixed-influence and plot somewhere within the triangle. Many modern deltas, such as the Danube, show a mixture of delta types both between and

within discrete lobes (Fig. 9). This can create problems for interpreters, especially in subsurface studies, where the nature of a depositional system is typically determined based on sparse core data, which may not represent the whole system. Furthermore, the terms “dominated” versus “influenced” have never been adequately distinguished. These could be quantified in terms of wave or tide energy expended at the coast versus sediment discharge or sedimentation rate at the river mouth or other measurable parameters (e.g., as attempted by Coleman and Wright, 1975, Hayes, 1979, and Dalrymple, 1992), but these are virtually impossible to measure or determine in an ancient system (see, however, Bhattacharya and Tye, 2004). Another approach would be to measure the physical proportion of facies that were formed by wave, tide, or fluvial processes. This approach is more applicable to ancient systems, and it may be especially applicable to reservoir or aquifer modeling, where facies architecture may have a first-order control on flow behavior. In a study of the Baram and Trusan deltas of Borneo, Lambiase et al. (2003) showed that despite a smooth-fronted external geometry, suggestive of wave domination, the internal facies show a strong tidal signature. The plan-view shape of the modern Brazos delta has long been cited as a classic example of a wave-dominated delta, but recent coring studies show a dominance of river-flood deposits and have led Rodriguez et al. (2000) to reclassify the Brazos as a river-flood-dominated, wave-influenced delta. Similar work on the Burdekin delta in Australia shows that despite its smooth-fronted external appearance, which historically led it to be classified as wave-dominated,

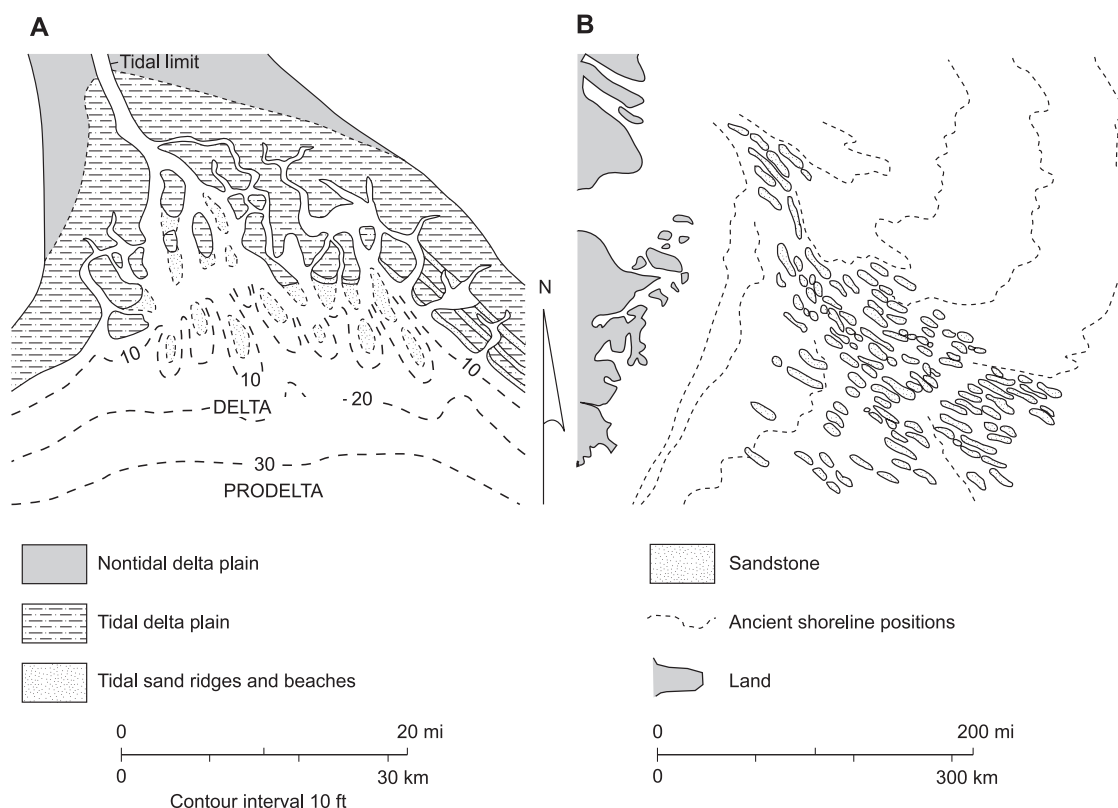


FIG. 20.—Tide dominated deltas. **A)** Tidally reworked mouth bars are highly elongated. Central part of delta is predominantly sandy bars whereas mud is partitioned along sides of system (from Dalrymple, 1992). **B)** During transgression of the East China Sea, mouth bars are reworked into elongate, shore-normal shelf ridges (after Yang and Sun, 1988).

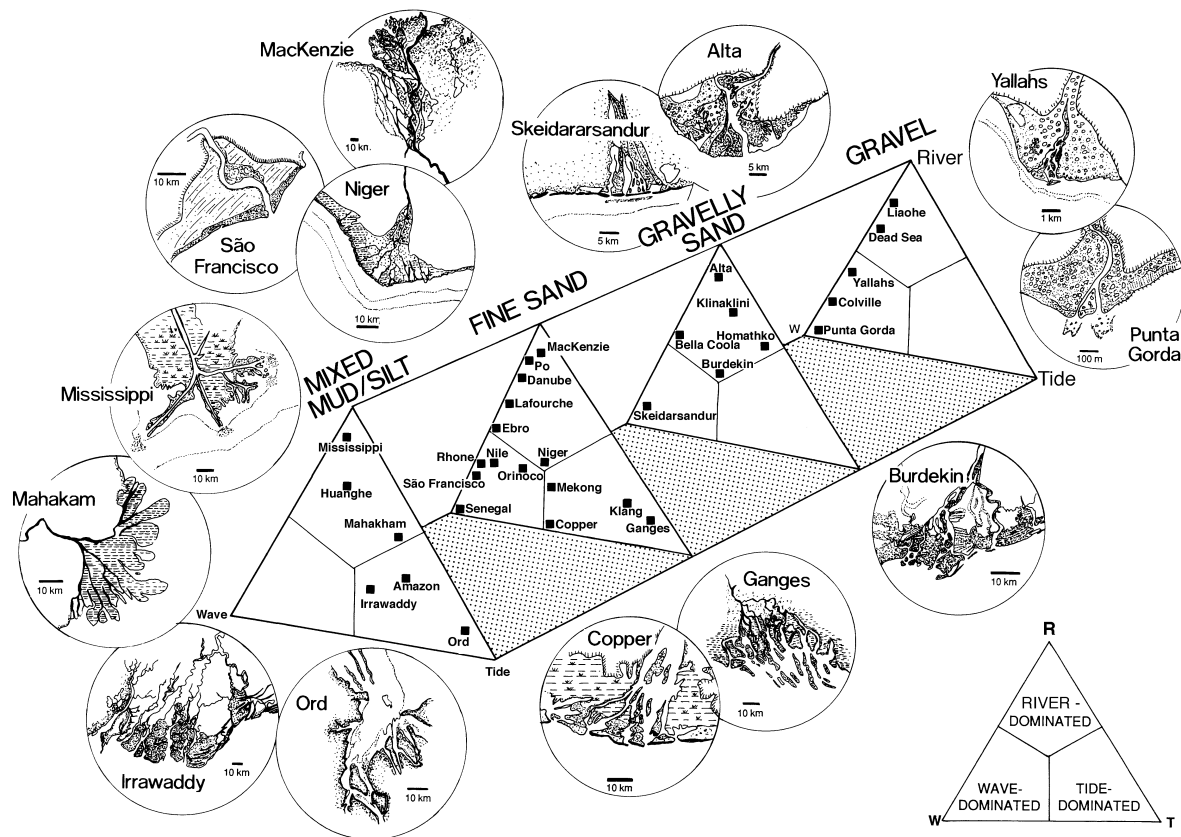


FIG. 21.—Delta triangle of Galloway (1975) is extended to include sediment caliber as a fundamental control (from Reading and Collinson, 1996, after Orton and Reading, 1993).

internal facies from cores and outcrops show a predominance of river-flood deposits (Fielding et al., 2005a, 2005b)

River dominated deltas display an overall digitate or lobate morphology (Figs. 4, 5, 8, 11, 17, 18). In contrast, wave-influenced deltas show smooth-fronted lobes with arcuate to cusped margins (Figs. 1, 4, 5, 6, 9, 17, 18). Tidal processes form sand bodies oriented parallel to the directions of the tidal currents (Figs. 4, 5, 17, 18E, 20), typically perpendicular to regional shorelines (Dalrymple, 1992; Maguregui and Tyler, 1991; Willis and Gabel, 2001). The area of the marine-influenced lower delta plain and delta front may be quite extensive where tidal range is large. In tide-dominated settings, mud is partitioned along the margins of the system (Fig. 20), in contrast to wave-dominated estuaries, where a wave-formed barrier-beach complex protects the estuary mouth and mud accumulates in the lagoon behind the barrier.

Coleman and Wright (1975) recognized that the geometry of a delta sand body should reflect the relative importance of fluvial and marine processes. All of their delta geometries (Fig. 4) emphasize narrowing and thickening of sands towards a point (i.e., fluvial) source, but the seaward margins differ, as explained above. Orton and Reading (1993) extended the Galloway classification to include sediment type (Fig. 21). Postma (1990) presented an independent classification scheme based on the type of feeder system, water depth, and mouth-bar process (Fig. 22). This classification scheme does not, however, include waves or tides as key parameters.

The term “braid delta” or “braidplain delta” has been used to refer to a sandy or gravelly delta front fed by a braided river and characterized by a fringe of active mouth bars (e.g., McPherson et al., 1987). This term must be used with caution. There are many examples of braided rivers that feed highly wave-influenced deltas (e.g., the São Francisco and Paraíba do Sul in Brazil) or tide-influenced deltas (e.g., the Ganges–Bhramaputra) that bear little resemblance to the so-called “braid deltas” of McPherson et al. (1987).

Dalrymple et al. (1992) extended the delta triangle of Galloway (1975) into three dimensions to include the sequence stratigraphic concept that the abundance of different depositional systems is a function of relative sea-level change (Fig. 23). They emphasize the relationship between “regressive” delta-type systems and “transgressive” depositional systems, such as estuaries and barrier-lagoons. Although this is a valuable extension of Galloway’s work, missing from this diagram is the fact that many deltas contain barrier-island–lagoon systems, tidal flats, and even drowned abandoned distributaries, which may exhibit a strongly estuarine-type fill (Fig. 9).

Seismic-stratigraphic studies and sequence-stratigraphic studies of deltas led to a recognition that depositional systems change their character as a function of their physical and temporal position. For example, shelf-edge deltas tend to form at sea level lowstands (Fig. 24; e.g., Edwards, 1981; Posamentier et al., 1992; Tesson et al., 1993). Deltas deposited at the shelf edge are commonly unstable and develop impressive growth faults. Sand

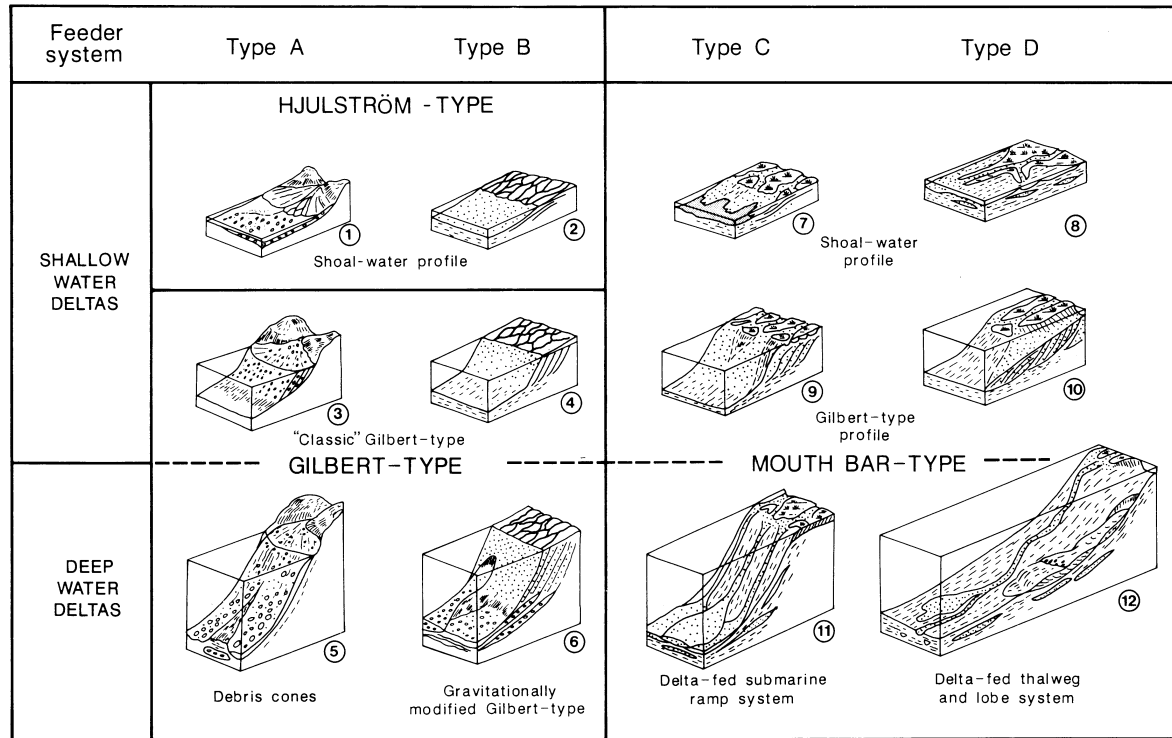


FIG. 22.—Classification of coarse-grained delta types incorporating type of feeder system, water depth, and type of mouth-bar process (from Reading and Collinson, 1996; after Postma, 1990).

bodies in these settings are often aligned along strike (Fig. 25), but this elongation is controlled by subsidence along the growth faults rather than by wave processes.

VERTICAL FACIES SUCCESSIONS

Coleman and Wright (1975) presented, in addition to sand body shapes, a series of composite vertical facies successions through the prodelta, delta-front, and delta-plain environments of each of their delta types. These idealized facies successions represent “norms” and may be very useful points of reference for outcrop studies where three-dimensional control may be limited. Although a single vertical profile may not be representative, especially in deltas that show extreme lateral facies variability, 1D vertical profiles remain the most common data type on which any study of depositional systems, be it core, well log, or outcrop data, is based. Lateral facies relationships, in contrast, can be interpreted directly from seismic data (e.g., papers in Anderson and Fillon, 2004) or in continuous cliff panoramas (e.g., Willis et al., 1999; Willis and Gabel, 2001; Soria et al., 2003; Barton, 2004; Garrison and van den Bergh, 2004). Typical facies successions through the dominantly marine (prodelta and delta front) and dominantly nonmarine (delta plain) parts of deltas, mostly in river- and wave-dominated settings, are outlined below.

Prodelta and Delta Front Successions

Progradation of a delta commonly produces a coarsening-upward facies succession (Fig. 26) showing a transition from muddier facies of the prodelta into the sandier facies of the delta-front and mouth-bar environments (Elliott, 1986; Coleman and

Wright, 1975). Thick mudstone deposits that do not display a distinctive upward-coarsening facies succession may occur in areas away from the river mouth, such as where there is significant alongshore diversion of a muddy river plume.

Depending on the shoreline trajectory during progradation, significant delta-plain facies can accumulate above and behind the migrating and subsiding delta front (Fig. 27). Thicknesses of upward-coarsening facies successions may range from a few meters to a hundred meters depending on the scale of the delta, the water depth, the shoreline trajectory and the subsidence rate. Delta-front sands may subsequently be partially eroded as the distributary channel migrates over its own mouth bar (Fig. 26). Recent studies of deltas that build into very shallow water (e.g., Kroonenberg et al., 1997; Overeem et al., 2003; Fielding et al., 2005a, 2005b) show that upward-coarsening successions are more difficult to produce, because there is little space to accumulate a thick underlying prodelta platform over which the delta can build. Where channel flow depths are on the same scale as water depth, the channel flows more easily cannibalize underlying muddy facies and the facies succession is dominated by sharp-based coarsening-upward mouth bars or fining-upward distributary channel fills (e.g., Holbrook, 1996). These sharp-based mouth bars may produce facies succession that are very similar to those produced in other low-accommodation settings, such as during forced regressions (Plint, 1988; Posamentier et al., 1992; Fielding et al., 2005a, 2005b).

In shelf-edge deltas (Fig. 24), thick upward-coarsening delta-front successions can be preserved within the hanging wall of growth faults, although they show increasing dip with depth. In the shallower landward portions, greater reworking by shallow-marine processes can result in more complex facies successions (Winker and Edwards, 1983).

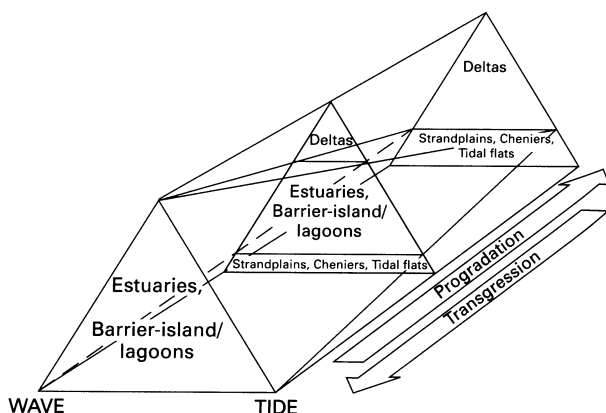
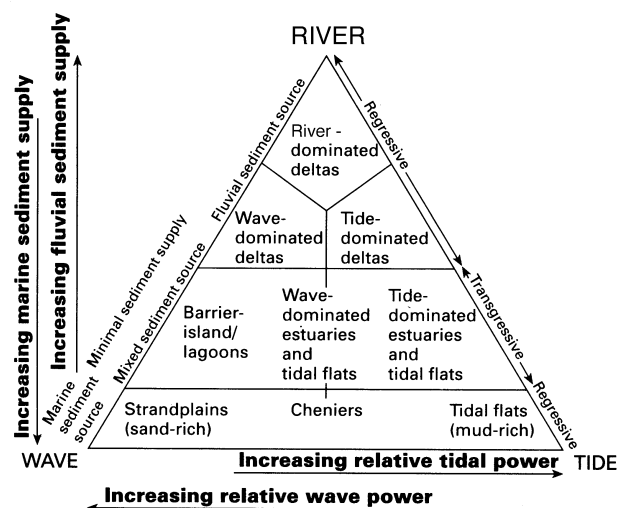


FIG. 23.—Delta triangle of Galloway (1975) is extended by Dalrymple et al. (1992) to reflect changes in sediment supply (from Reading and Collinson, 1996).

The specific nature of the facies and beds in prograding prodelta and delta-front successions depend on the processes influencing sediment transport, deposition, and reworking. Upward-coarsening facies successions can be produced by the progradation of wave-formed shorefaces. However, unless there is a significant supply of sediment, which almost invariably must come from rivers, significant progradation does not occur.

River-Dominated Delta-Front and Prodelta Successions.—

In river-dominated deltas, prodelta sediments are typically heterolithic laminated to thin-bedded mudstones with or without sandstones (Fig. 14A, B). Siltstones and sandstones are typically massive to well stratified and may show graded bedding (Fig. 28, river-dominated column). The graded beds may reflect deposition from hypopycnal density underflows generated at the river mouth during high-discharge floods (Wright et al., 1988; Mulder and Syvitski, 1995). The amount of bioturbation can be variable, depending on rates of sedimentation and the grain size of the sediment supplied (MacEachern et al., 2005;

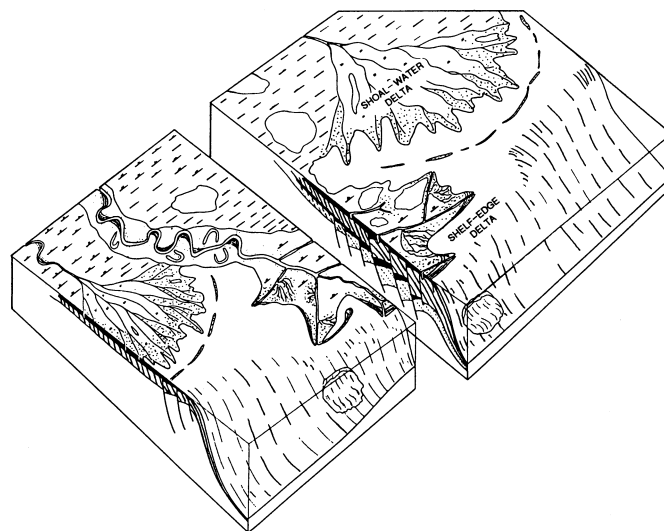


FIG. 24.—Block diagram contrasting lobate shoal-water (or shelf-phase) deltas and shelf-edge deltas. Note thickening of facies across growth faults in the shelf-edge delta (From Bhattacharya and Walker, 1992; after Edwards, 1981).

Neill and Allison, 2005). Wave-formed structures may occur at the tops of graded sandstone beds, but are less abundant than in a more wave-influenced setting. If floods occur during major coastal storms, sets of highly aggrading wave-rippled sandstone beds may occur (Fig. 14A) and hummocky cross stratification may be abundant. Soft-sediment deformation features result from high sedimentation rates and are common in river-dominated deltas (Fig. 29). Deposition of overpressured prodelta muds may cause remobilization of the overlying delta-front sand (Figs. 29, 30; Coleman et al., 1983; Bhattacharya and Davies, 2001; Wignall and Best, 2004). Cores from prodelta and delta-front deposits of the Mississippi show a complex facies architecture dominated by upward-coarsening facies successions (Fig. 31).

