Quaternary Depositional Environments of the East Texas Coast and Shelf: Analogs for Ancient Deposits



SEPM/AAPG Field Trip #12
April 13, 2006, Houston, Texas

Field Trip Leader
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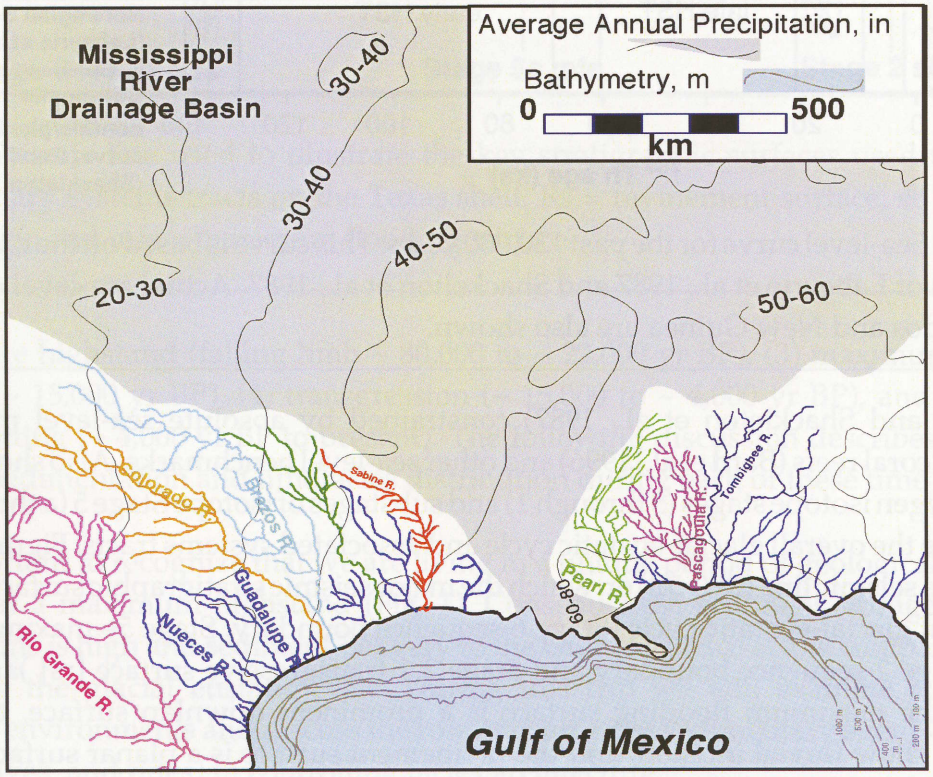
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INTRODUCTION

During this field trip we will visit and / or discuss several depositional environments, and examine sediment cores and high-resolution seismic records from these and other offshore environments. The objective is to understand the facies architecture of these environments, the distribution of these facies on the continental shelf during the last glacial eustatic cycle, and their preservation potential.



Drainage basins of Gulf Coast rivers.

The Late Quaternary is the one interval of geological time when sea level history is well enough constrained to relate coastal and shelf sedimentation to rising and falling sea level. Climatic changes in Texas are also reasonably well known, although there is a need for more detailed paleoclimatic reconstructions, particularly as they relate to sediment erosion and transport within fluvial drainage basins and into the Gulf of Mexico basin. Given the constraints on eustasy, we are able to examine depositional systems using a sequence stratigraphic framework. The sea-level curve that we will use to study sedimentation and strata formation is illustrated in Figure 1. It is derived from oxygen isotope curves (Labeyrie

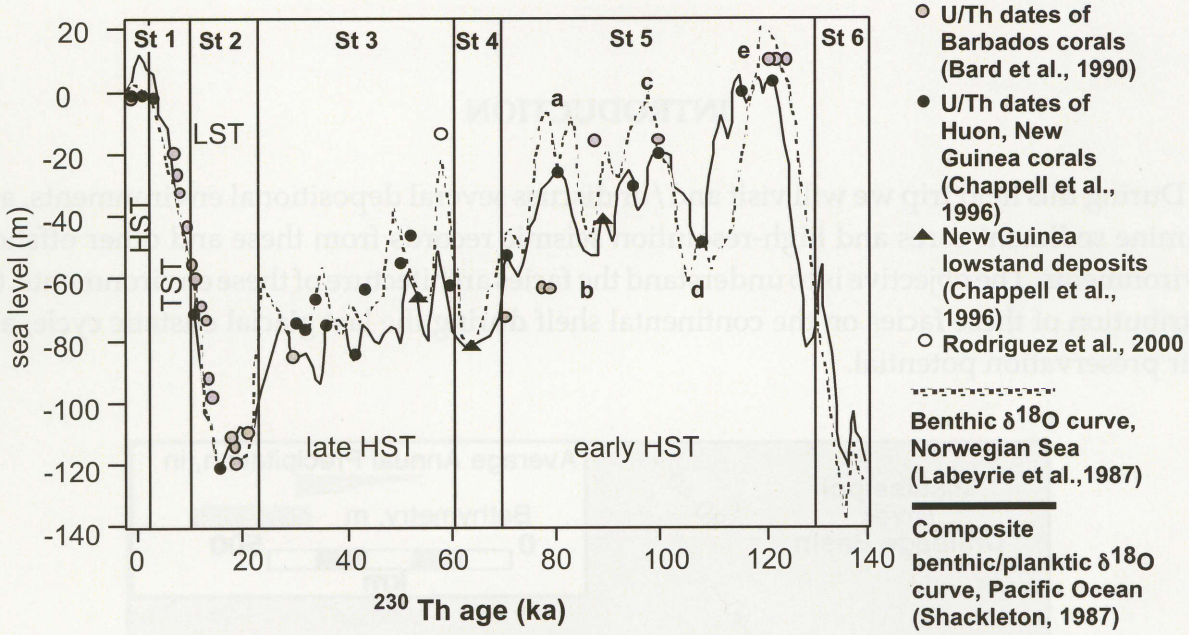


Figure 1. Sea-level curve for the past 120,000 years. This curve is based on the oxygen isotope curves of Labeyrie et al., 1987 and Shackelton et al., 1987. Actual sea-level datums from Barbados and New Guinea are also shown.

et al., 1987 and Shackleton et al., 1987) constrained by absolute sea-level markers from submerged coral reefs (Bard et al., 1996) and other sea-level benchmarks. Also shown in Figure 1 are the oxygen isotope stages (5 through 1) and substages of isotope Stage 5 (a through e) used to subdivide the overall glacial-eustatic cycle and associated systems tracts. Figure 2 shows an example of a seismic line (R90-5B) on which the major seismic stratigraphic surfaces associated with the last glacial-eustatic cycle (Stage 6 sequence boundary, Stage 5e maximum flooding surface, Stage 2 sequence boundary and Stage 2-1 transgressive surface [rs])are identified. Note that the maximum flooding surface is a prominent downlap surface, the sequence boundaries show fluvial incision, and the ravinement surface is a planar surface of erosion. Identification of these surfaces allows subdivision of the stratigraphic section into systems tracts.

Paleogeographic maps depicting the distribution of major depositional systems on the shelf at various time intervals during the glacial eustatic cycle were generated using this stratigraphic framework (see our web site for more detail, http://gulf.rice.edu and Anderson and Fillon, 2004). These maps will be used to illustrate the important influence of long-term forcing agents (sea-level and climate change, tectonics and subsidence) on sedimentation and stratigraphic evolution. Several thousand kilometers of seismic data, platform boring descriptions (over 300), and sediment cores (over 400) were used to produce these maps (Anderson et al., 1996, in press) (Fig. 3). The maps show depositional systems for the (1) early highstand (~ 120,000 to ~ 80,000

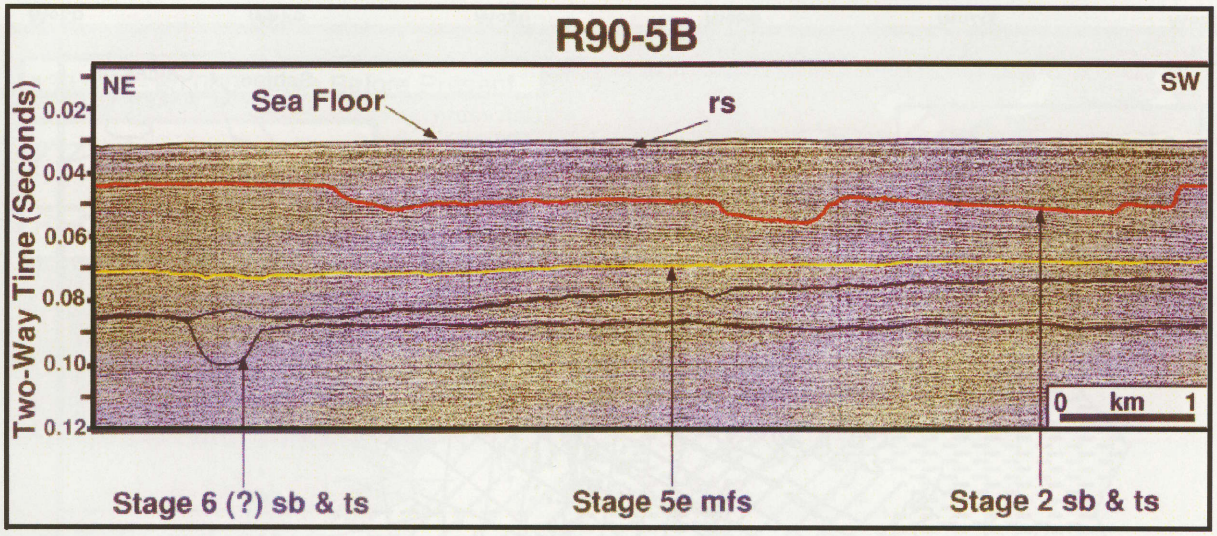


Figure 2. Seismic line used to illustrate the key stratigraphic surfaces used to map late Quaternary systems tracts on the Texas shelf. Rs = ravinement surface, sb = sequence boundary, and mfs = maximum flooding surface.

yr BP), (2) late highstand (falling limb ~ 80,000 to ~ 22,000 yr BP), (3) maximum lowstand (~ 22,000 to ~ 15,000 yr BP), (4) transgression (~ 15,000 to ~ 4,000 yr BP), and (5) present highstand setting (~ 4,000 yr BP to present). The following discussion describes the factors influencing sedimentation and lithofacies distribution during each of these time intervals.

It is important to recognize that we are living in a unique period of geological time when sea level is at a maximum highstand (Fig. 1). Consequently, sediment storage in alluvial drainage basins is high and sediment delivery to the Gulf of Mexico is low compared to other times during the glacial eustatic cycle. During this trip we will examine a number of sedimentary environments and discuss the role of short-term sedimentary agents (e.g. fluvial drainage, waves, and tides) in controlling facies architecture

Early Highstand (~ 120,000 to ~ 80,000 yr BP)

During the previous maximum highstand (120,000 yr BP), sea level was approximately 5 m above present (Fig. 1) and the shoreline was located approximately 20 km north of the present shoreline (Graf, 1966). Otherwise, the environmental setting along the east Texas coast was similar to the present and an extensive coastal barrier system evolved. These former shoreline deposits are known as the Ingleside Barrier and they extend across virtually the entire coastal plain.

As the great ice sheets of both hemispheres began to expand, sea level started to fall and rivers began to incise their channels. Fluvial terraces within relatively broad fluvial channels

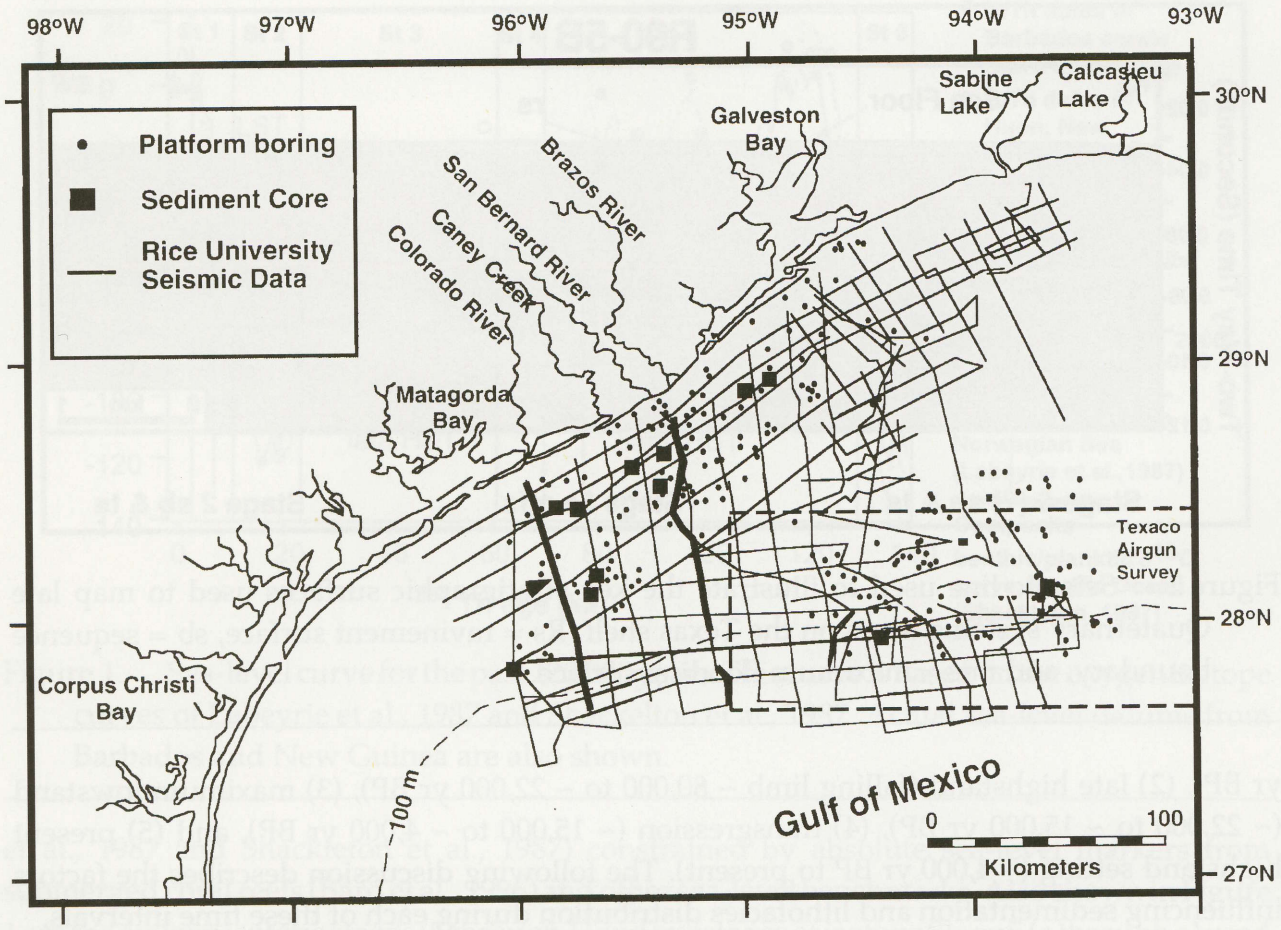


Figure 3. East Texas data set used in this study.

record the episodic fall in sea level (Thomas and Anderson, 1994). Fluvial incision resulted in increased quantities of sediment being transported from alluvial basins into the Gulf to construct vast highstand deltas from west Florida to the Rio Grande. The Brazos, and Colorado rivers, the two largest rivers in east Texas, constructed large highstand deltas on the east Texas continental shelf (Figs. 4, 5 and 6). Sediment supply to these deltas was large enough that they prograded into up to 70 meters of water, as indicated by the thickness of clinoforms (Fig. 5.).

The ancestral Colorado River formed a sandy, lobate delta on the inner shelf (Fig. 4). In contrast, the ancestral Brazos River deposited a more elongate muddy delta. Three phases of progradation of the Brazos delta are recognized. Based on detailed chronostratigraphic work, these deltas are assigned to stages 5e, 5c, and 5a, respectively (Fig. 4), (Abdulah et al., 2004; Anderson et al., 2004). Minor transgressions (5b and 5d) caused backstepping of the deltas. There is a virtual absence of large sand bodies within the early highstand Brazos delta complex. Sandy mouth bar deposits associated with these early highstand deltas were eroded as these deltas were subaerially exposed.

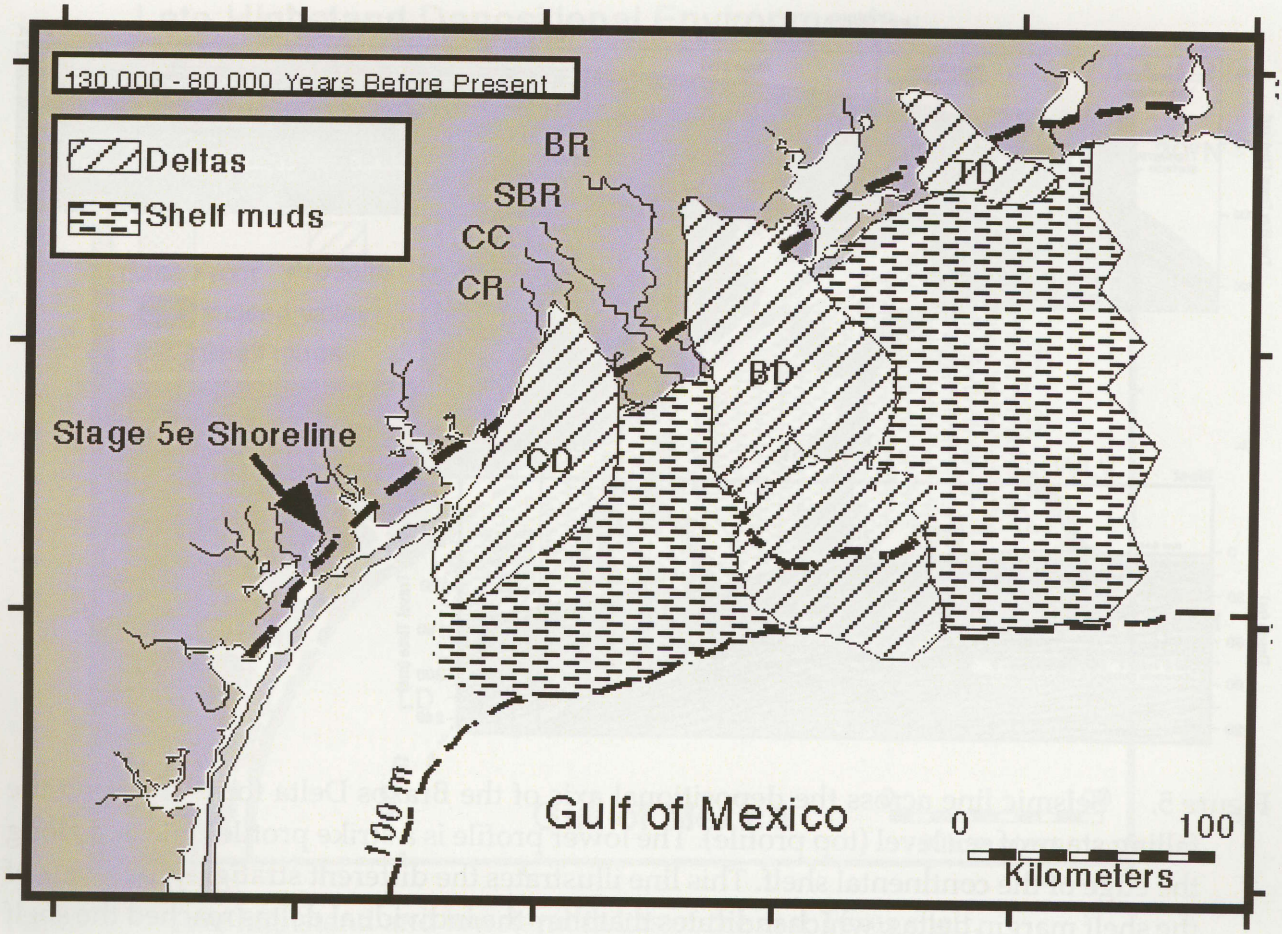


Figure 4. Early highstand paleogeography. BD = Brazos Delta and CD = Colorado Delta.

Late Highstand (Falling Limb ~ 80,000 to 22,000 yr BP)

From about ~ 80,000 to ~ 22,000 yr BP, the paleoshoreline episodically retreated from the mid-shelf to the shelf margin. This overall regression was interrupted by a major flooding event during oxygen isotope stage 3 (~ 60,000 to ~ 40,000 yr BP) that forced the paleoshoreline landward to approximately –15 meters (Rodriguez et al., 2000). During this flooding event, the ancestral Brazos and Colorado rivers, as well as ancestral rivers of western Louisiana, constructed large highstand deltas that prograded across virtually the entire shelf, filling the newly created accommodation (Figs. 5 and 6). A significant portion of the sediment supply of these deltas was derived by cannibalization of early highstand deltas. Detailed analyses of oil company platform borings, long sediment cores and seismic facies analyses revealed that these deltas included large, sandy delta fronts and that these sand bodies are preserved on the shelf.

