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SEISMIC GEOMORPHOLOGY AND STRATIGRAPHY OF DEPOSITIONAL ELEMENTS IN DEEP-WATER SETTINGS

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ABSTRACT: Analyses of 3-D seismic data in predominantly basin-floor settings offshore Indonesia, Nigeria, and the Gulf of Mexico, reveal the extensive presence of gravity-flow depositional elements. Five key elements were observed: (1) turbidity-flow leveed channels, (2) channel-overbank sediment waves and levees, (3) frontal splays or distributary-channel complexes, (4) crevasse-splay complexes, and (5) debris-flow channels, lobes, and sheets. Each depositional element displays a unique morphology and seismic expression. The reservoir architecture of each of these depositional elements is a function of the interaction between sedimentary process, sea-floor morphology, and sediment grain-size distribution.

(1) Turbidity-flow leveed-channel widths range from greater than 3 km to less than 200 m. Sinuosity ranges from moderate to high, and channel meanders in most instances migrate down-system. The high-amplitude reflection character that commonly characterizes these features suggests the presence of sand within the channels. In some instances, high-sinuosity channels are associated with (2) channel-overbank sediment-wave development in proximal overbank levee settings, especially in association with outer channel bends. These sediment waves reach heights of 20 m and spacings of 2–3 km. The crests of these sediment waves are oriented normal to the inferred transport direction of turbidity flows, and the waves have migrated in an up-flow direction. Channel-margin levee thickness decreases systematically down-system. Where levee thickness can no longer be resolved seismically, high-sinuosity channels feed (3) frontal splays or low-sinuosity, distributary-channel complexes. Low-sinuosity distributary-channel complexes are expressed as lobate sheets up to 5–10 km wide and tens of kilometers long that extend to the distal edges of these systems. They likely comprise sheet-like sandstone units consisting of shallow channelized and associated sand-rich overbank deposits. Also observed are (4) crevasse-splay deposits, which form as a result of the breaching of levees, commonly at channel bends. Similar to frontal splays, but smaller in size, these deposits commonly are characterized by sheet-like turbidites. (5) Debris-flow deposits comprise low-sinuosity channel fills, narrow elongate lobes, and sheets and are characterized seismically by contorted, chaotic, low-amplitude reflection patterns. These deposits commonly overlie striated or grooved pavements that can be up to tens of kilometers long, 15 m deep, and 25 m wide. Where flows are unconfined, striation patterns suggest that divergent flow is common. Debris-flow deposits extend as far basinward as turbidites, and individual debris-flow units can reach 80 m in thickness and commonly are marked by steep edges. Transparent to chaotic seismic reflection character suggest that these deposits are mud-rich.

Stratigraphically, deep-water basin-floor successions commonly are characterized by mass-transport deposits at the base, overlain by turbidite frontal-splay deposits and subsequently by leveed-channel deposits. Capping this succession is another mass-transport unit ultimately overlain and draped by condensed-section deposits. This succession can be related to a cycle of relative sea-level change and associated events at the corresponding shelf edge. Commonly, deposition of a deep-water sequence is initiated with the onset of relative sea-level fall and ends with subsequent rapid relative sea-level rise.

INTRODUCTION

The understanding of deep-water depositional systems has advanced significantly in recent years. In the past, much understanding of deep-water sedimentation came from studies of outcrops, recent fan systems, and 2D reflection seismic data (Bouma 1962; Mutti and Ricci Lucchi 1972; Normark 1970, 1978; Walker 1978; Posamentier et al. 1991; Weimer 1991; Mutti and Normark 1991). However, in recent years this knowledge has advanced significantly because of (1) the interest by petroleum companies in deep-water exploration (e.g., Pirmez et al. 2000), and the advent of widely available high-quality 3D seismic data across a broad range of deep-water environments (e.g., Beaubouef and Friedman 2000; Posamentier et al. 2000), (2) the recent drilling and coring of both near-surface and reservoir-level deep-water systems (e.g., Twichell et al. 1992), and (3) the increasing utilization of deep-tow side-scan sonar and other imaging devices (e.g., Twichell et al. 1992; Kenyon and Millington 1995). It is arguably the first factor that has had the most significant impact on our understanding of deep-water systems. Three-dimensional seismic data afford an unparalleled view of the deep-water depositional environment, in some instances with vertical resolution down to 2–3 m. Seismic time slices, horizon-datum time slices, and interval attributes provide images of deep-water depositional systems in map view that can then be analyzed from a geomorphologic perspective. Geomorphologic analyses lead to the identification of depositional elements, which, when integrated with seismic profiles, can yield significant stratigraphic insight. Finally, calibration by correlation with borehole data, including logs, conventional core, and biostratigraphic samples, can provide the interpreter with an improved understanding of the geology of deep-water systems.

The focus of this study is the deep-water component of a depositional sequence. We describe and discuss only those elements and stratigraphic successions that are present in deep-water depositional environments. The examples shown in this study largely are Pleistocene in age and most are encountered within the uppermost 400 m of substrate. These relatively shallowly buried features represent the full range of lowstand deep-water depositional sequences from early and late lowstand through transgressive and highstand deposits. Because they are not buried deeply, these stratigraphic units commonly are well-imaged on 3D seismic data. It is also noteworthy that although the examples shown here largely are of Pleistocene age, the age of these deposits should not play a significant role in subsequent discussion. What determines the architecture of deep-water deposits are the controlling parameters of flow discharge, sand-to-mud ratio, slope length, slope gradient, and rugosity of the seafloor, and not the age of the deposits. It does not matter whether these deposits are Pleistocene, Carboniferous, or Precambrian; the physical “first principles” of sediment gravity flow apply without distinguishing between when these deposits formed. However, from the perspective of studying deep-water turbidites it is advantageous that the Pleistocene was such an active time in the deep-water environment, resulting in deposition of numerous shallowly buried, well-imaged, deep-water systems.

Depositional Elements Approach

This study is based on the grouping of similar geomorphic features referred to as depositional elements. Depositional elements are defined by

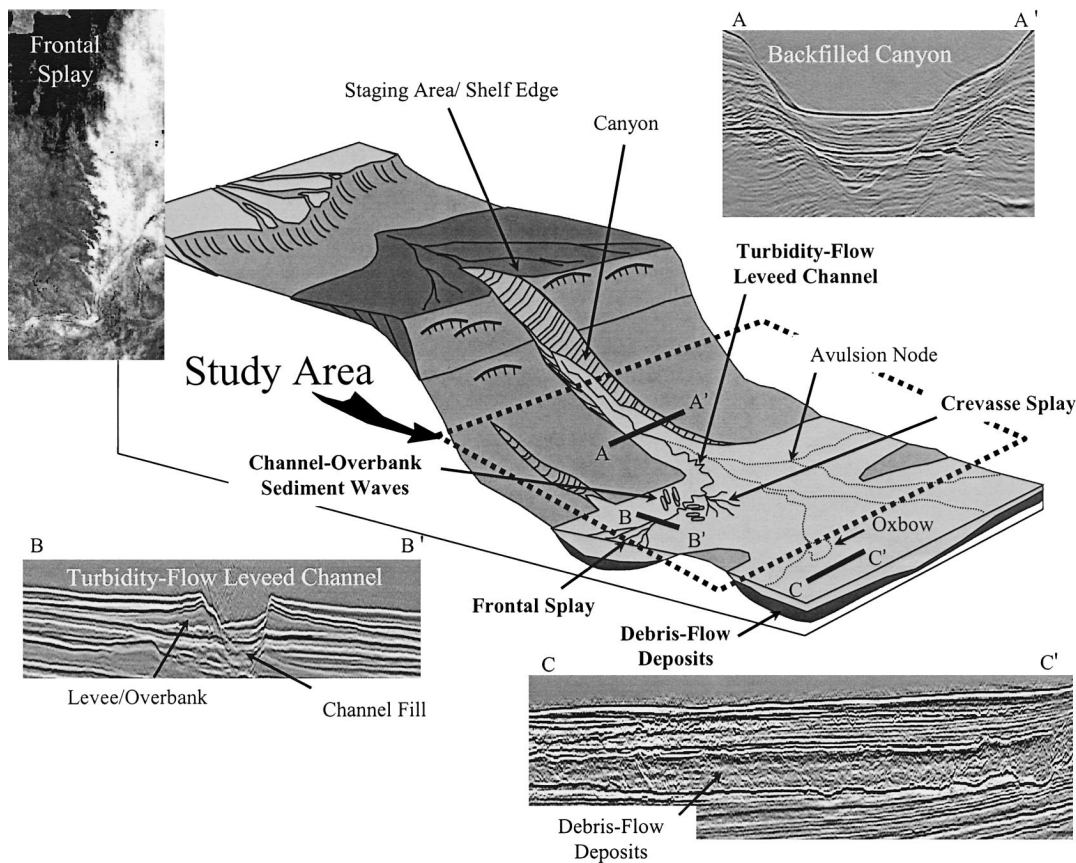


FIG. 1.—Schematic depiction of principal depositional elements in deep-water settings.

Mutti and Normark (1991) as *the basic mappable components of both modern and ancient turbidite systems and stages that can be recognized in marine, outcrop, and subsurface studies*. These features are the building blocks of landscapes. The focus of this study is to use 3D seismic data to characterize the geomorphology and stratigraphy of deep-water depositional elements and infer process of deposition where appropriate. Depositional elements can vary from place to place and in the same place through time with changes of environmental parameters such as sand-to-mud ratio, flow discharge, and slope gradient. In some instances, systematic changes in these environmental parameters can be tied back to changes of relative sea level. The following depositional elements will be discussed: (1) turbidity-flow leveed channels, (2) overbank sediment waves and levees, (3) frontal splays or distributary-channel complexes, (4) crevasse-splay complexes, and (5) debris-flow channels, lobes, and sheets (Fig. 1). Each element is described and depositional processes are discussed. Finally, the exploration significance of each depositional element is reviewed.

Examples are drawn from three deep-water slope and basin-floor settings: the Gulf of Mexico, offshore Nigeria, and offshore eastern Kalimantan, Indonesia. We utilized various visualization techniques, including 3D perspective views, horizon slices, and horizon and interval attribute displays, to bring out the detailed characteristics of depositional elements and their respective geologic settings. The deep-water depositional elements we present here are commonly characterized by peak seismic frequencies in excess of 100 Hz. The vertical resolution at these shallow depths of burial is in the range of 3–4 m, thus affording high-resolution images of depositional elements. We hope that our study, based on observations from the shallow subsurface, will provide general insights into the reservoir architecture of deep-water depositional elements, which can be extrapolated to

more poorly resolved deep-water systems encountered at deeper exploration depths.

DEPOSITIONAL ELEMENTS

The following discussion focuses on five depositional elements in deep-water environments. These include turbidity-flow leveed channels, overbank or levee deposits, frontal splays or distributary-channel complexes, crevasse splays, and debris-flow sheets, lobes, and channels (Fig. 1).

Turbidity-Flow Leveed Channels

Leveed channels are common depositional elements in slope and basin-floor environments. Leveed channels observed in this study range in width from 3 km to less than 250 m and in sinuosity (i.e., the ratio of channel-axis length to channel-belt length) between 1.2 and 2.2. Some leveed channels are internally characterized by complex cut-and-fill architecture. Many leveed channels show evidence of having grown by lateral and down-system migration, whereas other channels seem to have remained fixed in one location through extended periods and are characterized by vertical stacking (i.e., aggradation). In some instances, leveed channels are associated with overbank sediment waves. Many leveed channels are associated with frontal splays and crevasse splays (organized as distributary-channel complexes).

Channel-Floor Erosion and Deposition

At the base of a channel complex, most leveed channels are characterized by incision into the immediately underlying substrate (Fig. 2). The degree

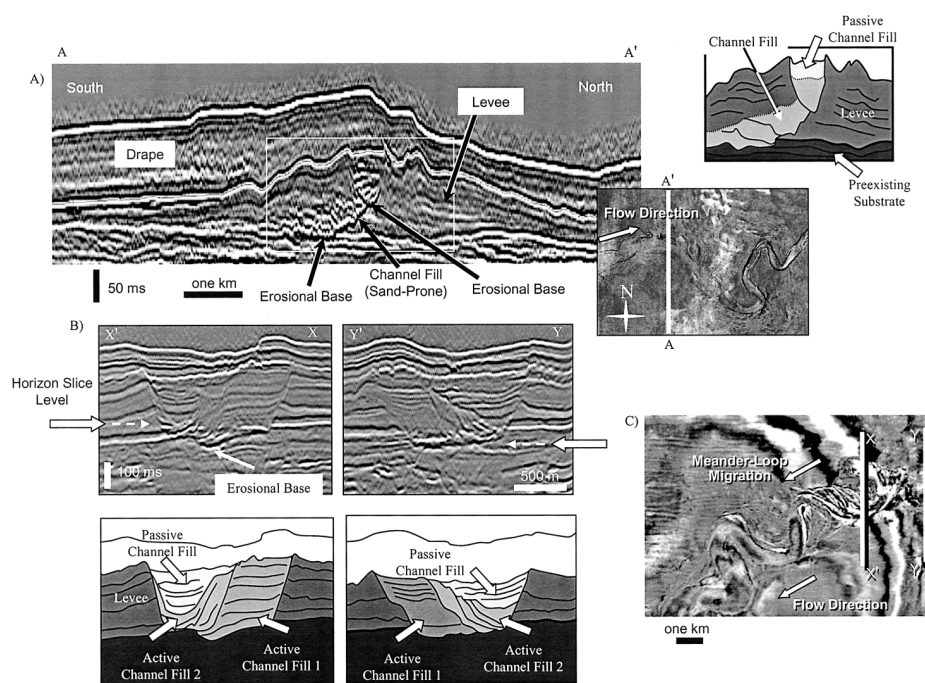


FIG. 2.—**A**) Transverse seismic profile across a turbidity-flow leveed channel offshore eastern Borneo, Kalimantan, Indonesia (map view at lower right indicates line location). The channelized deposits are expressed as high-amplitude reflections that are most likely indicative of the presence of sand-prone deposits. The channel facies are characterized by cut-and-fill architecture, and migrate both laterally, towards the outer channel bend, as well as upward. The base of the channel fill complex is characterized by erosion into the preexisting substrate. The inset contains an interpretation of the stratigraphic architecture. **B**) Transverse seismic profile across a leveed channel offshore Nigeria. Here also, the channelized deposits are expressed as high-amplitude reflections that migrate laterally as well as vertically, and are also erosional at the base of the channel complex. An interpretation of the stratigraphic architecture is shown on the inset. Two sets of levees are observed: master (channel belt) levees and inner (thalweg) levees. **C**) Horizon slice (i.e., flattened time slice) through leveed channel shown in part B illustrating down-system meander-loop migration.

of initial incision is variable (Figures 2A, B). In some instances the initial erosion is deep enough so that high-amplitude seismic reflections, inferred to be the seismic expression of sand-prone channel fill, do not come in contact with the levee walls. Rather, the sand-prone part of the channel fill can be fully confined within the incised part of the leveed channel (Fig. 2B). In other instances, the sand-prone part of the channel fill can in part be confined to the incision and in part be in contact with the levee walls. In still other instances there is only minimal incision, and the sand-prone part of the channel can be in complete communication with constructional levee walls (Fig. 2A). From an exploration or a reservoir compartmentalization perspective the extent to which the sand-prone part of a channel fill is in contact with the associated levee is important. If the sand-prone channel fill is fully confined within erosional walls, then these sands would not be in direct communication with associated constructional levees and could form isolated flow units. In contrast, in more aggradational systems, where the sand-prone channel fill is in direct communication with levee walls, the sands of the channel fill and those of the associated levees would form a single flow unit.

