12-17-2010

Configuration of the Pleistocene Surface Beneath Cat Island, Mississippi and Implications for Barrier Island Formation and Evolution

Kathryn Rose

University of New Orleans

Configuration of the Pleistocene Surface Beneath Cat Island, Mississippi, and Nearshore: Implications for Barrier Island Formation and Evolution

A Thesis

Submitted to the Graduate Faculty of the
University of New Orleans
in partial fulfillment of the
requirements for the degree of

Master of Science
in
Earth and Environmental Sciences
Coastal Geology

by

Kathryn Virginia Georgina Rose

B.G.S., University of New Orleans, 2007

December 2010

Acknowledgements

The U.S. Geological Survey and the University of New Orleans funded this study cooperatively, and many people contributed to its completion. First, I would like to thank my co-advisors, Dr. Dawn Lavoie, U.S. Geological Survey, and Dr. Mark Kulp, University of New Orleans for their mentoring, assistance in arranging the geophysical surveys, interpreting data and reviewing this manuscript. The input of my committee members, Dr. Mike Miner and Dr. Ioannis Georgiou, has also been invaluable. I would also like to thank Phil McCarty, Dallon Weathers and Dane Fischer with UNO, and Nancy DeWitt and Dana Weise with the USGS St. Petersburg Coastal and Marine Science Center for their help in carrying out the surveys; Arnelle Harrison, also with the USGS St. Petersburg Coastal and Marine Science Center for processing the SB-424 chirp data; and Elizabeth Pendleton with the USGS Woods Hole Seafloor Mapping Group for processing the bathymetry data and lending her expertise on processing methods. I am extremely grateful to Jim Flocks with the USGS St. Petersburg Science Center for making equipment and staff available. Finally, I would like to thank George Boddie for providing access to Cat Island during fieldwork.

Table of Contents

List of Figures

Abstract

The mechanism of Holocene barrier formation aids in determining island geomorphologic responses to modifying climatic processes of the surrounding environment. The geometry and composition of local antecedent topography plays a role in barrier formation by providing an elevated base, nucleus for sedimentation and local sediment supply. Investigation of barriers' subsurface geology provides insight into island formation and evolution. High-resolution shallow seismic data acquired in the island's nearshore zone and interior canals, correlated with existing drillcore data, reveal that Cat Island, MS is situated over an Oxygen Isotope Stage 3 Phase 3 paleochannel located between two topographic high-grounds of the Pleistocene surface. Beach ridge strandplain sets on Cat Island provide additional evidence supporting the island's formation over a relict depocenter. A new, 4-stage model for Cat Island development and evolution incorporating the influence of pre-existing topographic high-grounds and abundant local sediment supply provided by a backfilling fluvial channel is presented here.

Keywords: Cat Island, MS, Holocene barrier, Pleistocene surface, paleochannel, beach ridges

Introduction

The inception, evolution and fate of mid-Holocene coastal plain barrier islands along the Gulf of Mexico and the Atlantic coasts has been the subject of ongoing debate in scientific literature (Curray and Moore, 1963; Swift, 1975; Field and Duane, 1976, 1977; Hoyt, 1967, 1970; Otvos, 1970; Otvos and Giardino, 2004; Schwartz, 1971; Swift, 1975; Tanner, 1990). After the 2005 Atlantic hurricane season when Hurricane Katrina inundated mainland Mississippi with storm surge of as much as 8.5 m at Pass Christian, the highest reported for that storm (National Hurricane Center [NHC], 2005), local, state and federal agencies have partnered to evaluate the condition of the deteriorating Mississippi barrier islands and develop restoration and management measures as part of a comprehensive storm protection and ecosystem restoration program (the Mississippi Coastal Improvement Program, or MSCIP, U.S. Army Corps of Engineers [USACE], 2009). Future decisions relating to the islands' management depend on their present state and the integrity and condition of the sand resources available for restoration, prompting a need for detailed geologic characterization of the region.

The Mississippi-Alabama (MS-AL) barrier islands form a linear chain approximately 104 km long along the north-central Gulf of Mexico coast 9-19 km-wide Mississippi Sound. From east to west, the chain includes Dauphin, Horn, East and West Ship, and Cat Islands (figure 1). These islands (excluding Dauphin Island, which is developed, and part of Cat Island which is privately owned) are part of the National Park System's Gulf Islands National Seashore (GUIS), which also includes parts of Santa Rosa Island in Florida. In 2002, the National Park Service (NPS) purchased approximately 40% of Cat Island from the family that has owned the island since 1910 (the Trust for Public Land, 2002). The seashore provides habitat for numerous wildlife species as well as those commercially important to the U.S. fishing industry. It is also a

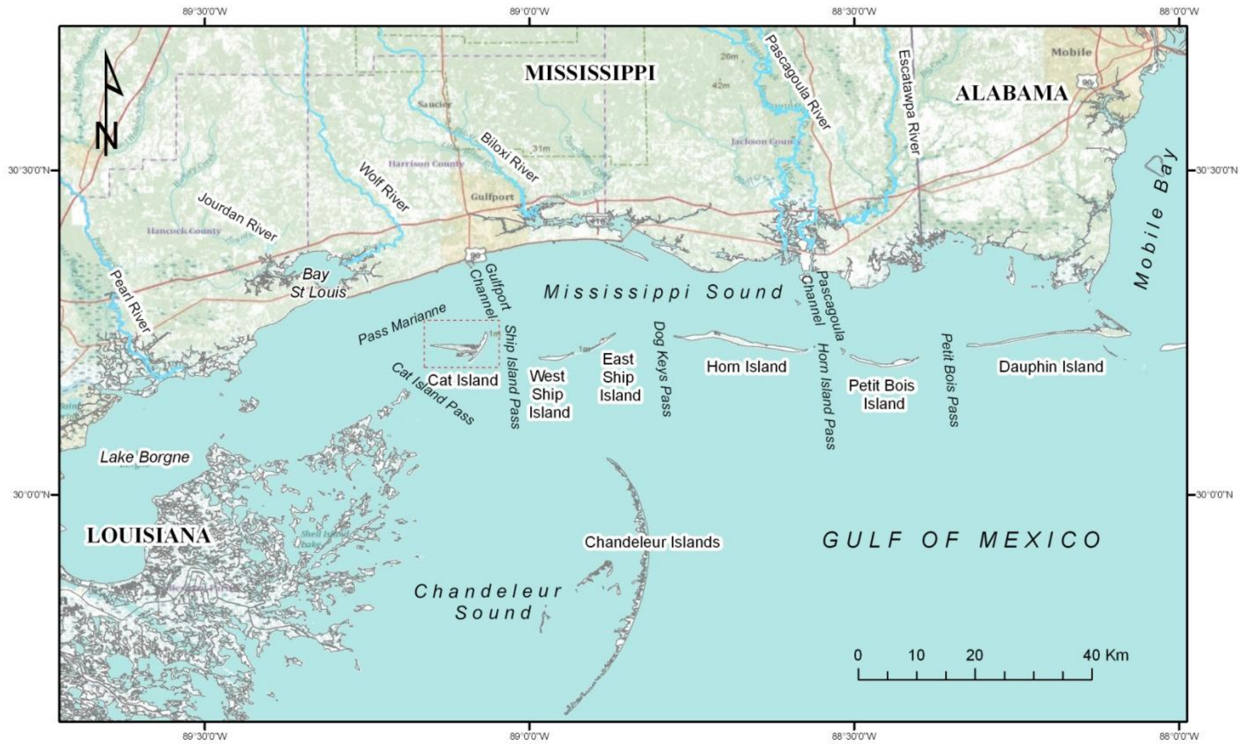


Figure 1. Map of the Northern Gulf of Mexico and Mississippi Sound. Study area is within red dashed line.

popular recreation area containing several historical and archaeological sites documenting the settlement and colonization of the Gulf region (NPS, 2009).

Perhaps because of its private ownership status, few studies have been conducted on Cat Island and its near-shore area, despite its unique position where processes associated with both Holocene coastal plain and Mississippi River delta plain evolution have influenced various stages of its geomorphic evolution producing well-preserved progradational beach ridge sets and the north-south trending spits that result in the island's distinctive T-shaped morphology (Otvos and Giardino, 2004). This study incorporates shallow high-resolution chirp seismic data from surveys conducted in 2008 and 2009 around Cat Island with existing drill core data and vibracore data taken on the island and in the immediate offshore. The previously acquired data are used to help define the depth and geometry of the underlying Pleistocene surface. This study proposes that Cat Island may have originated as mid to late Holocene sediments aggraded over an existing Pleistocene ridge that provided a stable island core, anchoring the successive sedimentary units that evolved into the modern barrier island. Presence of an underlying topographic high may explain the island's anomalous evolution and geomorphology relative to the other units of the MS-AL barrier island system. This information will enable managers to anticipate the island's geomorphologic response to future climatic stressors, such as sea level rise and extreme storms and to prioritize the restoration needs of the barrier island system. This study also contributes to ongoing research efforts to refine the geologic history and development of the northern Gulf coastal region and to understand and predict coastal response to tropical cyclone impacts and sea-level rise.

Hypothesis

This study proposes that the lateral stability and unique shape of Cat Island may be controlled by the presence of a pre-Holocene topographic high, presently underlying Cat Island, that provided a nucleus for sand accretion and continues to influence island position and stability. Improved understanding of the configuration of the Pleistocene surface underlying and in the vicinity of Cat Island will help to resolve questions regarding island formation and evolution as well as predict future trends.

Background

Quaternary Evolution of the Northern Gulf of Mexico

The geologic evolution of the Northern Gulf of Mexico has been controlled by eustatic glacial-interglacial cycles and corresponding changes in sea level (Anderson et al., 2004; Boyd et al., 1989; Curray, 1960; Fisk, 1947; Fisk, et al, 1954; Frazier, 1967; Kindinger, 1988, 1989; Kindinger et al, 1989; Kulp et al., 2002; Lopez et al., 1997; McBride et al., 2004; Morton and Boyd, 1996; Otvos, 2005 b; Shepard, 1960 b; Wilkinson, 1975). Sea level fluctuation is a forcing mechanism responsible for erosional and depositional sequences that affect the geomorphology of coastal and marine sedimentary systems. Transgressing and regressing shorelines alter the hydrodynamic environment across marine basins by re-working and re-distributing surficial sediment. Depocenters migrate landward or seaward and laterally as changing shelf gradients cause rivers to avulse (Scruton, 1960). Submerged surfaces can be eroded or preserved while subaerially exposed surfaces are weathered creating distinct horizons between unconformable sedimentary packages. The resulting lithofacies found in the northern Gulf are the product of the interactions between sea level fluctuation, sediment supply and the underlying receiving basin geometry (Fisk, 1947; Fisk et al, 1954; Frazier, 1967; Morton and Suter, 1996; Scruton, 1960). Subsurface mapping of the northern Gulf of Mexico continental shelf has identified multiple paleochannel and deltaic complexes constructed during the major stages of glacial-interglacial cycles of the late Quaternary (Anderson et al., 2004; Boyd et al., 1989; Fisk, 1947; Fisk, et al, 1954; Frazier, 1967; Greene et al, 2007; Kindinger, 1988, 1989; Kindinger et al, 1989; Kulp et al., 2002; Lopez et al., 1997; McBride et al., 2004; Morton and Boyd, 1996; Otvos, 2005 b; Shepard, 1960 b) (figure 2). Sedimentary units representing successive geologic events can be interpreted using a sequence stratigraphic framework, where the erosional surfaces identify the

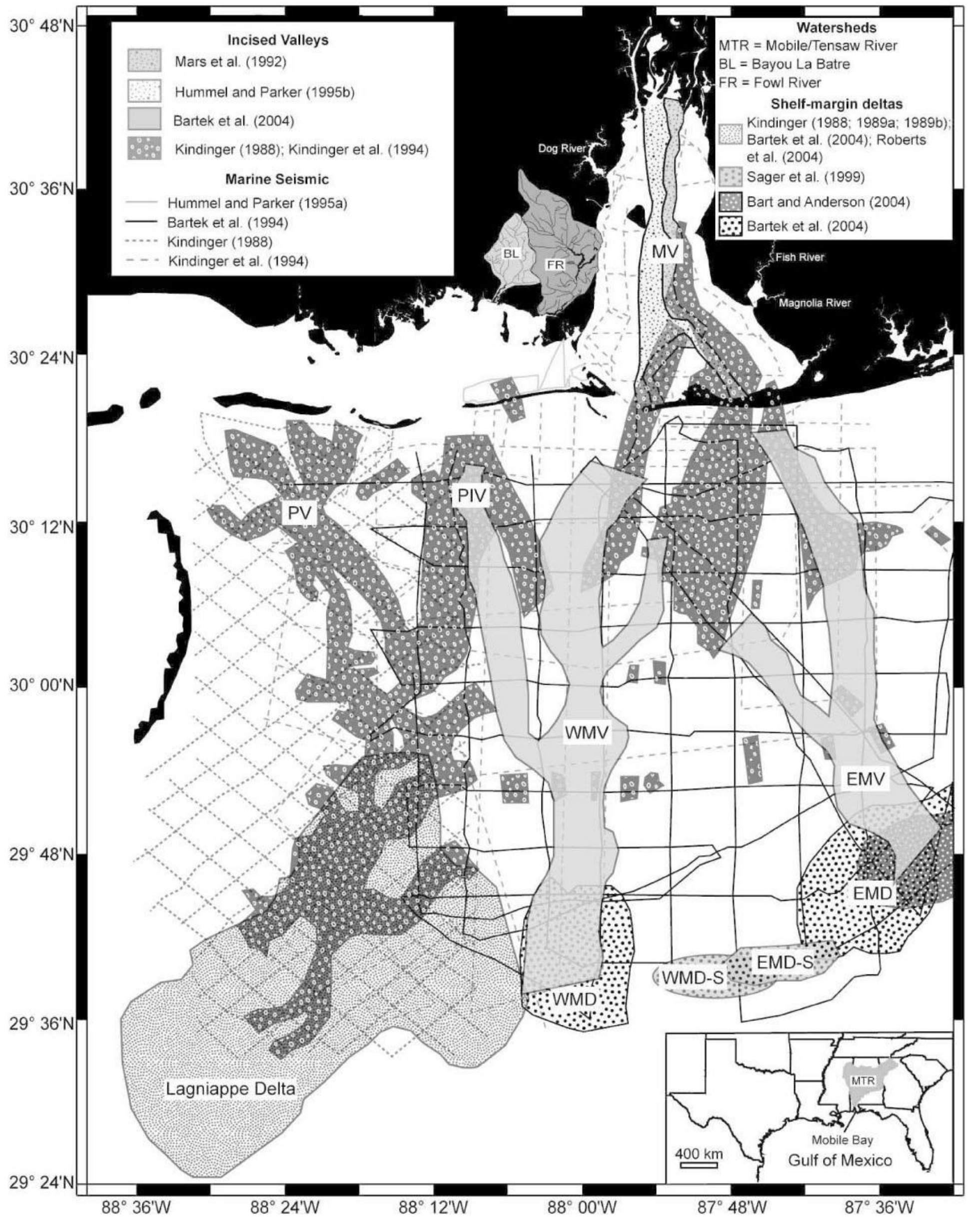


Figure 2. Composite map showing the geographic locations and extent of incised valleys of the most recent sea level lowstand, Oxygen Isotope Stage 2, PV: Pearl Incised Valley (Kindinger, 1988; Kindinger et al., 1994); PIV: Pascagoula Incised Valley (Kindinger, 1988; Kindinger et al., 1994); MV: Mobile Incised Valley (Hummel and Parker, 1995 b; Kindinger et al., 1994; Mars et al., 1992); WMV: West Mobile Valley (Bartek et al, 2004); EMV: East Mobile Valley (Bartek et al., 2004), and shelf edge deltas, EMD (Bart and Anderson, 2004; Bartek et al., 2004) and EMD-S (Sager et al., 1999): East Mobile Deltas; WMD (Bartek et al., 2004), WMD-S (Sager et al., 1999): West Mobile Deltas; Lagniappe Delta (Kindinger, 1988; 1989 a; 1989 b; Bartek et al., 2004; Roberts et al., 2004), (Greene et al, 2007)

base and upper limit of transgressive-regressive sequences constrained by glacial cycles (Kindinger, 1988; Kindinger et al., 1989; Morton and Suter, 1996) enabling basin-wide, temporally constrained correlation of the superimposing strata. Throughout the late Quaternary

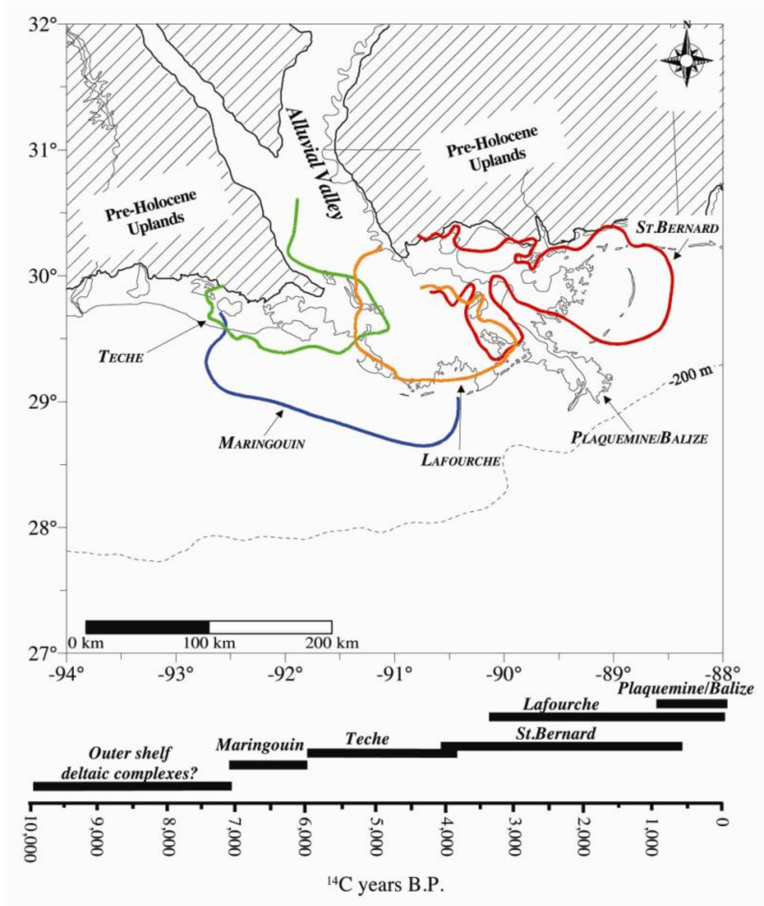


Figure 3. Geographic distribution and proposed chronology for Holocene delta complexes of the Mississippi River (modified from Frazier, 1967)

(~1.8 Ma ybp) northern Gulf of Mexico geology has been dominated by the Mississippi River system, which, by means of delta switching, deposited multiple overlapping deltas across the continental shelf out to the shelf break. River avulsion, forced by changes in the gradient of the coastal plain providing more favorable paths to the basin, shifted depocenters to new locations. As active distributaries became progressively abandoned, vertically stacked units of deltaic sediments formed (Fisk et al., 1954; Frazier, 1963; Scruton, 1960). Abundant sediment supply or falling sea level allowed deltas to prograde out into the basin, whereas abandonment (sediment starvation), rising sea level and subsidence leads to the “destructive” phase of shoreline submergence and erosion. Fringing barrier islands or shoals in alignment with the former delta

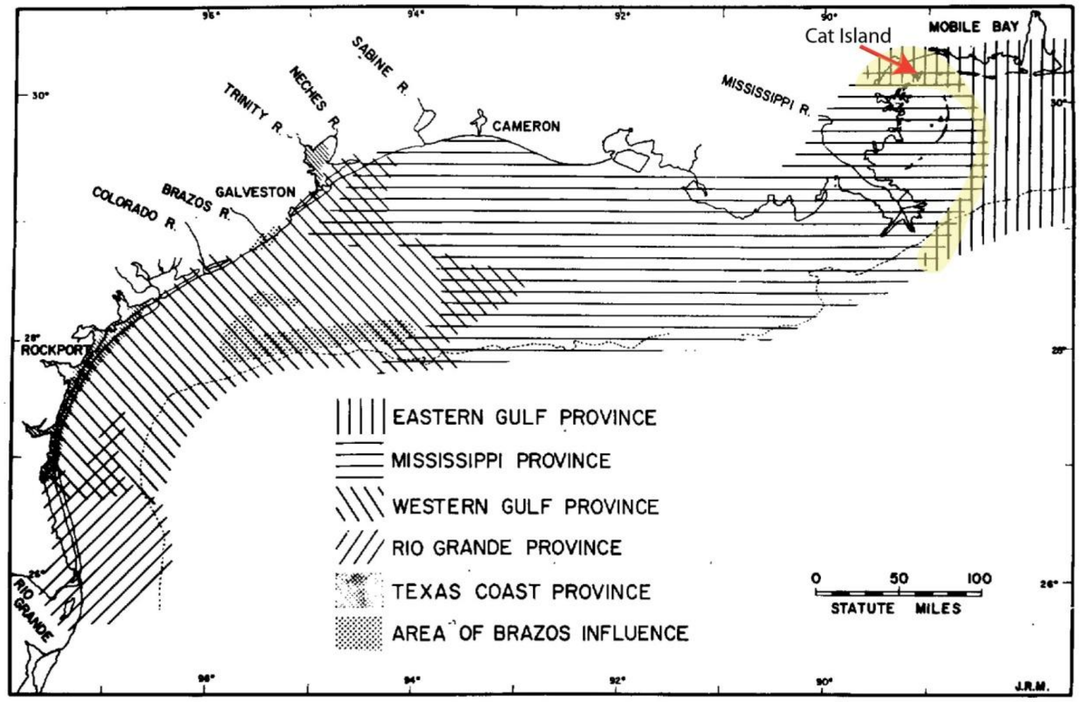


Figure 4. Depositional provinces in the northern Gulf of Mexico by mineralogy type, with Eastern Gulf and Mississippi Province transition zone highlighted (modified from Van Andel, 1960)

shoreline may be constructed from re-worked deltaic deposits (Penland et al., 1989). Five major delta complexes of the late Holocene (~7000 ybp) extending from southeast to southwest of the modern Mississippi River form a delta complex that comprises the subaerial land mass of southern Louisiana have been identified in sedimentary facies (Frazier, 1967; Boyd, et al., 1989; Roberts, 1997; Kulp et al., 2002) (figure 3). Underlying the southeastern margin of the Holocene delta complex is an earlier stage delta from the late Wisconsinan, the Lagniappe delta (Kindinger, 1989; Kindinger et al., 1994) (figure 2).