Sandy delta front facies predominantly reflect deposition from rapidly decelerating unidirectional flows in distributary-mouth-bar environments. Structures may include unidirectional current ripples and cross bedding, flat-stratified sandstones, or massive graded beds (Fig. 14B, C), depending on the importance of frictional versus inertial processes (Martinsen, 1990). High rates of deposition result in rapid burial and preservation of structures formed by unidirectional or oscillatory flows. Sorting, especially in gravelly systems, may be poor to moderate (Arnott, 1992; Bridge, 2003). Variations in discharge of the fluvial feeder system may produce an irregular coarsening-upward succession, with interbedded mudstones throughout (Figs. 26, 28, 31).

Fresh-water influence may be indicated by an abundance of syneresis cracks, reflecting flocculation and contraction of clays as a result of salinity changes (Plummer and Gostin, 1981), and early diagenetic siderite, which commonly requires fresh-water influx to reduce sulfate activity (Coleman and Prior, 1982; Bhattacharya and Walker, 1991b).

Ichnological Effects of Fluvial Processes.—

Periods of low discharge (e.g., hypopycnal) may result in intense faunal colonization of the substrate alternating with

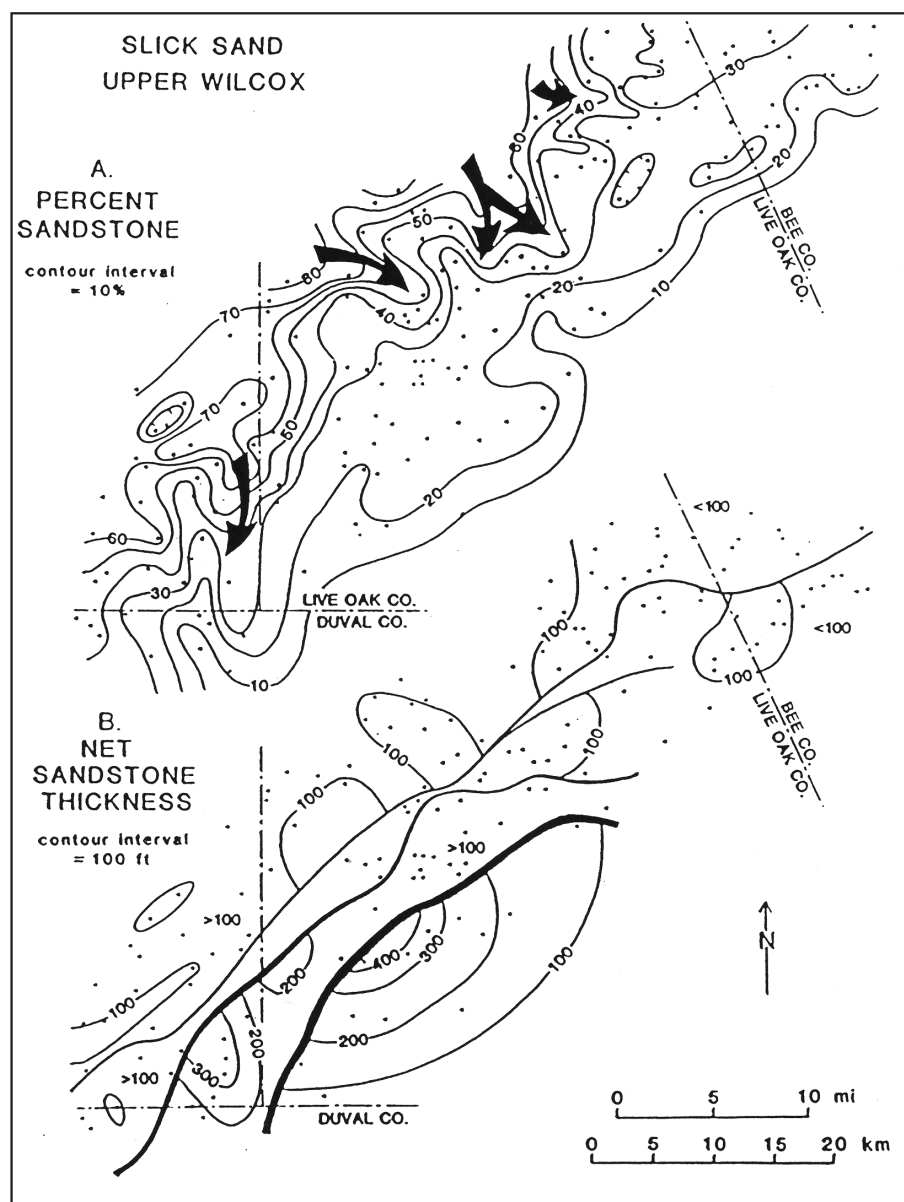


FIG. 25.—Percent-sandstone and net-sandstone thickness maps of the Eocene Slick Sand (a shelf edge delta), Texas Gulf Coast, U.S.A. Percent-sandstone map gives the best indication of the lobate nature of the delta. Growth faults (heavy lines on net-sandstone-thickness map) have a fundamental control of facies distribution, with thickening on the hanging wall. After Winker and Edwards (1983).

sparingly burrowed flood deposits, resulting in a highly irregular, or even cyclic, bioturbation index (MacEachern et al., 2005). Preserved organic matter is commonly high in river-dominated delta fronts, reflecting numerous phytodetrital pulses of deposition (MacEachern et al., 2005). Fresh-water or brackish-water influence may be reflected in the trace faunal assemblages (Moslow and Pemberton, 1988; Bhattacharya and Walker, 1991b; Gingras et al., 1998; MacEachern et al., 2005). Fluvial-dominated substrates produce the most stressful conditions for infauna. Salinity stress leads to conditions that can be exploited only by trophic generalists. Ichnological suites are dominated by low diversity and locally high abundance of generally diminished forms. High amounts of suspended sediment inhibits filter

feeders, such as are common in the *Skolithos* ichnofacies, and simple deposit feeders are more common. Buoyant plumes of sediment also create a sunlight stress, further reducing biogenic reworking.

Wave-Influenced Delta Fronts.—

Wave-influenced deltas commonly consist of a series of prograding beach and shoreface complexes, with sand fed from a nearby river (e.g., Rhone, Danube, Paraíba do Sul; Figs. 9, 17, 18). Many wave-influenced deltas show an asymmetry that results from oblique wave approach, with amalgamated sandy beach-ridge and shoreface deposits on the updrift side, and muddier

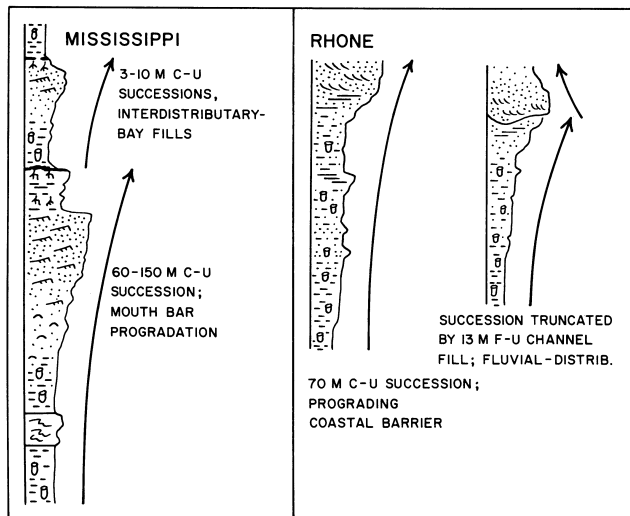


FIG. 26.—Typical coarsening-upward facies successions formed as a result of prograding deltaic lobes and mouth bars. Mississippi example shows a composite of a thicker mouth-bar succession below and the more irregular bay-fill successions above. (From Bhattacharya and Walker, 1992; after Elliott, 1986, and Coleman and Wright, 1975). Rhone example (left) shows progradation of a mouth bar forming an asymmetrical coarsening-upward succession, whereas Rhone example (right) shows truncation by a distributary channel, and has a more symmetrical profile. Rhone examples are from Bhattacharya and Walker (1992); modified by Elliott (1986) after Oomkens (1970).

facies on the downdrift side (Figs. 6, 9). Updrift areas are usually characterized by a relatively continuous coarsening-upward facies succession representing a wave-dominated shoreface (e.g., as in Fig. 28 wave-dominated column; Figs. 32, 33). The proportion of wave-produced structures (such as wave ripples and hummocky cross stratification) are greater updrift, whereas indicators of high sedimentation rates and fresh-water influence (e.g., soft-sediment deformation, climbing current ripples, brackish fauna) are fewer. Sandy sediment may be texturally more mature and better sorted than on the downdrift side, where fluvial influence is greater (Dominguez et al., 1987). Prodelta mudstones may be more bioturbated, thinner, and sandier than in river-dominated settings (Fig. 14D).

In the geological record, a single vertical facies succession of this type indicates a prograding wave-dominated shoreface. Good three-dimensional control may be necessary before such a shoreface can be positively ascribed to a delta. However, recent studies of many so-called “classic” shoreface successions are showing high ichnological stress in intervening mudstones, which may be a direct indicator of a brackish, fluvial plume nearby (e.g., Hampson and Howell, 2005).

In asymmetric wave-influenced deltas, the sandy mouth bar may be elongated downdrift to form a barrier island, such as seen in the Brazos, Danube, and São Francisco deltas. The barrier may temporarily inhibit delta progradation, because the back-barrier lagoon area acts as an important area for trapping of river-derived sediment. Vertical facies successions may appear more like the irregular river-dominated examples described above. River-borne mud is deposited in greater proportions in the downdrift than in the updrift areas (Fig. 32).

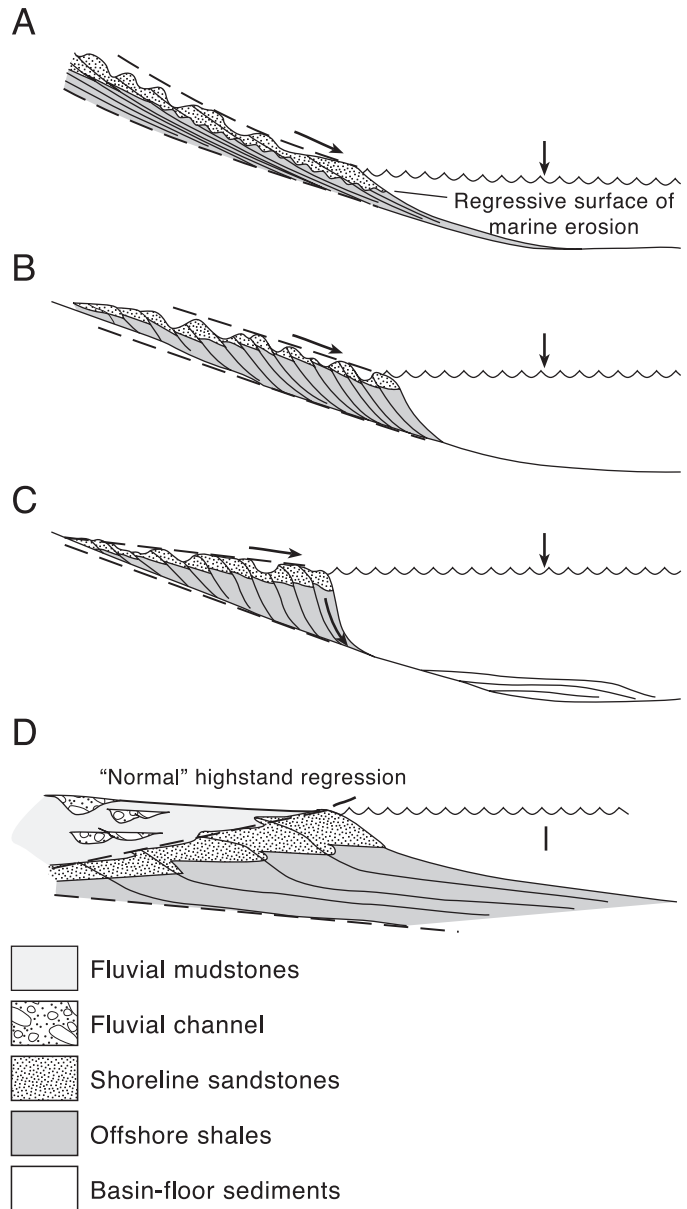


FIG. 27.—Examples of “forced” and “normal” regression (modified after Helland-Hansen and Gjelberg, 1994). **A**) Sharp-based shoreline deposits are produced when the trajectory of a falling shoreline is greater than sea-floor slope. **B, C**) Gradationally based deposits, are predicted when falling shoreline trajectory is equal to or less than sea-floor slope. **C**) Oversteepening can cause sediment gravity flows that are deposited on the basin floor. In all cases of forced regression (**B, C, D**) there is no subaerial accommodation and delta topset facies are thin to absent. Thin topset facies may easily be reworked or eroded during subsequent transgression. **D**) This contrasts with normal regression where shoreline trajectory is opposite of the basin slope. As a consequence, subaerial accommodation is positive and significant accumulation of delta topset facies (i.e., fluvial channels, mudstones) can occur. Thick paralic and nonmarine facies thus accumulate and are more likely to be preserved.

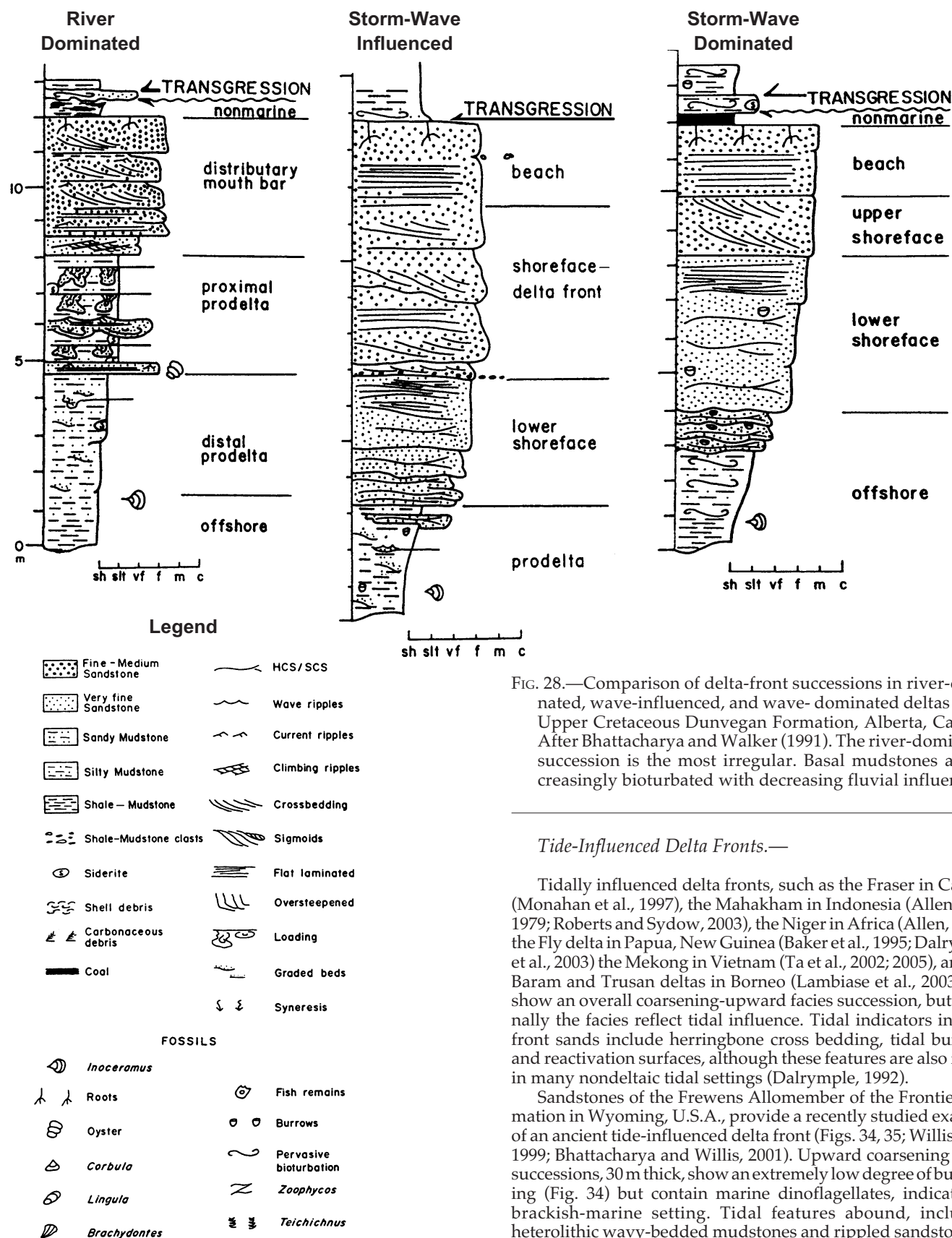


FIG. 28.—Comparison of delta-front successions in river-dominated, wave-influenced, and wave-dominated deltas in the Upper Cretaceous Dunvegan Formation, Alberta, Canada. After Bhattacharya and Walker (1991). The river-dominated succession is the most irregular. Basal mudstones are increasingly bioturbated with decreasing fluvial influence.

Tide-Influenced Delta Fronts.—

Tidally influenced delta fronts, such as the Fraser in Canada (Monahan et al., 1997), the Mahakham in Indonesia (Allen et al., 1979; Roberts and Sydow, 2003), the Niger in Africa (Allen, 1970), the Fly delta in Papua, New Guinea (Baker et al., 1995; Dalrymple et al., 2003) the Mekong in Vietnam (Ta et al., 2002; 2005), and the Baram and Trusan deltas in Borneo (Lambiase et al., 2003) also show an overall coarsening-upward facies succession, but internally the facies reflect tidal influence. Tidal indicators in delta front sands include herringbone cross bedding, tidal bundles, and reactivation surfaces, although these features are also found in many nondeltaic tidal settings (Dalrymple, 1992).

Sandstones of the Frewens Allomember of the Frontier Formation in Wyoming, U.S.A., provide a recently studied example of an ancient tide-influenced delta front (Figs. 34, 35; Willis et al., 1999; Bhattacharya and Willis, 2001). Upward coarsening facies successions, 30 m thick, show an extremely low degree of burrowing (Fig. 34) but contain marine dinoflagellates, indicating a brackish-marine setting. Tidal features abound, including heterolithic wavy-bedded mudstones and rippled sandstones at the base, passing into thicker cross-bedded sandstones with

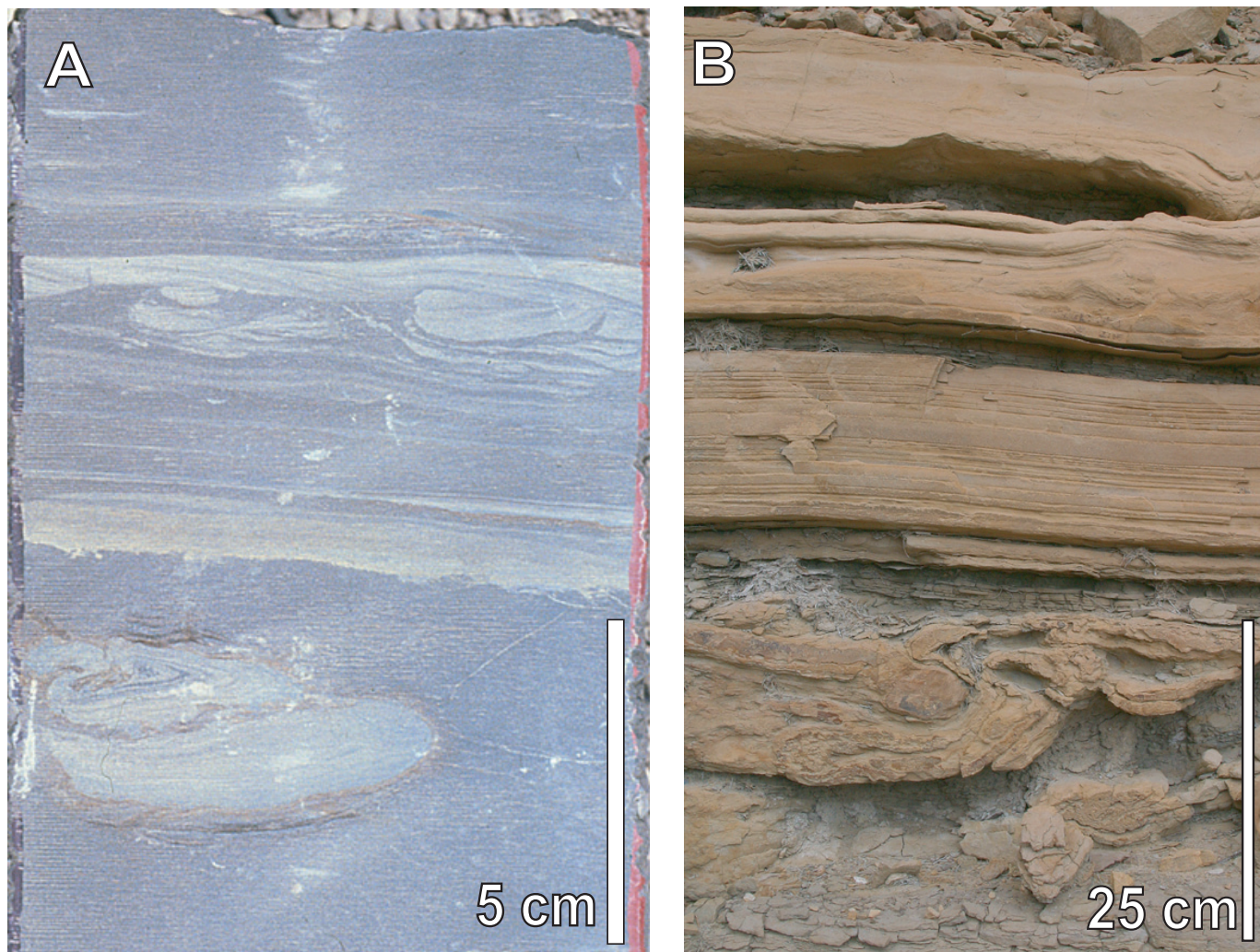


FIG. 29.—Deformation structures (load casts) in: **A)** prodelta mudstones of the Kavik Formation, Prudhoe Bay Field, Alaska, U.S.A. **B)** Deformed sandstone bed overlain by parallel-laminated to rippled delta front splays interpreted as distal delta front, sediment gravity flow deposits, Cretaceous Ferron sandstone, Utah, U.S.A.

abundant double mud drapes and reactivation surfaces (Fig. 35). Mudstones show abundant subaqueous shrinkage cracks, which may reflect salinity changes. However, these “syneresis” cracks are also associated with small-scale interstratal deformation, which may indicate diastasis rather than syneresis (Cowan and James, 1992).