Channel-fill deposition commonly is aggradational, but in many instances it also is characterized by meander-loop migration. Figures 2 and 3 illustrate channels that are characterized by significant laterally as well as vertically accreted sand-prone fill. In plan view the lateral migration is expressed as down-system meander-loop migration as illustrated by horizon slices through the sand-prone part of the channel fills (Figs. 2, 3, 4). Some channels, however, show significant aggradation with subordinate lateral migration (Fig. 5).

Leveed channels can exist as isolated single entities on one extreme or as multi-stage channel complexes at the other. Figure 5A illustrates a relatively simple, single, leveed channel. Such channels can show minimal erosion at their base and can have an aggradational character that can result in them being perched as high as 100 m or more above the surrounding sea floor (Fig. 5B). Internal to these relatively simple systems, channel scour can be observed. In contrast with these simple channels, some channels exhibit multiple cut-and-fill architecture and associated channel rejuvenation (Fig. 6). In such instances, complex intercutting and nesting of channels (commonly confined to a master channel or valley) occurs and results in valley-confined, leveed-channel complexes (Fig. 6). In plan view,

such complex channels can have the appearance of meandering leveed channels within larger leveed channels or valleys. The fill of the narrow channel shown in Figure 6 at or just below the sea floor is characterized by relatively high reflection amplitude (Fig. 6B) suggesting the presence of sand within the thalweg of the channel. The overall channel fill is characterized by multiple channel cuts and fills, each characterized by the presence of high-amplitude reflections (and therefore likely sands) at the respective channel bases (Fig. 6).

Excellent images of leveed sinuous channels have been reported from the deeper subsurface (i.e., from exploration depths) based on 3D seismic analyses (e.g., Mayall and Stewart 2000; Sikkima and Wojcik 2000; Kolla et al. 2001). They illustrate that significant additional information regarding channel architecture can be derived from map view compared with section view alone, highlighting the need for integrating seismic geomorphological (plan-view interpretations) with seismic stratigraphic analyses (section-view interpretations). The level of detail to be seen in these deeper subsurface settings is nonetheless only a fraction of what can be imaged from the near subsurface, both in section as well as plan view. The interpretations of features in the deeper subsurface benefit greatly from insights derived from analyses of shallowly buried features that provide comprehensive analogs for the deeper ones.

Evolution of Channel Meander Loops

Leveed-channel planform patterns range from moderate to high sinuosity. Meander cutoffs are not common but have been observed in isolated instances (Fig. 7). Straight channels associated with levee construction were not commonly observed in the data sets analyzed.

It has long been recognized that many deep-water leveed channels on recent fans are of high to moderate sinuosity (Flood and Damuth 1987; Damuth et al. 1988; Clark et al. 1992). However, because only widely spaced 2D seismic data were utilized in the past, the temporal evolution of channel pattern remained unclear (Kolla and Coumes 1987; McHargue 1991). On the basis of closely-spaced 2D seismic data across the Mississippi channel, Stelling et al. (1985), Kastens and Shor (1986), and Peakall et al. (2000) suggested the presence of both lateral and down-system channel drift (coupled with aggradation), accompanying increased channel sin-

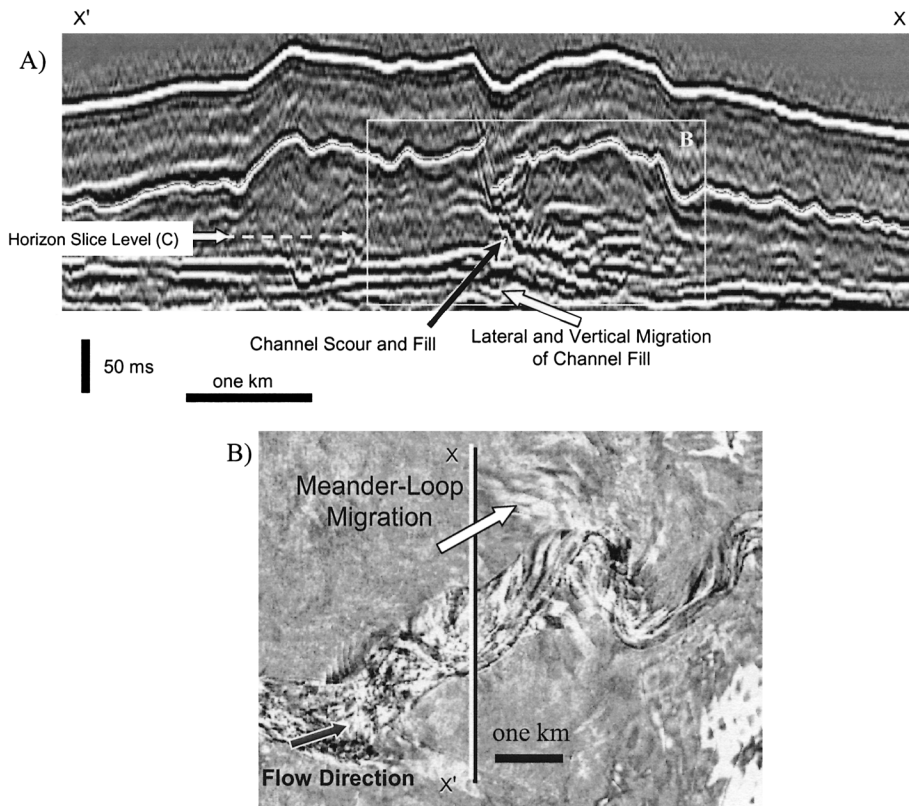


FIG. 3.—Example of a turbidity-flow leveed channel shown in Figure 8 characterized by both lateral channel migration and down-system meander-loop migration. **A)** Transverse seismic profile across the leveed channel offshore eastern Borneo, Kalimantan, Indonesia. Scour as well as lateral migration and aggradation characterize the channel fill. **B)** Seismic horizon slice through the leveed channel illustrating down-system meander-loop migration.

uosity through time. Striking similarities exist between deep-water channels and alluvial channels with respect to channel sinuosity versus terrain gradient, and the relationship between meander wavelength and both channel width and radius of meander curvature (e.g., Leeder 1999). However, morphological differences exist as well (Pirmez 1994; Pirmez and Flood 1997). In deep-water systems, channel widths and depths decrease down-system

in contrast with alluvial systems, where the opposite is true (Flood and Damuth 1987). In addition, the degree of channel aggradation is commonly far higher in deep-water systems than in fluvial systems (e.g., Imran et al. 1999; Kolla et al. 2001).

The differences between landforms associated with subaerial versus deep-water flows can be attributed largely to the far smaller density contrast

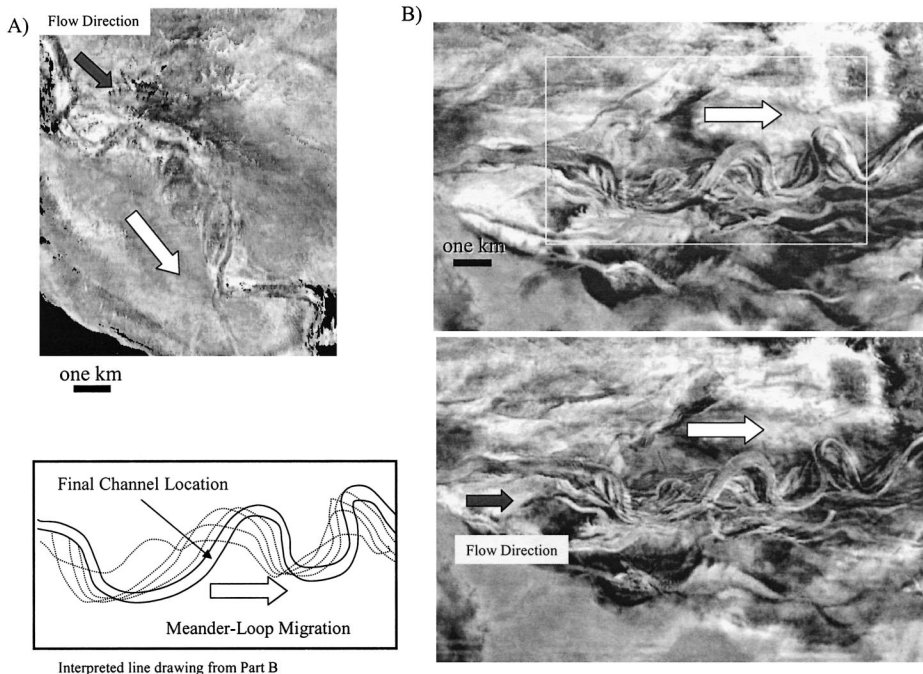


FIG. 4.—Seismic horizon slices through two leveed channels offshore Gulf of Mexico in the **A)** Green Canyon and **B)** De Soto Canyon areas, illustrating variations of down-system meander-loop migration (migration direction indicated by white arrow). The two horizon slices shown in Part B are approximately 10 m apart.

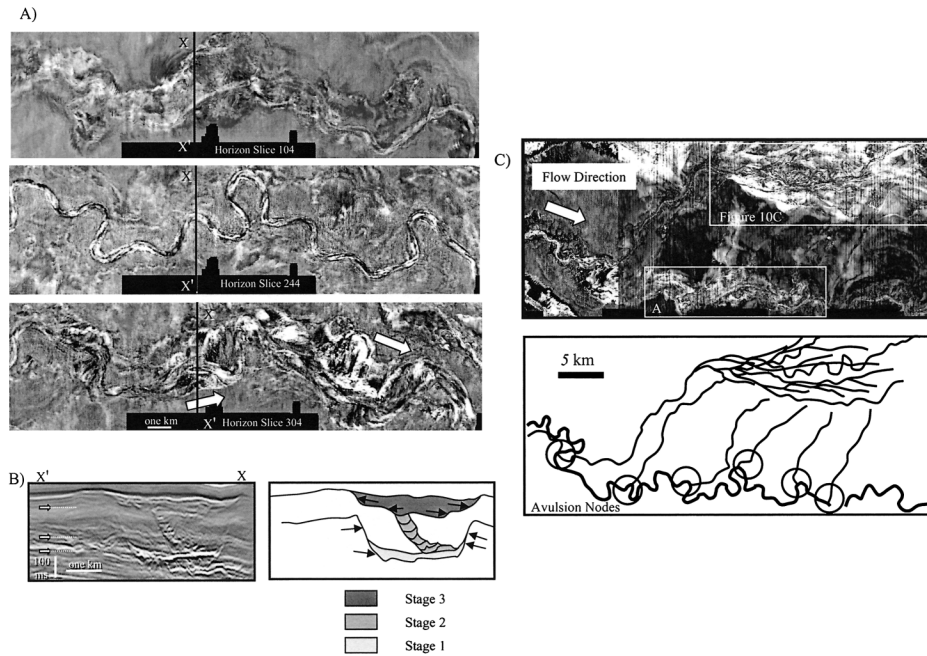


FIG. 5.—A) Three seismic horizon slices across a Pleistocene leveed-channel complex, De Soto Canyon area, offshore Gulf of Mexico, illustrating channel avulsions as well as progressive changes of channel pattern through time. Horizon slice 304 cuts through the lowest part of the channel complex (i.e., stage one) and illustrates a moderate-sinuosity channel characterized by significant aggradation and some lateral migration. Horizon Slice 244 cuts through the middle section of the complex (i.e., stage two) and illustrates a high-sinuosity channel characterized by significant down-system meander-loop migration. Horizon Slice 104 cuts through the central part of the complex (i.e., stage three) and illustrates a low- to moderate-sinuosity channel characterized by meander cutoffs and channel straightening. B) Seismic section across this channel complex illustrating three stages of channel evolution. C) Seismic interval attribute map showing multiple avulsion channels originating from this channel within the upper part of the complex. The overbank areas proximal to the avulsion channels are characterized by amplitude anomalies, suggesting the possible presence of sand-prone deposits there. The seismic attribute shown in Part C is the average amplitude for a 20 ms interval bracketing the horizon of interest. The westernmost avulsion channel is observed to feed a frontal-splay deposit (see Fig. 10). Inset A corresponds to Part A of this figure.

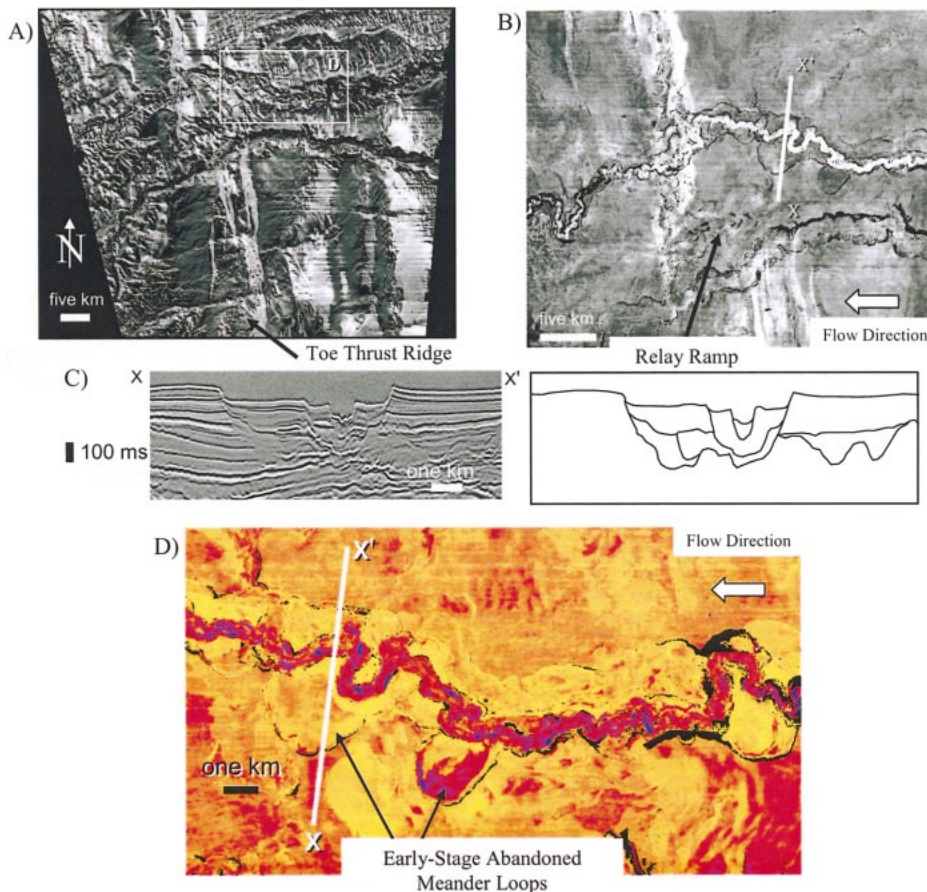


FIG. 6.—Late Pleistocene leveed channels offshore Nigeria in plan view (A, B, and D), and section view (C). The plan-view images are a sea-floor reflection dip azimuth map (A), sea-floor reflection amplitude map (B), and interval attribute (maximum negative amplitude) map of a near sea-floor interval 100 ms thick (D). Image D is a detail of image A, the location of which is shown on the inset in Part A. The dip azimuth map illustrates the physiography whereas the reflection amplitude and interval attribute maps provide insight to the lithology. Apparent early-stage meander loops are incised by later rejuvenation of the channel system. The latest-stage meandering leveed-channel system (D) seems to be fully confined within the larger early stage channel feature (C). The long linear slope-parallel features, which influence channel locations, correspond to the surface expression of toe thrust structures.

