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RICE UNIVERSITY

LATE QUATERNARY MARINE DEPOSITS, OFFSHORE CENTRAL TEXAS: PROCESSES CONTROLLING GEOMETRY, DISTRIBUTION, AND PRESERVATION POTENTIAL

by

MICHELLE LEE FASSELL

A THESIS SUBMITTED IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE MASTER OF ARTS

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Abstract

Late Quaternary Marine Deposits, Offshore Central Texas: Processes Controlling Geometry, Distribution, and Preservation Potential

by
Michelle Lee Fassell

The interplay of sediment supply, accommodation space, and the rate of sea-level rise and fall, determine the thickness and overall extent of offshore deposits along the central Texas shelf. Analysis of near-shore sediment cores suggests that transgressive and highstand shorelines prograded during the Holocene. The preservation of these deposits may be likely given the geometry and distribution of older, offshore shorelines. The Stage 5d and Stage 3 highstand shorelines prograded during an overall sea-level fall. High sediment supply from longshore transport preserved thick, aerially extensive shoreline deposits.

Fluvial incision varied significantly during lowstand Stage 2. As the rate of base-level fall increased from Stage 3 to 2, low sediment supply rivers incised narrow, shallow channels with no deltaic deposition, whereas sediment bypass and deeper incision prevailed for the high sediment supply rivers.
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Chapter 1: INTRODUCTION

1.1 Study Objectives

This study resolves the sedimentary processes active in the formation of older, near-shore marine sequences offshore central Texas by analyzing the distribution of sediments in modern shoreface profiles along the coast. Therefore, the first objective is to analyze the vertical progression of facies in modern shoreface cores, taking into account sediment type, grain size, and fossil content. The second objective is to determine the lateral extent of these facies. The third objective is to determine the extent and distribution of older near-shore marine sequences offshore. The final objective is to relate the lateral facies heterogeneity observed in the modern shoreface core transects to older late Pleistocene sequences offshore.

High sediment supply combined with low subsidence along the Texas coast, as well as high frequency sea level fluctuations during the late Pleistocene, yield stacked highstand, lowstand, and transgressive systems tract depositional packages that are within the resolution of cores and high-resolution seismic data. This makes the Texas continental shelf an ideal location to study the influence of shelf processes on highstand, lowstand, and transgressive depositional facies during the Pleistocene. In fact, the nature of sediment accumulation, fluvial sediment input, and shelf gradient varies significantly between east Texas, central Texas and south Texas, with widely differing depositional regimes. Sediment dispersal patterns along the central Texas shelf differ from those of the east and south Texas shelves due to the convergence of
longshore currents. One result is the deposition of extensive marine sheet-sands on the central Texas shelf (Eckles, 1996). Previous studies have recognized that similar marine processes have influenced sediment deposition along the central Texas coast since the Cretaceous (e.g. Winker, 1979; Meckel and Galloway, 1996; Eckles, 1996). Two of these studies have analyzed the regional distribution of late Pleistocene highstand prograding shorelines offshore (Eckles, 1996) as well as lowstand coalescing fluvial valleys on the central Texas inner shelf (Berryhill et al., 1986). These studies provide a regional depositional framework but do not explain the depositional processes responsible for facies heterogeneity observed in offshore platform borings (Eckles, 1996). In this study, facies variability is examined.

1.2 Organizational structure

This thesis consists of five chapters. Chapter 1 includes a general introduction to the methodology, sequence stratigraphic framework and implications for this study, overview of the study area, and previous research. Chapter 2 includes a comprehensive methodology used in this study. Chapter 3 addresses results from the modern shoreface transect study, including regional shoreface trends, definitions of shoreface facies, and vertical stacking patterns. The discussion focuses on a comparison of grain size trends, interpretation of transgressive ravinement processes responsible for shaping the modern shoreface, and a comparison of central Texas shoreface preservation potential versus east Texas shoreface preservation potential. Chapter 4 reexamines
offshore seismic data and platform borings in relation to fluvial deposits on the
shelf. Stratigraphic concepts, such as the effects of base level and gradient on
the nature of fluvial incision during lowstands, are addressed. In addition,
highstand shoreline extent and lithologic character is examined. Chapter 5
compares the modern shoreface thickness and extent to the Stage 5 and Stage
3 highstand shorelines. Evidence is presented to suggest that localized high
sediment supply is the key factor influencing the preservation of thick highstand
shorelines.

1.3 Study Area

The central Texas coast is located between Matagorda Peninsula and
North Padre Island, extending 80 km from latitude 29°30' to 27°. The inner shelf
is a low gradient (~1.2 m/km) shelf from the shoreface to the 100-meter isobath

1.3.1 Wave and Current Regime

An inherent cyclicity in sedimentation on the shelf is due in part to the
curvature of the coastline and the nature of longshore current transport. Short-
term, seasonal variability in ocean circulation patterns and wave intensity on the
shelf is due to the nature of the predominant wind direction. Fair-weather
summer months are characterized by southeasterly winds with moderate wave
intensity producing onshore-directed currents. Storm-induced onshore-directed
southwesterly winds generate steep, asymmetric waves that strike the eastern and southern Texas coast at oblique angles (McGowen et al., 1977). Winter months are characterized by alternating onshore, fair-weather southeasterly's combined with offshore directed northeasterly storm fronts (Shideler, 1981) establishing variable onshore and offshore directed currents (Curray, 1960). Due the curvature of the Texas coastline, northerly fronts set up high intensity waves offshore in east Texas, along shore waves in central Texas, and onshore-directed waves in south Texas; therefore, erosion is the most intense in the central and south Texas shelves during the winter (McGowen et al., 1977). These high intensity waves set up longshore current transport towards the west along the east Texas coast and toward the north along the south Texas coast, converging near central Padre Island at approximately 27°N latitude. The convergence of these currents along the central Texas shelf controlled sedimentation patterns throughout the Cenozoic (Curray, 1960) (Figure 1-1).

The converging longshore currents entrain suspended sediment from the east (particularly from the Brazos and Colorado Rivers) and Rio Grande River to the south. These currents are thus effective in accumulating thick Holocene deposits, locally in excess of 40 m on the central Texas shelf (Eckles, 1996). The different density water masses flowing from the east and south converge along the central Texas shelf forming a turbid benthic nepheloid layer (Shideler, 1981). The nepheloid layer entrains fine-grained silts and clays in the convergence zone due to internal breaking waves associated with onshore
Figure 1-1: Map showing the direction of current convergence along the Texas coast. Winds are predominantly from the southeast. These winds set up longshore currents which converge just south of the study area, near Baffin Bay (modified from Lohse, 1956).
directed converging currents and offshore directed ebb-tide flow and coastal runoff. The relative intensities of these onshore and offshore fronts control seasonal fluctuations in current intensity and sediment supply (Shideler, 1981).

1.3.2 Storm-related Sedimentation

Storm intensity has a dramatic effect on sediment movement along the coast. Over 60 hurricanes have struck the Texas coast this century. In 1961, hurricane Carla struck the Texas coast at Pass Cavallo (central Texas shelf), causing a tidal surge of 4.1 m above sea level (Hayes, 1967; Morton and Pieper, 1976). Hayes (1967) sites dune erosion up to 50 m. Sediment was likely transported offshore in storm surge channels and deposited over a wide geographic area with a thickness up to 6 cm (Hayes, 1967). A recent study by Snedden and Nummedal (1991) determined that the Carla bed still extends over most of the central and south Texas shelf (Figure 1-2). The wide distribution of the Carla bed, as well as thicknesses up to 6 cm, suggests that the effectiveness of storm transport of silts and sands onto the shelf is high. Moreover, the likelihood of short-term preservation of these sediments is also high (Snedden and Nummedal, 1991).

The process by which the Carla bed and other storm beds are deposited is typically by offshore-directed bottom currents. This process is termed geostrophic flow. Onshore-directed hurricanes and tropical storms from the southeast set up intense onshore-directed oceanic flow. The resultant build up in pressure along the coast causes offshore directed bottom flow. Sediment
Figure 1-2: Present-day Carla Storm Bed isopach thickness map. The contour interval is 2 cm. In less than -20 m water depth, significant bioturbation obscures bed identification. After 30 years, up to 6 cm of silty sand remains in the lower shoreface. (Modified from Snedden and Nummedal, 1991).
dispersion is deflected offshore and to the south due to the Coriolis force. These geostrophic currents cause very fine sand to be transported onto the lower shoreface as storm beds (Swift, 1976).

1.3.3 Tidal Regime

Tidal range varies between 30 to 80 cm (1-2.5 feet) and is diurnal to semidiurnal. The Texas coast is microtidal with ebb-current velocities slightly greater than flood currents (Morton and McGowen, 1980). Flood tidal currents range from 1 to 3.5 knots and ebb currents range from 1 to 4.3 knots. A high tidal range, up to 3 m, is typical during large hurricanes, whereas a low tidal range predominates during northerly fronts (Bernard and LeBlanc, 1965).

1.3.4 Subsidence

Regional uplift onshore Texas is controlled by isostatic uplift due to sedimentary loading offshore. The hinge line, or the zone between uplift of fluvial terraces inland (Pleistocene and older deposits) and subsidence offshore Texas, has shifted 1-3 km seaward since Marine Oxygen Isotope Stage 5 (about 120,000 years) due to deltaic deposition (Bernard and LeBlanc, 1965). In contrast, the hinge line offshore the Mississippi River has shifted approximately 80 km seaward since Stage 5 due to high sediment accumulation rates and associated high subsidence rates (Bernard and LeBlanc, 1965).

In general, subsidence rates are low along the Texas shelf with rates on the order of 0.1 mm/yr (Winker, 1979). Subsidence rates are much higher at the
shelf edge and within incised valleys, with rates on the order of 5 mm/yr along the shelf edge (Winker, 1979) and 1.0 mm/yr within incised valleys (Paine, 1993). Abdullah (1995) calculated subsidence rates from Quaternary sequences along the east Texas shelf. Rates of 0.13 mm/yr for the inner shelf and 2 to 4 mm/yr for the outer shelf were determined and are similar to rates found by other workers (Abdulah, 1995).

1.3.5 Fluvial Systems

Deposition on the central Texas shelf is strongly affected by the long-lived fluvial discharge patterns in the region. The east Texas shelf is supplied with sediment from three large rivers, the Trinity, Brazos and Colorado Rivers and the combined drainage area extends 300,000 km² from eastern New Mexico to coastal Texas (Houston Embayment Drainage area) crossing from humid to semiarid climatic zones (LeBlanc and Hodgson, 1959; Winker, 1979). Rivers with small drainage basins (40,000 km²) and small sediment yields, characterizing a vast inter-deltaic setting dominate the central Texas coast. The Rio Grande River (semiarid climatic zone) in south Texas has occupied approximately the same location throughout the Tertiary, receiving sediments shed from the Colorado plateau (400,000 km² drainage area) (LeBlanc and Hodgson, 1959) (Figure 1-3).
Figure 1-3: Drainage area map of central Texas rivers. In black are the central Texas rivers. In gray are the Colorado and Rio Grande rivers. River names are shown. Bays are abbreviated. MAB: Matagorda bay. SAB: San Antonio Bay. ACB: Aransas/Copano bays. CCB: Corpus Christi bay. BB: Baffin bay. (modified from LeBlanc and Hodgeson, 1959).
1.4 Sequence stratigraphic framework for the Texas shelf

The sequence stratigraphic framework of the east, central, and south Texas shelves has been developed over the past 10 years based on four criteria: determination of 1) stratal bounding surfaces, 2) the character of seismic facies contained within stratal boundaries, 3) facies identification based on core material, and 4) the actual timing of stratal formation.

Initially, stratal boundaries were identified based on sequence stratigraphic slug models developed by Vail et al. (1977) and Posamentier et al. (1988) (Figure 1-4). Typically, the bounding seismic surfaces constrain highstand, lowstand, and transgressive depositional units based on stratal geometries, such as prograding clinoforms, fluvial incision, and onlap, respectively (Thomas, 1990; Abdullah, 1995; Banfield, 1998; Snow, 1998). The lower bounding surface of a highstand unit is called the downlap surface (Vail et al., 1977) or maximum flooding surface and is formed as sea level floods the shelf and then begins to fall (Figure 1-5; HST - highstand systems tract). This surface is generally characterized by a high amplitude reflection. The upper boundary of a highstand unit is the sequence boundary, or lowstand surface. This surface is identified based on truncation of underlying highstand reflectors and represents the maximum fall of sea level, typically to the shelf break (Vail et al., 1977) (Figure 1-5; LST: lowstand systems tract). The transgressive surface (or ravinement surface) is the first marine erosional surface that floods the shelf, causing coastal deposits to step landward (Nummedal and Swift, 1987) (Figure 1-5; TST: transgressive systems tract). This surface has been shown to form
Figure 1-4: Block diagram illustrating stratigraphic position of HST (highstand systems tract), LST (lowstand systems tract), and TST (transgressive systems tract) deposits. The sea-level curve is a generalized sinusoidal curve illustrating the timing of stratal formation (modified from Posamentier et al., 1988).
Figure 1-5: Two sea-level curves from 140,000 years to present are shown. These curves are oxygen isotope curves constrained using U-Th dates from corals in Barbados and New Guinea. Triangles indicate lowstand deposits on raised reef terraces. Terminology is HST (Highstand Systems Tract); LST (Lowstand Systems Tract); TST (Transgressive Systems Tract). Marine oxygen isotope stages are written at the top. (data compiled from Bard et al., 1990 and Chappell et al., 1996).
steps upward and landward due to increases in the rate of sea-level rise (Thomas and Anderson, 1994).

After the establishment of a sequence stratigraphic framework, a detailed examination of the facies architecture within sequences can be performed using platform borings, long cores, and shallow hammer cores. Seismic facies interpretations were corroborated using core lithofacies.

1.4.1 Global oxygen isotope curves and stratigraphic studies

Imbrie et al. (1984) used oxygen isotope analyses ($\delta^{18}$C per mil) conducted on planktonic foraminifera in selected cores from low latitude marine environments to deconvolve the control of Milankovitch forcing on climate. High correlation was determined between orbital variations and climate, resulting in a detailed Pleistocene ice-volume reconstruction. The SPECMAP curve of Imbrie et al. (1984) is used as a proxy for relative sea-level fluctuations in the Pleistocene (Williams, 1983; Anderson et al., 1996; Eckles, 1996). Correlation of actual sea-level curves derived from seismic stratigraphic studies to the SPECMAP curve has met with varying success in the Gulf of Mexico, due to autocyclic events on the shelf (Thomas and Anderson, 1991). A relatively precise sea-level curve from 9,000 to present has been created using Carbon 14 dates of corals from the Caribbean (Lighty et al., 1982). This curve has been combined with a Uranium-Thorium sea-level curve from 9,000 to 18,000 yBP created from coral material in Barbados (Bard et al., 1990) (Figure 1-6). The
Figure 1-6: Sea-level curve from Stage 2 to 1. The curve in gray is U/Th dates of Barbados corals (modified from Bard et al., 1990). The curve in black is a Carbon-14 curve of corals from the Bahamas, Florida, Martinique, Panama, Puerto Rico, and St. Croix, dated from 8,500 yBP to present (modified from Lighty et al., 1982). U/Th dates of coral terraces are more accurate from 9,000 yBP to 18,000 yBP compared to Carbon 14 dates and are shown here. HST is highstand systems tract; TST is transgressive systems tract; LST is lowstand systems tract. This sea-level curve is used in this study.
latter curve constrained the original Carbon 14 sea-level curve from 9,000 to 18,000 yBP from Fairbanks (1989). This composite curve is used in this study.

The oxygen isotope curves developed for the east, central, and south Texas shelves were developed to link the oxygen isotope stages (sea-level highs and lows) to depositional units and erosional surfaces observed in seismic data from the Texas shelf. The oxygen isotope curves developed along the Texas coast show reasonable correlation with the seismic stratigraphic interpretations of these studies. Comparisons to the global sea-level curves suggest good correlation during oxygen isotope Stage 5, Stage 2, and the overall transgression from Stage 2 to 1. The sea-level position established for oxygen isotope Stage 3 in this study is inconsistent with the sea-level position that is established in global sea-level curves.

Several global sea-level curves place the position of Stage 3 at -60 m (Labeyrie et al., 1987) and -75 m (Shackleton, 1987) (Figure 1-5). Uranium-Thorium dates of exposed coral terraces in New Guinea (Chappell et al., 1996) and Barbados (Bard et al., 1990) support these sea-level positions. Other sea-level curves suggest a shallower sea-level position for Stage 3. Curray (1965) proposed a sea-level position along the Texas coast of -10 m at 30,000 yBP, based on Carbon 14 dates of shell material. A detailed oxygen isotope curve from the western Pacific Sulu Sea suggests sea-level was at -40 m at 40,000 yBP (Linsley, 1996). Similarly, Chappell and Shackleton (1986) established that sea level was at -41 m at 40,000 yBP based on U/Th dates of New Guinea coral terraces.
Detailed stratigraphic studies as well as Carbon 14 analyses of shell material along the Atlantic and Texas coasts suggest sea level was at or near the present position during Stage 3. Finkelstein and Kearney (1988) dated a preserved mid-Wisconsinan Stage 3 lagoonal peat deposit along the Delmarva Peninsula. They suggest sea level was at -2.5 m, 33,000 yBP. A similar study by Wellner et al. (1993) suggested a sea-level position of -20 m along the New Jersey coast during Stage 3.

