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Coastal lithosome evolution and preservation during an overall rising sea level: East Texas gulf coast and continental shelf

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Rice University, 1993
RICE UNIVERSITY

COASTAL LITHOSOME EVOLUTION AND PRESERVATION DURING AN OVERALL RISING SEA LEVEL: EAST TEXAS GULF COAST AND CONTINENTAL SHELF

by

FERNANDO PASCUAL SIRINGAN

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

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ABSTRACT

COASTAL LITHOSOME EVOLUTION AND PRESERVATION DURING AN OVERALL RISING SEA LEVEL: EAST TEXAS GULF COAST AND CONTINENTAL SHELF

by

FERNANDO PASCUAL SIRINGAN

Present coastal systems along the east Texas coast evolved during the past 3.5 ky of relative sea-level stillstand following a rapid sea-level rise 4 ka. Closure of proto-Galveston Bay, caused by spit accretion of Galveston Island and Bolivar Peninsula, formed Bolivar Roads approximately 3.3 ka. Increased tidal prism and entrenchment over the Trinity River incised valley led to inlet stabilization and intensification of tidal influence.

The present shoreface and inner shelf package is characterized by a paucity of storm deposits. Strong along-shelf storm currents, low sediment supply, and low effective accommodation space in the region are unfavorable for the preservation of storm beds. Higher sand supply during the early establishment of the present coastal lithosomes resulted in a greater occurrence of storm beds lower in the section. Amalgamated storm deposits on the east Texas shelf are associated with reworked coastal lithosomes.

Pods of tidal-inlet, tidal-inlet/spit, and tidal-delta deposits mark previous shoreline positions on the continental shelf. Their distribution mimics the along-strike variation of the present coastal systems, defines six relative sea-level stillstands, including the present, during the past 10.2 ky, and supports the model of a step-like sea-level rise. The seismic architecture of pre-8 ka coastal lithosomes provide evidence for greater tidal influence, greater accommodation
space, and higher sedimentation rates compared to the present.

The preserved coastal lithosomes indicate that the depth of shoreface ravinement decreases with decreasing shelf gradient, increasing rates of sea-level rise, and increasing sediment supply. Better preservation within incised valleys results from greater accommodation space and the soft valley-fill that allows incision of the inlets beyond the depth of shoreface ravinement.

The mechanism of shoreline translation (discontinuous erosional shoreface retreat, transgressive submergence, or in-place drowning), is a function mainly of shelf gradient and rate of sea-level rise. Gentle shelf gradient and rapid sea-level rise favor transgressive submergence. In regions with steep shelf gradient, aggradation may produce stratigraphic signatures consistent with in-place drowning and discontinuous erosional shoreface retreat.
ACKNOWLEDGMENTS

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finally, I thank my friend, Jesus, for giving me purpose, peace, and joy.
The earth is the Lord's, and everything in it,
the world, and all who live in it;
for he founded it upon the seas
and established it upon the waters.

- Psalm 24:1-2
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PREFACE

This Ph.D. dissertation is composed of four chapters written as separate manuscripts to be submitted for publication, therefore, some duplication was unavoidable. The first and second manuscripts examine the stratigraphy and architecture of extant coastal systems. Emphasis is placed on tidal inlet/tidal delta and lower shoreface deposits because they have the highest preservation potential among coastal lithosomes. The first manuscript relates the development of a tidal inlet/delta system to the evolution of adjacent coastal lithosomes. The second manuscript focuses on the shoreface and inner shelf development in relation to storm sedimentation. Results from these studies provide some insights into: 1) the factors influencing formation and modifications of coastal and inner shelf systems; and 2) the seismic facies, seismic architecture, and lithofacies most likely to be encountered as preserved sequences on the continental shelf.

The third manuscript documents the occurrence, distribution, state of preservation, and nature of Holocene preserved coastal lithosomes on the continental shelf. The distribution and state of preservation of coastal lithosomes are used to define paleoshoreline trends (altitude and location) corresponding to sea-level stillstand events during the overall Holocene transgression. Variations in the architecture and degree of preservation of the preserved sequences provided insights into: 1) the individual impact of and interplay between the factors that influence the formation and preservation of coastal lithosomes; and 2) temporal and spatial variations of these factors.

The final manuscript, presented in the fourth chapter, documents the morphology of the ravinement and pre-transgressive surfaces and the nature of two of the several shelf sand banks. The efficiency of shoreface ravinement,
sediment budget during transgression, the controls on the mechanism of
migration of coastal lithosomes across the shelf, and the resultant shelf sand
bank formation are examined. These are then related to the development of
present coastal systems.

The submitted manuscripts will have John Anderson as co-author; the many
hours of discussion with him led to the development and refinement of the ideas
presented in these papers. Also, his patience in sorting through jumbled
thoughts led to their present readable form.
INTRODUCTION

Shorelines have migrated across the world's continental shelves and inland seas several times throughout earth's history. Shoreline migration is linked to the rise of sea level during the last 18 ky. As the environments of coastal deposition shifted landward, deposition of a transgressive sequence of sediments occurred across the continental shelf. Several important questions related to this shift can be addressed: 1) how do the present coastal lithosomes fit into the overall transgression?; 2) what is the sedimentary record of the transgression on the continental shelf?; 3) what are the factors that influence coastal lithosome preservation?; and 4) are there differences between the types of shorelines and the sedimentary sequences that developed during migration of the shorelines across the shelf? Answers to these questions would improve our understanding of: 1) the nature of Holocene sea-level rise; 2) the character of the induced coastal retreat; 3) the influence of sea-level rise on coastal and estuarine evolution and inner shelf sedimentation; and 4) the lithofacies and facies architecture of estuarine, coastal, and inner shelf lithosomes and processes.

The study area is within the northern Gulf of Mexico, along the east Texas coast and adjacent continental shelf (Fig. 1). It extends from Sabine Pass at the Texas-Louisiana border to San Luis Pass at the western end of Galveston Island, and includes the inner to middle continental shelf, coastal, and estuarine systems. The east Texas coast and adjacent continental shelf is an excellent site to address the questions cited above for several reasons. First, the present coast provides a variety of well-studied lithosomes that may behave differently in response to sea-level rise and that may serve as analogues for preserved sequences on the shelf. Some of these lithosomes include chenier plains (Gould and Mc Farlan, 1959; Byrne et al., 1959; Johnson, 1979), barrier islands (LeBlanc
and Hodgson, 1959; Bernard et al, 1959; 1970; Morton and McGowen, 1980), tidal-inlets and tidal-deltas (Eyer, 1984; Israel et al., 1987); shoreface deposits (Bernard et al., 1959; 1970; Williams et al., 1979), wave-dominated deltas (Bernard et al., 1970; Bartek et al., 1990; 1991), and bay/estuarine systems (Kane, 1959; Rehkemper, 1969; Smyth, 1991; Anderson et al., 1991a, 1991b).

Second, incised valleys on the continental shelf, previously mapped by Nelson and Bray (1970), Pearson et al. (1986), and Thomas (1990), have the greatest potential of yielding preserved coastal lithosomes (Kraft et al., 1987; Belknap and Kraft, 1981, and 1985) as documented in the study area by Thomas (1990).

Third, studies of Holocene sea-level changes in the region may provide a reference for the ages of sea-level events (Curray, 1960; Rehkemper, 1969; Nelson and Bray, 1970; Frazier, 1974; Thomas, 1990) (Fig. 2). Finally, varying shelf gradient along the Texas coast (-40 m water depth), from 0.21 m/km in the east to 0.49 m/km in the west, provides an opportunity to evaluate the role of shelf gradient in the evolution, preservation, and migrational mechanism of coastal lithosomes.
CHAPTER 1

SEISMIC FACIES, ARCHITECTURE AND EVOLUTION OF THE
BOLIVAR ROADS TIDAL INLET/DELTA COMPLEX,
EAST TEXAS GULF COAST
CHAPTER SYNOPSIS

The seismic facies, facies architecture, and stratigraphy of Bolivar Roads tidal inlet and tidal delta deposits and their relation to the development of adjacent coastal lithosomes are examined using high resolution seismic profiles, vibracores, and bore hole descriptions. The Bolivar Roads tidal inlet/delta complex formed approximately 3.3 ka due to spit accretion across the baymouth following a rapid sea-level rise 4 ka. An increase of tidal prism through time and entrenchment of the tidal inlet over the Trinity River incised valley caused inlet stabilization and intensification of tidal processes on the tidal inlet/delta complex.

The tidal inlet facies exhibits channel-stacking and cut-and-fill structure. Stacked clinoforms dip westward across the inlet. The spit/inlet facies is characterized by oblique-tangential clinoforms that build outward and deepen from the valley edge toward the valley center. The flood-tidal delta facies has a base that abruptly shallows bayward. As the flood-tidal delta facies thins bayward, it interfingers with bay sediments. The flood-tidal delta inlet-proximal region exhibits channel cut-and-fill with an overall channel stacking pattern. On the seaward side, the channels have a trough-like geometry. Bayward, the channels broaden and shallows. The channels exhibit a prograded-fill pattern. The ebb-tidal delta facies exhibits gently inclined clinoforms prograding over the ravinement surface.

The tidal inlet deposits are composed of sand, shell, and mud interbeds. Sand and clay interlaminations are ubiquitous in the tidal-deltas. Sand and shell beds are common in the inlet-proximal regions. Overall, the Bolivar Roads tidal inlet/delta complex is mud-dominated as a result of a high influx of fine sediment into the bay.
INTRODUCTION

High resolution seismic (3.5 kHz and uniboom) data, coupled with vibracore and borehole data, allow 3-dimensional study of a tidal inlet/tidal delta complex (Fig. 1.1). In this paper, the seismic facies, lithofacies, facies architecture, and stratigraphy of the Bolivar Roads tidal inlet and tidal delta deposits are examined. Their development is related to the overall evolution of Galveston Bay and adjacent Galveston Island and Bolivar Peninsula.

Tidal inlet and inlet-related deposits may account for 30-50% of Holocene subsurface strata of several modern barrier islands (Moslow and Tye, 1985). Their high preservation potential (Belknap and Kraft, 1981; 1985) makes it probable that they represent a significant percentage of the facies preserved within coastal systems in the rock record (Barwis and Makurath, 1978; Boothroyd, 1985). Thus, recognition of inlet and inlet-related sediments in subsurface and in outcrop is extremely important. An understanding of sedimentary characteristics and variations of inlet-fill sequences in modern environments facilitates identification of ancient equivalents and the processes responsible for their deposition, and aids in prediction of lateral facies changes within subsurface marine shoreline sand-bodies.

Coastal Processes

The study area is located along the east Texas coast (Fig. 1.1). It is characterized by dominantly diurnal tides with a mean tidal range from 45 to 60 cm (Morton and McGowen, 1980). Off Galveston Island, in water depths of approximately 5 m, mean significant wave height is 43 cm and mean wave period is 6 s (Thompson, 1977). Wave heights are less than 1 m for 77% of the year (Hall, 1976). These parameters fall under the mixed energy, wave-
Figure 1.1. a) Geographic and location map of study area. b) Locations of high resolution seismic lines (uniboom and 3.5 kHz), vibracores, gravity cores, and selected boreholes used in this study. Abandoned dredged-spoil dump sites are indicated by boxes. Segments of seismic data (bold lines) and cores with figure numbers refer to seismic profiles and cores shown in subsequent figures.
dominated field of Davis and Hayes (1984). The configuration of the east Texas shoreline, coupled with prevailing southeast winds and wave approach, commonly produces southwesterly littoral drift along the northern and western Texas coast (Lohse, 1955). During winter months, relatively high-velocity winds with strong northerly components are common. They are associated with cold fronts that come from the northwest and move southeast along the Gulf Coast. An average of 47 cold fronts pass through the Texas coast each year (Henry, 1979). Strongest winds occur during tropical storms and hurricanes that strike the Texas coast about once every 1.5 years (Hayes, 1967).

Bolivar Roads, the largest tidal inlet on the Texas coast, is the main tidal pass to the Galveston Bay system (Galveston, Trinity, East, and West bays) (Fig. 1.1), the largest lagoonal estuary on the Texas coast. It is about three kilometers wide and historically had a maximum depth of almost 16 m (Mason, 1981; Eyer, 1984). Jetty construction reduced its depth to a maximum of 7 m. Now, it is maintained by dredging to a minimum depth of 11 m (Mason, 1981).

The mean tidal range in the Bolivar Roads area is 43 cm (Eyer, 1984). Tide measurements since 1852 indicate no net increase or decrease in mean tidal range (Mason, 1981). The average maximum tide velocities are 140 cm/sec for ebb and 94 cm/sec for flood (Eyer, 1984; Harwood, 1973). Harwood (1973) calculated lower tidal current velocities for Bolivar Roads prior to jetty construction. Tidal flow segregation (different flow pathways during ebb and flood flows) occurred before and after jetty construction, but is more evident in the pre-jettied state (Fig. 1.2).

The average diurnal tidal prism passing through Bolivar Roads is approximately 3.0 ×10^8 m³ (Mason, 1981), or 85% of the tidal prism of the Galveston Bay system (Fisher et al., 1972). During a cold front passage, the
Figure 1.2. Morphology of Bolivar Roads tidal inlet and tidal deltas.
a) Present morphology (bathymetric data is from NOAA 11326, 1976 edition). Boxed area denotes approximate area covered on Fig. 1.2b. b) 1867 morphology prior to man-made structures (from Eyer, 1984). Tidal current flow directions are indicated by arrows (compiled from Hall, 1976 and Eyer, 1984).
tidal prism can be five times greater (Mason, 1981); during hurricanes it can increase by a factor of 11 (Corps of Engineers, 1942 in Eyer, 1984).

**Geomorphic Features**

The natural morphology of the Bolivar Roads tidal inlet and associated delta complexes, outlined by bathymetric contours, show well-developed ebb- and flood-tidal deltas with comparable areal extents (Fig. 1.2b). The present morphology resulted from human modifications (Fig. 1.2a). Pelican Island, a natural emergent portion of the flood-tidal delta, was enlarged with dredged-spoil. Bayward extension of the bifurcating channels in the flood-tidal delta may be due to enhanced scouring caused by increased velocities and decreased sediment input from the Gulf of Mexico into the bay. Eyer (1984) and Paine and Morton (1986) present a summary of anthropogenic alterations and ensuing morphologic changes within the tidal deltas and tidal inlet.

Stunted, drumstick-shaped terminations of Galveston Island and Bolivar Peninsula at Bolivar Roads (Fig. 1.2a), and a robust pre-jetty ebb-tidal delta (Fig. 1.2b) are features normally associated with mixed-energy tide-dominated shorelines (Fitzgerald et al., 1984). A robust ebb-tidal delta at Bolivar Roads (Fig. 1.2) results from a large tidal prism, ebb-dominance (ebb current velocities can be 33% to 50% greater than flood currents), and frequent frontal passages that push water out of the bay (Hall, 1976; Eyer, 1984). Tidal current dominance over wave energy helps to confine the inlet, resulting in its stability (Price, 1952; Mason, 1981; Eyer, 1984).
**Surface Sediment Distribution**

A surface sediment distribution map generated by White et al. (1985) shows that deeper portions of the inlet are covered with sandy mud, while shallower regions generally are covered by sand with varying proportions of clay and shell. Muddy and shelly sand also occurs in the bayward throat of the inlet and its bifurcating channels. Sandy muddy shell to shelly sand covers the flood ramp. Muddy sand, the dominant surface sediment in the more proximal region of the flood-tidal delta, is replaced by sandy mud in the more distal areas.

The present ebb-tidal delta exhibits asymmetry in surface sediment distribution; it is sand-dominated southwest of south jetty and mud-dominated northeast of north jetty. The asymmetry reflects the difference in the wave energy expended on either side of the jetties. Wave statistics show breaker heights off Galveston Island consistently larger than those off Bolivar Peninsula (Hall, 1976). The jetties attenuate and block waves from the south and southeast, protecting the area northeast of the north jetty. The ebb-delta itself may act as a protective barrier on its eastern side. A two-fold increase in shelf gradient off Bolivar Peninsula, as compared to offshore Galveston Island, is also a contributing factor. Another component influencing asymmetric sediment distribution is the amount and direction of sand input. Sediments from the northeast are derived from bays and eroding clayey segments of the coastline, while sediments from the southwest, brought in during longshore current reversals, consist of sands from the eroding segments of Galveston Island and the Brazos Delta.

**Late Holocene Sea-Level Changes and General Stratigraphy**

The estimated mean regional historical sea-level rate of rise for the Gulf of Mexico is 0.23 cm/y (Gornitz and Lebedeff, 1987). The mean global eustatic
rise ranges from 0.12 cm/y (Gornitz and Lebedeff, 1987) to 0.18 cm/y (Douglas, 1991).

The historical subsidence rate for the Texas coast (Galveston Area), calculated from tide gauges, is 0.62 cm/y for the period between 1908 to 1980 (Penland et al., 1987). An increase in subsidence rates to 1.17 cm/y during the early 1960's (Swanson and Thurlow, 1973; Penland et al., 1987) is attributed to fluid withdrawal (Gabrysch and Bonnet, 1975; White et al., 1985). Compaction of the valley-fill and barrier sediments should be a major component of the measured subsidence. Thus, estimates of subsidence rates from Galveston Island may not be applicable to the entire east Texas coast because the tide gauges are located above the Trinity River incised valley (Fig. 1.3 and 1.4). Sediment loading of the valley fill by Galveston Island and Bolivar Peninsula and tidal delta deposits contribute to compaction of the valley fill. A long-term subsidence rate for the Texas inner continental shelf is 0.01 cm/y or less (Winker, 1979; Paine, 1991). This probably is more in line with subsidence rates in the interfluve areas of the shelf. Overall, the east Texas coast is retreating in response to relative sea-level rise (Morton, 1977; Paine and Morton, 1989).

Bolivar Roads and Galveston Bay formed above the Trinity River incised valley (Fig. 1.3 and 1.4). The valley was last reincised during oxygen isotope Stage 2 lowstand (Thomas, 1990), when sea-level fell to approximately -126 m below its present level (Fairbanks, 1989). The valley thalweg lies approximately 55 m below present sea level. Offshore, the depth of incision of is 35 to 40 m below the sea floor (Thomas, 1990). Back-filling of the incised valley occurred during the Holocene transgression. Under Bolivar Roads, fluvial sand fills one-third of the valley. Overlying estuarine sediments contain
Figure 1.3. Structure contour map of the Holocene-Pleistocene surface depicting the trace of Trinity River incised valley beneath Galveston Bay and adjacent offshore areas (modified from Smyth, 1991). Tide gauges used by Swanson and Thurlow (1973) to calculate the rate of relative sea-level in the Galveston area are marked (a) Pleasure Pier and (b) Galveston Channel Pier 21.
Figure 1.4. Cross-section from Bolivar Peninsula across Bolivar Roads to Galveston Island showing valley-fill stratigraphy and superposition of Bolivar Roads over the Trinity River incised valley based on borehole data from the U. S. Army Corps of Engineers (modified from Morton and McGowen, 1980). Figure 1.1b shows the profile location.
peat layers and pods of gray, silty to clayey sands, interpreted as bay-head delta deposits. Tidal-inlet/flood-delta sand, capped by barrier/peninsula and extant tidal-inlet sand, occurs in the upper part of the estuarine sequence.

Studies on the Texas continental shelf and coast indicate an episodic or step-like rise of sea level during the Holocene transgression (Curray, 1960; Rehkemper, 1969; Nelson and Bray, 1970; Frazier, 1974; Anderson and Thomas, 1991). This episodic sea-level rise produced backstepping parasequences within incised valleys on the east Texas shelf (Thomas and Anderson, 1989; Anderson and Thomas, 1991). Relative sea-level stillstands are recorded by the aggradation of coeval bayhead delta and tidal inlet lithosomes; rapid sea-level rises are represented by flooding surfaces that separate backstepping parasequences (Anderson and Thomas, 1991). A rapid sea-level rise submerged the broad shallow portions (~6 m and above) of Galveston Bay approximately 4 ka (Fig. 1.3) resulting in the establishment of the present broad, shallow estuarine/bay system (Anderson et al., 1991a; 1991b; Smyth, 1991). Present coastal lithosomes along the east Texas coast may have evolved during the past 3.5 ky of relative sea-level stillstand (Gould and McFarlan, 1959; Bernard et al., 1970; Cole and Anderson, 1982).

**DATA BASE AND METHODOLOGY**

High resolution (Uniboom and 3.5 kHz) seismic profiles were obtained over the present Bolivar Roads tidal inlet and delta complex (Fig. 1.1). Seismic data were acquired with an EG&G uniboom and an EDO 3.5 kHz subbottom profiling system using a towed transducer. Some of these data were acquired digitally. Output power for the uniboom varied between 200 and 500 joules with the band pass filter set at 300 to 2500 Hz. Signals were received either by a single
hydrophone or a seven hydrophone single channel streamer. At times, stratigraphic resolution of 30 cm was achieved. More commonly, resolution was approximately 50 cm. Uniboom lines from a sand resources study conducted by the Corps of Engineers Research Center (CERC) augment the data set (Williams et al., 1979). An average interval velocity of 1525 m/s for valley-fill sediments, calibrated by correlation to borings, was used for depth conversions.

Vibracores and gravity cores provided lithofacies control (Fig. 1.1). Descriptions of borings from navigational channel construction projects by the U. S. Army Corps of Engineers (USCE), CERC vibracores (Williams et al., 1979) and lithologic data from previous investigations in the area (Bernard et al., 1970; Rehkemper, 1969) augmented the core data base. Seismic profiling and vibracore collection were conducted aboard the R/V Lone Star, operated by Rice University. Navigation consisted of both GPS and Loran C. Cores and seismic profiles obtained prior to 1990 were collected aboard R/V Matagorda, using Loran C and radar for navigation.

RESULTS and DISCUSSION

Seismic Facies, Lithofacies, and Facies Architecture

Seismic resolution of 30 to 50 cm allows recognition of structures (e.g., channels, bedsets, master beds) that portray the facies architecture, geometry, and lateral facies changes of the Bolivar Roads tidal inlet/delta complex (Fig. 1.5). The following results and discussions are based on the analysis of seismic data, cores, and borehole descriptions.

