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LATE QUATERNARY GLACIAL HISTORY OF THE PENNELL COAST REGION, ANTARCTICA, WITH IMPLICATIONS FOR SEA-LEVEL CHANGE AND CONTROLS ON ICE SHEET BEHAVIOR; AND, LATE QUATERNARY STRATIGRAPHIC EVOLUTION OF THE WEST LOUISIANA CONTINENTAL SHELF

by

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

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ABSTRACT

Late Quaternary Glacial History of the Pennell Coast Region, Antarctica, with Implications for Sea-Level Change and Controls on Ice Sheet Behavior; and, Late Quaternary Stratigraphic Evolution of the West Louisiana Continental Shelf

by

Julia Smith Wellner

The Pennell Coast continental shelf is isolated from West Antarctic Ice Sheet drainage; for an ice sheet to ground in this region it must flow over the Transantarctic Mountains from East Antarctica. Features observed indicate that ice grounded on the Pennell shelf. Cores from the shelf sampled till, a pelleted unit, glacial-marine sediments, contourite deposits, and diatomaceous muds. The timing of ice sheet grounding is revealed by radiocarbon dates that indicate the ice sheet was grounded on the shelf during the Last Glacial Maximum and has a retreat history that differs from nearby drainage areas. Comparison to sea-level curves suggests that melting ice from the region contributed to the Holocene sea-level rise and that formation of that ice contributed to the fall in sea level immediately prior to the Last Glacial Maximum.

Comparison between the Pennell Coast and drainage outlets of the West Antarctic Ice Sheet allows examination of controls on ice sheet behavior. There is a consistent pattern of erosional features on the crystalline bedrock of the inner shelf, mega-scale glacial lineations on the sedimentary strata of the outer shelf, and drumlins between the
two. The troughs in the areas of sedimentary substrate are interpreted to have been occupied by fast-flowing ice and those in the areas of crystalline substrate by slower-moving ice. The Pennell shelf differs in that it has both crystalline and sedimentary substrates but no drumlins or lineations. Possible reasons for this difference include the size of the drainage basin, the narrow continental shelf, and the high sand content of the tills.

Core and seismic data were used to conduct an analysis of the west Louisiana outer shelf depositional systems formed during the last glacial-eustatic cycle. Differences in deltaic deposition in the area illustrate the complex relationship between depositional patterns and sea-level change. Particularly salient is the difference between the two primary sequence boundaries. The oldest sequence boundary is a major erosional surface. The youngest sequence boundary is characterized by much smaller channels and is primarily an interfluve feature. The observed variations in each system can be used to refine sequence-stratigraphic models.
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PREFACE

This dissertation consists of five chapters that cover two different research projects. Chapter One is an introduction to the Antarctic research. Chapters Two and Three are results from projects conducted in the Antarctic. Chapter Four is an introduction to the Gulf of Mexico project, similar to Chapter One. Chapters Five is results from the Gulf of Mexico study area. The references for each section are combined.
CHAPTER 1: INTRODUCTION TO ANTARCTIC STUDY AREA

Chapter Context

This chapter presents an introduction to the study area and a general overview of the geography, geology, and glacial setting of the area. It also presents the motivation for this study.

Introduction

The Antarctic ice sheet currently has a volume of $30 \times 10^6$ km$^3$, covering an area of $13.6 \times 10^6$ km$^2$ and is divided into two parts by the Transantarctic Mountains (TAM) (Fig. 1-1; Denton et al., 1991). The West Antarctic Ice Sheet (WAIS) contains the ice equivalent of 6 m of global sea-level rise. The WAIS rests primarily below sea level and is a marine ice sheet (Fig. 1-2); for this reason it is considered to be the less stable of the two ice sheets (Hughes, 1973). The base of the East Antarctic Ice Sheet (EAIS) is primarily above sea level and it is a terrestrial ice sheet (Fig. 1-2); the EAIS contains the ice equivalent of about 60 m of global sea-level rise. An understanding of how these two ice sheets have behaved is critical to our ability to predict how they may behave in the future and how they may affect global sea-level change in the next centuries.
Figure 1-1. Satellite image of the Antarctic content. Major areas that are discussed in the text are labeled. Image from the USGS TerraServer.
Figure 1-2. Profiles of the West Antarctic and East Antarctic ice sheets illustrating the differences between the marine and terrestrial based ice sheets. Locations shown in Figure 1-1. Adapted from Bentley, 1964.
The primary study area of this project is the Pennell Coast, North Victoria Land (Fig. 1-1) which drains part of the EAIS. The first person to spend an austral winter on the Antarctic continent stayed on the Pennell Coast. A Norwegian, Carsten Borchgrevink, and 30 members of his crew spent the winter of 1899 at Ridley Beach, on the west side of Cape Adare, in two huts that still stand. Cape Adare was also visited by a field party during R. F. Scott’s 1910 expedition. Despite these early explorations, the area offshore of the Pennell Coast remained poorly studied, even for an Antarctic coastline. The first scientific exploration of the offshore region was conducted as part of the Deep Freeze 1980 expedition (Brake and Anderson, 1983). Following that cruise, the only other scientific study of the continental shelf of the region was the cruise conducted in 1998 as part of this study. One of the reasons for the paucity of ships’ tracks through the region is the heavy sea ice that often covers the area (Figs. 1-3).

The Antarctic continent consists of the relatively stable East Antarctic craton and several smaller land masses that compose West Antarctica. The study area is on the extinct plate boundary between the east and west parts of the Antarctic continent (Fitzgerald and Stump, 1997; Cande et al., 2000). The TAM have continued to rise during the late Neogene at a rate of about 100 m per million years based on apatite fission track data (Ishman and Rieck, 1992). The fact that the area is still active tectonically makes estimates of glacial extent based on isostatic rebound curves difficult.
Figure 1-3. Satellite image of North Victoria Land region showing sea ice in the Ross Sea and along the Pennell Coast margin. The image was made in January 1998. Lighter gray shades in the water indicate ice.
The onshore area of the Pennell Coast is dominated by valley glaciers that flow directly to the sea from the northern limit of the TAM (Fig. 1-4). Only an expanded EAIS that flowed through and likely over these mountains could result in grounded ice on the Pennell shelf. The offshore region is characterized by rugged bathymetry and transverse troughs (Fig. 1-5), which suggest that ice has grounded on the shelf. However, subglacial deposits were not previously recovered from the region (Brake, 1982). Modern sedimentary deposits on the continental shelf are dominated by volcanic sands that have been sorted by marine currents and by diatomaceous deposits that indicate high productivity (Brake, 1982).

**Glacial and Climate History**

Controversy remains about the timing of the origin and the stability of the EAIS through the Cenozoic. The EAIS began to form prior to the Oligocene and achieved continental scale by the middle Miocene (Abreu and Anderson, 1998; Wilson et al., 1998; Anderson, 1999; Bart and Anderson, 2000). Antarctic glaciation was made possible in part by the separation of Australia and Antarctica and the subsequent isolation and cooling of the Antarctic continent near the pole (Kennett, 1977). Outlet glaciers began to breach the TAM and ground in the Ross Sea at the end of the Oligocene (Wilson et al., 1998). Some researches have argued for significant reduction of the EAIS during the Pliocene (e.g., Ishman and Rieck, 1992), but mounting evidence suggests that no major reduction occurred in the Neogene (Stroeven and Prentice, 1997; Harvey et al., 1998).
Figure 1-4. Antarctic Sketch Map of North Victoria Land showing the mountains and valley glaciers that drain into the Pennell shelf region.
Figure 1-5. Bathymetry of the Pennell Coast continental shelf based on this study.
The type of ice sheet in Victoria Land has changed in the Neogene. Studies of landscape evolution with age control provided by dates from volcanic rocks indicate that the glaciers in the TAM were warm-based until about 8 million years ago and since then the region has been dominated by cold-based glaciers (Armienti and Baroni, 1999).

**Research Goals**

The main purpose of this study is to determine the recent glacial history offshore of the Pennell Coast. The Pennell shelf is important because of its position relative to the EAIS and the TAM. Grounded ice on the Pennell shelf must come from ice flowing through or over the TAM. Only an expanded EAIS could account for ice flowing through or over the TAM. Thus, the Pennell continental shelf can act as a ‘dipstick’ for a much larger area of the EAIS. The rugged bathymetry and transverse troughs on the continental shelf suggest that ice grounded in the area at some point (Brake, 1982). The goals of this project are to map the maximum grounding line reached on the shelf, to sample subglacial sediments from the area, and to use radiocarbon dating to constrain the time period during which the ice sheet was grounded on the shelf. The results of this project can be used to help constrain models of ice sheet extent and changes in global sea level (Fig. 1-6).
Figure 1-6. Sea-level curve for the last glacial-eustatic cycle. Modified from Bard et al., 1990.
Geophysical and geologic datasets were collected from the RV IB Nathaniel B. Palmer in 1998. Chapter 2 presents results from the Pennell region with an emphasis on radiocarbon dates and timing of ice sheet advance and retreat. Chapter 3 presents a comparison between the Pennell study area and drainage basins of the West Antarctic Ice Sheet, surveyed in 1999, and possible controls on ice sheet behavior.
CHAPTER 2: GLACIAL HISTORY OF THE EAST ANTARCTIC ICE SHEET OFFSHORE NORTH VICTORIA LAND – EVIDENCE FOR GROUNDED ICE DURING THE LATEST GLACIAL MAXIMUM

Chapter Context

This chapter is based on a paper co-authored with John Anderson and submitted to Geology. It is based on the Pennell Coast study area that was surveyed during the 1998 cruise.

Overview

The continental shelf offshore of the Pennell Coast, North Victoria Land, Antarctica, is isolated from the West Antarctic Ice Sheet drainage; for an ice sheet to ground in this region it must flow over the Transantarctic Mountains from the East Antarctic drainage system. Multibeam swath bathymetry records collected during the 1998 season show glacial troughs with rugged relief and grooves on the inner continental shelf. Seismic data show these features occur on crystalline bedrock. The multibeam records show that the middle and outer shelf is characterized by less relief and seismic data show sedimentary deposits at the sea floor. Gullies were observed at the shelf break. The described features indicate that ice was grounded on the Pennell shelf at some point. Piston cores sampled a generalized stratigraphy consisting of, from bottom to top, sub-glacial till, a pelleted lift-off unit, glacial-marine sediments, and traction current deposits and diatomaceous muds. Accelerator mass-spectrometry radiocarbon dates on
foraminifera from within the till suggest that the ice advanced onto the shelf after approximately 35,000 $^{14}$C years ago (uncorrected dates). Dates from marine units above the till show that the shelf was ice-free by about 13,000 $^{14}$C years ago, and possibly by 15,645 $^{14}$C years ago (uncorrected dates). These dates suggest that melting ice from the region contributed to the Holocene sea-level rise and that expansion of that ice sheet may have contributed to the relatively sudden drop in sea level immediately prior to the Last Glacial Maximum.

Introduction

Controversy remains about the extent of the Antarctic ice sheets during the last glacial maximum (LGM) (Domack et al., 1991; Colhoun et al., 1992; Anderson et al., submitted) and there is a discrepancy between the amount of sea-level rise during the most recent transgression and the locations of ice masses that could have provided the water (Tushingham and Peltier, 1991; Andrews, 1992; Colhoun et al., 1992; Ackert et al., 1999; Nakada et al., 2000). The problem is further complicated by recent results that suggest that parts of the Antarctic ice sheet were retreating well before the maximum lowstand of sea level (Berkman et al., 1998; Anderson, 1999) and by data that suggest a sea level that was between 130 and 135 m lower than today at the LGM, lower than some previous estimates (Yokoyama et al., 2000).

Considerably more is known about the retreat of the West Antarctic Ice Sheet (e.g., Anderson and Shipp, 2001; Steig et al., 2001) than about the retreat of the East
Antarctic Ice Sheet (EAIS). Stuvier and others (1981) envisioned an EAIS that reached the shelf break in their models. Colhoun (1991) argued that the EAIS did not expand as far as the shelf break during the LGM. Recent data show that the EAIS has different advance and retreat histories along its margin (Berkman et al., 1998; Anderson, 1999). Most of the controls on EAIS expansion have come from terrestrial studies (e.g., Berkman et al., 1998; Goodwin and Zweck, 2000) and results from marine studies are sparse. Research on the East Antarctic continental shelf is desperately needed to resolve these problems.

Based on the rugged, foredeepened bathymetry of the continental shelf offshore of the Pennell Coast, North Victoria Land (Fig. 2-1), it was assumed that the EAIS expanded onto the continental shelf in the region, but the timing was not known (Brake, 1982; Brake and Anderson, 1983). The fact that ice was not grounded to the shelf edge in the western Ross Sea during the LGM (Domack et al., 1999b; Licht et al., 1999; Shipp et al., 1999) implies that an advance and retreat of the EAIS occurred in the region in order to account for the isostatic rebound observed in the Cape Adare region (Fig. 2-1: Colhoun et al., 1992). Detailed studies of the glacial history from the continental shelf west of Cape Adare and east of Wilkes Land have not previously been conducted. Therefore, no reliable glacial reconstruction for this sector of East Antarctica exists.
Figure 2-1. Location map of the Pennell Coast, North Victoria Land, study area. Polar projection map from http://usarc.usgs.gov/antarctic_atlas.
This paper presents geophysical and geologic evidence that the EAIS did advance on the continental shelf offshore of the Pennell Coast during the LGM. The Pennell Coast was chosen for this study because of its location next to the Transantarctic Mountains (TAM). In order for ice to ground on the Pennell shelf, it must flow through and over the TAM. Thus, evidence for grounded ice on the Pennell shelf indicates a much expanded EAIS. Radiocarbon dates obtained from reworked carbonate material within till help constrain the timing of ice sheet advance in the area. Ice sheet retreat is constrained by a suite of radiocarbon dates from glacial-marine sediments.

Methods

Data used in this study were collected aboard the RV/IB Nathaniel B. Palmer during cruise NBP9801. This new data set augments the cores and bathymetric data collected in the area during the Deep Freeze 1980 expedition (Brake and Anderson, 1983). Heavy sea ice in the area during the 1998 cruise limited data collection. However, eleven cores, three side-scan sonar surveys, moderate quality multibeam data, and some seismic data were collected (Fig. 2-2).
Figure 2-2. Map of the study area showing the NBP9801 cruise track and data distribution.
Seismic data were collected with either a 50 in$^3$ (820 cm$^3$) or a 210 in$^3$ (3,442 cm$^3$) generator-injector air gun and a single-channel streamer. The data were recorded with an Elics seismic acquisition system. The seismic data were band-pass filtered. The resolution of these data is on the order of 10 m. CHIRP 3.5 kHz data was also collected. The 3.5 kHz data are higher resolution, but do not penetrate beyond ~10 meters. The 3.5 kHz data were not processed. Multibeam swath bathymetry data were collected with a SeaBeam 2100 hull-mounted system and consist of 120 beams of 12 kHz data. Data editing removed anomalous beams. Processing included applying a corrected sound-velocity profile and gridding and displaying the data in shaded-relief maps. Vertical resolution of the multibeam data is on the order of a few meters. Horizontal resolution of the multibeam data varies based on the grid size used to produce each image. The multibeam images from this area have an average grid size of about 20 m. Deep-tow side-scan sonar data were collected, using a Datasonics SIS 1000, to image finer-scale bathymetric features. These data were not processed.

The geological data set consists of piston cores that penetrated to a maximum of 8.15 m and Kasten cores that are up to 2.5 m long. Magnetic susceptibility measurements were taken on all of the cores on board (Appendix 1). Kasten cores were opened, described, and photographed on board. Piston cores were shipped to the Florida State University Antarctic Research Facility where they were split, described, photographed, and x-radiographed. Core samples selected for radiocarbon analysis were washed and sieved. Clean samples were then centrifuged in a sodium-polytungstate solution to
separate sand material from foraminifera and other carbonate material. The lighter portion with the concentrated carbonate material was used to pick foraminifera for dating. The foraminifera were dated by accelerator mass-spectrometry (AMS) methods (Domack et al., 1989; Jull, 1998). Organic material from three other samples was also analyzed. These were also dated by AMS methods. There is much uncertainty about the carbon reservoir effect in the Antarctic (Licht et al., 1996; Domack et al. 1999a, b); for this reason, all $^{14}$C dates reported here are uncorrected.