The western Louisiana fluvial system also constructed a sizeable delta on the outer shelf, originally mapped by Lewis (1984) and Suter and Berryhill (1985) (Fig. 6). Sarzalejo's (1993)

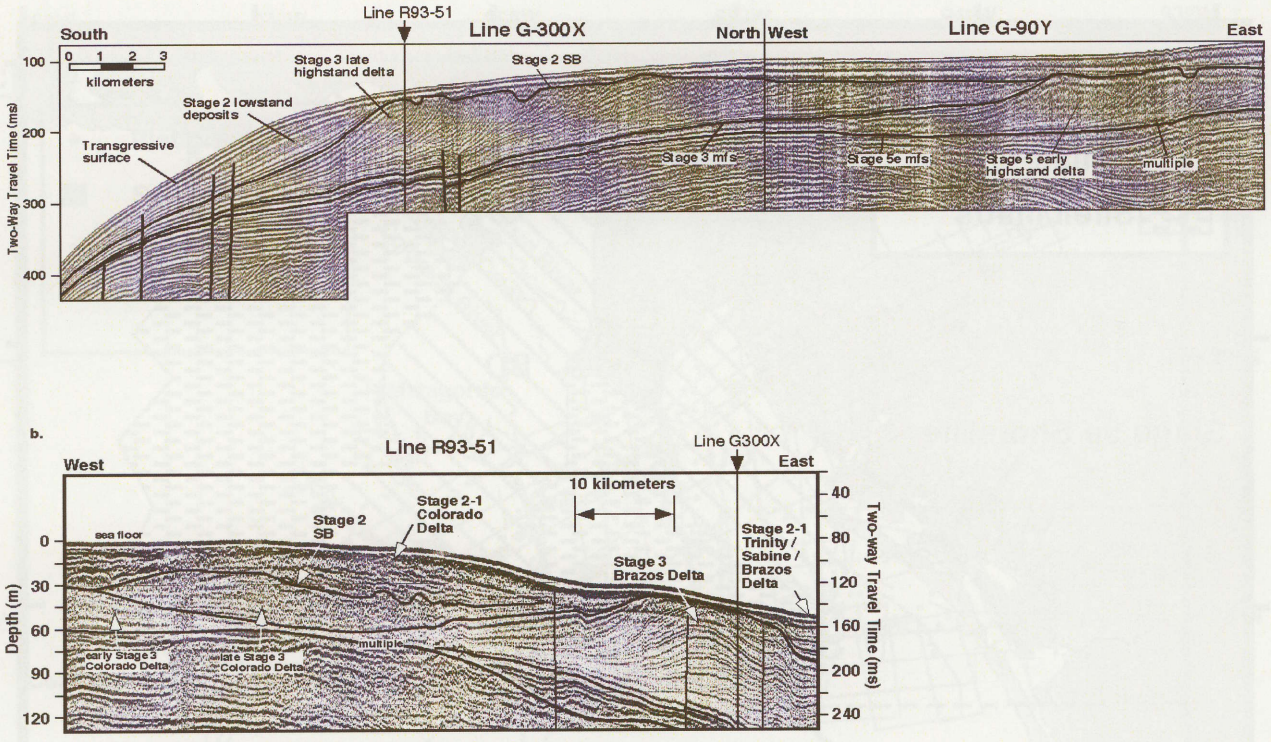


Figure 5. Seismic line across the depositional axis of the Brazos Delta formed during the falling stage of sea level (top profile). The lower profile is a strike profile collected along the edge of the continental shelf. This line illustrates the different stratigraphic levels of the shelf margin deltas, which indicates that they the individual deltas reached the shelf margin at different times in the overall sea level fall.

detailed work on this delta showed that the Western Louisiana Delta prograded across the shelf following the Stage 5e maximum highstand; it downlaps the maximum flooding surface. The delta reached the shelf edge by about 80,000 yr BP and was abandoned prior to the Stage 2 lowstand; it is cut by the sequence boundary (Fig. 5). Thus, it is an example of a shelf-margin delta formed during the falling limb of sea level. This explains why there is no associated lowstand delta or slope fan.

Sarzalejo's (1993) work showed that the western Louisiana delta consists of four delta lobes that overlap to form a widespread sandy mouth bar that is over 1600 km^2 in area. Because the delta is associated with the falling limb of sea level (forced regression of Posamentier et al., 1992), the mouth bar is incised into fine-grained distal bar and prodelta deposits. The result is a sharp contact between these facies as opposed to the idealized fining upwards succession.

Throughout the highstand and falling limb of sea level, the Trinity and Sabine rivers occupied the same fluvial valleys, which merge on the mid-shelf. A detailed analysis of the Sabine/Trinity valley system by Thomas and Anderson (1994) showed that the valley is

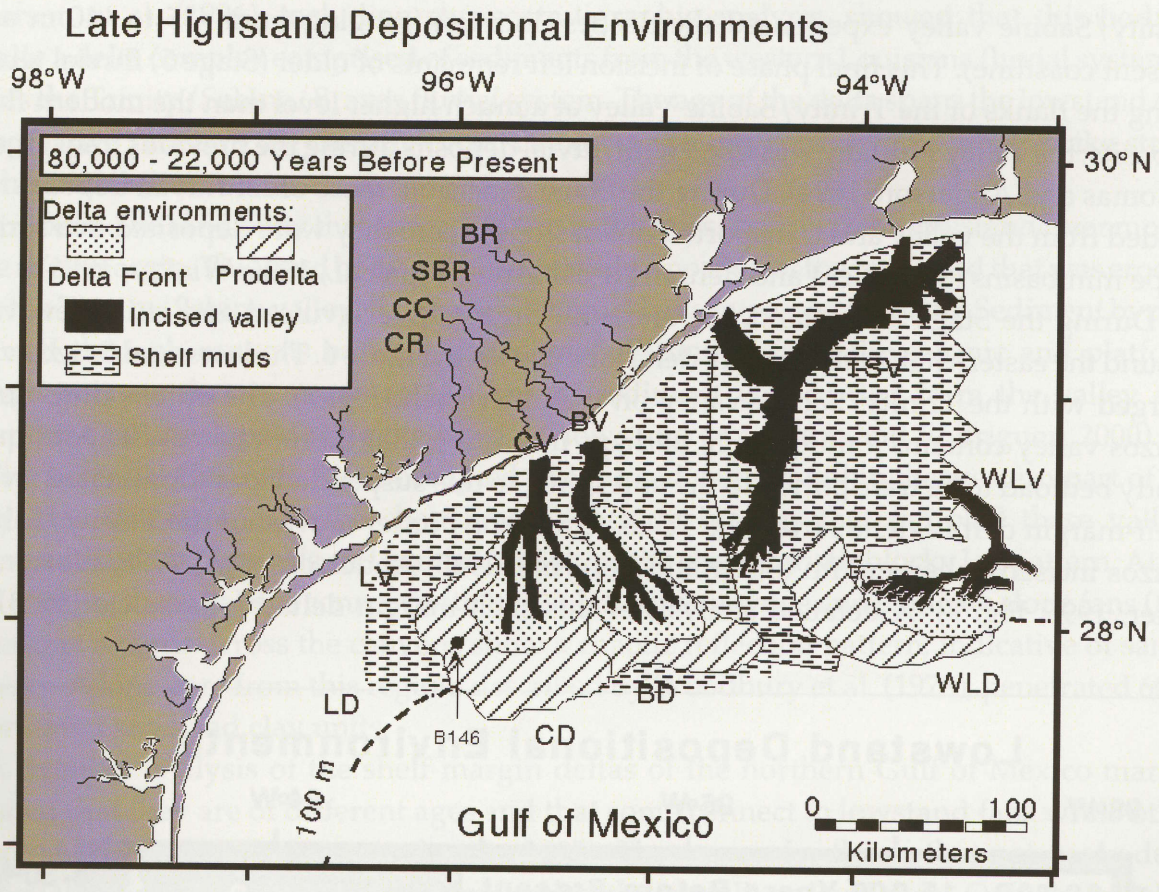


Figure 6. Late highstand and lowstand deltas of the east Texas shelf.

flanked by terraces that record subsequent stages of incision and filling caused by the episodic falls and rises in sea level during this time interval. Sediment supply to these rivers was never so great as to form deltas large enough to be preserved on the shelf.

Lowstand (22,000 to 15,000 yr BP)

Figure 7 shows the distribution of major depositional systems on the east Texas shelf and upper slope during the lowstand (~ 22,000 to ~ 15,000 yr BP). Prominent features of the lowstand include incised fluvial valleys, shelf-margin deltas, lowstand slope fans, and other sediment gravity flow deposits.

Initial incision of the fluvial valleys is believed to have occurred during oxygen isotope Stage 5d (~ 110,000 yr BP), based on seismic stratigraphic analysis by Thomas (1990), Sarzalejo (1993) and Abdulah (1995). Recall that the ancestral Colorado and Brazos rivers experienced several episodes of avulsion during the regression, hence there are several fluvial channels cut to various depths that extend seaward into deltas. In contrast, the Trinity and Sabine rivers occupied the same valleys throughout this time interval. During the Stage 2 lowstand, the

Trinity/Sabine valley experienced the deepest incision (ranging from –35 to –40 m at the present coastline). This final phase of incision left remnants of older (Stage 5) fluvial terraces along the flanks of the Trinity/Sabine Valley at a much higher level than the modern fluvial deposits; the valley virtually was filled with fluvial deposits during the previous transgression (Thomas and Anderson, 1994). During the Stage 2 incision, these older fluvial deposits were eroded from the valley and transported down slope where they were deposited in a series of slope minibasins (Satterfield and Behrens, 1990; Anderson et al., 1996; Winker, 1996).

During the Stage 2 lowstand, the ancestral Brazos River avulsed and cut a new valley around the eastern half of the muddy delta lobe shown in Figure 6. The lowstand Brazos valley merged with the Trinity/Sabine valley on the outer shelf (Fig. 7). The outer portion of the Brazos valley contains a muddy fill, perhaps indicating that there was insufficient time for sandy bedload to be transported to the outer shelf in this suspended load-dominated river. A shelf-margin delta extends from the Texas–Louisiana border to west of the Trinity/Sabine/Brazos incised valley mouth (Suter and Berryhill, 1985, Anderson et al., 1996; Morton and Suter, 1996, Fig. 7). A detailed analysis of this shelf-margin delta by Sarzalejo (1993) and

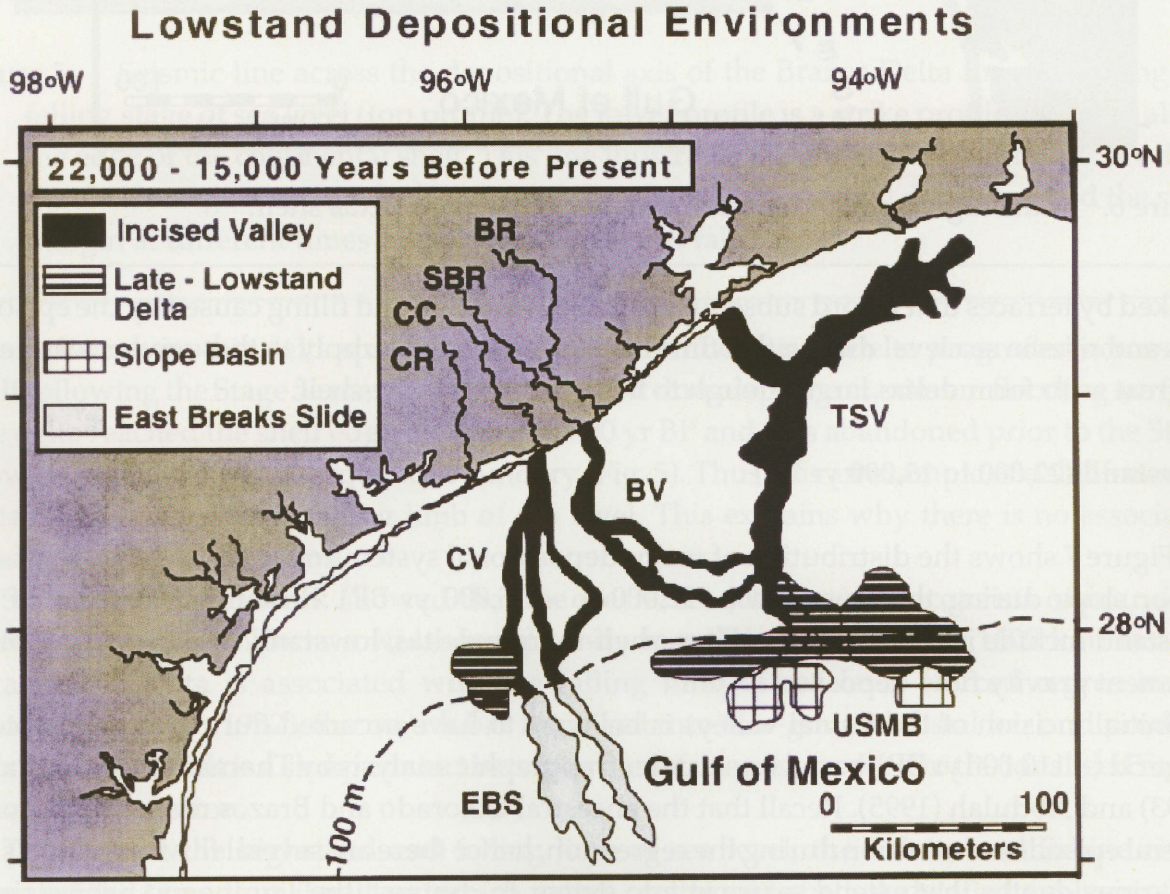


Figure 7. Lowstand depositional systems of the east Texas shelf.

Anderson et al. (1996), including chronostratigraphic analysis, showed that this body is actually a delta complex comprised of sediments from the western Louisiana fluvial system as well as the Trinity / Sabine / Brazos fluvial system. The age of the delta spans the lowstand and transgressive phase of the eustatic cycle. The delta exhibits considerable along-strike stratigraphic variability.

An extensive sand body that averages 20 m in thickness occurs at the lowstand river mouth at the shelf margin. This sand body accounts for only a portion of the total sand that was eroded from the Trinity / Sabine valley; the remainder must have bypassed the shelf. Sediment bypass occurred through canyons between salt domes. Estimates from the seismic and platform borings indicate that as much as 120 km^3 of sediment was eroded from the valley and transported down-slope during the previous lowstand (Anderson and Rodriguez, 2000).

The ancestral Colorado River occupied two incised valleys within the western part of the study area that extend to the shelf break. Platform borings that sampled these valleys recovered sands and gravels and gamma logs show the characteristic blocky log pattern. At the shelf margin, these valleys connect with two submarine canyons and elongate slope fans (Fig. 7). Seismic records across the canyons show a chaotic reflection pattern, indicative of sandy deposits. A long core from this region, described by Woodbury et al. (1978), penetrated 60 m of contorted sand and clay units.

A detailed analysis of the shelf-margin deltas of the northern Gulf of Mexico margin revealed that they are of different ages and that some connect to lowstand fans while other do not (Anderson et al., 2004, Anderson, 2005). The reason for this is that not every delta reached the shelf margin at the same time within the overall sea level cycle. Only those deltas that where actively being sourced during the maximum lowstand are connected to lowstand fans. Thus, not every shelf-margin delta is a lowstand delta. The Brazos River, for example, nourished a delta that prograded to the shelf margin prior to the maximum lowstand and avulsed from that location prior to the lowstand. So, its shelf margin delta is not connected to a lowstand fan.

The Colorado River, on the other hand, remained in the same location throughout the sea level cycle and it shelf-margin delta is connected to a lowstand fan. So, only the incised river valleys that are associated with the sequence boundary are apt to extend down dip into lowstand fans.

Transgressive Systems Tracts (~ 15,000 to ~ 4,000 yr BP)

Figure 8 shows the prominent elements of the transgressive systems tract. These include incised valley-fill deposits, shelf sand bodies, and backstepping deltas.

Rates of sea-level rise during the Stage 2 to 1 transgression have ranged from 0.7 cm / a during the first 15,000 years of the transgression to 0.3 cm / a during the last 3,500 years (Fairbanks, 1989). The overall rise is punctuated by episodic rapid rises on the order of 2 to 5 cm / a. While these rapid rises are small in magnitude (probably less than 3 m within a few

centuries), they manifest themselves as dramatic landward shifts in facies. The episodic nature of sea-level rise during the last transgression is believed to be characteristic of glacial eustatic rises and is, therefore, representative of the past 35 million years of Earth history, or since the Antarctic Ice Sheet evolved.

Many of the transgressive deposits of the continental shelf are situated at or near the sea floor and, therefore, within reach of shallow sediment coring methods. We have conducted several studies of these transgressive deposits and will discuss them during the course of the field trip. From a reservoir standpoint, the main depositional systems that include reservoir-scale sand bodies include transgressive deltas, sand banks and incised valleys.

Present Highstand (~ 5,000 yr BP to Present)

The modern depositional systems associated with the present highstand, which began approximately 5,000 yr BP, are shown in Figure 9. This period is characterized by a slow rise in sea level that permitted the development of an extensive chain of barrier islands, peninsulas and chenier plains. This period is also characterized by significant aggradation,

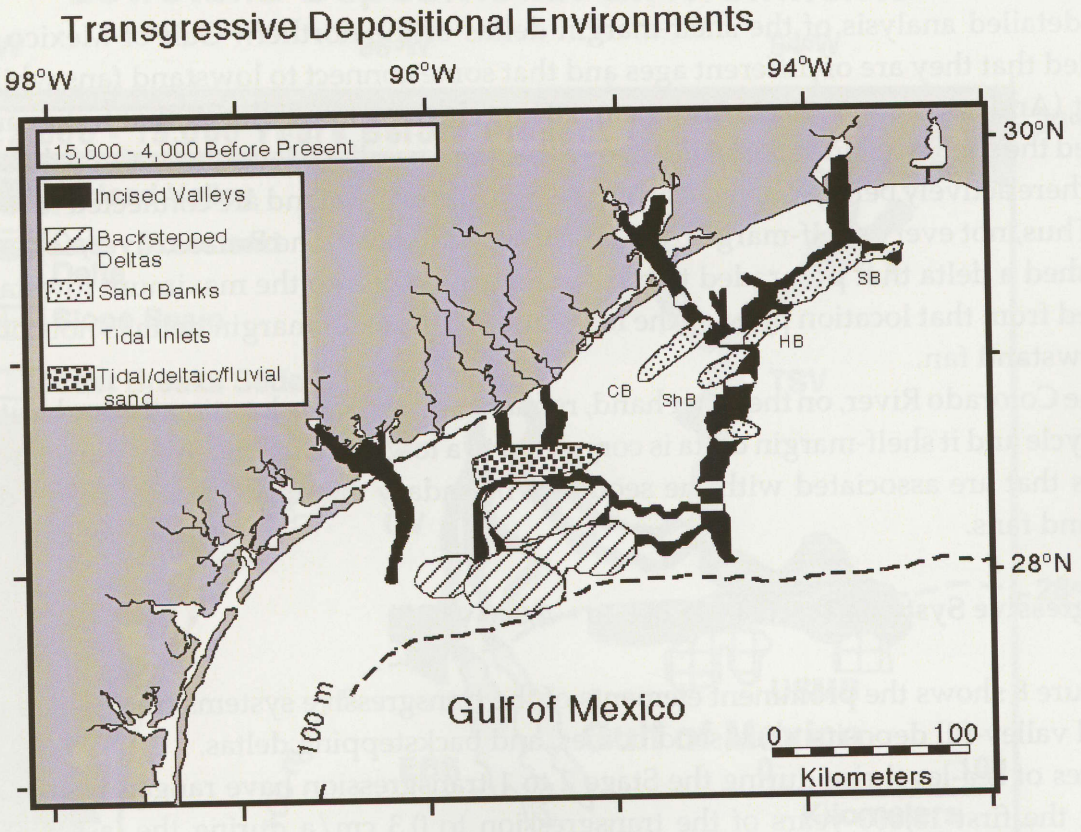


Figure 8. Paleogeographic map showing major depositional systems that formed during the last transgression.

with fluvial stockpiling of sediment within drainage basins. The higher sediment supply of the Brazos and Colorado rivers relative to the Sabine, Trinity, and Lavaca rivers continues to exert a strong influence on the style of sedimentation in the fluvial valleys. Bays occupy the upper reaches of the Trinity and Sabine valleys (Galveston Bay and Sabine Lake respectively), but the Brazos and Colorado rivers, with their larger sediment supplies, have begun to reconstruct prominent deltaic headlands into the Gulf of Mexico. Perhaps these rivers have reached the point where they will again begin to construct fluvial deltas across the continental shelf.

