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SEDIMENTARY FACIES AND EVOLUTION OF LATE PLEISTOCENE TO RECENT COASTAL LITHOSOMES ON THE EAST TEXAS SHELF

by

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A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE DOCTOR OF PHILOSOPHY

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ABSTRACT

Late Pleistocene to Recent Coastal Lithosomes on the East Texas Shelf

by

Antonio Beyra Rodriguez

Examination of Late Pleistocene to recent coastal lithosomes on the east Texas continental shelf can help refine current models aimed at understanding how coastal environments respond to a variety of forcing mechanisms including changes in subsidence rate, sediment supply (climate), and eustasy. High-resolution seismic data, sediment cores, paleontologic data, and radiocarbon ages were examined from bay complexes (fluvial, bay-head delta, middle bay, tidal delta, and barrier shoreline environments), offshore banks, incised valleys, and the Brazos Delta. Within the study area, preservation of these deposits on the shelf has been variable. Coastal lithosomes had a high preservation potential in the eastern portion of the study area (around the Trinity/Sabine incised valley) and a low preservation potential in the western portion of the study area (offshore of Follets Island and the Brazos Delta).

Sabine, Heald, Shepard, and Thomas banks, located above and adjacent to the Trinity and Sabine incised fluvial valleys, represent submerged barrier shorelines. The shoreline submergence events have been correlated
with flooding surfaces located within the Trinity incised valley. Maps of the paleoenvironments bound by these flooding surfaces indicate that each paleoshoreline submergence event is associated with estuarine environments being shifted tens of kilometers landward. Rising sea level during the Holocene was the forcing mechanism behind these events.

Freeport Rocks Bathymetric High, a bank located further to the west, also represents a submerged barrier shoreline. However, this feature was not emplaced during the Holocene transgression; rather, it was deposited during the middle Wisconsin (oxygen isotope stage 3) sea-level highstand. The shoreline associated with the bank was mapped regionally at -15 m ± 2 m. This suggests sea level was around 15-30 m shallower than what oxygen isotope and coral records indicate for stage 3.

A detailed sedimentary and geomorphologic study was undertaken on the Brazos Delta, Texas to better define the facies architecture and controlling processes on wave-dominated delta evolution. The Brazos Delta is primarily composed of fine-grained sediments; prodelta clay composes more than half of the sediment volume. The facies architecture is not representative of the classic strandplain model for wave-dominated deltas due to the strong influence of floods on deltaic evolution.
ACKNOWLEDGMENTS

The encouragement, guidance, and generosity Dr. John B. Anderson provided throughout the past five years has helped me tremendously. I could not have asked for a better advisor. I thank Drs. Dale Sawyer, Peter Vail, and Frank Fisher for serving on my committee. Dr. John Bradford helped with the collection and processing of the geophysical data, and Dr. Marco Taviani helped with the paleontological data. I learned a great deal from both of these people. Dr. Virginia Sisson flew me over the Brazos Delta many times to photograph the morphologic changes that occurred in response to the 1992 flood and wave erosion. Her support is much appreciated. I have benefited a great deal from interacting with students in the sedimentology lab. Our discussions have shaped many of my ideas. I thank Patricia, my wife, for her understanding and support. She never once complained about my long hours and low paychecks. She agrees that the easiest, most lucrative road is not always the best one to take, and with her I feel anything can be accomplished.

This project was funded by a consortium of oil companies, and a GEM fellowship.
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CHAPTER 1

SEDIMENTARY FACIES AND EVOLUTION OF LATE PLEISTOCENE TO RECENT COASTAL LITHOSOMES ON THE EAST TEXAS SHELF: AN INTRODUCTION

Examination of Late Pleistocene to recent coastal lithosomes can help refine current models aimed at understanding how coastal environments respond to a variety of forcing mechanisms including changes in subsidence rate, sediment supply (climate), and eustasy. Results from this work will help petroleum geologists predict and characterize hydrocarbon reservoirs more effectively, and help coastal planners predict how modern coastal systems will evolve.

The study area is located in the northern Gulf of Mexico, along the east Texas coast and continental shelf. It spans from Sabine Pass, which marks the Texas-Louisiana border, to Matagorda Island (Fig. 1-1). Specific environments investigated include: bay complexes (fluvial, bay-head delta, middle bay, tidal delta, and barrier shoreline environments), offshore banks, incised valleys, and the Brazos Delta.

This document is divided into seven chapters. Chapter one provides a layout of the document and an introduction to each topic presented. Chapters two through six were written as separate manuscripts to be submitted for
Figure 1-1. Map of the Texas gulf coast. The study area is located on the east Texas coast and inner continental shelf.
publication; therefore, some redundancies exist. Chapter seven summarizes the thesis by discussing the differences in coastal lithosome preservation throughout the study area.

**Holocene sand banks located on the east Texas continental shelf: Chapter 2**

Studies of modern and ancient shelf sand bodies have produced two rather different schools of thought concerning the origins of these bodies. One school argues that formation is the result of hydrodynamic processes that occur on the continental shelf (Swift et al., 1974; Huthnance, 1982; Swift et al., 1984; Swift and Rice, 1984; Boczar-Karakiewicz and Bonas, 1986; Tillman and Martinsen, 1984, 1987; Rine et al., 1991; Fig. 1-2a); and the second school contends that formation is the result of coastal lithosomes being overstepped and stranded on the shelf during transgression (Curry, 1960; Sanders and Kumar, 1975; Stubblefield et al., 1984; Penland et al., 1988; Walker and Bergman, 1993; Bergman, 1994; Wagle and Veerayya, 1996; Fig. 1-2b). These two schools of thought represent end-member type models. A study of Sabine, Heald, Shepard, and Thomas banks, which are located adjacent to and over the Trinity/Sabine incised valley on the east Texas shelf, was undertaken to examine the influence of these two end-member type models on bank formation and evolution. The investigation focused on the facies architecture,
Figure 1-2. Two end-member models of bank formation. A shows formation as the result of hydrodynamic processes that occur out on the continental shelf. B shows formation as the result of coastal lithosomes being overstepped and stranded on the shelf during transgression.
stratigraphic occurrence relative to key seismic surfaces, distribution relative to the Trinity/Sabine incised fluvial valley, and origin of the banks. Specific questions that were addressed include: 1) What are the facies that make up the banks?, 2) How did the banks form?, 3) What is the relationship between the banks and the Trinity/Sabine incised valley?, and 4) How do the banks evolve, and what is their preservation potential?

Holocene to recent tidal delta complexes of the Trinity incised valley: Chapter 3

Examination of Holocene and modern incised valley fill deposits aids in the identification, prediction, and characterization of ancient equivalent type deposits. The facies architecture and evolution of the modern Bolivar Roads tidal delta complex was compared to Holocene tidal-delta deposits preserved within the Trinity incised valley on the east Texas shelf (Fig. 1-3). Due to the depth of the offshore tidal delta complexes within the Trinity incised valley, sampling was not possible. However, in the modern Bolivar Roads tidal delta complex direct correlation between seismic facies and lithologic units has been accomplished. Therefore, lithologies of the offshore tidal deltas were inferred by comparing their seismic facies to the modern Bolivar Roads tidal delta seismic facies. Specific questions addressed in this study include: 1) Are
Figure 1-3. Bathymetric map of the east Texas shelf and Bolivar Roads tidal delta complex. The Trinity/Sabine incised river valley (gray, Thomas and Anderson, 1994) and tidal delta complex deposits (striped) preserved within it are also shown.
sandy proximal tidal delta facies preserved within the Trinity incised valley?, 2) How do tidal delta complex deposits differ from adjacent shoreface deposits?, and 3) How does the rate of sea-level rise impact facies preservation.

**Impact of Holocene sea-level rise on coastal systems: Chapter 4**

One of the key issues raised in the 1990 Intergovernmental Panel on Climate Change report concerns the potential impact of global warming on sea level, specifically the magnitude and rate of sea-level rise that will occur over the next century. It is clear that an increased rate of sea-level rise will have a number of adverse impacts on world coasts, such as wetlands loss and accelerated coastal erosion, however the magnitude of these changes remains unpredictable (Warrick et al., 1996; Fig. 1-4). The best prediction of future sea-level rise from global warming is around 49-55 cm/century (Warrick et al., 1996). This is close to the average rate of sea-level rise during the Holocene (10,000 BP to Present). This being the case, the geological record of coastal change during the Holocene was examined in the northern Gulf of Mexico to better predict coastal response to accelerated sea-level rise. Specific questions addressed in this study include: 1) How do coastal systems respond to sea-level rise?, 2) Did coastal systems respond to a continuous rate of sea-level rise, or did the rate of sea-level rise vary throughout the Holocene?, and 3) Which coastal environments are most sensitive to sea-level rise?
Figure 1-4. East Texas study area today (above) and after a 2 m rise in sea level (below). These two figures illustrate the impact of the IPCC worse case 200 year future prediction of sea-level rise on the east Texas study area.
Position of the Middle Wisconsin shoreline on the Texas inner continental shelf: Chapter 5

Oxygen isotope data from deep sea cores have been widely applied as sea-level proxies (Mix and Ruddiman, 1984; Shackleton, 1987). During isotope stage 3 (24-59 ka), the SPECMAP curve (Imbrie et al., 1984) and the temperature corrected benthonic oxygen isotope record of east Pacific core V19-30 (Shackleton, 1987) predict sea level was at ~85 m and ~75 m respectively (Fig. 1-5). These estimates do not compare well with much of the direct geologic evidence for the position of the middle Wisconsin (stage 3) maximum sea-level shoreline (Fig. 1-5). The types of direct geologic evidence used to constrain the position of the stage 3 highstand of sea level are: radiocarbon dates from shallow-water fauna (Curry, 1965; Milliman and Emery, 1968; Blackwelder et al., 1979; Finkelstein and Kearney, 1988); uranium series ($^{234}$U-$^{230}$Th) dating of reef terraces in Barbados (Mesolella et al., 1969; Steinen et al., 1973; Mathews, 1984; Bard et al., 1990), New Guinea (Bloom et al., 1974; Chappell, 1974, 1983; Bloom and Yonekura, 1985; Chappell et al., 1996), and Indonesia (Chappell and Veeh, 1978 and Bard et al., 1996); and high-resolution seismic stratigraphy (Suter et al., 1987; Thomas and Anderson, 1989,1991; Wellner et al., 1993; Roy et al., 1997). This evidence for the position of the stage 3 maximum sea-level highstand is inconsistent, for it has been
Figure 1-5. Position of the stage 3 shoreline based on oxygen isotope data (A, B), coral age depth data (A, B), and seismic stratigraphic studies (C). These studies are not in agreement on the depth of this shoreline.
interpreted as being anywhere from at or above present day levels (Curry, 1965; Milliman and Emery, 1968; Blackwelder et al., 1979; Finkelstein and Kearney, 1988; ) to -85 m (Bard, et al., 1990). In order to better constrain the elevation of the stage 3 shoreline, we conducted a focused investigation on an interfluvial portion of the east Texas shelf where the paleoshoreline can be projected to the modern sea floor. Specific questions addressed in this study include: 1) At what depth is the stage 3 shoreline on the Texas shelf?, and 2) What is the preservation potential of highstand shorelines?

**Evolution and facies architecture of the modern Brazos Delta, Texas: Chapter 6**

The well-known ternary delta classification system indicates delta morphology and three-dimensional structure is dominantly controlled by river discharge, tidal range, and wave energy (Fisher 1969; Fisher et al. 1969, Coleman and Wright 1975; Galloway 1975; Fig. 1-6). According to this classification, the subaerial shape of wave-dominated deltas is cuspatate, while fluvial-dominated deltas are elongate (Fig. 1-6). Based on these criteria, the Brazos delta has previously been classified as a wave dominated delta (Galloway 1975; Fig. 1-6). Although this classification scheme accurately describes the gross controlling factors for subaerial delta morphology, it makes
Figure 1-6. Schematic diagram illustrating the threefold division of deltas into fluvial, wave, and tide-dominated types. The Brazos Delta has been classified as a wave-dominated delta (modified after Galloway, 1975).
inaccurate assumptions concerning the mechanism for emplacing sands, and the character and extent of delta facies, particularly offshore facies. In order to better define the facies architecture and controlling processes on wave-dominated delta evolution, a detailed sedimentary and geomorphologic study was undertaken on the Brazos Delta, Texas. Specific questions addressed in this study include: 1) What is the facies architecture of the delta?, 2) How has the delta evolved since formation in 1929?, and 3) What are the relative impacts of floods and storms on delta morphology?
CHAPTER 2

SEDIMENTARY FACIES AND GENESIS OF HOLOCENE SAND BANKS ON THE EAST TEXAS INNER CONTINENTAL SHELF

Summary

Sediment cores and high-resolution seismic and side-scan sonar data were collected from four shelf banks on the east Texas inner continental shelf. Sabine, Heald, Shepard, and Thomas banks all have similar sediment facies, structure, and genesis. The banks are comprised of three facies (top to bottom): A) an interbedded shell hash and sand unit; B) a muddy-sand unit characterized by a seaward-prograding and chaotic seismic facies, and C) an interbedded sand and mud unit characterized by landward-dipping seismic reflectors. These three sediment facies represent amalgamated storm beds, lower-shoreface or ebb-tidal delta, and back-barrier/flood-tidal delta environments respectively. Facies B and C were deposited during a time of relatively slow sea-level rise and were stranded on the shelf during a rapid sea-level rise. Facies A is the result of storms and wind-driven currents reworking the paleoshoreline deposits on the shelf. The banks are drowned paleo-shorelines restricted to the area above and immediately adjacent to the Trinity and Sabine incised fluvial valleys. This association is explained by the greater
thickness of sands within the valleys (larger sources of sands), greater accommodation space, and greater subsidence rate within the valleys (greater preservation potential).

**Introduction**

Previous studies of modern and ancient shelf sand bodies have produced two rather different schools of thought concerning the origins of these bodies. One school argues that formation is the result of hydrodynamic processes that occur on the continental shelf (Swift et al., 1974; Huthnance, 1982; Swift et al., 1984; Swift and Rice, 1984; Boczar-Karakiewicz and Bonas, 1986; Tillman and Martinsen, 1984, 1987; Rine et al., 1991); and the second school contends that formation is the result of coastal lithosomes being overstepped and stranded on the shelf during transgression (Curry, 1960; Sanders and Kumar, 1975; Stubblefield et al., 1984; Penland et al., 1988; Bergman, 1994; Walker and Bergman, 1993; Wagle and Veerayya, 1996). These two schools probably do not explain the origin of all shelf sand ridges and banks; in fact, they represent end-member type models.

Debate about the origin of the lower Campanian Shannon Sandstone in the Powder River Basin is an example of a recent controversy concerning the origin of shelf sand bodies (Tilliman and Martinsen, 1984, 1987; Gaynor and Swift, 1988; Walker and Bergman, 1993; Bergman, 1994). The Shannon
Sandstone consists of a series of elongated sand bodies contained within marine shales. One group of investigators believes that these sand bodies were deposited as a shelf ridge complex 160 km from the shoreline (Tillman and Martinsen, 1984, 1987; Gaynor and Swift, 1988; Rine et al., 1991). Another group of investigators argues that the Shannon sandbodies are shoreface deposits stranded and preserved on the shelf due to eustatic fluctuations (Walker and Bergman, 1993; Bergman, 1994).

Sand ridges on the New Jersey continental shelf have been studied as possible modern analogs to ancient shelf sand bodies. However, interpretations are mixed and indicate that the ridges were either deposited as a result of hydrodynamic processes occurring on the shelf (Swift et al., 1978; Swift et al., 1984; Rine et al., 1991) or as barrier islands were “drowned” and modified during transgression (Stubblefield et al., 1984).

Snedden et al. (1994) conducted a detailed study of Peahala ridge located on the New Jersey continental shelf. The formation of this ridge was interpreted as the result of overstepping of a coastal lithosome (ebb-tidal delta) during transgression and subsequent reworking of this feature by hydrodynamic processes (Snedden et al., 1994, Snedden et al., in this volume).

The east Texas inner continental shelf is a mud-dominated shelf. The only sands on the shelf are in the form of banks or are located within the Trinity/Sabine incised valley. This type of setting is very different from that at any other modern bank study site, but it is similar to the lower Campanian
Shannon Sandstone in the Powder River Basin, for which east Texas banks are considered a reasonable modern analog.

The origin and evolution of four shelf sand banks (Thomas, Shepard, Heald, and Sabine banks) on the east Texas continental shelf were studied to better understand the mechanisms for shelf sand body formation (Fig. 2-1). These sand banks currently exist at water depths of -29 m, -17 m, -15 m, and -12 m, respectively, and are surrounded by marine muds. The shallower banks (Sabine, Heald, and Shepard) exhibit signs of active sediment transport and deposition, while it is unclear whether the outer, deeper bank (Thomas Bank) is currently active or is being buried in shelf muds.

Curry (1960) was the first to study the four shelf sand banks and speculated that they were formed during the Holocene transgression. Nelson and Bray (1970) performed a subsequent study utilizing high-resolution seismic data, sediment cores, and surface grab samples from Sabine and Heald banks. They concluded that these banks formed in a shallow marine environment due to the reworking of shoreline deposits. Our investigation focused on the facies architecture, stratigraphic occurrence relative to key seismic surfaces, distribution relative to the Trinity/Sabine incised fluvial valley, and origin of the banks.

Study Area
Figure 2-1. Bathymetric map of the east Texas shelf showing the locations of seismic lines, cores, and cross-sections. The Trinity/Sabine incised river valley (gray) and tidal inlet deposits (striped) preserved within it are also shown (Thomas and Anderson, 1994; Siringan, 1993).
The Texas coast is characterized by fair weather astronomical tides ranging from 45 to 60 cm and relatively low-amplitude waves with periods commonly within 4-6 s (Morton and McGowen, 1980). The maximum mean significant wave height reported for this area is 2.1 m, and the maximum significant wave period is 8.3 s. Historical wave data indicate that wave heights can reach up to 7 m (Armstrong, 1980).

Unidirectional bottom flows produced by wind forcing are probably the most important mechanism for shelf sediment transport in the Gulf of Mexico (Morton, 1981). Fair weather wind- and storm-generated currents on the Texas shelf generally flow along-shore to the southwest (Morton, 1977; Snedden et al., 1988). However, Snedden et al. (1988) observed that on the shallow portions of the central Texas shelf (southwest of the study area) the southwest current switches to the northeast during summer months, coinciding with a southerly shift in prevailing winds. Average current speeds reported for the east Texas shelf range from about 10 to 30 cm/s. Current speeds measured during the passage of tropical cyclones ranged from 53 cm/s to 180 cm/s, with speeds farther offshore reaching in excess of 250 cm/s (Armstrong, 1980). During tropical storm Delia, bottom currents up to 200 cm/s at -18 m water depth were recorded approximately 30 km west-southwest of Sabine Bank (Forristall et al., 1977). The center of tropical storm Delia passed within 5 km west of the current meter.
Late Pleistocene to Holocene History

A brief overview of the response of depositional systems in the study area to the rise of sea level during the last glacial eustatic cycle (18,000 yr BP to present) is necessary before bank formation and origin can be understood. A more detailed description of the late Quaternary stratigraphy of shelf sediments in the study area can be found in Thomas and Anderson (1994) and Anderson et al. (1992, 1996).