Independent studies along the Texas coast suggest sea level ranged between -40 and -15 m during Stage 3 time (Banfield, 1998; Snow; 1998; Rodriguez, 1999; this study). This range in sea-level position is similar to stratigraphic studies along the Atlantic coast but is significantly shallower than most global sea-level curves suggest. The stage 3 sea-level position is based on mapping the landward pinchout of Stage 3 highstand deposits along the east, central, and south Texas shelves. The timing was constrained by oxygen isotope curves (Abdulah, 1995; Eckles, 1996; and Banfield, 1998) and Carbon-14 dates on articulated mollusk shells within Stage 3 deposits (Snow, 1998; Rodriguez, 1999).

1.4.2 Regional oxygen isotope curves

Abdulah (1995) constrained the timing of deposition along the east and central Texas shelves based on oxygen isotope curves generated from benthic (Quinqueloculina sp.) and planktonic foraminifera (Globigerinoides ruber) from platform boring B-146 (Figure 1-7). Sequence timing was also determined based
Figure 1-7: Oxygen isotope curve and paleobathymetry curves from the study of Abdullah (1995). Good correlation exists between this chronology and the central Texas dataset. The location of Platform boring B-146 is shown on Line 20, Figures 4-8 and 4-24.
on the paleo-water depths derived from the benthic foraminifera *Quinqueloculina* sp. It was shown that increases in paleo-water depth correspond to decreases in δ¹⁸O and, similarly, decreases in paleo-water depth correspond to increases in δ¹³O. Stage boundaries were determined based on water depth and planktic and agglutinated foraminifera abundance. The Stage 2 sequence boundary has the heaviest δ¹⁸O value (2.0), lowest foraminifera abundance and reflects the shallowest water depth (Figure 1-7). Maximum flooding surfaces 3 and 5e have a high foraminifera abundance and negative δ¹⁸O values (Abdullah, 1995). Substages 5c and 5a have moderate water depths (middle neritic) with low to moderate planktonic and agglutinated foraminifera abundance (Figure 1-7). The surfaces identified by Abdullah (1995) were correlated to surfaces identified within the central Texas dataset.

Additional timing constraints were provided by oxygen isotope curves generated by Eckles (1996) and Banfield (1998), from the central and south Texas shelves, respectively. Eckles (1996) used the planktonic foraminifera *Globigerinoides ruber* from platform boring B-92 to generate her oxygen isotope curve and Banfield (1997) also used *Globigerinoides ruber* to generate an oxygen isotope stratigraphy from core B-2 (Figure 1-8). The chronostratigraphy developed by Banfield (1998) and Abdullah (1995) correlates well with that established for this study. However, comparison with Eckles' (1996) isotope curve suggests good correlation from Stage 2 to present. Older isotope stages,
Figure 1-8: The oxygen isotope curve modified from the study of Banfield (1998). The curve was generated from *Globogerinoides ruber* foraminifera. Timing of events as well as well as Ericson- Wollin zones are shown. This chronology and isotope stage identification shown here ties well with the central Texas seismic dataset.
identified by Eckles (1996), did not correlate to the surfaces identified in this study.

1.4.3 Lithologic units and bounding surfaces

Seismic surfaces, facies, and lithologic units identified in this study were compared to those of Eckles (1996) and Siringan (1993). The lowstand surface, or sequence boundary, was identified based on incision and truncation of underlying reflectors. Erosional surfaces, commonly manifested by erosion and shell lags, are common in cores. Pleistocene deposits consist of stiff, oxidized clays, owing to exposure during sea-level fall. Several transgressive ravinement surfaces are identified in this study. The outer shelf transgressive ravinement surface is identified in seismic data as an erosional surface above a back-stepping transgressive delta. Snow (1998) first identified this surface in her study of the marine Stage 2 to 1 transgressive systems tract deposits of the Colorado River. In near-shore sediment cores, the transgressive ravinement surface is manifested by a shell lag. Several highstand sequences were identified based on downlapping reflectors onto a high amplitude surface as well as truncation by overlying lowstand sediments.

1.5 Previous Work

1.5.1 Coastal erosion along the Texas coast

A large portion of this study focuses on understanding the evolution of the central Texas shoreface during the overall transgression (Stage 2-1) and
during the current highstand. Very little nearshore data exists along the Texas coast, especially central Texas. The following is a review of the current knowledge that exists on barriers of the Texas coast.

Over the past 150 years a series of shoreline studies have been performed along the Texas coast by the Bureau of Economic Geology. Morton and Pieper (1976), Morton et al. (1976), and Paine and Morton (1989), Morton (1997), to name a few, have documented shoreline movement using a series of monitoring stations along the Texas coast. Techniques used to determine shoreline position tend to be fairly subjective, with considerable variability in chosen shoreline location (high water line, vegetation line, berm crest, etc.).

Morton and McGowen (1980) differentiate between high-profile accretionary barriers and low-profile transgressive barrier Islands based on barrier growth seaward and retreat landward, respectively. East Texas barriers are thought to be transgressive (Morton and McGowen, 1980). Galveston Island, on the east Texas coast, is considered a complex accretionary and transgressive island (Morton and McGowen, 1980). Galveston Island shows considerable seaward growth on the eastern side of the Galveston seawall (Bolivar peninsula and east beach) and significant retreat at the west end of the Island (Morton, 1997; Paine and Morton, 1989).

Erosion is predominant along deltaic headlands and peninsulas due to low profiles offshore of these deltas, absence of significant dunes, and low sediment supply. West of Galveston Island, Follets Island to Brown Cedar Cut is an eroding deltaic headland, having eroded at rates up to 14.5 m/yr from 1974 to
1982. Brown Cedar Cut to Matagorda Peninsula is considered a transgressive barrier that has stepped landward due to shoreline erosion and storm washover (Morton and McGowen, 1980; Paine and Morton, 1989), with retreat rates from 0.5m to 1.0m with a maximum up to 8m/yr (McGowen et al., 1977). Sediment removal from the deltaic headlands has been most severe during the Holocene due to climatic influenced decreases in sediment delivery to the coast (Morton and Pieper, 1976; Snow, 1998).

Accretion occurs west of the Matagorda ship channel and significant erosion occurs down-drift of the jetties (Paine and Morton, 1989). Matagorda Island, extending from Pass Cavallo to Cedar Bayou, is a high profile accretionary barrier that has grown seaward up to 2.4 m/yr over the last 30 years. In contrast, San Jose Island, extending from Cedar Bayou to Aransas Pass, has retreated at rates up to 3.7m/yr (Paine and Morton, 1989). North Padre Island, extending from Port Aransas to Yarborough Pass, is stable to slightly recessional and is a narrow barrier Island formed mainly by upward and landward growth (Morton and McGowen, 1980). Central Padre Island showed a net advance averaging 0.2m/yr from 1974 to 1982, due to converging northerly and southerly directed longshore currents in this location (Paine and Morton, 1989).

Tidal inlets of the Texas coast initially formed near lowstand drowned River valleys (e.g. Bolivar Roads) (Siringan and Anderson, 1993). The migration of these inlets is mainly controlled by longshore currents (McGowen et al., 1977). Pass Cavallo, separating Matagorda Island from Matagorda Peninsula, has
migrated 11.2 km west of the Lavaca River lowstand drowned river valley, where it initially formed (McGowan et al., 1977).

1.5.2 Effects of sea-level rise on the shoreface

Numerous articles have been written discussing the effects of sea-level rise on the evolution of barrier islands and shoreface profiles (Swift, 1968; Sanders and Kumar, 1975; Swift, 1976; Belknap and Kraft, 1985; Everts, 1985; Everts, 1987; Pilkey and Davis, 1987). Bruun’s early work on shoreface profile response to sea-level rise (1962) is the foundation for most of these shoreface profile studies. The following is a discussion of the drawbacks of the “Bruun rule”, as well as alternative concepts proposed to determine shoreface profile response to rising sea level.

Bruun (1962) proposed a theory for shoreline evolution that suggests that the shoreline maintains a “longshore quantitative equilibrium” where “the same quantity of sediment that is passing in from the updrift side is also passing out downdrift” during sea-level rise (Figure 1-9). The shoreline will shift landward and upward, eroding the upper shoreface and aggrading the lower shoreface, thus creating equilibrium conditions (Bruun, 1962). Bruun (1962) considers sediment removal from the shoreface by currents as a geologically long-term process that does not need to be considered in landward migration equations. This is based on a study done on the California coast showing little sediment transport below 18 m water depth. This same ‘closure depth’ is applied to all coasts regardless of profile gradient or the depth of the lower shoreface (Bruun,
Figure 1-9: Bruun model of shoreface translation landward during sea-level rise. The shoreline shifts upward and landward with a zone of shoreface erosion and zone of aggradation offshore. (modified from Swift, 1976).
1962). The recession rate, proposed by Bruun (1962), is indirectly proportional to the shape of the shoreface slope, therefore, low shoreface gradients equal high recession rates and high gradients equal slow recession rates.

Swift (1976) agrees that shore retreat approximates equilibrium but only when the shoreline is mature (rate of change of sediment input, wave conditions, and sea level are constant). He contends that the equilibrium profile undergoes evolution during rapid sea-level rise. If the coast is initially flat, the effects of rapid sea-level rise, such as shoaling waves, will cause the profile to become flatter. On the other hand, a stillstand in sea level will allow the profile to attain the maximum gradient, depending on sediment type, input, and wave condition, approximating an equilibrium shoreline (Swift, 1976).

Everts (1985, 1987) determined analytical shoreline equations that suggest shoreline translation landward causes the shoreline gradient to change. Everts (1985, 1987) agrees that the shoreface profile is in equilibrium with the position of sea level, similar to Bruun (1962). The difference is that Everts (1985, 1987) considers removal or addition of sediments from/to the shoreface separately from the effects of sea-level rise. Thus, the equilibrium profile does not simply involve onshore and offshore movement of sediments (as the Bruun model suggests) but also alongshore movement, storm-dominated movement, and removal of sediments through tidal inlets and overwash. Everts (1985, 1987) suggests that in order to maintain the equilibrium profile relative to sea-level position, the shoreface sediment must be sand-size or greater. Sediment less than 0.063 mm (4.0 φ) will not be stable in an active (upper) shoreface
environment and will be removed by offshore/onshore currents and longshore transport. The resulting shoreface profile after sea-level rise will be lower than the initial profile (Figure 1-10a). Moreover, sand-sized sediment deposition onto the shoreface will cause an increase in the original shoreface gradient relative to the original lower sea-level position (Everts, 1985; Everts, 1987) (Figure 1-10b).

In summary, early equilibrium shoreface profile concepts suggested that shoreface sediments had to be conserved updrift or downdrift of the shoreface (Bruun, 1962), whereas later studies (Everts, 1987; Swift, 1976) contend that the equilibrium shoreface profile is controlled by an interplay of sediment removal offshore and alongshore, sediment input from longshore drift, and variability in the rate of sea-level rise. The shoreline will shift landward or seaward depending on all of these factors and will eventually achieve equilibrium conditions. Other studies contend that the concept of an equilibrium profile is too simplistic given the rapidly changing conditions in the shoreface through time as well as the variability in the older substrate below the shoreface (Williams and Meisburger, 1987; Pilkey and Davis, 1987; Pilkey et al., 1993).

Given a stillstand or slow rise in sea level (present conditions), retreat rates of barriers and the associated shoreface depends on grain size of available sediment in the shoreface environment (Everts, 1987; Pilkey and Davis, 1987; Pilkey et al., 1993), the number of inlets to trap sediment from littoral drift (Everts, 1987; Pilkey and Davis, 1987), and the initial shape of the shoreface (Everts, 1987).
Figure 1-10: Everts (1985) model of shoreface erosion during shoreline retreat. Figure A suggests shoreface gradient increase when the sedimentation rate is high. Figure B shows a gradient decrease due to erosion and low sediment supply.
Everts (1987) contends that portions of east coast barriers (Ocean City, Maryland; Sandbridge, Virginia; and the northern Outer Banks) are narrowing and retreating slowly due to overwash, loss of sand through inlets, and offshore movement of sand during storms. Everts (1987) also suggests that continual narrowing to a critical width (possibly 350 m) will cause Islands to migrate landward at high rates - possibly up to 5 to 8 times faster.

Grain size appears to be a critical factor in shoreface stability. The shoreface shape is controlled by an interplay of grain size and wave energy. The bottom orbital wave energy is highest in the upper shoreface, typically moving sand-sized sediment onshore during fair-weather times and offshore in friction-dominated bottom flow during storms (Swift, 1976). The threshold velocity needed to maintain fine sand in suspension is higher than for silt and clays, therefore, sand-sized sediment typically falls out of suspension in the upper shoreface (during fair-weather). On the other hand, silts, clays, and very fine sand are carried in suspension farther offshore into the lower shoreface. Movement offshore of the mud-sized and very fine sand-sized sediment contributes to shoreface sediment loss and potential recession.

The amount of sand versus mud in the shoreface can be critical for maintaining the shoreface profile given a slow sea-level rise. A shoreface with a low sand/mud ratio will tend to have higher coastal and shoreface recession rates (Pilkey and Davis, 1987; Swift, 1976). In the high-energy upper shoreface, muddy sediments will be taken into suspension and transported along-strike and/or offshore into the lower shoreface.
Similarly, the initial shoreface profile shape (before coastal recession) is dependent on sand/mud content (Everts, 1987). Sandy shorefaces are as steep as 10 degrees and have a higher angle of repose than muddy shelves (<1 degree) (Swift, 1976). Therefore, given a slow sea-level rise a mud-dominated shoreface will retreat faster than a sand-dominated shoreface due to: 1) muddy suspended sediment removal at higher rates than sand-sized sediment, and 2) a low angle mud-dominated shoreface will flood faster than a high angle sandy shoreface (Pilkey and Davis, 1987).

1.5.3 Late Pleistocene studies

The second portion of this study focuses on understanding the temporal and spatial stratal stacking patterns in offshore highstand and lowstand sequences. Previous studies by Berryhill et al. (1986) of late Pleistocene lowstand deposits of the region have shown an extensive network of coalescing fluvial valleys offshore central Texas (Figure 1-11). These valleys are thought to run along-strike toward the south Texas shelf. Poor chronostratigraphic control existed in the study of Berryhill et al. (1986), making correlation with the seismic dataset in this study difficult, if not impossible.

1.5.4 Cenozoic depositional stacking patterns

Central Texas Cenozoic sedimentary formations, such as the Paleocene-Eocene Wilcox, the Eocene Yegua and the Oligocene Frio formations, all display
Figure 1-11: Late Wisconsinan fluvial valleys and deltas along the central and south Texas shelf. (Modified from Berryhill et al., 1986). In gray are the fluvial valleys and in black are the reefs.
facies patterns characteristic of storm-dominated coastlines (Fisher and McGowen, 1967; Galloway et al., 1982; Meckel and Galloway, 1996). These formations have a similar spatial geometry as the late Pleistocene highstand deposits of the Texas shelf (Eckles, 1996; Berryhill et al., 1986). The following discussion focuses on the similarities and differences observed in Cenozoic formations.

Fisher and McGowen (1967) examined the Lower Eocene Wilcox formation (stratigraphically lower than the Yegua Formation and farther updip). They found three main systems - the Rockdale delta system along the paleo east Texas shelf, the San Marcos strandplain/bay system along central Texas, and the Cotulla Barrier bar system/Indio bay-lagoon system in central to south Texas (Fisher and McGowen, 1967). The deltas of the Rockdale deltaic system are in a similar location to modern river locations (i.e. Brazos, Colorado, and Guadalupe deltas). The Cotulla barrier bar system/ Indio bay-lagoon system contains thick, very well sorted sand deposits, up to 30 m thick. Updip these sand deposits grade into muddy lagoonal facies. Processes controlling the thickness and distribution of the sand barrier bar is from longshore transport of reworked sands from the strandplain to the east. Deltaic influence appears to be minimal in this unit (Fisher and McGowen, 1967).

Meckel and Galloway (1996) performed a study of the extensive Eocene Yegua deposits along the Texas coast. In general, wave-dominated deltaic deposition prevailed along the central Texas shelf during initial sea-level fall (early highstand) with high sediment supply resulting in progradation. In
contrast, fluvial-deltaic sedimentation prevailed along the east Texas shelf during early highstand. Both fluvial-to wave-dominated deposition was prominent during the late highstand with extensive progradation due to high sediment supply. Early lowstand deposition occurred at the paleo-shelf edge. Wave-dominated deltas existed along central/south Texas, and a mix of fluvial-and wave-dominated deltas existed along the east Texas outer shelf. During late lowstand to early transgression, wave-dominated deltas generally prevailed along the Texas coast, with extensive linear shorelines situated along the central to south Texas coast. Finally, during the late transgression, when sediment supply was low and wave energy was high, wave-dominated deltas and a strandplain developed along central to south Texas; fluvial deltaic deposition was still active along the east Texas shelf during this time (Meckel and Galloway, 1996).