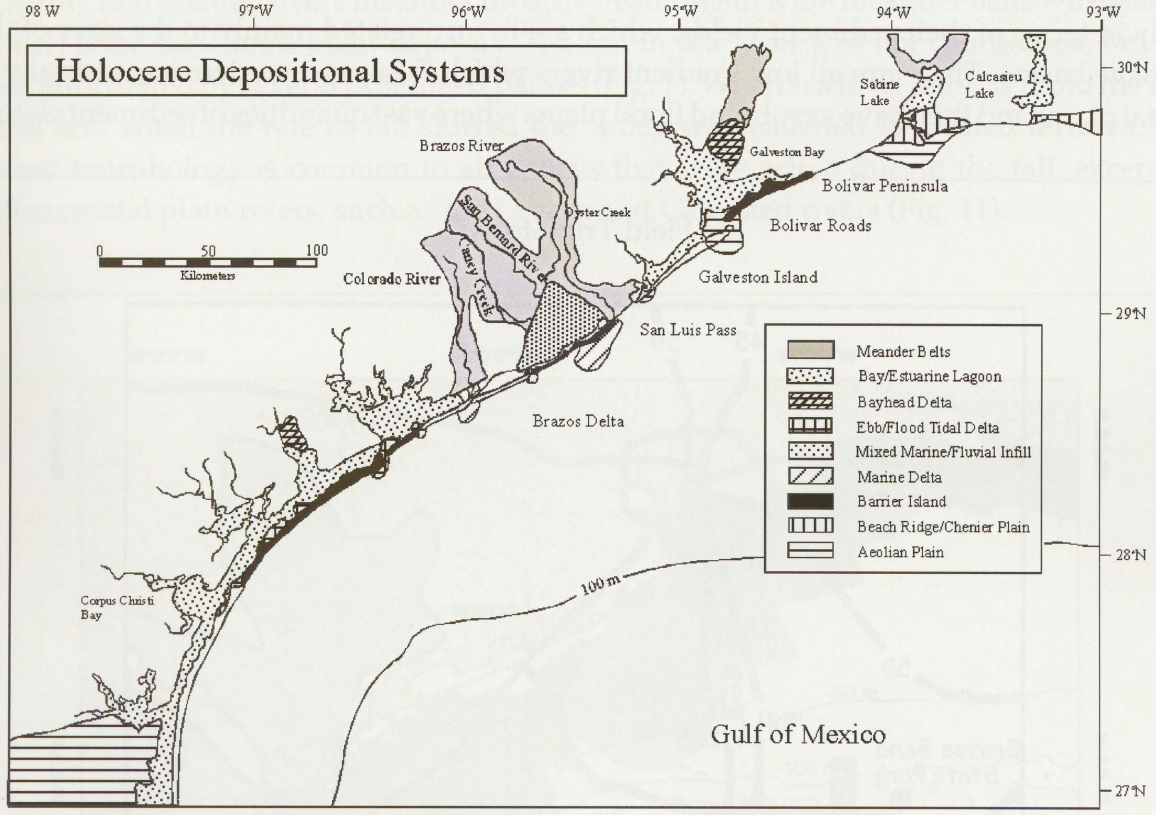


Figure 9. Holocene depositional environments of the of the upper Texas coast.

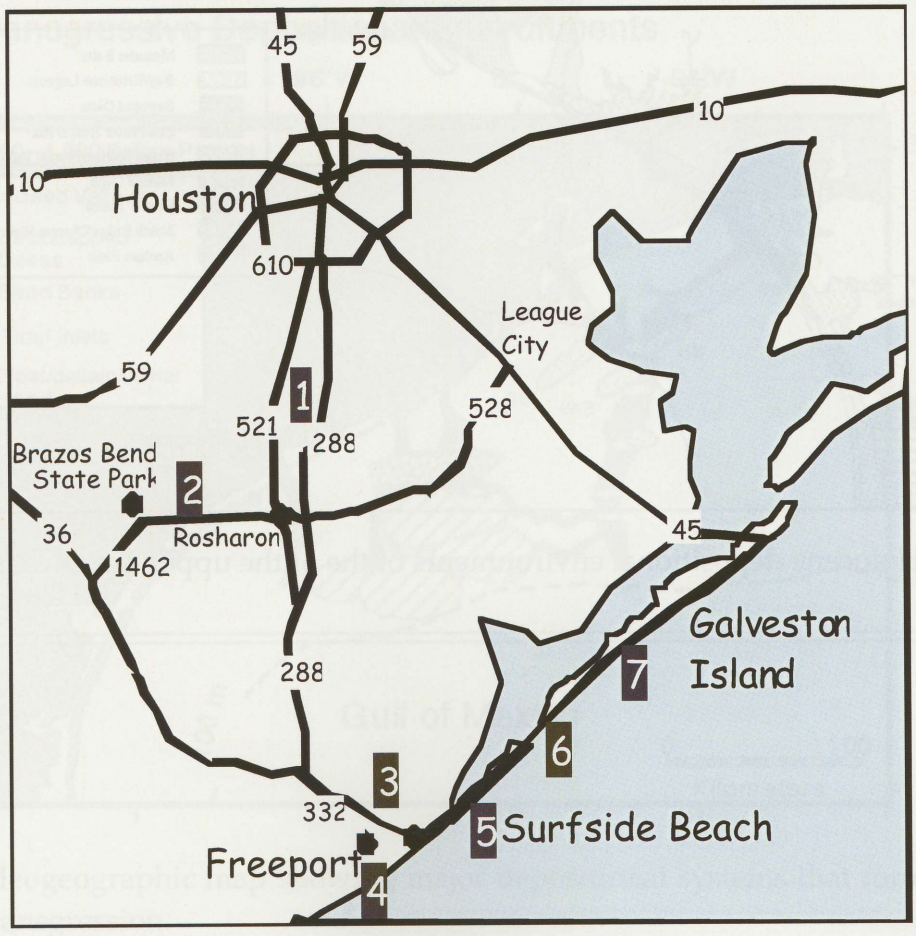
DEPOSITIONAL SYSTEMS

We will begin our trip by looking at fluvial and deltaic deposits associated with the modern and ancient Brazos River. The day will end with a stop at the San Luis Pass tidal delta.

Stop 1
Ancient (Highstand) Fluvial Deposits of the Brazos River

The rivers of east Texas (Sabine, Trinity, Brazos, Colorado, La Vaca rivers) vary significantly in terms of their sediment yields, which are in turn related mainly to the sizes of their drainage basins. These are all low gradient rivers which flow across extensive, virtually flat coastal plains and they have very broad flood plains where vast quantities of sediment is stored

Field Trip Stops



during highstands (Fig. 8). Indeed, these alluvial valleys are currently filled virtually to capacity with sediments. The exact response of these rivers to climate and base-level change, in terms of sediment transport from the alluvial basins to the Gulf, is still poorly understood. We do know that the sediment yields of these rivers today is significantly less than it was during most of the last eustatic cycle. Today, the Brazos and Colorado rivers form small wave-dominated and bay-head deltas, respectively. During the previous highstand, both rivers nourished vast fluvial dominated deltas (Figs. 4, 5, and 6). The fluvial channels of these ancestral rivers are preserved on the shelf and represent large reservoir-scale sand bodies within the highstand systems tract (Fig. 10).

During highstands, rivers meander broadly, resulting in wide meander belts with stacked channel/point bar/flood plain deposits. The fall in sea level was not continuous, but was rather punctuated by several prolonged pauses (Fig. 1). When sea level fall was rapid the river incised and when the rate of fall slowed the valley was widened to created terraces. This terraced morphology is common to all valleys that were active during the fall, except for smaller coastal plain rivers, such as the Lavaca and Calcasieu rivers (Fig. 11).

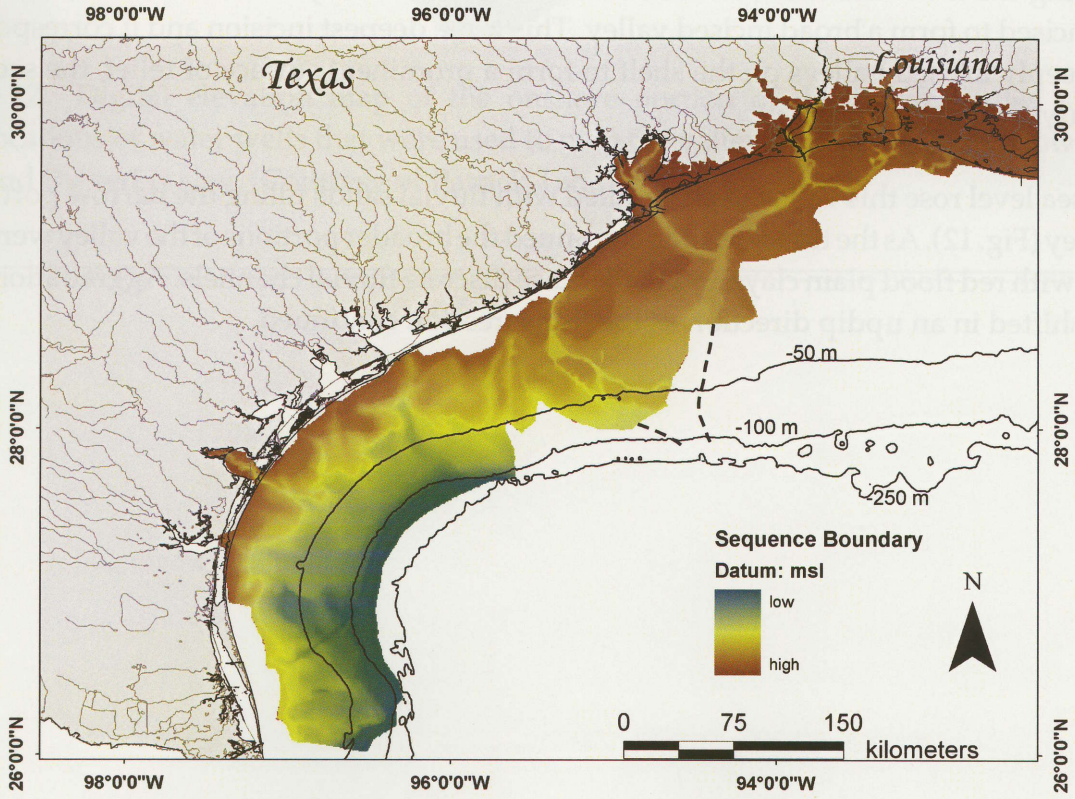


Figure 10. Digital elevation map of the Stage 2 sequence boundary. The major river incised valleys are shown in blue (from Simms et al., in press).

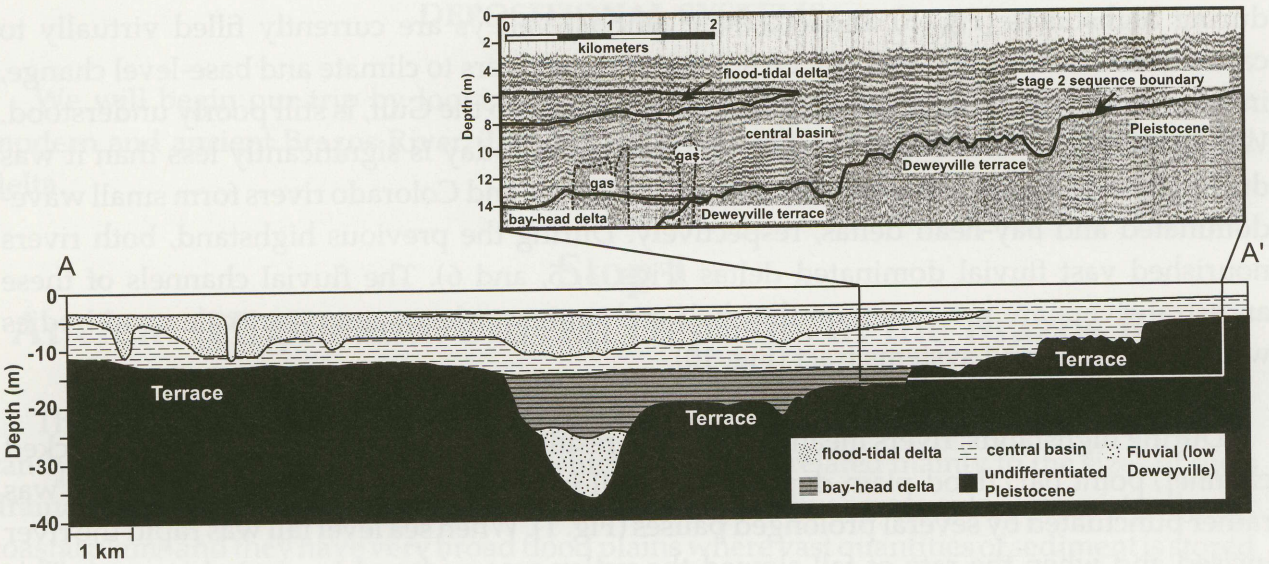


Figure 11. Seismic line and interpreted seismic line across Galveston Bay (Trinity Valley) illustrating terraced morphology of the Trinity incised valley.

During the lowstand that occurred between approximately 22 ka and 17 ka, the Brazos River incised to form a broad incised valley. This is the deepest incision and it corresponds to other deeply incised valleys on the shelf to form a prominent surface of relief, the sequence boundary (Fig. 11).

As sea level rose this valley initially filled with fluvial sands filling the narrow portions of the valley (Fig. 12). As the transgression continued the broader portions of the valley were filled mostly with red flood plain clays and that encase discrete fluvial channels. Aggradation of the valley shifted in an updip direction as the transgression continued.

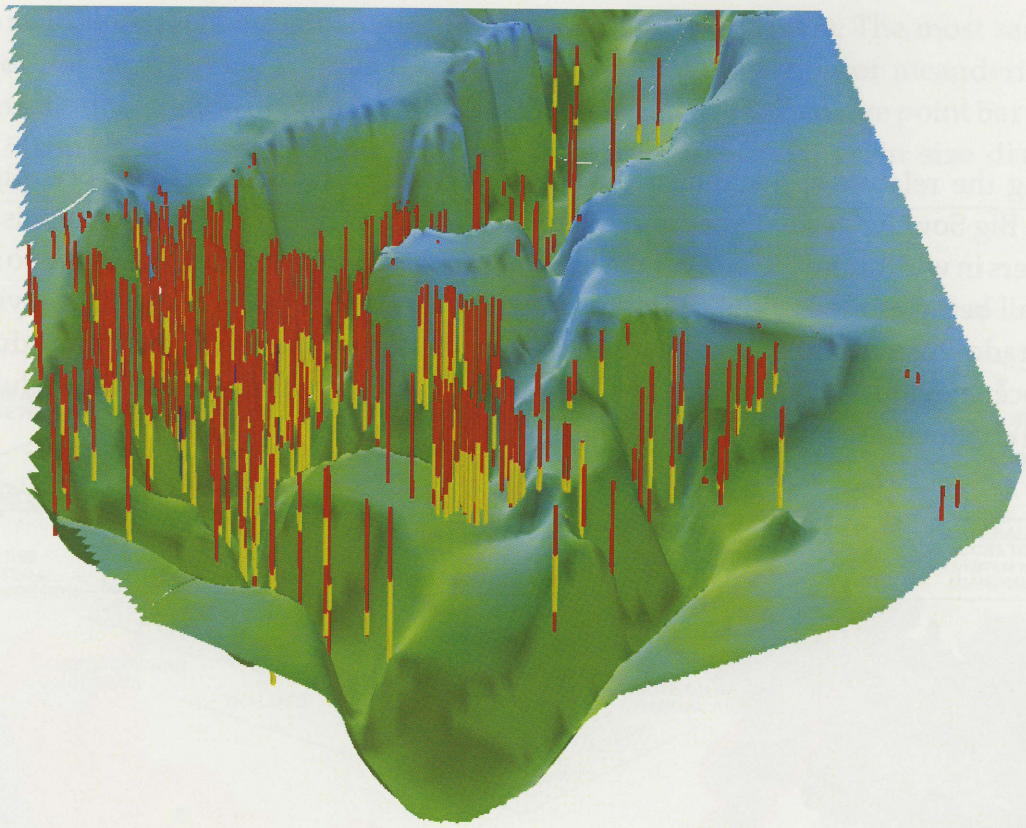


Figure 12. Digital elevation map of the onshore portion of the Brazos valley showing locations of water wells that were used to map the valley and its fill. The yellow is sand and the red is clay. (Courtesy of Patrick Taha)

Stop 2
The Modern Brazos River

During the relative still stand of the past 6,000 years the Brazos River occupied three channels, Big Sough, Oyster Creek and the modern channel (Fig. 13). These channels average 5 kilometers in width, and are on average less than 15 meters deep. Because they are so shallow, they are all being eroded as the shoreline advances landward via transgressive ravinement. The Colorado River occupied two channels, Caney Creek and the modern channel, during the late Holocene. Sand is confined to the basal portions of the Brazos channels, whereas the Colorado channels are mostly filled with sand.

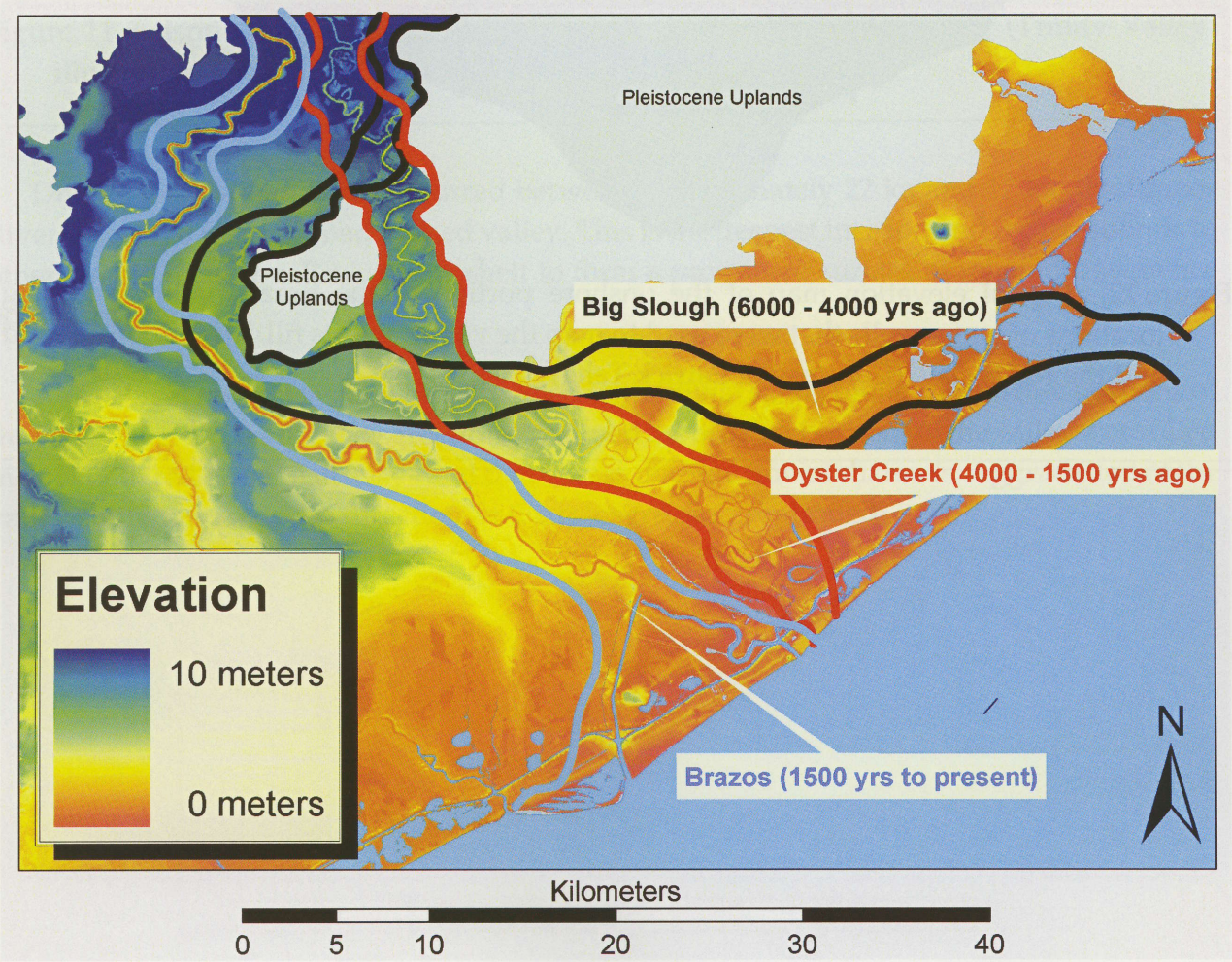


Figure 13. Map showing the locations of the ancestral Brazos River channels that were occupied during the Holocene transgression. (Courtesy of Patrick Taha)

Figure 14 illustrates the subenvironments of meandering rivers. The most sand-prone portion of the meandering river is the point bar. Lateral migration of meandering rivers results in lateral accretion of point bars and the development of extensive point bar deposits. Figure 15 shows an idealized point bar sequence and associated grain size distribution

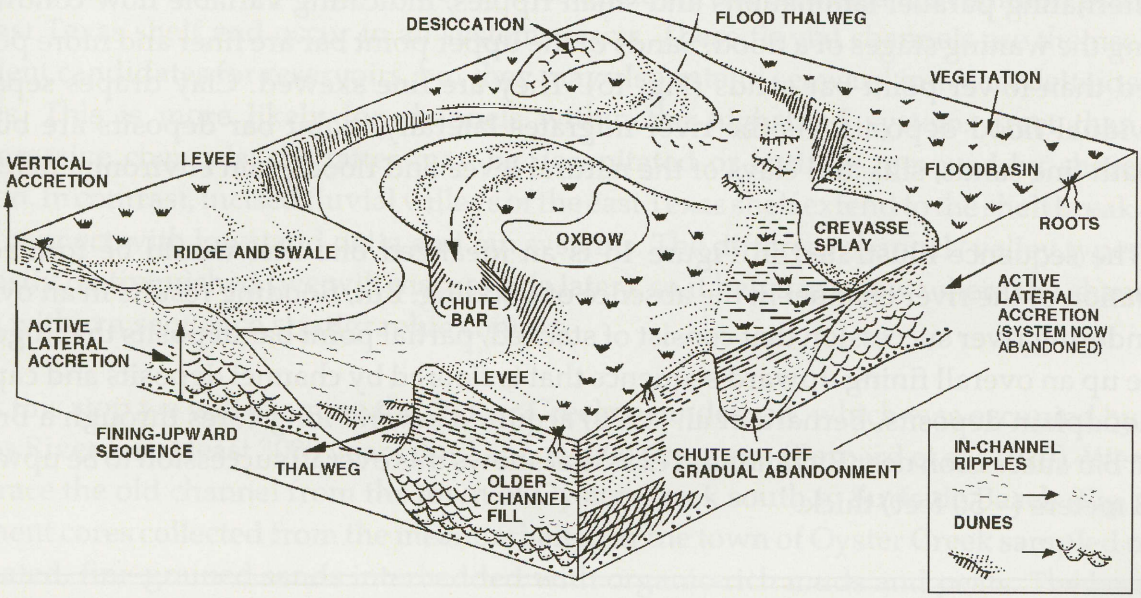


Figure 14. Subenvironments and facies of a meandering river.