Tidal Inlet Facies. - Seismic profiles across Bolivar Roads exhibit broad-based (>4 km wide) channel cut-and-fill structure (Fig. 1.6a). The youngest channel displays the deepest incision (about 23 m). Pleistocene sediments to
Figure 1.5. Oblique line drawings of seismic profiles BL-1 (a), CERC-Line 885-972 (b), and BR-4 (c). Portions of seismic profiles presented in succeeding figures are indicated by the boxes. Figure 1.1b shows locations of profiles.
Figure 1.6. Seismic profiles (uniboom) across Bolivar Roads tidal inlet/delta complex. a) Across the present tidal inlet. b) Strike-section of the spit/inlet facies behind Bolivar Peninsula. c) Strike-section of the flood-tidal delta inlet-proximal facies. Figures 1.1b and 1.5. show locations of profiles.
the west appear to be inhibiting farther westward thalweg migration. Westward dipping clinoforms (0.5° - 4.6°, usually approximately 1°) define lateral accretion. Clinoforms may downlap abruptly into the channel base. To the west, clinoforms downlap the Holocene-Pleistocene surface. Along-dip, the inlet base is scalloped (width of approximately 1 km, relief of up to 3 m), but displays an overall undulating nature. Landward and seaward dipping clinoforms occur.

U. S. Army Corps of Engineers' boreholes within the inlet penetrated a clayey silty sand to silty sand inlet-fill. A pre-jetty surface sediment distribution map shows that the shallow and deep portions of the inlet were covered with sand, while the intermediate depths were covered with mud (Morton and McGowen, 1980). Vibracores taken within the present inlet exhibit silty to clayey shelly sand with sandy clay interbeds (Fig. 1.7). BRFTD-10 to 12-92, with a maximum recovery of more than 3 m, penetrated interlaminated to interbedded fine sand and clay in approximately equal quantities (Fig. 1.7). Vibracore BRFTD-10-92, taken in a water depth of about 8 m on the eastern flank of the present tidal channel, also penetrated medium- to thin-bedded shell layers separated by medium to very thin beds of clay in the lower third of the core. Shell beds consist of a mixed faunal assemblage, are grain supported, and exhibit inverse grading.

The occurrence of mud in the Bolivar Roads inlet-fill is not unique. Moslow and Tye's (1985) model of a tide-dominated tidal-inlet sequence includes interbedded sand and mud. In Doboy Sound inlet, Georgia, Hoyt and Henry (1967) documented the occurrence of silt and clay layers in the inlet-fill. Finklestein (1988) conjectured that a paucity of sand on the barriers and nearshore zone most likely will result in permanent inlet-fill deposits dominated by mud.
Figure 1.7. Lithologic logs of vibracores from the Bolivar Roads flood-tidal delta and tidal inlet. Inset shows the locations of cores. Stippled area indicates extent of present flood-tidal delta
The common occurrence of mud in the Bolivar Roads inlet-fill is probably related to the dominance of tidal processes over wave processes within the inlet and to the abundance of fine-grained sediment supplied to the inlet. It has been shown that increasing concentrations of fine-grained sediments in an estuary increases flocculation exponentially (Mehta, 1989; Pejrup, 1991).

**Spit/Inlet Facies.** - On the backside of the western end of Bolivar Peninsula, a package of westward-dipping clinoforms prograde and thicken from the valley edge toward the valley center (Figs. 1.5a, 1.5b, and 1.6b). The base of this facies progressively deepens toward the incised valley axis. Along-strike, this facies consists of dipping (0.5° - 2.5°, usually about 1°) reflectors that terminate updip by toplap at the nearly horizontal upper surface. Reflectors either pass laterally into thinner bottomset segments or downlap against the channel base. Reflector geometry defines oblique-tangential progradation. Along-dip, this facies displays a more parallel reflector pattern with gentler dips. These are onlapped by more horizontal reflectors that represent aggrading bay deposits. The clinoform package is interpreted as a spit/inlet facies, the subaqueous portion of a recurved spit platform, that reflects westward spit growth. The upper bounding surface shows channel-like features with 1 to 2 m of relief and tens of meters of width, interpreted as possible storm overwash channels. Similar seismic facies occur behind Matagorda Peninsula and San Jose Island along the south Texas coast.

The spit/inlet facies consists of clay with sand laminations grading up to fine sand with clay laminations (Fig. 1.7). This sequence, as indicated by the seismic data (Fig. 1.6b), represents spit/inlet progradation over its distal toe. The spit/inlet facies is overlain by approximately 2.5 m of bay mud.
Flood-Tidal Delta Facies. - Bathymetric and surface sediment maps of Galveston Bay outline a small sub-aqueous portion of the flood-tidal delta. Seismic profiles show a flood-tidal delta sequence extending almost 9 km bayward (Figs. 1.5c, 1.6c, and 1.8a). The flood-tidal delta inlet-proximal facies exhibits channel stacking and cut-and-fill (Figs. 1.5b and 1.6c). The channels have a trough-like geometry on the seaward side (2 to 2.5 km wide, 8 m deep), becoming broad and shallow (>4 km wide, <5 m deep) toward the bay (Fig. 1.5a). The channels have a prograded fill which may represent lateral accretion. Clinoforms dip at 1° to 3° on the seaward side and 1° to 0.5° on the bayward side. Accretion is predominantly to the southwest. Some channels exhibit bi-directional accretion. Broad and shallow channels also exhibit divergent fill.

In dip-section, the basal channel exhibits an initial deepening (to -16 m), then an abrupt shallowing in the bayward direction (Figs. 1.5c and 1.8a). In less than a kilometer, the channel shallows to -7 m. The thalweg cuts into stiff Pleistocene sediments, which probably control the depth of erosion. The channel-fill exhibits ebb-oriented clinoforms. A set of flood-oriented prograding clinoforms (3.3° - 0.5°) with a slight aggradational element occurs on the thinning limb of the delta toward the bay. This is overlain by another set with ebb-directed clinoforms on the bayward side and flood-directed clinoforms on the seaward side (1° - 0.5°). An erosional surface separates the upper and lower sets. The upper bounding surface of the upper set, the present bay floor, also is erosional. Prograding clinoforms in the lower set can be sub-divided into subunits. Each subunit is characterized by an initially steep reflector
Figure 1.8. a) Dip-section on the flood-tidal delta showing bayward thinning and a concave-upward base. b) Older flood-tidal delta facies correlative to the tidal-inlet package beneath Bolivar Peninsula. c) and d) Uninterpreted and interpreted seismic profile offshore of Bolivar Peninsula across the tidal inlet/ebb-tidal delta channel, respectively. This sequence is correlative to the inlet sequence beneath Bolivar Peninsula and the older flood-tidal delta. Figures 1.1 and 1.5 show the locations of the profiles.
V.E. = 25x

Bay / Estuarine Facies

Landward-Dipping Clinoforms

Holocene-Pleistocene Boundary
followed by a train of progressively flattening reflectors that may represent episodic delta build up and/or lateral shifting of bayward growing delta lobes.

Vibracores indicate a mud-dominated flood-tidal delta (Fig. 1.7). In general, shell beds are rare, whereas clay and fine sand interlaminations are common. BRFTD-13-92, taken on the flood-ramp, penetrated almost a meter of densely packed shells underlain by clay with fine sand laminations and shell fragments. The shell bed, consisting of granule to pebble-sized shells and shell fragments, exhibits imbrication and may represent an amalgamation of three shell beds, each exhibiting inverse graded bedding. The only other thick shell bed encountered occurs as an almost 50 cm thick unit at the base of core BRFTD-1-91 (Fig. 1.7). This shell bed also is packed densely and imbricated, and consists of granule to pebble-sized shell and shell fragments. Abrupt lateral facies change within the channel facies, over a distance of approximately 150 m, is recorded by cores BRFTD-1-92 and BRFTD-15-92 (Fig. 1.7). BRFTD-1-92, taken in slightly more than 5 m of water, penetrated a sand-dominated sequence. BRFTD-15-92, taken in approximately 4 m of water, penetrated almost 2 m of bioturbated mud with few sand laminations, underlain by clay and sand interlaminations inclined at about 5°.

Cores at the inlet-distal portions of the delta (BRFTD-3 to 7-92) penetrated clay and fine sand interlaminations overlain by approximately 2 m of bioturbated bay mud with sand-filled burrows (Fig. 1.7). The package exhibits coarsening-upward cycles in an overall fining-upward trend. In the more intermediate and proximal regions, the cores show a coarsening upward trend; clay with fine sand laminations grade-up to clay and fine sand interlaminations capped by shelly, clayey fine sand (e.g. BRFTD-3-91 and BRFTD-2-92) (Fig. 1.7).
An older flood-tidal delta sequence (Figs. 1.5b and 1.8b), probably correlative to the inlet sequence beneath Bolivar Peninsula (Fig. 1.4), is preserved partially beneath the present flood-tidal delta. The imaged segment of this flood-tidal delta is an inlet-proximal facies. Radiocarbon dates from samples near the base of this sequence (at -9.3 m of borehole TGB-A) yielded ages of 3112 ±240 yBP and 2375 ±180 yBP (Rehkemper, 1969), which suggest that Bolivar Roads may have formed at least 3.3 ka. Core BRFTD-8-92, with a penetration of approximately 4 m, may have sampled the upper portion of this flood-tidal delta (Fig. 1.7). The core shows more than 2 m of bioturbated bay mud underlain by flood-tidal delta deposit consisting of clay and fine sand interlaminations/interbedding.

The southwestward migration of the tidal inlet and the flood-tidal delta produced the fining-upward trend in the inlet-distal regions; shifting of channels over previously distal areas produced the coarsening-upward trend in the inlet-proximal regions.

The unusual mud-dominance of the Bolivar Roads flood-tidal delta, as with the tidal inlet fill sequence, is attributed to the high influx of fine-grained sediments into the bay. Vibracores from the Sabine Pass flood tidal delta, located east of Bolivar Roads, shows a similar dominance of mud.

**Bay/Estuarine Facies.** - The bay/estuarine facies interferes with and envelopes the flood-tidal delta. It is characterized by even to wavy reflectors (Fig. 1.6b and 1.8a). A high amplitude reflector defines a surface with some relief and is interpreted as a flooding surface. Relief is inherited from the flooded subenvironments. The upper portions of BRFTD-3 to 9-92 are dark greenish gray, slightly to moderately bioturbated, bay/estuarine muds with few
sand laminations. Bay deposits underlying the flood-tidal delta package are sticky to cohesive, dark greenish gray to olive gray clay with varying amounts of sand and carbonaceous material. Faunal assemblage varies depending on the bay/estuarine subenvironment. Key subenvironment faunal indicators, such as *Rangia cuneata*, characterize the upper bay to bay-head delta environment. *Crassostrea virginica* dominates the middle bay assemblage, and a diverse fauna characterizes the lower bay, with *Mulinia lateralis* being most common. *Mulinia*-dominated and *Crassostrea*-dominated shell beds occur occasionally.

**Ebb-Tidal Delta Facies.** - The present-day bathymeric expression of the ebb-tidal delta extends 13 km offshore to the -12 m isobath (Fig. 1.2a). However, an isopach map of marine sediments indicates that ebb-tidal delta influence may extend 25 km offshore (Fig. 1.9). An asymmetry in the thickness of the sediments (thickest section occur on the eastern side of Galveston Island) probably is a result of predominant southwestward longshore transport.

Core CERC-14 reflects ebb-tidal delta progradation over the inner shelf; the ebb-tidal delta package shows a coarsening upward trend (Fig. 1.10). Inner shelf sediments are underlain by a ravinement surface marked by a transgressive lag. CERC Line 62-173 and core CERC-12, from offshore Bolivar Peninsula, indicate the presence of an inlet/ebb-delta channel sequence (Fig. 1.8c and 1.10). This sequence correlates with the inlet sequence beneath Bolivar Peninsula (Fig. 1.4) and the older flood-tidal delta (Figs. 1.5b and 1.8c). CERC Line 62-173 shows northwest, gently dipping clinoforms (fix points 81-86) that become eastward-prograding clinoforms with steeper gradients as a sharp turn is made at fix point 86 from a northwest (landward) to an easterly (seaward) heading (Fig. 1.8c). The geometry implies that actual dips are more
Figure 1.9. Isopach map of Holocene marine sediments defines the extent of the ebb-tidal delta. Cores used in measuring the thickness of marine sediments are shown. Boreholes are not included.
Figure 1.10. CERC vibracores from the Bolivar Roads ebb-tidal delta. CERC-14 is from the distal portion of the ebb-tidal delta lobe and CERC-12 is from the eastern end of the preserved inlet/ebb-tidal delta channel offshore of Bolivar Peninsula. Figure 1.1b shows the locations of the cores.
east-northeast, parallel to Bolivar Peninsula's coastline. The channel base shallows seaward and laterally. It exhibits a maximum incision of approximately 16 m. A ravinement surface separates the inlet/ebb-channel package from the overlying extant ebb-tidal delta. Between fix points 81-86, the present ebb-tidal delta downlaps the ravinement surface. In contrast, reflectors between fix points 86-91 define an onlap. CERC-12 shows that the inlet/ebb-channel fill consists of a stack of thick-bedded shelly sands with basal lags and erosive bases (Fig. 1.10). Basal lags consist of large and mixed shell fragments, calcareous nodules, and other granule- to pebble-sized lithic fragments. The ebb-tidal delta package above the ravinement surface exhibits a fining upward trend.

The present asymmetric surface sediment distribution on the ebb-tidal delta, described earlier, is also indicated by cores and boreholes. Although the delta is mainly mud-dominated, there is greater sand accumulation on its western side. This suggests that the asymmetric surface sediment distribution pre-dates jetty construction.

**Evolution of Tidal Inlet/Delta Complex and Adjacent Lithosomes**

The evolution of Bolivar Roads tidal inlet/delta complex is intricately linked to the development of adjacent coastal lithosomes. A 3.3 ka age for the formation of Bolivar Roads is consistent with the establishment of present coastal lithosomes along the east Texas coast during the past 3.5 ky of relative sea-level stillstand, as proposed by Gould and McFarlan (1959), Bernard et al. (1970) and, Cole and Anderson (1982). Core and seismic data, and the surface morphology of Galveston Island and Bolivar Peninsula, indicate that Bolivar Roads tidal inlet evolved by bay-mouth closure through spit growth, possibly from both sides of the paleo-Galveston embayment (Fig. 1.11).
Figure 1.11. Reconstruction of events leading to the formation of Bolivar Roads: a) approximate bay configuration immediately after the 4 ka rapid sea-level rise. Position of pre-4 ka inlet, based on unpublished data; b) Bolivar Roads approximately 3.0 ka. The system is more "wave-dominated", and recurved beach ridges trace the spit progradation from both sides of the baymouth; c) present (pre-jetty) configuration. The inlet has migrated southwest, incised deeper and became more "tide-dominated".
Bolivar Peninsula. - Stages in the development of Bolivar Peninsula are reflected by its surface morphology (Figs. 1.11b and 1.11c). An early stage of development is represented by a series of recurved-slip ridges indicating southwest migration, washover features, and washover-associated channels on the back side of the peninsula. LeBlanc and Hodgson (1959) proposed a spit accretion origin for the peninsula while Eyer (1984) proposed an emergent bar origin. Southwest lateral accretion seen on seismic data behind Bolivar Peninsula supports the former (Figs. 1.5a and 1.5b). The channels that later truncated the ridges may have served as ephemeral tidal inlets at a time when the peninsula was narrower than at present. Hayes (1976) and Pierce (1970) documented the formation of similar narrow, shallow, ephemeral inlets along narrow regions of a barrier island activated during storms. Two prominent lobate features connect with these channels on the back side of Bolivar Peninsula (Fig. 1.1). They are comprised of 1 to 2 m of fine sand and sand and clay interlaminations underlain by bay/lagoonal bioturbated mud with oyster shells and occasional oyster-dominated shell beds. They are capped by marsh deposits. Cores from East Bay (Fig. 1.1), north of the lobate features, indicate that fine sand and clay interlaminations extend just a few hundred meters into the bay. The sand and sand/clay interlaminations are interpreted as overwash/ephemeral flood-tidal delta deposits. Shells from the underlying bay/lagoonal deposits yielded radiocarbon ages of 1980 ± 70 (Beta-43465) and 2160 ± 70 (Beta 43466) yBP.

A later stage in the development of Bolivar Peninsula is represented by a series of nearly continuous beach ridges that parallel the present shoreline (Fig. 1.11c). The ridges are slightly concave seaward, diverge to the southwest, and have recurved ends. Older beach ridges appear to have been eroded at their
northeastern end. The erosion may also have removed the morphologic
evidence for initial spit accretion from the peninsula's northeastern end.

**Galveston Island.** - Beach ridges of the back side of the northeastern half
of Galveston Island (Figs. 1.11b and 1.11c) coalesce into a fan-shaped feature
that widens toward the bay. The geometry indicates that the island initially grew
north-eastward, tracking the shallow portions of the Holocene-Pleistocene
surface, similar to a spit bar. The ridges later rotated toward the Gulf and
seaward accretion occurred. Prior to, and during the early periods of seaward
accretion, breaching of beach ridges occurred frequently and at several sites,
as indicated by numerous washover channels on the backside of the island.

**Bolivar Roads Tidal Inlet.** - With the south-westward growth of Bolivar
Peninsula and north-eastward growth of Galveston Island, the bay mouth
became constricted and an inlet formed approximately 3.3 ka. The 3.5 ka age
for the oldest portion of Galveston Island (Bernard et al., 1970) implies rapid
closure of the baymouth. This is supported by the rapid establishment of
bay/estuarine subenvironments in the Galveston Bay complex (Anderson et al.,
1991a). Reworking of previous coastal lithosomes provided a higher sediment
supply during the early phase of relative stillstand, enhanced spit progradation
and, later, seaward accretion. During bay mouth closure, deepening probably
occurred to maintain the tidal prism passing through the inlet, as predicted by
the tidal prism/inlet area relationships of O'Brien (1976) and Jarret (1976).

The tidal inlet under Bolivar Peninsula and its correlative tidal delta indicate
that Bolivar Roads initially formed a few kilometers to the east of its present
location (Fig. 1.11). Migration to the southwest created vertical and lateral
facies changes in the tidal deltas, including a fining-upward trend in distal
regions of the flood-tidal delta, coarsening-upward in intermediate area, channel stacking in inlet-proximal facies, and ravinement of the inlet/ebb-delta channel off Bolivar Peninsula. The second set of beach ridges on Bolivar Peninsula implies that seaward accretion occurred after inlet migration. Remnants of the distal ebb-tidal delta to the east of Bolivar Roads are non-existent, implying efficient reworking by the ravinement that truncated the inlet/ebb-channel package. Part of the eroded material may have accreted to Bolivar Peninsula and to the pre-jettied ebb-tidal delta.

The inlet migrated southeastward under the impetus of a predominantly southeast longshore current. When the migrating inlet encountered stiffer and more cohesive Pleistocene sediments of the western edge of Trinity River incised valley, it was forced to migrate southwestward, along the valley edge (Fig. 1.3). Continued southwestward growth of Bolivar Peninsula caused the inlet to narrow and incise more deeply into softer estuarine valley-fill (Fig. 1.6) to maintain the tidal prism. The depth of the inlet increased from -16 m to deeper than -20 m. It has since behaved as a mixed-energy, tide-dominated inlet/delta complex.

In addition to the influence exerted by superposition within an incised valley, deepening of Bolivar Roads inlet also may be caused by an increased tidal prism, and consequently, increased tidal currents (Escoffier, 1940). Relative sea-level rise and closure of smaller inlets as the barriers widened, contributed to the increase of the tidal prism.

Price (1952) and Price and Parker (1979) noted that stable inlets along the Texas coast are associated with incised valleys. Sabine Pass, located east of Galveston Bay (Fig. 1.1b), is another example of a stable inlet positioned above an incised valley (Mason, 1981). However, tidal inlet/incised valley association
will not lead necessarily to inlet stability. Seismic lines behind Matagorda Peninsula and San Jose Island (Fig. 1.1b) indicate the presence of abandoned tidal inlets occurring above incised valleys. Corpus Christi, San Antonio, and Baffin bays, are other Texas bays located over incised valleys (Wright, 1980; Morton and McGowen, 1980) that presently do not have tidal inlets over the valley. One possible factor contributing to stability of Bolivar Roads and Sabine Pass is efficient passage of tidal prism through the inlets. Along the east Texas coast, wave approach is more perpendicular to the coastline, and the north-south bay orientation enhances the effectiveness of strong northerly winds in pushing bay waters out of these inlets during cold front passages. Along the south Texas coast, wave approach is more oblique. This may cause higher longshore currents that would result to less stable inlets. Also, the maximum fetch of the bays along the south Texas coast trend northwest-southeast, hence strong northerly winds are less effective in pushing water out of the bays. This may partly be the reason for the small and blunted ebb-tidal deltas of the south Texas coast.

SUMMARY AND CONCLUSIONS

Bolivar Peninsula initially was a thin, narrow and probably low feature, formed by southwestward spit accretion. The peninsula later was breached by storm overwash channels that served as ephemeral tidal inlets. Reactivations ceased as the peninsula widened by beach ridge accretion.

A rapid sea-level rise 4 ka, flooded the Trinity River incised valley. During the ensuing relative sea-level stillstand starting at approximately 3.5 ka, Bolivar Peninsula and Galveston Island grew across the bay mouth by spit accretion, resulting in the formation of Bolivar Roads around 3.3 ka. Rapid closure of the bay mouth led to quick establishment of bay subenvironments. Under the impetus of a dominant southwestward longshore current, the inlet migrated to the southwest. However, incised valley flank obstruction impeded inlet migration and caused further constriction of the inlet. In addition, relative sea-level rise and closure of smaller inlets increased the tidal prism resulting in enhanced tidal current velocities and deeper inlet incision and, ultimately, stabilization of the inlet. Initially wave processes were dominant over tidal processes in the inlet: constriction of the inlet and the increase of tidal prism caused the influence of tidal processes to exceed wave processes in the inlet/delta complex.

The present tidal inlet fill exhibits channel-stacking and cut-and-fill structure composed of sand, shell, and mud interbeds. Individual channels are filled with westward-dipping clinoforms that prograde and downlap the channel base. The spit/inlet facies is characterized by oblique-tangential progradation, building-out and deepening from the valley edge toward the valley center. In dip-section, prograding clinoforms are more parallel and display gentle dips. The flood-tidal delta facies has a base that abruptly shallows bayward. As the flood-tidal delta
facies thins bayward, it interfingers with bay sediments. The flood-tidal delta inlet-proximal region exhibits channel cut-and-fill with an overall channel stacking pattern. On the seaward side, the channels have a trough-like geometry. Bayward, the channels broaden and shallows. The channels exhibit a prograded-fill pattern that may represent lateral accretion. Sand and clay interlaminations are ubiquitous in the tidal-deltas. Sand and shell beds are common in inlet-proximal regions. The ebb-tidal delta exhibits gently inclined clinoforms prograding over the ravinement surface. Inlet/ebb-delta channel facies show lateral migration with the channel base shoaling in the direction of migration. The inlet/ebb-channel consists of stacked sand beds with basal shell lags; distal portions consist of mud with sand laminations to interbeds. Mud-dominance of Bolivar Roads inlet/delta complex predominantly results from high influx of mud in the area.

