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RICE UNIVERSITY

THE EVOLUTION OF THE BRAZOS AND COLORADO FLUVIAL/DELTAIC SYSTEMS DURING THE LATE QUATERNARY:
AN INTEGRATED STUDY, OFFSHORE TEXAS

by

KENNETH CHRISTOPHER ABDULAH

A THESIS SUBMITTED
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ABSTRACT

The Evolution of the Brazos and Colorado Fluvial/Deltaic Systems During the Late Quaternary: An Integrated Study, Offshore Texas

by

Kenneth Christopher Abdullah

The evolution of the Brazos and Colorado fluvial/deltaic systems over the last 190,000 years is examined. The data set includes high-resolution seismic data, lithostratigraphic data from offshore platform borings, cores and gamma-ray logs, biostratigraphic data, oxygen isotope data, and radiocarbon dates. Major periods of fluvial/deltaic deposition for the Brazos and Colorado fluvial/deltaic systems have been mapped through time, and across 20,000 square kilometers of the Texas continental shelf and upper slope. Chronologic control has allowed an independent assessment of the timing and nature of fluvial/deltaic deposition relative to the SPECMAP oxygen isotope curve.

The Brazos and Colorado fluvial/deltaic systems have responded in phase with eustasy and facies distribution is related to fourth-order and fifth-order eustatic cycles. The Brazos fluvial/deltaic system developed deltas with highstand geometries during each of the oxygen isotope stages 5e, 5c, 5a, and 3. Fluvial incision and sediment bypass characterize the lowest sea-level stands. The Stage 2 to Stage 1 transgression is marked by a planar,
time-transgressive ravinement surface. During the transgression, the Brazos deltas backstepped across the shelf.

An oxygen isotope record, generated from benthic foraminifera, records the Mississippi meltwater pulse during the Stage 2 to Stage 1 transgression. Identification of isotope Stage 5 and Stage 3 is made possible by correlating the isotope data with paleoenvironmental data and regional downlap surfaces from seismic data. The interpretation of sea-level history and its role in fluvial/deltaic sedimentation has, therefore, been arrived at using three branches of stratigraphy (i.e., oxygen-isotope stratigraphy, biostratigraphy, and seismic stratigraphy). Lithologic data provide the ground truth for the seismic facies analyses.

An additional aspect of this study was the acquisition of a low-cost, high-resolution 3-dimensional seismic survey. The experimental 3-D survey successfully imaged an incised valley and its sedimentary fill. The survey covered a 750 m X 2000 m area over the Stage 2 Brazos incised valley.
ACKNOWLEDGMENTS

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The experimental 3-D seismic survey was processed by Mr. Shane Coperude and Dr. Anat Canning at HARC. I would like to thank them for
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INTRODUCTION

Objectives

The primary aim of this research project was to further our understanding of the response of fluvial/deltaic systems to global changes in sea level. This has been achieved through the study of two Quaternary fluvial/deltaic systems from the Texas Gulf Coast, i.e. the Brazos and the Colorado fluvial/deltaic systems, and their comparison to the higher sediment supply Western Louisiana fluvial/deltaic system studied by Sarzalejo (1993).

An additional objective was to test our ability to acquire low cost, 3-Dimensional seismic data. Through this experiment, we have gained further insight into the question of vertical and spatial seismic resolution as applied to stratigraphic studies.

Layout of the Thesis

Following this brief introduction, the thesis is presented in five chapters. Chapter 1 describes the study area, the geological setting, and defines the problems being addressed. Chapter 2 describes the methods used to address these problems. The data are presented in Chapter 3, and the integrated interpretation and discussion are presented in Chapter 4. The experimental, high-resolution 3-Dimensional seismic survey, and a
discussion of seismic resolution, are presented in Chapter 5. Separate summaries and conclusions follow Chapter 4 and Chapter 5.

CHAPTER 1 - THE STUDY AREA, THE GEOLOGICAL SETTING, AND THE NATURE OF THE PROBLEM

1.1 The Study Area and the Data Base for the Project

The study area lies on the east Texas continental shelf from Galveston Island to Matagorda Island, and offshore to the upper continental slope (Fig. 1-1). The study area covers approximately 20,000 square kilometers of the east Texas continental shelf, and part of the upper continental slope.

Over this area, a total of 3,400 kilometers of 2-Dimensional, high-resolution seismic data were recorded using Rice University's research vessel, the R/V Lone Star. The data on the shelf were recorded primarily with an EG&G Uniboom, a seismic source with a frequency bandwidth of 300-2400 Hertz (Hz), and penetration of 150-200 milliseconds (ms) two-way travel time (TWT), i.e., approximately 112.5-150 meters (m). On the upper slope, the data were recorded primarily with a 15 cubic inch SSI water gun which generates frequencies of 40-2000 Hz, and has a penetration of 500-600 ms (TWT), or approximately 375-450 m. Two regional strike lines, and one regional dip line, were recorded using a 50 cubic inch SSI Generator Injector gun. This source generates frequencies of 40-1400 Hz, and is
Figure 1-1  Study area and regional seismic grid. The area covered by the Texaco survey is also shown. The Texaco grid is shown in Fig. 1-2.
capable of imaging the first 1000 ms (TWT), or approximately 750 m of the uppermost sedimentary section.

A 1.5 square kilometer experimental 3-Dimensional survey was recorded in 1993. The target for the survey was a portion of the Brazos fluvial channel, incised during the last glacial maximum approximately 18,000 years before present (ybp). The data were recorded on the inner shelf using a 15 cubic inch SSI water gun. The objective of this experimental survey was to test our ability to acquire, at low cost, a high-resolution 3-D seismic volume for use in calculating the volume of coarse-grained valley fill.

An additional 1500 kilometers of high-resolution seismic data, recorded by Texaco, were also used in this study. The Texaco seismic data were recorded in late 1972 and early 1973 using a 40 cubic inch air gun (100-500 Hz). The Texaco seismic lines are part of the data set used by Lewis (1984) in her interpretation of the Galveston South Addition. The area covered by the Texaco seismic survey is shown in Figure 1-1, and the Texaco seismic lines used in this study are shown in Figure 1-2.

The seismic data have been integrated with 140 platform boring descriptions, and cores from 10 platform borings. The core data have been provided through Fugro-McClelland Engineers, and represent geotechnical data for prospective offshore production platforms. A typical platform boring description includes generalized lithologies and descriptions of color, stiffness, abundance of macro-fossils, and the occurrence of gravels and nodules (Fig. 1-3). Down-hole gamma-ray logs were available for 23 platform borings.
Figure 1-2 Grid of selected Texaco seismic lines used in this study
Figure 1-3  An example of a typical platform boring description. The description includes color, stiffness of clays, presence of calcareous nodules, and shell fragments.
The core data have provided the lithofacies control for the seismic facies interpretations, and have also provided benthic and planktic foraminifera for the generation of oxygen isotope records. The cores have also yielded organic carbon samples and molluscs for bulk radiocarbon dating, and macrofauna for paleoenvironmental interpretations. Detailed micropaleontologic analyses were carried out on cores from two platform borings, by Dr. Martin Lagoe and Lynette Holdford, as a separate study at the University of Texas at Austin.

Biostratigraphic data, and lithostratigraphic data, provided by Shell Development Company, from their 1965-67 Eureka Project, have also been integrated in this study. The Eureka cores were taken on the Texas upper continental slope, and average 305 m in length.

The data base for this study, therefore, consists of a regional grid of 4,900 kilometers of high-resolution seismic data, lithologic and biostratigraphic data from cores, and additional geochronologic control from radiocarbon dates and oxygen-isotope geochronology. Temporally, the study incorporates data over the last 190,000 years, i.e., from oxygen isotope stage 6 to the present. Figure 1-4 shows the locations of the regional seismic grid and core data sets.

1.2 A Brief Geologic History

The early history of the Gulf of Mexico is not as clearly defined as many ocean basins. This is due to the fact that, until recently, there has been little evidence for magnetic anomalies that typically date and define the
development of oceanic lithosphere. Although linear magnetic anomalies have recently been reported over a limited survey area in the eastern Gulf of Mexico (Hall and Najmuddin, 1992), linear magnetic anomalies characteristic of sea-floor spreading are yet to be identified over the central Gulf. In the absence of direct evidence of sea-floor spreading, a number of different theories for the opening of the Gulf of Mexico have developed.

The Gulf of Mexico began its development in the Late Triassic-Early Jurassic following the suture of the North American, South American, and African continents that created the super continent, Pangea. As Pangea broke up, the early rift phase left its record in the continental red beds of the Eagle Mills Formation and the Newark Group, both of North America, the Huizachal Formation of Mexico, the Todos Santos Group of Central America, and the La Quinta and Giron Formations of northern South America (Maze, 1984, Dengo, 1985, Pindell, 1985, Salvador, 1991).

The initial riftting phase was followed by a second phase of extension in the Middle-Late Jurassic that saw the thinning of the continental crust, and the development of extensive salt deposits (Buffler and Sawyer, 1985, Wu et al., 1990, Sawyer et al., 1991). Ebeniro et al. (1988) used large capacity air guns, and ocean bottom seismometers (OBS's) to develop velocity models in order to arrive at a velocity structure for the Gulf of Mexico. Relatively thin crust (i.e., 5 km in the region of the mid-slope) of variable velocity, led the authors to conclude that the opening of the Gulf of Mexico was characterized by non-uniform extension. Figure 1-5 (from Sawyer, 1991) shows the distribution of crustal types and the distribution of salt in the Gulf of Mexico. The basin asymmetry, seen in Figure 1-5, tends
Figure 1-5  Crustal types in the Gulf of Mexico after Sawyer (1991). The area enclosing the pre-marine evaporites is shown by a dashed outline.
to support the interpretation of Ebeniro et al. (1988) regarding non-uniform extension.

The Middle-Late Jurassic extension was followed by a relatively short (5-10 m.y.) period of sea-floor spreading during the Late Jurassic (Sawyer et al., 1991). In Figure 1-5, the salt deposits can be seen separated by oceanic crust, although there is still some debate as to whether this separation was pre-depositional, or post-depositional (Salvador, 1991). This brief period of sea-floor spreading was followed by cooling and subsidence that created the space for the deposition of 10-15 km of sediments on top of the Late Jurassic oceanic crust (Salvador, 1991).

During the Early Cretaceous, the Gulf of Mexico extended from the Pacific to the Atlantic Oceans. Carbonate deposition dominated the western, eastern and southern areas, and siliciclastic deposition was restricted to the northern margins of the basin (Salvador, 1991). The Late Cretaceous saw the lowering of sea level, and the development of the mid-Cenomanian unconformity. Siliciclastic sediments prograded south off the Ouachita and Appalachian Mountains, and carbonate platform sedimentation became restricted to Mexico, Florida and the Yucatan (Salvador, 1991). Connection to the Cretaceous Interior seaway became cut off at this time, although it was temporarily re-established during the Turonian transgression.

The Turonian saw the re-connection of the Gulf of Mexico with the Western Interior seaway, along the area of the Mississippi embayment. This connection continued up until the Santonian (Salvador, 1991). During this period, Turonian to the Santonian, the western margin of the Gulf of Mexico was connected with the Pacific Ocean with extensive carbonate platform
development along this western margin. Clastic sedimentation, during this interval of time, was restricted to a narrow zone along the northeastern portions of the basin, i.e., along the flanks of the Ouachita and Appalachian Mountains.

The western margin of the Gulf of Mexico basin saw a dramatic change in the Campanian-Maastrichtian as a result of the Laramide orogeny. Carbonate platform development ceased along the western margin during this time and siliciclastic deposition occurred in response to the development of the western cordilleras of North America and Mexico. The carbonate deposition that characterized most of the Lower Cretaceous rock record became restricted to the Yucatan peninsula, in the south. In more westerly parts of the basin carbonate sedimentation was abruptly replaced by siliciclastic deposition in response to the Laramide orogeny.

The Laramide orogeny was also responsible for increased siliciclastic input from the northern and northwestern flanks of the Gulf of Mexico basin. Progradation of thick, offlapping sedimentary wedges characterized the Tertiary deposits of most of the basin (Martin, 1978, Winker, 1981, Winker, 1982, Galloway et al., 1991, Salvador, 1991). The exceptions being the Yucatan peninsula, where carbonate sedimentation continued throughout the Cenozoic, and the Florida peninsula, that was also dominated by carbonate deposition, except during the Miocene and Pliocene when renewed uplift of the Appalachians resulted in a period of dramatic siliciclastic progradation (Salvador, 1991). Carbonates were eventually re-established in the Florida peninsula area during the Pleistocene.
The Tertiary sequences were sourced by three main drainage basins that entered the Gulf of Mexico through three embayments. These embayments are, from east to west, the Mississippi, Houston and Rio Grande Embayments (Winker, 1981). Figures 1-6a and 1-6b, after Winker (1981), illustrate these three main drainage basins. Major fluvial/delta systems were associated with these three drainage basins, and these were the fore-runners to today's Mississippi delta, Brazos and Colorado deltas, and the Rio Grande delta, respectively.

Tertiary deltaic systems, associated with the Houston embayment, include the Rockdale delta (late Paleocene), the Fayette delta (late Eocene), and the Houston delta (late Oligocene) (Galloway et al., 1991). Major canyon systems, such as the Paleocene Lavaca and Yoakum canyons and the Oligocene Hackberry canyon, are also associated with the Tertiary development of this northern portion of the continental shelf and slope of the Gulf of Mexico basin (Galloway et al., 1991).

The large influx of Tertiary sediments into the northern and northwestern parts of the Gulf of Mexico basin resulted in flexural loading and associated subsidence (Sawyer et al., 1991). Sediment loading also mobilized the underlying, Mesozoic, salt (Martin, 1978, Amery, 1978, Worrall and Snelson, 1989, Wu et al., 1990). The Tertiary tectonic history of the Gulf of Mexico has therefore been closely related to the effects of rapid sedimentation. On the longer time scale, the tectonic history of the basin is characterized by subsidence due, initially, to continued thermal cooling of the Mesozoic oceanic crust and, later, to flexure loading due to sediment loading. On the shorter time scale, however, growth faulting, slope
Figure 1-6a  Tertiary drainage basins after Winker (1981)

Figure 1-6b  Tertiary paleo-shelf margins after Winker (1981)
instability and failure, shale diapirism, and salt tectonics, dominated the
tectonic style and persisted into the Quaternary.

The tremendous input of siliciclastic sediments from the northern
portions of the Gulf of Mexico basin continued into the Quaternary. Figure
1-7, from Galloway et al. (1991), summarizes this continued clastic
progradation. Glacial-eustatic cycles were also present in the Tertiary, but
are not easily separated from the tectonic pulses associated with the
Laramide orogeny. However, glacial cycles became more dominant in the
Pleistocene. High amplitude, high frequency Pleistocene glacial-eustatic
fluctuations have imparted a dominant signature to the stratigraphic record
of the Gulf of Mexico.

The outbuilding of the continental shelf of the northern and north-
western Gulf of Mexico, during the Quaternary, has been largely a result of
siliciclastic deposition from the Mississippi, Brazos and Colorado
fluvial/deltaic systems. The major factors controlling the nature, and timing,
of the regressions and transgressions have been changes in global sea level
and changes in sediment supply. These fluctuations have left their signature
in the stratigraphic record of the continental shelf and upper slope. This
signature has been frozen by the record of paleo-shorelines, the position and
nature of paleo-deltas, interfluvial deposits, and incised valleys and their
sedimentary fill.

The late Quaternary history of the Brazos and Colorado fluvial/deltaic
systems is presented in this thesis. The study area covers the present marine
area from the innermost continental shelf to the upper continental slope.
Using an integrated approach to the study of stratigraphy this project
Figure 1-7  Regional north-south cross section, after Galloway (1991)
attempts to illustrate the separate effects of eustasy, and sediment supply, on fluvial/deltaic deposition.

1.3 The Nature of the Study, and the Quaternary Setting

1.3.1 The Role of Tectonics, Eustasy, and Sediment Supply.

Today, it is widely accepted that the major factors controlling siliciclastic deposition are tectonics, eustasy and sediment supply (Vail et al., 1977, Bally, 1982, Winker, 1982, Pitman and Golovchenko, 1983, Posamentier et al., 1988, Cloetingh, 1988, Galloway, 1989, Perlmutter and Matthews, 1989, Embry, 1989, Van Wagoner et al., 1990, Christie-Blick, 1991, Vail et al., 1991, Posamentier and Allen, 1993). It is also widely accepted that each of these factors operate at different temporal scales (Vail et al., 1977, Haq et al., 1987, Jervey, 1988, Embry, 1989, Galloway, 1989, Lopez, 1990, Mitchum and Van Wagoner, 1991, Vail et al., 1991, Embry, 1995). Orogenic events typically occur over the time span of millions of years, whereas glacio-eustasy has been seen to follow changes in solar insolation and, by extension, the periodicity of orbital forcing (tens to hundred of thousands of years) (Emiliani, 1955, Broecker, 1966, Hays et al., 1976, Imbrie et al., 1984, Imbrie et al., 1993). Changes in sediment supply can occur with a periodicity that ranges from one year (annual wet and dry seasons) to thousands of years (climatic patterns linked to glacial and interglacial controls on ocean circulation), and perhaps even millions of years (climatic changes due to the alteration of atmospheric circulation by major orogenies).
Although there is agreement that tectonics, eustasy and sediment supply are the main factors that shape the sedimentary rock record, controversy exists as to the relative roles of these three factors (Bally, 1981, Burton et al., 1987). This debate is by no means new, as pointed out by Dott (1992). Dott (1992) refers to the early works on the influence of eustasy, by Suess in 1888, and on the effects of tectonics, by Chamberlin in 1898 and 1909. Dott (1992) also reminds us that the theory of orbital forcing and climate change was already documented by Croll in 1864, and Milankovitch in 1920. Limitations of geochronology (Marshall, 1990, Ricken, 1991) will probably ensure that this debate will continue for many more years to come, particularly for the ancient record.

Any attempt at predicting the spatial distribution of sedimentary facies through time requires further understanding of the relative roles of tectonics, eustasy, and sediment supply. However, the fact that sediment supply is often so closely related to both tectonics and climate, and the fact that glacio-eustasy is also linked to climate, makes the separation of these factors non-trivial. This statement becomes even more relevant when we consider the fact that our ability to unravel the relationship between tectonics, eustasy, and sediment supply is limited both by our observations, and by our geochronologic control. Older formations are often bracketed by boundaries that are separated by millions of years, i.e., at the temporal scale of tectonic events, even though most of the sedimentation may have occurred over relatively short time periods. Since it is usually impossible to constrain these higher frequency depositional events, potential climatic and eustatic effects remain embedded in the lower frequency tectonic record.
One way around these problems is to focus our attention on periods of time, and geographical areas, where we have more control on, or at least a better understanding of, the factors that produced the stratigraphic record. It is for this reason that the Quaternary geology of the east Texas continental shelf has become a focus of study at Rice University.

The objective of this study is not to prove, or disprove, the relative roles of tectonics, eustasy and sediment supply in the formation of ancient sedimentary deposits. Rather, it is an attempt to document the role that these factors have played in a study area where it is possible to constrain these important variables. The Quaternary rock record of the Texas shelf, when placed in context, may help to clarify some of the stratigraphic relationships observed in the ancient rock record. The nature of these three variables, during the Quaternary, are discussed briefly below.

1.3.2 The Quaternary Tectonic Setting

The tectonic setting of the Quaternary is very similar to that of the late Tertiary. Thermal subsidence is no longer active in the Gulf of Mexico basin, but the margin can be seen to be far from "passive". The tectonic factors affecting the sedimentary section, the relative rates of which show considerable variation, are:

1) subsidence due to flexure of the underlying lithosphere, in response to both the sediment load and the water column,
2) subsidence due to sediment compaction,
3) subsidence and uplift, due to salt and shale flowage, in response to
rapid sediment loading, and
4) growth faulting and shelf margin failure, or "mass wasting", due to
oversteepening of the shelf margin during periods of rapid
progradation.

Subsidence due to flexure loading (Walcott, 1972) varies across the
Texas continental shelf. Estimates from Winker (1979), averaged over the
last 120,000 years, indicate a range of values from 0.1 mm/year to 1.2
mm/year, from the inner to outer shelves, respectively. These values show
good agreement with estimates from this data set based on paleo-shoreline
positions for the latest Quaternary, and geochronology from the SPECMAP
oxygen isotope curve. Data from the study area indicate late Quaternary to
Holocene subsidence rates of 0.13 mm/year for the inner shelf, and from 2-4
mm/year for the outer shelf. Of course, all of these estimates include the
effect of sediment compaction.

Sediment compaction on the Texas shelf, by itself, is considered to
make only a very minor contribution to the overall space available for
deposition (Siringan, 1993). Exceptions do occur, locally, and very high
subsidence rates for the Texas inner shelf have been documented over
historical times. These high subsidence rates, up to 6 mm/year, are
attributed to compaction of barrier and valley-fill sediments within the
Trinity incised valley (Siringan and Anderson, 1993).

Salt tectonics, in response to rapid sediment loading, has produced
intraslope basins, and piercing salt diapirs (Lehner, 1969, Tatum, 1977,
Snelson, 1989, Satterfield and Behrens, 1990, Wu et al., 1990). Salvador (1991) estimates that some 3.8 km of sediments were deposited offshore Louisiana during the Pleistocene. However, over 7.5 km of Pleistocene sediments have accumulated, locally, due to subsidence in salt-related mini-basins of the Louisiana slope (Worrall and Snelson 1989). Local subsidence rates of 4.6 mm/year can, therefore, occur within these salt-withdrawal basins.

Although salt tectonics are not a major factor in the study area, data from the study area show that diapiric uplift can occur, locally, at minimum rates of 5 mm/year. With maximum sedimentation rates in the study area estimated at 3 mm/year, diapirs are more than able to pond sediments on the upper slope. An example of ponding has been described for the East Breaks Slide (Lehner, 1969). Shale diapirism (sedimentary volcanism) has not been identified, although it has been well documented offshore Louisiana (Neurauter and Bryant, 1990).

1.3.3 The Quaternary Sea-Level Record

The record of the Pleistocene glaciation, as evidenced by Gulf coast stratigraphy, was first documented by Fisk during his work on the Mississippi River and its alluvial plain (Fisk, 1940, 1944, 1952). This early work related the Pleistocene stratigraphy to the glacial and interglacial stages documented for the Laurentide ice sheets, i.e., the Nebraskan, the Kansan, the Illinoian, and the Wisconsin glacial stages, and the Aftonian, Yarmouth, Sangamon, and Peorian interglacials. Fisk related the development of the Williana, Bentley, Montgomery, and Prairie terraces of Louisiana to deposition during successively younger interglacials. Figure 1-8, from Bernard and LeBlanc (1965) (after Fisk, 1944) shows the relationship between glacial, interglacials and lithostratigraphy. This concept of interglacial deposition, and glacial valley entrenchment set the stage for the further understanding of the role of transgressions and regressions in the stratigraphic record of the Gulf Coast.