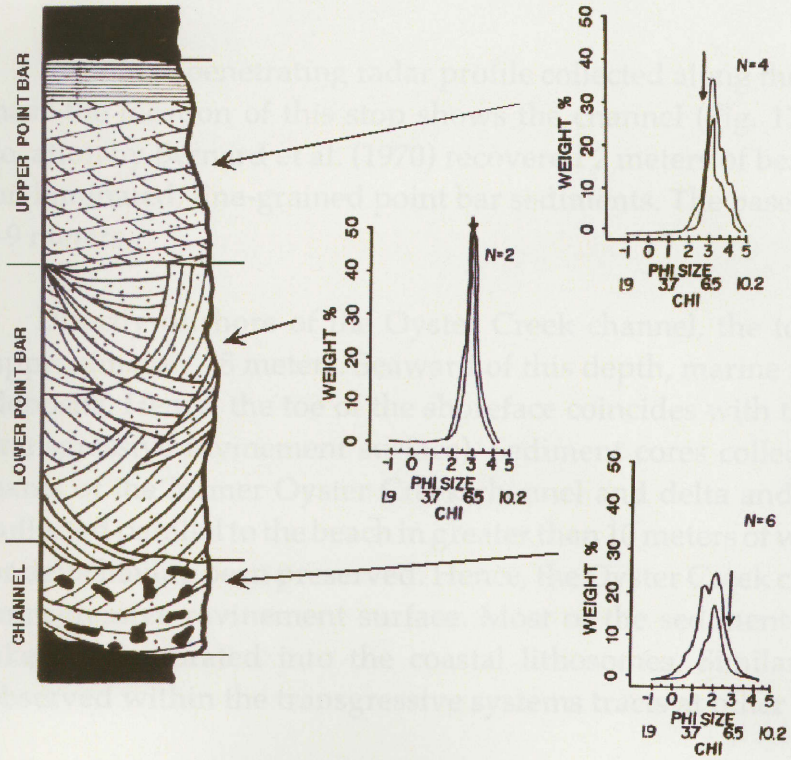


Figure 15. Sedimentary structures and grain size data for a representative point bar deposit.

curves. Cut-and-fill structures and large-scale transport indicators (i.e., sand waves and troughs) dominate the basal channel deposit with clay rip-ups. The grain-size distributions of sands from the channel are coarse, poorly sorted, and negatively skewed. The lower point bar sequence includes a basal unit with giant ripple forsets and trough stratification. Sands within this part of the sequence are well-sorted (Fig. 15). The upper point bar is characterized by alternating parallel laminations and small ripples, indicating variable flow conditions during the waning stages of a flood. Sands of the upper point bar are finer and more poorly sorted than lower point bar sands (Fig. 15). They are fine skewed. Clay drapes separate individual flood deposits. As the river migrates laterally, point bar deposits are buried beneath fine sands, silts and clays of the natural levee and flood plain environments.

The sequence illustrated in Figure 15 is an idealized one that would be formed by migration of the river channel in the absence of flooding. But, flooding results in an overall meandering river succession that consist of stacked, partial point bar deposits that together make up an overall fining upward sequence that is floored by channel deposits and capped by floodplain deposits. Bernard et al. (1970) acquired several long cores through a Brazos Point Bar succession near Richmond, Texas that shows this overall succession to be upwards of 18 meters (~ 50 feet) thick.

Stop 3
The Oyster Creek Meanderbelt and Transgressive deltas

To an exploration geologist, the difference between fluvial channels and incised fluvial valleys is non-trivial. As we have already discussed, fluvial channels are common place on the east Texas shelf and occur in all systems tracts. These fluvial channels are themselves excellent candidates for reservoirs, but they may also extend seaward into vast deltaic sand bodies. This is more likely for channels within the highstand systems tract than for transgressive channels; the latter may be decapitated or entirely removed by shoreface erosion. In contrast, incised fluvial valleys of the east Texas shelf extend to the shelf break and some connect with lowstand delta and fan systems. The different channel/valley types are not easily distinguished (we will discuss this later), so it helps to know where the channels occur within a sequence stratigraphic context.

At this stop we will examine the Oyster Creek meanderbelt, which was occupied by the Brazos River for at least 3000 years until about 1000 years ago (Bernard et al., 1970). We will first trace the old channel from the town of Oyster Creek south to Surfside Beach (Fig. 16). Sediment cores collected from the meanderbelt near the town of Oyster Creek sampled only laminated, fine-grained sands interbedded with organic-rich muds and peats. The base of the channel at this location is at about 10 meters. Thus, the fluvial succession on this relatively flat portion of the coastal plain is rather different from that which occurs just 20 kilometers to the north.

A ground-penetrating radar profile collected along the highway and across the channel near the location of this stop shows the channel (Fig. 17). A sediment core taken at this location by Bernard et al. (1970) recovered 2 meters of beach sand with shell layers resting on laminated, fine-grained point bar sediments. The base of the channel was penetrated at –9 meters.

Directly offshore of the Oyster Creek channel, the toe of the shoreface is situated at approximately –8 meters. Seaward of this depth, marine muds rest directly on Pleistocene deposits. Hence, the toe of the shoreface coincides with the level of shoreface erosion (the transgressive ravinement surface). Sediment cores collected offshore sampled only remnants of the former Oyster Creek channel and delta and high-resolution seismic profiles collected parallel to the beach in greater than 10 meters of water show no signs of the channel or delta having been preserved. Hence, the Oyster Creek channel terminates abruptly at the transgressive ravinement surface. Most of the sediment removed from the channel was likely incorporated into the coastal lithosomes. Similar stratigraphic relationships are observed within the transgressive systems tracts at other locations on the shelf.



Figure 16. Aerial photograph of Oyster Creek at Surfside Beach. The yellow line marks the location of the GPR profile shown in figure 17.

Not all fluvial channels within the transgressive systems tract are removed by transgressive erosion, which is to say they do not terminate down dip at the transgressive ravinement surface. Each time the Brazos and Colorado rivers avulsed during the transgression these new river channels adjusted to a higher base-level. Thus, younger channels are shallower and more prone to removal by transgressive ravinement. Of course, the lowstand incised valley will be the deepest and, therefore, have the greatest preservation potential. Figure 18 shows an example of a fluvial channel, in this case a Brazos River channel, that has been cut by the transgressive ravinement surface (rs). Note that the sands eroded from the channels were deposited as a transgressive sand body that consists of westward dipping cross-beds

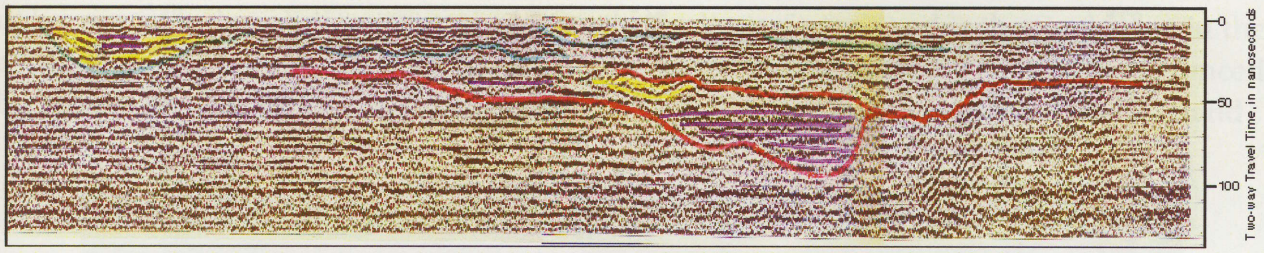


Figure 17. GPR survey across Oyster Creek paleochannel near Surfside Beach

that downlap the transgressive surface (Fig. 18). This mechanism of mining sands from paleo-channels is believed to have played a major role in sourcing modern coastal lithosomes of Texas (Anderson et al., 1996).

About 60 kilometers west of this location is where the ancestral Colorado River crosses the shoreline. Approximately 30 kilometers offshore, a large sediment body, which Snow (1998) interpreted as a transgressive delta of the Colorado River, exists (Fig. 19). Platform borings and long sediment cores indicate that the delta is composed primarily of sand. The age of the delta was radiometrically constrained to 11,500 and 9,500 radiocarbon years BP, which is consistent with its location on the shelf and published sea-level curves. The total volume of the delta is 6.3 km^3, which equates to a sediment flux of 3.2 km^3 / 1000 yr, compared to a modern sediment yield for the Colorado River of 0.3 km^3 / 1000 yr. From where did all this sediment come? Snow (1998) found that the formation of the delta coincided with a climatic transition to dryer and warmer conditions, which may have led to widespread stripping of soils and possible re-organization of the drainage system due to lowering of the

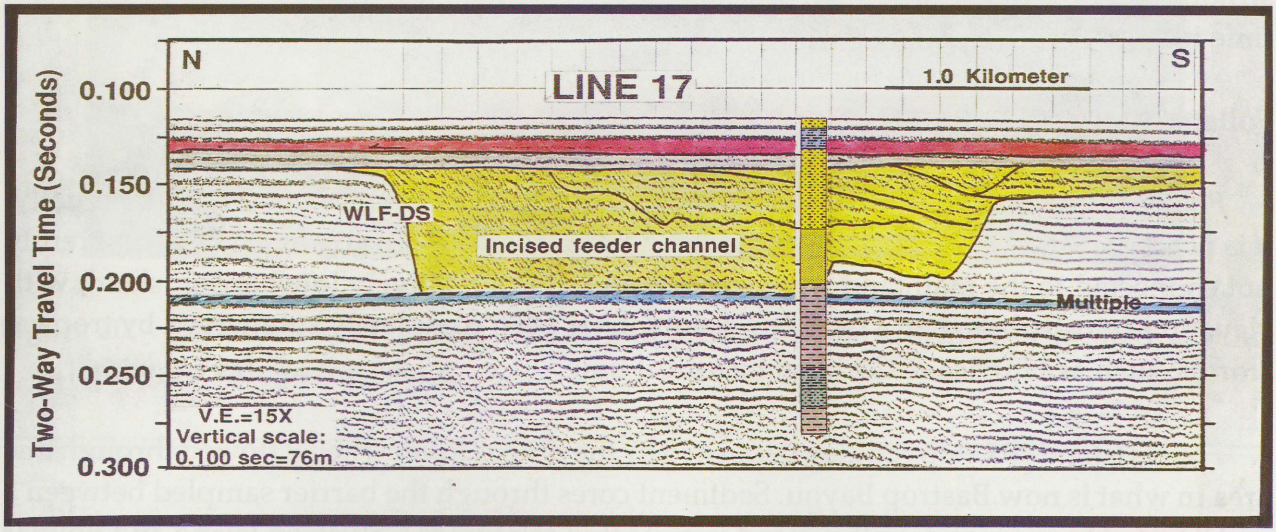


Figure 18. Seismic line across from the western Louisiana inner shelf that crosses an incised valley.

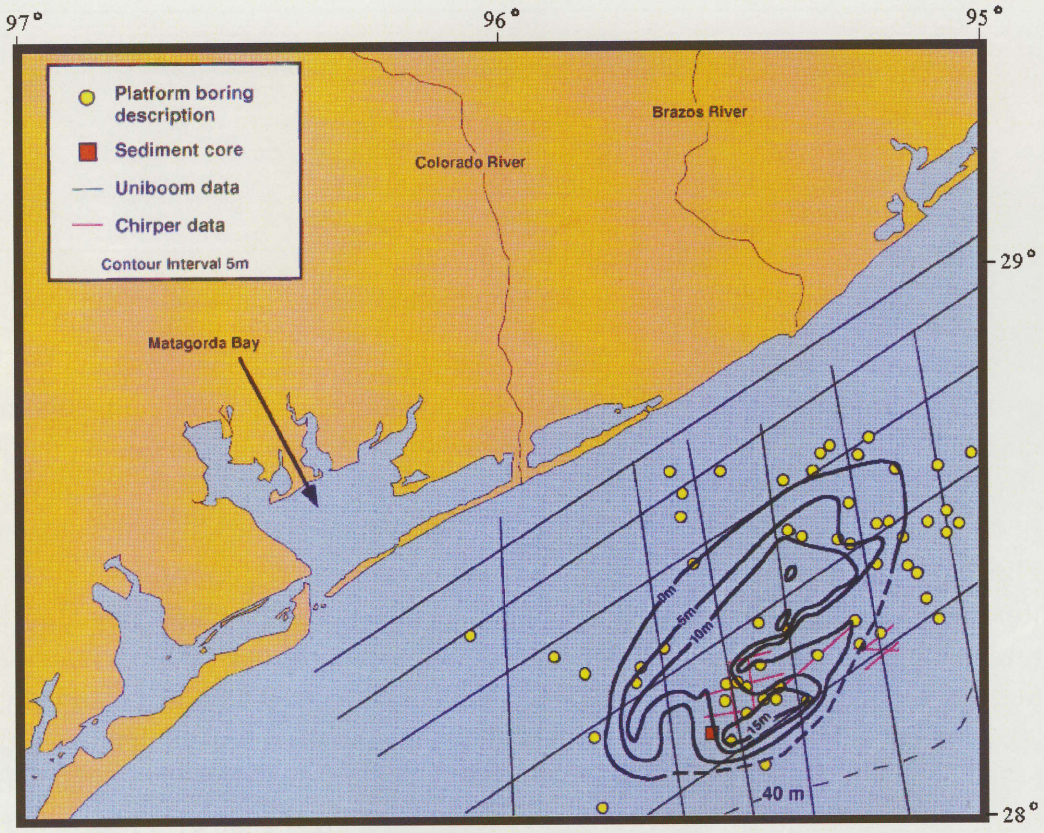


Figure 19. Isopach map of the transgressive Colorado Delta.

water table (base-level). Regardless of the exact mode of origin of the delta, it serves as striking example of the formation of a large (3200 km^2) sand body within a relatively small time interval.

Follets Island

Follets Island is a small barrier island located west of Galveston Island (Fig. 20). Actually, it is no longer an island, but when it was first mapped over a century ago it was apparently not connected to the mainland at its western end. It is a relatively low barrier island, with elevations mostly less than 4 feet. As a result, it has been completely submerged by tropical storms, such as Hurricane Carla, in historical time.

Follets Island sits atop red clay deposited by the Brazos River when it flowed through this area in what is now Bastrop Bayou. Sediment cores through the barrier sampled between 3 and 8 feet of barrier sand, so the island is quite thin relative to Galveston Island and Bolivar Peninsula. Unlike Galveston Island, Follets Island is narrow and a significant amount of the sand that is being eroded from the island washes over into back-barrier wetlands. This stage

of barrier island evolution is referred to as the "rollover" stage and is the final stage in the evolution of the barrier. The island is highly prone to breaching during storms, as evidenced by numerous storm channels and wash-over deposits on the landward side of the island.



Figure 20. Aerial view of Follets Island and Christmas Bay, which is located on the landward side of the island.

Stops 4 and 5
Brazos Delta

In 1929, the U.S. Corps of Engineers diverted the Brazos River mouth to its new channel west of Freeport. The present Brazos Delta, subsequently referred to as the new delta, was constructed and marine processes destroyed the old delta. The new delta is approximately 35 square kilometers in area and extends offshore to water depths of –20 meters (Fig. 21). The

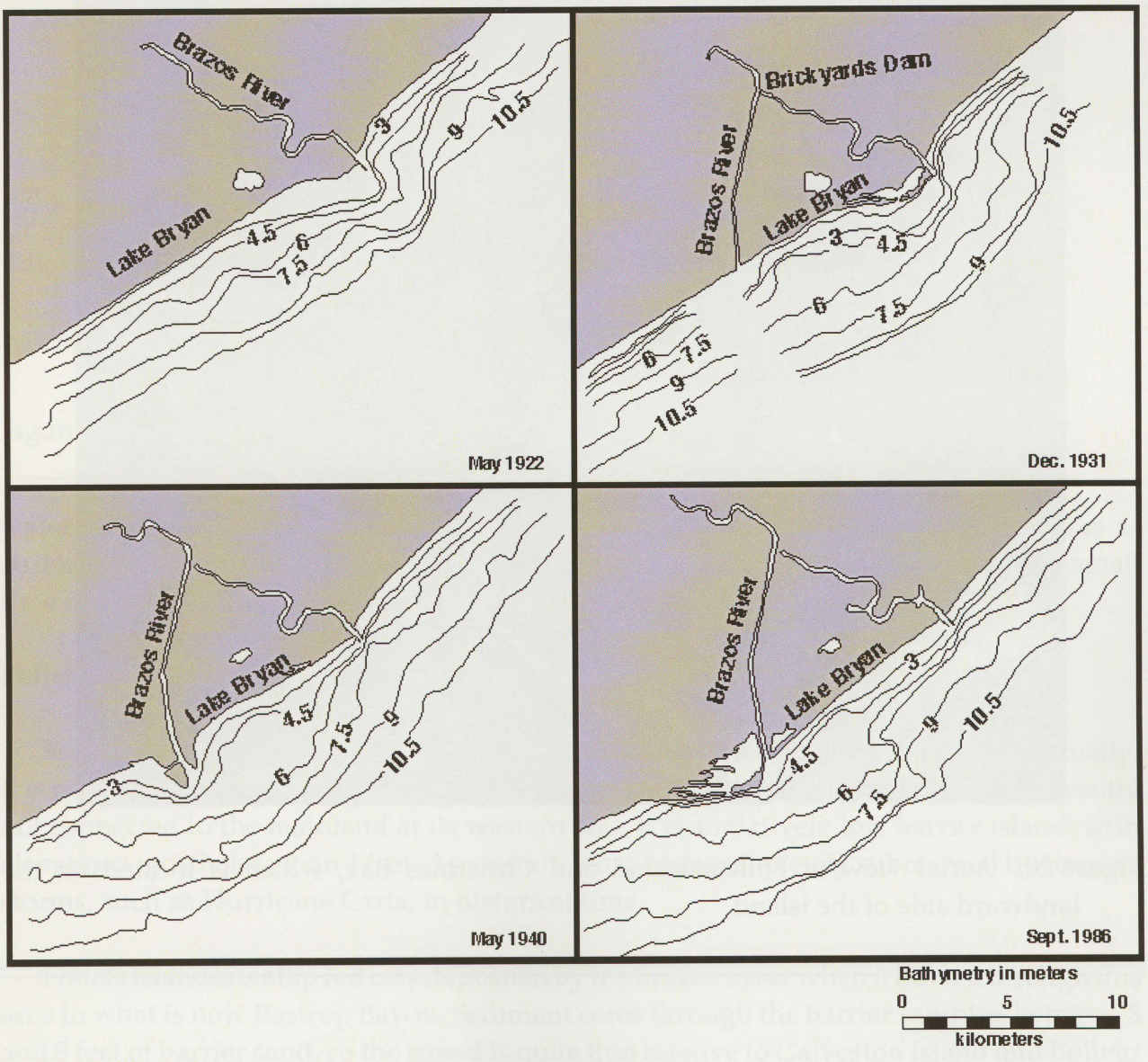


Figure 21. Bathymetric maps constructed over several decades illustrate the changes in the offshore portions of the Brazos delta before and after the river's diversion in 1929.

morphologic outline of the onshore deltas is lobate to lunate, with a significant headland on the western flank of the delta. Deposition is asymmetrical and accretes to the west along the projecting headland (Fig. 22).

The new Brazos delta is mostly situated above the level of shoreface ravinement, so it has little preservation potential during a transgression. However, we know that during the last highstand, and during previous highstands throughout the Neogene, the Brazos River constructed sizable deltas on the continental shelf. Therefore, preservation potential of the delta is high during highstands.