The upward coarsening and low bioturbation indicates deltaic progradation, but the tidal features throughout indicate significant tidal modulation. The top of the Frewens sandstone is characterized by meter-thick sets of angle-of-repose cross beds, commonly floored by mud chips. Vertical cliffs expose seaward-dipping clinoforms (see Fig. 42) interpreted as seaward-migrating, tidally influenced mouth bars (Willis et al., 1999). In the case that only cores were available through this system, these features would look like erosionally based distributary channels. Paleocurrents in the Frewens are dominantly unidirectional, and strongly ebb-dominated, also suggesting tidal modulation of river flows.

Detailed work on the mixed tide- and wave-influenced Cretaceous Sego sandstone in the Book Cliffs of Utah, U.S.A. shows

numerous erosional features interpreted to have been produced by tidal scours (Willis and Gabel, 2001). The tops of tide-influenced deltaic successions are commonly reworked by tidal processes, producing deep tidal scours that might be mistaken for fluvial or distributary-channel erosion surfaces. The Mahakham delta contains 12 terminal distributary channels but over 20 distinct bars. Bars not fed by active channels are bounded by landward-narrowing tidal channels that scour up to 30 m deep (Fig. 17; Allen et al., 1979). Other ancient examples of tide-influenced deltas have been presented by Mutti et al. (1985), Maguregui and Tyler (1991); Nummedal and Riley (1999), and Jennette and Jones (1995).

Delta-Plain Successions

Distributary Channels.—

Distributary channels are erosionally based (Fig. 36). Filling commonly takes place after channel switching and lobe abandonment. At this time, the distributary channel may develop into an

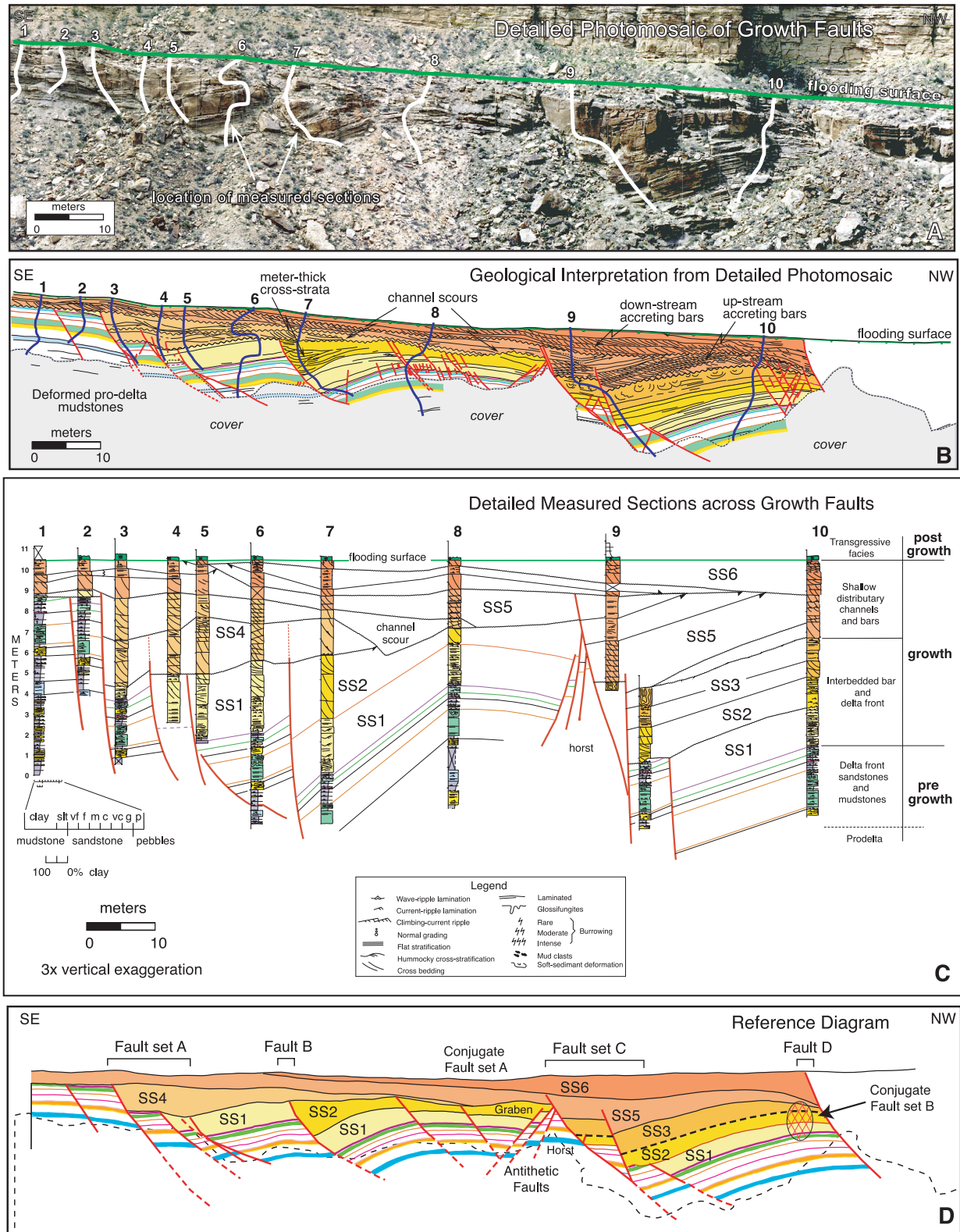


FIG. 30.—Cross section and interpretation of growth faults formed in prodelta and delta-front strata of the Cretaceous Ferron Sandstone exposed along Muddy Creek, Utah, U.S.A. **A)** Detailed photomosaic, **B)** geological interpretation of structure, **C)** detailed measured sections, and **D)** a reference diagram. The growth interval consists of upstream- and downstream-accreting cross-bedded sandstones deposited in shallow distributary channels and proximal distributary-mouth bars. Successive sandstones in the growth section are labeled SS1 to SS6 (from Bhattacharya and Davies, 2001, 2004).

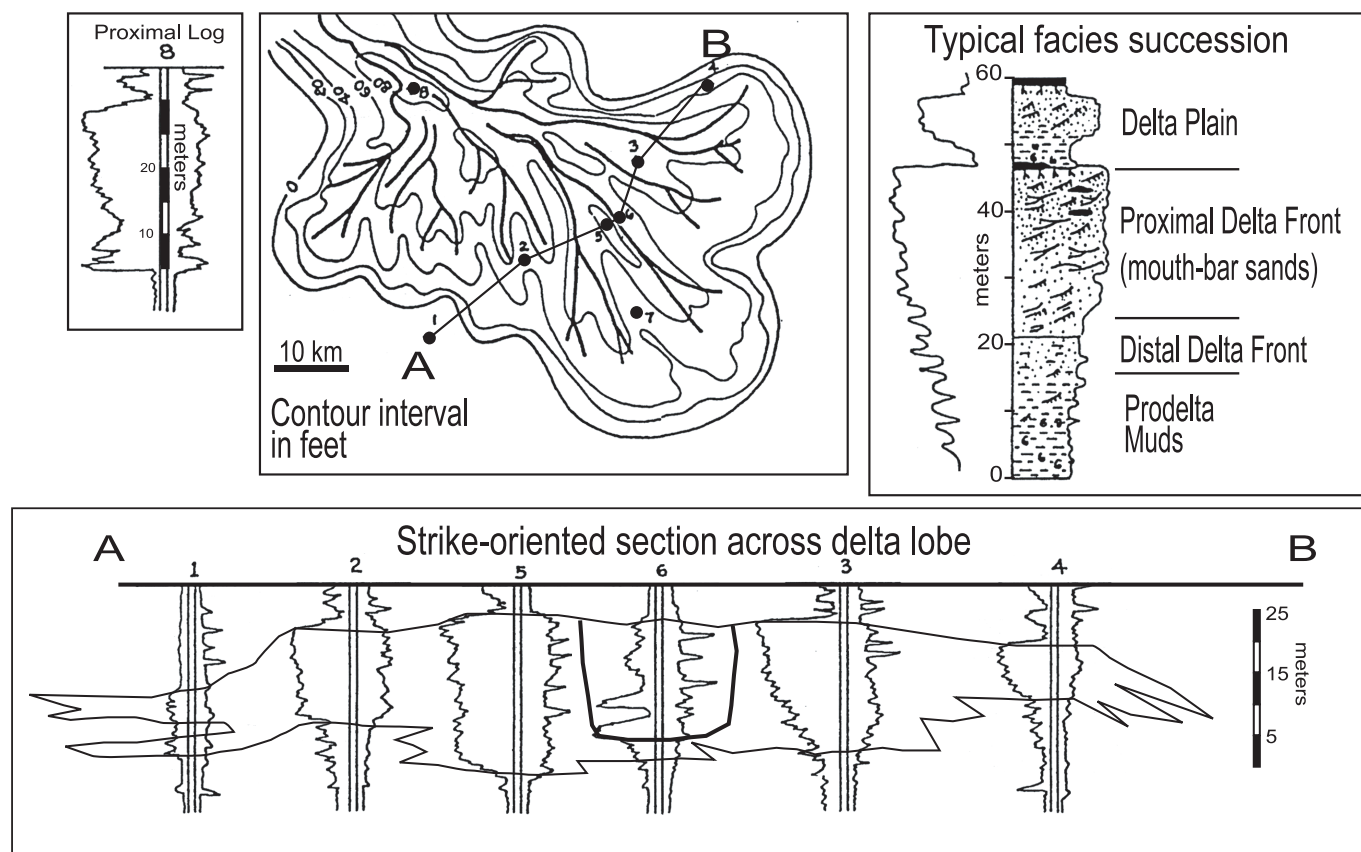


FIG. 31.—Cores and well logs from the modern Mississippi delta show a variety of upward-coarsening facies successions. Sharper-based, blockier-appearing log patterns (e.g., profiles 5 and 8) lie in the more proximal portions of the delta lobe. Interlobe succession (profile 1) is irregular. From Coleman and Prior (1982).

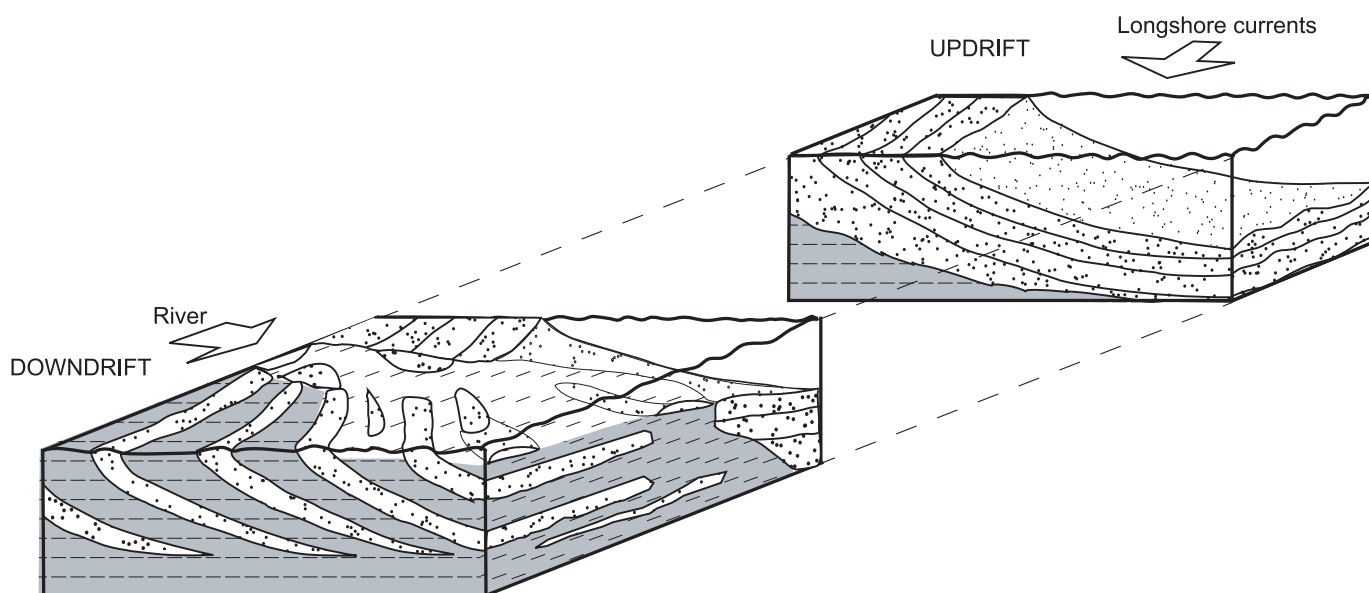


FIG. 32.—Block diagram illustrating the hypothesized three-dimensional facies architecture of an asymmetric delta. Significant prodelta mudstones are associated with downdrift portion of the delta where sandy barrier-bar complexes occur within lagoonal mudstones and bayhead-delta deposits. The updrift side of the delta comprises a sandy beach-ridge plain (from Bhattacharya and Giosan, 2003).

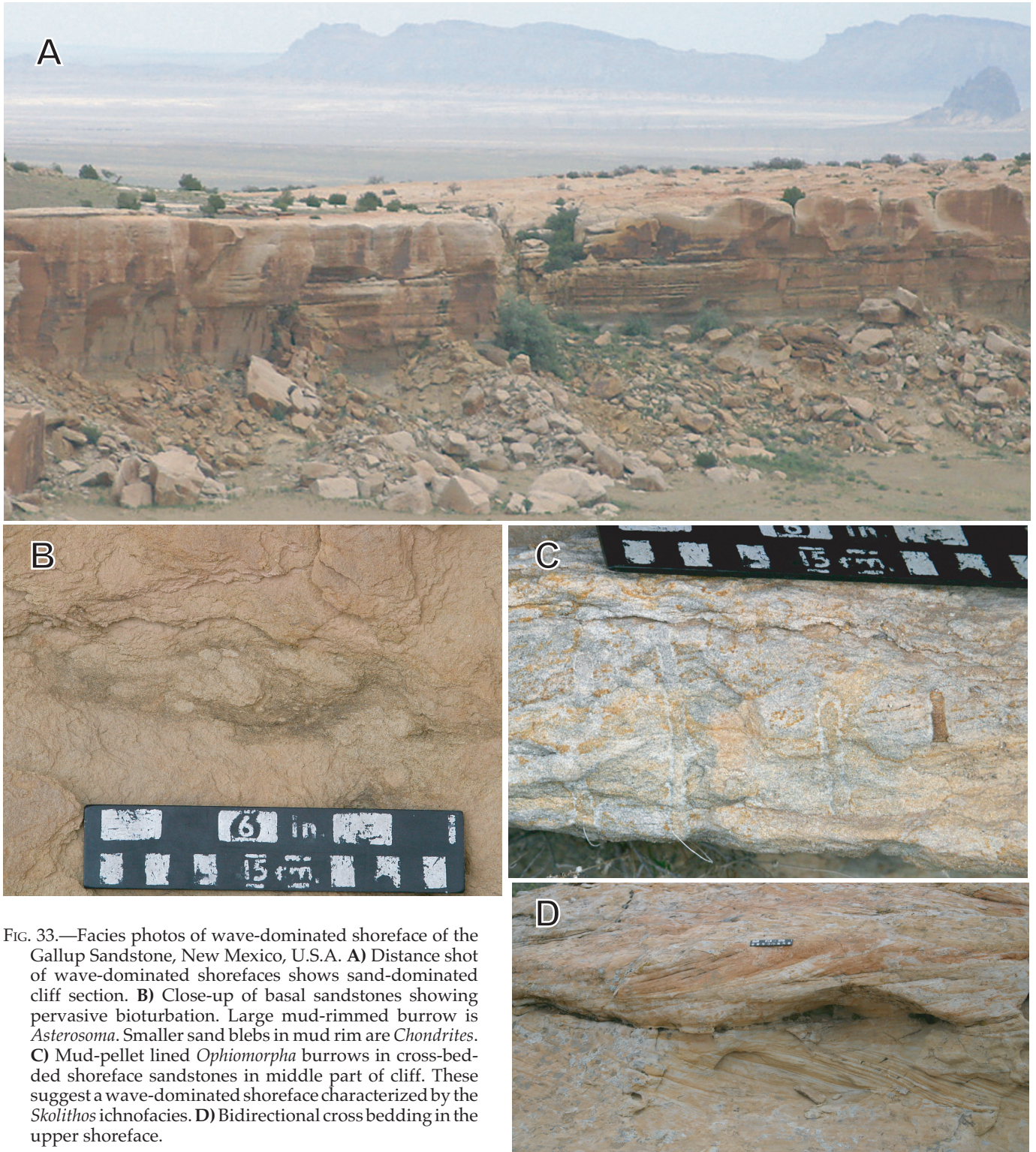


FIG. 33.—Facies photos of wave-dominated shoreface of the Gallup Sandstone, New Mexico, U.S.A. **A)** Distance shot of wave-dominated shorefaces shows sand-dominated cliff section. **B)** Close-up of basal sandstones showing pervasive bioturbation. Large mud-rimmed burrow is *Asterosoma*. Smaller sand blebs in mud rim are *Chondrites*. **C)** Mud-pellet lined *Ophiomorpha* burrows in cross-bedded shoreface sandstones in middle part of cliff. These suggest a wave-dominated shoreface characterized by the *Skolithos* ichnofacies. **D)** Bidirectional cross bedding in the upper shoreface.

estuary, and the fill is commonly transgressive with strong tidal indications. The facies succession tends to fine upward, with some preserved river-derived facies at the base, and a greater proportion of marine or brackish facies in the upper part of the channel fill. The extent of brackish to marine facies development

depends on the degree of river dominance and the position of the channel. Distributary channels within the nonmarine upper delta plain look entirely fluvial in nature (see Bridge, 2003, and Bridge, this volume), although they tend to be single-story rather than multi-story, compared to valley fills.

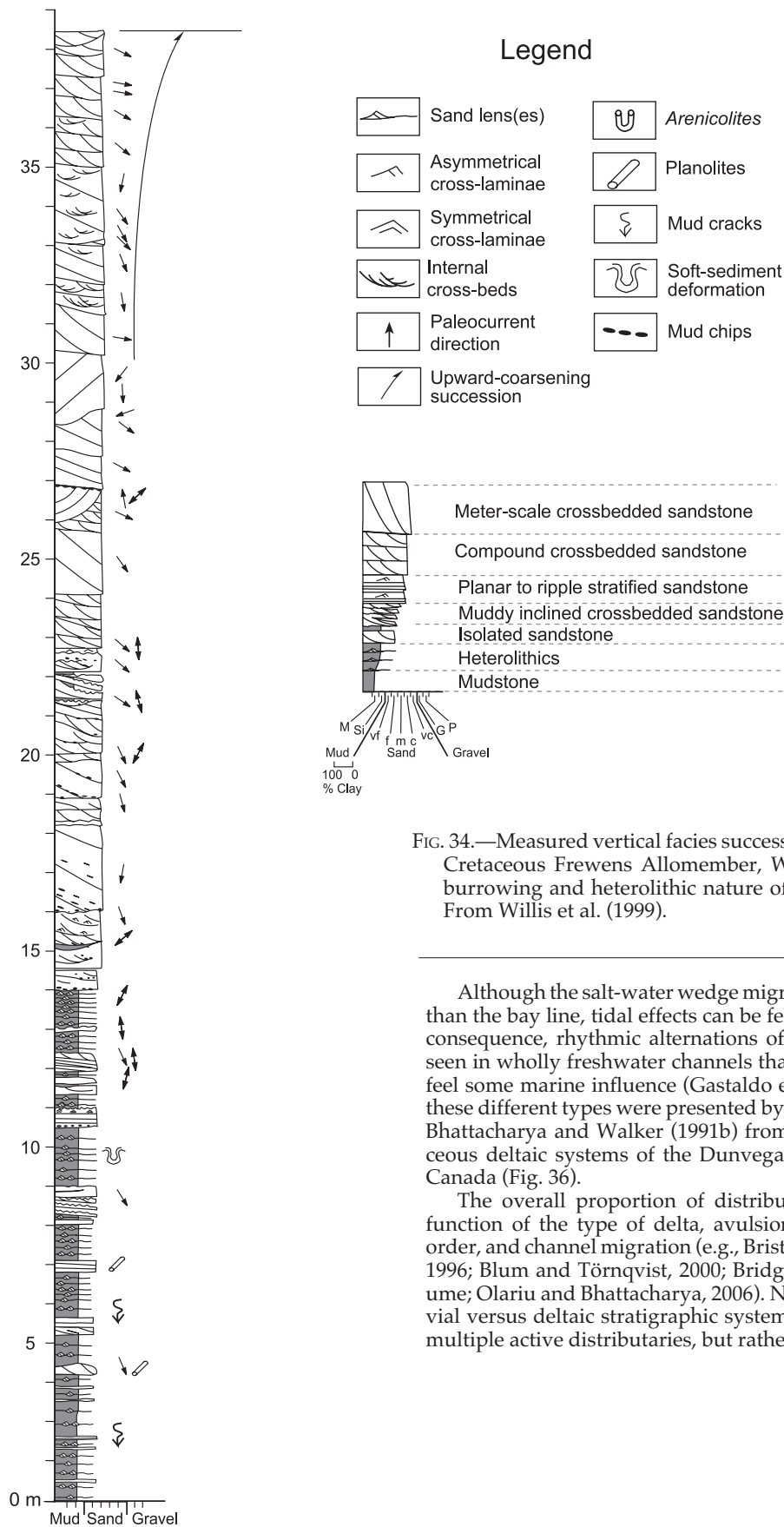


FIG. 34.—Measured vertical facies succession through tide-influenced delta front of the Cretaceous Frewens Allomember, Wyoming, U.S.A., emphasizes low degree of burrowing and heterolithic nature of interbedding. Photos of facies in Figure 35. From Willis et al. (1999).