Sediments from the Mississippi River dominate the Western Gulf region, extending from the Mississippi River plain westward to central Texas, are distinct from the sediments of the adjacent Northeastern Gulf region. The Gulf of Mexico encompasses six depositional provinces based on sediment mineralogy and texture, used to identify provenance and age, and amount of weathering and erosive re-working. Transition zones occur between provinces where type minerologies can be mixed vertically and laterally (Hsu, 1960; Isphording, 1989; Van Andel, 1960) (figure 4). The Mississippi Province is composed of sediment derived from the continental sources of the Mississippi River's extensive watershed. Mississippi Province sediments, dominated by a relatively young, unstable petrologic assemblage, have undergone very little weathering or diagenetic transformation relative to the Eastern Gulf Province sediments (Van Andel, 1960). The Mississippi River sediments are composed of amphiboles, dolomite, pyroxenes, epidote, ilmenite and biotite, abundant feldspar and a montmorillonite-illite-kaolinite suite of clays and (Hsu, 1960; Foxworth et al., 1962; Isphording, 1989; Van Andel, 1960).

The northeastern Gulf of Mexico region encompasses the gently sloping continental shelf south of the present-day Louisiana, Mississippi, and Alabama shorelines out to the shelf break, which lies approximately 75 m below sea level (bsl) (Kindinger, 1988; Kindinger et al., 1989). The late Quaternary sedimentary record is ordered into five chronostratigraphic stages based on sequences of transgressive-regressive stratigraphic packages (Kindinger, 1988; Kindinger et al., 1989). The major erosional surface horizons and sedimentary boundaries within are correlated by Kindinger (1988) and Kindinger et al. (1989) with oxygen isotope dates of the glacial-interglacial cycles from the early Wisconsinan lowstand, ~180,000-150,000 ybp to present (Kindinger, 1988; Kindinger et al., 1989; Morton and Suter, 1996) (figure 5). During Stage 1, at the base of the earliest unit is an erosional surface boundary that formed on the then-subaerial coastal plain, when sea level was at its last major lowstand or regression (early Wisconsinan). The entire Gulf of Mexico basin was exposed and few incised fluvial channels and deposits were evident. As the climate warmed, North American glacial melt inundated the basin, initiating a transgressive phase. Rapid sea level rise during the middle Wisconsinan (Stage 2, ~120,000 ybp to 85,000 ybp) is indicated by facies thickening landward as depocenters retreated.

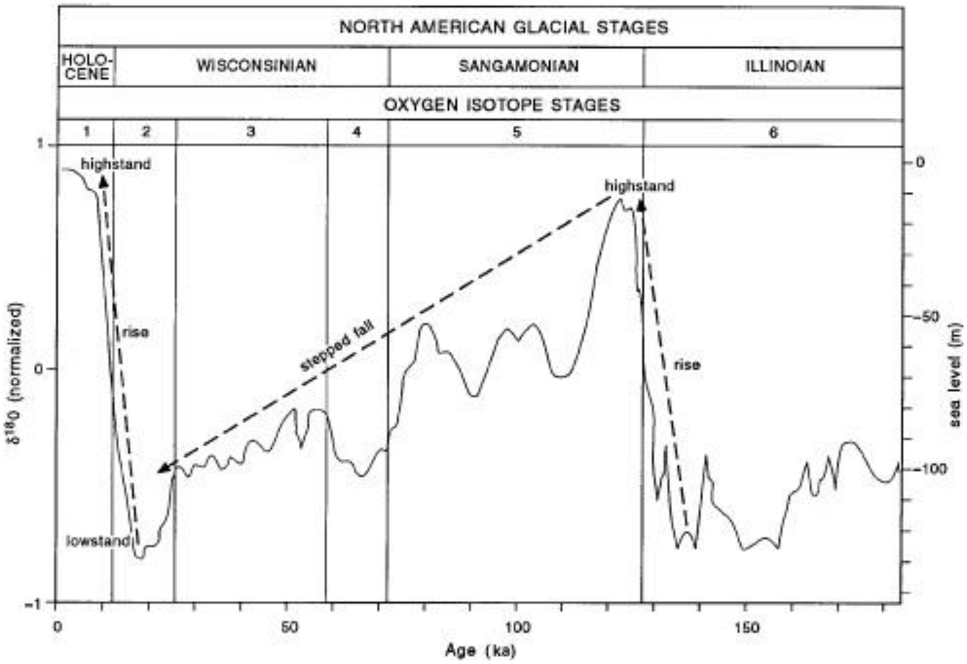


Figure 5. Oxygen-Isotope Stages and inferred sea level fluctuations for the past 180,000 years (Morton and Suter, 1996)

During Stage 3 (late Wisconsinan, ~85,000 to 24,000 ybp), the northeastern Gulf entered another period of sea level fall. Three phases of Stage 3 are evident in the sedimentary sequences of this unit, suggesting a stair-stepped regression with a brief stillstand interval. During Phase 1, multiple fluvial networks cut through the mid-shelf, eroding sediment, and in Phase 2 the regression slowed to a stillstand, allowing for a period of deltaic progradation as alluvial sediments built out to the shelf break (Kindinger, 1988; Kindinger et al., 1989; Greene et al., 2007) (figure 2). The provenance of these alluvial sediments is the metamorphic-intruded rocks of the Southern Appalachian region. This mineralogic suite (the Eastern Gulf Province) is characterized by a high ratio of stable heavy minerals to quartzose sand and the same suite of clays, but with decreasing abundance of kaolinite and is predominant through Phase 2 to present-day Northern Gulf sediments (Hsu, 1960; Isphording et al., 1989; Foxworth et al., 1962; Kwon, 1969).

Maximum lowstand of Stage 3 occurred during Phase 3, marking the transition between the Pleistocene and Holocene epochs (Kindinger 1988; Kindinger, 1989). The alluvial plain extending across the entire Gulf basin was exposed and weathered creating a heavily dissected paleosol, the Type 1 regional unconformity that serves as the base for the subsequent Holocene facies, also referred to as the Pleistocene surface. Incisions in the surface are the result of erosion by fluvial networks cutting across the inner shelf to the shelf edge, where shelf edge (Type 1) deltas are constructed (Brooks et al, 1995; Greene et al, 2007; Morton and Suter, 1996). The Pleistocene surface is composed of heavily oxidized, non-fossiliferous clays of the Beaumont formation (Curray and Moore, 1963; Shepard, 1960 a; Brooks et al, 1995; Lopez et al, 1997). Regionally, the horizon dips to the southwest, following the configuration of the Gulf of Mexico basin (Fisk, 1947) and in the Mississippi Sound, slopes toward the west from 3 m bsl at Horn Island to an average depth of 10 m bsl north of Ship Island (Curray and Moore, 1963). Beneath Matagorda and St. Joseph Islands, TX in the northwestern Gulf, it has been found at depths of ~7 - ~18 m bsl (Shepard, 1960 a). Incisions in the lowstand surface are the remnant of fluvial channels that prograded out to the shelf, constructing shelf edge or Type 1 deltas during shoreline regressions and lowstands (Kindinger, 1988; Kindinger 1989; Kindinger et al., 1989; Boyd et al., 1989; Brooks et al, 1985; Anderson et al, 2004; Greene et al., 2007; McBride, et al, 2004).

Transgression resumed during the late Pleistocene-early Holocene (Stage 4, ~24,000 to 7000 ybp) (Kindinger, 1988; Kindinger, 1989). The shoreline retreated at a steady pace, eroding surface sediments and re-depositing them in a thin, relatively smooth sheet as depocenters shifted to the inner shelf. This transgressive unit, the MAFLA sandsheet, extends from the far eastern to central northern Gulf, where it interfingers with the eastern terminus of the Mississippi River's

St. Bernard delta lobe deposits, creating a transition zone between the two regions (Kindinger, 1989; Kindinger et al., 1994; McBride et al., 2004). The MAFLA sandsheet is composed of clean, fine to medium quartz sand, is up to 5 m thick, extends up to 400 km long along strike with the present-day shoreline and extends out to the outer shelf with widths between 25-100 km (Rodriguez et al., 2004). Some minor stream incisions are found at this position in the sedimentary record, indicating that another minor sea level fluctuation may have occurred, shifting depocenters seaward again (Kindinger, 1988). Sea level rose to its current position during Phase 5 (~7000 ybp to present) (Kindinger, 1988; Kindinger, 1989). The major geologic event of this stage is marked by the progradation of the St. Bernard delta complex from ~3800 to ~2000 ybp (Frazier, 1967; Kindinger, 1988; Kindinger et al, 1989; Roberts, 1997; Brooks et al., 1995). The volume of sediment transported by the Mississippi River at this time was large enough to outpace rising sea level and allow the delta to prograde. The fine-grained, prodeltaic Western Gulf province sediments extended along the sea floor out to the central mid-shelf. Continued sea level rise forced the depocenter and the delta front to shift further shoreward, initiating the abandonment and destructive phase of the delta cycle.

As the northwestern Gulf of Mexico basin was transgressed during the early Pleistocene, a succession of Plio-Pleistocene sedimentary units began to onlap the Cenozoic Citronelle Formation of the southern continent. A series of basin-ward sloping, stepped terraces across the coastal plain indicate faulting of the basin margin that occurred in response to regional sediment loading (Fisk, 1947; Flocks et al., 2009; Otvos, 2005 b; Saucier, 1963). These terraces constitute the Plio-Pliestocene Prairie Formation extending across the continent above the northern Gulf, downwarping at the seaward margin where it is onlapped by recent Holocene deposits in the Gulf basin. The downwarping occurred in response to Quaternary sediment loading of the basin, prior to the Holocene Mississippi River deltaic phases and construction of the southern Louisiana land mass (Flocks et al., 2009; Lopez et al, 1997; Otvos 2005 b; Saucier, 1963). This flexure created a system of faults extending west of the Pearl River to Baton Rouge, LA, and marked the location of the early Holocene Pontchartrain Basin shoreline, roughly the same position as its present-day northern shoreline, at ~7000 to ~4000 ybp (Lopez et al., 1997; Otvos, 2005 b). As sea level rose, embayments formed as estuaries along the Gulf coast flooded. Depressions in the Pontchartrain Basin floor seaward of the shoreline, formed by sediment loading and faulting marked the future sites of lakes Maurepas, Pontchartrain and Borgne (Saucier, 1963; Treadwell, 1955). At ~4000 ybp, the Mississippi River avulsed from its western course to the east and began constructing the St. Bernard delta. The 4th and 5th St. Bernard sub-delta lobes built out ~ 80 miles eastward, enclosing the Pontchartrain embayment and establishing the southern shore of Lake Pontchartrain and the southeastern Louisiana shoreline (Fisk, 1944; Frazier, 1967; Saucier, 1963).

Barrier Islands

Holocene barriers occur worldwide on tectonically active and passive margin coasts with wide, gently sloping continental shelves (Hayes, 1979; Stutz and Pilkey, 2002). Defined as "long, narrow strips of detritus raised above sea level and extending some distance from the original land parallel to the general trend of the coast" (Leont'yev and Nikiforov, 1965), barrier islands are situated off the mainland shores of approximately 7 % of the world's coastlines, on every continent except Antarctica (Stutz and Pilkey, 2002; Hayes, 1979). Due to their widespread distribution across varying climates and geologic settings, however, they are subject to a spectrum of modifying physical processes (Field and Duane, 1976; Hayes, 1979; Hoyt, 1967, 1970; Hoyt and Henry, 1967; Otvos, 1970, 1977; Schwartz, 1971; Shepard, 1960; Swift,

1975) and are not restricted to shore-parallel trends or elongate morphology (Field and Duane, 1976). Barrier islands are distinguished from offshore barrier bars, which are submerged at high tides, and from coral reefs, which are composed of carbonates (Hoyt, 1967). They can extend to over 150 kilometers in length and are usually a few miles wide, or they may form a chain of islands separated by tidal inlets that accommodate tidal exchange between the backbarrier estuary and open ocean. Typically, the subaerial portion of the island is composed of dune and beach ridge complexes, demonstrating island progradation, backed by a flat or marsh. The subaerial elevations of mature, degrading islands can be as low as 1-2 m. They are separated from the mainland by a shallow bay, sound or lagoon that differs markedly in hydrodynamics and ecology from the open ocean and so have a profound effect on the mainland coasts they fringe (Johnson, 1919; Shepard, 1960 a; Hoyt, 1967; Otvos, 1970; Ovianki, 1998). Swift (1975) proposes that the existence, geometry and behavior of all barrier sub-environments are dependent on the response of the shoreface to the prevailing hydrodynamic regime (Swift, 1975). The shoreface is the site where the barrier and the marine environment interface and where sediment is transported to the barrier from its surrounding environment. Modification to the shoreface alters sand transport dynamics to the subaerial barrier, controlling variable deposition or erosion on the barrier, thereby determining barrier geomorphology (Swift, 1975).

Understanding the conditions and mechanisms of barrier island genesis provides insight to past global climate and local variations. The contemporaneous appearance of Holocene barriers ~7000 - ~3000 years ago (Stutz and Pilkey, 2002; Stanley, 1997) implies a geologic response to eustatic sea-level rise and stabilization, but the variations among their subsequent geomorphologies are the result of regional and local geology and hydrodynamics and climate. These physical processes control the inception of barriers, determine their geologic structure and sedimentary composition, and thus, how they respond to the modifying forces in their environment. Investigation of barriers' geologic evolution provides insight as to previous environmental conditions, such as sea level, during barrier formation (Rodriguez, et al., 2004; Schwartz, 1971). In many cases, the importance of formation processes may be secondary to the influence of the island's current surroundings. Most Holocene barriers have been removed from their original constructive environment and geological setting by island migration, or the original environments may no longer exist, as in transgressed deltaic systems (Hayes, 1979; Hoyt and Henry, 1967; Otvos, 1970; Schwartz, 1971; Swift, 1975). The current local climatic and hydrologic conditions and underlying geology are, therefore, the dominant factors in continuing barrier evolution and survival (Field and Duane, 1976; Rosati and Stone, 2009).

Conceptual Models of Barrier Island Formation

Early barrier formation hypotheses

Johnson (1919) reviewed several early models of barrier island formation and concluded that three of the proposed models provided credible mechanisms for initiation of major Holocene barriers. Johnson (1919) favored the hypothesis proposed by de Beaumont in 1845, which suggests that emergent bars built up from the seafloor by shore-normal wave action in the nearshore swash zone. A shallow nearshore slope in equilibrium dissipates incoming wave energy, allowing deposition of sediment and vertical accretion and lateral progradation. As waves break on the shoreline, energy is transferred from the waves to the beach. Wave energy dissipates allowing sediment deposition shoreward of the breaker zone into an initial ridge. The nearshore slope initially increases as the bar aggrades but receding wave currents also transport some sediment seaward of the ridge, maintaining the shoreface profile. Johnson (1919) favored this hypothesis because sediment is derived from the local seafloor, which he considered the only source of sufficient volume and continuity to build and maintain large barriers as they migrate across the shelf. A mechanism for maintaining an equilibrium profile, and thus continual sedimentation and growth is also provided by the de Beaumont (1845) model.

Bruun (1962) later articulated the concept of shoreface equilibrium as a mechanism for shoreface maintenance during periods of sea-level rise (Bruun, 1962). Fair-weather, constructive waves favor deposition on the upper shoreface, so that the shoreface progrades and displays a gently sloping profile. During major storms, high-energy waves break further up the shoreface, eroding sediment and depositing it in the nearshore, steepening the shoreface profile but shallowing the nearshore bottom, which then provides sediment for onshore transport during following fair weather intervals (Bruun, 1962). Equilibrium exists as the net volume of sediment in the barrier system is maintained, but transferred between the shoreface and the nearshore environments (Bruun, 1962; Swift, 1975). Shoreface equilibrium is also maintained during long-term sea level rise by the same process, but with sea level transgressing up the shoreface replacing storm waves as the process by which sediment is removed from the subaerial barrier (Bruun, 1962; Swift, 1975).

Similar but more complex hypotheses were developed, including a higher stillstand scenario during which subaqueous bars are formed and then exposed during sea level fall (Leont'yev and Nikiforov, 1965). Barriers formed by either of these models develop *in situ* in an open marine environment and would therefore be positioned over marine sediments both seaward and landward. The Leont'yev and Nikiforov (1965) model suffers the same limitations as de Beaumont's emerging barriers. As subaqueous bars build up to sea level they are impacted by high-energy waves that prevent deposition and terminate vertical growth at highest high tide elevations, and during sea level fall, subaqueous bars are subject to wave erosion, truncating the tops of bars, preventing vertical growth (Hoyt, 1967, 1970). Additionally, the de Beaumont model requires a smooth slope as a base, yet barriers are found on shelves where the surface topography is highly variable (Hoyt, 1967; Schwartz, 1975). Barriers are also frequently situated between lagoonal and open marine facies, sometimes resting directly over lagoonal facies. Johnson had also not considered the contribution of alluvial systems in supplying sediment (Hoyt, 1967, 1970; Shepard, 1960 a).

The Gilbert detached spit model (1885) describes the buildup of sediment at the downdrift margins of coastal headlands, bays and estuaries where the nearshore slope provides a shallow platform and reduced wave energy for deposition. Separation from the mainland occurs when the spit is breached during storms, creating an inlet, which may then be widened by erosive tidal currents and waves. As long as the updrift sediment supply is uninterrupted the spit will continue to grow at the downdrift margins. Johnson (1919) thought this could be a viable method of smaller island formation or progradation, but thought that sediment sources were not available to create and maintain major barrier islands. A third hypothesis put forth by McGee (1890) and mentioned only briefly by Johnson (1919), involved submergence of mainland shorelines as coasts subsided. As low-lying coastal areas are flooded, the elevated ridges and dunes on the upper shoreface would remain above water, becoming islands as they were separated from a receding mainland.