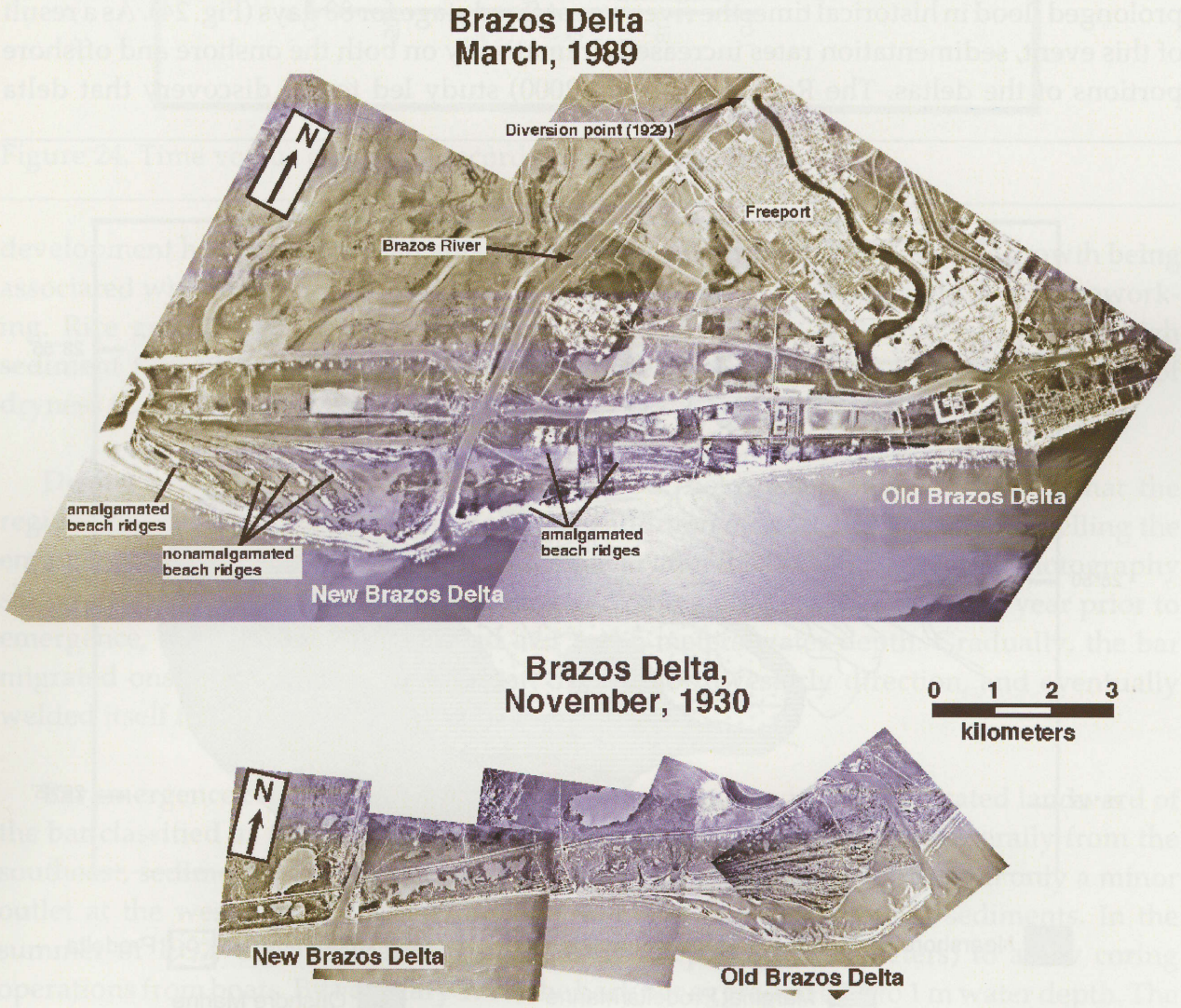


Figure 22. Historical photographs showing changes in the Brazos delta after the river was diverted in 1929.

Following diversion of the river in 1929, the delta experienced a rapid phase of growth that was due largely to sand being eroded from the pre-1929 delta and transported alongshore to the new delta. During this phase of growth, the Brazos delta was strongly wave-influenced and the facies architecture of the delta was comprised of well sorted sand of stacked beach ridges (Bernard et al., 1970).

Rice graduate students Mike Hamilton and Tony Rodriguez studied the Brazos Delta (Rodriguez et al., 2000). More than a hundred sediment cores were collected onshore and offshore during these two studies. The facies map shown in Figure 23 is a product of their work.

During December through February of 1992, the Brazos River experienced the most prolonged flood in historical time; the river was at flood stage for 80 days (Fig. 24). As a result of this event, sedimentation rates increased dramatically on both the onshore and offshore portions of the deltas. The Rodriguez et al. (2000) study led to the discovery that delta

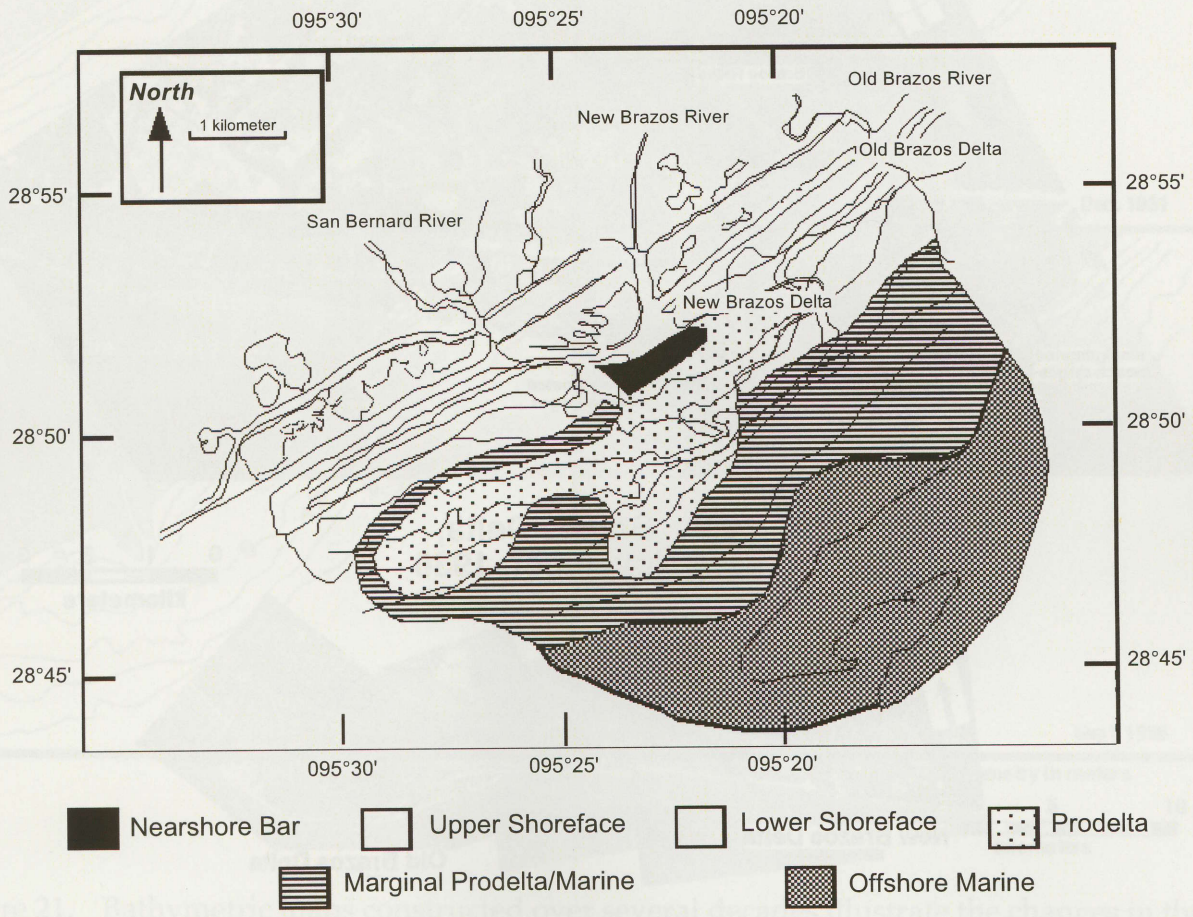


Figure 23. Sedimentary facies of the modern Brazos Delta.

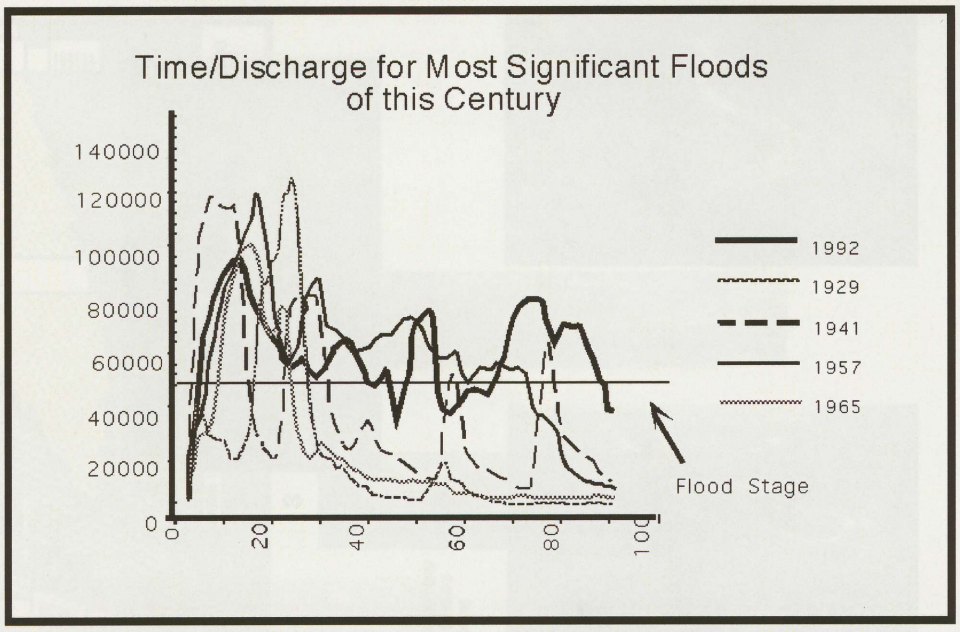


Figure 24. Time versus discharge record for the Brazos River.

development has occurred in an episodic fashion, with episodes of dramatic growth being associated with major floods of this century (Fig. 25), followed by periods of wave reworking. Rice graduate student Carmen Fraticelli later argued (2003) that episodes of high sediment flux occurred during El Niño (wet) years that followed prolonged periods of dryness (La Niña years).

During and immediately after the 1992 flood, field observations indicated that the regions closest to shore had experienced a net addition of sand, but nothing foretelling the emergence of a large offshore bar. Almost one year after the flood began, aerial photography showed the emergence of a bar approximately 1.9 km offshore (Fig. 25). One year prior to emergence, the bar zone was situated at 1 to 1.5 meters water depth. Gradually, the bar migrated onshore and alongshore, in an overall northwesterly direction, and eventually welded itself to the shoreline (Fig. 25).

Bar emergence created a second distinct depositional environment located landward of the bar classified as the "back-bar lagoon". Because wave approach is generally from the southeast, sediment-laden river waters were deflected into this lagoon (with only a minor outlet at the west-end of the bar), rapidly filling it with fine-grained sediments. In the summer of 1992, the back bar was sufficiently deep (1.5 to 2.5 meters) to allow coring operations from boats. By February 1993, the back bar aggraded to .5 to 1 m water depth. The sediments in this environment are silty clays with washover and eolian sands nearest the shorelines (Fig. 26).

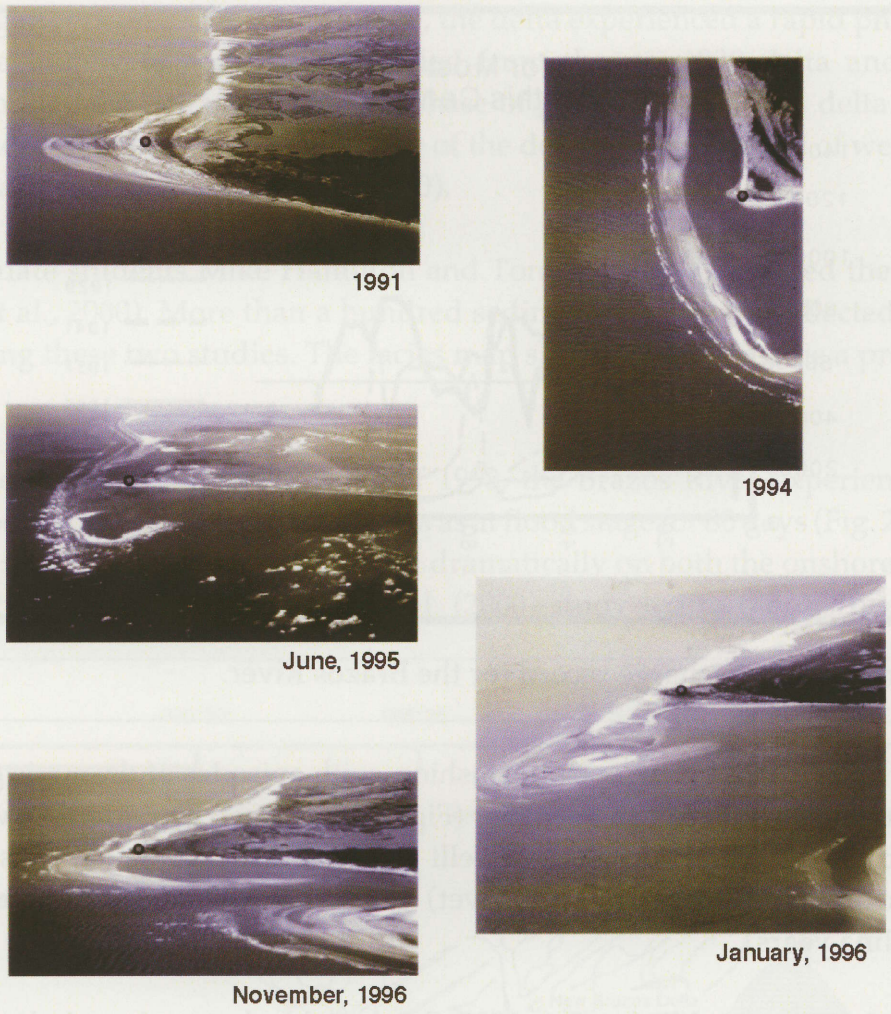


Figure 25. Aerial photographs showing changes in morphology of the Brazos Delta prior to and after the 1992 flood.

Flood-associated deposition also occurred on the main shores of the delta, both east and west. Significant accumulation occurred on a headland of the west flank (Fig. 25). Spit accretion around this headland accelerated progradation of the shore locally. On the east flank of the delta, upper shoreface bars have recently become emergent and significant progradation of the eastern shore has occurred. Prior to the flood, the beach on the eastern shore was approximately 25 meters wide. By February 1993, the beach had widened to nearly 100 meters.

During the flood, a tremendous amount of debris was brought to the delta. The river was nearly non-navigable due to the abundance of floating trees. This debris was piled on the shorelines, defining boundaries of flood related-accumulation. Each major flood results in a new debris zone on the shore. Successive floods can be identified from dates on the bottles and other garbage so that ridges can be correlated to the flood history shown in Figure 26.

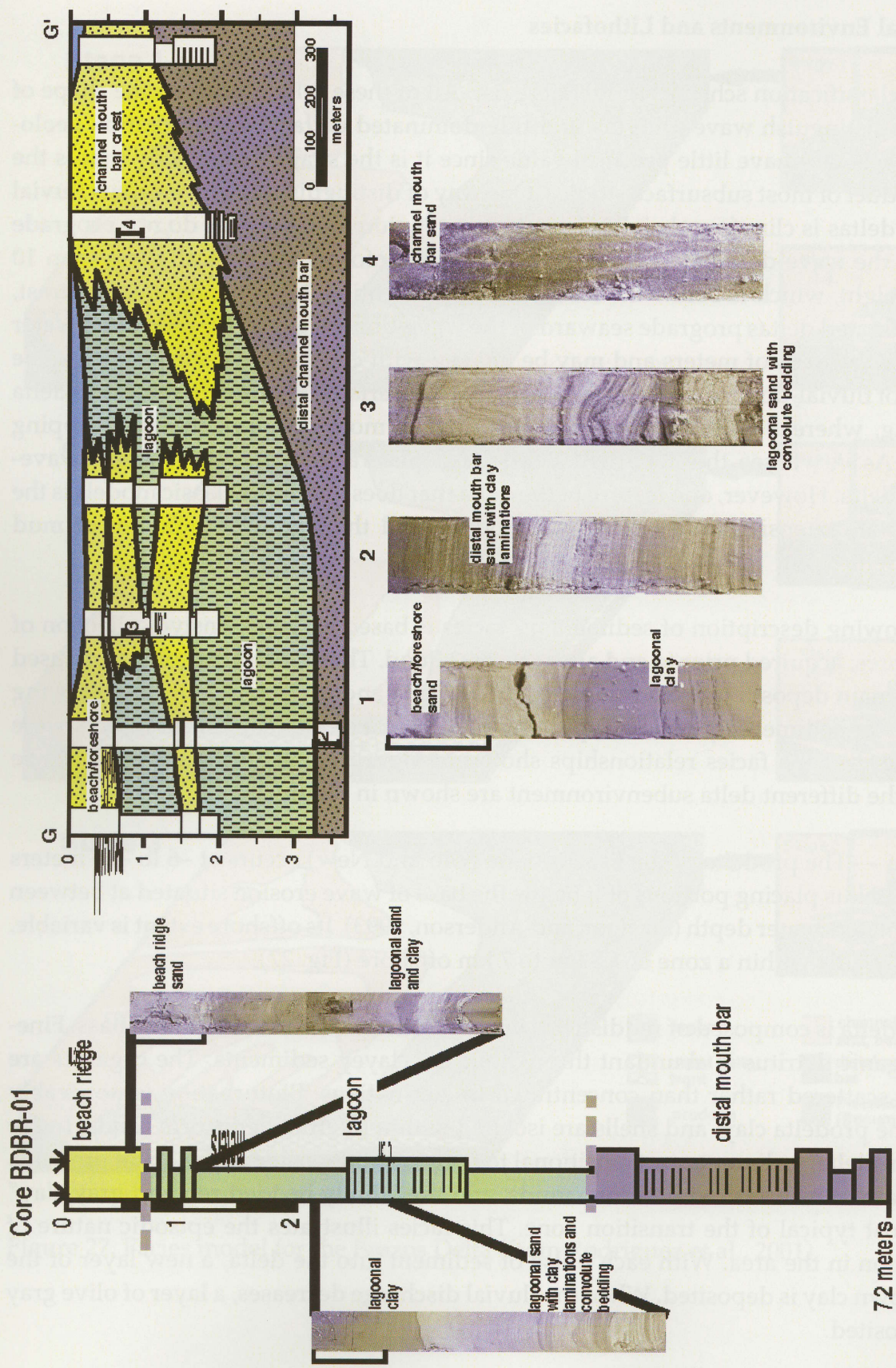


Figure 26. Nearshore core transect across the Brazos Delta

Depositional Environments and Lithofacies

Several classification schemes for details exist. All of these rely heavily on the shape of the delta to distinguish wave-, fluvial- and tide-dominated deltas. For exploration geologists, these schemes have little practical value since it is the shape of the body that is the desired product of most subsurface studies. One way of distinguishing wave versus fluvial dominated deltas is clinoform height. Since true wave-dominated deltas do not prograde seaward of the wave-dominated coastal zone, their clinoforms are typically less than 10 meters in height, which is below the resolution of conventional seismic data. In contrast, fluvial-dominated deltas prograde seaward of the wave-dominated coastal zone into water depths of several tens of meters and may be imaged with conventional seismic data. The clinoforms of fluvial-dominated deltas will also display variable dip directions, due to delta lobe shifting, whereas wave-dominated deltas display more uniform, seaward dipping clinoforms. As we will see, the modern Brazos delta displays all the characteristics of a wave-dominated delta. However, one feature of the delta that does not fit the classic models is the existence of an extensive mud-dominated prodelta and the significant amount of mud within the delta front.

The following description of sedimentary facies is based on an extensive collection of sediment cores, acquired prior to and after the 1992 flood. These sediment cores were used to map the main depositional environments of the delta and is the basis for the following descriptions of sedimentary facies. Dip-oriented and strike-oriented core transects were used to construct the facies relationships shown in Figure 27. Representative grain-size curves for the different delta subenvironment are shown in Figure 28.

Prodelta — The prodelta of the Brazos delta (Old and New) occurs at –6 to –15 meters water depth, thus placing portions of it below the base of wave erosion situated at between –8 and –10 meters water depth (Siringan and Anderson, 1993). Its offshore extent is variable, but generally falls within a zone of 1.2 km to 7 km offshore (Fig. 27).

The prodelta is composed of reddish brown clay with less than 10% sand by mass. Fine-grained organic detritus is abundant throughout the clayey sediments. The organics are most often scattered rather than concentrated in laminations. Bioturbation is noticeably absent in the prodelta clay, and shells are isolated within interbedded storm sands. In the proximal prodelta, sediments are transitional to delta front deposits. In the distal prodelta, sediments are transitional with marine muds, and the "thinly bedded red and gray clay" facies is most typical of the transition zone. This facies illustrates the episodic nature of sedimentation in the area. With each pulse of sediment into the delta, a new layer of the reddish brown clay is deposited. When the fluvial discharge decreases, a layer of olive gray clay is deposited.