Distinctive architecture and geometry of facies comprising Bolivar Roads tidal inlet/delta complex may serve as analogues for subsurface and outcrop interpretations and facilitate predictions of their lateral and vertical facies changes. It provides an example of an inlet sytem that formed through baymouth closure by spit accretion and an inlet system that shows an increase of tide-influence through time as a result of increased tidal prism and influence of pre-Holocene geology.
CHAPTER 2

SHOREFACE AND INNER SHELF STORM DEPOSITS,
EAST TEXAS GULF COAST
CHAPTER SYNOPSIS

Holocene shoreface and inner shelf sedimentation off the east Texas coast on the Gulf of Mexico was examined using high-resolution seismic data and sediment cores. Bathymetric changes that occurred during a span of 26 years or more indicate that erosion has deepened the shoreface and inner shelf of the study area. This, coupled with a relatively low sediment input, has resulted in the overall thinness of the Holocene marine sediment cover.

A paucity of storm deposits characterize the present lower shoreface and inner shelf. This appears to be inconsistent with the abundance of cross-cutting channels on the shoreface wedge, interpreted as storm return flow channels, and the high frequency of storms and hurricanes that have impacted the region during historical times. Strong along-shelf storm currents, low sediment supply, and low effective accommodation space has caused the scarcity of storm deposits in the region. However, greater sand supply to the lower shoreface and inner shelf during the early phase of the establishment of the present coastal lithosomes resulted in a greater occurrence of storm beds lower in the stratigraphic section.

Amalgamated sand beds on the east Texas shelf, potentially of storm origin, are possible preserved lower shoreface deposits and reworked Holocene coastal lithosomes. Amalgamation of storm beds led to the formation of shelf sand banks. Onshore and shore-parallel transport of sand during storms seems to be equally as, or possibly more, important than offshore transport of sand.
INTRODUCTION

The transgressive history and Holocene sedimentation of the east Texas coast and adjacent inner shelf were studied using an extensive grid of vibracores, gravity cores, and piston cores, coupled with high-resolution seismic data (3.5 kHz and boomer) (Fig. 2.1). The specific objectives of this investigation were to examine the distribution and geometry, sourcing and formation, and preservation of storm deposits.

Previous studies of storm deposition on the Texas shelf focused on the impact of specific storm events and on the actual process of storm sedimentation (Hayes, 1967; Morton, 1981; 1988; Snedden, 1985; Snedden, et al., 1988). These studies were conducted on the central Texas shelf where sedimentation rates are the highest for the Texas shelf (Shideler, 1978). The cores used in these studies were either limited by shallow penetration, thus prohibiting any stratigraphic analysis of storm deposits, or had limited coverage, thus regional analysis was not possible.

This study was expected to provide a contrast in storm deposition between the east and central Texas shelf regions; the central Texas shelf has a higher sedimentation rate. Although the present study concentrates on the modern shoreface and inner shelf systems, the data set extends to the outer shelf (Fig. 2.1). Evaluation of Holocene storm sedimentation and the preservation of storm deposits in relation to the overall transgressive history of the region was possible because of the regional core coverage and penetration of the entire Holocene marine section in numerous cores.

This study indicates that major differences in the storm record of the central and east Texas shelf regions reflect variation of storm processes and factors of storm bed preservation affecting the two adjacent regions.
Figure 2.1. a) Geographic and bathymetric map of the Texas coast and shelf. b) Locations of high-resolution seismic lines (boomer and 3.5 kHz), cores, and selected boreholes used in the study area.
STUDY AREA

The Texas coast is characterized by fair weather astronomical tides ranging from 45 to 60 cm and waves of relatively low amplitude with periods most commonly within 4-6 s (Morton and McGowen, 1980). Off Galveston Island, wave heights are lower than 1 m 77% of the year (Hall, 1976). Prevailing winds are from the south and southeast during the spring and summer months. In the fall, winds are more easterly. During winter months, relatively high-velocity offshore winds with strong northerly components are common. They are associated with cold fronts that come from the northwest and move southeast along the Gulf Coast. An average of 47 cold fronts pass through the Texas coast each year (Henry, 1979). The strongest winds occur during tropical storms and hurricanes that strike the Texas coast about once every 1.5 years (Hayes, 1967). Nearly a century of hurricane records indicate that Galveston Island and Bolivar Peninsula are two of the three most hurricane-prone areas along the Gulf Coast (Simpson and Riehl, 1981). Water currents in the area are influenced primarily by wind and tides. The wind, however, is the dominant factor controlling the currents; nearly all major changes in current speed can be related to wind speed or direction (Armstrong, 1979).

The study area extends from San Luis Pass, at the western end of Galveston Island, to Sabine Pass at the Texas-Louisiana border and from the shoreline to water depths of 18 m, approximately 35 km offshore (Fig. 2.1). The shelf gradient (to -18 m) varies from 0.24 m/km to 1.06 m/km from east to west. This shelf gradient is inherited from several Pleistocene episodes of delta progradation and westward shifting of depositional lobes (Abdulah and Anderson, 1991). Present coastal lithosomes along the east Texas coast are believed to have evolved during the past 3.5 ky, or during the latest sea-level
stillstand (Gould and McFarlan, 1959; Bernard et al., 1970; Cole and Anderson, 1982).

The historical subsidence rate for the Galveston area, calculated from tide gauges, is 0.62 cm/y for the period between 1908 to 1980 (Penland et al., 1987). A long-term subsidence rate for the Texas inner continental shelf is 0.01cm/y or less (Winker, 1979; Paine, 1991). This is probably more in line with subsidence rates in the interfluve areas of the shelf.

The estimated mean regional rate of sea-level rise for the Gulf of Mexico is 0.23 cm/y (Gornitz and Lebedeff, 1987). Mean global eustatic rise ranges from 0.12 cm/y (Gornitz and Lebedeff, 1987) to 0.18 cm/y (Douglas, 1991). Overall, the east Texas coast is retreating in response to this relative sea-level rise (Morton, 1977).

**METHODS**

Vibracores (50) and gravity cores (75) were collected on the shoreface and inner shelf with the intent of sampling the entire Holocene marine section to document the overall vertical and lateral facies variations of the Holocene shoreface and inner shelf sequences (Fig. 2.1). Piston cores (143) collected on the inner shelf to the outer shelf provide regional control. These were augmented by more than 100 vibracores from Williams et al.'s (1979) sand resources study of the inner shelf, and Pearson et al.'s (1986) archeological investigation of the offshore Sabine River incised valley. Foundation and navigational channel borehole descriptions by the U.S. Army Corps of Engineers also aided this study. Grain size analysis of the sand-sized fractions were conducted using an automated settling tube (Anderson and Kurtz, 1979).
Approximately 1,300 km of high-resolution seismic data (uniboom and 3.5 kHz) were available for this study (Fig. 2.1). An EG&G uniboom and an EDO 3.5 kHz subbottom profiling system with a towed transducer were utilized for seismic data acquisition. Some of the data were acquired digitally. Uniboom lines from Williams et al. (1979) augment the data set. An average interval velocity of 1525 m/s, calibrated by correlation to borings, was applied for conversion of acoustic travel time to depth within the sediments.

Seismic profiling and core collection were conducted aboard the R/V Lone Star and the R/V Gyre. GPS and Loran C were used for navigation for both vessels. Seismic profiles and cores obtained prior to 1989 were collected aboard R/V Matagorda; Loran C and radar were used for navigation.

RESULTS

**Historical Shoreface and Inner Shelf Erosion**

Bathymetric changes off Galveston Island and Bolivar Peninsula, mapped to a distance of 20 km by comparing 1:80,000 scale 1973 and 1986 bathymetric charts, indicate a general deepening of the region (both charts referenced soundings to mean low water; comparable accuracy of positioning and depth measurements for these two time periods are assumed) (Figs. 2.2 and 2.3). The survey dates for the charts are unverified, thus the actual time period spanned by these two charts is uncertain. However, comparison of the 1973 chart with a 1960 chart indicates that the regions near tidal passes were resurveyed for the 1973 chart. Thus, the bathymetric changes, away from the tidal passes, may reflect changes during a period of 26 years or more.

The bathymetric changes define zones that are parallel to, or slightly oblique to, the coastline (Fig. 2.2). In general, deepening increases towards the
Figure 2.2. Water depth changes based on comparative overlay of bathymetric charts. a) 1973 (NOAA 1282) and 1986 (NOAA 11323) bathymetric charts off Galveston Island and Bolivar Peninsula. b) 1960 (C&GS 1282) and 1973 (NOAA 1282) bathymetric charts off Bolivar Roads. c) 1960 (C&GS 1282) and 1973 (NOAA 1282) bathymetric charts off San Luis Pass. Location of Figures 2.2b and 2.2c are indicated on Figure 2.2a. Also, Figure 2.2a shows the bathymetric transect locations on Figure 2.3. Boxed areas seaward of Bolivar Roads are dredged-sediment dumping grounds. Dashed line indicated by (d) off the western half of Galveston Island outlines the seaward toe (at -16 m) of Big Slough Bathymetric High.
Figure 2.3. Shoreface to inner shelf bathymetric transects from the 1973 and 1986 charts. Location of these transects are indicated on Figure 2.2a.
shoreface base, then decreases farther offshore (the shoreface limit is
demarcated by the first observable break in slope of the shoreface profile where
the slope decreases from ~1:200 to ~1:2000 according to Friedman and
Sanders, 1978 and Swift, et al., 1985). At the shoreface base, maximum
deepening is slightly greater than 2 m. Commonly, deepening is slightly less
than 1 m (Figs. 2.2 and 2.3). Overall, zones of bathymetric deepening on the
shoreface and the inner shelf correlate with more rapidly retreating segments of

Off Bermuda Beach (Fig. 2.2), where the maximum shoreface base and
shoreline translations occur, landward shoreface base retreat of one kilometer
is matched by 0.1 km shoreline erosion (Fig. 2.3). In other areas, shoreface
base translation is more than 10 times greater than shoreline translation.
Extensive deepening at the shoreface base, resulted in the general steepening
of the shoreface profile. This steepening of the shoreface profile may be related
to the decreased rates of shoreline erosion during the 1974 to 1982 period, a
time of low storm incidence (Paine and Morton, 1989), and probably reflects a
time of net onshore transport of sand (similar to Pilkey and Field, 1972).
Regrading of the shoreface during fair weather conditions, in an attempt to
reach ideal wave-grade profile, steepens the shoreface profile (Moody, 1964).

The bathymetric changes show also a general increase of water depth from
east to west. The zone between 5 to 15 km off Bolivar Peninsula has an
average deepening of approximately 0.25 m. Off Galveston Island, away from
the jetties, this same offshore zone has an average deepening of greater than
0.5 m. The offshore decrease in deepening off Galveston Island is interrupted
by a 2 km wide zone, 22 to 24 km offshore in 12 to 16 m water depth,
characterized by deepening of as much as 1.5 m and with an average of greater
than 0.5 m (Fig. 2.2). This zone is the seaward flank of a bathymetric high (2 m relief, 15 km long, and 5 km wide), Big Slough Bathymetric High (Fig. 2.2). Some segments of this flank display a landward shift of as much as 2 km. The landward flank of the bank displays an average deepening of less than 0.25 m.

Concentric zones of erosion centered on the Bolivar Roads jetty mouth show that deepening decreases away from the jetty mouth. Offshore, and southwest of the jetties, there is evidence of shoaling of as much as 1.5 m. Comparison of the 1973 and 1960 charts shows a similar concentric pattern, but deepening is greater (as much as 2 m) and the zones are broader (Fig. 2.2b). This reflects the tripartite pattern of offshore deposition, nearshore erosion, and beach accretion, characteristic of all of the jettied inlets of the Texas coast (Morton, 1977; Mason, 1981). The ebb-tidal delta at San Luis Pass shows deepening of more than 1 m on its western side while its eastern side shows deposition of as much as 2 m (Fig. 2.2c).

The patterns of bathymetric changes cannot be ascribed directly to relative sea-level rise. Although subsidence rates in the study area are highest over the Bolivar Roads because of compaction of Holocene incised valley fill, deepening in this sector is not the highest in the region. Instead, the bathymetric changes can be related to tidal flow dynamics in and around the jetty mouth. Deposition of sediments entrained by ebb-tide currents may account for the shoaling southwest of the jetties: ebb-tide currents are deflected to the southwest by longshore currents. The surface sediment distribution map shows that this area is covered with mud (White et al., 1985) (Fig. 2.4). In contrast, zones on the inner shelf that showed deepening correlate with increased sand content (Fig. 2.4). These patterns, plus other factors discussed in later sections, suggests that observed bathymetric changes are real and
Figure 2.4. Surficial sand distribution (from White et al., 1985).
define zones of erosion and deposition. Morton (1979) and Paine and Morton (1989) point out that the erosion occurring along the Texas coast and bay shorelines is caused by relative sea-level rise. Similar patterns of bathymetric changes on the shoreface and the inner shelf elsewhere, such as off Smith Island, Virginia (Everts, 1987) and off the Louisiana coast (Penland et al., 1985), have also been attributed to relative sea-level rise.

Erosion along the shoreface and inner shelf off the east Texas coast is not a recent trend. During an 80 year monitoring period (1878 to 1958), the U. S. Army Corps of Engineers estimated $3.8 \times 10^6$ m$^3$/y of sediment was eroded offshore of Galveston Island (LeBlanc and Hodgson, 1959). Likewise, from 1853 to 1933, the San Luis Pass ebb-tidal delta was reduced by $8.41 \times 10^6$ m$^3$ (Mason, 1981). The shoreline also was undergoing net erosion during this period (Morton, 1974, 1975).

Currents capable of eroding the inner shelf are common. Fair weather bottom currents in about -17 m water depth, measured 1 m above the sea floor and 10 km southwest of the Bolivar Roads jetty mouth, yielded velocities ranging from 0 to 45.3 cm/s with an average velocity of 15.4 cm/s (Hall, 1976). Directions of these currents were concurrent with the tidal cycle; maximum bottom currents flowed toward the southwest and occurred during the ebb tide of a diurnal tidal cycle. Flood currents flow in a north to northeasterly direction. Tropical storm Delia, with sustained winds less than 250 cm/s, produced alongshore (southwest) bottom currents up to 200 cm/s in 18 m of water (Forristall et al., 1978). Summaries of storm induced currents within the study area and adjacent regions are presented by Morton (1981; 1988) and Snedden et al. (1988).
**Thickness of Holocene Marine Sediments**

An isopach map of Holocene marine sediment shows an overall thinness of present lower shoreface and adjacent inner shelf sediments (Fig. 2.5). The thickest Holocene sediment cover occurs offshore of Galveston Island and adjacent ebb-tidal delta, and contrasts with a very thin veneer of Holocene sediment to the east. Off Galveston Island, the Holocene sediment cover is greater than 25 cm thick to a distance of 17 km from shore. In contrast, off Bolivar Peninsula and High Island, the Holocene sediment cover decreases to less than 25 cm in less than 2 km from shore. Farther to the east, it thickens again because of the Sabine Pass ebb-tidal delta lobe.

One reason for the thinness of the Holocene marine deposits is the erosion documented by historical data (Figs. 2.2 and 2.3). Another reason is the lack of direct sediment input by rivers along the east Texas coast. Sediment loads of the Trinity and Sabine rivers are trapped in their respective estuaries (Galveston Bay and Sabine Lake) (Fig. 2.1).

Contrasts in sediment thickness between the eastern and western regions of the study area suggests that sediment supply is higher offshore of Galveston Island (Fig. 2.5). This results from the westward deflection of the sediment-laden ebb currents coming out of Bolivar Roads by the predominantly southwest longshore currents. Thus, there is a southwestward thickening of the Bolivar Roads ebb-tidal delta (Figs. 2.4 and 2.5) (Chapter 1). This also is indicated by the broad depositional zone located southwest of the jetties (Fig. 2.2). Occasional reversals in longshore current direction delivers sediment from the Brazos River Delta. An additional sediment source is from dredged sediment from Bolivar Roads that is dumped south of the jetty mouth (Fig. 2.2).
Figure 2.5. Isopach map of Holocene marine sediments. Location of cross-sections in Figure 2.6 are indicated by dashed lines.
Another factor leading to the thin sediment cover on the east Texas shelf is the relatively shallow Holocene-Pleistocene surface. Because of this shallow surface, the effective accommodation space, accommodation space below storm wave-base, is relatively small. The thinner sediment cover to the east of Bolivar Roads may be due to shallower Holocene-Pleistocene surface in that area (Fig. 2.6a to 2.6f). Greater effective accommodation space to the west has allowed thicker sediment accumulation.

**Shoreface and Inner Shelf Sediments**

The present observed depth of the shoreface along the east Texas coast is -4 m off High Island, -5 to -7 m off Bolivar Peninsula, and -9 m off Galveston Island. Off Galveston Island, the upper shoreface limit is at -3 to -4 m, while off Bolivar Peninsula it is shallower than -3 m. These depths correspond to Hallermier’s (1981) littoral limit, defined as the seaward limit of extreme surf related effects. The limit is calculated based on wave and sand characteristics. Sediments on the physiographically defined shoreface commonly consists of >60% sand (Fig. 2.4).

The upper shoreface sediments of the study area consist of amalgamated shelly sand beds, normally with less than 5% silt and clay content. The beds commonly have erosional bases and shell lags (Fig. 2.7). Sand-sized fractions have a mean grain size range of 2.4φ to 2.9φ and a modal range of 2.75φ to 3.25φ, respectively. Very thin shell beds consisting mostly or entirely of *Mulinia* shells are common. Burrows and inclined laminations are rare.

The lower shoreface consists of slightly to moderately bioturbated, dark greenish gray, interbedded sand and mud (Fig. 2.7). Downcore and offshore, sand layers are usually very thin-bedded (3 to 1 cm thick) to laminated (less
Figure 2.6. Shoreface/inner shelf wedge cross-sections. Location of transects are indicated on Figure 4. Vertical exaggeration of cross-sections is 376X.

**LEGEND:**

- **Brazos River Mud**
- **Shoreface Sand**
- **Inner Shelf Mud**
- **Ebb-Tidal Delta**
- **Freshwater Marsh**
- **Outer Shelf Sands**
- **Ravinement Surface**
- **Holocene / Pleistocene Ravinement Surface**
Diagram e: Depth in Meters vs Distance From Shoreline for OBP-11, OBP-12, OBP-13, OBP-14, OBP-15, OBP-16, OBP-90-3, and OBP-90-14.

Diagram f: Depth in Meters vs Distance From Shoreline for OHI-5, OHI-6, GY-121, BOREHOLE, PLEISTOCENE SEDIMENT, and VE=376x.
than 1 cm thick) and may consist of as much as 40% silt and clay. Sand-sized fractions have a mean grain size range of 2.35φ to 3.4φ and modal range of 2.75φ to 3.75φ. Very thin- to medium-bedded (30 to 10 cm thick) shell layers with mixed shell assemblages (Mulinia, Anadara, Donax, Crassostrea and others) are common. Shell beds with mostly or entirely Mulinia shells are rare. Shell layers may grade up to sand. Off High Island, the lower shoreface consists of brownish green mud with occasional silty very fine sand laminations.

Inner shelf sediments of the western portion of the shelf are bioturbated, dark greenish gray mud (50 to 80% water content) with variable amounts of shell and sand (Fig. 2.7). Burrows are filled with sand and/or shell fragments. On the eastern portion of the shelf, toward Sabine Pass, the inner shelf sediments are dark yellowish green and exhibit consistently high (~80%) water content. The color change suggests a change in sediment source, and the very high water content implies recent deposition.

Surficial and patchy concentrations of sand-sized sediment, commonly between 40 to 80%, correlate with the areas where erosion has occurred: there is enrichment of heavy minerals, glauconitic sand, and shell hash which indicates winnowing of fine sediments. Sand fractions generally have a mode of either 2.75φ or 3.00φ; mean grain size averages 2.74φ. In contrast, in regions where accumulation occurred, surface sediments have lower sand content; sand fractions typically have a mode of 3.50φ or 3.75φ and a mean grain size average of 3.08φ.

Within the study area, storm deposits, defined mainly by an erosional base and a fining upward trend, typically are laminated to very thin-bedded. The thickest storm bed encountered is a 25 cm thick, tightly packed, imbricated, shell bed. Generally, sand fractions have a mode of either 3.00φ or 3.75φ, and
Figure 2.7. Lithologic logs of cores along transects 2.6d and 2.6e of Fig. 2.6.
Profile E

Core Length (m)

0.00 0.25 0.50 0.75 1.00 1.25 1.50

Clay and Silt Lamina
Sand Lamina

Holocene-Pleistocene Surface
Bioturbation
Shell Fragments
Peat/Plant Fragments

Normal Grading
rarely 2.75ϕ. Beds with a sand fraction mode of 3.75ϕ commonly have parallel or inclined laminations. Thick (≤ 30 cm), amalgamated storm beds are absent in the lower shoreface and on the inner shelf. In general, shell components of storm beds decrease offshore. However, thickness and frequency of shell lags initially increases from upper shoreface to lower shoreface. Seaward, *Mulinia* becomes less abundant to absent, while *Anadara* becomes more abundant. *Rangia* and *Crassostrea* are common in regions with very thin sediment cover, indicating reworking from underlying non-marine sediments. Basal shelly layers are common in the lower portions of the inner shelf package.

Morton and Winker's (1979) study of the coarse components of inner shelf surficial sediments on the Texas shelf showed that there are more reworked materials derived from the underlying Pleistocene sediments off Bolivar Peninsula than off Galveston island. This may indicate that the sediments off Bolivar Peninsula are constantly reworked. Perhaps one reason why there are less bathymetric changes on the inner shelf off Bolivar Peninsula (less than 0.25 m deepening) is because of the thin sediment cover and its constant reworking.

**Facies Architecture of Holocene Marine Sediments**

Shoreface and inner shelf sediments of the study area prograded over Holocene bay/estuarine sediments within incised valleys and over Pleistocene fluvial-deltaic sediments in interfluve areas. The shoreface and inner shelf wedge shows interdigitation of shoreface sand and inner shelf mud that is most evident on Galveston Island (Figs. 2.6 and 2.7). Seaward, sand layers are thin and isolated. These sand lenses consist mostly of sharp-based, normal-graded sand and shell, indicating that they are storm deposits. Overall, the
shoreface/inner shelf wedge cross-sections show a coarsening upward trend typical of progradational systems (Figs. 2.6 and 2.7). However, an increasing landward encroachment of inner shelf mud produces a fining upward trend in the upper sections of the shoreface wedge (Figs. 2.4 and 2.6).