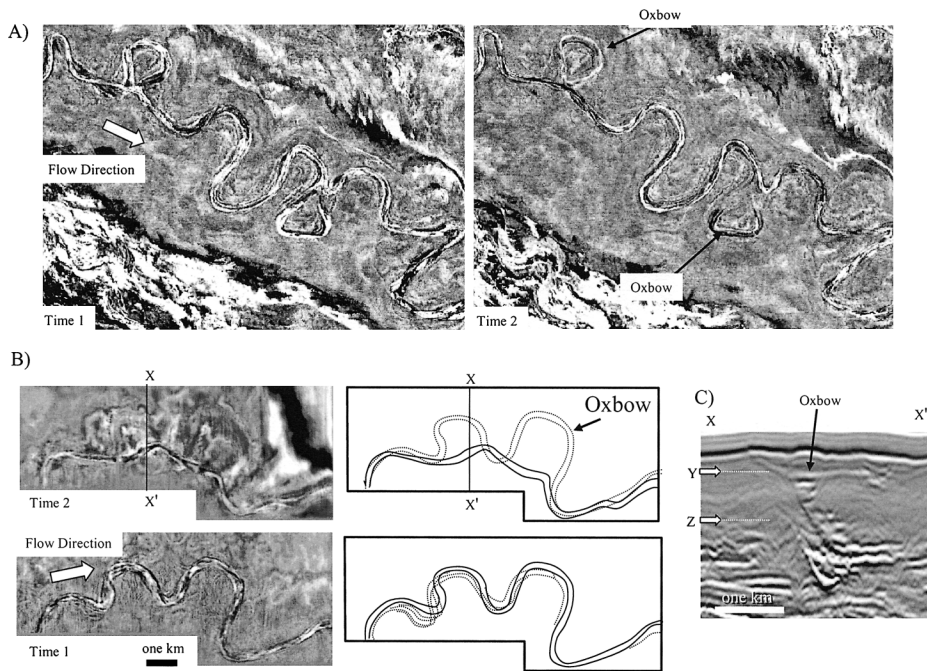


FIG. 7.—Examples of Pleistocene meander-loop cutoffs in the De Soto Canyon area of the Gulf of Mexico. **A, B**) Cutoff meander loops of a high-sinuosity leveed channel, forming two oxbows. **C**) Seismic profile illustrating a section view of the meander cutoff. This is the same channel system as that shown in Figure 5.

and greater friction at the upper boundary of turbidity flows (where they come in contact with overlying lower-density, ambient water) compared with the much greater density contrast and negligible friction between sub-aerial flows and air above (Imran et al. 1999; Peakall et al. 2000; Kolla et al. 2001). The observed morphological similarities, however, suggest that channel-pattern evolution, with both types of flow, is the result of a process that involves dynamic interaction of flows, sediments, and terrain gradient (e.g., Leopold and Wolman 1957; Kneller et al. 1999; Peakall et al. 2000; Kolla et al. 2001).

Several deep-water, sinuous-channel systems have been described above (Figs. 2, 3, 4, 5, 6, 8). The images show that these channels are commonly characterized by meander-loop migration, in most instances with down-system drift accompanying overall aggradation, as the channel pattern evolved through time. Seismically detectable scour and fill accompanying channel drift commonly can be observed, suggesting that turbidity currents of various velocity and magnitude were involved. Overall channel-fill sedimentation was punctuated by episodic erosional events. In most instances this resulted in amalgamation of high-amplitude channel-fill reflections. In some instances, channel migration through time is manifested by progressive expansion of meander loops, resulting in increased sinuosity (Fig. 9).

In isolated instances, we encountered highly sinuous channels characterized by significant aggradation and some lateral migration, but with minor down-system migration of meander loops. One such example is shown in Figure 5A (horizon slice 244). This channel system seems to have experienced three stages of development, with each stage change likely attributable to a major change of flow regime and type of sediment load. Stage one is characterized by moderate- to high-sinuosity channels (average sinuosity of 1.6), ca. 0.5 km wide, which exhibit distinct down-system meander-loop migration and high-amplitude seismic reflection character (Fig. 5A, Horizon Slice 304) within a larger valley (Fig. 5B). We interpret this stage to be characterized by higher-energy turbidity-flow deposits than during subsequent stages. The middle stage (stage two) is characterized by high-sinuosity channels accompanied by significant aggradation and significantly less meander-loop expansion and down-drift migration than during stage one. What distinguishes this stage of the channel system from stage one is (1) its high sinuosity (its sinuosity of 2.2 is the highest of all channels observed), and (2) the nearly reflection-free character of the as-

sociated levee deposits. Stage two of this system appears to be an underfit channel; the channel is significantly narrower than that of stage one, and the channel as well as its associated levee are fully confined within the levees formed during stage one. This apparent underfit situation likely was associated with a decrease in flow volume and possibly lower energy turbidity flow deposits than during stage one. The channel is significantly narrower at this second stage and the channel-levee complex is, at least initially, fully confined within the precursor valley system.

Schumm (1963) suggested that for river flows highly sinuous channels tend to be more muddy. Although we have no borehole calibration and have limited number of examples upon which to draw conclusions with regard to the relationship between channel pattern and grain size or flow parameters, we nonetheless speculate that this system is relatively mud-rich on the basis of the seismic reflection character of the overbank deposits. A reflection-free or transparent seismic facies suggests sediments of relatively uniform acoustic property. It is reasonable to assume that rather than consisting of variable lithologies (i.e., sand and mud) that happen to be characterized by nearly identical acoustic properties, the overbank more likely consists of deposits of uniform lithology. Following on this, a relatively uniform mud composition for the overbank seems a reasonable inference. Whether by extension one can conclude that those systems characterized by significant meander loop migration (such as illustrated in stage one of Fig. 5) are more sand prone is not yet sufficiently clear.

Stage three (Fig. 5) is characterized by a more difficult-to-image channel system. However, where a channel can be imaged within this stage, it is characterized by markedly lower sinuosity and in places forms meander cutoffs and oxbows (Fig. 7). This channel-levee system is progressively abandoned by virtue of a succession of avulsion events (Fig. 5) (see discussion below). The presence of meander cutoffs and oxbows as well as avulsion channels suggests that stage three likely was associated with a significant increase in flow volume and an overfit channel.

Meander patterns can be significantly influenced by sea-floor gradient (Flood and Damuth 1987; Pirmez and Flood 1997; Clark and Pickering 1996a, 1996b). Structural features such as toe thrusts and salt domes can have expression as sea-floor morphological elements, and can exert first-order control on channel location, channel pattern, and associated reservoir sweet spots. Although it is not a focus of this paper to address the issue

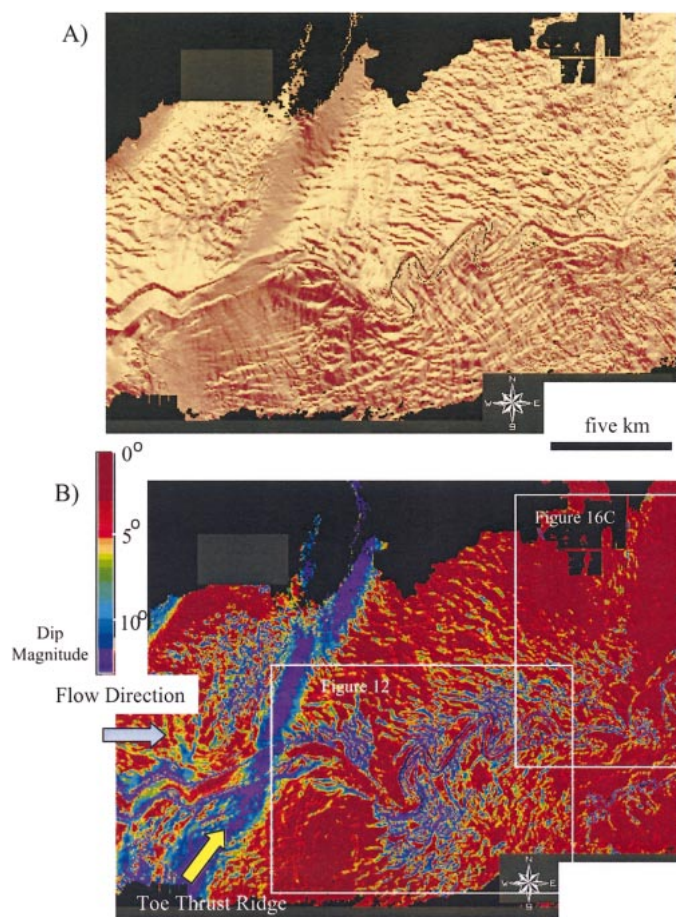


FIG. 8.—**A)** Horizon dip azimuth map and **B)** dip magnitude map of the top of a Pleistocene leveed-channel system, offshore eastern Borneo, Kalimantan, Indonesia. The dip azimuth map extracted from the upper bounding surface of the channel–levee complex illustrates the physiography along this horizon. The dip magnitude map illustrates the slope variations along this horizon, highlighting such features as channel edges, structural sills, and sediment waves. Note the marked change of channel morphology just down system from the significant gradient break shown on the dip magnitude map as a broad blue swath oriented SSW–NNE; the channel width narrows from 1 km to less than 0.25 km and the sinuosity increases markedly. The linear ridge-like features on either side of the channel are interpreted as sediment waves (see Fig. 12). The strike-parallel lineaments on the south side of the channel are interpreted as erosional furrows likely related to ambient bottom currents.

of structural and/or physiographic control on depositional-element morphology, two examples presented here do bear mention. The channel–levee system shown in Figure 8 illustrates the effect of an uplifting structural sill (i.e., a toe thrust ridge) on channel morphology. The sea-floor gradient inboard as well as outboard of this steep slope is approximately 0.6° , whereas the slope of the steep segment today is approximately 9.4° . The slope of this steep section likely has been enhanced by recent structuring; the slope at the time the channel was active likely was less than today's 9.4° . This channel is characterized by low sinuosity (i.e., 1.1) up-system and moderate to high sinuosity (i.e., 1.6) down-system of the sill. Furthermore, the channel width decreases markedly down-system of the sill, from approximately 1 km to less than 0.25 km in width. The absence of any tributary or distributary channels at this location suggests that the sill alone has induced this effect. Likely, the cause of these morphological changes is the accelerated flow that occurred as the flow traveled down this short but steep slope. Rivers have been shown to respond to increased slope by increasing their channel depth and their sinuosity (Schumm 1993). A similar cause and effect is postulated for the deep-water channel system shown

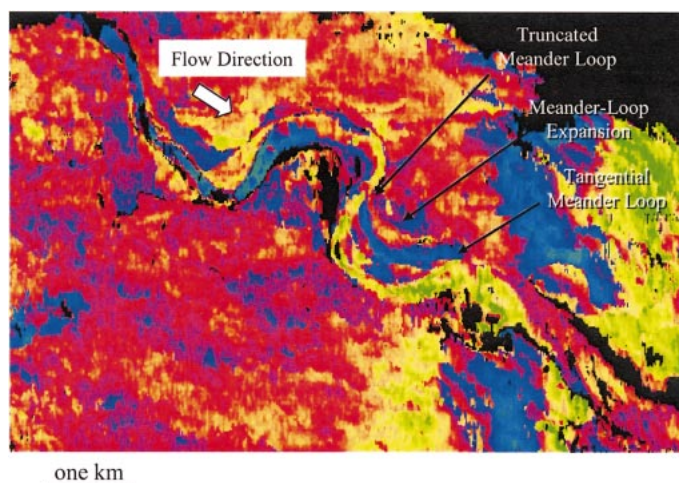


FIG. 9.—High-sinuosity leveed channel characterized predominantly by meander-loop expansion rather than down-system migration in the Green Canyon area of offshore Gulf of Mexico. This is in contrast with the predominantly down-system meander-loop migration illustrated in Figures 2, 3, 4, and 5. Minor down-system meander-loop migration is indicated by the truncated meander loop on the up-system end and tangential meander loop on the down-system end.

here. Pirmez and Flood (1997) infer a similar relationship on the middle Amazon fan.

Another example of structural effects on channel morphology is illustrated in Figure 6. In this slope setting, channels are influenced by the sea-floor expression of toe thrusts, which form ridges transverse to flow direction. The channels appear to be clustered in the area where one toe thrust ridge dies out and another begins, forming a relay-ramp morphology. In this way, the channels sought out the pathway of lowest topography. Some smaller channels (i.e., those to the south, Figs. 6A and B) reach toe thrust ridges and parallel them for a certain distance before breaching them and continuing down slope.

Exploration Significance

The channel-fill deposits in the examples studied here are characterized primarily by high-amplitude seismic reflections and are interpreted to be sand-rich. The areal extent, the distribution of these sand-prone deposits, and their reservoir architecture are dependent to some degree on the extent of meander-loop migration. Channel-fill deposits not associated with meander-loop migration are preserved as narrow ribbons of reservoir facies, the width of a single channel. Reservoir facies in these channel systems can, however, be vertically stacked. In contrast, channel-fill deposits associated with meander-loop migration produce a swath or fairway of reservoir facies commonly several times the width of a single channel (Figs. 2, 3, 4, 5, 6, 8, 9).

Mud-rich systems may be associated with narrow, aggradational, sand-prone channels. Within such mud-rich, lower-energy systems some preservation of shale baffles to fluid flow may occur, in contrast with the comparative rarity of such deposits in higher-energy migrational settings. Another related effect is the muddiness of associated levees and therefore the absence of reservoir quality in the overbank facies. In addition, associated frontal splays tend to be relatively small and less well developed, given the paucity of sand.

The incision that characterizes the initial phase of nearly all meandering leveed-channel systems suggests that at least some of the channel fill lies below the surface into which the channel has been incised. Associated levee deposits may lie significantly higher and in some instances might not be in contact with the sand-prone, basal channel fill (e.g., Figs. 2A and B;

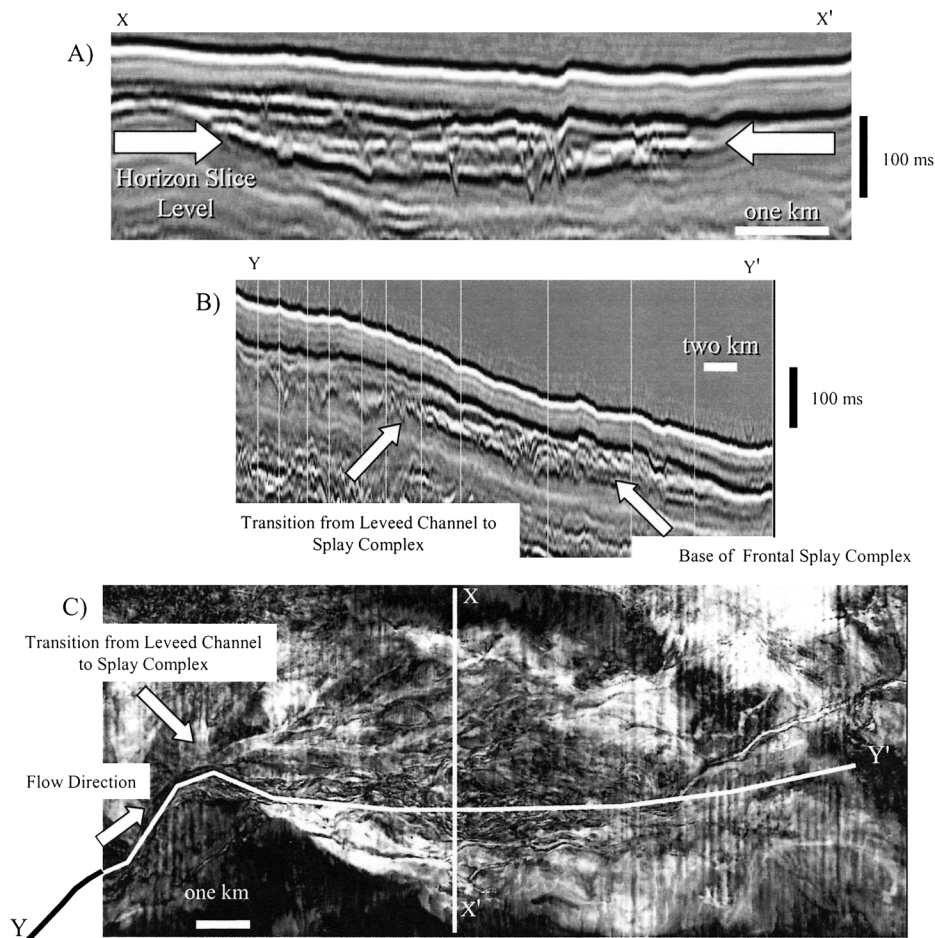


FIG. 10.—Frontal-splay deposits in the De Soto Canyon area of the Gulf of Mexico (see Fig. 5 for location), where there is no apparent gradient break at the transition from turbidity-flow leveed channel to frontal splay. The splay complex is shown both in A, B) section and C) plan view. The seismic facies pattern of the frontal-splay deposits is characterized by high-amplitude continuous reflections (A) and a distributary to braided channel pattern (B). The transition from leveed channel to frontal splay is illustrated in the traverse in part B, which tracks the axis of the leveed channel through the transition zone and across the frontal splay.