During the Stage 2 sea-level lowstand, the Trinity and Sabine rivers incised into older highstand fluvial-deltaic deposits on the continental shelf; these valleys converge approximately 50 km offshore (Fig. 2-1). During the subsequent rise of sea level, the incised valleys backfilled with fluvial, estuarine, and coastal (flood and ebb tidal delta) deposits. Seismic and lithologic data used to map lithofacies within the incised valley indicates incomplete valley-fill sequences bounded by intermediate flooding surfaces (Thomas and Anderson, 1994). The valley-fill facies architecture is characterized by backstepping parasequences that have been attributed to an episodic or step-like rise of sea level (Anderson et al., 1991; Thomas and Anderson, 1994). Periods of relatively slow sea-level rise are manifested by tidal inlet and associated flood and ebb tidal delta deposits. The banks lie adjacent to or directly over these tidal inlet deposits (Fig. 2-1). The interfluve portions of the study area are characterized by a thin (< 1m) layer of bioturbated marine mud resting directly on stiff, varigated Pleistocene (Beaumont
Formation) clay (Fig. 2-2). The Beaumont formation consists of mainly coastal and delta plain silts and clays deposited on the subaerially exposed shelf during the previous highstand. The surface separating these units is an amalgamated stage 2 sequence boundary and transgressive ravinement surface (Siringan and Anderson, 1994).

Methods

Approximately 4200 km of high-resolution seismic data, 15 km of side-scan sonar data, and 57 sediment cores were collected aboard the R/V Lone Star from the east Texas shelf (Fig. 2-1). A Datasonics CAP 6000 Chirper system using a 10 ms linear sweep from 1.5-7 kHz was used to collect the seismic data, which were then processed using ProMAX software. The processing consisted of applying a bandpass filter (500-800-2000-4000 Hz), automatic gain control (3 ms operator length), and trace mixing (over 7 traces). The method used for seismic interpretation centered around the recognition of important bounding surfaces and the identification of variations in seismic character. The acoustic travel time to depth conversion was computed under the assumption that the velocity of the upper few meters of sediment is the same as that of salt water (1525 m/s).

The side-scan sonar data were collected using a Datasonics SIS 1000 system. Sediment cores provide ground truth for the seismic data. They were
Figure 2-2. Representative core photographs that characterize the sedimentary units that make up the banks, and the interfluvial portions of the study area.
collected using a pneumatic hammer coring device and a piston coring device, both capable of recovering cores up to 5 m in length. Radiocarbon dating (radiometric and accelerated mass spectrometry (AMS) techniques) of shell and peat material sampled from the cores was performed by Beta Analytic Inc.

Results

Seismic Data Analysis

The banks are large bathymetric features. Sabine Bank, the largest feature (600 km²), rests 7.5 m above the surrounding seafloor. Echo sounding profiles oriented SE-NW across and perpendicular to the long axis of the banks show steeply dipping seaward flanks, irregular tops, and gently dipping landward flanks (Figs. 2-3, 2-4, 2-5, and 2-6). Figure 2-7 shows an Echo sounding axial profile oriented SW-NE across Heald Bank. Directly on top of the bank, very large (1-m-high and 700-m-long) two-dimensional or three-dimensional subaqueous dunes (Ashley, 1990) are shown migrating toward the NE. Side-scan sonar data acquired along the same transect did not image any smaller-scale bedforms. The NE flank of Heald Bank is the steepest feature in the study area (Fig. 2-7).

The banks contain three seismic facies units (SFU), A through C. A is the uppermost unit capping each bank (Figs. 2-3, 2-4, 2-5, and 2-6) and is characterized by a chaotic to acoustically reverberating seismic reflection
Fig. 2-3. Cross-section A-A', echo-sounding profile, and a portion of a seismic (chirper) line through Sabine Bank. Five cores collected along the transect are indicated on the cross-section.
Figure 2-4. Cross-section B-B', echo-sounding profile, and a portion of a seismic (chirper) line through Heald Bank. Six cores collected along the transect are indicated on the cross-section.
Figure 2-5. Cross-section C-C', echo-sounding profile, and a portion of a seismic (chirper) line through Heald Bank. Four cores collected along the transect are indicated on the cross-section.
Figure 2-6. Echo-sounding profile trace D-D', and a portion of a seismic (Chirper) line through Thomas Bank. Only two cores were collected along this transect; therefore, no cross-section is shown.
Figure 2-7. Echc-sounding profile across Heald Bank showing large-scale (2-m-high and 700-m-long) sand waves migrating toward the northeast.
pattern. Imaging the internal stratal geometries within SFU A is difficult due to the high amplitude sea-bottom reflection. This facies either shows a sharp erosional contact with, or is difficult to distinguish from, SFU B. SFU B is characterized by seaward-dipping reflectors. SFU B is most discernible on the seaward side of the banks, where it is exposed at the seafloor. This facies generally has a sharp erosional contact with SFU C, which is characterized by landward-dipping reflectors and large-scale cross-bedding. This facies is thickest directly over the incised valley.

Lithologic and Paleontologic Data Analysis

Only cores collected from the banks or the Trinity/Sabine incised valley contain any significant amount of transgressive deposits. Cores from interfluvial areas of the shelf penetrated a thin layer (tens of centimeters thick) of Holocene bioturbated, olive-gray sandy mud with basal clay rip-ups lying on mottled green and red, very stiff Pleistocene (Beaumont Formation) clay (Fig. 2-2). The shear strength of the clay is >1.0 kg/cm².

The correlation of seismic facies units with lithologic units is straightforward; the banks are comprised of three general lithologic units, each of which corresponds with one of the seismic facies units. From top to bottom these are: A) an interbedded shell hash and sand unit; B) a muddy sand unit, and C) an interbedded sand and mud unit (Fig. 2-2). Unit A, the shelly sand facies, is approximately 2-5 m thick at the crest of the banks. The shell hash
beds reach up to 25 cm in thickness while the sand beds are up to 45 cm thick. Contacts between the beds are generally sub-horizontal. Shells from unit A were sourced from a variety of environments, including bay-lagoonal (Crassostrea virginica), back-barrier/nearshore (Crassinella lunulata, Mulinia lateralis, Natica pusilla, Anchis obesa), and open marine (Plicatula gibbosa, Strigilla mirabilis, Semele bellastriata; Parker, 1960; Andrews, 1992). The boundary between units A and B (the muddy sand facies) varies from sharp to gradational.

Unit B is 2-3 m thick and comprised of a clean to <10% mud-bioturbated sand. Grain size of the sand ranges from 2.5 to 2.0 phi. This facies subcrops on the bank flanks. Most of the shells from unit B (Natica pusilla, Olivella dealbata, Ervilia concentrica, Caecum johnsoni, and Anadara transversa) represent shoreface/inlet environments (Parker, 1960; Andrews, 1992). The boundary between unit B and unit C, the interbedded sand and clay facies, is gradational and difficult to identify from cores alone. The seismic data help to resolve this problem.

Unit C is slightly burrowed with thin (<2 cm), intercalated layers of sand and silt. Shells from this unit include Ostrea equestris, Nassarius acutus, Nuculana concentrica, and Mulinia lateralis. Although all the units contain Mulinia lateralis, this species is most abundant in unit C, where some layers are monospecific. This shell assemblage is typical of bay/inlet environments (Parker, 1960; Andrews, 1992).
Nelson and Bray (1970) obtained radiocarbon dates for 15 samples from Sabine Bank and for 20 samples from Heald Bank. An additional three radiocarbon dates from Heald Bank and two from Sabine Bank were obtained for the recent study. Marine shells collected from Sabine Bank unit A yielded ages ranging from 215 ± 150 yr BP to 383 ± 210 yr BP. Back-barrier shells from Sabine Bank unit A yielded ages ranging from 2,210 ± 243 yr BP to 7,500 ± 330 yr BP (Nelson and Bray, 1970) indicating that the shells are relict material. Nelson and Bray (1970) reported a radiocarbon date of 4,630 ± 215 yr BP from an *Anadara transversa* shell acquired from Sabine Bank at the base of unit B. An AMS date of 4490 ± 50 yr BP was measured from an articulated *Ostrea equestris* shell sampled from Sabine Bank at the top of unit C. Freshwater peats sampled from below Sabine Bank at the base of unit C yielded a radiocarbon age of 7,800 ± 70 yr BP. *Mulinia lateralis* shells from Heald Bank in unit B yielded an age of 7065 ± 275 yr BP. Freshwater peats sampled from below unit C in Heald Bank yielded an age of 8570 ± 70 yr BP.

**Discussion**

The sand banks are presently being modified by currents and storms as indicated by bedforms on Heald Bank (Fig. 2-7). High gradients associated with seaward (southern) sloping bank flanks are attributed to the fact that storms typically approach the east Texas coast from the SW (Snedden et al., 1988).
Unit A is mostly composed of reworked sand and shells. Radiocarbon dating of marine and back-barrier shells from Sabine Bank unit A yielded quite variable ages, \(215 \pm 150\) yr BP to \(7,500 \pm 330\) yr BP (Nelson and Bray, 1970), indicating a contribution of modern and reworked material. The gradational nature of the contact between units A and B, as observed from core and seismic data, indicates reworking of unit B into unit A. The older shells from unit A in both Sabine Bank and Heald Bank are always back-barrier or shoreface/inlet species, which indicates that they were sourced from excavation of earlier deposits. The presence of both back-barrier and marine mollusc shells indicates reworking and bed-load transport at current water depths. Thus, unit A represents amalgamated storm beds presently being modified and deposited at water depths of up to -24 m.

Unit B is only exposed at the seafloor on the banks' seaward flanks, which are most heavily affected during storms. SFU B shows seaward-dipping reflectors and an inlet or shoreface faunal assemblage, which indicates that this unit represents either an ebb-tidal delta or lower shoreface environment. Unit B lacks distinct sand mud interbeds and laminations, characteristic of the modern Bolivar Roads ebb-tidal delta (Siringan and Anderson, 1993). The lithology instead resembles the bioturbated shoreface deposits offshore of Galveston Island (Siringan and Anderson, 1994).

SFU C shows landward-dipping reflectors and a bay or tidal inlet faunal assemblage, which indicates that this unit represents either a flood-tidal delta or
other back-barrier environment. The seismic expression of the modern Bolivar Roads flood-tidal delta shows channel stacking and cut-and-fill geometries near the inlet, and landward-dipping reflectors near the bay (Siringan and Anderson, 1993). This compares well to the geometries observed in SFU C. Siringan and Anderson (1993) collected 18 cores from the modern Bolivar Roads flood-tidal delta. The distal portions of the flood-tidal delta were described as mud-dominated (interlaminated clay and fine sand) with few shell beds (Siringan and Anderson, 1993). Unit C has similar lithologies.

The examination of Thomas Bank was intended to provide an example of a bank situated below storm influence. Thomas Bank was first studied by Thomas and Anderson (1994), although they referred to it as -29 m bank. Thomas and Anderson (1994) suspected that preserved barrier island/lagoon or strandplain facies directly underlie Thomas Bank, similar to Sabine, Heald, and Shepard banks. Our hypothesis was that Thomas Bank would be draped in marine muds because of its depth. Although the burial process has clearly begun, Thomas Bank is not currently draped in marine muds. Two piston cores were collected from Thomas Bank. The first core location was on the northern bank flank and sampled 133 cm of olive-gray marine mud and the second location was on the top of the bank and sampled 20 cm of well-sorted, shelly sand 1.5-2.0 phi (Facies A; Fig. 2-6). The thick marine mud unit sampled from the northern bank flank is evidence that Thomas Bank may be in the process of becoming completely buried.
Sabine Bank appears to be the youngest reworked paleoshoreline deposit in the study area. Radiocarbon dates from fresh water peats below unit C and an Ostrea equestris shell at the top of unit C indicates that Sabine Bank is between 7,800 ± 70 yr BP and 4,490 ± 50 yr BP in age. The gradational contact between units B and C makes it difficult to ascertain whether the dates from the Anadara transversa shell (4,630 ± 215 yr BP) in the base of unit B and the Ostrea equestris shell (4,490 ± 50 yr BP) in the top of unit C are actually from those units or from the gradational contact. Because the Ostrea equestris shell was articulated, it was preserved in-situ and therefore the 4,490 ± 50 yr BP date measured from it is the best youngest age estimate for this paleoshoreline. Modern and 7,500 ± 330 yr BP ages from unit A also indicate active reworking at present water depths. Fresh water peats below unit C show the age of the Heald Bank paleoshoreline to be between 8570 ± 70 yr BP and 7065 ± 275 yr BP (from Mulinia lateralis shells in unit B). The timing of Thomas Bank and Shepard Bank formation is unknown but by inference from the Holocene sea level curve (Bard et al., 1990), the paleoshoreline associated with Thomas Bank is approximately 9,500 years old. Shepard Bank either formed before or simultaneous with Heald Bank.

**Bank Formation and Evolution**

Bank formation is closely tied to changing rates of sea-level rise during transgression, the Trinity/Sabine incised valley, and the hydrodynamic setting of
the shelf. During the last lowstand, an irregular surface—a sequence boundary-formed on the shelf. This surface marks the top of the Beaumont Formation. During the subsequent transgression this surface was reshaped by wave erosion and the incised valleys were infilled with back-stepping parasequences (Thomas and Anderson, 1994). The back-stepping parasequences within the Trinity/Sabine incised valley record flooding events during rapid sea-level rises. Tidal inlet deposits within the valley and sand banks above and adjacent to the valley represent episodes when sea-level rose slowly, allowing barriers and associated inlets to evolve.

Figure 2-8 illustrates the four general stages of bank evolution in the study area. During stage one, sea-level rose relatively slowly, enabling stabilization of the shoreline and construction of barrier islands and tidal inlets. Units B and C were deposited at this time. Unit C was deposited in back-barrier bays situated within and adjacent to the Trinity/Sabine incised valley. Hence, these deposits were formed in an environment very similar to modern East Bay and West Bay (Fig. 2-1). The base of unit C represents a flooding surface—a bayline—as evidenced by peat near the contact, back-barrier muds above the contact, and Pleistocene clay below the contact. During stage 2, the rate of sea-level rise accelerated. The barrier island responded by retreating shoreward, depositing unit B over unit C as the shoreface reestablished itself. The contact between unit C and unit B is the ravinement surface. Unit B represents the preserved shoreface deposits of the prograding shoreline.
Figure 2-8. Block diagram illustrating barrier island formation, retreat, overstepping, and reworking, which are the four stages of bank genesis and evolution.
Subsidence rates are much faster over the incised valley (0.62 cm/yr based on tide gauge records from Galveston Bay) relative to the interfluvial areas of the shelf (0.01 cm/yr; Paine, 1993); this ultimately helped to preserve unit C along the flanks of the valley. Stage 3 was a time of rapid sea-level rise, which caused the coastal lithosomes to be “drowned,” or overstepped, leaving them isolated on the shelf. This final flooding produced a second ravinement surface that truncated the shoreface (unit B) and is located within or at the base of unit A.

An examination of how Galveston Island and Bolivar Peninsula would respond to a rapid rise in sea level illustrates the above sequence of events. The Bolivar Peninsula and Galveston Island sand bodies both thicken toward the Trinity River incised valley (Fig. 2-9). This is due to Bolivar Roads tidal inlet, a stable, mixed-energy, tide-dominated tidal inlet/delta complex that sits directly over the incised valley and the easily compacted/eroded soft estuarine valley fill deposits (Siringan and Anderson, 1993). The depth to the modern ravinement surface is marked by the onlap of offshore marine muds onto shoreface deposits and corresponds closely to the toe of the shoreface (abrupt change in the shoreface profile). The ravinement surface increases from -7 m off Bolivar Peninsula to -9 m off Galveston Island (Siringan and Anderson, 1994; Fig. 2-9). Offshore Bolivar Peninsula the ravinement surface cuts deeply into the shoreface profile, leaving only a thin layer of Holocene marine mud resting on Pleistocene clay (Fig. 2-9). Offshore Galveston Island the ravinement surface
Figure 2-9. Cross-section E-E’ through Galveston Island and Bolivar Peninsula shows thick barrier island and shoreface sands located within and adjacent to the Trinity River incised valley. Profiles G-G’ and F-F’ are representative core transects that illustrate how shoreface ravinement has removed virtually all coastal deposits on the inner shelf. The ravinement surface is marked by offshore marine muds onlapping either Pleistocene deposits (offshore Bolivar Peninsula) or lower shoreface deposits (offshore Galveston Island) and occurs at -7m to -9 m water depths. With continued sea-level rise and current depths of ravinement, coastal lithosomes will be eroded everywhere except in the incised valley (modified from Cole and Anderson, 1982, Siringan and Anderson, 1993, and Siringan and Anderson, 1994).
has cut the shoreface at a higher level, preserving shoreface deposits beneath marine muds (Fig. 2-9). When the ravinement surface cuts through the expanded barrier island/inlet complex; the shoreline deposits located adjacent to and directly over the Trinity River incised valley—a relatively thick section of the coastal lithosome—will be preserved (Fig. 2-9). Bolivar Roads tidal inlet will continue to redistribute sediment to the tidal delta complex and the adjacent shoreline during the initial sea-level rise. Hence, the tidal delta complex and the shoreline adjacent to it will remain in place longest before the entire barrier island complex is removed or truncated and finally submerged. It is this portion of the barrier island complex that will be stranded on the shelf as a bank. In the case of offshore banks, initial flooding created the lower ravinement surface, and then progradation and aggradation of coastal lithosomes created the bank topography. Galveston Island sits atop a ravinement surface and has a similar history (Bernard et al., 1970).

The overstepping events correspond to episodes of significant flooding, as indicated by backstepping (tens of kilometers) of the valley fill facies in the Trinity/Sabine incised valley (Thomas and Anderson, 1994). Subsequent reworking of units B and C by storms and wind-driven currents led to the formation of unit A. With time, unit A continues to thicken at the expense of the lower units. Heald, Shepard, and Thomas banks are older and have thicker unit As than Sabine Bank, indicating that the longer these features are stranded on the shelf, the sandier they become. The banks also migrate up-dip
(landward) through time as the lower units are reworked into unit A. Thomas Bank has migrated up-dip approximately 5 km from its associated tidal inlet deposit (Thomas and Anderson, 1994). The banks may evolve to a point where only unit A is preserved as a shelf sand body encapsulated within marine shelf muds; however, none of the banks studied have reached that point. The 133 cm of marine mud sampled from the northern flank of Thomas Bank are evidence that it has started to become encapsulated. Holocene sea-level rise was very rapid, certainly in comparison with nonglacial intervals of time, such as the Mesozoic. Given a slower rate of sea-level rise, bank isolation and encapsulation in marine muds on the shelf are more likely.

This sand bank formation and evolution model is similar to what Berne et al. (1994) evoke in the southern North Sea for the formation of Middelkerke Bank, and to what Snedden et al. (in this volume) evoke for the formation and evolution of sand ridges on the Atlantic shelf. There are important differences though. Snedden et al. (in this volume) interpret a very rigorous final stage of ridge evolution where a ridge loses almost all of its original characteristics through extensive migration and sediment volume increases. Thomas Bank is reaching its final stage of evolution by becoming encapsulated in marine muds; however, its seismic facies have remained similar in character to those of the younger banks, and it is still situated adjacent to the Trinity/Sabine incised valley. Thomas Bank may still evolve to a point where only facies A is present; however, given the present shelf setting, it will most likely be entirel/
encapsulated in marine muds before it is completely reworked. Hence, these
banks on the east Texas shelf will always remain in a predictable position:
adjacent to the Trinity/Sabine incised valley and a tidal delta complex
preserved within the valley (Thomas and Anderson, 1994).