Off High Island and Bolivar Peninsula, a thinner and narrower shoreface/inner shelf wedge shows sparse interfingering of shoreface sand and inner shelf mud (Figs. 2.6e and 2.6f). The shoreface becomes mud-dominated farther to the east (Fig. 2.6f). Off the western end of Bolivar Peninsula, remains of a decapitated tidal inlet/ebb-channel sequence occur in the shoreface; distal ebb-tidal delta sediments overlie the ravinement surface (Chapter 1).

Southwestward migration of Bolivar Roads to its present position caused the abandonment and subsequent ravinement of the inlet/ebb-channel sequence (Chapter 1).

Seismic records across Galveston Island's shoreface and Bolivar Roads ebb-tidal delta exhibit long, seaward-dipping reflectors that represent the seaward growth of the island (Fig. 2.8). 3.5 kHz records within the upper shoreface also show variations in reflector angle and geometry. Several trough-like features, 75 to 100 m wide and 1 to 2 m deep and probably shore-parallel, are interpreted as shoreface runnels (Fig. 2.8).

Seismic profiles (3.5 kHz), collected parallel to the coast, from Galveston Island's upper shoreface (3 to 4 m water depth) define cross-cutting channels of varying sizes (50 to 300 m wide and 1 to 5 m deep) (Fig. 2.9). The channels are completely healed by on-lap fill. Prograded-fill is rare. The channels do not have any surface bathymetric expression. These channels are ubiquitous on Galveston Island's shoreface but were not imaged on Bolivar Peninsula, probably because of a thinner shoreface wedge. The channels are interpreted
Figure 2.8. Line drawing of a 3.5 khz record across Galveston Island's shoreface showing prograding reflectors.
Figure 2.9. Line drawing of a 3.5 kHz record along Galveston Island's upper shoreface showing cross-cutting channels interpreted as storm return-flow channels.
as storm-return flow (ebb-surge or coastal-downwell) channels that may have served as sand conduits to the inner shelf during storm events (Fig. 2.10). Rarity of a prograded-fill pattern, typical of inlet fills (Chapter 1), favors a storm-channel origin as opposed to an inlet origin. Channels at the backside of Galveston Island, oriented perpendicular to the shoreline, could be the overwash continuations of some of the shoreface channels (Figs. 2.2 and 2.10). Abundance of channels within the shoreface contributes to the difficulty in correlating beds along the shoreface and, perhaps, on the inner shelf as well.

Because of the common occurrence of the channels in the shoreface wedge, it is likely that a reflector with a gentler gradient or concave geometry along the dip-lines represents an axial channel trace (Figs. 2.8 and 2.10). However, it may also be a storm shoreface profile: the steeper and more parallel clinoforms are the fair weather profiles (Fig. 2.8). This is based on Moody's (1964) observation that the shoreface profile gradient is reduced during storm events.

**Storm Deposit Stratigraphy**

Cored lower shoreface and inner shelf deposits display an increase in the occurrence of storm beds with depth (Figs. 2.6 and 2.7). Some of the older storm beds appear to be shoreface-attached. Furthermore, both Galveston Island and Bolivar Peninsula have numerous surface morphological features, including overwash channels and fans, indicating a significant storm influence during their early development (Chapter 1). These combined data may imply greater frequency and/or magnitude of storms in the past. More likely, the barriers were breached more frequently in the past because they were much narrower and had a lower profile. Greater frequency of barrier breaching resulted in greater channelized offshore transport, and thus a larger
Figure 2.10.  a) Model for formation and subsequent modification of storm-return flow channels on the shoreface. Channels are formed during a storm by coastal downwelling and cyclic loading of the substrate by storm waves. Some channels may extend into the inner shelf but probably terminate at the lower shoreface. Some sediments are deposited at the shoreface base and the rest are entrained by along-shore geostrophic currents (after Snedden et al. (1988), Swift et al. (1985), and Walker (1985)).  b) The relatively steep shoreface profile is flattened during a storm; regrading occurs during fair weather conditions (after Moody (1964)). Channels are healed and get truncated on their updip ends. Sediment is entrained by longshore currents away from the plane of section during the storm and recovery period resulting in net loss of sediment.
representation of storm beds. Abundant supply of sand during this period also may account for the greater occurrence and thickness of storm beds. Progradation of Galveston Island and Bolivar Peninsula widened the barriers: this led to less frequent breaching (Chapter 1).

The abundance of sediment supply during the early development of coastal systems along the east Texas coast is indicated by their accelerated establishment after the occurrence of a rapid sea-level rise 4 ka (Anderson et al., 1991a; 1991b; Siringan and Anderson, 1991; Chapter 1). Cole and Anderson (1982), based on the variation of mineralogy and texture of sediments comprising Galveston Island and Bolivar Peninsula, proposed that the sediment were derived from an offshore source. Eventual depletion of this sand source contributed to the erosion in the region and the enhancement of mud signature on the upper sections of the shoreface sequence.

DISCUSSION

Storm Deposition

The more widely recognized models for the formation of storm deposits involve offshore sand transport during major storms. These include: 1) direct-wind-forced currents and geostrophic flows (Morton, 1981; Swift et al., 1985; Snedden et al., 1988); 2) storm surge ebb-currents (Hayes, 1967); and 3) turbidity currents (Walker, 1985). The abundance of channels on the shoreface appears to support massive offshore sand transport during storms. Coastal downwelling currents generated by the coastal set-up (Swift, et al., 1985) may have formed the channels (Fig. 2.10a). Together with cyclic loading of the substrate by storm waves, sediments may have been moved, en masse, seaward (Walker, 1985). As the channels broaden and shallow offshore, some
of the sediments get redeposited within the lower shoreface-inner shelf transition, the rest are entrained by along-shelf currents. The distance of offshore transport depends on the intensity of downwelling currents. Walker (1985) argues that liquifaction of shoreface sediments may lead to the formation of turbidity currents that carry sediments far onto the inner shelf. During fair weather, regrading of the shoreface truncates the channels downdip and sediments are moved back onto the shoreface (Fig. 2.10b).

Given the abundance of channels within the shoreface and on Galveston Island, and the high frequency of storms and hurricanes that have impacted the region during historical times (Simpson and Riehl, 1981; Paine and Morton, 1989), a thick amalgamation or high frequency of storm beds on the lower shoreface and inner shelf is expected. But there is a paucity of such deposits in these regions, indicating that sediment eroded from the beach and the upper shoreface was not transported far offshore. Eventually, this material was transported back onshore or alongshore.

During Hurricane Alicia in 1983, Morton and Paine (1985) estimated that nearly $1.5 \times 10^6 \text{ m}^3$ of sand was eroded from a 30 km stretch of west Galveston Island. After two years, they were able to account for 60% of this volume in beach recovery and washover deposits; the remaining sand is presumed to have been transported southwestward along the shoreface. This is supported by data from current meters located on both sides of Alicia landfall. Predominantly unidirectional currents flowed alongshore and slightly offshore before and after the storm crossed the shelf and made landfall at the western tip of Galveston Island (Morton, 1988). Overall, there seems to be no evidence for massive sand transport offshore during Hurricane Alicia. This is supported by the patchy distribution of sand on the inner shelf. Since erosion on the inner
shelf tends to winnow fine sediment and concentrate the sand and coarser fraction, greater sand accumulation should be found if large quantities of sand have been pumped onto the inner shelf. It is more likely that small amounts of sand were deposited on the lower shoreface and inner shelf, but the storm beds were not preserved because of reworking.

There are more storm beds in the lower sections of the shoreface, probably because of better preservation. Greater sand supply during their formation may have allowed deposition of thicker storm beds. This combined with higher sedimentation rates during fairweather would have increased their preservation potential.

On the central Texas shelf, discrepancies in the estimated volume of sand eroded from the beach versus the volume stored in storm beds suggest that each storm bed is not necessarily a product of sand eroded from the beach and shoreface during a single storm (Morton, 1988). Instead, they also may result from sand being transported offshore by a series of moderately intense events and later selectively sorted during a single extreme storm lasting several days. This is supported by an increasing documentation of \textit{in situ} resuspension and net onshore sand transport (Nittroer et al., 1986; Gagan et al.1988).

Within the study area, the modes of 3.00$\phi$ and 3.75$\phi$ exhibited by storm beds may imply two sources of sand. Beds with 3.75$\phi$ mode are more likely to have been winnowed from the inner shelf mud, or derived from the muddy shoreface and inner shelf to the east. Storm beds with modes of 3.00$\phi$ probably were derived from the adjacent shoreface and/or by \textit{in situ} resuspension. Grain size studies show that the shoreface is dominated by sand within the 3.0$\phi$ (upper shoreface) to 3.5$\phi$ (lower shoreface) size range and that the inner shelf mud
contains sand with 3.50φ or 3.75φ modes (Cole and Anderson, 1982; Anderson et al., 1982; this study).

*Distribution of Storm Deposits on the Continental Shelf*

Amalgamated sand beds of possible storm origin on the continental shelf are associated with sand banks and incised valleys (Fig. 2.11) (Siringan and Anderson, 1991). A possible preserved lower shoreface deposit occurs above the Trinity/Sabine incised river valley, located seaward of Shepard Bank. The rarity or possible absence of preserved lower shoreface sequences on the shelf attest to their low preservation potential. Shoreface ravinement (Swift, 1975) appears to have been very efficient in removing shoreface and inner shelf sequences during shoreface transgression across the shelf. Storm beds comprising the sand banks are believed to have been reworked from previous Holocene coastal lithosomes (Thomas, 1990; Siringan and Anderson, 1991). The sand bodies are characterized by coarsening upward trends, a ravinement surface at their base, linear and along-coast orientations, steeper seaward flanks, and interfingering of sand and mud on their landward sides.

The geometry and internal architecture of the sand banks on the inner and middle shelf indicate landward and shore-parallel (to the southwest) growth (Siringan and Anderson, 1991). This implies that, during storms in this region, onshore and along-coast transport of sand is more dominant than offshore movement.

*Central Texas Shelf Holocene Sedimentation and Storm Deposits*

Contrast in sediment thickness between the eastern and western regions of the study area, on a smaller scale, is similar to the difference between the east Texas and the central Texas shelf regions. Holocene marine sediment cover on
Figure 2.11. Isopach map of amalgamated sand beds of possible storm origin on the east Texas shelf.
the central Texas shelf is considerably thicker (Berryhill, 1976; Morton, 1981) than on the east Texas shelf. Although rivers of central Texas also are unable to deliver sediments into the Gulf, convergence of net southward longshore currents from the east and net northward longshore currents from the south bring sediments onto the shelf, resulting in a higher rate of sedimentation (Shideler, 1978). Also, effective accommodation space on the central Texas shelf is greater because of a steeper shelf gradient and the Holocene-Pleistocene boundary is deeper (Berryhill, 1976; Siringan et al., 1991). The central sector stood as a clastic-free embayment during the Wisconsinan sea level lowstand while deltaic environments in the northern and southern sectors of the Texas shelf prograded across the shelf to the shelf break (Suter and Berryhill, 1985; Abdulah and Anderson, 1991).

The high influx of fine-grained sediment and greater effective accommodation space account for the higher preservation of storm deposits in this region. Cores show a greater abundance of storm beds (Berryhill, 1975; Morton, 1981; Snedden, 1985). The sand beds commonly are thin- to very thin-bedded and are interbedded with mud. Medium-bedded sand is common in the lower shoreface, but thick (> 30 cm), amalgamated storm beds are absent on the shelf. The sand beds can cover wide areas, as demonstrated by Hayes (1967) and Snedden (1985). Laterally persistent, solitary, thin storm beds are more likely to occur on the central Texas shelf than on the east Texas shelf as indicated by the results of Hayes (1967) and Snedden (1985) and this study. Thin storm beds with a patchy distribution would probably be common in both regions, but are more abundant on the east Texas shelf.

Offshore decrease of storm bed frequency, thickness, and amalgamation on the central Texas shelf are consistent with Aigner's (1985) proximality trends.
On the east Texas shelf, Aigner's (1985) proximity trends are valid only to within the inner shelf adjacent to the present shoreface (to ~25 km offshore); beyond this zone, the sand banks disrupt the trends.

**SUMMARY and CONCLUSIONS**

1. An overall historical deepening on the shoreface and the inner shelf off Galveston Island and Bolivar Peninsula is attributed to erosion. In general, deepening increases toward the shoreface base, then decreases farther offshore; water depth increases from east to west. Greater shoreface-base translation, in relation to shoreline translation, resulted in the general steepening of the shoreface profile. It reflects a time of net onshore sand transport.

2. The Holocene marine sediment cover of the east Texas shelf is thin because of low sediment input and low effective accommodation space.

3. The shoreface is characterized by cross-cutting channels of varying sizes that are interpreted as storm-return flow (ebb-surge or coastal-downwell) channels. These may have served as sand conduits to the inner shelf during storms.

4. Despite the high frequency of storms and hurricanes that have impacted the region during historical times and the abundance of channels within the shoreface, there is a scarcity of storm beds in these regions. Greater occurrence of storm beds during the early phase of the development of present coastal lithosomes results from higher sand supply. Greater sand supply resulted when barriers were narrower and breaching of these barriers was more common.
5. Storm deposits in the study area typically are laminated to very thin-bedded, in contrast to the very thin- to medium-bedded storm deposits off the central Texas coast.

6. There is no evidence of massive offshore sand transport in the study area during storms, mainly because of poor preservation. The along-shore component of storm flow appears to be an important control on the extent of offshore transport of sand during storms. Possibly, a large portion of the sand contained in storm beds of the central Texas shelf was derived from the east Texas shelf.

7. Amalgamated sand beds on the east Texas shelf, potentially of storm origin, are possible preserved lower shoreface deposits and reworked Holocene coastal lithosomes. They appear to be absent in the central Texas shelf, despite the abundance of storm beds. Amalgamation of storm beds seems to favor the formation of sand banks. Onshore and shore-parallel transport of sand during storms seems to be equally as, or possibly more, important than offshore transport of sand.

8. The contrast of storm deposition between the east and central Texas shelf may help explain the differences in the character of storm deposits in the rock record. This study shows that differences of bed thickness, lateral continuity, and storm bed frequency are not necessarily related to varying storm frequency and magnitude.
CHAPTER 3

PRESERVED COASTAL LITHOSOMES: VARIATIONS AND PRESERVATION, EAST TEXAS CONTINENTAL SHELF
CHAPTER SYNOPSIS

Preserved coastal lithosomes on the east Texas inner to middle continental shelf were identified and mapped using approximately 1,300 km of high resolution seismic data coupled with more than 400 sediment cores. Discrete pods of tidal-inlet, tidal-inlet/spit, and tidal-delta facies occur within incised valleys. Tidal-delta facies also occur outside of incised valleys. The distribution of preserved coastal lithosomes mimics the along-strike variation of the present coastal system and supports a step-like nature of sea-level rise. Also, the distribution defines six relative sea-level stillstands during the past 10.2 ky, including that of the present. The seismic architectures of the preserved lithosomes indicate that sediment supply and wave/current energies have not varied much during past 8 ky. But pre-8 ka coastal lithosomes indicate greater landward tidal influence, greater accommodation space, and higher sedimentation rates. Preserved coastal lithosomes indicate that the depth of ravinement decreases with decreasing shelf gradient, increasing rates of sea-level rise, and increasing sediment supply. High preservation potential within incised valleys is due to greater accommodation space and the soft valley-fill that allows incision of the inlets beyond the depth of shoreface ravinement.
INTRODUCTION

A better comprehension of Holocene preserved coastal lithosomes would lead to improved understanding of the factors that control their preservation, the nature of Holocene sea-level rise, the character of the induced coastal retreat, and the influence of sea-level rise on coastal and estuarine evolution. This study examines a variety of preserved coastal lithosomes within a broad region. In this paper, the occurrence, distribution, state of preservation, and nature of Holocene preserved coastal lithosomes on the east Texas inner to middle continental shelf are documented. Five paleoshoreline trends demarcating the shoreline retreat path during transgression are established. Variations in state of preservation and architecture indicate temporal and spatial variations of factors that influence the formation and preservation of coastal lithosomes and their interplay.

The study area is within the northern Gulf of Mexico, along the east Texas coast and adjacent continental shelf (Fig. 3.1). It extends from the Texas-Louisiana border at Sabine Pass to San Luis Pass at the western end of Galveston Island, from the shoreline to -35 m water depth.

The east Texas continental shelf is an excellent site for the examination of preserved coastal lithosomes for several reasons. First, the present coast provides a variety of well-studied lithosomes that may serve as analogues for preserved sequences. Some of the lithosomes include chenier plains (Gould and Mc Farlan, 1959; Byrne et al., 1959; Johnson, 1979), barrier islands (LeBlanc and Hodgson, 1959; Bernard et al, 1959; 1970; Morton and McGowen, 1980), tidal inlets and tidal deltas (Eyer, 1984; Israel et al., 1987; Chapter 1); shoreface deposits (Bernard et al., 1959; 1970; Williams et al., 1979; Chapter 2), wave-dominated deltas (Bernard et al., 1970; Bartek et al.,
Figure 3.1. Geographic and bathymetric map of the study area and location map of high-resolution seismic lines (boomer and 3.5 kHz) and cores used in this study. Boreholes are not included.
and bay/estuarine systems (Kane, 1959; Rehkmemper, 1969; Smyth, 1991; Anderson et al., 1991a, 1991b). Second, incised valleys on the continental shelf have the greatest potential of yielding preserved coastal lithosomes (Kraft et al., 1987; Belknap and Kraft, 1981, 1985) and this have been documented in the study area by Thomas (1990). Third, studies of Holocene sea-level changes in the region may provide a reference for the ages of sea-level events (Curray, 1960; Rehkmemper, 1969; Nelson and Bray, 1970; Frazier, 1974; Thomas, 1990) (Fig. 3.2). Finally, a twofold increase of the shelf gradient (to -40 m) from 0.21 m/km in the east to 0.49 m/km in the west, provides an opportunity to evaluate shelf gradient as a variable of coastal lithosome evolution and preservation (Fig. 3.1).

**BACKGROUND**

Bolivar Roads and Galveston Bay, and Sabine Lake and Sabine Pass are within the Trinity River and Sabine River incised valleys, respectively (Fig. 3.3). The valleys were last incised during the $\delta^{18}O$ Stage 2 lowstand (Thomas, 1990), when sea-level fell to approximately -126 m below present sea level (Fairbanks, 1989). The valley thalweg under Bolivar Roads lies approximately 55 m below present sea level (Chapter 1). Offshore, the depth of valley incision is 35 to 40 m below the sea floor (Thomas, 1990). Back-filling of the incised valley occurred during the Holocene transgression. Under Bolivar Roads, fluvial sand fills one third of the valley. Overlying estuarine sediments contain peat layers and pods of gray, silty to clayey sands, interpreted as bay-head delta deposits. Tidal-inlet/flood-delta sand, capped by barrier/peninsula and present tidal-inlet sand, occurs in the upper part of the estuarine sequence.

Studies on the Texas continental shelf and coast indicate an episodic or
Figure 3.2. Holocene sea-level curves for the Gulf of Mexico. Curray (1960) interpreted bathymetric ridges and sand banks as former shoreline positions. Frazier (1974) also interpreted the banks as shorelines during relative sea-level stillstands. Thomas (1990) identified parasequences in the Trinity-Sabine incised valley-fill interpreted to have formed during relative sea-level stillstands. Flooding surfaces that separate these parasequences were construed to represent rapid sea-level rises. Included are sea-level curves from Barbados (Fairbanks, 1989 and Bard et al., 1990). Both curves were derived from the same Acropora palmata samples collected off Barbados. However, for the chronology, Fairbanks (1989) used $^{14}$C ages while Bard et al. (1990) used $^{230}$Th-$^{234}$U ages.
Figure 3.3. Structure contour map of the Holocene-Pleistocene surface and distribution map of preserved coastal lithosomes and paleoshoreline trends on the continental shelf. Additional data points for the Holocene-Pleistocene surface map were compiled from Kane (1959), Rehkemper (1969), Nelson and Bray (1970), Israel (1983), Thomas (1990), and Smyth (1991).
step-like rise of sea level during the Holocene transgression (Curray, 1960; Rehkemper, 1969; Nelson and Bray, 1970; Frazier, 1974; Anderson and Thomas, 1991) (Fig. 3.2). The episodic sea-level rise produced backstepping parasequences within incised valleys on the east Texas shelf (Thomas and Anderson, 1989; Anderson and Thomas, 1991). Relative sea-level stillstands are recorded by the aggradation of coeval bayhead-delta and tidal-inlet lithosomes. Rapid sea-level rises are represented by flooding surfaces that separate backstepping parasequences (Anderson and Thomas, 1991). Four major rapid sea-level rises during the past 10 ky were recognized by Thomas (1990) within the Trinity-Sabine valley-fill. The most recent event is approximated to have occurred 4 ka (Anderson et al., 1991a, 1991b). This sea-level rise submerged the broad shallow portions (~6 m and above) of Galveston Bay and Sabine Lake, resulting in the establishment of the present broad, shallow estuarine/bay systems (Anderson et al., 1991a; 1991b). Present coastal lithosomes along the east Texas coast are believed to have evolved during the past 3.5 ky, or during the latest sea-level stillstand (Gould and McFarlan, 1959; Bernard et al., 1970; Cole and Anderson, 1982, Chapter 1).

METHODS

Approximately 1,300 km of high-resolution seismic data (uniboom and 3.5 kHz) coupled with vibracores (48), gravity cores (75), and piston cores (143), were used in this study (Fig. 3.1). The seismic lines were collected with the intent of mapping the incised valleys and examining their valley-fills (Thomas, 1990). More than 100 vibracores from Williams et al.'s (1979) sand resources study of the north Texas inner continental shelf, and Pearson et al.'s (1986) archeological investigation of the offshore Sabine River incised valley augment
the data set. Foundation and navigational channel borehole descriptions by the U.S. Army Corps of Engineers also aided the present study.

Seismic profiling and core collection were conducted aboard the R/V Lone Star, Rice University's research vessel, and aboard the R/V Gyre. GPS and Loran C were used for navigation for both vessels. An EG&G uniboom and an EDO 3.5 kHz subbottom profiling system with a towed transducer were utilized for seismic data acquisition. Some of the data were acquired digitally. A velocity of 1525 m/s was applied for conversion of acoustic travel time to depth within the sediments. Seismic profiles and cores obtained prior to 1989 were collected aboard R/V Matagorda; Loran C and radar were used for navigation.