Later works include those of LeBlanc and Hodgson (1959), Curray (1960), Bernard and LeBlanc (1965), Frazier (1967 and 1974), McGowen et al. (1976), Wilkinson and Basse (1978). LeBlanc and Hodgson (1959) documented the role of river entrenchment during the last glacial, followed by the flooding of these older valleys to develop the Texas estuaries. They also documented the Holocene regressive delta plains that were developed by the Brazos and Colorado rivers, and later mapped in detail by McGowen et al. (1976). This regressive phase was associated with the latest sea-level stillstand. Curray (1960) dated relict offshore features that he interpreted as paleo-shorelines. Frazier (1967) documented the phases of development of
Figure 1-8  Glacial cycles and the Quaternary stratigraphy of the Gulf coastal plain, from Bernard and LeBlanc (1965) (after Fisk, 1944)
the Mississippi delta lobes, and the role of depositional episodes in the stratigraphy of the Gulf Coast (Frazier, 1974). These studies all focused on the onshore, or nearshore, distribution of sedimentary environments and facies in time and space.


Difficulties in obtaining an accurate sea-level record from the stratigraphic record, highlighted by Bernard and LeBlanc (1965), still hold true today. The list of Bernard and LeBlanc (1965) included local subsidence rates, inaccuracies in paleo-water depth estimates for organisms used in determining paleobathymetry, the range of water depths for peats, ability to recognize marsh or swamp peats, and their reworking, and errors in $^{14}$C dates. Preservation potential of coastal deposits can be added to this list (Siringan and Anderson, 1991, and Siringan, 1993). Thomas (1990) and Thomas and Anderson (1991, and 1994) focused on the record of preserved fluvial terraces within the Trinity/Sabine incised valley, in order to estimate paleo-sea levels. They attributed the episodic nature of their sea-level curves
to decoupling of marine ice sheets in Antarctica (Anderson and Thomas, 1991).

The relationship between the oxygen isotope record and global ice volume (Shackleton and Opdyke, 1973) provides us with a proxy for changes in global sea level (Shackleton, 1987). In fact, it is this relationship that opens up interesting opportunities for deductive, and predictive, analyses in stratigraphy (Mathews, 1990). The SPECMAP curve (Imbrie et al., 1984) serves as the proxy in this study, and provides an independent indicator of sea-level changes over the time frame of this study, i.e., over the last 190,000 years. The SPECMAP curve is shown in Figure 1-9a and it is shown converted to depth, by G. Haddad, in Figure 1-9b. The conversion to depth assumes that sea level was 6 m higher than today during the last interglacial, and 120 m lower than the present level during the last glacial. A 10 m sea-level change for every 0.1 per mil change in $\delta^{18}O$ has been applied, after Shackleton and Opdyke (1973).

The results of this study show good agreement with the SPECMAP curve, with the exception of the amplitude of isotope Stage 3. Data from this study show a sea-level stand of -30 m to -35 m, below present sea level, for the highest sea-level position during isotope Stage 3. A higher sea-level stand during Stage 3, of -40 m, also was documented by Thomas and Anderson (1991). In this regard there is better correlation with coral terrace data from the Huon Peninsula of New Guinea (Bloom et al., 1974, Bloom and Yonekura, 1985). Figure 1-9c shows an updated version of the Huon Peninsula sea-level curve from Bloom and Yonekura (1985).
Figure 1-9a The SPECMAP curve
(Imbrie et al., 1984)

Figure 1-9b The SPECMAP curve
converted to depth by G. Haddad

Figure 1-9c Sea-level curve from the
Huon Peninsula, after Bloom and
Yonekura (1985)
1.3.4 The Quaternary Climate Record

The Quaternary climate is perhaps one of the more difficult variables to constrain, since it is dependent on an understanding of what Imbrie et al. (1993) refer to as "both the "fast physics" of the atmosphere and the "slow physics" of the oceans and cryosphere". On the long-term scale, there are the major glacial and interglacial cycles that are related to the 100 ky, 41 ky, 23 ky, and 19 ky variations in solar insolation, the Milankovitch cycles (Imbrie et al., 1984). For these we rely on the oxygen isotope record as our indicator of paleoclimate as far as global temperatures and ice volume are concerned. Emiliani (1955) provided the basis for the relationship between the oxygen isotope record, glacialis and interglacials, and the link to wet and dry cycles.

On the shorter time scale, we are confronted with the fact that the stratigraphic record is punctuated by catastrophic climatic events that may occur at periods as short as 10 years. The flooding cycles of the Brazos River, and their dramatic effect on the stratigraphic record of the modern Brazos delta, is such an example (Hamilton, 1995).

Blum (1993) and Blum and Price (1994) describe terrace deposits from the Colorado River that were deposited over time intervals that varied from 2 ky to 7 ky. Blum (1993) and Blum and Price (1994) attribute the development of these fluvial terraces to periods of increased fluvial discharge. The authors use this record as evidence for climatic, rather than eustatic, controls on incised-valley fill architecture (Blum and Price, 1994). However, the preservation of these terraces in the longer term geologic
record is debatable. High-frequency terrace records are vulnerable to future episodes of stream erosion, and may prove to be ephemeral.

The state of Texas spans five major climatic belts that run roughly NNW-SSE across the state (Fig. 1-10a). From east to west the climatic belts are humid, moist subhumid, dry subhumid, semiarid, and arid (Russell, 1945, Thornthwaite, 1948). The drainage patterns of the Brazos and Colorado rivers are shown in Figure 1-10b. Wet and dry phases linked to glacials and interglacials, respectively, would have a marked effect on the climatic belts, fluvial discharge, and on the resulting deltaic deposition.

The correlation between glacial cycles and the stratigraphic record is not straightforward. Thomas (1990), and Thomas and Anderson (1994), show evidence for increased fluvial deposition during isotope Stage 5, an interglacial. This is supported by the presence of large Stage 5 Brazos deltas identified in this study. This project also presents evidence of high levels of fluvial discharge in the study area during both isotope Stage 6 (a glacial), and isotope Stage 3 (an interstadial).

The fluvial response to changes in climate and base level is in fact very complex (Schumm and Brakenridge, 1987). The same river can show different responses along its course in response to changes in gradient, and changes in sediment load as it flows through different climatic belts (Schumm and Brakenridge, 1987). In addition, the susceptibility of arid areas to intermittent periods of large scale erosion must be weighed against increased precipitation, but also increased vegetation, during more humid conditions. Further complexities develop when we consider increased surface run off during lowstands of sea level. Complex dendritic drainage
Figure 1-10a  The Climate belts of Texas (after Russell, 1945, and Thornthwaite, 1948)

Figure 1-10b  The Drainage patterns of the Brazos and Colorado Rivers
can develop on the newly exposed continental shelf, providing a source for significant quantities of sediment, independent of the main fluvial source (Koss et al., 1994). All we can say is that the study area shows evidence of periods of increased sediment supply that are most likely due to climate changes.

Biostratigraphic records also provide us with an indication of paleotemperatures. For example, the coiling ratios between left and right coiled forms of the planktonic foraminifer *Globorotalia truncatulinoides* indicates either cold (left-coiling dominant) or warm (right-coiling dominant) intervals (Kennett and Huddleston, 1972, Beard et al., 1982). The foraminifers *Globorotalia menardii* and *Pulleniatina obliquiloculata* are associated with warm water (interglacials), and the foraminifers *Globorotalia inflata* and *Globigerina falconensis* with cool water (glacials), (Kennett and Huddleston, 1972). The biostratigraphic report from the upper slope Shell Eureka project (Anderson and Parrott, 1969) has proven invaluable in this regard.

1.4 Stratigraphic Models and Predictive Stratigraphy

Predicting the distribution of sedimentary facies requires that the roles of tectonics, eustasy, and sediment supply be quantified, in the case of computer models, or at least semi-quantified, in the case of conceptual models (Jervey, 1988, Posamentier et al., 1988, Tetzlaff, 1989, Perlmutter and Mathews, 1989, Posamentier et al., 1992, Posamentier and Allen, 1993, Bowman, 1994, Ross et al., 1994, Ross et al., 1995). However, difficulties
exist in separating the tectonic and the eustatic components (Burton, et al., 1987, Christie-Blick, 1991). Tetzlaff (1989) points out the complexity of the issue and demonstrates the importance of using 3-dimensional, rather than 2-dimensional, models to predict stratigraphy. There is also the need for models to include an aspect of the random behavior typical in nature (Tetzlaff, 1989).

One would think that, with the tectonic and eustatic signatures constrained during the Quaternary, the study area would be a simple candidate for simulation. However, this study demonstrates the complexities involved in predicting the spatial and temporal distribution of sedimentary facies. This complexity is a result of the interplay between allocyclic events, and autocyclic events. Separation of these events requires an integrated approach to mapping the three-dimensional distribution of sedimentary facies through time. The problem is, at best, four-dimensional.
CHAPTER 2 - METHODS

Introduction

Data from five branches of stratigraphy (i.e., seismic stratigraphy, lithostratigraphy, biostratigraphy, oxygen isotope stratigraphy, and chronostratigraphy) have been integrated in this project. In some cases the data were acquired and analyzed at Rice University. In other cases (e.g., the biostratigraphic data) the analyses have come from existing reports from Shell Development Company, or through collaboration with Dr. Martin Lagoe, from The University of Texas, at Austin. The methods applied within each subdiscipline are discussed separately, below.

2.1 Seismic Stratigraphy

Seismic stratigraphic techniques, developed and published by Vail et al. (1977), are applied in this study. The method centers on the recognition of important bounding surfaces, and on the identification of variations in seismic character. Key bounding surfaces allow us to bracket genetically related sedimentary strata (Vail et al., 1977, Mitchum et al., 1977, Vail et al., 1987). The variation of seismic character, when placed within the context of the sedimentary sequence, allows for the classification of distinct seismic facies that can be related to depositional environment and lithology (Mitchum et al., 1977, Mitchum et al., 1978, Sangree et al., 1978).
High-resolution seismic data allow us to image important surfaces within the stratigraphic section. These surfaces are defined on the seismic sections by the manner in which the seismic reflections, either above or below the surface in question, are terminated. Five types of surfaces have been recognized within the study area. These surfaces are:

i) Erosional surfaces associated with periods of rapid sea-level fall, i.e., the sequence boundaries of Vail et al. (1977). On dip sections, reflections below the sequence boundaries show high-angle, irregular, erosional truncation (Fig. 2-1). This surface becomes conformable in a down-dip direction. On strike sections, the sequence boundary is often characterized by fluvial incision (Fig. 2-2).

ii) Erosional surfaces associated with periods of rapid sea-level rise (i.e., the ravinement surface of Swift, 1968, Demarest and Kraft, 1987, and Nummedal and Swift, 1987). On dip sections, reflections below the ravinement surface can show high-angle, planar, erosional truncation (Fig. 2-3). The ravinement surface is a relatively smooth surface, it is not associated with fluvial incision, and it dips gently seawards.

iii) Downlap surfaces associated with periods of low sedimentation rates linked to times of maximum transgression, i.e., the maximum flooding surfaces (mfs) of Posamentier et al. (1988). The reflections
Figure 2-1  Sequence boundary (SB) on line R91-4. It forms the base of the chaotic reflections associated with fluvial channels. (10 ms = 7.625 m; VE = X37)
Figure 2-2  Sequence boundary (SB) on seismic line R90-9. This is a strike line and shows a 30 m deep incision. (10 ms = 7.625 m; VE = X 25)
Figure 2-3  The ravinement surface is a relatively smooth, planar surface that dips gently seawards. In contrast, the sequence boundary is characterized by fluvial incision. (10 ms = 7.625 m; VE = X20)
immediately above the maximum flooding surface show low-angle downlap. The maximum flooding surface is a smooth, laterally continuous, widespread event, characterized by high amplitudes and an absence of incisions (Fig. 2-4).

iv) Onlap surfaces that define the landward extent of deltaic deposition (Fig. 2-5).

v) Onlap surfaces that define the lateral extent of deltaic deposition (Fig. 2-6).

Surfaces (i), (ii), and (iii) allow us to separate regressive deposits from transgressive deposits. The regressive deposits are bounded either by a sequence boundary above and a maximum flooding surface below, or by a sequence boundary below and a ravinement surface above. Transgressive deposits are bounded below by either the ravinement surface or a sequence boundary, and above by the maximum flooding surface.

Surfaces (iv) and (v) allow us to constrain the areal extent of different episodes of deltaic deposition. The age of surfaces (i), (ii), and (iii), and the timing of the episodes of fluvial/deltaic deposition, both in relation to an independent sea-level curve, have been the primary focus of this study. High-resolution seismic data are the means by which these surfaces can be mapped at submeter levels of vertical resolution over large areas. In this study, the area covers 20,000 sq. km of the Texas continental shelf and uppermost slope.
Figure 2-4  Maximum flooding surface (mfs) on seismic line R90-5. On the inner shelf, the maximum flooding surface is a near-horizontal surface that separates the transgressive deposits, below, from the overlying, downlapping regressive deposits. (10 ms = 7.625 m; VE = X30).
Figure 2.5
Estimated positions of paleoshorelines can be determined from onlap. Minor incisions interpreted as distributary channels are also visible on this dip oriented seismic line. (10 ms = 7.625 m; VE = X35)
Figure 2-6  The lateral limit of delta lobes can be determined by onlapping relationships between younger and older lobes. The delta's areal extent can then be mapped. (10 ms = 7.625 m; VE = X200)
The interpretation of depositional environments from seismic data requires a link between the character of the seismic data and sedimentary facies. Groups of reflections, comprising seismic reflections with similar characteristics, are analyzed in terms of their external form, and in terms of the configuration, continuity, amplitude, frequency, and interval velocity of their reflections (Mitchum et al., 1977). Interval velocity is the only parameter, listed above, that cannot be examined on the high-resolution single-channel seismic data used in this study.

When a group of like seismic reflections can be clearly defined, and is mappable in three dimensions, it constitutes a seismic facies unit of Mitchum et al. (1977). It is the three-dimensional association of seismic facies units that allows for the interpretation of depositional environments, processes, and lithology (Mitchum, et al., 1977). The link between seismic data and depositional environment becomes reinforced when the seismic facies units can be tied directly to lithologic data.

Sangree and Widmier (1977) applied this approach, which they termed "seismic facies analysis" to define nine seismic facies units characteristic of the continental shelf, shelf margin, slope, and basin floor. The authors recognized that vertical resolution was likely a limiting factor in their classification of the seismic facies units on the shelf. Stuart and Caughey (1977) carried out a detailed analysis of the relationship between seismic facies and sedimentary facies using an extensive grid of high-resolution seismic data, and cores, from the northwestern and central portions of the Gulf of Mexico. Sangree et al. (1978) described six seismic facies units for the continental slope, offshore Texas-Louisiana. The seismic
stratigraphic methods used in this study are based on the work of Vail et al. (1977), Sangree and Widmier (1977), Stuart and Caughey (1977), and Sangree et al. (1978).

Seismic stratigraphy has provided an important tool in other studies of the outer shelf and upper slope offshore Texas-Louisiana. Lewis (1984) studied in detail the stratigraphy of the Galveston South Addition lease area. Lewis (1984) mapped an area that overlaps with the southeastern quadrant of this study area, and covered a large part of what is presented in this project as the Brazos shelf-margin delta. Lewis' study also overlapped the southwestern portion of an area of the outer shelf and upper slope studied by Sarzalejo (1993).

Sarzalejo (1993) studied the downdip extension of the Trinity/Sabine incised valley system, and the western Louisiana shelf-margin and upper-slope deposits. The western Louisiana shelf-margin delta progrades into the southeastern part of this study area, from the northeast (Sarzalejo, 1993). These shelf-margin deltas were previously identified by a regional study of the northern Gulf of Mexico that included high-resolution seismic data on the Texas outer shelf and upper slope by Berryhill et al. (1986). The updip portion of the Trinity/Sabine incised-valley system, located on the inner shelf, was mapped in detail by Thomas (1990).

The four previous studies, listed above, used seismic facies analysis, and core data, to interpret sedimentary facies. This study adds further detail to the shelf margin deltas mapped by Berryhill et al. (1986). It also is based on a data set comprising 4,900 km of high-resolution seismic data that includes the inner and outer portions of the continental shelf, and the upper
slope. The areal extent of the seismic data coverage is important since it has allowed us to understand the link between the fluvial systems, the shelf-margin deltas, and the turbidites of the upper slope.

2.2 Data Acquisition Methods

In general, the inner and outer shelf data were recorded with an EG&G Uniboom. On the upper slope, due to the increased water depth, the data were acquired using a 15 cubic inch SSI water gun. A regional dip line, and regional strike line on the outer shelf, were recorded with a 50 cubic inch SSI Generator Injector gun. The Uniboom and water gun data were recorded using an Elics, PC based, Delph 1 acquisition system. These data were sampled at 1/8 ms, and 1/6 ms sample intervals. The 50 cubic inch GI gun data were recorded using an Elics, PC based, Delph 2 acquisition system. These data were sampled at 1/4 ms sample intervals. Examples of these data, and the seismic facies units used in this study are presented in Chapter 3.

Positioning accuracy for the seismic lines varied from 40 m when using Loran C, to 15 m when using GPS (Global Positioning System), and 3 m for the experimental 3-D survey, when DGPS (Differential GPS) was used (Chapter 5).
2.3 Lithostratigraphy

Lithostratigraphic control, or ground truth, is provided by 140 petroleum industry offshore platform boring descriptions, cores from 10 of these platform borings, down-hole gamma-ray logs from 23 of these boreholes, lithologic descriptions from 15 cores from Shell Development Company's Eureka Project on the upper slope, and descriptions from 3 cores from the upper slope (acquired by Exxon, Standard Oil Co. of California, and Mobil) published by Woodbury et al. (1978). The borehole and core data are listed in Appendix 1.

The lithologic data are tied to the seismic data using an average velocity of 1525 m/sec. These conversions are based on check-shot surveys from three 300 m core holes from the upper slope. These surveys were carried out by Exxon and gave average velocities of 1504 m/sec, 1525 m/sec, and 1547 m/sec down to 150 m. Average velocities down to 300 m increased to 1592 m/sec, 1567 m/sec, and 1626 m/sec, respectively, however these depths are below the level of investigation of this study. The velocity data are included in Appendix 2.

Although it is recognized that the true velocities will vary from the value of 1525 m/sec, remarkably good correlation between lithofacies boundaries and changes in seismic character occurs using this approximation. Very similar conversions have been used by other authors (e.g., Sidner et al., 1978, used 1525 m/sec for the first 100 m of sedimentary section, Thomas, 1990, used 1550 m/sec, Sarzalejo, 1993, used 1524 m/sec, Sydow and Roberts, 1994, used 1500 m/sec). Sydow and Roberts (1994)
Figure 2-7a Typical marine clay from the boring B146 at 46.8 m. This boring is discussed in Chapter 3, see Fig 3-1b for the core location.

Figure 2-7b Typical marine silts from the boring B152 at 73.6 m.
Figure 2-8a Fine-grained sands from the boring B151 at 68.4 m.

Figure 2-8b Medium-grained sands from the boring B150 at 87.3 m. This boring is discussed in Chapter 3, see Fig 3-1b for the core location.
Figure 2-9a Coarse-grained sands from the boring B149 at 69 m.

Figure 2-9b Gravels from the boring B147 at 35.5 m.
Figure 2-10a Red clays typical of Brazos River influence from the boring B150 at 81.1 m.

Figure 2-10b Red, Brazos River clays interbedded with marine clay from the boring G147 at 8.8 m.
Changes in stiffness are often documented in the descriptions of the geotechnical properties of the platform borings. The changes in stiffness usually occur across exposure surfaces, and can be used to separate the Holocene sediments from the underlying Pleistocene strata (Winker, 1979, Thomas, 1990). Subaerial exposure surfaces are often characterized by a mottled green and red color, due to oxidation, and by the occurrence of calcareous and ferrous nodules (Winker, 1979, Van Siclen, 1991). Examples of these features are shown in Figures 2-11a and 2-11b. Organic-rich clays, characteristic of fresh-water and low-salinity marshes, are shown in Figures 2-12a and 2-12b.

Fine-grained sands occur on the upper slope in the southwestern and southeastern corners of the study area. Their occurrence has been reported in the upper-slope cores of Shell and Exxon, where they have been described as turbidites. The spectrum of lithofacies in the study area ranges from low salinity marsh deposits to turbidites of the upper slope. There is no direct evidence for carbonates in the study area, no doubt due to the dominant influence of the terrigenous sediments from the Brazos and Colorado Rivers.

No attempt has been made to carry out any mineralogical analyses of the sediments in the study area. Only very general relationships will be noted, such as relating the presence of large, angular feldspars within coarse-grained sands and gravels to the Colorado River. Detailed analyses of the mineralogy and distribution of the sediments on the shelf were carried out by van Andel (1960), and Curray (1960). Their maps are discussed in Chapter 4, during the discussion and interpretation of the results of this study.
Figure 2-11a Change of stiffness shown on platform boring logs is often indicative of the older, Pleistocene clays. The reference to numerous calcareous nodules at 216 ft (sample shown in accompanying Figure 2-11b) is indicative of subaerial exposure.

Figure 2-11b Mottled green and red color due to oxidation, note also the presence of calcareous and ferrous nodules
very stiff to hard olive grey clay
- shell fragments 39-47
- silts 46, 56
- shell fragments 65-76
- sand 66
- siltstones and calcareous nodules at 86
- silty 95-107 greenish grey 97
- shell fragments 115
- brownish yellow below 125
- silty 126
olive grey silty fine sand
- brownish yellow to 137
- greenish grey clay and calcareous nodules at 136
- fine sand 145
- clay and sand and brown banding 155
- shell fragments 166
- sandy silt 176
grey fine sand
- silty at 185
- silt at 198
- clay and shell fragments 205
hard greenish grey clay
- sandy clay and numerous calcareous nodules at 216
- brown fine sand 225
- silty 235-260
- mottled with brown at 236
- grey, mottled with greenish grey calcareous and ferrous nodules 246
- silt pockets and partings, ferrous streaks and yellowish red banding 256
- yellowish red with brown silts, calcareous nodules and light grey and calcareous at 266
- olive grey mottled with yellowish red and calcareous nodules at 276
greenish grey fine sand
- olive grey and silt and a few clay pockets and seams at 296
- clay seams and shell fragments below 305
- claystones at 315
hard olive grey clay
- with silty fine sand seams and shell fragments 326
- greenish grey with silt pockets 341
- silt pockets and marbled with yellowish red at 349
Figure 2-12a Organic rich clay typical of fresh-water and low-salinity marshes with the Gastropod *Littorina irrorata* from the Brazos B151 boring at 5.6 (m). The organic material was dated by the radiocarbon method and its significance is discussed under chronostratigraphy.