Conclusions

Holocene sand banks of the east Texas shelf are comprised of three
facies (top to bottom): A) an interbedded shell hash and sand unit; B) a muddy-
sand unit characterized by a seaward-prograding and chaotic seismic facies,
and C) an interbedded sand and mud unit characterized by landward-dipping
seismic reflectors. Unit C represents a back-barrier/flood-tidal delta
environment and unit B represents a lower-shoreface or ebb-tidal delta
environment. These coastal lithosomes were stranded on the shelf during rapid
transgression. Unit A was deposited as storms and wind-driven currents
reworked ancestral paleoshoreline deposits on the shelf. It is possible that the
lower units (B and C) will become completely reworked into unit A, resulting in a
shelf-sand body encapsulated within shelf muds. Thus, sand banks on the east
Texas shelf are formed through a combination of hydrodynamic processes
occurring on the shelf and overstepping and stranding of coastal lithosomes
during transgression.
CHAPTER 3

HOLOCENE TIDAL DELTAS OF THE TRINITY INCISED VALLEY: ANALOGS FOR EXPLORATION AND PRODUCTION

Summary

The facies architecture and evolution of modern and Holocene tidal-delta deposits were examined using high-resolution seismic (chirper) data, sediment cores, and boring descriptions. During the last transgression (18,000 yr. BP to present) the Trinity/Sabine incised valley backfilled with continuous fluvial and bay-head delta facies and discontinuous middle-bay and tidal delta facies. The tidal delta facies preserved within the valley extend up to 22 km in dip-direction and are up to 12 m thick. The modern Bolivar Roads tidal delta complex is similar in size to those preserved within the offshore portion of the valley, and is a mud-dominated system. The overall size of the tidal delta complex decreased around 1,500 yr. BP as the peninsula accreted and the inlet grew narrower. Prior to this time the tidal delta complex was two times larger.

The tidal delta seismic facies is characterized by stacked landward and seaward dipping reflectors overlain by a ravinement surface. Lithofacies have been described using sediment cores for the Holocene tidal delta complexes.
Isopach maps of these tidal deltas provide reservoir analogs for subsurface exploration and production.

**Introduction**

Significant volumes of hydrocarbons are contained in the fluvial and estuarine facies deposited within incised valleys (Zaitlin and Shultz, 1990; Barwis, 1990). Examination of Quaternary and modern incised valley fill deposits aids in the identification, prediction, and characterization of ancient equivalent type deposits. This paper describes the seismic facies, lithofacies and facies architecture of the modern Bolivar Roads tidal inlet and associated tidal delta complex as well as tidal delta complexes preserved within the Trinity/Sabine incised valley.

**Study Area**

The east Texas coast is characterized by fair weather astronomical tides ranging from 45 to 60 cm and relatively low amplitude waves with periods commonly within 4-6 s (Morton and McGowen, 1980). Fair-weather wind and storm generated currents on the east Texas shelf generally flow along-shore, to the southwest (Morton, 1977; Snedden et al., 1988). Ebb-current velocities can be 33-50% greater than flood-current velocities through Bolivar Roads tidal inlet
due to a large ebb dominated tidal prism and the frequent passages of fronts that push water out of the bay (Hall, 1976; Eyer, 1984).

Bolivar Roads tidal inlet is 3 km wide making it the largest tidal inlet on the Texas coast. The inlet and associated tidal delta complex has been significantly impacted by human occupation. Historically, the tidal inlet had a maximum depth of 16m (Mason, 1981). Jetty construction initially reduced this depth to a maximum of 7 m creating the need for dredging, which currently maintains a minimum depth of 11 m (Mason, 1981). Pelican Island is a natural emergent part of the flood-tidal delta; however, it has been significantly enlarged with dredge spoil. Eyer (1984) and Siringan and Anderson (1993) demonstrated from bathymetric charts that the natural (prior to jetty construction and dredging) morphology of the tidal inlet/delta complex had well-developed ebb-tidal and flood-tidal deltas of comparable areal extent.

_Late Pleistocene to Holocene History and General Stratigraphy_

The evolution of Bolivar Roads and Galveston Bay were directly related to a rise in sea level which flooded the Trinity River incised valley (Fig. 3-1). Therefore, it is necessary to present a brief overview of the response of depositional systems in the study area to the rise of sea level during the last 18,000 years. More detailed descriptions of the response of east Texas coastal systems to this transgression can be found in Siringan and Anderson (1993), Thomas and Anderson, (1994), and Rodriguez and Anderson (1999). During
Figure 3-1. Bathymetric map of the east Texas shelf showing the locations of seismic lines, cores, and borings. The Trinity/Sabine incised river valley (gray, Thomas and Anderson, 1994) and tidal delta complex deposits (striped) preserved within it are also shown. Segments of seismic data (bold lines) refer to seismic profiles shown in subsequent figures.
the Stage 2 sea-level lowstand, the Trinity and Sabine rivers incised into older
highstand fluvial-deltaic deposits on the continental shelf; these valleys
converge approximately 50 km offshore (Fig. 3-1). Beneath Bolivar Roads, the
Trinity incised valley thalweg lies approximately 55 m below present sea level
(Siringan and Anderson, 1993) while on the inner shelf, incision is 35-40 m
below the sea floor (Thomas and Anderson, 1994). During the subsequent rise
of sea level, the incised valleys backfilled with fluvial, estuarine, and coastal
(flood- and ebb-tidal delta) deposits. Seismic and lithologic data used to map
lithofacies within the incised valley indicates incomplete valley-fill sequences
bounded by intermediate flooding surfaces (Thomas and Anderson, 1994). The
valley-fill facies architecture is characterized by backstepping parasequences
that have been attributed to an episodic or step-like rise of sea level (Anderson
et al., 1991; Thomas and Anderson, 1994). Periods of relatively slow sea-level
rise are manifested by tidal delta deposits. Three discontinuous tidal delta
complexes have been mapped within the valley (Fig. 3-1).

The modern Galveston Bay system initially formed around 4,000 yr. BP
(Bernard et al., 1970; Anderson et al., 1991; Smyth, 1991). Bolivar Roads tidal
inlet initially formed around 3,000 yr. BP (Siringan and Anderson, 1993),
although at this time it was much wider than at present. As the inlet became
narrower due to westward accretion of Bolivar Peninsula, it incised more deeply
into the softer estuarine valley fill sediments and became stabilized in stiff
Pleistocene clays located along the flanks of the valley.
Methods

Approximately 200 km of high-resolution seismic (chirper) data, and 24 sediment cores were collected aboard the R/V Lone Star from Galveston Bay and the east Texas shelf (Figs. 3-1, 3-2). A Datasonics CAP 6000 Chirper system using a 10 ms linear sweep from 1.5-7 kHz was used to collect the seismic data. Seismic data were processed using ProMAX software. The processing consisted of applying automatic gain control (3 ms operator length), and trace mixing (over 7 traces) and in some cases a bandpass filter (500-800-2000-4000 Hz). Sediment cores provide ground truth for the seismic data. They were collected using a pneumatic hammer coring device capable of recovering cores up to 5 m in length.

Results

Modern Tidal Delta Complex and Adjacent Environments

The Bolivar Roads tidal delta complex consists of two general sedimentary facies; a proximal facies located closest to the inlet and a distal facies located seaward and bayward of the inlet. Each of these facies have a distinct seismic character. Middle bay facies, upper shoreface facies, and lower shoreface facies are also important due to their proximity to the tidal delta complex.
Figure 3-2. Bathymetric map of Bolivar Roads tidal inlet and tidal deltas showing the locations of seismic lines and cores. Segments of seismic data (bold lines) refer to seismic profiles shown in subsequent figures.
Tidal Delta Complex Facies.—The proximal facies of the tidal delta complex is composed of laminated to massive well sorted (mean grain size 3.0 phi) sands (Figs. 3-3a, 3-4a). Some cores collected from the flood-tidal delta penetrated nearly a meter of densely packed granule-size to pebble-size shells and shell fragments including a diverse molluscan fauna, bryozoans, and corals (Astrangia). The seismic facies associated with this unit is characterized by a cut and fill pattern in both strike and dip lines (Fig. 3-5a, 3-5b). Reflectors associated with this facies are of high amplitude. Figure 3-6a shows the subsurface distribution of this facies. Cores indicate that this facies is approximately 3 m thick.

The distal facies is composed of interbedded sand and mud couplets (Fig. 3-3b). The sand layers are 0.5-2.0 cm thick and are well sorted having a mean grain size of 3.5 phi (Fig. 3-4a). The seismic facies associated with this unit is characterized by gently dipping (0.4 degrees) parallel laminations; dip direction is toward the northwest in dip lines, and toward the southwest and northeast in strike lines (Fig. 5a, 5b). Cores indicate that this facies is approximately 3.5 m thick.

No ebb-tidal delta sediments were sampled offshore Bolivar Peninsula on the eastern side of north jetty (Fig. 3-6a). Here, marine bioturbated mud with centimeter thick shelly sand beds throughout lie directly on Pleistocene
Figure 3-3: Representative core photographs that characterize the sedimentary units that make up the tidal deltas.
Figure 3-4. Grain size distributions for distal ebb-tidal delta (3-4a, black line) proximal ebb-tidal delta (3-4a, gray line), upper shoreface (3-4b, Black line), and lower shoreface (3-4b, gray line) sands.
Figure 3-5. Interpreted seismic (chirper) profiles through Bolivar Roads flood-tidal delta. Figure 3-2 shows locations of profiles.
Figure 3-6. Map showing the distribution of the sandy proximal tidal delta facies (3-6a) and locations of cross-sections A-A' through the flood-tidal delta (3-6b) and B-B' through the ebb-tidal delta (3-6c). Cross-sections were constructed from core and seismic data.
(Beaumont) stiff clay (Fig. 3-3c). Shell fragments predominately include *Mulinia, Anadara,* and *Oliva.*

**Middle Bay Facies.**---Middle bay deposits consist of dark-greenish-gray, slightly to moderately bioturbated muds with few sand laminae. *Crassostrea* and *Mulinia* shells are common. The seismic facies associated with this unit is characterized by parallel reflectors. Bay muds onlap and drape the flood-tidal delta facies (Fig. 3-5a, 3-5b).

**Upper and Lower Shoreface Facies.**---The shoreface sediments of Bolivar Peninsula and Galveston Island were studied in detail by Siringan and Anderson (1994). The upper shoreface consists mainly of medium (10-30 cm) to thick (30-100 cm) sand beds with scattered shell fragments (Siringan and Anderson, 1994). The lower shoreface consists mainly of slightly to moderately bioturbated interbedded sand and mud. Sand layers are thin-bedded to medium-bedded (10-30 cm, Siringan and Anderson, 1994). Thin shell beds comprised dominantly of *Mulinia* are common. Upper shoreface and lower shoreface sands both have a mean grain size of 3.5 phi (Fig. 3-4b).

*Holocene Tidal Delta Complex Deposits*

Following the study by Thomas and Anderson (1994), a seismic line was collected along the axis of the Trinity/Sabine incised valley in an attempt to better image the seismic facies and bounding surfaces of the valley fill. Three discontinuous tidal delta complexes were recognized, each separated by
middle bay deposits. The tidal delta complex seismic facies is characterized by northeast and southwest gently dipping (0.3 degrees) reflectors (Fig. 3-7a). The middle-bay seismic facies is characterized by parallel reflectors (Fig. 3-7b). Lithologic control within the valley is in the form of boring descriptions which are vague and scarce. Borings 4 and 8 both penetrated middle-bay and tidal delta complex seismic facies (Fig. 3-7a). The middle-bay deposits are clay while the tidal delta complex deposits are silty to sandy clay. Note the absence of tidal delta complex seismic facies in that portion of the valley shown in figure 3-7b, which illustrates the discontinuous distribution of tidal deposits within the valley.

**Discussion and Conclusions**

It can be difficult to distinguish between shoreface deposits and ebb-tidal delta deposits due to their similar lithologies, facies architecture, and proximity. Yet, proper identification of these deposits is important because of their vastly different shapes and orientations. Tidal complex facies are elongate in a dip direction and confined to the incised valley. Shoreface deposits occur along the flanks of the valley and may be sheet-like or shore-parallel elongate outcrops if the rate of sea-level rise and fall is episodic.

Grain size for individual sand layers can be a powerful tool in distinguishing tidal and shoreface deposits. Upper and lower shoreface sands have similar grain size (very fine sand, Fig. 3-4b), while proximal ebb-tidal delta sands are
Figure 3-7. Seismic (chirper) profiles through the Trinity/Sabine incised valley showing the tidal delta complexes (3-7a) and middle bay deposits (3-7b) preserved within the valley. Figure 3-1 shows locations of profiles.
coarser (fine sand) than distal ebb-tidal delta sands (very fine sand, Fig. 3-4a). This is probably due to the difference between tide and storm currents; which are the primary mechanisms of offshore sand transport in these two environments. Tidal currents decrease in velocity away from the tidal inlet, from the proximal facies to the distal facies. In contrast, storm generated currents maintain their strength throughout the upper and lower shoreface. One of the most diagnostic features of the tidal delta complex is couplets of mud and fine sand.

Cross-section A-A' through Bolivar Roads flood-tidal delta (Fig. 3-6) shows that this environment was twice as large in the past. Westward accreting and northward prograding proximal facies were mapped beneath the distal facies 4.5 km bayward from where the proximal facies is currently at the sea-floor (Fig. 3-5b, 3-6b). This marks the maximum extent of the flood-tidal delta before the inlet began to narrow around 3,000 yr. BP (Siringan and Anderson, 1993). As the inlet narrowed the proximal and distal facies initially retreated towards the inlet. Once the modern inlet was stabilized by incision into the Trinity incised fluvial valley, northward progradation of the proximal flood tidal delta began. This more recent episode of flood tidal delta development has been characterized by delta lobe shifting, as indicated by granule- to pebble-sized shell units representative of old channels located within the modern sandy proximal facies. Currently, the proximal facies is being buried in distal facies and bay mud (Fig 3-6b). This could be due to decreased sediment supply to
this flood tidal delta lobe as it initially responds to a lobe shifting event, or a change in the tidal regime of the area, probably related to channel and jetty construction.

Cross-section B-B' through Bolivar Roads ebb-tidal delta indicates that this environment was also larger in the past (Fig 3-6c). Proximal facies underlie distal facies, marking the maximum seaward extent of the ebb-tidal delta prior to when the inlet narrowed. As the inlet narrowed the ebb-tidal delta retreated and distal facies stepped landward over proximal facies. Currently the ebb-tidal delta is prograding. The most recent episode of ebb tidal delta growth is associated with seaward growth of the beach and shoreface on the western (leeward) side of the south jetty. The absence of ebb-tidal delta deposits on the eastern side of the north jetty probably results from accelerated erosion of these deposits following jetty construction. Navigation charts of the area that predate jetty construction show an ebb-tidal delta in the area where some of our cores penetrated marine and shoreface muds resting directly on Pleistocene deposits.

The seismic facies and lithologic descriptions of the Holocene tidal delta complexes within the Trinity/Sabine incised valley resemble the modern Bolivar Roads distal tidal delta environments. Both are characterized by gently dipping reflectors (0.3-0.4 degrees) and they are mud dominated. The absence of proximal tidal delta facies within the incised valley is due to erosion during transgression; shoreface and tidal ravinement have removed those deposits
that occupied the upper portion of the valley fill succession, including proximal tidal delta deposits.

The poor preservation potential of proximal tidal facies is due to the fact that offshore segments of the Trinity incised valley were subject to continuous and rapid transgression. Rates of sea-level rise during the interval between 18,000 and 4,000 yr. BP have been much faster than for most of geologic time. Recall that the modern Bolivar tidal delta complex evolved during the last 3,000 years, after the rate of post-glacial sea-level rise slowed substantially. Therefore the modern Bolivar Roads tidal delta complex provides a better analog for ancient tidal deltas. However, the observed backstepping or detached distribution of tidal deposits within the Trinity valley may be typical of many ancient valley fill successions where the rate of transgression and/or sediment supply to the valley has been episodic.
CHAPTER 4

IMPACT OF HOLOCENE SEA-LEVEL RISE ON COASTAL SYSTEMS: IMPLICATIONS FOR THE FUTURE OF OUR COASTS

Summary

Sedimentologic and seismic data from Gulf of Mexico bays and the east Texas continental shelf indicate that two episodes of rapid shoreline retreat occurred during the Holocene in response to sea-level rise. Flooding surfaces at ~10 m and ~14 m were recognized and mapped regionally. Maps of the paleoenvironments bound by these flooding surfaces indicate that each event resulted in estuarine environments being shifted tens of kilometers landward and barrier shorelines being submerged and stranded on the continental shelf as banks. These destructive events were followed by periods of environmental growth and stability. The threshold at which coastal systems respond catastrophically to sea-level rise is currently unknown, however is important in light of recent indications that sea level is again starting to rise more rapidly.

Introduction
One of the key issues raised in the 1990 Intergovernmental Panel on Climate Change Report concerns the potential impact of global warming on sea level, specifically the magnitude and rate of sea-level rise that will occur over the next century. There is little question that the rate of sea-level rise will increase, and indeed has already begun to increase based on long-term tide gauge records (Hicks, 1978; Gornitz and Lebedeff, 1987; Douglas, 1991) and radiocarbon dating of salt marsh deposits (Varekamp and Thomas, 1998). It is also clear that an increased rate of sea-level rise will have a number of adverse impacts on world coasts, such as wetlands loss and accelerated coastal erosion, but the magnitude of these changes remains unpredictable (Warrick et al., 1996).

Modern coasts and estuaries evolved during an interval of relatively slow sea-level rise (15-25 cm/century) that spanned the past 4,000 years. Prior to this time, sea level was rising at a faster rate (average 50 cm/century) in response to ice sheet melting in both hemispheres. Early coastal inhabitants were conditioned to rapidly retreating shorelines and associated changes in coastal environments. During the past 10,000 years, the east Texas coast has retreated landward approximately 50 kilometers, or at an average rate of 5m/yr. This is five times the current rate of coastal retreat. The Louisiana coast experienced even greater changes during this time, with substantial loss to coastal wetlands (Penland et al., 1988). The Louisiana and Texas coasts are
the most vulnerable coasts in North America to accelerated sea-level rise, due
to high rates of subsidence (locally in excess of 100 cm/century).