Clemenceau et al. 2000). In those instances, there is potential for stratigraphic trapping and compartmentalization of reservoir units. This can occur both within levee walls, where levee deposits would be sealed by mud-prone passive channel fill deposits, and within channel fill deposits, where channel sands would be juxtaposed against previously deposited mud-prone substrates.

Meandering Channels and Avulsion

The avulsion process has been documented in a variety of deep-water settings (Damuth et al. 1983; Damuth et al. 1988; Kolla and Coumes 1987; Manley and Flood 1988; Pirmez 1994; Kenyon et al. 1995). In the areas analyzed here only one clear example of a series of channel avulsion events can be documented. Figure 5A (Stage Two, Horizon Slice 244) illustrates a high-sinuosity leveed channel that is characterized by extensive aggradation accompanied by minimal down-system meander loop migration. Aggradation of over 150 m of channel–levee sediment can be observed. Near the top of this aggradational system an abrupt change in depositional style can be observed; several new channels developed as a result of levee breaching at avulsion nodes along the original high-sinuosity channel (Fig. 5C). The cause of avulsion most likely was a change in flow parameters associated with changes higher up on the slope or outer shelf (i.e., the staging area). Such changes could have included increased flow discharge and/or increased sand-to-mud ratio (Pirmez et al. 2000).

These avulsion channels likely formed systematically, with the most down-system channel forming first, and avulsion occurring progressively farther up-system shortly thereafter. The evidence for this order of events is somewhat circumstantial. Presumably once an avulsion channel forms,

the flow discharge within the main channel down-system of the avulsion node will be significantly decreased. Consequently, any additional avulsion events are more likely to occur up-system of the first avulsion node, where flow discharge has not yet been affected. The result is a succession of avulsion nodes along the main channel, with the youngest located most proximally and the oldest, most distally.

The most up-system avulsion channel observed seems to have been the longest lived, as evidenced by the recognizable, well-developed splay complex that it feeds (Figs. 5C, 10). It is noteworthy that this splay complex was not deposited in close proximity to the levee crest of the parent channel, but rather, significantly farther down-system, approximately 25 km away from the avulsion node. Similarly, the splay complex described by Twitchell et al. (1992) lies tens of kilometers away from the associated avulsion node of the parent deep-water Mississippi channel.

Exploration Significance

In the areas examined, the relative rarity of avulsion events (with the exception of those small-scale avulsions that characterize frontal splays and crevasse splays) is striking. Although avulsion events clearly must be part of overall fan construction (Flood et al. 1991; Kenyon et al. 1995; Pirmez and Flood 1997; Pirmez et al. 1997), avulsion frequency appears to be sufficiently low that for areas of the sizes involved in this study the presence of avulsion features would be uncommon. However, where they are present, avulsion channel fills and associated frontal splays, expressed as high-amplitude reflection packages (i.e., the ‘‘Harps’’ of Pirmez and Flood 1997), may comprise good reservoir lithologies (Pirmez et al. 1997).

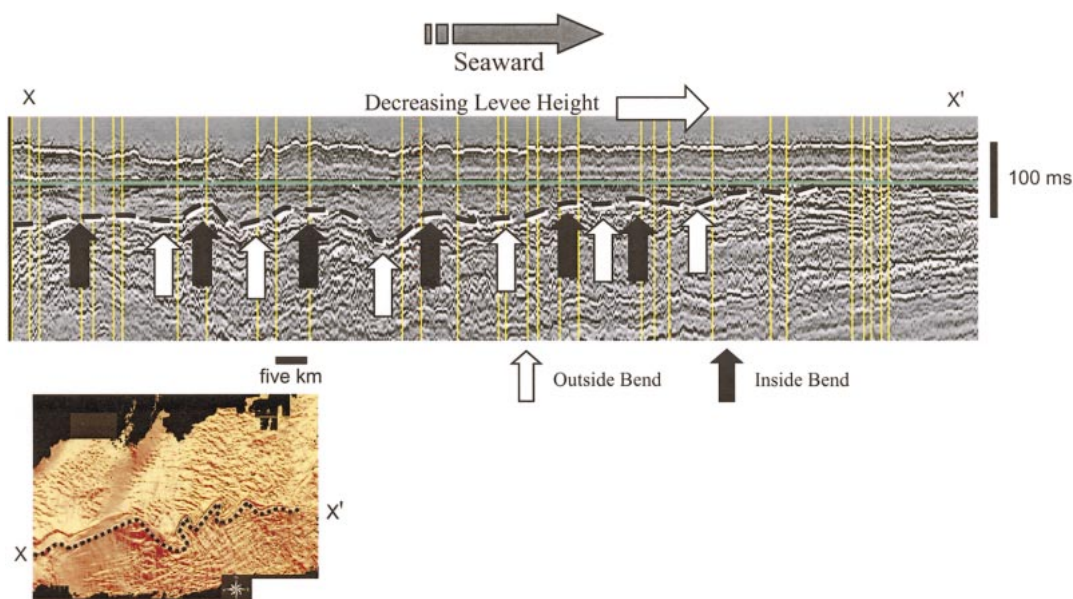


FIG. 11.—Seismic traverse tracking along the right-hand levee crest of the leveed channel shown in Figure 8, flattened at the levee top. The seismic facies of the levee is characterized by low-amplitude discontinuous seismic reflections. The thickness of the levee wedge is variable, being thicker along outer channel bends and thinner along inner channel bends. Overall, the levee thickness progressively decreases down system, approaching the limit of seismic resolution towards the end of the section.

Overbank or Levee Deposits

Aggradational deep-water channels are associated with levee-overbank construction. Such levees commonly are an order of magnitude or more wider than their associated channels (Fig. 2A). Levee height above the channel floor can be highly variable along the course of a channel, with levee height varying by as much as a factor of two from inner to outer bends of a given channel (Fig. 11; Posamentier 2001a). Levees commonly diminish in height from their landward to seaward reaches (Fig. 11). Levee thickness in highly aggradational systems can be upwards of 150–200 ms (equivalent here to approximately 120–160 m) (Fig. 5). In contrast, in highly erosional systems, levees can be completely absent and channelized flow can be completely confined by the erosional walls of the channel (or canyon).

Some levees are characterized by extensive sediment-wave development, whereas in other systems sediment waves are not present. Figure 8 illustrates sediment waves developed on levees on both sides of a Pleistocene leveed channel in the Makassar Strait offshore eastern Kalimantan, Indonesia. The trends of the sediment-wave crests and troughs are subparallel to the respective adjacent channel bends. Note that the sediment waves are best developed outboard of the outer bends of the channel (Fig. 12B). The area of well-developed sediment waves that characterizes the proximal overbank is approximately four by four kilometers. As Figure 12A illustrates, sediment waves can be observed through the entire overbank depositional unit, suggesting that sediment-wave development is an integral part of levee construction. Seismically, levees are characterized by low- to moderate-amplitude, continuous to discontinuous reflections, and in some instances are completely transparent. Fine to medium sand grain size has been reported by Migeon et al. (2000) for similar sediment waves observed on the Var system along the Mediterranean coast.

Typical dimensions of overbank-related sediment waves as illustrated by those observed offshore Nigeria and Indonesia (Figs. 12, 13) are up to 20 m high with wavelengths of 0.5 to 0.8 km. Some sediment waves show an apparent upstream accretion (Fig. 13). In Figure 8A we note the presence of strike-parallel lineaments, transverse to the sediment waves on the south side of the channel. These may be sediment waves associated with later

currents reworking the overbank top, or alternatively may be erosional furrows.

Sediment Waves on Levees

Sediment waves related to turbidity-current, overbank flow in deep-water settings are relatively common on the modern sea floor (Damuth 1975; Normark et al. 1980; Twichell et al. 1991; Piper and Savoye 1993; Migeon et al. 2000; McHugh and Ryan 2000; Wynn et al. 2000a; Wynn et al. 2000b). Likewise, sediment waves have been observed on levees in each deep-water setting studied here. Normark et al. (1980), Piper and Normark (1983), and Migeon et al. (2000) suggested that these features form in response to spillover and flowstripping from channels of the predominantly dilute fine-grained upper part of turbidity flows. These authors discussed the flow parameters and turbidity-current conditions and mechanism of sediment wave formation in detail. Figure 14 illustrates this process schematically. The part of the flow that travels higher than the levee constitutes the sediment budget available for levee construction. As the flow travels progressively farther down-system, the flow steadily becomes impoverished of its original mud content by flowstripping and general overbank spillover. With diminishing sediment budget available for levee construction, levee height steadily decreases down system (Hiscott et al. 1997).

Figure 12 illustrates overbank sediment waves on the levee of a Pleistocene channel. On this dip magnitude map the sediment waves are best developed along levees just outboard of outer channel bends, similar to sediment waves reported by McHugh and Ryan (2000) from the overbank areas of the Monterey fan. This observation is consistent with the notion that these waves are deposited by turbidity flows escaping the confinement of the leveed channel by the process of flowstripping towards outer bends of channels due to superelevation of turbidity flows resulting from centrifugal forces (Piper and Normark 1983; Imran et al. 1999). Transverse profiles illustrate that these sediment waves progressively diminish in height with distance from the levee crest. The furrows or ridges oriented transverse to the sediment waves on the south side of the leveed channel shown in Figure 8A may be related to bottom currents (Mutti 1992).

Figure 13 illustrates sediment waves in the ultra-deep offshore Nigeria.

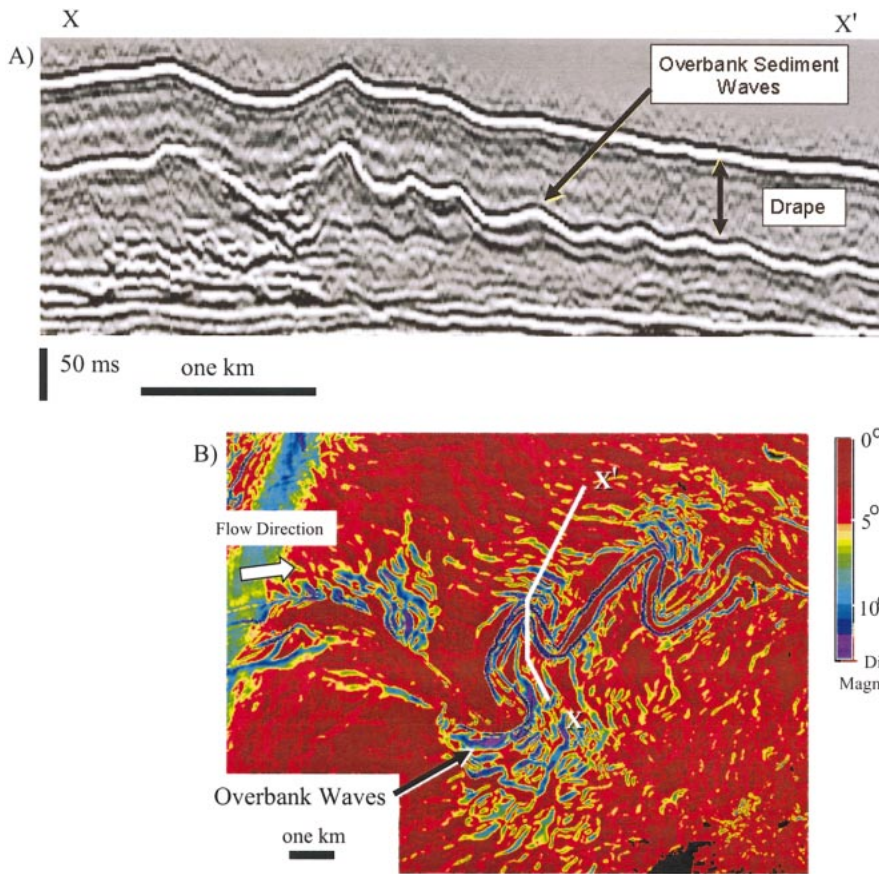


FIG. 12.—A) Seismic section oriented normal to a sediment wave field, offshore eastern Borneo, Kalimantan, Indonesia. Sediment waves are observed within as well as at the top of the levee wedge. B) Seismic reflection dip magnitude map of the associated channel–levee complex. The largest-scale sediment waves are distributed as clusters or fields in close proximity to leveed-channel outer bends (see Fig. 8 for location).

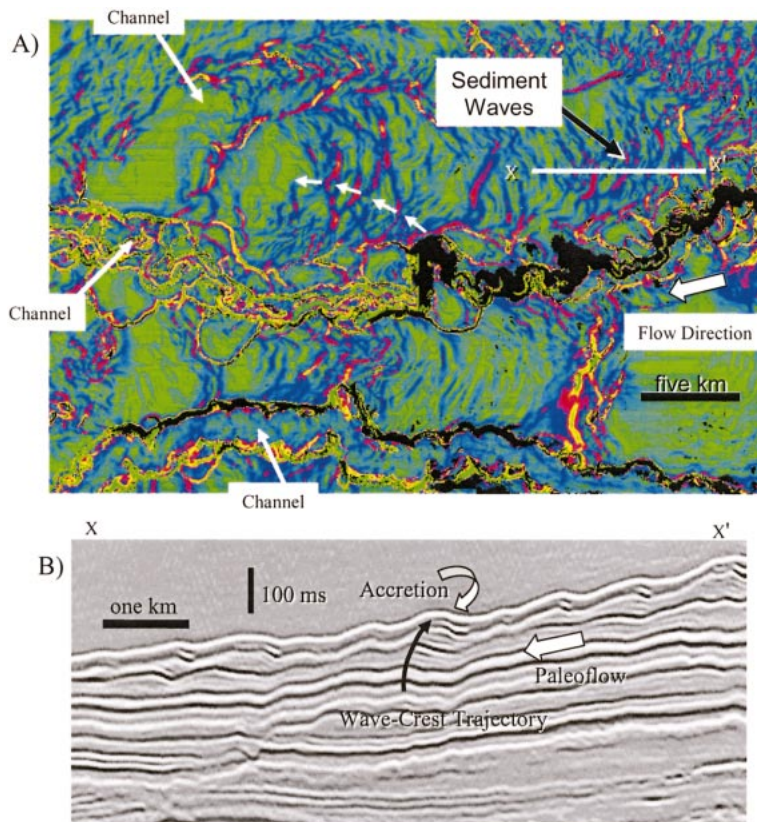


FIG. 13.—A) Dip magnitude map view and B) section view across sediment waves associated with overbank turbidity current flow offshore Nigeria. Line location is shown in Part A. The seismic section illustrates accretion on the upslope-facing sides of the sediment waves. The trajectory of the sediment wave crests is characterized by an upward convex path, supporting an interpretation of an aggradational and depositional origin rather than a rotated-fault-block or slump origin for these features. Small white arrows indicate overbank flow direction away from the channel.