**Results**

Multibeam swath bathymetry data, in conjunction with previously published bathymetric data (Brake and Anderson, 1983), were used to map the bathymetry of the area (Fig. 2-3). Rugged relief characterizes the inner shelf. The outer shelf is less rugged, with two large troughs being the dominant features on the shelf. Detailed swath bathymetry data were used to image erosional grooves and map the glacial troughs (Fig. 2-4). Seismic records show that the rugged inner shelf is floored by crystalline bedrock and a wedge of sedimentary strata exists on the outer shelf (Figs. 2-5 and 2-6). The shelf wedge is truncated by an unconformity that extends to the shelf break (Fig. 2-5). The landward slope of the unconformity indicates a glacial origin. The geomorphic features within the troughs consist of grooves (Fig. 2-4), indicating that the ice sheet on the inner shelf was resting directly on rugged, crystalline bedrock. Mega-scale glacial lineations, which are indicative of ice flowing across a deforming bed (C. D. Clark, 1993; Shipp et al., 1999; Wellner et al., in press), do not occur in the seaward portions of the troughs.
Figure 2-3. Bathymetric map of the study area. Locations of cores that are discussed in the text are marked.
Figure 2-4. Multibeam swath bathymetry image of the primary transverse glacial trough in the study area.
Figure 2-5. Seismic line from the outer continental shelf of the Pennell Coast illustrates the wedge of offlapping sedimentary strata that occur from the middle shelf to the continental shelf break.
Figure 2-6. Seismic line from the inner continental shelf of the Pennell Coast illustrating the rugged relief and crystalline bedrock.
Gullies were observed on multibeam images from the shelf break (Fig. 2-7). Gullies occur elsewhere around the Antarctic margin and are interpreted as having been cut by down slope flow of turbulent sediment-laden water coming from the base of an ice sheet that was grounded at the shelf break (Anderson, 1999).

Several different facies were identified in cores based on lithology, shear strength, pebble content, sedimentary structures, and magnetic susceptibility data. See Anderson (1999) for a more detailed discussion of criteria used to distinguish different sediment types. The oldest unit sampled by the cores is a diamicton with uniform magnetic susceptibility, no preferred pebble orientation, and rare and poorly preserved fossils. This unit is interpreted as till. In the two cores that sampled till, a pelletized lift-off unit occurs immediately above the till. This unit consists of molded sediment pellets and contains only agglutinated foraminifera. The foraminifera are *Miliammina earlandi*, which were identified by Milam and Anderson (1981) as being associated with the ice edge. This unit is interpreted as a grounding zone deposit formed shortly after the ice sheet decoupled from the bed (Domack et al., 1999). Above the pelletized unit, glacial-marine and open-marine deposits occur. These consist primarily of diatomaceous muds and volcanic sands. The latter are interpreted as traction current deposits derived from the Cape Adare region and transported across the shelf by westward flowing currents (Anderson, 1999).
Figure 2-7. Multibeam swath bathymetry image showing gullies at the shelf break. The ship was breaking through sea ice during this part of the survey. Data gaps are caused by ice being below the ship sonar systems.
Radiocarbon dates are reported in Table 2-1 and Figure 2-8. Six samples from the tills in cores NBP9801-PC17 and -PC19 were dated; these cores are from the transverse trough on the west of the study area (Fig. 2-3). These dates range in age from 35,830 ± 890 to 39,600 ± 1,200 uncorrected $^{14}$C years BP. None of the samples were beyond the accepted range of radiocarbon dating (about 46,000 years BP, Linick et al., 1986; 50,000 years BP, http://www.physics.arizona.edu/ams/index.html) and the maximum error reported from the six till samples is 1,200 years.

Fifteen samples from glacial-marine and marine units in cores DF80-153 and -158 and NBP9801-PC17, -PC19, -PC22, and PC26 were dated (Fig. 2-3). The oldest date directly above a glacial unconformity in core NBP9801-PC19 from the middle shelf is 13,200 ± 80 $^{14}$C years BP. Core NBP9801-PC17 also sampled subglacial till. The date from above unconformity is 8,200 $^{14}$C years BP; however, this date is from organic matter and represents a slightly inverted stratigraphy with the carbonate date above it (Table 2-1, Fig. 2-8). The oldest date from the outer shelf, from core NBP9801-PC26, is 15,645 ± 95 $^{14}$C years BP and is from a sandy diamicton. This unit is interpreted as a sub-ice shelf or proximal glacial-marine deposit. The youngest dates, 3,895 ± 50 and 6,425 ± 55, were obtained from the upper volcanic sand unit that is interpreted as a traction current deposit. The middle part of NBP9801-PC26 is composed of alternating layers of these two types of deposits; the contacts between the layers are disturbed and the two units are partially mixed. Three radiocarbon dates obtained from this zone are all about 10,000 $^{14}$C years BP (Fig. 2-8).
Table 2-1. Information about radiocarbon dates obtained in this study. Appendix 2 contains information about the foraminifera used for the dates.
<table>
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<th>Core Number</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Depth (cm)</th>
<th>Lab Number</th>
<th>Material Dated</th>
<th>Mass (mg)</th>
<th>Age (yr. B.P.)</th>
<th>Error (+/- yr.)</th>
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<td>10,475</td>
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<td>169°35'14.6&quot;E</td>
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<td>AA31726</td>
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<td>15,645</td>
<td>95</td>
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Figure 2-8. Cartoon representation of dated cores with date locations marked.
Discussion

The results from this investigation indicate that the EAIS underwent significant expansion during the LGM. The evidence for grounding on the shelf include the presence of a glacial unconformity that extends to the shelf break (Fig. 2-5), troughs with erosional grooves (Figs. 2-3 and 2-4), upper slope gullies (Fig 2-2), and the presence of till and pelletized sub-ice shelf deposits on the shelf (Fig. 2-8). This requires that the EAIS was flowing through and over the TAM and onto the shelf. Because the maximum grounding line position in the Ross Sea during the LGM was inland of the shelf break (Domack et al. 1999b; Licht et al., 1999; Shipp et al., 1999) ice could not have come from around Cape Adare (Fig. 2-1) to ground in this area.

Dates from foraminifera obtained from till units represent the age of the sediment over which ice advanced, and thus constrain ice sheet advance. The formation of till by ice advance mixes different aged sediments. Thus, dates from reworked foraminifera within the till represent the earliest date the ice could have grounded in the region (Domack et al., 1999). Of the six dates obtained from till samples, none of the samples was radiocarbon isotope dead. Thus, we interpret the till dates to mean that ice grounded in the location of the cores after about 35,000 $^{14}$C years BP.

The retreat history of the EAIS following the LGM is poorly constrained. The outer shelf in the western Ross Sea, which receives drainage from the EAIS, remained
free of grounded ice at the LGM (Domack et al., 1999b; Licht et al., 1999; Shipp et al., 1999): the maximum LGM grounding line position in the Ross Sea was near Coulman Island, south of Tucker Glacier (Domack et al., 1999; Shipp et al., 1999). Far to the west of the study area, in the region of the Vestfold and Bunger Hills, radiocarbon dates have indicated that most of the ice retreat occurred between 10,000 and 6,000 calendar years BP (Colhoun, 1991). Domack and others (1989) determined that the retreat from the continental shelf offshore the Adélie Coast area of Wilkes Land, between the Pennell study area and the Vestfold and Bunger Hills, occurred between 9,000 and 2,500 calendar years BP.

Our results show that the retreat of grounded ice from the Pennell shelf occurred by about 13,000 $^{14}$C years BP and probably by 15,645 $^{14}$C years BP (Fig. 2-8). The 13,000 year date is from pristine, undisturbed sediments above a glacial unconformity and clearly shows ice-free conditions by that time. The series of radiocarbon dates from the middle of NBP9801-PC26 (Table 2-1) show little age variability between 67 and 113 cm (10,265-10,475 respectively). This possibly indicates reworking of sediment by iceberg furrowing (Fig. 2-9). An iceberg that plowed into the sediments would have mixed older and younger materials together. The 15,645 year date is from a unit underneath the disturbed zone and does not appear to have been disturbed by the iceberg plowing; there is no evidence in the magnetic susceptibility or x-radiograph data that suggest the lowest unit has been disturbed. We thus interpret this date as indicating ice-free conditions by 15,645 $^{14}$C years BP.
Figure 2-9. Side-scan sonar record illustrating iceberg furrows. This image is used to show how icebergs can plow into and disturb the surface sediments. (a) Furrows with internal ridges and (b) highly irregular furrow.
In addition to simply finding locations where enough ice could have been stored to account for the LGM lowstand of sea level, timing is problematic (Colhoun et al., 1992; Ackert et al., 1999). Models that simply reconstruct the LGM ice distributions may miss the additional constraints on the timing of the formation of that ice. The maximum sea level during Marine Oxygen Isotope Stage 3, also known as the mid-Wisconsin highstand, is a subject of debate. Some evidence suggests that sea level was higher than indicated by oxygen isotope curves and possibly as high as -15 m (e.g., Wellner et al., 1993; Rodriguez et al., 2000). Other authors have presented geologic evidence that the sea levels that are indicated by oxygen isotope curves are correct (Chappell et al., 1996). The timing of this possible highstand is not yet well constrained. A sea-level peak of 42 m below present may have occurred as late as 28,000 years BP (Bloom and Yonekura, 1985). If the Bloom and Yonekura (1985) estimates are correct, nearly 90 m sea-level equivalent of ice would have formed on land in about 10,000 years. This implies at least some contribution from East Antarctica. The results from this investigation indicate that the EAIS did experience significant expansion after about 35,000 $^{14}$C years BP. This expansion may have contributed to the sea-level fall that occurred prior to the LGM.

Because the Pennell Coast and the northwest Ross Sea receive drainage from the same area of the TAM, it was expected that the two areas would share relatively similar advance and retreat histories. The outer shelf offshore of the Pennell Coast has very shallow banks (Fig. 2-3). Ice may have become pinned on the banks. Once the banks
had grounded ice, that ice may have blocked ice from the mountains from easily moving offshore. This would then cause the ice to fill the deeper troughs. The outer shelf in the Ross Sea is deeper (Shipp et al., 1999); this may have prevented ice from grounding in the region during the LGM and be the reason for the differences in the glacial history of the two regions. In some sense the Pennell dates that indicate grounded ice during the latest LGM are transitional between the Ross dates and the Wilkes Land dates. The differences in the dates from each area also point out the complex retreat of the ice.

**Conclusions**

The EAIS grounded offshore of the Pennell Coast during the LGM. AMS radiocarbon dates from foraminifera picked from the till suggest that the ice advanced to the Pennell Coast region after approximately 35,000 $^{14}$C years ago. Dates from marine units above the till show that the shelf was ice-free by about 13,000 $^{14}$C years ago and probably by 15,645 $^{14}$C years BP. Confirmation that ice was grounded offshore of the Pennell Coast during the LGM allows the region to be used in models as a place to store ice during the sea-level lowstand. As radiocarbon dates from different areas of the Antarctic are reported, the complexity of the timing of the advance and retreat of the ice is becoming apparent.
CHAPTER 3: DISTRIBUTION OF GLACIAL GEOMORPHIC FEATURES ON THE ANTARCTIC CONTINENTAL SHELF AND CORRELATION WITH SUBSTRATE: IMPLICATIONS FOR ICE BEHAVIOR

Chapter Context

This chapter is based on a paper of the same name in press in the *Journal of Glaciology* co-authored with Ashley Lowe, Stephanie Shipp, and John Anderson. It is based on the results of two cruises on the RV/IB *Nathaniel B. Palmer* in 1998 and 1999.

Overview

Surveys were conducted seaward of all the major drainage outlets of the Antarctic Ice Sheet from the Pennell Coast of North Victoria Land to Marguerite Bay of the Antarctic Peninsula. The results show that the West Antarctic Ice Sheet extended onto the outer shelf. Glacial troughs occur offshore of all major glacial outlets. Where the substrate is crystalline bedrock, ice flow tended to follow the structural grain of the bedrock, deposited little sediment, and eroded the underlying bedrock. Where ice flowed over relatively soft, more easily eroded sedimentary strata, the direction of ice flow was more directly offshore and depositional features characterize the seafloor. In these areas the signature of the grounded ice consists of till deposits and large-scale geomorphic features with mega-scale glacial lineations as the dominant features. Drumlins occur within the region of contact between crystalline and sedimentary substrates.
The different geological substrates are interpreted to have exerted a fundamental control on the behavior of past ice sheets. The troughs in the areas of sedimentary substrate are interpreted to have been occupied by relatively fast-flowing ice, ice streams, and the troughs in the area of crystalline substrate are interpreted to have been occupied by slower-moving ice. The area between these two zones was characterized by ice acceleration and is marked by drumlins.

**Introduction**

Currently much attention is focused on sea-level rise and the potential for ice-sheet collapse. Ice streams have been cited as the potential weak link that could lead to this collapse (e.g., Hughes, 1973). Ice streams account for the majority of ice discharge from the West Antarctic Ice Sheet (WAIS) (Bentley, 1987) and they influence the behavior and stability of ice sheets by providing a means for rapid discharge from the interior of the ice sheet.

Field measurements indicate that neighboring West Antarctic ice streams vary widely in behavior (Shabtaie and Bentley, 1987; Whillans et al., 1987), and satellite observations have recorded significant changes in activity (Bindschadler and Vornberger, 1998; Rignot, 1998). Studies of the WAIS indicate that irregular ice-stream behavior is, in large part, controlled by subglacial bed conditions and processes occurring at and within the bed (Alley et al., 1986, 1987; Alley, 1989; Blankenship et al., 1986, 1987, 2001, Rooney et al., 1987; Kamb and Engelhardt, 1991; Anandakrishnan et al., 1998;
Bell et al., 1998; Tulaczyk et al., 1998). Troughs in the underlying topography may
delineate ice-stream margins, and a step in ice-surface topography typically marks the
onset of streaming (Shabtaie and Bentley, 1987). Aerogeophysical imaging (Bell et al.,
1998) and seismic profiling (Anandakrishnan et al., 1998) indicate that the area of Ice
Stream B, which flows into the Ross Ice Shelf (Fig. 3-1), coincides with a geologic
boundary between a sedimentary basin and crystalline bedrock. Bell and others (1998)
and Anandakrishnan and others (1998) asserted that a transition from crystalline bedrock
to softer sediments is necessary for onset of streaming to occur. Drilling at an upstream
portion of Ice Stream B (Fig. 3-1) recovered a weak, unconsolidated sediment layer
located at the base of the streaming ice (Engelhardt et al., 1990; Kamb, 1991; Kamb and
Engelhardt, 1991). Detailed examination of weak tills retrieved from Ice Stream B
determined that the till formed from recycled poorly indurated sediments. Therefore,
movement of ice streams via a deforming-till layer may be controlled by locations of
sediment sources (Tulaczyk et al., 1998).