Figure 2-12b Organic rich clay, showing preserved rootlet in the a sample from the boring B152 at 6.0 m. These organic-rich marsh deposits are overlain by Holocene marine muds.
2.4 Biostratigraphy

The biostratigraphic data for this project have come from four sources, i.e.:

i) Published results from cores taken by Exxon, Standard Oil Co. of California, Gulf, and Mobil (Woodbury et al., 1978). Paleobathymetry and evidence of shallow water fauna displaced into deeper environments were reported in this paper.

ii) Shell Development Company's biostratigraphic reports from their Eureka Project cores (Anderson and Parrott, 1969). These are very detailed reports. Depositional environments, paleotemperature indicators, and the presence or absence of age diagnostic species are reported. The cored intervals are placed within the major glacial and interglacial stages.

iii) Results from the analyses of two cores that formed part of the Master's Thesis data set of Lynette Holdford, a graduate student of Dr. Martin Lagoe, at The University of Texas at Austin. Cluster sample groups were defined statistically for the benthic foraminifera. Paleobathymetry was defined using these cluster groups together with plots of percent planktic foraminifera, and percent agglutinated foraminifera.

iv) Macrofossils recovered from the platform borings discussed above.
The macrofossils *Littorina irrorata*, *Rangea cuneata*, and *Mercenaria campechiensis texana* provide direct evidence for low-salinity marshes, bayhead, and open-bay environments, respectively (Parker, 1960, Andrews, 1981).

The Gulf Coastal plains, and the Gulf of Mexico, have been the study areas for some of the pioneering work in Quaternary benthic and planktic biostratigraphy. Early work on the Louisiana coast set the stage for biostratigraphic analysis of the Louisiana chenier plains (Byrne et al., 1959). Byrne et al. (1959) used benthic foraminifera, molluscs, ostracods, and diatoms to define 4 faunal zones and 7 sedimentary environments.

Phleger (1960), studied in detail the foraminiferal faunas from seven Gulf of Mexico environments that ranged in water depth from marine marshes to the lower continental slope and deep sea. Ericson and Wollin (1968) provided a critical link between the frequency of occurrence of the foraminifer *Globorotalia menardii*, and climate. They correlated the Yarmouth interglacial, Illinoian glacial, Sangamon interglacial, Wisconsin glacial, and postglacial stages to a series of climatic zones labeled T through Z.

Kennett and Huddlestun (1972) analyzed the planktic faunal assemblages associated with the Ericson and Wollin Zones, and the coiling ratios of *Globorotalia truncatulinoides*. They also recognized the extinction of the planktic foraminifera *Globorotalia menardii flexuosa* as occurring at the X-Y boundary of Ericson and Wollin (1968). This extinction has been
identified by the Shell stratigraphers in their Eureka cores and is an extremely important datum in this study.

Poag and Valentine (1976) used planktic and benthic foraminifera, together with calcareous nanofossils, to document the transgressions and regressions related to glacial eustasy. Other examples of this methodology can be seen applied in the Gulf of Mexico by Poag and Sidner (1976), Beard et al. (1982), and Thunell (1984). The Shell biostratigraphic reports are based on these methodologies, and have provided independent constraints to global sea level. Figure 2-13, from Lamb et al., 1987, summarizes one of the existing correlations between the Pleistocene glacial and interglacial cycles and the biostratigraphic record.

It should be noted that a major difference exists between the number of cold and warm cycles documented by Poag and Valentine (1976) and by Sidner et al. (1978). This has resulted in much older ages being assigned by Sidner et al. (1978) to their reflection B than the age assigned using the Shell biostratigraphy, which is in keeping with that of Poag and Valentine (1976). Sidner et al. (1978) place their reflection B at 320-350,000 ybp due to use of warm and cool cycles constrained below by *Psuedoemiliania lacunosa* (stage 12) and *Globorotalia menardii flexuosa* (stage 5b/5a), above. The Shell biostratigraphy places reflection B from core 67-45 at stage 6 (190,000 ybp). The data set from this project support the zonations of Poag and Valentine (1976). The additional cycles of Sidner et al. (1978) appear to have led to an incorrect correlation with the oxygen isotope curve of Emiliani and Shackleton (1974), and hence incorrect chronology. This will be discussed further in Chapter 4.
Figure 2-13  Glacials and interglacials, and Pleistocene Biostratigraphy (after Lamb et al., 1987)
The other biostratigraphic data used in this project are the detailed paleoenvironmental analyses from two cores that formed part of the data set analyzed by Holdford (1995). Holdford (1995) applied cluster analysis to define faunal similarity between benthic foraminifera and combined these data with the percentage of agglutinated foraminifera and the percentage of planktic foraminifera to analyze of the paleoenvironments recorded in the cores. The paleobathymetric curves, produced from her analyses, provide an independent record of sea-level changes over the time interval of this study.

2.5 Oxygen Isotope Stratigraphy

Higher concentrations of $^{18}$O in sea water during glacial times are the result of the preferential evaporation of the lighter $^{16}$O molecule, and the preferential precipitation of the heavier $^{18}$O molecule (Emiliani, 1955). The lighter $^{16}$O molecules are then trapped within continental and marine ice sheets during glacial periods. The relative proportions of the heavy, $^{18}$O, and the light, $^{16}$O, oxygen molecules in the CaCO$_3$ tests of benthic and planktic foraminifera provide us with an indication of paleotemperatures (Emiliani, 1955) and, by extension, paleo-ice volumes (Shackleton and Opdyke, 1973).

The oxygen isotope record provides us with an indication of paleoclimate, and in this project the SPECMAP curve of Imbrie et al. (1984) is used to provide independent evidence of glacial and interglacial times (Fig. 1-9a). The SPECMAP curve also provides a proxy to sea level in keeping with the concept that a significant part of the oxygen isotope record
represents ice volume (Shackleton and Opdyke, 1973, Shackleton, 1987). It should be noted, however, that the relative effects of temperature versus ice-volume, on the $\delta^{18}$O record, is still debatable (Mix, 1987). In this project, the SPECMAP curve is used as an accurate indicator of the timing, and frequency of eustatic change, but only as a guide to the absolute magnitude of these changes (Fig. 1-9b).

Oxygen isotope records in the Gulf of Mexico also record the influx of $^{16}$O rich meltwater during the retreat of the Laurentide ice sheet. This $^{16}$O rich meltwater flowed down the Mississippi River between 14,000-12,000 ybp (Leventer et al., 1982) and resulted in a negative spike in the oxygen isotope records of the Gulf of Mexico (Kennett and Shackleton, 1975, Leventer et al., 1982). This southerly flow of glacial meltwater was short lived. At about 12,000 ybp the outflow of meltwater shifted to the Atlantic Ocean, through the Champlain-Hudson Valley (Teller, 1987). The drainage shifted northeastwards at 9,500 ybp to the Gulf of St. Lawrence, via the St. Lawrence and Ottawa Rivers, and by 8,000 ybp a northerly drainage system was established to Hudson Bay (Teller, 1987).

The southerly flow of meltwater, along the Mississippi River, dated by Leventer et al. (1982) at 14,000-12,000 ybp, provides a distinctive chronomarker for the Gulf of Mexico (Fig. 2-14). Older meltwater spikes have been identified in the Gulf of Mexico by Williams (1984), Trainor and Williams (1990), and Joyce et al. (1993). Williams (1984) identified a meltwater event associated with oxygen isotope Stage 5 (Fig. 2-15). Meltwater events that immediately precede isotope Stage 1, and Stage 5, are recorded in the study area.
Figure 2-14  The Mississippi meltwater pulse of Leventer et al. (1982). The second major discharge is dated at 12,000-14,000 ybp. The negative pulse in the core B146 from the study area has been correlated to this second major meltwater pulse.
Figure 2-16 The isotope record for the core TR126-23 from the Gulf of Mexico from Williams (1984). Williams has interpreted meltwater (mw) pulses for both Stage 1 and Stage 5.
Samples were processed from the B146 platform boring for this study, and an isotope record generated over the entire cored interval using the benthic foraminifera *Quinqueloculina sp.* The upper 7.5 m (25 ft) were also analyzed using the planktic foraminifera *Globigerinoides ruber* in an effort to record the Stage 2 to Stage 1 meltwater pulse in both the benthics and the planktics.

The samples were soaked in Calgon solution for 1 day to dissociate the clays and then washed through a 63 micron sieve. The samples were then dried at temperatures less than 50 deg. C. and sieved to allow the picking of foraminifera between 300 and 355 micron sieve sizes. Only if 15-20 monospecific foraminifera could not be found were sizes greater than 355 microns considered. The foraminifera picked (whenever present) were the planktics *Globigerinoides ruber* and *Globigerinoides sacculifer*, and the benthics *Elphidium sp.* and *Quinqueloculina sp.*

Initially, the samples were examined without further preparation, but problems with ammonia contamination resulted in a preparation sequence that included ultrasonification of picked foraminifera in a 3% solution of hydrogen peroxide. The samples were then rinsed and placed under vacuum where they were kept at 250 deg. C. for 1 hour. This sequence successfully removed spurious spikes that previously occurred during both the carbon and oxygen analyses. The isotope values are tabulated in Appendix 3.
2.6 Chronostratigraphy

Geochronologic control over the time period of this study, i.e., over the last 190,000 years, was achieved using a combination of the methods and events, listed below. The ability to understand the response of the fluvial/delta systems to glacio-eustasy hinges on accurate chronocorrelation. This statement follows from the fact that the objective is to place the evolution of these systems within the temporal framework of an independent sea-level proxy, in this case the oxygen isotope record.

i) Correlation of an isotope record, generated within the study area, to the orbitally tuned SPECMAP record, of Imbrie et al. (1984). The relationship between the cycles of solar insolation, the Milankovitch cycles, climatic cycles, and the oxygen isotope record allows for the creation of an isotope record that has been tuned to the solar insolation cycles (Broecker, 1966, Hays et al., 1976, Imbrie et al., 1984). Correlation of other records to this orbitally tuned record provides a means of dating the stratigraphic section (Williams et al., 1988).

ii) Correlation with the regional seismic reflection associated with the Stage 5e condensed section (Thomas, 1990). Thomas (1990) tied a regional seismic reflector, offshore Texas, with the base of the Stage 5e Ingleside Barrier, and also with a similar reflection mapped in Louisiana by Coleman and Roberts (1990). The age of this condensed section would be similar to the Stage 5e coral terraces from Barbados,
dated at 125,000 ybp (Fairbanks and Matthews, 1978, Bard et al., 1990). The top of this condensed section is the Stage 5e maximum flooding surface (mfs).

iii) The extinction of *Globorotalia menardii flexuosa* (Kennett and Huddleston, 1972, Poag and Valentine, 1976). This extinction is currently dated at 85,000 ybp (Kohl, 1986) (Fig. 2-16). Dates are based on Tephrochronology (Ledbetter, 1984). The disappearance of the planktic foraminifer *Globorotalia menardii flexuosa* was documented in the Shell Eureka Project cores from the upper slope (Anderson and Parrott, 1969). This extinction level occurs at the boundary between Ericson's X and Y zones, i.e., at the transition from oxygen isotope Stage 5b to Stage 5a.

iv) Bulk radiocarbon (\(^{14}\text{C}\)) dating of carbonate shell material, and organic carbon from marsh deposits. This method provides accurate dates for ages less than 30,000 ybp (Bard et al., 1990). Dates outside of the range of the method provide a minimum age that eliminates erroneous correlation with younger sediments. The radiocarbon dates obtained in this study are shown in Appendix 4.

v) The dating of the last glacial maximum at 18,000 ybp (Fairbanks, 1989). The erosional surface associated with the last sea-level minimum has been used to constrain the age of sediments below this surface to older than 18,000 ybp, and above this surface, to younger
Figure 2-16 Stratigraphic chart, from Kohl (1986). The correlation of an Early Wisconsin Glacial with isotope Stage 6 is supported by this study.
than 18,000 ybp.

vi) The Mississippi meltwater pulse of Kennett and Shackleton (1972) and Leventer et al. (1982) provides a chrono-marker that spans the period from 14,000-12,000 ybp (Leventer et al., 1982) (Fig. 2-14). This negative excursion in the $\delta^{18}$O record was identified in the B146 oxygen isotope record, and was used to date a major transgressive event in the study area.
CHAPTER 3 - DATA AND RESULTS

Introduction

In this chapter the data that form the basis of this project are presented and discussed. Nine seismic facies units identified within the study area are defined in terms of their external form and reflection configuration, following Vail et al. (1977). The lithologic data are presented with the discussion of the seismic facies units, and add support to the interpretation of depositional environments. The lithologic data are either descriptions of petroleum industry platform borings, core data, or down-hole gamma-ray logs.

The biostratigraphic data, oxygen isotope data, and geochronologic data are presented in separate subsections. The integrated interpretation of the data set is reserved for Chapter 4.

Figures 3-1a and 3-1b show the locations of the seismic lines, platform boring descriptions, gamma-ray logs, and the cores discussed in this chapter.

3.1 Seismic Facies Units and their Lithology

The seismic data have provided the means to identify and map distinct seismic facies units over the study area. The approach is based on the principles of seismic stratigraphy (Vail et al., 1977). Nine seismic facies units have been identified. The descriptions follow the formats of Mitchum
Figure 3-1a Location map for seismic lines described in Chapter 3
Figure 3-1b Location map for cores described in Chapter 3
et al. (1977), and Sangree and Widmier (1977), i.e., external form, reflection configuration, modifying term (if applicable), and bounding surfaces.

3.1.1 Seismic Facies Unit 1 (SFU 1)

**Description:** Lobate mound to lobate lens, prograding clinoforms - from sigmoid to oblique tangential. Upper surface often modified by erosion, clinoforms gently downlap or become tangential to, and grade into, the underlying parallel, near-horizontal reflectors of SFU 3 or SFU 4 (Figs. 3-2 and 3-3).

SFU 1 is an areally extensive facies unit and occurs as five distinct lobate mounds. These lobate mounds are labeled 1 through 5, from oldest to youngest, respectively (Figs. 3-4a, 4b, 4c, 4d, and 4e). As the mounds prograde from the inner to the outer continental shelf the angle of the clinoforms increase from 0.1 to 2.5 degrees (Figs. 3-2 and 3-3). The shape of the mounds also change from elongate in a dip direction on the inner shelf, to more strike aligned on the outer shelf (Figs. 3-4a, 4b, 4c, and 4d).

Strike-oriented seismic lines demonstrate that the lobate mounds 1 through 4 show almost continuous lateral migration (Fig. 3-5). Mound 5, the youngest, shows little evidence for lateral accretion or migration. Mound 5 appears to have been physically restricted during its development by older mounds on both sides (Figs. 3-4d and 4e).

The clinoforms of all of these lobate mounds show a sigmoid to oblique geometry, indicative of aggradation and progradation, to more rapid progradation (Mitchum et al., 1977). Mitchum et al. (1977) attribute the
Figure 3-2  Seismic line R91-4 shows the progradational clinoforms associated with SFU 1. The lower boundary of SFU 1 is a downlap surface of regional extent. The upper boundary of SFU 1 is often erosional. In this figure, chaotic reflections are present directly above the erosional boundary. This chaotic character is described later as SFU 7a. (Sequence boundaries (SB) and maximum flooding surface (mfs) are shown here for later reference only). (10 ms = 7.625 m; VE = X37)
Seismic line R91-8 is a dip line showing the oblique tangential clinoforms that can occur within SFU 1. The Stage 2 to Stage 1 ravinement surface is shown for future reference. (10 ms = 7.625 m; VE = X26)
Figure 3-4a  One of the five mounds that define the areal distribution of SFU 1. The other mounds are shown in Figs. 3-4b, 4c, 4d, and 4e. Refer to Fig. 3-1a for examples of SFU 1.
Figure 3-4b  The second of five lobate mounds that characterize SFU 1. The other mounds are shown in Figs. 3-4a, 4c, 4d and 4e. Refer to Fig. 3-1a for seismic lines that illustrate SFU 1.
Figure 3-4c  The third of five mounds that delineate SFU 1. See also Figs. 3-4a, 4b, 4d and 4e. Seismic lines that illustrate SFU 1 are shown in Fig. 3-1a.
Figure 3-4d Areal distribution of SFU 1. SFU 1 occurs as 5 lobate mounds the other four mounds are shown in figures Figure 3-4a, 4b, 4c, and 4e.
Figure 3-4e Areal distribution of SFU 1. SFU 1 occurs as 5 lobate mounds, the other four mounds are shown in figures 3-4a, 4b, 4c, and 4d.
Figure 3-5 Seismic line R90-5 demonstrates lateral migration within SFU 1. Also shown is the mounded form of SFU 8.
sigmoid geometry to possible combinations of low sediment supply, rapid subsidence, and/or rapid sea-level rise. The authors attribute an oblique geometry to possible combinations of high sediment supply, minimal subsidence, and a sea-level stillstand.

The lobate mounds prograde over near-horizontal parallel reflectors (Fig. 3-6). The prograding clinoforms can also be seen to grade laterally, as well as down dip, into near-horizontal, parallel reflectors (Figs. 3-5 and 3-7). The lower boundary of SFU 1 is conformable with the near-horizontal, parallel reflectors.

The upper boundary is always erosional. Clinoforms show updip truncation against both irregular, and planar, erosional surfaces (Fig. 3-8). The angular nature of the truncations indicates that SFU 1 is always bounded above by an erosional unconformity.

The lithology within SFU 1 is consistently identified from platform boring descriptions, and cores, as grey clay to silty clay, with laminae and small lenses of silt and very fine sand (Figs. 3-9a and 3-9b). Fine organics are often present, as are small shell fragments, and occasional silty, mud-filled burrows. The fine grain size, grey color, nearshore to outer shelf foraminiferal faunas, and occasional burrows, all indicate a marine origin. A down-hole gamma-ray log for platform boring G 73 is shown in Figure 3-10. A clay-rich interval with silty laminations can be inferred from the character of the gamma-ray log. Figure 3-10 shows also the seismic character of SFU 1 over the logged interval.

The seismic data and lithologic data indicate that SFU 1 is representative of the aggradational to progradational portions of a
Figure 3-6 Seismic line G260X demonstrates downlap of SFU 1 onto near-horizontal, parallel reflectors. Also shown is the internal configuration of SFU 9.
Figure 3-7 Seismic line R91-4 demonstrates the downdip transition, from prograding clinoforms, to horizontal reflectors.
Figure 3-8 Seismic line R92-26 demonstrates truncation of clinoforms against both irregular, and planar, erosional surfaces.
Figure 3-9a Fugro-McClelland boring description for G148 illustrating the lithofacies associated with SFU 1

Figure 3-9b Core photograph of G 148 at 10.5 m, within SFU 1
Figure 3-10 Down-hole gamma-ray log for G73 and the seismic character over the logged interval.
prograding delta. SFU 1 represents distal-delta front to prodelta deposits, and the lobate mounds of Figures 3-4a to 3-4e are interpreted to represent distinct, aggradational to progradational delta lobes. Lobes 1 through 4 are interpreted to represent delta lobes deposited by the late Quaternary Brazos fluvial/deltaic system, and lobe 5 by the late Quaternary Colorado fluvial/deltaic system.

The largest Brazos delta lobe, labeled 1 in Figure 3-4a, is also the thinnest Brazos lobe. This lobe covers approximately 5,000 sq km and has a maximum thickness of approximately 30 m. In contrast, lobe 4 (Fig. 3-4d) covers an area of approximately 2,000 sq km but reaches a maximum thickness of some 80 m. Overall sediment volumes of both lobes appear similar, with the sediments becoming more areally restricted as the delta lobes prograde into deeper water and onto steeper shelf gradients. The gradient of today's inner shelf is on the order of 0.05 degrees, changing to the order of 0.2 degrees towards the outer shelf. The change in lobe shape, from the dip-elongate lobe 1 (Fig. 3-4a) to the more strike-elongate lobe 4 (Fig. 3-4d), is attributed to the change in accommodation from the inner to the outer shelf.

Indications that sediment supply may not have been much different during the deposition of lobes 1 through 4 have far reaching implications. With near-constant sediment supply, a change in progradational clinoform pattern from sigmoid to oblique within the course of deposition of a single lobe can only be due to eustasy.

Changes in rates of subsidence can be ruled out as the controlling factor in clinoform geometry since the subsidence rates in the study area
increase from the inner to the outer shelf, i.e., they increase in the direction of progradation. Under conditions of constant sediment supply and in the absence of eustatic changes, the effect of subsidence would be the opposite to what is observed. Instead of a change from aggradation to progradation, we would expect to see a change from aggradation to retrogradation.

In addition, the eustatic change required to generate the observed clinoform configuration is one of a relatively slow continuous rise, followed by a relatively slow continuous fall. These observations have important implications regarding the nature of sea-level fluctuations. Once the seismic stratigraphic interpretations have been placed within a chronostratigraphic framework, the stratal geometry will be compared to an independent sea-level proxy, the SPECMAP oxygen isotope curve of Imbrie et al. (1984).

3.1.2 Seismic Facies Unit 2 (SFU 2)

Description: Thin lens, prograding clinoforms - from hummocky to shingled. Upper surface irregular and erosional, bounded below by parallel, near-horizontal reflections (Figs. 3-11 and 3-12).

SFU 2 occurs as a relatively thin lens with maximum thicknesses of 15 m. Figure 3-13 shows the areal extent of this lens and its projected possible extent towards the southwest. The prograding clinoforms are hummocky to shingled. Mitchum et al. (1977) attribute both hummocky and shingled clinoform configurations to progradation into shallow water.

The lens is underlain by near-horizontal parallel reflectors shown in Figure 3-12, a strike line. The lens appears to thin against these parallel
Figure 3-11 Seismic line R93-42 showing prograding, shingled clinoforms and upper erosional surface of SFU 2
Figure 3.12 Seismic line R90-7 showing parallel, near-horizontal reflectors underlying SFU 2
Figure 3-13 Areal extent of SFU 2, showing its inferred southwesterly orientation
reflectors in the dip direction (Fig. 3-11). Figure 3-11 also shows the erosional nature of the upper bounding surface of this lens.

The distal portions of the lens are characterized by grey, silty clays with silt partings and occasional organics. The gamma-ray response of this lithology is shown in the down-hole logs for B129 and B130 (Fig. 3-14 and 3-1b). The more proximal and central parts of the lens show an overall coarsening upwards pattern on the gamma-ray log for B125 and B77 (Fig. 3-15 and 3-1b). The upper portion of the coarsening upwards sequence is sand rich (Fig. 3-15).

The external form and internal reflection configuration of SFU 2, together with its related lithology and gamma-ray response, indicate that SFU 2 represents a delta lobe prograding into shallow water. The fact that the lobe does not develop the thickness of the Brazos lobes, nor their steeper clinoform angles, is further evidence of progradation into shallow water.