Galloway and others (1982) studied the stratigraphy and sedimentology of the Oligocene-Miocene Frio Formation. The paleogeography that existed during this time was similar to the Paleocene-Eocene Wilcox Formation (Fisher and McGowen, 1967), and the Eocene Yegua Formation (Meckel and Galloway, 1996). During Frio time, the Texas coast was dominated by two deltaic systems - the Houston delta system on the east Texas shelf and the Norias delta system on the south Texas shelf (Galloway et al., 1982). Along the eastern-most portion of the coast was the Buna barrier/strandplain system and the Greta/Carancahua barrier/strandplain system (Figure 1-12). Minor sediment input for the Greta/Carancahua strandplain system (along central Texas) was from the Choke
canyon/Flatonia streamplain. Large volumes of sediment are thought to have been sourced from the Houston delta system and Norias delta system by longshore current processes and were deposited onto the strandplain, similar to processes active in the modern Gulf of Mexico (Galloway et al., 1982) (Figure 1-12).

The paleogeography and depositional patterns described above are similar to those that existed during the late Pleistocene. Paleo-reconstructions from the Late Pleistocene from Winker (1979) and from the Cenozoic Frio Formation from Galloway et al. (1982) are shown in Figure 1-12. Both maps show an interdeltaic embayment across the central Texas shelf with associated strandplain deposits.
Chapter 2: Comprehensive Methodology

2.1 Core dataset

The shoreface study incorporates 105, 5 meter pneumatic hammer cores, acquired aboard the R/V Lone Star during the summer of 1997. Each of the 15 core transects consist of seven cores, with a 5 km spacing between transects. Each of the seven cores in the transects were taken in approximately 3, 5, 6, 7.5, 10.5, 12, and 13.5 m water depths (Figure 2-1). Sediment analyses included an initial examination and documentation of sediment type, grain size, color, texture, and sphericity. Bounding surface identification is based on the character of the lithologic contact; whether the contact is sharp and erosive or gradational. Lithofacies designation is based on sediment type, grain size, and fossil content as well as fining upward and coarsening upward trends. Shear vane tests were performed on stiff muds to aid in facies identification. The core dataset is presented and discussed in Chapter 3.

Approximately 100 samples were selected for statistical grain size analyses based on uncertainty of visual grain size assessments as well as proximity to key bounding surfaces. The <63 μm mud fraction was sieved from the sand using deionized water. The samples were then dried at 50°F in a oven overnight. Mineralogy and fossil content was assessed using a binocular microscope. Grain size of the sand-fraction was determined using an automated settling tube. The settling tube has a diameter of 17 cm and length of 140 cm. Settling velocities are converted to grain diameter (Phi units) analytically and are
Figure 2-1: Generalized study area map showing the core dataset, seismic lines and platform boring description. The oxygen isotope curve from Abdullah (1995) is designated by a star.
reported as phi-size, mean, mode, sorting, and skewness. To ensure statistical accuracy of the grain size measurements, the first, middle, and last samples of every 10 samples were re-analyzed.

Five Carbon 14 analyses were performed on selected fossil samples at key sedimentological boundaries. The specimens were sent to Beta Analytic, Inc. where they were acid etched with HCl to remove impurities. The available Carbon 14 was extracted using standard techniques. Isotopic dates are reported in conventional years before present.

2.2 Seismic dataset

Seismic data acquired during the summer of 1995, aboard the *R/V Lone Star*, was used in this study. The seismic dataset, initially interpreted by Eckles (1996), consists of 1200 km of 2-D, single channel, 15 in³ water gun data. Eckles (1996) applied Automatic Gain Control (AGC) during seismic collection as well as a bandpass filter of 90-180-720-1440. This study refines the initial interpretations by incorporating additional offshore platform boring descriptions and modern shoreface cores (Figure 2-1). Seismic facies identification is based on Mitchum et al. (1977) and Sangree and Widmier (1977) and stratal boundary identification is based on Vail et al. (1977). Seismic facies correlate to lithofacies identified in offshore platform borings.

Approximately 68 offshore platform boring descriptions and sediment cores, used in this study, were donated by Fugro-McClelland Engineers (Figure
2-1). Platform boring descriptions are typically 100 m long. Cross-sections have been created based on lithofacies and seismic facies comparisons.
Chapter 3: Preservation Potential of shoreface deposits

3.0: RESULTS: Central Texas shoreface

3.1 Introduction

The core dataset consists of 15 profiles, extending from Matagorda Peninsula to North Padre Island (Figure 3-1). Regional trends are presented in Section 3.1.1. Sections 3.1.2 and 3.1.3 focus on descriptions of sedimentary facies and bounding surfaces, respectively. In section 3.2, sedimentary facies are divided into facies units based on vertical stacking patterns. Grain size and radiocarbon results are also presented in this section.

The discussion focuses on the extent of shoreface units and the timing of facies unit formation. The factors influencing the preservation potential of shoreface deposits are also discussed.

3.1.1 Regional shoreface trends

Profiles 1 to 15 are shown in Appendix 1; 1-15 and are discussed in Section 3.2. This section provides a brief description of regional shoreface trends. A fence diagram for profiles 1-15 is shown in Figure 3-2 and in Plate 1.

There are four regional shoreface trends observed across the study area. The first trend is the occurrence of four distinct shoreface facies. These facies are upper shoreface, proximal lower shoreface, distal lower shoreface, and tidal delta complex. In addition, a marine facies and a Pleistocene facies are present. Grain size, sediment type, bed thickness, and fossil content are the key criteria
Figure 3-1: Core dataset map. Core profiles are numbered. NPI is North Padre island, MUI is Mustang Island, SJI is San Jose Island, and MAP is Matagorda Peninsula. Each profile is shown in Appendix 1: 1-15.
Figure 3-2: Fence diagram showing the trends of shoreface profiles 1-15. The stage 2 sequence boundary is deepest in profiles 1-4, 12. The sequence boundary depth is related to the presence or absence of fluvial incision. The Lavaca and Nueces rivers (profiles 1-4, 12, respectively) incised up to -36 meters within the bays during the Stage 2 lowstand. High accommodation space combined with high sediment supply has resulted in thick successions of shoreface 1 sediments preserved within these valleys below ravinement surface 2. Subsequent progradation related to the modern shoreface 2 has prograded and downlapped onto marine and distal lower shoreface sediments.
used to differentiate facies types. These characteristics will be discussed in
Section 3.1.2.

The second trend is the variability in the amount of shoreface versus
marine sediments present across the profiles. The volume of shoreface sand is
the greatest in profiles 11-15, offshore Mustang and Padre Islands. Shoreface
thickness ranges from 1 to 4 m in the upper shoreface facies, decreasing to 1-2
m farther offshore in the lower shoreface (Figure 3-2). The thickness of
shoreface sediments at the base of profiles 1-4 is similar to the thickness
observed in profiles 11-15. In contrast, only one meter of shoreface facies is
present in profile 8-10. The shallowest Pleistocene deposits observed in the
study area are found in profiles 8-10 (Figure 3-2).

The marine mud facies seen in the most distal cores is the thickest in
offshore profiles 1-5, up to 4 m thick in 13.5 m water depth (Figure 3-2). Marine
muds decrease in thickness to less than a meter in profiles 8-14 and were not
sampled in profiles 7 and 15 (Figure 3-2).

The third shoreface trend is the gradual increase in profile gradient from
profile 1 to profile 15, from east to west (Appendix 1: 1-15). Gradients increase
from 6 m/km in the upper shoreface facies of profile 1 to 10 m/km in profile 15.

The final profile trend is a similarity in the vertical stacking pattern across
all profiles. The succession of facies consists of basal Pleistocene deposits with
a sharp, erosional contact with an overlying shoreface succession. A sharp
contact separates these shoreface deposits from overlying marine to distal lower
shoreface facies. The top of the succession is composed of shoreface sediments (Figure 3-2).

3.1.2 Facies characteristics

The shoreface is divided into four distinct sub-environments. Terminology related to sub-environments is presented in Table 3-1. The upper shoreface (USF) is typically massive fine sands with rare thin silty sand beds (Figure 3-3). The sands are moderately to intensely burrowed, often by *Skolithos* traces (*Ophiomorpha* is the most common). Burrows are filled with muddy sand. Coarse shell hash is pervasive throughout the fine sand, with isolated beds of 100% shell hash (*Mulinia* are the most common mollusks). The USF extends from the base of the suri zone in one meter of water up to 12 m of water. 3 km offshore. Upper shoreface grain size is very fine upper (vfU) (3.0-3.5 φ) to fine lower (fL) (3.0-2.5φ). Rare 1-5 cm-thick muddy sand to clayey sand beds are present in massive very fine to fine sands.

The lower shoreface is divided into the proximal lower shoreface and distal lower shoreface. This division is based on sand bed thickness. The proximal lower shoreface contains 5-20 cm-thick, very fine to fine sand beds alternating with 5-10 cm-thick silty sand beds or mud (Figure 3-4). The distal lower shoreface contains less than 5 cm-thick, very fine sand beds encased in mud (Figure 3-5). Grain size ranges from very fine upper (vfU) to fine lower (fL) with a higher percentage of very fine upper sands.
<table>
<thead>
<tr>
<th>Term</th>
<th>USF</th>
<th>PLSF</th>
<th>DLSF</th>
<th>RS</th>
<th>SB</th>
<th>TDC (PTDC/DTDC)</th>
<th>Interpretaion</th>
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<tr>
<td></td>
<td>Upper shoreface</td>
<td>Proximal lower shoreface</td>
<td>Distal lower shoreface</td>
<td>Ravinement surface</td>
<td>Sequence boundary</td>
<td>Tidal delta complex (Proximal/Distal)</td>
<td>SF 1 (USF) / SF 2 (LSF) / RS 1 / SB 1 / DLSF 1, TDC 1</td>
</tr>
<tr>
<td></td>
<td>SF 2 (USF) / SF 2 (LSF) / RS 2 / SB 2 / DLSF 2, TDC 2</td>
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<thead>
<tr>
<th>Term</th>
<th>Grain size</th>
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<tr>
<td></td>
<td>FU</td>
<td>V/L/MU</td>
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<td></td>
<td>Fine lower (3.5-0.0 phi)</td>
<td>Fine upper (2.5 to 0.0 phi)</td>
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</table>

Table 3.1: Abbreviations and terminology used for core profile descriptions.
Figure 3-3: An example of the variability of USF facies. Black bars indicate a 5 cm scale.
Figure 3-4: Grayscale digital photos of variability observed across the Proximal lower shoreface (PLSF) facies.
Figure 3-5: Grayscale digital photos of variability observed in the Marine and DLSF facies. Photos A and B illustrate marine facies and C-E illustrate distal lower shoreface facies.
The distal lower shoreface (DLSF) contains thinly laminated, 1-5 cm-thick, very fine sand beds with shell hash encased in mud (Figure 3-5c-e). Grain size ranges from very fine upper to very fine lower. Hummocky cross-stratification is rarely observed. *Glossifungites* trace fossils are observed in muddy, compacted DLSF muds.

The tidal delta complex is divided into proximal and distal facies. The proximal facies contains predominantly coarse shell hash with medium to fine sands. The distal facies is similar to the proximal or distal lower shoreface facies, being comprised dominantly of mud with thin sand beds (Figure 3-6). Grain size of the sand fraction varies significantly from medium lower to very fine upper. Typically, the medium to fine sands are found in the proximal tidal delta, whereas the fine to very fine sands are found in the distal tidal delta.

Two additional facies are present in the shoreface cores. These are the marine facies and the Pleistocene facies. The marine facies is identified based on >20 cm-thick clay and silt layers. Rare mm-scale very fine sand beds are present (Figure 3-5a,b).

The Pleistocene facies consists of stiff mottled green and red clays and well rounded, clean sands. Shear strengths of the stiff clay range between 3.0 kg/cm² to 10.8 kg/cm². The sediments are often oxidized with calcium carbonate nodules present. Large *Ophiomorpha* burrows are present in the sandy intervals (Figure 3-7).
Figure 3-6: Grayscale digital photos of proximal and distal tidal facies. Core photo A-C indicate proximal tidal facies. Core photo D shows the distal tidal facies.
Figure 3-7: Grayscale digital photos of Pleistocene sediments.
3.1.3 Facies boundaries

Three facies boundaries have been identified across all core transects. Boundary terminology is presented in Table 3-1. The lowest facies boundary is an exposure surface, which separates Pleistocene facies from Recent shoreface sediments. This bounding surface is named SB 2/RS 1 (Figure 3-8a). Typically, a shell bed is present directly above this boundary.

A second facies boundary separates fine to very fine sand of the shoreface from overlying marine muds to distal lower shoreface deposits. This facies boundary is called RS 2 (Figure 3-8b,c).

The shallowest surface bounds shoreface sediments above from underlying marine to distal lower shoreface sediments. This bounding surface is named the downlap surface (Figure 3-8d,e).

3.2 Depositional environment interpretations

3.2.1 Vertical stacking patterns:

Facies unit terminology is based on similar vertical stacking patterns across all profiles (Figure 3-2). The terminology and abbreviations used in this section are shown in Table 3-1. These lithofacies units are bounded above and below by stratal surfaces identified in Section 3.1.3. These lithofacies are described from top to bottom.

The shoreface 2 depositional unit is the uppermost (youngest) shoreface unit which is composed of upper shoreface 2 (USF 2) facies, proximal lower shoreface 2 (PLSF 2) facies, and distal lower shoreface 2 (DLSF 2) facies. The
Figure 3-8: Grayscale digital photographs of the bounding surfaces. SB 2 is sequence boundary 2. RS 1 is ravinement surface 1. RS 2 is ravinement surface 2.
upper boundary is the seafloor. The lower stratal boundary is the downlap surface (e.g. Figure 3-9).

Grain size ranges for the USF 2, PLSF 2, and DLSF 2 are shown in Appendix 2-1 to 2-3. A summary of all grain size measurements is presented in Figure 3-10a, which is a plot of mean (phi) grain size values versus the standard deviations. Overall, the USF 2 deposits are well sorted, with an average mean of 3.0 φ (fine to very fine). The PLSF 2 and DLSF 2 deposits are more uniform in size, about 3.3 φ (very fine sand). Sediment variability of SF 2 is shown in Figure 3-3c-e, 3-4c-e, and 3-8d-e.

The shoreface 2 unit averages 2.5 m thick, from USF 2 to DLSF 2. The offshore extent of shoreface 2 averages 7 km. This is a minimum extent because cores were not collected seaward of this point.

The marine unit occurs below the SF 2 deposit and was sampled in most profiles (Figure 3-2). The upper stratal boundary is the downlap surface and the lower boundary is RS 2 (ravinement surface 2).

Below the marine unit is the shoreface 1 unit (e.g. Figure 3-9). This unit is composed of upper shoreface 1 (USF 1), proximal lower shoreface 1 (PLSF 1) and distal lower shoreface 1 (DLSF 1) facies. Shoreface 1 (SF 1) is bounded below by the SB 2/RS 1 stratal boundary and above by RS 2. The stratigraphic position of SF 1 is shown in profiles 1, 3-5, 7, 9, 11-15 (Figure 3-2; Appendix 1; Plate 1). Sediment variability of SF 1 is shown in Figure 3-3,a, b; 3-4a,b; 3-5c-e; and 3-8a,c.
Figure 3-9: Cross-section of profile 1, located offshore Matagorda Peninsula showing major surfaces and facies.
Figure 3-10: Comparison plot of mean versus standard deviation for the upper and lower shoreface deposits. Upper shoreface 2 deposits show the greatest correlation (A). Lower shoreface 2 deposits are distinctly finer than USF 2 deposits (A). The distribution of shoreface 1 is similar to that of SF 2 but more scattered (B). The grain size of USF and PLSF deposits (C, D) has significant overlap (gray boxes).
Grain size frequency percent curves for selected samples are shown in Appendix 2-4 and 2-5. Comparison of the sorting and mean of all USF 1 and PLSF 1 grain size samples are shown in Figure 3.10b. USF 1 samples have a mean of 2.8 φ to 3.4 φ and are well sorted (less than 0.5 standard deviation). PLSF 1 samples have a wider grain size range (generally 3.0 to 3.5 φ) and display more variable sorting (Figure 3-10b).

Potentially, the SF 1 unit extends farther offshore than SF 2, due to the predominance of USF 1 and PLSF 1 in the most distal core of Profiles 1-4, 7, 11-12, and 14-15 (Figure 3-2). The distal extent of SF 1 could be up to 10 km in these profiles.

The USF 2 and USF 1 sands show similar mean values of 3.0 φ (Figure 3-10c). Similarly, the PLSF 2 versus PLSF 1 show significant overlap around 3.3 φ (Figure 3-10d). The basal unit is the Pleistocene unit. The upper stratal boundary is the SB 2/RS 1 surface.

3.2.2 Radiocarbon ages

Shell material from the USF 2, the marine, and the Pleistocene facies were radiocarbon dated (Table 3-2). An oyster shell in the USF 2 facies yielded a radiocarbon age of 6,450 ± 80 yBP. Three shells from the marine facies were dated. An oyster shell in profile 8 yielded a radiocarbon age of 7,980± 100 yBP. An *Eontia* mollusk shell in profile 10 yielded a radiocarbon date of 1,320 ± 100 yBP. An *Oliva* shell in profile 11 yielded a radiocarbon age of 2,840± 60 yBP.
<table>
<thead>
<tr>
<th>Radiocarbon sample</th>
<th>Facies location</th>
<th>Age date (Conventional Years BP)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Profile 1, core 2, 90 cm Oyster shell</td>
<td>USF 2</td>
<td>6,450 ± 80 yBP</td>
</tr>
<tr>
<td>Profile 8, core 54, 350 cm Oyster shell</td>
<td>Marine</td>
<td>7,980 ± 100 yBP</td>
</tr>
<tr>
<td>Profile 10, core 70, 135 cm <em>Eontia</em> (mollusk)</td>
<td>Marine/ RS 2</td>
<td>1,320 ± 100 yBP</td>
</tr>
<tr>
<td>Profile 10, core 70, 240 cm Articulated oyster</td>
<td>Pleistocene</td>
<td>&gt; 41,920 yBP</td>
</tr>
<tr>
<td>Profile 11, core 76, 168 cm <em>Oliva</em> shell</td>
<td>Marine/ RS 2</td>
<td>2,840 ± 60 yBP</td>
</tr>
</tbody>
</table>

Table 3-2: Radiocarbon samples from select facies environments. Locations are shown in appendix 1: 1-15. RS 2 is Ravinement surface 2.
One articulated oyster shell from the Pleistocene deposits yielded a radiocarbon dead age of >41,920 yBP.