Although the salt-water wedge migrates no farther landward than the bay line, tidal effects can be felt farther upstream. As a consequence, rhythmic alternations of mud and sand may be seen in wholly freshwater channels that nevertheless indirectly feel some marine influence (Gastaldo et al., 1995). Examples of these different types were presented by Bhattacharya (1989) and Bhattacharya and Walker (1991b) from distributaries in Cretaceous deltaic systems of the Dunvegan Formation in Alberta, Canada (Fig. 36).

The overall proportion of distributary-channel facies is a function of the type of delta, avulsion frequency, bifurcation order, and channel migration (e.g., Bristow and Best, 1993; Miall, 1996; Blum and Törnqvist, 2000; Bridge, 2003; Bridge, this volume; Olariu and Bhattacharya, 2006). Numerical models of alluvial versus deltaic stratigraphic systems do not typically allow multiple active distributaries, but rather model a single channel

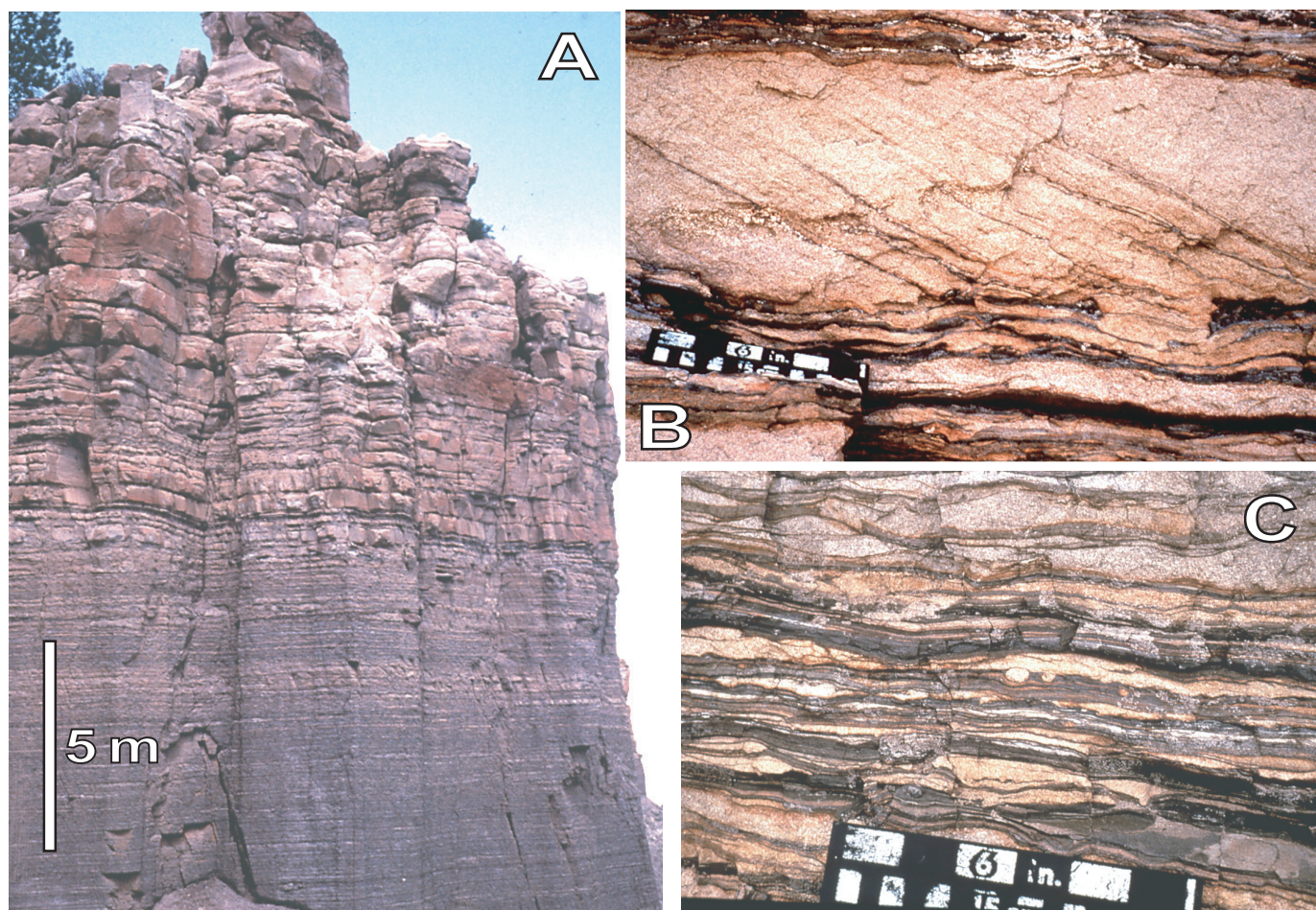


FIG. 35.—Tide-dominated delta front of the Cretaceous Frewens sandstone, Wyoming, U.S.A. **A)** Upward-coarsening facies succession (see measured section in Figure 34). **B)** Double mud drapes indicative of tidal modulation. **C)** Heterolithic, lightly burrowed subtidal prodelta facies at the base of the succession. Bedding architecture is shown in Figure 42.

that avulses and migrates (e.g., Paola, 2000; Mackey and Bridge, 1995; Overeem and Weltje, 2001; Overeem et al., 2005; Olariu and Bhattacharya, 2006). Some of the fundamental characteristics of distributive channel systems, particularly downstream decreases in discharge, channel width, and channel depth, are thus not predicted by present numerical models, and the increase in bedload-related form friction is not generally accounted for (Giosan and Bhattacharya, 2005; Overeem et al., 2005).

In general, the more wave-dominated the delta, the greater is the proportion of lobe sediment, with more limited amounts of interlobe and distributary-channel facies. It is also emphasized that there is no such thing as “a distributary channel” of a unique width or depth. Several scales of distributary channel may occur within any given delta (Fig. 18).

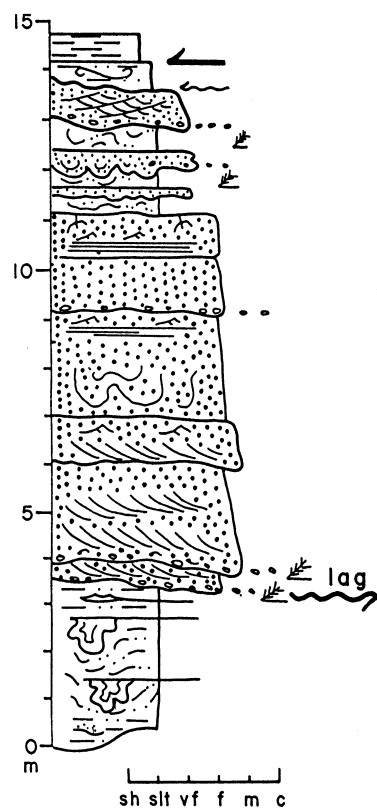
In general, valley fills are much thicker than associated delta-front successions and consist of multi-story sandstones (Figs. 19, 37; Reynolds, 1999). Plan-view maps of valleys should show a tributive rather than a distributive pattern (Fig. 38; e.g., Plint and Wadsworth, 2003). While these may be difficult to map in outcrops or in sparse subsurface data sets using 1D log or 2D seismic data, horizon mapping through 3D seismic cubes has revealed complex dendritic drainage patterns within buried paleovalleys (e.g., Brown, 2005). Also, valley fills do not inter-

finger with adjacent floodplain or delta-plain facies but exhibit an erosional relationship. This may be more difficult to observe in well log data, and large valleys may internally contain flood-plain or delta-plain facies. Interfluvial paleosols may also provide key evidence of sediment bypass, floodplain starvation, and avulsion frequency (McCarthy, 1999; Kraus, 2002).

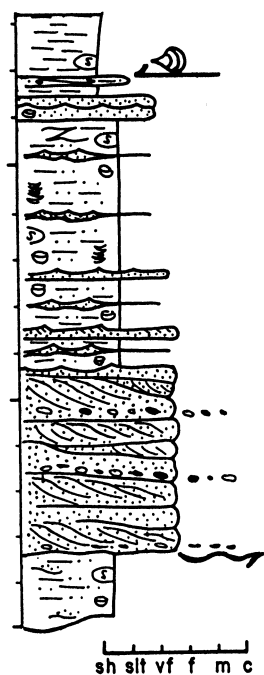
Interdistributary Areas.—

Interdistributary and interlobe areas are less sandy, and commonly contain a series of relatively thin, stacked coarsening- and fining-upward facies successions (Figs. 31, 39). These are usually less than ten meters thick, and they do not show as thick or as well-developed coarsening-upwards facies successions as are found in prograding deltaic lobes (Elliott, 1974; Tye and Coleman, 1989). The proportion of lobe versus interlobe successions depends on the nature and type of delta system, the stability of distributary channels, and the amount of nonmarine accommodation. Wave-influenced systems, like the Danube, can contain significant lagoonal and bay mudstones in regions downdrift of the river mouth and, depending on shoreline trajectory, thick accumulations of mud-prone paralic and nonmarine facies can accumulate behind an aggrading shoreface or delta front (Fig. 27D).

River - Dominated



Marine - Dominated



Interdistributary Areas in River-Dominated Deltas.—

An interdistributary bay is filled by overbank spilling of fine-grained material from the river during flood stages (Fig. 8). There is an overall shallowing-upward facies succession, associated with a trend from more marine to more nonmarine facies, but commonly without the deposition of thick sands (Fig. 39). This represents the transition from offshore prodelta mudstones into delta-top facies without the development of a sandy shoreline (Walker and Harms, 1971; Bhattacharya and Walker, 1991b). The muddy nature of interdistributary-bay successions may be punctuated by sandy crevasse-splay or channel deposits that may produce thin coarsening or fining successions (Coleman and Prior, 1982; Elliott, 1974). The succession may grade into rooted coaly mudstones or coals representing a variety of swamp, marsh, and lacustrine environments. Beach sands, associated with the development of barrier strandplains, spits, or cheniers, may be present at the tops of these successions, although they are relatively thin compared to the delta front. Interlobe areas may also act as the locus for progradation of a subsequent lobe and may be erosionally truncated by younger distributary channels.

FIG. 36.—Comparison of distributary-channel-fill successions in river- and marine-dominated deltas of the Dunvegan Formation (Cretaceous, Alberta, Canada). In the marine-dominated system, the distributary fill reflects transformation of the distributary into an estuary. After Bhattacharya and Walker (1991b). Legend in Figure 28.

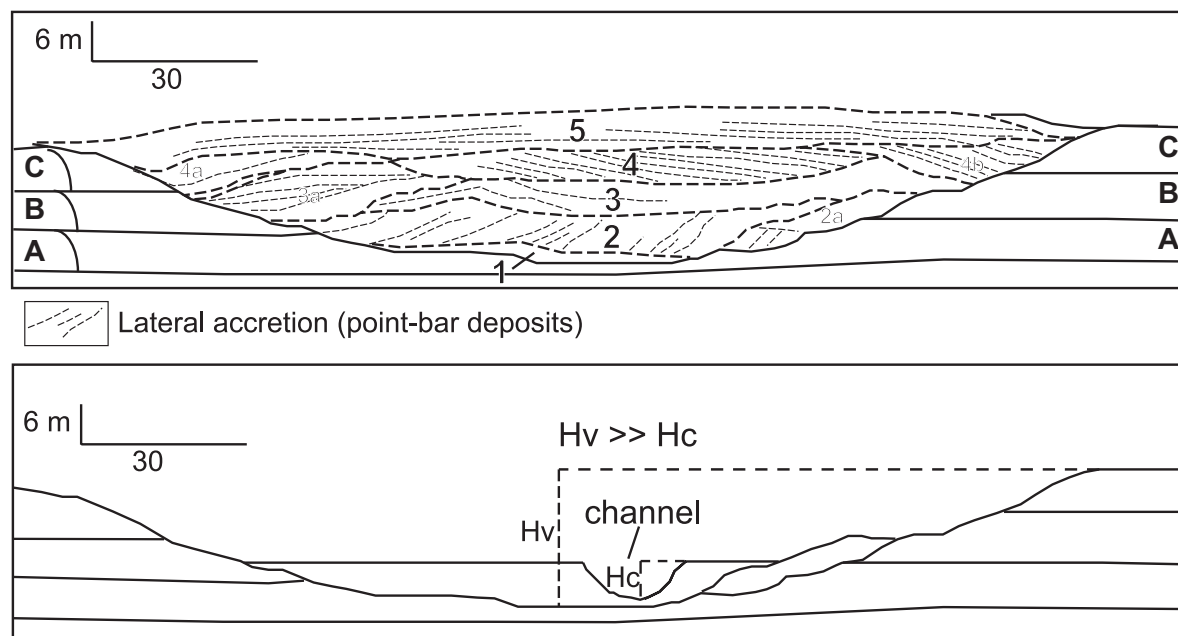


FIG. 37.—Facies architecture of an interpreted valley fill in the Cretaceous Ferron Sandstone member of the Mancos Shale, Utah, U.S.A. Base of valley erodes into several upward-coarsening parasequences A, B, C. Valley depth (H_v) is about 21 meters. In contrast associated channel depths (H_c) are only about 6 m. Valley is filled with 5 channel stories (1–5). Lowest channel-belt deposit (1) is largely eroded by migration of younger channels. Predominance of laterally accreting bars defines the internal facies architecture of each channel-belt deposit. The bedding geometry shows that the rivers were single-thread, meandering streams that gradually filled the larger valley. Calculation of water depth from dune-scale cross strata within the bar deposits suggest maximum bankfull depth of about 9 m. From Bhattacharya and Tye (2004), modified after Barton et al. (2004).

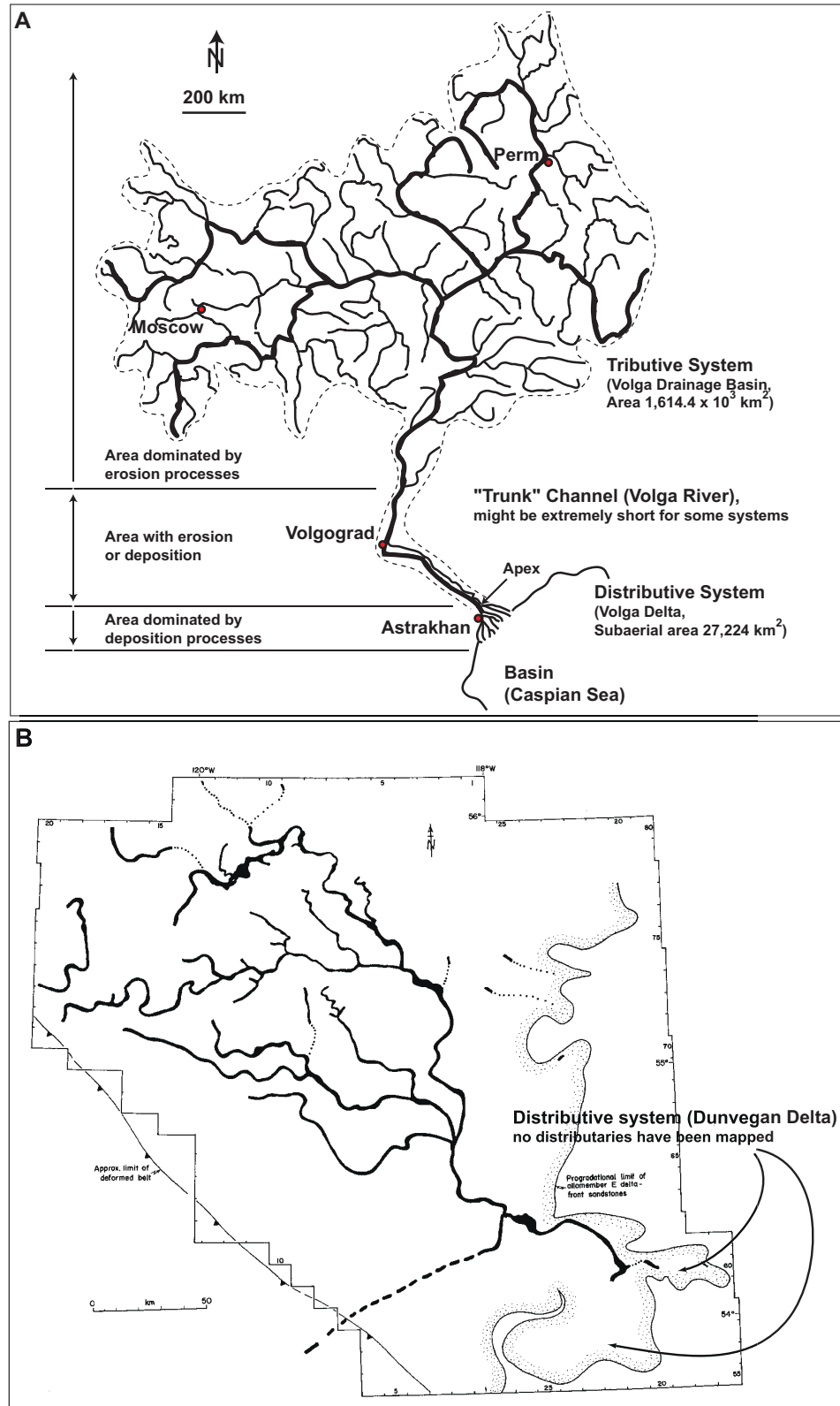


FIG. 38.—**A**) Example of a tributive–distributive system, Volga basin. The tributive pattern is an order of magnitude larger (tens to hundreds of times) than the distributive pattern, and the main “trunk” valley connects the two patterns. Modified after Payne et al., 1975. **B**) Tributive–trunk system in Dunvegan lacks details of distributive pattern because distributary channels are too small to image. (From Flint and Wadsworth, 2003).

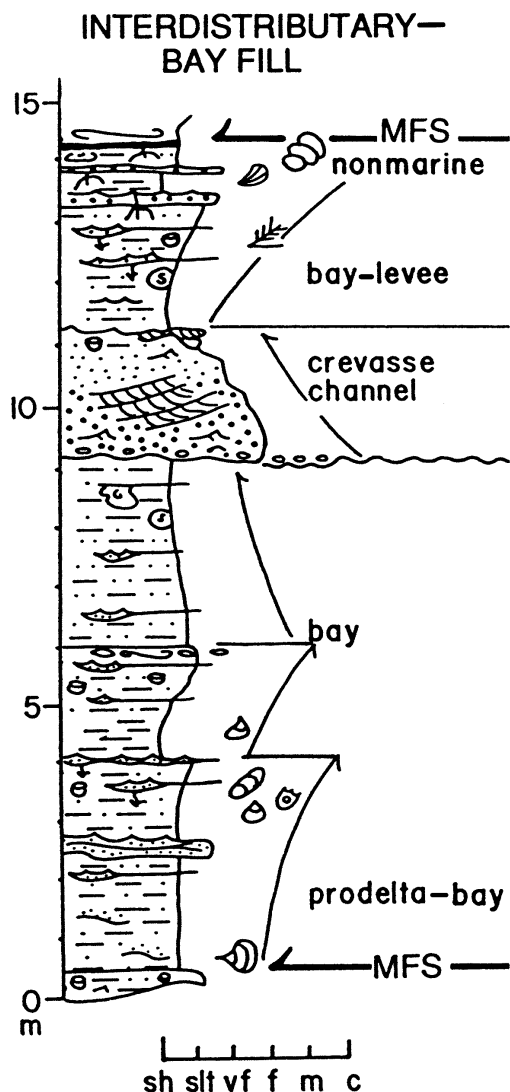


FIG. 39.—Interdistributary-bay fill in a river-dominated delta lobe in the Dunvegan Formation (Cretaceous, Alberta, Canada), showing thin irregular cycles and overall increase in proportion of nonmarine facies upwards. Compare with profile 1 in Figure 31. Legend in Fig. 28. After Bhattacharya and Walker (1991).