Initially, characterization of the seismic facies and facies architecture of present coastal systems was conducted (Siringan et al., 1989; Siringan and Anderson, 1991; Chapter 1). Results from these studies were used to identify and map preserved coastal lithosomes on the inner shelf.

RESULTS

Preserved coastal lithosomes on the east Texas inner continental shelf occur as discrete pods of tidal-inlet facies, tidal-inlet/spit facies, ebb-tidal delta facies, and flood-tidal delta facies (Fig. 3.3). Initially identified only within the Trinity and Sabine incised valleys (Thomas, 1990), re-examination of the Thomas (1990) data set and additional seismic and core data enabled the identification of other preserved coastal lithosomes within and outside the incised valleys. The greater occurrence of preserved coastal lithosomes within incised valleys supports the idea that valley-fill sediments have the greatest preservation potential (Belknap and Kraft, 1981; 1985).
Preserved tidal-inlet and tidal-inlet/spit facies occur only within the incised valleys, while preserved ebb- and flood-tidal deltas occur within and outside the incised valleys, although they are more common within the valley. Identification of some of the preserved ebb- and flood-tidal delta facies was based solely on core data. The preserved sections are thin, making them difficult to see in seismic profiles. Shoreface deposits appear to have been completely removed during transgression. The only possible preserved lower shoreface, based on core data, occurs on the seaward flank of Shepard Bank. The rarity of preserved shoreface deposits attests to their low preservation potential during transgression.

Five paleoshoreline trends are delineated based on the distribution of the preserved coastal lithosomes and reconstruction of sea-level during their formation (Fig. 3.3). Reasonable age estimates of the shoreline trends are achieved using \(^{14}C\) age dates (adjusted to U/Th age (Bard, 1992)) from Thomas (1990) and estimates of sea level compared to Thomas' (1990) and Bard et al.'s (1990) sea-level curves.

**Paleoshoreline Trend 1**

The landwardmost paleoshoreline trend (PST) is located 10 to 30 km offshore between San Luis Pass and Sabine Pass. It is defined by, from east to west, Sabine valley tidal inlet 1/flood-tidal delta (SVTI 1/FTD), Trinity valley tidal inlet 1 (TVTI 1), and Big Slough ebb-tidal delta (BSETD).

BSETD is located 10 to 15 km southeast of San Luis Pass, in 13 to 15 m water depths (Figs. 3.3, 3.4, and 3.5). It consists of soft, dark greenish gray, slightly bioturbated clay with thin interbeds to laminations of fine sand that probably were distal deposits of the paleo-ebb-tidal delta. The delta prograded
Figure 3.4. Location map of subsequent figures.
Figure 3.5. Decapitated ebb delta complex, BSETD, along Paleoshoreline Trend I. Figure 3.4 shows the location of the profile.
over the Holocene-Pleistocene boundary; in some areas it prograded over fresh water marsh deposits (Fig. 3.5). Two meters of this paleo-ebb-tidal delta, spared from shoreface ravinement, has served as a nucleus for the Big Slough Bathymetric High (BSH), a muddy shelf sand bank. The ravinement surface, at -15 to -17 m, completely removed the associated tidal-inlet and flood-delta facies. Transgressive lags consisting of shell fragments with muddy matrix, formed in some places.

To the east of BSETD lies TVTI 1, located 9 to 15 km downdip of Bolivar Roads in 14 to 16 m water depths (Fig. 3.3). Seismic profiles show a tidal-inlet sequence, approximately 10 m thick, that has been ravined at -17 m (Fig. 3.6). The preserved inlet is incised to -27 m, cutting into a seismically defined middle-bay to upper-bay/bay-head delta deposits. The inlet base is scalloped, but overall shallows landward and seaward. The preserved section exhibits channel cut-and-fill with a general stacking pattern (Fig. 3.6).

Landward-dipping clinoforms, downlapping the tidal inlet channel base at the northwestern end of Figure 3.5, could be the inlet-proximal facies of the correlative flood-tidal delta. The rest of the flood-tidal delta seems to have been removed by ravinement. A possible remnant of the ebb-tidal delta is penetrated by core GY-153, taken downdip of TVTI 1 at 16.5 m water depth (Fig. 3.7). The deposit consists of slightly to moderately bioturbated, dark greenish gray, very fine sand grading up to clay with occasional silt laminations and shell lag layers comprised mostly of *Mullinia lateralis* shells. Two ravinement surfaces separate the preserved ebb-tidal delta sequence from the overlying inner shelf mud and underlying lower estuarine deposit (Fig. 3.7). The lower ravinement surface was produced during an earlier transgression; the upper ravinement surface is associated with the shoreline translation to its present position.
Figure 3.6. Decapitated tidal inlet, TVTI 1, under the lower shoreface/ebb-tidal delta and inner shelf wedge. Figure 3.4 shows the profile location.
Figure 3.7. Core GY-153. Ravinement surfaces separate the possible TVTI 1 preserved ebb-tidal delta sequence from the overlying inner shelf mud and underlying lower estuarine deposit. In parenthesis is $^{14}$C date adjusted to U/Th age (after Bard, 1992). Figure 3.4 shows the core location.
The easternmost preserved coastal lithosome along PST 1 is a tidal-inlet and flood-tidal delta pair, SVTI 1/FTD (Figs. 3.3 and 3.8). The tidal-inlet facies is incised to -28 m, and the ravinement surface occurs at or near the sea-floor at -12 to -13 m. The flood-tidal delta facies was mapped previously by Thomas (1990) as tidal-inlet facies. However, the seismic architecture is more similar to the flood-tidal delta facies of the present Bolivar Roads tidal-inlet/delta complex (Chapter 1) (Fig. 3.8). SVTI 1's flood-tidal delta shows bi-directional cliniforms with a more dominant landward component. The flood-tidal delta facies is approximately 11 m thick, and is truncated at -13 m.

The present observed depth of the shoreface toe along the east Texas coast is -4 m off High Island, -5 to -7 m off Bolivar Peninsula, and -9 m off Galveston Island. These depths, estimated using the first observable break in slope of the shoreface profile where the slope decreases from ~1:200 to ~1:2000 (Friedman and Sanders, 1978; Swift, et al., 1985) as the shoreface limit, were used as first approximation of the depth of ravinement during trangression. Based on this, sea level for PST 1 was at approximately -7 m. Within the Texas bays, a relative sea-level stillstand that may have lasted until about 4 ka was estimated at -6 m (Anderson et al., 1991a; 1991b, Siringan et al., 1991). The sea-level stillstand estimated from the depth of the ravinement surface is deeper by 1 m probably because of additional erosion caused by shelf processes. Subsidence due to valley-fill compaction after the shoreface has passed through the region may have also contributed to this difference.

**Paleoshoreline Trend 2**

PST 2, located 30 to 40 km seaward of the present coastline, approximately parallels the present shoreline (Fig. 3.3). It is defined by, from east to west,
Figure 3.8. Seismic profile of the inlet-proximal flood-tidal delta facies of SVTI 1/FTD (modified from Thomas (1990)). Figure 3.4 shows the location of the profile.
Sabine valley tidal-inlet 2 (SVTI 2), Trinity valley tidal-inlet 2 (TVTI 2), and Curtis Bank flood-tidal delta (CBFTD).

Curtis Bank flood-tidal delta, defined by piston cores, occurs on the landward side of Curtis Bank (Figs. 3.3 and 3.9). It consists of a meter or less of thin- to medium-bedded, well-sorted fine sand (colors range from light olive gray to moderate olive brown to tan) with small shell fragments sometimes occurring as shell layers. Compared to the present Bolivar Roads and San Luis Pass flood-tidal deltas (Chapter 1, Israel et al., 1987), the preserved section of CBFTD is consistent with an inlet-proximal facies. It is incised into dark greenish gray, slightly cohesive, moderately bioturbated estuarine mud (Fig. 3.9). The -20 m ravinement surface that truncated the delta is overlain by inner shelf mud and sand.

There are no indications of CBFTD's correlative tidal-inlet and ebb-tidal delta facies, although piston cores within the area penetrated, on average, almost 2 m (Fig. 3.9). The inner shelf sediments thicken seaward and to the west indicating that ravinement was deeper in these regions. This raises the possibility that CBFTD's correlative tidal-inlet and ebb-tidal delta facies have been eroded completely. The Holocene-Pleistocene surface high, located seaward of CBFTD, may have protected the flood-tidal delta from complete reworking (Figs. 3.3 and 3.9).

Approximately 20 km downdip and to the east of TVTI 1, in 17 m water depth, is TVTI 2 (Figs. 3.3 and 3.10). It is incised to -31 m. Decapitation of TVTI 2 occurred at -19 m. Similar to TVTI 1, TVTI 2 has incised into seismically defined upper-bay/bay-head delta deposits (Fig. 3.10). A strike-line shows channel stacking and a general westward inlet migration (Fig. 3.10a). A dip-line illustrates the shallowing of the inlet channel in the seaward direction (Fig.
Figure 3.9. Cross-sections showing the geometry of CBFTD and its location relative to a Holocene-Pleistocene surface high. Figure 3.4 shows the locations of the transects.
Figure 3.10. Seismic profiles showing the truncated TVTI 2. a) Strike-section. b) Dip-section. Figure 3.4 shows the locations of the profiles.
A set of seaward-dipping clinoforms, interpreted as a flood-tidal delta facies, occurs on the landward edge of the channel (Fig. 3.10a).

Slightly seaward dipping reflectors, interpreted as ebb-tidal delta facies, downlap the seaward edge of the tidal inlet channel base (Fig. 3.10b). Piston cores in this area penetrated shell-rich sand beds grading-up to mud with sand laminations separated from underlying bay/estuarine sediments by the inlet channel base (Fig. 3.11). The ebb-tidal delta package's fining-upward trend may represent lateral migration of the ebb tidal delta away from the core locations. A possible correlative inlet-distal flood-tidal delta facies to TVTI 2 is the lower estuarine sequence underlain by TVTI 1's ebb-tidal delta package on GY-153 (Fig. 3.7). This flood-tidal delta package consists of light greenish gray clay with highly dispersed plant fragments, very fine sand laminations and well-defined 1 cm thick *M. lateralis*-dominated shell beds (Fig. 3.7).

Farther to the east, along the southern flank of the Sabine River incised valley, SVTI 2 can be traced for more than 5 km along-strike (Figs. 3.3 and 3.12). It has a scalloped base that is incised into Pleistocene sediments to as much as 8 m. The maximum depth of the tidal-inlet scour is -28 m; the ravinement surface occurs at -16 to -19 m. Inlet migration shift, from westward to eastward, is indicated by the change in the predominant westward dip direction of the clinoforms in the lower section to a predominant eastward dip in the upper section. SVTI 2's coeval flood- and ebb-tidal deltas were not recognized. It is possible that they were ravined completely. A continuation of the tidal-inlet facies to the east, nestled in one of the tributaries of Sabine River incised valley, was mapped by Thomas (1990). It was incised to -30 m and was decapitated at -15 to -17 m.

A first approximation of the depth of ravinement during trangression based
Figure 3.11. Series of piston cores (GY-141 to 143) across the Trinity River incised valley and on the seaward side of TVTI 2 defining the TVTI 2 ETD. In parenthesis is $^{14}$C date adjusted to U/Th age (after Bard, 1992). Figure 3.4 shows the locations of the cores.
Figure 3.12. Seismic profile of SVTI 2. Figure 3.4 shows the location of the profile.
on the observed depth of the present shoreface toe along the east Texas coast, yields a sea level for PST 2 at approximately -10 m. This corresponds to a flooding surface at -11 to -10 m, indicated by seismic lines on Galveston Bay and Sabine Lake, respectively (Anderson et al., 1991a). Thomas's (1990) sea-level curve indicates a sea-level stillstand at -14 m (Fig. 3.2). A stillstand at -14 m implies that the thickness of sediments removed during transgression was on the order of 3 to 4 m. However, the preserved sequences along PST 2 indicate greater depth of decapitation.

Samples from the lower and upper parts of two peat layers within the Trinity River incised-valley fill that correlate to the -11 m flooding surface yielded $^{14}$C ages of 7165±259 to 6785±243 and 7689±280 to 6402±228 yBP (Rehkemper, 1969). Shells at or just above the ebb-tidal delta/tidal inlet base yielded $^{14}$C dates of 7,480±110 yBP and 8340±130 yBP. Organic mud from the underlying bay-head delta deposit yielded a date of 8,550±140 yBP (Thomas, 1990). These ages, converted to U/Th ages (after Bard, 1992), yield an age estimate of 8 ky to 6 ky for PST 2.

**Paleoshoreline Trend 3**

Paleoshoreline Trend 3, located approximately 50 km seaward of Bolivar Roads, is defined by a tidal inlet (TVTI 3) and its correlative ebb-tidal delta (TVTI 3 ETD) (Figs. 3.3, 3.13, and 3.14). It is located at the confluence of Trinity River and Sabine River incised valleys. There are no indications of a preserved TVTI 3's correlative flood-tidal delta facies.

TVTI 3 is incised to -48 m (Fig. 3.13). The channel geometry is asymmetric, it deepens landward and it has a steep landward edge. This flank may represent the flood-tidal delta ramp. The upper half of the inlet-fill shows channel stacking
while the lower half shows prograded- to onlap-fill. TVTI 3 was decapitated at -20 m.

TVTI 3 ETD is 4 to 5 m thick. Seaward-dipping (∼0.4°) clinoforms characterize the delta (Fig. 3.14). In places where the clinoforms are not defined clearly, the reflectors are wavy (amplitudes of 2 to 3 m and wavelengths on the order of 100 m). Updip, more continuous horizontal reflectors replace this reflector pattern. TVTI 3 ETD prograded more than 20 km seaward over the Holocene-Pleistocene surface, updip and over a ravinement surface downdip (Fig. 3.14). TVTI 3 ETD's base is relatively flat. The ravinement surface that decapitated TVTI 3 ETD at -20 m, also is relatively flat. The geometry of the preserved clinoforms indicates minor decapitation; some of the topsets seem to have been preserved (Fig. 3.14). Decapitation probably was on the order of 2 to 3 m.

A possible 2 to 3 m deep ravinement of TVTI 3 ETD places sea-level for PST 3 at ∼17 m. TVTI 3 and its associated tidal delta probably formed 8.5 ky to 8 ky ago, based on the maximum age of PST 2 and on Bard et al.'s (1990) sea-level curve from Barbados.

**Paleoshoreline Trend 4**

Paleoshoreline Trend 4, located 60 km downdip of Bolivar Roads, is defined by TVTI 4 and its associated flood-tidal delta (TVTI 4 FTD) (Figs. 3.3, 3.14b, and 3.15).

TVTI 4's inlet-base geometry resembles TVTI 3, but the deep portion of the channel is broader (Fig. 3.15). It is incised to -35 m, into seismically defined upper bay/bay head delta deposits. The lower half of the inlet fill consists of landward dipping clinoforms that downlap onto the channel base. The upper
Figure 3.13. Seismic profile of TVTI 3. Figure 3.4 shows the location of the profile.
Figure 3.14. a) TVTI 3 FTD, b) TVTI 3 ETD and TVTI 4 FTD. Figure 3.4 shows the locations of the profiles.
half exhibits small channels.

Thomas (1990) identified and mapped TVTI 4 FTD, based on a sequence of landward-dipping clinoforms (Fig. 3.14b). Along a dip-line, it is characterized by prograding and aggrading parallel clinoforms that downlap onto the underlying seismically defined upper-bay/bay-head delta deposits. The topsets and foresets are long and they progressively lengthen up the section (maximum dip of >0.4°). Along-strike, tangential clinoforms dip westward with a gentler gradient (>0.2°). TVTI 4 FTD and its coeval bay-head delta consists of several aggradation and progradation phases, each a meter or more in amplitude, interpreted to represent small but episodic rises of sea level (Thomas, 1990). Initially however, a large sea-level rise is needed to superimpose the flood-tidal delta facies on upper bay facies.

TVTI 4 and TVTI 4 FTD were truncated by a ravinement surface at -22 m at its updip limit, and at -26 m at its seaward extent (Figs. 3.14b and 3.15). No more than 3 m of the TVTI 4 FTD is estimated to have been truncated by ravinement, based on the preservation of toplap reflectors. Possible lower shoreface deposits located at the toe of Shepard Bank, consisting of sand and clay interbeds, seem to support this. Although it appears that decapitation was minimal, no related ebb-tidal delta deposits were found.

At its last stillstand position, sea-level was at -20 m, based on the estimate that only 3 m of TVTI 4 FTD's upper section was removed by ravinement. This is the same paleo-sea-level Thomas (1990) derived based on TVTI 4 FTD's coeval bay-head delta elevation. $^{14}$C age dates of *Mulinia* shell hash above the ravinement surface yielded dates of 8680±100 yBP (GY-151). Shells within TVTI 4 FTD yielded an age of 8640±120 yBP (GY-149). Organic muds overlain by TVTI 4 Flood-TD yielded $^{14}$C ages of 13670±180 yBP (GY-146) and
Figure 3.15. Seismic profile of TVTI 4. Figure 3.4 shows the location of the profile.
13760±180 yBP (GY-149). Based on the estimated -20 m sea-level elevation, Bard et al.'s (1990) sea-level curve, and 14C (adjusted to U-Th ages based on Bard's (1992) conversion equation), TVTI 4 and TVTI 4 FTD are 9.5 ky to 8.5 ky old.

A possible coeval system of TVTI 4 and TVTI 4 FTD is the Freeport Rocks wave-dominated delta, located to the west of the study area (Fig. 3.1). It occurs approximately 15 km offshore of the present Brazos River Delta, in about -20 m water depth. Bartek et al. (1991) estimated that only the upper 2 to 3 m of the paleo-delta was reworked into what is now a strike-aligned sand body that defines Freeport Rocks Bathymetric High (FRBH). Roughly 7 m of the delta distributary channel and mouth bar facies are preserved beneath the ravinement surface (Bartek et al., 1991). Bartek et al. (1991) estimates an age of 8 to 9 ky for FRBH based on Winchester's (1971) data and Gulf of Mexico sea-level curves. Also, along the central Texas shelf ~7 km off Matagorda Island in 13 m water depth, Shepard (1956) found a preserved 20 m thick barrier facies buried under a ~3.5 m thick inner shelf deposit (Fig. 3.1). This preserved barrier facies probably represents stacked barriers that correlate to the stacked TVTI 4 FTD and TVTI 3 ETD.

**Paleoshoreline Trend 5**

The outermost paleoshoreline trend within the data limit occurs at 30 m water depth about 85 km downdip of Bolivar Roads (Fig. 3.3). It is defined by a paired tidal-inlet/spit facies and flood-tidal delta facies, mapped by Thomas (1990) (Figs. 3.3 and 3.16). The tidal-inlet/spit facies (Outer TI/Spit), along-strike, consists of tangential clinoforms that define westward lateral accretion into the valley center (Fig. 3.16a). The clinoforms downlap onto a slightly
irregular channel base that deepen toward the valley center and incise into a seismically defined bay head delta facies. From the west, parallel clinoforms (long foresets with no bottomset) accrete eastward. A gap between the two clinoform sets, approximately 2.5 km wide and 10 m deep, is onlap-filled. The tidal-inlet/spit facies has a maximum thickness of about 14 m. A dip-section shows the landward progradation of the Outer TI/Spit and the seaward dipping clinoforms of the flood-tidal delta facies (Outer FTD) (Fig. 3.16b). The Outer FTD, as mapped by Thomas (1990), is 14 m thick and extends to 35 km updip of the Outer TI/Spit (Fig. 3.3). Sediments were derived from the east, based on the overall northwestward growth of the delta (Thomas, 1990).

The ravinement surface that decapitated the Outer TI/Spit and FTD occurs at -32 to -35 m. An estimate of sea level for PST 5, based on flooding surface elevation, is -29 m (Thomas, 1990). Correlation with Bard et al.'s (1990) sea-level curve yields a minimum age of 9.5 ky and maximum age of 10.2 ky. This is constrained by the age of PST 4 and a possible earlier sea-level stillstand at -36 m (Thomas, 1990) respectively.

DISCUSSION

Nature of Sea-Level Rise

The discrete nature of the preserved coastal lithosomes, the deeply incised tidal inlets, and the potential of estimating paleo-sea-levels based on the depth of shoreface ravinement, supports the occurrence of relative sea-level stillstands and episodic rises (Fig. 3.2). A continuous rise of sea-level is less likely to leave preserved coastal lithosomes because of the efficiency of shoreface ravinement. If coastal lithosomes are preserved, they would have a sheet-like geometry, rather than a discrete pod geometry. The development of deeply
Figure 3.16. a) Outer TI Strike Section, b) Outer TI Dip-Section. Figure 3.4 shows the locations of the profiles.
incised tidal inlets in the Gulf Coast are related to the enhancement of tidal processes due to stabilization and increase of tidal prism (Suter and Penland, 1987; Chapter 1). Stabilization is achieved during relative stillstands.

The possibility of establishing coeval coastal systems by estimating the paleo-sea level based on the observed variation of shoreface depth along the present coast and the depths of ravinement, implies that the preserved coastal lithosomes were formed during relative sea-level stillstands.

**Distribution and Variations**

The distribution of the preserved coastal lithosomes on the east Texas continental shelf reflects the along-strike variations of the present coastal systems in the region (Fig. 3.3). Tidal-inlet complexes characterize the bay mouths of bays that have occupied, and presently are occupying, the former valleys of Trinity and Sabine rivers. In contrast, fluvial deltaic depositional systems are associated with the Brazos River. The distribution also attests to the influence of pre-transgressive topography in the evolution of coastal systems. The observed incised valley/tidal inlet associations are similar to those on other coastlines (Morton and Donaldson, 1973; Halsey, 1979; Price and Parker, 1979; Tye, 1981; Suter and Penland, 1987; Levin, 1990; Chapter 1), on present continental shelves (Hine and Snyder, 1985; Penland and Suter, 1983; Penland et al., 1985; Suter and Penland, 1987; Thomas, 1990), and in the rock record (Ricketts, 1991).

The seismic architecture of preserved coastal lithosomes defining PSTs' 1 and 2 (e.g., channel stacking and a highly irregular channel base of the tidal-inlet facies) suggests that coastal parameters (wave/tidal energy, sedimentation rates) and accommodation space during these periods were similar to the
present. Meager preservation of TVTI 1 and TVTI 2 flood-tidal delta deposits probably is due partly to their geometry, extent, and depths of shoreface ravinement (11 and 7 m deep, respectively). Bolivar Roads flood-tidal delta exhibits a concave-up base that abruptly shallows bayward (Chapter 1) (Fig. 3.17). The base barely exceeds a depth of 10 m, even in the inlet-proximal region. A 9 m deep shoreface ravinement would be sufficient to erode almost the entire flood-tidal delta sequence.