Despite great interest in the connection between subglacial geology and ice
behavior, there remains much to be learned from field observations. The Antarctic
continental shelf, previously covered by expanded ice sheets, contains a relatively
undisturbed record of subglacial conditions that is more accessible than the current
subglacial environment.
Figure 3-1. (a) Satellite image of Antarctica with study areas marked. EAIS is the East Antarctic Ice Sheet and WAIS is the West Antarctic Ice Sheet. AVHRR image from http://TerraWeb.wr.usgs.gov/TRS/projects/Antarctica/color/index.html. (b) Distribution of data used in this study. Black lines mark positions of cruise tracks. Multibeam data and 3.5 kHz data were collected for all of each cruise.
This paper presents geophysical and sedimentological observations from the continental shelf offshore of each of the major ice drainage outlets between North Victoria Land and the Antarctic Peninsula (Fig. 3-1). All major drainage outlets of the WAIS are included, except those of the Weddell Sea. In addition, drainage areas of the East Antarctic Ice Sheet (Pennell Coast and western Ross Sea) and the Peninsula Ice Cap (Marguerite Bay) are included. The research focused on the large glacial troughs that occur in each drainage area. The data set includes seismic imaging of the substrate over which the ice flowed, high-resolution seismic imaging of the upper bed, multibeam swath bathymetry imaging of large geomorphic features on the shelf, deep-tow side-scan sonar imaging of smaller bedforms, and geological sampling of the bed. Relationships between geomorphic features and substrate geology were observed. These correlations have implications for the control that the substrate geology may exert on ice behavior.

Methods

Our analysis of individual drainage outlets followed a systematic approach whereby seismic records, swath bathymetry data, and sediment cores were acquired along the axis of troughs. Data between troughs were collected primarily in transit. The seismic data were collected with either a 50 in³ (820 cm³) or a 210 in³ (3,442 cm³) generator-injector air gun and a single-channel streamer. The data were recorded with an Elics seismic acquisition system. The seismic data were band-pass filtered. The vertical
resolution of these data is on the order of 10 m. The 3.5 kHz data are higher resolution, but do not penetrate beyond ~10 meters. The 3.5 kHz data were not processed. The multibeam swath bathymetry data were collected with a SeaBeam 2100 hull-mounted system and consist of 120 beams of 12 kHz data. Data editing removed anomalous beams. Processing included applying a corrected sound-velocity profile, and gridding and displaying the data in shaded-relief maps. Grid size varied for each image based on the water depth and area included in the image; an average grid size is about 30 by 30 m. Vertical resolution of the multibeam data is on the order of a few meters. The Datasonics deep-tow side-scan sonar images finer-scale bathymetric features. These data were not processed. The geological data set consists of piston cores that penetrated to a maximum of 8.15 m and kasten (gravity) cores that are up to 2.5 m long.

Substrate differentiation was based on seismic facies descriptions and consisted largely of determining if the subcrop in each area is crystalline or sedimentary. Continuous to semi-continuous parallel to divergent seismic reflections, which onlap landward onto more chaotic seismic units, represent sedimentary substrates. Chaotic and discontinuous reflection patterns were interpreted as crystalline substrates. This seismic reflection pattern also can represent a diamicton. The two are distinguishable because till and glacial-marine diamictons form a thin deposit packaged with marine strata and rest on an erosional unconformity whereas the basal contact is not observed in crystalline bedrock where there is generally much greater relief. In addition, sedimentary strata of different ages onlap crystalline basement rocks. Onshore outcrops also were used to help interpret the observed seismic units. Glacial landforms were identified and divided into
categories of depositional, erosional or mixed based on mode of formation (Table 3-1: Fig. 3-2). Table 3-2 presents sedimentary units identified in cores and seismic profiles. Analysis of the data followed Anderson’s (1999) summary of sedimentological and seismic criteria for defining deposits that occur on the Antarctic continental shelf and his description of geomorphic features imaged by side-scan sonar and multibeam techniques.

Results

North Victoria Land

The Pennell Coast is located west of Ross Sea in North Victoria Land, where the Transantarctic Mountains meet the coast (Fig. 3-1). Unlike the other areas in this study, drainage along the Pennell Coast is wholly within East Antarctica. The continental shelf, relative to the other regions in this study, is very narrow (Fig. 3-3a). Rugged bathymetry and crystalline substrate, similar to the onshore portion of the study area, characterizes the inner shelf (Fig. 3-3b) (Brake and Anderson, 1983). Multibeam data from the inner shelf show steep troughs and erosional grooves. A narrow sedimentary wedge occurs on the outer shelf (Fig. 3-3c): no lineations were observed on the outer shelf. Piston cores collected from the trough flanks contain diamicton overlain by a thin pelletized unit (Table 3-2), and topped by diatomaceous deposits with varying amounts of ice-rafted debris. The basal diamicton is interpreted as till based on its occurrence below the pelletized unit, moderate shear strengths, lack of pebble fabric visible on x-ray, and lack of pristine microfossils. The presence of till in piston cores indicates that the ice sheet
<table>
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<th>Landform</th>
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<tr>
<td>DEPOSITIONAL FEATURES</td>
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</tr>
<tr>
<td>Mega-scale glacial lineations</td>
<td>Large flutes, up to 70 km long, that are thought to form in deforming till layers and represent rapidly flowing ice (C.D. Clark, 1993, 1994).</td>
<td>Figure 3-2c</td>
</tr>
<tr>
<td>EROSIONAL FEATURES</td>
<td></td>
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<tr>
<td>Grooves</td>
<td>Irregular lineations in bedrock formed by scouring particles embedded in ice (Smith, 1948). They are oriented parallel to each other and indicative of ice flow direction.</td>
<td>Figure 3-2b</td>
</tr>
<tr>
<td>Roches moutonnées</td>
<td>Asymmetric bedrock bumps or hills with abraded stoss faces and quarried lee faces (Carol, 1947). They range in size from meters to several hundreds of meters across.</td>
<td>Figure 3-2b</td>
</tr>
<tr>
<td>Glacial troughs</td>
<td>Large canyon-like depressions carved by ice flow. Troughs can be hundreds of kilometers long and tens of kilometers wide and hundreds of meters deep. When formed in crystalline bedrock they tend to have narrow, deep, steep-sided u-shaped profiles and follow the structural grain. In sedimentary substrates, they are shallower and broader.</td>
<td>Figure 3-2b</td>
</tr>
<tr>
<td>Gullies</td>
<td>Submarine valleys carved in the continental slope by underflows of debris-rich meltwater from ice grounded at the shelf break (Anderson, 1999).</td>
<td>Figure 3-2a</td>
</tr>
<tr>
<td>Meltwater channels</td>
<td>Channel forms cut by glacial meltwater. These are erosional features cut into bedrock and consolidated sediments by subglacial drainage.</td>
<td>Figure 3-2d</td>
</tr>
<tr>
<td>INTERMEDIATE FEATURES</td>
<td></td>
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<tr>
<td>Drumlinso</td>
<td>Smooth, oval-shaped hills resembling a spoon along its long axis with the steep end pointing in the up-ice direction and the gentler, sloping end facing the down-ice direction (Menzies, 1979). Drumlins may form around crystalline or sedimentary bedrock irregularities and in that sense are erosional. They may also form by the deposition and erosion of glacial sediments.</td>
<td>Figure 3-2e</td>
</tr>
</tbody>
</table>

Table 3-1. Summary of geomorphic features identified in this study. This study uses a wide range of geomorphic features to interpret glacial history and basal conditions. The features can be grouped on the basis of whether they form in depositional zones, erosional zones or both.
Figure 3-2. Multibeam swath bathymetry: (a) Central Ross Sea showing gullies at the shelf break; (b) Sulzberger Bay showing the trough carved by grounded ice and the associated erosional grooves and roches moutonnées; (c) Pine Island Bay showing mega-scale glacial lineations that become overlain by an iceberg-furrowed sediment wedge further offshore; (d) Pine Island Bay showing meltwater channels; and (e) Eltanin Bay showing drumlins. See Table 3-1 for further discussion of geomorphic features and Figure 3-1 for locations. The multibeam data has a vertical resolution of a few meters. Two of these images are modified from Anderson and Shipp, 2001.
<table>
<thead>
<tr>
<th>Sedimentary Unit</th>
<th>Definition</th>
</tr>
</thead>
<tbody>
<tr>
<td>Lodgement till</td>
<td>Diamicton with shear strengths &gt; 0.5 kg/cm² formed in the subglacial environment by sediment plastering onto the substrate through pressure melting processes (Dreimanis, 1988; Lundqvist, 1988).</td>
</tr>
<tr>
<td>Deformation till</td>
<td>Diamicton formed in a subglacial deforming layer (Benn and Evans, 1998). Deformation till is characterized by a wide-range of shear strengths, typically greater than associated glacial marine sediment and less than lodgement till (usually between 0.2 and 0.5 kg/cm²) (Anderson, 1999).</td>
</tr>
<tr>
<td>Pelletized unit</td>
<td>Granulated sandy facies characterized by pellets or balls of sediment grains and agglutinated foraminifera. This unit occurs just above diamicton and below a muddy unit and is believed to be a grounding line feature (O’Brien and Harris, 1996; Domack and others, 1998; 1999).</td>
</tr>
<tr>
<td>Sub-ice-shelf (transitional) and compound glacial-marine sediment</td>
<td>Includes a range of sediment types, ranging from silts to diamictons formed beneath an ice shelf or on the open shelf (Anderson, 1999). They typically have lower shear strengths and more varied mineralogical content than associated tills and are characterized by occurrences of fossils (Anderson, 1999). X-ray analysis does not show a preferred clast orientation.</td>
</tr>
<tr>
<td>Diatomaceous mud/ooze</td>
<td>Identified by its olive-green color. Where this unit occurs it is always the surficial unit. Its maximum thickness is not known, but in some areas it is greater than 8 m thick. Diatomaceous mud contains 10-30% diatoms, and ooze contains greater than 30% diatoms.</td>
</tr>
</tbody>
</table>

Table 3-2. Sedimentary units described in this study. Shear strengths reported in this paper are a factor of 10 lower than values reported in some sources. It appears that these older values were taken directly from the Torvane dial without applying the appropriate correction factor.
Figure 3-3. (a) Bathymetric map of the continental shelf offshore of the Pennell Coast region showing the steep, narrow trough on the inner shelf that grades into a broad trough on the outer shelf. The distance between the crystalline-sedimentary contact and the shelf break is narrow, 35 km, compared to other areas. Contour interval is 100 m. (b) 50 in$^3$ air gun seismic line from the inner shelf showing the crystalline basement and (c) from the outer shelf showing progradational sediments. Locations are shown in (a).
grounded on the shelf. Limited multibeam data collected seaward of the shelf break indicate the presence of gullies.

Ross Sea

The Ross Sea, located between the Pennell Coast and Sulzberger Bay, is bound on the west margin by the Transantarctic Mountains, to the south by the Ross Ice Shelf, and to the east by Marie Byrd Land (Fig. 3-1). The Ross Sea receives drainage from both the East and West Antarctic ice sheets. The continental shelf, wide relative to the rest of the study regions, is geographically divided into western, central and eastern sectors. Broad glacial troughs characterize the continental shelf. These troughs, occupied by paleo-ice streams over several glaciations, trend northeast-southwest (Hughes, 1977: Anderson, 1999: Shipp et al., 1999).

On the basis of seismic data, the continental shelf substrate is predominantly seaward-dipping sedimentary strata. Deep Sea Drilling Project (DSDP) Site 270 recovered Early Paleozoic marble and gneiss close to the sea floor in a single location on the inner shelf of the central Ross Sea (Hayes and Frakes, 1975). Based on interpretations of seismic data, the innermost shelf in this area has crystalline bedrock at the seafloor just south of the DSDP 270 site. Excluding the innermost western and the central Ross Sea shelves, unlithified sedimentary wedges and till sheets overlie the lithified sedimentary substrate. Geomorphic features, in addition to core data, have been used to define the extent of ice during the last glacial maximum (LGM). In the western
Ross Sea, ice extended to the outer shelf but did not reach the shelf edge (Licht et al., 1996; Domack et al., 1999; Shipp et al., 1999). Ice in the central and eastern Ross Sea extended to the continental shelf edge (Domack et al., 1999; Licht, 1999; Shipp et al., 1999). Multibeam data show mega-scale glacial lineations within troughs. Drumlins occur on the inner shelf of the central Ross Sea in association with the transition from crystalline to sedimentary substrate (Fig. 3-4). Subcrops of sedimentary strata characterize the innermost western Ross Sea shelf. Side-scan sonar and subbottom profiler data reveal thin patchy deposits overlying an erosional surface in this region (Shipp, 1999; Shipp et al., 1999).

Cores from the Ross Sea typically recovered diatomaceous mud overlying a soft diamicton with moderate numbers of microfossils. The latter is interpreted as an ice-shelf facies. A thin pelletized unit locally marks the base of the ice-shelf facies (Domack et al., 1999). Below the pelletized unit or directly below the ice-shelf facies is till. The transition between the ice-shelf facies and the till is variably marked by an increase in shear strengths, color change, and decrease in microfossil abundance. Locally, cores sampled a stiff diamicton, interpreted as lodgement till, underlying the sequence (Anderson et al., 1980; Shipp et al., 1999). Radiocarbon ages indicate that these strata were deposited during and since the last glaciation (Licht et al., 1996; Domack et al., 1999; Licht, 1999). In the western Ross Sea, lodgement till occurs on the innermost shelf in association with the thin patchy deposits overlying the eroded sedimentary substrate.
Figure 3-4. (a) Drumlins in the transition zone in paleo-ice stream 3 in the central Ross Sea shown on swath bathymetry data. Darker shades of gray indicate shallower depths and lighter shades of gray indicate greater depths. (b) Line drawing of a full seismic profile showing substrates upon which features in (a) formed. Crystalline bedrock occurs on the inner shelf and seaward-dipping strata occur farther offshore. The drumlin field occurs near the contact between crystalline and sedimentary substrates. (c) Seismic data showing three substrate zones in detail. C1 is seaward-dipping strata, C2 shows strata onlapping onto crystalline bedrock, and C3 shows smooth crystalline bedrock on inner shelf.
Softer diamictons occur in association with mega-scale glacial lineations (Domack et al., 1999; Shipp et al., 1999).

**Sulzberger Bay**

Sulzberger Bay, located east of the Ross Sea (Fig. 3-1), is bounded by the Edward VII Peninsula. A prominent trough appears aligned with the structural grain of the Marie Byrd Land coast. Multibeam data collected from the region show an irregular sea floor characterized by erosional grooves (Fig. 3-2b). Seismic data acquired during previous cruises indicate that the bay is floored by crystalline bedrock (L. R. Bartek, seismic data, 02/99). Based on the extent of the trough and glacial grooves, it is apparent that the WAIS extended to the shelf break during a previous glacial cycle. Depositional geomorphic features (Table 3-1) were not observed.

Attempts during 1999 to core in this area largely were unsuccessful and yielded only bagged samples of sand and gravel. Several cores were obtained in Sulzberger Bay during a 1983 expedition (Deep Freeze 83). A core from inner Sulzberger Trough penetrated a light olive brown pebbly mud (Cassidy, 1984). A core from the outer trough penetrated a light olive gray mud with few pebbles (Cassidy, 1984). Shear strength was not measured on either core. However, both cores were significantly deformed and washed in the liner, implying that stiff material, which often serves to plug the core, was not penetrated.
Getz Ice Shelf (Wrigley Gulf and Bakutis Coast)

Ice streams of the Getz Ice Shelf region currently flow around a group of small islands into Wrigley Gulf and across the Bakutis Coast (Fig. 3-1). Large troughs occur in both of these regions (Fig. 3-5a). Multibeam data from both troughs show subglacial geomorphic features extending across the shelf, indicating that the WAIS grounded in locations currently at water depths over 1,000 m. The data record erosional grooves on the inner shelf and mega-scale glacial lineations extending across the outer shelf (Fig. 3-5b). On the inner shelf the troughs are aligned with the structural grain of the area. On the outer shelf, the troughs are oriented more directly offshore. Seismic data from the Bakutis Coast (Fig. 3-5c) show crystalline bedrock on the inner shelf. The outer shelf is characterized by seaward-dipping sedimentary strata, which toplap against an unconformity. The relief of the seafloor on the inner shelf where older strata are exposed suggests that the strata are heavily lithified (Fig. 3-5c).