Interdistributary Areas in Wave-influenced Deltas.—

Interdistributary bays may often be closed off by barrier-beach complexes in wave-dominated deltas resulting in extensive back-barrier lagoons — e.g., the Nile (Fig. 1), the São Francisco, the Danube (Fig. 9), the Brazos, and the Po. These may be filled from the landward side by progradation of bayhead deltas or from the barrier side by storm washovers (Bhattacharya and Giosan, 2003) and have been regarded as local estuaries within a larger delta system. Deposits are commonly organic-rich, with marsh vegetation or mangroves. These areas are typically more pronounced on the downdrift sides of highly asymmetric deltas (Bhattacharya and Giosan, 2003). However, identification of estuarine-type facies does not necessarily mean deposition within a valley (e.g., MacEachern et al., 1998).

Interdistributary Areas in Tide-influenced Deltas.—

Tidal processes may be important in interdistributary bays (even in river-dominated deltas) resulting in tidally influenced facies such as tidal flats or tidal channels (Allen et al., 1979; Ramos and Galloway, 1990). These are especially common in modern tidally influenced deltas such as the Niger, Fraser, Mahakham, Fly, Mekong, Ganges–Brahmaputra, Ayerawady, and Orinoco deltas. Ancient examples of tidally influenced facies in delta-plain settings include those published by Ramos and Galloway (1990), Eriksson (1979), and Rahmani (1988). Ebb versus flood tidal currents may move down different pathways, such that current directions are unidirectional at any one place but differ between ebb-dominated versus flood-dominated channels (Harris, 1988; Dalrymple et al., 2003). In tide-dominated deltas, mud is partitioned along the sides of the system as well as in the prodelta area (Fig. 20; Dalrymple, 1992; Willis et al., 1999; Willis, 2005).

FACIES ARCHITECTURE OF DELTAS

Bedding geometry and lateral facies variability can be addressed by the use of seismic data (e.g., Hart and Long, 1996; Anderson et al., 2004), ground-penetrating radar (Jol and Smith, 1991, 1992; Smith et al., 2005; Lee et al., 2005), continuous outcrop data (e.g., Willis et al., 1999; Soria et al., 2003; Gani and Bhattacharya, 2005; numerous papers in Chidsey et al., 2004), and interpolation of well data (e.g., Bhattacharya, 1991, 1993, 1994; Ainsworth et al., 1999; Tye et al., 1999; Plint, 2000; Bhattacharya and Willis, 2001). Facies architectural studies of deltas significantly lag behind those of fluvial, deep-water, and eolian systems, but more recent studies of deltaic systems are becoming available (Willis et al., 1999; Knox and Barton, 1999; Willis and Gabel, 2001; Olariu et al., 2005).

Although it may be premature to fully characterize the architectural elements that make up deltaic depositional systems, some generalizations can be made. Sandy architectural elements in the delta plain include channels at various scales, which may migrate or stack to form channel bodies or channel belts. Internally channel bodies consist of bars (macroforms) and smaller-scale bedforms, analogous to the architectural elements described in the fluvial literature (Miall, 1995, 1997; Bridge, 2003; this volume). The number of different scales of channels relates to the bifurcation order, which can be high in river-dominated deltas and low in wave-dominated settings. Unfortunately, bifurcation order is very difficult to determine in outcrop or subsurface examples, although this may be possible in selected settings or particularly good outcrop exposures (e.g., Bhattacharya and Tye, 2004). Areas away from distributary channels may include crevasse splays and levee deposits. The delta plain also consists of numerous mud-prone wetland environments, although there have been few studies that compile the typical dimensions of the associated muddy facies elements.

The distal delta plain and proximal delta front consist of mouth-bar elements, which in turn build bar assemblages and form depositional lobes. There may be several scales of bar assemblage and lobe clustering, especially in continental-scale river-dominated delta systems like the Mississippi. Mouth bars are in turn intimately associated with terminal distributary channels. A variety of sandy bedforms may be associated with the upstream sides of these bars. In river-dominated, shallow-water, friction-dominated deltas, these channels are typically only a few meters in depth and a few tens to a few hundred meters wide. The distal margins of bars are commonly formed by frontal splay elements or subaqueous channels and chutes. Channels typically scour only a few meters and may be intimately associated with

frontal splays. Dimensions of splays are largely unknown, although individual beds should scale to the generative flow. These frontal splays may coalesce to form a fringe of distal delta-front sand, which scales to the size of the depositional lobe.

Wave-formed architectural elements include barrier-island sand bodies and shorefaces. Width of barrier islands typically scales to the width of the initial mouth bar, although they may extend for several kilometers down-drift. Shorefaces can reach several to tens of kilometers in width and several tens of kilometers in length. In many wave-dominated coastlines, the area occupied by shoreface “wings” can greatly exceed the area occupied by the fluvial-dominated mouth bar (e.g., Paraíba do Sul, Fig. 18B). This has led many to question the value of calling river-fed shorefaces, deltas at all (Dominguez, 1996). Recent studies of the shoreface successions in the Book Cliffs, Utah, U.S.A., suggests that rivers were widely spaced, up to 50 kilometers apart, along strike (Hampson and Howell, 2005). The shorefaces are characterized by river-plume deposits that effectively “puncture” the otherwise rather uniform shoreface sand body. These studies predict that shoreface sandstones typically extend several tens of kilometers along strike before they are punctured. This invites the question: are there any shorefaces that truly extend,

unbroken, for hundreds of kilometers along depositional strike? The position of river plumes in these outcrops is indicated by the appearance of steep clinoform strata, a marked decrease in diversity and abundance of ichnofauna in underlying mudstones, and the appearance of laminated to graded prodelta facies (Hampson and Howell, 2005).

Tidal reworking may produce a bewildering variety of tidal bedforms and bars (see Dalrymple, 1992; Boyd et al., this volume; Willis, 2005). Tidal processes also result in greater winnowing of distributary channels, which may be stable for considerably longer periods than terminal distributary channels in river-dominated environments. This can result in significantly more elongate bars and bar assemblages than in nontidal settings.

Muddy elements associated with the subaqueous realm include prodelta muds, bay muds, and bar drapes. Prodelta muds may cover vast areas of the shelf and may migrate for thousands of kilometers along strike. Although muddy wave-formed clastic coastlines are common in the modern, there is a paucity of well documented ancient examples such as cheniers. Tidal flats are significantly better recognized but are largely discussed in the context of tidal depositional systems.

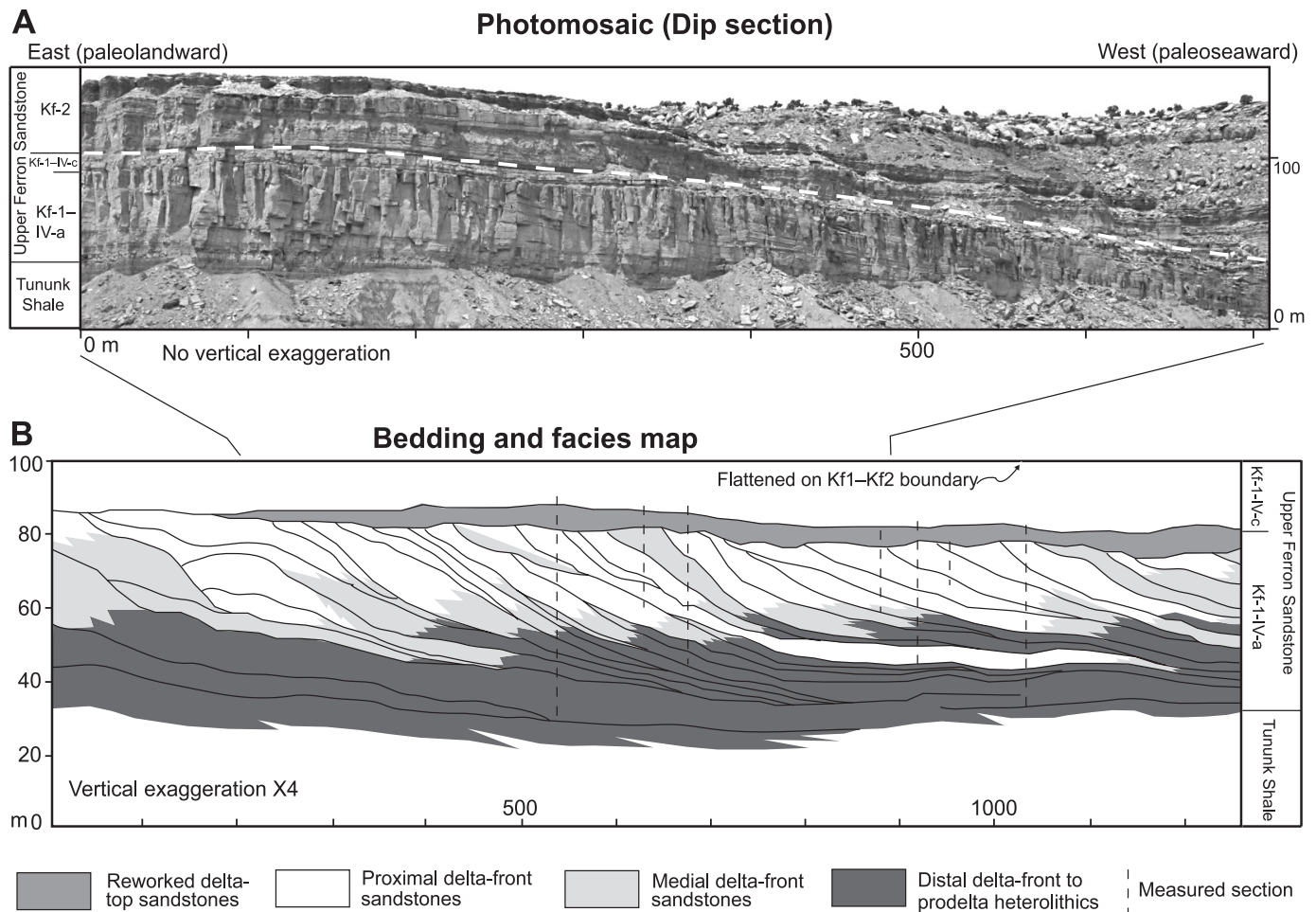


FIG. 40.—Inclined bedding (clinoforms) and facies in a river-dominated delta front of the Cretaceous Ferron sandstone member, Utah, U.S.A. **A**) Photomosaic of a cliff face. **B**) Bedding and facies geometry of the same cliff face (along depositional dip), Ivie Creek amphitheater, Emery County, Utah. The diagram shows prominent seaward-dipping clinoforms. From Gani and Bhattacharya (2005), modified after Anderson et al. (2004) and Mattson (1997).

A comprehensive compilation of the sizes and dimensions of these various architectural elements is simply beyond the scope of this review, and in this regard architectural-element analysis of deltaic systems remains significantly behind the deep-water, fluvial, and eolian systems described in other chapters in this volume. Many of the elements described in these other systems are also found in deltas (e.g., channels, bars, shorefaces, barriers, tidal bars, cheniers), but clearly a great deal of compilation is required to determine whether the dimensions of these elements is fundamentally different if they are associated with a delta system. Tye (2004) compiled data on the dimensions of mouth-bar sandstones in several arctic deltas, as well as the Atchafalaya delta in the Gulf of Mexico, and Reynolds (1999) compiled data on dimensions of a variety of paralic sandstone bodies, including mouth bars and distributary channels.

Dip Variability

In cross-sectional dip view, deltas can be divided into three distinct regions: topset, foreset, and bottomset (Figs. 2, 3, 40). Foresets are typically associated with the distal delta front and show dips that can range from a few degrees up to the angle of repose in Gilbert-type deltas (Fig. 41). Bottomsets dip less steeply than foresets, typically $< 1^\circ$, and are usually associated with prodelta sediments. Topset facies are typically flat to undulating and are built out of the proximal delta-front and delta-plain facies. Subaqueous topset facies (i.e., proximal delta front) can show extreme variability, depending on whether they are constructed from mouth bars, tidal bars, or shoreface deposits. Shoreface elements produce the simplest bedding geometries (e.g., Hampson, 2000) and essentially consist of seaward-dipping beds, with dip angles typically less than 1° . Mouth bars and tidal bars produce far more complex deposits (Fig. 42). The seaward migration of these elements builds the vertical facies successions detailed above Fig. 34).

Deltaic deposits are characterized by a prograding clinoform geometry (Figs. 40–46; (Gilbert, 1885; Barrell, 1912; Rich, 1951; Scruton, 1960; Berg, 1982). This geometry can be seen in downdip seismic profiles of modern and ancient deltas (see Figures 43 and 44). Many superb examples are given in a recent volume on Quaternary deltas of the Gulf of Mexico (Anderson and Fillon, 2004), as well as numerous examples from Southeast Asian deltas presented in Sidi et al. (2003). This clinoform geometry can also be reconstructed in core and well-log cross sections (Figs. 45, 46; Bhattacharya, 1991; Ainsworth et al., 1999; Plint, 2000). It can also be seen in some outcrops (Fig. 40–42); Chidsey et al. (2004); Gani and Bhattacharya, 2005). Berg (1982) discussed typical seismic facies in deltaic depositional systems and suggested that sandy wave-dominated systems are characterized by a shingled pattern, whereas muddier deltas show an oblique-sigmoidal pattern. Sigmoid-shaped portions are characteristic of the mud-dominated prodelta facies (Kuehl et al., 1997; Liu et al., 2002; Roberts and Sydow, 2003; Hiscott, 2003; Anderson et al., 2004; Roberts et al., 2004; Neill and Allison, 2005; Kuehl et al., 2005), whereas the more flat-lying or oblique reflectors represent the sandier delta-front and delta plain facies. Frazier (1974) showed a similar clinoform geometry on the basis of geological studies of the Mississippi delta plain.

Offlapping clinoformal geometries have also been recognized in Late Quaternary deltas around the world (Brown and Fisher, 1977; Suter and Berryhill, 1985; Tesson et al., 1993; Sydow and Roberts, 1994; Hart and Long, 1996; Hiscott, 2003; Roberts and Sydow, 2003; Roberts et al., 2004; Bart and Anderson, 2004; and others in Anderson and Fillon, 2004). In shelf-edge systems, clinoforms commonly steepen towards the shelf edge. The steepening reflects the progradation into progressively deeper water. At the shelf edge, the delta front can no longer build seaward, so it steepens and then fails (e.g., Fig. 24).

Climoform gradients have a wide range of values in different settings. Clinoform gradients of shelf-edge deltas in the Gulf of

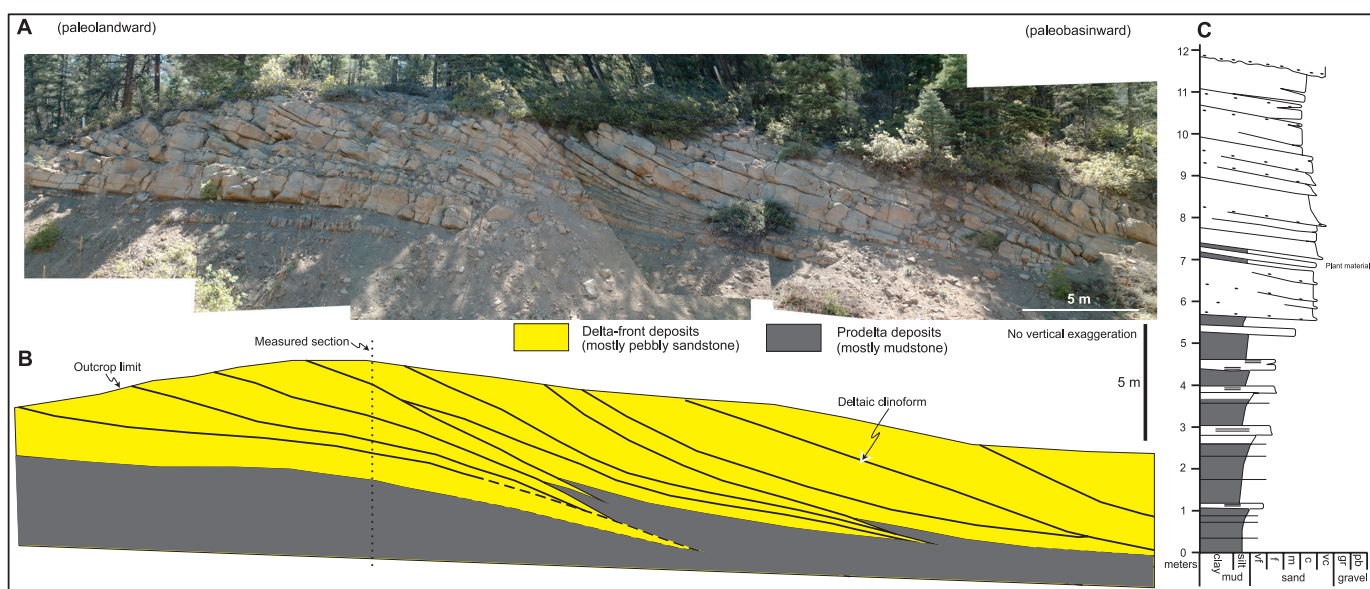


FIG. 41.—Details of facies interfingering at the base of a small-scale outcrop example of a gravelly, Pennsylvanian “Gilbert” delta, Taos Trough, New Mexico, U.S.A. **A)** Outcrop photomosaic. **B)** Line drawings of beddings with facies interpretation. Note that clinoforms are steeply dipping (average 13°). **C)** Lithologic column of this coarse-grained delta (position of the measured section is shown in Part B). From Gani and Bhattacharya (2005).

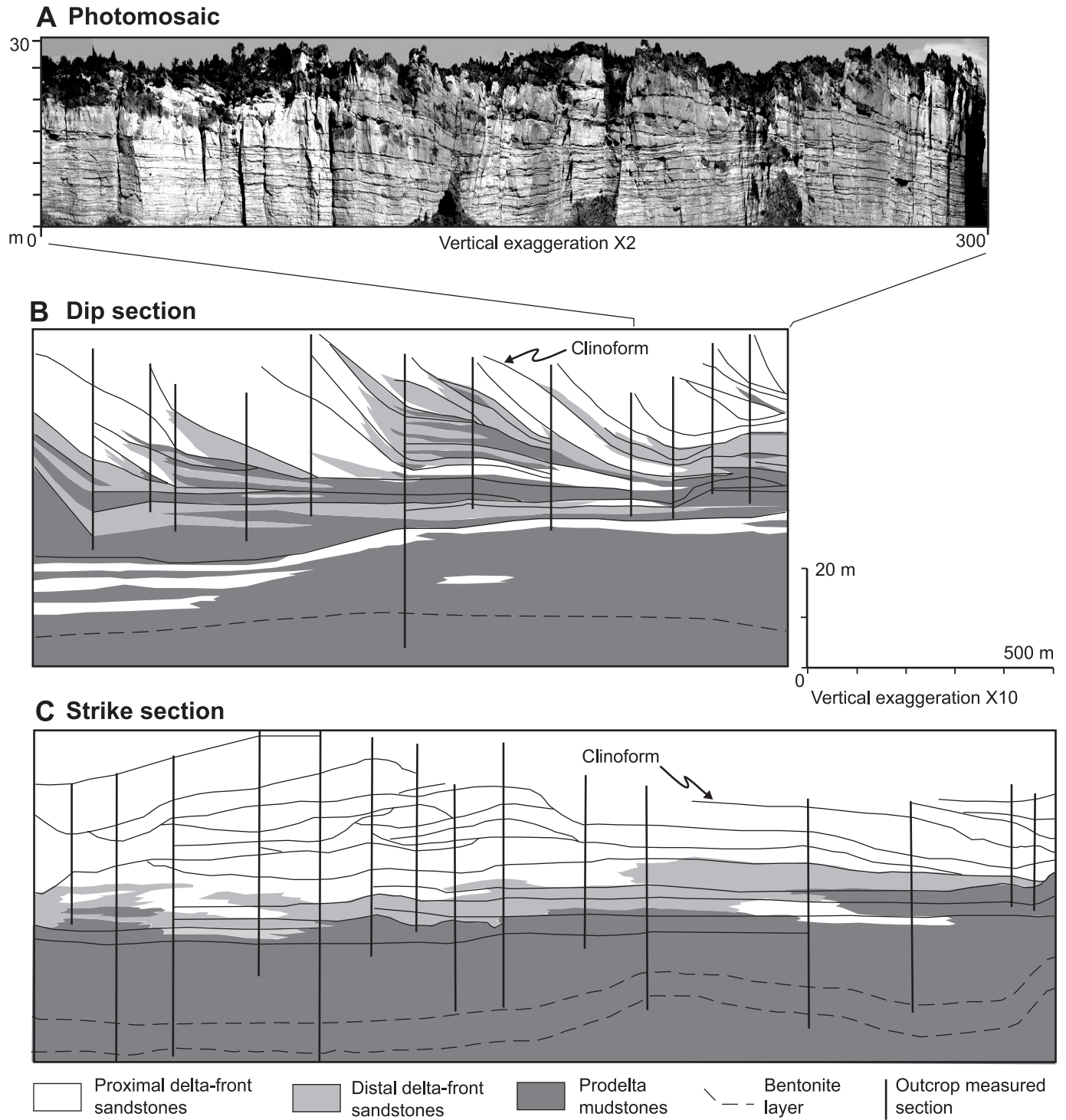


FIG. 42.—Outcrop example of complex internal architecture in the Cenomanian (Upper Cretaceous) tide-influenced river delta of the Frewens Allomember, Frontier Formation, central Wyoming, U.S.A. Dip view (AB) of the prograding delta shows the seaward-dipping clinoforms, whereas in strike view (BC) these clinoforms show bidirectional downlap, forming a classical lens-shaped geometry. In both cases, muddy bottomset facies interfinger with the sandy foreset facies forming a shazam-type facies boundary. Note that clinoform dip varies from 5° to 15° . Detailed facies shots and measured sections are shown in Figures 34 and 35 (modified from Willis et al., 1999).