Shoreline engulfment hypothesis of barrier island development

Hoyt (1967, 1970) articulates the shoreline engulfment hypothesis, building on work by Zenkovitch (1962) and first suggested by McGee (1890). Hoyt (1967) observes four conditions found in present day barrier systems that would have to be explained by any model of formation: 1) the absence of open marine beach or shallow neritic sediments and fauna landward of the barriers; 2) the ability of barrier systems to re-form after termination by emergences; 3) the absence of a eustatic, higher than present sea level during the Holocene; 4) the development and maintenance of barriers during a slow rise in sea level. Sea-level rise is the necessary mechanism for initiating shoreline engulfment and island formation. To explain a sequence of barrier development and response, Hoyt (1967) used the Curray-Shepard sea level curve (derived from radiocarbon dates of barrier sediments from the Texas coast) advocating a rapid late Holocene sea level rise following a stillstand at ~7000 to 8000, then slowing at ~3000 to ~4000 ybp until reaching present levels (Curray, 1960). During shoreline transgression, the zone of high-energy breaking waves at the land-water interface moves landward across the shelf. Entrained surficial shelf sediment is deposited on the mainland coast. As the rate of sea-level rise slows, the coastline stabilizes and progrades and offshore bars, beach ridges and berms form in the foreshore as dunes grow landward. When the shoreline is completely engulfed by rising waters the ridge-dune complexes become the remaining subaerial expression of the former coastal plain. A shoreface equilibrium is established as these newly formed barriers derive sediment from the underlying beach and nearshore deposits of the former coast that is then re-worked and deposited by on-shore wave action leading to vertical and lateral accretion, allowing them to survive and recover from extreme storms (Hoyt, 1967). Finer sediment accumulates in the flooded, protected backbarrier lagoon, overlying former back-beach dune and ridge sands. The length of the island is initially determined by the length and continuity of the pre-submergence, landward ridge. Lower elevations along the ridge crests provide breaching sites for erosive currents and may lead to barrier segmentation, developing a chain of barrier islands rather than a single island (Hoyt, 1967; Hoyt and Henry, 1967; Shepard, 1960 a). During periods of high rates of sea-level rise, ridges may be eroded and submerged if their aggradation cannot outpace submergence. Conversely, if engulfment proceeds too slowly the initial lagoon may silt in, reversing submergence (Hoyt, 1967, 1970; Shepard, 1960 a). Shepard (1960 a) cites the central Laguna Madre, TX barrier as an example of this case, where he describes the former lagoon site infilled with sand and covered by algal flats and migrating dunes. The Intracoastal Waterway traverses the infilled central lagoon, connecting the formerly continuous backbarrier lagoons to the north and south (Shepard, 1960 a). Morton et al. (2000) attribute the lagoonal infilling of the Bolivar Peninsula, TX to a combination of recurved spit growth into the backbarrier and storm-driven washover fan deposition (Morton et al., 2000). With continued transgression and geomorphologic modification, successful barriers migrate landward as the mainland shoreline continues to retreat (Field and Duane, 1976; Hoyt, 1967; Hoyt and Henry, 1967; Shepard, 1960 a) and may eventually be positioned atop fine-grained lagoonal sediments (Hoyt, 1967, 1970).

In advancing the shoreline engulfment model, Hoyt (1967) stresses his first condition that the underlying lagoon sediments must be devoid of high-salinity open marine organisms; a transgressed coast would never have been fronted by open waters and the back-barrier sediments would therefore comprise bay-sound and lagoonal facies (Hoyt, 1967, 1970). De Beaumont's emergent shoal model, which would result in barriers overlying a marine environment, was dismissed based on core interpretations from Padre, Matagorda and Galveston Islands in Texas, the Georgia and Dutch coasts that showed no evidence of an open marine environment (Hoyt, 1967, 1970). Hoyt (1967) also questions whether a valid mechanism for bar emergence exists, citing wave tank studies in which bars aggraded vertically only as high as the high-water level, at which point washover erosion prevented further deposition and eroded any accumulated subaerial mass. He fails to find any cases of field evidence for emergent shoals beyond small, ephemeral islands and subaerial shoals but no intermediate or advanced landforms that should exist since the same hydrodynamic forces exist presently as they did in the past. As to barriers forming via spit accretion, Hoyt (1967) concurs that they, too, exist, but like emergent shoals would require backbarrier open marine sediment, and the limited examples of such barriers precludes spit growth as valid model of major Holocene barriers.

While Otvos (1970) agreed with Hoyt (1967, 1970) that conditions for barrier formation via shoreline submergence existed during the late Holocene he disputed the validity of Hoyt's first criterion (1967). Absence of open marine beach and shallow marine sediments in the lagoon can be explained by alternative conditions prior to barrier formation. Coasts with numerous estuaries and bays of sufficient discharge, such as those along the Mississippi mainland, are able to sustain low-salinity zones without the presence of offshore barriers to impede exchange of ocean waters (Otvos, 1970, 1981, 2005 a). Barriers fringing coast with abundant freshwater inflow, therefore, could have evolved from mechanisms other than shoreline ridge engulfment. Presence of open marine sediments in the lagoon, however, would indicate an emergent shoal mechanism only if that were original environment in which the islands formed (Otvos, 1977). Field and Duane (1976) similarly rejected Hoyt's first condition as exclusive evidence against an emergent shoal hypothesis, explaining that the sedimentary composition of the surrounding environment was relevant to barrier formation only if the barrier was still situated close to its present site. Islands have been observed to migrate significant distances during sea-level transgression indicating that they likely originated out on the shelf, seaward of their present environment (Otvos, 1977; Field and Duane, 1976, 1977; Swift, 1975; Wilkinson, 1975). To further contest the engulfment model, Otvos (1970, 1977) cites core data taken on Horn Island, MS showing the barrier sands resting over a 3 m thick layer of Holocene mud over the

Pleistocene erosional surface: if the barrier were the result of engulfment, it should sit directly on the Pleistocene surface (Otvos, 1970, 1977). Otvos (1970) interpreted the Galveston Island facies sequence as evidence to support the de Beaumont hypothesis of *in situ* bar aggradation as the mode of genesis of barrier islands (Otvos, 1970). The facies arrangement on Galveston described by Bernard (1962) showing offshore to shore sediments in vertical succession from bottom to top, indicated to Otvos (1970) a simultaneous aggradation of a shoal and progradation of the Holocene facies over the Pleistocene surface (Otvos, 1970). Hoyt (1970) and Hoyt and Henry (1967) attribute both of these cases to landward island migration moving the island from its formation site, over the original lagoonal facies, which is descriptive of modification, not formation processes (Hoyt, 1970; Hoyt and Henry, 1967).

Progressive understanding of the complex relationship between coastal geomorphology and the marine environment led to the dismissal of simplistic, single-mechanism island evolution models. A synergistic approach, acknowledging that several interacting mechanisms contributed to the development of barrier islands in any geographic setting, was advocated for future studies (Schwartz, 1971; Swift, 1975). Research along the Atlantic coast continental shelf revealed that the Holocene barriers originated as outer shelf shoals, and then migrated landward during transgression (Field and Duane, 1976, 1977; Swift, 1975). Swift (1975) describes the topography of the seafloor off the mid-Atlantic coast as exhibiting a series of relict, terraced deltaic sequences, with erosional scarps at their seaward margins and low gradient slopes landward of the terraces (Swift, 1975). Field and Duane (1976) discuss the widespread presence of shoals on the inner shelf surface and propose that relict shoals act as a source of sediment for landward migrating barriers as the barriers fused with inner shoals or ridges deposited by fluvial networks during earlier eustatic fluctuations (Field and Duane, 1976, 1977; Swift, 1975). Either late

Holocene still-stand or welding of the transgressing shoals with existing higher ridges facilitated stabilization, vertical accretion and progradation of the barrier shorelines. Given the extensive reworking and modifications that barriers have sustained during thousands of years of eustatic sea-level change, however, the sedimentary structure of their genetic sites and internal structure have been eradicated (Hoyt and Henry, 1967; Field and Duane, 1976, 1977).

Composite barrier island development hypotheses

Matagorda Island, located off the central Texas coast, fronting the mouth of the Colorado River and Matagorda Bay, also illustrates a complex history of island genesis and modification. Barrier island sands 25 ft deep were found overlying bay, not open marine, deposits in a series of beach borings (Shepard, 1960; Wilkinson, 1975). Core data from cross-sections laterally along the island strike and its perpendicular dip showed two distinct sandy lithosomes (Wilkinson, 1975). Back-island sands were identified as early Holocene, and the fore-island sands as late Holocene, and were present in nearly equal volume (Wilkinson, 1975). The back-island sediments comprised the original, constructive material and the modern island originated as a shoal on the lower shelf, constructed from Pleistocene strand plain and early Holocene fluvial-deltaic sediment. Rising sea level during the early Holocene transgression inundated Pleistocene deltas on the outer shelf, transforming deltaic environments into estuaries, and river-dominated, freshwater environments became brackish. Eroded muddy deltaic and alluvial sediment backfilled fluvial valleys as shoals and offshore bars associated with the distributaries migrated landward over the reworked deltaic facies (Wilkinson, 1975). These conjoined processes explain the absence of open marine sediments beneath the barrier sands described by Shepard (1960 a). Sediment from the re-worked seafloor renourished and maintained the shoals as they "rolled over" toward the receding mainland. At late Holocene stillstand, shoal migration ceased and the

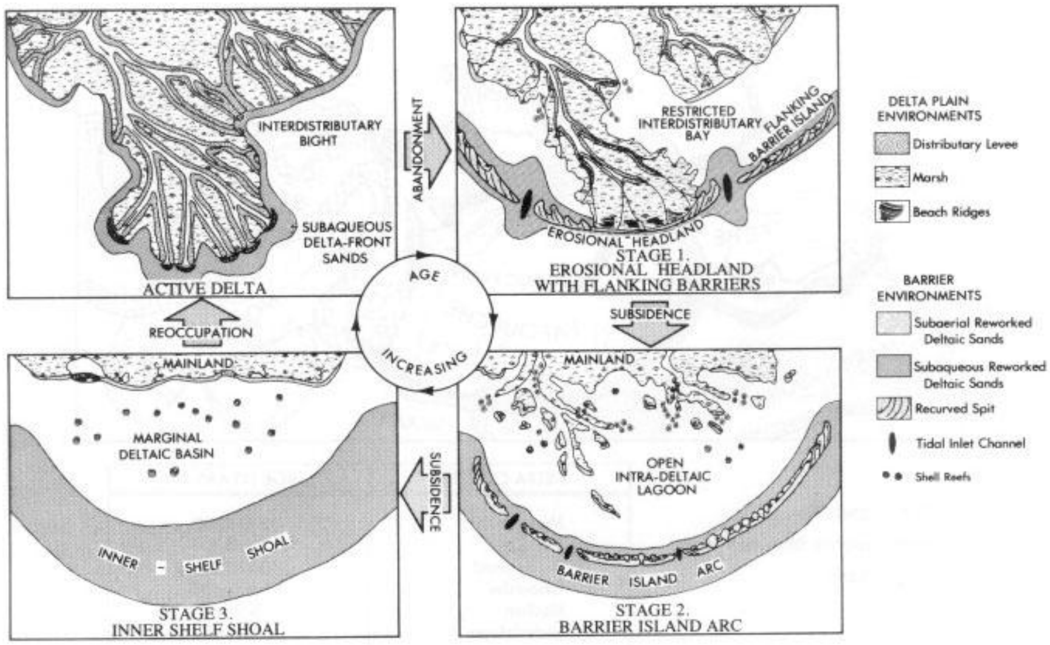


Figure 6. Three-Stage Transgressive Barrier Island model (modified Penland and Boyd, 1981)

stabilized shoal, supplied with material from the contemporary Colorado and Brazos Rivers via longshore currents, prograded seaward. These late Holocene alluvial sediments overly the original shoal base in the fore-island (Wilkinson, 1975), indicating Recent deposition over an early Holocene topographic high point was the mechanism of island construction.

The 3-stage transgressive barrier island model of Penland and Boyd (1981) (Penland and Boyd, 1981; figure 6) acknowledges the contributions of all three of the earlier models at various stages and levels of influence. They identify shoreline engulfment as the “dominant formative process for the transition of [the Cailou headland] from Stage 1 to Stage 2” of their model, where early transgression of the headland results in detachment of flanking barriers and bars from the mainland. Gilbert-type spit growth and de Beaumont-type bar aggradation are considered secondary contributors to the growth and maintenance of the transgressing islands (Penland and Boyd, 1981).

Beach Ridges

Beach ridges are accretionary features that form on moderate to low energy coasts. Beach ridges usually occur in sets of successive ridges, usually constructed over decades, forming a beach ridge strandplain (Otvos, 2000; Taylor and Stone, 1996; Tanner, 1995). Beach ridge strandplains exhibit a ridge and swale geomorphology characterized by sandy, vegetated ridges, often less than meter high, interspersed by swales that can be flooded, intertidal, or subaerial (Otvos, 2000; Taylor and Stone, 1996; Tanner, 1995). Ridge construction provides a mechanism for shoreline progradation and vertical aggradation of coasts and barriers (Hine, 1979; Otvos, 2000; Rodriguez and Meyer, 2006; Rodriguez et al., 2004; Tanner, 1995; Taylor and Stone, 1996). Beach ridge strandplains form where abundant fine to medium sandy sediment is available in the nearshore swash zone of a shallow, gently sloping shoreface (Otvos, 2000; Tanner, 1995). Straight-crested ridge sets are constructed perpendicular to dominant shore-normal wave approach, while the ridges formed by longshore transport and recurved spit development exhibit recurved ridge crests (Otvos, 2000).

Ridge formation can be initiated during neap tides, when a sea level is temporarily lowered nearshore berms are exposed (Hine, 1979; Otvos, 2000). The berms aggrade vertically as sediment eroded from the seaward slope of the berm is transported and deposited up the berm seaward slope face. Lateral aggradation of the berms occurs during higher tides as sediment is transported over the berm crest and deposited down the berm shoreward slope (Hine, 1979). Over several tidal cycles during extended fair weather periods, the berms aggrade to a permenantly subaerial height. Continued deposition of sediment landward of the berm crest infills the zone between the berm and the previous shoreline, establishing a new, seaward shoreline (Hine, 1979). As the ridge construction cycle repeats, successive ridges weld to the shoreface, ultimately removing the inland ridges from the nearshore environment. Inland ridges are essentially relict shorelines (2000). Offshore winds enable landward eolian sediment transport and deposition over the ridges, developing dunes deposits over the ridges (Otovs, 2000; Tanner, 1995). Colonization of ridges and dunes by vegetation enhances ridge stabiity and preservation (Otvos, 2000; Tanner, 1995).

Due to their low elevation and coastal locations, beach ridge strandplains are vulnerable to erosion from climatic stressors such as storm wave breaching, tidal erosion and sea level rise. In the northern Gulf of Mexico, however, beach ridge strand plains can be observed at multiple sites along the coast proximal to abundant sediment sources: the Morgan-Perdido strandplain complex that extends from the eastern headland of Perdido Bay, FL westward to the Morgan Peninsula that juts into Mobile Bay, AL from the eastern bay headland (Otvos and Giardino, 2004; Otvos, 2005 a; Rodriguez and Meyer, 2006); relict beach ridge sets on Point Clear and Campbell Islands on the eastern margin of the Pearl River, MS estuary (Otvos and Giardino, 2004; Otvos, 2005); and beach ridges on the Bolivar Peninsula and Galveston Island that flank Galveston Bay, TX (Morton et al., 2000; Rodriguez et al., 2004) are a few examples. As individual beach ridges form at the land-water interface in response to small-scale changes in sea level, their original elevations are approximate analogues of sea level. Sea level rise histories derived from beach ridge set elevations (Morton et al., 2000; Tanner, 1992) have disputed, as they do not acount for the eolian deposition that occurs on maturing ridges (Lopez and Rink, 2008; Otvos, 2000; Rodriguez and Meyer, 2006). Elevations and dates of the interface between ridge and dune lithofacies do record the progression of sea level fluctuations (Lopez and Rink, 2008; Rodriguez and Meyer, 2004). Rodriguez and Meyer (2006) used profiles of the Edith Hammock and Little Point Clear beach ridge strand plains on Morgan Peninsula, AL aquired with ground penetrating radar (GPR) to map the interface between the ridge and dune lithofacies (Rodriguez and Meyer, 2006). An overall seaward-increasing elevation of the beach ridge-dune interface validate a continuous sea level rise to present (Rodriguez and Meyer, 2006).

The orientation of beach ridges to each other and to the shoreline is principally determined by sediment supply and dominant wave approach (Lopez and Rink, 2008; Otvos, 2000; Rodriguez and Meyer, 2006; Rodriguez et al., 2004; Tanner, 1995; Taylor and Stone, 1996). Beach ridge orientation has been used to infer the direction of historical shifts in dominant incident wave approach intervals (Lopez and Rink, 2008; Otvos, 2000; Rodriguez and Meyer, 2006; Rodriguez et al., 2004; Tanner, 1995; Taylor and Stone, 1996). Abundant sediment supply is crucial to the development of beach ridge sets, and must be able to sustain the multiple cycles of ridge formation over multi-decadal intervals (Otvos, 2000; Tanner, 1995; Taylor and Stone, 1996). The site of ridge construction is therefore constrained by sediment availability. Timmons et al. (2008) observed paleochannels in seismic data acquired along Bogue Banks and Pine Knoll barrier shorelines of the Outer Banks, NC at sites corresponding to progradational beach ridge sets along the otherwise narrow, elongate barrier (Timmons et al., 2008).

Present-day Regional Background and Conditions

The Mississippi Sound is a shallow body of water along the Northern Gulf of Mexico coast from Mobile Bay, AL to Waveland, MS from east to west, and enclosed by the barrier islands that parallel the mainland 9-19 km offshore. The Sound encompasses an area approximately 2100 km^2 with average depths ranging from 2-5 m, and up to 25 m in the passes and channels. The MS-AL barrier island chain is currently comprised of Dauphin, Petit Bois, Horn, East and West Ship and Cat Islands, from east to west. The barrier trend is fragmented by the tidal inlets that run from the mainland estuaries Gulfward between the islands (from east to west): the Mobile Ship Channel on the eastern side of Dauphin Island, Petit Bois Pass, Horn Island Pass, Dog Keys Pass, and the convergence of Pass Marrianne and Cat Island Pass west of Cat Island. The Mobile, Pascagoula and Ship Island navigation channels are routinely dredged to maintain depths required for navigation channels (figure 1).

The Mississippi Sound is currently the receiving basin of several major fluvial systems: the Mobile, Pascagoula, Biloxi, Wolf, Jourdan and Pearl Rivers from east to west, as well as smaller tributaries that drain the coastal plain (Boone, 1973; Curray and Moore, 1963; Isphording et al., 1989). Mobile Bay has the fourth largest average annual discharge in the United States at 79,300 cfs delivering 6.35 million tons of sediment/year (Isphording et al., 1989). The Pascagoula and Pearl Rivers combined discharge 43,600 cfs and 4.58 million tons/year (Isphording et al., 1989). The infusion of fresh water results in lower average salinities in the Sound (0 to 30 parts per thousand) than in the open Gulf (30 to 40 parts per thousand) that support the estuarine biota found there (Boone, 1973). A diurnal, microtidal range of up to 0.6 m (Isphording et al., 1989) transports water from the Gulf through the tidal passes between the islands, resulting in locally higher salinities (Boone, 1973). The major rivers have been active throughout the Quaternary, delivering weathered sediment of Appalachian provenance, abundant in heavy minerals (Hsu, 1960; Isphording, 1989; Van Andel, 1960).

The transitory nature of this coastal environment is underscored by evidence of former and emerging barrier features. Dog Island was a small islet, part of a string of shoals, Dog Keys, located to the west of Horn Island in Dog Keys Pass that periodically aggraded and dissipated (Otvos and Carter, 2008). Dog Island had been the site of a military reservation since 1847 (Otvos and Carter, 2008) until a hurricane in 1852 apparently reduced it to a subaqueous shoal, as shown in U.S. Coast Survey chart H-489, published in 1854 (Otvos and Carter, 2008; Rucker and Snowden, 1988). An islet re-emerged between 1855 and 1877 at that site (U.S. Coast and Geodetic Survey chart T-4021, 1917), perhaps assisted by the westward migration and coalescing of smaller shoals directly to the east over the submerged remnant of Dog Island (Otvos and Carter, 2008). The island was re-named the Isle of Caprice and reached up to 2.5 km in length by 1926 when it was breached by a hurricane (Otvos and Carter, 2008). Segmentation of the island accelerated wave-induced erosion, until its total disappearance in 1939 (Otvos and Carter, 2008). In the 1960's, along the western margin of Horn Island Pass between Petit Bois and Horn Islands, a small man-made island emerged as dredge-spoil deposits accelerating construction of an intertidal shoal. By 1971, an island ~400 m in length had formed (Otvos and Carter, 2008). Continued dredge-spoil deposition has been reworked by tidal and westward littoral currents, constructing recurved spits and ridges, welded on to the original shoal increasing the island's length to ~1.2 km in 2005 (Otvos and Carter, 2008).