In Wrigley Gulf and the Bakutis Coast region the transition between the crystalline and sedimentary substrate is delineated by a change from grooves to mega-scale glacial lineations (Fig. 3-5). Drumlins occur at the crystalline-sedimentary bed transition. Piston cores from the area of mega-scale glacial lineations sampled gray to olive-gray diamicton characterized by low shear strengths (from 0.22 to 0.24 kg/cm²). These deposits are interpreted to be deformation till based on their stratigraphic relationships, shear strengths, x-ray character, and low abundance of microfossils. Overlying the till are thin deposits of gray to brown diamicton with lower shear strengths than the till; this is interpreted to be glacial-marine sediment. Cores from the area of
Figure 3-5. (a) Location of paleo-glacial drainage trough offshore Bakutis Coast. Box shows area of b. (b) Multibeam data and (c) seismic data from offshore of a portion of Getz Ice Shelf showing the change from crystalline to sedimentary substrate and the corresponding change in the geomorphic features on the shelf. Seismic location is shown on multibeam data.
crystalline bedrock sampled dark gray diamicton with shear strengths from 0.1 to 0.19 kg/cm$^2$. These sediments are interpreted, primarily based on the abundance of microfossils, as glacial-marine deposits.

**Pine Island Bay**

Pine Island Bay (Fig. 3-1) serves as catchment to a major convergent drainage system of the WAIS. Several deep and narrow troughs emerge from the coast (Fig. 3-6). These converge on the middle shelf to form one main trough that extends to the shelf break (Fig. 3-6b). The bathymetry of the continental shelf changes from a rugged inner shelf to a smooth outer shelf (Kellogg and Kellogg, 1987). The inner shelf displays relief with depths between 400 and 1,700 m, while the outer shelf averages 500 m in depth. A seismic profile collected along the axis of the main trough reveals the transition in bathymetry corresponds to a change in substrate. An irregular crystalline bedrock surface characterizes the inner shelf and most of the middle shelf. Farther offshore, this surface dips below seaward-dipping strata, which toplap against an erosional surface at or near the seafloor (Fig. 3-6c). North of Burke Island, a sedimentary wedge, composed of prograding sequences of chaotic and layered deposits, overlies these strata. The occurrence of this wedge and other deposits above the glacial unconformity creates the broad, smooth bathymetry common on the outer shelf.

Multibeam bathymetric images show channels on the rugged crystalline basement of the inner shelf and the preliminary interpretation is that these features are meltwater
Figure 3-6. (a) Location of paleo-glacial trough, marked with a 600 m contour line, in Pine Island Bay. Shaded areas within the trough represent areas deeper than 1,400 m. Location of (b) is shown. (b) Detailed bathymetric map of Pine Island Bay showing the steep, narrow trough on the inner shelf and the broader shallower trough on the mid-outer shelf. Contour interval is 200 m. (c) Seismic line from Pine Island Bay. Location is shown in (b).
channels (Fig. 3-2d). Glacial erosional features include grooves, gouges, and drumlins. Drumlins occur just landward of the contact between crystalline bedrock and sedimentary substrate bedrock. Cores from the inner shelf are short, but sampled alternating units of clay and low shear strength (< 0.1 kg/cm²) diamicton. This is interpreted to be a sub ice-shelf unit. Within the deeper portions of the trough on the outer shelf, mega-scale glacial lineations are present (Fig. 3-2c). In locations shallower than 700 m depth, such as the younger sedimentary wedge, iceberg furrows obliterate the lineated surface (Fig. 3-2c). Gullies occur on the continental shelf offshore of Pine Island Bay.

**Eltanin Bay**

Eltanin Bay (Fig. 3-1) is a broad bay located in the Bellingshausen Sea west of the Abbot Ice Shelf and east of the Ronne Entrance. In the LGM reconstruction of Denton and Hughes (1981), Eltanin Bay receives drainage from a large area of Ellsworth Land. Multibeam data reveal a large trough extending across the shelf. The data show erosional grooves in the inner portion of the bay. Farther offshore, mega-scale glacial lineations characterize the geomorphology. No data were collected from the outer shelf due to severe weather conditions. A drumlin field separates the zones of erosional grooves and mega-scale glacial lineations (Fig. 3-2e). Individual drumlins are 5-10 km long. The multibeam data suggest there may be channel-like depressions around the heads of the drumlins indicating that formation of the drumlins may have included pressure melting and water flow around the drumlins. Multibeam data show that the WAIS grounded where water depths are currently up to 1,200 m on the middle shelf, although the
grounding line probably was seaward of the data coverage. Piston cores collected from the middle shelf sampled till, with shear strengths less than 0.34 kg/cm², beneath thin units of very soft diamicton with greater abundance of microfossils. The area of diamicton is coincident with the area of mega-scale glacial lineations.

**Marguerite Bay**

Marguerite Bay is the most northern drainage area surveyed in this study: it is located north of Alexander Island at the southern end of the Antarctic Peninsula. It receives drainage from the Antarctic Peninsula Ice Cap. A deep trough, greater than 1.650 m, occupies the inner shelf of Marguerite Bay and extends seaward across the shelf (Pope and Anderson, 1992). The inner shelf seismic data from this region is not from the same location as the multibeam data. However, the seismic data that is available does show basement rocks and folded forearc rocks at the sea floor (Bart and Anderson, 2000). Based on that seismic data and comparison of the multibeam data in other areas, the substrate in the inner shelf is interpreted to be crystalline bedrock with highly irregular topography. Grooves occur on the trough floor. Steep slopes, especially on the western flank, mark the trough boundaries. A drumlin field lies farther seaward. Seismic profiles collected offshore of the drumlin region show smooth middle and outer shelves floored by prograding seaward-dipping strata (Bart and Anderson, 1995) and sampling at ODP site 1079 recovered lithified sediments (Barker et al., 1998).
Cores collected from the eastern portion of the bay sampled thick muds with ice-rafted debris and a moderate number of microfossils, underlain by sediment-gravity-flow deposits (Kennedy and Anderson, 1989). Cores from the western region sampled thin glacial-marine units underlain by transitional glacial-marine deposits, or in rare cases, thick sections of stiff till occurring in water depths greater than 700 m (Kennedy and Anderson, 1989). A single core collected from within the drumlin field sampled glacial-marine sediment underlain by sediment-gravity-flow deposits.

**Correlation between Geomorphic Features and Geological Substrate**

In each of the study areas, large glacial troughs were mapped on the continental shelf offshore of major drainage outlets. The troughs vary in size and extent across the study areas. However, their general shapes and bathymetry are similar in that they are narrower and deeper on the inner shelf over crystalline bedrock and broader and shallower on the outer shelf where they cut sedimentary strata. This change in trough geometry, from narrow and deep to broad and shallow, partially may be controlled by the foredeepened nature of the Antarctic continental shelf in that the troughs must broaden on the outer shelf to allow for somewhat constant volumes of ice.

Analysis of the data from across the region reveals a range of substrates on which former expanded ice sheets were grounded. Two end members include slightly erodible crystalline bedrock and more easily eroded sedimentary strata. In locations where ice is interpreted to have flowed over both end members, a transition zone between the two
exists. Each substrate zone displays characteristic geomorphic features and sedimentary deposits that are consistent throughout the drainage areas.

**Crystalline Bedrock**

The inner shelf of most study areas is characterized by crystalline bedrock with highly irregular bathymetry. Deep erosional grooves and steep-sided troughs are the dominant features (Fig. 3-2b). These features indicate the direction of ice flow, which appears to have followed the structural grain of the underlying bedrock, displaying sharp turns and an orientation that is not always directed in an offshore direction. The North Victoria Land (Fig. 3-3), Sulzberger Bay (Fig. 3-2b), Getz Ice Shelf (Fig. 3-5), and Pine Island Bay (Fig. 3-6) study areas are the primary examples of these features. Other landforms associated with crystalline bedrock include meltwater channels such as those seen in Pine Island Bay (Fig. 3-2d) and roches moutonées such as those seen in Sulzberger Bay (Fig. 3-2b). Erosional formation of the landforms may be controlled by either grounded ice (e.g., grooves and roches moutonées) or by meltwater (e.g., incised channels) (Table 3-1).

**Sedimentary Substrates**

Sedimentary strata that dip offshore characterize the outer shelf in most areas. Troughs widen and shallow as they extend across the softer substrate, and mega-scale glacial lineations are the dominant geomorphic feature. Mega-scale glacial lineations form in material eroded from the substrate. Individual lineations are 10s of kilometers
long and are about 0.4 km wide. Sets of lineations can be seen to deviate in orientation as ice is able to spread and the flow direction changes.

In general, as the ice sheet advanced across the continental shelf it sampled crystalline bedrock, lithified sedimentary strata, and unlithified Plio-Pleistocene deposits. The extent of these geological domains varies from one area to the next.

Sedimentary substrates characterize the majority of the Ross Sea (Fig. 3-4) and the outer shelves of North Victoria Land, Wrigley Gulf and Bakutis Coast (Fig. 3-5), Pine Island Bay (Fig. 3-6), Eltanin Bay (Fig. 3-2e), and Marguerite Bay. Sulzberger Bay is the only study area where no sedimentary substrate is known to exist. The North Victoria Land drainage area is the only area that has a sedimentary wedge not characterized by mega-scale glacial lineations. Elsewhere, lineations exist just seaward of the crystalline bedrock-sedimentary strata contact. Thus, it appears that mega-scale glacial lineations can form on lithified sedimentary strata, as well as soft Plio-Pleistocene sediments.

The distribution of deformation till (Table 3-2) appears to correlate to the occurrence of mega-scale glacial lineations. This relationship previously has been documented in Antarctica in the Ross Sea (Licht, 1999; Shipp et al., 1999). In most study areas, cores collected from the region of mega-scale glacial lineations consist of glacial-marine deposits overlying relatively soft till. Variations include the eastern Ross Sea where lodgement till, in addition to soft till, is occasionally found in the region of mega-scale glacial lineations. Also, cores from areas such as bank margins between adjacent
ice streams in the Ross Sea sampled material described as a deformation till, but lack associated geomorphic features. Few cores from the areas of crystalline substrate recovered diamicton. However, when diamicton is recovered in cores from areas of crystalline substrate, it is not significantly different from that in cores farther out on the shelf.

It is not possible from this data set to measure accurately the thickness of the deforming bed in the different study areas. The complete thickness of till is at least as thick as the amplitude of the mega-scale glacial lineations. These average about 15 - 20 m in height across the study areas. Seismic profiles from the Ross Sea image a distinct unit, in which mega-scale glacial lineations occur, that ranges from 15 to 35 m thick (Shipp et al., 1999). Cores did not penetrate the bottom of this deposit. Hence, it is beyond the range of the data to determine whether the entire unit is deformed, or only its upper surface.

The extent of deformation till varies from basin to basin, but more data are needed to quantify its distribution. The minimum dimensions of such a deposit can be inferred by analyzing the extent of mega-scale glacial lineations and the distribution of deformation till sampled in cores. In the Ross Sea, where a sedimentary substrate floors most of the area, the area of lineations extends almost 200 km from the inner shelf to the outer shelf. In those areas where multibeam data were collected across the troughs, mega-scale glacial lineations are imaged across the entire trough. Limited bathymetric data show no mega-scale glacial lineations in the area of the inter-ice-stream banks and
boundary ridges in the Ross Sea (Shipp et al., 1999), but cores indicate deformation till is present.

**Transition Zone**

A geomorphic transition zone occurs where the substrate changes from crystalline bedrock to sedimentary strata. The extent of this zone varies. It includes portions of the seafloor seaward and landward of the point where strata begin to onlap onto the crystalline basement. Drumlins occur within the transition zone. Only one area, the North Victoria Land shelf, has crystalline and sedimentary substrates, but lacks drumlins within a transition zone.

Offshore of the Getz Ice Shelf, drumlin-like features occur landward of the crystalline-sedimentary substrate contact (Fig. 3-5). These drumlins have their heads in crystalline bedrock and grade seaward into thin onlapping sediments. Different types of drumlins include those in the central Ross Sea. These are rock-cored drumlins, with seismic data showing underlying bedrock highs serving as obstructions to ice flow (Fig. 3-4). In Eltanin Bay drumlins are believed to be composed of sediment. Deep troughs surrounding the heads of the drumlins in Eltanin Bay may indicate erosion by water that flowed around the drumlins during their formation (Fig. 3-2e). Drumlins in Pine Island Bay display the same character, but these are bedrock drumlins formed in the crystalline portion of the transition zone.
Discussion

Many studies have examined the role of subglacial geology on the behavior of Northern Hemisphere ice sheets (e.g., Alley, 1991; MacAyeal, 1993; Alley and MacAyeal, 1994). P. U. Clark (1994) proposed that subglacial bed conditions, not climate, caused rapid fluctuations in behavior of the Laurentide Ice Sheet. Boulton (1996) attributed large-scale till sequences formed by the North American, British, and European ice sheets during the last glacial cycle to the process of subglacial sediment deformation over beds of un lithified sediments. Marshall et al. (1996) used geologic and topographic data to model areas of potential fast flow in the Laurentide Ice Sheet and suggested that interior plains and the continental shelf, rather than the crystalline shield, favor ice stream development. Patterson (1998) proposed that geology and topography beneath the Laurentide Ice Sheet exerted control on the areas of fast ice flow. Her results indicated that ice-covered sedimentary valleys contained thicker ice and reached the pressure-melting point before areas of positive relief. This mechanism facilitated faster flow. A geologic control on modern-day Antarctic ice streams has also been proposed (e.g., Bell et al., 1998; Anandakrishnan et al., 1998).

A consistent distribution of geomorphic features was observed in each of the drainage areas. A typical distribution includes erosional features such as grooves or roches moutonnées on the inner shelf, drumlins in the transition zone, mega-scale glacial lineations on the outer shelf, and gullies on the slope (Fig. 3-7a). This distribution pattern of geomorphic features correlates with changes in the substrate (Fig. 3-7). Additionally,
Figure 3-7. Diagram summarizing the inferred relationship between geomorphic features, substrate, and subglacial ice flow velocity. (a) Distribution of geomorphic features in plan view. Black represents areas of erosion and gray represents areas of deposition. (b) Corresponding change in substrate from the inner to outer shelf. (c) Relative speed of ice flow.
the features observed on the continental shelf may be interpreted to reflect the behavior of
the ice sheet that was grounded on the continental shelf. The crystalline, transition and
sedimentary substrate zones are interpreted to represent three different conditions of ice
flow (Fig. 3-7c). Table 3-3 summarizes the flow conditions in each zone.

A correlation between a deforming bed and streamlined bedforms has been made
by several workers including C. D. Clark (1993; 1994), Hart and Smith (1997), Shipp and
others (1999), Stokes and Clark (1999) and Canals and others (2000). On the Antarctic
continental shelf, troughs in areas of sedimentary substrate have been interpreted as areas
of convergent ice flow and streaming ice (Anderson, 1999). This is based on the
configuration of the bathymetry, the concentration of sediment within portions of the
troughs, and the lineated nature of the upper surface. The mega-scale glacial lineations
visible on multibeam swath bathymetry data have been interpreted as representing areas
where ice was streaming (Shipp et al., 1999). Areas of mega-scale glacial lineations also
have been correlated to deposits of soft diamicton (Domack et al., 1999; Shipp et al.,
1999). Slower moving ice is interpreted to have occupied banks, although no absolute
rates are implied.

Flow over the crystalline substrate is thought to have been relatively slow and
resulted in erosional geomorphic landforms. Flow over the sedimentary substrate, where
there is a deforming bed, is interpreted to have been relatively fast. It is important to note
that although the data show that a sedimentary substrate is required for the presence of
mega-scale glacial lineations, there are areas of sedimentary substrate that, despite having
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Table 3-3. Geomorphic features observed in this study and their interpretations. PIB is Pine Island Bay, and R. S. is Ross Sea.
supported grounded ice, do not have mega-scale glacial lineations, particularly the inner shelf of the western Ross Sea and the outer shelf of North Victoria Land. Therefore, a sedimentary substrate is required for fast flow by bed deformation, but fast flow does not occur on all sedimentary substrates.