This lens is interpreted to represent a late Quaternary lobe of the lower sediment supply, Colorado fluvial/deltaic system that slowly prograded into shallow water. This is supported by the low clinoform angles and the sandy nature of SFU 2, seen in B77 and B125 (Fig. 3-15). The rate of deltaic progradation probably just kept pace with falling relative sea-level, so that unlike the Brazos delta, the Colorado SFU 2 delta never built out into deep water.

The southwestwards orientation of the delta lobe in Figure 3-13 is in keeping with the southwestwards trend of the fluvial drainage mapped by Berryhill (1986). The southwestwards orientation is in response to the steeper gradients of the south-central Texas shelf.
Figure 3-15 Down-hole gamma-ray logs for B77 and B125
3.1.3 Seismic Facies Unit 3 (SFU 3)

**Description:** Sheet to sheet drape, parallel even to subparallel. Upper surface can be downlap surface for overlying SFU 1 or SFU 2. SFU 3 grades updip, and laterally into SFU 1, or SFU 2 (Figs. 3-7 and 3-16).

SFU 3 occurs over large areas of the continental shelf. SFU 3 underlies SFU 1 and SFU 2. SFU 1 and SFU 2 also grade laterally, and down dip, into SFU 3 (Figs. 3-7 and 3-16).

Lithologically, SFU 3 is often comprised of grey clay, interbedded with occasional silt laminae and shell fragments. Gamma-ray logs for B54 and B124 demonstrate the laminated, clay-rich character of the sediments within SFU 3 (Fig. 3-17). Figure 3-18a shows a core sample of the grey clay from B148, at 33 m. Grey clays are typical of an open marine environment.

The example of SFU 3 in Figure 3-18a is interpreted to represent the distal prodelta sediments deposited seaward of the prograding delta lobe, i.e., within the fondo environment. However, SFU 3 also occurs within the low energy bays between major deltaic headlands. In Figure 3-18b, the brownish-grey color of the clays from B150, at 35 m, is indicative of an interdeltaic, rather than an open-marine, environment. Figure 3-19 shows the gamma-ray response for B128 and B150 and the corresponding seismic character within this interdeltaic environment.

The top of SFU 3 is often a downlap surface for the overlying prograding delta lobes (Fig. 3-6), and a change from transgression to regression occurs as we move upwards across this boundary. Following periods of major transgression, the top of the time equivalent SFU 3
Figure 3-16 Seismic line R90-5 showing parallel, near-horizontal reflectors of SFU 3 grading laterally into SFU 1. SFU 3 are the distal, lower angle equivalents of SFU 1.
Figure 3-17 Down-hole gamma-ray logs for B54 and B124 demonstrate the clays, with silty interbeds, of SFU 3
Figure 3-18a Core sample from B148 at 32.9 m within SFU 3, showing
typical grey marine clays

Figure 3-18b Core sample from B150 at 35.4 m indicative of an interdeltic
setting for SFU 3
Figure 3-18a Core sample from B148 at 32.9 m within SFU 3, showing typical grey marine clays
becomes an important regional marker, i.e., the maximum flooding surface (mfs).

3.1.4 Seismic Facies Unit 4 (SFU 4)

Describeion: Sheet to sheet drape, parallel even. Can be overlain, and underlain by, chaotic reflections (Fig. 3-20).

SFU 4 differs from SFU 3 in that it is restricted to the slope where it can be seen to both overlie, and underlie, chaotic reflections (Fig. 3-20). The reflections within SFU 4 are extremely even and parallel.

An example of the typical lithology within SFU 4 comes from Shell core 13, and consists of greenish-grey clay with lighter laminations, common foraminifera and rare shell fragments. Occasional echinoid spine fragments also occur. The gamma-ray logs that accompany the lithologic descriptions show very little character, i.e., the clay interval is fairly massive.

Sangree and Widmier (1977) interpret this facies as hemipelagic clays that, on occasion, grade into the fondoform reflections of an updip progradational unit. Figure 3-21 is an example of this updip gradation of SFU 4 into SFU 3 and SFU 1. A hemipelagic interpretation for SFU 4 is supported by the core data from the upper slope. Occasionally, fine-grained turbidites occur interbedded within this predominant clay interval. These fine-grained turbidites likely are related to the transport of fine sands over the clinoforms of the time equivalent progradational units. However, the turbidites are only a very minor component of SFU 4.
Figure 3-20 Seismic line G10Y illustrating the parallel, even, seismic reflections that characterize SFU 4, and their relationship to underlying, and overlying chaotic reflections.
Large-scale slumps, slides, and turbidites do occur on the shelf but not within SFU 4. These features cut across, and erode into, SFU 4 (Fig. 3-20) and are described below.

3.1.5 Seismic Facies Unit 5 (SFU 5)

Description: Basin fill, chaotic. Lower surface erosional, underlain and overlain by SFU 4 (Fig. 3-20).

SFU 5 can be seen in the southeastern corner of the study area where it occurs as a chaotic fill that scours into the underlying parallel reflections of SFU 4 (Fig. 3-20). The chaotic reflections in Figure 3-20 reach thicknesses of 150 m, and are overlain by parallel reflectors of SFU 4.

Shell core 7 indicates that SFU 5 is comprised predominantly of grey-green clays, although intervals of mottled, reddish-grey clay occur. Contorted intervals in the core have been described as due to either the coring process or to burrows, although in one case it has been described as a possible slump structure. A 4 m core in the basal section of these chaotic reflectors penetrated medium to dark grey clay described as "silty to very finely sandy throughout" (Anderson and Parrott, 1969).

This particular type of seismic facies unit has been attributed to mass-transport processes by Sangree and Widmier (1977). These chaotic reflections were previously described by Lewis (1984) and Sarzalejo (1993), who attributed them to mass-transport deposition following shelf-edge instability and collapse. The Shell core data support these interpretations. These mass-transport deposits are ponded in upper-slope salt withdrawal
basins (Satterfield and Behrens, 1990, Sarzalejo, 1993). The approximate areal extent of these chaotic events is shown in Figure 3-22.

3.1.6 Seismic Facies Unit 6 (SFU 6)

Description: Mounded fan, chaotic. Lower surface erosional, underlain by SFU 4 (Fig. 3-23).

Figure 3-23 shows the chaotic seismic reflection character of SFU 6. SFU 6 can be seen to erode into the underlying parallel, near-horizontal, reflections of SFU 4. Most of this unit lies outside of the study area and the areal distribution of SFU 6, shown in Figure 3-22, is from previous work by Lehner (1969), Tatum (1977), Woodbury et al. (1978), and Rothwell et al. (1991).

The chaotic reflections seen in Figure 3-23 reach thicknesses of at least 100 m. Core data published by Woodbury et al. (1978) indicate that a minimum of 60 m of the chaotic reflections are comprised of contorted beds with intercalated sands. These authors have interpreted these as displaced sediments, and Lehner (1969) named the deposit the East Breaks Slide.

One difference between SFU 6, in the southwestern corner of the study area, and SFU 5, in the southeastern corner of the study area, is the sandier nature of SFU 6. As we will see later, this is due to its relationship to the sandy Colorado fluvial/deltaic system. The other difference is its external form. Instead of the basin-fill geometry of SFU 5, SFU 6 has developed an elongate, mounded external geometry.
Figure 3-22 Areal distribution of chaotic reflections of SFU 5 and SFU 6
Figure 3.23 Seismic line R92-35 showing the chaotic reflection character of SFU 6.
The difference in shape is due to the fact that SFU 6 is less affected by salt withdrawal. Active salt tectonics in the southeastern corner of the study area is due to the influence of the very high sediment supply western Louisiana fluvial/deltaic system, studied by Sarzalejo (1993). In the western Louisiana area, sediments are more likely to become trapped within bowl-shaped minibasins on the upper slope.

3.1.7 Seismic Facies Units 7a and 7b (SFU 7a and SFU 7b)

Description: Channel fill, chaotic fill (SFU 7a) to onlap fill (SFU 7b). Lower surface is highly erosional, can be overlain by SFU 3 or the elongate mounds of SFU 8 (Figs. 3-24 a and 24b, and 3-5).

Both SFU 7a and SFU 7b are characterized by scour and incision and are described as subsets of SFU 7. The channel fill can be either chaotic (SFU 7a) or consist of laminated, onlapping reflectors (SFU 7b), (Figs. 3-24a and 24b). The onlap fill in Figure 3-24b is associated with a larger channel complex of type SFU 7a.

Figure 3-25 shows the areal distribution of SFU 7. The generally dip-oriented ribbons, that bracket the occurrence of these incised channel features, run from the innermost shelf to the outermost shelf. The underlying and surrounding facies, incised by SFU 7, are either a prograding delta lobe (i.e., SFU 1 or SFU 2) or the clays of SFU 3. As shown in Figures 3-24a and 3-5, the overlying facies is either SFU 3, or the mounded chaotic reflectors of SFU 8.
Figure 3-24a Seismic line R90-7 (westernmost end of line) showing the chaotic channel fill of SFU 7a overlain by SFU 3
Figure 3-24b: Seismic line R90-9 showing the laminated, onlap channel fill of SFU 7b.
Figure 3-25 Map showing the areal distribution of SFU 7
Platform boring descriptions, gamma-ray logs, and cores show that the chaotic reflections of SFU 7a consist mainly of medium to coarse-grained sands and gravels. Figure 3-26a (location shown in Fig. 3-1b) is an example of SFU 7a and its lithology as described in B148 (Fig. 3-26b). The channel fill can be seen to be predominantly sand with a coarse basal-lag.

A gamma-ray log from B 51, down dip of B148, is shown in Figure 3-27 (see Fig. 3-1b for core locations). The sharp-based, blocky, gamma-ray character typical of sandy channel fill is evident in this log and a gravel seam was recorded between 17.2-17.4 m (56.5-57 ft) (Fig. 3-27). The basal, coarse-grained deposits are shown in Figure 3-28a, from boring A24. This sample shows gravels at 15.5 -16.2 m within the same channel shown in Figure 3-25. The angular, gravel sized quartz and feldspars in the sample indicate a Colorado River origin for these deposits. The Colorado River drainage includes the igneous and metamorphic province of the Llano Uplift, a source of coarse-grained sediments rich in quartz and poorly-weathered feldspars. The significance of the large mollusc fragments, seen in Figure 3-28a, will be discussed in Chapter 4.

Figure 3-28b shows a core from B148 within the upper part of the channel at 9.5 m. The reddish-brown color of the clays, and the high organic content indicate Colorado River influence. Above these reddish-brown clays, the uppermost portion of the valley fill is made up of grey clays with occasional burrows and shell fragments, and corresponds to the onlap fill of SFU 7b. Figures 3-29a and 29b illustrate the lithology and seismic character of this facies.
Brazos 148

0
very soft to soft brown silty clay
brown silty fine sand
soft to firm olive grey clay
olive grey silty fine sand
light brown fine to medium sand
-coarse at 57

50
stiff to very stiff olive grey clay
-silt pockets 119
-siltstones 68
-shell fragments below 118

100
olive grey silty fine to fine sand
-silt 158
-silt 168
-silt below 166
-clay seams and shell fragments 197

150
very stiff olive grey clay
-loculated with silt pockets at 208
-silt seams at 218
-shell fragments 228
-silt seams 237-249
-sand silt pockets and silt partings 238
-sandy silt pockets and shell fragments 268
-silt partings 278

200
-green fine sand with silt
-shell fragments 288
-clay pockets 299

WATER DEPTH 123'
TOTAL LENGTH 300'

Figure 3-26b The lithologic description for B148
Figure 3-27 Down-hole gamma-ray log for B51 shows the sharp-based blocky character typical of a channel fill.
Figure 3-28a Sample of gravels, rich in quartz, feldspars, and large shell fragments. Sample from A24, at 15.5-16.2 m

Figure 3-28b Core from B148, at 9.5 m, the reddish-brown, organic-rich clays indicate Colorado River influence
Figure 3-29a Laminated onlap fill in upper portion of channel as shown on seismic line R90-5, representative of SFU 7b

Figure 3-29b Grey clays, with occasional silt-filled burrow, shown in core sample from B 148, taken at 6.7 m, within SFU 7b
SFU 7 is interpreted to represent the channel facies infilling major incised valleys that cut across the entire continental shelf during sea-level lowstands. The chaotic fill (SFU 7a) characterizes most of the Colorado incised valley system and consists of coarse, medium, and fine sands and gravels. This character can be traced all the way to the canyons feeding the East Breaks Slide (Fig. 3-25 and 3-30).

The laminated, onlap fill (SFU 7b) is characteristic more of the abandoned portions of the Brazos incised valley system (Fig. 3-24b) than of the Colorado incised valley system. Although there are no platform borings within the laminated Brazos fill, comparison to the laminated fill present in the Colorado valley system indicates that SFU 7b, in the Brazos area, should be predominantly clay (Fig. 3-29a and 3-29b). In fact, coherent, laminated seismic reflections within the study area are always associated with clays, often with finely laminated silts and fine sands.

As sea-level falls, and the fluvial/deltaic systems prograde, the fluvial channels incise in response to the fall in base level. A continuous down-dip progression of this process results in SFU 7 down cutting into the prograding prodelta facies, and the deeper marine facies of SFU 3. However, whenever major river capture occurs, the fluvial system makes a dramatic lateral jump and flows into the interdeltaic bays (Figs. 3-13 and 3-25). The result is that SFU 7 cuts into the shallow water, interdeltaic facies of SFU 3 (Fig. 3-18b). The differences in water depth and lithofacies for these two members of SFU 3 were discussed above.

The spatial and vertical relationships between the seismic facies units, therefore, allow us to document periods of major river capture, and
Figure 3-30 Seismic line R9-41, a strike line on the outermost shelf, shows the chaotic fill of the feeder channels for the East Breaks Slide.
accompanying delta-lobe abandonment. River capture during sea-level fall is interpreted to be the cause of the shift of the Colorado River from a more southwesterly position, during the deposition of SFU 2 (Fig. 3-13), to a southerly orientation at the time of incision across the shelf (Fig. 3-25).

The lower sediment supply of the Colorado River, in contrast to the Brazos River, has made it more susceptible to capture in the upper reaches of its trunk stream. The easterly change in direction of the Brazos River (Fig. 3-25) is also interpreted to be due to river capture during a time of falling sea-level. However, the higher sediment supply Brazos River had already built its delta to the outer shelf before river capture, and delta-lobe abandonment, occurred.

The chronology and significance of these events are discussed in Chapter 4, as is evidence for lobe shifting within the Brazos fluvial/deltaic system that is driven by marine transgressions, rather than by river capture during sea-level lowering.

3.1.8 Seismic Facies Unit 8 (SFU 8)

Description: Elongate mound to thin lens, from predominantly chaotic to prograding clinoforms - hummocky. Bounded below by a planar lower surface, onlapped and overlain by SFU 3 (Figs. 3-5 and 3-31).

SFU 8 occurs as oblate, strike-oriented mounds or thin lenses, generally with chaotic internal configuration (Figs. 3-5, 3-31, and 3-32). These units are bounded below by parallel, near-horizontal reflectors (Fig. 3-31), or by a relatively planar surface with occasional channels (Fig. 3-5).
Figure 3-31 Seismic line R91-8, a dip line across the mounded form of SFU 8
Figure 3-32 Areal distribution of SFU 8
Platform boring G93 identifies the interval shown in Figure 3-5 as olive grey, fine sand with clay (Fig. 3-33). This interval also contains wood and shell fragments. The boring B76 and its gamma-ray log show the fine sands present in a number of the borings through a thin sandy lens in the eastern part of the study area (Fig. 3-34).

The oblate mound shown in strike view in Figure 3-5 is interpreted to represent a reworked, transgressive sand body. The mound sits on the planar ravinement surface, the result of modification of a previously subaerially exposed surface by transgressive erosion. Remnant channels are still visible (Fig. 3-5), and these channels were likely the source for the basal, transgressive sand body. Further transgression has led to the onlap and subsequent burial of the sand body by SFU 3.

The oblate sand bodies in the central and western parts of the study area (Fig. 3-31 and 32) are interpreted as deltas that are stepping landwards in response to marine transgression. The basal contact of these sands is erosional, and represents short periods of regression during the overall transgression (Fig. 3-34). The result is stacked channels that are separated by a ravinement surface (Fig. 3-35).

SFU 8 is interpreted to represent both overstepped, basal, transgressive sand bodies, and back-stepped deltas. The generally chaotic internal configuration of the back-stepped deltas implies that these deltas may be more wave dominated than the regressive, fluvial-dominated deltas of SFU 1, and SFU 2.
Figure 3-33 Boring description for G73, showing the sandy lithology associated with SFU 8. The seismic character over the cored interval also is shown in Fig. 3-30
Figure 3-34 Boring description and down-hole gamma-ray log for B76, located on seismic line R92-28.

The uppermost olive, grey fine sand corresponds to SFU 8
Figure 3-35 Strike line R90-5 showing vertically-stacked channels related to different periods of progradation
3.1.9 Seismic Facies Unit 9 (SFU 9)

**Description:** Elongate lens, prograding clinoforms - oblique to complex sigmoid. Downdip, the clinoforms become tangential, and the fondoforms grade into SFU 4. Updip, the lower portion of SFU 9 can often be seen to onlap SFU 1 (Fig. 3-6). The upper boundary is an erosional surface, parallel to subparallel with the underlying and overlying reflections (Figs. 3-6 and 3-36).

SFU 9 is areally restricted to the outer shelf and upper slope. It pinches out before reaching the inner shelf (Figs. 3-6 and 3-36). Its progradational to aggradational geometry indicates deposition during a period of relative sea-level rise. The upper boundary of SFU 9 is a near-horizontal erosional surface that marks a minor break in sedimentation, i.e., the ravinement diastem of Nummedal and Swift (1987). Shell cores that penetrated SFU 9 characterize its lithology as grey, moderately firm clays, with some fine sand, common fossil fragments, and echinoid spines.

This progradational unit is interpreted to be due to clastic input from fluvial-dominated shelf-margin deltas during a relative rise in sea level. The extent of progradation and the complexity of the reflection configuration are dependent on the relative roles of eustasy and sediment supply. A comparison of the reflection geometry shown in Figures 3-6, versus the geometry shown in Figure 3-36, illustrates a significant difference in sediment supply between these two areas.
Figure 3-36 Seismic line R92-24 showing a complex clinoform configuration within SFU 9
3.2 Biostratigraphy

3.2.1 Micropaleontology (Anderson and Parrott, 1969)

The paleontologic report from the Shell Eureka Project cores (Anderson and Parrott, 1969) divided the stratigraphic section into the Upper and Lower Pleistocene. From the youngest to the oldest, Shell's Upper Pleistocene consisted of the Late Wisconsin Glacial, the Peorian Interglacial, the Early Wisconsin Glacial, the Sangamon Interglacial, and the Illinoian Glacial. Shell's Lower Pleistocene consisted of the Yarmouth Interglacial, the Kansan Glacial, the Aftonian Interglacial, and the Nebraskan Glacial. The Upper Pleistocene subdivisions were very important in this study.

A comparison of the ages assigned by various authors to the subdivisions of the Upper Pleistocene shows a lot of variation. For instance, if one focuses on the Sangamon interglacial the following ages exist:

130,000 - 90,000 ybp (Poag and Valentine, 1976)
300,000 - 70,000 ybp (approx.) (Beard et al., 1982)
420,000 - 300,000 ybp (Lamb et al., 1987)

The age assigned to the Sangamon affects the age of the underlying Illinoian, which in turn leads to differing correlation with the oxygen isotope record.

The Sangamon was defined by the Shell stratigraphers as being equivalent to an interval characterized by the first appearance, and an abundance, of Globorotalia menardii flexuosa. The Shell stratigraphers also recognized two zones with Globorotalia menardii flexuosa. The lower zone
being the Sangamon, and the upper zone the Peorian Interglacial. The Shell stratigraphers assigned ages of 355,000 - 135,000 ybp and 115,000 - 80,000 ybp, to the Sangamon and Peorian Interglacials, respectively.

It is interesting that the total range in ages between the first appearance and last appearance of the *Globorotalia menardii flexuosa*, as assigned by Shell, i.e., from 355,000 - 80,000 ybp, is similar to the total range of ages assigned by different authors for the Sangamon, listed above, i.e., from 420,000 - 70,000 ybp. The discrepancies in the age of the Sangamon may be a function of whether the authors correlated the Sangamon to the upper, or lower, occurrence of *Globorotalia menardii flexuosa*, i.e., with Shell's Sangamon, with the younger Peorian, or across both intervals.

The stratigraphic table of Kohl (1986), shown in Figure 2-16, shows good agreement with the Shell biostratigraphy. Kohl (1986) shows Wisconsin divided into Late and Early glacials separated by an Interstadial, that corresponds to the Ericson zone X (approx. 130,000 - 85,000 ybp). The age of Shell's Peorian Interglacial (approx. 115,000 - 80,000 ybp) shows good agreement with this Wisconsin Interstadial of Kohl (1986). Both use the extinction of *Globorotalia menardii flexuosa* to mark the top of this interglacial.

Kohl (1986) correlates the Early Wisconsin Glacial with isotope Stage 6, and places the top of the Sangamon close to the boundary between oxygen isotope Stages 6 and 7 (Fig. 2-16). Kohl (1986) therefore considers the top of the Sangamon to be younger than the Stage 6/7 boundary, i.e., younger than 186,000 ybp. Shell estimated the age for the top of the Sangamon at
135,000 ybp, and placed the Sangamon directly below the Early Wisconsin Glacial. Shell's estimated age is a bit young, but the basic stratigraphic succession agrees well with Kohl (1986) who had the advantage of tephrochronology and isotope stratigraphy for absolute age dating.

The Shell stratigraphy, together with the stratigraphic table from Kohl (1986) and associated oxygen isotope stages have been used in this study. Other correlations, such as Sidner et al. (1978) and Lamb et al. (1987) (Fig. 2-13) that correlate the Sangamon with isotope Stages 9 and 11, appear to have included an additional glacial and interglacial cycle within the Wisconsin. These additional cycles can be seen in a comparison of Figures 2-13 and 2-16.

The data from this study do not support these additional cycles. Figure 3-37 shows the seismic line G300X with the Shell glacial and interglacials from the Eureka Project core 67-45. The Peorian correlates with parallel reflectors that are associated with distal shelf-margin deposits during an interglacial. Both the Early and the Late Wisconsin Glacials, of Shell, correlate with a periods of shelf-margin progradation. The prograding clinoforms are associated with fluvial-dominated shelf-margin deltas. The older shelf-margin deposits are identified by Shell as being made up predominantly of sand deposited in inner neritic to middle neritic water depths.

A Stage 6 age (186,000 - 128,000 ybp) for the Early Wisconsin shelf-margin progradation agrees extremely well with estimates of subsidence rates for the outer shelf. There are approximately 128 m of sediment above
Figure 3-37 Seismic line G300X illustrates tie with Eureka core 67-45
the older shelf-margin seen in Figure 3-37 which equates to subsidence rates of 1 mm/year, if the top of the prograding sequence is taken as end Stage 6, i.e., 128,000 ybp.