The predicted rates of sea-level rise from global warming by the year
2,100 range from 20-23 cm/century to 86-96 cm/century (Warrick et al., 1996;
Fig. 4-1), with the main uncertainty in this prediction being the stability of polar
ice sheets, particularly the West Antarctic Ice Sheet. A more conservative
estimate of 49-55 cm/century (Warrick et al., 1996; Fig. 4-1) is close to the
average rate of sea-level rise during the Holocene (10,000 BP to Present). This
being the case, we should be able to better predict coastal response to
accelerated sea-level rise by examining the geological record of coastal
change during the Holocene. This is not a simple task because the shoreface,
the area located just seaward of the beach, erodes deeply into the coast as it
retreats, thus removing regressive shoreline deposits. On the east Texas coast,
for example, the depth of shoreface erosion is in the range of 6 to 9 meters,
which is below the level of most modern coastal barriers (Siringan and
Anderson, 1994). The exception to this is those areas that were formerly
occupied by river valleys. During the last lowstand in sea level, most rivers
eroded deep valleys on the continental shelves in response to falling base
level. As sea level rose, these incised fluvial valleys were flooded to create
bays, and these bays should contain one of the best records of coastal
response to rising sea level (Belknap and Kraft, 1981). The record of sea-level
rise is recorded as a stratigraphic succession of fluvial and estuarine sediments.
Figure 4-1. High, middle, and low projections of global sea level rise over the period 1990 to 2100 (after Warrick et al., 1996).
If tidal and wave energy in the bay are low, erosion of this sedimentary record should be minimal. However, not all bays are well suited for analysis of sea level change. The best stratigraphic records of coastal flooding will occur in bays where the sediment supply to the river and bay was high enough to keep pace with the rate of sea-level rise but not so high as to completely fill the valley with fluvial deposits.

We report here the results of an experiment conducted in the incised fluvial valleys of the northern Gulf of Mexico, from Apalachicola Florida to the Rio Grande. Our results, and the results of previous workers in the region, show that rivers with relatively high sediment supply, such as the Mississippi, Brazos, and Rio Grande rivers, filled their incised valleys predominately with fluvial sediments and that these rivers occupied more than one fluvial valley during the late Pleistocene-Holocene transgression. Hence, these river valleys are poorly suited for this work. However, other river valleys in the region, specifically those now occupied by Mobile Bay, Lake Calcasieu, Sabine Lake, Galveston Bay, and Corpus Christie Bay (Fig. 4-2) are characterized by stratigraphic sequences showing superposition of more open marine (distal) deposits over coastal, estuarine, and fluvial deposits. The contacts between these deposits are typically quite abrupt, suggesting that landward shifts in coastal environments and their associated ecosystems occurred rapidly. We present mainly the results of a study of the ancestral Trinity River Valley, now occupied by Galveston Bay, which bears a particularly good record of coastal response to
Figure 4-2. Map of the Gulf Coast showing the locations of Corpus Christie Bay, Galveston Bay, Sabine Lake, Lake Calcasieu, and Mobile Bay. Contours (in gray) are mean annual precipitation, which varies significantly across the region.
sea-level rise. We will demonstrate that the retreating coastal environments that have occupied this valley, which include the river, its bay-head delta, the middle bay environment, and the lower bay tidally influenced environment have retreated landward in a series of abrupt shifts of many kilometers. We will also present preliminary data from other bays of the northern Gulf of Mexico that indicate similar flooding histories and data from offshore east Texas which indicates abrupt drowning and overstepping of barrier islands.

Results and Discussion

A compilation of over 4000 km of high-resolution seismic data and 300 sediment cores from Rehkemper (1969), Thomas (1990), Smyth (1991), Siringan (1993), and Rodriguez et al. (1999) were used to map the coastal environments that fill the Trinity and Sabine incised valleys on the east Texas shelf (Fig. 4-3). As sea-level rose during the last deglaciation, the Trinity/Sabine incised valley backfilled with continuous fluvial and bay-head delta sediments and discontinuous middle bay, and coastal (tidal delta complex facies) deposits (Fig. 4-4). Thomas and Anderson (1994), identified four packages of sediments bound by contacts (flooding surfaces) separating proximal coastal environments (bay-head delta and fluvial environments) from overlying distal coastal environments (middle bay and tidal delta environments) within the incised valley. These contacts were mapped as flat surfaces
Figure 4-3. Bathymetric map of the east Texas shelf, and map of Galveston Bay indicating the locations of cross-sections shown in subsequent figures. The Trinity/Sabine incised river valley (gray, Thomas and Anderson, 1994; Rehkemper, 1969; Smyth, 1991) and tidal delta complex deposits (striped) preserved within it are also shown.
Figure 4-4. Cross-section A-A' showing the discontinuous middle bay and tidal delta complex environments within the Trinity/Sabine incised valley. Figure 4-3 shows location of cross-section.
interpreted to represent episodes of rapid environmental change (Thomas and Anderson, 1994). The packages of sediments bound by these flooding surfaces were interpreted to represent periods of environmental growth and stability (Thomas and Anderson, 1994). The last two episodes of rapid environmental change recognized within Galveston Bay and the Trinity/Sabine incised valley were studied in detail as part of this investigation.

Each environment identified within the Trinity/Sabine incised valley is characterized by a distinct seismic, lithologic, and paleontologic facies. Fluvial deposits are not well imaged on seismic profiles, although the data wipe-out zone beneath the incised valleys can be used as an aid in their identification. Geotechnical borings and sediment cores indicate that fluvial facies fine upwards from fine gravel to fine sand (Thomas and Anderson, 1994). The bayline surface is the contact between fluvial deposits and bay-head delta deposits and is imaged as a continuous, high amplitude reflector (Thomas and Anderson, 1994). The top of the bay-head delta is imaged as an irregular very high amplitude reflector (Thomas and Anderson, 1994; Rodriguez et al., 1998; Fig. 4-4). Cores that penetrated bay-head deltas sampled a muddy sand unit with abundant organic material, peat layers, and Rangia shells. Rangia currently lives along the bay shore and is most abundant near the Trinity bayhead delta. The elevation of the bay line is a sea level marker ± one to two meters. The top of the bay-head delta surface was mapped regionally within the incised valleys and is characterized by a series of flat steps and inclined
rizers (Fig. 4-4). These features have also been recognized in the lower bayline surface (Thomas and Anderson, 1994; Fig. 4-5).

Stratigraphically above the bay-head delta, middle bay and ebb/flood tidal delta facies are discontinuous; these facies alternate up the axis of the incised valley (Fig. 4-4). The middle bay facies is characterized by parallel, flat lying seismic reflectors (Fig. 4-4) and is comprised of a homogeneous mud unit with oyster (Crassostrea) reefs in places (Rodriguez et al., 1998). The ebb/flood-tidal delta deposits are characterized by gently seaward dipping and landward dipping reflectors respectively (Fig. 4-4). Sediment cores from this facies sampled sandy mud, shelly sand, and distinctive thin interbedded sand and mud (Siringan and Anderson, 1993; Rodriguez et al., 1998).

In places on the inner shelf, banks (Sabine, Heald, and Shepard banks) lie adjacent to and over the incised valley (Fig. 4-3). Sabine and Heald banks were first interpreted as reworked paleoshoreline deposits by Nelson and Bray (1970). Detailed work by Rodriguez et al. (1999) indicate that these banks represent submerged paleoshorelines composed of three facies (from bottom to top): (1) a back barrier estuarine facies characterized by landward dipping seismic reflectors and consisting of interbedded sand and mud unit; (2) a fore-barrier, lower shoreface/ebb-tidal delta facies characterized by seaward prograding to chaotic seismic reflectors and consisting of a muddy-sand unit; and (3) a storm reworked facies characterized by a chaotic to acoustically
Figure 4-5 Cross-section B-B' through Galveston bay based on seismic and lithologic data. Locations of two seismic examples (shown below) are indicated in gray on the cross-section. Flooding surfaces are outlined on the representative seismic examples.
reverberating seismic reflection pattern consisting of a interbedded shell hash and sand unit (Fig. 4-6).

The observed risers and landward shifts in the bayline flooding surface and the discontinuous distribution of middle bay, tidal delta complexes, and shoreline deposits within the incised valley indicate episodes of dramatic environmental change. The landward shifts in the bayline flooding surface represent periods of landward translation of coastal environments (Figs. 4-7 and 4-8). The risers represent periods of aggradation and therefore, adjustment of the fluvial and bay-head delta environments to a higher sea level position (Fig. 4-8). Two prominent flooding surfaces exist within Galveston Bay, at around -10 m and -14 m (Figs. 4-5 and 4-6). The deposition of middle bay, tidal delta and shoreline deposits, i.e. the formation of a bay, indicate environmental stability, while the overstepping and stranding of these bay and coastal barrier facies on the continental shelf represent periods of massive environmental change (Fig. 4-8). Sabine and Heald banks currently exist at water depths of -12 m and -15 m respectively (Figs. 4-3 and 4-6). The shoreface/ebb-tidal delta facies was the only portion of the barrier preserved; the beach and all other shallower environments were eroded and reworked into the upper storm influenced facies. The depths of the ravinement surfaces that separates shoreface/ebb-tidal delta facies from the storm influenced facies are --10 m for Sabine Bank and --14 m for Heald Bank (Fig. 4-6,) matching the depths of the two flooding surfaces within Galveston Bay (Fig. 4-5).
Figure 4-6. Cross sections C-C’ through Sabine Bank and D-D’ through Heald Bank. The ~10 m (Sabine Bank) and ~14 m (Heald Bank) flooding surfaces are indicated by dashed lines.
Figure 4-7. Structure contour map of the top of the bay-head delta flooding surface showing the location of the inclined riser within Galveston Bay.
Figure 4-8. Cartoon showing formation of the flat step and inclined riser identified within Galveston Bay, and the linkage between the flooding surfaces within the bay and those observed in the offshore banks. Between time 1 and time 2, sea-level rose drowning the barrier shoreline and the bay-head delta. At this time the flat step is formed in the top of the bay-head delta flooding surface. Because the barrier shoreline was submerged at the same time as the bay-head delta, the offshore bank will contain a ravinement surface at about the same depth as the flooding surface within the bay. Between time 2 and time 3 there is environmental stability. During this time the bay-head delta progrades and aggrades forming the inclined riser, a new barrier shoreline forms, and the old barrier shoreline continues to be reworked offshore by storm waves.
Radiometric and AMS age dates from the shoreface/ebb-tidal delta deposits (below the ravinement surface) of Sabine and Heald banks were measured to be $4655 \pm 107$ cal. BP and $8325 \pm 38$ cal. BP respectively. These ages closely resemble radiometric ages from upper bay deposits of $4555 \pm 122$ cal. BP and bay-head delta deposits of $8371 \pm 315$ cal. BP (Rehkemper, 1969) sampled above the $\sim 10$ m and below the $\sim 14$ m flooding surfaces within the bay. Thus, the linkage between the offshore and onshore flooding events is reasonably well constrained. Age dates below the $\sim 14$ m flooding surface within the bay and Heald Bank are similar, indicating that both the shoreline and the bay-head delta stepped landward simultaneously. During the later ($\sim 10$ m) flooding event, the barrier shoreline and bay-head delta did not step landward simultaneously. Because the age date above the $\sim 10$ m flooding surface within the bay is similar to the age date below the $\sim 10$ m flooding surface within Sabine Bank, the landward shift in the bay-head delta must have occurred prior to barrier shoreline submergence. Hence, Sabine Bank remained as a barrier island, similar to modern Galveston Island, separated from the new paleoshoreline by a distance of over 30 km. The bay-head delta shifted landward a similar distance. This landward shift in coastal environments would have transformed the low-salinity upper bay ecosystem into a higher-salinity middle bay, and the lower-bay ecosystem would have been inundated with marine waters and transformed into an open marine environment. The question of how rapid did these environmental changes take place still remains,
and the answer is important for determining how ecologically destructive this event was.

Conclusions

The most likely forcing mechanism for shifting the bay and shoreline landward such great distances is sea-level rise. These landward shifts were followed by intervals of aggradation, upward steps in the bayline, and associated increase in sediment accommodation space. This increase in accommodation space is not likely to have been created by a sudden increase in coastal subsidence because the subsidence rate for this area is too low (0.1 mm year\(^{-1}\)) (Paine, 1993). Furthermore, a decrease in the rivers sediment supply could not have caused submergence of the ancestral Heald and Sabine coastal barriers because river discharge occurred 30 km landward of these barriers. The barriers were predominately nourished by sands derived from offshore and along-shore, similar to Galveston Island whose fate by “drowning” seems inevitable.

If sea-level rise truly is the forcing mechanism behind the observed environmental changes in Galveston Bay, then flooding surfaces with approximately the same elevation and age should be present in other bays. We have imaged the -14 meter flooding surface within Corpus Christie Bay, Texas (Fig. 4-9). Anderson et al. (1991) imaged the -10 meter flooding surface within
Figure 4-9. Map showing location of the Nueces incised valley within Corpus Christie Bay (after Wright, 1980). The seismic section indicates the presence of a -14 m flooding surface within Corpus Christie Bay. Buried oyster reefs are testament to the impact of flooding events on coastal ecosystems.
Sabine Lake, and the sediments just above this surface were dated at 4,415 ± 362 cal. BP (Nelson and Bray, 1970; Fig. 4-10). LeBlanc (1949) and Nichol et al. (1994) recognized the -14 m flooding surface from a transect of cores through the Calcasieu incised valley, Louisiana. We have also imaged the -10 meter flooding surface within Mobile Bay, and the sediments near this surface were dated at between 3,690 ± 130 BP and 5,560 ±130 BP (Mars, 1992; Fig. 4-11).

Did the observed flooding events in these bays take place in response to rapid sea-level rises, or is there a threshold at which coastal systems respond catastrophically to a continuous rate of sea-level rise? If flooding occurred in response to an increase in the rate of sea-level rise, how dramatic was this increase? Holocene sea level curves based on coral age depth data do not show rapid sea-level rise events after 8,000 yr. B.P. (Lighty et al., 1982; Fairbanks, 1989; Bard et al., 1990; Montaggioni et al., 1997; Toscano and Lundberg, 1998; Fig. 4-12). But this simply tells us that the magnitude of sea-level rise events needed to profoundly impact coastal systems is probably below the resolution of these curves (± 5 m). The observed steps in the bayline flooding surface in Galveston Bay are in the range of 3 to 5 meters, below the resolution of modern sea level curves.

The modern Galveston Bay system formed around 3,300 yr. BP (Siringan and Anderson, 1993) and at this time sea level was rising slower than at any other time during the Holocene (1-2 mm/year). If the rate of sea-level rise does
Figure 4-10. Map showing location of the Sabine incised valley. The seismic section indicates the presence of a -10 m flooding surface within Sabine Lake (modified after Anderson et al., 1991).
Figure 4-11. Map showing location of the Mobile incised valley within Mobile bay (after Kindinger et al., 1994). The seismic section indicates the presence of a -10 m flooding surface within Mobile Bay.
Figure 4-12. Sea level curve for the last 18,000 years based on coral age-depth data from Barbados, Tahiti, and New Guinea (after Bard et al., 1996).
revert back to more typical Holocene rates (4-5 mm/yr), Gulf of Mexico coastal environments and ecosystems may again experience a rapid landward shift. Ongoing research is aimed at trying to constrain the timing and magnitude of flooding events of the past 10,000 years and at better documenting the environmental and ecological impacts of these events.
CHAPTER 5

POSITION OF THE MIDDLE WISCONSIN SHORELINE ON THE
TEXAS INNER CONTINENTAL SHELF

Summary

High-resolution seismic, lithologic, paleontologic, and radiocarbon data from the east Texas continental shelf were used to map the position of a middle Wisconsin (oxygen isotope stage 3) shoreline. The shoreline was placed in regional context with a study of the Texas shelf, based on seismic profiles, core, and platform boring descriptions that identifies the stage 3 maximum flooding surface and the stage 2 sequence boundary. These surfaces were then traced up-dip on the east Texas shelf where a shore parallel escarpment was recognized. Landward of the escarpment the stage 3 maximum flooding surface is amalgamated with the stage 2 sequence boundary. Therefore, the location of the escarpment marks the up-dip limit of stage 3 deposits and the stage 3 shoreline. Lagoonal sediments with in-situ brackish water fossils on-lap the escarpment. Seaward of the escarpment, remnants of barrier island facies have been preserved as a bank, Freeport Rocks Bathymetric High, located in -
18 m of water. The shoreline (adjusted for subsidence) is located at -15 m ± 2 m, which is shallower than what oxygen isotope and coral records indicate.

Introduction

Oxygen isotope data from deep sea cores have been widely applied as sea-level proxies (Mix and Ruddiman, 1984; Shackleton, 1987). During isotope stage 3, the SPECMAP curve (Imbrie et al., 1984) and the temperature corrected benthonic oxygen isotope record of East Pacific core V19-30 (Shackleton, 1987) predict sea level was at ~-85 m and ~-75 m respectively. These estimates do not compare well with much of the direct geologic evidence for the position of the middle Wisconsin (stage 3) maximum sea-level shoreline. The types of direct geologic evidence used to constrain the position of the stage 3 highstand of sea level are radiocarbon dates from shallow-water fauna (Curray, 1965; Milliman and Emery, 1968; Blackwelder et al., 1979; Finkelstein and Kearney, 1988); uranium series (²³⁴U-²³⁰Th) dating of reef terraces in Barbados (Mesoella et al., 1969; Steinen et al., 1973; Mathews, 1984; Bard et al., 1990), New Guinea (Bloom et al., 1974; Chappell, 1974, 1983; Chappell et al., 1996; Bloom and Yonekura, 1985), and Indonesia (Chappell and Veeh, 1978 and Bard et al., 1996); and high-resolution seismic stratigraphy (Suter et al., 1987; Thomas and Anderson, 1989, 1991; Wellner et al., 1993; Roy et al.,
1997). This evidence for the position of the stage 3 maximum sea-level highstand is inconsistent, for it has been interpreted as being anywhere from at or above present day levels (Curry, 1965; Milliman and Emery, 1968; Blackwelder et al., 1979; Finkelstein and Kearney, 1988; ) to -85 m (Bard, et al., 1990). However, the Huon Peninsula of New Guinea reef sequence indicates that global sea level was -73 to -55 m around 44 ka (Chappell et al., 1996), which is within the range of estimates from core V19-30 but still deeper than what high-resolution seismic stratigraphic studies indicate.

High-resolution seismic stratigraphic studies of the Louisiana (Suter et al., 1987), New Jersey (Wellner et al., 1993), southeast Australia (Roy et al., 1997) and east Texas (Thomas and Anderson, 1991) continental shelves have revealed a stratigraphic unit resting above a surface interpreted as the stage 3 maximum flooding surface (MFS). The up-dip pinch out of this unit occurs between -20 and -40 meters, and was interpreted as indicating that the stage 3 shoreline was situated within this depth interval (Suter et al., 1987; Thomas and Anderson, 1991; Wellner et al., 1993; Roy et al., 1997). With the exception of Roy et al. (1997) who used the thermoluminesence technique to date a stage 3 prograding shoreline, the other studies lacked the chronostratigraphic data needed to document the age of the inferred stage 3 unit.

More recent high-resolution sequence stratigraphic studies conducted on the Texas continental shelf (Abdulah, 1995; Banfield, 1998; Snow, 1998, Fig. 5-
1) have also led to the recognition of the stage 3 maximum flooding surface and an overlying stage 3 highstand unit. These studies include oxygen isotope data, micropaleontological data, and radiocarbon dates that constrain the ages of stratigraphic units and bounding surfaces. These regional studies formed the chronostratigraphic framework for this investigation.