The transition zone spanning the crystalline-sedimentary substrate contact represents a zone of acceleration. Drumlin{s are the most prominent features within transition zones, including the central Ross Sea, Getz Ice Shelf, Pine Island Bay, Eltanin Bay, and Marguerite Bay. The acceleration zone was an area of dynamic change in the ice sheet and thus, was also the area of maximum relief in the ice surface. Bell et al. (1998) described an area of high relief in the ice surface at the modern-day onset of Ice Stream B.

All locations with crystalline substrate on the inner shelf and mega-scale glacial lineations on the outer shelf have drumlins in the transition zone. Drumlin{s were not observed in any other setting. Because drumlins occur between regions of slow ice flow over crystalline substrate and regions of fast flow over sedimentary substrate, drumlins are interpreted to be indicative of acceleration. Mickelson and others (1983) and Attig and others (1989) suggested that drumlins formed in areas where the ice was partially frozen to the bed and locally thawed. This may be comparable to the association we have made between drumlins and ice acceleration. Drumlins also have been interpreted as forming in areas of extensional flow (Lundqvist, 1989, van der Wateren, 1995), but to our
knowledge, this study provides the first correlation between drumlins and the interpreted onset of streaming flow for Antarctic ice streams.

Meltwater also is interpreted as having played an important role in controlling ice behavior. The data indicate that the amount of meltwater present varies from basin to basin. Abundant channels in Pine Island Bay, which are interpreted as meltwater channels based on the morphological expression (Fig. 3-2d), represent the only evidence for channelized meltwater on the inner part of the shelf. Drumlin heads (Shaw, 1988; Shaw et al., 1989). Similarly, the drumlins in Eltanin Bay (Fig. 3-2e) and the Ross Sea (Fig. 3-4) appear to have channels, or moats, around their edges. These channels possibly were formed by localized flowing water. Water would have formed by pressure melting of the ice as it flowed over the core of the drumlin. Where drumlins grade into mega-scale glacial lineations, this water was available to be incorporated into deforming till. Meltwater affected ice-stream dynamics until the ice reached deformable sediments on the shelf, where it became incorporated into the sediments to produce deforming till. No geomorphic features indicative of large amounts of meltwater were imaged on the sedimentary substrate. Because water must be incorporated into till to allow deformation to occur, free meltwater probably does not occur in areas of deforming till. Gullies at the shelf break in the Ross Sea, Pennell Coast shelf, Sulzberger Bay, and Pine Island Bay represent erosion by debris-rich meltwater expelled when the ice grounded near the shelf break (Anderson, 1999). The fact that there is evidence for meltwater both landward and seaward of the mega-scale glacial lineations indicates that
available meltwater was incorporated into the deforming till in the areas of mega-scale glacial lineations.

Bathymetry itself is a control on ice flow. Banks and adjacent troughs in the Ross Sea have primarily sedimentary substrates. However, ice flow over bank tops is interpreted to have been relatively slow in contrast to the fast flow in the troughs (Shipp et al., 1999). A certain thickness of ice is required to create enough pressure on the underlying sediments to initiate fast flow (Tulaczyk et al., 2000a, b). Therefore, once a trough has been formed, ice will consistently be thicker in those areas, and streaming will be concentrated in troughs and away from banks.

The shelf offshore of Pennell Coast is one area where the zone of sedimentary substrate does not have any evidence for a deforming bed, and there is no transition zone. There are a few linear, mound ed features visible on the multibeam data in the area of sedimentary substrate. Because of their small scale, they are interpreted to be, not mega-scale glacial lineations, but rather erosional grooves. Why there is no evidence for a deforming bed on the Pennell shelf is unclear; three possibilities are presented here. One reason may have to do with the width of the shelf. The Pennell study area has a narrow shelf with little sedimentary substrate; the maximum distance ice flowed over sedimentary substrate is about 35 km. In other areas, such as Pine Island Bay (Fig. 3-6) and offshore of the Getz Ice Shelf (Fig. 3-5), the acceleration zone covers only a few tens of kilometers and forms right at the crystalline-sedimentary substrate transition. However, seaward of the transition zone there is over 300 km of sedimentary substrate in
Pine Island Bay and at least 85 km in the Bakutis Coast area of the Getz Ice Shelf. There may be a minimum distance over which the ice flows on sedimentary substrate to allow either for an increase in speed or else for geomorphic features to be formed. Alternatively, the dominant grain size and associated porosity of the till may not be appropriate to hold the water needed to form deforming till. The Pennell Coast shelf sediments are dominated by volcanic sand and the tills may be too sandy to develop into deformation till. Lastly, the Pennell Coast receives drainage from a small area of the Transantarctic Mountains and there may not have been enough ice feeding the system for ice streaming at the scale recorded by mega-scale glacial lineations.

**Implications**

Ice that is underlain by deformable till may be prone to fast retreat or to collapse (Engelhardt et al., 1990; MacAyeal, 1992). If this is true, is ice in places like Sulzberger Bay and the Pennell Coast inherently more stable than in places like the Ross Sea? Geomorphic features indicative of retreat, such as corrugation moraines, are present in the western Ross Sea (Shipp et al., 1999). These retreat features are evidence that the ice stayed in contact with its bed during retreat. Radiocarbon dates from the western Ross Sea also indicate a gradual retreat (Conway et al., 1999; Domack et al., 1999). There are no retreat features visible in any of the other study areas. This may indicate decoupling during ice-sheet retreat, but it may well be due either to a lack of sediment to form such features or to incomplete data coverage. If the lack of retreat features is indicative of fast retreat, a potentially rapid retreat occurred in areas characterized by crystalline substrate...
and no deforming bed, such as Sulzberger Bay, and a slow and gradual retreat is indicated in the western Ross Sea, which had an extensive deforming bed. Do these data suggest that a deforming bed may not have led to instability? Tulaczyk and others (2000a, b) proposed that deformation of the subglacial bed may be a process that operates at end members of very fast or very slow flow, without taking place in the conditions between those two end members. Thus, ice resting on a deforming bed may have an internally regulated mechanism by which fast flow may be stopped.

As ice advances and retreats across the shelf, it comes into contact with different substrates. This implies that, throughout a glacial cycle, ice behavior will be subject to different controls that should be included in models of ice behavior. Conditions for stability in the ice sheet also may change through multiple glacial cycles. The substrate and acceleration zones are not stationary through time. Throughout multiple glacial cycles more and more erosion takes place, shifting the crystalline-sedimentary substrate contact farther seaward. This requires ice to be grounded farther onto the shelf in each cycle before it reaches the sedimentary substrate and is able to achieve fast flow.

Conclusions

This study provides the first documentation of the range of geomorphic features present on the Antarctic continental shelf and their correlation with substrate type. Where crystalline bedrock is the substrate, ice tended to follow the structural grain of the bedrock and carved deep troughs. Where ice flowed over relatively soft sedimentary
strata the direction of ice flow was more directly offshore. In these areas the signature of
the ice consists of till deposition and mega-scale glacial lineations that formed in the
deforming bed. Drumlins are the characteristic feature observed in the transition zone
that spans the crystalline-sedimentary substrate. Ice is interpreted to have flowed
relatively slowly in the area of crystalline substrates and relatively fast in the area of
mega-scale glacial lineations. In this interpretation, drumlins would indicate a zone of
acceleration. Because the presence of a deforming bed affects the rate at which ice can
flow, the substrate below an ice sheet may also exert a fundamental control on the
stability of that ice sheet.
CHAPTER 4: GULF OF MEXICO INTRODUCTION

Chapter Context

This chapter presents an overview of the Gulf of Mexico study area and the motivation for this work. It also provides an overview of the climate history for the study area and the drawbacks that occurred when trying to correlate climate to sedimentary deposits in a very large drainage basin.

Introduction

The Gulf of Mexico began to form in the late Middle to early Late Jurassic as a result of the breakup of the supercontinent Pangea (Salvador, 1991). The rifting event resulted in four types of crustal material that comprise the basin (Sawyer et al., 1991). Continental crust pre-dates the rifting period and occurs mainly on the periphery of the basin. Continental crust that has been thinned during the rifting of the basin is divided into thick or thin transitional crust based on the amount of thinning. Oceanic crust in the deep part of the basin formed during a short period of sea-floor spreading and is the fourth type of crust (Sawyer et al., 1991). Five main events characterize the Mesozoic development of the basin (Worrall and Snelson, 1989). First, there was a period of rifting and redbed deposition. This was followed by Middle Jurassic evaporite deposition and formation of a major unconformity and then the Late Jurassic and Early Cretaceous carbonate deposition and progradation of terrigenous clastics. In the Early and Middle
Cretaceous there were two periods of shelf-margin reef development. Finally, after the
mid-Cretaceous, the reefs were drowned and an unconformity formed (Worrall and
Snelson, 1989).

Abundant growth faults characterize the basin. On the Louisiana shelf these
faults are arcuate, dip landward and basinward, and are associated with near-surface salt
bodies. The growth faults are attributed to salt movement triggered by sedimentary
loading (Worrall and Snelson, 1989).

In the early Tertiary a great volume of sediment entered the basin due to the
Laramide orogeny to the west (Worrall and Snelson, 1989). Throughout the Cenozoic
the basin has been continually fed with sediments. The depocenters shifted during this
time but accumulated up to 16 km of sediment. The Mississippi system in some form has
been emptying into the Gulf of Mexico since at least Cretaceous time (Coleman, 1982).

The modern Mississippi River drainage basin is 3,344,560 km² (Fig. 4-1) and is
an order of magnitude greater in area than the Rio Grande drainage basin (Coleman,
1982). The water discharge is 15,360 m³/s (Geraghty et al., 1973) and the annual
sediment yield to the Gulf of Mexico is 2.4 x 10¹¹ kg, 70% of which is clay (Coleman,
1982). The modern delta is 28,568 km², 23,900 km² of which is subaerial. The delta is
actually small for its drainage basin compared to other deltas (Coleman, 1982).
Figure 4-1. Topography of the modern Mississippi drainage basin with the main branch of the Mississippi River and tributaries. Data provided by Conoco.
The Mississippi fluvial-deltaic system has been extensively studied beginning with Fisk (1945) and later by many others including Lehner (1969), Coleman and Roberts (1988a, b, 1990), Suter and Berryhill (1985), and Berryhill (1985). Petroleum production from the Louisiana continental shelf has provided the data and the incentive for study of the area and the Quaternary sediments of the area provide a excellent opportunities to demonstrate some of the concepts of sequence stratigraphy (Suter and Berryhill, 1985; Coleman and Roberts, 1988a, b; Boyd et al., 1989). Coleman and Roberts (1988a, b) based their studies on platform borings from the west Louisiana continental shelf and Suter and Berryhill (1985) based theirs on high-resolution seismic data. Sarzalejo (1993) and Anderson and others (1996) studied the outer shelf along the Texas-Louisiana border with integrated core and seismic data but a similar study has not yet been conducted on the inner west-Louisiana continental shelf.

**Study Motivation**

Since its introduction in the 1970s, sequence stratigraphy has become a standard tool in both academics and the oil industry. The majority of research in sequence stratigraphy has been done either with low-resolution seismic data (e. g., Vail et al., 1977) or in outcrops of Mesozoic rocks (e. g., Van Wagoner, 1995). Detailed climate and sea-level data and precise dating are not available for those types of studies. Studies of late Quaternary depositional systems can help to modify sequence stratigraphic models
and add details that would not otherwise be included. One of the goals of this work is to
determine the depositional history of sediments from the last glacial-eustatic cycle on the
west Louisiana continental shelf and to examine any implications the area may have for
sequence stratigraphic models. Stratigraphic correlations were completed following the
methods of Vail and others (1977) and sequence terminology is that of Van Wagoner and
others (1988, 1990) and Reynolds (1996). A variety of published literature was used for
sea-level and climate information. Implications for sequence stratigraphic models were
made in large part by comparing to other systems of the same age in the Gulf of Mexico
(Anderson et al., 1996).

Sediment supply, which is in large part a function of climatic controls, and
eustasy together control how depositional systems form (Posamentier and Allen, 1993;
Ethridge et al., 1998). There is still controversy about whether the majority of sediment
comes to basins during the lowstand or highstand of sea level (Morton and Suter, 1996).
A probable answer is that it varies. Much work has focused on the response of
depositional systems to sea level and climate, but separating the effect of one factor from
the other is difficult. The literature is divided such that work in the marine realm has
focused on the control of eustasy, (e. g., Vail et al., 1977) and work in the fluvial systems
has focused on climatic controls (e. g., Blum, 1990). Examining the effects of climate on
sediment deposition is one of the goals of this project and is discussed in more detail
below.
Climate Effects on Sediment Yield

Climate is known to be an important control on sediment supply (e.g. Langbein and Schumm, 1958; Hall, 1990; Bettis and Autin, 1997; Ethridge et al., 1998). Garner (1959) showed the control that vegetative cover, which is controlled principally by precipitation, could have on retaining sediment in a modern alluvial system. Other studies have emphasized the importance of overall climatic regime (for example, wet-dry vs. year-round precipitation), rather than simply mean annual precipitation (Wilson, 1973). These types of studies have shown that the relationship between climate and sediment supply is complicated (Wellner et al., 2000).

The northern Gulf of Mexico was chosen to study climate effects on sediment yield because it provides a range of depositional systems whose response to changing climate and eustasy can be investigated, and is a region that is intrinsically sensitive to climate variability. The late Quaternary sedimentary systems in the Gulf of Mexico provide an excellent opportunity to study the relative effect that climate and eustasy have on depositional systems. These can be linked to work in the fluvial systems. The sea-level record for the late Quaternary, particularly since Marine Oxygen Isotope Stage 5e (Figure 4-2; 125,000 years BP: Bard et al., 1990) is better known than for any other period. Additionally, the depositional environments from the Rio Grande to the Western Louisiana system span a dramatic range in climatic regimes. Published climate reconstructions focus on the period since the last glaciation (Knox, 1983; Dorale et al., 1992; Toomey et al., 1993) but are also available for the late Pleistocene. Because of
Figure 4-2. Sea-level curve for the late Quaternary, modified from Bard et al., 1990.
good dating control and availability of modern settings as an analog, reconstructing the climate for these relatively recent sediments is done much more easily than for pre-Quaternary units.

**Climate and Sediment Deposition in the Mississippi Drainage Basin**

The Mississippi River drainage basin crosses several climatic zones and includes areas that were glaciated during the last several glacial cycles. The modern climate is humid subtropical with North-South variations (Autin et al., 1991). Most of the area receives the majority of precipitation during the winter from cyclonic storms. The coastal area of Louisiana has its precipitation maximum during the summer from convective thunderstorms (Autin et al., 1991).

The Mississippi drainage basin extends through so many different climatic zones that it is difficult to simplify the paleoclimatic history of the river. Additionally, the fact that the drainage includes runoff from the Laurentide ice sheet that almost certainly altered the sediment load (Saucier, 1974; Brown and Kennett, 1998) makes the climate signature hard to determine. Most of the strata analyzed in this study were deposited during the last highstand (Fig. 4-2). Below is a summary of climate and sediment deposition within the Mississippi drainage basin during the last highstand.
During the last interglacial, Stage 5, there was a long period of slow aggradation and formation of the Prairie terraces in the valleys of the Mississippi drainage basin (Saucier, 1974, 1981). The earliest of these deposits consists of sand and gravel followed by fine-grain sediments from backswamp and meander deposits (Saucier, 1974). Precipitation was significantly less in the central part of the drainage basin at the end of Stage 3 and during the LGM (Barry, 1983). During the late stages of glaciation (Stage 3 to 2) braided streams formed and increasing amounts of glacial outwash material were deposited in the alluvial valley (Saucier, 1974, 1981; Baker, 1983).