Holocene Evolution of Mississippi Sound

Central and Eastern Mississippi Sound

Several hypotheses for the MS-AL Barrier Islands have been proposed. Shepard (1960 a) suggested, in the absence of substantial headlands to provide sand through erosion, the islands are the product of both local and longshore-transported sediment from the east. Shepard (1960 a) thought it unlikely for sediment from the major regional fluvial source, the Mobile River to the east, to be carried in sufficient quantities across the shelf to construct the barrier chain, proposing that offshore shoals formed in situ during shoreline transgression, aggrading into a barrier platform (Shepard, 1960 a). With upward bar growth and rising sea level, however, shelf sediment was submerged below the constructive breaker zone and was no longer available to nourish the barrier system. Fluvially derived sediment transported by longshore currents was

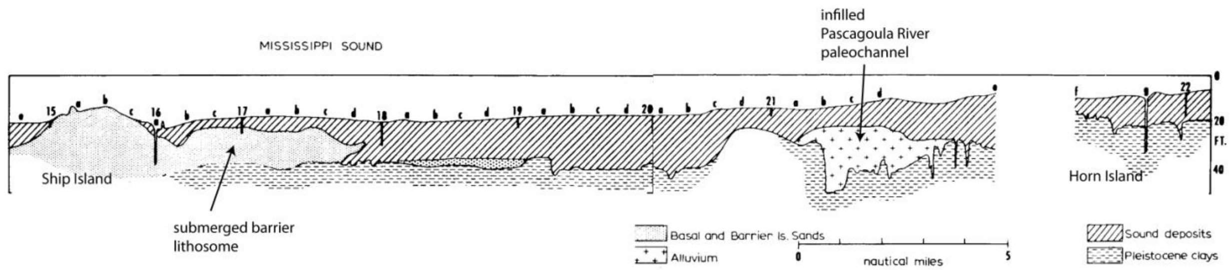


Figure 7. Geologic cross-section of Mississippi Sound showing interpreted lithofacies from acoustic reflectance survey (Curray and Moore, 1963)

necessary to assist in island growth and maintenance (Shepard, 1960 a; Kwon, 1969). Based on a minerologic analysis of barrier beach sands, Foxworth (1962) suggested subaqueous bars sands developed from the eroded and re-worked underlying Pleistocene deposits (Foxworth, 1962). These Pleistocene sediments originated as deltaic deposits accumulated during the glacial lowstand, delivered by the fluvial systems of Mississippi and Alabama (Foxworth, 1962). Storms then provided the necessary energy and temporary rise in sea level to elevate the bars above sea level. Island growth proceeded under fair weather conditions with longshore currents as the predominant source of sediment and geomorphic modification (Foxworth, 1962).

Based on interpretations of seismic and core data from the northern Gulf, Curray and Moore (1963) identified a weathered lowstand horizon, the Pleistocene surface, approximately 6-12 m bsl from east to west, underlying the Sound facies (Curray and Moore, 1963; figure 7), incised by paleochannels associated with relict fluvial networks of the modern Pearl, Wolf, Biloxi and Pascagoula Rivers traversing the inner Mississippi-Alabama continental shelf (Curray and Moore, 1963). The overlying Sound facies were determined to have originated from alluvium deposited by the local rivers (Curray and Moore, 1963). The rapid early to mid Holocene transgression precluded delta formation and landward-shifting depocenters first deposited sediment in the accommodation space provided by the paleochannels, and as these valleys filled, alluvium was distributed across the inner shelf. Winnowing of fines from the alluvium by the landward migrating wave breaker zone resulted in a coarser-grained basal sand layer that aggraded into offshore shoals, forming the subaqueous barrier platform (Curray and Moore, 1963). Continual sediment delivery from the Pearl, Pascagoula and Wolf river systems during the late Holocene in addition to the re-worked alluvial deposits provided the material for barrier island aggradation by shore-normal constructive waves. Seismic transects between Cat and Ship Islands and Ship and Horn Islands showed a sandy ridge ~9 - ~12 m bsl thick on strike with the islands (Curray and Moore, 1963; figure 7), suggesting a continuous sandy lithosome, possibly an earlier-stage barrier, underlying the length of the barrier chain, grading laterally into the basal, or barrier platform sands of the modern islands. The base of the sands superimposes a unit of marine clays over Pleistocene clays. Correlating the depth of the layer with the Curray (1960) sea level curve, Curray and Moore (1963) proposed that the underlying, then-continuous ridge began forming ~7000 ybp. Vertical and seaward growth via a "depositional regression" was facilitated by abundant underlying sediment and a slowing rate of sea level rise. Breaching by inlets subsequently developed widening channels segmenting the ridge resulting in the contemporary chain of islands (Curray and Moore, 1963).

Otvos and Giardino (2004) proposed an evolutionary model for the Mississippi Sound that incorporates ridge engulfment in the eastern Sound and an emergent bar model in the central Sound. In the central Mississippi Sound the underlying barrier island platform was constructed by nearshore sediment aggradation over muddy-sandy Holocene nearshore deposits (Otvos, 1981, 1985 a, 1985 b, 2005 a; Otvos and Giardino, 2004). This interpretation was supported by core interpretations showing increased sorting and decreased silt and clay concentrations upward in the 3-12 m thick muddy, brackish Holocene layer (Otvos, 1981, 1985 a, 1985 b, 2005 a) with open nearshore, inner shelf fauna assemblages present throughout the unit (Otvos, 1985 a, 1985 b). Over the muddy layer lies the barrier platform, composed of 7-12 m of poorly- to moderately-sorted sandy-to-muddy deposits with few fauna (Otvos, 1985 a, 1985 b). Interspersed lenses of moderate to well-sorted sands in this unit were interpreted as transient intertidal shoals (Otvos, 1981). The stabilizing barrier platform began to enclose the Sound, protecting the backbarrier environment from the high-energy waves of the open Gulf. Transition to a restricted bay hydrodynamic regime allowed fine-grained sediment to settle and accumulate as lagoonal-type facies. A low-salinity nearshore marine environment had been maintained by high levels of freshwater influx from estuaries along the mainland prior to the enclosing of the Sound and was reinforced by the newly restricted exchange with the open Gulf waters, explaining the absence of open marine fauna (Otvos, 1985 b, 2005 a; Otvos and Giardino, 2004).

In the eastern Sound, Dauphin Island illustrates the control of the original geologic structure over geomorphic evolution, exhibiting the stratigraphic sequence of a composite barrier island (Otvos, 1979, 1985 a, 2005 a). The intervening Holocene mud unit is absent, and the underlying Pleistocene deposits form the base of the island's core and are exposed as the subaerial island surface approximately 2 m above sea level (Otvos, 1979, 1985 a, 2005 a). The sandy barrier platform is presented on the Gulf side of the island onlapping the Pleistocene core and forming a foundation for the intertidal shoreface deposits surface. On the landward side, lagoonal deposits pinch out on to the Pleistocene base (Otvos, 1981, 1985 a, 2005 a) (figure 8). Material for Dauphin Island's barrier platform originates at the eroding Morgan Peninsula that extends from the eastern margin of the Mobile Bay headland toward Dauphin Island, and the Mobile Bay ebb tidal delta (Otvos, 1981, 2005 a). Locally reduced tidal currents enable settling of entrained sediment that coalesces around the semi-consolidated core. With sustained, abundant sediment supply and diminishing rate of sea level rise during the late Holocene, the

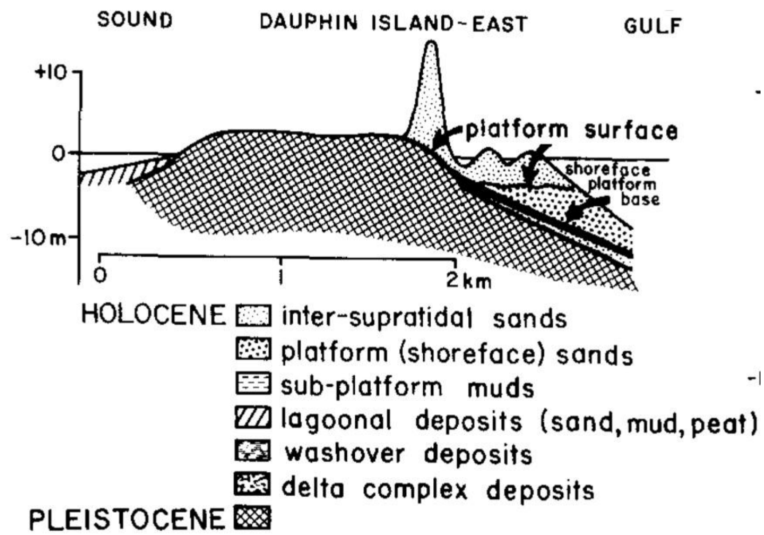


Figure 8. Interpreted geologic cross-section, north to south, through Dauphin Island, AL (Otvos, 1985 a)

island prograded seaward through the development of a beach ridge strandplain and westward, as longshore currents transported sand from the western tip of the island to the eastern terminus forming a long, thin subtidal barrier platform (Morton, 2008; Otvos, 2005 a). The prevailing east-to-west longshore currents "had a primary influence on the vertical aggradation of the platform" (Otvos, 1981, 1985 a, 1985 b, 2005 a) and presumably carried sediment across the length of the Mississippi Sound, providing sediment to the downdrift aggrading shoals and emerging barriers (Otvos, 1981, 1985 a, 2005 a). Beach ridge strand plains on Dauphin and Cat Islands and relict ridges on Horn and Petit Bois Islands attest to a substantial volume of sediment delivered to the Mississippi Sound barriers during the late Holocene (Otvos, 1985 a; Otvos and Giardino, 2004; Rucker and Snowden, 1989, 1990).

Western Mississippi Sound

Discovery of a series of sandy barrier island deposits, the Pine Island Barrier trend, buried under deltaic sediments along the southern shoreline of Lake Pontchartrain led to a hypothesis of Cat Island's formation (Flocks, 2009; Otvos, 1978, 2005 a; Otvos and Giardino, 2004). The Pine Island barrier trend was first identified and mapped by Treadwell (1955), who described them as a band of “cheniers” (Treadwell, 1955). Cores taken from the New Orleans barrier showed a quartzose sandy lithosome, 10-11 m thick, resting entirely within prodelta silts and clays, protruding slightly into an overlying peaty marsh layer (Corbeille, 1962; Saucier, 1963; Treadwell, 1955). The Pleistocene surface in adjacent drillcore sites was found at ~13.5 m and ~14.4 m below the land surface (Treadwell, 1955). The barrier was interpreted to be a relict former shoreline, based on the high degree of sorting and moderate grain size, and of similar mineralogy to the Eastern Gulf Province (Corbielle, 1962; Treadwell, 1955; Saucier, 1963). The barriers were estimated to have formed contemporaneously with the Mississippi River Margouin and Teche delta phases ~7000 - ~4000 ybp (Corbielle, 1962). Further investigation constrained the initiation of the Pine Island trend to between ~5300 - ~4400 ybp (Otvos, 1978). Although the Pine Island and Cat Island barriers likely formed at the same time, the buried Pine Island lithosomes gave older dates as the sediments had been preserved from reworking and uncontaminated by later sedimentation (Otvos and Giardino, 2004). Otvos (1978) also surmised that deposition occurred in a subtidal environment, as the coarse grain size and lack of subaerial dune morphology and internal cross-bedded strata are indicative of offshore shoals rather than stabilized barrier islands (Otvos, 1978).

At ~7000 ybp, the Mississippi River began prograding westward out over the newly-submerged Pleistocene surface, while the Pontchartrain Basin remained an open bay maintaining a high-salinity environment that supported oysters and other saline-tolerant fauna (Saucier, 1963; Otvos, 1978). As sea level rise slowed and shorelines stabilized, at around ~5000 - ~4600 ybp, transgressive offshore shoals began to aggrade, supplied with sediment re-worked and transported from the inner shelf as well as by fluvial sediment carried by the rivers emptying into the Gulf (Frazier, 1967; Otvos, 1978; Saucier, 1963; Treadwell, 1955). Sea level rise in the

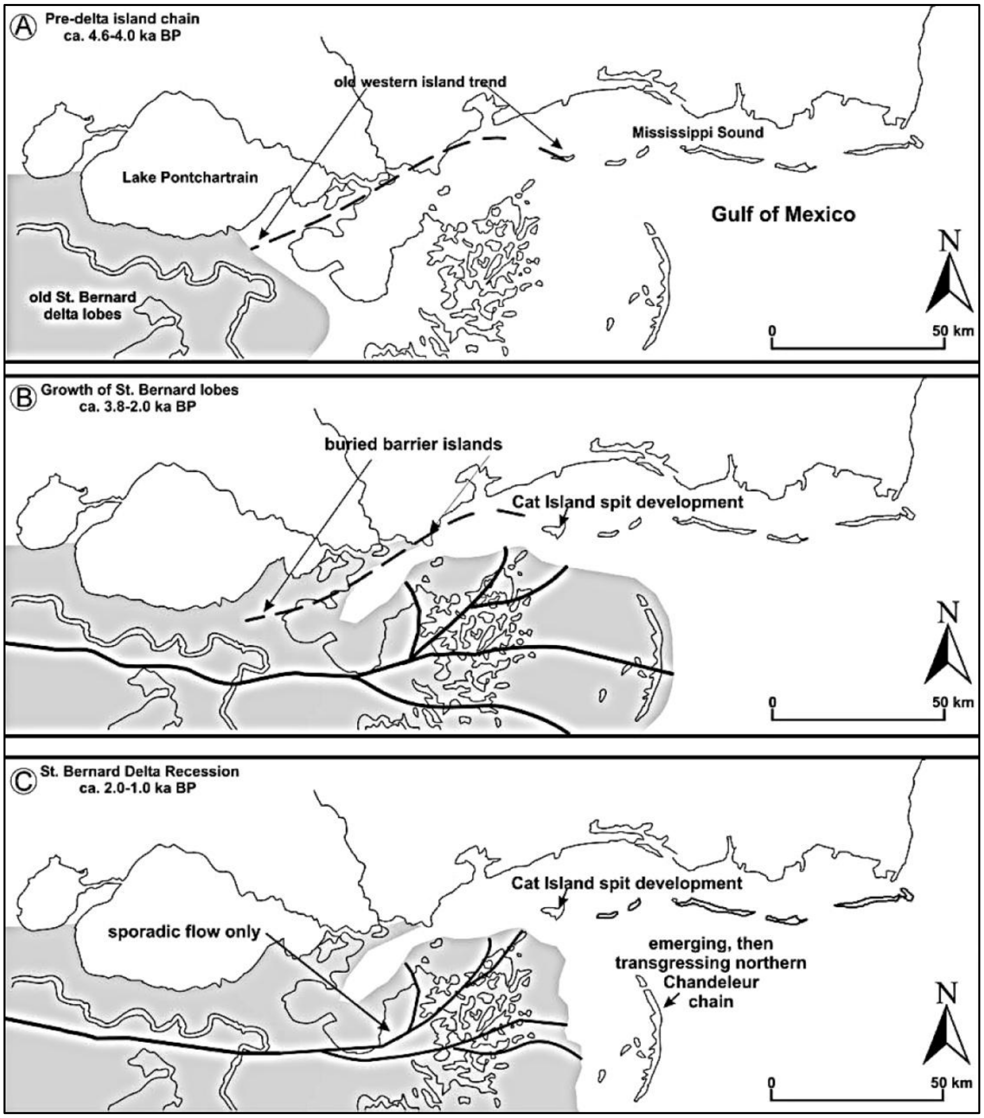


Figure 9. Conceptual model for barrier island chain and associated delta plain development (Otvos and Giardino, 2004)

vicinity of the Pontchartrain Basin may have proceeded more slowly than along the adjacent coast, as the faulting in that area created a zone where the slope was steeper, favoring shoal aggradation into stable barrier islands (Saucier, 1963). A hypothesized brief fluctuation in sea level at ~4100 ybp may have allowed for barrier stabilization as sea level fell, then for continued onshore transport of sediment as transgression resumed (Stapor and Stone, 2004). Prevailing east-west longshore currents in the northern Gulf provided delivered sediment and helped shaped the elongate morphology of the emerging barriers (Otvos, 1978; Otvos and Giardino, 2004).

As the St. Bernard delta lobe prograded eastward, deltaic plain and pro-deltaic marshland were constructed in the formerly open waters of the eastern Mississippi Sound, halting and re-directing the westward flowing longshore current. Otvos and Giardino's conceptual model of barrier island chain and associated delta plain development shows the St. Bernard delta lobe reaching its maximum extent at ~2000 ybp (Otvos and Giardino, 2004; figure 9). Its fringing marshes stranded the Hancock County barriers along the mainland, and its easternmost shoreline reached beyond the present location of the Chandeleur Islands, southeast of Cat Island (Otvos, 2005 a; Otvos and Giardano, 2004). As the eastern shoreline of Louisiana prograded, westward flow of the northern Gulf and Mississippi Sound was obstructed and longshore currents decelerated as they approached Cat Island. Prodelta deposition shallowed the waters around Cat Island, transforming the hydrodynamic regime from open marine to low-energy nearshore and re-directing the dominant wave approach to the island's southeast shoreline (Rucker and Snowden, 1989; 1990). At this stage, Cat Island retained its original elongate, mainland-parallel configuration (Otvos and Giardino, 2004; Otvos, 2005 a). Analysis of sediment retrieved from a vibracore south of the present-day N-S spit (core BAR03, Barnhardt, 2003; figure 10) confirms the extension of the St. Bernard delta complex to this position (Barnhardt, 2003). The upper 0.45 m of the core contained quartz sand, moderately to well sorted and fining with depth, and was interpreted as sediment eroded and transported from the original eastern island terminus (Barnhardt, 2003). The lower core, from ~0.45 - ~0.75 m bsf was composed of clays and sands, with some sandy lenses, and below ~0.76 m to the base, at ~2.13 m bsf, consisted of a stiff, gray clay and was interpreted as St. Bernard delta facies deposited during the delta complex's active phase, grading upward into intermingled St. Bernard and barrier sand facies (Barnhardt, 2003).

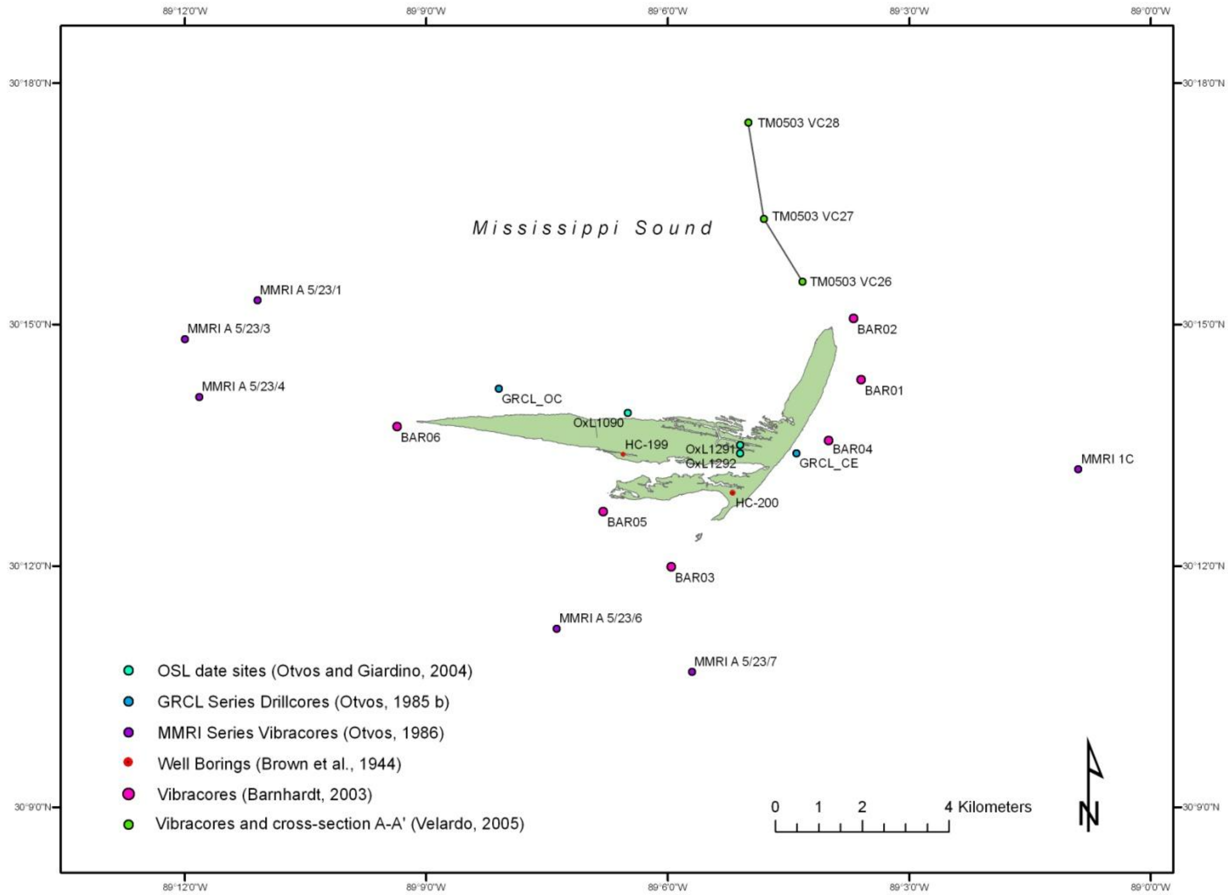


Figure 10. Locations of vibracores (Barnhardt, 2003; Otvos, 1986; Velardo, 2005) and well borings (Brown et al., 1944; Otvos, 1985 b) acquired during previous studies on and around Cat Island, MS.