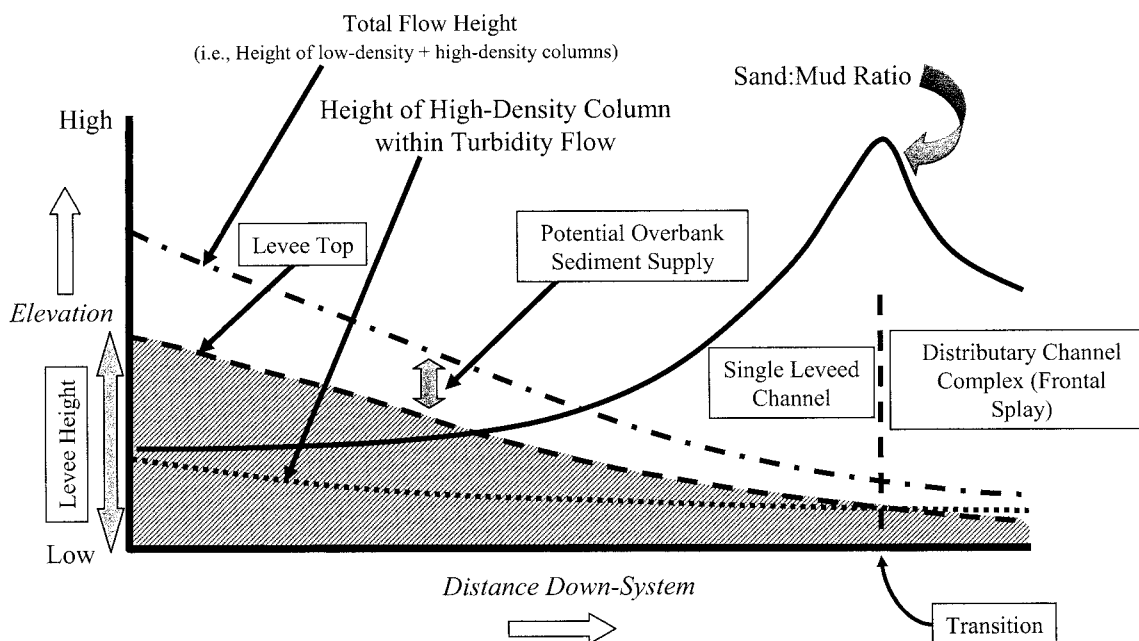


FIG. 14.—Schematic and conceptual depiction of the relationship between total turbidity-flow height, height of the high-density part of the flow, and levee height. The transition from leveed channel to frontal splay occurs where the curve of high-density flow height crosses the curve of levee height. The flow sediment budget for the levee construction is represented by the space between the total-flow-height curve and the levee-height curve. Because of the flowstripping process, much of the decreased total flow down system is associated with decreasing volume of mud-prone sediments, which are prone to overbanking, rather than sand-prone sediments, which are largely channel confined within high-density flows. Because of decreased volume of muddy sediments down system, the net sand within the total flow increases, reaching a peak at the transition between leveed channel and frontal splay. Down system from the transition, accelerated deposition of high-density deposits results in decreasing net sand within the total flow.

These sediment waves are oriented oblique to the associated channel axis and are interpreted to be associated with overbank flow out of the adjacent channels. The waves are arcuate in plan form as a result of flow vectors that are directed away from the channel in close proximity to the levee crest, turning progressively to face down regional dip with distance away from the levee crest (Fig. 13). The thickened section occurs at the stoss side of each sediment wave (Fig. 13), resulting in migration of these waves in an antidune-like, up-slope direction (compare with Normark et al. 1980).

Exploration Significance

Proximal overbank, or levee, deposits have been documented to contain reservoir-quality thin-bedded sandstones (e.g., Piper and Savoye 1993; Clemenceau et al. 2000; Migeon et al. 2000). These types of thin-bedded sandstone deposits have been described as “low-resistivity” pay zones (Clemenceau et al. 2000; Kendrick 2000). Such pay zones would be best developed in proximal levee settings, especially along outer channel bends (Pickering et al. 1989; Clark and Pickering 1996a; 1996b; Manley et al. 1997). In addition, in northern and southern latitudes, where Coriolis effects can be significant, preferential deposition of sand-prone overbank can be anticipated along right and left banks, respectively (Menard 1964; Chough and Hess 1980). The distribution of reservoir-quality sands might also be uneven owing to the apparent thickening of section on the stoss sides of sediment waves, resulting in a possible “striping” of sand isopach thicks in plan view, normal to the direction of overbank flow. Because of the extensive lateral extent of shale beds and inferred minimal erosion by successive unconfined flows within the overbank environment, significant barriers and baffles to vertical fluid migration likely characterize these deposits (Hiscott et al. 1997).

Frontal Splay, Lobeforms, or Distributary-Channel Complexes

Some levee-confined channels feed relatively unconfined splay complexes (also referred to as sheet sand deposits; Hackbarth and Shaw 1994;

Mahaffie 1994) that commonly display a distributary-channel pattern. The transition from leveed channel to splay complex is associated with a marked reduction in channel width, channel depth, channel sinuosity, and levee height. Figures 10, 15, and 16 illustrate examples of splay complexes from a variety of settings. In some instances, splay complexes seem to be associated with significant breaks in slope (i.e., abrupt decreases in gradient) (Figs. 16A, B), whereas this is not so in others (Figs. 10, 16C). The distribution of splay complexes tends to reflect the irregularity of the floor of the basin within which they are deposited. In settings characterized by relatively featureless sea-floor topography, splay complexes tend to develop an oblong or “paddle” shape (Figs. 10, 14C). Splay-complex thicknesses of up to 65 m and widths of up to 10 km have been observed. Seismically these deposits are characterized by high-amplitude, continuous to slightly discontinuous reflections. In cross-section view, these deposits appear relatively flat topped.

Transition from Leveed Channel to Frontal Splay

In many instances, leveed channels can be observed to directly feed lobes or lobe complexes, herein referred to as *frontal-splay complexes*. The term *frontal-splay complex* is used here because it implies a process of flaring out from a confined to a largely unconfined system. This splay process commonly produces a lobate planform. The term *frontal-splay complex* therefore implies both a process of formation as well as a morphometric plan shape. The term *distributary-channel complex*, also commonly applied to these deposits, implies the presence of a distributary-channel pattern, which is a product of the splay process (Figs. 10, 16) (Twichell et al. 1991; Twichell et al. 1992). However, in some instances anastomosing or braided patterns can characterize splay complexes (Kenyon and Millington 1995). Consequently, the more general term *frontal-splay complex* with its process connotation is preferred here. From a lithologic perspective, these deposits likely are synonymous with the sheet sands of Mahaffie (1994) and Hackbarth and Shaw (1994).

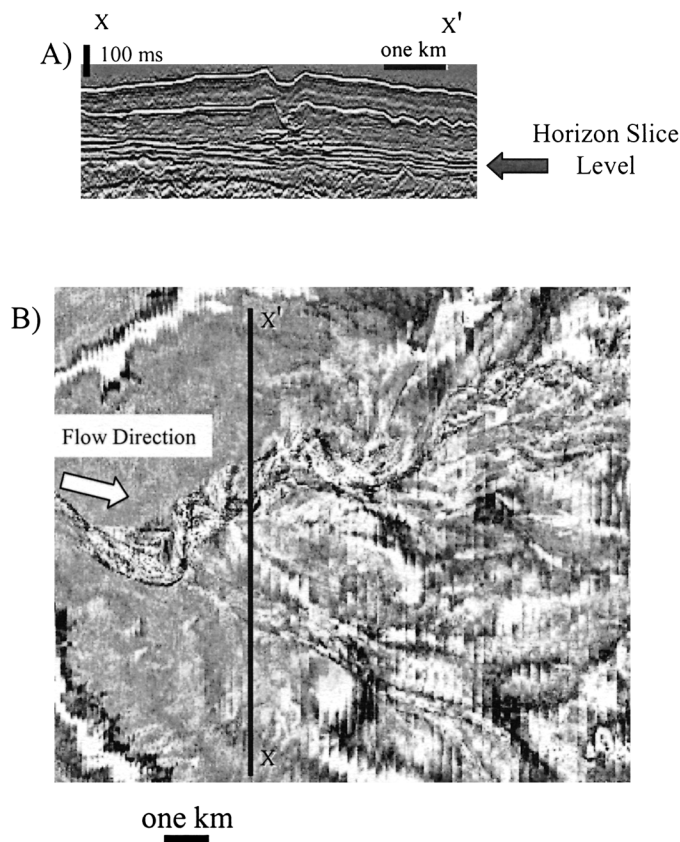


FIG. 15.—A) Section view and B) horizon slice map view of frontal-splay complex offshore eastern Borneo, Kalimantan, Indonesia, with the distinctive distributary-channel pattern that characterizes this feature (B). The seismic section illustrates the distinctively high-amplitude continuous reflection pattern that commonly characterizes frontal-splay deposits.

The transition from relatively confined flow within leveed channels to relatively unconfined flow within splay complexes can be related to (1) a progressive lowering of levee height down system to where the high-density part of the turbidity flow is no longer effectively confined, or (2) a marked reduction in channel gradient. The precise location of this transition in the former case is primarily the result of turbidity-flow characteristics including sand-to-mud ratio, flow volume and height, and flow velocity. All else being equal, where no marked reduction in slope gradient is present, such as would characterize basin floors, *the more mud rich a flow, the farther down system the transition from confined leveed channel to splay complex*. Such a shift of transition down system would be associated with a progressive overall grain-size decrease in the supply system at the associated updip shelf margin (i.e., at the staging area).

Figure 14 schematically illustrates the relationship between total flow height, the height of the high-density part of the flow, and levee height, and the resulting transition between leveed channel and frontal splay in the absence of a break in slope gradient. As a result of successive overbank events that tap into the upper parts of turbidity flows as these flows travel basinward, flows progressively are impoverished of their low-density, mud-prone component as they travel seaward. Coincident with this there is comparatively minimal loss of sand within these flows, because the high-density part of the flow remains effectively confined by levee walls. What loss of sand there is would be associated with sedimentation at the channel base and with mixing with the upper low-density part of the flow and subsequent dispersion into overbank areas. Volumetrically, flows progressively diminish in magnitude down system and, conversely, flow sand-richness increas-

es down system. Kolla and Coumes (1987), Pirmez and Flood (1997), and Hiscott et al. (1997), have observed that with increased distance down-system there is a gradual increase of net sand deposited on levees, consistent with a progressive impoverishment of mud within flows in the down-system direction. Levees, which depend upon overbanking for their construction, gradually diminish in height down-system as flows become progressively more sand rich. The top of the high-density part of the flow lies progressively closer to the levee top with distance down-system. At some point the levees are not sufficiently high so as to effectively confine the high-density part of the flow, whereupon the flow splays out radially, producing a distributive or braided pattern, similar to that which characterizes deltaic distributary-mouth bars. Note that an abrupt change in slope does not necessarily characterize this transition. An example of such a transition from leveed channel to frontal splay is illustrated in Figure 10.

The transition from confined leveed channel to splay complex also can be facilitated by a marked break in slope (Fig. 16B). Where flows hit a break of slope, flow vectors with increased magnitude are directed both downward and outward onto the channel floor and against the levees, respectively. Levee construction at this location suffers from the increased stresses directed outwards. At this location, also, increased downward-directed stresses and consequent increased turbulence results in erosion and are associated with the occurrence of a hydraulic jump (Komar 1971). Consequently, the loss of levee height, coupled with reduced flow velocities associated with the hydraulic-jump process, results in rapid deposition of sands down system of the break in slope. This increased rate of deposition tends to produce shallow channels characterized by frequent avulsion events, and hence a distributary or braided channel pattern. In general, the larger the flow, the less that flow “feels” a break in slope, so that very large flows require relatively larger gradient changes to trigger this transition.

In the absence of a sharp gradient break, but where a smooth but strongly concave-upward longitudinal profile exists, it is possible that a transition from leveed channel to frontal splay also can be induced. Under those circumstances of rapidly decreasing gradient, there would be a tendency for flow vectors again to be directed both downward and outward onto the channel floor and against the levees, respectively. As was the case when flows encounter a profile break, this increased lateral stress militates against the formation of levees and favors the formation of frontal splays. The greater the curvature, the greater the stress directed laterally against the levees and therefore the greater the likelihood of a transition from leveed channel to frontal splay. This can occur near the base of slope where channels that have traversed the slope encounter the lower-gradient basin floor. Another example of a topographic setting where this effect can be significant constitutes areas characterized by intraslope mini-basins associated with salt withdrawal. These mini-basins commonly are characterized by strongly concave-upwards longitudinal profiles.

Intraslope mini-basins are examples of where longitudinal profiles can vary with time and how, despite keeping flow discharge and sediment caliber constant, the transition from leveed channel to frontal splay can nonetheless systematically migrate down-system. In this way, the transitions from leveed channel to frontal splay within intraslope basin settings represents a combination of the effect of progressive depletion of mud down-system and the effect of concave-up longitudinal gradients. Prather et al. (1998) described a process of fill and spill in salt-supported intraslope basins. They described a fill succession from sheet-bedded deposits at the base to leveed-channel deposits near the top. This succession can be explained by analysis of the evolution of longitudinal profiles. Intraslope basin relief is highest after a period of sediment starvation during which time continued salt withdrawal results in relatively high relief. Flows entering such an intraslope basin would initially encounter a markedly concave-up topography. This morphology can engender a transition from leveed channel to frontal splay. All else being equal, as the basin gradually fills, the topography progressively becomes less concave upwards, and in response

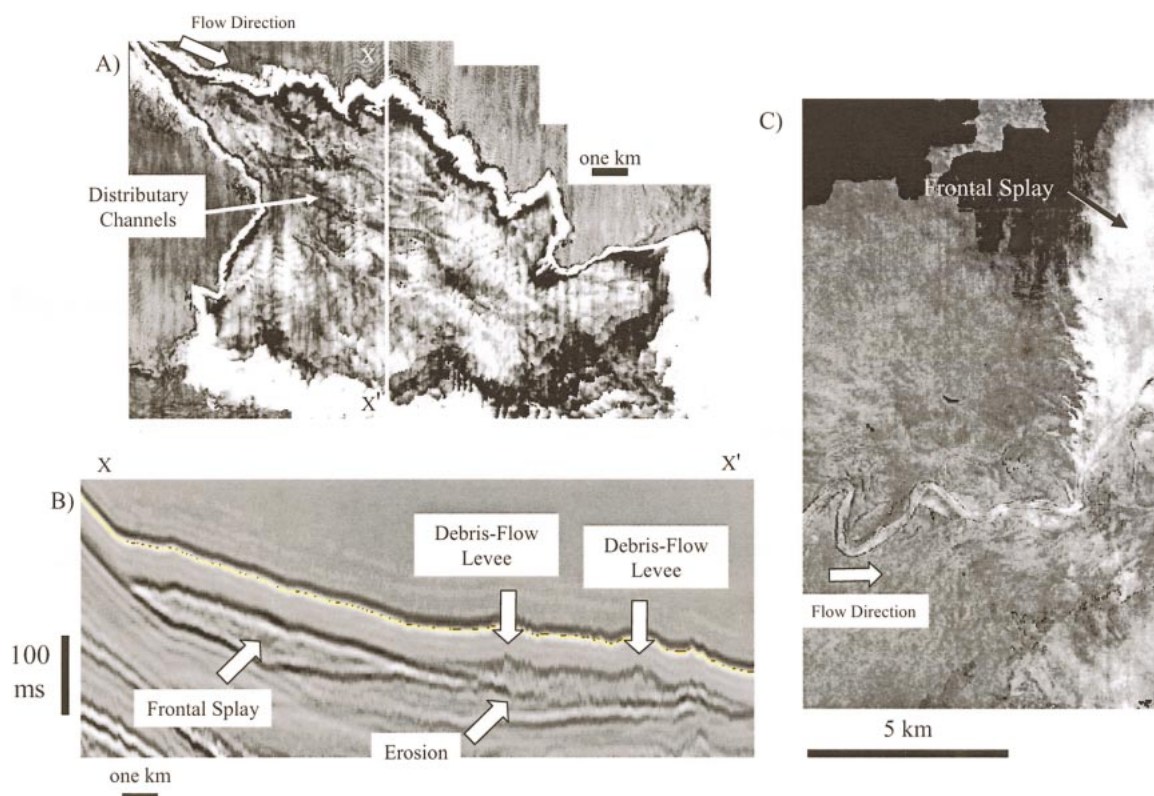


FIG. 16.—Frontal-splay deposits in the De Soto Canyon area of the Gulf of Mexico, where there is a significant gradient break at the transition from turbidity-flow leveed channel to frontal splay. The splay complex is shown both in **A**) plan and **B**) section view; a distributary-channel pattern characterizes this splay deposit. A leveed debris-flow channel overlying and eroding into the splay complex can be observed on the seismic profile (**B**). **C**) Frontal-splay deposits on the basin floor offshore eastern Borneo, Kalimantan, Indonesia (see Fig. 8 for location). A strong amplitude field characterizes these deposits. Although no channels can be identified within the interpreted frontal-splay deposit shown here, the linear fringing features at the left-hand edge of the feature suggests the presence of channeling all the way to its distal extremity, similar to what has been observed by Twichell et al. (1991) at the edges of the Mississippi fan system.

the transition from leveed channel to frontal splay can be expected to migrate down-system (Fig. 17). This basin-fill evolution from frontal-splay dominated to leveed-channel dominated is an autocyclic phenomenon, in contrast with a similar succession caused by a progressive decrease in sand-to-mud ratio, which would constitute an allocyclic phenomenon (Fig. 17).