By the last glacial maximum, much of the glacial outwash was draining to the Gulf of Mexico through the Mississippi River (Leventer et al., 1982); the Mississippi River carried more proglacial drainage than any other river system in North America (Baker, 1983). The glacial outwash from the Laurentide ice sheet contained greater amounts of fine-grain material than the typical bedload of the Mississippi (Brown and Kennett, 1998), which was a braided system during the glacial maximum (Saucier, 1974). During the maximum lowstand the gradient of the lower portion of the Mississippi steepened and the system was entrenched south of approximately Vicksburg, Mississippi (Saucier, 1974). The flux of glacially derived fine-grained quartz flour decreased as the ice sheet disintegrated (Brown and Kennett, 1998). At approximately the same period, the discharge of glacial meltwater through the Mississippi decreased. Fisk and McFarlan (1955) noted a change from sand and gravel to finer grained sediments on the alluvial plain of Louisiana at approximately this time (Leventer et al., 1982).
Just before and during the glacial maximum, extreme wind conditions during the winter season resulted in the removal of fine-grained material from stream channels. This material was redeposited as loess (Saucier, 1974; Follmer, 1983). During the summer season, pronounced rainy conditions, paired with cooler temperatures, resulted in increased discharge through tributaries (Saucier, 1991).

The signature of influence from the Laurentide Ice Sheet overwhelms smaller scale climatic effects in the highstand deposition offshore west Louisiana. For this reason, individual climatic changes cannot be correlated to events in the offshore sediments. However, even if the Mississippi drainage basin had not been partially glaciated, it may not be an appropriate system for studying the effects of climate on offshore sedimentation. Wellner and others (2000) cataloged the climatic changes that occurred during the last glacial-eustatic cycle in the drainage basins of the northwestern Gulf of Mexico (Fig. 4-3). They showed the complexity associated with correlating climatic changes to offshore sedimentation in drainage systems such as the Brazos and Colorado rivers of Texas that are two orders of magnitude smaller than the Mississippi drainage basin. In a drainage system the size of the Mississippi that crosses so many different climatic zones (Fig. 4-4), it becomes that much more difficult to correlate climate to offshore sediment deposition.
Figure 4-3. Summary of climate studies from Texas and Louisiana from the last glacial-eustatic cycle. Modified from Wellner et al., 2000. Numbers on chart refer to references in Appendix 1.
Figure 4-4. Map of the Mississippi drainage basin and other rivers that drain into the Gulf of Mexico and average annual rainfall across the drainage basins.
CHAPTER 5: LATE QUATERNARY STRATIGRAPHIC EVOLUTION OF THE WEST LOUISIANA/EAST TEXAS CONTINENTAL SHELF

Chapter Context

This chapter is based on a paper co-authored with John Anderson and Sabrina Sarzalejo that was prepared for a special publication of the Society of Economic Paleontologists and Mineralogists.

Overview

High-resolution seismic data, boring descriptions, and core samples were used to conduct a sequence stratigraphic analysis of the west Louisiana and east Texas outer shelf and upper slope depositional systems formed during the last glacial-eustatic cycle. The main objective of this research was to see how these systems responded to falling and rising sea level and how the delivery of sediment to the shelf responded to climatic fluctuations.

During the Stage 5 to Stage 2 highstand, the relatively high-sediment-supply Western Louisiana fluvial-dominated delta reached the outer shelf and formed an extensive sand body. A nearly continuous ridge of salt diapirs on the shelf edge blocked offshore sediment transport, forcing the delta to prograde to the west. During the last glacial maximum, the Western Louisiana fluvial system shifted to the east so that only prodelta clays were deposited in the study area. The Trinity, Sabine, and Brazos rivers
merged on the shelf and formed an incised valley that extended to the outer shelf. Sediment bypass to upper slope minibasins occurred at this time. The rise in sea level during the early Holocene resulted in the development of thick and complex shelf edge deltas associated with the Trinity/Sabine/Brazos fluvial system. Prograding deltaic sediments and sediment gravity flows were deposited in minibasins situated between diapiric uplifts. Remobilization of salt caused considerable displacement of these deposits. The Brazos River then shifted to a new valley situated west of the study area, greatly reducing sediment supply to the shelf margin delta. Continued sea-level rise resulted in overstepping of this shelf margin delta.

Our results show that there is no simple relationship between sea level and outer shelf-upper slope deposition in the study area. The two deltas that existed were active at different times and the sand bodies associated with these deltas have different sequence-stratigraphic settings. The western Louisiana delta is a highstand delta that is situated between the Stage 5e maximum flooding surface and the Stage 2 sequence boundary. The latter surface is poorly defined since this was an interfluve during the Stage 2 lowstand. The Trinity/Sabine/Brazos shelf margin delta formed mainly during the early transgression and its sandy distributary-mouth bar complex is situated above the Stage 2 sequence boundary.
Introduction

In this study, an extensive grid of high-resolution seismic reflection profiles (Fig. 5-1), oil company platform boring descriptions, and platform boring samples (Fig. 5-2) were used to study the depositional systems of the west Louisiana—east Texas middle to outer shelf and upper slope (Fig. 5-3). These systems were deposited during the last 125,000 years, the last glacial-eustatic cycle. This study focused on deposits that were derived in part from the paleo-Mississippi (Abdulah and Anderson, 1994; Morton and Suter, 1996) and from the combined Trinity/Sabine/ Brazos rivers (Thomas and Anderson, 1994). The objectives of this study were to (1) characterize the depositional systems associated with the last glacial-eustatic cycle using seismic facies and platform borings, (2) examine the influence of salt diapirs and related structures on sediment deposition, (3) determine when, within the sea-level cycle, shelf margin deltas developed and when sediment bypass of the continental shelf occurred, (4) compare and contrast the sedimentologic response of two fluvial systems, a relatively high sediment yield system (Western Louisiana) and a relatively low sediment yield system (Trinity/Sabine/Brazos) to the fall and rise of sea level, (5) examine the role of climatic controls on the type of sediment delivered to the shelf, and (6) determine the extent to which sequence stratigraphic information would help in prospecting for these sand bodies.

Lehner (1969) first identified and described the shelf-edge deltas and submarine slides of the study area using high-resolution seismic reflection profiles and cores. His
Figure 5-1. Locations of seismic lines used in this study. Bold lines are those line segments shown in the text. More examples of seismic data are provided at http://gulf.rice.edu/. Also shown are locations of cores 349-B and 343. Bathymetry is from Bryant et al., 1990.
Figure 5-2. Map of platform boring data set offshore Louisiana with bathymetry and federal lease fields.
Figure 5-3. Bathymetry of the northern Gulf of Mexico shelf and slope. From http://www.ngdc.noaa.gov.
pioneering work controlling the morphology of the continental slope and described and mapped salt-related features such as salt pillows, swells, diapirs, and growth faults.

Sidner et al. (1978), relying on high-resolution seismic and core data from the slope, reconstructed the late Quaternary history of the western-most part of the study area. They recognized and dated two major phases of shelf margin progradation, which commenced during lowstands of sea level at 350,000 and 80,000 years BP. Their chronological control is based on biostratigraphic data (foraminifera) that was further constrained using radiocarbon ages and correlation to oxygen isotope curves. These authors also recognized a series of aggradational packages that were correlated to transgressions and highstands. They argued that the rapid eustatic changes in Pleistocene sea level accelerated the progradation and aggradation of the continental margin. Large slides were recognized and mapped on the upper slope. The slides were interpreted as the products of instability in areas of shelf margin progradation. Diapiric uplifts also were considered responsible for slides and for slower creep-like movement.

Lewis (1984) and Suter and Berryhill (1985) described shelf-margin deltas on the Texas and Louisiana margin, including a large shelf margin delta in the study area. Berryhill (1987) summarized work from the Louisiana-Texas continental margin in an atlas that displays seismic data and line drawings along with interpretations of the Wisconsin shelf phase deltas and shelf margin deltas. He mapped distributary channels across the majority of the West Louisiana shelf (5-4) and interpreted to represent the last glacial
Figure 5-4. Map of distributary channels on the west Louisiana continental shelf from Berryhill, 1986.
maximum. Upper slope mass movement deposits are also described and illustrated in this atlas.

Coleman and Roberts (1990) documented the Louisiana shelf and upper slope stratigraphy using platform borings and some high-resolution seismic data. They described cycles composed of expanded sections, formed during periods of low sea level, and condensed sections, formed during periods of high sea level. The rapidly accumulated expanded sections are characterized by coarse-grained (sand and gravel) deposits and show well-defined depositional facies. The slowly accumulated condensed sections are characterized by calcareous-rich deposits (hemipelagic sediments and shell hash) and display wide lateral continuity.

Morton and Suter (1996) reported on the stratigraphy from the outer-shelf portion of our study area based on high-resolution seismic data and platform borings. They argued that the shelf-margin delta was Wisconsin (Oxygen Isotope Stages 4 to 2) in age, but were unable to put further constraints on the timing of delta development. Anderson and others (1996) called it the Western Louisiana Delta and interpreted it as a highstand delta.
Data Set and Methods

Sedimentologic Data Set

Descriptions of 306 platform boring descriptions and material from 52 platform borings were used in this study (Fig. 5-2). The samples and descriptions from the borings were provided by Fugro-McClelland Inc. These data were used to map the major depositional systems on the middle-outer shelf and upper slope. Two-way travel time was converted to depth using an acoustical velocity of 1,550 m/s. This is slightly less than the 1,675 m/s used by Lehner (1969) for the Pleistocene of the Gulf of Mexico but is closer to the velocities suggested by Orsi and Dunn (1991) for similar materials. Gamma ray logs of some borings were also available. These were used to further characterize the sedimentologic facies.

This study used an oxygen-isotope curve (Bard et al., 1990) for the past 140,000 years as a sea-level proxy (Fig. 5-5). Paleontological studies, described below, provided paleobathymetric curves and an independent measure of sea-level change. Radiocarbon dates of selected samples were correlated with the isotope curve and paleobathymetric curves to obtain chronostratigraphic control for the interpretations. Radiocarbon dating was performed at Beta Analytic. Conventional $^{14}$C dating techniques were used when sufficient material was acquired, commonly in shell layers. Accelerator mass spectrometry (AMS) $^{14}$C dating was performed on smaller samples of foraminifera and shells.
Figure 5-5. Sea level curve modified from Bard and others, 1990.
Paleontologic Data Set

A total of 60 samples from boreholes 343 and 349-B were analyzed for foraminifera by Dr. Martin Lagoe of the University of Texas at Austin. Sample spacing averaged 3 m. Sediments were soaked in a Calgon solution to disaggregate them and then sieved at 150 μm. Census estimates were made on picked populations of 300 benthic specimens and associated planktic foraminifera. Identification and interpretation of faunas relied on previous work from the Texas shelf including Phleger and Parker (1951), Phleger (1951, 1956), Post (1951), Lankford (1966), Tipsword et al. (1966), Mello and Buzas (1968), Greiner (1974), and Poag (1978, 1981). Quantitative analysis followed the methods outlined in Holdford (1995).

Unlike some other studies of Quaternary sediments in the Gulf of Mexico (e.g., Abdullah, 1995) a δ18O curve was not constructed in this study. The study area received direct runoff from the Laurentide Ice Sheet and thus any 18O signature on the shelf would be strongly influenced by the 18O from meltwater. Also, the distribution pattern of planktonic foraminifera was too varied to allow for a single species to be used for δ18O analysis. Lastly, because of the rapid sedimentation in the area and the limited length of platform borings, our samples did not penetrate deep enough to produce a δ18O curve through the last entire glacial-eustatic cycle.
Seismic Data Set

Figure 5-1 shows the location of the grids of seismic lines used in this study. During the summers of 1991, 1992, 1997, and 1998 a little over 2,000 km of high-resolution seismic data were acquired using the Rice University research vessel *Lone Star*. Two seismic sources were used in the Rice University survey. The uniboom source produced frequencies between 300 and 2,400 Hz, and penetration to 400 msec. The calculated vertical resolution for this source is between 25 and 50 cm; however, a realistic estimate of the vertical resolution is about 2 m. The 15 in³ water gun produced frequencies between 40 and 2,000 Hz, penetration to 900 msec, and an idealized vertical resolution of 30-60 cm. In addition to the Rice University survey, Texaco provided about 5,000 km of high-resolution seismic data that were incorporated into this study. The Texaco data set was collected using two different sources. The 40 in³ air gun data were filtered between 100 and 500 Hz and produced penetration exceeding 700 msec and a vertical resolution between 1.5 and 3 m. The 10 in³ air gun source data were filtered between 150 and 1,000 Hz and produced penetration up to 700 msec and a vertical resolution of 1 to 2 m. Conoco provided some intermediate-resolution seismic data.

Seismic Facies

Seismic facies were identified using reflection character, similar to the procedure outlined by Sangree et al. (1978). We recognize six of the seismic facies identified by Sangree and his coworkers (divergent and parallel layered, sheet drape, onlapping fill, sigmoid progradational, oblique progradational, and mounded chaotic). Three additional
seismic facies were recognized on the mid- to outer shelf (divergent fill, chaotic-complex fill, and chaotic fill) (Sarzalejo, 1993). Table 5-1 summarizes the principal characteristics of the nine seismic facies used in this study and should be used as a reference for the facies designations shown in the interpreted seismic sections. Platform borings provided lithological information about the seismic facies and were used to aid in the depositional environment interpretations.

Incised fluvial valleys in the study area are characterized by a complex-chaotic and prograding fill of approximately 45 m in thickness. Figure 5-6 shows an example of an incised fluvial valley cutting into an acoustically layered deltaic facies. A platform boring though this valley sampled sand overlain by silty sand within the valley and clay and silty clay in the acoustically layered deltaic deposits. The valley shown in Figure 5-6 is truncated by the transgressive ravinement surface and overlain by marine muds with abundant shell material.

Amalgamated channels in the area are characterized by chaotic, complex fill (DC, Table 5-1). Figure 5-7 shows an example of amalgamated channels, interpreted as distributary channels, cut into older deltaic deposits characterized by divergent fill. The distributary channels are interpreted to be from a later stage of delta growth and represent the progradation of the distributary channels over the more distal deposits of the same delta system. Core B-349 and the platform borings that penetrated the DC seismic facies sampled sand. The channels are overlain by the transgressive ravinement surface and onlapping parallel-layered marine mud units.
<table>
<thead>
<tr>
<th>Reflection Pattern and External Form</th>
<th>Seismic Facies</th>
<th>Sedimentary Facies</th>
</tr>
</thead>
<tbody>
<tr>
<td>even-layered parallel or gently divergent; continuous reflections; layered sheets or divergent wedges</td>
<td>divergent and parallel layered (PL)</td>
<td>hemipelagic; prodelta turbidites; neritic</td>
</tr>
<tr>
<td>continuous, parallel; drapes topography; nearly constant thickness</td>
<td>sheet drape (SD)</td>
<td>hemipelagic</td>
</tr>
<tr>
<td>divergent fill reflection pattern; associated with complex fill in tectonized areas</td>
<td>divergent fill (DF)</td>
<td>deltaic</td>
</tr>
<tr>
<td>onlapping fill reflection pattern; continuous, nearly parallel reflections: conforms to onlapping surface</td>
<td>onlapping fill (OF)</td>
<td>marine muds; prodelta facies</td>
</tr>
<tr>
<td>sigmoidal progradational pattern in dip view; parallel even-layered pattern in strike view; gradual downlap; sigmoidal in shape</td>
<td>sigmoid progradational (SF)</td>
<td>neritic; littoral; turbidity current; pelagic facies associated with shelf-margin deltas</td>
</tr>
<tr>
<td>oblique-tangential progradational pattern; oblique-parallel progradational pattern in dip view; progradation from a common surface</td>
<td>oblique progradational (OP)</td>
<td>shelf-margin delta</td>
</tr>
<tr>
<td>several mounded-chaotic configurations; discordant and contorted discontinuous wavy subparallel; hummocky reflections fill topographic lows and in cases form as isolated mounds</td>
<td>mounded chaotic (MC)</td>
<td>sediment gravity flow</td>
</tr>
<tr>
<td>chaotic reflections filling well-defined incisions or poorly defined depressions</td>
<td>chaotic-complex fill (DC)</td>
<td>fluvial incision filled with fluvial or bay/estuarine</td>
</tr>
<tr>
<td>subparallel discontinuous or parallel disrupted reflection pattern: continuous parallel divergent reflection pattern</td>
<td>chaotic fill (MB)</td>
<td>deltaic distributary mouth bar</td>
</tr>
</tbody>
</table>

Table 5-1. Definitions and interpretations of seismic facies used in this study (modified from Sarzalejo, 1993).
Figure 5-6. Seismic profile 17 (N) shows the seismic facies associated with incised valley fill. A platform boring through this valley sampled sand overlain by silty sand within the valley and clay and silty clay in the acoustically layered deltaic deposits. Location shown in Figure 5-1.
Figure 5-7. Seismic profile HITL-4 shows the complex chaotic seismic facies (DC) in channels that are interpreted as distributary channels. These channels cut into stratified prodelta deposits. Location shown in Figure 5-1. OF = onlapping fill, DF = divergent fill.
Line HITL-4

B-349  Ravinement Surface

Two-way travel time (msec)

150  200  250

DC  5eMFS  DC  5eMFS  MULTIPLE

V. E. = 18x
1 km

- Sand
- Clay with some silt
- Shell fragments in dominantly silt
Locally, the distributary channel complex rests above and grades laterally into wavy subparallel discontinuous, parallel disrupted, and discontinuous parallel divergent reflection patterns (Fig. 5-8: MB facies). These reflection patterns define low-angle tabular cross-beds that are tens to hundreds of meters across. This facies is most common on the edges of delta lobes, where distributary channels are not deeply incised. This seismic facies is interpreted as a distributary mouth bar facies. Platform borings from this facies sampled mostly sand and silty sand. Delta lobe shifting has resulted in juxtaposition of different delta facies and abrupt lateral pinch-outs of distributary channel and mouth bar sands (Fig. 5-9).