By ~2900 to ~2600 ybp, the incipient barriers south of Lake Pontchartrain were completely stranded and buried (Otvos and Giardino, 2004). The Hancock County, MS barriers further to the west (Pine Island, Point Clear and Campbell Island) outcropped a few feet above the surrounding marsh surface (Treadwell, 1955; Saucier, 1963; Otvos, 1978; Otvos and Giardino, 2004). The St. Bernard delta was abandoned at ~1800 ybp and entered its destructive phase at ~1000 ybp (Frazier, 1967; Penland et al, 1989) as the Mississippi River changed course. Sediment delivery from the Mississippi River was drastically reduced, and transgressing Gulf waters eroded the subsiding delta, re-opening the pass between Cat Island and the Louisiana coast. Wave approach was then confined to a 25˚ window southeast of the island, focusing erosion on its eastern tip (Rucker and Snowden, 1989, 1990).

As the large volumes of material sequestered in Cat Island's extensive beach ridge plain eroded, the sediment was immediately re-deposited in the shallowing waters over the island platform, developing the north-south spit. The upper sandy unit of core BAR03 (Barnhardt, 2003; figure 10) originated from the eroded eastern end of the island as the St. Bernard Delta lobe receded and the area reverted to a high-energy nearshore regime (Barnhardt, 2003). During this transition from a lower to higher energy environment, the original deltaic and barrier sands were re-worked and re-deposited as the mixed sedimentary unit superimposing the deltaic deposits in lower core BAR03 (Barnhardt, 2003; figure 10). The coarsening upwards of the barrier sands in upper core 03 also indicates a transition to a higher energy environment (Barnhardt, 2003). A series of three vibracores taken in the Mississippi Sound northwest of the N-S spit (cores TM-VC28, TM-VC27, TM-VC26, Velardo, 2001; figure 10) revealed a mud-over-sand to sand-over-mud progression toward the island and is interpreted as a depositional sequence initiated by the transition from an open-water, high-energy environment to the present-day reduced energy environment on the lee side of Cat Island's N-S spit. The basal sandy units in cores TM-VC28 and TM-VC27 were deposited prior to disruption of the high-energy open marine environment to the west and the longshore currents in the Mississippi Sound that prevented fine grains from settling. As the spit prograded into the Sound, low energy waters behind the spit allowed fine sediment deposition, forming the muddy unit over the sand. The upper sandy facies present at core TM-VC26 likely indicate the distal edge of the N-S spit platform prograding over the muddy unit. The accumulation rate of 0.2 cm/year calculated for TM-VC28 was used to estimate initiation of this "shielding effect" between ~750 ybp and ~1000

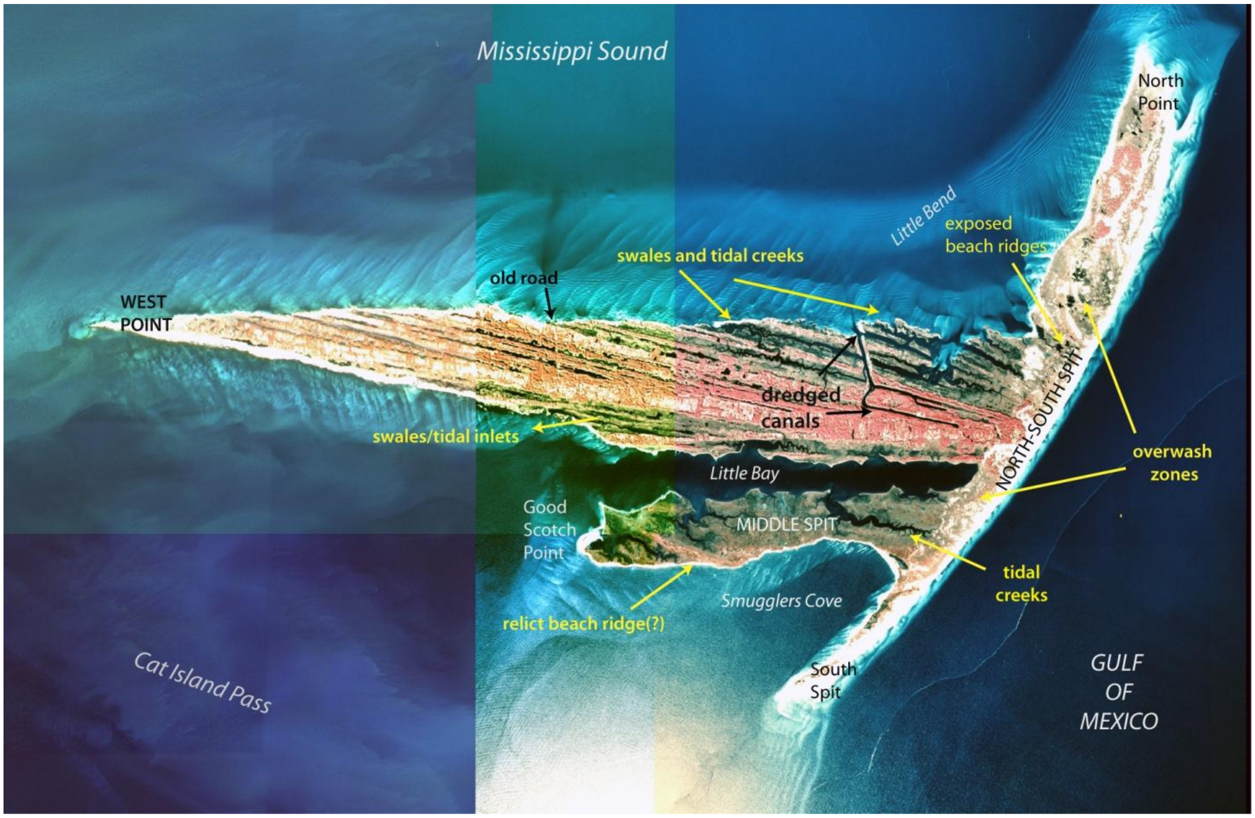


Figure 11. USGS aerial photography of Cat Island, MS acquired on June 29, 2007 showing present-day geomorphology (imagery courtesy of USGS Coastal and Marine Geology Program's Decision Support for Coastal Science and Resource Management Project, http://coastal.er.usgs.gov/remote-sensing).

ybp and approximately 500 ybp-700 ybp after the abandonment of the St. Bernard Delta, based on the conceptual model proposed by Otvos and Giardino (2004) (Velardo, 2005).

Present-day Geomorphology of Cat Island

Cat Island, MS is located at the western perimeter of the Mississippi Sound approximately 12 km south of the Mississippi mainland. The westernmost marsh fringe of Louisiana's relict St. Bernard delta complex lies roughly 10 km to the east, across the Chandeleur Sound (figure 1). The island consists of two east-west trending, vegetated segments separated by a narrow lagoon. The larger segment is the main remnant of the original island, currently measuring 7 km from the western tip (West Point) to the eastern shoreface, and 1 km across at its widest point tapering to the west (figure 11). A north-south trending sandy spit 5 km

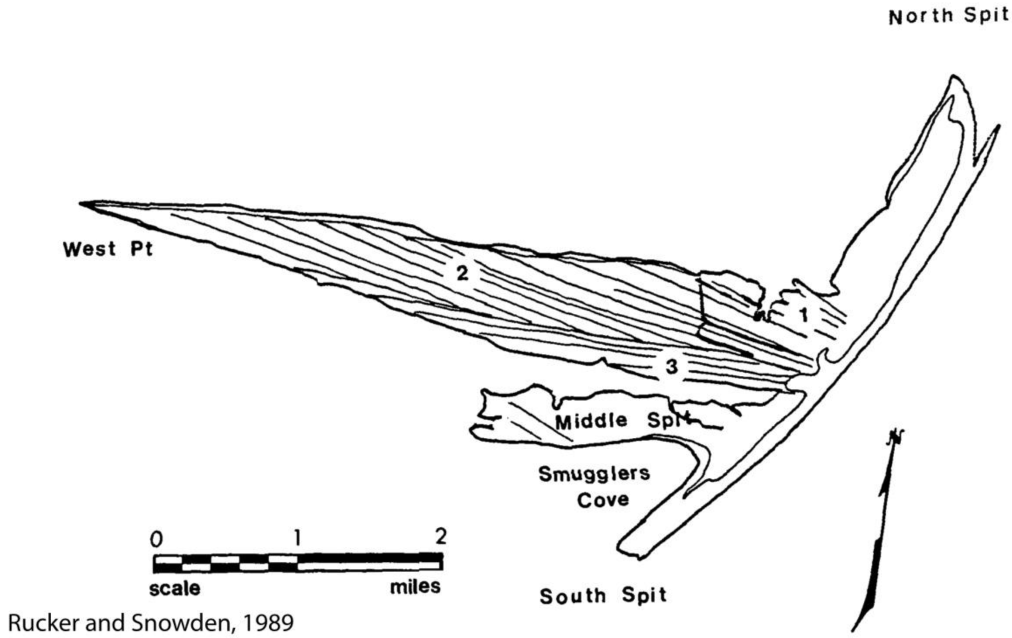


Figure 12. Beach ridge sets on Cat Island from oldest (1) to youngest (3) (Rucker and Snowden, 1989)

long (North Point to South Spit) abuts the main island at its eastern end and is oriented at an ~ 60 degree angle to the strike of the island. The second east-west trending segment (Middle Spit) extends west from the north-south spit approximately 2 km and roughly parallel to the main island's shoreline. It is separated from the main island by a shallow bay (0.5-2 m) approximately 300 m wide (Little Bay). Early ecological studies of Cat Island describe a dynamic system undergoing constant change. On the east-west body of the island, "between two to sixteen sand ridges from four to ten feet in height" separated by swales either intermittently or continuously submerged under up to 6 feet of water were identified by Penfound and O'Neill (1934). The ridges extended to the shoreline where they were either truncated or fronted only by a low berm and narrow sandy beach (Penfound and O'Neill, 1934; Treadwell, 1955). The beach ridge strand plain was later grouped into three to four distinct ridge sets, delineated by shifting angels of orientation (Rucker and Snowden, 1989; figure 12). Protected swales of the interior were populated with freshwater marsh vegetation. The ridges on the main island supported forests of slash pine and oak. The north-south spit, largely devoid of vegetation, consisted of a broad sandy beach transitioning into dunes (Penfound and O'Neill, 1934) and appeared then to be the only area where the island was actively aggrading. Initiation of the spit occurred as the island's eastern ridges were eroded by dominant longshore currents and redistributed westward. At the intersection of the spit and island, the ridges provided a supportive base for dune and beach stabilization, allowing dunes to grow as high as 9 m (Treadwell, 1955). Here, on the lee side of the spit, marsh vegetation along the margins of the forests was actively being buried under sand transported from the dunes (Penfound and O'Neill, 1934).

The southernmost two miles of the north-south spit is a low-relief stretch of beach subject to frequent overwash during storms, preventing dune-stabilizing vegetation growth. Tree stumps extending eastward into Ship Island Channel and the presence of organic marsh soil buried under the lower beach (Penfound and O'Neill, 1934) indicated that the spit was undergoing erosion as the island migrated westward (Penfound and O'Neill, 1934; Treadwell, 1955). The dunes at the southern stretch showed frequent occurrence of eolian blowouts (Penfound and O'Neill, 1934). Both studies concluded that the island was migrating westward (Penfound and O'Neill, 1934; Treadwell, 1955).

Middle Spit is composed of a few low ridges occupied by salt marsh vegetation, also found on the southern shore of the main island and in the swales of the ridges most exposed to open water. A crested ridge (figure 11) cuts across Middle Spit from the southern shoreline, sloping into the Little Bay marsh. The ridge is oriented at a steeper angle relative to the orientation of the ridges of the main island, and may be either the lone remnant of a former beach ridge set. Alternatively, Middle Spit may represent a stalled bar complex that failed to

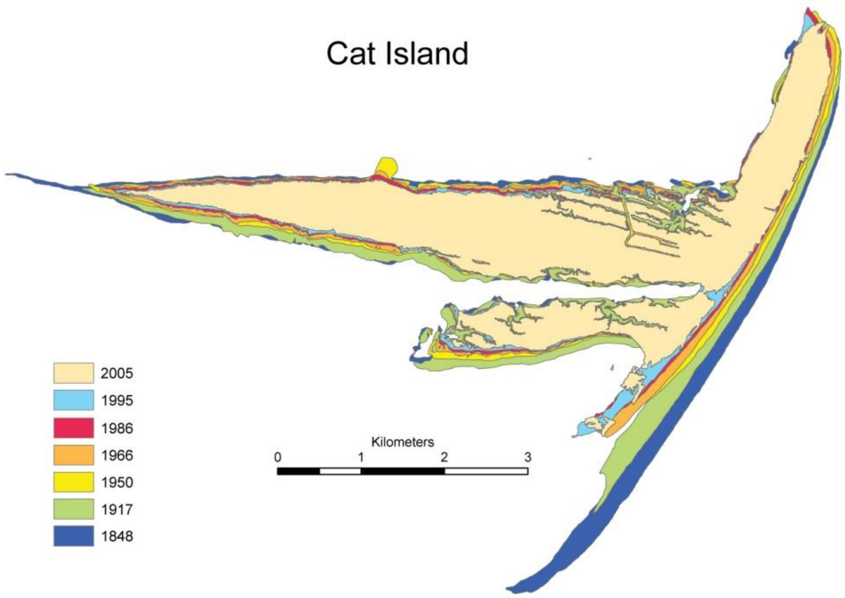


Figure 13. Time series of shoreline change on Cat Island, MS from 1848 to 2005 (Morton, 2008).

weld to the main island; the ridge would have formed as a result of lateral spit accretion along a laterally migrating shoreline (personal communication, Mike Miner).

Evidence of Cat Island's deterioration can be found in its receding shorelines (McBride, et al., 1995; Morton, 2008). The island lost 39% of its area between 1848 and 2005 (figure 13). Recent aerial photography (figure 11) shows tidal creeks cutting into the swales between the vegetated beach ridges of the main island and the marsh of Middle Spit. Continued rising sea level and temporary, local storm-driven elevated water levels will likely widen these creeks by increasing the tidal prism and thus, the velocity of erosive flood and ebb currents. As the channels expand and connect with flooded swales, wedges of the island may become isolated and completely detached from the mainland, which may have been the case with Middle Spit. Morton and Rodgers (2010) have recently classified the entire north-south axis of the spit above the juncture with the main island as an inactive overwash zone that is periodically inundated by extreme storm waves and surges, but not by seasonal or above average high water events (Morton and Rodgers, 2010). Within this zone, a ~500 m long section of the north-south spit immediately above the juncture with the main island (figure 11), is remarkably more degraded than the northernmost spit. Although the Gulf shoreline of the north-south axis is continuous along this degraded zone, the backbarrier has a much greater area of marsh and low-lying ridges at the expense of dunes and a backing berm and beach. Here, beach ridge and swale topography with intervening tidal creeks trend laterally across the north-south axis from the backbarrier to the Gulf shoreline are exposed as a result of dune erosion by Hurricane Katrina storm surge and waves in 2005 (Otvos and Carter, 2008). South of the juncture is an active overwash zone that is frequently flooded and is vulnerable to being permanently breached, detaching Middle Spit from the main island (Morton and Rodgers, 2010). On the main island segment, a canal has been dredged from the northern approach at Little Bend, behind the north-south spit, into the mid-interior where it meets with two east-west canals that provide access to private residences. A road located approximately mid-way along the main island from the northern shoreline to the interior is visible in the aerial photograph (figure 11).

Island Erosion Trends

Translational westward migration of the Mississippi barrier islands in response to the predominant westward littoral currents of the northern Gulf of Mexico has been well established in the literature (Penfound and O'Neill, 1934; Treadwell, 1955; Foxworth, 1962; Curray and Moore, 1963; Morton, 2008; McBride, et al., 1995; Otvos, 1970, 1980, 2006; Otvos and Giardino, 2004; Otvos and Carter, 2008; Cipriani and Stone, 2001; Tanner, 1990). Gulf barrier island degradation and modification by seasonal cold fronts and hurricanes have also been thoroughly discussed (McBride, et al., 1995; Morton, 2008, 2010; Schmidt, 2003). Hurricanes and seasonal storms are episodic and intense whereas fair weather waves, winds and tides are constant but usually produce gradual geomorphic change. Subsurface geology and changes in relative sea level and sediment supply also effect barrier modification (McBride, et al., 1995).

Island geomorphology is dependent on the configuration of the shoreface, which is constantly modified by hydrodynamics and climate (Swift, 1975). Long-term shoreline change is a manifestation of individual, though sometimes interdependent, geomorphic responses of a barrier to the physical processes acting upon it. McBride, et al. (1995) developed eight geomorphic response classifications of barrier coastlines based on shoreline change comparisons from historical maps, aerial photographs and GPS readings in the field acquired from 1847 to 1991 and included the Mississippi and Louisiana barrier islands and Cumberland (GA) and Amelia (FL) Islands on the Atlantic coast. The focus of the study conducted by McBride et al. (1995) was to determine mega-scale shoreline changes i.e. that take place on a large temporal and spatial scale (years to decades and tens to hundreds of kilometers). Within the Mississippi barrier island chain, Dauphin, Horn and Petit Bois Islands showed substantial westward migration in response to dominant, shore-parallel transport resulting in updrift erosion and downdrift lateral accretion, and recurved spit formation (McBride et al., 1995). Dauphin Island showed an average accretion rate on its western margin of 55.3 m/yr from 1848 to 1986 (McBride et al., 1995). Petit Bois lost 89.9 m/yr on its eastern flank and gained 31.3 m/yr on its western flank while retreating landward 3.3m/yr (McBride et al., 1995). Horn Island lost 39.3 m/yr from the eastern flank, grew 34.5 m/yr on the western flank with no landward retreat (McBride et al., 1995). East and West Ship Islands, collectively, displayed rotational instability, a complex class that describes the net effects of updrift erosion and downdrift advancement around a stable center point (McBride et al., 1995). Both islands migrated slightly westward (McBride et al., 1995). East Ship also retreated landward at a rate of 4.5 m/yr, while West Ship advanced seaward at 0.7 m/yr with a net effect of island rotation (McBride et al., 1995). Cat Island was classified as being in retreat, describing shoreline retrogradation due to lack of an adequate sediment supply or rising sea level (McBride, et al, 1995). Erosion occurred almost exclusively along the north-south spit while the island remained in place. The entire length of the spit shoreface, exposed to the westerly current, retreated at an average rate of 12.4 m/yr during the 144-year study period (McBride et al., 1995).