Sidner et al. (1978) considered this latter progradation to have occurred between 350,000 - 320,000 ybp, an interval that they correlate with isotope Stage 10. This translates to subsidence rates of 0.4 mm/year, which is less than half the value expected for the outer shelf. In addition, if these were Stage 10 lowstand deposits, there is no evidence for either Stage 8, or Stage 6 lowstands in the study area. Stage 8 appears to have been as severe a glacial as Stage 10, and Stage 6 more severe than Stages 8 or Stage 10 (Figs. 1-9a and 1-9b). It is unlikely that Stage 10 shelf-margin deposits would be present in preference to both Stage 6 and Stage 8. A Stage 6 interpretation for this lower phase of shelf-margin progradation seen in Figure 3-37 is much more consistent with the subsidence history for the area, and conforms with the Shell biostratigraphy and the stratigraphic table of Kohl (1986) (Fig. 2-16).

The upper limit for the Peorian Interglacial also provides a guide to the Stage 5a/5b boundary (Fig. 2-16). Figure 3-37 shows the top Peorian tied to the seismic data. Correlation of this level in an updip direction separates the older Stage 5e, 5d, 5c, and 5b, deposits, below the top Peorian, from the younger Stage 5a, Stage 3, Stage 2, and Stage 1 deposits, above the top Peorian. This is an important distinction since it places the younger shelf-margin delta, shown in Figure 3-38, in either Stage 3, Stage 2, or Stage 1.
Other sources of evidence will place this younger shelf-margin delta in Stage 3, and will also associate it with the Brazos River. However, it is interesting to note that the subsidence history for the area can again be used to rule out an older age for this shelf-margin delta. In this case, an older Stage 5 (128,000 - 71,000 ybp) age would imply at least 57 m of subsidence, at 1 mm/year, since the deposition of the delta and the subsequent rapid rise in sea-level 14,000 ybp (a period of 57,000 years). In Figure 3-38 the topset beds of the shelf-margin delta have been eroded by the process of transgressive erosion. These topsets would have been preserved had the delta enough time to subside by 57 m prior to the transgression. This indicates a relatively young age for this period of shelf-margin progradation.

3.2.2 Micropaleontology (Holdford, 1995)

Figures 3-39 and 3-40 show the paleoenvironmental summaries for core samples from the platform borings B146 and A24 from Holdford (1995). The results of the cluster analysis, percentage of planktic foraminifera, and percentage of agglutinated foraminifera are combined together to produce an interpretation of the paleobathymetry. Figures 3-41 and 3-42 show the paleobathymetric curves for B146 and A24 together with their respective seismic data.

The paleobathymetric curve for B146 shows four phases of deepening (Fig. 3-41). Two of these deepening phases are centered at approximately 67 m (218 ft), and 47 m (154 ft). The other two phases occur at or near the base of the boring, at approximately 92 m (300 ft), and at or near the sea floor. The deepening event at the base of the boring, at 92 m (300 ft), is
Figure 3-39 Paleoenviromental summary for B146 (from Holdford, 1995)
Figure 3-42 Paleobathymetry for platform A24 located on seismic line R92-28. Top of the core is at sea-level and the seismic data and core depths are shown at approximately the same vertical scale.
below the level of investigation of this study. The events at 67 m (218 ft) and 47 m (154 ft) are associated with high amplitude, near parallel, horizontal reflections (Fig. 3-41). The event at 47 m (154 ft) forms the downlap surface for a wedge of reflectors that thin towards the east. The upper surface of this wedge of reflectors correlates with a dramatic change in paleobathymetry at 33 m (108 ft) (Fig. 3-41). Environments change from outer neritic, below this surface, to marginal marine/embayment, immediately above the wedge.

The deepening events at 67 m (218 ft) and 47 m (154 ft) are candidates for maximum flooding surfaces associated with isotope Stage 5e and Stage 3, respectively. The wedge of reflectors seen downlapping the Stage 3 maximum flooding surface is interpreted as a the distal portions of a Stage 3 shelf-margin delta associated with the Colorado River. The upper surface of this delta is interpreted to represent an exposure surface during the Stage 2 lowstand, resulting in the dramatic change in paleo-water depths above and below this surface. The level of this exposure surface is consistent with the erosional surface seen in Figure 3-41 further to the east of B146. This highly erosional surface correlates with the location of the Stage 2 Colorado fluvial channels at the time of maximum incision of the Colorado River.

The uppermost deepening event is interpreted to immediately follow a period of rapid rise in sea level from isotope Stage 2 to Stage 1. This rapid rise produced the shallow ravinement surface that can be traced from B146 eastwards over the Brazos shelf-margin delta seen in Figure 3-38. As deepening continued the planktic foraminifera flooded into the area as
recorded in the paleobathymetry for B146. In the next section, we will see that this deepening event follows a negative excursion in the oxygen isotope record for B146 that is associated with meltwater from the Laurentide ice sheet that flowed down the Mississippi River between 14,000 and 12,000 ybp.

Figure 3-42 shows the seismic data and the paleobathymetry for the boring A24. The boring falls within the channel shown in Figure 3-24 but is approximately 7 km downdip from the seismic line, and 12 km updip of B146. Using today's water depth as a guide we can expect A24 to be shallower than B146 by approximately 15 m (50 ft), and 3.7 m (12 ft) stratigraphically lower than the reflectors seen in the accompanying seismic line (Fig. 3-42).

A comparison with B146 should, therefore, place the Stage 2 unconformity at 18 m (58 ft), the Stage 3 mfs at 32 m (104 ft), and the Stage 5 mfs at 51 m (168 ft). The Stage 2 boundary in A24 (Fig. 3-40) shows good agreement with this prediction using B146. In Figure 3-40 coarse-grained sand and gravel occur between 15.5 and 16.2 m (Fig. 3-28a), with the base of the sandy interval occurring at 18.3 m (60 ft) (Fig. 3-40). This shows excellent correlation with the predicted depth of 18 m (58 ft) using B146.

The Stage 3 mfs and Stage 5 mfs are not as clear cut. In Figure 3-40, the plot for percent planktic foraminifera shows two peaks, i.e., at 51 m (168 ft) and at 42.4 m (139 ft). The lower peak shows excellent correlation with the predicted depth for the Stage 5e mfs, using B146. However, neither the upper peak, at 42.4 m (139 ft) nor the interpreted paleobathymetric curve,
centered at 43 m (140 ft), correlate with the predicted depths of either the Stage 5e mfs or the Stage 3 mfs.

One possible interpretation is that the percentage of planktic foraminifera is accurately flagging the Stage 5e mfs at 51 m (168 ft), with the second peak, at 42.4 m (139 ft), being associated with Stage 5c, or 5a. The Stage 3 mfs may not be recognized due to the shallow position of this boring. The boring A24 is located in approximately 41 m (135 ft) water depth, and with the Stage 3 sea-level position estimated in this study at 30-35 m below today's sea level, water depths of 6 - 11 m would have been too shallow to produce a planktic record. This would not be the case for Stage 5e when water depths would have reached 47 m at this location.

Using the seismic data in Figure 3-42 to predict the maximum flooding surfaces helps to resolve some of the conflict. The seismic data show the Stage 5e mfs and the Stage 3 mfs at 44 m (145 ft) and 27 m (88 ft), respectively. When we add the 3.7 m (12 ft) to compensate for the updip projection onto the seismic line we get predicted depths of 47.7 m (157 ft) and 30.7 m (100 ft) for the Stage 5e mfs and the Stage 3 mfs, respectively. The Stage 5e reflectors represent regional events and place the Stage 5e mfs coincident with the peak in percentage of agglutinated foraminifera at 47.7 m (157 ft) in Figure 3-40. This also shows good agreement with the central portion of the percentage of planktic foraminifera if we ignore the central trough (Fig. 3-40).

The preceding analysis, using the paleoenvironmental analyses for B146 and A24, and the seismic data, allow us to separate the ambiguities and arrive at a consistent interpretation for the Stage 2 sequence boundary, the
Stage 3 mfs, and the Stage 5e mfs. This is important since these boundaries allow us to establish the relative timing of depositional events throughout the study area.

3.2.3 *Macropaleontology*

Figures 3-43a and 3-43b show samples of the bivalve *Rangia cuneata* in cores at 4 m (13 ft) and 2 m (7 ft) from the borings G148 and B148, respectively. *Rangia cuneata* is a brackish-water bivalve, characteristic of a river influenced, low salinity environment (Parker, 1960, Andrews, 1981). It is common near bay-head deltas.

These specimens were found in the upper sections of incised valleys, associated with the Brazos and Colorado Rivers. In these relatively high-sediment supply fluvial/deltaic systems, bay-head deltas exist for only short time periods following transgressions. Subsequent regression of the Brazos and Colorado fluvial/deltaic systems can completely fill their estuaries (LeBlanc and Hodgson, 1959).

The occurrence of *Rangia cuneata* within the Brazos and Colorado incised valleys, therefore, indicates periods of marine transgression. In addition, they provide shell material for radiocarbon dating. The specimen from G148, at 4m, was dated at greater than 45,430 ybp (i.e., outside of the range of confidence for the radiocarbon method). The specimen from B148, at 2 m, was dated at 10,760 ybp using the radiocarbon method. In spite of their similar subsurface depths, and environments, these specimens lived during different cycles of marine transgression.
Figure 3-43a The Bivalve *Rangia cuneata* at 4 m in G148

Figure 3-43b The Bivalve *Rangia cuneata* at 2 m in B148
The older specimen is interpreted to be within a Stage 5d channel that was infilled during the ensuing Stage 5c transgression. The channel is shown in Figure 3-2, within the Brazos Stage 5e delta lobe. The younger specimen was located within the Colorado Stage 2 channel that filled during the Stage 2 to Stage 1 transgression. The channel is shown in Figure 3-26.

Figure 2-12a shows a specimen of the gastropod *Littorina irrorata*, from a core in B151, at 5.6 m (18.5 ft). This gastropod occurred in conjunction with organic rich clays, and is characteristic of fresh-water and low-salinity marshes (Parker, 1960, Andrews, 1981). Its occurrence above fine to medium sands, and below Holocene marine mud shows its relationship to the Late Pleistocene-Holocene transgression. The organic rich clays were dated at 9,530 ybp, which would also be equivalent to the approximate age of the Gastropod.

Macrofossils have provided additional information regarding the paleoenvironments in the study area, and the relative ages of two valley fills. In addition, we can see the flooding of the Colorado incised valley at the location of B148, at approximately -40 m (-130 ft), 10,760 ybp. Progressive marine transgression resulted in the development of a low-salinity marsh at the location of B151, at approximately -28 m (-93 ft), 9,530 ybp. This gives a general rate of sea-level rise of 12 m over a period of 1,230 years, or approximately 1 cm/year, during that interval of time. This value is in good agreement with that of Bard et al. (1990), who estimate rates of 0.9 m/century during this period.

The distance between the core sites, B148 and B151, is 30 km. Rapid landward shifts of shorelines must have occurred during these
transgressions. In addition, the marsh deposits in B151 were the only example of preserved marsh deposits identified in the study area. Slow, continuous transgressions can be expected to completely remove shallow deposits from the stratigraphic record (Thomas and Anderson, 1994). Preservation may be related to one of the rapid sea-level rises documented by Thomas (1990), and Thomas and Anderson (1994).

3.3 Oxygen Isotope Stratigraphy

Figure 3-44 shows the oxygen isotope record for B146. This record was generated using the benthic foraminifer Quinqueloculina sp. Figure 3-45 shows the isotope record together with the seismic line R93-51. Figures 3-46a and 3-46b show detailed plots of the upper 7.6 m (25 ft) of the isotope records for B146 generated from the benthic foraminifer Quinqueloculina sp. and the planktic foraminifer Globigerinoides ruber, respectively.

Cores from the B146 boring were selected for oxygen isotope analyses since B146 represented the most seaward located boring that had cores, in the study area. The seaward position was important since samples from the inner shelf were more likely to contain hiatuses during even minor sea-level falls. Initially it was hoped that a complete planktic record may have been obtained, however, gaps in the planktic record occurred due to shallow water depths during sea-level minima.

A relatively continuous benthic record was obtained using the foraminifera Quinqueloculina sp. and this record is shown in Figure 3-44. The upper 7.6 m (25 ft) of the isotope record shows an extremely negative
Figure 3-44 Oxygen isotope record and boring description for B146
Figure 3-45 Oxygen isotope record for B146, and seismic line R93-51
Figure 3-46a Meltwater pulse generated using *Quinqueloculina sp.*

Figure 3-46b Meltwater pulse generated using *Globigerinoides ruber*
excursion (Figs. 3-44, 46a and 46b). This negative pulse is interpreted to represent the Mississippi meltwater pulse of Kennett and Shackleton (1975) and Leventer et al. (1982). Leventer et al. (1982) dated this pulse as occurring between 14,000 - 12,000 ybp (Fig. 2-14). In Figure 3-45 we can see that this pulse can be correlated to the base of the upper, parallel reflectors. The base of these reflectors corresponds to the ravinement surface associated with the Stage 2 to Stage 1 rise in sea-level. This surface is seen to truncate the top of the prograding clinofoms in Figure 3-38, and is interpreted to have developed at or near 14,000 ybp, based on its position relative to the meltwater pulse in B146. Bard et al. (1990) document a rapid rise in sea level from a study of coral terraces in Barbados that started at 14,000 ybp. These authors estimate the rate of sea-level rise at that time to be 3.7 m/century.

The start of the negative excursion just precedes the large influx of planktics shown in the paleoenvironmental chart for B146 in Figure 3-39. However, insufficient planktics were present to document the entire pulse. The isotope record, shown in Figure 3-46b, was generated using the planktic foraminifer *Globigerinoides ruber*.

The interval from 8 m (27 ft) to 26 m (85) ft was characterized by interlayered silty clays and silty fine sands, and was barren of fossils so that no isotope record exists for this interval (Fig. 3-44). This interval is interpreted to represent silty delta-front deposits associated with a slow rise in relative sea level, prior to the rapid rise at 14,000 ybp. The paleobathymetry for this interval is inner neritic (Fig. 3-39).
The isotope record for B146 (Fig. 3-44) shows other correlations with the paleobathymetry (Figs. 3-39 and 3-41) and with the seismic data (Fig. 3-45). Figure 3-47 shows the isotope record together with the paleobathymetric data. The exposure surface seen on the seismic and paleobathymetry data at 33 m (108 ft) correlates with a positive $\delta^{18}$O maximum at 31 m (102 ft). The Stage 3 deepening in the paleobathymetry and the Stage 3 downlap surface, both at 47 m (154 ft), correlate well with a negative isotope maximum at 47.6 m (156 ft). The Stage 5e flooding interpreted to be at 67 m (218 ft) in the paleobathymetry and seismic data can be correlated with a negative isotope maximum at 68.3 m (224 ft).

The strong correspondence between sea-level maxima and negative peaks, and the sea-level minimum with a positive peak, strengthens the ties between the oxygen isotope data, the biostratigraphic data, and the seismic data. This in turn provides us with a chronostratigraphic framework for the study area.

Other negative peaks in the isotope record of Figure 3-44 occur at 76.2 m (250 ft), 64.6 m (212 ft), 58.5 m (192 ft), and 36.6 m (120 ft). These negative peaks may represent a pre-Stage 5e meltwater pulse, Stage 5c, Stage 5a, and Stage 3a, respectively (Fig. 3-44).

### 3.4 Chronostratigraphy

Correlation between the biostratigraphic data, oxygen isotope data, seismic data, and radiocarbon age dates provides chronostratigraphic control
Figure 3.47 Isotope data and paleohydraulic data for the boring B146.
for this study. The important chronohorizons that can be established are, from oldest to youngest:

i) The downlap surface associated with the Stage 5e maximum flooding surface (approximately 125,000 ybp).

ii) The stratigraphic level that ties with the top of the Peorian Interglacial, i.e., the extinction of the planktic foraminifer Globorotalia menardii flexuosa (85,000 ybp, i.e., Stage 5a/b transition).

iii) The downlap surface associated with the Stage 3 maximum flooding surface (approximately 59,000 ybp).

iv) The Stage 2 exposure surface, or sequence boundary (approximately 18,000 ybp).

v) The ravinement surface associated with the rapid rise in sea level at 14,000 ybp.

vi) Absolute dates obtained from shell material and organic carbon over the last 11,000 years.
The evolution of the study area can then be addressed once the chronocorrelation has been established. The evolution is presented in Chapter 4.
CHAPTER 4 - INTEGRATED INTERPRETATION

Introduction

The integrated interpretation of the data set is presented in this chapter. Using the chronology established by the biostratigraphic data, the oxygen isotope data, and radiocarbon dates, the evolution of the study area is reconstructed. Where possible, reference is made to the oxygen isotope stages so that deposition is tracked through time in reference to global sea level. The discussion starts from the oldest, Stage 6, to the youngest, Stage 1. For reference, the SPECMAP oxygen isotope record of Imbrie et al. (1984) is shown in Figure 4-1, converted to depth by G. Haddad.

The relationship between the late Quaternary evolution of the shelf, the surficial sediments, and the present-day bathymetry is also discussed. This is followed by a discussion of the relevance of this study to sequence stratigraphy. The main conclusions of the thesis are listed at the end of this chapter, except for those conclusions related to the experimental 3-D seismic survey, which are presented in Chapter 5.

4.1 Stage 6 Deposition

Most of the Stage 6 deposits are obscured below the sea-floor multiple, and have not been routinely mapped over the study area. However, at the shelf edge the sea-floor multiple steepens in dip and Stage 6 shelf-margin deltas can be identified (Fig. 3-37). The biostratigraphy from the
Figure 4-1 SPECMAP curve of Imbrie et al. (1984) converted to depth by G. A. Haddad
Shell Eureka core 67-45 identifies the upper prograding package as deltaic, and places it within the Early Wisconsin (Anderson and Parrott, 1969). This corresponds to isotope Stage 6 using the Shell stratigraphy and the chronostratigraphic chart of Kohl (1986) (Fig. 2-16).

Two phases of progradation can be seen in Figure 3-37, and Lewis (1984) mapped these units within the Galveston South Addition lease area. Lewis (1984) referred to the upper and lower prograding units as Sequence A, and Sequence B, respectively. The lower prograding package has not been mapped over the study area, and its age at this time is uncertain. It is possible that it may be associated with an early Stage 6 delta, or it may be an older Stage 8 lowstand deposit.

Incised valleys can sometimes be identified on the inner shelf as possible Stage 6 valleys. Figure 3-24a showed an example of an older incision, capped by the parallel reflections of the Stage 5e mfs. The position of this valley, relative to the Stage 5e mfs, makes it a likely candidate for Stage 6 incision. Although the data quality is poor at this level, there are indications of older incisions within the central part of the study area. This trend is more clearly recognized by the platform borings that show thick sands, often with medium to coarse grain size, and occasional gravels.

Figure 4-2 shows two incised valley trends and their associated shelf-margin deltas. The eastern valley is interpreted to be the sediment source for the Stage 6 Brazos shelf-margin delta shown in the southeastern corner of the study area. The areal extent of the delta has been mapped from the seismic data, an example of which was shown in Figure 3-37. Figure 4-3 shows examples of the sandy valley fill within the Stage 6 Brazos valley, in
Figure 4-2 Stage 6 incised valley and shelf-margin deltas
Figure 4-3 Platform borings B35, B116, B150, and B56 within the Stage 6 incised valley
platform boring descriptions for B35, B116, B151, and B56. The gamma-ray log for this interval, shown in Figure 4-3 for B56, shows a sharp basal contact with the underlying clays, characteristic of incised channels. Figure 4-4 shows seismic line R93-42, recorded over the Stage 6 Colorado shelf-margin delta. The areal extent of the Stage 6 Colorado delta is not as well controlled as the Stage 6 Brazos delta, but its approximate extent is shown in Figure 4-2. The Stage 6 Colorado incised valley can be seen in the platform borings B2, B4, B77, and B148, shown in Figure 4-5.

Large amounts of sand occur within both the Brazos and Colorado Stage 6 incised valleys (Figs. 4-3 and 4-5). The sandy nature of the Brazos shelf-margin delta reported in the Shell 67-45 core is supported by evidence of large slumps seawards of the Brazos shelf-margin delta as seen by the chaotic seismic reflections on line G300X (Fig. 3-37). These factors indicate that Stage 6 was likely a period of high sediment supply, and sediment bypass at the shelf edge.

4.2 Stage 5 Deposition

The chronologic control for Stage 5 deposition is the identification of the Peorian interglacial, and the position of the stratigraphic section relative to the Stage 5e maximum flooding surface.

Figure 3-37 shows the thick fine grained sediments that were deposited during the Peorian Interglacial as identified by the biostratigraphy in the Shell core 67-45. The fine-grained nature of these sediments is recorded both by the lithostratigraphy from the core, as well as by the near
Figure 4-4 Seismic line R93-42 showing the Stage 6 Colorado shelf-margin delta
Figure 4-5 Platform borings B2, B4, B77, and B148 within the Stage 6 Colorado incised valley
parallel stratification of the seismic reflections. Some of these sediments were no doubt deposited during the rise in sea level from Stage 6 to Stage 5e.

The top of the Peorian lies close to the Stage 5a/5b transition at 85,000 ybp, as defined by the extinction of *Globorotalia menardii flexuosa* (Fig. 2-16). This level can be traced updip along the seismic reflections and is found to fall within the most seaward of three prograding delta lobes associated with the Brazos fluvial/deltaic system (i.e., lobe 3 of Figure 3-4c).

The oldest of the three Brazos lobes, lobe 1 of Figure 3-4a, downlaps a high-amplitude seismic reflector associated with the Stage 5e mfs (approximate age 125,000 ybp) (Fig. 2-4). The Colorado lobe shown in Figure 3-11 is also interpreted to down lap the Stage 5e mfs. The Stage 5e maximum flooding surface is a smooth, regionally extensive surface, characterized by high-amplitude seismic reflections. Thomas (1990) mapped this surface offshore Galveston Bay using high-resolution seismic data. The surface was tied into a regional reflector mapped offshore Louisiana by Coleman and Roberts (1988), and projects onshore directly below the Stage 5e Ingleside Barrier (Thomas, 1990).

The Stage 5e reflector within the study area has been tied to that of Thomas (1990), and forms a major downlap surface (Figs. 2-1 and 2-4). In addition, it can be traced offshore to the deepening event in B146 at 67 m (Fig. 3-39), and the negative oxygen isotope peak in B146 at 68.3 m (Fig. 3-44). The Stage 5e mfs has been correlated over the entire study area and the
structure contour map for the Stage 5e mfs is shown in Figure 4-6. Note that it is a gently sloping surface with little relief.