Exploration Significance

Although we have no borehole calibration for the examples shown, we infer sand presence on the basis of seismic reflection amplitude, continuity, and geometry. On the basis of seismic character, therefore, we infer that splay complexes can constitute significant exploration targets. Reservoir-prone beds can be more widespread and continuous by as much as an order of magnitude than associated up-system leveed-channel deposits. Because of the presence of numerous small, shallow channels that can be characterized by extensive lateral shifts (Figs. 10, 16), these deposits commonly are characterized by a sheet-like stratigraphic architecture (Fig. 10A), likely characterized by good to moderate fluid communication between reservoir-prone beds (Hackbarth and Shaw 1994; Mahaffie 1994; Chapin et al. 1996; Booth et al. 2000). This is especially true of the up-system apex area, just outboard of the transition from leveed channel to splay complex where successive shallow channels have less freedom to spread out and consequently tend to amalgamate (Winker and Booth 2000). In contrast, at the distal edge of splay complexes, reservoir compartmentalization associated with more widespread shale deposits as well as the digitate pattern of channel deposits at the fringes can be common (Fig. 16C) (Twichell et al. 1991; Twichell et al. 1992; Winker and Booth 2000). Moreover, there is an over-

all decrease in sediment grain size from the apex to the distal extremities of frontal splays.

Awareness of the controls on the location of the transition from leveed channel to splay complex can help mitigate the risk associated with reservoir presence and high grade the exploration effort. Such factors as gradient breaks and paleoslope profiles in general can be observed and subsequently used to predict the presence of splay-complex reservoir bodies. Changes in grain-size distribution within flows can also be predicted in certain instances (on the basis of conditions in the associated staging area) and can therefore assist in the prediction of transition location. Sediment supply systems that are shutting down in response to either local avulsion or changes of depositional systems in the upper-slope to outer-shelf staging areas may become more mud prone and therefore may result in turbidite systems characterized by a more down-system transition location (Figs. 14, 17).

Crevasse Splays and Overbank Splays

In some instances, where channel levees are breached to form crevasses, splay or distributary-channel complexes can form immediately outboard of the crevasse (Figs. 18A and B). The crevasse splays shown in Figure 18 cover areas approximately 50 km². Overbank splays comprise similar deposits not associated with a significant levee crevasse (Fig. 18C) but tend to cover smaller areas than crevasse splays. Both types of deposit are most common along outer channel bends where flow momentum results in an increased tendency for flows to overtop or breach the adjacent levee.

Crevasse splays involve processes similar to those responsible for the

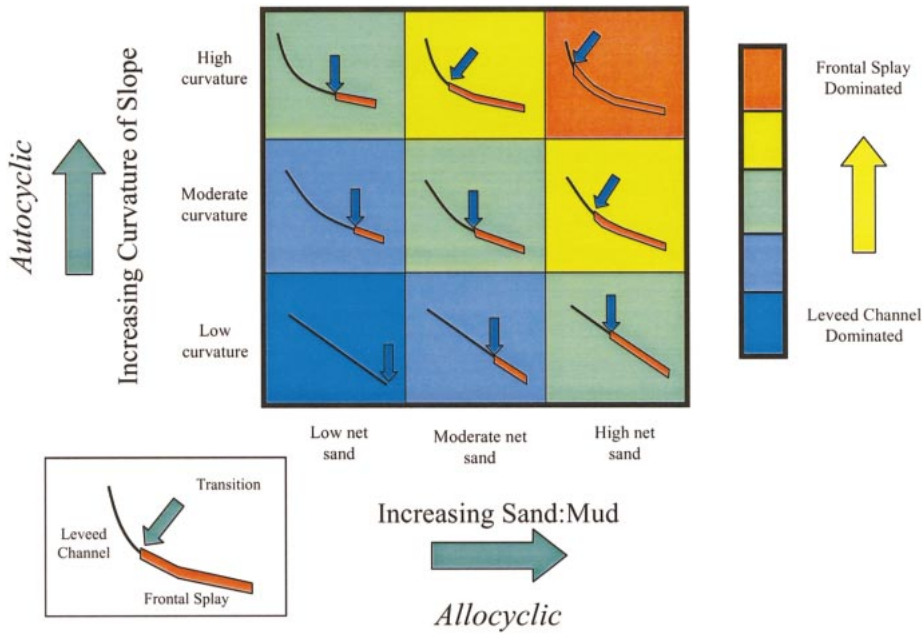


FIG. 17.—Matrix showing location of transition between leveed channel and frontal splay as a function of varying flow sand-to-mud ratio and varying slope curvature. Varying sand-to-mud ratio is a function of conditions in the staging area and is considered an allocyclic effect; varying slope curvature is a local factor and is considered an autocyclic effect (see discussion in text).

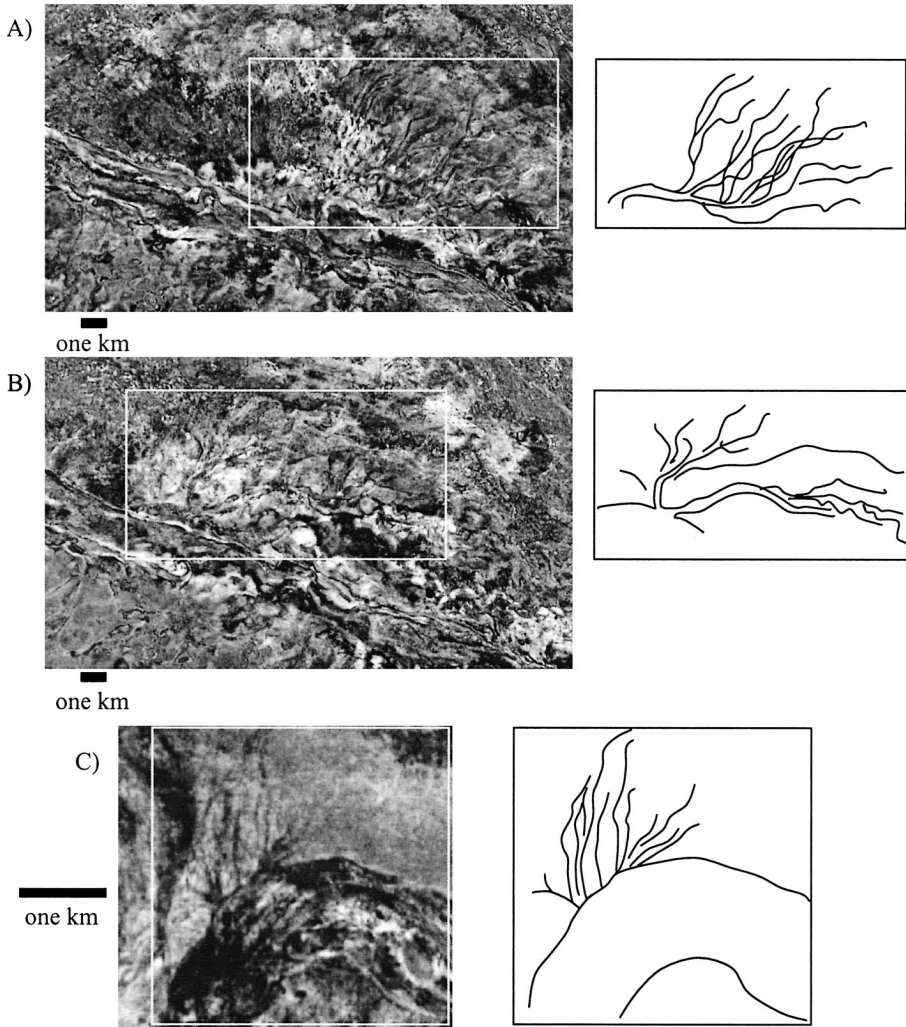


FIG. 18.—Examples of A, B) crevasse splays and C) overbank splay from the Pleistocene of the Gulf of Mexico. The images shown are amplitude extractions from seismic horizon slices.

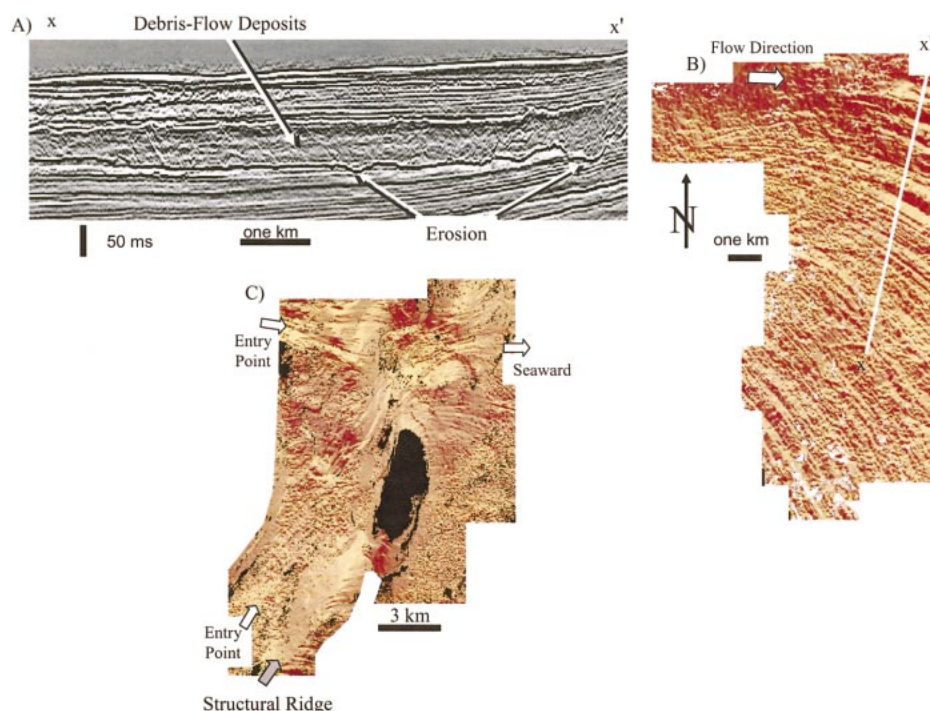


FIG. 19.—Two examples of erosional scour at the base of debris-flow deposits offshore eastern Borneo, Kalimantan, Indonesia. **A)** Seismic section across debris-flow deposit. Erosional scour at the base of the unit locally is greater than 30 m deep and 0.7 km wide. Location of section is shown in Part B. **B)** Seismic reflection dip azimuth map at the base of the debris-flow unit illustrating long, continuous, linear grooves, which show down-system divergence in map view. **C)** Seismic reflection azimuth map of the base of an extensive debris-flow complex at a lower stratigraphic level and south of that shown in Part B, offshore eastern Borneo, Kalimantan, Indonesia, illustrating the effect of irregularities in sea-floor topography on debris-flow direction. These debris flows were characterized by multiple entry points as well as divergence of basal grooves in the down-system direction.

formation of avulsion channels. In some instances, rather than a single avulsion channel a network of distributary channels forms immediately distal to a levee crevasse, forming a crevasse splay (Fig. 18). The avulsion channel connecting the principal leveed channel to the crevasse splay may be limited to just the distance across the levee crevasse itself. A possible explanation for the absence of an avulsion channel would be the sand-to-mud ratio of the flow coming through the crevasse. If the sand-to-mud ratio of the flow coming through the crevasse is relatively high, levees would be unlikely to form outboard of the crevasse, and, in response to the suddenly increased cross-sectional area as the flow passes through the crevasse, flow velocity abruptly decreases and rapid sedimentation occurs. The rapid sedimentation results in a distributive channel pattern. A similar process characterizes those situations where flow stripping results in sandy overbank splays. Overbank splays are similar to crevasse splays but lack an associated levee crevasse and are smaller in size.

Exploration Significance

Crevasse splays can produce sheet-like sand-rich deposits that thin away from the associated levee crevasse. Shallow channels giving way to sheets largely characterize these deposits. Lateral reservoir continuity can be excellent, and areal coverage can be extensive, with length and widths greater than 10 km. As with overbank splays, these deposits tend to be fan-shaped.

Debris-Flow Sheets, Lobes, and Channels

Debris-flow deposits take a variety of forms, ranging from sheets to lobate tongues to channel fills (Prior et al. 1984). Debris-flow sheets can be very extensive, in excess of tens of kilometers wide, and accumulate to substantial thickness in some instances (< 150 m). Such sheets can extend well onto basin floors beyond the toe of slope, in many instances extending as far basinward as associated turbidity-flow deposits (Fig. 19).

Debris-flow lobes commonly tend to be relatively straight and narrow, though some can be broad and arcuate. In either case, these deposits can be characterized by steep edges (Fig. 20) with slopes of 3–4 degrees. As a consequence, transverse profiles across debris-flow lobes can give the

appearance of mounding. In other instances, gradual thinning and gentle slopes can characterize debris-flow margins. In addition to sheet- and lobe-form geometries, channelized debris flows also have been observed. In a few instances the channelized flows are associated with levees (Fig. 16), whereas not so in other instances (Fig. 21). In either event, channelized debris flows tend to be characterized by lower sinuosity than most leveed turbidity-flow channels (Figs. 16, 21). In all instances the channel-fill deposits are expressed seismically by chaotic to transparent seismic facies, as is the case with debris-flow deposits in general, and their basal surfaces are grooved.

A distinctive attribute of debris-flow deposits, regardless of whether they are lobe, sheet, or channel form, is the pattern of long linear grooves that commonly characterizes the basal surface upon which they are deposited. Figure 19A shows a seismic reflection profile across a debris-flow deposit up to 55 m thick. Erosional scour, in some instances over 40 m deep and over 700 m wide, characterizes the base of these deposits. In plan view these grooves are long and linear (in some instances greater than 20 km long; Fig. 19) and tend to diverge in a seaward direction. Grooves at the base of channelized deposits are shown in Figure 21B; they, too, are characterized by seaward divergence in map pattern.

These grooves suggest the presence of large cohesive blocks of sediment lodged within the base of the flow and remaining there for long distances (in some instances greater than 10 km) before either being lifted off the flow base or being disaggregated. Similar scour at a smaller scale can be observed in outcrop (Posamentier 2001b). Note that the degree of groove development can be variable, from well-developed, as in the example from offshore Indonesia (Fig. 19), to poorly developed, as offshore Nigeria (Posamentier 2001b). This variability can be attributed either to differences in sediment caliber and the size of indurated blocks within the respective flows or differences in the degree of induration of the substrate across which the flows travel. The common divergence of grooves down system suggests divergent flow vectors in an unconfined setting.