Stratigraphic Framework

Banfield (1998) conducted a seismic and sequence stratigraphic analysis of Quaternary strata of the south Texas shelf using a grid of high-resolution seismic data, a core, and platform boring descriptions (Fig. 5-1). An oxygen isotope curve generated from core B-2 provided a chronostratigraphic scale for this area of the shelf (Fig. 5-2b). The curve was augmented with micropaleontologic data (Ericson-Wollin Zones), radiocarbon ages, and the position of two prominent meltwater anomalies, one at 42,800 to 44,300 and the other at 60,700 yrs BP. The pulses were previously recognized and dated (tephrachronology) by Williams and Kohl (1986) offshore Louisiana. Core B2 does not contain a continuous record of sedimentation; two erosional unconformities exist at times when the core was subaerially exposed (stage 2 and stage 4 sequence boundaries (SB); Fig. 5-2b). The oxygen isotope curve for core B-2 compares favorably with two oxygen isotope curves derived from
Figure 5-1. Map of the Texas gulf coast indicating the study areas of Banfield (1998; A), Eckles (1996; B), and Abdullah (1995; C). Data examined is shown in gray.
Figure 5-2. Seismic line 5 crosses core B-2 enabling correlation of seismic surfaces (a) to oxygen isotope stages (b) for the south Texas shelf (modified after Banfield, 1998). This enables seismic and lithologic units to be placed within a chronostratigraphic framework. Location of seismic line and core is indicated on figure 5-1.
analysis of *Globogerinoides ruber* from DSDP site 619 in the Pigmy Basin offshore Louisiana (Williams and Kohl, 1986) and core TR-125-23 from the southwestern Gulf of Mexico (Williams, 1984). The chronostratigraphic data from core B-2 are summarized in Figure 5-2b.

Banfield (1998) correlated a prominent downlap surface on the south Texas shelf to the stage 5e MFS of the central (Eckles, 1996) and east (Abdullah, 1995) Texas shelves (Fig. 5-2a). Banfield (1998) recognized that several phases of ancestral Rio Grande delta evolution occurred in the region in response to high frequency sea level fluctuations during the last glacial eustatic cycle and sediment supply variations. One of the largest of these deltas, up to 100 ms thick and approximately 112 km$^3$ in volume, prograded across the top of the stage 3 MFS (Fig. 5-2a). In core B-2, the older meltwater pulse (60,700 yrs BP) occurs approximately one meter above the stage 3 MFS, so progradation of the stage 3 delta is constrained by this age and that of the stage 2 sequence boundary (SB; 22,000 to 18,000 yrs BP), which truncates the top of the delta (Figs. 5-2a, b). The updip pinch out of the stage 3 delta is mapped at ~-40 to -30 m water depth where it is truncated by an erosional surface (stage 2 SB), indicating that the original elevation of the stage 3 shoreline is shallower than ~-40 to -30 m.

Banfield (1998) correlated her mapped surfaces with those identified by Eckles (1996) on the central Texas shelf. Eckles had linked her surfaces with
those mapped by Abdullah (1995) on the east Texas shelf (Fig. 5-1). During the last glacial cycle there was little fluvial sediment input to the central Texas shelf and therefore, no large deltas (Suter and Berryhill, 1985; Eckles, 1996). Sedimentation rates were lower here than on shelves to the east and south. The stage 3 MFS is amalgamated with the stage 2 SB on the majority of the central Texas shelf. Sedimentation rates increase toward the east Texas shelf where the ancestral Colorado and Brazos deltas are located.

Abdullah (1995) acquired a grid of high-resolution seismic data from the east Texas continental shelf (Fig. 5-1) and mapped a series of seaward prograding deltas, the ancestral Brazos and Colorado deltas (Abdullah and Anderson, 1994; Anderson et al., 1996). The oldest of these deltas downlaps a regional surface identified as the stage 5e MFS based on oxygen isotope stratigraphy and micropaleontological data from drill site B 146 (Figs 5-1, 5-3b). The micropaleontological data for this core shows two prominent peaks in the abundance of planktonic foraminifera. The lower peak is associated with the stage 5e MFS and is separated from the upper peak by the extinction of Globorotalia menardii flexuosa. The extinction is dated at ~85,000 yrs. BP (Kennett and Huddleston, 1972, Poag and Valentine, 1976), and marks the stage 5a/5b boundary (Fig. 5-3). A radiometric age date of >45,430 yr. BP (radiocarbon dead) was obtained from shells in the unit above this horizon. The upper peak in planktonic foraminifera abundance is associated with a deltaic
Figure 5-3. Regional seismic line R93-51 crosses core B-2 enabling correlation of seismic surfaces (a) to oxygen isotope stages (b) for the east Texas shelf (modified after Abdullah, 1995). Two peaks in abundance of planktonic foraminifera are associated with the stage 5e maximum flooding surface (MFS) and the stage 3 MFS. Location of seismic line and core is indicated on figure 5-1.
Figure 5-3 continued.
unit that progrades across and downlaps onto a flooding surface that Abdullah interprets as the stage 3 MFS. Abdullah (1995) interpreted this unit as a stage 3 highstand delta and its landward pinch-out occurs in water depths between -30 and -35 meters.

The stage 3 Brazos delta is partially onlapped by a younger Colorado delta (Fig. 5-3a). The Colorado Delta was studied in detail by Snow (1998) using the stratigraphic framework constructed by Abdullah (1995; Fig. 5-4). Two radiometric age dates (38,850 ±1,380 yrs. BP. and 39,320 ±480 yrs. BP) were acquired from laterally equivalent facies (delta front and interdistributary bay) within the up-dip portion of the delta. The similarity of the age dates from samples in different deltaic sub-environments but the same stratigraphic level further supports Abdullah’s (1995) identification of this deltaic unit as a stage 3 deposit. This delta lobe was traced landward to a modern water depth of -27 to -21 meters (Snow, 1998; Fig. 5-4). East of this area in an interfluvial portion of the shelf is the location where we chose to carry out a detailed investigation of the stage 3 shoreline (Figs. 5-1, 5-5).

In summary, three regional high-resolution sequence stratigraphic analyses of the Texas shelf (Abdullah, 1995; Banfield, 1998; Snow, 1998) have developed independent chronostratigraphic frameworks based on oxygen isotopes, micropaleontological data, and radiocarbon ages. These frameworks have been correlated using two regional seismic surfaces, the stage 5e MFS
Figure 5-4. Map of the stage 3 Colorado Delta (modified after Snow, 1998). This delta was traced landward to a modern water depth of -27 to -21 meters.
Figure 5-5. Study area map of the east Texas shelf. Data examined is shown in gray. The edge of an escarpment which marks the up-dip pinch out of stage 3 deposits has been mapped in black. Location of map is shown in figure 5-1.
and the stage 2 SB. The southern and eastern studies have demonstrated the existence of a stage 3 MFS and stage 3 deltas (of the ancestral Brazos, Colorado, and Rio Grande) prograding across and downlapping onto this surface. The deltaic deposits extend landward into modern water depths of between -20 and -40 meters, indicating that sea level must have been within this depth range during stage 3.

In order to better constrain the elevation of the stage 3 shoreline, we conducted a focused investigation east of the stage 3 Colorado Delta. The study area is located on an interfluval portion of the east Texas shelf where the paleoshoreline can be projected to the modern sea floor (Fig. 5-5). The following text describes this investigation.

Methods

Approximately 260 km of high-resolution seismic (Chirper) data and 150 sediment cores were collected aboard the R/V Lone Star from the east Texas shelf (Fig. 5-5). A Datasonics CAP 6000 Chirper system using a 10 ms linear sweep from 3 kHz to 7 kHz was used to collect the seismic data. Seismic data were processed using ProMAX software. The processing consisted of applying a bandpass filter (500-800-2000-4000 Hz), automatic gain control (3 ms operator length), and trace mixing (over 7 traces). Sediment cores provide a
direct link between seismic facies and lithofacies. The cores were collected using a pneumatic hammer coring device with a maximum core recovery length of 5 m. The acoustic travel time to depth conversion was made assuming that the velocity of the upper few meters of sediment is the same as that for water (1500 m/s). Radiocarbon dating of shell material sampled from the cores was performed by Beta Analytic Inc. and all dates are reported in radiocarbon years before present (yr. B.P.).

Coleman et al (1989) have argued that age dates approaching the limit of the radiocarbon dating method are problematic. One problem is that a radiocarbon dead sample with 1% contamination by secondary carbonate will produce an age of about 37,000 yr. B.P. (Grootes, 1983), which is near stage 3 time. To avoid this problem, lithologic units were interpreted using the regional chrono-stratigraphic and sequence-stratigraphic framework previously described. This framework is our primary dating method. All radiocarbon age dates obtained from below the stage 2 sequence boundary have been reported in this paper (no radiocarbon dead samples have been excluded) and are used only to support, not form our chronostratigraphy.

Results and Discussion

East Texas Shelf Physiography
Although the bathymetric map of the east Texas shelf (Fig. 5-5) indicates a gentle shelf gradient (0.28 m/km), there are some prominent features. The -10 m bathymetric contour outlines the proximal portion of the Brazos Delta lobe (Fig. 5-5). The lobe is a modern feature that was created in 1929 when the main river channel was diverted. A strike-aligned offshore bank referred to as Freeport Rocks Bathymetric High (FRBH) is outlined by the -18 m isobath (Fig. 5-5). The shelf stratigraphy east of the bank, bounded by the Brazos Delta lobe and Oyster Creek, will be described first using a representative cross section A-A’ (Figs. 5-5, 5-6).

*East of Freeport Rocks Bathymetric High*

**Seismic Data Analysis and Interpretation.**--- Seismic line FRBH-96-16 is a dip-oriented profile that extends from the present shoreface to nearly -20 m water depth (Fig. 5-5). A portion of this line is used to illustrate two seismic facies units (SFU1 and SFU2) that occur in the study area (Fig. 5-6b). The segment displayed in Fig. 5-6b shows a prominent erosional surface (E1) that forms an escarpment sloping offshore. E1 is a planar surface landward and seaward of the scarp throughout the study area. SFU1 is characterized by laminated reflectors that onlap the E1 surface. The unit becomes more chaotic to the southeast. SFU1 is cut by an erosional surface (E2) that amalgamates
Figure 5-6. Cross-section A-A' (a) and a portion of seismic line FRBH-96-16 (b) showing an escarpment cut into the Beaumont Formation clay. This escarpment marks the up-dip pinch out of stage 3 deposits. Location of cross-section is indicated on figure 5-5.
with E1 landward of the scarp. SFU2 is characterized by low-angle landward
dipping reflectors that downlap the E2 surface.

Seismic surfaces recognized in this study were correlated with surfaces
mapped by Abdullah (1995) and Snow (1998); to verify their chrono-
stratigraphic significance. E1 is the stage 3 MFS amalgamated with the stage 4
SB. Landward of the scarp this surface amalgamates with E2. E2 is the
transgressive ravinement surface (TRS) amalgamated with the stage 2 SB.
SFU1 lies below the stage 2 SB and above the stage 3 MFS, indicating
deposition during the stage 3 highstand. SFU2 lies above the stage 2 SB
indicating deposition during the stage 2 to 1 transgression.

**Lithologic Data Analysis and Interpretation.***--- A transect of
cores, A-A' (Figs. 5-5, 5-6), was collected across seismic line FRBH-96-16 to
correlate seismic facies with lithologic units. Core BD-96-2 was collected
landward of the scarp and penetrated sediments corresponding to SFU2. The
lithofacies consists of a Holocene moderate brown mud with a shear strength of
0.05-0.1 kg/cm2 (Fig. 5-7). This unit is underlain by a mottled green and red,
very stiff (shear strength >1.0 kg/cm2) clay of the Beaumont Formation (Figs. 5-6
and 7). The contact between these two units is sharp and erosional (other
cores sampled clay rip ups and shell lag deposits at the contact; Fig. 5-7). All
cores collected landward of the escarpment in the study area sampled
Holocene muds (SFU2) on top of Beaumont clay, separated by an erosional
Figure 5-7. Cores OFI-95-17 and BD-96-2 were collected landward of the escarpment. Each core sampled Holocene sediments (SFU2) overlying clay of the Beaumont Formation (~ 80-125 ka). The contact between these two units is the transgressive ravinement surface and is equivalent to the E1/E2 surface. Core locations are indicated on figure 5-5.
contact (the amalgamated E1/E2 surface). The Beaumont Formation clay (deposited during stage 5e) has been subaerially exposed. Effects of the exposure are the sharp change in sediment shear strength across the E1/E2 surface (Bernard and LeBlanc, 1965; Winker, 1979; VanSiclen, 1991) and the oxidized color. Radiometric analysis of a Crassostrea virginica shell in the Beaumont clay unit sampled from core OFI-95-17 indicates this sample is radiocarbon dead (Fig. 5-7).

Cores BD-96-4 and OBD-95-5 (Fig. 5-8) were collected seaward of the lapout of SFU1 against the escarpment (Fig. 5-6). They penetrated the sediments corresponding to SFU2 and SFU1 and the E2 boundary (the TRS amalgamated with the stage 2 SB). The contact between the Holocene muds of SFU2 and SFU1 is sharp and erosional, indicated by a coarse sand and shell lag deposit (and clay rip ups in other cores) at the boundary. The sediments corresponding to SFU1 consist of medium gray clay with an intermediate shear strength of 0.3-0.65 kg/cm². The fossil assemblage is dominated by small brackish water gastropods (Vioscalba louisianae and Littorina sphinctostoma), clams (RANGIA CUNEATA many still paired), and barnacles (Balanus sp., Fig. 5-8). All growth stages are represented. In addition to the dominant taxa, this unit also contains other brackish water mollusks such as Macoma mitchelli, fragments of Ischadium recurvum, and worn fragments of Crassostrea virginica. In core BD-96-4 a radiometric age date of 42,250 ± 2590 yr. B.P. was obtained
Figure 5-8. Cores BD-96-4 (a) and OBD-95-5 (b) were collected seaward of the escarpment. These cores sampled Holocene sediments (SFU2) overlying middle Wisconsin (stage 3) upper-bay deposits (SFU1). The contact between these two units is the transgressive ravinement surface (TRS) amalgamated with the stage 2 sequence boundary (SB) and is equivalent to the E2 surface. Core locations are indicated on figures 5-5 and 5-6.
Figure 5-8 continued.
from an articulated *Rangia cuneata* shell and an AMS age date of 45,450 ± 920 yr. B.P. was obtained from a *Littorina sphinctostoma* shell (analyzed as aragonite; appendix 2) also from SFU1 (Fig. 5-8a). In core OBD-95-5 a radiometric age date of 37,080 ± 630 yr. B.P. was obtained from an articulated *Rangia cuneata* shell (calcite) in SFU1 (Fig. 5-8b). All cores collected seaward of the scarp in the study area penetrated Holocene muds (SFU2) on top of medium gray clay (SFU1) with an erosional contact separating these two units (the E2 surface); no cores sampled Beaumont Formation clay or the E1 boundary.

Seismic data constrains SFU1 to within the time range of the stage 3 highstand. The radiometric age dates of 37,080 ± 630 yr. B.P. and 42,250 ± 2590 yr. B.P. from articulated *Rangia cuneata* shells and the AMS age date of 45,450 ± 920 yr. B.P. from *Littorina sphinctostoma* shells indicate SFU1 was deposited during stage 3 time. The scarp marks the up-dip pinch out of stage 3 deposits.

*Freeport Rocks Bathymetric High*

Bartek et al. (1991) interpreted FRBH (Fig. 5-5) as an overstepped transgressive Brazos Delta lobe. This conclusion was based only on seismic data, because no cores had been collected in the area. The sediments of the Brazos River have a distinctive red color and are characterized by fine grain
sizes; most of the sediment load is clay-sized. These sediments are derived primarily from Triassic red beds located in the upper reaches of the drainage basin in northwestern Texas and northeastern New Mexico. Cores from transect B-B' collected over FRBH (Figs. 5-5, 5-9) did not sample any Brazos river sediments above the stage 2 SB, indicating that FRBH is not an overstepped transgressive Brazos Delta lobe. The internal stratification of the bank was not successfully imaged with seismic data; therefore, lithologic data were used to identify surfaces and bank facies which were then correlated to the rest of the study area.

Lithologic Data Analysis and Interpretation.--- FRBH is situated just seaward of the previously discussed scarp. Core FRBH-25, collected landward of the scarp, penetrated Holocene bioturbated sandy mud (SFU2) over Beaumont Formation clay (Fig. 5-9). The surface separating these units is erosional and correlates with the amalgamated E1/E2 surface from Figure 5-6. Cores FRBH-26 and FRBH-27, collected from the bank top, penetrated Holocene bioturbated sandy mud over a well sorted fine grained (2.5 phi) quartz sand with spherical frosted grains and scattered roots and wood fragments (Fig. 5-9). In places an organic-rich clay is found below the Holocene mud and above the sand (Fig. 5-9). The sharp contact between the Holocene bioturbated sandy mud and the well sorted sand units is the TRS amalgamated with the stage 2 SB (the E2 surface). Seaward of the bank, cores FRBH-28 and
Figure 5-9. Cross-section B-B' through Freeport Rocks Bathymetric High (FRBH). FRBH has been interpreted as a middle Wisconsin barrier island. Location of cross-section is indicated on figure 5-5.
FRBH-29 penetrated Holocene bioturbated sandy mud over a medium dark gray clay unit (shear strength 0.3-0.65 kg/cm²) with sand layers and laminations (Fig. 5-9). The sharp erosional contact between these two units also represents the TRS amalgamated with the stage 2 SB (the E2 surface). Core CCBH-1 was collected west of cross-section B-B’, landward of the bank, and sampled Holocene olive gray sandy mud overlying a grayish yellow green to olive gray clay (Fig. 5-10). The contact between the two units is sharp and erosional representing the E2 surface. An oyster reef containing a monospecific assemblage of *Crassostrea virginica* (many still paired) was sampled between 143-197 cm in the grayish yellow green mud unit. A radiometric age date of 39,420 ± 820 yrs. B.P. was measured from an articulated *Crassostrea virginica* shell within this reef.

The amalgamated TRS and stage 2 SB (E2 surface) is not a flat surface; it mimics the bathymetric expression of the bank. This conformance with topography indicates that the bank is an older feature (pre Holocene) that was later covered with Holocene sandy mud. The E2 surface and the scarp have been correlated with cross section A-A’, and Snow’s (1998) Colorado dataset. The radiometric age date of 39,420 ± 820 yrs. B.P. from an articulated *Crassostrea virginica* shell sampled from core CCBH-1, and the stratigraphic position of the sediments located seaward of the scarp, including FRBH indicate that these are stage 3 highstand deposits.
Figure 5-10. Core CCBH-1 was collected landward of FRBH and sampled Holocene sediments (SFU 2) overlying middle Wisconsin (stage 3) middle-bay deposits (SFU 1). Location of core is indicated on figure 5-5.
A regionally extensive scarp was mapped across the east Texas shelf. The sediments located below the stage 2 SB seaward of the scarp represent the up-dip pinch out of stage 3 deposits. The age of these deposits were constrained primarily by their stratigraphic position. The four radiocarbon age dates acquired from this unit fall between 37,000 and 45,000 yrs. B.P.; no samples analyzed produced radiocarbon dead signatures (as determined by Beta Analytic Inc.). Landward of the scarp an age date from an older unit (Beaumont Formation clay) was radiocarbon dead (>37,870 yr. BP). All age dates acquired support the stratigraphy of the area. Before the depth of the stage 3 shoreline can be calculated the environment of deposition of stage 3 sediments must be constrained.