Using similar methodology, Morton (2008) expanded on this work by incorporating the island's shoreline change from 1986 to 2005. In contrast to McBride, et al. (1995), Morton (2008) investigated rates of change at various short-term intervals during the overall time span in order to isolate the dominating influence of morphodynamic processes. Long-term trends affecting the barrier chain include: 1) island narrowing caused by high-energy wave attack from both the Gulf and Sound; 2) unequal lateral migration as net updrift erosion is greater than downdrift accretion (the combined effects of westerly longshore currents and reduced sediment supply); 3) on Dauphin and Ship Islands, island segmentation via multiple breaching by storms; and 4) relative sea level rise (Morton 2008; 2010). Morton (2008; 2010) suggests that, although the region is affected frequently by extra-tropical storms, a significant increase in erosion rates observed between 1955 and 2005 may be the result of an increase in the frequency of large-magnitude tropical cyclones (Category 3 or higher) during this period (Morton, 2008; 2010). The Mississippi barrier islands have been in the paths of six major hurricanes during this time: Ethel (1960, Category 5), Camille (1969, Category 5), Elena (1985, Category 3), Georges (1998, Category 4-2), Ivan (2004, Category 4-3) and Katrina (2005, Category 5-3) (Morton, 2008; 2010). Coastal systems typically are able to recover from storm erosion during extended periods of fair-weather, a process indicated by differences in slopes of winter and summer beach profiles

(Bruun, 1962; Johnson, 1919; Swift, 1975). Shoreface equilibrium, with no net loss, requires that sediment stays in the littoral system to be re-deposited onshore by low-energy, constructive waves (Bruun, 1962; Johnson, 1919; Swift, 1975). Cipriani and Stone (2001) determined that present-day net sediment transport along the nearshore of each of the MS-AL barriers (excluding Cat Island) approaches zero at the islands' termini, indicating that, currently little sediment is transported between the islands (Cipriani and Stone, 2001). This net-loss of sediment within the littoral system of the MS-AL barrier islands is exacerbated by storm erosion (Morton, 2008). The storm channel breach on Ship Island was open and then filled in twice and has remained open since Hurricane Camille. The breach on Dauphin Island, likewise, closed several times (Morton, 2008, 2010; Otvos, 1981, 1985, 2005). Although storms associated with seasonal winter cold fronts are more frequent and exert more cumulative energy on coastlines than extra-tropical storms, the short periods between extreme storms since 1995 (Morton, 2008) leaves the islands in a weakened state and more vulnerable to destruction by subsequent storms (Morton, 2008; Schmidt, 2001).

From 1848 to 2005, Cat Island lost 40% of land area in comparison to 60% on Ship Island, 19% on Horn Island and 52% on Petit Bois (Morton, 2008). For the 1848-1917 interval, the averaged loss rate on Cat Island was 0.9 ha/yr; increasing from 1917-1959 to 4.9 ha/yr; and decreasing between 1950- 1986 to 3.4 ha/yr (Morton, 2008). From 1986-2005, the average loss rate increased again to 6.4 ha/yr, but a sub-interval from 2000-2005 the rate was 14.4 ha/yr, indicating that erosion spiked in the latter five years (Morton, 2008). The north-south spit, where land loss is greatest, has been eroding disproportionally at the southern end where low angle wave approach reaches the island (Morton, 2008; Rucker and Snowden, 1989). The relative lateral stability of Cat Island has been largely attributed to the protective environment sustained by the active St. Bernard Delta for a ~2000 year period (Otvos, 1985 a, 2005 a; Otvos and Giardino, 2004; Rucker and Snowden, 1989, 1990; Schmidt, 2001).

Project Rationale

The purpose of this project was to determine whether a pre-Holocene topographic high presently underlies Cat Island in an effort to increase understanding of the island's current position and stability. Mapping of the Pleistocene surface was accomplished through the acquisition of geophysical data in the unsurveyed area around the entirety of Cat Island, from the nearshore out to the adjacent environments of the Mississippi Sound, Ship Island Pass and Cat Island Channel and integrating this new information with existing local and regional data. This reconnaissance survey provides a spatially low-resolution representation of the subsurface geometry of the Pleistocene surface underlying the island and its surrounding environment, thereby contributing new information to the existing regional configuration of this major erosional unconformity. This survey has also defined acoustically impenetrable areas where chirp seismic-based technology employed here is ineffective, which can advise future survey planning.

Methods

Existing Datasets

The following data from previous studies undertaken on or in close proximity to Cat Island were used to assist interpretation of the seismic data acquired in this study. Gross core descriptions, sediment lithology, mineralogy, faunal assemblages and geologic interpretations are summarized in the Appendix.

Mississippi State Geological Survey Borehole Data

Brown et al. (1944) describes well borings drilled at two sites on Cat Island (HC-199 and HC-200; figure 10). Well HC-199 was drilled to a depth of 296 m; the analysis includes sediment lithology, mineralogy and faunal assemblages. Well HC-200 was drilled to a depth of 161.5 m; lithofacies descriptions are provided. The geologic interpretations of the sedimentary facies in both wells showed a sequence of Holocene to recent sands at the surface overlying the Pamlico Sand Member of the late Pleistocene Prairie Formation, Plio-Pleistocene Citronelle Formation, and Pliocene Graham Ferry Formation units. In well HC-199, the contact between Recent and Pamlico Sand units was undefined (Brown et al., 1944). Throughout the Mississippi coast, where Recent beach sands and dunes directly overlie the seaward margin of Pamlico Sand Member, the contact between the two units is difficult to identify due to similarities in lithology and faunal assemblages of the sediments (Brown et al, 1944). The data from drillcore HC-199 (Brown et al., 1944) does not provide a depth to the Pleistocene surface and, therefore, were not used in the mapping of the Pleistocene surface conducted in this study. In well HC-200, the Holocene to recent deposits were identified as 1.8 m of white sand overlying 5.8 m of marsh mud or blue clay over fine gray sands of the underlying Pamlico Sand unit (Brown et al, 1944).

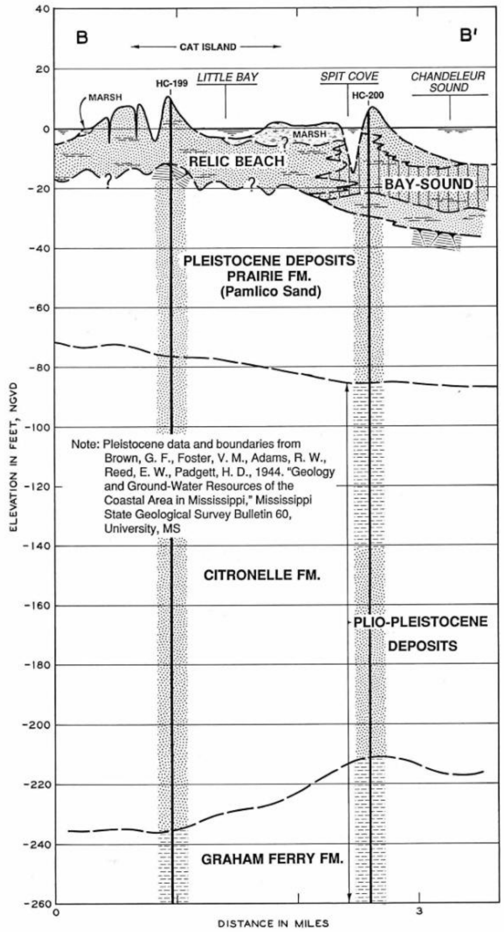


Figure 14. Interpreted cross section northwest-southeast across Cat Island based on cores HC-199 and HC-200 (Dunbar et al., 1995)

USACE Atlas Data

Dunbar et al. (1995) is an atlas composed of 1:62,500 quadrangle maps of the Mississippi River delta plain with corresponding geologic cross sections for each quadrangle (Dunbar et al, 1995). An interpreted stratigraphic cross-section (Dunbar et al., 1995; figure 14) of the depositional environments of the Holocene to recent sandy unit described in the two Mississippi State Geological Survey borings above. From the backbarrier marsh to the site of HC-199, the cross-section shows an undifferentiated upper relict beach sand unit ~6.7 m thick directly over the Pleistocene surface of the Prairie Formation at ~3.5 m bsl (Dunbar et al., 1995; figure 14). This upper beach sand unit is continuous beneath Little Bay and Middle Spit, where it interfingers with bay-sound deposits that extend under the N-S spit out into Chandeleur Sound. Across Middle Spit, the upper beach sand unit is capped by marsh, and at the site of HC-200, bay-sound deposits ~ 6 m thick both overlie and are capped by beach sands (Dunbar et al., 1995; figure 14). The upper beach sand unit of HC-200 is ~2.5 m thick and the lower beach sand unit is ~1.8 m thick (Dunbar et al., 1995; figure 14). The contact of the lower beach sand unit of HC-200 with the Pleistocene is interpreted at ~9 m bsl (Dunbar et al., 1995; figure 14). The criteria for establishing the depth to the Pleistocene surface by Dunbar et al. (1995) in core HC-199 (Brown et al., 1944) is not presented by the authors, and is not considered a reliable depth to the Pleistocene surface for the purposes of this study.

Mississippi Minerals Resource Institute (MMRI) Reports

Otvos (1985 b) synthesizes rotary drillcore data taken on the islands by the University of Mississippi Gulf Coast Research Laboratory (GCRL) during an approximate 10 year period prior to the publication of the report (Otvos, 1985 b; figure 10). Two cores were sampled in the Cat Island nearshore, both of which penetrated into Pleistocene facies (cores GRCL CE, GRCL OC, Otvos, 1985 b) (figure 10). These data were used by this study to validate the seismic data and in generating the Pleistocene surface contour map.

Otvos (1986 b) synthesizes vibracore data collected by the MMRI in 1985 in the passes and island offshore areas of the MS-AL Barrier Islands and Mississippi Sound (figure 10). None of these vibracores reached the Pleistocene surface, as they were unable to penetrate upper coarse sand and shell units (Otvos, 1986 b). The depths to the base of these cores are included in this study as an indication that the Pleistocene surface is necessarily deeper than the core base depths at each core site.

Velardo, 2005 Data

A series of three vibracores and an interpreted stratigraphic cross section in Mississippi Sound extending from ~1 km north of the North Point on Cat Island to ~4 km out into the Mississippi Sound from Velardo, 2005 (figure 10) were used to assist interpretation of Pleistocene surface contours in this study. Although the cores did not penetrate the Pleistocene surface, the cross-section was used in this study to indicate the sites where the Pleistocene surface is deeper than the given base depth of the core. Grain size was derived from gamma bulk-density profiles (Velardo, 2005). For core TM-VC28 a sediment accumulation rate of 0.2 cm/yr was calculated from excess ^{210}Pb and ^{137}Cs activities (Velardo, 2005). The cores were taken in water depths ranging from 3.7-4.5 m bsl (Velardo, 2005). These water depths were added to the base depths of the cores to give the below-sea-level elevation mapped here: TM-VC28 (-6.5 m), TM-VC27 (-7.5 m) and TM-VC26 (-6.7 m). The northernmost core, TM-VC28, is composed of an upper muddy unit transitioning to sand at ~1.3 m below sea floor (bsf) that continues to the base of the core at ~2.5 m bsf; core TM-VC27 shows an upper muddy unit to ~ 1.8 m (bsf) over sand to the base at ~3.0 m bsf; core TM-VC26 is composed of an upper sandy unit to ~1.0 m bsf overlying mud to ~3.0m bsf (Velardo, 2005).

Data Collection

Data collection consisted of a geophysical survey in the near to offshore waters surrounding Cat Island conducted with two chirp sub-bottom profiler systems employing different frequency ranges. Chirp sonar uses pulsed FM signals covering full spectrum frequency ranges to penetrate the shallow sub bottom sediment (LeBlanc et al., 1992). Higher frequency acoustic waves provide greater vertical resolution with less depth penetration while, lower frequency waves penetrate deeper but provide less vertical resolution. Profile images of

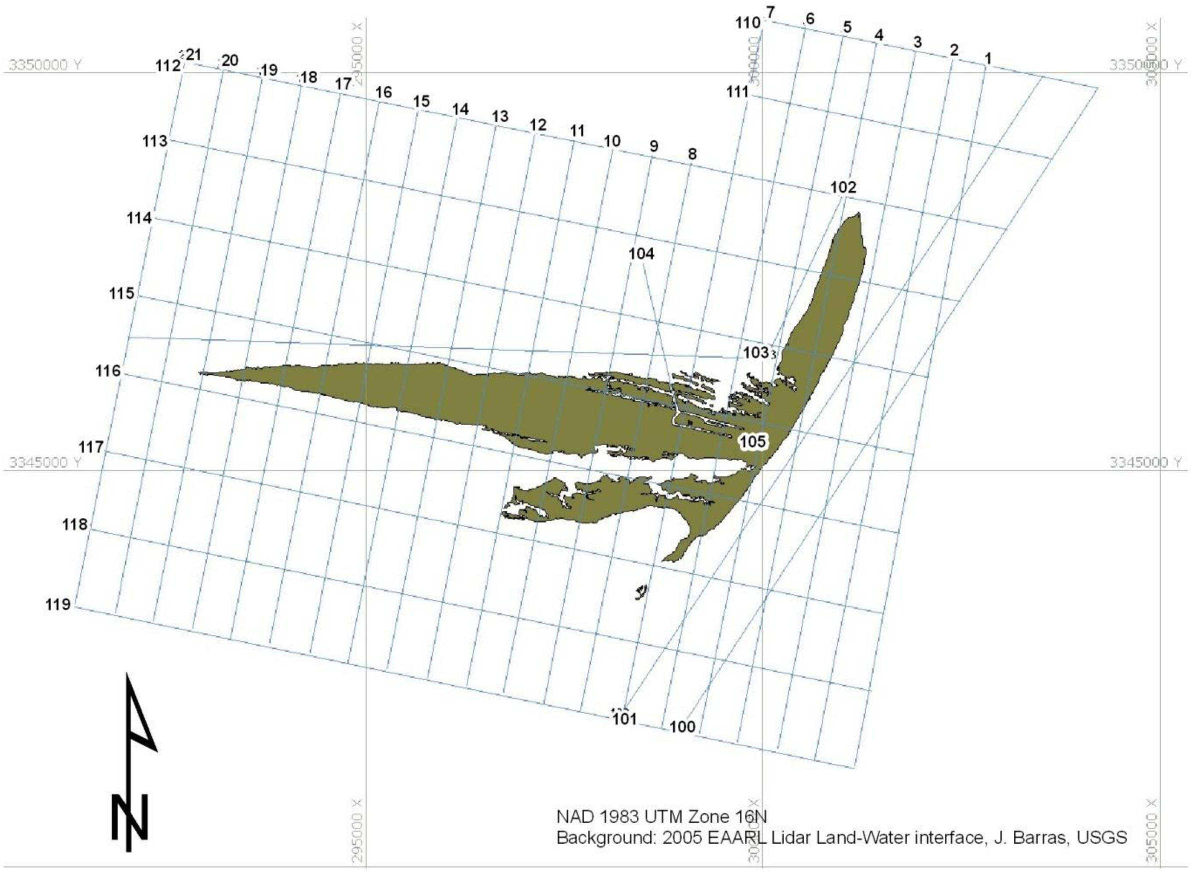


Figure 15. Planned survey trackline grid

stratigraphic layers are generated when pulsed signals emitted from a towfish reflect off sediment of different acoustic impedance values, or density. The reflected signal is received by the same towfish, and transmitted to a topside processing unit that calculates the location and depth of sediment horizons from two-way travel time and displays the density horizons in a two-dimensional visual profile.

EdgeTech towfish models SB-216S, provided by UNO, and SB-424, provided by the USGS St. Petersburg Coastal and Marine Science Center, were deployed on separate, multi-day cruises, each collecting data along different transects of the study area. Two different systems were used because of limited availability of either system, and of suitable weather conditions. Towfish were suspended off the stern or side, submerged approximately 1 to 2 m above the seafloor as the vessels cruised at approximately 2.6 m/s. The SB-216S unit simultaneously transmits pulses across three frequency ranges: 2-16 kHz, 2-12 kHz and 2-10 kHz. The SB-424 emits pulses across three higher frequency ranges: 4-24 kHz, 4-20 kHz and 4-16 kHz that have lower penetration potential but return signals with greater vertical resolution than the SB-216S (EdgeTech, 2010). The towfish units were connected via a tow cable to a Topside Processing Unit (TPU) utilizing EdgeTech Discover 3200-XS data acquisition software, which runs the chirp acquisition controls as well as receiving reflected signals from the towfish, converting them into a 2-D image that can be viewed in real-time on the topside laptop monitor (EdgeTech, 2010). Navigation data was provided by a Global Positioning System (GPS) receiver feeding a continuous string of coordinates to the acquisition software that was integrated with the seismic data file.

Survey Preparation

Prior to surveying, a grid of georeferenced tracklines was plotted using Hypack 2008© navigation software (figure15). The study was conceived as a reconnaissance survey that would identify major subsurface features, so the grid was designed to cover as much area as possible in the few days allowed. The resulting grid provided navigation for a spatially coarse resolution survey. The tracklines extended approximately 3-4 km north and south of the main island body, approximately 2 km north and south of the eastern spit and approximately 1 km east and west of the island termini. Twenty-one parallel lines spaced 500 m apart from east to west form the vertical grid lines. These N-S lines were angled slightly to the shoreline but normal to the strike of the subaerial beach ridges that comprise the east-west trending main island with the intent of capturing cross sections of any preserved shoreface deposits associated with relict beach ridges. Ten parallel east-west lines, spaced 1 km apart, were plotted perpendicular to the N-S

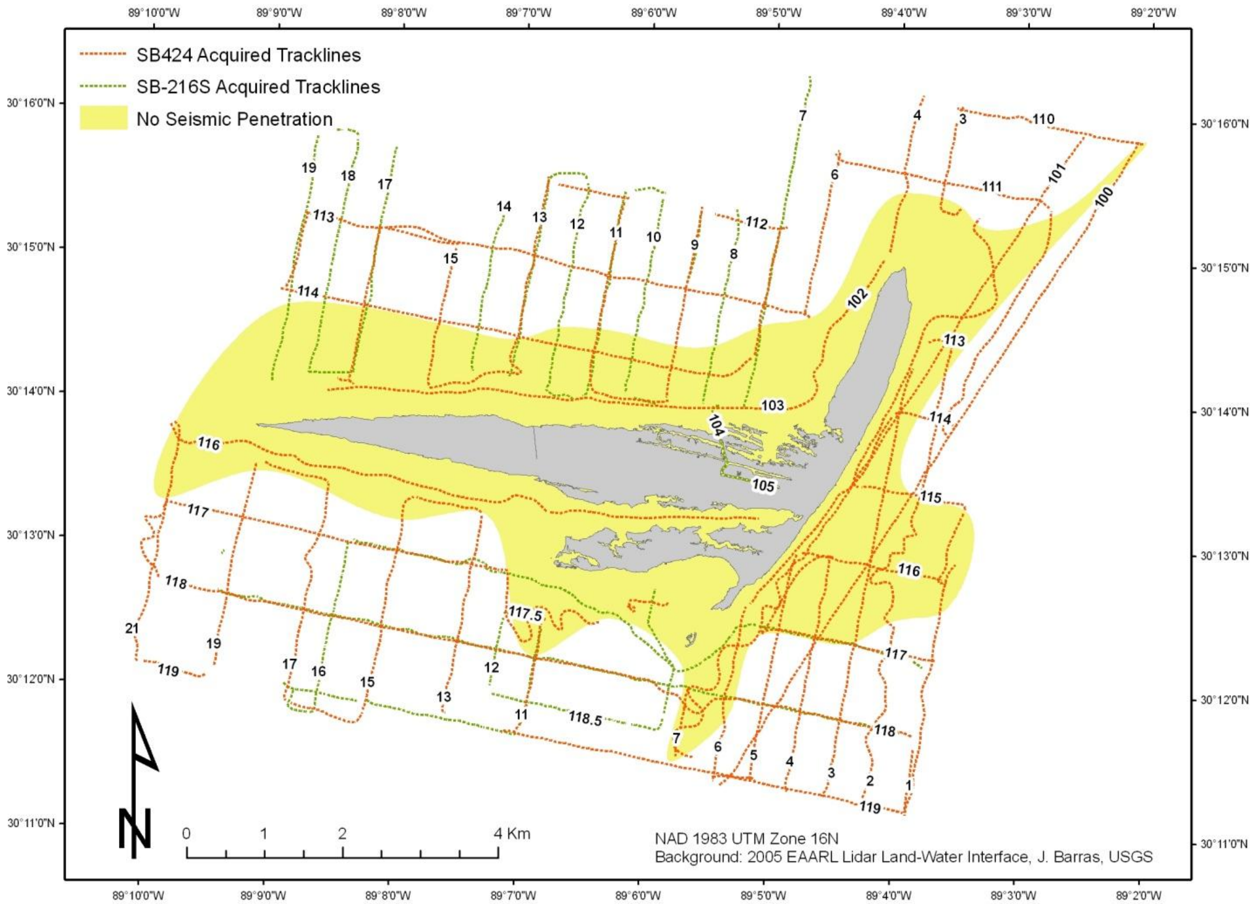


Figure 16. Acquired survey lines and extent of area of poor seismic penetration (barrier island platform).

lines. Additional lines were plotted along the main island and spit shorelines in order to establish the landward extent of the nearshore zone and two lines oblique to the main grid but shore-parallel to the N-S spit ran the length of the spit. Another line was plotted through the man-made navigation channel on the northeastern island that cuts across the beach ridges into the island interior in an attempt to detect features beneath the island. The trackline grid shows the intended coverage of the survey area. Optimal surveying conditions require relatively calm waters that so that scattering of the acoustic signal, resulting in noisy or totally obscured data, is minimal. The order in which tracklines were covered was determined by the weather and sea conditions each day, and not all of the planned lines were surveyed (figure 16). A listing of data files associated with each trackline is shown in Appendix Table 1.