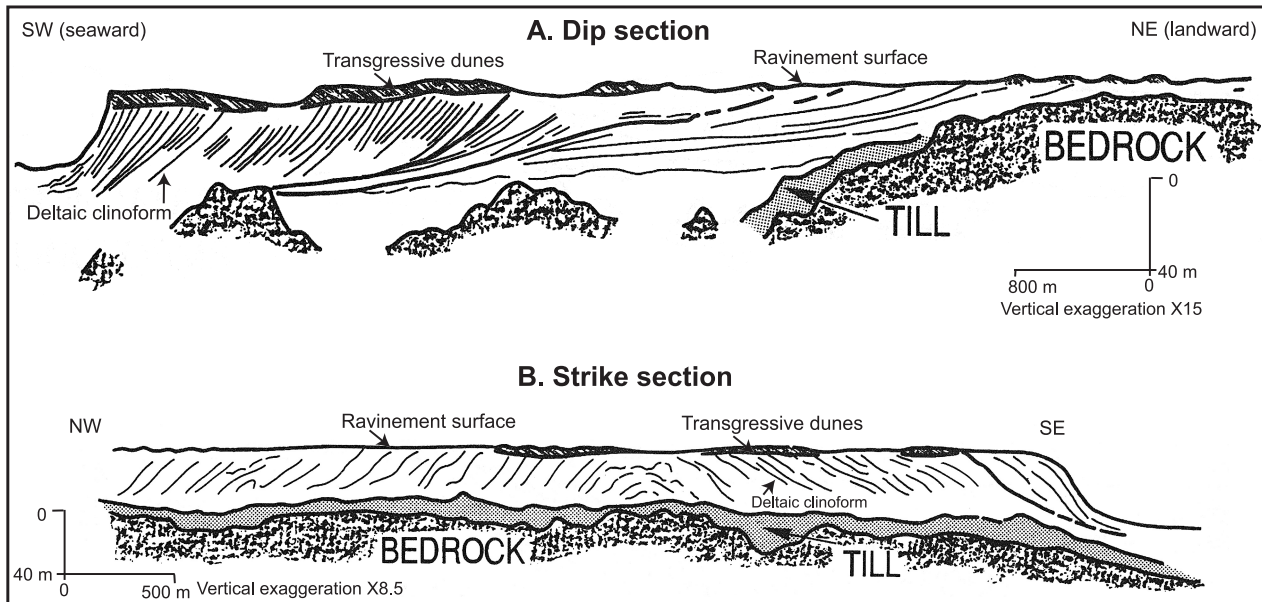


FIG. 43.—**A**) Dip-oriented and **B**) strike-oriented views showing bedding geometry of a top-truncated, lowstand delta, based on shallow seismic profiles off the Natashquan River, Gulf of St. Lawrence, Canada (after Hart and Long, 1996). Note reworked sediments on top of deltas.

Mexico average between 4° and 8° (Suter and Berryhill, 1985). Gradients of the Rhone shelf-edge deltas average about 1° . The high slopes in the Gulf Coast and Rhone shelves result in significant instability of the shelf-edge sediments, where large-scale synsedimentary deformation features are common. Slopes are typically much lower in ramp-type depositional margins, and soft-sediment deformation features in the Alberta examples are mostly limited to loading rather than large-scale slumps, slides, or growth faults. Foreset dips of subaqueous prodelta clinoforms are typically less than 0.1° (e.g., Liu et al., 2002; Neill and Allison, 2005) as opposed to sandy foresets, which are usually an order of magnitude higher (i.e., $> 1^{\circ}$).

Strike Variability

Along strike, facies relationships may be less predictable and depositional surfaces may dip in different directions (Figs. 42, 43B, 44B, 47). This is particularly so in more river-dominated deltas where along-strike reworking is not significant and abrupt facies transitions may occur between distributaries and interdistributary areas (Bhattacharya, 1991). Overlapping delta lobes result in lens-shaped stratigraphic units that exhibit a mounded appearance on seismic lines (Figs. 43B, 44B). Regional mapping of delta lobes in the Cretaceous Dunvegan Formation, Alberta, Canada (Bhattacharya, 1991; Plint, 2000), the Frontier Formation in Wyoming, U.S.A. (Bhattacharya and Willis, 2001) and Pennsylvanian deltas in Kentucky (Horne et al., 1978) shows similar lateral overlap of lens-shaped delta lobes (e.g., Fig. 47).

Although the older seismic and sequence stratigraphic literature is rife with dip-oriented cross-sectional depictions of shelf depositional systems, newer studies emphasize strike-oriented variability (e.g., Anderson et al., 2004). Strike variability (i.e., the timing and spacing of overlapping lenses or lobes) is dependent on the number, spacing, and avulsion frequency of distributary channels. It also depends on the shape of the sea floor, especially if there is differential subsidence or uplift related to tectonics,

because deltas commonly fill low areas on the sea floor (Bhattacharya and Willis, 2001; Martinsen, 2003). Tectonics can be related to salt or shale mobility, or be related to larger-scale lithospheric deformation (e.g., plate tectonic).

Sequence Stratigraphy of Deltas

Delta systems have been an important focus of research in the development of new allostratigraphic and sequence stratigraphic concepts (e.g., Boyd et al., 1989; Van Wagoner et al., 1990; Posamentier et al., 1992; Bhattacharya, 1993; Miall, 1997; Posamentier and Allen, 1999; Anderson et al., 2004). Sequence stratigraphy provides a very different view of both the large-scale and small-scale architecture of sedimentary systems (e.g., Van Wagoner et al., 1990; Bhattacharya and Posamentier, 1994; Miall, 1997).

Allostratigraphy and sequence stratigraphy involves the correlation of bounding discontinuities through potentially varying lithologies that yields a fundamentally different picture of genetic stratal relationships than older lithostratigraphic techniques, and allows far more accurate paleogeographic maps to be constructed (Fig. 48). This is best illustrated by example. Historically the term delta has been generally applied to many clastic wedges, such as the Devonian–Carboniferous Castkill delta wedge in the northeastern U.S.A. (Woodrow and Sevon, 1985), and the Cretaceous Dunvegan, Ferron, and Frontier formations in Western North America. Previous lithostratigraphic maps of these undifferentiated wedges show broadly lobate geometries, especially at the distal margins of the wedges, but it was practically impossible to determine which lobe belongs to which channel without more detailed sequence stratigraphic correlations (Fig. 49).

The Dunvegan Delta

The Dunvegan Formation of Alberta, Canada, represents a heterolithic wedge of mudstones and sandstones, up to 300 m

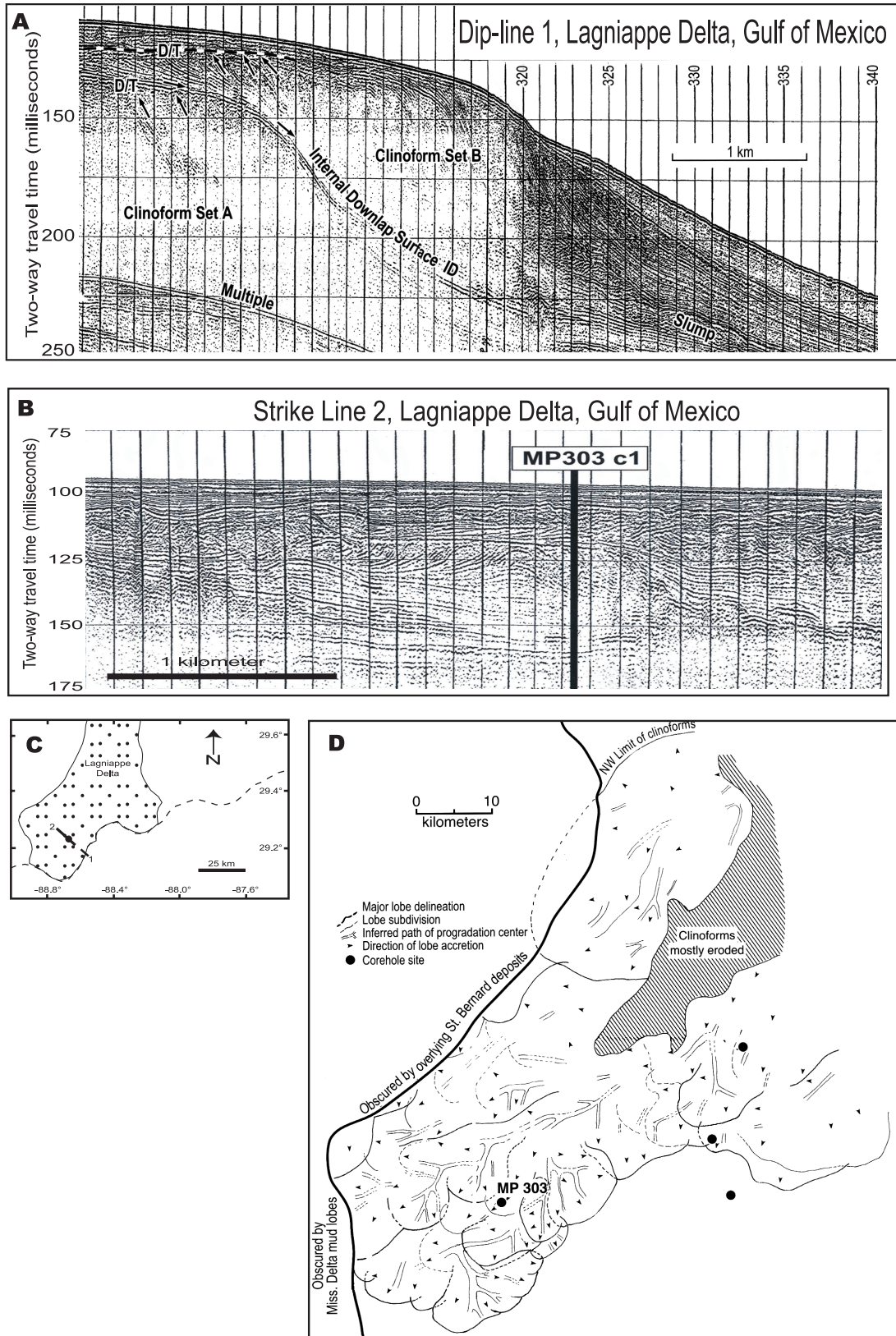


FIG. 44.—Seismic geometry of the Lagniappe delta, Gulf of Mexico. **A**) Dip line showing clinoforms; **B**) strike line showing lens-shaped cross sections; **C**) base map showing outline of lobe; **D**) detailed seismic facies mapping shows sub-lobes and distributary channels. Core MP303 c1 shows a predominantly upward-coarsening facies succession. After Roberts et al. (2004).

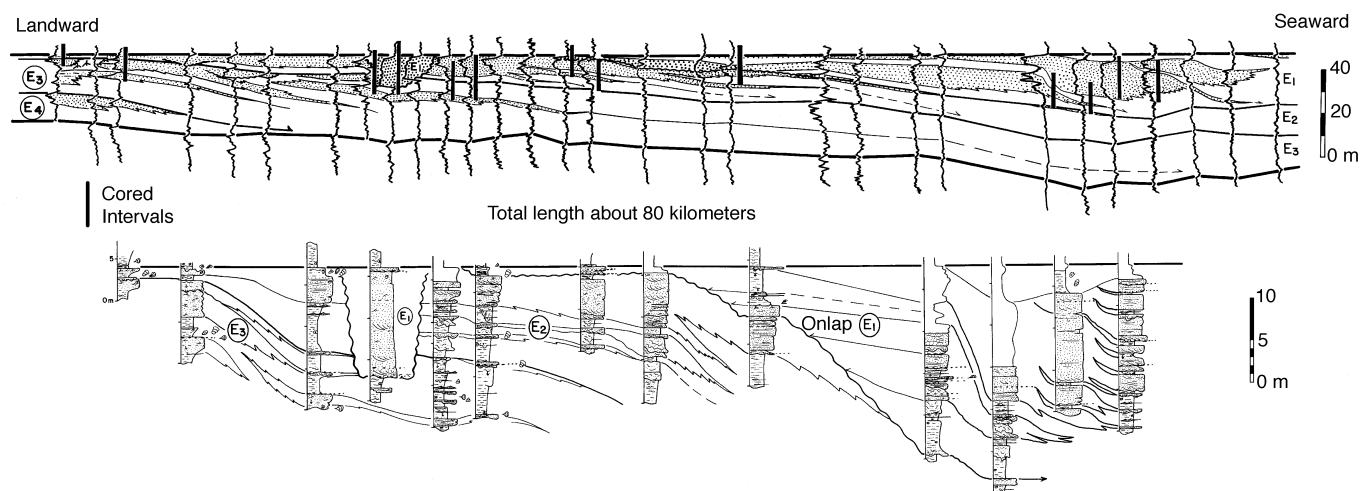


FIG. 45.—A) Dip-oriented well log and core cross sections within Allomember E of the Upper Cretaceous Dunvegan Formation, Alberta, Canada, showing offlapping clinoforms. Paleogeographic maps of the various offlapping shingled units are shown in Figure 51. Modified after Bhattacharya (1991).

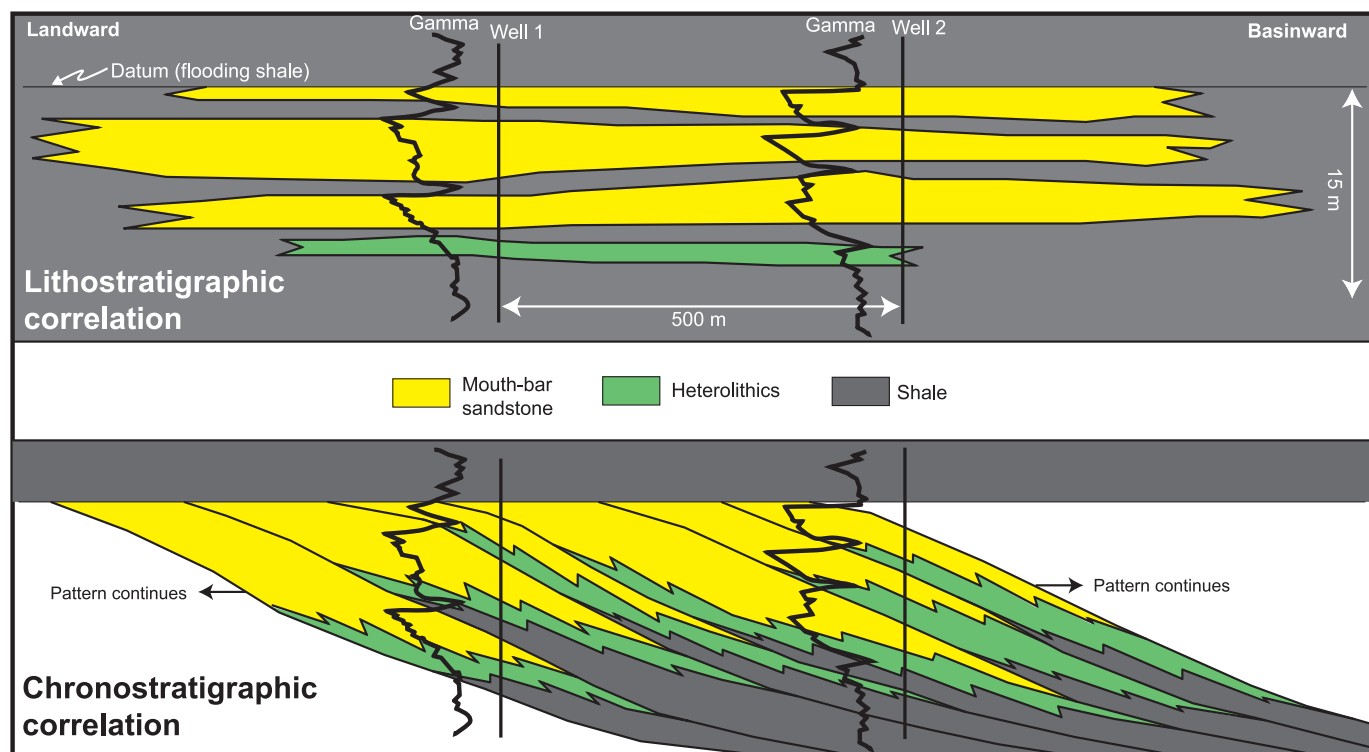


FIG. 46.—Subsurface model of well-log correlation in a lacustrine deltaic environment along depositional dip. Lithostratigraphic correlation (upper diagram) assumes no dip in sand bodies towards basin, whereas chronostratigraphic correlation (lower diagram) assumes basinward-dipping clinoforms. Chronostratigraphic model better predicts reservoir behavior. Note that correlation lengths of many beds are below the well spacing (from Gani and Bhattacharya, 2005, modified after Ainsworth et al., 1999).

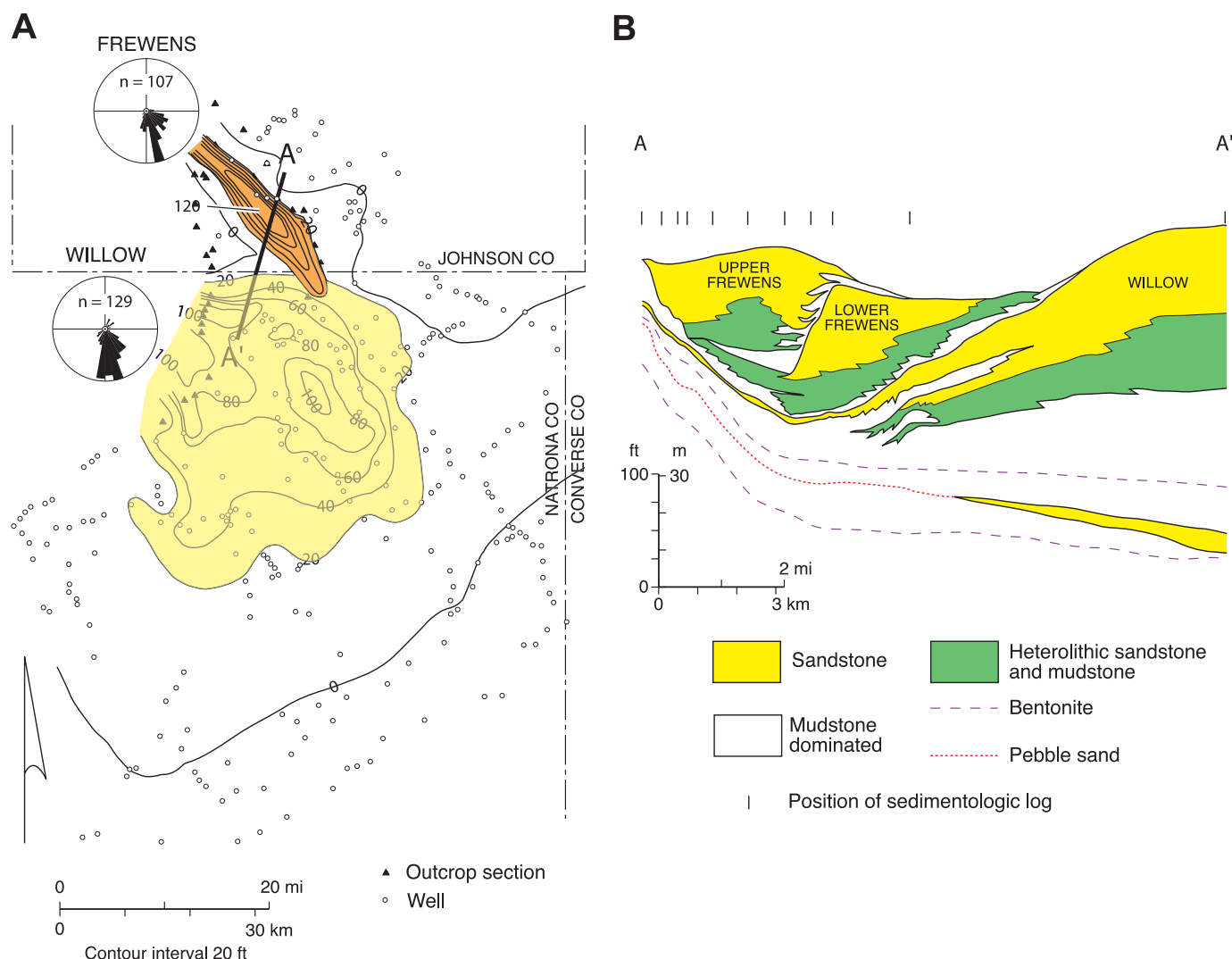


FIG. 47.—**A**) Maps and **B**) cross section of delta lobes and fingers in the Cretaceous Lower Belle Fourche Member, Frontier Formation, Wyoming, U.S.A. Strike section (b) shows overlapping lens-shaped delta bodies. Facies details of the Frewens tide-dominated delta sandstone are shown in Figures 34, 35, and 42. (After Willis et al., 1999, and Bhattacharya and Willis, 2001).

thick, deposited from the actively rising Western Cordillera into the adjacent Cretaceous Interior Seaway. The term “Dunvegan Delta” has been applied to this entire undifferentiated sedimentary package (Figs. 48, 49).