*Paleoenvironmental Analysis of Stage 3 Deposits on the East Texas Shelf*

The large core database in the study area brackets the regionally extensive, shore-parallel escarpment illustrated in cross sections A-A’ and B-B’ (Figs. 5-5, 5-6, 5-9). This feature formed as waves eroded Beaumont Formation clay during the stage 3 maximum height of sea level. A scarp with similar geometry exists beneath Galveston Island (Bernard, 1970). The Galveston scarp was cut into Beaumont Formation clay 3,000 yrs. B.P. as sea level reached its present level, prior to barrier island formation (Siringan and Anderson, 1993).
East of FRBH a medium gray clay unit corresponding to SFU1 fills the accommodation space seaward of the scarp. Cores BD-96-4 and OBD-95-5 reveal that all growth stages of dominant taxa in this unit (SFU1) are represented and some of the bivalves are still paired, indicating in-situ preservation with no post-mortem mechanical size selection. The faunal assemblage represents a river-influenced, low-salinity environment; there are no offshore marine species. Similar fossil assemblages are known to exist in modern Texas bays (Parker; 1959, Calnan, 1980; White, et al., 1985) and in bays and lagoons associated with the Mississippi delta (Parker, 1956). The mollusks found in these modern communities inhabit very shallow depths (normally less than 2 m) and thrive in salinities between 5-10\% on average (White et al., 1985). Thus, this unit is interpreted as a low salinity back-barrier bay/lagoon deposit.

Cores FRBH-26 and FRBH-27 from transect B-B' across FRBH (Fig. 5-9) illustrate a unit composed of well-sorted, fine-grained (2.5 phi) quartz sand with frosted spherical grains and scattered roots and wood fragments that exists below the TRS. The frosted spherical quartz grains indicate transportation by eolian processes (during some point in their history). The scattered roots and wood fragments are probably not associated with sand deposition because this feature was exposed during the stage 2 lowstand and was probably vegetated. This sandy unit is interpreted as a barrier island deposit. Seaward of the
barrier, cores FRBH-28 and FRBH-29 sampled a clay unit with sand layers and laminations. This unit was deposited in a shoreface environment and is typical of modern shoreface sediments that have been sampled northeast and southwest of the Brazos Delta (Rodriguez et al., in press). Landward of the barrier, core CCBH-1 sampled oyster beds containing a monospecific oyster assemblage of *Crassostrea virginica* with many paired valves indicating in-situ preservation. This assemblage indicates deposition in a mid-bay environment (Parker, 1955).

Figure 5-11 illustrates the depositional environments that existed during the stage 3 highstand on the east Texas shelf. In the southwest portion of the study area a 6,700 km² Colorado Delta was mapped by Snow (1998; Fig. 5-11). The Colorado River had a high sediment yield at this time and produced a delta that prograded to the shelf break as sea-level fell (Abdulah, 1995). Adjacent to this delta, in an interfluvial portion of the shelf, a barrier island with an associated back-barrier lagoon existed (Fig. 5-11). This barrier was probably nourished by adjacent Colorado and Brazos delta sediments. The depth of the landward pinchout of the lagoonal facies approximates the position of the middle Wisconsin maximum sea-level shoreline.

The long-term subsidence rate for the Texas inner continental shelf is 0.01 cm/yr or less (Paine, 1993). Taking the location of the paleoshoreline and the subsidence rate for the area into account yields a total subsidence of ~4 m.
Figure 5-11.
Paleoenvironmental map of the stage 3 highstand on the east Texas shelf. A barrier island with an associated back-barrier lagoon lies to the northeast of the Colorado Delta (Snow, 1998).
This amount of subsidence places the maximum highstand associated with the middle Wisconsin at -15 m ± 2 m. The shoreline may actually have been a few meters shallower because the Holocene shoreface ravinement surface eroded some upper stage 3 deposits; however the presence of the escarpment indicates the shoreline was very close to this location.

Conclusions

Two regional high-resolution sequence stratigraphic analysis of the Texas shelf (Abdulah, 1995 and Banfield, 1998) have developed independent chronostratigraphic benchmarks from oxygen isotopes, micropaleontological data, and radiocarbon ages acquired from long sediment cores on the shelf. These chronologies have been correlated using two regional seismic surfaces, the stage 5e MFS and the stage 2 SB. Both studies have demonstrated the existence of a stage 3 MFS and stage 3 fluvial dominated deltas (the ancestral Brazos, Colorado, and Rio Grande deltas) prograding across and downlapping onto the 5e MFS. Using a very high resolution seismic tool (Chirp) and pneumatic hammer cores, the up-dip pinch out of the stage 3 deposits imaged below the stage 2 SB and above the stage 3 MFS were examined on an interfluvial portion of the east Texas shelf.
The data collected across the east Texas inner shelf imaged an escarpment. Sediment cores across the escarpment show a change from the subaerially exposed Beaumont formation Clay (~ 80-125 ka) landward, to lagoonal and barrier island deposits seaward. Based on this data, a paleoshoreline has been mapped throughout the study area. The fossils collected from lagoonal deposits filling the escarpment were preserved in-situ. The assemblage represents a low salinity back-barrier bay/lagoon environment with a water depth of no greater than 2 meters. The stratigraphic position of this unit indicates deposition during the middle Wisconsin maximum sea level highstand. The elevation of this shoreline corrected for regional subsidence was -15 m. This depth is about 70 m shallower than what is predicted by the SPECMAP curve, but only about 15 m to 30 m shallower than new $d^{18}O$ curves of Chappell et al. (1996) and Linsley (1996). The reason for the discrepancy between our data and the proxy ($d^{18}O$) sea level curve remains problematic.
CHAPTER 6

EVOLUTION AND FACIES ARCHITECTURE OF THE MODERN
BRAZOS DELTA, TEXAS: WAVE VERSUS FLOOD INFLUENCE

Summary

In order to better define the facies architecture and controlling processes on wave-dominated delta evolution, a detailed sedimentary and geomorphologic study was undertaken on the Brazos Delta, Texas. The delta has formed since 1929 when the Brazos river was diverted by the U.S. Corps of Engineers. The Brazos Delta is primarily composed of fine-grained sediments. Prodelta clay composes more than half of the sediment volume. Thick sands are isolated to the narrow delta front environment, while back-bar lagoonal clays comprise a significant component of the delta plain sequence. The facies architecture is not representative of the classic strandplain model for wave-dominated deltas. This is due to the strong influence of floods on deltaic evolution.

In early 1992, statewide flooding facilitated a major constructional phase of the delta. Significant quantities of fine-grained sediments were deposited in the prodelta. One year after the onset of flooding, a channel mouth bar emerged offshore of the river mouth and enabled progradation of the delta.
Similar flood events that occurred during 1941, 1957, and 1965 were recorded as ridge-trough pairs in the delta headland. The Brazos delta is therefore fluvially-influenced during floods, and wave-influenced during intervening periods.

**Introduction**

The well-known ternary delta classification system indicates delta morphology and three-dimensional structure is dominantly controlled by river discharge, tidal range, and wave energy (Fisher 1969; Fisher et al. 1969, Coleman and Wright 1975; Galloway 1975). According to this classification, the subaerial shape of wave-dominated deltas is cuspatate, while fluvial-dominated deltas are elongate. Based on these criteria, the Brazos delta has previously been classified as a wave dominated delta (Galloway 1975). Although this classification scheme accurately describes the gross controlling factors for subaerial delta morphology, it makes inaccurate assumptions concerning the mechanism for emplacing sands, and extent of prodelta facies.

Sedimentological investigations of modern wave-dominated deltas have concentrated mainly on the onshore portions of these deltas (Van Straaten 1959; Lagaaij and Kopstein 1964; Bernard et al. 1970; Psuty 1967; Oomkens 1970; Coleman and Wright 1975; Dominguez 1996). Based on these studies, the dominant facies of wave-dominated deltas include beach-ridge, interridge swell,
and eolian deposits. In most cases, offshore facies relationships for these deltas were inferred because of a lack of data collected from the subaqueous delta environments. The results of outcrop and subsurface studies indicate that ancient wave-dominated deltas may have distinct and widespread marine facies and that the preservation potential of these marine facies is high (Fisher 1969; Fisher and McGowen 1969; Ricoy and Brown 1977; Bebout and Loucks 1978; Foss 1979; Weise 1979).

The objective of this study was to identify and describe the active (onshore and offshore) depositional environments, facies architecture, and stratigraphy of the modern Brazos wave-dominated delta. The facies architecture and idealized depositional sequence for the delta were derived by examining the near surface lithofacies of the delta's active environments. This stratigraphic model was tested by examining the onshore portion of the delta where, due to delta progradation, a complete succession of delta facies has evolved.

The modern Brazos Delta is located in east Texas (Figs. 6-1, 6-2A). Detailed sedimentological work on the onshore portion of the delta during early development was conducted by Bernard et al. (1970), and provided an important framework for our work. During the course of our investigation, the Brazos River experienced the longest lived flood in historical time and the delta underwent a series of dramatic changes in response to this event. This investigation was drawn out over a period of five years so that the influence of this flood could be studied.
Figure 6-1. (A, B) Study area maps showing locations of cores and transects referred to in this paper. Bathymetry taken from the National Oceanic and Atmospheric Administration nautical chart number 1283 dated 1989.
Figure 6-2. Scaled aerial photographs of the Brazos delta taken in 1989 (A) and 1930 (B). The 1930 delta photo was taken shortly after river diversion and depicts the early stage of delta reworking. The photos illustrate the significant amount of sediments eroded from the old delta and deposited at the new delta following river diversion.
Study Area

The coastal zone encompassing the Brazos Delta is a climatically dynamic region. Since 1900, on average, the zone experienced the effects of tropical storms or hurricanes once every two years. Average annual rainfall ranges from 103.7 centimeters to 124.87 centimeters, with large deviations during flooding and tropical storm-induced precipitation. Winds are dominantly southeasterly, averaging 5-15 knots. Dominant southeasterlies set up waves trending northeast-southwest, which approach the shore in a northwesterly direction. The prevailing longshore currents flow from east to west.

The Brazos River leads all Texas rivers in rates of flow. Its watershed encompasses 118,000 square kilometers and includes large sections of central-north Texas and eastern New Mexico. The average annual suspended sediment yield of the Brazos is the highest of all rivers in Texas, 39 metric tons per square kilometer (Curtis et al. 1973). The sediments of the Brazos River have a distinctive red color and are characterized by fine grain sizes; most of the sediment load is clay-sized. These sediments are derived primarily from Triassic red beds located in the upper reaches of the drainage basin in northwestern Texas and northeastern New Mexico.

The Brazos Delta is located immediately southwest of Freeport, Texas at coordinates 28°52'N and 095°23'W (Figs. 6-1A, 6-2A). It is approximately 35 km² in area and extends seaward to water depths of 20 meters. The delta has
formed since 1929, when the Brazos River was diverted by the U.S. Corps of Engineers in an effort to reduce flooding and shoaling at Freeport (Seelig and Sorensen 1973).

The morphologic outline of the coast at the delta is lobate, with a significant headland on the western flank of the delta (Figs. 6-1A, 6-2A). Deposition is asymmetrical, and the delta accretes to the west along the projecting headland. Overall morphology is controlled by sediment flux and wave energy and is extremely dynamic. The east flank of the delta is composed of a series of amalgamated beach ridges at the former (pre-river diversion) shoreline position, a large lagoon immediately seaward, and another set of beach ridges parallel and adjacent to the shore (Fig. 6-2A). The west flank of the delta is composed of a series of ridges that occur as amalgamated and nonamalgamated ridge sets (Fig. 6-2A). In addition, tidal channels and washover channels are common throughout the area. Water depths in the interridge troughs are tidally controlled, averaging 0.25 to 0.5 meters. The orientation of the beach ridges is a result of wave refraction around the headland. The subaqueous delta is located mainly to the west of the river mouth and is present along shore for approximately 10-15 kilometers (Fig. 6-3). In all, the subaqueous delta comprises nearly 70% of the total area of the delta.

Methods
Figure 6-3. Evolution of the Brazos delta since 1922 based on the National Oceanic and Atmospheric Administration nautical chart number 1283. These charts show rapid erosion of the old delta at depths greater than 10 meters following river diversion. By 1940 the old delta has no bathymetric expression.
Approximately 120 cores were collected for this study (Fig. 6-1A, B). Vibracores were collected onshore and in shallow waters and a pneumatic hammer coring device was used offshore. Cores were described immediately after cutting and stored in a refrigerated storage facility. Analyses of sediments included descriptions of color, sedimentary structures, determination of total water content, grain size analysis, determination of gross mineralogy, and cursory paleontologic examination. An automated settling tube was used for grain size analysis.

Historical coastal navigation maps dating to the early 1900's and historical aerial photographs were used to examine the delta's development from its inception in 1929 to the present. Additional aerial photographs were acquired as part of this study. Photographs were taken at altitudes of 152.4 meters (500 ft) to 2438 meters (8000 feet).

**Deltaic Processes**

Prior to 1929, the Brazos Delta was located east of Freeport near Surfside Beach at coordinates 28°55.5'N and 095°17.5'W (Fig. 6-2B). More detailed descriptions of shoreline changes since 1850 in the vicinity of the Brazos River Delta can be found in Seelig and Sorensen (1973) and Morton and Peiper (1975). A synopsis of the historical evolution of the delta is presented here to highlight periods of delta progradation and reworking, and to link periods of progradation to significant floods.
The first detailed bathymetric survey of the area was carried out by the National Ocean Survey in 1852 and indicates that the Brazos River had only a small subaerial delta with a major channel mouth bar to the west of the river mouth (Seelig and Sorensen 1973). After the Freeport Jetties were constructed in 1881, a delta immediately began to expand (Seelig and Sorensen 1973). Although wave energy has significantly reworked this old delta¹, an aerial photograph from 1930 reveals a similar morphology to the new delta (Fig. 6-2A, B). The old delta was approximately 30-35 km² in area and extended seaward to 20 meters water depth. Similar to the new delta, the old delta was asymmetrical to the west, with amalgamated ridges on the eastern and western flank (Fig. 6-2A, B). The morphology of the shoreline was also lobate with a significant headland on the west flank.

Time reconstructions based on the National Oceanic and Atmospheric Administration nautical chart number 1283 (Fig. 6-3) trace the development of the delta over the last six decades. Prior to diversion, as seen in the 1922 map, the Brazos entered the Gulf southeast of Freeport. The delta had an asymmetrical lobate shoreline with the west flank dominant. This morphology was largely due to the construction of the Freeport Jetties and longshore current direction (east to west). A significant subaqueous lobe also existed and prograded into water depths of at least -10.5 meters. The river diversion occurred in 1929, and by December,

¹ “Old” is used to describe the Brazos Delta location prior to river diversion in 1929, while “New” is used to describe its current location.
1931 the subaerial portion of the pre-1929 delta was already being reworked (Fig. 6-3). Sands from the old delta lobe were moved along shore and deposited near the east flank of the new delta, creating an extensive low lying subaerial delta plain. By 1940, few remnants of the old delta remained and headland progradation of the new delta was extensive (Fig. 6-3). Net headland accretion of the new delta has occurred more or less continuously since that time, but at a slower rate. The onshore portion of the delta has prograded approximately 6.5 kilometers since 1929.

Throughout the history of the new delta, a prominent channel mouth bar and associated back bar lagoon have appeared and disappeared on several occasions. Examination of historical aerial photographs and United States Geological Survey river discharge data (Richmond, TX gauging station number 08114000) revealed that episodes of channel mouth bar expansion and emergence were associated with major floods that occurred in 1941, 1957, 1965, and 1992 (Figs. 6-4, 6-5, 6-6, 6-7). Figures 6-6 and 6-7 show emergent channel mouth bars which were enlarged as a result of the 1965 and 1992 floods. There appears to be a lag time of over a year between flood initiation and channel mouth bar emergence.

Other floods in addition to the 1941, 1957, 1965, and 1992 events have impacted the Brazos River. The reason these floods did not significantly prograde the delta is primarily due to their smaller duration. The more time the river is above flood stage, the more pronounced effect the flood has on delta progradation. The
Figure 6-4. Time versus discharge for the Brazos River. Data acquired during the largest floods of this century are plotted. Discharge data was obtained from the United States Geological Survey Richmond, Texas gauging station number 08114000.
Figure 6-5. Scaled and oriented aerial photographs of the New Brazos delta illustrating delta progradation. Following each large flood a new ridge/trough pair was preserved in the western delta headland.
New Brazos Delta
May, 1967

Channel mouth bar enlarged as a result of the 1965 flood.

New Brazos Delta
August, 1969

Channel mouth bar migrating to the west soon to form shoreline 4.

Figure 6-6. Scaled aerial photographs of the New Brazos delta illustrating the progradation event that resulted from the 1965 flood (modified after Seelig and Sorensen 1973).
Figure 6-7. Unscaled aerial photographs of the offshore emergent bar. The bar is located to the southwest of the river mouth (A) and is made up of smaller bars amalgamated together (B). Washover deposits frequently inundate the back bar lagoon (B). The southwestern portion of the bar contains large subaqueous 2-D dunes migrating towards shore as the bar welds to the shoreline (C). By June, 1995 (D) the bar has completely welded to the shoreline and is in the process of becoming the new shoreline. The dot located on the western delta headland (A, D) mark the same location in both photos.
floods that occurred in 1941, 1957, 1965, and 1992 are among the longest duration floods of this century. The Brazos River was at flood stage for about 80 days during the 1992 flood.

Following each episode of bar development, the bar was reworked shoreward and westward by waves and eventually its western edge welded to the shoreline (Figs. 6-6, 6-7). As the offshore bar migrated landward, the pre-flood shoreline was preserved within the headland as a ridge; the offshore bar became the new shoreline. The back bar lagoon was also preserved within the headland as a trough, separating the ridge from the newly formed shoreline. Each ridge within the delta headland represents the location of a previous shoreline (Fig. 6-5). One complete flood cycle for the delta spans from flood initiation to the preservation of a ridge/trough pair within the western delta headland and typically takes about six years (Figs. 6-5, 6-6). During intervals between large floods, shoreline growth is characterized by the amalgamation of beach ridges due to wave reworking and redistributing sands on the western delta headland (Fig. 6-5). River discharge data indicates that between 1965 and 1992 no significant floods impacted the area. This is expressed as amalgamated beach ridges located on the western delta headland in the March, 1989 aerial photograph (Fig. 6-5).

Storm frequency is another important factor affecting delta progradation. A well timed hurricane or tropical depression may remove an emergent channel mouth bar before it has time to evolve into a new shoreline. Seelig and Sorensen (1973) examined the impact of storms on delta morphology and concluded that the
onshore environment (the ridges, troughs, and beach) have not been altered by storms; however storms have impacted the delta front environment. Hurricane Carla struck the Brazos Delta in 1961 and swept away sand bars from the delta front environment (Seelig and Sorensen 1973). If this storm had struck in 1969 it might have removed the emergent channel mouth bar created from the 1965 flood (Fig 6-7D). Two tropical depressions impacted the Brazos Delta in 1998. At this time the channel mouth bar had welded to shore but had not formed a new shoreline yet. The channel mouth bar was completely removed by these storms. Therefore, floods are important progradational mechanisms for the delta, however a well timed storm may remove all traces of the flood event.