The seismic architecture exhibited by the preserved coastal lithosomes older than 8 ky, TVTIs 3 and 4 and Outer TI/Spit and FTD, indicates that coastal parameters and accommodation space were different during their formation. The overall size (thickness and extent) and seismic architecture of Outer and TVTI 4 FTDs are very different from the Bolivar Roads system. Their thicknesses of 14 m and landward extents of more than 30 km, are more than twice the size of the present Bolivar Roads flood-tidal delta (Fig. 3.17). Bolivar Roads flood-tidal delta, the largest flood-tidal delta along the Texas coast, has an average thickness of less than 5 m (channels within the inlet-proximal facies are slightly more than 10 m thick), and only extends 10 km bayward (Chapter 1). Along a dip-line, the base of the Bolivar Roads flood-tidal delta is "ladle-shaped" (with the handle pointing landward) (Fig. 3.17). Along a strike-line, it is more tabular, but it exhibits interfingering and gradation with estuarine/bay facies. Internally, none of the Bolivar Roads flood-tidal delta features (e.g., channel stacking, prograded channel-fill, bi-directional clinoforms) are exhibited by Outer and TVTI 4 FTDs. They also exhibit a well-defined northwesterly progradation and have a more flat base (Figs. 3.14b). Reflector geometry along strike is more similar to the inlet/spit facies behind Bolivar Peninsula.
Figure 3.17. a) Geometry of Bolivar Roads flood tidal delta. b) Geometry of TVTI 4 FTD and Outer FTD.
Lithologically, the Outer FTD and TVTI 4 FTD are similar to the mud-dominated Bolivar Roads flood-tidal delta (Chapter 1), although there is an apparent absence of *Crassostrea* shell fragments. Piston cores that penetrated the upper 1 to 3 m of the Outer FTD and TVTI 4 FTD sampled soft to firm, slightly bioturbated, dark greenish gray to olive gray to green gray clay with silt and very fine sand laminations and occasional very thin-beded shell layers (Fig. 3.18). Commonly, the shell beds consist predominantly of *M. lateralis*. Very thin- to thin-beded sand layers are rare.

The anomalous extent and thickness of TVTI 4 FTD, compared to the modern Bolivar Roads flood-tidal delta, is due partly to the stacking of several flood-tidal deltas. However, the landward reach of TVTI 4 FTD can be due only to greater tidal reach. The lengthening of the topsets within TVTI 4 FTD, as described earlier, may indicate that the shoreline position has migrated only short distances landward of its previous positions during the episodic rises of sea-level. This is supported by the geometry and fill architecture of TVTI 4. The deep, broad channel base and landward prograding inlet-fill probably record the small landward migrations of the shoreline. It is likely that during this period, the coeval barrier facies would have undergone aggradation.

The rates and amplitudes of sea-level rises during the deposition of TVTI 4 and Outer TI/Spit and their flood-tidal delta sequences were faster and greater, as indicated by Holocene sea-level curves. Combined with increased shelf gradient at about -20 m (shelf gradient is more than doubled), for each episodic rise greater accommodation space is created. It also is easier to superimpose widely separated sub-environments during backstepping, as attested by the thick flood-tidal deltas and their superposition with upper-bay/bay-head delta deposits. Wave and current energies also would have been different. Wave
Figure 3.18. Outer FTD. Figure 3.4 shows the core location.
energy along the coastline would have been higher because the shelf was
narrower and steeper. Longshore sediment transport rates also may have been
faster. This partly may explain the thick sequences deposited over very short
periods of time (1,000 years and less). Also, wetter climates within the upper
drainage basins of fluvial systems along the Texas coast during these times
(Gustavson et al., 1991; Toomey, 1992), indicated by paleoclimate records, may
have resulted in greater amounts of sediment discharged into the bays and the
Gulf. This is supported by the thicker and more expansive bay-head deltas of
the Trinity and Sabine paleo-estuaries (Thomas, 1990) and channel
aggradation and/or floodplain construction in the upper Colorado River, located
west of the study area (Blum, 1991), during this time. Rivers debouching
directly into the Gulf may have provided more sediment, enabling the formation
and preservation of Freeport Rocks wave-dominated delta. In addition, greater
sediment supply may have resulted in longshore current redistribution of sand
to source and allow aggradation of Shepard's (1956) preserved barrier facies.

Although the steeper and narrower shelf should have decreased the tidal
range, higher tidal ranges are indicated by the greater landward influence of
tidal currents. This may be caused by tidal convergence (Komar, 1976)
because the estuaries had funnel-like configurations. The long topsets and
foresets indicate tidal influence into the estuary as far landward as 35 km.
Examples of present systems with a similar extent of tidal influence are the
macrotidal Gironde estuary in France (Allen, 1991), and the mesotidal
Delaware Bay (Knebel, 1989).

High sedimentation rates may have contributed to the apparent absence of
_Crassostrea_ in the sediments of TVTI 4 FTD and the Outer FTD. This absence is
indicated also by the lack of seismic expression of oyster reefs within the
estuarine sequence. Conditions (salinity, turbidity) conducive to *Crassosstrea*’s growth probably were never attained in these estuaries. Also, the higher discharge of fresh water into the estuary, magnified by the confluence of Sabine and Trinity valleys, and a strong tidal influence may have created an environment unfavorable for *Crassosstrea*’s growth.

The apparent absence of ebb-tidal delta facies coeval to TVTI 4 FTD and the Outer FTD suggests the insignificance of ebb-directed tidal flow. The absence of these deposits probably is not due to ravinement, indicated by the very shallow shoreface ravinements and preservation of lower shoreface deposits. The occurrence of ebb-tidal delta facies in the TVTI 3 ETD, represents a change to ebb-tide dominance. Perhaps this change resulted from TVTI 3’s occurrence near the valley confluence and an estuary that was relatively shallow due to the flood-tidal delta build-up during the previous relative sea-level stillstand. The sigmoidal geometry of the TVTI 3 ETD clinoforms also may indicate deposition in shallow water. The overlapping wavy reflector pattern, seen in strike- and dip-sections, may represent subtidal migrating dunes.

**Determinants of Preservation**

The preservation of a coastal lithosome 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 and 1985; Sha, 1990). The amount of separation between these surfaces is determined by several variables, of which, pre-transgressive topography and relative sea-level rise (rate and magnitude) are most important.

**Relative Sea-Level Rise.** - Creation of accommodation space, in large part, depends on the amplitude and rate of sea-level rise. Rapid and large sea-
level rises create greater accommodation space that allows thicker sequences to develop, increasing their preservation potential. Location within an incised valley, coupled with a rapid and large sea-level rise, provides the greatest accommodation space. This is exemplified by the Outer FTD and TVTL 4 FTD. Also, the rate of relative sea-level rise controls the rates of landward and upward shoreface shift, which in turn, are factors in the efficiency of shoreface ravinement. Comparison of the pre-8 ka and post-8 ka preserved coastal lithosomes shows that the average depths of ravinement of the latter are shallower than those of the former (9 m versus 4 m). Potentially, the depth of shoreface ravinement could have been deeper during the pre-8 ka period because of higher wave energy. However, faster rates of landward translation due to faster rates of sea-level rise may have prevented deeper shoreface ravinement. During a rapid landward shoreface retreat, the shoreface-inner shelf profile may flatten. This is based on the flattening of the shoreface profile during stormy periods observed by Moody (1964) off the Delaware coast. The flattening of the shoreface-inner shelf profile may cause a shallowing of the shoreface depth resulting in shallower depth of ravinement during transgression. Greater amplitudes of sea-level rise also would allow abandonment of a coastal lithosome considerably below the depth of ravinement. However, a rapid sea-level rise still does not guarantee preservation of coastal lithosomes. The rate of relative sea-level rise associated with the 4 ka flooding event could have been as much as 4.5 cm/y. This is based on the estimated sea-level at the onset of the establishment of present coastal systems (unpublished data), and a possible 100 years duration of rapid sea-level rises (Anderson and Thomas, 1991). But even with this rate, almost all of the previous coastal lithosomes have been stripped away. This
illustrates the importance of the other factors involved in preservation of coastal lithosomes.

**Pre-Transgressive Geology (Gradient, Topography, and Lithology).** - Shelf gradient and its influence on the effective depth of ravinement is exemplified along the present east Texas coast. The westward increase of the shoreface depth along the east Texas coast is related to varying wave energy. Breakers off Galveston Island typically are a few centimeters higher than off Bolivar Peninsula (Hall, 1976). This can be attributed to the almost doubled shelf gradient from east to west. Generally increasing depth of ravinement of the coastal lithosomes, from east to west, along PSTs 1 and 2 confirms this relationship. The shallower depths of ravinement prior to 8 ka indicates that the rate of sea-level rise may be of greater importance than shelf gradient in controlling the depth of ravinement during transgression.

Seismic profiles across incised valleys exhibit the influence of lithology on the depth of shoreface ravinement; the ravinement surface deepens over the valleys (Fig. 3.8). Shallower ravinement outside the valley results from the stiffness and cohesiveness of Pleistocene sediments that are more difficult to erode. Although ravinement was efficient in decapitating, or even completely removing, coastal lithosomes during transgression, coastal lithosomes behind Holocene-Pleistocene surface highs were preserved (CBFTD, SVTI 1, and SVTI 2).

The relation of preservation to the rate of landward and upward translation of the shoreface, as a function of shelf gradient, still is unclear. A given amplitude of sea-level rise on low gradient shelves, potentially, would lead to a faster landward shoreline translation. Rates of shoreline translation may have varied
from 9.3 m/y at the western end of Galveston Island to almost 92.6 m/y at
Sabine Pass (faster transgression off Sabine Pass is due to a gentler shelf
gradient). This is calculated from the shoreline position prior to and
immediately after the rapid sea-level rise 4 ka and assuming a duration of
transgression of approximately 500 years. Faster landward shoreline
translation, as discussed earlier, may result in shallower depths of shoreface
ravinement. This suggests that the degree of preservation of coastal lithosomes
along the Texas coast should decrease from east to west. The increasing depth
of ravinement from east to west along PSTs 1 and 2, and the apparent paucity
of correlative preserved Brazos River wave-dominated deltas (Bartek et al.,
1991), suggests this. Occurrence of Shepard's (1956) 20 m-thick preserved
barrier facies, correlative with PSTs 3 and 4, implies that similar facies
associated to TVTI 3 and TVTI 4 also should be preserved. This is true for TVTI
4, where an associated shoreface facies has been recognized, but not for TVTI
3.

**Extent of a Coastal Lithosome.** - The seaward extent of the lower
boundary of a coastal lithosome predominantly is influenced by pre-
transgressive topography, wave and tidal energies, and sediment supply.
Redistribution of sediment by waves may prevent seaward progradation beyond
the depth of shoreface erosion. A large tidal prism and ebb-tide dominance,
coupled with a high sedimentation rate, favor the formation of a large ebb-tidal
delta prograding into depths beyond the depth of shoreface ravinement. This is
the case for the Bolivar Roads ebb-tidal delta, which extends 15 km offshore to
15 m water depth (Chapter 1); shoreface depth in this area is approximately -9
m. The more common occurrence of preserved ebb-tidal delta facies than
shoreface and flood-tidal delta facies reflects the greater preservation potential. Large tidal prisms and high sedimentation rates also are the reasons for the preservation of ebb-tidal deltas in the Wadden Sea where the preserved sections of ebb-tidal deltas were constructed beyond the depth of shoreface ravinement (Sha, 1991).

The ability of tidal inlets to incise beyond the depth of shoreface ravinement may account for their high preservation potential. In the Bolivar Roads tidal-inlet/delta complex, the maximum depth of incision is -23 m (Chapter 1); a ravinement depth of -10 m essentially will remove all of the barrier facies, but 13 m of the inlet deposit still would be preserved. TVTI 1 and TVTI 2 display such a relationship; none of the associated barrier facies were preserved. Tidal inlets that were not incised deeply enough were removed by shoreface ravinement. The tidal inlet associated with the BSETD, unable to incise the stiff Pleistocene sediments, has been ravined.

**SUMMARY AND CONCLUSIONS**

1) Preserved coastal lithosomes on the east Texas inner continental shelf are discrete pods of tidal-inlet and tidal-delta facies. Preserved lower shoreface deposits are rare. Preserved tidal-inlet facies occur only within incised valleys, while preserved ebb- and flood- tidal deltas occur within and outside the incised valleys, although they are more common within the valley.

2) There are six relative sea-level stillstands, including the present, during the past 10.2 ky; present sea-level from ~ 4 ka to present, at - 6 m from 6 to 4 ka, at -11 m from 8 to 6 ka, at -17 m from 8.5 to 8 ka, at -20 m from 9.5 to 8.5 ka, and at -29 m from 9.5 to 10.2 ka.
3) The distribution of preserved coastal lithosomes on the east Texas continental shelf reflects the along-strike variations of the extant coastal systems along the east Texas coast.

4) During the past 8 ky, sediment supply, effective accommodation space, and wave and tidal energies are lower than they were for the period prior to 8 ka. Steeper shelf gradient beyond -20 m allowed greater wave energies to impinge on the coast; combined with faster rates and greater magnitudes of sea-level rise prior to 8 ka, greater accommodation space was produced and tidal energies were focused within flooded incised valleys resulting in greater landward reach. Also, sediment supply was greater prior to 8 ka as a result of a wetter climate. Thus, the pre-8 ka flood-tidal delta and tidal-inlet facies and fauna are different from those of the present.

5) Depth of ravinement may increase with increasing shelf gradient and decrease with faster rates and greater amplitudes of sea-level rise. Rapid sea-level rises do not guarantee preservation of shoreface lithosomes.

6) The ability of tidal inlets to incise beyond the depth of shoreface ravinement, amplified within the valley because of softer valley-fill, accounts for the association of preserved tidal inlets with incised valleys.
CHAPTER 4

HOLOCENE TRANSGRESSIVE HISTORY OF THE EAST TEXAS
GULF COAST AND ADJACENT CONTINENTAL SHELF
CHAPTER SYNOPSIS

Approximately 1,300 km of high-resolution seismic data, coupled with more than 400 sediment cores were used to study the Holocene transgressive history of the east Texas shelf and coast. The Holocene transgressive record is characterized by: 1) continuity of the ravinement surface on the shelf and under the present coast; 2) occurrence of scarps on the Holocene-Pleistocene surface and double ravinement surfaces; 3) discrete pods of preserved tidal-inlet and tidal-delta lithosomes and poor to non-preservation of shoreface lithosomes; 4) poor to non-preservation of bay/lagoonal sediments in interfluvial areas; and 5) backstepping parasequences. Although these patterns are observed throughout the study area, transgressive shelf sand bodies, differing degrees of coastal lithosome preservation, and other features indicate temporal and spatial differences of regional transgressive histories. Shelf gradient, rate of sea-level rise, and rate of sediment supply are the major variables that control the mechanism of shoreline transgression.

On the eastern part of the east Texas shelf, the record of shoreline translation is consistent with the transgressive submergence mechanism; on the western half, it is consistent with the discontinuous erosional shoreface retreat mechanism. Prior to 8 ka, perhaps due to higher rates of sea-level rise, the transgressive record on the east Texas shelf is consistent with transgressive submergence. On the central Texas shelf, during this period, the transgressive record is consistent with both in-place drowning and discontinuous erosional shoreface retreat mechanisms.
INTRODUCTION

Sea-level change studies on the Texas shelf and adjacent Louisiana shelf indicate that episodic rapid sea-level rises and relative sea-level stillstands characterize the Wisconsinan-Holocene transgression (Curray, 1960; Nelson and Bray, 1970; Frazier, 1974; Anderson and Thomas, 1991; Penland et al., 1991, Chapter 3) (Fig. 4.1). Episodic sea-level rises produced backstepping parasequences within the incised valleys on the east Texas shelf (Thomas and Anderson, 1989; Anderson and Thomas, 1991). The distribution of preserved coastal lithosomes on the continental shelf indicates that six relative sea-level stillstands, including that of the present, have occurred during the past 10.2 ky: the stillstands occurred from ~ 4 ka to present at the present level, at - 6 m from 6 to 4 ka, at -11 m from 8 to 6 ka, at -17 m from 8.5 to 8 ka, at -20 m from 9.5 to 8.5 ka, and at -29 m from 9.5 to 10.2 ka (Chapter 3) (Fig. 4.2). The latest rapid sea-level rise that occurred approximately 4 ka, submerged the broad shallow portions (~-6 m and above) of Galveston Bay and Sabine Lake, resulting in the establishment of the present broad, shallow estuarine/bay system (Anderson et al., 1991a; 1991b; Smyth, 1991). Present coastal lithosomes along the east Texas coast are believed to have evolved during the past 3.5 ky, or during the latest sea-level stillstand (Gould and McFarlan, 1959; Bernard et al., 1970; Cole and Anderson, 1982).

This paper examines the Holocene transgressive record on the east Texas continental shelf. Coastal lithosome response to rapid sea-level rise is evaluated. The character of the shoreface ravinement surface, pre-transgressive surface, and nature of sand banks are documented. These are evaluated with respect to the efficiency of shoreface ravinement, sand budget, shelf sand bank formation, and the mechanism of shoreface translation, with a
Figure 4.1. Holocene sea-level curves for the Gulf of Mexico. Curay (1960) interpreted bathymetric ridges and sand banks as former shoreline positions. Frazier (1974) also interpreted the banks as shorelines during relative sea-level stillstands. Thomas (1990) identified parasequences in the Trinity-Sabine incised valley-fill interpreted to have formed during relative sea-level stillstands. Flooding surfaces that separate these parasequences were construed to represent rapid sea-level rises. Included are sea-level curves from Barbados (Fairbanks, 1989 and Bard et al., 1990). Both curves were derived from the same *Acropora palmata* samples collected offshore the island of Barbados. However, for the chronology, Fairbanks (1989) used $^{14}$C ages while Bard et al. (1990) used $^{230}$Th-$^{234}$U ages.
Figure 4.2. Distribution of preserved coastal lithosomes, paleo-shoreline trends, and shelf sand banks superimposed on the structure contour map of the Holocene Pleistocene surface (TV = Trinity valley, SV = Sabine valley, TI = tidal inlet, CBFTD = Curtis Bank flood-tidal delta, BSBH = Big Slough bathymetric high, BSETD = Big Slough ebb-tidal delta, PST = paleoshoreline trend). Also indicated is the location of transect shown on Figure 4.10.
focus on the latest episode of transgression related to the 4 ka rapid sea-level rise. Detailed discussion of the nature and distribution of preserved coastal lithosomes on the shelf are presented elsewhere (Chapter 3).

Results of this study suggest that erosional shoreface retreat, transgressive submergence, and in-place drowning occurred during different episodes of shoreline migration across the Texas shelf: they may have occurred simultaneously on different segments of the shelf during an episode of migration.

The study area is within the northern Gulf of Mexico (Fig. 4.3). It extends from the shoreline to 80 km offshore, between Sabine Pass at the Texas-Louisiana border to San Luis Pass at the western end of Galveston Island, and includes the inner continental shelf, coastal, and estuarine systems. The study area is an excellent site for the examination of coastal lithosome response to transgression because: 1) the east Texas coast provides a variety of well-studied coastal lithosomes that may behave differently in response to sea-level rise and that may serve as analogues for preserved coastal lithosomes on the shelf; 2) preserved coastal lithosomes and paleo-shoreline trends identified on the shelf provide an opportunity to examine several and specific episodes of transgression (Chapter 3) (Fig. 4.2); and 3) increasing shelf gradient within the study area (to -40 m water depth - 0.28 m/km off Sabine Pass, 0.49 m/km off San Luis Pass, and 1.15 m/km off Matagorda Island) should allow the evaluation of shelf gradient as a factor controlling the migrational nature of coastal lithosomes (Fig. 4.3).
Figure 4.3. Geographic and bathymetric map of the study area and location map of high-resolution seismic lines and cores used in this study. Borehole data are not included.
METHODS

Vibracores (148), gravity cores (75), and piston cores (143) were collected in the bays, on the coast, and offshore (Fig. 4.3) with the intent of sampling the entire Holocene marine section and penetrating the Holocene-Pleistocene surface within the interfluvies. In addition, vibracores from a prior sand resources study (Williams et al., 1979) of the inner shelf and archeological investigation of the Sabine River incised valley flanks (Pearson et al., 1986) augmented the core data base. Borehole data also were obtained from the U.S. Army Corps of Engineers.

Approximately 1,300 km of high-resolution seismic data were acquired with an EG&G uniboom and an EDO 3.5 kHz subbottom profiling system using a towed transducer (Fig. 4.3). Some of these data were acquired digitally. Output power for the uniboom varied between 200 and 500 joules with the band pass filter set at 300 to 2500 Hz. Signals were received either by a single hydrophone or a seven hydrophone single-channel streamer. At times, stratigraphic resolution of 0.3 m was achieved. More commonly, the resolution was approximately 0.5 m. Uniboom lines from Williams et al. (1979) augment the data set. An average interval velocity of 1525 m/sec, calibrated by correlation to borings, was used for depth conversions.

Seismic profiling and vibracore collection were conducted aboard the R/V Lone Star, operated by Rice University. Navigation consisted of both GPS and Loran C. Cores and seismic profiles obtained prior to 1990 were collected aboard R/V Matagorda, using Loran C and radar for navigation.

The first phase of the study concentrated on characterization of the seismic facies and facies architecture of present coastal systems along the east Texas
coast (Siringan, et al., 1990; Chapter 1, Chapter 2). Subsequently, preserved coastal lithosomes on the continental shelf were mapped and paleo-shoreline trends were identified (Chapter 3) (Fig. 4.2). Results from these studies, combined with previous investigations (e.g., Bernard et al., 1959; 1970; Gould and McFarlan, 1959; Cole and Anderson, 1982), are used to establish the migrational history of coastal lithosomes across the continental shelf.

RESULTS

Pre-Transgressive Surface

A structure contour map of the Holocene-Pleistocene surface outlines the incised valleys and interfluvies in the study area (Fig. 4.2). The Trinity valley is incised to -55 m beneath Bolivar Roads (Chapter 1) and 35 m to 40 m from the sea-floor in the offshore areas (Thomas, 1990). As noted by Nelson and Bray (1970), this paleo-relief is not defined bathymetrically on the present sea-floor. Mound-like features with heights on the order of 2 to 4 m occur at or near the river confluences (Fig. 4.2).