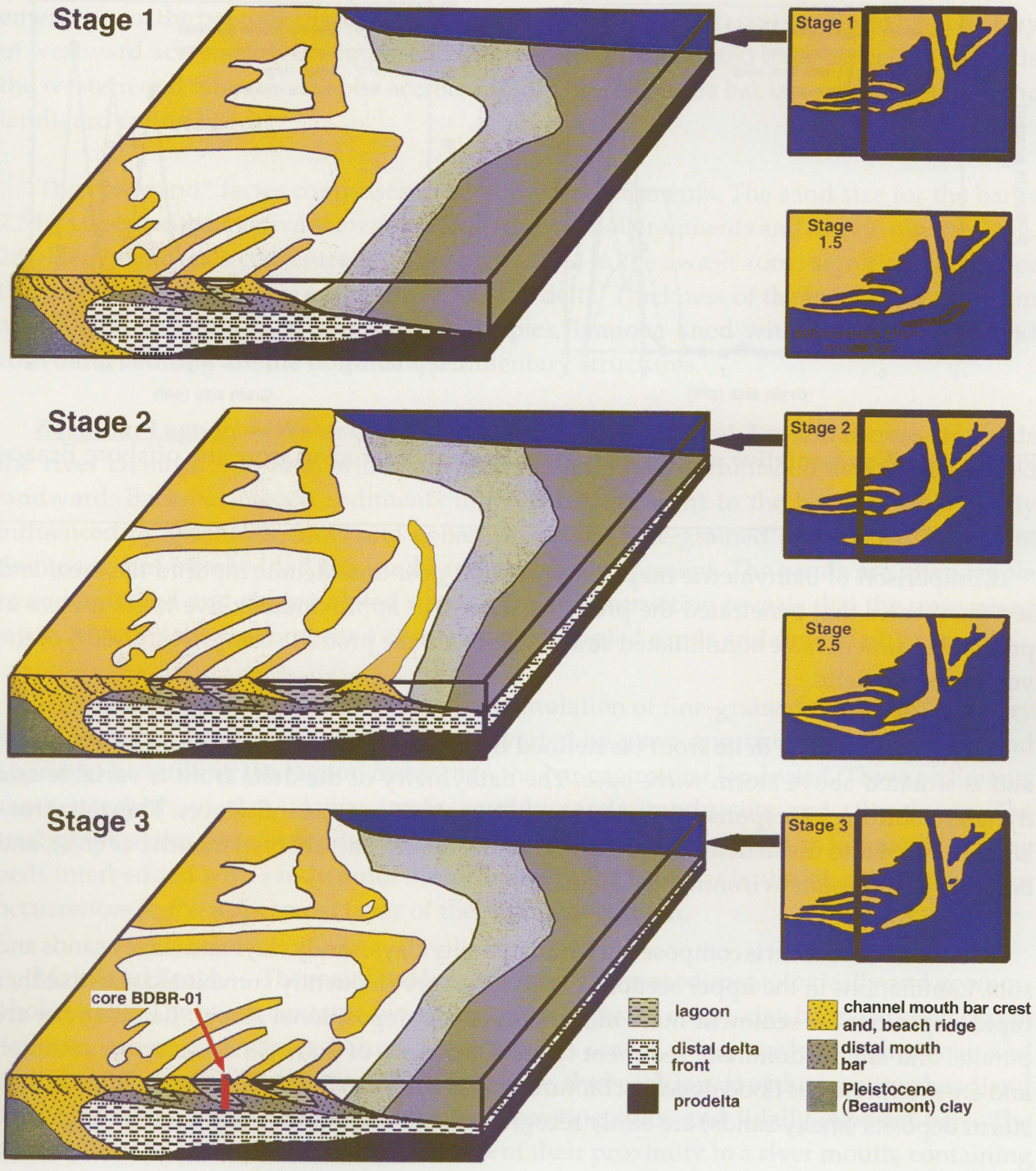


Figure 27. Facies model for the Brazos Delta (from Rodriguez et al., 2001).

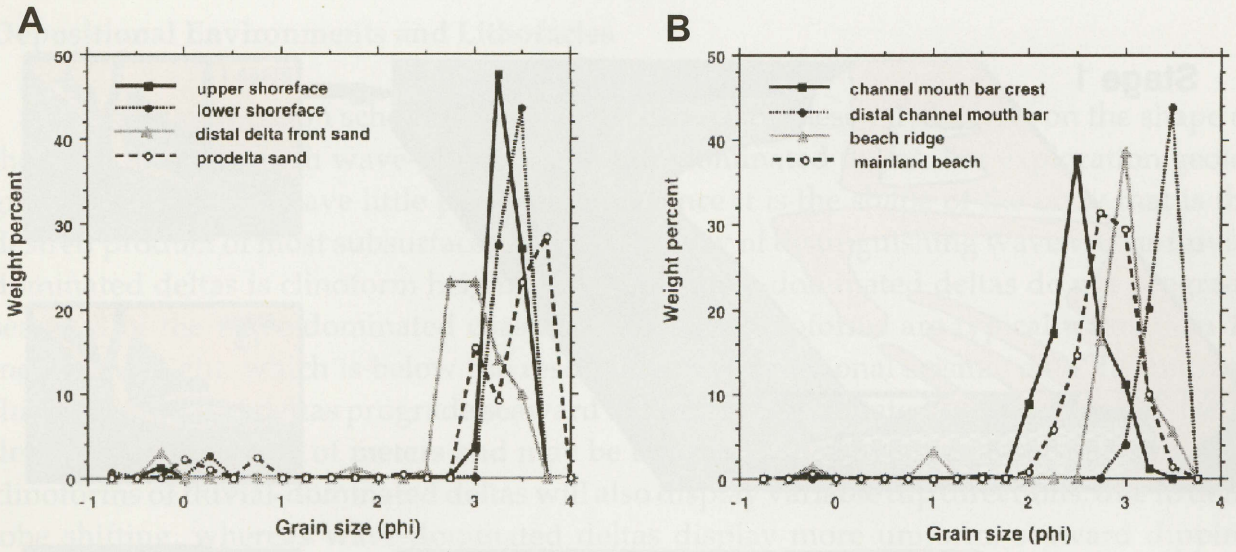


Figure 28. Representative grain size frequency curves for sand from the offshore Brazos Delta.

Comparison of bathymetric maps that pre- and post-date deltaic input to the area, and sediment cores that penetrated the prodelta, show that approximately five to six meters of prodelta sediment have accumulated since 1929. In all, the prodelta comprises 50–60% of the volume of the delta.

Delta Front — The delta front (as defined here) occurs just seaward of the river mouth and is situated above storm wave-base. The bathymetry of the delta front is variable and dynamic, shifting in response to variations in fluvial and marine influences. The delta front is subdivided into distal delta front, distal channel mouth bar, channel mouth bar crest, and back bar lagoon subenvironments.

The distal delta front is composed of laminated silty clays, sandy clays and clayey sands and silts. Laminations in the upper sections of this facies are frequently convoluted due to either rapid deposition or sediment mass movement (steep depositional slope). Laminations are parallel and are the dominant sediment types. Thick beds of homogenous clay are common and are interpreted as flood deposits. Bioturbation is absent in this environment. Interbedded storm deposits (shelly sands) are easily recognized, and organics are scattered throughout.

Distal Channel Mouth Bar — The distal channel mouth bar facies exist seaward of the river mouth and delta headland out to approximately –4 m water depth. Sediments of the distal channel mouth bar are less muddy than distal delta front sediments and are laminated sands and interbedded sands and muds. The sediments are (by mass) 85% sand. This facies underlies the channel mouth bar crest and the back-bar lagoon.

Channel Mouth Bar Crest — The emergent bar is a highly dynamic feature. The emergence of the bar took place .9 to 1.2 km from the mainland. It is arcuate and has a history of westward accretion. It is composed of numerous amalgamated bars, especially towards the western end where new spits accrete rapidly. Relief on the bar is nearly 1 meter on the landward facing beach.

The "bar sand" facies comprises most of the bar sediments. The sand size for the bar is 2.5 to 3.0 phi, with negative skewness values due to shell fragments and heavy minerals (Fig. 28). Heavy minerals concentrations are high (15%) in the swash zone, significantly higher than in foreshore deposits east and west of the delta. Thickness of the bar ranges from 1 m at the west-end to 3 m at its east end. Ripples, laminae lined with organic debris, and convolute bedding are the dominant sedimentary structures.

Back-Bar Lagoon — Water depths in the lagoon are –0.5 to –1.5 m and increase towards the river channel. The total width of this zone gradually diminished as the bar moved landward. Back-bar lagoon sediments immediately adjacent to the bar may be strongly influenced by washover, tidal and eolian processes. Fine-grained muddy sediments are dominant, but interbedded bar sands and clay is also present. The sands are often ripple cross-laminated and show isolated ripples. Direct observation reveals that the uppermost layers of sediment are composed of interbeds of rippled sands and sandy/silty clays.

The central lagoon is the site of rapid accumulation of fine-grained, very wet, and very muddy sediments. Fluvial sediments are diverted by wave approach into the lagoon and appear to be infilling the lagoon faster than the bar can retreat landward. These sediments show interbeds of clay, clayey sands, muddy sands, sandy silts and silty sands. The stratigraphy preserves individual flood events as distinct beds, such as homogenous clay beds interbedded with sandy mud. Bioturbation in the back bar is absent, except for minor occurrences in the uppermost layer of the sediment column.

Mainland Beach — The mainland beach is very diverse sedimentologically and geomorphologically. The beach contains swash lags, ripples, beach cusps, clay balls, exposed marsh deposits, storm and winter berms, berm erosion scarps, logs, washover channels, and scattered debris of human origin. On a larger scale, the beach is part of the accreting headland on the west flank, and includes a series of accreting spits, and tidally exposed bars. The sediments of this environment are reflective of their proximity to a river mouth, containing concentrated micas and heavy minerals in the swash zone. Organic debris is also present in the mainland beach deposits. Mineral segregation is very distinct in laminae exposed in the wave-cut scarps. Overall, sands are similar to channel mouth bar crest sands, but slightly less mature texturally. The beach is currently prograding seaward and facilitating the enclosure

of the back-bar lagoon. Beach progradation results in coarsening upward units of sand with interbedded back-bar mud.

Beach Ridges — Beach ridges are oriented parallel and oblique to the coast within a zone from the modern beach to the pre-1929 coastline (Fig. 24). Individually, numerous tidal channels within the marsh cut the beach ridges. The beach ridges are vegetated, and their crests are above high tides. Eolian processes influence the sediments, and local morphology shows dune development on some crests.

Beach ridge sediments are very sandy and more mature (textural and mineralogical maturity) than modern beach materials. They have a dominant grain-size mode of 2.75 – 3.0 phi (Fig. 28). The intertidal sediments on the margins of the ridge are sandy, but have approximately 10% mud by mass.

Inter-ridge Trough — The inter-ridge troughs are the lagoon environments on the flanks of the delta. Water depth in the troughs is less than .5 meters and is primarily controlled by tidal range. Tidal flow has created a web of channels between troughs.

Sediments of the troughs are variable from the margins to the axes but consist mainly of clay with abundant laminae and very thin beds of sand. Nearest the ridges, sands are flushed into the trough by washovers, abnormally high tides and eolian transport. These sands are interbedded with very fine-grained suspension deposits. In the centers of the troughs, the sediments are extremely fine-grained, with plentiful organic material, and are heavily bioturbated. Also, oysters have colonized the troughs and act as a trap for tidally-swept sediments. In areas where the inter-ridge troughs are tidally exposed, crusty algal mats, bird and crustacean fecal matter, and thin evaporitic laminae are observed.

Facies Architecture

The first flood affected the new Brazos delta 12 years after river diversion. Prior to this, significant delta progradation occurred largely in response to erosion of pre-1929 delta sediments and deposition of these sediments in the New Brazos Delta. It was not until after the 1941 flood that the first lagoon was preserved in the western delta headland. After the old delta had been eroded, floods became the primary mechanism enabling delta progradation.

Figure 27 illustrates delta evolution through one complete flood cycle. Prior to flooding (stage 1) no emergent offshore bar exists. The shoreline is lobate, with a series of beach ridges separated by inter-ridge troughs on the western delta headland. After the flood, the delta-front environment expands and the channel mouth bar emerges, enabling formation of the back-bar lagoon (stage 2). The channel mouth bar rests above distal mouth-bar facies. The

emergent channel mouth bar is subsequently reworked by waves and migrates landward, welding its western end to the shoreline. During stage 3, the pre-flood shoreline and back-bar lagoon become preserved as a ridge-trough pair within the western delta headland, and the bar becomes the new shoreline.

Core BDBR-01 (Fig. 26) was collected from a beach ridge that was preserved in the delta headland following the 1941 flood. This core tests the model presented in Figure 27 by illustrating the stacking pattern of nearshore facies after a complete flood cycle. The location of this core with respect to the model is indicated in Figure 27. The core sampled beach-ridge sands resting on lagoonal facies. This was the result of an offshore bar being reworked landward. The distal mouth bar sediments at the base of the core were deposited prior to bar emergence and represent the pre-flood shoreline. Bernard et al. (1970) examined a 10-m-long core collected from the beach on the western side of the Brazos delta (core R3660). This core, collected in 1954, shows the same facies succession as core BDBR-01, but the core collected by Bernard and his colleagues was long enough to penetrate the entire delta succession. Although Bernard et al. (1970) used different facies terminology, this core sampled approximately 1 m of distal delta-front sediments and 2.6 m of prodelta sediments below the distal channel-mouth-bar facies, and bottomed out in the Pleistocene Beaumont Formation (Bernard et al. 1970).

The facies relationships described above are very different from the classic wave-dominated delta model where well sorted sands are stacked in a succession that may appear quite similar to a prograding clastic shoreline succession (Coleman and Wright, 1975; Dominguez, 1996). This has implications for the interplay of waves and fluvial sediment discharge. The lagoons are notable depocenters of previously unrecognized fine-grained sediments.

One of the most striking characteristics of the delta is its large prodelta. The abundance of prodelta muds is dependent on proximity to the river mouth. Proximity, however, is not the only constraint on prodelta deposition. Longshore currents dramatically affect the depositional patterns, and the resulting architecture is strongly asymmetric to the west. During our study, significant reworking of the 1992 flood deposit was observed, resulting in a more strike-oriented delta geometry.

The prominence of prodelta and back-bar lagoonal silts and clays within the overall Brazos Delta succession is a fundamental difference between the Brazos facies architecture and other wave-dominated deltas (e.g., Sao Francisco delta. Coleman and Wright, 1982; Dominguez, 1996). Most studies attribute more significance to nearshore facies. The Brazos Delta analog serves to emphasize that subtle fluvial influence can result in deltas that, while having the overall shape and appearance of wave-dominated deltas, can exhibit facies architectures that are quite heterogeneous.

Stop 6
San Luis Tidal Inlet/Delta

Because barrier islands are so extensive today, we tend to overestimate their contribution to the coastal stratigraphic record, certainly with regard to the transgressive systems track. In reality, landward and along strike migration of barriers and tidal inlets results in very extensive flood tidal deltas which, if deposited within incised fluvial valleys, have a much higher preservation potential than coastal barriers. This has certainly been the case on the east Texas shelf. During this field trip we will examine two flood tidal deltas, the San Luis and Bolivar flood tidal deltas.

San Luis Pass is a deep tidal inlet that separates Galveston Island from Follets Island (Fig. 31). It is incised into Pleistocene sediments to about –10 meters (Bernard et al., 1970; Williams et al., 1979; Israel et al., 1987). As we cross the bridge between Follets Island and Galveston Island, look to the right and you will see waves breaking over the San Luis ebb tidal delta. To your left you can see shallow tidal flats that comprise part of the flood tidal delta.

A detailed sedimentological analysis of the flood tidal delta was conducted by Israel et al. (1987). Their study showed that the tidal delta complex consists of a number of sub-environments including oyster reefs, washovers, distal tidal delta, proximal tidal delta, inlet, and marsh subenvironments. The tidal inlet consists mainly of graded shell beds and shelly sands. The aerial photograph of the delta (Fig. 29) shows the sandy lobes and tidal channels that comprise the proximal flood tidal delta. Note that the flood tidal delta is significantly larger than the ebb tidal delta, which is under constant attack of large waves.

Sediment cores from the proximal portion of the delta sampled well-sorted sand and shelly sand and cores from the distal part of the delta sampled mostly mud with diagnostic thinly bedded sand/mud beds (tidal couplets). Sedimentary structures are not observed in cores from the proximal portion of the delta, but laterally shifting of tidal channels and sandy lobes occurs frequently and undoubtedly produces large-scale trough cross bedding and migration of large bedforms to create other large-scale structures. Both proximal and distal flood tidal delta facies interfinger with bay muds and organic-rich marsh deposits.

The ebb tidal delta forms a prominent bathymetric bulge that extends offshore about 300 meters. No detailed sedimentological work has been conducted on the San Luis Pass ebb tidal delta.

Growth and development of the San Luis tidal inlet/delta complex was caused by westward accretion of Galveston Island; the flood tidal delta is thought to extend along the

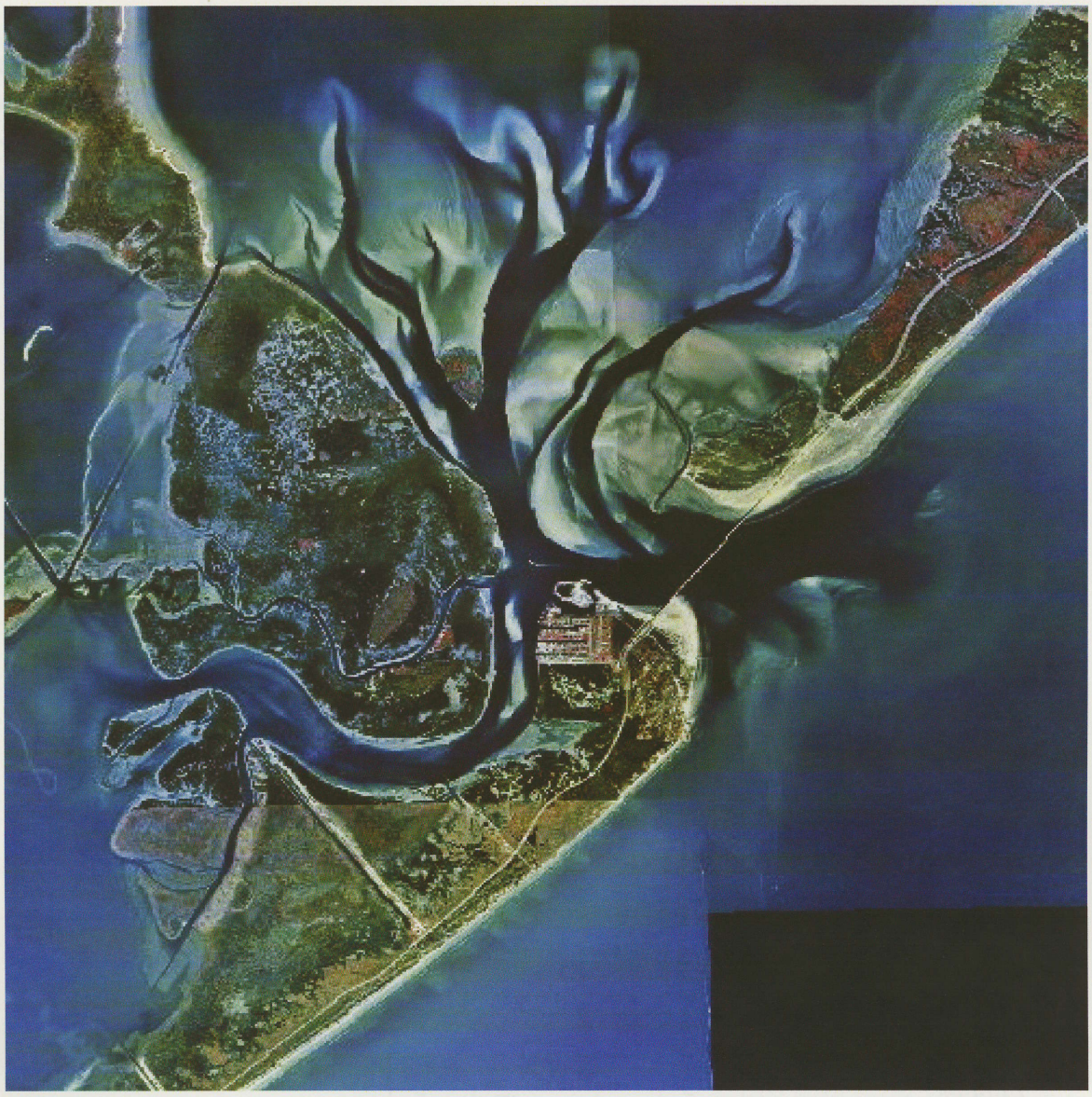


Figure 29. The San Luis Pass tidal delta.

back side of the island for several kilometers. Within the past several centuries, westward migration of the inlet slowed when it reached an ancestral channel of the Brazos River, the Chocolate Bay channel. The inlet has now incised into the channel, lateral accretion of the flood tidal delta has been slowed and the flood tidal delta has begun to prograde in a landward direction.