The most recent episode of bar emergence occurred eight months after the 1992 flood (Figs. 6-1B, 6-7). During this event, the bar formed in water depths of between 3 and 4 meters. In January 1995 it was approximately 0.3 kilometer wide and 1.5 kilometer long. Relief on the bar exceeded 2.5 meters, with average elevation at 0.5 meters above sea level (Fig. 6-7A). Bar evolution during the period between 1992 and 1997 was monitored by routine reconnaissance flights and trips to the delta, including re-occupation of coring sites and GPS transects across the bar. These combined observations have revealed extensive changes, particularly to the bar area and onshore sections. These include large-scale bedform migration, tidal channel creation/destruction, washover during storms, and eventually the bar welding itself to the mainland (Fig. 6-7D). These observations,
coupled with detailed descriptions of each delta facies, were used to examine the sedimentological evolution and facies architecture of the delta.

**Depositional Environments and Lithofacies**

Detailed examination of approximately 120 vibracores and pneumatic hammer cores (located on Fig. 6-1A, B) revealed both the modern facies and the overall stratigraphy of the Brazos delta. This study primarily focused on the offshore and nearshore portions of the delta (Fig. 6-1A, B), since that is the least understood part of this and other wave-dominated deltas. The sediment cores show a fairly systematic change in facies with increasing distance from shore and water depth. Figure 6-8 illustrates the offshore distribution of the various depositional environments and figures 6-9 and 6-10 show representative core transects across the delta. Many cores bottomed out in a mottled green and red, very stiff (shear strength >1.0 kg/cm²) Pleistocene (Beaumont Formation) clay (Fig. 6-9). This unit is unrelated to the Brazos Delta and therefore was not examined in detail. Seafloor gradients increase rapidly west and east of the river mouth (Figs. 6-1A, 6-9). The following is a general description of the main sedimentary environments and facies.

*Onshore Environments*
Figure 6-8. Surficial facies of the Brazos Delta with core transect locations indicated. All data collected during a two week period prior to the 1992 flood (see Fig. 6-1A).
Figure 6-9. Offshore core transects through the Brazos Delta with representative core photographs of each facies. These cores were collected prior to the 1992 flood. Transect locations are indicated in Figures 6-1A and 6-8.
Figure 6-9 continued.
Figure 6-10. Nearshore core transect through the Brazos Delta acquired after the 1992 flood when the offshore bar was emergent, but prior to significant landward migration of the bar. Representative core photographs of each facies are indicated by number on the transect. Transect location is indicated in Figure 6-1B. Notice that the vertical exaggeration of this cross-section is 160.
The onshore portion of the delta comprises beach ridges separated by interridge troughs and associated tidal creeks, washover fans, and the beach. Beach ridges are the dominant geomorphic element (Fig. 6-5).

**Beach Ridge.**—Beach ridge sands are more mature (texturally and mineralogically) than beach sands, as a result of alteration by colian and washover processes. They have a dominant grain size mode of 2.75-3.0 phi and are moderately sorted (Fig. 6-11B). Root casts are common. The intertidal sediments on the margins of the ridge are sandy, but have approximately 10% mud by mass.

**Interridge Trough.**—The interridge troughs are tidally influenced with water depths of less than 0.5 meters. Tidal flow has created a web of channels between troughs (Fig. 6-5). The sediments of the interridge troughs are predominantly clay, with abundant laminae and very thin beds of sand. Large sand-filled burrows cross cut interbed boundaries, often making boundaries less distinct. Small (generally less than two meters in diameter) oyster colonies are widespread.

**Mainland Beach.**—The mainland beach is very diverse sedimentologically and geomorphologically. The beach contains ripples, beach cusps, shell-armored clay balls, exposed marsh deposits, storm berms, abundant 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 (Fig. 6-5). Beach sands have a mean grain size of 2.68 phi (modal size = 2.0 phi) and are poorly sorted (sd=2.06 phi, Fig.
Figure 6-11. Grain size frequency curves for sands from the offshore (A) and nearshore (B) Brazos Delta environments.
Burrows are common. The beach is currently prograding seaward and facilitating the enclosure of the back bar lagoon (Figs 6-7A, 6-10).

**Delta Front**

The delta front encompasses the area just seaward of the river mouth and beach (Fig. 6-8). This sandy environment is extremely dynamic, including the channel mouth bar which periodically emerges creating a back bar lagoon. Prior to the 1992 flood, the nearshore setting of the new Brazos Delta was characterized by a gently dipping offshore profile extending to a subaqueous channel mouth bar situated approximately 90 meters offshore in about 2 meters water depth. Eight months after the 1992 flood, and during extreme low (spring) tides, the offshore channel mouth bar initially emerged into a prominent sand body that was located between 0.9 and 1.2 km offshore and extended for approximately 1.5 km along shore (Fig. 6-7A). The emergent channel mouth bar was a highly dynamic feature. It was composed of numerous amalgamated bars and underwent dramatic changes with seasonal tidal cycles (Fig. 6-7B). During spring tides the channel mouth bar was partially submerged and large subaqueous two dimensional sand dunes formed on its western end (Fig. 6-7C). Storm overwash was common (Fig. 6-7B). During neap tides the channel mouth bar attained relief of nearly 1 meter on the landward facing beach. Several core transects were collected from the onshore beach to the seaward side of the emergent channel mouth bar (Figs. 6-1B, 6-10).
**Channel Mouth Bar Crest.**---Channel mouth bar crest sands comprise the coarsest sands in the region (2.5 to 3.0 phi, Fig. 6-11B). They are moderately sorted with negative skewness values due to shell fragments and heavy minerals. Heavy mineral concentrations in the swash zone can reach up to 15%, compared to <1% for adjacent beach sands. Thickness of the bar deposit ranged from 1 meter on the west end to 3 meters in the east. Ripples, laminations lined with organic debris, and convolute laminations are the dominant sedimentary structures (Fig. 6-10). Bioturbation and silt and mud filled burrows are common throughout. Armored mud balls are readily found along the seaward side of the bar.

**Distal Channel Mouth Bar.**---Distal channel mouth bar facies exist seaward of the river mouth and delta headland out to approximately -4 m water depth. This facies is characterized by laminated sands and interbedded sands and muds (Fig. 6-10). The sediments are (by mass) 85% sand. This facies underlies the channel mouth bar crest and the back bar lagoon (Fig. 6-10).

**Back Bar Lagoon.**---Associated with emergence of the channel mouth bar in 1992 was the formation of a back-bar lagoon. Water depth within the lagoon averaged between 0.5 to 1.5 meters. The central lagoon was the site of rapid accumulation of fine-grained, very wet, muddy sediments. The muds have less than 10 % sand, and up to 50% water by mass. Bioturbation is noticeably absent, and significant amounts of logs that accumulated in the lagoon often prevented core penetration. The lagoonal environment is also strongly influenced by fluvial
input, washover, tidal and eolian processes which are responsible for the deposition of sand layers within the lagoon (Fig. 6-7, 6-10).

**Offshore Environments**

**Shoreface.---**The shoreface deposits recovered in cores consists of upper shoreface massive sands that grade offshore into lower shoreface laminated sands interbedded with mud and muddy sand layers (Fig. 6-9). Shoreface deposits are more extensive to the west and east of the delta. Generally, shoreface sands are texturally mature, with rounded to subrounded well sorted grains. Upper shoreface sands are similar in mean grain size to lower shoreface sands (around 3.2 phi) but have a slightly coarser dominant mode (3.25 versus 3.5 phi, Fig. 6-11A). Bounding surfaces between sand and mud layers in the lower shoreface are highly erosional. Muddy sands and sandy muds of the lower shoreface possess a mixture of properties: variable muddiness, high bioturbation, and variegated color. Shell fragments are scattered throughout the shoreface and within vertically oriented burrows. In the absence of grain size data, the lower shoreface may be difficult to distinguish from the distal delta front environment (Fig. 6-11A).

**Distal Delta Front.---**Reddish brown clay with very fine laminations of silty clay is the dominant sediment type in the distal delta front. This clay generally has less than 10% silt and sand, but is often interbedded with sandier and siltier units (Fig. 6-9). Sands within the distal delta front are coarser than shoreface sands (2.8 phi, Fig. 6-11A) and occur as thin beds (Fig. 6-9). Organic material
(mostly wood fragments) is scattered throughout the clay and occurs as individual beds (<5 cm).

**Prodelta.** Delta front facies grade offshore into thinly-interbedded olive gray clay and reddish brown clay facies of the prodelta (Fig 6-9). This facies illustrates the episodic nature of sedimentation in the delta. With each pulse of sediment into the delta, a new layer of reddish brown clay is deposited in the prodelta. When the fluvial discharge decreases, a layer of offshore marine sediments consisting of shell-bearing, bioturbated, olive gray clay is deposited. Although the prodelta is characterized dominantly by mud, thin (<1 cm) very fine sand layers were sampled (Fig. 6-11A).

In all, prodelta muds comprise 50-60% of the total sediment volume of the delta. The distribution of prodelta, fine-grained sediments is mostly dependent on proximity to the river mouth; however, longshore currents also affect prodelta deposition and produce an architecture asymmetric to the west (Fig. 6-3). Areas east of the river mouth only experience appreciable prodelta deposition during times of increased discharge. The prodelta occurs at -6 to -15 meters water depth. This is below the level of wave erosion for most of the east Texas coast, which is situated at between -8 and -10 meters water depth (Siringan and Anderson 1993). The offshore extent of the prodelta is variable, but generally occurs between 1.2 km to 7 km.

Comparison of bathymetric maps that pre- and post-date deltaic input (Fig. 6-3) to the area, and sediment cores that penetrated the prodelta, show that
approximately five to six meters of prodelta sediment have accumulated since 1929. This equates to an estimated total sediment volume of 84,000,000 m³. Cores collected after the 1992 flood in the prodelta indicate 50 cm of vertical accumulation in some places (Hamilton 1995).

**Facies Architecture**

The combined offshore core transects (Fig. 6-9) show a typical progradational deltaic succession comprised of proximal facies overlying distal facies, and an overall coarsening upward lithologic sequence. Sea floor gradients decrease toward the river mouth due to increased accumulation of thick delta front and prodelta sediments (Figs. 6-1A, 6-9). The delta front and distal delta front facies are the most confined, extending only 7 km northeast and 13 km southwest of the river mouth. The distal delta front facies represents a transitional unit separating the sand dominated environments of the delta front from the clay-dominated prodelta.

The first flood impacted 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 within the New Brazos Delta. It was not until after the 1941 flood that the first lagoon was preserved within the western delta headland (Fig. 6-6).
After the old delta had been eroded, floods became the primary mechanism enabling delta progradation.

Figure 6-12 presents a simplified model illustrating delta evolution through one complete flood cycle. Prior to flooding (Fig. 6-12, stage 1) no emergent offshore bar exists. The shoreline is lobate with a series of beach ridges separated by interridge 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 (Fig. 6-12, stage 2). The channel mouth bar rests above distal mouth bar facies.

The Richmond gauging station indicated that during the 1992 flood 90% of suspended sediments were finer than sand. However, significant erosion of point bars did occur following the 1992 flood. This, plus the grain size data, indicates channel mouth bar sands were derived from the river and transported to the delta via traction or saltation. Sand transported during the 1992 flood was most likely deposited near the river mouth where decreased competence of the flow facilitated sand deposition. The high concentration of suspended silts and clays increased the viscosity of the flood plume to a point where it dampened waves, thus protecting nearshore sands from being reworked (Fig. 6-13). The lag time between flood initiation and channel mouth bar emergence can be explained by the relatively slow transport rate of sand via bedload processes, and the time needed for waves to rework nearshore sands into an emergent channel mouth bar. The emergent channel mouth bar is
Figure 6-12. Brazos Delta evolutionary model illustrating the morphologic, and sedimentologic changes that occur during one complete flood cycle.
Figure 6-13. Aerial photograph of the Brazos Delta flood plume taken from approximately 150 meters looking south. The viscosity of the plume dampens the incoming 1 to 1.5 meter waves.
subsequently reworked by waves and migrates landward welding its western end to the shoreline (Fig. 6-7D).

As the bar migrates landward over the back bar lagoon, the seaward portion of the underlying distal mouth bar facies is extensively reworked, and a portion of the seaward side of the back bar lagoon is eroded. However, the landward portions of these environments are preserved below the delta headland (Figs 6-10, 6-12). Significant headland accretion and progradation occurs on both the eastern and western sides of the river mouth. During stage 3 (Fig. 6-12) 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. 6-14) was collected from a beach ridge that was preserved within the delta headland following the 1941 flood (Fig. 6-1B). This core tests the model presented in Figure 6-14 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 6-12, stage 3. 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. This core was collected in 1954 and shows the same facies succession as core BDBR-01, however, the core collected by Bernard and
Figure 6-14. Core BDBR-01 illustrates the stacking pattern of nearshore environments after one complete flood cycle. Core location is indicated in Figure 6-1B. Figure 6-12 illustrates the delta processes leading to this facies stacking pattern and the location of the core.
his colleagues was long enough to penetrate the entire delta succession. 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).

Conclusions

The modern Brazos Delta has experienced two phases of evolution. During the first two decades following diversion of the river, sands swept from the old delta were transported by longshore currents, reworked shoreward, and deposited as amalgamated ridges near the mouth of the new delta. This early phase of evolution was more typical of the classic wave-dominated delta (Bernard et al 1970). Once the old delta’s sand supply was depleted, the delta assumed a new morphology, characterized by alternating ridges and interridge lagoons. This phase of delta evolution has been highly episodic, characterized by rapid phases of sediment discharge and growth during and immediately after major floods, and slower delta growth and reworking during intervening periods. Major floods occurred in 1941, 1957, 1965 and 1992. During each of these events the channel mouth bar experienced significant growth and a back bar lagoon was formed. Following each flood the bar migrated onshore and to the west, welded to the shore, and became the new shoreline. The old shoreline, in addition to part of the back-bar lagoon, was preserved within the
western delta headland. These flood cycles have resulted in the alternating beach ridge sands and mud-filled troughs that characterize the delta headland. Prodelta development was also enhanced by floods. During non-flood periods, shoreline growth is characterized by accreting sand ridges which amalgamate at the western delta headland; no troughs are created during non-flood periods. Since 1965 a wide area of accreted sand ridges has formed in the westernmost portion of the delta headland (Figs. 6-2, 6-6). These ridges are sourced from reworked delta front sands and/or sands carried from the east by longshore currents.

The new Brazos delta contains a suite of depositional environments from clayey prodelta to sandy delta front. The most striking feature of the Brazos delta is its large prodelta. The offshore prodelta comprises nearly 60% of the delta's volume. The prodelta and distal delta front environments have the highest preservation potential given the present day depth of wave erosion (-8 to -10 m). However, under different conditions, for example where subsidence rates are higher, the sandy delta front environment would have a high preservation potential.

Although the coastal outline is similar to the classically-postulated wave-dominated delta, the Brazos Delta has internal complexities that are not immediately recognizable strictly from geomorphology. These features include the back bar lagoonal deposits preserved within the onshore troughs, the continuous distal mouth bar facies which underlies the onshore portion of the
delta, and the extensive prodelta environment. Psuty (1967) and Coleman and
Wright (1975) contend that wave dominated systems are composed of flanking
strandplains, thereby inferring that beach ridges are amalgamated. The Brazos
Delta does not follow this model, for it is dominantly composed of preserved
shorelines (ridges) separated by troughs. These troughs contain previously
unrecognized lagoonal clay. Lack of sand body connectivity in our model
versus the classical wave-dominated delta model is important to resevoir
quality. Although the dominant process controlling Brazos delta morphology is
waves, at times (during floods) it is fluvially dominated. The floods and
associated delta prograding events are episodic. Preservation of these events
within the delta headland is largely dependent upon the duration of the flood
and the timing of post-flood storms. Thus, the Brazos Delta may justifiably be
termed wave dominated (Galloway, 1975); however, it is the combination of
wave, fluvial, and to a lesser extent even tidal processes that influence the
facies architecture.
CHAPTER 7

PRESERVATION OF LATE PLEISTOCENE TO RECENT COASTAL LITHOSOMES ON THE EAST TEXAS SHELF: A DISCUSSION AND SUMMARY

Introduction

This study of Late Pleistocene to recent coastal lithosomes on the east Texas coast and continental shelf focused on characterizing sedimentary facies and developing evolutionary models for bay complexes (fluvial, bay-head delta, middle bay, tidal delta, and barrier shoreline environments), offshore banks, incised valleys, and the Brazos Delta. Within the study area, preservation of these deposits on the continental shelf has been variable. Coastal lithosomes had a high preservation potential in the eastern portion of the study area (around the Trinity/Sabine incised valley; Fig. 7-1) and a low preservation potential in the western portion of the study area (offshore of Follets Island and the Brazos Delta; Fig. 7-2).

Most shoreline migration models relate coastal lithosome preservation to changes in the rate of sea-level rise (Fig. 7-3). The continuous shoreface retreat model of Brunn (1962) and Swift (1975, 1976) indicates that as sea level rises, erosion at the shore causes barriers to migrate continuously landward.
Figure 7-1. Map of the eastern portion of the study area showing the locations of preserved Holocene coastal lithosomes on the shelf (in gray).
Figure 7-2. Map of the western portion of the study area showing the absence of preserved Holocene coastal lithosomes on the shelf. Core data indicates that the Holocene Brazos Deltas (dot pattern) from Bartek et al. (1990) do not exist.
Figure 7-3. Four models illustrating shoreline retreat during transgression. The Brunn theory indicates that a rise in sea level causes the equilibrium profile to rise and move landward causing erosion of the shoreface and deposition seaward of a null point (A; after Schwartz, 1967). Swift (1975) concluded that periodic still stands cause the vertical component of shoreface translation to increase at the expense of the Horizontal component forming steps in the ravinement surface (B). Kumar and Sanders (1975) indicate that barriers are "drowned" in place as sea-level rises (C). Penland et al. (1988) showed that inner-shelf shoals can be generated by the reworking of a submerged barrier island (D).
Figure 7-3 continued.
leaving only a thin layer of reworked sand behind on the shelf (Fig. 7-3 A, B). During periodic stillstands in the overall transgression, the shoreline aggrades/progrades ultimately producing steps in the ravinement surface (Swift, 1975, 1976; Fig. 7-3 B). The in-place drowning model of Sanders and Kumar (1975) and Rampino and Sanders (1980, 1982) indicates that during rapid sea-level rise, barrier shorelines are submerged and preserved in-place on the continental shelf (Fig. 7-3 C). Penland et al. (1988) concluded that during relative sea-level rise a barrier island will submerge and become completely reworked into an inner shelf shoal (Fig. 7-3 D). These models only examine changes in the rate of sea-level rise as a mechanism for coastal lithosome preservation during transgression; hence, they are simplistic. Belknap and Kraft (1981) concluded from data collected on the Delaware and Maryland shelf that preservation of coastal lithosomes is dependent on seven factors: pre-existing topography, depth of erosion, wave energy, sediment supply, erosion resistance, tidal range, and rate of relative sea-level change. The relative importance of each factor in coastal lithosome preservation can not be assessed within an area where most of these factors are constant (except for the rate of sea-level rise) or poorly constrained. The high density of data and the variable preservation of coastal lithosomes makes the east Texas continental shelf ideal for examining the factors that influence preservation during transgression.
Eastern Portion of the Study Area

Coastal lithosomes deposited within the Trinity/Sabine incised valley, in the eastern portion of the study area, had the highest preservation potential. As sea-level rose during the last deglaciation, the Trinity/Sabine incised valley backfilled with continuous fluvial and bay-head delta sediments and discontinuous middle bay, and coastal (tidal delta complex) deposits (Fig. 4-4). Thomas and Anderson (1994) identified four packages of sediments bound by flooding surfaces that separated proximal coastal environments (bay-head delta and fluvial environments) from overlying distal coastal environments (middle bay and tidal delta environments) within the incised valley. The top of the bay-head delta surface was mapped regionally within the incised valley and is characterized by a series of flat steps and inclined risers (Fig. 4-4 and 4-5). These features have also been recognized in the lower bayline surface (Thomas and Anderson, 1994). In places on the inner shelf, banks (Sabine, Heald, Shepard, and Thomas banks) lie adjacent to and over the incised valley (Fig. 4-2). These banks represent submerged paleoshorelines composed of three facies (from top to bottom): (A) a storm reworked facies; (B) a fore-barrier, lower shoreface/ebb-tidal delta facies; and (C) a back barrier estuarine facies (Rodriguez et al., 1999; Fig. 2-8).