3.2.3 Tidal Delta deposits

Tidal delta deposits are present in the SF 1 and SF 2 units (i.e. profile 11; Appendix 1-11). The lower SF 1 tidal complex contains medium-grained sand and abundant shell hash, is bioturbated, and contains frosted quartz grains. A unit of regularly alternating fine sands and silts is shown in Figure 3-6b. Abundant wood fragments and bioturbation are present. The SF 1 and SF 2 tidal delta complexes are similar in character except the SF 1 unit extends farther onto the shelf than the SF 2 tidal complex (Appendix A1-11).

Profile 2 is located offshore of the modern Pass Cavallo ebb tidal delta. This tidal complex is similar to that present in Profile 11 except that a distal tidal facies is observed (Figure 3-6d). The distal facies contains regularly spaced alternating laminations of very fine sand and mud.

In addition to sedimentological considerations, identification of the tidal delta complex in Profiles 2-4 is based on the profile morphology (Figure 3-11). Concave downward profiles appear to be related to high sedimentation from the modern ebb tidal delta (Figure 3-11).
Figure 3-11: Cross-section and plan view of the Pass Cavallo ebb tidal delta. The modern-day profiles are overlain and the amount of delta sediment is filled with hatch patterns. The upper profile is profile 2. The upper stippled section is concave downward, indicating a thick section of tidal delta deposits. Core markers are for profile 2 for scale. PETD is proximal ebb tidal delta. DETD is distal ebb tidal delta. MAB is Matagorda Bay.
3.3 Discussion

The discussion focuses on the use of grain size data as a tool for facies identification. The vertical stacking patterns observed suggest multiple episodes of shoreface progradation during the transgression and highstand. The extent and timing of shoreface 1, ravinement surface 2, and shoreface 2 are discussed. The end of the discussion focuses on a comparison of central Texas and east Texas shoreface extent and preservation potential.

3.3.1 Grain size variability across the shoreface

A vertical progression of grain sizes provides insight into depositional environments (Anderson et al., 1982). Anderson et al. (1982) found that the foreshore to upper shoreface sands across Galveston Island have a grain size range of 3.0 to 3.25 φ (fine to very fine). The lower shoreface is chiefly composed of 3.5φ (very fine sand) (Anderson et al., 1982). The fining offshore trend is related to different hydrodynamic effects on sand-sized versus mud-sized sediments, where fine sands are usually transported as bedload, and very fine sands to muds are transported in suspension (Anderson et al., 1982).

Generally, a decrease in grain size is also observed from the upper shoreface to the lower shoreface across central Texas (Figure 3-10). The dominant grain size of the upper shoreface ranges from 2.8 to 3.2 φ. The dominant grain size of the lower shoreface (proximal and distal facies) is finer, ranging from 3.2 to 3.6 φ. Similar grain size trends are observed in shoreface 2
and shoreface 1 (Figure 3-10a, b). Grain size data are thus a key component of shoreface facies identification. Hence, the vertical grain size progression consists of shoreface 2 deposits, marine facies, and basal shoreface 1 deposits.

3.3.2 Timing and extent of Shoreface 1, Ravinement surface 2, and Shoreface 2

3.3.2.1 The extent of shoreface 1 (SF 1)

There are two episodes of shoreface progradation observed in the core transects. The first is SF 1 progradation over an amalgamated sequence boundary and ravinement surface (SB 2 and RS 2, respectively). SF 1 extends farther offshore than the modern SF 2 (Figure 3-2; Plate 1). Core profiles 1 and 14 show the greatest progradation, with the USF facies extending to the most distal cores (Figure 3-2). In addition, profile 1 shows the progradation of USF over PLSF (Figure 3-9). Profiles 2-4, 7, 12, and 15 contain PLSF facies at the most distal core (Figure 3-2). Profile 11 contains a preserved tidal delta complex and profiles 5 and 6 contain DLSF facies. There is no evidence of shoreface 1 deposits in profiles 8 or 10. Profile 9 has a very thin shoreface that does not extend to the distal core.

Two sources of sediment existed. The first was from reworking of older shoreface units offshore. The sands were likely reworked and either transported to the south or deposited in barrier islands. A second sand source was from the reworking of sands from adjacent shorelines. Ravinement of deltaic or shoreline
sands and transport of these sands by littoral drift also contributed to the overall distribution of shoreface 1 deposits.

3.3.2.2 Development of the transgressive ravinement surface (RS 2)

The ravinement surface is marked by a flooding surface, representing a rapid rise in sea level during which erosion occurs. This surface separates the lower SF 1 from overlying marine to distal lower shoreface sediments (Figure 3-2: Plate 1). In general, this ravinement surface slopes upward from the most distal core, 7 km offshore to about 1.0 km offshore, where it flattens and amalgamates with the sequence boundary (SB 2). The flattening of the ravinement surface indicates that shoreface sediments have been eroded as sea-level rose rapidly.

The range in slope of the ravinement surface is 0.6 to 2 m/km, with a mean of 1.2 m/km across profiles 1-15. Thus, RS 2 is much flatter than the modern shoreface profiles, which range from 6-8 m/km. The surface was not sampled in profiles 2 or 6 and is amalgamated with the sequence boundary in profiles 8 and 10 (no shoreface sediments are preserved).

The termination and amalgamation of RS 2 with SB 2 is observed in profiles 5, 7, 9, 11, 13, and 15 (see Appendix 1). The point at which RS 2 amalgamates with SB 2 ranges from -8.5 m to -11.5 m with a mean of -10.3 m. The average depth of ravinement (-10.3 m) where RS 2 flattens and amalgamates with SB 2 possibly represents the maximum water depth during the time of deposition of shoreface 1 (see profiles 5, 7, 9, 11, 13, 15). An
estimate of sea-level position during the deposition of SF 1 is made by noting the present water depth of the same facies in the modern shoreface, above the ravinement surface (profiles 5, 7, 9, 11, 13, and 15). The possible range of water depth during time T1 is large, ranging from -4 to -10 m. based on this approach (Figure 3-12a).

The process by which ravinement occurs is described diagrammatically in Figure 3-12a. The position of the shoreface at time T1 was between -4 to -10 m. A rise in sea level caused erosion of a portion of SF 1. The flattening of RS 2 and subsequent amalgamation with SB 2 is related to wave erosion and removal of shoreface sediments during a rapid sea-level rise. The hatched pattern in Figure 3-12a indicates the area of eroded sediment. The eroded shoreface was likely reworked updip and either added to the modern barriers or was transported southward by longshore currents. The sea-level position at time T2 (present-day) represents maximum flooding of the shelf (Figure 3-12a).

This model is similar to the shoreline retreat model proposed by Everts (1985, 1987) in that the shoreline position shifts landward and upward during a sea-level rise. Everts (1985, 1987) proposed that the shoreface gradient would increase if net deposition occurs. It is difficult to assess whether shoreface gradients have increased from SF 1 time to SF 2 time due to ravinement and removal of the updip portion of the upper shoreface 1 (Appendix 1; 1-15). Generally, SF 1 gradients are similar to SF 2 gradients across profiles where the sequence boundary is deeper than core penetration (e.g. profiles 1,3, and 4; Appendix 1). Where the sequence boundary is penetrated (profiles 5,7,9,11,13,
Figure 3-12: Diagramatic representation of ravinement surface 2 (RS 2) across central Texas (A) and shoreface retreat across east Texas (B). A) RS 2 is a ravinement surface associated with a rapid sea-level rise event. As sea-level rises, wave erosion ravines a portion of shoreface 1 (hatched area). A portion of SF 1 (identified above) is preserved because it is located below the depth of shoreface ravinement. The value -10.3 meters is the average depth of RS 2 flattening and amalgamation with SB 2 and RS 2. The modern shoreface is prograding onto RS 2. B) RS 2 in east Texas is amalgamated with SB 2 and RS 1. Initial SF 2 progradation is followed by shoreface retreat and marine onlap. The marine section onlaps at -7 to -10 meters.
and 15), the shoreface gradient appears to approximate the slope of the SB 1/RS 1 stratal surface (Appendix 1; 1-15).

3.3.2.3 Timing of SF 1 and RS2

The initial drowning of the most distal cores (-15 m) across the central Texas inner shelf dataset occurred about 6,500 years ago based on the Stage 2 to 1 composite sea-level curves (Figure 1-6). The most updip extent of SF 1 is located at about 1.5 km offshore, in -4 to -10 m water depth (farther updip SF 1 has been ravined). Lighty et al. (1982) and Fairbanks (1989) suggest that sea level reached -4 to -10 m, between 5,000-7,000 yBP. Therefore, shoreface 1 progradation potentially commenced after the initial drowning of the shelf, between 6,500 and 5,000 yBP.

The timing of SF 1 deposition can be further constrained by comparison with east Texas shelf, where the timing of shoreface progradation is well constrained (Rodriguez et al., 1999). Rodriguez et al. (1999) conducted a facies analysis of preserved offshore shorelines using high-resolution seismic data and cores. These shorelines were formed during stillstands in sea level and were drowned and overstepped during rapid rises in sea level (Rodriguez et al., 1999). The youngest preserved shoreline observed offshore east Texas, is found on Sabine Bank, situated 15 km offshore in 8 to 12 m of water. Three facies occur within this bank. The lowest facies (facies unit C) is interpreted as a flood tidal delta or back-barrier facies based on landward dipping seismic reflectors as well as lithofacies observed in cores. The boundary above this unit is a flooding
surface representing a rapid rise in sea level. This surface occurs at -10 to -12 m water depth (Rodriguez et al., 1999). Above this surface, are seaward dipping seismic reflectors as well as a shoreface faunal assemblage indicating an ebb tidal delta or lower shoreface environment. The flooding surface above unit B is a ravinement surface, representing a rapid sea-level rise and reworking into the storm unit A at the seafloor. This ravinement surface ranges between -8 and -10 m below sea level (Rodriguez et al., 1999).

The youngest age estimate for Sabine Bank is \(4,490 \pm 50\) yBP, obtained from a preserved oyster shell at the top of unit C/base of unit B. The lower age of the paleo-shoreline is \(7,800 \pm 70\) yBP, from a peat preserved at the base of unit C. An upper age limit for the ultimate drowning and overstepping of the paleo-shoreline (ravinement surface located between units A and B) is not known due to reworking of unit A (Rodriguez et al., 1999).

The depth of the ravinement surface (-8 to -10 m) separating unit B and A (separating the lower shoreface facies from the overlying reworked storm unit A), is similar to the range in depth of the second ravinement surface (RS 2) of central Texas (-8.5 to -11.5 m). The latter separates shoreface from underlying marine deposits.

Likewise, the timing of shoreface 1 progradation along the central Texas shelf is similar to the age of lower shoreface deposits (unit B) of Sabine Bank. An age of the onset of shoreface development for Sabine Bank is about 4,500 yBP (age estimate from the base of unit B; lower shoreface facies), but shoreline
formation could extend as far back as 7,800 yBP (date from peats in unit C). If SF 1 off central Texas is the same paleo-shoreline as that of Sabine Bank, SF 1 progradation in central Texas would range from 4,500 yBP to 7,800 yBP. Additional radiocarbon dates from bay sediments preserved near the base of central Texas incised valleys may corroborate this correlation. Two radiocarbon dates at the RS 2/marine boundary of 1,320 ± 100 yBP and 2,840 ± 60 yBP (Table 3-2) potentially represent the youngest possible age of RS 2.

Anderson et al. (1991) determined that the terminal flooding event of Galveston Bay occurred at around 4,000 yBP. Assuming that the east and central Texas shelves responded at the same time to a rapid sea-level rise, ravinement surface 2 on the central Texas shelf likely formed about 4,000 years ago. Additional age dates are needed to confirm the timing of shoreface progradation and drowning events of central Texas.

3.3.2.4 Timing and extent of Shoreface 2 deposition (SF 2)

A marine to DLSF facies lies above RS 2 in all profiles. This facies was likely deposited after 4,000 yBP (potential age of RS 2) but before deposition of the modern shoreface. Generally, this facies thins updip against USF 1 and USF 2 deposits. Two radiocarbon dates were obtained from the marine deposits (Table 3-2), yielding ages of 6,450 ± 80 yBP and 7,980 ± 100 yBP. These older dates imply that this is a reworked shell lag.
Currently, the modern shoreface (SF 2) is prograding, downlapping onto the marine facies. In profiles 3, 7, and 15 sedimentation rates have been high enough to prograde DLSF facies seaward of the seaward-most core. Proximal lower shoreface has prograded over the DLSF facies. No marine facies were observed in these profiles.

SF 2 did not prograded to the extent of shoreface 1 (SF 1) (Figure 3-2; Plate 1). Profile 1 represents the least amount of progradation, with marine clays exposed at the seafloor in core 7. All other profiles contain the DLSF facies in the most distal core, except for profiles 6, 11, 12, 14 and 15 that contain PLSF facies.

The modern shoreface (SF 2) deposits are derived from three sources. The main source of sand in the shoreface is from sediment reworked by transgressive ravinement (RS 2). These sands potentially were eroded from the upper shoreface and subsequently incorporated into the barrier complex. The upper shoreface prograded seaward as the rise in sea level slowed approximately 3,500 to 4,000 yBP. Longshore current transport of sand from the east Texas shelf is likely a second source. It is difficult to ascertain the relative contributions of these two sources.

The third source of sand for SF 2 is from storm deposition. Deposition in the lower shoreface is thought to occur during storms (Snedden et al., 1988; Snedden and Nummedal, 1991). Geostrophic current flow offshore and to the southwest is likely the main process responsible for deposition of lower shoreface sand beds, such as the Carla Bed which was mapped by Snedden
and Nummedal, 1991 (Figure 3-13). Hurricane Carla made landfall near Pass Cavallo, between Matagorda Peninsula and Matagorda Island, in 1961.

A lower shoreface storm bed, from top to bottom, consists of bioturbated muds grading down into planar laminae (high flow regime). Below is a fining upwards sand to silty sand bed. The entire sequence rests on an erosional surface. The overall fining upward sequence is due to decreasing storm energy (Snedden and Nummedal, 1991). Identification of isolated storm beds in water depths less than 20 m is difficult, if not impossible (Snedden and Nummedal, 1991). This is due to amalgamation of storm deposits in the near-shore environment as well as intense bioturbation of the sandy beds (Snedden and Nummedal, 1991).

General comparisons have been made between storm beds observed in cores collected for this study and the Carla Bed (Figure 3-13). Storm beds exist in core profiles 3 and 4 (Figure 3-13; 3-2). These profiles are located offshore of the landfall of Hurricane Carla. Core 21 of profile 3 (5 km offshore, -12 m) contains a similar fining upward sequence to that identified by Snedden and Nummedal (1991). The basal contact is erosional, with a shell lag containing whole mollusks. The unit fines upward from fine upper sand to very fine sand with shell hash. The unit is capped by silty clay. A similar sequence is observed in core 27, profile 4 (5.5 km offshore, -11 m). An erosional sharp contact with an overlying coarse shell lag also defines the base of this bed. The unit fines upward from fine sands to dusky brown silty clay.
Figure 3-13: Hurricane Carla Storm Bed distribution and thickness across the study area. Central Texas profiles are numbered and are located updip of the mapped Carla Bed. The Carla bed is observed in profiles 3 and 4. Other profiles contain stacked storm bed deposits.
Direct evidence for Carla Bed deposition related to the ‘thins and thicks’ identified in Figure 3-13, are not readily apparent in profiles 5-15. Updip of the thick units identified by Snedden and Nummedal (1991), in profiles 5-15, are fine to very fine sands with no evidence of storm lag deposits, fining upward sequences, or a sharp basal contact. Profiles 5-15 are located south of the landfall of Hurricane Carla. In contrast, profiles 3 and 4 contain storm beds, possibly from Hurricane Carla, and are located proximal to the hurricane landfall. This suggests that the manifestation of individual storm bed deposits, such as the Carla Bed, may only be identified proximal to the landfall location of the hurricane, in water depths less than 20m.

3.3.2.5 Sequence stratigraphic implications

Potentially, lowstand, transgressive, and highstand systems tract deposits exist in the central Texas core dataset. Lowstand systems tract deposits would be located only at the base of incised valley fills. No lowstand fluvial sediments were cored in this study, but the Lavaca River incised valley (profiles 1 and 2) as well as the Nueces River incised valley (profile 12) may have preserved fluvial sediments at their base above sequence boundary 2 (Plate 1).