Core SLP-91-1 (Fig. 30), taken at the backside of the western tip of Galveston Island, penetrated an overall coarsening upwards trend that reflects the westward migration of the tidal inlet, flood tidal delta and barrier island and infilling of the Chocolate Bay fluvial channel. A highly cohesive, organic-rich, olive-black clay interpreted as a freshwater

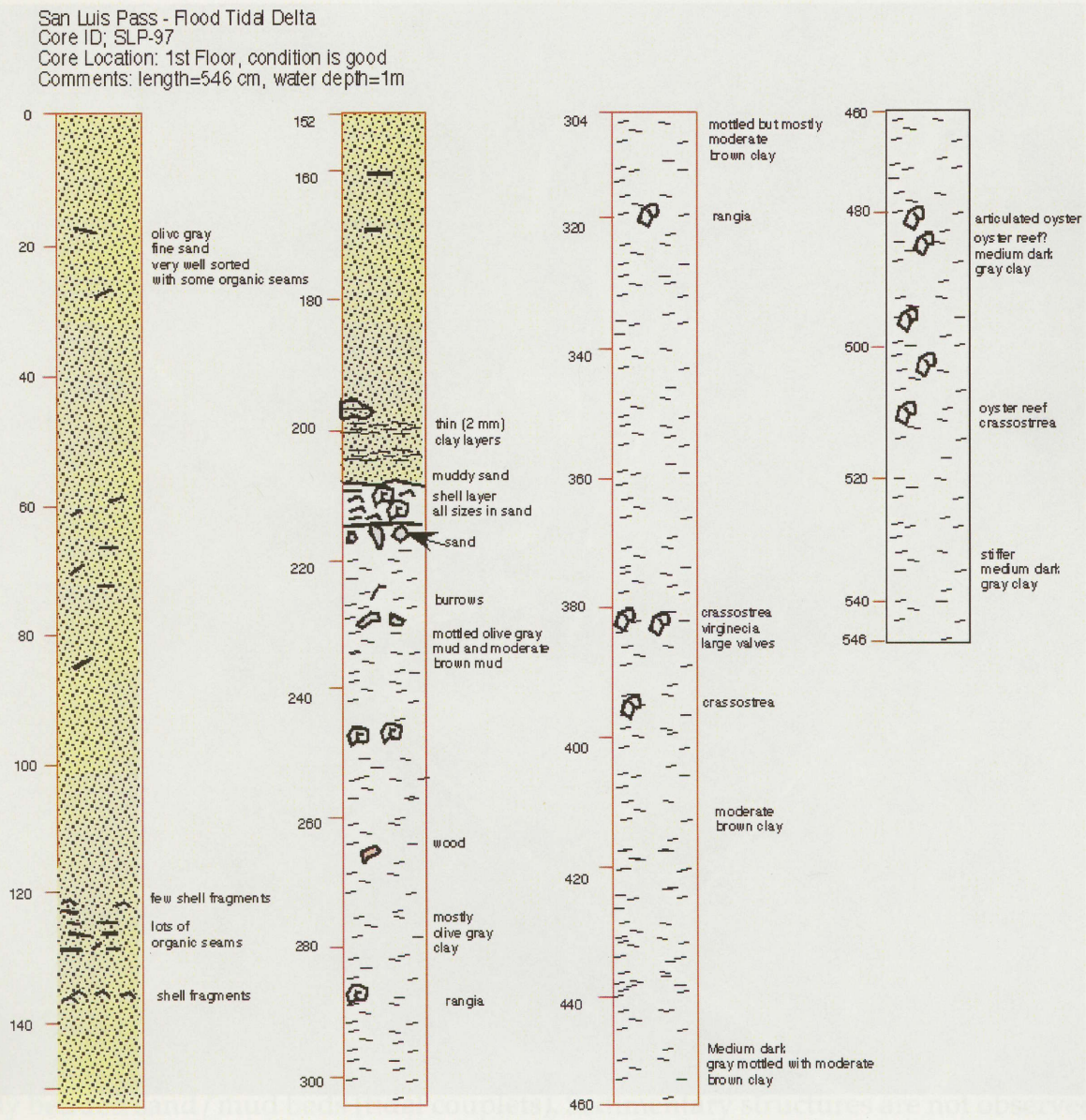
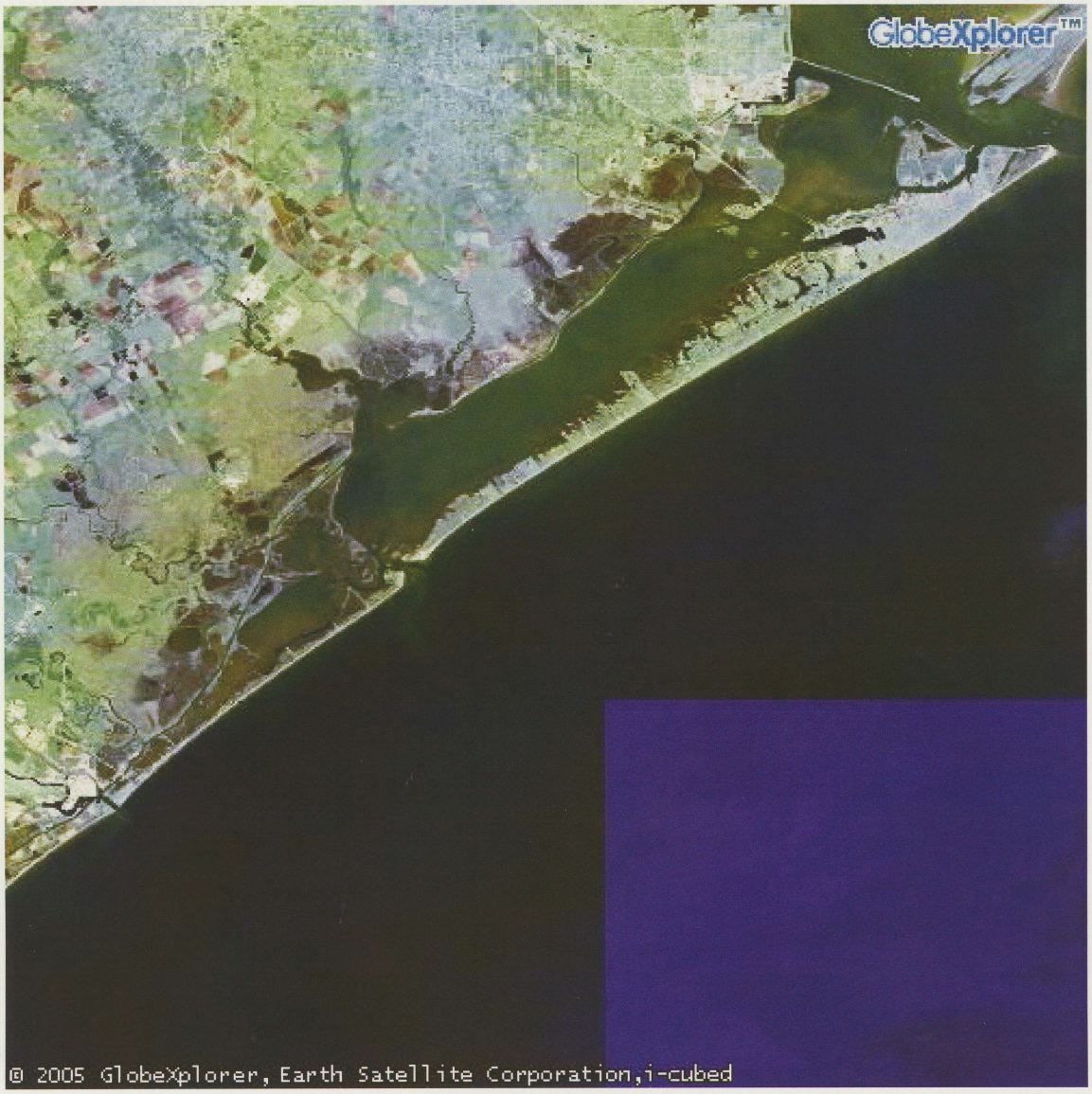


Figure 30. Description of Core SLP-1 from the west end of Galveston Island.

marsh deposit underlies the sequence. A radiocarbon age date of 7,500 ± 75 cal BP was acquired from this organic-rich sediment and constrains the initial flooding of the area (bayline surface). This is overlain by dark greenish gray bioturbated clay unit, interpreted as bay/estuarine sediments, grading up to brown to reddish brown clay unit, interpreted as a prodelta deposit of the Brazos River. The upper clay unit is moderately bioturbated and contains inarticulated *Crassostrea* shells. The Brazos River clay unit is overlain by stacked fine sands and clay laminations with shell lag bases, interpreted as flood-tidal deposits.

A radiocarbon age date of 4,150 ± 191 cal BP was obtained from an articulated *Cyrtopleura costata* (Angel Wing) shell sampled at the base of the flood-tidal delta deposits, in the Brazos river clay unit. This bivalve is indicative of tidal inlet areas, and may burrow up to 0.5 m into clay substrate. This unit is capped by a coarsening upward sequence of bay / lagoon clay with sand and shelly sand laminations and interbeds with abundant plant debris, interpreted as washover deposits. These deposits grade up into stacked graded shell lag to fine sand beds representing the shallowing upwards and emergence of the flood-tidal delta.

Stop 7
Galveston Island-Back Barrier



Barrier islands and peninsulas are common features of modern low-gradient coasts. They are considered transgressive features. Galveston Island is one of the best studied barrier islands in the world and its evolution is well established. The island began to form about 5,500 years ago and, for most of its history it grew and advanced seaward. The growth of the island left an imprint of the landscape that is seen in aerial photographs, similar to coastal landscapes along many of the shorelines of the world (Fig. 31). This landscape consists of linear ridges and swells between these ridges. The ridges are the highest natural features on the island and the swells provide most of the islands limited freshwater habitats.

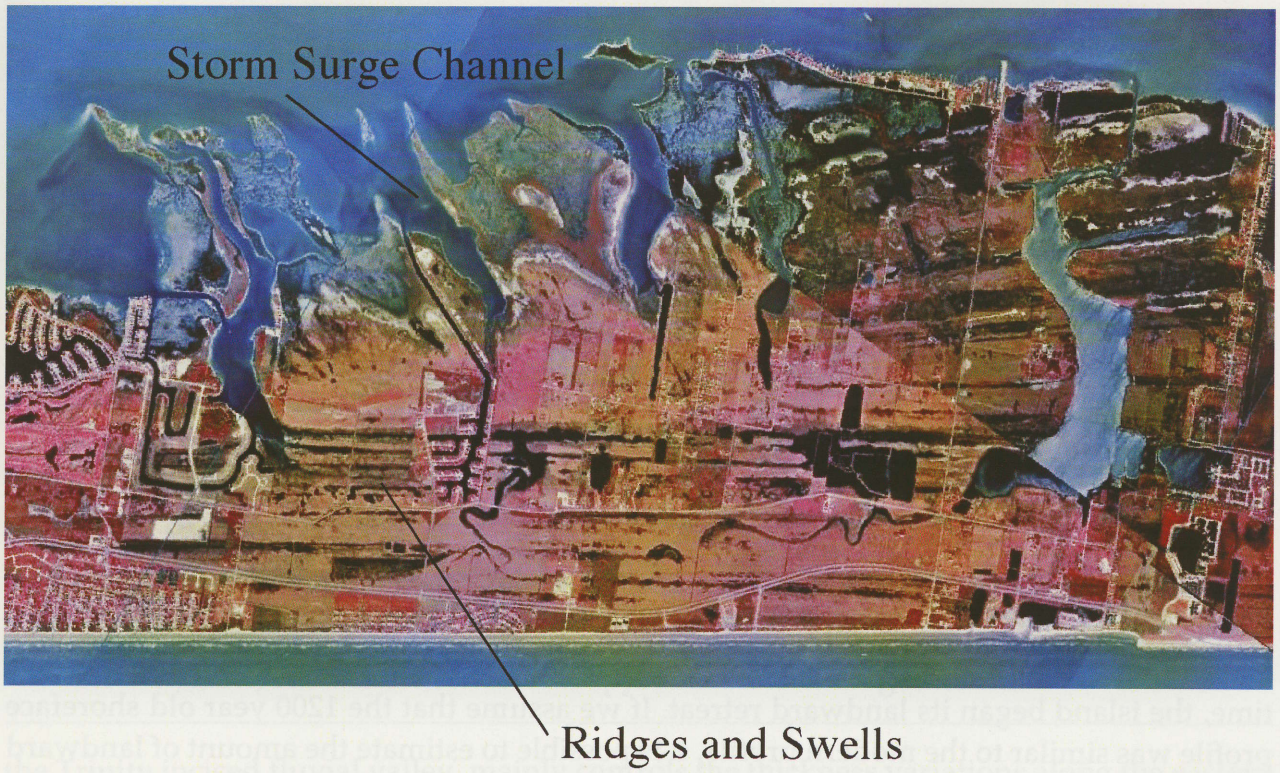


Figure 31. Ridge and swell topography of Galveston Island.

Borings through Galveston Island and radiometric age dating were used to construct one of the first, and most widely cited, models for barrier island formation (Bernard et al., 1959, 1970). This and more recent work (Rodgriguez et al., 2004) showed that the island had prograded seaward during the past 5,000 years, when the rate of sea-level rise slowed. Rodriguez et al.'s (2004) cross-section through the island is shown in Figure 32. This section shows the Pleistocene–Holocene boundary forming a ramp that shallows to a depth of –3 m on the landward side of the island. This ramp served as a nucleus for growth of the island. Bernard and his colleagues interpreted the island as having formed from a small bar that emerged and grew seaward by beach accretion and southwestward by spit accretion (Bernard et al., 1959, 1970). The geometry of the beach ridges at the backside of the northeastern half of the island defines a set of ridges that fan-out towards the bay. This indicates that, initially, the island grew northeastward towards the bay, similar to a spit bar, using the shallow portions of the Pleistocene–Holocene surface as an "anchor". The ridges later rotated out towards the Gulf. Storm surge channels dissect older beach ridges and mark episodes when the younger, narrower barrier was breached.

The radiocarbon dates obtained by Rodriguez et al. (2004) were acquired largely from monospecific samples, mostly *Donax* shells. This species is known to live only in the

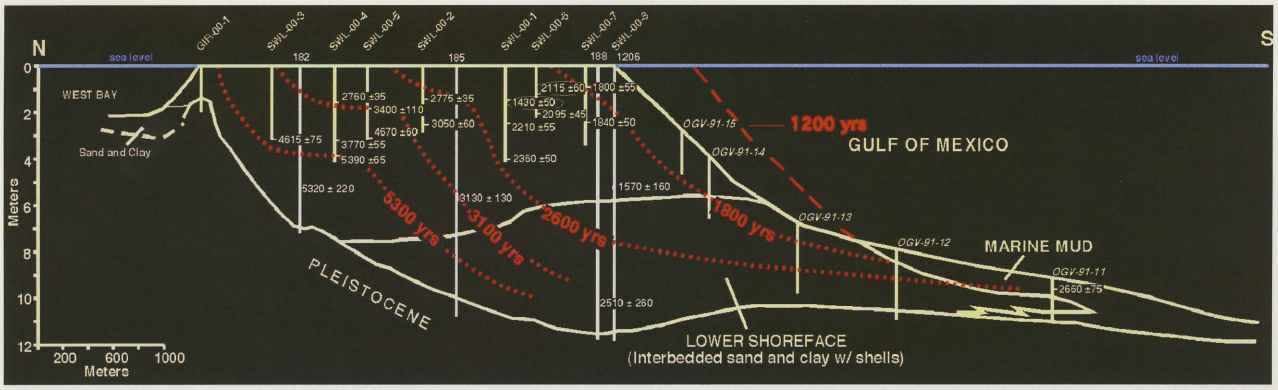


Figure 32. Geological cross section through Galveston Island based on the combined results of Bernard et al. (1970) and Rodriguez et al. (2004).

intertidal swash zone. In general, the newer data support the Bernard et al. (1970) results, indicating that the progradation of the island took place between approximately 5,300 years ago and 1200 years ago, based on the age of the most seaward shoreface deposits. After this time, the island began its landward retreat. If we assume that the 1200 year old shoreface profile was similar to the modern profile, it is possible to estimate the amount of landward retreat of the shoreline during the past several centuries (Fig. 32). This yields a rate of shoreline and shoreface migration of 1m/yr, which is the current rate of shoreline migration in the area.

What is the origin of the sand that now comprises Galveston Island? Detailed mineralogical analyses by Cole and Anderson (1982) showed that the island is made up of sand derived from the Mississippi, Trinity/Sabine, and Brazos rivers, in almost equal proportions (Fig. 33).

Currently, the only sand that exist offshore of the shoreface occurs within the Trinity/Sabine incised fluvial valley. Thus, transgressive erosion has been highly efficient at transporting sand onshore to nourish barriers. These sands were stacked in shingle-like fashion against a Pleistocene surface whose profile mimics the modern shoreface profile (Fig. 32). Ridge and swell topography marks the episode of barrier growth (Fig. 31). A higher sediment supply during the early phase of relative stillstand, provided by reworking of previous coastal lithosomes, hastened spit progradation and seaward accretion of barriers. When the offshore sand supply was exhausted, the barrier began to erode. The modern beach is an erosional surface cut into beach ridges that range in age from several centuries old to nearly 2,000 years old.

Figure 34 shows an isopach map of the sand bodies that comprise the Galveston and Bolivar barriers. The relief on the underlying Pleistocene surface, which is mainly related to

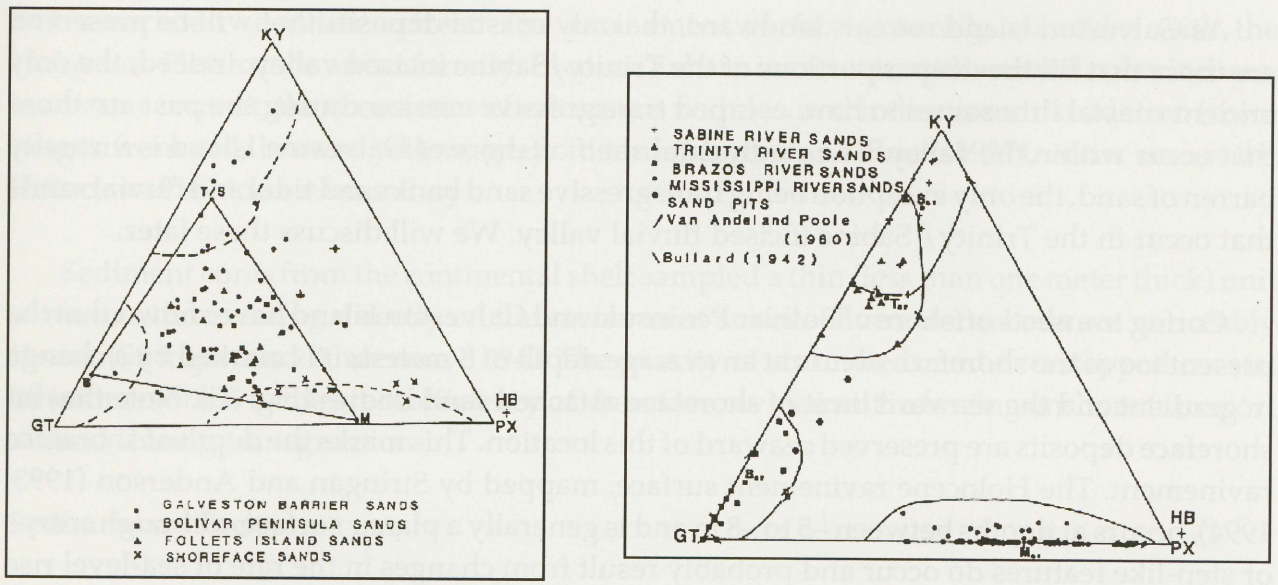


Figure 33. Mineralogical composition of Galveston Island (from Cole and Anderson, 1982). KY = kyanite, HB = hornblende, PX = pyroxene and GT = garnet.

the Trinity incised fluvial valley, mainly controls the thickness variations along the eastern side of the island. Recall that fluvial terraces formed during the falling stage of sea level characterize the valley. For the most part, the sands that fill the flanks of the valley and interfluvial areas are less than 8 meters thick. Thus, most of the barrier does not survive transgressive erosion.

Galveston Island is a little over 3 kilometers wide at the location where the profile shown in Figure 32 was constructed. At the current rate of landward retreat of the shoreline, the barrier will be gone in about 1,500 years. Actually, the island may be removed earlier if the rate of sea level rise increases.

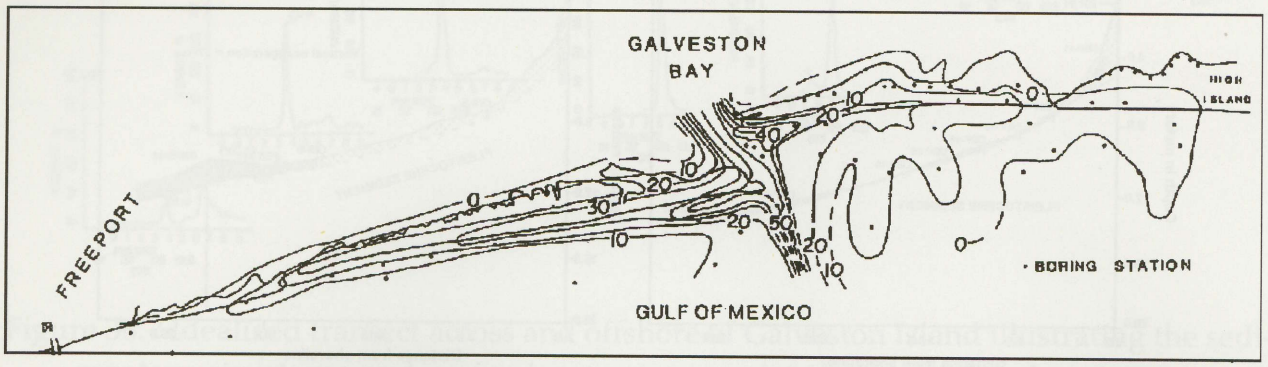


Figure 34. Contour map of depth to Pleistocene surface based on U.S. Army Corps of Engineers borings (Williams et al., 1978) and information taken from Bernard et al. (1959).