The deposition of middle bay, tidal delta, and shoreline deposits, i.e. the formation of a bay complex, indicates a time of environmental stability. During
this time, coastal lithosomes prograde/aggrade. The overstepping and
stranding of bay and coastal barrier facies on the continental shelf represent
periods of environmental change. The periods of environmental stability and
environmental change were most likely caused by changes in the rate of sea-
level rise. During the overstepping events, wave and current (storm and tidal)
processes (ravinement processes) eroded into coastal lithosomes forming a
ravinement surface. The depth of ravinement controls the preservation potential
of coastal lithosomes. For example, within the banks, the ravinement surface is
located at the contact between the storm reworked unit (facies A) and the fore-
barrier, lower shoreface/ebb-tidal delta unit (facies B); the upper portions of the
barrier Island, including all environments from the upper shoreface to the
backshore, have been completely removed and/or reworked. In the eastern
portion of the study area, coastal lithosomes are preserved only within and
adjacent to the Trinity/Sabine incised valley. A thin layer (<100 cm) of
Holocene marine mud lies directly on Pleistocene clay in the interfluvial areas
of the shelf.

Western Portion of the Study Area

A study of the modern Brazos Delta was carried out not only to better
constrain the processes and facies architecture of a wave-dominated delta, but
also to aid in the recognition of Holocene Brazos Deltas that may have been
preserved on the continental shelf during transgression. The Brazos River leads all Texas rivers in rates of flow. The watershed of the river encompasses 118,000 square kilometers and includes large sections of central-north Texas and eastern New Mexico. The average annual suspended sediment yield of the Brazos is the highest of all rivers in Texas, 39 metric tons per square kilometer (Curtis et al. 1973). The sediments of the Brazos River have a distinctive red color and are characterized by fine grain sizes; most of the sediment load is clay-sized. Previous studies indicate that during the Holocene the Brazos River flowed through the Big Slough meander belt, and the Oyster Creek meander belt (McGowen et al., 1976; Bernard et al., 1970; Bartek et al., 1990; Fig. 7-2). Bartek et al. (1990) concluded that three overstepped Holocene Brazos Deltas exist on the continental shelf at the locations of Freeport Rocks Bathymetric High (FRBH), San Louis Pass Bathymetric High (SLPBH), and Oyster Creek Sand Body (OCSB; Fig. 7-2). These conclusions were based only on seismic data; no cores were collected from the area. During this study sediment cores were collected from FRBH, SLPBH, and OCSB, none sampled a transgressed Holocene Brazos Delta. If an overstepped Holocene Brazos Delta exists in the western portion of the study area, it should have been recognized given the large core database of this study and the distinctive characteristics of Brazos delta facies. In fact, no Holocene coastal lithosomes have been preserved in the western portion of the study area. All cores
collected sampled Holocene marine mud lying directly over Pleistocene deposits.

Siringan (1993) examined sediment cores from SLPBH and determined this feature to be a transgressed flood tidal delta, no Brazos Delta sediments were found. FRBH represents a highstand barrier shoreline deposit associated with the middle Wisconsin (stage 3) maximum sea-level highstand (Fig. 5-9). Cores collected from the OCSB penetrated <100 cm of Holocene marine mud lying directly over Pleistocene clay. One of these cores did penetrate a thin (<20 cm) sand layer at the sea floor verifying that there is sand exposed at this location. Given the high sediment load of the Brazos River, each avulsion event must have abandoned a delta on the shelf; however, these potentially large deltas were not preserved. To understand why preservation of Holocene coastal lithosomes within the study area is variable, it is necessary to examine the dominant factors that control preservation, and explore how these factors vary throughout the study area.

**Preservation Potential of Coastal Lithosomes**

The three dominant factors that control the preservation potential, distribution, and internal geometry of Holocene coastal lithosomes along the east Texas coast and continental shelf are the rate of sea-level rise, the rate of sedimentation, and the pre-transgressive geology (topography of the stage 2
sequence boundary, shelf gradient, and lithology; Belknap and Kraft, 1981, 1985; Siringan, 1994). Ravinement removes coastal lithosomes as the shoreface retreats during sea-level rise and forms a prominent surface of erosion on the shelf. Preservation of coastal lithosomes during transgression is determined by the separation between the ravinement surface, formed during shoreface retreat, and the lower bounding surface of the coastal lithosome (Belknap and Kraft, 1981, 1985). When these two surfaces meet there is no preservation, total preservation exists when the distance between these two surfaces is at a maximum.

Currently, the depth of ravinement in the study area is -7 to -9 m (Siringan and Anderson, 1994). This surface corresponds to a change in the offshore profile and is the point where shoreface deposits pinch out. Seaward of this location, marine muds lie directly overly Pleistocene deposits. One way to maximize the preservation of a coastal lithosomes is to minimize the depth of ravinement. An increase in the rate of sea-level rise will decrease the depth of the ravinement surface. This is because coastal lithosomes will become submerged below the depth of ravinement at a faster rate; thus, minimizing the amount of time exposed to ravinement processes. Although changes in the rate of sea-level rise may play a large role in bank formation and facies architecture in the eastern study area, it can not be the only limiting factor controlling preservation because no Holocene coastal lithosomes were preserved in the interfluvial areas of the eastern study area or in the western study area.
Sediment supply also effects the preservation potential of coastal lithosomes. A coastal lithosome with a high sediment supply generally has a higher preservation potential than a coastal lithosome with a low sediment supply. This is because as ravinement processes remove sediments from a lithosome, new sediment is available to be deposited in its place. When sediment supply is low, there is not enough sediment replacement and the rate of erosion increases. The Brazos River had a higher discharge than the Trinity River during the Holocene. This should indicate that Holocene Brazos deltas had a high preservation potential on the continental shelf; however, no Holocene coastal lithosomes were preserved in the western portion of the study area.

Deltas and barrier shorelines are nourished in fundamentally different ways. These differences partially explain the greater preservation of shoreline deposits in the eastern study area as compared to delta preservation in the western study area. Deltaic sediments are point sourced (from a river’s discharge), while barrier shoreline sediments are sourced from offshore (reworking of sediments landward) and alongshore (transport of sediments by longshore currents). Each Brazos River avulsion event shut off sediment supply to the delta. In the absence of a sediment source, ravinement processes were able to erode down to the lower bounding surface (Pleistocene clays). This is similar to what happened to the pre-1929 Brazos Delta after river diversion by the army core of engineers (Fig. 7-4). In less than 15 years the old Brazos Delta
Figure 7-4. Isopach map of eroded Old Brazos Delta sediments. The map was created by comparing the 1922 National Oceanic and Atmospheric Administration nautical chart number 1283 with the same numbered 1986 nautical chart.
was completely removed by ravinement processes. If each avulsion event was associated with a rapid sea-level rise event, then the preservation potential of these Holocene Brazos deltas may have been greater; however, no data exists to support this association. Preservation potential is at a maximum when sediment supply to a coastal lithosome is constant (preferably constantly high), and sea-level is rising rapidly. The barrier shorelines in the eastern portion of the study area were nourished by longshore currents and by Pleistocene sands eroded by the advancing shoreface throughout the transgression. This more or less continuous input of sand contributed to the high preservation potential of coastal lithosomes. Although rapid sea-level rise and constant sediment supply are important conditions necessary for coastal lithosome preservation, they do not necessarily ensure preservation. For example, if these were the only two controlling factors then the preserved barrier shorelines in the eastern portion of the study area (the banks) would be more extensive, when in fact, they are only preserved adjacent to and over the Trinity/Sabine incised valley.

The pre-transgressive geology (topography of the stage 2 sequence boundary and shelf gradient) and lithology of the deposits being eroded also needs to be addressed, for this played a key roll in the preservation of banks in the eastern portion of the study area. During the last lowstand, the Trinity/Sabine rivers incised on average ~40 m (into the Pleistocene; Thomas and Anderson, 1994) providing the accommodation space necessary for bay complexes to form and become preserved. Sedimentary environments
deposited within incised valleys are protected from ravinement processes (Belknap and Kraft, 1985). Barrier shorelines extend over interfluvial areas of the shelf as well as over the incised valleys. Complete removal of barrier shoreline deposits occurred over the interfluvial areas, while partial preservation occurred over the Trinity/Sabine incised valley. Barrier shoreline features are more likely to be preserved over incised valleys because subsidence rates are greater (0.62 cm/yr over the valley versus 0.01 cm/yr over the interfluve) due to sediment compaction of young Holocene deposits. A higher subsidence rate submerges the lower portions of a barrier shoreline below the level of ravinement at a faster rate, increasing preservation.

Although no Holocene coastal lithosomes were preserved in the western portion of the study area, a middle Wisconsin highstand shoreline was preserved on the continental shelf in around -18 m of water (Fig. 5-9). The shoreline was subaerially exposed for ~ 35 ka. During this time the upper and lower bay clays became stiff (from dewatering and diagenetic alterations) and portions of the shoreline were cemented (Curry, 1960). These shoreline sediments were deposited within a wave cut scarp, and subsided at least 3 meters prior to transgression. The above factors helped assure preservation of these deposits during the Holocene transgression. Preservation of highstand shoreline deposits during subaerial exposure is not rare. The Ingleside shoreline trend, a stage 5e highstand shoreline deposited when sea-level was
+5 m around 120 ka, occurs intermittently from SW Louisiana to Tamaulipas, Mexico (Price, 1934, 1958; Winker, 1979; Morton and price, 1987).

**Future Work**

Correlating flooding surfaces between Corpus Christi, Galveston, and Mobile bays will aid in distinguishing climatic from eustatic events. A series of long (up to 50 m) sediment cores needs to be collected up the axis of the Trinity, Nueces, and Mobile incised valleys. Flooding surfaces can then be radiocarbon dated from each core site. By dating a flooding surface in various locations, the rate of environmental change, the magnitude of sea-level rise that caused the change, the time necessary for environments to reestablish equilibrium, and the timing of these events, can be constrained. Once the rate, and magnitude of sea-level rise necessary to cause bay complexes to respond catastrophically is known, accurate models can be created that will predict how these and other coastal systems will respond to future sea-level rise.
LIST OF REFERENCES


Anderson, J.B., Abdulah, K., Sarzalejo, S., Siringan, F., and Thomas, M.A., 1996, Late Quaternary sedimentation and high-resolution sequence


Bernard, H.A., Major, C.F., Jr., Parrot, B.S., and LeBlanc, R.J., 1970, Recent sediments of southeast Texas-a field guide to the Brazos alluvial and deltaic
plains and the Galveston barrier island complex: The University of Texas at Austin, Bureau of Economic Geology Guidebook 11, 132 p.


Colman, S. M., Mixon, R. B., Rubin, M., Bloom, A. L., and Johnson, G. H., 1989, Comments and Reply on “Late Pleistocene barrier-island sequence along
the southern Delmarva Peninsula: Implications for middle Wisconsin sea levels": Geology, v. 17, p. 84-85.


Finkelstein, K., and Kearney, M. S., 1988, Late Pleistocene barrier-island sequence along the southern Delmarva Peninsula: Implications for middle Wisconsinan sea levels: Geology, v. 16, p. 41-45.


Montaggioni, L. F., Cabioch, G., Camoinau, G. F., Bard, E., Ribaud-Laurenti, A.
over the past 14 k.y. on the mid-Pacific island of Tahiti, Geology, v. 25, no. 6,
p. 555-558.

Morton, R. A., and Pieper, M. J., 1975, Shoreline changes in the vicinity of the
Brazos River Delta (San Luis Pass to Brown Cedar Cut) an analysis of
historical changes of the Texas Gulf Shoreline: The University of Texas at
Austin, Bureau of Economic Geology, Geological Circular 75-4, 47 p.

Morton, R.A., 1977, Historical shoreline changes and their causes, Texas Gulf
Coast: Gulf Coast Association of Geological Societies Transactions, v. 27, p.
352-364.

the Texas Coast: The University of Texas at Austin, Bureau of Economic
Geology, Guidebook 20, 167 p.

Morton, R.A., 1981, Formation of storm deposits by wind-forced currents in the
Gulf of Mexico and the North Sea, in Nio, S.D., ed., Holocene Marine
Sedimentation in the North Sea Basin: International Association of
Sedimentologists Special Publication 5, p. 385-396.

sedimentary phases of the Texas coastal plain and shelf, in Nummedal, D.,
Pilkey, O. H., and Howard, J. D., eds., Sea-Level Fluctuation and Coastal


Parker, R.H., 1955, Changes in the invertebrate fauna, apparently attributable to salinity changes in the bays of central Texas: Journal of Paleontology, v. 29 p. 193-211.


Suter, J. R., Berryhill, H. L., and Penland, S., 1987, Late Quaternary sea-level fluctuations and depositional sequences, southwest Louisiana continental


Varekamp, J. C., and Thomas, E., 1998, Climate change and the rise and fall of sea level over the millennium, EOS, v. 79, no. 6.


## Appendix 1. Radiocarbon Age Date Information

Analysis performed by:
Beta Analytic, Inc.
4985 SW 74th Street
Miami, FL 33155
tel: (305) 667-5167 fax: (305) 663-0964
beta@analytic.win.net

<table>
<thead>
<tr>
<th>Lab code</th>
<th>Sample name</th>
<th>Material</th>
<th>Conventional C14 age</th>
<th>Calibrated age</th>
</tr>
</thead>
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<tr>
<td>Beta-118457</td>
<td>TV-93-3-425 cm</td>
<td>Rangia</td>
<td>5790 ± 60 BP</td>
<td>4315 to 4205 cal.BC</td>
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<tr>
<td>Beta-116394</td>
<td>BRETD7 330-340 cm</td>
<td>Crassostrea</td>
<td>6610 ± 100 BP</td>
<td>5245 to 5035 cal.BC</td>
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<tr>
<td>Beta-112201</td>
<td>OGV91-28c 215-229 cm</td>
<td>Peat</td>
<td>&gt;46160 BP (radiocarbon dead)</td>
<td>-</td>
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<td>Beta-111400</td>
<td>HB-93-7-S4-310 cm</td>
<td>Mulineae</td>
<td>7900 ± 50 BP</td>
<td>6410 to 6335 cal.BC</td>
</tr>
<tr>
<td>Beta-111401</td>
<td>SH-93-2 352-355 cm</td>
<td>Mulineae</td>
<td>7390 ± 50 BP</td>
<td>5915 to 5795 cal. BC</td>
</tr>
<tr>
<td>Beta-111402</td>
<td>TV-93-16-S4-435 cm</td>
<td>Abra aequalis</td>
<td>1900 ± 50 BP</td>
<td>455 to 585 cal. AD</td>
</tr>
<tr>
<td>Beta-106196</td>
<td>BD-96-4TS 246-252 cm</td>
<td>Littorina</td>
<td>45450 ± 920 BP</td>
<td>-</td>
</tr>
<tr>
<td>Beta-106197</td>
<td>SB93-7S10 288-292 cm</td>
<td>Ostrea equestris</td>
<td>4490 ± 50 BP</td>
<td>2835 to 2621 cal. BC</td>
</tr>
<tr>
<td>Beta-102665</td>
<td>CB-1 155-159 cm</td>
<td>Crassostrea</td>
<td>40310 ± 1370 BP</td>
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<tr>
<td>Beta-100751</td>
<td>BD-96-4 246-252 cm</td>
<td>Rangia</td>
<td>42250 ± 2590 BP</td>
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<tr>
<td>Beta-95293</td>
<td>OFI-95-17 73 cm</td>
<td>Rangia</td>
<td>&gt;37870 BP (radiocarbon dead)</td>
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<td>Beta-95294</td>
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<td>37080 ± 630 BP</td>
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<tr>
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<td>Beta-92337</td>
<td>FRBH-15 78-83 cm</td>
<td>Mixed shell assemblage</td>
<td>&gt;40830 (radiocarbon dead)</td>
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<tr>
<td>Beta-74445</td>
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<td>Beta-74598</td>
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<td>HB-93-7 324-337 cm</td>
<td>peat</td>
<td>8570 ± 70 BP</td>
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</tbody>
</table>
Appendix 2. Geochemical and SEM analysis of *Vioscalba louisiana* and *Littorina sphinctostoma* shells.

A. *Vioscalba louisiana*
Appendix 2 cont.

B. Littorina sphinctostoma
Appendix 2 cont.

C.

X-ray diffraction data for *Vioscalba louisiana* and *Littorina sphinctostoma*

- Vioscalba louisiana
- Littorina sphinctostoma

100 peak for aragonite
52 peak for aragonite

D-spacing: 4.4394 to 2.9785
2-theta: 20-30 degrees
in 500 steps of .02 degrees

The highest intensity peak (100 peak) for aragonite is when 2-theta = 26.22; the 52 peak is when 2-theta = 27.24. The 100 peak for calcite is when 2-theta = 29.34. From the data presented above, *Vioscalba louisiana* and *Littorina sphinctostoma* shells are aragonitic.
Appendix 2 cont.

D. We attempted to U-Th date the *Littorina sphinctostoma* and *Vioscalba louisianae* shells. The analysis was performed by Dr. Robert Anderson at the Lamont-Doherty Earth Observatory. The U content of the shells are extremely low, making these species unsuitable for U-Th dating. The samples were prepared by gentle crushing, sonicating, and decanting to remove any clays. “Cleaned” specimens were then dissolved in nitric acid, and U and Th were purified on a single anion exchange column. The results are shown below:

<table>
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<tr>
<th>Sample name</th>
<th>Inferred age</th>
<th>232 Th (ug/g)</th>
<th>232 Th (dpm/g)</th>
<th>238 U (ug/g)</th>
<th>238 U (dpm/g)</th>
</tr>
</thead>
<tbody>
<tr>
<td>TV-7</td>
<td>modern</td>
<td>0.826</td>
<td>0.2</td>
<td>0.492</td>
<td>0.370</td>
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<td>BD-96-4LS</td>
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<td>0.293</td>
<td>0.133</td>
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<td>stage 3</td>
<td>1.410</td>
<td>0.342</td>
<td>0.145</td>
<td>0.109</td>
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</table>
IMAGE EVALUATION
TEST TARGET (QA-3)