Above the amalgamated sequence boundary 2 (SB 2) and transgressive ravinement surface (RS 1) is the first prograding shoreface unit (SF 1). This unit is bounded at the top by RS 2 (Figure 3-9). Progradation of SF 1 potentially ranged in age from 7,800 yBP to 4,500 yBP. If so, SF 1 is a parasequence-scale
deposit (Van Wagoner, 1995) contained within the transgressive systems tract Stage 2 to 1 sea-level rise (Figure 1-5; 1-6).

A downlap surface at the base and the seafloor at the top bound Shoreface 2. The SF 2 deposit is a parasequence contained within the highstand systems tract (Van Wagoner, 1995). It is difficult to assess whether the current highstand shoreline will be eroded, due to continuing sea-level rise, or if, in fact, the highstand shoreline will continue to aggrade and prograde seaward. In the latter case, the downlap surface identified in this study is the lower boundary of the highstand systems tract. The preservation potential of both transgressive and highstand deposits is discussed in the next section.

3.3.3 Preservation potential of coastal deposits

3.3.3.1 Preservation Potential of Central Texas SF 1 (Shoreface 1 sediments)

The preservation potential of coastal lithosomes during transgression is related to the amount of accommodation space, sediment supply, and the rate of sea-level rise. Of all factors, accommodation space is the most critical for lithosome preservation.

High accommodation is related to lows formed by fluvial incision during the Stage 2 lowstand. Subsidence rates are presumably much higher within incised valleys due to sediment compaction of the valley fill deposits. As sea level rose and transgressed the inner shelf, SF 1 deposition commenced. Shoreface 1 sediments presumably were sourced from alongshore and offshore and were of a uniform thickness and extent offshore during this time. A rapid
sea-level rise (RS 2) occurred approximately 4,000 yBP, drowning and eroding SF 1 deposits. Thick SF 1 deposits were preserved below the depth of shoreface ravinement within incised valleys. Sediments on inter-fluvial highs were subject to transgressive ravinement and had a lower preservation potential.

Profiles 1-4, 6, 8, and 12 (Figure 3-2; Plate 1) are located within incised valleys. Profiles 1-4 did not penetrate sediments below sequence boundary 2. Incision during the Stage 2 lowstand was up to ~36 m within Matagorda Bay (Wright, 1980). Therefore, it is likely that the incision depth offshore 7 km is at least greater than 20 m, probably up to 30 m (Plate 1). The deepest portion of the most distal core offshore penetrated a meter or so of USF 1 and PLSF 1 deposits. These deposits are the thickest within the Lavaca River valley, about 2 m thick, and thin southward after profile 4 (Plate 1).

The Guadalupe River Stage 2 valley, located within San Antonio Bay, is incised up to 24 m (Wright, 1980). Incision 7 km offshore in profile 6 can be constrained to be slightly greater than 20 m. Sequence boundary 2 occurs at 17.5 m in profile 5, 18-20 m in profile 7, and 24 m within the bay. Shoreface sediments within this valley are not as thick (1 m) as in the Lavaca River valley (2 meter), and are comprised of a more distal facies of PLSF 1 overlying DLSF 1 facies.

The Aransas Bay and Copano Bay Stage 2 fluvial valleys are possibly from the Aransas River (largest river now emptying into the bay). The fluvial drainage area was possibly small and discharge rates may have been low, resulting in about 18 m of incision within the bay (Wright, 1980). Offshore
incision does not exceed 17 m (Profile 8; Plate 1). No SF 1 deposits are preserved at the distal core in profile 8 (Plate 1).

Similarly, no SF 1 deposits are preserved in the distal offshore cores of the inter-fluvial profiles 9 and 10 (Figure 3-2; Plate 1). The Stage 2 surface is at 15 and 14.5 m depth, in cores 63 and 70, respectively. Updip in cores 61 and 62 of profile 9, minimal PLSF 1 is preserved below the ravinement surface (RS 2).

Farther to the south in profiles 11 and 12, the Stage 2 sequence boundary is deeper. An increase in the depth is related to incision from the Nueces River, located in Corpus Christi Bay (Wright, 1980). Incision within the bay was 36 m (Wright, 1980). Offshore in profile 12, valley depth could be from 17-30 m depth (no Stage 2 sediments were penetrated). A thick (3 meter) PLSF 1 deposit was preserved in core 84 of profile 12 (Appendix 1). Similarly, updip core 83 contains 2 m of USF 1.

Profiles 14 and 15 also contain thick sequences of USF 1 and PLSF 1, respectively. No fluvial valley was found updip of these profiles (Figure 3-2). The thick section of SF 1 deposits above the sequence boundary (SB 2 is possibly 16-20 m deep) is potentially related to recycling of thick Pleistocene sands into SF 1 deposits.

It is clear, from the above discussion, that the thickest accumulation of SF 1 deposits is located within the areas of the greatest accommodation space, or within the incised fluvial valleys. The deepest fluvial incision depth (Lavaca River and Nueces River valleys) contains the thickest accumulations of SF 1 deposits. The shallower incision depths of the Guadalupe and Aransas Rivers created
minimal accommodation space offshore, resulting in moderate to no preservation of SF 1 deposits (none preserved in the Aransas River valley offshore). Where no fluvial valley exists above the sequence boundary (profiles 9 and 10), little to no SF 1 deposits are preserved. This suggests that shallow fluvial incision (Aransas River, -18 m) does not create enough accommodation space to preserve coastal lithosomes during a rapid sea-level rise.

It is possible, from the above discussion, to create a preservation potential depth range for SF 1 deposits. Aransas River valley (profile 8, Plate 1) has a SB 2 depth of -17 m, with no preservation of SF 1 deposits. Profiles 9 and 10 contain no shoreface sediments at the distal core, and are located in an interfluvial high with a SB 2 depth of -15 and -14.5 m, respectively. Therefore, the shallowest depth of antecedent topography possible for preservation of shoreface 1 sediments during the transgression is -14.5 m (interfluvues) to -17 m (Aransas River valley) (Figure 3-2; Plate 1).

3.3.3.2 Preservation potential of shoreface 2 versus the east Texas modern shoreface

Central Texas modern shoreface (SF 2) deposits are located above the ravinement surface 2 (RS 2) and range in thickness between 1-10 (?) m and extend from 6 to 10 (?) km offshore. In contrast, east Texas modern shoreface deposits range in thickness between >1 m to 9 m and extend only 5 km offshore (Siringan, 1993) (profile C, Figure 3-14, this study).
Figure 3-14: Profile C is offshore Galveston Island, located on the east Texas shoreface. The basal SB 2 is amalgamated with RS 1 and RS 2. Shoreface 2 initially prograded onto SB 2. Currently, the modern shoreface 2 is back-stepping with marine sediments onlapping SF 2.
Profile C (Figure 3-14), located on the west end of Galveston Island, is a representative example of the facies observed in the east Texas modern shoreface. The original study by Siringan (1993) only differentiated between upper shoreface and lower shoreface. For the purposes of a direct comparison with this study, facies have been grouped into USF, PLSF, and DLSF based on bed thickness and grain size.

There are four main differences observed in central Texas profiles versus east Texas profiles. The first is the amount of sediment present above the sequence boundary. Across the central Texas coast, there were two episodes of progradation, SF1 and SF 2 (Figure 3-2; Plate 1), whereas across the east Texas shoreface there was only one progradational event (SF 2) (Figure 3-14). Across east Texas, Ravinement surface 1 and 2 (RS 1 and 2) are probably amalgamated with the sequence boundary (Figure 3-14). Shoreface 1 progradation on the east Texas coast probably occurred farther offshore and may be related to the Sabine Bank paleo-shoreline.

The second difference observed across the Texas coast concerns the amount of sand present in the coastal system. Comparing the amount of sediment above RS 2 in both study areas suggests that the sediment supply to the central Texas shelf is higher than that of the east Texas shelf. The central Texas shoreface has prograded up to 4 km seaward of the comparable shoreline in east Texas (Figures 3-2; 3-14).

An early episode of shoreface progradation across east Texas (Siringan, 1993) is shown in profile C (Figure 3-14), with PLSF 2 and DLSF 2 prograding
out over SB 2. A similar progradational event is seen in profiles 2, 3, 4, and 6 of central Texas (Appendix 1). The interval above the prograding shoreline of east Texas is composed of marine sediments, demonstrating that the current shoreface is backstepping (Figure 3-12b; 3-14) (Siringan, 1993). The gradient of the shoreface was steeper during the initial progradation. Subsequently, the shoreface slope has decreased since the deposition of marine clays. Everts (1985, 1987) proposed a similar scenario for recessional coasts (Figure 1-9a). He suggested that during a slow sea-level rise, the shoreface gradient would decrease if a predominance of muddy sediments were removed from the shoreface and transported offshore or along-shore.

Two episodes of progradation are recorded across several central Texas profiles (profiles 2, 3, 4, and 6). Early shoreface progradation is observed in these profiles, potentially due to high accommodation near incised fluvial valleys. In addition, these profiles are located offshore Matagorda Island which is currently prograding seaward (Paine and Morton, 1989). The earlier prograding shoreface had a similar gradient to that of the modern shoreface (Appendix 1; 1-4, 1-6). Perhaps a relatively constant sediment supply combined with a slow sea-level rise allowed the coastline to maintain its constant gradient. This implies that the central Texas coastline may be stable despite a slow sea-level rise, where the amount of sand transported into the system is equal to or slightly greater than the amount of sand removed. This scenario is similar to that proposed by Swift (1976), who suggested that a stillstand in sea level would allow the shoreline to attain the maximum gradient, approximating an equilibrium shoreline.
The east Texas shoreface and central Texas shoreface are out-of-phase with one another. The east Texas shoreface is being transgressed, whereas the central Texas shoreface is stable to prograding. This difference is potentially due to the availability of sand to these coasts. Fine sands are relatively stable in a high energy upper shoreface environment, whereas very fine sand to muddy sediments are unstable and are carried in suspension offshore or along-shore. The sand-sized sediment supply for the east Texas shoreface is low; the shelf is mud-dominated (Siringan, 1993), with westward-directed longshore transport contributing less sand than is removed from the system. Therefore, a predominance of muddy sediments across the east Texas shoreface equates to high rates of sediment removal and high rates of shoreface retreat.

In contrast, central Texas shoreface sediment supply is high. Currently, the central Texas outer shelf is mud-dominated, with an extensive mud blanket draping offshore transgressive deposits (?) and the sequence boundary (Shideler, 1981). The inner shelf, extending to about 10 km offshore, is sand-dominated. During transgression, reworking of SF 1 likely contributed a large portion of the sand contained within the modern SF 2. Currently, longshore sediment supply is from the south and east Texas shelves. The constant sediment supply to the central Texas shoreline, coupled with minimal loss of shoreface sands, results in progradation of the shoreface.

The third factor contributing to the different amount of shoreface deposits above the Stage 2 sequence boundary offshore east Texas versus central Texas relates to the amount of accommodation space available. Modern shoreface
gradients systematically increase from east to south (Figure 3-15). High Island, on the east Texas Bolivar Peninsula, has a gradient of 4 m/km, 1 km offshore. The gradient increases toward Galveston Island to 5 m/km in the upper shoreface, 1 km offshore. Central Texas gradients are consistently about 6 m/km in the upper shoreface, 1 km offshore. In general the shoreface toe is located at a shallower depth offshore east Texas, deepening gradually across central Texas. East Texas shoreface extent is from 2 km offshore Bolivar Peninsula to 4 km offshore Galveston Island. In contrast, the shoreface extent for central Texas ranges from 5.8 km offshore in profile 1 (Matagorda Peninsula) to 7 km offshore of Matagorda Island and San Jose Islands. Profiles 11-15. offshore Mustang Island and Padre Island, are significantly steeper with the shoreface extending to only 4.5 km offshore (Appendix 1).

The break in shoreface profile, observed across east and central Texas, is related to the depth of active fair-weather wave erosion (Siringan, 1993). The break in shoreface profile across east Texas is coincident with the depth of the modern shoreface (-7 m to -10 m) (Figure 3-14). In contrast, the break in profile across central Texas ranges from -8 m across Matagorda and San Jose Islands, -10 m across Mustang Island, and -12 m across the North Padre Island shoreface (Appendix 1: 1-15). The break in shoreface profile across central Texas is not coincident with the toe of the lower shoreface. In fact, the upper shoreface, proximal lower shoreface, and distal lower shoreface extend farther offshore, below the break in shoreface profile. Therefore, a significant portion of the modern central Texas shoreface is not exposed to fair-weather wave erosion.
Figure 3-15: Profile gradient comparison across the Texas coast. Profile gradient increases gradually from east to central Texas, from 4 m/km to 6 m/km, respectively.
The cause of the steeper gradients observed across central Texas versus east Texas is two-fold. First, the amount of accommodation space above the sequence boundary is greater in central Texas than in east Texas. Profile D, of east Texas, is located near Bolivar Roads tidal inlet, proximal to the paleo-Trinity incised valley. The depth of sequence boundary 2 is the deepest observed across the east Texas shoreface (-13.5 m, 6 km offshore). The depth of the shallowest Stage 2 SB in central Texas is -14.5 m, 6 km offshore in profile 10 (Appendix 1; 10). No older shoreface 1 (SF 1) deposits were preserved in the area of profile 10 (Appendix 1; 10).

The second factor that influences the profile steepness is sediment supply. The east Texas shoreface is mud-dominated whereas the central Texas shoreface is sand-dominated, with a high influx of sand via longshore drift. A sand-prone shoreface will have a greater angle of repose compared to a mud-prone shoreface (Swift, 1976).

Both the amount of accommodation space and sediment supply across the east and central Texas shoreface suggest that the preservation potential of the central Texas shoreface is higher than that of the east Texas shoreface, given a fast rate of sea-level rise. East Texas shoreface gradients are flatter than central Texas shoreface gradients, with less accommodation space above SB 2. Profile 10 (Appendix 1; 10) of central Texas shows no SF 1 sediments preserved above SB 2. This means that shoreface erosion by ravinement surface 2 (RS 2) was effective in removing all shoreface 1 sediments above a depth of -14.5 m. The accommodation space above the Stage 2 sequence
boundary across east Texas is less than this, up to 13.5 m. If the rate of sea-level rise increased today, ravinement would probably decapitate the modern east Texas shoreface (SF 2) (Figure 3-14), similar to ravinement and removal of SF 1 sediments in profile 10.

3.4 Summary

There are two episodes of shoreface progradation observed across the central Texas shoreface, SF 1 and SF 2. The first episode of progradation (SF 1) possibly occurred 7,800 yBP and ended from a rapid rise in sea level at about 4,000 yBP. The SF 1 shoreline may be time correlative to a paleo-shoreline at Sabine Bank on the east Texas shelf.

The modern shoreface (SF 2) likely prograded across the shelf over the last 3,500 years, downlapping onto marine sediments. This surface represents the downlap surface of the modern highstand systems tract.

The preservation potential of transgressive and highstand sediments is high across the central Texas shelf and low across the east Texas shelf. This is due to a low accommodation space and low gradients across east Texas and a high accommodation space and steep gradients across central Texas. The amount of accommodation space is related to the depth of antecedent topography. Across central Texas, the minimum depth of shoreface preservation given a rapid sea-level rise is -14.5 m. All east Texas shoreface profiles have antecedent topography shallower than this depth. Therefore, the potential for
preservation of shoreface sediments across east Texas is low given a rapid sea-
level rise.
Chapter 4: Preservation potential of Late Pleistocene fluvial and coastal deposits

4.1 Results

Eckles (1996) acquired the regional offshore seismic grid used in this study (Figure 4-1). The results section begins with a description of lithologic characteristics of facies identified in this study as well by Eckles (1996). Changes to the original interpretations by Eckles (1996) are noted. Seismic stratal boundaries are identified in section 4.1.2. Finally, combining lithologic descriptions and seismic bounding surfaces, 5 seismic facies units will be presented.

4.1.1 Lithologic units

Five lithologic units are observed in this study and in the study of Eckles (1996). These units were identified based only on platform boring descriptions. No sediment materials were examined. Offshore borings are labeled and cross-section lines are shown on Figure 4-1, with corresponding cross-section figure locations. Platform boring lithologies were converted from depth to time based on seismic velocities ranging from 1500 m/s, 1550 m/s, and 1600 m/s based on the work of Abdullah (1995) on the east and central Texas shelves.

4.1.1.1 Mud unit 1:

This unit is the youngest (shallowest) observed (Eckles, 1996). It consists of very soft to firm olive gray clay with scattered shell fragments. Silt and sand pockets were identified in the boring descriptions. The term 'pocket' may
Figure 4-1: Study area map showing offshore boring locations, seismic line locations as well as seismic figures shown. Figure locations are shown in (). Platform boring cross-sections are shown in Figures 4-2, 4-3, and 4-6.
refer to bioturbation, with sand and silt infilled burrows. The thickness range of
this unit is significant across the study area. Mud unit 1 is less than 15 m thick in
the central portion of the study area and the unit thickness increases to 30 m in
the southern portion of the study area (e.g. Figure 4-2; 4-3). The greatest
thickness (45 m) is observed in seismic line 17.

4.1.1.2 Sand / mud unit 2:

This unit is present only within fluvial valleys (not observed in the study of
Eckles, 1996). The typical valley fill succession is floored by a basal gray muddy
sand to fine sand. Organic-rich bioturbated clays overlie the basal sand. Deep
valleys, such as the Lavaca/Guadalupe valley contain about 20 m of light gray
fine sand overlain by stiff bioturbated gray clay, and are capped with gray muddy
fine sand (Figure 4-2). In shallow valleys, such as that identified in seismic line 3
and in boring B-566 (Figure 4-3), the valley succession is capped with fine
sands.