In a study area of about 30,000 km² in the subsurface of Alberta, Bhattacharya and Walker (1991) recognized seven throughgoing transgressive surfaces (Fig. 48B). These were used to subdivide the Dunvegan wedge into seven allomembers (Fig. 48B). Each of the allomembers could be further subdivided into several shingled offlapping units separated by less extensive surfaces of transgression and regression (Figs. 45, 48). The discontinuity-bounded shingles and allomembers provided the stratigraphic basis for more detailed facies mapping and paleogeographic reconstruction than had ever before been possible (Figs. 50–52). Continued work in a more landward direction by Plint (2002) has demonstrated older shingles and allomembers (Fig. 53) below those mapped by Bhattacharya and Walker (1991a).

Sandbody geometries within individual shingles (Figs. 50, 51) revealed a wide range of deltaic to shoreline-related depositional

systems, including some superb examples of ancient river-dominated deltas (Bhattacharya, 1991). The abundance of core data allowed reconstruction of the lateral facies relationships both down dip (Fig. 45) and along depositional strike. The cores also facilitated the development of summary vertical facies successions for the various components in the different deltaic systems (Figs. 28, 36, 39).

Individual delta lobes could not be mapped without the detailed correlation of the offlapping, shingled units (Figs. 45, 50, 51). If sandstones within each allomember are mapped together, the isolith patterns do not show narrowing and thickening towards a point fluvial source in a landward position (Fig. 49). Only the seaward deltaic promontories can be seen (e.g., Fig. 49). Also, the highstand and lowstand deltas within each allomember could not be mapped without the detailed correlation of the shingle boundaries. This is an especially acute problem in high-accommodation settings, where lowstands tend to be attached to the highstand clastic wedge, versus low-accommodation settings, where lowstands are highly detached (Ainsworth and Pattison, 1994).

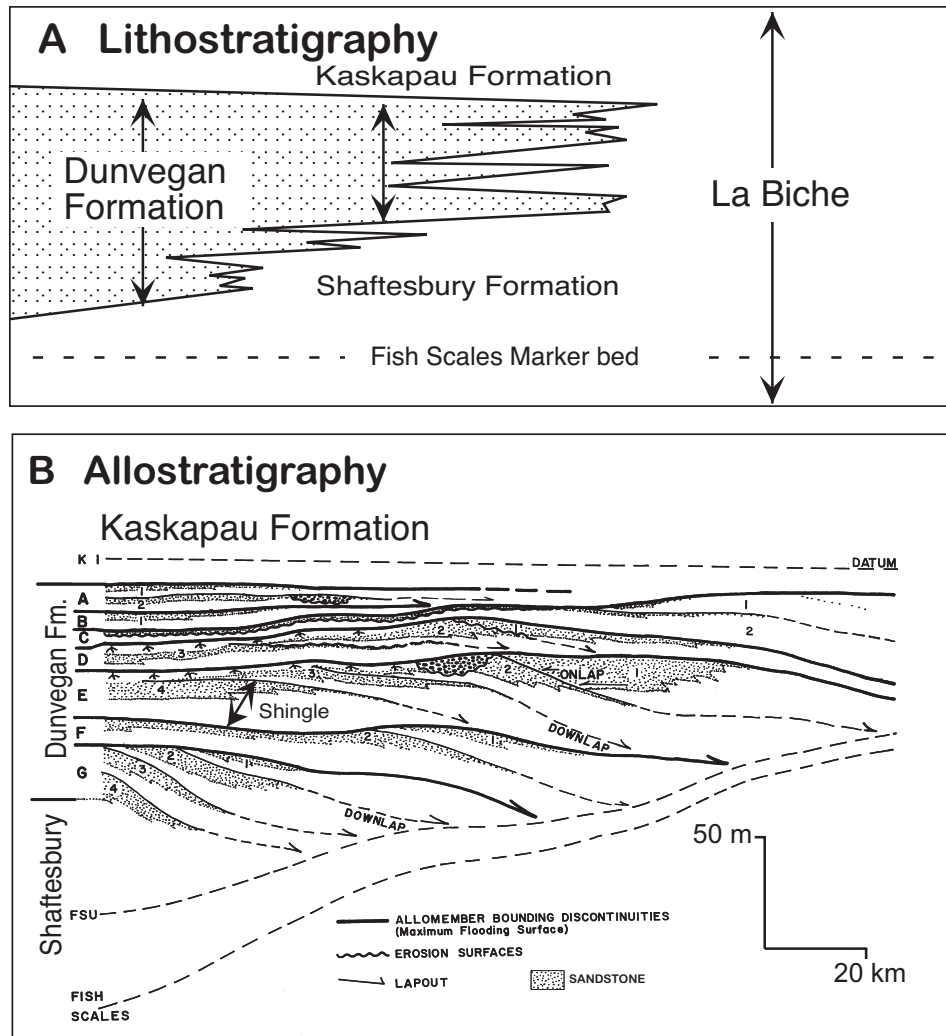


FIG. 48.—Regional cross section across the Alberta Foreland Basin, Canada, illustrating the difference between a lithostratigraphic and an allostratigraphic interpretation of the Upper Cretaceous Dunvegan Formation (from Bhattacharya, 1993). **A)** The lithostratigraphic interpretation depicts a homogeneous, wedge-shaped sandstone body that tapers to the right. Some interfingering of the distal end of the Dunvegan Formation into the La Biche Formation shales is shown. **B)** The allostratigraphic interpretation shows that the Dunvegan comprises several stacked allomembers (A to G). Each allomember consists of several smaller-scale, offlapping, shingled units that map as delta lobes (e.g., Figs. 50 and 51). Each allomember is bounded by a regional transgressive flooding surface. These regional flooding surfaces and smaller-scale “shingle” boundaries show that the Dunvegan consists of numerous sandy compartments, bounded by mudstones. Oil and gas reservoirs occur within these smaller-scale shingled units.

More recent work (Plint and Wadsworth, 2003) in the landward realm has allowed mapping of the nonmarine facies within the Dunvegan, including superb examples of tributive valley systems (Fig. 38B).

Ferron Example

A recent correlation of the Ferron sandstone member, in central Utah, U.S.A., based on nearly continuous outcrop exposures, shows a complicated series of seaward-stepping, offlapping, to aggrading and finally backstepping shorelines (Fig. 54; Gardner, 1995; Gardner et al., 2004; Barton et al., 2004; Garrison and van den Bergh, 2004). The correlation is based on tracing various bounding discontinuities, including flooding

surfaces, erosional surfaces, coal beds, bentonites, and ammonite horizons across the outcrop belt. The Ferron has been subdivided into seven major transgressive-regressive “stratigraphic cycles”, each of which is bounded by a regionally traceable flooding surface and associated coal (Ryer, 1984; Gardner, 1995). The lower two stratigraphic cycles consist of strongly seaward-stepping shoreline and delta deposits. The middle three cycles aggrade, and the last two cycles backstep. Regionally, the Ferron delta prograded to the northeast, but locally, individual delta lobes prograded at high angles to this general northeast direction.

Internally, the stratigraphic cycles consist of a series of lens-shaped to lobate offlapping, shingled delta-front and shoreface sandstone bodies that show upward-coarsening facies succes-

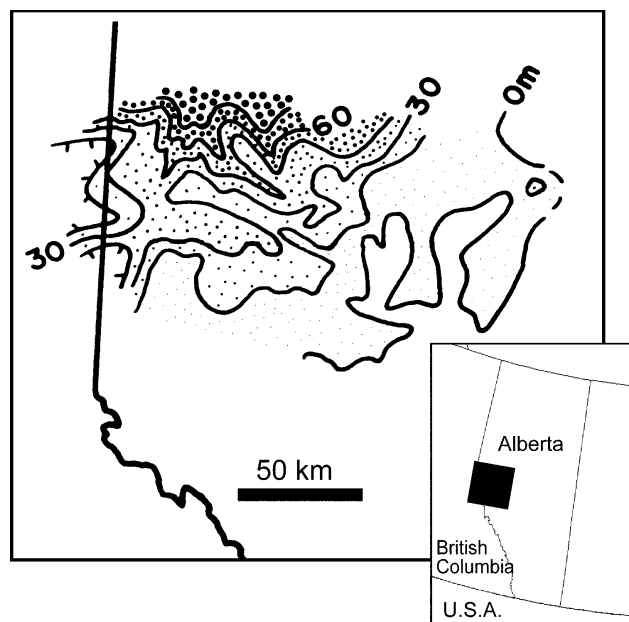


FIG. 49.—Sandstone isolith mapping of undifferentiated Dunvegan wedge (after Burk, 1963).

sions. Wave-influenced and fluvial-dominated successions can be recognized. The shingled units are locally bounded by less extensive flooding surfaces, and they thus resemble “parasequences” (e.g., Van Wagoner et al., 1990). The “stratigraphic cycles” effectively are “parasequence sets”. The shoreline-type sandstone bodies correlate landward into a thick delta-plain succession, consisting of single-story distributary channels and multi-story valley fills (Fig. 37; Garrison and van den Bergh, 2004; Barton, 2004). Highly asymmetric subsidence, characteristic of the foreland-basin setting, results in high accommodation, high sediment supply, a broadly positive shoreline trajectory, and consequently accumulation and preservation of thick paralic, delta-plain topset facies (compare Fig. 54 and Fig. 27).

The integration of detailed facies analysis within a well developed stratigraphic framework allows specific statements to be made about how depositional systems evolve as a function of position within the wedge. The seaward-stepping deltaic parasequences in the lowest stratigraphic cycle show greater river influence than the more wave- and tide-influenced parasequences in the upper cycles, although there is considerable facies variability at all scales.

The lower parasequences in the Ferron contain normal growth faults well exposed along the highly accessible walls of Muddy Creek Canyon in Central Utah (Fig. 30; Bhattacharya and Davies, 2001, 2004). Distinctive pre-growth, growth, and post-growth strata indicate a highly river-dominated crevasse delta that prograded northwest into a large embayment of the Ferron shoreline. The growth section comprises medium- to large-scale cross-stratified sandstones deposited as upstream- and downstream-accreting mouth bars in the proximal delta front. Deposition of mouth-bar sands initiates faults. Because depositional loci shift rapidly, there is no systematic landward or seaward migration of fault patterns. During later evolution of the delta, foundering of fault blocks creates an uneven sea-floor topography that is smoothed over by the last stage of deltaic progradation. Faults are inferred to occur within less than 10 m water depths in soft, wet sediment (Bhattacharya and Davies, 2004). Detailed examination

of the fault zones shows that deformation was largely by soft-sediment mechanisms, such as grain rolling and by lubrication of liquefied muds, causing shale smears. Mechanical attenuation of thin beds occurs by displacement across multiple closely spaced small throw faults. Analogous river-dominated deltaic subsurface reservoirs may be compartmentalized by growth faults, even in shallow-water, intracratonic, or shelf-perched highstand deltas. Reservoir compartmentalization would occur where thicker homogeneous-growth sandstones are placed against the muddy pre-growth strata and where faults are shale-smeared, and thus potentially sealing.

Although the Dunvegan and Ferron are deposited into a very similar tectonic and paleogeographic setting, the Ferron shows considerably more complexity in cross section (compare Figures 53 and 54). This likely reflects the fact that the Ferron is nearly continuously exposed in outcrop, allowing greater resolution of individual delta lobes than is possible with the largely well-log-based cross section of the Dunvegan. Outcrop examples, like the Ferron, indicate the fundamentally complex nature of reservoir compartmentalization within fluviodeltaic reservoirs.

Top-Truncated Deltas and the Shelf Sand Problem

One of the central debates in facies sedimentology in the past few decades has been the interpretation of seemingly isolated basin-distal, “shelf” sandstones (cf. Snedden and Bergman, 1999). These deposits comprise meters-thick to tens-of-meters-thick, upward-coarsening facies successions that are surrounded by “shelf” mudstone. Coeval paralic facies, such as typify the high-accommodation and unequivocally deltaic Dunvegan and Ferron sandstones described above, are commonly found tens to hundreds of kilometers away. In the 1970s and 1980s these basinally isolated sandstones were widely interpreted as offshore shelf bars, molded by shelf processes unrelated to the shoreline, or deposited as turbidites, implying deeper-water depositional systems (see summary in Bhattacharya and Willis, 2001).

In the late 1980s and the 1990s, sequence stratigraphic concepts forced a reconsideration of these units, and it was recognized that the sandstones, in many cases, were underlain by erosional surfaces, implying no genetic link between the underlying shelf mudstones and the newly interpreted overlying shelf sand deposits, interpreted to be deposited by the process of “forced regression” during major drops of sea level (Plint, 1988; Posamentier et al. 1992).

The next level of debate then centered on the magnitude of sea-level drops and the origin of the erosional surfaces. Several studies went the ultimate step and interpreted the erosion surfaces to be fluvially eroded, incised-valley fills (e.g., Jennette and Jones, 1995). Thus, sediment bodies once interpreted as deeper-water “shelf” turbidites became fluvially incised valley fills.

An unwillingness to interpret these systems as strictly deltaic can be traced directly to the application of Barrell’s 1912 criterion, namely that subaerial facies are *required* to define a delta. Strict adherence to this rule does not consider the fact that during forced regression, shoreline trajectory is negative, and accommodation for delta-plain facies is extremely limited (Fig. 27). Also, ravinement is a very efficient process that truncates the tops of these systems.

The Cretaceous Frontier Formation, in the Powder River Basin of Wyoming, U.S.A., is a recently described example that contains numerous basin-distal sandstones separated by mudstones (Bhattacharya and Willis, 2001). Previous interpretations of these sandstones, based on only limited data, ranged from

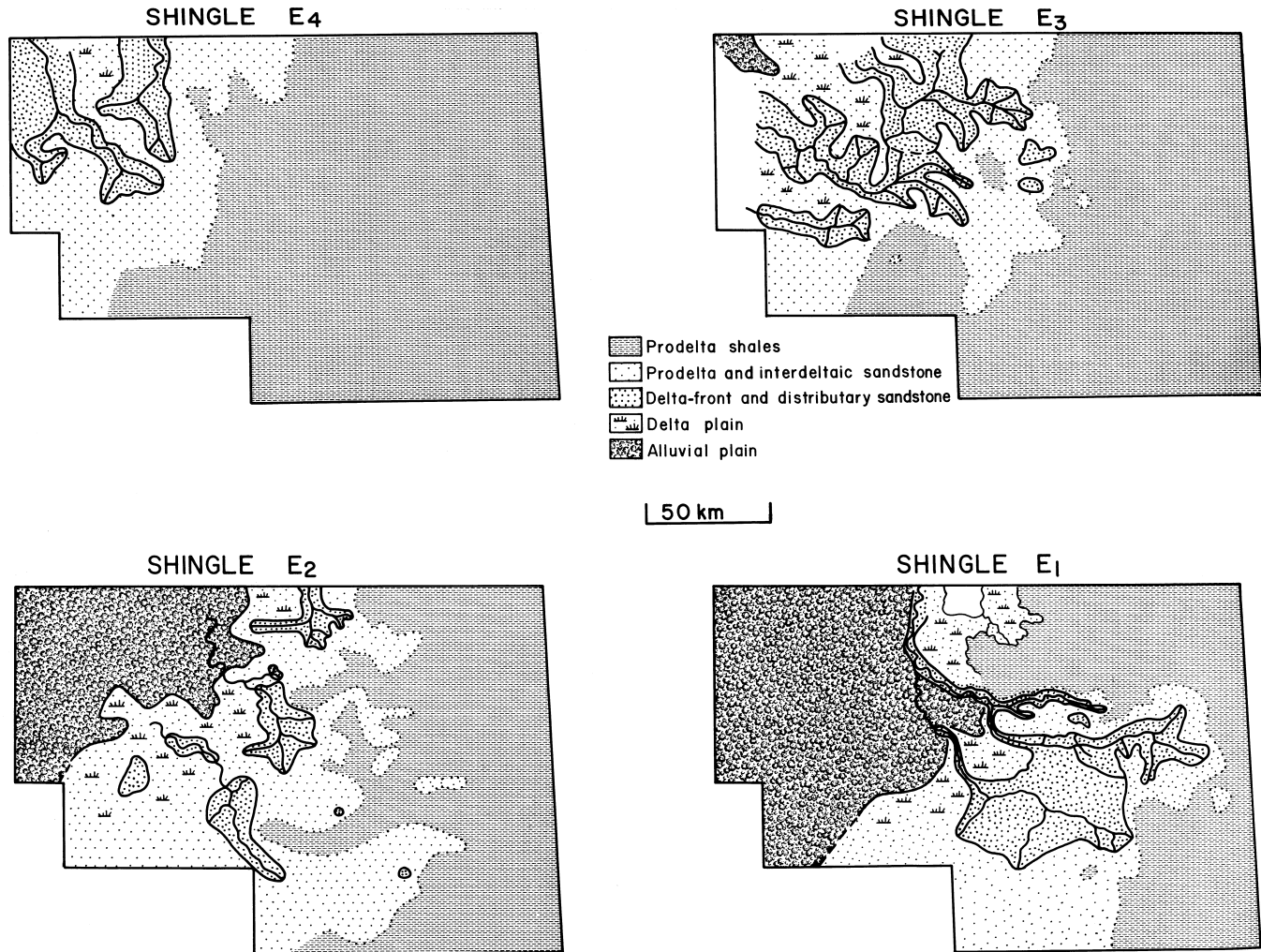


FIG. 50.—Paleogeographic maps of successive offlapping shingles within Allomember E of the Dunvegan Formation (see also Figures 45 and 48). Note numerous small deltas associated with the highstand shingles E4 to E3. The youngest shingle, E1, is accompanied by fluvial entrenchment, incision of an incised valley, and development of a single, massive delta lobe. E1 is thus interpreted as a lowstand delta. From Bhattacharya (1991).

prodelta shelf plumes (Winn, 1991), to tidally modified shelf ridges (Tillman and Merewether, 1998), to incised estuarine valley fills (Tillman and Merewether, 1998). A detailed allostratigraphic study of the lower portion of the Frontier incorporated lithofacies, ichnofacies, palynofacies, paleocurrent data, bedding relationships, and isolith maps from nearly 100 measured outcrop sections and about 550 subsurface well logs. Four allomembers, interpreted as major delta lobes, were identified and mapped (Fig. 47). Each allomember has a lobate to elongate geometry, basinward-dipping internal clinoform bedding (Fig. 42), radiating paleocurrents, and low to moderate degree of shallow marine burrowing. Facies successions show variable wave- and tide-influenced lithofacies. Delta-plain, paralic, and nonmarine facies have been eroded from the top of every deltaic succession, at least in the Powder River Basin. Erosion surfaces capping progradational deltaic successions are the only stratal discontinuities that can be mapped regionally (Fig. 55) and were used to define the four allomembers within the Formation. The bounding discontinuities record transgressive ravinement that was enhanced over areas of structural uplift, rather than low-

stand surfaces of erosion recording the bypass of sediments basinward (Bhattacharya and Willis, 2001; Martinsen, 2003). Chert pebble and cobble lags associated with these surfaces are the primary evidence that coarser-grained rivers fed these shorelines (Fig. 55). The fluvial deposits have been completely reworked into a transgressive lag.

The low-accommodation setting left little room for sandstones to stack vertically, and successive episodes of delta progradation were offset along strike (Fig. 47). More tide-influenced delta deposits of the Frewens Allomember (Willis et al., 1999) formed within shoreline embayments defined by the topography of older wave-influenced delta lobes and subtle syndepositional deformation of the basin floor (Bhattacharya and Willis, 2001).

San Miguel Formation

In a final example, Weise (1980) used somewhat more traditional lithostratigraphic and facies techniques to correlate and map sandstones within the Cretaceous San Miguel Formation in

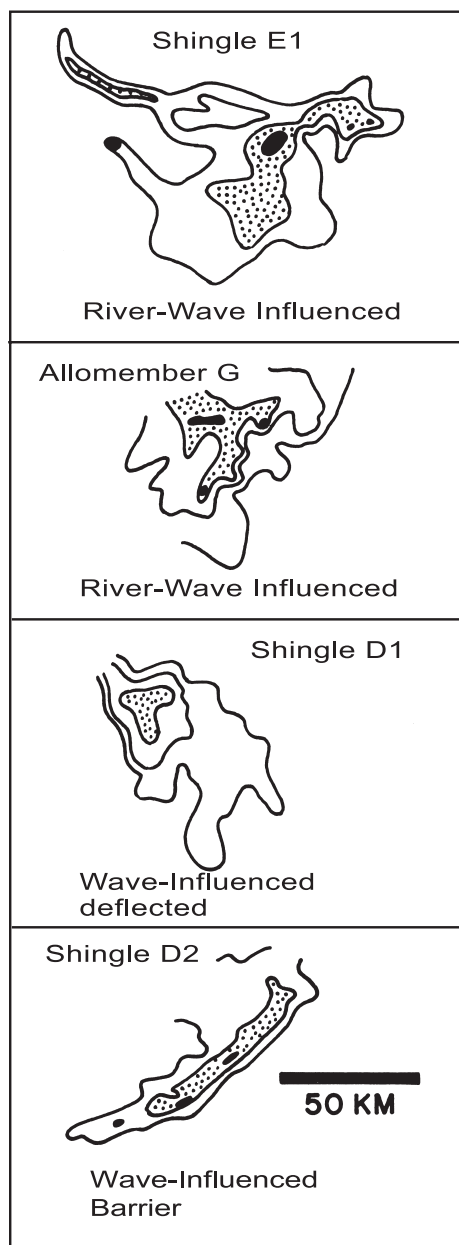


FIG. 51.—Spectrum of sandbody geometries shown by sandstone isolith maps of shingles within various allomembers in the Upper Cretaceous Dunvegan Formation (Alberta, Canada). From Bhattacharya and Walker (1992), based on data presented by Bhattacharya and Walker (1991b). Mapping of isoliths within entire allomember does not reveal deltaic shape as clearly.

east Texas, U.S.A., (Fig. 56). Cores showed highly bioturbated shelf and shoreface type, upward-coarsening facies successions, with abundant hummocky cross stratification and a typical wave-dominated *Cruziana* ichnofacies grading into an *Ohiomorpha*-dominated *Skolithos* ichnofacies. No fluvial or paralic facies were noted in the subsurface. Her maps (Fig. 56) showed classic deltaic morphologies, similar to the models of Coleman and Wright (1975). The sand-body geometries were readily interpreted to be wave-influenced deltas, and the paper has become a classic.