The sigmoidal progradational and oblique progradational seismic facies, typical of the shelf margin deltas, are illustrated in Figure 5-10. This portion of the delta is characterized predominantly by silt and clay.

There are four upper-slope minibasins in the study area, the central basin being just down-slope from the Trinity/Sabine/Brazos incised valley. Mounded-chaotic fill seismic facies commonly occur near the toes of the shelf margin delta clinoforms in the minibasins (Fig. 5-10). These are interpreted as sediment gravity flow deposits, mainly debris flows and turbidites. No platform borings penetrated the mounded-chaotic seismic facies within the study area, but Sangree and others (1978) described the lithology of five cores that penetrated this seismic facies just south of the study area. Three cores sampled clays and the other two penetrated clays with interbedded fine, well-sorted sands. The
Figure 5-8. Part of Texaco Profile 510 that shows the MB seismic facies (Table 5-1). Location shown in Figure 5-1.
Figure 5-9. Seismic line 16 demonstrating delta lobe shifting. Location shown in Figure 5-1. DF = divergent fill, DC = chaotic-complex fill, MB = chaotic fill.
Figure 5-10. Seismic profile OS-91-12 shows the progradational seismic facies of the shelf margin deltas. This seismic line also crosses an upper slope minibasin that is filled with a mounded-chaotic seismic facies (MC) that is interpreted as sediment gravity flow deposits. Location shown in Figure 5-1. SF = sigmoid prodational, OP = oblique progradational. OF = onlapping fill, MC = mounded chaotic.
thickness of the units varied from 30 to 150 cm for the clays and from a few cm to a maximum of 350 cm for the sands.

Mounds with chaotic reflection patterns, located along the flanks of the minibasin, are interpreted as slumps and debris flows. These slumps and debris flows were probably initiated by diapiric uplift along the flanks of the basin. Intrabasin channels occur within the central part of the basin.

Figure 5-11 illustrates another mounded-chaotic facies. This facies is characterized by a continuous, contorted reflection-pattern. This facies fills a deep incision cut into the clinoforms of the shelf margin delta. It occurs seaward of a linear series of diapiric uplifts that trend along the shelf margin in the eastern part of the study area. This particular feature resembles massive deposits that occur elsewhere along the upper slope offshore Louisiana (Berryhill, 1987) and is interpreted as a slump deposit initiated by diapiric uplift. No platform borings penetrated this seismic facies but it is believed to be comprised of recycled muddy prodelta deposits. Figure 5-11 also illustrates the sheet drape facies (SD, Table 5-1) that is characterized by strong, continuous reflections. Cores that penetrated this seismic facies recovered hemipelagic clays.
Figure 5-11. Seismic profile 18 crosses the shelf break in the eastern part of the study area (Fig. 5-1) where a diapiric uplift extends along the shelf margin. The uplift acted as a barrier to offshore sediment transport during the lowstand. Note also the thick, mounded-chaotic seismic facies that fills the deep erosional feature. The continuous nature of the reflectors within the chaotic unit indicates that it is a slump deposit. The slump deposit is buried beneath a draping seismic facies that is characterized by continuous, parallel reflections and is interpreted as hemipelagic drape. $SD = $ sheet drape.
Sequence Stratigraphic Terminology

Terminology used in this paper follows that defined by Van Wagoner and others (1988). The Stage 6 (Fig. 5-5) sequence boundary was mapped in part of the area; the Stage 6 lowstand dates from around 140,000 years BP (Fig. 5-5). Regional correlations were made using the Stage 5e (Fig. 5-5) maximum flooding surface (MFS); the peak sea level during Stage 5 was at approximately 125,000 years BP. All of the sediments between the Stage 5e MFS and the next sequence boundary are included in the highstand systems tract. We sometimes divide the HST into early and late components; these are informal terms and do not necessarily refer to stratigraphically significant surfaces. The main sequence boundary mapped across the study area is the Stage 2 (Fig. 5-5) sequence boundary. The maximum low in sea level during Stage 2 occurred between 22,000 and 17,000 years ago (Bard et al., 1990).

A sequence boundary is an unconformity and the associated correlative conformity. Sequence boundaries are dated at the correlative conformity and that surface represents a time-line surface. The updip erosional component of a sequence boundary represents the time period of erosion and thus a period of time that brackets the date of the correlative conformity. This characteristic is particularly notable in the western Louisiana study area. The deltaic deposits between the Stage 5e MFS and the Stage 2 sequence boundary were formed by a system that contained much of the Mississippi drainage during that time (Morton and Suter, 1996). By the peak lowstand, the paleo-Mississippi had shifted farther east. Therefore, the main incision associated with the
paleo-Mississippi is not represented in the study area and the Stage 2 sequence was cut by distributary channels as the delta prograded across the shelf.

**Seismic Stratigraphy**

Diapiric uplifts and associated faults have strongly influenced sediment transport across the Texas-Louisiana continental shelf. For example, the thickest deltaic deposits occur in minibasins adjacent to salt-cored structural highs (Fig. 5-12). Consequently, both strike-oriented and dip-oriented profiles acquired only a few kilometers apart are different in terms of thickness, seismic character, and overall stratatal stacking patterns of seismic units.

The Stage 6 (Fig. 5-5) sequence boundary is the deepest surface mapped. It represents significantly more erosion and larger channels than the Stage 2 sequence boundary (Fig. 5-13). This indicates that the fluvial system that was delivering sediment to the western Louisiana shelf during Stage 6 captured more of the paleo-Mississippi drainage than the Stage 2 fluvial system.

A regional downlap surface was traced throughout the study area and correlated to the Stage 5e MFS of Thomas and Anderson (1994) on the inner shelf. In West Cameron Block 586, this surface also correlated to the Stage 5 condensed section of Coleman and Roberts (1990). However, the 5e MFS visible on the seismic data in the northeast portion of the study area does not follow the Stage 5 condensed section of Coleman and Roberts
Figure 5-12. Line OS-91-8 showing the minibasins and salt influence on the outer shelf. Location shown in Figure 5-1.
Figure 5-13. Texaco seismic profile 180 showing the Stage 6 sequence boundary.
(1990), which was determined only with platform borings. Our results indicate that the majority of the sand body mapped by Coleman and Roberts (1990) and identified as the Stage 6 delta (Fig. 5-14) is actually the updip component of the Stage 5 delta mapped in this study. The Stage 5e MFS (Fig. 5-15) is the principal surface used in this study to correlate stratigraphic packages of the last glacial eustatic cycle across the shelf and upper slope.

The oldest seismic unit, Unit a, is a thick westward prograding unit associated with the outer shelf delta in the eastern part of the study area. It downlaps onto the Stage 5e MFS (Fig. 5-16). The top of Unit a is marked by a regional onlap surface and the base of Unit b (Fig. 5-16). The boundary between units a and b marks a major shift in the locus of deposition at this time, from east-west to north-south. The upper surface of Unit b is an erosional surface that is interpreted as a sequence boundary (2SB in Fig. 5-16). Figure 5-16 shows an onlapping wedge above 2SB. This wedge is are interpreted as lowstand delta deposits.

Comparison of Figures 5-10 5-11, and 5-16 illustrate how deposition has varied across the shelf break. Such a comparison also serves to demonstrate that significant seismic reflections are difficult to carry between closely spaced mini-basins on a tectonically active shelf. It is thus difficult to make sequence stratigraphic interpretations in isolated areas. For example, shelf margin progradation in the western minibasin ceased after Unit g was deposited but continued in the minibasins to the east (Fig. 5-17). Thus, a shift in the locus of sedimentation caused one minibasin to look much different
Figure 5-14. Isopach map of the Stage 6 sand body from Coleman and Roberts (1990). This unit correlates to the Stage 5 delta defined in this study. Contours in meters.
Figure 5-15. Seismic profile G410 is a dip-oriented profile used to illustrate the seismic units of the shelf-margin delta. It crosses an expanded portion of the Trinity/Sabine/Brazos delta in a large minibasin in the western part of the study area.
Figure 5-16. Seismic line OS-92-35 with Core 343 shown. Location in Figure 5-1. Legend the same as Figure 5-3.
Figure 5-17. Seismic line 29 showing the shelf edge delta. Location given in Figure 5-1.
from the adjacent basin. Sequence stratigraphic surfaces could not be picked from an individual basin (Fig. 5-12).

**Chronostratigraphy**

Chronostratigraphic data for individual units were obtained by radiocarbon dating and comparison to the oxygen-isotope sea-level curve (Bard et al., 1990) and by using foraminifera to construct paleobathymetric curves. These comparisons provide an independent means of relating depositional patterns to sea-level changes. Radiocarbon dating and paleontological analysis were performed in core samples from platform borings 349-B and 343 (Figs. 5-18 and 5-19).

Foraminiferal distributions from boreholes 349-B and 343 form striking down-core patterns (Figs. 5-18 and 5-19). Because of the shelfal position of these cores, relative sea level falls are marked by shallow marine benthic faunas and a lack of planktic foraminifera. Relative sea level rises are characterized by deeper benthic faunas and moderate to high numbers of planktic foraminifera. Agglutinated fauna tend to track planktic foraminiferal patterns. Common upper bayhead, fluvially influenced agglutinated faunas were not detected in any samples. Locations significantly farther updip (embayment or coastal plain) or downdip (slope) would have very different faunal patterns in response to the same sea-level changes.
Figure 5-19. Micropaleontological data for core 343. Organization of chart follows that of Figure 5-18. TUB = Trifarina-Uvigerina-Bolivina, AEN = Ammonia-Elphidium-Nonionella, AE = Ammonia-Elphidium, N = Nonionella. Data from Martin Lagoë.
Core 349-B (Fig 5-1) was acquired from the outer shelf in the eastern portion of the study area and sampled deltaic facies and onlapping marine deposits (Fig. 5-7). Several paleontological intervals were recognized in core 349-B. The Stage 5e condensed section is marked by an abrupt change from a marginal marine or inner neritic faunal zone overlain by an upper bathyal to outer neritic zone (Fig. 5-18). Above the condensed section, the acoustically layered deltaic facies is characterized by shallowing upward faunal transition and a distributary channel complex that is mostly a barren interval. The youngest onlapping fill seismic facies is characterized by inner to middle neritic foraminifera and thus marks a transgressive event.

The interpretation for this faunal distribution is that the sampled interval extends through a prograding fluvial-dominated delta and into its shallow delta front distributaries. There, fresh water input and high sediment supply would produce an adverse environment, and high sedimentation rates would mask microfossils. This would result in the observed barren intervals. A radiocarbon date of 48,000 years BP, which probably indicates a radiocarbon-dead sample, was obtained from a sample from the upper part of the delta. A radiocarbon age of 33,000 years BP (Stage 3) was obtained from a sample in the base of the onlapping fill above the delta. These data provide an upper age limit on the delta.

Core 343 was collected on the outer shelf in the western part of the study area (Fig. 5-1). It sampled the acoustically layered seismic units b through g in the shelf
margin delta (Figs. 5-16 and 5-17) and Unit h, the onlapping fill. Three paleobathymetric intervals were identified in the core (Fig. 5-19). The bottom interval, which corresponds to Unit b, ranges from outer to middle neritic. The boundary between these lower and middle faunal zones is sharp and is interpreted to correlate to the Stage 2 sequence boundary. Unit c marks the transgressive phase of shelf margin delta development in the western part of the study area. The upper faunal interval is characterized by a shift from a marginal marine to outer neritic benthic foraminifera and a significant increase in planktonic foraminifera (Fig. 5-19) and thus marks an increase in water depth. This faunal change coincides with aggradation in the shelf-margin delta. Thus, the foraminiferal data from Core 343 indicate a single lowstand episode when marginal marine to inner neritic conditions prevailed on the outer shelf.

A shell sample and a foraminiferal sample from the upper deltaic units in Core 343 were used for radiocarbon dating. Both samples yielded ages of about 14,000 years BP (Fig. 5-18). These dates indicate that the shelf margin delta was forming during the early stages of transgression. This is consistent with the interpretation of 2SB, just below Unit c, as the Stage 2 sequence boundary.

The transgressive ravinement surface associated with the Stage 1 transgression is imaged in most of the seismic records as a relatively flat erosional surface. It is typically marked in platform borings as a sharp increase in marine shells. In core 349-B, the interval above the ravinement surface contains glauconite (Fig. 5-16) and core 343 (Fig.
5-18) shows a sudden change from marginal marine to middle and outer neritic foraminifera.

**Discussion**

An objective of this research was to determine the relationship between sea level and the late Quaternary evolution of the mid-outer shelf and upper slope deposits within the study area. Our seismic stratigraphic analysis has shown that considerable complexity exists within the strata deposited during a single glacial-eustatic cycle. General patterns of deposition, however, can be reconstructed.

Once the different seismic facies within the study area were characterized and their ages established, maps were made of the correlative facies. These maps show two main depositional systems, a Stage 5 to 3 fluvial-dominated delta on the middle to outer shelf in the eastern part of the study area (Fig. 5-20 and 5-21) and a Stage 2 to early Stage 1 shelf-margin delta. The Stage 5 to 3 delta is here referred to as the Western Louisiana Delta (WLD) by Anderson and others (1996) who concluded that it was formed by a river which included much of the paleo-Mississippi drainage. The Stage 2 to 1 shelf-margin delta was formed by the joined Trinity, Sabine, and Brazos rivers and is here called the Trinity/Sabine/Brazos Delta (TSBD).
Figure 5-20. Late Stage 5 to 3 depositional map for the Western Louisiana Delta.
Figure 5-21. Stage 2 through early Stage 1 depositional map for the Trinity/Sabine/Brazos system. By this time, the WLD had shifted to the east of the study area.
Both Anderson and others (1996) and Morton and Suter (1996) hypothesized that the Stage 5 to 3 sedimentary deposits on the west Louisiana continental shelf had been formed by the paleo-Mississippi. Suggestions that point to such an origin for the WLD include the scale of the delta, which must have been formed by a large drainage system, and the lack of a highstand deposit formed by the Mississippi elsewhere on the shelf. By the LGM the Mississippi River had shifted to the area of the Mississippi Canyon (Fig. 5-22) and was no longer depositing sediments on the west Louisiana shelf. The modern Mississippi delta has formed several different lobes in the last 6,000 years (Fig. 5-23). These lobes indicate periodic avulsion of the river. By comparison, it is easy to imagine the Mississippi drainage being directed to western Louisiana during the last highstand of sea level.