Data Acquisition

EdgeTech SB-216S survey

Seismic data collected with the SB-216S chirp system was acquired on November 17th, 2008, April 8th, 2009 and June 12th, 2009 the UNO *RV Mudlump*, a 22 ft aluminum workboat. The vessel was retrofitted with a davit and outrigger on the bow center from which the towfish wassuspended in order to mitigate interference from the boat's outboard motors. The cable feed from the towfish connected to an Edgetech 3100P topside processing unit that received and processed the seismic return signal. The processed data was then sent via ethernet connection Edgetech Discover 3200-XS software running on a laptop computer. The processing software is the interface controlling the settings and operation of the chirp system, and also displays the image in real-time, and stores the data in files (.jsf or .sgy format) on the laptop's hard drive. Geographic coordinates were streamed from a Thales ZMAX dual frequency GPS receiver to the laptop computer, and synced and recorded by the Discover 3200-XS software with the seismic data files.

EdgeTech SB-424 and single beam bathymetry survey

Seismic data collected with the SB-424 chirp system was collected during a three-day period (June 21st, 22nd and 23rd, 2009) with the U.S. Geological Survey on the *R/V Catboat*, a 25 ft Glacier Bay Coastal Runner catamaran. The vessel is outfitted with a sliding aluminum truss, which sits on a pair of fixed rails that run forward from the stern between the hulls. Instruments can be mounted on the truss and then repositioned under the vessel. This arrangement minimizes interference with other surveying equipment, so that multiple instruments can be deployed simultaneously. During this survey, the echosounder transducers were attached to the truss and secured approximately midship, and the chirp towfish was suspended from the stern. The SB-424 chirp system was set up in essentially the same configuration as in the SB-216S survey. All the SB-424 data were recorded in SEG-Y (.sgy) file format.

Bathymetric soundings were collected using a SEA SWATHplus-H 468 kHz dual transducer interferometric echosounder system. Swath bathymetry systems send and receive acoustic signals in an arc from the transducer to the seafloor, so that coverage is obtained in a continuous sweep along the width and the track of the swath to either side of the vessel nadir. The swath width at the seafloor is increases with water depth, and is therefore dependent on the height of the transducer above the seafloor. Acoustic signals emitted from the swath system's transducer travel through the water column and reflect off the seafloor. The incident angle and two-way travel time of the return phase of the backscatter waveforms are received by the second transducer and sent to the Transducer Interface Unit (TIU). The converted data are sent to a topside laptop computer running the processing software. The data were stored in eXtended Triton File (XTF) file format. In addition to processing and storing the data, the acquisition software displays real-time imagery of the seafloor and functions as the interface for the controlling settings for and receiving feedback from the system. Vessel attitude and positional data, from an auxiliary motion sensor recording pitch, heave and roll, and from a RTK differential GPS antenna, recording vessel vertical and horizontal position were also recorded.

Ideal bathymetric surveys plan trackline spacing so that swath margins overlap and complete coverage of the seafloor is obtained. The survey grid used here covered only ~5% of the survey area, with swath widths ranging from 8 m to 40 m wide and gaps between data ranging from ~100 m to over 900 m (personal communication, Elizabeth Pendleton, USGS Woods Hole Coastal and Marine Science Center).

Post-Survey Processing

Seismic Data Post-Processing

SB-424 seismic data were processed using Seismic Unix, a public domain seismic processing software package (Center for Wave Phenomena, Colorado School of Mines) and computational scripts written by the USGS Seafloor Mapping Group. The SEG-Y and navigation files were read and converted into Seismic Unix format. Optimal gain settings were determined by trial and error on a few select lines, and were then applied to all the data using an automatic gain control script. Gained files were then exported back into SEG-Y format (personal communication, Arnelle Harrison, USGS St. Petersburg Science Center).

The gained SB-424 and the SB-216S data were imported into SonarWiz.MAP v.4.04.0074 (Chesapeake Technology, Inc., 1999-2010) projects. SB216S files were imported using gain settings ranging between 16x to 256x; SB-424 files were imported using the 2x gain setting. Within each project, each file, containing seismic and position data for a segment of trackline, was mapped and projected in Universal Transverse Mercator (UTM) Zone 16N using the 1983 North American Datum (NAD83). Seismic data were displayed in the digitizer viewer where gain levels, image resolution and intensity and, in some cases, band pass filtering were applied in order to achieve the clearest image or to resolve reflectors of interest. Also from within the digitizer viewer, a graphic seafloor reflector was automatically generated from the towfish bottom track signal and manually corrected as needed. Graphic annotations (i.e., digitized features) were traced over significant reflectors in the seismic image. The digitized features were categorized as “seafloor reflector”, “Pleistocene surface”, “truncated clinoforms” or “channel”, and stored as georeferenced files in the feature manager database. The digital features were then exported from SonarWiz.map as ASCII XYZ format files, where the X and Y values are the easting and northing coordinates and the Z value is the depth below the towfish position. The gained images were exported as JPEGs.

Bathymetry Data Post-Processing

The bathymetry data were post-processed using Computer Aided Resource Information System (CARIS) software (http://www.caris.com). The navigation data were inspected and edited for accuracy and applied to the raw bathymetric soundings so that anomalous soundings could be eliminated. Elevation data, recorded by the RTK GPS during the survey, were also applied. The processed bathymetric data were then interpolated to produce a grid surface (5 m grid-node spacing) and exported as an ASCII XYZ text file. The XYZ file was then rasterized and a 20 m resolution continuous grid was interpolated from the rasterized base surface using Generic Mapping Tools software (GMT, http://imina.soest.hawaii.edu/gmt/) (personal communication, Elizabeth Pendleton, USGS Woods Hole Coastal and Marine Science Center).

Pleistocene Surface Mapping

In order to compare sedimentary horizons identified in this survey with those from other surveys, deriving a depth below sea level is essential. Sea level provides a consistent regional zero elevation and is the standard means of describing the relief of features on or below the earth’s surface. The elevations of the subsurface features in this survey as calculated by the acquisition software represent the distance from the chirp system’s towfish, which is variable throughout the survey, to the subsurface digitized reflectors. To obtain the elevation below sea level for the Pleistocene surface features, the elevation of the Pleistocene reflectors below the sea floor was added to the bathymetry elevation using the following equation:

$Z_{PS} = Z_1 + Z_2$ where

- Z_1 = the distance from the seafloor to the digitized Pleistocene reflectors
- Z_2 = the distance from MSL to the seafloor (bathymetry)

This equation was executed in ArcMap using the “Raster Math” tool. Using the previously interpreted depths of the Pleistocene surface from drillcores and acoustic surveys ranging from 3 m bsl (Curray and Moore, 1963) to 14 m bsl (GRCL CE, Otvos, 2005) as a reference range, seismic reflectors originating within that range were identified as the Pleistocene surface and digitized. To obtain the Pleistocene reflector depth below the seafloor, a unit thickness was calculated within SonarWiz.MAP by measuring the distance from the digitized seafloor feature to the digitized Pleistocene features. The unit thicknesses, measured from the depth below the seafloor, were exported as an ASCII XYZ file. The XYZ file was converted into a multipoint feature shapefile in ArcGIS . The shapefile was then converted into a raster using the ArcGIS “feature to raster” tool, assigning the Z value to each pixel. The Pleistocene thickness raster (Z_1) was then imported into ArcMap along with the bathymetry raster (Z_2), and the ArcGIS “Raster Math” tool was used to derive the sum of the pixel values of the two datasets together, giving an output raster of below sea level depths for the Pleistocene surface. Attempts to use contour these data using contouring software were unsuccessful, perhaps because of the spatial distribution of the data. The corrected Pleistocene depth raster was converted back to a multipoint shapefile so that each point represented the depth to the Pleistocene surface below MSL. The points were displayed in ArcMap and were contoured manually to create the map of Pleistocene surface contours (figure 17).

Due to a lack of coincident bathymetric data, the seismic data collected with the SB-216S chirp system could not be interpolated to a “below sea level” depth. Subsurface reflectors for the

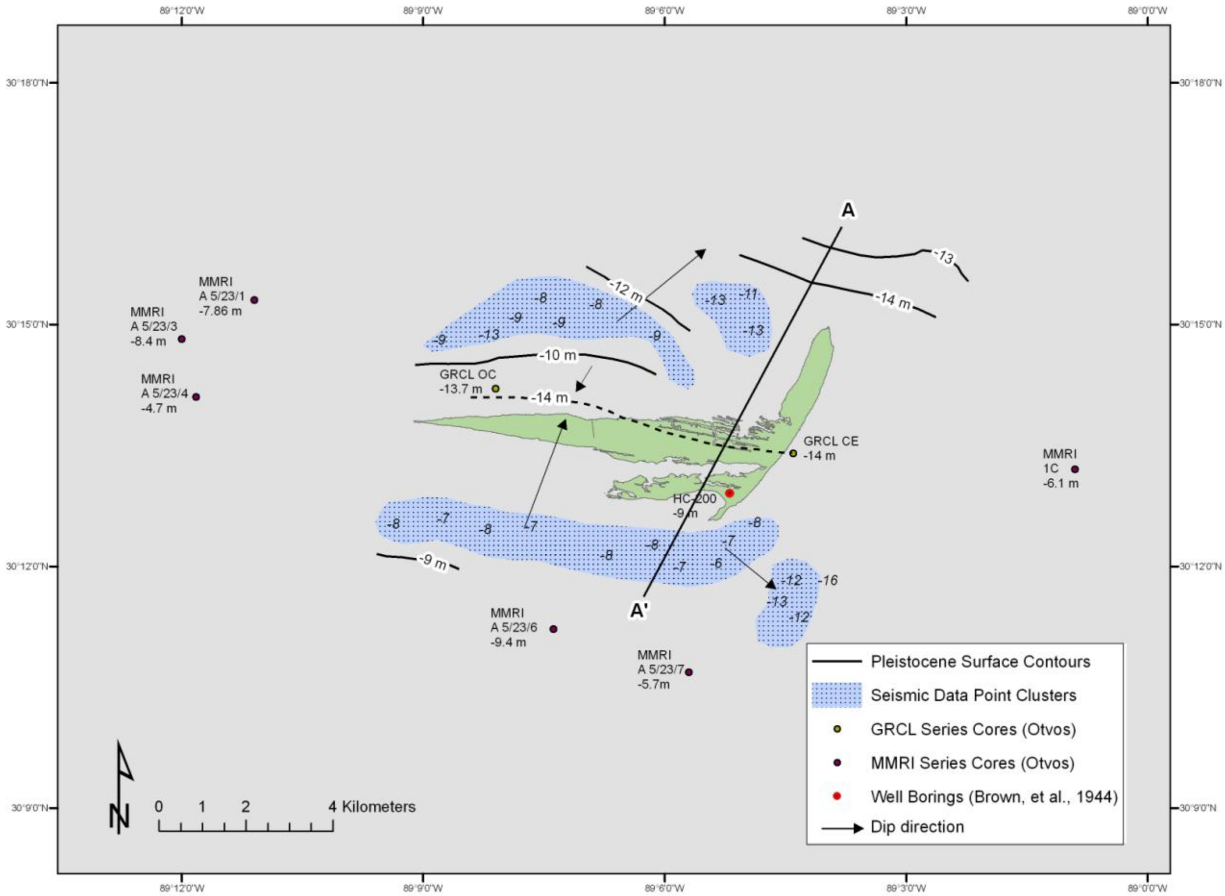


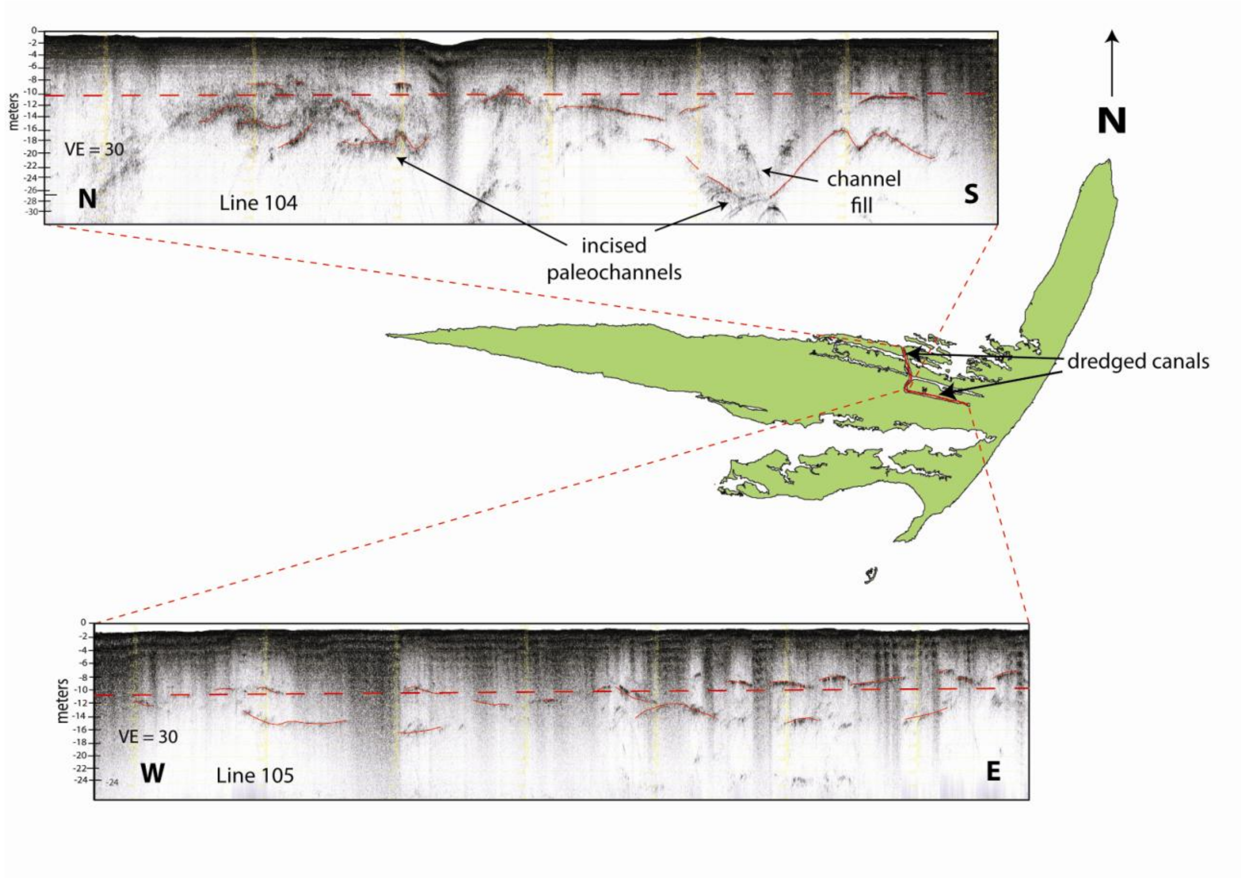
Figure 17. Contour map of the Pleistocene surface surrounding and underlying Cat Island, MS and locations and depths of vibracores (Otvos, 1986; Velardo, 2005) and well borings (Brown et al., 1944; Otvos, 1985 b) acquired during previous studies on and around Cat Island. Shading represents areas where clusters of depth-value points derived from the seismic data are present, but unable to be contoured. A to A' indicates the strike of barrier cross-section shown in figure 24.

SB-216S dataset can therefore only be described in terms of the depth below the seafloor, geometry, horizontal extent, and sedimentary layer thickness. Although reflectors were present in the SB-216S data that appeared likely to represent the Pleistocene surface based on a rough estimate of depth below sea level, the characteristic visual appearance of the reflectors, and positional agreement with the SB-424 data where overlap occurred, only the data acquired with the SB-424 system, concurrently with single-beam bathymetric data, could be adequately mapped.

Results

Pleistocene Surface Contours

The seismic survey covered a zone encompassing the island and extending seaward ~1.5 km to ~2.5 km, roughly coinciding with the -2 m to -2.5 m bathymetric contour from the island shoreline where neither the SB-216S nor SB-424 chirp system was able to penetrate the substrate (figure 16). This lack of penetration could be attributed to interference from the sandy barrier island platform or from the shallow water depth; lines 102, 103 and 116 returned no data (figure 16). The acoustic properties of a thick, well-sorted sandy lithosome (high density and acoustic impedance of quartz grains, and low porosity of the lithosome) such as a barrier platform increase impedance and decrease penetration of the chirp signal (LeBlanc, et al., 1992). Penetration depth was generally greater along the Sound side of the island. Few reflectors at potential Holocene/Pleistocene contact depths were found on the Gulf side, except along the E-W trending line 118 (figure 16). Although water depth increases more rapidly from the north-south spit shoreline toward Ship Island channel (from to -3 m in less than 500 m), the seismic data returned only seafloor multiple reflectors along survey lines 100 and 101 until the ends of the lines, approximately 2-2.5 km from the tips of the north and south ends of the spit. Core GRCL CE (figures 10 & 17) reported a 4.2 m thick fine sandy lithosome underlain by a medium sand and silty-muddy-fine sand unit 8 m thick (Otvos, 1985 b), which would likely cause impedance of SB-424 acoustic penetration in this area. Line 5 and the shoreward 1000 m segment of line 6 were not surveyed as the water was too shallow to accommodate the vessel. Rough sea conditions may have contributed to the unusable data returned by the SB216-S system along lines 7, 8 and 9 (figure 16). On the island's northern side in the Mississippi Sound, penetration of the acoustic signal to the Pleistocene erosional surface began to occur at a



threshold distance of

Figure 18. Pleistocene surface reflectors in seismic profiles from survey lines 104 (north to south) and 105 (east to west) in the canals. The surface is irregular, and incised with paleochannel reflectors in the north-south line

~1.5 to ~2.5 km from the island shoreline and continued northward to the termination of the survey grid.

The Pleistocene surface mapped in this study display an irregular topography with depths ranging from ~6 - ~16 m bsl. The highest Pleistocene surface elevations found in this study, at ~6 - ~8 m bsl returned from the seismic data, are located along the southern nearshore margin on the Gulf side of the island and at ~8 - ~9 m bsl along the northern, Sound side of the island, between the -10 and -12 m contours (figure 17). The lowest elevations of the Pleistocene surface mapped in this study range from ~14 - ~6 m bsl: the Pleistocene/Holocene contact derived from drillcore data, at ~13.7 - ~14 m bsl are located off the northwestern and central eastern island shorelines (GRLC drillcores, MMRI Open-File Report 85-6F, Otvos 1985 b; figures 10 & 17); the Pleistocene surface reflectors returned from the seismic data at ~14 m bsl are located north of the N-S spit (figure 17) and in the interior canals (figure 18), and at ~16 m bsl southeast of the South Spit terminus (figure 17). Depths of the Pleistocene surface reflector points returned from the seismic data and depths to the Pleistocene/Holocene contact from existing drillcore data (Brown et al., 1944; Dunbar et al., 1995; Otvos, 1985 b) were used to generate a contour map for the Pleistocene surface (figure 17). Shaded polygons were drawn to the north and northeast of the main island between the -10 m and -12 m depth contours and between the -12 m and -14 m contours and along the southern nearshore zone. The shaded polygons represent areas where clusters of seismically derived Pleistocene-depth data points are present but have highly variable Z values, so that no trend of the Pleistocene surface geometry could be discerned. The values within the shaded areas indicate depths of representative data points. Arrows indicate the general downdip direction of the contours.