Figure 4-7 shows the delta lobes interpreted to have been deposited during Stage 5. The lobate outlines of the oldest deltas are projected onshore. The eastern onshore area shows good agreement with the occurrence of Stage 5 fluvial systems mapped by Van Siclen (1991) for the Brazos River. The projection of the western, Stage 5 Colorado delta lobe onshore shows good agreement with the Stage 5 Colorado fluvial system mapped by Blum and Price (1994).

Three Stage 5 Brazos delta lobes have been mapped and one Stage 5 Colorado delta lobe (Fig. 4-7). Each of the Stage 5 Brazos lobes are characterized by an aggradational to progradational clinoform configuration. Plate 1 shows seismic lines R91-4 and G380X which illustrate the reflection configuration within these lobes. Each younger lobe onlaps the older lobe during a period of aggradation, prior to more rapid progradation (Plate 1).

The periods of onlap and aggradation are interpreted to indicate periods of relative sea-level rise. Evidence for a minor transgression is the occurrence of the bivalve *Rangia cuneata* within a fluvial incision cut into the oldest Stage 5 Brazos lobe. This bivalve, *Rangia cuneata*, was found at 4m in G148 (Fig. 3-43a) and is characteristic of the brackish-water environment typical of the bayhead delta. The radiocarbon age of this specimen was greater than 45,430 ybp, i.e., definitely not Holocene. The flooding of the fluvial channel must have taken place after the deposition of the delta lobe, but before the Stage 2/1 rise in sea level. Further, since the
Figure 4-6 Stage 5e structure contour map
Figure 4-7 Map showing the distribution of Stage 5 delta lobes
delta lobe downlaps the Stage 5e mfs, the transgression must have occurred post-Stage 5e.

As seen on seismic line G380X (Plate 1), the upper surface of the youngest Stage 5 Brazos delta lobe is characterized by a high-amplitude reflection. Seismic line G400X (Plate 2) also illustrates this high-amplitude boundary. The high amplitudes recorded off the upper surface of the youngest Stage 5 Brazos delta lobe is an indication of subaerial exposure, or at least a break in sedimentation. This interpretation is reinforced by the subsequent onlap of a younger delta seen on seismic line G400X (Plate 2).

The clinoform configuration, evidence of minor transgressions that includes the presence of bayhead delta environments within incised channels, and evidence for falling sea level, indicate that the Stage 5 Brazos deltas were deposited during three periods characterized by rising and subsequent falling sea levels. The oldest lobe downlaps the Stage 5e mfs, and the youngest lobe appears to have been active during the Stage 5b/5a transition.

The three Stage 5 Brazos delta lobes are interpreted to associated with oxygen isotope Stages 5e, 5c, and 5a (Fig. 4-1), based on the factors discussed above. These lobes will be referred to as Stage 5e, Stage 5c, and Stage 5a lobes in following discussions. Should this correlation be correct, then the Brazos fluvial/deltaic system provides evidence for delta-locale shifting in response to the high-frequency eustatic fluctuations indicated by the oxygen isotope record. The minor transgressions put an end to the earlier phase of progradation, and the subsequent regression is associated with a new delta lobe. This process is different from that of delta-locale
switching in response to upstream river capture, which is related more to valley incision and headward erosion of minor streams.

Only one Stage 5 Colorado delta has been mapped (Fig. 4-7), and it is interpreted as a Stage 5e Colorado lobe. The Colorado Stage 5e lobe shown in Figure 4-7 appears to migrate to the southwest, probably in response to the steeper shelf gradients (greater accommodation) along the south-central Texas shelf. Therefore, it is possible that other lobes exist outside of the study area, towards the southwest. Alternatively, younger Stage 5 lobes may have prograded southwards in a similar manner to the Stage 5 Brazos lobes. The gamma-ray log for B51 (Fig. 3-27) shows two coarsening upwards cycles between the Stage 5e mfs and the Stage 3 mfs. One could speculate that these represent the additional Stage 5 Colorado delta lobes. However, the seismic data at this level is of extremely poor quality and additional lobes cannot be identified.

4.3 Stage 4 Deposition

The extent of deposition during Stage 4 is not clear. The only area where Stage 4 deposits appear to exist is in the vicinity of seismic line G400X (Plate 2). Here we can see a relatively thick lowstand to transgressive deposit underlying the subsequent prograding clinoforms (G400X, Plate 2). The occurrence of a relatively thick Stage 4/3 record here may be a function of increased accommodation due to the fault.

Elsewhere, Stage 4, and for that matter Stage 5d and Stage 5b, deposits are restricted to thin transgressive deposits associated with the Stage
4/3, Stage 5d/c, and Stage 5b/a transgressions, respectively. True lowstands appear to be missing. Whether they have been removed by longshore currents, or make up part of the regressive delta lobe is not clear.

4.4 Stage 3 Deposition

The Stage 5a Brazos delta lobe is onlapped by a younger phase of deltaic deposition seen on seismic lines G380X (Plate 1) and G400X (Plate 2). Seismic line G400X (Plate 2) shows that a period of transgression preceded the subsequent progradation of the delta to the extent that the lower portion of the Brazos Stage 5a delta was completely transgressed. This transgression resulted in the development of a well defined downlap surface for the subsequent prograding clinoforms (G400X, Plate 2).

The prograding clinoforms show a progression from aggradational and progradational to strongly progradational (G380X, Plate 1, G400X, Plate 2), suggestive of a period of gradual rise in sea level, followed by gradual fall. The fact that deposition of this prograding delta lobe is followed by a fall in sea level is also seen in the fact that the upper surface of the delta lobe is characterized by a high-amplitude reflection often associated with a depositional hiatus. The seaward surface of the lobe is also onlapped by reflectors indicating that deposition on the seaward side of the lobe occurred during subsequent rising sea level, i.e., following a fall.

On seismic line G300X (Plate 2 and Fig. 3-38) we see that the upper surface of this prograding delta is characterized by fluvial incisions. The upper surface has also been eroded during the late Pleistocene-Holocene
transgression, and this ravinement surface has been tied to the platform boring B146 and the rapid rise that occurred 14,000 ybp.

This prograding delta, therefore, was deposited following a fall in sea level after Stage 5a. It was deposited during a period of sea-level rise and later fall that resulted in the incision of its upper surface. It is also onlapped by sediments that pre-date the 14,000 ybp rapid rise that produced a ravinement surface at its upper surface, and removed most of the topset beds of the delta. All of these factors identify this delta as being deposited during isotope Stage 3 (Fig. 4-1). Figure 4-8 shows the areal extent of this delta lobe, which is interpreted as a Stage 3 Brazos delta lobe.

Plate 3 shows regional strike line R93-51. This line was recorded to allow us to tie the biostratigraphy and oxygen isotope stratigraphy from B146 to the regional seismic grid shown in Figure 1-4. The Brazos Stage 3 delta lobe is seen in the central portion of line R93-51 (Plate 3). The lobe downlaps the Stage 3 mfs tied to the deepening event in B146 (Fig. 3-41), and to a negative peak in the isotope record for B146 (Fig. 3-45). At the western end of line R93-51 (Plate 3) is a thin sedimentary wedge that downlaps the Stage 3 mfs and thins to the east. This wedge is interpreted as a Stage 3 Colorado lobe. It is shown in Figure 4-8 although its full areal extent is not known.

The main Stage 3 Colorado delta lobe is also shown in Figure 4-8. From Plate 3 we see that this lobe onlaps the Stage 3 Brazos lobe and, therefore, must be slightly younger than the Brazos lobe. However, it is bounded above by an erosional surface that is at the same level as the Stage 2 exposure surface identified in B146 (Fig. 3-41 and Fig. 3-45). This
Figure 4-8 Map showing the distribution of Stage 3 delta lobes
Colorado lobe is definitely pre-Stage 2 and in an updip location; it shows aggradation and progradation associated with the shape of the Stage 3 sea-level proxy (Fig. 4-1). The smaller sediment supply of the Colorado River, and its bedload mode of transport versus the higher sediment supply, suspended load of the Brazos River are considered the reasons for the later arrival of the Stage 3 Colorado delta at the shelf margin.

Figure 4-8 also shows the landward extent of the Colorado Stage 3 delta lobe. Its landward extent corresponds to present-day water depths of approximately 40 m. If we assume that these sediments were deposited at the maximum Stage 3 highstand, then they would be approximately 53,000 years old. At subsidence rates in the inner shelf of 0.13 mm/year, we would expect this location to be 7 m lower today than at 53,000 ybp, i.e., the Stage 3 shoreline may have been as high as -33 m below today's sea level. An estimate of -30 m to -35 m for Stage 3 agrees more with the amplitude of sea-level events predicted from work on coral terraces (Bloom and Yonekura, 1985) (Fig. 1-9c) than the sea-level stand estimated from the SPECMAP curve (Fig. 4-1).

This discrepancy implies that temperatures were likely colder than predicted, since cooler temperatures will result in the enhancement of the positive values in the d$^{18}$O curve. The lack of a pre-Stage 3 meltwater pulse in our oxygen isotope record implies that either the Laurentide ice sheet did not advance sufficiently far south to have its meltwater directed down the Mississippi drainage (i.e., the meltwater went down the Champlain-Hudson Valley or the St. Laurence River) or the Laurentide ice sheet advanced, but did not melt. Cooler temperatures seem to favor the latter explanation but
either way the evidence suggests that the Stage 3 sea-level rise was predominantly a result of Antarctic deglaciation.

4.5 Stage 2 Deposition

During the rapid fall in sea-level from Stage 3 to Stage 2 (Fig. 4-1) the Brazos River was diverted eastwards (Figs. 4-9a and 9b) and entered the incised valley system of the Trinity and Sabine Rivers, previously mapped by Thomas (1990) and Sarzalejo (1993). Seismic line R93-51 (Plate 3) shows that the Brazos Stage 3 delta lobe is not incised at the shelf margin. Its crest lies above the level of the exposure surface in B146 indicating that the top of the lobe, seen in Plate 3, was subaerially exposed. The lack of fluvial incisions on R93-51 (Plate 3) coupled with an absence of major turbidites or slumps directly downdip of the lobe in any of the upper-slope cores, supports the interpretation of an eastwards diversion of the Brazos River.

Sediment bypass during this time was probably occurring down the joint Trinity/Sabine incised valley system and out onto the upper slope. Evidence exists for chaotic predominantly fine-grained sediments within the upper-slope minibasin directly downdip from the incised valleys (Figs. 4-9a and 9b). The site for the deposition of the coarser grained sediments is not known but they must have made their way further downdip. Sarzalejo (1993) suggested that the sands from the Trinity/Sabine incised valley system may have made their way to the Gyre and Horseshoe Basins further down the slope.
Figure 4-9a Map showing the Stage 2 incised valleys and Stage 3 delta lobes
Figure 4-9b Map showing Stage 2 incised valleys and late Stage 2 deposits
Curray (1960) predicted an eastwards trend for the Brazos River during the last lowstand from bathymetric data and grain size analysis from surficial samples. Parker (1960) used Curray's data set to show an eastwards trend for the Brazos River based on the occurrence of shell material containing low-salinity fauna (i.e., *Rangia, Mulinia* and *Crassostrea*). These fauna are likely to have developed during the subsequent transgression as the Brazos Stage 2 incised valley was flooded.

The Colorado River continued to flow directly south during the Stage 3 to Stage 2 sea-level fall (Fig. 4-9a). The upper surface of the Colorado Stage 3 delta is irregular due to fluvial incisions (Plate 3) that can be traced all the way to the shelf edge. At the shelf edge, the incisions merge with the submarine canyons updip of the East Breaks Slide (Figs. 4-9 and 3-30). The East Breaks Slide was previously mapped by Lehner (1969), Tatum (1977), Woodbury et al. (1978), and Rothwell et al. (1991). Woodbury et al. (1978) described the contorted bedding and displaced inner neritic, shallow-water fauna from the core 14-6 taken within the slide.

4.6 Stage 2 to Stage 1 Transgression and Deposition

The deposition associated with the early rise in sea level from Stage 2 to Stage 1 is poorly recorded for the Brazos fluvial/deltaic system. This may be due to the fact that most of the deposition was occurring within the lower portion of the combined Brazos/Trinity/Sabine incised valley. The seismic data within the lower portion of this incised-valley system is extremely poor.
Fine-grained sediments are seen to onlap the Stage 3 Brazos delta lobe as seen on seismic lines G380X (Plate 1), G300X (Plate 2), G400X (Plate 2), R91-8 (Plate 4) and R92-24 (Plate 4). The lower onlapping wedge of sediments is associated with late Stage 2 Brazos sediments (Fig. 4-9b). However, the upper wedge of sediments will be seen to be associated with the western Louisiana fluvial/deltaic system that prograded into the study area from the northeast, and was studied by Sarzalejo (1993).

Figure 4-10 shows the distribution of facies during the transgression. The Brazos and Colorado deltas both backstepped onto the continental shelf during this time. The base of the backstepped delta is the ravinement surface associated with the rapid rise in sea level that can be correlated with the Mississippi meltwater pulse dated at 14,000 ybp. Figure 3-45 shows the correlation with the strike line R93-51 (see figure 3-1 for location).

Seismic line R93-51 is shown in Plate 3. The ravinement surface can be seen truncating the upper surface of the large central delta lobe on this line. This is the Stage 3 Brazos lobe and it can be seen in Plate 3 to be onlapped from the east by a thick, young wedge of coherent reflectors. These reflections are from the western Louisiana fluvial/deltaic system previously mapped by Sarzalejo (1993).

Sarzalejo (1993) had radiocarbon dates that identified the lower most reflectors of this wedge as being approximately 14,000 ybp. This date shows good agreement with the chronostratigraphy from this study that places the wedge directly above the time correlative seismic events at the time of first transgression, i.e., at 14,000 ybp.
Figure 4-10 Map showing the Stage Figure 2- Stage 1 Transgression
The thick wedge of sediments is expressed in the dip direction as a thickening of the youngest prograding clinoforms on dip lines as one goes further east. Plate 4 is an illustration of this effect. Seismic line R91-8 intersects the Stage 3 Brazos lobe just off its western flank (see Plate 3 for the intersection of line R91-8 with the strike line R93-51). The youngest wedge of sediments seen onlapping the Stage 3 Brazos lobe has a slightly convex upwards geometry at the shelf edge (seismic line R91-8, Plate 4). In comparison, seismic line R92-24 on Plate 4 has a very thick and complex series of prograding wedges that can be seen to correlate in a landwards direction with reflections that lie above the ravinement surface in line R92-24 (for the position of the ravinement surface on line R92-24 see also Fig. 2-3).

The intersection of line R92-24 with R93-51, and its eastwards position relative to line R91-8, can be seen on Plate 3. On Plate 3 we see that R92-24, with the young onlapping wedges, crosses the thick young wedge of sediments that onlaps line R93-51. In fact, these packages tie together and we are seeing the oblique progradation of a young, fine-grained delta across the southeastern corner of the study area, i.e., from ENE to WSW. Figure 4-10 demonstrates this thick wedge that trends almost east-west in this figure.

Further updip within the Brazos and Colorado systems, on the shelf, dates from a *Rangia cuneata* shell in B148 (see Fig. 3-1 for core locations) was dated as 10,760 ybp. This specimen is shown in Figure 3-43b. Marsh deposits also developed on the shelf during the transgression. Examples of the organic rich clays and characteristic low-salinity marsh gastropod
*Littorina irrata* are shown in Figure 2-12a. The organics from these clays were dated at 9,530 ybp, and the samples were from B151 and B152, both located very close to B129, whose location is shown in Figure 3-1 for the boring locations. So, the transgression of the Brazos and Colorado fluvial/deltaic systems can be traced across the shelf (from an early date of 14,000 ybp dated with the meltwater pulse and positioned by the first occurrence of the ravinement surface) to later positions and dates marked by bayhead delta and marsh macrofossils, dated by radiocarbon methods, and positioned by the core locations.

The western Louisiana fluvial/deltaic system was not backstepping at this time, but continued to prograde seawards. This is an important point since it demonstrates the ability for sediment supply to overwhelm the immediate eustatic effect of a rapid rise in sea level, and increased accommodation. It is not known exactly when the western Louisiana fluvial/deltaic system ceased to prograde. However, we know that its base is defined by radiocarbon dates from Sarzalejo (1993) at approximately 14,000 ybp. We also see at least 100 m of sediment within this young prograding wedge (see line R92-24 in Plate 4 and line R93-51 in Plate 3). Given even dramatically high sedimentation rates for the Gulf of Mexico of 10 mm/year, 100 m of sediment represent some 10,000 years of deposition. It is speculated, therefore, that the western Louisiana fluvial/deltaic system continued to prograde into the eastern part of the study area up until 4-5,000 ybp.

An examination of the oxygen isotope record shown in Figure 4-1 provides us with an independent sea-level proxy during the times indicated
by our chronologic data. Figure 4-1 shows that the Brazos and Colorado fluvial/deltaic systems backstepped during the steep rising limb of the late Pleistocene to Holocene sea-level rise, i.e., the Stage 2 to Stage 1 transgression. This same time interval was a period of dramatic progradation for the western Louisiana fluvial/deltaic system.

The transgression is seen also to be the time when most of the infilling of the fluvial-incised valleys, cut during the preceding lowstand, occurred. Evidence for this is seen in the frequent occurrence of shell material together with coarse-grained sands and gravels of the basal-channel lag deposits (Fig. 3-28a). Shell material also appears in the platform boring descriptions for the borings shown in the cross-sections of the Stage 6 incised valleys (Figs. 4-3 and 4-5). The shell material is interpreted to be associated with marine influence at the mouth of the channels during the transgression. One would not expect this mixed lithology in the upper continental reaches of today's Brazos or Colorado Rivers.

The marine influence in the basal, coarse-grained channel deposits becomes more dominant higher in the section of the valley-fill facies. The fine-grained, laminated valley fill seen in Figures 3-24b are most certainly estuarine to marine muds. Figure 3-29a and 3-29b show this laminated upper-valley fill character with a silt-filled burrow. The transgressive valley-fill facies and architecture were well documented by Thomas (1990) and Thomas and Anderson (1994) for the adjacent Trinity/Sabine incised valley.
4.7 Summary of the Evolution of the Brazos and Colorado Fluvial/Deltaic Systems

The evolution of the Brazos and Colorado fluvial/deltaic systems can be summarized as follows.

1) Stage 6 (Figs. 4-2 and 4-11)
   
   Stage 6 was a period of valley incision, and deposition of lowstand deltas. The plan view is seen in Figure 4-2, and the corresponding schematic in Figure 4-11.

2) Stage 5 (Figs. 4-7 and 4-12)
   
   High stand deposition occurred on the shelf with the deposition of Stage 5e Brazos and Colorado deltas. Higher-order cyclicity produced Stage 5c and Stage 5a Brazos delta lobes. It is not clear where the Colorado fluvial/deltaic system was during Stages 5c and 5a, but it seems likely that it was located further to the south west, outside of our data control. The plan view is seen in Figure 4-7, and the corresponding schematic in Figure 4-12.

3) Stage 3 (Figs. 4-8 and 4-13)
   
   Stage 3 saw the development of two Colorado delta lobes, one early Stage 3 and one late Stage 3. The younger Colorado delta lobe onlaps an earlier Stage 3 Brazos delta lobe (Figs. 4-8 and 4-13). Both the early Stage 3 Colorado lobe and the early stage 3 Brazos lobe downlap the maximum flooding surface associated with the Stage 3 higher order eustatic cycle.
Figure 4.12 Schematic diagram for Stage 5
This relationship is seen in Figures 3-41 and 3-45, and in Plate 3. The upper surfaces of both the early and the late Stage 3 Colorado deltas have been tied to the erosional Stage 2 sequence boundary (Fig. 3-41). The upper surface of the Stage 3 Brazos delta has been identified as a clearly defined ravinement surface (Figs. 2-3 and Plates 3 and 4).

4) Stage 2 (Figs. 4-9a, 4-9b and 4-14)

Stage 2 was a period of rapid sea-level fall, valley incision and sediment bypass (Figs. 4-9a, 4-9b and 4-14). However, bypass did not always occur directly downdip of the delta lobes. The Stage 2 incisions for the Brazos shifted to the east in response to the shorter distance and steeper gradients provided by the flank of an earlier Stage 5a delta lobe and the presence of the Trinity/Sabine incised valley to the east. The slower progradation of the more bedload dominated Colorado fluvial/deltaic system allowed it to maintain a straight path to the shelf edge and to the East Breaks Slide (Fig. 4-9 and 4-14).

5) Stage 2 to Stage 1 Transgression (Figs. 4-10 and 4-15)

The transgression produced backstepping of the Brazos and Colorado fluvial/deltaic systems (Figs. 4-10 and 4-15). The transgression is marked by infilling of the incised valleys with a coarse-grained basal deposit often showing marine influence. The upper portions of the valleys often have environments characteristic of the bay-head delta environment. Marsh deposits occur locally.
Figure 4.15 Schematic diagram for Stage 2 to Stage 1 Transgression
The western Louisiana fluvial/deltaic system prograded into the study area during the period of rapid sea-level rise and continued to prograde, despite the increasing rate of accommodation (Figs. 4-10 and 4-15). The continued progradation of this system is a function of its very high sediment supply.

4.8 Surficial Sediments and Present-Day Bathymetry

Curray (1960) published a series of surficial sediment maps that were the result of an eight year study of the recent sediments of the Northwest Gulf of Mexico. It is useful to review these surficial sediment maps and the detailed bathymetry, also from Curray (1960), within the context of the interpretations presented in this study.

Figures 4-16a, 16b, 16c, and 16d, show the distribution of four grain-size modes over the northwestern Gulf. The first grain-size mode (Fig. 4-16a) includes the range from 0 - 2.9 phi (1.00 - 0.134 mm). Following the Wentworth size classification, this range includes coarse sand, medium sand, and fine sand. The arrows shown in Figure 4-16a are Curray's interpretation for the direction of sediment transport.

The Brazos and Colorado rivers are both candidates for the source of sand-sized sediments on the east Texas shelf, seaward of the 40 m isobath (Fig. 4-16a). In addition, an easterly trend for the Brazos River sediments is shown in Figure 4-16a. This easterly trend for the distribution of coarse to fine Brazos sediments is in good agreement with the easterly trend for the
Figure 4-16a Distribution of grain size modes 0-2.9 phi (1.00 -0.134 mm) after Curray (1960)

Figure 4-16b Distribution of grain size modes 3.0-4.2 phi (0.125-0.054 mm) after Curray (1960)
Figure 4-16c Distribution of grain size modes 4.3-7.1 phi (0.051-0.0073 mm) after Curay (1960)

Figure 4-16d Distribution of grain size modes > 7.2 phi (0.0068 mm) after Curay (1960)
Stage 2 Brazos incised valley mapped in this study (Figs. 4-9a, 4-9b and 4-14).