Because it has been modified by shoreface ravinement and tidal-inlet scouring during transgression, the Holocene-Pleistocene surface only approximates the pre-transgressive topography. Landward bifurcating channels, attributed to tidal-inlet scouring, are incised into Pleistocene sediments beneath the flood-delta and tidal-inlet of San Luis Pass (Fig. 4.2). Similar topography occurs east of the confluence of Trinity and Sabine incised valleys. A paleoshoreline trace over this feature suggests that it may have been tidally carved. The modifications caused by shoreface ravinement are discussed later.
Holocene sequences overlie weathered Pleistocene substrate of firm deltaic clay and rare sand interbeds. Sandy Pleistocene fluvial terraces occur along the Trinity River and Sabine River incised valleys (Thomas, 1990) (Fig. 4.4). A sand-rich channel, probably a continuation of a Pleistocene Brazos distributary channel, occurs off the western end of Galveston Island. Its landward continuation currently is occupied by Chocolate Bayou (Fig. 4.4).

In cores, a sharp increase in stiffness and cohesiveness, low water content, presence of calcareous nodules, and a change of color to brown, yellow, red, or tan, occur beneath the Holocene-Pleistocene boundary. In seismic records, where Holocene sediments are at least 2 m thick, a prominent high amplitude reflector, traceable over much of the study area, tracks the Holocene-Pleistocene boundary (Fig. 4.5). This reflector also defines incisions into Pleistocene sediments a few meters deep and a few to several hundred meters wide. Pleistocene sediments are represented by a series of parallel, even, continuous reflectors. Incised valleys are identified predominantly by abrupt truncation of these reflectors. The bases of the incised valleys are difficult to image, probably because diffusion of gas from decaying organic material within the valley-fill prevents acoustic penetration (Fig. 4.5). Shells from Pleistocene sediments exposed on the inner shelf yielded radiocarbon dates ranging from >26,000 to >40,000 years BP (Jeffrey S. Williams, pers. comm.).

**Ravinement Surface**

The ravinement surface and the Holocene-Pleistocene surface are amalgamated within most of the interfluvial areas (Fig. 4.6). Over incised valleys, 35 to 40 m of fluvial and estuarine valley-fill separates the two surfaces (Thomas, 1990). Generally, the ravinement surface is deeper over incised
Figure 4.4. Subcrop map beneath the ravinement surface (modified from Thomas, 1990). Also indicated are locations of a seismic profile shown on Figure 4.5 and transect along the Trinity River incised valley on Figure 4.9.
Figure 4.5. Seismic profile across an incised valley (RAV = ravinement surface, SF = shelf deposit, FTDF = flood-tidal delta facies, EF = estuarine deposit). Location of profile is indicated on Figure 4.4.
Figure 4.6. Structure contour map of the basal ravinement surface. In dashed line is a trace of the Trinity-Sabine incised valley on the shelf.
valleys (Fig. 4.5). Shallower ravinement outside the valley probably results from the stiffness and cohesiveness of Pleistocene sediments; they are more difficult to erode than the softer, less cohesive valley-fill sediments. Holocene estuarine sediments that escaped shoreface ravinement on the inner shelf usually are associated with incised valleys and their tributaries (Fig. 4.4). To the west, pods of estuarine deposits probably are associated with the Pleistocene channel. This indicates that the ravinement process was efficient in stripping previous coastal lithosomes and estuarine/lagoonal deposits that may have formed within the interfluvial areas.

Transgressive lag deposits, normally associated with ravinement surfaces, are not pervasive features. In some cores, marine mud directly overlies estuarine mud without an intervening shell/pebble lag or a "coarse" sand layer. Shell/pebble lags consist mostly of shells and shell fragments from the species *Rangia, Crassostrea*, and *Anadara*. Calcareous nodules typically comprise the occasional lithic pebble encountered in the cores. The above components constitute the coarse fractions of surficial sediments on the inner shelf (Morton and Winker, 1979; McGowen and Morton, 1979). *Crassostrea*-bearing sediments occur mostly over or near incised valley-fills, while *Rangia*-bearing sediments and lithic fragments are common in interfluvial area offshore of Sabine and High Island (Morton and Winker, 1979) (Fig. 4.7). *Anadara*, although more common, has a more random distribution. Radiocarbon dating shows that most of the *Rangia* are Pleistocene in age, while most of the *Crassostrea* and *Anadara* are Holocene (Morton and Winker, 1979). The distribution of fossil pelecypods and radiocarbon dates strongly indicate that the coarse components were derived from underlying sediments, and that redistribution was limited.
Figure 4.7. Distribution of coarse components of inner shelf sediments (from Morton and Winker, 1979).
The Holocene-Pleistocene surface overlain by Galveston Island exhibits a profile that mimics the shoreface profile (Fig. 4.8). Cores on Galveston Island indicate that this surface is a continuation of the ravinement surface, and that it underlies the shoreface and inner shelf wedge. Scarps or step-like features, with 2 to 3 m of relief, are superimposed on the overall concave Holocene-Pleistocene surface; these divide the concave profile into several segments. The profile could have been inherited from the pre-transgressive topography. The Holocene-Pleistocene surface underlying the southwestern Louisiana chenier plain is an example of an inherited topographic surface. The Holocene-Pleistocene surface is separated from the ravinement surface by a thin veneer of marsh and bay deposits. This indicates that the topography exhibited by the Holocene-Pleistocene surface was not carved or modified by shoreface ravinement.

A trace of the ravinement surface along the axis of the Trinity River incised valley shows the occurrence of two ravinement surfaces in some areas on the shelf (Fig. 4.9). The lower ravinement surface is produced during an initial transgression associated with a relative sea-level rise. Ebb-tidal-delta progradation over the lower ravinement surface occurs during a subsequent relative sea-level still-stand. The upper ravinement surface is formed during another rapid sea-level rise that translates the shoreline to a more landward position. If there is only one recognizable ravinement surface on the shelf, it could represent an amalgamation of two ravinement surfaces. The scarp on the Holocene-Pleistocene surface would be associated with the most updip shoreface position prior to the progradation of coastal lithosomes. Thus, intense sculpturing of Pleistocene sediments occurs during transgression. There is no
Figure 4.8. Cross-sections of Galveston Island (modified from Bernard et al., 1970).
Gulf of Mexico

- A - Anomalous Sample (contains reworked older shell fragments)
- Insufficient Sample (diluted with coal)

Sand and Clay
Sand with Thin Shell Beds

Source: Ward et al.,
Figure 4.9. Section along the Trinity River incised valley. Location of transect is indicated on Figure 4.4.
indication that the advancing toe of the shoreface is erosive during progradation.

Bergman and Walker (1987), Walker and Eyles (1991), and Pattison and Walker (1992), documented similar step-like features on a transgressive surface within the Cardium Formation in Alberta, Canada. They attributed the features to shoreface ravinement. They suggest that shoreface scarp incisions form during relative stillstands.

Sand Banks

Several large-scale, elongate, asymmetrical sand bodies with strong along-strike orientations, and bases that lie upon the ravinement surface (Holocene-Pleistocene surface in the interfluves) occur on the east Texas shelf (Siringan and Anderson, 1991) (Fig. 4.2). Two of these sand bodies, Sabine and Heald banks, are described below.

Sabine Bank, the largest sand bank on the Texas shelf, accounts for almost half of the arcuate-shaped, ~90 km long ~18 km wide, chain of banks that parallels the east Texas - west Louisiana coastline. It occurs in water depth of 12.5 to 5 m, has 7.5 m of relief, and is asymmetric with a steep and highly irregular seaward side characterized by several northwestward trending grooves and finger-like seaward-projecting lobes. Grab samples indicate that the bank is covered by \( \geq 95\% \) sand (mode of 2.75 \( \phi \), mean grain size range of 1.95 \( \phi \) to 2.58 \( \phi \)) with a small fraction of gravel-sized shell (Hamilton, 1991). The shell fraction increases on the seaward projecting lobes. Vibracores at the crest of the bank penetrated almost 2 m of amalgamated, graded shell to sand units (2.5 \( \phi \) - 3.0 \( \phi \) mode), interpreted as storm deposits, in an overall coarsening upward sequence (Fig. 10). Platform borings penetrated interbedded sand and
mud (Nelson and Bray, 1970; Thomas, 1990). Sand beds thin rapidly on both sides of the bank (Fig. 4.10). Cores on the seaward side sampled a thin veneer of sand resting on the Pleistocene-Holocene surface, sand on the landward side interfingers with shelf mud and is underlain by estuarine/lagoonal clay. The base of the bank, at -12 to -14 m, is a ravinement surface (Thomas, 1990). Bedforms observed over most of the bank include small ripples and sand waves with wavelengths of a few meters located below the crests and on the seaward slope of the bank (Nelson and Bray, 1970). Thomas (1990), likewise, reported diver observations of low-amplitude (1-3 cm) symmetric ripples and asymmetric sand waves with 1 to 3 m amplitudes and 600 to 1200 m wave lengths, measured on seismic profiles and depth records. The crests of the sand waves appear oriented NW-SE. Core and seismic data indicate that the landward portion of the bank overlies the Holocene estuarine Sabine River incised valley-fill (Fig. 4.10b). The segmented seaward portions rest directly upon a Holocene-Pleistocene surface high (Figs. 4.2 and 4.10a).

Brackish water shells taken from Sabine Bank yielded $^{14}$C ages ranging from 12,360±540 yBP to 6,195±350 yBP, with a cluster around 7.5 to 6.5 kyBP. Marine shells yielded $^{14}$C ages ranging from 215±150 yBP to 4,697±265 yBP, with a grouping around 4.6 kyBP (Nelson and Bray, 1970). The highly mixed shell assemblage and wide range of ages imply that the materials comprising Sabine Bank are reworked. This is supported by the ravinement surface beneath the bank and associated preserved coastal lithosomes. The northwestward trending grooves on the bank and the bank's overall geometry suggest oblique west-northwest landward reworking or growth. Graded sand and interfingering inner shelf mud indicate episodic growth, triggered during storms and hurricanes (Fig. 4.10b). $^{14}$C age dates of shells (converted to U/Th
Figure 4.10.  a) Cross-section of Sabine Bank based on seismic and core data. b) Lithologic log of vibracores along the section. Location of transect indicated in Figure 4.2.
ages after Bard, 1992), imply that the bank may have been derived from a 6 kyBP to 4 kyBP paleobarrier/strandplain.

Heald Bank lies in water depths of 14 to 8 m (Fig. 4.2). A vibracore from the bank penetrated less than 0.5 m of a densely packed shell layer that grades up to medium- to fine-grained sand. Heald Bank has a maximum thickness of 8 m, indicated by a borehole that terminated in Pleistocene sediment and penetrated a coarsening upward trend consisting of clayey sand to clean sand (Nelson and Bray, 1970). Most of the bank is located directly on top of a Holocene-Pleistocene surface high (Fig. 4.2). Radiocarbon dates of marine shells from the borehole yielded an age range of 7520 yBP to <3180 yBP, while brackish species gave an age range from 10,952±420 to 5350±300 yBP (Nelson and Bray, 1970). The similarities of Heald Bank with Sabine Bank, suggest a similar origin.

**DISCUSSION**

The Holocene transgressive record on the east Texas shelf and coast is characterized by: 1) the continuity of the ravinement surface on the shelf and under the present coastal systems; 2) the occurrence of scarps on the Holocene-Pleistocene surface and double ravinement surfaces; 3) discrete pods of preserved tidal-inlet and tidal-delta lithosomes and poor to non-preservation of shoreface lithosomes (Chapter 3); 4) poor to non-preservation of bay/lagoonal sediments in interfluvial areas (Chapter 3); and 5) backstepping parasequences within the incised valleys (Thomas, 1990). Although these patterns are observed throughout the study area, the presence of other features indicate temporal and spatial differences of regional transgressive histories.
Discontinuous erosional shoreface retreat (Swift, 1975, 1976), in-place drowning (Sanders and Kumar, 1975; Rampino and Sanders, 1980, 1982), and transgressive submergence (Penland et al., 1988) models have been presented to relate shoreline migration and shelf sand body formation with changing rates of sea-level rise (Fig. 4.11). Shoreface retreat refers to a process whereby the base of the shoreface translates landward, truncating pre-existing facies (Fig. 4.11a). Discontinuous erosional shoreface retreat may occur when a period of stillstand (during the process of long-term barrier transgression with sea-level rise) allows the barrier to stabilize. Stability may cause the vertical component of shoreface translation to increase at the expense of the horizontal component (Swift, 1975, 1976; Swift and Moslow, 1982). The stratigraphic signature of discontinuous erosional shoreface retreat is a ravinement surface overlain by a thin, often discontinuous, sand sheet (Swift, 1976). The barrier, if preserved, would have its upper part decapitated (Stubblefield et al., 1983).

In-place drowning describes a process in which both barrier and lagoonal sediments accrete vertically, keeping pace with relative sea-level rise (Fig. 4.11b). Rapid relative sea-level rise results in overstepping of the barrier and generation of a new shoreline along the landward edge of the former lagoon. The stratigraphic signature of in-place drowning consists of a thickened lagoon and a barrier sand sequence largely preserved intact and perhaps overlain by an erosional unconformity.

Transgressive submergence describes a process of shoreline and shelf sand generation in which the horizontal component of reworking occurs during shoreface retreat and in conjunction with a vertical component of submergence acting to preserve the sequence (Fig. 4.11c). The submerged barrier later is
Figure 4.11. Models of shoreline response to rapid sea-level rise; a) discontinuous erosional shoreface retreat (Swift 1975), b) in-place drowning (Sanders and Kumar, 1975; Rampino and Sanders, 1980), and c) transgressive submergence (Penland et al., 1988).
reworked by shelf processes to form a shelf sand body. The stratigraphic signature of transgressive submergence consists of a lagoonal deposit separated by a ravinement surface from an overlying coarsening upward shelf sand body. These 3 models were examined to evaluate which model best fits the east Texas continental shelf.

4 ka Transgression

The mineralogical and textural variations of the present coastal lithosomes record an important aspect of the latest transgressive history (the past 4 ky) of the region. Cole and Anderson's (1982) mineralogical and textural analyses of Galveston Island and Bolivar Peninsula sand fractions showed that: 1) they are comprised of mixed Mississippi, Brazos, and Trinity-Sabine river sediments; 2) the older portions of Galveston Island are composed of Trinity River derived-sand; Bolivar Peninsula shows a similar trend, although less well-defined because of fewer samples; 3) the modern shoreface contains a higher proportion of Mississippi River sand with accreting segments of the shoreline exhibiting the highest concentrations; and 4) the younger sediments have a dominant mode of 3.00 φ, and the rest have a prevalent mode of 2.75 φ. Cole and Anderson (1982) concluded that sediments comprising the bulk of Galveston Island and Bolivar Peninsula were derived from an offshore source and not directly from fluvial sources acting in conjunction with longshore currents.

The predominantly Trinity River derived-sand of the older portions of Galveston Island probably were reworked from Pleistocene sediments through shoreface ravinement or tidal-inlet scouring. A Pleistocene fluvial terrace, truncated by the ravinement surface, is located on the eastern flank of the Trinity
River incised valley (Thomas, 1990) (Fig. 4.4). It extends from the seaward to the bayward side of Bolivar Peninsula. Direct sourcing from Trinity River was not possible. Radiocarbon dates of shells and peat deposits show that the flooding of the Trinity River incised valley that led to the formation of Galveston Bay, occurred between 12 ka and 4 ka, well before Galveston Island and Bolivar Peninsula formed. Also, valley-fill studies have shown that Trinity and Sabine rivers were never able to build fluvial deltas that encroached into the gulf (Rehkemper, 1969; Nelson and Bray, 1970; Pearson et al., 1986; Thomas, 1990; Smyth 1991). Instead, the fluvial portion of the valley-fill is overlain by a thick muddy estuarine sequence. Sand accumulates within bay-head deltas and in more up-dip regions. Engineering borehole logs and seismic data indicate that the ravinement surface, underlain by the Bolivar Roads ebb-tidal-delta, was not deep enough to have cut through the bay-head delta sediments. However, ravined upper bay deposits also may have contributed sand to the present coastal lithosomes.

The predominantly Trinity River-derived sand of the oldest portions of Galveston Island and correlation of the shell components of transgressive lags on the inner shelf with the underlying ravined sediments, suggests that sediments mined through shoreface ravinement are not dispersed or carried too far landward or seaward during transgression. This is consistent with observations on the Atlantic shelf where sediments derived through shoreface ravinement appear to travel for only a short time before being deposited on the trailing inner shelf wedge (Niederoda et al., 1985).

Absence of preserved barrier lithosomes associated with the 6 to 4 ka paleo-shoreline trend offshore of Galveston Island and Bolivar Peninsula (Chapter 3) (Fig. 2) supports Cole and Anderson's (1982) proposition that a bulk of their
sediments were offshore-derived. It suggests that a large portion of the previous barrier lithosomes were reworked onshore during transgression. Common occurrence of >6 ky to >4 ky $^{14}$C shell ages from Galveston Island supports this (Bernard et al., 1959, 1970) (Fig. 8). Encroachment of mud in the upper section of the shoreface resulted in a fining upward trend and may represent the depletion of the offshore sand source: depletion enhanced the signature of mud and the Mississippi-derived sand (Cole ands Anderson, 1982; Chapter 3).

Unlike Galveston Island and Bolivar Peninsula, sediments comprising the Sabine Pass and southwestern Louisiana chenier plains were not derived from an offshore source, they were delivered by longshore currents from the east (Gould and McFarlan, 1959; Roberts et al., 1989). This is supported by the young $^{14}$C age dates of shells within the chenier complex (~3.5 ky or younger. Sand from previous coastal lithosomes were stranded on the inner shelf and reworked by marine processes to form Sabine Bank. The stranding of the sediments of the previous coastal lithosomes may have influenced which coastal environments developed (e.g., chenier plain instead of a strandplain), but it was inconsequential to their rapid establishment because of a high sediment input from longshore currents from the Mississippi Delta area and low wave energy, which favored chenier plain formation.

The reworked nature of sediment comprising Galveston Island and Bolivar Peninsula and the above features are compatible with a discontinuous erosional shoreface retreat (Swift, 1975, 1976). To the east, Sabine Bank and the long shore-derived nature of sediment comprising the Sabine Pass chenier plain suggest a mechanism more similar to transgressive submergence (Penland et al., 1988). However, the discrete nature of preserved coastal
lithosomes, in particular the tidal-inlet facies (Fig. 4.2) (Chapter 3), can be used as an argument for *in-place drowning* mechanism (Sanders and Kumar, 1975; Rampino and Sanders, 1980; 1982; 1983). Lack of similar sediments between the previous and present shoreline positions may be interpreted to mean that the barrier never existed in the intermediate areas. An underlying assumption is that the depth of tidal-inlet incision is uniform along the coast and that this depth is maintained during transgression. In the in-place drowning model, truncation of the barrier and back-barrier deposits is minor, presumably because the shoreface does not pass continuously across the area submerged (Sanders and Kumar, 1975; Rampino and Sanders, 1980). However, on the east Texas shelf, decapitation of coastal lithosomes at or lower than the wave-base allowed preservation of only the deeply incised tidal-inlets (Fig. 4.9) (Chapter 3). This is exemplified by the lack of the tidal-inlet facies associated with the Big Slough ebb-tidal delta and Curtis Bank flood-tidal delta deposits (Chapter 3) (Fig. 4.2). Thus, lack of tidal-inlet indicators may imply that the tidal-inlets that formed were not incised deep enough to have ample portions of the inlet-fill escape ravinement. It also may indicate that deep tidal-inlet systems, even if associated with incised valleys, may evolve into shallow inlets during rapid transgression. Foundering also can occur leading to the formation of embayments.

**Discontinuous Erosional Shoreface Retreat.** Along the east Texas coast, relative sea-level stillstand during the past 3.5 ky has been characterized by a large seaward shoreface progradation and a minor vertical shoreface translation (Fig. 4.8). Contemporaneous lithosomes along the south Texas coast exhibit a more dominant upward shoreface translation (Fisk, 1959).
Presently, along the east Texas coast, the accelerating rise of sea-level has caused renewed landward shoreface and shoreline translation and bayshore erosion, resulting in the narrowing of Galveston Island and Bolivar Peninsula (Morton, 1979; Morton and Paine, 1985; Chapter 2). As pointed out by Leatherman (1979), a barrier island needs to reach a critical width, through erosion of its bay and seaward shores, before barrier translation becomes operative (Fig. 4.12a). Once this width is reached, transfer of sediment over the island becomes efficient via inlet and overwash processes and the barrier "rolls over" onto it, a process termed "barrier roll-over" (Leatherman, 1979; 1983) (Figs. 4.12 b and 4.12c). At this time, transgression would be accelerated. Sea-level rise rates in excess of 2 cm/y have been estimated as the threshold for rapid transgressions (Penland et al., 1991). The rate of sea-level rise associated with the flooding event 4 ka may have been as high as 3 to 5 cm/y (Anderson and Thomas, 1991).

The erosional shoreface mechanism proposes that, during shoreface retreat, the trailing inner shelf wedge becomes a repository of sediments eroded from the shoreface (Figs. 4.11a and 4.12c) (Swift, 1975). However, recent bathymetric changes off Galveston Island and Bolivar Peninsula, complemented by core data, indicate erosion on the inner shelf (Chapter 2). Yet, the reworked nature of sediments comprising the present coastal systems supports the model. The apparent disparity suggests that offshore transport of sediment is more efficient during rapid transgression (Fig. 4.12c). Similarly, rapid and extensive erosion due to storms produces a flattened shoreface profile and causes massive deposition on the inner shelf (Moody, 1964; Swift et al., 1985; Niederoda et al., 1985). Rapid transgression may have the same effect on the shoreface profile and inner shelf deposition. During this period,
Figure 4.12. General sequence of a discontinuous erosional shoreface event. a) Barrier configuration during a relative sea-level stillstand. Availability of sediment allows the seaward progradation of the shoreface. b) Narrowing of barrier island occurs during a slow rise. Sediments are transported largely away from the region by longshore currents, others may be deposited in the inner shelf. c) Barrier is narrow and has a very low profile. The rate of landward shoreline migration is rapid and the shoreface-inner shelf profile is flattened. The inner shelf wedge is thickened during this period. Rapid sea-level rise is probably related to this stage and the previous one. d) Stabilization of barrier island occurs during the deceleration of the rate of sea-level rise. The timing of stabilization may vary along the coast depending on the shelf gradient: in regions with steeper shelf gradient, reestablishment may occur earlier. Irregularities within the Holocene-Pleistocene surface may serve as nucleation sites for the new barrier islands. The shoreface profile steepens during barrier progradation.
shoreface regrading becomes inefficient, thus, more sediment gets stored on the inner shelf (Fig. 4.12c).