The first phase of evolution corresponds to the deposition of the WLD deltaic and incised channel facies on the middle to outer shelf in the eastern part of the study area (Fig. 5-20). The WLD prograded across the Stage 5e MFS (Fig. 5-5). This is consistent with the Stage 3 radiocarbon dates from the outer portion of the delta and with the paleontologic data showing a shift from inner neritic to upper bathyal conditions across the surface interpreted as the Stage 5e MFS (Fig. 5-15).

The seismic data show that the WLD prograded to the southwest across the middle shelf, then began to prograde to the west on the outer shelf (Fig. 5-21). A series of diapiric uplifts extend along the shelf break (Fig. 5-12). These uplifts prevented the
Figure 5-22. Swath bathymetry map of the Louisiana slope showing the Mississippi Canyon. Image from http://deeptowserver.tamu.edu/deeptow/gom_bathymetry.html.
Figure 5-23. The main lobes of the Mississippi Delta that have formed during the last 6,000 years numbered sequentially.
delta from prograding directly offshore and forced the change to westward progradation. The sandy delta lobes constructed by the delta are situated landward of this barrier. Only muddy prodelta deposits occur seaward of the diapiric ridge. The distributary mouth bar deposits associated with the delta lobes form an almost continuous sand body of variable thickness that extends across the study area from the northeast to the southwest. By the end of the highstand, the river feeding the WLD shifted to the east and out of the study area.

During the Stage 5 to 2 highstand, the Brazos River constructed a fluvial-dominated delta to the west of the study area (Abdulah, 1995). Only after the Brazos river merged with the Trinity/Sabine system (Fig. 5-24) did the shelf-margin delta begin to form in this study area. During the lowstand, the Trinity, Sabine, and Brazos incised valleys merged on the central shelf and extended to the outer shelf. Unit b, which is situated below the Stage 2 sequence boundary, marks the initial southward progradation of the TSBD. The TSBD spans the period from the late highstand systems tract, through the lowstand, and into the beginning of the transgression. The majority of the delta, however, lies above the Stage 2 sequence boundary (Fig. 5-15).

A detailed study of the Trinity/Sabine incised valley was conducted by Thomas and Anderson (1989, 1994). They presented evidence for multiple periods of fluvial incision and subsequent valley filling within the Trinity/Sabine incised valley system. They predicted that these episodes of reincision resulted in significant downstream transport of sand-prone sediment that may extend to the outer shelf and slope. The
Figure 5-24. Stage 2 lowstand map of the East Texas shelf showing the combined Trinity/Sabine/Brazos delta. From Anderson and others (1996).
TSBD identified in this study may be a repository for much of that sand. Seismic data collected slightly downdip from the Trinity/Sabine incised valley show seismic facies indicative of a distributary channel and mouth bar complex between the Stage 2 sequence boundary and the sea floor. Seven platform borings penetrated this facies and recovered an average of 30 m of sand.

During the Stage 2 lowstand, sands from the fluvial-deltaic system are believed to have bypassed the shelf and accumulated in upper slope minibasins (Anderson et al., 1996; Winker, 1996; Anderson and Rodriguez, 2000). These sands were transported through a depression in the shelf break south of the Trinity incised valley. Only those slope basins connected to that depression would have received sand from the shelf.

As sea level rose, a large shelf-margin delta developed and Units c through l were deposited (Fig. 5-15). This delta was fed by the combined Trinity/Sabine/Brazos River. The radiocarbon dates show that the development of the shelf-margin delta continued well into the transgression (Fig. 5-19). The shelf edge delta shows both progradational and aggradational geometries that reflect delta lobe shifting events. The lobe switching events may have been tied to 5th order fluctuations in sea level.

As transgression continued, the TSBD contracted and began to fill a single minibasin that is situated immediately down dip of the Trinity/Sabine/Brazos incised valley (Fig. 5-12). Eventually, transgression resulted in overstepping of the shelf margin delta and sedimentation was confined to the Trinity/Sabine/Brazos valley. The high
sediment supply of the Brazos river enabled it to quickly fill its lower valley and shift to the west out of the study area (Abdulah, 1995; Anderson et al., 1996). The lower sediment supply of the Trinity and Sabine rivers resulted in that valley being underfilled with fluvial sediments and later filled with estuarine deposits (Thomas and Anderson, 1994).

The stratigraphic development in the study area was controlled by the different sediment supplies of the two separate fluvial systems that occupied the area. The much higher sediment supply of the WLD enabled that system to prograde across the shelf as sea level fell. In contrast, it took much longer for the lower sediment supply Trinity/Sabine/Brazos system to deliver a significant volume of sediments to the outer shelf and the TSBD did not begin to form until the very end of the regression. A similar situation exists today between the Mississippi delta, which has already prograded to the outer shelf, and smaller rivers such as the Brazos and Colorado, which are constructing small wave-dominated deltas that have not prograded across the shelf.

Implications

Sequence stratigraphic models such as those of Vail and others (1977) and Van Wagoner and others (1988) have become widely popular for both oil industry and academic stratigraphic studies. Vail and others based their models on low- to intermediate-resolution seismic data and Van Wagoner and others based their models primarily on outcrop and well-log data. These models work very well for studies at
similar scales and are useful as a starting point for higher resolution studies. Detailed studies of Quaternary strata are able to further refine sequence stratigraphy models. These refinements are not just applicable to other Quaternary examples, but also to larger scale studies. A study of the late Quaternary evolution of the west Louisiana continental shelf shows that, overall, sequence stratigraphic models are applicable to modern systems. However, the detail provided by such a study presents a few lessons that can be applied to other areas.

Sea level was significantly lowered at three different periods during the deposition of the units in this study (Fig. 5-5). The system response to each of these periods is dramatically different ranging from a major sequence boundary, to a minor interfluvue sequence boundary, to now response at all. The three highstands following the formations of each of these boundaries also show striking differences. Lastly, the shelf margin delta and its associated sand bodies do not occur in the systems tract predicted by basic models.

The Stage 6 is the only sequence boundary in the area that represents a major fluvial-erosional surface (Fig. 5-13). This contrasts with the Stage 2 sequence boundary in the area, which consists only of amalgamated distributary channels (Fig. 5-13). The reasons for this are simply that the paleo-Mississippi River occupied the area during the Stage 6 lowstand and shifted to the east during Stage 2 where it formed the Mississippi Canyon (Weimer, et al., 1998). Sea level during both of these lowstands was below the shelf break yet the surfaces look quite different over a few hundred kilometers. Sequence
boundaries have been broken into two different types, based on the amount of erosion (Vail et al., 1984) and it has been argued that the two types of sequence boundaries correlate to different amounts of sea-level lowering. The difference between the Stage 2 and Stage 6 sequence boundaries in the area point out the lack of significance of defining two fundamentally different types of boundaries. Some authors (e.g., Posamentier and Allen. 1999) have suggested dropping terminology that differentiates between types of sequence boundaries. This study indicates the validity of such an approach.

The Stage 4 sequence boundary was not identified in this study. Sea level did not fall below the shelf break during Stage 4 (Fig. 5-5). Apparently, the paleo-Mississippi system continued to prograde through the Stage 4 lowstand and the subsequent sea-level rise. This presents another line of evidence against trying to define sequence boundaries for periods sea level did not fall below the shelf break. It is also why the entire period from Stage 5e to Stage 2 (Fig. 5-5) is referred to as the highstand in this study.

The 5e maximum flooding surface was used to correlate throughout the area and to other studies. The Stage 3 flooding surface was not identified in this study. The Stage 3 flooding surface has been identified as a major surface elsewhere in the northern Gulf of Mexico (Rodriguez, et al., 2000). The paleo-Mississippi system seems to have carried enough sediment that, even with the increased accommodation space created during Stage 3, it continued to prograde across the shelf. This points out the difficulty of correlating any but the most significant surfaces from one drainage basin to another.
Conclusions

The nine seismic facies identified in the area combined with the core data set indicate sandy incised fluvial valley, sandy distributary channel and mouth bar, and prodelta facies associated with the WLD. The vertical succession of these facies is that of a prograding, sandy fluvial-dominated delta. This delta prograded to the southwest and then westward landward of a salt barrier at the outer shelf. Only prodelta deposits exist seaward of the shelf break. A second delta, the TSBD, occurs at the shelf break. An aerially restricted, but thick, sandy distributary channel and mouth bar complex formed at the mouth of the incised valley and prograded across the prodelta. The shelf-margin delta is characterized by continuous reflection patterns and sediments recovered from this delta are silts and clays. Upper slope minibasins contain mounded chaotic seismic facies that are interpreted as sediment gravity flow deposits. The deltaic and sediment gravity flow deposits are draped by a hemipelagic facies that is characterized by strong, continuous parallel reflections.

Two distinct phases in the evolution of the deltas in the study area were reconstructed using the oxygen isotope curve as a proxy for sea-level change, sequence stratigraphic techniques for correlating the deposits, and paleontological and $^{14}$C dates for chronostratigraphic control. The WLD formed during the highstand and prograded across the shelf onto the Stage 5e MFS until the end of Stage 3 when it shifted to the east. The younger TSBD was formed by the combined Trinity/Sabine/Brazos River. Development of this delta began at the latest stage of the highstand, only after the
moderate sediment supply Brazos River combined with the low sediment supply Trinity/Sabine system, and then continued to form during the transgression. The Stage 2 sequence boundary is contained within the deltaic package.

During the maximum lowstand, sediment bypass of the shelf and into the slope minibasins occurred. Sand seems to be limited to the minibasins directly downdip of the Trinity/Sabine/Brazos incised valley.

Sand delivery to the study area was diachronous. The WLD became progressively less sand-dominated through the highstand. The combined Trinity/Sabine/Brazos system delivered sand across the shelf only during lowered sea level. Knowledge of the basin characteristics and climate in the drainage basin, in addition to basic sequence stratigraphy, is necessary to predict when sand will be delivered to a system.
COMBINED REFERENCES


Blankenship, D. D., Morse, D. L., Finn, C. A., Bell, R. E., Peters, M. E., Kempf, S. D.,
Hodge, S. M., Studinger, M., Behrendt, J. C., and Brozena, J. M., 2001 Geologic
controls on the initiation of rapid basal motion for West Antarctic ice streams: a
geophysical perspective including new airborne radar sounding and laser
altimetry results: Antarctic Research Series, v. 77, p. 105-121.

Bloom, A. L., and Yonekura, N., 1985, Coastal terraces generated by sea-level change
and tectonic uplift, in Woldenberg, M. J., ed., Models in Geomorphology: Allen

Blum, M. D., 1990, Climatic and eustatic controls on Gulf Coastal plain fluvial
sedimentation: an example from the late Quaternary of the Colorado River: in
Armentrout, J. M., and Perkins, B. F., eds., Sequence Stratigraphy as an
Exploration Tool: Concepts and Practices in the Gulf Coast: GCSSEPM
Foundation Eleventh Annual Research Conference Program and Abstracts, p. 71-
83.

Blum, M. D., 1994, Genesis and architecture of incised valley fill sequences: a late
Quaternary example from the Colorado River, gulf coastal plain of Texas, in
Weimer, P., and Posamentier, H. W., eds., Siliciclastic Sequence Stratigraphy:
Recent Developments and Applications: American Association of Petroleum
Geologists Memoir 58, p. 259-283.


Holdford, L., 1995, Quantitative Sequence Biostratigraphy of Late Quaternary Foraminifera on the Texas Continental Shelf [unpub. M. A. Thesis]: The University of Texas at Austin, 160 p.


Phleger, F. B., 1956, Significance of living foraminiferal populations along the central Texas coast: Contributions from the Cushman Laboratory for Foraminiferal Research, v. 7, p. 106-151.


cycle: Gulf Coast Association of Geological Societies Annual Meeting Transactions. v. L. CDROM.


APPENDIX 1: MAGNETIC SUSCEPTIBILITY MEASUREMENTS FOR NBP9801 CORES

Magnetic susceptibility data was collected on all of the NBP9801 cores. This data is useful for differentiating between glacial-marine sediments and subglacial till (Anderson, 1999). Till tends to have constant susceptibility measurements because it is a diamicton with all of the different materials evenly mixed. Glacial-marine sediments tend to have magnetic susceptibility spikes in layers of concentrated ice rafted debris. Diatomaceous muds tend to have low susceptibility but may also have spikes. Because this data is averaged over about 5 cm, each end of the core tends to have low susceptibility values. Units are SI magnetic susceptibility units.
TC-17 and PC-17*
The measurements for this core had to be redone with a different tool. The units are cgs.
PC-19*
The measurements for this core had to be redone with a different tool. The units are cgs.
TC-21
## APPENDIX 2: INFORMATION ABOUT FORAMINIFERA USED FOR RADIOCARBON DATING

<table>
<thead>
<tr>
<th>Core Number</th>
<th>Depth (cm)</th>
<th>Mass (mg)</th>
<th>Foraminifera Found in Sample</th>
</tr>
</thead>
<tbody>
<tr>
<td>DF80-153-1</td>
<td>8</td>
<td>2.9</td>
<td>Globoratalia, Ehrenbergina, Uvigerina</td>
</tr>
<tr>
<td></td>
<td>121</td>
<td>12.8</td>
<td>Uvigerina, Globoratalia</td>
</tr>
<tr>
<td>NBP9801-17</td>
<td>218</td>
<td>7.2</td>
<td>Pachyderma, Uvigerina, Globoratalia, Ehrenbergina, Lagena, Orbulina</td>
</tr>
<tr>
<td></td>
<td>270</td>
<td>6.7</td>
<td>Uvigerina, Orbulina, Fissuarina, Lagena</td>
</tr>
<tr>
<td></td>
<td>370</td>
<td>5.5</td>
<td>Uvigerina, Pachyderma, Globoratalia, Ehrenbergina, Orbulina</td>
</tr>
<tr>
<td></td>
<td>390</td>
<td>5.9</td>
<td>Pachyderma, Globoratalia, Ehrenbergina, Orbulina, Lagena</td>
</tr>
<tr>
<td>NBP9801-19</td>
<td>117</td>
<td>6.5</td>
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</tr>
<tr>
<td></td>
<td>152</td>
<td>6.9</td>
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</tr>
<tr>
<td>NBP9801-26</td>
<td>20</td>
<td>12.3</td>
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<tr>
<td></td>
<td>67</td>
<td>6.0</td>
<td>Uvigerina, Globoratalia, Ehrenbergina, Pachyderma, Fissurina</td>
</tr>
<tr>
<td></td>
<td>107</td>
<td>15.5</td>
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</tr>
<tr>
<td></td>
<td>113</td>
<td>7.3</td>
<td>Fissurina, Pachyderma, Uvigerina, Globoratalia, Ehrenbergina, Pyrgo, Fissurina</td>
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<tr>
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<td>139</td>
<td>12.2</td>
<td>Pyrgo, Fissurina</td>
</tr>
</tbody>
</table>
APPENDIX 3: CLIMATE REFERENCES FROM FIGURE 4-3


Holliday, V.T., and Haynes, C. V., Jr., 1994, Geoarchaeology and geochronology of the Miami (Clovis) Site, Southern High Plains of Texas: Quaternary Research. v. 41, p. 234-244. (10)


Labeyrie, L. D., Duplessy, J. C., and Blanc, P. L., 1987, Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years: Nature. v. 327, p. 477-482. (13)