4.1.1.3 Sand unit 3:

Sand unit 3 is only present in water depths less than -50 m. The
distribution of sand unit 3 is shown in Figure 4-4 and 4-5. Sand unit 3 is divided
into the upper sand facies and the lower sand facies based on stratigraphic
position. The upper sand facies (Figure 4-4) is composed of gray to olive gray
silty fine sand with shell hash. The lower sand facies is composed of fine sand in
the central portion of the study area to silty fine sand in the southern portion.
Section 1

Figure 4-2: Platform boring cross-section 1. The offshore Lavaca River and Guadalupe river valleys are shown here in borings B-623 to B-620.4. Lithofacies are shown above. Boring depths are reported in feet below the seafloor.
Figure 4-3: Platform boring cross-section 2 across the highstand and transgressive deltas. The asterick (*) is the downdip limit of silty sand during stage 3 (Sand unit 3). Inferred downlap is shown on the downdip extent of Stage 3 deposits. Onlap is shown onto the Stage 5c bounding surface. Depth is in feet below the seafloor.
Figure 4-4: Map showing the offshore extent of silty sand unit 3 associated with Stage 3 deposition. Cross-section 1-6 locations are also shown.
Figure 4-5: Map showing the distribution of the lower sand facies of sand unit 3. Boring cross-section locations are shown.
Sand is not identified across the eastern portion of the study area, only olive gray muds are present. The offshore limit of the lower sand is at -52 m water depth in the south and -35 m water depth in the central portion (Figure 4-5). The offshore gradation from sand to mud (mud unit 4) is observed across the entire study area. This is shown in Figure 4-6.

4.1.1.4 Mud unit 4:

Mud unit 4 is present downdip of Sand unit 3 (Eckles, 1996). This unit is very similar to mud unit 1. Instead of soft olive gray clays, observed in mud unit 1, stiff olive gray clays with calcareous nodules are present in mud unit 4. Bioturbation is prevalent in this unit (Figure 4-6).

4.1.1.5 Sand Unit 5:

This unit is composed of gray silty fine sand to fine sand with scattered shell fragments (Figure 4-3). Sand unit 5 is divided into upper sand unit 5 and lower sand unit 5 based on stratigraphic position (Figure 4-3). Core control is sparse in this unit. Boring 51A shows a thickness of 110 feet (33 m) for this unit (Figure 4-3). The thickness, combined with the high sand content, as well as the position near the shelf edge, suggests a deltaic origin for sand unit 5.

4.1.2 Seismic bounding surfaces

Several bounding surfaces were recognized in the seismic dataset. These surfaces bound systems tract deposits as well as the lithologic facies
Figure 4.6a: Platform boring cross-section 3 across Stage 2 Guadalupe River channels. Depths are in feet below seafloor.
Figure 4-6b: Offshore platform boring cross-section 4. Downlap of Stage 3 sediments onto the Stage 3 surface is shown (see Line 9, Figure 4-13 for seismic facies). See Figure 4-4 and 4-5 for section location.
Figure 4-6c; Offshore boring cross-section 5. Profile location is shown in Figure 4-4 and 4-5.
Figure 4-6d: Offshore boring cross-section 6. See Figure 4-4 and 4-5 for section location.
described above. The surfaces are presented from top to bottom (youngest to oldest). The stratal surfaces identified in this study were tied to those identified by Abdullah (1995) and Snow (1998) (Figure 4-7).

4.1.2.1 Transgressive surface (Ts):

The transgressive surface locally overlies upper sand unit 5 in the eastern portion of the study area (Figure 4-8). This surface is time correlative to the upper sand unit -ts of Snow (1998). This surface shows a fair degree of channelization, only observed in water depths of -30 to -45 m (Figure 4-9a). Locally, this surface is flat and inclined seaward from -45 m water depth to about -120 m water depth (Figure 4-8; 4-9a, b).

4.1.2.2 Stage 2 Sequence boundary (SB 2):

The Stage 2 sequence boundary is present throughout the study area (Figure 4-8, 4-9, 4-9). This surface separates the underlying sand unit 3, mud unit 4, and lower sand unit 5 from the overlying mud unit 1, sand/mud unit 2, and upper sand unit 5. Up to 37 m of fluvial relief occurs on this surface. The time structure map of this surface is shown in Figure 4-10.

Sequence boundary 2 is correlated to the Colorado Stage 2 incised valley as mapped by Snow (1998) and Abdullah (1995). The Colorado valley is up to 35 m deep on the inner shelf (Snow, 1998). Several smaller valleys (20 to 26 m deep) cross line 1 from -30 m to -120 m water depth (Figure 4-9b).
Figure 4-7: Compressed strike line R93-51 modified from Abdullah (1995). Stratal surfaces have been tied from Louisiana in central Texas based on this regional strike line. The stratal surfaces identified in this study were tied to surfaces identified by Abdullah (1995). Good correlation exists from Stage 3 to Stage 1. The early Stage 3 delta is shown on Figure 4-22 and was mapped in detail by Snow (1998). The late Stage 3 delta is described in this study and is shown on Figure 4-23. The oxygen isotope curve generated from platform boring B-146 is shown in Figure 1-7 and Figure 4-24 and correlates well with surfaces identified in this study.
Figure 4-9a: Uninterpreted and interpreted Line 1. Bounding surfaces are shown in gray. Examples of prograded and chaotic fill within Stage 3 and Stage 2 fluvial valleys, respectively. Downlap is seen on maximum flooding surface 3. Line location is shown on Figure 4-1.
Figure 4-9b: Dip line 1b illustrates bounding surfaces and seismic facies at the shelf margin. Location is shown on Figure 4-1.
Figure 4-10: The Stage 2 time structure map was constructed using Stage 2 fluvial incision depths from Wright (1980), who mapped this surface in bays of central Texas, through the near-shore core dataset, and the shelf seismic dataset.
West of line 1, incision related to the Stage 2 surface can be traced offshore to -38 to -52 m water depth, which occurs between 38 to 50 km offshore (Figure 4-10). Seaward of this point, the SB 2 surface has up to 10 ms (7.5 m) of relief (Figure 4-10).

4.1.2.3 Stage 3 downlap surface:

This bounding surface is a downlap surface (Figure 4-8; 4-9). The surface is flat with faint downlap of sand unit 3 observed in dip-oriented seismic profiles. The time structure map of the Stage 3 surface is shown in Figure 4-11.

4.1.2.4 Sub-Stage 5c:

The 5c stratal boundary is located offshore of the Stage 3 boundary. The updip limit of this surface is about 40 km offshore (-50 m water depth) (Figure 4-12). Surface 5c slopes gradually basinward. The downdip limit is at the shelf break (-120 m). Overlying reflectors onlap the stratal surface updip (Figure 4-13) and are conformable offshore (Figure 4-14). Mud unit 4 is present above and below this boundary.

4.1.2.5 Sub-Stage 5d/5e:

The basal surface is the Stage 5e/5d downlap surface. The overlying sediments are lower sand unit 3, located updip in water depths less than -52 m (Figure 4-5), and mud unit 4, located in water depths from -30 m to the shelf
Figure 4-11: Stage 3 time structure map. Gray areas represent erosion of Stage 3 sediments by Stage 2 fluvial incision.
Figure 4.12: Stage 5c time-structure map. Surface 5c amalgamates with the Stage 3 surface about 40 km offshore (-50 m water depth).
Figure 4-13: Dip line 9 shows valley fill and SFU reflector configurations. Arrows within seismic facies indicate onlap and downlap. Line location is shown on Figure 4-1. Boring cross-section 4, Figure 4-6b, overlaps line 9.
Figure 4-14: Dip line 11 showing seismic facies and bounding surfaces. Seismic facies units A and B contain subparallel clinoform reflectors within an overall wedge-shaped geometry. Erosion is present on the stage 2 sequence boundary. Downlap and toplap are shown by arrows. Line 11 location is shown on Figure 4-1.
break (Figure 4-6a-d). Minor channelization (less than 7 m) exists in less than 30 m water depth. The time structure map is shown in Figure 4-15.

4.1.3 Seismic facies units

Eckles (1996) identified 4 seismic facies units (SFU). Two additional seismic facies units as well as 3 valley fill facies are identified in this study. The following is a brief overview of SFU A-D of Eckles (1996) and new results from this study.

4.1.3.1 SFU A:

SFU A is composed of subparallel reflectors of low to moderate amplitude. This unit is bounded above by the seafloor and onlaps a high amplitude reflector at the base (Eckles, 1996). The basal bounding surface is the Stage 2 sequence boundary. This facies is shown on Figures 4-8, 4-13, 4-14.

4.1.3.2 SFU B:

In dip view SFU B is composed of discontinuous low to high amplitude reflectors. It has a chaotic seismic reflection pattern with faint high angle shingled clinoforms (Eckles, 1996; Figure 4-13, 4-14). This seismic facies downlaps against a high amplitude reflector at the base that is the 5e/5d downlap surface and is bounded above by the Stage 3 and Stage 5c reflectors (Figure 4-13). SFU B is also present above the Stage 3 surface (Figure 4-13). Downlap surface 3 is the basal boundary and SB 2 is the upper boundary.
Figure 4-15: Stage 5e/5d time structure map. Minor incision is observed up to 50 km offshore. The gray area represents localized erosion by Stage 2 fluvial incision.
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Figure 4-16: Time thickness map of SFU B and SFU E from the Stage 2 SB to the Stage 3 maximum flooding surface. The overlapping gray and black contours in the east are two phases of deltaic progradation, Phase 1 (Snow, 1998) and Phase 2 (this study). Gray areas represent erosion of SFU B by Stage 2 fluvial incision.
Figure 4-17: Time thickness map of SFU B and SFU C from the amalgamated Stage 5e/5d surface to the Stage 5c surface. The gray area represents erosion and removal of Stage 5d deposits by Stage 2 fluvial incision. SFU B thins offshore, grading into SFU C. The dashed line shows the approximate SFU B/C boundary at -50 m water depth.
suggesting a marine origin (mud unit 4). The time thickness map of SFU C from Stage 5d to 5c is shown on Figure 4-17.

SFU C is the only facies present from Stage 5c to Stage 3 and is shown in Figure 4-18. Thickness is greatest in the south, up to 50 ms (37 m), thinning toward the east. The water-bottom multiple obscures facies identification in the eastern part of the study area.

4.1.3.4 SFU D:

SFU D is a low amplitude chaotic to transparent facies with a pinnacle shape (Figure 4-14). This is a reef facies identified within the Stage 2-1 transgressive systems tract and older strata (Eckles, 1996).

4.1.3.5 SFU E:

Eckles (1996) did not recognize SFU E. In dip view this facies consists of tangential oblique prograding clinoforms within an overall wedge-shaped geometry (Mitchum et al., 1977) (Figure 4-9b). In strike view, SFU E contains a combination of complex sigmoid oblique clinoforms within a mounded external geometry (Figure 4-8) and subparallel clinoforms with a lens-shaped geometry (Mitchum et al., 1977). Reflectors are moderate to high amplitude. SFU E progrades onto a downlap surface within Stage 3 and is truncated above by the Stage 2 sequence boundary (Figure 4-9b). A time thickness map of this facies is shown in Figure 4-16. The thickest portion is 60 ms (46.5 m). SFU E overlies and is downdip of the Phase 1, Stage 3 delta mapped by Snow (1998) (Figure 4-
Figure 4-18: Time thickness map of SFU C from Stage 5c to the Stage 3 maximum flooding surface. The thickest deposits are located in the South.
Strike line 20 shows two distinct lobes, the western oblique prograding lobe and the eastern subparallel lobe (Figure 4-8). This seismic facies is thought to form during times of high sediment supply (Mitchum et al., 1977). Lower sand unit 5 is present within SFU E (Figure 4-3).

4.1.3.6 SFU F:

This seismic facies is the clinoform facies of the USU (upper sand unit) that was mapped by Snow (1998). Upper sand unit 5 found within SFU F is probably correlative to the upper sand unit of Snow (1998). SFU F contains complex sigmoid oblique prograding clinoforms that are moderate amplitude (Figure 4-8). A time-thickness map for this seismic facies is shown in Figure 4-19. This facies is present stratigraphically over and somewhat updip of SFU E (Figure 4-3, 4-19). SFU F downlaps the Stage 2 sequence boundary and is locally truncated by the Ts surface (Figure 4-8). Sigmoid prograding reflectors are thought to form during times of low sediment supply and / or rapid sea-level rise (Mitchum et al., 1977). Therefore, it is likely that SFU F was deposited during an overall sea-level rise (Snow, 1998).

4.1.3.7 Valley-fill seismic facies:

There are three types of valley fill seismic facies found in the study area: Chaotic, complex, and prograded. The chaotic valley fill seismic facies characterizes many of the Stage 2 valleys (Figure 4-9a; 4-13). This fill type is thought to be associated with repeated cut and fill of the valley (Mitchum et al.,
Figure 4-19: Time thickness map of SFU F. SFU F is in black and is located updip of SFU E (in gray). SFU F is interpreted as a transgressive delta. Arrows indicate Ancestral Colorado distributary channels (Snow, 1998).
1977). The complex valley fill contains hummocky clinoform reflectors juxtaposed against subparallel clinoform reflectors (Figure 4-20). This lateral gradation from hummocks to subparallel reflectors is interpreted as sediments building outwards into shallow water or an interdeltaic setting (Mitchum et al., 1977). The prograded fill seismic reflection pattern is only observed in Line 1 (Figure 4-9a) and is thought to be associated with the outbuilding of the Stage 3 highstand delta mapped by Snow (1998).

4.2 Discussion: Depositional history

The first part of this discussion focuses on paleoenvironmental reconstructions of fluvial incision patterns, as well as highstand and transgressive deposition. Following this discussion is an analysis of the relative role of eustacy on the extent and distribution of the mapped deposits.

4.2.1 Stage 5d deposition:

The Stage 5e/5d to 5c depositional package contains shingled clinoforms of SFU B and downlaps onto the basal surface landward of -52 m depth. Sand unit 3 is present in this interval. Offshore, seismic reflections are conformable with the 5e/5d surface. The upper bounding surface is the Stage 3 maximum flooding surface. Farther offshore, the upper bounding surface is Stage 5c. Below -52 m depth, mud unit 4 is present. This unit is generally thin (10 ms, 7.8 m) on the inner shelf, and up to 30 ms (23 m) thick on the outer shelf. SFU B of Stage 5d is interpreted as an early highstand prograding shoreline. Offshore,
Figure 4-20: Seismic line 42 located about 20 km offshore (Figure 4-1). Hummocky fill grades laterally into subparallel fill. Subparallel clinoforms are shown for SFU C (5e/d to SB 2). Stage 3 deposits are absent here due to Stage 2 incision by the Lavaca/Guadalupe rivers.
mud unit 4 is interpreted as a coeval deposit (Eckles, 1996). The Stage 5d deposit thickens toward the east, possibly due to deposition of the Stage 5 Colorado delta mapped by Abdulah (1995). Sand content increases toward the south (Figure 4-5). The sand limit map shows that sand prograded offshore about 40 km (sand unit 3). East of Mustang Island, progradation extended to 20 km offshore.

The localized distribution of sands in the southern portion of the study area suggests the main source of sediment feeding this shoreline was located to the south (Figure 4-21). A Rio Grande wave-dominated delta existed during this time (Banfield, 1998). The elongate lens geometry, steep seaward slope, and evidence of longshore movement of delta front and prodelta sediments, suggest that deltaic sands were transported northward during this time (Banfield, 1998).

Updip of the Stage 5d highstand shoreline is the Ingleside barrier Island chain. This Island chain is thought to have been deposited during the maximum rise in sea level, Stage 5e. It is difficult to assess whether the Ingleside shoreline is connected offshore to the mapped highstand shoreline, due to the absence of seismic and core information. Regardless, the Ingleside barrier chain represents the most landward highstand shoreline in the study area.

4.2.2 Stage 5c to Stage 3 deposition:

The unit above the Stage 5d unit is the Stage 5c to Stage 3 depositional package. This unit thinns and pinches out at approximately -50 m and is onlapped by Stage 3 deposits (Figure 4-13). Farther offshore, the upper
Figure 4-21: Depositional environments of the Stage 5c shoreline. Thick, extensive Stage 5c sands were deposited on the central Texas shelf, primarily from longshore transport of delta front sands from the ancestral Rio Grande HST 1 delta.
bounding surface is the amalgamated Stage 3/2 surface. Lithologic samples suggest clay to mud composition (mud unit 4). Seismic reflectors are subparallel and onlap only the updip basal 5c surface. Down dip, the reflectors downlap or are conformable with the 5c surface. This seismic facies (SFU C) is interpreted as a transgressive and highstand deposit. The updip onlapping reflectors are associated with Sub-stage 5c to 5a highstand and transgressive deposition. The offshore and overlying subparallel facies are associated with sea-level fall from Stage 5a to 3. A shoreline position of -50 m during 5a time is consistent with the updip limit of the Brazos Stage 5a delta mapped by Abdullah (1995) at -45 m. Moreover, similar to the unit mapped here, transgressive onlap followed by progradation and downlap offshore occurred in the Stage 5a Brazos delta (Abdullah, 1995).