Another distinguishing characteristic of debris-flow deposits is their seismic expression, commonly characterized by low-amplitude, generally chaotic reflection patterns (Figs. 19–22). The upper bounding surface can be

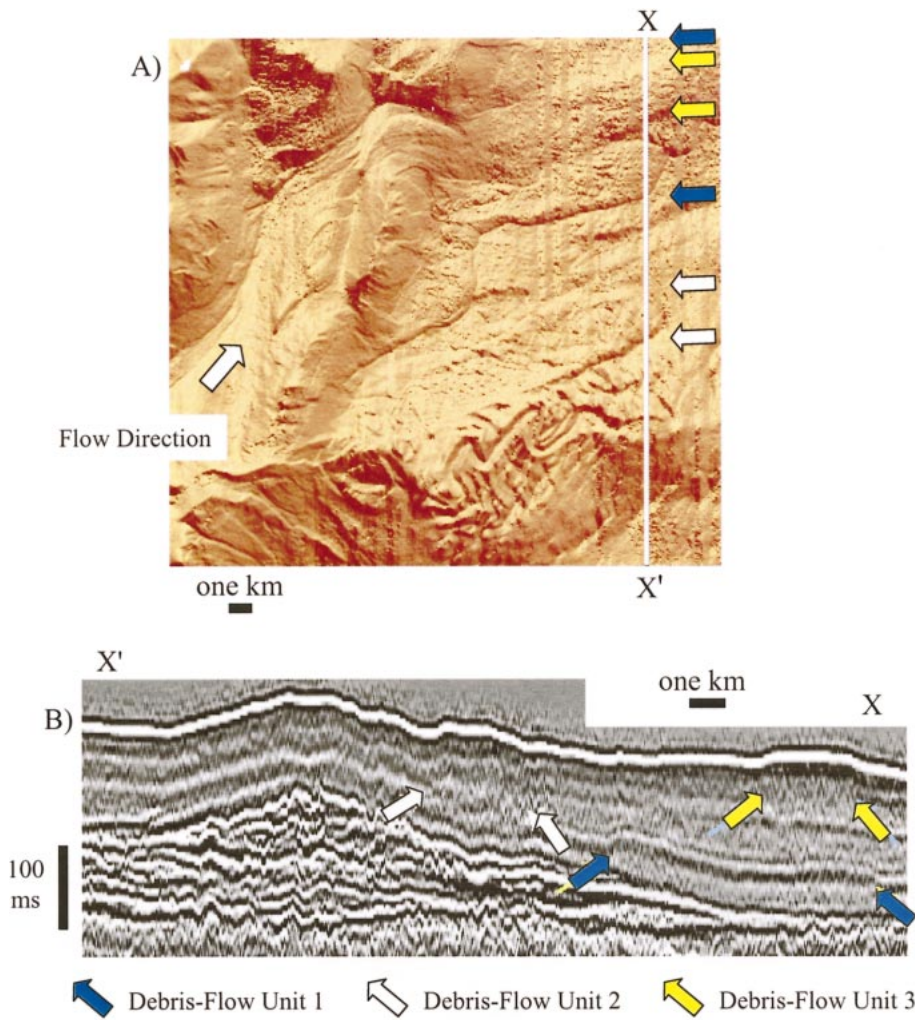


FIG. 20.—A) Seismic reflection dip azimuth map of the sea floor offshore eastern Borneo, Kalimantan, Indonesia, illustrating debris-flow lobes and tongues of various shapes. B) The seismic section shows that the stacked debris-flow units, which have expression at the current sea floor, actually correspond to buried debris-flow units of varying ages, which have been covered by drape deposits so as to preserve their current sea-floor expression. These debris-flow units are difficult to recognize in section view. The arrows are color-coded as to relative age, with debris-flow unit 1 being the oldest and debris-flow unit 3 being the youngest.

uneven; in some instances pressure ridges indicative of thrust faults commonly observed near flow termini can be observed at the top of debris-flow units (Fig. 22) (e.g., Brami et al. 2000). Debris-flow deposits, with their typically chaotic to reflection-free seismic character, are interpreted to be largely mud prone. As such they constitute relatively poor reservoir targets. However, some interpreted debris flows containing reservoir-quality deposits have been described (Jennette et al. 2000).

INTERPRETATION AND DISCUSSION

The following section focuses on controls on deep-water sedimentation and subsequent packaging of depositional elements and stratigraphic units into depositional sequences. First-order controls include sea-level change and the sand-to-mud ratio present at the shelf edge. Understanding the dominant controls on the spatial and temporal distribution of depositional elements can substantially improve prediction of reservoir and seal facies important in the exploration for hydrocarbons.

Controls on Deep-Water Sedimentation

The controls on deep-water sedimentation can be grouped into two categories. The first group controls the *occurrence* of sand-rich deep-water deposition, whereas the second controls the *style* of deep-water deposition. Factors in the first group include (1) the presence of sand in the shelf-margin staging area, (2) the presence of steep slopes immediately outboard

of the staging area, and (3) the occurrence of sea-level falls sufficient to expose most of the shelf. Factors in the second group include (1) sand-to-mud ratio within the material delivered to the deep water, (2) rugosity of the sea floor (i.e., basin physiography), (3) magnitude, duration, and frequency of depositional events originating in the shelf-margin staging area, (4) sediment delivery mechanism (e.g., turbidity flow, debris flow, slump, etc.), (5) single point source (e.g., canyon) vs. multiple small point sources (e.g., slope gullies), and (6) rheology and concentration of sediment gravity flows, and (7) bottom-current activity. Reading and Richards (1994) presented an excellent overview of controls on deep-water systems. The subsequent discussion follows their approach to sedimentation control factors.

Controls on Presence of Sand-Prone Depositional Elements

The presence of sand in the staging area is of primary importance in the occurrence of sand-prone deep-water deposits (Posamentier and Allen 1999). The staging area can be defined as the area at the outer shelf and upper slope where sediments are temporarily stored before later remobilization and delivery to the deep-water environment beyond. Most coarse-grained deep-water sediments originate or pass through this staging area. If no sand is present in the staging area, no sand is delivered to the deep water from that location, and unless sand is transported in from another staging area along strike, there will be no sand-prone deposits downdip from this staging area. In some instances sand can be delivered directly from river systems by hyperpycnal flow when rivers are in flood, thus not

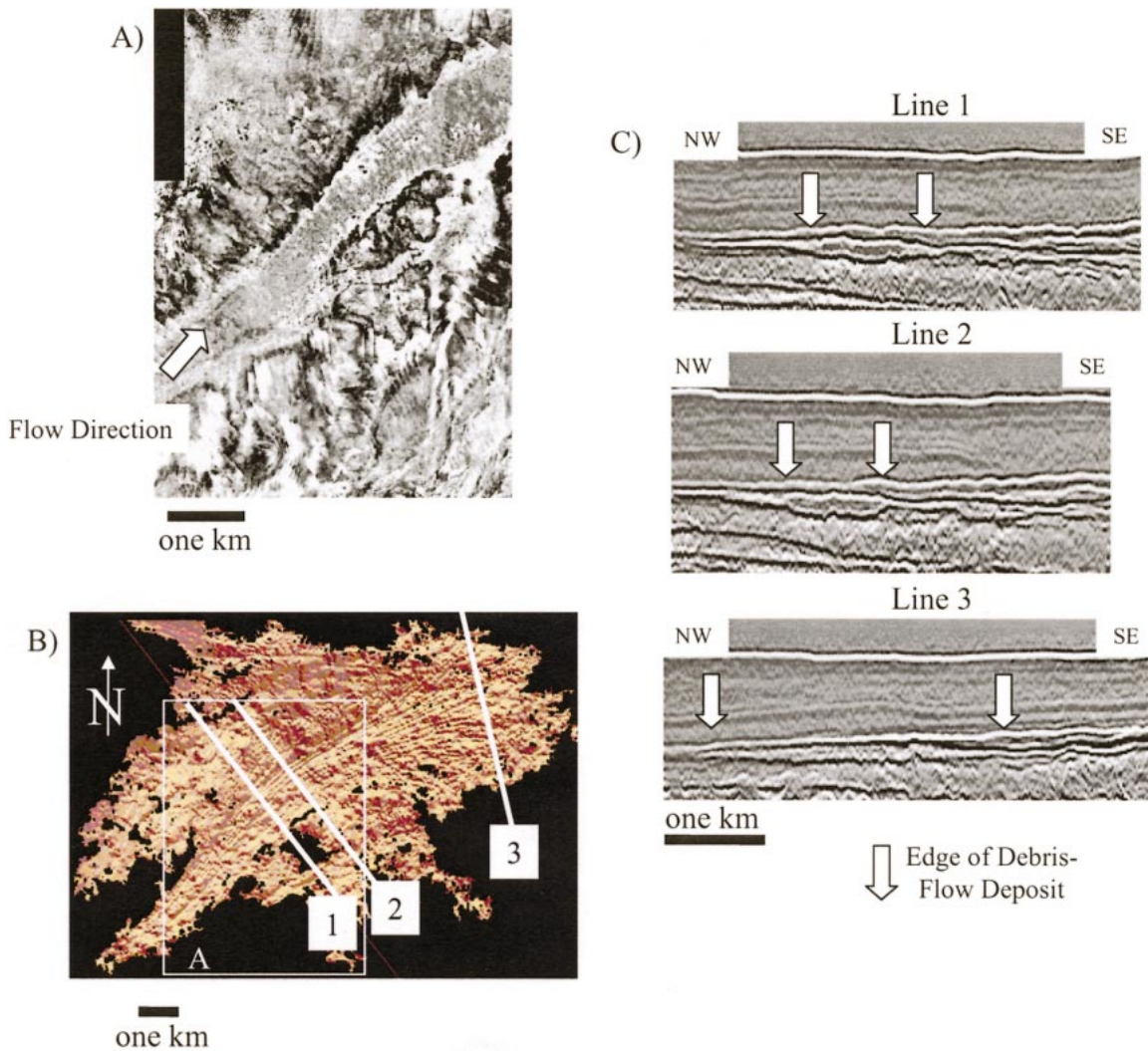


FIG. 21.—A) Seismic reflection horizon slice illustrating a debris-flow channel in the deep-water Makassar Strait offshore Kalimantan, Indonesia. Location of Part A shown on inset in Part B. B) Seismic reflection azimuth map on the reflection at the base of the debris-flow channel showing extensive linear grooves within the confines of the channel. The grooves diverge down system (northeastward) where the debris flow goes from confined to unconfined. C) Seismic reflection profiles across the debris-flow channel. Absence of levees characterizes this debris-flow channel (contrast with leveed debris-flow channel shown in Fig. 16B).

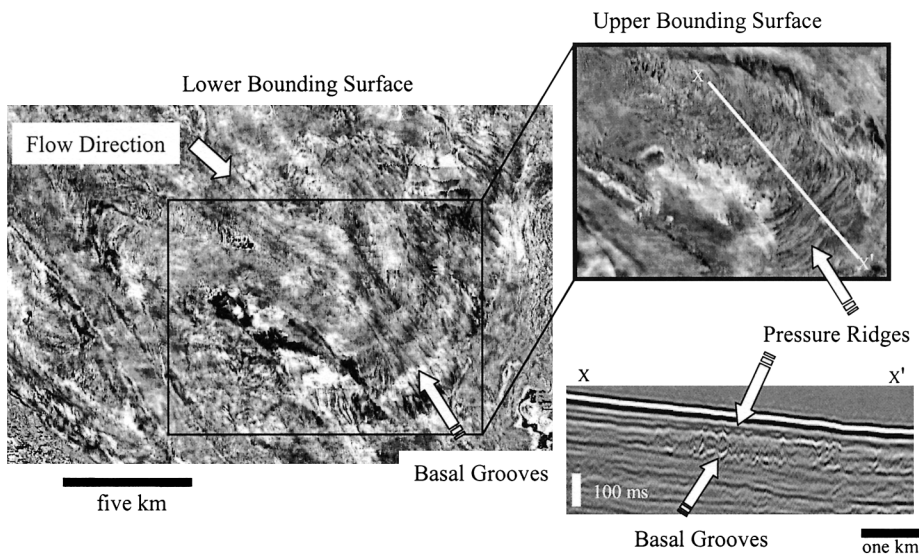


FIG. 22.—Debris-flow deposits offshore Gulf of Mexico in the Green Canyon area imaged on reflection horizon slices as well as vertical profile. This debris-flow deposit is characterized by grooves at the base and pressure ridges at the top as seen on the reflection horizon slices. On the vertical profile, both the basal grooves and the pressure ridges are imaged as a corrugated irregular surface. The pressure ridges are inferred to be associated with thrusts within the debris-flow deposit.

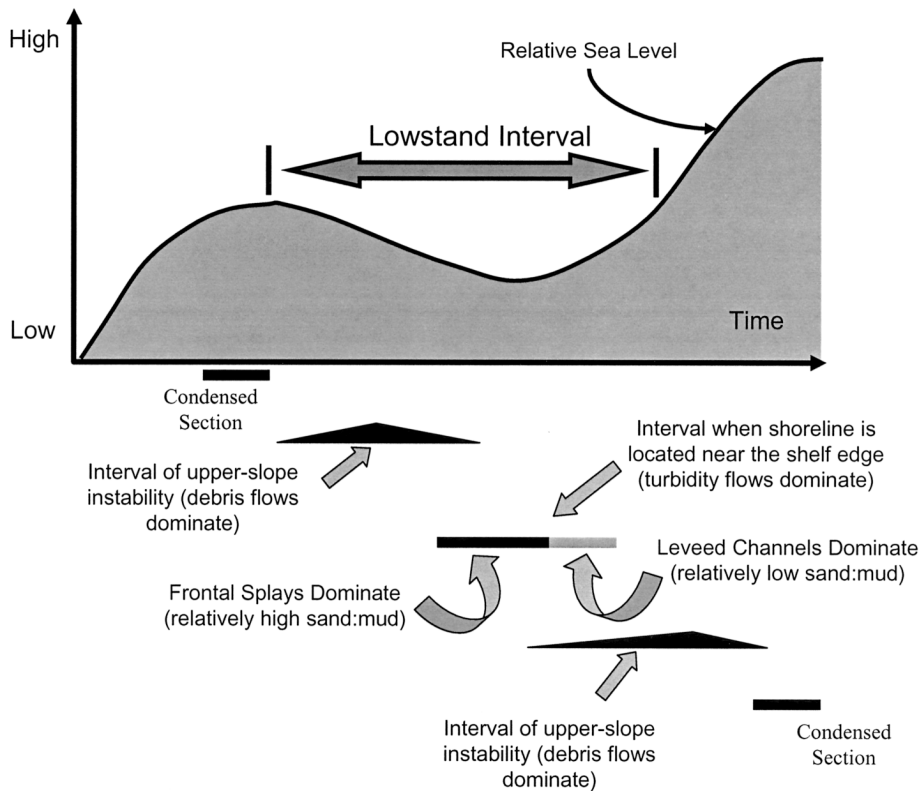


FIG. 23.—Schematic depiction of the relationship between relative sea level and type of dominant mass-flow process. The succession comprises debris-flow deposits at the base (corresponding to the initial period of relative sea-level fall), overlain by frontal-splay-dominated and then leveed-channel-dominated sections (corresponding to the subsequent period of early and late relative sea-level lowstand respectively). The succession is capped by deposition of debris-flow and condensed-section deposits (corresponding to periods of rapid sea-level rise and highstand, respectively).

coming to rest in the staging area during these times of peak discharge (Mulder and Syvitski 1995).