When the barrier island becomes narrow and low, more tidal-inlets will form. The inlets probably will be shallow, migratory, and ephemeral. Nevertheless, the inlets would contribute to the scouring of the Pleistocene and other Holocene paralic deposits (Fig. 4.12c). Kumar and Sanders (1975) and Rampino and Sanders (1982; 1983) have used the absence of preserved tidal inlet deposits on the inner shelf between the present coast and what they identified as a former barrier island position as an argument against the mechanism of erosional shoreface retreat. However, the shallower incision of migratory tidal inlets during rapid transgressions may explain the absence of blankets of preserved tidal-inlet deposits on the inner shelf.

Reestablishment of present coastal lithosomes probably occurred during the deceleration of the rate of sea-level rise. Initially, they were narrow, thin, and highly segmented (e.g., the numerous overwash channels on the backside of Galveston Island and Bolivar Peninsula) (Chapter 2). The mixed sand provenance of the bulk of Galveston Island and Bolivar Peninsula suggests that during the stabilization and subsequent progradation of coastal lithosomes, the sediment source shifts to the trailing inner shelf wedge (Fig. 4.12d). Sediments may have been driven onshore, to the shoreface, by asymmetry of shoaling waves and tidal- and storm-induced currents (Pilkey and Field, 1972; Niederoda et al., 1985). A similar model describes the barrier evolution sequence for New South Wales (Roy and Thom, 1981). Availability of a large amount of sediment during this phase resulted in the rapid establishment and progradation of coastal lithosomes.
Big Slough Bathymetric High (BSH), the most landward and smallest of the banks on the east Texas shelf (Figs. 4.2 and 4.13), developed on the inner shelf off Galveston Island. It has a 2 m relief, a 15 km length, and a 5 km width. A decapitated ebb-tidal delta cores BSH (Chapter 3). A small kink on the seafloor, produced by the decapitated ebb-tidal-delta section, may have been sufficient to create flow divergence, as described by Huthnance (1982), and serve as a nucleation site for the muddy bank (Fig. 4.13).

**Transgressive Submergence.** - Off High Island and Sabine Pass, very rapid rates of transgression caused the stranding of sediments on the shelf; the roll-over mechanism of sediment transfer was overtaken by shoreline migration (Fig. 4.11b). Rates of shoreline translation may have varied from 9.3 m/y at the western end of Galveston Island to almost 92.6 m/y at Sabine Pass. This is approximated from the shoreline position prior to and immediately after the rapid sea-level rise 4 ka and a duration of transgression of approximately 500 years. The faster transgression off Sabine Pass is due to a gentler shelf gradient. The rate of sea-level rise associated with this transgression may have been 3 to 5 cm/y (Anderson and Thomas, 1991). The above rates of transgression are not unrealistic values. In Louisiana, the average barrier shoreline retreat rates are less than 20 m/y with a relative sea-level rise of 0.5 cm/yr (Penland et al., 1988). Stranding of the barrier occurs because of faster mainland shoreline retreat (rates in excess of 25 m/y) (Penland et al., 1988).

The apparent continuity of the ravinement surface, from the previous shoreline position to the landwardmost edge of transgression, indicates that the shoreface passed through the entire area (Fig. 4.10a). The absence of backbarrier deposits on the interfluvial areas of this region attests to ravinement.
Figure 4.13. Big Slough Bathymetric High and Big Slough ebb-tidal delta.

Location of section is shown on Figure 4.4.
At the place where the stranding occurred, what is identified as the ravinement surface could be a product of marine erosion.

Sabine Bank's overall geometry and position with respect to the probable source of sediment, indicated by the decapitated tidal-inlet and flood-tidal-delta pair near the eastern flank of the bank, suggests that the bank has a slight landward and a more dominant southwest, along-shelf, growth components (Figs. 4.2 and 4.10a). The position of Sabine Bank may have been controlled in part by the Holocene-Pleistocene surface (Fig. 4.2 and 4.10a). Protuberances from this surface on the southern flank of Sabine River incised valley may have served as nucleation sites (sand accumulation) and/or slowed the landward migration of the bank.

The east and west geographical variation of the nature of the latest shoreline translation seems to hold true for the shoreline migration of the 8 to 6 ka paleoshoreline trend to the 6 to 4 ka paleoshoreline position. Heald Bank is the equivalent of Sabine Bank for this period of transgression. Similar to Sabine Bank, a Holocene-Pleistocene surface high may also have served as a nucleation site and as an "anchor" (Fig. 4.2).

**Pre-8 ka Transgressions**

The shelf sand banks located landward of the paleoshoreline trends (an unnamed bank north of paleoshoreline trend 5, the southwestern segment of Shepard Bank landward of paleoshoreline trend 4, and Curtis Bank for paleoshoreline trend 3) suggests transgressive submergence as the mechanism of shoreline translation during this period (Fig. 2). The 4 m average depth of ravinement is comparable to the depth of ravinement on the eastern half of the study area during the transgressions of the past 8 ky.
The stratigraphic signature for in-place drowning, aggradational barrier and lagoonal deposits with minor decapitation, occurs along the central Texas shelf. A preserved 20 m thick barrier sequence, buried under a ~3.5 m thick inner shelf deposit, was penetrated by a borehole located approximately 7 km off Matagorda Island in -13 m water depth (Shepard, 1956) (Fig. 4.14). The sequence correlates to the preserved coastal lithosomes on the east Texas shelf that define paleoshoreline trends 3 and 4 (Chapter 3). The barrier sequence consists of stacked barrier flat/inlet facies, dune facies, then back to inlet facies. If Shepard's (1956) facies interpretations are correct, the stacked barrier sequence may represent two episodes of overstepping. The first event is indicated by the overstepping of the barrier flat/inlet facies and the second event by the overstepping of the dune facies. Thus, the preserved barrier sequence represents stacked barriers. These correlate to the stacking observed in paleoshoreline trends 3 and 4, represented by an ebb-tidal delta overlying a flood-tidal delta (Chapter 3) (Figs. 4.2 and 4.14). These systems also may correlate with Stubblefield et al.'s (1983) series of mid-shelf coast-parallel ridges, interpreted as partially degraded barriers drowned by rapidly rising sea-level off New Jersey. These barriers lie at depths of -20 to -40 m and are estimated to have formed between 8 and 13 ka.

The occurrence of an aggradational sequence on the central Texas shelf is related to a steeper shelf gradient and higher rate of sedimentation on this region. On the east Texas shelf, the coastal lithosomes show a strong progradational architecture (Bernard et al., 1959; Gould and McFarlan, 1959) because of a gentle shelf gradient that provides limited accommodation space. In contrast, present coastal lithosomes along the central and south Texas shelf regions exhibit strong aggradational architecture (Fisk, 1959; Shepard, 1956)
because the steeper shelf gradient provided greater accomodation space. Also, greater sediment supply on the central Texas shelf allowed the aggradation of coastal lithosomes during sea-level rise. At present, convergence of westward flowing longshore currents from the east, and northward flowing longshore currents from the south brings large amounts of sediment into this region (Shideler, 1978). High sediment supply enabled Matagorda Island and Padre Island to aggrade and prograde (Shepard, 1956; Fisk, 1959; Wilkinson, 1974) (Fig. 4.14).

Even during the past 8 ky, in-place drowning may have occurred on the central Texas shelf. Shepard's (1956) barrier facies interpretation of Matagorda Island indicates stacking of barrier facies (Fig. 4.14). Landward encroachment of the nearshore gulf facies (Shepard, 1956), probably equivalent to lower shoreface, may represent the 4 ka transgression.

Prior to 8 ka, sediment supply to the central Texas shelf may have even been higher. During this period, sediment supply to the Gulf was greater because of higher fluvial discharges resulting from more humid climates (Chapter 3). Higher wave energies also may have increased longshore sediment transport rates.

Kumar and Sanders (1975) and Rampino and Sanders (1980, 1982) interpret the apparent minor decapitation of aggradational sequences as indicative of a shoreface jump during transgression. However, even if decapitation during transgression is greater than 5 m, aggradational barrier and lagoonal deposits may still have preserved sections that may indicate minor truncation. Also, shallowing of the shoreface during rapid transgression as a result of the flattening of the shoreface-inner shelf profile may also lead to minor decapitation. In the central Texas shelf setting, high sediment supply (derived
Figure 4.14. Preserved barrier island sequence off Matagorda Peninsula (from Shepard, 1956).
alongshore) may allow the thickening of the inner shelf sediment wedge without necessitating erosion of the shoreface wedge as might have occurred on the east Texas shelf during the 4 ka transgression.

**SUMMARY and CONCLUSIONS**

1) The ravinement surface appears to be continuous on the shelf and under the present coast, indicating that the shoreface passed through the entire shelf during transgression. Generally, the ravinement surface is deeper over incised valleys.

2) Transgressive lag deposits, normally associated with ravinement surfaces, are not pervasive features. The coarse components of the transgressive lag are derived from underlying sediments; redistribution is limited.

3) Although the overall effect of shoreface ravinement is planation, scarps or step-like features were carved on the Holocene-Pleistocene surface during transgression. Estuarine/lagoonal deposits that escaped shoreface ravinement on the inner shelf usually are associated with incised valleys and their tributaries. This indicates that the ravinement process was efficient in removing previous coastal lithosomes and estuarine/lagoonal deposits that may have formed within the interfluvial areas.

4) If there is only one recognizable ravinement surface on the shelf, it could represent an amalgamation of two ravinement surfaces, a product of progradational phases that occur during the relative stillstands of the step-like rises of sea-level.

4) During transgression, barrier sand may be: 1) lost permanently through removal by longshore currents; 2) stored temporarily on the inner shelf during
rapid transgression and eventually reworked onshore during a relative stillstand; or 3) get stranded on the inner shelf to be reworked later by shelf processes.

5) The dominant source of sand for coastal systems may change through time. Shoreface ravinement and tidal-inlet scouring provide sand from Pleistocene and Holocene sediments during barrier migration and early stabilization of coastal systems. Once stabilized, onshore transport of sand provides shelf-derived sediment. Depletion of this sand source increases the importance of longshore-derived sand. However, in other areas, longshore-derived sediments may be dominant throughout.

6) Shoreline translations along the east Texas shelf, after 8 ka, were characterized by discontinuous erosional shoreface retreat on the western half of the study area, and by transgressive submergence on the eastern half. The very gentle shelf gradient in the east resulted in rapid rates of transgression and submergence.

7) Sabine Bank and Heald Bank are products of transgressive submergence. They are associated with the 6 to 4 ky and 8 to 6 ky paleo-shoreline trends, respectively. Holocene-Pleistocene surface highs contributed to the sequestering of sand, ultimately leading to the formation of these banks.

8) The transgressive record prior to 8 ka are consistent with transgressive submergence on the east Texas shelf and in-place drowning on the central shelf. Aggradation of sequences on the central Texas shelf are caused by steeper shelf gradient and higher sedimentation rates.

9) Shelf gradient, rate of sea-level rise, and rate of sediment supply are the major variables that control the mechanism of shoreline transgression.
SUMMARY OF CONCLUSIONS

The nature of sea-level rise, rate of sedimentation, and pre-transgressive geology (shelf gradient, topography, and lithology) are the three dominant factors that control the distribution, internal geometry/stratigraphy, nature of translation, and preservation of Holocene coastal lithosomes along the east Texas coast and adjacent continental shelf.

Following a rapid sea-level rise 4 ka, the present coastal systems along the east Texas coast evolved during the relative sea-level stillstand of the past 3.5 ky. These relative stillstands and rapid rises are superimposed on the overall rising Holocene sea-level. Bolivar Peninsula and Galveston Island initially were thin, narrow and probably low features, formed by spit accretion. They experienced frequent breaching by storm overwash channels that may have served as ephemeral tidal inlets. Reactivations of the channels ceased as the features widened through beach ridge accretion. Bolivar Roads inlet/delta complex formed approximately 3.3 ka as the spits accreted bayward. Incised valley flank obstruction impeded inlet migration and caused further constriction of the inlet. In addition, relative sea-level rise and closure of smaller inlets increased the tidal prism resulting in enhanced tidal current velocities, deeper inlet incision, and ultimately, stabilization of the inlet. Subsequently, inlet migration has occurred within a narrow zone. Enhanced tidal currents increased the tidal influence resulting in the formation of a robust ebb-tidal delta and drumstick-shaped barrier terminations in the inlet.

The present tidal inlet exhibits channel-stacking and cut-and-fill structure composed of sand, shell, and mud interbeds. Individual channels are filled with westward-dipping clinoforms that downlap the channel base. The tidal-inlet/spit
facies is characterized by oblique-tangential clinoforms, building-out and
deepening from the valley edge toward the valley center. In dip-section, the
clinoforms are more parallel and display gentle dips. The flood-tidal delta
facies has a concave upward base, thins bayward, and interfingers with bay
sediments. Inlet-proximal portions exhibit channel stacking and cut-and-fill, with
trough-like channel geometry on the seaward side becoming broad and
shallow bayward. Channels exhibit prograded-fill that may represent lateral
accretion. Sand and clay interlaminations are ubiquitous in the tidal-deltas.
Sand and shell beds are common in inlet-proximal regions. The ebb-tidal delta
exhibits gently inclined clinoforms prograding over the ravinement surface.
Mud-dominance of Bolivar Roads inlet/delta complex predominantly results
from high influx of mud in the area.

The distinctive architecture and geometry of facies comprising Bolivar Roads
tidal inlet/delta complex may serve as analogues for subsurface and outcrop
interpretations. The Bolivar Roads system provides an example of an inlet
system that formed through baymouth closure by spit accretion.

Concomitant with Bolivar Roads tidal inlet/tidal delta sequence development
is the deposition of the shoreface and inner shelf lithosomes. The shoreface is
characterized by cross-cutting channels of varying sizes that are interpreted to
be storm-return flow channels that may have served as sand conduits to the
inner shelf during storms. Despite the abundance of these channels and the
high frequency of storms and hurricanes that have impacted the region during
historical times, there is a paucity of storm beds on the shelf. The storms
deposits that do exist are patchy in distribution and typically are laminated to
very thin-bedded. Indicators of massive offshore sand transport to the shelf
during storms are lacking. This probably results from strong along-shelf storm
currents that control not only the extent of offshore transport of sand but also the amount of sand deposited in the region. The strong along-shelf storm currents, combined with low sediment input and low effective accommodation space, resulted in a thin marine sediment cover and poor preservation of storm beds on the east Texas shelf. Greater occurrence of storm beds during the early development of extant coastal lithosomes resulted from more frequent breaching of the narrower barriers and higher sand supply.

Amalgamated sand beds on the east Texas shelf, potentially of storm origin, are possible preserved lower shoreface deposits or reworked Holocene coastal lithosomes. The geometry and internal architecture of the sand banks on the inner and middle shelf indicate landward and shore-parallel (to the southwest) growth. This implies that, during storms, onshore and along-coast transport of sand probably is more dominant than offshore movement.

Greater preservation of storm beds on the central Texas shelf results from the high influx of fine-grained sediment and greater effective accommodation space. The east and central Texas shelf shows that along-shelf and across-shelf variations of bed thickness and lateral continuity of storm deposits may be caused by: 1) differences in preservation and not necessarily variations of storm frequency and magnitude; and 2) differences in the sediment source and transport mechanism -- seaward and along-coast transport from shore zone, in-situ reworking, or onshore transport.

The poor preservation of the lower shoreface and inner shelf deposits during a slow transgression is exhibited on the east Texas shelf. Rising sea-level during historic times has caused extensive erosion off Galveston Island and Bolivar Peninsula. In general, erosion increases toward the shoreface base, then decreases farther offshore, and increases from east to west. Greater
shoreface-base translation, in relation to shoreline translation, resulted in the general steepening of the shoreface profile. This extensive erosion has caused as much as a kilometer of shoreface retreat within a span of 26 years.

Previous episodes of shoreface retreat on the inner to middle continental shelf left behind discrete pods of tidal inlet, tidal inlet/spit, and tidal delta deposits; lower shoreface facies are rare. Preserved tidal inlet facies occur only within incised valleys, while preserved ebb- and flood-tidal delta facies occur within and outside the incised valleys, though they are more common within the valley. The distribution of these preserved coastal lithosomes mimics the along-strike variation of the extant coastal system and supports the model of a step-like nature of sea-level rise. In addition, their distribution defines six relative sea-level stillstands, including the present, during the past 10.2 ky; present sea-level from ~ 4 ka to present, at -6 m from 6 to 4 ka, at -11 m from 8 to 6 ka, at -17 m from 8.5 to 8 ka, at -20 m from 9.5 to 8.5 ka, and at -29 m from 9.5 to 10.2 ka.

The seismic architectures of the preserved coastal lithosomes indicate that sediment supply, effective accommodation space, and wave and tidal energies were lower during the past 8 ky than the period prior to 8 ka. Steeper shelf gradient beyond -20 m allowed greater wave energies to impinge on the coast. Faster rates and greater magnitudes of sea-level rise prior to 8 ka, created greater accommodation space, and tidal energies were focused within flooded incised valleys resulting in greater landward tidal reach. In addition, sediment supply was greater prior to 8 ka because of a more humid climate. Thus, the pre-8 ka flood-tidal-delta and tidal-inlet facies and fauna differ from those of the present.

The preserved coastal lithosomes indicate that the depth of ravinement decreases with decreasing shelf gradient, increasing rate of sea-level rise, and
increasing sediment supply. But they also show that a rapid sea-level rise does not guarantee preservation. Better preservation within incised valleys results from greater accommodation space and the soft valley-fill that allows incision of the inlets beyond the depth of shoreface ravinement.

The ravinement surface appears to have passed across the entire shelf. On the shelf, if only one ravinement surface is recognized, that surface could be an amalgamation of two ravinement surfaces, a product of progradational phases that occur during the relative stillstands of the step-like rises of sea-level. Although the overall effect of shoreface ravinement is planation, scarps or step-like features were carved on the Holocene-Pleistocene surface during the episodes of rapid transgression. Estuarine/lagoonal deposits that escaped shoreface ravinement on the shelf predominantly are associated with incised valleys and their tributaries. This indicates that the ravinement process was efficient in removing previous coastal lithosomes and estuarine/lagoonal deposits that may have formed within the interfluvial areas.

The transgressive record in the western half of the study area, after 8 ka, are consistent with discontinuous erosional shoreface retreat. To the east, a very gentle shelf gradient caused rapid transgressions that favored transgressive submergence. Sabine and Heald banks, examples of transgressive shelf sand bodies on the east Texas shelf, formed during separate episodes of transgressive submergence. They are associated with the 6 to 4 ka and 8 to 6 ka paleoshoreline trends, respectively. The transgressive record on the east Texas shelf prior to 8 ka is consistent with transgressive submergence; on the central Texas shelf, with both in-place drowning and discontinuous erosional shoreface retreat. Aggradation of sequences on the south and central Texas shelf were due to greater sediment supply and steeper shelf gradient.
REFERENCES CITED


FISK, H. N., 1959, Padre Island and the Laguna Madre flats, coastal south Texas: Louisiana State University, 2nd Coastal Geography Conference, p. 29-52.


MORTON, R. A., 1975, Shoreline changes between Sabine Pass and Bolivar Roads: The University of Texas at Austin, Bureau of Economic Geology, Geological Circular 75-6, 43 p.


Department of the Interior, 314 p.

in a microtidal estuary: in Smith, D. G., Reinson, G. E., Zaitlin, B. A., and

PENLAND, S., MCBRIDE, R. A., SUTER, R. J., BOYD, R., AND WILLIAMS S. J., 1991,
Holocene development of shelf-phase Mississippi River delta plains: in
Coastal Depositional Systems in the Gulf of Mexico, Gulf Coast Section
SEPM, Twelfth Annual Research Conference, p. 182-185.

PENLAND, S., BOYD, R., SUTER, J. R., 1988, Transgressive depositional systems
of the Mississippi delta plain: A model for barrier shoreline and shelf sand

PENLAND, S., BOYD, R., AND MCBRIDE, R. A., 1987, Delta plain development and
sea level history in the Terrebonne coastal region, Louisiana: American
Society of Civil Engineers, Proceedings, Coastal Sediments '87, p. 1689-
1705.

PENLAND, S., SUTER, J. R., BOYD, R., 1985, Barrier island arsc along abandoned

PENLAND, S., AND SUTER, J. R., 1983, Transgressive coastal facies preserved in
barrier island arc retreat paths in the Mississippi River delta plain:
Transactions Gulf Coast Association of Geological Societies, v. 33, p. 367-
382.

230-234.

PILKEY, O. H. AND FIELD, M. E., 1972, Onshore transportation of continental shelf
sediment: Atlantic southeast United States: in Swift, D. J. P., Duane, D.
Dowden, Hutchinson and Ross, Stroudsburg, Pa., p. 429-446.

PRICE, W. A., 1952, Reduction on maintenance by proper orientation of ship
channels through tidal inlets: Contributions Oceanography and Meteorology,
v. 1, no. 12, Texas A & M University, College Station, Texas, p. 75-107.

PRICE, W. A. AND PARKER, R. H., 1979, Origins of permanent inlets separating
barrier islands and influence of drowned valleys on tidal records along the
Gulf Coast of Texas: Transactions Gulf Coast Association of Geological


SHEPARD, F. P., 1956, Late Pleistocene and Recent history of the central Texas coast: Journal of Geology, v. 64, p. 56-69.


TOOMEY, R. S., 1992, Central Texas climates and environments 25,000 years B.P. to present: the vertebrate evidence: Geological Society of America, 26th Annual South-Central Section Conference Abstracts, p. 49.


Appendix 1

$^{14}$C dates from this study.
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Equation used for adjustment to U/Th Age (from Bard, 1982):

$$U/Th \text{ Age} = 1.24 \times (14C \text{ Age}) - 840$$
Appendix 2

List of Rice University Cores
(Core Type (G=gravity, VC=vibracore);
Depth and Length are in cm; NP=no penetration;
H/P=Holocene/Pleistocene Surface; Rav.= Ravinement Surface).
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PB = Pirates Beach  
SLP = San Luis Pass

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CL = off Clear Lake  
DP = off Dollar Point
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Appendix 3

Location Maps of Rice University Cores.
Location Map of OHI-90 Cores

Gulf of Mexico

Sabine Pass

94°00' 94°20'

29°40' 29°20'

Kilometers