The sub-stage 5c designation for the basal bounding surface is based on correlation with 2 oxygen isotope curves, the curve of Banfield (1998) in south Texas and Abdullah (1995) in central Texas. The curve of Abdullah (1995) suggests an age range between 5a to 5c. The isotope curve of Banfield (1998) suggests that the surface is sub-Stage 5c. Based on the similarity between central and south Texas stratal boundaries and seismic facies, the south Texas sub-Stage 5c interpretation is preferred.

The predominance of muds in the Stage 5c to Stage 3 deposits, suggests that depositional processes similar to today's may have been active along the central Texas shelf. The Mississippi River loop current may have been active, bringing in silts and clays from the Mississippi River and depositing thick
blankets of mud, similar to the modern ‘Texas mud blanket’ (Shideler, 1981).

The lack of sands may also be related to the absence of significant Rio Grande and Colorado deltas which would have been sources for sand. In addition, ravinement of the thin shoreface could have occurred during Sub-stage 5b to 5a sea-level rise.

4.2.3 Stage 3 Deposition – Phase 1:

The Stage 3 shingled, prograding unit is up to 10 ms (7.8 m) thick with extensive updip erosion associated with Stage 2 fluvial incision. Downdip, the Stage 3 unit pinches out and amalgamates with the Stage 2 sequence boundary (Figure 4-13). This facies (SFU B) is interpreted as a prograding highstand shoreline (Eckles, 1996; this study). A thick (up to 30 ms, 23 m) area to the east is associated with the early Stage 3 Colorado delta (Snow, 1998). The thick (60 ms, 46.5 m) area in the south is associated with the late Stage 3 Colorado and Lavaca/Guadalupe deltas. The paleo-shoreline position for the Colorado delta during early Stage 3 (Phase 1) time was -27 m, based on the updip termination of prograding medium to high amplitude subparallel reflectors (Snow, 1998). A paleo-environmental reconstruction at the time of deposition is shown in Figure 4-22.

The southern portion of the study area is dominated by reworked prodelta clays from the Stage 3 Rio Grande delta (Banfield, 1998). Banfield (1998) determined the shoreline position to be about -40 m from the updip termination of prograding oblique reflectors. This is a best estimate due to updip erosion
Figure 4-22: Depositional environments across the east to south Texas shelves during Stage 3. The ancestral Colorado and Rio Grande Rivers deposited fluvial-dominated deltas. Deltaic sands were likely transported along-shore toward the central Texas shelf. Three protuberances in the central Texas shoreline indicate fluvial deposition may have been responsible for greater progradation in these locations.
associated with Stage 2 fluvial incision (Banfield, 1998). Updip erosion of Stage 3 sediments across the central Texas shelf is common. The shallowest Stage 3 deposits in the area are found in -21 m water depth (corrected for subsidence), although the shoreline position may be higher, updip of cross-section 6 (Figure 4-6d).

Potential sources of sediments for the central Texas shoreface include the Hst 2 Rio Grande delta and the Stage 3 Colorado delta. Wave reworking of delta front sediments is suggested to be partly responsible for the overall lobate shape of the Colorado delta (Snow, 1998). These wave-reworked sediments were likely transported by longshore drift toward the central Texas shoreface, similar to longshore processes occurring today.

Three protuberances are located on the seaward side of the shoreline. These portions of the shoreline have prograded from 5 to 10 km farther seaward than the surrounding shoreline. These lobate-shaped protuberances may be wave-dominated deltas that were fed by low discharge rivers.

4.2.4 Stage 3 Highstand Delta – Phase 2:

Climinoform shape, a shelf-margin position, and the high sand content (lower sand unit 5) indicate that SFU E is a highstand delta associated with the Colorado, Lavaca, and Guadalupe valleys (Figure 4-23). Progradation is generally toward the southwest, based on clinoform orientations. The overall lobate shape suggests that the delta was a fluvial-dominated delta, but
Figure 4-23: Depositional environments of the shelf-edge highstand delta. The delta is composed of delta plain, delta front, and prodelta sediments and was sourced from the ancestral Colorado and Lavaca/Guadalupe rivers. Lowstand fan deposits are sourced from the highstand delta. Two slope canyons (East Breaks Slides) are shown. The western canyon is sand-filled and was sourced from the delta front. The eastern canyon is mud-filled and was sourced from the prodelta.
longshore drift and wave reworking affected the overall delta shape, as seen in the southwestward extent of the prodelta (Figure 4-23).

The highstand delta was nourished by up to 4 distributary channels (Figure 4-23). The western portion of the delta is very sandy and contains abundant shell hash (Figure 4-3). Line 20 (Figure 4-8) crossed two sublobes. The portion of the delta located between the Lavaca/Guadalupe distributary channel and the Colorado distributary channels is composed dominantly of clay and is likely part of the delta plain environment (Figure 4-23). Platform boring B-146 is located in the position of the delta (Figure 4-24) and was used to construct the oxygen isotopic stratigraphy of Abdulah (1995) (Figure 1-7).

Delta front sediments are only inferred, because of a lack of core information. Clinoform thickness is greatest near the center of the delta front, up to 60 ms (46 m), thinning to the west and east (Figure 4-23). The prodelta facies is composed of low angle prograding reflectors. This portion of the prodelta is likely composed of clay (Figure 4-23). The delta front and prodelta sediments source the slope fans during Stage 2 (Figure 4-23).

4.2.5 Lowstand slope canyons

Two submarine canyons are found downdip of the shelf edge (Figure 4-23). These canyons are called East Breaks slides and were mapped by Sidner et al. (1978), Woodbury et al. (1978), and Abdulah (1995). Sidner et al. (1978) and Woodbury et al. (1978) incorporated core material from oil industry sediment
Figure 4-24: Platform boring B-146 overlain on central Texas strike line 20. Oxygen isotope stratigraphy was performed by Abdullah (1995). Oxygen isotope stratigraphy from B-146 correlates well with stratal boundaries identified in this study. Location of B-146 is shown on Figure 4-1.
cores to assess the lithologic content of the canyon sediments as well as the timing of formation.

Slope core 14-6 is located in -549 m water depth, on the western fan downdip of the thickest portion of the Stage 3 highstand (Phase 2) delta front facies (Figure 4-23). Core descriptions describe distorted and contorted bedding with thick sand zones and thin interbedded muds (Woodbury et al., 1978). This unit occurs from 0 to 180 m core depth. Laminated clays occur between 180 and 270 m (Woodbury et al., 1978). Core 67-112 is located in the eastern fan in -200 m water (Figure 4-23). The upper 20 m of core 67-112 is composed of terrigenous sandy clay (Sidner et al., 1978). Below 20 m, cored sediments are described as clay (Sidner et al., 1978). Therefore, the slope fan proximal to the thickest delta front (46 m) contains a thick succession of slumped sands and muds (180 m), whereas the eastern slope fan, which was likely sourced from prodelta deposits, contains mostly clay. The timing of highstand delta progradation and associated slope canyon formation will be discussed at the end of Chapter 4.

4.2.6 Stage 2 fluvial incision:

The Colorado River has the largest fluvial drainage area in central Texas (110,000 km²). The other central Texas rivers have a combined drainage area of 40,000 km² (LeBlanc and Hodgson, 1959). Currently, the Lavaca River empties into Matagorda Bay, the Guadalupe River, empties into San Antonio Bay, the Aransas and Mission Rivers, empty into Copano and Aransas bays, the Nueces
River empties into Corpus Christi Bay, and San Fernando Creek empties into Baffin bay. These same rivers likely incised the exposed shelf during the Stage 2 lowstand.

A reconstruction of Stage 2 fluvial drainage patterns is presented in Figure 4-25. Incised valleys, in the western part of the study area, occur in water depths up to about -50 m. In deeper depths erosive topography characterizes the Stage 2 surface (Figure 4-13; 4-14). The offshore incised valleys have been tied updip to the valleys mapped by Wright (1980).

The Nueces, Mission, Aransas, and Guadalupe incised valley depths decrease from about -35 m within the bays to -19-28 m, 30 km offshore (Figure 4-25). Seaward of about -45 to -50 m water depth (30 km offshore), valley incision appears to terminate. The deepest valley incision west of the Lavaca/Guadalupe River confluence is -28 m (Figure 4-25). Incision depth decreases dramatically toward the west, near the Nueces River valley which is incised to -19 m (30 km offshore). Localized erosive topography occurs between -50 and -90 m water depth (38 to 80 km offshore) (Figure 4-13; 4-14). This erosive topography is interpreted to be related to bifurcation of the river channels during the overall sea-level fall from Stage 3 to 2. The downdip extension of channelization is inferred in water depths deeper than -90 m due to the absence of erosive topography in deeper water depths.

The Lavaca/Guadalupe and Colorado Stage 2 valleys incised up to -30 m, 30 km offshore. Farther offshore, valley incision decreased (-20 to -26 m) and the valleys bifurcated into narrower channels to the shelf edge (Figure 4-25).
Figure 4-25: Stage 2 fluvial valleys are superimposed on the Stage 2 time structure map. Valley depths decrease offshore in the western portion of the study area. These valleys bifurcate, shallow, and become narrower channels offshore. In the east, valley depths increase to approximately 30 km offshore. Farther offshore, valley depths decrease. The Lavaca/Guadalupe and Colorado river deposits bypass the shelf, feeding the slope fans.
A study by Berryhill et al. (1986) suggested a different scenario for the late Wisconsinan Stage 2 lowstand. The valley courses mapped to 30 km offshore are similar to those of this study (Figure 1-10; 4-25). However, at 30 km offshore, Berryhill et al. (1986) suggested that the fluvial valleys turned 90 degrees toward the south draining into one valley. This valley linked the central Texas lowstand drainage to the late Wisconsinan Rio Grande delta. A higher resolution seismic dataset as well as a detailed chronostratigraphic framework established in this study, suggests that a mid-shelf fluvial valley was not present during the Stage 2 lowstand. Furthermore, fluvial incision and channelization to the shelf edge is more plausible given a large base-level fall.

The exact nature of valley incision during falling sea level may be due to an interplay of several factors. These factors include the amount of sandy sediments available to the river and discharge rates. Given sparse core coverage, it is difficult to assess the amount of sand in the base of the Nueces, Aransas Bay, or Guadalupe incised valleys. Based on the central Texas core dataset (Chapter 3), the upper portions of these incised valleys are composed predominantly with fine sands. The basal fluvial sediments have not been cored, but based on a significant amount of sand reworked from underlying eolian deposits (Fisk, 1959) near Padre Island, high sand content within these valleys seems likely.

Currently, the overall discharge rates of the Nueces River, Guadalupe and San Antonio River, and the Aransas Bay rivers, are low, about 500,000 acre-feet per year for the Nueces (LeBlanc and Hodgson, 1959), and 150,000 acre-feet
per year for the combined Guadalupe and San Antonio rivers (White and Morton, 1987). Aransas Bay Rivers (Mission and Aransas rivers) have even lower discharge rates. These values are an order of magnitude lower than the Colorado River discharge (3,167,000 acre-feet/year) (LeBlanc and Hodgson, 1959). During the transgression from Stage 2 to 1, central Texas fluvial discharge rates are thought to have been high due to frequent flash floods (Blum et al., 1995). In contrast, discharge rates during glacial times (Stage 2) were probably much lower due to extensive soil formation in floodplains as well as thick vegetation cover due to wetter conditions (Blum et al., 1995). Therefore, low discharge rates likely predominated during Stage 2.

4.3 The role of eustacy, climate, and sediment supply:

The relative roles of eustacy, climate, and sediment supply on the facies architecture and stratal stacking patterns can be assessed based on the mapped thickness and spatial distribution of lowstand, highstand, and transgressive systems tract deposits on the central Texas shelf and slope. The following is a discussion of the interplay of these three factors.

4.3.1 Highstand systems tract (HST):

The HST can be divided into early highstand, from the Stage 5e/5d amalgamated bounding surface to the Stage 5c bounding surface; middle highstand, from the Stage 5c reflector to the Stage 3 maximum flooding surface;
and late highstand, from the Stage 3 maximum flooding surface to the Stage 2 sequence boundary. The early highstand prograding shoreline (Stage 5d) was nourished by wave reworking of the Hst 1 wave-dominated Rio Grande delta. The middle highstand Stage 5c to Stage 3 strata consists of onlapping and downlapping marine muds. The late highstand, Stage 3, was characterized by a lower sediment supply, with minimal along-shelf transport of sand, resulting in an overall thinner sediment package.

All three depositional packages were deposited during the falling limb of sea level, yet have widely differing lithologic facies. The early highstand Stage 5d prograding shoreline is 11.7 ms-thick (9.4 m), with the bulk of the sand-sized sediment sourced from the reworking of Rio Grande delta. In contrast, sand supply was low during Stage 5c to 3 due to the absence of significant deltas, whereas the mud sediment supply was high (up to 50 ms, 38 m).

Climate in the updip drainage basins of the Colorado and Rio Grande Rivers played a significant role in downdip delta development (Banfield, 1998; Snow, 1998) and ultimately in sediment supply to the central Texas shelf. Although direct evidence for Stage 5 and Stage 3 highstand climatic conditions in the updip drainage basins is lacking, warm and dry climatic conditions characterize the current highstand (Banfield, 1998). During highstand, the floodplain sediments that were sequestered in the drainage basin during the cooler, moister lowstand were excavated and released downstream. This is believed to have resulted from stripping of soil cover under dry climatic conditions (Blum, 1995; Snow, 1998; Banfield, 1998). Indirectly, an increase in
sedimentation on the east and south Texas shelves translates to increased sand available for longshore drift. Moreover, it is thought that the delta front sands of the Rio Grande were subsequently reworked alongshore and deposited across the central Texas shoreline during the early highstand.

This scenario is similar for the late highstand (Stage 3 shoreline), however, there is an overall thickness decrease of sandy sediments from Stage 5d to Stage 3. The cause of the thickness decrease from early to late highstand may be three-fold. First, the amount of wave reworking of the fluvial-dominated Colorado and Rio Grande deltas was minimal (Snow, 1998; Banfield, 1998), with limited transport of sand-sized sediment into the study area during Stage 3 time. Second, the possible updip fluvial feeders of the Stage 3 shoreline probably contributed only a small amount of sediment to the shoreline. Finally, an increase in the rate of sea-level fall from Stage 3 to Stage 2 combined with a lower overall sediment supply may have played a significant role in a thinner Stage 3 highstand shoreline deposits.

If the rate of sea-level fall was rapid during late highstand (Stage 3) then the low sediment supply shoreline would not have been able to keep up with the rate of fall and continue to prograde to the shelf break. This scenario is supported by flume experiments performed by Koss et al. (1994). They suggest that during a rapid base-level fall (highstand), low sediment supply systems deposit bedload sediment fraction (sands) on the shelf, with across shelf transport of suspended sediments.
A rapid base level fall from 10 km offshore (-21 m water depth) to 30 km offshore (-52 m) water depth occurred from early to late Stage 3 (from Phase 1 to 2). The Phase 1 delta (mapped by Snow, 1998) shifted seaward during late Stage 3 in response to low accommodation. The oblique progradational reflectors of the Phase 2 delta suggest high depositional rates during an overall rapid sea-level fall (Figure 4-9b).

4.3.2 Lowstand systems tract:

   Lowstand system tract Stage 2 fluvial incision was most pronounced during a rapid sea-level fall from Stage 3 to Stage 2. The updip rivers that fed the highstand Stage 3 shoreline excavated and incised the coastal plain as base-level fall increased associated with the sea-level fall to the shelf break.

   As discussed earlier, the ancestral Guadalupe River, the Mission and Aransas rivers, and the Nueces River formed shallow narrow channels that extended to the shelf break (-120 m). In contrast, the ancestral Lavaca/Guadalupe River and Colorado River valleys incised up to 20-30 m, 100 km offshore. The high sediment yield from these rivers likely bypassed the shelf and contributed to the sedimentary deposits that filled the lowstand slope fans.

   The striking difference in depositional patterns across the central Texas shelf is likely related to the sediment supply from the updip fluvial feeders which, in turn, is related to climate in the updip drainage basin. Climate during lowstand time is thought to have been cool and moist, sequestering sediments in the
updip fluvial drainage basin, with a low overall influx of fluvial sediments to the coastal plain (Blum et al., 1995). Given the same gradient shelf and the same rate of Stage 2 sea-level fall across the study area, sediment supply must have been the driving factor in the variable depositional patterns.

In the western portion of the study area, the river sediment supply may have been very low. No delta or slope fans are found downdip of the channelized system. It is likely that a low fluvial sediment supply combined with a rapid sea-level fall would produce minimal to no downdip deltaic sedimentation. Similarly, no lowstand delta is found downdip of the Lavaca/Guadalupe rivers or the Colorado River. It is likely that high sediment supply combined with a rapid sea-level fall caused sediment to bypass the shelf, accumulating in the slope fans.

The exact timing of slope canyon formation can only be inferred. Typically, slope canyons form during the maximum lowstand in sea level (Mitchum and Widmier, 1977). It is likely that rapid deposition of the highstand Colorado/Lavaca/Guadalupe delta near the shelf edge increased the water saturation of the delta front sediments. This led to sediment instability and subsequent slumping (Woodbury et al., 1978).

4.3.3 Transgressive systems tract:

Transgressive systems tract sedimentation began about 12,000 years ago, flooding the outer. Snow (1998) found that transgressive deltaic deposition did not begin on the central Texas shelf until after melt water pulse 1B,
approximately 11,500 years ago, and ended at approximately 10,760 yBP. The time-thickness map of the western-most lobe of Snow (1998) is shown in Figure 4-19. Two other eastern lobes have been mapped and are discussed in detail in the work of Snow (1998).