The presence of steep slopes outboard of the staging area facilitates the transfer of sediments to the deep-water environment. The steeper the slope the greater the velocity of sediment gravity flows, and therefore the greater the sediment discharge. Sea-level lowstands sufficient to expose most of the shelf also enhance the likelihood of sand-prone deep-water sedimentation. During times of falling sea level, shorelines migrate seaward, towards the outer shelf. In fact, in most instances sea-level fall is necessary in order to get shoreline depocenters out to the margins of even moderately wide shelves (Posamentier and Allen 1999; Muto and Steel 2002). This seaward shift of the shorelines, and consequently the depocenters, results in the replenishment with sand and the reactivation of the shelf-margin staging areas.

Controls on Depositional Style

Once shelf-margin conditions are in place for the delivery of sand to the deep water, the depositional style in the deep water will depend on a variety of factors. The sand-to-mud ratio within sediment gravity flows plays a significant role in determining whether a single leveed channel or a network of shallow distributary channels will dominate (Posamentier 2001a). The rugosity or irregularity as well as the gradient of the sea floor across which flows travel play an important role in determining the shape and distribution of the resulting depositional elements and influence the type of depositional element that might form. Insofar as sediment gravity flows prefer bathymetric lows, the form of sand-prone deep-water depositional elements closely follows the shape of sea-floor irregularities (Nardin et al. 1979). Abrupt changes in gradient across which flows travel can trigger either erosion or deposition as well as significantly modify the style of deposition (Kneller et al. 1999).

The flow magnitude of a depositional event controls the size of the associated depositional elements. The effects of flow duration and frequency

are not well understood, but it is likely that they can also influence the morphology of depositional elements (Piper and Normark 1983). Bottom-current activity can play a variable role in the morphology of deep-water depositional elements. Active deep-water currents (e.g., contour currents and tidal currents) can in some instances significantly modify sediment-gravity-flow deposits (Mutti 1992). The extent to which this occurs depends on current flow velocity, and the grain size and degree of cementation of these earlier-deposited elements.

The delivery mechanism, whether it is turbidity flow, slurry flow (Lowe and Guy 2000), debris flow, slump, slide, etc., plays a significant role in determining the type and distribution of depositional elements (Mutti and Normark 1991). The suite of depositional elements associated with each delivery mechanism is unique and predictable, and can therefore provide insights ultimately into their stratigraphic and reservoir architecture.

Vertical Successions—Sequence Stratigraphic Significance

Stratigraphic sections observed in the three deep-water areas studied suggest a systematic development of deep-water sequences that can be related to cycles of relative sea-level change (Fig. 23; Posamentier et al. 2000). It should be noted that we draw our conclusions on the basis of observations of vertical successions and temporal relationships, without the benefit of age dating with which to calibrate the model. These observations suggest that cyclic successions are common. In places, inferred mass-transport deposits, characterized by chaotic seismic reflections, directly overlie thick successions of inferred hemipelagic and pelagic sediments indicating the abrupt onset of a period of high sedimentation rate following an extended period of slow sedimentation rate and interpreted as the base of a depositional sequence. These mass-transport deposits commonly are overlain by interpreted turbidite successions and eventually capped by hemipelagic and pelagic deposits (Fig. 24). These successions typify the deep-water expression of a depositional sequence. Although mass-transport deposits are common at the base of such successions, in certain circumstances they can

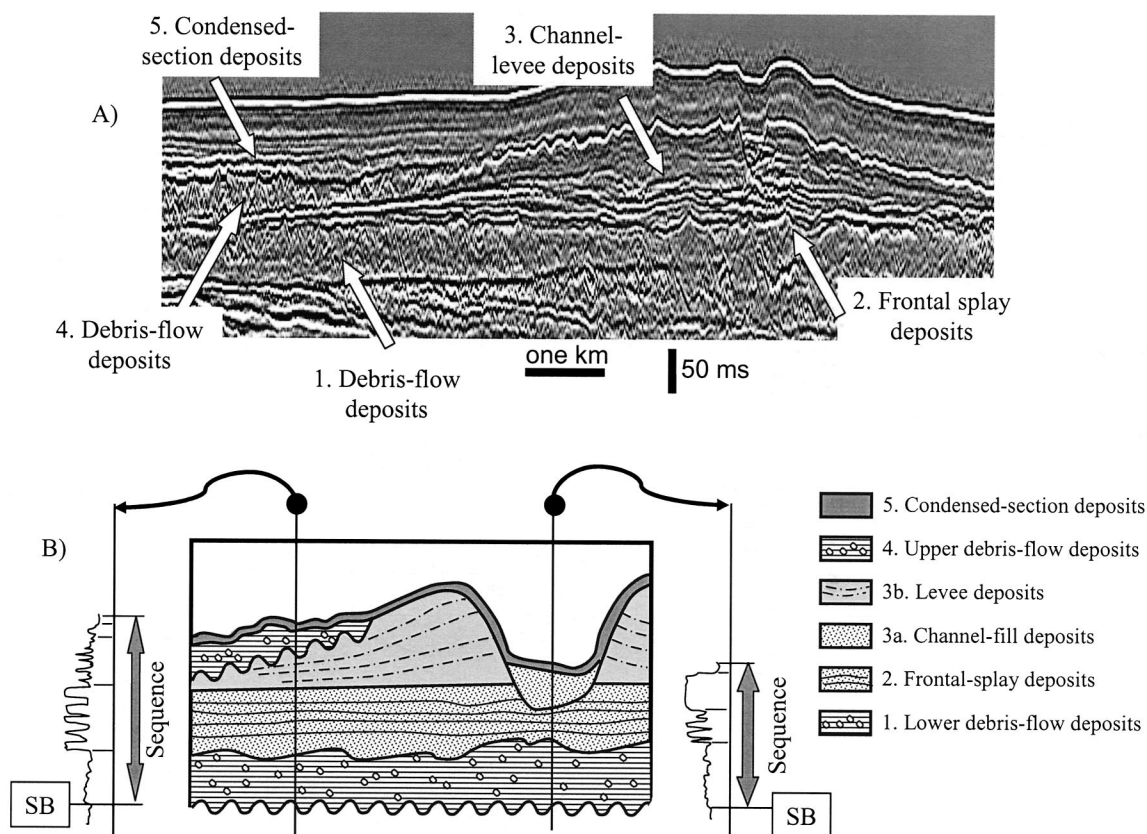


FIG. 24.—A) Seismic reflection profile from offshore Indonesia illustrating the stratigraphic succession of a deep-water sequence. Debris-flow deposits (1) are overlain by frontal-splay deposits (2), channel-levee deposits (3), and again debris-flow deposits (4). The entire succession is then inferred to be mantled by a thin veneer of condensed-section deposits (5). B) Schematic depiction of an idealized deep-water depositional sequence, with two hypothetical log profiles shown.

form mid cycle as well (Pirmez and Flood 1997), suggesting the need for caution when applying general models. Ultimately, age calibration remains an essential but missing piece to the puzzle, which awaits future research. In the following section we take a first-principles approach in constructing the general model, an approach that leaves the door open to appropriate variations on the general theme.

Sea-level change plays an instrumental role in shifting depocenters towards the shelf edge, where river-borne sediments can then be supplied directly to the deep water to form extensive sediment-gravity-flow deposits. The succession in deep-water sections observed by us as well as by others to varying degrees (Weimer 1991; Piper et al. 1997; Pirmez et al. 1997; Manley and Flood 1998; Maslin et al. 1998; Beauboeuf and Friedmann 2000; Brami et al. 2000; Winker and Booth 2000) comprises debris-flow deposits at the base, overlain by turbidite deposits, and subsequently capped by another section of debris-flow deposits (Fig. 24). Further, the turbidite deposits can be subdivided into a lower splay-complex-dominated section overlain by a leveed-channel-dominated section (Fig. 24). However, because of the complexities inherent to deep-water depositional environments, this succession is not observed at every location insofar as conditions in the staging area can be highly variable through time and because of spatial variations in the basin as well. For example, in slope settings debris-flow and splay components can be thin to absent, whereas in basin floor settings the leveed channel component can be thin to absent. In deep-water settings, where no sand-rich shelf-margin staging area is present upslope, a deep-water environment may completely lack turbidite deposits and be represented only by debris-flow deposits (Posamentier and Allen 1999).

Nonetheless first-order control by relative sea-level change can be inferred and commonly has expression in the stratigraphic record. At the very

least, relative sea-level fall results in depocenter shifts from the middle and inner shelf to the outer shelf. When depocenters are located at the shelf margin, an active sediment staging area is most likely to develop and result in accelerated sediment supply to the deep water beyond. In contrast, when depocenters are located at the inner to middle shelf, significantly less sediment is delivered to outer shelf staging areas and, by extension, to deep-water systems. During these times staging areas become relatively dormant and are sites of minimal sedimentation (Coleman and Roberts 1988). Concomitantly, during these times deep-water environments are characterized by deposition of condensed sections comprising primarily hemipelagic and pelagic sediments (Fig. 23). Consequently, it can be said that in this sense relative sea-level change creates an on-off switch to deep-water systems, and further, that sea-level highstands result in deposition of condensed sections over deep-water areas, which punctuate and subdivide the stratigraphic record.

Following periods of relative sea-level highstand characterized by inactive outer-shelf staging areas, relative sea-level fall causes reactivation of staging areas even before depocenters reach the outer shelf. Falling relative sea level results in a progressive stripping of "overburden" (in this instance, layers of water), which causes disequilibrium conditions to develop on the outer shelf and upper slope. These disequilibrium conditions on the outer shelf and upper slope, which in some instances are accompanied by gas hydrate dissolution (Kayen and Lee 1991; Haq 1993, 1995; Maslin et al. 1998), results in inherently unstable upper-slope conditions because of the potential for increased pore pressure within the sediment column. This, in turn can lead to slope failure, resulting in mobilization of cohesive predominantly fine-grained deposits producing mass-transport complexes where mud-prone slope sediments are involved (Fig. 23). Where failures

occur at or near the shelf edge, and where previous lowstand shelf-edge deltas can be involved, these mass-transport complexes can contain sand as well.

When depocenters ultimately reach shelf-edge staging areas, direct (by hyperpycnal flow) or indirect (by remobilization of temporarily deposited sediments) supply of turbidity currents into lower-slope and basin-floor environments can occur (Fig. 23). During the time of relative sea-level fall, fluvial erosion of the shelf with associated development of incised valleys and shelf-edge canyons commonly occurs. Subsequently, during times of relative sea-level stillstand and slow rise, preferential sedimentation of the sand fraction commonly occurs in association with fluvial and estuarine deposition within incised valleys. This results in a more sand-rich supply to staging areas during periods of sea-level falls, when all sediment supplied from hinterlands plus that which is cannibalized from the substrate is delivered to the staging area, and a more mud-rich supply to staging areas during periods of relative sea-level stillstand and rise (Fig. 23). A variation on this theme is observed along margins where deep-water systems can continue to remain active during highstands of sea level because of interception by canyons of active longshore drift (e.g., the Hueneme fan offshore California; Piper and Normark 2001). This situation is most common in basins with narrow shelves. The sediment input changes from sea-level highstand to lowstand, from longshore drift to more direct fluvial input. In these instances, though the sediment volume delivered to the deep-water likely is higher, the sand likely is cleaner (i.e., less mud) during highstand because of the inherently better sorted sediments supplied by longshore drift.

In response to changes of sediment caliber delivered to staging areas during late sea-level fall, stillstand, and early rise, there will be systematic and predictable responses in the style of associated turbidite deposition. Initially, when staging areas are experiencing direct supply of sediment from fluvial systems during sea-level fall, flows tend to be more sand rich, whereas during the late stages (i.e., during periods of relative sea-level stillstand and slow rise) flows tend to be more mud rich. Consequently, the transition from leveed channel to splay complex, in the case where it is not a break in slope that has precipitated this transition location (see discussion above), tend to be located farther up system during the early stages of turbidite deposition, and farther down system during the late stages of turbidite deposition (Fig. 17). As a consequence, at any given location this shift of transition seawards through a sea-level cycle could result in a superposition of leveed-channel deposits (i.e., late-stage deposits) over splay-complex deposits (i.e., early-stage deposits) (Fig. 24). This pattern has been observed on the Amazon fan and is linked there to the sand-to-mud ratio of the sediment load (Pirmez et al. 2000). The description of this succession is essentially consistent with that of earlier sequence stratigraphic models (Vail et al. 1977; Mutti 1985, 1992; Posamentier et al. 1991).

A cycle of deep-water sedimentation commonly ends with a resumption of rapid relative sea-level rise and an associated landward shift of shoreline-associated depocenters. This landward shift results in progressive deactivation of shelf-edge staging areas. Slow shelf-margin regression during the late lowstand gives way to shoreline transgression at that time. Turbidity flows progressively become more mud rich, producing a dominance of leveed-channel deposits (transitions from leveed channel to splay complex commonly shift far down system at this time) (Figs. 17, 23). Ultimately the delivery of fresh sediment to staging areas ceases and turbidity flows to the deep-water environment become highly infrequent. However, during these periods of rising relative sea level, loading of the outer shelf and upper slope, by the addition of new overburden comprising sediment as well as increased water column, results again in disequilibrium conditions on the upper slope. This, in turn, can result in another period of slope instability and slope failure, and deposition once again of mass-transport complexes (Fig. 23) (Paull et al. 1991; Haq 1993; Maslin et al. 1998).

Finally, when relative sea level stabilizes at a highstand position, slope stability is restored, shelf-edge areas (i.e., staging areas) become sites of

nondeposition, and deep-water environments are dormant as the deep-water sequence is capped by a highstand condensed section (Fig. 23). The typical deep-water sequence is depicted in Figure 24, with debris-flow deposits at the base of the section, overlain by splay-complex-dominated turbidite deposits, leveed-channel-dominated deposits, debris-flow deposits again, and ultimately capped by condensed-section deposits.

Relating the succession of depositional elements to specific stages in a sea-level cycle, as done by us here, and others elsewhere (e.g., Weimer 1991; Beaubouf and Friedmann 2000), is yet to be confirmed by extensive coring and detailed sedimentological and isotopic studies. Autocyclic factors such as depocenter shifts associated with delta lobe switching can modify the idealized succession illustrated in Figure 24. For example, Piper et al. (1997) reported four large mass-transport complexes on the Amazon fan within the last glacial–interglacial cycle. These mass-transport complexes were interbedded with major channel–levee deposits. On the basis of ODP data, one of these mass-transport complexes was deposited during the time of maximum sea level lowstand and not during early sea-level fall (Piper et al. 1997). It is obvious that autocyclic factors, in addition to allocyclic factors (e.g., relative sea-level change) can play a significant role in determining the succession of deep-water depositional elements. Consequently, the stratigraphic succession shown in Figure 24 must be considered an idealized case that must be adapted to accommodate effects of local autocyclic factors.

SUMMARY AND CONCLUSIONS

Deep-water depositional elements from several deep-water settings were analyzed. Key depositional elements described included turbidity-flow leveed channels, levees or overbank settings with sediment waves, frontal splays (also described as lobeforms and distributary-channel complexes), crevasse flows, and debris-flow channels, lobes, and sheets.

Both lateral and vertical relationships between depositional elements vary systematically. Leveed channels commonly are characterized by decreasing levee thickness down system. Where levee heights become ineffective at confining the high-density parts of turbidity flows, a morphologic transition between leveed channel and frontal splay occurs. The transition from leveed channel to frontal splay also can be associated with abrupt breaks in sea-floor gradient. Frontal splays are observed to be largely channelized and commonly form distributary-channel complexes.

Crevasse splays are observed most commonly along outer channel bends. These deposits are associated with levee breaching and form lobate deposits with distributary-channel patterns just outboard of the crevasse. Overbank splays, also most common along outer channel bends, are not associated with levee crevasse, but are also lobate in form. They are commonly smaller than crevasse splays.

Debrites are observed in a variety of forms, including lobes, sheets, and channels. Two distinguishing characteristics are (1) the transparent to chaotic seismic reflection pattern within the deposits, and (2) the presence of pronounced grooves and striations at the base of the deposits. The grooved ubiquitous patterns commonly are divergent down system in map view, indicating flow spreading.

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