The shelf-margin parallel trend for the coarse to fine-grained Brazos and Colorado sediments (Fig. 4-16a) are also in good agreement with the location of Brazos and Colorado Stage 3 shelf-margin deltas (Figs. 4-8, 4-9a, 4-13, 4-14 and 4-17). The prominent bulge seen in the southwestern corner of bathymetric map of Figure 4-17 is the Stage 3 Colorado shelf-margin delta. Curray (1960) identified this shelf-margin delta using the surficial sediment maps and present-day bathymetry. Since the modern Brazos and Colorado rivers deliver sand-sized sediments only to the innermost shelf, the coarse to fine sands mapped by Curray (1960) (Fig. 4-16a) represent pre-modern deposition. Stage 3 and Stage 2 rivers and deltas were the likely source for these surficial sediments.

Figures 4-9a, 4-13 and 4-14 show that the Stage 3 Colorado delta prograded north to south throughout the fall in sea level from Stage 3 to Stage 2. This continued north-south progradation resulted in the oversteepening of the Stage 3 delta at the shelf edge and resulted in the prominent bulge seen in the bathymetry (Fig. 4-17). The higher sediment supply of the Brazos River allowed it to construct a larger delta than the Colorado delta, as evidenced by the strike line R93-51 (Plate 3). However, the eastwards shift of the Brazos River into the embayment that would become the future site for the Trinity/Sabine incised valley (Figs. 4-9a and 4-14) prevented the continued north-south progradation of the Brazos Stage 3 delta to the shelf-edge. The result is that the Stage 3 Brazos delta lies
Figure 4-17  Present-day bathymetry, after Curray (1960). Contour interval is approximately 2m (6 fathoms). Depths shown from 10m on the inner shelf to 100m on the outer shelf.
landward of the previous Stage 6 shelf edge (Plate 2) and did not produce a conspicuous bathymetric bulge.

The Stage 3 Brazos delta is, therefore, not oversteepened. This fact is supported by the absence of chaotic seismic events seaward of the Stage 3 Brazos delta lobe, as well as the absence of major turbidites in the upper-slope core data. In contrast, both the Stage 6 Brazos delta and the Stage 3 Colorado delta were oversteepened, as evidenced by the slumping seen on line G300X (Plate 2) and by the occurrence of the East Breaks Slide, respectively (Fig. 3-23). Displaced shallow-water fauna reported in upper-slope core data (Anderson and Parrott, 1969, and Woodbury et al., 1978) support this interpretation.

Figures 4-16 b and 4-16c show the distribution of the 3.0 - 4.2 phi (0.125 - 0.054 mm) and the 4.3 - 7.1 phi (0.051 - 0.0073 mm) size factions, respectively. These two grain-size modes approximate the very fine sand and silt (coarse silt, medium silt and fine silt) sizes, based on the Wentworth classification. The distribution of both of these grain-size modes point to the Brazos and Colorado Rivers as sources for the sediments of the east Texas shelf (Fig. 4-16b and 16c). The southeasterly trends of the Brazos sediments and the southwesterly trends of the Colorado sediments, as seen by the arrows in Figures 4-16b and 4-16c, show good agreement with the trends of the Stage 5 delta lobes mapped in this study (Figs. 4-7 and 4-12). The fine-grained sediments of the inner shelf are likely to have a large component that is sourced from the reworking of the Stage 5 delta lobes. These Stage 5 delta lobes would have been subaerially exposed during the Stage 2 lowstand
and their crests would have experienced transgressive erosion during the Stage 2 to Stage 1 transgression.

Figures 4-16b and 4-16c also show the distinct separation of the Brazos and Colorado sediments from the Mississippi sediments further to the east. A separate Mississippi source can also be seen in Figure 4-16a, for the coarser factions. The separate provenances for the surficial sediments are supported by the distribution of source specific heavy mineral assemblages as mapped by van Andel (1960) (Fig. 4-18).

The areal distribution of the Mississippi influence on the inner shelf of east Texas restricts the Mississippi delta influence to east of the Trinity/Sabine incised valley of Thomas (1990). This provides additional support for the Stage 5e, 5c and 5a delta lobes, shown in Figures 4-7, being of Brazos origin. The Brazos origin is already supported by the onshore correlation with the Brazos meanderbelts mapped by Van Siclen (1991), and this onshore extension is shown in Figure 4-7. However, the easternmost limit of the Stage 5 lobes lies outside of the grid of seismic data (Figs. 3-4a, 3-4b and 3-4c). It is, therefore, important that independent sedimentologic data place the spatial occurrence of the Stage 5 delta lobes (Figs. 3-4a, 3-4b, 3-4c and 4-7) within an exclusively Western Gulf Province, rather than a Mississippi Province (Fig. 4-18).

A comparison of Figure 4-7 and Figure 4-9a shows that the Stage 2 Brazos incisions occur along the western flank of the Stage 5e Brazos delta. A comparison of these two figures also indicates that it is likely that the Trinity/Sabine incised valley, mapped by Thomas (1990), is constrained by the eastern flank of the Stage 5e Brazos delta lobe. Paleogeomorphology
Figure 4-18 Heavy mineral provinces in the Northern Gulf of Mexico, from van Andel (1960) (after van Andel and Poole, 1960)
was, therefore, very influential in the location of incised valleys and in surficial sediment distribution.

The paleotopography of the east Texas continental shelf at the time of the maximum Stage 2 lowstand (18,000 ybp) was strongly influenced by the Stage 5 and Stage 3 delta lobes, and the Stage 2 incisions. The Stage 5 and Stage 3 delta lobes are shown together with the Stage 2 incisions in Figure 4-19. Superimposing the present-day bathymetry on to Figure 4-19 shows that remnants of the Stage 2 paleotopography are preserved in today's bathymetry, despite the subsequent effects of transgressive erosion (Fig. 4-20).

Figure 4-16d shows the distribution of grain-size modes less than 7.2 phi (0.0068 mm) from Curray (1960). This grain-size range includes very fine silts and clays. Figure 4-16d shows that the Mississippi River is the main source for these fine-grained sediments. In addition, their occurrence offshore Texas is restricted primarily to the outer shelf and upper slope. The east to west transport direction shown in Figure 4-16d for these Mississippi sediments is in keeping with the interpretation of fine-grained sediments prograding into the study area from the western Louisiana fluvial/deltaic system, previously studied by Sarzalejo (1993). This relationship is illustrated in Figures 4-10 and 4-15.

4.9 Sequence Stratigraphy

Seismic stratigraphy provides the basis for the recognition of unconformity bounded depositional sequences, and the criteria for seismic
Figure 4-19 Composite diagram showing all the major delta lobes (i.e., the Brazos and Colorado Stages 5 and 3) together with their Stage 2 incised valleys. Also shown is the Trinity/Sabine incised valley mapped by Thomas (1990) and Sarzalejo (1993).
Figure 4-20 Present-day bathymetry shown superimposed on the composite delta lobe diagram shown separately in Figure 4-19.
facies analysis (Vail et al., 1977). Sequence stratigraphy allows for the further subdivision of genetically related, unconformity bounded depositional sequences into systems tracts (Vail, 1987, Van Wagoner et al., 1988). Depositional sequences and systems tracts have been tied directly to relative sea level (Vail, 1987, Jervy, 1988, Posamentier et al., 1988, Posamentier and Vail, 1988). In addition, systems tracts have been associated with specific depositional environments and lithofacies, the suggestion being that sequence stratigraphy provides the framework for predictive stratigraphy (Vail, 1987).

Sequence stratigraphic methodology was developed using petroleum industry data sets that consisted of seismic data of much lower frequencies (10's of Hz) than the frequencies used in this study (100's to 1,000's of Hz). The chronostratigraphic control of this study (1,000's of years) is also better constrained than that of traditional industry studies (100 thousand's of years to 1,000,000's of years). In addition, although this study documents the evolution of fluvial/deltaic systems during very high-frequency, fourth-order (100,000 yr) and fifth-order (20,000 yr) cycles, the subjects of study are naturally occurring fluvial/deltaic systems. The results of this study provide useful insights into the factors that influence the timing and distribution of sedimentary facies over a period of 190,000 years and over a complete glacial-eustatic cycle.

The time frame of this study (190,000 years) also fills an important gap between the time frames of industry studies (100 thousand's of years to 1,000,000's of years) and those of laboratory scale (Posamentier, 1992, Wood et al., 1993, and Koss et al., 1994) that span hours to days. It is,
therefore, useful to document the results of this study in relation to the principles of sequence stratigraphy.

4.9.1 Lithofacies and Seismic Stratigraphy

Nine seismic facies units and their lithologies were identified in this study. The statements that follow summarize our ability to predict lithology from the seismic data. Examples of these seismic facies units are provided in Chapter 3.

i) Well defined, prograding and/or aggrading clinoforms (SFU 1 and SFU 9) are composed predominantly of clays with interlayered silts and occasional fine sands.

ii) Shingled clinoforms (SFU 2) prograding into shallow water are also composed predominantly of clay but show higher sand percentages.

iii) Parallel, near-horizontal seismic reflections are fine-grained prodelta deposits on the shelf (SFU 3), and fine-grained hemipelagic deposits on the upper slope (SFU 4).

iv) Chaotic seismic character on the upper slope is indicative of displaced shelf sediments that could be either clay rich (SFU 5), or sand rich (SFU 6), depending on the nature of the shelf-margin delta with which these deposits are associated, and whether, or not, the
slope deposits were fed directly by the fluvial system during the lowstand.

v) Chaotic seismic character within channels and incised valleys on the shelf indicates sand with occasional gravel (SFU 7a).

vi) Laminated onlap fill within channels indicates clay with occasional silt (SFU 7b).

vii) Mounded external forms with generally chaotic to slightly progradational clinoforms (SFU 8) are indicative of transgressive sand bodies and backstepping deltas that appear to be more wave-dominated. Both of these features are sand prone.

4.9.2 Ravinement Surfaces and Sequence Boundaries

The nature of the ravinement surface, relative to the sequence boundary, can be documented due to the chronologic control provided by the oxygen isotope stratigraphy and radiocarbon dates, and the paleoenvironmental data obtained from cores. The high-resolution seismic data provided the means to correlate these surfaces across the entire shelf.

Figure 4-21, a schematic north-south cross section, illustrates the relationship between the sequence boundary and the ravinement surface. The location of this cross section is shown in Figure 4-22. The southern end of the cross section starts at the shelf margin and the section crosses the Stage 2 Colorado incised valley. It continues landwards across the area
Figure 4-21 Schematic north-south cross-section summarizes the data that illustrates the time-transgressive nature of the Stage 2 to Stage 1 ravinement surface. The ravinement surface is dated at the outer shelf at 14,000 ybp and estimated to be 3,500 years old near today's shoreline. Along this transect, the ravinement surface can be seen to lie above Holocene coastal plain deposits that rest unconformably on top of the Stage 2 sequence boundary. The separation of the ravinement surface from the sequence boundary is accommodation dependent.
Figure 4-22 Location map for cores and schematic cross section shown in Figure 4-21. The line of section runs almost north-south from the present-day Colorado delta, to the position of the Stage 3 Colorado shelf-margin delta.

Line of section shown in Figure 4-21. Core locations for B151, B152, B148, and A24 also are shown.
between the Brazos and Colorado Stage 5e delta lobes and ends at the mouth of the modern Colorado River.

The Stage 2 sequence boundary is shown in Figure 4-21 as a thick, wavy line. The Stage 2 sequence boundary is coincident with today's subaerially exposed land surface, and can be traced southwards as the base of the incised valley, and the upper surface of the prograding Stage 3 shelf-margin delta. The sequence boundary can be traced further seawards, where its correlative conformity forms the seaward limit of the Stage 3 delta and a distinctive onlap surface (Fig. 4-21).

The ravinement surface is first documented as the planar surface that erodes the top of the Stage 3 shelf-margin deltas (Fig. 2-3). The Mississippi meltwater pulse in the oxygen isotope record for B146 dates the initiation of the ravinement surface at 14,000 ybp at this outer shelf site. At the shelf margin, the ravinement surface forms the upper surface of the Stage 3 deltas and can be considered to be coincident with the sequence boundary. This statement is not strictly correct, however, since the process of shoreface erosion has removed some of the topsets of the Stage 3 deltas and, therefore, has eroded down below the level of the previously subaerially exposed surface. Figure 4-21 shows this relationship.

Landward of the Stage 3 shelf-margin delta, the ravinement surface can be seen to occur above the bayhead delta deposits cored by B148. The Rangia cuneata shell found in B148 (Fig. 3-43b) was dated at 10,760 ybp, which places a maximum age for the ravinement surface at this location. Still further landward, the ravinement surface lies above the low-salinity marsh deposits cored by B151 and B152 (Figs. 2-12a and 2-12b). The marsh
deposits were dated at 9,530 ybp, which once again provides a maximum date for the ravinement surface at this location.

Near to today's shoreline, the ravinement surface becomes coincident with the development of the Stage 1 maximum flooding surface. The turnaround from transgression to regression will vary along the coast as a function of sediment supply, but can be considered to have occurred somewhere around 3,500 ybp for the Texas shoreline. For instance, Bernard and LeBlanc (1965) record the onset of progradation of Galveston Island as occurring at 3,500 ybp based on radiocarbon dates. This progradation would have postdated the turnaround of the Trinity bayhead delta, and also postdates the progradation of the Mississippi delta. Boyd et al., (1989) interpret the St. Bernard delta lobe as the first Holocene Mississippi highstand delta complex, and tentatively date the time of maximum flooding at 4,000 ybp. For the purpose of the schematic diagram (Fig. 4-21) the time of maximum flooding has been taken to be 3,500 ybp.

The ravinement surface can, therefore, be seen to be a time transgressive surface constrained by an outer-shelf date of 14,000 ybp, and constrained by successive inner-shelf dates of 10,760 ybp, 9,530 ybp, and 3,500 ybp. The ravinement surface can also be seen to initially lie above the sequence boundary when it forms the upper surface of the sedimentary wedge that onlaps the seaward extent of the Stage 3 deltas (Fig. 4-21). It then "merges" with the sequence boundary when it forms the upper erosional surface of the Stage 3 delta lobe (Fig. 2-3), and then lies above the sequence boundary as the retreating shoreline moved landward over coastal-plain deposits that aggraded in response to rising base level. Transgressive
erosion ceases at the time of maximum flooding. The relationship between
the ravinement surface and the subaerial unconformity was addressed by
Demarest and Kraft (1987) and has recently been summarized in a

It should be noted that the development of coastal-plain deposits
above the sequence boundary, and below the ravinement surface, is a
function of accommodation. The marsh deposits cored in B151 and B152
occur between the Stage 5e Brazos and Colorado delta lobes that formed
topographic highs; the Stage 3 and Stage 2 Brazos and Colorado Rivers
flowed between these highs (Fig. 4-19). During the Stage 2 to Stage 1
transgression, coastal plain aggradation was centered around the fluvial
systems within this relatively low area, and not on the crests of the Stage 5e
delta lobes.

Cross sections directly over the Stage 5e delta lobes would be quite
different from that shown in Figure 4-21. For instance, offshore of
Galveston Island the ravinement surface cuts across the top of the Stage 5e
Brazos delta lobe, resulting in the coincidence of the ravinement surface and
the top of the underlying Pleistocene clays. Siringan (1993) describes
numerous cores taken offshore Galveston Island that penetrated Holocene
muds that rest directly above stiff, reddish-grey mottled clays of the
Pleistocene Beaumont Formation.

The temporal and spatial distribution of delta lobes can be seen,
therefore, to determine the areas of maximum accommodation during the
subsequent transgression. The amount of accommodation, which is a
function of antecedent topography, subsidence rates, and the nature of
eustatic rise, will determine the degree to which the ravinement surface lies above the underlying sequence boundary. The separation of the ravinement surface from the sequence boundary is a necessary prerequisite for the preservation of coastal-plain deposits.

4.9.3 The Evolution of the Study Area Placed Within a Sequence Stratigraphic Framework

Figure 4-23 is a chronostratigraphic diagram for the study area. This figure represents an idealized dip section. In reality, the complexity of the distribution of delta lobes will result in considerable variability in dip-oriented cross sections. For instance, a dip section down the axis of the late Stage 3 Colorado delta lobe will show the entire Stage 5 interval as a condensed section comprised of prodelta clays (compare Figs. 4-12 and 4-13). In contrast, a dip section down the axis of the Stage 5 Brazos delta lobes will show three distinct delta lobes (Figs. 4-7 and 4-12), each separated by short periods of transgression. However, a section down the axis of the Stage 5 Brazos delta lobes will not cross the backstepping Brazos deltas that developed during the last transgression (Figs. 4-10 and 4-15), nor will it cross the preserved Holocene marsh deposits.

In Figure 4-23 the oxygen isotope curve is shown with horizontal time lines that separate periods of regression, transgression and sediment bypass. The oxygen isotope stages are also shown in Figure 4-23, however, the horizontal boundaries shown do not represent the Stage boundaries. The isotope stage numbers are shown only as a visual aid for the reader. The
Figure 4-23 Idealized chronostratigraphic diagram for the study area. The evolution of the Brazos and Colorado fluvial/deltaic systems is shown within the context of sequence stratigraphy. Systems tract boundaries are shown relative to the SPECMAP oxygen isotope curve. Note, however, that the horizontal lines in this figure separate the systems tracts and are not the isotope stage boundaries. The stages in the column next to the isotope curve are for visual reference only.
reader may refer to Imbrie et al. (1984) for the ages of the oxygen isotope stage boundaries.

Figure 4-24 shows a portion of seismic line G300X, which crosses the Stage 6 Brazos shelf-margin delta. Its aggradational to progradational geometry indicates that the delta was deposited during a period of gradual rise to gradual fall in relative sea level. The biostratigraphy and lithostratigraphy for Shell Eureka core 67-45 identifies a sandy, inner to middle-neritic interval overlying prodelta clays. These intervals are shown in Figure 4-24.

The top of the underlying prodelta clays is a strong seismic reflector that correlates updip with an erosional surface (Fig. 4-24). This erosional surface is interpreted as a ravinement surface that cuts into and across the Stage 6 sequence boundary (Fig. 4-24). The strong reflector at the top of the prodelta clays is interpreted to be the correlative conformity to the updip, erosional sequence boundary.

The overlying sandy inner to middle-neritic section does not onlap the Stage 6 sequence boundary. Instead, the updip segments of the clinoforms within this sandy interval are sub-parallel to the underlying prodelta deposits indicating continued progradation. This sandy, inner to middle-neritic interval is interpreted to represent a forced regression (Posamentier et al., 1992). The sandy sediments of the forced regression are associated, downdip, with chaotic events indicative of slumps and sediment bypass.

The aggradational to progradational clinoform configuration of the main Stage 6 Brazos shelf-margin delta (Fig. 4-24) indicates that it likely was deposited during the fifth-order highstand shown on the oxygen isotope
Figure 4-24 Seismic line G300X illustrates a forced regression during the deposition of the Stage 6 Brazos shelf-margin delta. The isotope curve is inset for reference.
curve inset in Figure 4-24. The upper surface of the sandy, shallower-water interval is approximately 20 ms (15 m) lower than the top of the main Stage 6 delta lobe (Fig. 4-24). The top of the main Stage 6 delta lobe is a near-horizontal, planar, erosional surface that is interpreted to be a ravinement surface. If we make the assumption that shoreface erosion removed 7-10 m of sediment from the top of the Stage 6 delta lobe, then the fall in sea level that gave rise to the forced regression was on the order of 22-25 meters. An examination of the depth-converted SPECMAP curve (Fig. 4-1) shows that a 22-25 m sea-level fall is consistent with the straight line segment of the SPECMAP curve as sea level fell to the Stage 6 maximum lowstand. In addition, the minor rise that followed the maximum lowstand can be seen in Figure 4-24 as a minor period of onlap that preceded the deposition of the Peorian sediments. The Peorian interval is one of rising sea level, and the Peorian sediments initially onlap, and subsequently cover, the sandy deposits of the forced regression (Fig. 4-24).

The extended duration of the maximum lowstand during Stage 6 versus Stage 2 (Fig. 4-1) appears to have been one of the factors responsible for the forced regression during Stage 6. The other factor may have been the fact that sea level did not fall quite as low during the Stage 6 lowstand as it did during Stage 2. The rapid rate, brief duration and lower level of the Stage 2 sea-level fall resulted in sediment bypass rather than the development of attached lowstands.

The Stage 5 interval can be separated into three periods of regressive delta-lobe development separated by minor transgressions (Fig. 4-23). Each successive delta lobe is not interpreted as a lowstand deposit associated with
a forced regression, but rather as a fifth-order highstand deposit. This follows from the fact that successive lobes are not initiated by falling sea levels and do not rest directly on the sequence boundary of the underlying lobe. Instead, each successive delta lobe is preceded by an interval of transgression that results in onlap and transgression of the previous delta lobe (Fig. 3-2) and creates bay-head delta environments within the older fluvial channels (Fig. 3-43a). The prograding delta lobe, therefore, downlaps a maximum flooding surface associated with a fifth-order eustatic cycle rather than a sequence boundary.

Each of the Stage 5 highstand deltas continue to prograde up to the point of their maximum lowstand. Progradation, therefore, continues during the sea-level falls from Stages 5e to 5d, 5c to 5b, and possibly from 5a to 4, although isolated Stage 4 deposits do occur in one part of the study area. These possible Stage 4 deposits are shown on seismic line G400X (Plate 2) and may owe their presence to the increased accommodation due to the growth fault. Elsewhere, deposition during these fifth-order lowstands (Stages 5d, 5b and 4) was occurring on the shelf which indicates that sea level never fell below the level of the shelf edge. Within these stratigraphic intervals, the lowstand deposits appear to be attached to the higher-order highstand deposits and have not been separated out in the chronostratigraphic chart (Fig. 4-23). The Stage 5 interval is, therefore, illustrated as being comprised of highstand and transgressive systems tracts with the proviso that the latter parts of the highstands may include a brief period of forced regression (Posamentier et al., 1992).
The Stage 3 deltas are also preceded by a period of transgression (Fig. 4-23). The increased accommodation results in thicker transgressive and regressive deposits than those on the inner shelf, and the maximum flooding surface associated with this fifth-order cycle (the Stage 3 mfs) is easily recognized on the seismic data (seismic line G300X, Plate 2). Although the deposition of "lowstand" shelf-margin deltas is occurring during Stage 3, this interval is identified as a highstand systems tract in Figure 4-23. This is due to the fact that the interval is characterized by an aggradational to progradational clinoform package that downlap a maximum flooding surface. This is a highstand geometry associated with a fifth-order eustatic cycle.

The Stage 3 to Stage 2 fall in sea level resulted in exposure of the shelf edge and sediment bypass to the East Breaks Slide and to upper slope salt withdrawal basins. Deeply incised valleys and sediment bypass characterize the Stage 2 sea-level fall, which is interpreted as a fourth-order fall related to the 100,000 year Pleistocene glacio-eustatic cycles. The subsequent Stage 2 to Stage 1 transgression resulted in backstepped deltas and coastal plain aggradation (Figs. 4-10, 4-15, 4-21, 2-12a and 2-12b). It was also the time during which most of the incised-valley fill was deposited (Figs. 4-21 and 3-28a).