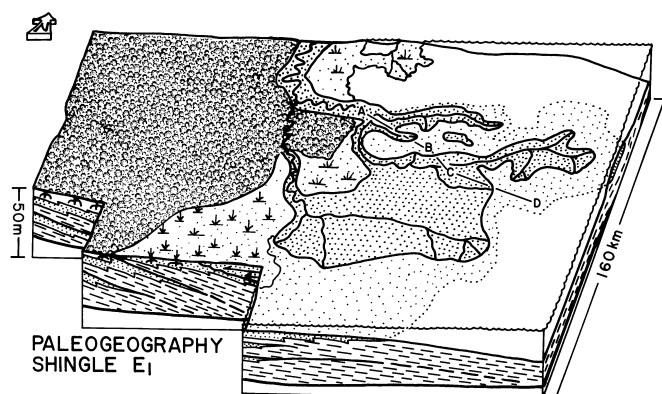


FIG. 52.—Block Diagram showing paleogeography of delta lobe of shingle E1, Dunvegan Formation. Allostratigraphic correlations allow channels and associated delta lobes to be separated and linked. Compare with Figure 48 and Figure 49 (From Bhattacharya, 1991).

Although the paper was published before the sequence stratigraphic revolution, Weise suggests that the lack of delta-plain facies resulted from transgressive reworking and removal of topsets.

The San Miguel is one of the earliest well-documented examples of a top-truncated delta system. It illustrates the key point that the lower the accommodation rate, the greater the degree of erosion resulting in partial preservation of an ancient depositional system. These examples show that detailed sedimentology, ichnology, facies mapping, and paleocurrent data must all be integrated to make an unequivocal interpretation, as was recently elucidated by Martinsen (2003). Although fluvial influence can be identified in some single vertical successions, many deltas may contain little evidence of fluvial effects (such as the San Miguel example), which can then be identified only on the basis of correlation and mapping.

Controls on Sequence Stratigraphic Organization

The distribution of lithologies and overall facies architecture within a given depositional system is highly sensitive to allocyclic controls, such as eustasy, tectonics, and sediment supply (Jervey, 1988). Sequence stratigraphy provides a means of interpreting these controls on sediment partitioning within and between different depositional systems, as well as providing a means for understanding the origin of key bounding discontinuities that are critical to correlate and map depositional systems and systems tracts (Posamentier et al., 1988; Bhattacharya, 1993; Anderson, et al., 2004). For example, when base level is low or falling (lowstand and falling-stage (forced regressive) systems tracts), river-dominated deltas may form, and the fluvial systems may be sandier and may be predominantly erosional or incised (Bhattacharya and Walker, 1992; Shanley and McCabe, 1994). During times of lowered sea level in the Cretaceous Western interior, the effects of subtle sea-floor topography may be enhanced, causing complex physiography and numerous embayments of the shoreline (Bhattacharya and Willis, 2001; Nummedal and Riley, 1999). Tidal effects, in particular, can be enhanced within these embayments, especially during the initial turnaround of sea level. Units originally interpreted as shelf sands, such as the Shannon, Tocito, and Frewens sandstones, have been reinterpreted as tidally influenced lowstand deltas and shorefaces (Bergman, 1994; Sullivan et al., 1997; Nummedal and

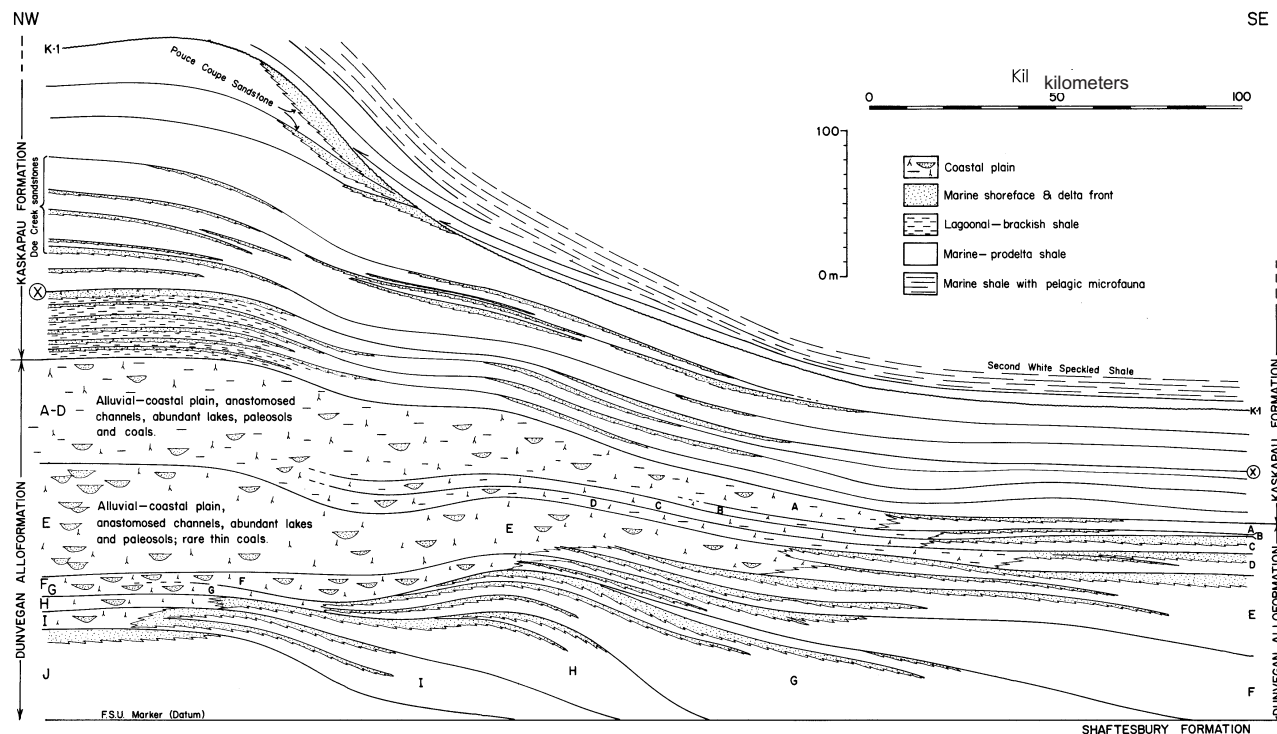


FIG. 53.—Allostratigraphy of the Dunvegan Alloformation. Plint (2000) extended correlation of marine flooding surfaces (initiated by Bhattacharya, 1993) into the nonmarine portion of the clastic wedge. This included mapping of tributive valley systems (see Figure 38). Nonmarine facies show a distinct thickening, interpreted as driven by increased tectonic subsidence to the northwest.

Riley, 1999; Willis et al., 1999; Bhattacharya and Willis, 2001; Willis and Gabel, 2001).

In contrast, during rising base level (e.g., transgressive and highstand systems tracts) mud tends to be trapped within the estuary or within the alluvial realm, resulting in more heterogeneous fluvial reservoirs (see Ferron example, above). Coeval shorelines may tend to be wave- or tide-dominated (Bhattacharya, 1993; Barton, 1997; Plint, 2000).

Galloway (1975) suggested that many regressive-transgressive clastic wedges in the Texas Gulf Coast show a transition from elongate river-dominated types in the lower parts of a wedge to more wave-modified deltas in the upper parts. Other examples include the Norias delta system (Duncan, 1983). This idea was recognized earlier by Barrell (1912), who noted that the Catskill wedge showed a similar transition.

An understanding of the allostratigraphic framework potentially allows the geologist to predict relative shale proportions, geometries, and distributions in different depositional environments as a function of stratigraphic position and basin setting (e.g., Gardner, 1995). This is of great value in constructing detailed geo-cellular models in reservoir characterization.

The concept of systems tracts also provides a way of linking the different types of depositional systems in various positions in the basin. Growth of deep-water submarine fans is commonly related to the development of shelf-edge deltas in lowstand systems tracts and the development of hyperpynal flow at the river mouth (Porebski and Steel, 2003; Anderson, 2005; Plink-Bjorklund and Steel, 2005). Development of fans may also correlate with the incision of alluvial systems in a landward direction. Submarine fan growth commonly ends with a relative rise of sea

level, and may coincide with development of shoal-water deltas in a transgressive or highstand systems tract. The generalizations embodied in the definitions of systems tracts allow the tracts to be used predictively, and in that sense they can be used as basin-scale facies models. One part of a given systems tract may be important in predicting the appearance and nature of a related part.

The differences in interpretations and nomenclature in sequence stratigraphy results from the distinction between systems tracts defined purely on geometric character and physical position within the sequence versus definition in the context of time and relative changes in sea level. In most papers on sequence stratigraphy, concept-driven genetic models and field observations are complexly intertwined. Nomenclature problems of the same kind exist in other areas of sedimentology, such as the use of the term facies to describe rock properties (e.g., cross-bedded sandstone facies) versus paleodepositional environments (shoreface facies). The way that most scientific systems are named and analyzed reflects how we think about them, and to some degree all observations are model driven. The use of systems tracts in the context of relative sea-level change seems to be more widespread than the use favored by Van Wagoner (1995) but in general it is critical to separate observation from interpretations.

CONCLUSIONS

Deltas are complex three-dimensional progradational depositional systems that form primarily as a consequence of the interaction of a river plume with basin processes, chiefly tides and waves. Longer-term changes in controlling parameters

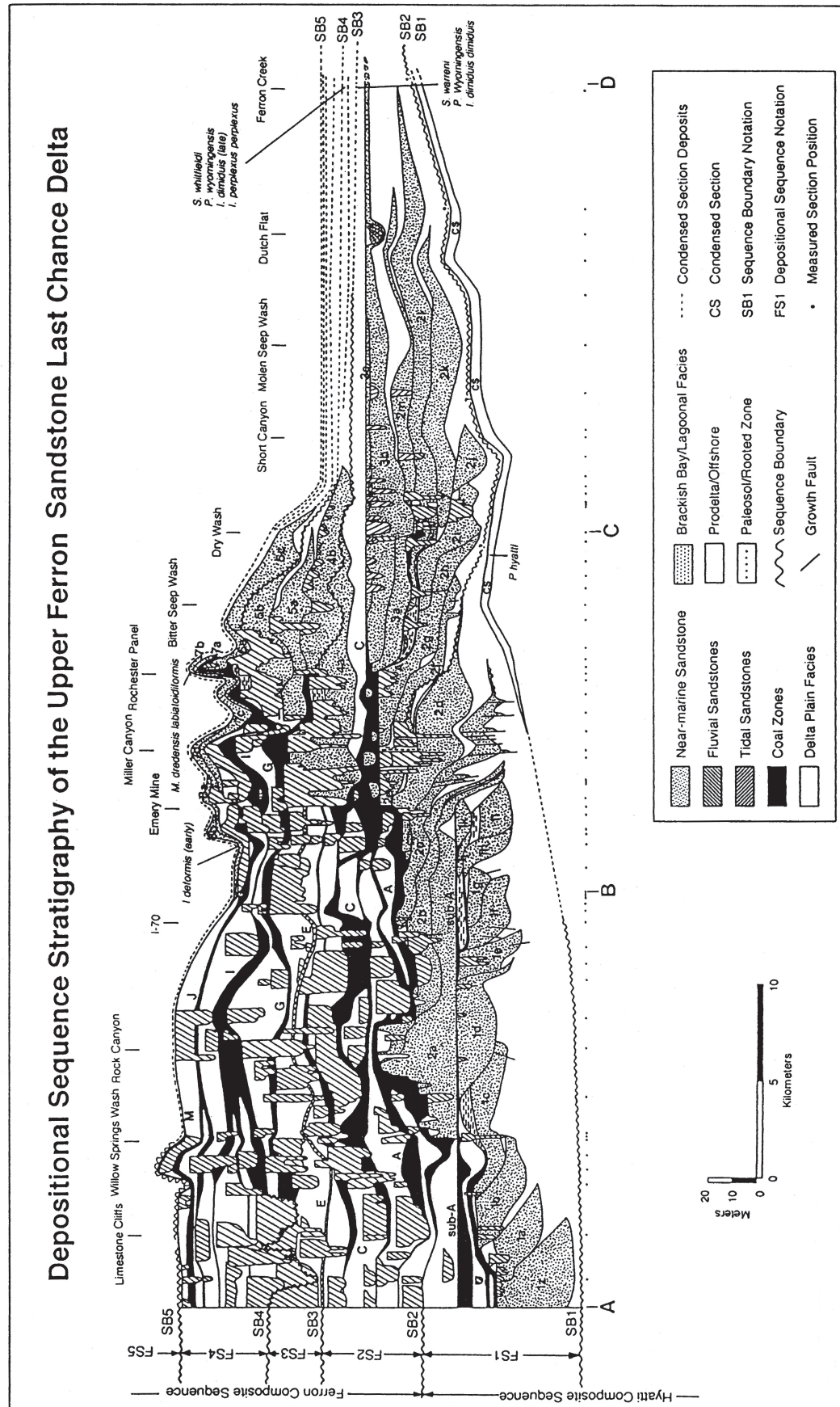


FIG. 54.—Sequence stratigraphic cross section of the Cretaceous Ferron sandstone. The cross section is based on nearly 100 measured sections and correlations of coals, bentonites, and other key flooding surfaces in nearly continuous cliff exposures along the Coal Cliffs and Wasatch Plateau in central Utah, U.S.A. (from Garrison and van den Bergh, 2004).

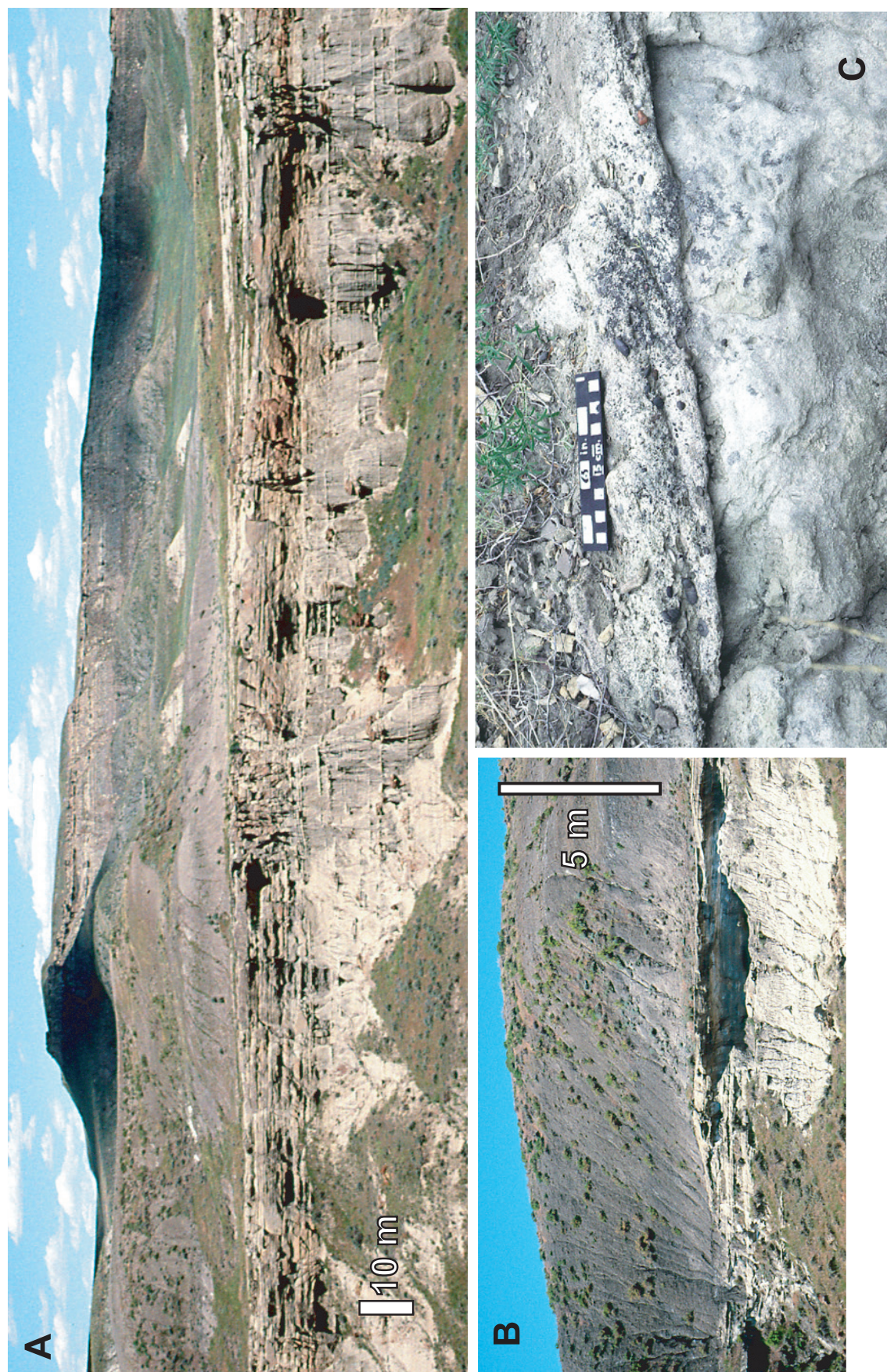


FIG. 55.—Photos of erosional discontinuity above Belle Fourche sandstone bodies of the Frontier Formation, Wyoming, U.S.A.. A) Regional photo shows dipping delta-front sandstones truncated on top and overlain by shale. B) Close-up of erosional contact between shales and underlying delta-front sandstones. C) Locally, a cross-bedded pebbly sandstone lag containing sharks' teeth and oyster fragments marks the transgressive erosion (see Bhattacharya and Willis, 2001, for more details).

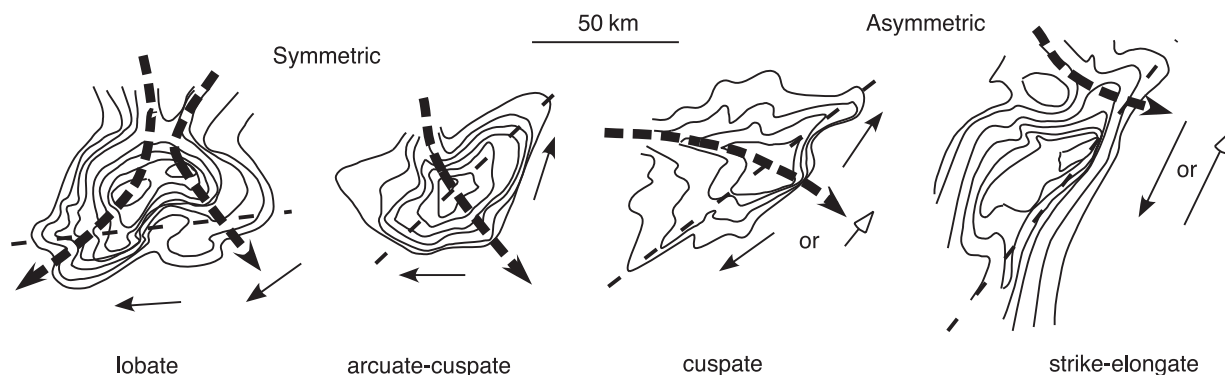


FIG. 56.—Spectrum of river-dominated (left) to wave-dominated (right) deltas in the Upper Cretaceous San Miguel Formation (Texas, U.S.A.) after Weise (1980). The isopach maps are contoured in 20 foot (6.1 m) intervals, with maximum thicknesses of about 120–140 feet (36–43 m). Symmetrical and asymmetrical deltas can be identified. The model for the formation of asymmetrical deltas suggests that the original interpretations of longshore-drift directions may be reversed from that originally interpreted. Sandier deposits may occur on the updrift rather than on the downdrift side.

may cause deltas to be transformed into other depositional systems. A delta may transform into a barrier-island system during a major transgression, for example (Fig. 7). Conversely, an increase in sedimentation rate, or a channel avulsion, could transform a prograding strandplain into a delta. A steady rise of relative sea level may cause a river channel or distributary to widen into an estuary with tidally influenced sand bars. The estuary might in turn be drowned, and the system would evolve into a series of shallow-marine, tidal sand ridges (Fig 20; Yang and Sun, 1988).

In other settings, waves are important in modifying deltaic sediments during transgression. Around the margins of the Mississippi delta, such modification has produced winnowed sandbodies that began as beach-ridge complexes, were detached as barriers, and have now been drowned to produce shelf shoals (Fig. 7). It is now recognized that waves, rivers, and tides continuously interact, resulting in deposits that may show an interfingering of facies architectural elements formed by these different processes (Bhattacharya and Giosan, 2003). Depending on the larger stratigraphic context, these can be regarded as components of deltaic systems or as distinct depositional systems.

The definition and understanding of a depositional system depends largely on the scale of observation and objectives of the study. The distribution of sediment at the surface in many modern deltas may not characterize the sandbody geometry in the immediately underlying deposits. In many modern deltas, transgressive reworking by waves and tides of the uppermost veneer of sediments (i.e., destructional phase) may not reflect the sedimentary processes in the major proportion of the underlying regressive deltaic package (i.e., constructional phase).

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