As Galveston Island retreats landward, the only coastal deposits that will be preserved are those that fill the deeper portions of the Trinity/Sabine incised valley. Indeed, the only ancient coastal lithosomes to have escaped transgressive erosion during the past are those that occur within the valley. The continental shelf offshore of Galveston Island is virtually barren of sand, the only exception being transgressive sand banks and tidal and fluvial sands that occur in the Trinity/ Sabine incised fluvial valley. We will discuss these later.

Coring transects offshore of Bolivar Peninsula and Galveston Island have shown that the present toe of the shoreface occurs at an average depth of 8 meters; it is marked by a change in gradient and the seaward limit of shoreface-attached sand bodies (Fig. 35). Note that no shoreface deposits are preserved seaward of this location. This marks the depth of shoreface ravinement. The Holocene ravinement surface, mapped by Siringan and Anderson (1993; 1994), occurs at depths between –5 to –8 m and is generally a planar surface, although scarps or step-like features do occur and probably result from changes in the rate of sea-level rise

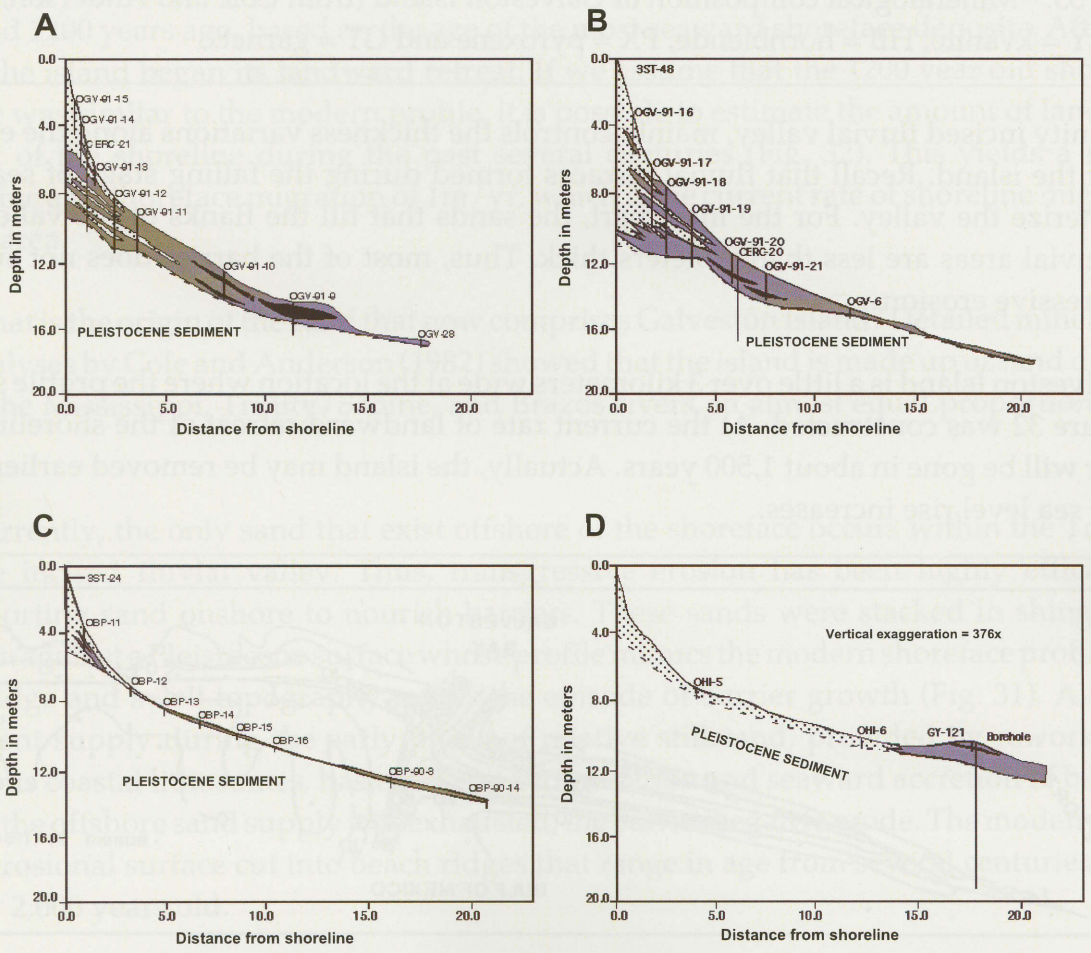


Figure 35. Representative shoreface transects for offshore Galveston Island (A and B) and Bolivar Peninsula (C and D).

and associated shoreline retreat. The ravinement surface generally coincides with the Holocene–Pleistocene unconformity. In cores, a sharp increase in stiffness and cohesiveness, low water content, the occurrence of calcareous nodules, and indications of oxidation (color ranges from red, brown, yellow, gray, to tan) (Morton and Price, 1987), occur across the Holocene–Pleistocene boundary.

Sediment cores from the continental shelf sampled a thin (less than one meter thick) unit of marine mud resting directly on the ravinement surface. Even storm beds are noticeably absent (Siringan and Anderson, 1994). The paucity of storm beds and overstepped coastal lithosomes attest to the great efficiency of shoreface ravinement in forcing onshore transport of sand during transgression.

Sedimentary Facies

Siringan and Anderson (1994) acquired a number of sediment core transects across the shoreface and onto the inner shelf along the east Texas coast. Figure 36 shows the idealized

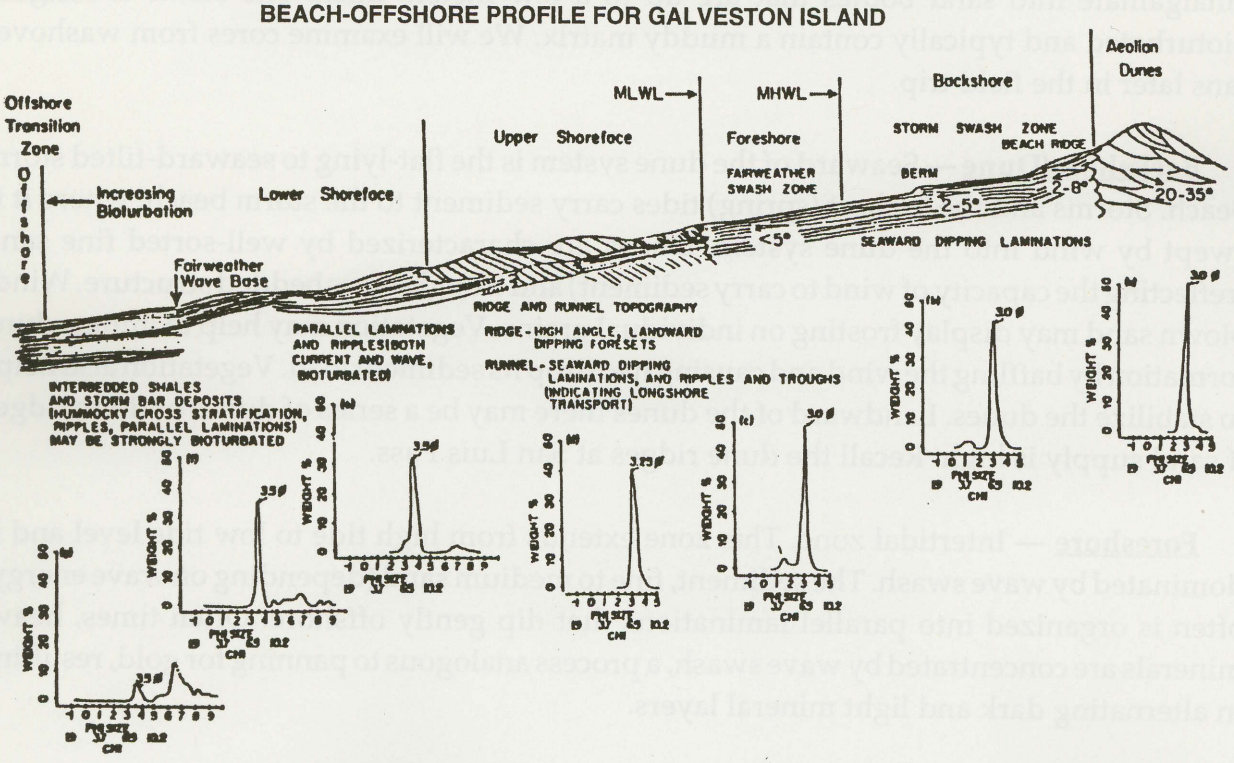


Figure 36. Idealized transect across and offshore of Galveston Island illustrating the sedimentary structures and grain size curves associated with each subenvironment. A prograding shoreline succession will consist of these units stacked on top of one another with marine sediments at the base and eolian dune sediments at the top of the sequence.

sequence of sedimentary structures and grain sizes along one of these transects. This sequence can be used to develop an idealized stratigraphic succession that would result from shoreface-barrier progradation during a highstand. We don't see this succession today offshore of Galveston Island, but progradational coastal deposits do occur off the central Texas coast where convergence of longshore currents deliver larger quantities of sand to the coast (Rodriguez et al., 2001). Later this morning we will visit another prograding shoreline at the west end of Bolivar Peninsula. There, shoreline progradation is occurring because the north jetty traps sand that is transported to the west by longshore currents. The following is a brief description of the coastal environments and facies.

Back-Barrier Environments — Landward of the barrier islands are lagoon systems. These are relatively quiet, protected waters. During storms, wave energy may carry sediment over the barrier, and dump it into the lagoon as washover deposits. In this manner, barrier islands and their associated lagoon systems move landward during periods of rising sea level — providing there is sufficient sediment available to the system and that it is in balance with the rise of sea level. This is one of the main mechanisms for landward migration of barrier islands. Where these washovers fill bays on the backside of the islands, they amalgamate into sand bodies that are up to a few meters thick. The sands are highly bioturbated and typically contain a muddy matrix. We will examine cores from washover fans later in the field trip.

Backshore/Dune — Seaward of the dune system is the flat-lying to seaward-tilted storm beach. Storms and the highest (spring) tides carry sediment to the storm beach, where it is swept by wind into the dune system. Dunes are characterized by well-sorted fine sand (reflecting the capacity of wind to carry sediment) and trough-cross bedded structure. Windblown sand may display frosting on individual grains. Vegetation may help to initiate dune formation by baffling the wind and causing it to drop its sediment load. Vegetation also helps to stabilize the dunes. Landward of the dunes there may be a series of dunes or dune ridges if sand supply is high. Recall the dune ridges at San Luis Pass.

Foreshore — Intertidal zone. This zone extends from high tide to low tide level and is dominated by wave swash. The sediment, fine to medium sand (depending on wave energy) often is organized into parallel laminations that dip gently offshore. Often times, heavy minerals are concentrated by wave swash, a process analogous to panning for gold, resulting in alternating dark and light mineral layers.

Shoreface — The shoreface is constantly underwater and lies between the low tide mark to the deepest location where fair weather waves touch the bottom (about 10 meters). Because of the presence of wave activity, the sands often are fairly well sorted and rounded. Sediment is reworked into longshore bars that parallel the shoreline and migrate landward.

Migration of the bar results in landward dipping foresets. Within the runnel, troughs that are aligned roughly parallel to shore, exists along with wave and current ripples. The bars form a breaker zone that commonly can be seen offshore. Trace fossils are common. High-resolution seismic records acquired parallel to the shore and across the shoreface show that offshore-directed storm channels of varying sizes characterize the older part of the shoreface succession. These channels were active when the barrier was narrower, making it more vulnerable to breaching by storms. The thickest sands occur within these channels and there is considerable along-strike variability in sand body thickness due to the irregular nature of the Pleistocene surface (Siringan and Anderson, 1994).

The preservation potential of the idealized sequence outlined above is minimal during a transgression. This is because of shoreface ravinement, which along the upper Texas coast occurs to a water depth of about 8 to 10 meters. Anything above this depth is not preserved, and that includes part of the lower shoreface and everything above it (Fig. 37 on following page). This is why sediment cores collected a few kilometers seaward of the shoreline off Galveston Island and Bolivar Peninsula have sampled marine mud resting directly on Pleistocene deposits. The boundary between these units is the transgressive ravinement surface.

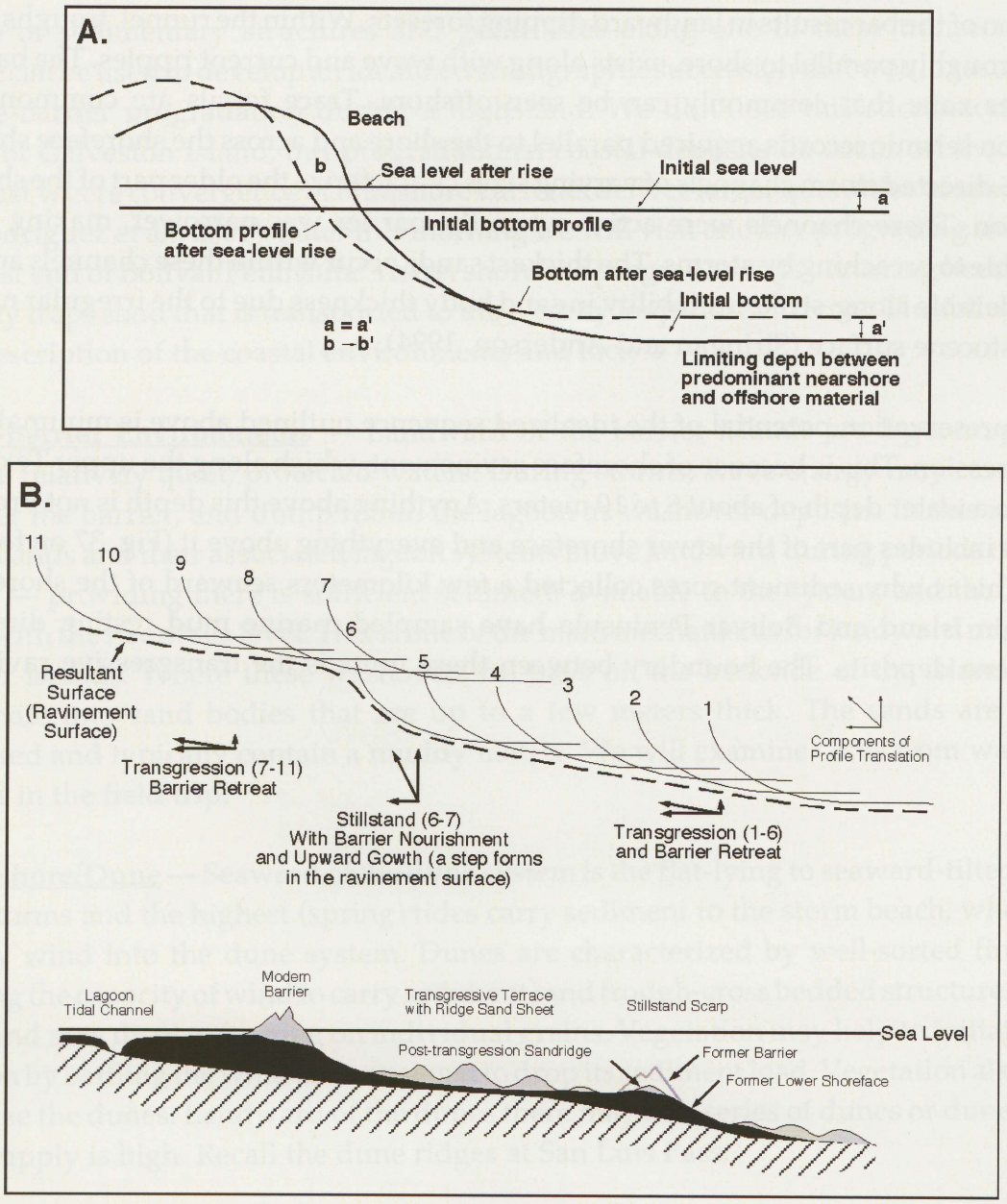


Figure 37. Examples of models that attempt to capture how the shoreface and beach migrate landward in response to relative sea level rise (from Brunn, 1975 and Swift, 1975). Note that in model A a constant shoreface profile is maintained during retreat. Sand that is eroded from the shoreface is either moved landward or remains in the longshore transport system. Model B illustrates the manner by which the shoreface moves landward and cuts a new profile. The shape of this profile is controlled by the rate of sea level rise and by sediment supply. During a fast rise, or a significant reduction in sediment supply, steps are created in the profile as the shoreface shifts rapidly landward. Changes in the offshore profile also occur where waves encounter different substrates that are either easier or harder to erode.

Coastal Erosion

All along the coast we will observe signs of beach erosion. The average rate of erosion for the upper Texas coast is 1 to 1.5 meters per year. As geologists, we know that this transgression has been going on for thousands of years. In fact, the rate of shoreline retreat is significantly



slower now than it was a few thousand years ago. Try to tell this to the people who own houses on the beach; they are convinced that erosion is a recent phenomenon and somebody is to



This photograph shows one of the few remaining seawalls on the upper Texas coast at Surfside. Note that the beach is virtually gone and the elevation of the beach has been lowered by about 1.5 meters since the wall was constructed.

blame. The sad part is that if predictions about increased sea-level rise in the next century prove accurate, the rate will increase.

When it comes to beach erosion, Texans have two options: 1) allow the beaches to erode, or 2) nourish the beaches with sand. Current state law requires that as the beach advances landward and overtakes private property, the property owner must remove their house at their own expense. This is called the "Open Beaches Act", meaning that all citizens must have free access to the beaches. Many property owners think that this law is unfair and that they have a right to protect their property, even if that necessitates constructing seawalls and other "hard" structures.

Recent efforts on the part of beach-front property owners to save their property has included the construction of "straw dunes" and Geotubes. The latter are large tubes of fiberglass fabric filled with sand. The straw dunes lasted only until the first small tropical storm and the Geotubes are being undercut by waves in several locations. Where they still stand, the beach is becoming narrower, which will eventually result in a lack of access to the beach, a violation of Texas Law.

Many property owners feel that beach nourishment is the answer to their problems. But, this is a losing battle - the sand is placed in a setting where wave reworking eventually will



The Geotubes placed along this stretch of the beach clearly have not worked.

remove it from the beach and deliver it to the deeper water environment or farther along shore. It is interesting to note that the continental shelf offshore of Galveston Island is virtually barren of large, concentrated sand deposits, with the exception of small concentrated sand banks deposited along the margins of the Trinity/Sabine incised fluvial valley during an earlier phase of sea-level rise. In short, there are no offshore sand resources available for beach nourishment and re-supplying the beaches is a short-term solution. The reality is, if Texans truly want to have free access to their beaches, we will need codes the prohibit construction within one hundred meters of the beach.

The wetlands behind the barrier systems are valuable and protected areas. Where sand is not available to the barrier system, the dunes are not developed, the beach is narrow, and the roads wash out. Rebuilding of the highways often is not an available option. The advancing sea in about a decade will overrun the road we are driving on, unless a major storm destroys it first. Note how much property is for sale along the highway.

Optional Stop
The Galveston Seawall and the 1900 Storm

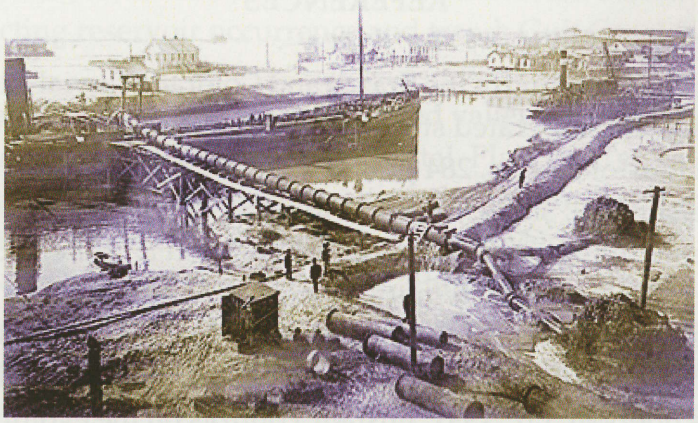
Around the turn of the century, Galveston was a thriving port city with a population of almost 38,000. Too a large degree, the prosperity of Galveston at this time can be attributed to the deep tidal channel that served as a ship channel for deep draft vessels. On September 8, 1900 the deadliest natural disaster in the history of the United States struck the island. A storm with winds over 190 kilometers per hour (120 miles per hour) and an overwhelming tidal surge decimated the island and killed over 6,000 people. One-third of the city was completely destroyed.



Galveston in the aftermath of the 1900 storm.

In response to the devastating event, city built a seawall 11 kilometers long and 5 meters high. The wall is 1 to 2 meters wide at the top and expands to 5 to 6 meters wide at the base. Accompanying the building of the seawall, the city raised the level of part of the island by 1 to 2 meters. When the Galveston seawall was complete in 1910, the beach was ~ 30 meters seaward of the present coastline. During the early 1900s the city continued to prosper. As we drive through the older part of the city you will see many of the beautiful Victorian houses that were constructed during this era.

Ironically, it was not the 1900 hurricane, but the construction of the Houston Ship Channel that led to the economic downfall of Galveston during the depression and post-World War II era. Galveston suffered while Houston grew. In recent years Galveston's economy has been re-energized by tourism. But, now the city faces another problem. Its beaches are eroding at a rate of more than a meter per year. An attempt to nourish the beaches met with only modest success and there is a serious shortage of sand needed for future nourishment projects. In the end, Galveston will probably loose its battle with nature. The retreat of the island can best be seen where the seawall ends; a dramatic break occurs between the seaward "anchored" seawall and the more landward placement of the "natural" beach.



The raising of Galveston Island.

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