As sea-level rose, the shallow valleys of the lowstand river system were ravined. Channel depths may have been up to -20 m offshore the Nueces, Mission, Aransas, and Guadalupe incised valleys and -35 to -45 m offshore the Lavaca/Guadalupe and Colorado incised valleys prior to transgressive ravinement. Ravinement depths of the modern highstand and transgressive shoreface are thought to range between -8 and -12 m. This range in ravinement depth may be similar to the depth of ravinement on the outer shelf during the Stage 2 to 1 sea-level rise.

In addition to remnant river valleys on the outer shelf, a series of offshore pinnacle reefs form a linear trend at approximately -60 water depth (Figure 4-19). The exact timing of reef development is not known, but is thought to be during the early phase of transgression (Eckles, 1995; Berryhill et al., 1986). Berryhill et al. (1986) suggests reef growth commenced concurrently with a sea-level stillstand and shoreline development.

Timing of reef development can be constrained in this study based on the location of reefs outside of lowstand river channels and their position far from terrigenous sediment input. This is shown in figures 4-19 and 4-25. The reefs are positioned to the west and updip of the highstand shelf edge delta. In addition, reef growth is limited to inter-fluvial highs. Figure 4-19 shows the reef
position relative to the transgressive phase 1 delta mapped by Snow (1998). It is unlikely that reef growth would commence in the same location and at the same time as deltaic deposition. Therefore, an upper limit of reef growth during the transgression is prior to 11,500 yBP (before transgressive delta development). A lower limit of reef growth is about 12,000 yBP based on the composite Stage 2-1 sea-level curve (Figure 1-6).

An updip transgressive wave-dominated Colorado delta was mapped by Snow (1998) at -26 m water depth. It is likely that transgressive shoreline deposits are located within the incised valleys at -26 m water depth. Analysis of core material will help to identify the location and preservation potential of these deposits.

4.4 Summary

Highstand, lowstand, and transgressive systems tract deposits are located offshore central Texas. The highstand systems tract is divided into early, middle and late. Prograding sand unit 3 and coeval mud unit 4 offshore deposits characterized early highstand (Stage 5e/5d to 5c) deposition. Sediment input from erosion of a highstand Rio Grande delta was the dominant sediment source for the prograding shoreline.

Stage 5c to Stage 3 was characterized by onlap and downlap, with no coastal lithosome preservation. The absence of a thick highstand shoreline is associated with the absence of significant Colorado or Rio Grande deltas during this time. Late highstand Stage 3 was characterized by a thin prograding sandy
shoreline (sand unit 3) as well as a shelf-margin delta. The mapped distribution of the shoreline suggests that sediment supply was low, with minimal reworking of along-strike deltaic systems.

Lowstand systems tract deposition consists of late lowstand slope fans (East Breaks slides). The slope fan located downdip of the highstand delta front is dominantly sand-prone, whereas the fan adjacent to the prodelta is clay-prone.

Prominent features of the transgressive systems tract include a backstepping delta, sourced from the ancestral Colorado River and a linear reef trend on the middle shelf. Ravinement of the shallow Stage 2 fluvial channels occurred as sea-level rose. It is possible that transgressive shoreline deposits exist at the base of central Texas incised valleys on the inner shelf.
5.1 Modern and Pleistocene shoreline volume estimates

Difficulties exist when trying to relate the modern shoreline to preserved highstand shorelines offshore. The preserved highstand shorelines offshore prograded during the falling limb of sea level from Stage 5 to Stage 2. Currently, we are close to the maximum highstand, where sea-level rise is slow, about 3 mm/yr. Moreover, active deltaic deposition into the Gulf of Mexico is limited in extent, whereas during the Stage 5 to 2 sea-level fall, extensive Colorado and Rio Grande deltas existed as sediment sources for shoreline progradation. The shoreline prograded seaward due to a high sediment supply combined with a low accommodation space.

Regardless of these differences, several general insights into the long-term development and potential offshore progradation of the modern shoreline can be made. First, several volume calculations have been made, comparing the modern shoreface extent with the older Stage 5c to 3 shoreline, Stage 5d shoreline, and the Stage 3 shoreline. These calculations provide insight into the importance of sediment supply as well as accommodation space for shoreline aggradation and progradation. The duration of each systems tract deposit is shown in Figure 5-1.

The modern shoreface is an average of 2.5 m thick and has prograded an
Figure 5-1: The timing of highstand deposition on the central Texas shelf. The following volumes are growth rates of the shorelines: A) 0.4 km$^3$/1000 yrs B) 0.1 km$^3$/1000 yrs C) No shoreline observed D) 1.2 km$^3$/1000 yrs. (curves modified from Bard et al., 1990 and Chappell et al., 1996).
average of 7 km offshore. This modern shoreface has existed for about 3,500 years (see chapter 3). This equates to an average rate of growth of shoreface deposits of about 0.4 km$^3$/1000 years.

Similar volume estimates for the Stage 3 shoreline suggest a shoreline position of -21 m with a maximum progradation of 8.1 km offshore. The average unit thickness is 7.8 m and the amount of time available for progradation was 35,000 years. This equates to a volume of shoreline deposits of 0.1 km$^3$/1000 years. Erosion of the Stage 3 deposits by Stage 2 fluvial incision may have removed a significant portion of the shoreline, therefore this estimate is the lowest possible volume.

The Stage 5c to 3 shoreline is mud-prone. As discussed in Section 4.2.4, this shoreline may be similar to the modern shoreline. A sandy shoreline may have existed during this time, but due to a minimal sediment supply - no Colorado or Rio Grande River deltas emptied into the Gulf during this time - the shoreline may have been eroded during the overall fall from Stage 5c to 3 or the overall rise from Stage 5b to 5a.

Shoreface thickness calculations have been made for the Stage 5d shoreline. Assuming the shoreline position was about -20 m at the beginning of progradation (Banfield, 1998; Hst 1 shoreline position), the shoreline prograded an average of 16.5 km offshore. The unit averages 9.4 m thick and prograded offshore for about 10,000 years (Stage 5d to 5c). This equates to a rate of net growth of 1.2 km$^3$/1000 years.
The maximum net growth of preserved shoreline deposits associated with the Stage 5d shoreline is 1.2 km$^3$/1000 years. This value represents only 10,000 years of progradation as compared to the Stage 3 shoreline which represents about 35,000 years of progradation with a rate of 0.1 km$^3$/1000 years.

5.2 The effect of sediment supply on shoreline preservation

Three extremes of shoreline preservation are found offshore central Texas. The Stage 5d shoreline is thick and extensive. The Stage 5c to 3 shoreline is nonexistent and the Stage 3 shoreline is thin and limited in extent. The wide range of preservation offshore leads to possible hypotheses about the preservation potential of the modern highstand shoreline.

Presumably, the longer length of time that the Stage 3 shoreline prograded (Figure 5-1) should equate to thicker, more extensive shoreline deposits relative to the Stage 5d shoreline. There are two possible explanations why this is not the case. First, the amount of longshore current influx of sand-sized sediments from the surrounding Stage 3 deltas was probably low, considering both the Rio Grande and Colorado deltas were fluvial-dominated (Banfield, 1998; Snow, 1938). Similarly, the amount of sediment transported by the central Texas rivers may have been low. Secondly, the rate of sea-level fall was potentially very high from Stage 3 to Stage 2, up to 1.5 m/1000 years (Labeyrie et al., 1987: Figure 5-1). The Colorado and Rio Grande systems had a high sediment supply during that time and were able to keep pace with the rapid fall and fill their respective accommodation space, whereas the central Texas
shoreline could not keep pace with rapid base-level fall and was later incised by Stage 2 rivers.

On the other hand, the Stage 5d shoreline prograded for only 10,000 years (Figure 5-1), with an overall average net growth of 1.2 km$^3$/1000 years. This high volume, as compared to the Stage 3 volume, may be related to high sediment supply. The Stage 5d Hst 1 delta of the Rio Grande River (Banfield, 1998) was wave-dominated and was a major source of sand to the central Texas shelf at this time. In addition, sea level was rising during this time (Figure 5-1). This scenario may imply that during a parasequence-scale transgression in the overall highstand, high sediment supply was able to overwhelm the system leading to progradation of the shoreline and a high preservation potential.

The modern highstand shoreline has experienced average net growth of 0.4 km$^3$/1000 years. This is comparable to the Stage 3 shoreline growth. Sediment supply to the modern shoreline is mostly from reworking of former and current deltaic headlands, such as the Brazos and Colorado deltas, as well as reworking of east Texas upper shoreface deposits. These deposits are transported alongshore by longshore currents. The preservation potential of this shoreface is high, given the current slow rise of sea level. This is indicated by the fact that a significant portion of shoreface sediments currently occurs below the depth of fair-weather wave base.

In general, the Stage 3 highstand shoreline was unstable due to a rapid sea-level fall and a low sediment input. The Stage 5c to 3 shoreline was unstable, largely because sediment supply was very low. The Stage 5d
shoreline had a high sediment input and prograded despite rising sea level. As a result, a thick highstand shoreline was preserved.

It follows that there is a high preservation potential of a highstand shoreline if the sediment supply is great enough to overcome erosion and ravinement associated with sea-level rise and fall. It appears that the critical factor that determines the thickness and overall extent of the shoreline is the amount of sediment input into the system.

Given that the modern highstand shoreline has a low sediment input compared to Stage 5d time, and continues to prograde despite this low sediment input, the likelihood of preservation is uncertain. It may be inferred that the modern shoreline will be preserved if sediment supply to the system increases.

5.3 Comprehensive conclusions

1) The interplay of sediment supply, accommodation space, and the rate of sea-level rise and fall, determine the thickness and overall extent of highstand, lowstand, and transgressive deposits along the central Texas coast.

2) The central Texas shoreface experienced two episodes of shoreface progradation. The first episode (SF 1) appears to have occurred between 7,800 yBP to 4,500 yBP, based on correlation to the Sabine Bank paleo-shoreline offshore east Texas. Shoreface 1 deposition occurred in the transgressive systems tract. Rapid sea-level rise ravined SF 1. Later progradation of SF 2 likely occurred during the overall slow rise in sea-level rise from 3,500 yBP to present.
3) The amount of accommodation space determines how much shoreface preservation occurs during an overall sea-level rise. Shoreface sediments within valleys are effectively below the depth of transgressive ravinement and subject to faster subsidence and are therefore preserved during sea-level rise.

4) Central Texas shoreface sediments have a higher preservation potential than east Texas shoreface sediments due to a higher accommodation space, higher sediment supply, and steeper gradients.

5) Highstand, lowstand, and transgressive systems tract deposits are found offshore central Texas. The highstand systems tract contains sand-prone prograding shorelines and a prograding shelf margin delta. During lowstand oversteepening of the highstand shelf margin delta front and prodelta deposits led to extensive slope fan deposition. The transgressive systems tract consists of backstepping deltas, reefs, and preserved shoreline deposits within offshore incised valleys.

6) The thickness and extent of highstand shoreface deposits was controlled by the rate of base-level fall and sediment supply.

7) A comparison between the modern highstand shoreline and Pleistocene highstand shorelines offshore suggests that the preservation potential of the modern highstand shoreline is uncertain. The modern highstand shoreface is currently prograding below wave base, suggesting a high preservation potential. In contrast, comparisons to preserved highstand shorelines offshore suggest that sediment supply to the modern shoreline may not be
enough to preserve thick shoreface deposits.
References


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


LeBlanc, R. J., and Hodgson, W. D., 1959, Origin and development of the Texas shoreline, Transactions - Gulf coast association of geological societies, p. 197-220.


Winker, C. D., 1979, Late Pleistocene fluvial-deltaic deposition, Texas coastal plain and shelf Masters thesis: University of Texas at Austin. 187 p.


Appendix 1
Profile Cross-sections
A1-1: Cross-section of profile 1, located offshore Matagorda Peninsula. Stars and the letter C indicate grain size measurements. One radiocarbon sample is at T1-2, 90 cm (R label).
Figure A1-2: Profile 2 is situated at Pass Cavallo, at the southern limit of Matagorda Peninsula. See figure A-1 for legend information. Stars (*) and the letter C are grain size measurement and core photo locations, respectively.
Figure A1-3: Cross-section of profile 3, located offshore Matagorda Island.
Figure A1-4: Cross-section of Profile 4, located offshore Matagorda Island. The legend is located on Figure A-1. Stars (*) indicate a grain size measurement location. A C is core photo location.
Figure A1-5: Cross-section of Profile 5, located offshore Matagorda Island. Core photographs locations are labelled 'C'. Grain size analyses are shown for certain locations (*).
Figure A1-6: Cross-section of Profile 6, located offshore Matagorda Island. Legend information can be found on A1-1.
Figure A1-7: Cross-section of profile 7, located offshore the southern end of Matagorda Island. A similar pattern of deposition can be observed here and in profile 5. Legend information can be found on A1-1. One grain size measurement is shown.
Figure A1-8: Cross-section of Profile 8, located offshore of San Jose Island. Legend information can be found in profile 1,A-1. Two grain size analyses are shown. Two core photographs are identified as 'C'. One radiocarbon sample was taken at T8-54,350cm.
Figure A1-9: Cross-section of Profile 9, located offshore San Jose Island. Legend information can be found on profile 1, A1-1.
Figure A1-10: Cross-section of Profile 10, located offshore San Jose Island. Legend information can be found on profile 1, A1-1. Core photos ('C') are shown. Radiocarbon date locations are shown.
Figure A1-11: Cross-section of Profile 11, offshore Mustang Island.
Figure A1-12: Cross-section of Profile 12, located offshore Mustang Island.
Figure A1-13: Cross-section profile 13, located offshore the southern end of Mustang Island.
Figure A1-14: Cross-section of Profile 14, located offshore North Padre Island.
Figure A1-15: Core cross-section of Profile 15, located offshore North Padre Island.
Appendix 2
Composite Grain Size Curves (Sample locations shown in Appendix 1)
Figure A2-1: Comparison of grain size variability observed in USF 2 (modern) sediments. The samples are well sorted around a mean of 3.0 phi. Grain size sample location is designated by an asterick in Appendix 1 profiles.
Figure A2-2: Comparison of grain size variability observed in PLSF 2 sediments. The samples are labelled with a 'P' indicating profile location. Grain size varies from 3.0 phi to 3.3 phi. The PLSF 2 samples are moderately well sorted.
Figure A2-3: DLSF 2, grain size comparison in phi versus frequency weight percent. Samples are identified as P2-14, 39 cm (Profile 2, core 14, 39 cm). Grain size varies significantly across the DLSF 2 facies. P8-56, 270 cm is the only sample found to be 2.0 phi and poorly sorted. All other samples ranged from 3.0 to 3.5 phi.
Figure A2-4: Comparison of grain size variability observed in USF 1 deposits. Samples are labelled P1-5, which means, profile 1, core 5, cm location. The samples are well sorted, about 3.0 phi.
Figure A2-5: Grain size comparison of the variability observed across the PLSF 1 facies. Samples are labelled as Profile 1, core 7, centimeter location. Grain size ranges from 2.5 to 3.5 phi. The samples are well sorted.
NOTE TO USERS

Oversize maps and charts are microfilmed in sections in the following manner:

LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

The following map or chart has been microfilmed in its entirety at the end of this manuscript (not available on microfiche). A xerographic reproduction has been provided for paper copies and is inserted into the inside of the back cover.

Black and white photographic prints (17” x 23”) are available for an additional charge.
Core dataset map. Core profiles are numbered. NJM is Mustang Island, SJ is San Jose Island, and MAP Appendix 1: 1-15.
are numbered. NPI is North Padre Island, MUI is Island, and MAP is Matagorda Peninsula. Each profile is shown in
Plate 1: Central Texas

B

Nueces River
Stage 2 incised valley

Mustang Is
Lavaca River
Stage 2 incised valley

Matagorda Peninsula

0.0 2.0 4.0 6.0 8.0
1.0 2.0 3.0

17.5m >20 m? >20 m?
8-20m?

>20 m?

5 6
Shoreface profile gradients across east and central Texas, the central Texas shelf.

Profile 1: Matagorda Peninsula
Gradients increase from the east Texas shelf to...
Approximately 5 kilometers
Profile 7: Matagorda Island
Central Texas fence diagram showing the trends of shoreface 12, respectively) incised up to -36 meters within the bays during Subsequent progradation related to the modern shoreface 2 h
of shoreface profiles 1-15. The stage 2 sequence boundary is deepest in pro-
the bays during the Stage 2 lowstand. High accommodation space combined
shoreface 2 has prograded and downlapped onto marine and distal lower sho

stang Island
 deepest in profiles 1-4, 12. The sequence boundary depth is related to the presence combined with high sediment supply has resulted in thick successions of shallow lower shoreface sediments. The selected profiles below show many of the

Profile 15: North Padre Is
related to the presence or absence of fluvial incision. The Lavaca and Nueces successions of shoreface 1 sediments preserved within these valleys below rats show many of these regional trends.

Set: North Padre Island
face of fluvial incision. The Lavaca and Nueces rivers (profiles 1-4, fragments preserved within these valleys below ravinement surface 2. lands.
Profile 7: Matagorda Island
Profile 11: Mustang Island
Profile 15: North Padre Island