The Stage 1 maximum flooding is considered to have occurred along the Texas coast at approximately 3,500 ybp, based on radiocarbon dating of the onset of progradation of Galveston Island (Bernard and LeBlanc, 1965). The Stage 1 highstand systems tract is, therefore, considered to have its base at 3,500 ybp.
CONCLUSIONS

The main conclusions from this study are listed, below.

1) The temporal and spatial distribution of facies within the Brazos and Colorado fluvial/deltaic systems are controlled by fourth-order (100,000 yr) and fifth-order (20,000 yr) eustatic cycles.

2) Seismic facies analysis provides a means of predicting lithologies within the study area, and the following correlations exist.

   i) Well defined, prograding and/or aggrading clinoforms (SFU 1 and SFU 9) are composed predominantly of clays with interlayered silts and occasional fine sands.

   ii) Shingled clinoforms (SFU 2) prograding into shallow water are also composed predominantly of clay but show higher sand percentages.

   iii) Parallel, near-horizontal seismic reflections are fine-grained prodelta deposits on the shelf (SFU 3), and fine-grained hemipelagic deposits on the upper slope (SFU 4).

   iv) Chaotic seismic character on the upper slope is indicative of
displaced shelf sediments that could be either clay rich (SFU 5), or sand rich (SFU 6), depending on the nature of the shelf-margin delta with which these deposits are associated, and whether, or not, the slope deposits were fed directly by the fluvial system during the lowstand.

v) Chaotic seismic character within channels and incised valleys on the shelf indicates sand with occasional gravel (SFU 7a).

vi) Laminated onlap fill within channels indicates clay with occasional silt (SFU 7b).

vii) Mounded external forms with generally chaotic to slightly progradational clinoforms (SFU 8) are indicative of transgressive sand bodies and backstepping deltas that appear to be more wave-dominated. Both of these features are sand prone.

3) A forced regression occurred during the deposition of the Stage 6 Brazos shelf-margin delta.

4) The Brazos fluvial/deltaic system deposited deltas with highstand geometries during Stages 5e, 5c, 5a, and 3.

5) The oxygen isotope Stage 3 maximum highstand is estimated to have been 33 m below today’s sea level.
6) The Stage 2 lowstand is associated with a fourth-order (100,000 yr) eustatic cycle and was a time of major sediment bypass to the upper slope.

7) During the Stage 2 maximum lowstand the Brazos River flowed eastwards into the adjacent Trinity/Sabine incised-valley system, and the Colorado River actively fed the East Breaks Slide.

8) Almost all of the sediments that infill the Stage 2 incised valleys were deposited during the subsequent transgression.

9) The Mississippi meltwater pulse of Kennett and Shackleton (1975) and Leventer et al. (1982) was recorded in an isotope record for the core B146 using the benthic foraminifer *Quinqueloculina sp.*

10) The time transgressive nature of the ravinement surface and its position relative to the sequence boundary has been documented. Provided that there is sufficient accommodation, the ravinement surface can cut across the top of Holocene coastal-plain deposits that rest unconformably above Pleistocene clays.

11) A comparison of the Brazos and Colorado fluvial/deltaic systems with the western Louisiana fluvial/deltaic system, previously studied by Sarzalejo (1993), highlights the effect of sediment supply on facies distribution. The higher sediment supply western Louisiana fluvial/deltaic system continued
to prograde during the Stage 2 to Stage 1 transgression. During the same time interval, the Brazos fluvial/deltaic system developed a series of backstepping deltas on the inner shelf. The Colorado delta also backstepped and infilled a number of fluvial channels on the shelf.

12) Finally, the study has re-emphasized the three-dimensional nature of the distribution of sedimentary facies and illustrates the variability that can occur between adjacent, dip-oriented, transects. Reservoir prediction is dependent on understanding the regional evolution of fluvial/deltaic systems through time.
CHAPTER 5 - AN EXPERIMENTAL, HIGH-RESOLUTION 3-D SEISMIC SURVEY

Introduction

An experimental, high-resolution 3-D seismic survey was successfully acquired over a 1.5 square kilometer section of an incised fluvial valley. The data were acquired as a near-zero offset, single-channel survey using a 15 cubic inch water gun as the source, and differential GPS for navigation and positioning. The objective was to acquire a 3-D seismic data volume suitable for calculating the volume of shallow sand deposits. Horizontal time sections from the 3-D volume clearly show the flanks of the incised valley, as well as high-amplitude reflections interpreted as coarse-grained channel-lag deposits. The volume of this lag deposit can be calculated using the combination of the horizontal and vertical sections from the high-resolution 3-D seismic data set.

5.1 Objectives and Background

In recent years, there has been increased interest in the commercial value of shallow sand and gravel deposits on continental shelves. Shallow, nearshore sand deposits are used to replenish beach sands removed by coastal erosion. Sands and gravels can be used as aggregate in the construction industry and, in the case of placer deposits, they are the site of heavy minerals. These deposits can be imaged using high-resolution
seismic data (Anderson et al., 1994) and successful mining will depend on accurate estimates of their volume.

The objective of this project was to acquire a low cost, high-resolution 3-D seismic survey suitable for volumetric calculations. In this experiment we applied 3-D seismic methods to acquire a three-dimensional seismic data volume of an incised valley, and its coarse-grained channel lag. Time slices provide continuous horizontal images of the subsurface and can therefore be used to estimate the areal extent of a deposit. When the time slices are combined with the vertical sections, the economic potential of a deposit can be determined without the expense of extensive coring.

The incised valley, shown in Figure 5-1, was selected as the target for three reasons. The first and most important was that it was a large feature, over 500 meters wide and 30 meters deep. The size of the feature increased the likelihood that we would obtain an image for analysis. The second reason was that it was located in deep enough water (35 meters) to ensure that the water-bottom multiple would not interfere with the target. The third reason was that the valley fill, previously imaged using the EG&G Uniboom (300-2400 Hz), showed contrasting seismic character, i.e., a laminated upper fill versus a basal chaotic fill (Fig. 5-2). The Uniboom line therefore provided us with an upper level benchmark as far as vertical resolution was concerned.

The size of the target, and the questions to be addressed by the 3-D volume are critical to the survey design. On one end of the spectrum is the need for thin beds to be resolved, and this requires the use of high
Figure 5-1 Site of the experimental 3-D seismic survey
frequencies (Sheriff, 1985). On the other end of the spectrum are the constraints placed on spatial sampling as a result of the cost and logistics involved in offshore navigation and positioning. Very high frequencies require very small sampling intervals, not only in time, but also in space (Gardner, 1993). The decimeter levels of accuracy required to position source and receivers in a marine survey may be available, but the cost would be prohibitive for the assessment of sand and gravel deposits.

5.2 Survey Design and Data Acquisition

5.2.1 Temporal and Spatial Sampling

Numerous 3-D seismic surveys have been successfully acquired by the petroleum industry, over the last 17 years (Nestvold, 1992). However, it has only been since the introduction of digital seismic data acquisition systems capable of sampling frequencies in the 1000's of Hz (Lericolais et al., 1990, Lericolais et al., 1991) that truly high-resolution 3-D seismic surveys have become possible.

The spatial sampling interval was determined to a large degree by field conditions. The marine environment requires dynamic positioning and therefore surveying at sea has its unique challenges. Difficulties in steering a vessel in a straight line coupled with the inherent sideways motion of ship-based antennae relative to source and receiver locations, would require the use of an extremely accurate tracking system to achieve decimeter accuracy. We therefore considered it impractical to attempt to obtain the level of vertical resolution seen on the Uniboom line. The
importance of spatial sampling in 3-D seismic surveying is demonstrated by the comparison of two high-resolution 3-D surveys, one from Belgium and the other off New Jersey.

The Belgian survey was carried out in the river Scheldt, and targeted a shallow shale diapir (Henriet et al., 1992). The area covered was 50 x 180 square meters, with a bin size of 1 x 1 square meters, and source frequencies from a few hundred to more than 2000 Hz. Positioning was carried out using a shore-based, laser auto-tracking theodolite with decimeter accuracy. The survey was very successful. The target was adequately sampled and the diapir is clearly visible in the horizontal time slices (Fig. 5-3).

A high-resolution 3-D seismic survey was also carried out offshore New Jersey (Davies et al., 1992). The objective of the survey was to investigate changes in the regional acoustic stratigraphy within a 0.5 x 5 square kilometer area of the outer shelf. The data were acquired using frequencies of 500 - 3500 Hz and were recorded with a bin size of 2.5 x 10 square meters, and processed using 2.5 x 18 square meter bins. Positioning accuracy of less than 5 meters was achieved using a commercial satellite positioning system. The results of the New Jersey survey are shown in Figure 5-4. Although details are visible in the vertical section, the horizontal section is very noisy. This is mainly due to the large bin size relative to the frequencies recorded. The use of a seismic source with frequencies from 500 - 3500 Hz requires very small bins to ensure adequate sampling.
Figure 5-3 Horizontal time slices through a shallow shale diapir in the river Scheldt, Belgium (from Henri et al., 1992)
Figure 5-4 Horizontal and vertical sections through a 3-D seismic volume recorded offshore New Jersey (from Davies et al., 1992).
5.2.2 Data Acquisition

Our experimental survey was based on the principle that short offset 3-D seismic surveys are capable of yielding high data quality, despite being acquired at low fold (Newman, 1984, Gardner, 1993). The low fold simplifies data acquisition and reduces the volume of data to be processed, thereby reducing the survey costs.

The survey was recorded using the dual-channel Delph 2 acquisition system. Two single-channel, 10 element, hydrophone arrays were towed 26 meters apart, using outriggers. The seismic source, a 15 cubic inch SSI water gun (40 - 2000 Hz), was towed directly behind the vessel. This configuration allowed us to record two single-channel seismic lines, spaced 13 meters apart, with each pass of the vessel (Fig. 5-5). The vessel tracklines were pre-plotted at 25 meter spacings. The data were recorded at 1/4 ms sample interval allowing adequate recording over the bandwidth of the source. Shots were fired every 3 seconds which approximated to a shot interval of 6 meters for a constant ship speed of 4 knots.

Navigation and positioning were conducted using a Trimble 4000 DL II GPS system with accompanying Navbeacon that provided differential GPS co-ordinates with 1-3 meter accuracy. The GPS antenna was attached to an 'A' frame over the stern of the vessel, and the antenna position was written to the header of each seismic trace.
Figure 5-5 Post plot of the seismic track lines
5.3 Data Processing

Prior to 3-D processing, the data were reformatted from the Delph 2, Elics format, to SEGY. The antenna positions, recorded in the trace headers were then interpolated since our navigation program had introduced a delay that prevented the updating of the trace position at every shot. The common mid-point position (CMP) for each trace was then computed and this value was written into the trace header. Noisy traces were removed at this stage and the data set, comprising 17,765 traces, was processed at the Houston Advanced Research Center (HARC) using Landmark’s ITA data processing package.

The 3-D data processing included the computation of shot and receiver positions, gain equalization (to compensate for differences in hydrophone response), gain recovery, 3-D binning, stacking, filtering and display. Displays were generated using Silicon Graphics' GeoVoxel View.

5.4 Results

A total of 60 2-D seismic lines, each a minimum of 2 kilometers long, were recorded over an area of 0.75 x 2 square kilometers. Figure 5-5 shows a post plot of the line locations. The lines were recorded along a north-south grid. Two receiver lines can be seen for each shot line. Holes in the survey, i.e., where empty bins occur, were not excessive but did develop whenever the boat moved in opposite directions at the same latitude. On the other hand, overlap also occurred, increasing the fold of
some bins up to a maximum of four fold. Figure 5-6 shows the fold distribution when the data were binned using a bin size of 15 x 15 square meters. To reduce the number of holes in the data, the survey was eventually re-binned using a bin size of 25 x 25 square meters (Fig. 5-7).

The effect of the larger bin size reduces the holes in the final 3-D volume, but also smears the events since traces are being summed over a larger subsurface area. Fewer holes results in improved horizontal time slices, but smearing decreases the vertical resolution. Figure 5-8 shows a vertical section from the 3-D volume. The southern flank and the base of the channel are clearly seen, but the internal stratification of the channel fill has been lost to the binning process. Figure 5-9 shows a comparable line before binning. In this case the shot spacing is at the original spacing of approximately 6 meters, and the internal stratification of the valley fill is visible.

Binning allows us to produce a fairly complete 3-D volume, and to view the data as continuous horizontal sections, or time slices. Slices can be made at 0.25 ms time increments and we can see changes in the geometry of the channel, and in the amplitude of the valley fill. Figure 5-10 is a horizontal slice across the volume at 55.25 ms, near the top of the valley. The display is an amplitude display and the high amplitude reflections from the walls of the valley are shown in black. The location of the valley across the survey area is clearly seen.

Figure 5-11 shows a horizontal time slice near to the base of the valley at approximately 84 ms. In Figure 5-11, the channel fill can be seen as the high-amplitude white fill occupying the area between the valley sides
Figure 5-6 A fold distribution map of the survey area with the data binned with 15 X 15 square meter bins
Figure 5-7 A fold distribution map of the survey area with the data binned with 25 X 25 square meter bins
Figure 5-8 A vertical, in-line section selected from the 3-D volume, and located close to the eastern edge of the survey area
Horizontal Time Slice
@ 55.25 ms (221 samples)

Figure 5-10 Horizontal section through the 3-D volume at 55.25 ms
Horizontal Time Slice
@ 84 ms (336 samples)

Figure 5-11 Horizontal section near to the base of the valley at 84 ms
seen as high-amplitude black events in Figures 5-11 as well as in Figure 5-10. The high amplitude white events are interpreted to be due to a basal channel lag consisting of coarse-grained sands and gravels.

The volume can also be viewed as an entire cube, as shown in Figure 5-12. In Figure 5-12, the channel can be seen flowing out towards the viewer, i.e. out through the southeastern corner of the 3-D volume.

5.5 Conclusions and Recommendations

In spite of the data holes that occurred during acquisition, the experiment was very successful. We have been able to acquire an inexpensive, high-resolution 3-D seismic data volume. The survey was conducted using a PC-based acquisition system, and differential GPS. Navigational accuracy of 1-3 meters was more than adequate, but gaps will occur due to the natural movement of a vessel at sea. The latter could be compensated for by shooting infill lines where required. In this particular case, channel-fill deposits can be accurately mapped and, because the areal extent is now known, the volume of the deposit can be calculated.

Constraints in dynamic positioning dictate bin sizes that result in reduced resolution when compared to high-resolution 2-D data. High-resolution 2-D seismic data, acquired during the 3-D data acquisition, can provide the vertical resolution lacking in the vertical 3-D sections. However, any further increase in resolution will require the use of seismic sources such as the Uniboom that typically do not contain the lower frequencies needed for this type of 3-D survey. In addition, the higher
Figure 5-12 The data presented as an entire 3-D volume or cube
frequencies demand submeter positioning accuracy which at this time is impractical for low cost marine 3-D seismic surveys.

Attempts should be made in future surveys to build some redundancy into the trackline program. The overlap will help compensate for the inevitable data holes. The data set can then be processed with a smaller bin size, which will greatly enhance the resolution. Using hydrophone arrays with identical responses will also improve the data quality, which suffered as a result of the dampened response of one of the arrays.
REFERENCES;


Anderson, J.B., and M.A. Thomas, 1991, Marine ice-sheet decoupling as a mechanism for rapid, episodic sea-level change: the record of such events and their influence on sedimentation: Sedimentary Geology, v. 70, p. 87-104.


Bartek, L.R., J.B. Anderson, and K.C. Abdulah, 1990, The importance of overstepped deltas and "interfluvial" sedimentation in the


New $^{230}$Th/$^{234}$U dates from the Huon Peninsula, New Guinea: Quaternary Research, v. 4, p. 185-205.


Blum, M.D., and D.M. Price, 1994, Glacio-Eustatic and Climate Controls on Quaternary Alluvial Plain Deposition, Texas Coastal Plain, Gulf Coast Association of Geological Societies Annual Convention, Transactions 44, Austin, Texas, p. 85-92.


Dengo, G., 1985, Mid America: Tectonic setting for the Pacific margin from southern Mexico to northwestern Columbia, in A.E.M. Nairn, F.F.


Fisk, H.N., 1944, Geological investigation of the alluvial valley of the lower Mississippi River: Mississippi River Commission,


Frazier, D.E., 1974, Depositional episodes: Their relationship to the Quaternary stratigraphic framework in the Northwest portion of the Gulf of Mexico: Geological Circular 74-1, Texas Bureau of Economic Geology,


Jervey, M.T., 1988, Quantitative geological modeling of siliciclastic rock sequences and their seismic expression, in C.K. Wilgus, B.S.

Joyce, J.E., L.R.C. Tjalsma, and J.P. Prutzman, 1993, North American glacial meltwater history for the past 2.3 m.y.: Oxygen isotope evidence from the Gulf of Mexico: Geology, v. 21, p. 483-486.


Recent Sediments, Northwest Gulf of Mexico: The American Association of Petroleum Geologists, Tulsa, Oklahoma, p. 267-301


Geology No. 27, The American Association of Petroleum Geologists, Tulsa, Oklahoma, p. 1-10


APPENDIX 1  Borehole and Core Data
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**Average Velocity Core #6**

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### Average Velocity Core #8

![Average Velocity Core #8](image1)

### Time Depth Curve - Core #8

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<td>5201.11</td>
<td>288.4</td>
<td>750</td>
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<tr>
<td>1936</td>
<td>980</td>
<td>956</td>
<td>206</td>
<td>0.1789</td>
<td>0.0347</td>
<td>5936.60</td>
<td>5343.77</td>
<td>357.8</td>
<td>956</td>
</tr>
</tbody>
</table>

**Average Velocity Core #10**

![Average Velocity Graph](image1)

**Time Depth Curve - Core #10**

![Time Depth Curve Graph](image2)
APPENDIX 3  Oxygen Isotope Data (B146)
### Oxygen isotope results for B146 (Quinqueloculina sp)

<table>
<thead>
<tr>
<th>Depth (ft)</th>
<th>d18O</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>-0.205</td>
</tr>
<tr>
<td>2</td>
<td>0.655</td>
</tr>
<tr>
<td>3</td>
<td>0.525</td>
</tr>
<tr>
<td>4</td>
<td>0.134</td>
</tr>
<tr>
<td>5</td>
<td>-0.984</td>
</tr>
<tr>
<td>6</td>
<td>0.245</td>
</tr>
<tr>
<td>7</td>
<td>-3.11</td>
</tr>
<tr>
<td>8</td>
<td>-2.777</td>
</tr>
<tr>
<td>9</td>
<td>-5.012</td>
</tr>
<tr>
<td>10</td>
<td>163.5</td>
</tr>
<tr>
<td>11</td>
<td>164.0</td>
</tr>
<tr>
<td>12</td>
<td>183.0</td>
</tr>
<tr>
<td>13</td>
<td>184.0</td>
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<tr>
<td>14</td>
<td>193.0</td>
</tr>
<tr>
<td>15</td>
<td>194.0</td>
</tr>
<tr>
<td>16</td>
<td>194.0</td>
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<tr>
<td>17</td>
<td>203.5</td>
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<tr>
<td>18</td>
<td>212.5</td>
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<tr>
<td>19</td>
<td>213.5</td>
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<tr>
<td>20</td>
<td>214.0</td>
</tr>
<tr>
<td>21</td>
<td>222.5</td>
</tr>
<tr>
<td>22</td>
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<td>23</td>
<td>253.0</td>
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<td>26</td>
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<tr>
<td>27</td>
<td>283.0</td>
</tr>
<tr>
<td>28</td>
<td>284.0</td>
</tr>
</tbody>
</table>

### Oxygen isotope results for B146 (Globigerinoides ruber)

<table>
<thead>
<tr>
<th>Depth (ft)</th>
<th>d18O</th>
</tr>
</thead>
<tbody>
<tr>
<td>1.5</td>
<td>-1.989</td>
</tr>
<tr>
<td>2</td>
<td>-1.421</td>
</tr>
<tr>
<td>3</td>
<td>-1.411</td>
</tr>
<tr>
<td>4</td>
<td>-1.887</td>
</tr>
<tr>
<td>5</td>
<td>-2.314</td>
</tr>
<tr>
<td>6</td>
<td>-2.376</td>
</tr>
<tr>
<td>7</td>
<td>-3.926</td>
</tr>
</tbody>
</table>
APPENDIX 4  Radiocarbon dates
### Summary of radiocarbon data (analyses by Beta Analytical Inc.)

<table>
<thead>
<tr>
<th>Lab Number</th>
<th>Sample No.</th>
<th>C-14 Age ybp</th>
<th>C13/C12</th>
<th>C13 Adjusted Age ybp</th>
</tr>
</thead>
<tbody>
<tr>
<td>Beta-65492</td>
<td>B148-S7 shell</td>
<td>10420 +/- 110</td>
<td>-4.2 o/oo</td>
<td>10760 +/- 110</td>
</tr>
<tr>
<td>Beta 65493</td>
<td>B149-S19 shell</td>
<td>Greater than 43030</td>
<td>+0.3 o/oo</td>
<td>Greater than 43540</td>
</tr>
<tr>
<td>Beta 65494</td>
<td>B151-S19S20 sediment</td>
<td>9480 +/- 60</td>
<td>-21.8 o/oo</td>
<td>9530 +/- 60</td>
</tr>
<tr>
<td>Beta 65495</td>
<td>G148-S12 shell</td>
<td>Greater than 45100</td>
<td>-4.9 o/oo</td>
<td>Greater than 45430</td>
</tr>
</tbody>
</table>
PLEASE NOTE:

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

LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

The following map or chart has been refilmed in its entirety at the end of this dissertation (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.
clinoforms show downlap above Stage 4 sequence boundary
Top of Stage 5e Brazos Delta lobe
(for detail see Figs. 2 - 5)
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SOUTH
Plate 2
Seismic lines
G400X
G300X
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young prograding Western Louisiana sediments from Mississippi Province (see Figs. 4-16d and 4-18)

maximun flooding surface
Plate 3
Seismic line
R93-51
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Stage 2 to Stage 1 ravine surface near seafloor at this location but is buried below thicker sediments towards the east (see line R92-24 below and R93-51, Plate 3).
convex upwards wedge of sediment develops into overlying wedge of clinoforms towards east (see line R92-24, below)
Plate 4

Seismic lines
R91-8
R92-24
Stage 3 Brazos Delta

Stage 5a Brazos Delta

Top of Stage 5a Brazos Delta lobe

event horizon surface
Plate 1

Seismic lines
R91-4
G380X
Plate 3
Seismic line
R93-51

young prograding Western Louisiana sediments
from Mississippi Province (see Figs. 4-161 and 4-162)
Stage 2 to Stage 1 erosion surface near seafloor at this location but is buried below thicker sediments, towards the east (see line R92-24 below and R91-8, Plate 3)