Contours of the Pleistocene surface underlying Cat Island strike east to west. Overall trends derived from the seismic and existing drillcore data (Brown et al., 1944; Dunbar et al., 1995; Otvos, 1985 b) suggest that the Pleistocene surface dips from both the Gulf and Sound sides of the island toward a topographic low beneath the island. This elongate depression of the Pleistocene surface beneath the present-day island trends east to west from the northwestern island nearshore, passing through the position of the north-west canal, to the central eastern shoreface (figure 17). Presence of this trough is supported by the position of Pleistocene sediments in drillcores GRCL OC at 13.7 m bsl and GRCL CE at 14 m bsl (MMRI Open-File Report 85-6F, Otvos, 1985 b; figure 17) and Pleistocene surface reflectors ranging from ~11 - ~29.5 m bsl found in the seismic data acquired in survey line 104 (figures 17 & 18). Topographic

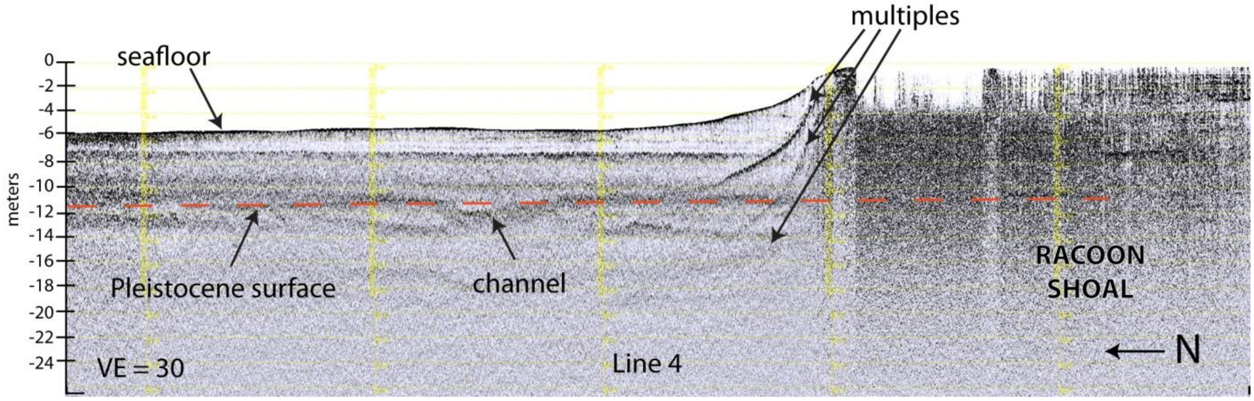


Figure 19. Seismic profile of survey line 4 showing the Pleistocene surface reflector onlapped by Racoon Shoal facies.

highs of the Pleistocene surface underlying Cat Island are located along the northern and southern nearshore margins of the island (shaded areas, figure 17). The Pleistocene surface also downdips soundward from the topographic high on the island's north side soundward out to the northern-most extent of the survey grid and gulfward from the topographic high located beneath the island's southern nearshore zone (figure 17).

On the northern, Sound side of the island, the Pleistocene surface downdips toward the island from the -10 m Pleistocene surface contour that extends along the nearshore margin (figure 17). To the north of the -10 m Pleistocene surface contour, the Pleistocene surface downdips soundward toward the -14 m Pleistocene surface contour located north of the North Point (figure 17). Depths of the reflectors between the -10 m and -12 m Pleistocene surface contours, where orientation trends of the Pleistocene surface could not be discerned (shaded area, figure 17), range from ~13 m bsl to ~8 m bsl. Data points north of North Point were dense enough to discern -13 m and -14 m Pleistocene surface contours. Figure 19 shows a Pleistocene surface seismic reflector in survey line 4 (figure 16), approximately 2 km offshore from the North Spit terminus, at ~14m bsl onlapped by the retrograded spit platform that forms the base of modern-day Raccoon Shoal (figure 19). To the northwest of North Point, soundward of the survey grid, a series of three vibracores reached depths of 6.5 m bsl, 6.7 m bsl and 7.5 m bsl without penetrating Pleistocene clays (Velardo, 2005; figures 10 & 17), indicating that the Pleistocene surface at these sites is deeper than the base of the cores.

Pleistocene surface relief trends on the southern, Gulf side of the island were difficult to identify, as very little data was returned from the shore-perpendicular cross-tie seismic survey lines. The seismic data returned from line 118 (shore-parallel) returned points alternating between ~9 - ~7 m bsl, and ~12 - ~16 m bsl in the south-westernmost area of the survey grid, approximately 2 km offshore (shaded areas, figure 17). Although the dip orientation of the Pleistocene surface topology cannot be resolved from these data, core HC-200 (Brown et al., 1944, figure 17) indicates Pleistocene sediments at ~9 m bsl, confirming an islandward deepening trend similar to the islandward deepening trend observed on the island's Sound side. Core MMRI A 5/23/6 (MMRI Open-File Report 86-1F, Otvos, 1986 b; figures 10 & 17), ~1.5 km south of the ~7 - ~8 m bsl zone penetrated to 9.4 m bsl without reaching the Pleistocene, suggesting that dipping continues further offshore toward the Gulf.

Drillcores GRCL OC ~500 m from the northern island shoreline and GRCL CE (MMRI Open-File Report 85-6F, Otvos, 1985 b) off the central eastern spit shoreface show Pleistocene sediments at ~13.7 m bsl and ~14 m bsl respectively. A -14 m contour is interpolated (dashed line, figure 17) between these two core sites. Pleistocene surface reflectors were also found in the seismic data acquired in the dredged canals (figure 18) cutting into the island behind the N-S spit. These seismic data was not adjusted to MSL as bathymetric data was lacking, but the observed water depth during the survey was approximately 1 m throughout the canal. A correction of -1 m was added to depths of the Pleistocene surface reflectors in order to make the data more compatible with the bathymetry-corrected seismic data. The resulting data from survey line 104 in the north-south trending canal showed several paleochannel incisions through

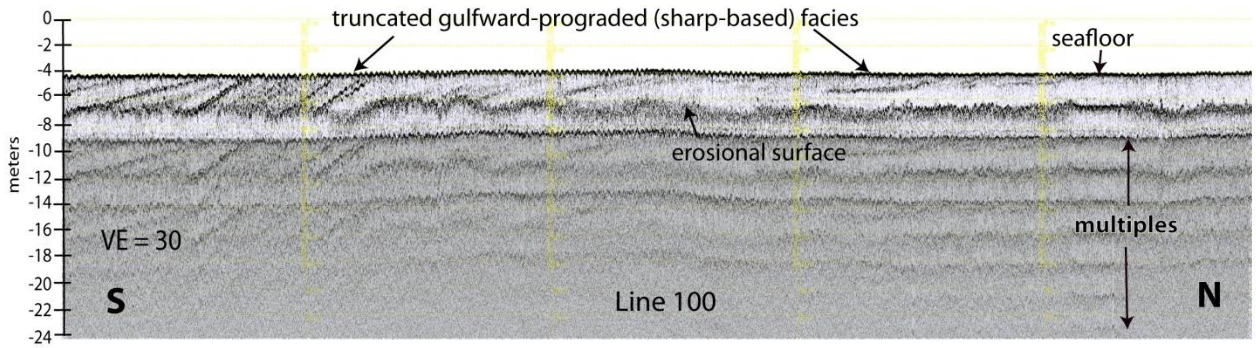


Figure 20. Seismic profile (SB-424) of line 100 showing truncated gulfward-prograded sharp-based facies.

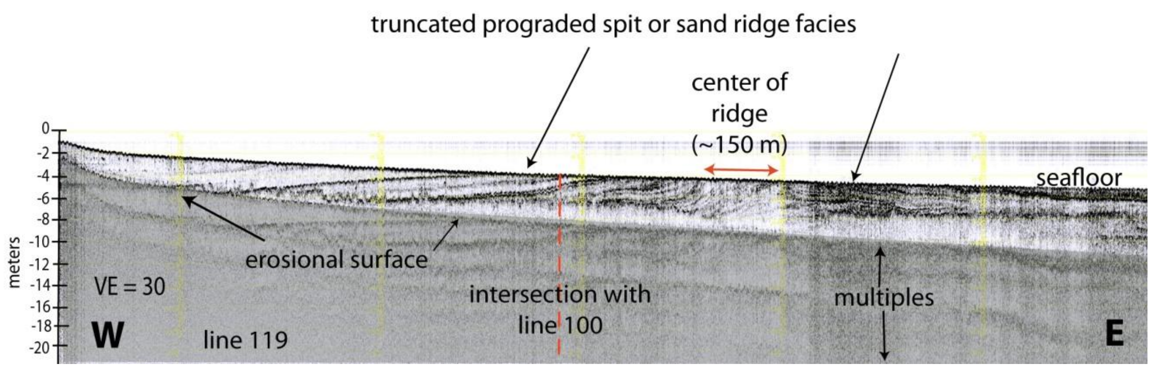


Figure 21. Seismic profile (SB-424) of line 119 showing east and west dipping truncated prograded facies dipping away from each other.

the Pleistocene surface. The uppermost depths of the channel levee reflectors are present at ~11 m bsl. The deepest channel thalweg extended to ~29.5 m bsl (figure 18). Seismic data from line 105 in the lower east-west trending canal showed an irregular, hummocky surface at ~11 - ~15 m bsl (figure 18). The seismic data from the canals were used to support the interpolated strike of the -14 m contour, and, therefore, the presence of a topographic low beneath the island.

Recent Holocene Features

At the southwestern margin of the survey grid, along the southernmost ~500 to ~600 m of lines 5, 6 and 100 (figure 16), a series of steeply angled, sharp-based southward dipping reflectors extend from the seafloor surface to a sedimentary horizon ~1.4 - ~3.5 m bsf (figure 20). Seismic line 119, trending perpendicular to line 100 (figure 16), shows similar but less-steep

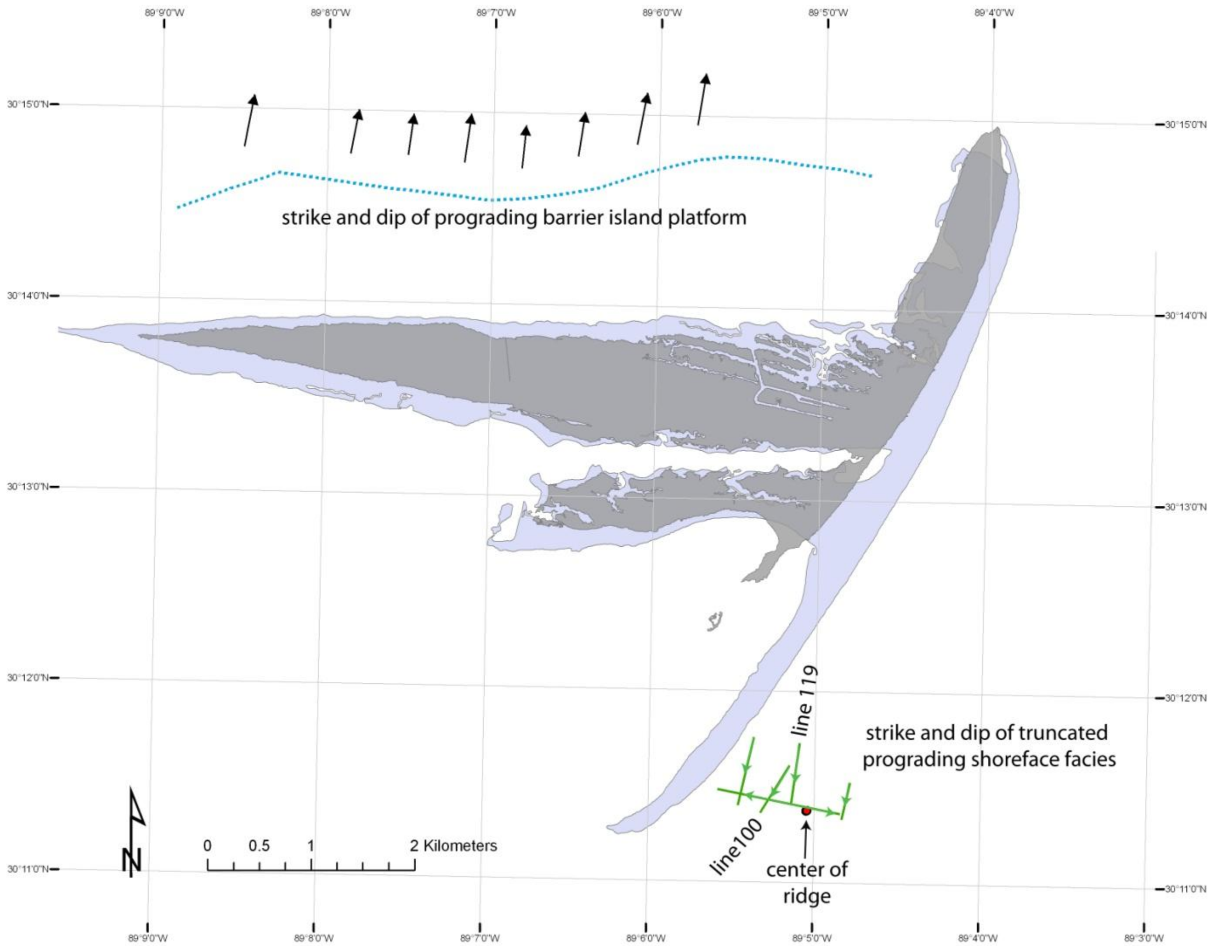


Figure 22. Locations of truncated ridge facies and retrograded barrier island facies

reflectors dipping to the east and west (figure 21). These reflectors are located in close proximity to the 1849 shoreline of the south spit and may be either remnants of stacked, prograding spit facies or an offshore sand ridge truncated by erosion (figure 22).

On the north side of the island in all the north-south oriented survey lines, at ~1 - ~2 km offshore, steep clinoform seismic reflectors onlap an underlying relatively flat sedimentary horizon and indicate the location and orientation of a previous soundward barrier retrogradation (figures 22 & 23). This retrograded lithofacies unit is overlain by the most recently deposited lithofacies unit that forms the present-day seafloor. Shoreward of the retrograded clinoforms, the lithofacies are acoustically transparent. This indicates that the upper and retrograded barrier

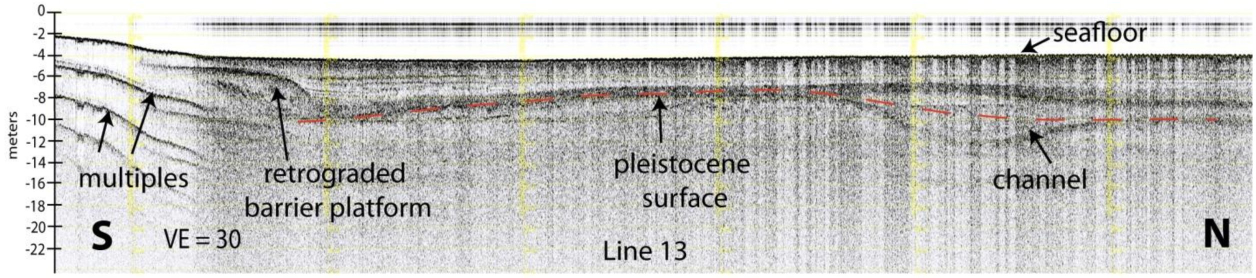


Figure 23. Seismic profile (SB-424) of line 13 showing retrograded barrier platform facies.

facies contain a high percentage of sand and is confirmed by core GRCL OC (MMRI Open-File Report 85-6F, Otvos, 1985 b; figures 10 & 17), situated shoreward of the location of where the retrograded clinoforms were mapped, which reported an upper unit ~4.2 m thick overlying a silty-muddy-very fine sand unit

.

Discussion

The new seismic data acquired for this study, combined with existing core data, document a vertically and laterally irregular Pleistocene surface underlying Cat Island and its nearshore area. These data indicate that the present-day island is situated over a topographic low, or trough, ~14 m bsl, rather than a topographic high as originally proposed by this study (dashed line, figure 17). Elevations of the Pleistocene surface mapped in this study (figures 17 & 24) indicate topographic highgrounds of the Pleistocene surface at ~8 - ~9 m bsl beneath the northern, sound-side nearshore margin and ~6 - ~8 m bsl beneath the southern, gulf-side nearshore margin of the present-day island (shaded areas, figure 17). The topographic highs of the Pleistocene surface to the north and south of the island strike roughly parallel to each other and to the mainland shoreline. These ridges define the boundaries of the trough (figures 17 & 24) that extends beneath the island from the northwestern shoreline to the central eastern nearshore. The Pleistocene surface also dips gulfward from the topographic highground beneath the southern, gulf side of the island, and soundward from the topographic highground beneath the northern sound side of the island. Striking north to south (A to A’, figures 17 & 24), the configuration of the Pleistocene surface underlying Cat Island and its nearshore can be visualized as a saddle, with the lowest point, the seat of the saddle, between the two east-west oriented topographic high-grounds north and south of the island (figure 24).

Seismic profiles acquired during this study identify infilled paleochannels incising the Pleistocene surface situated beneath the north-south trending canal that cuts into central Cat Island and between the southern gulf-side and northern sound-side topographic highgrounds. This suggests that prior to and during mid to late Holocene transgression (Oxygen Isotope Stage 3, Phase 1), the low-stand surface geomorphology was influenced by a fluvial system. Such an

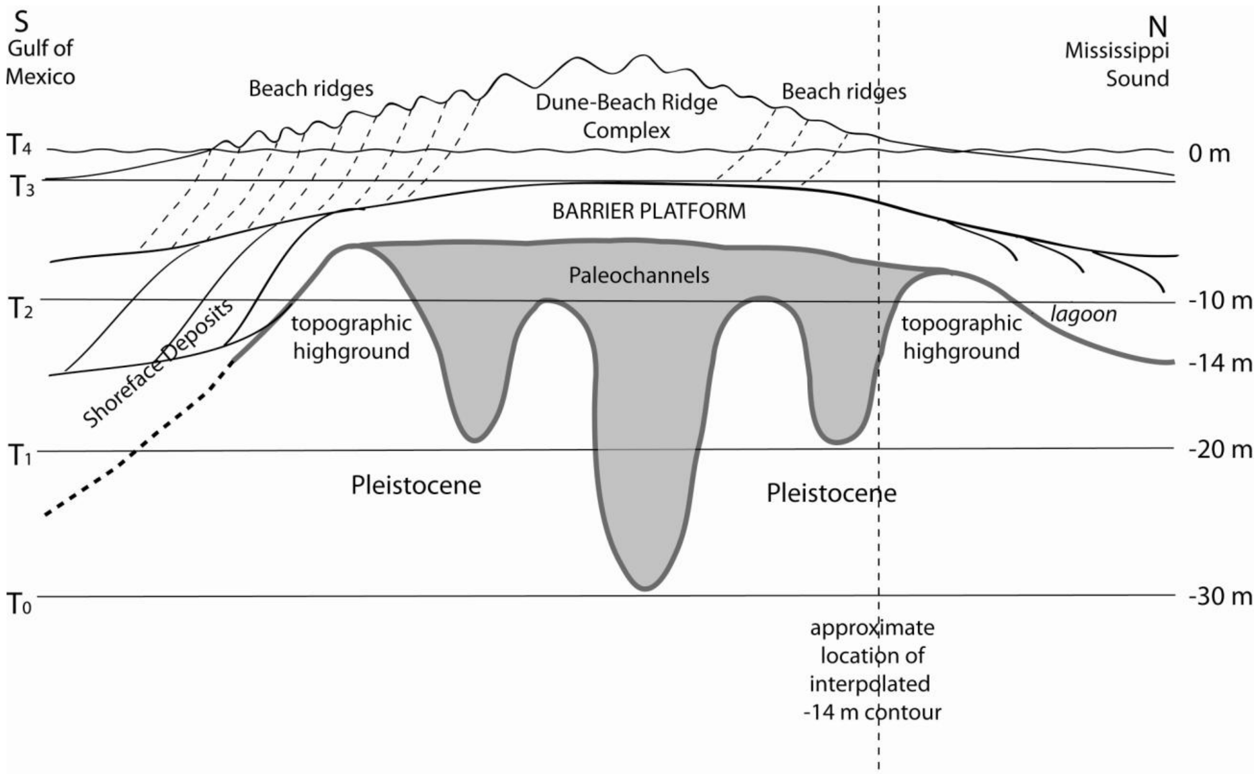


Figure 24. Geologic cross-section from north to south across Cat Island and nearshore zones. Time intervals (T_0 toT_4) correspond to the four stages of Cat Island development (figure not to scale). Location of cross-section is shown in figure 17 (A to A').

environment would presumably construct the associated features of backstepping fluvial depocenters and drowned estuaries modified by transgressive-regressive sequences during late Wisconsinan sea level rise. Backfilled tributary and distributary network channels, levee deposits, crevasse splays, bar finger sands, distributary mouth bars and beach ridge strandplains (Fisk et al., 1954; Frazier et al., 1963; Brooks et al., 1995; Morton and Suter, 1996; Roberts, 1997; Rodriguez et al., 2004) could result in the locally variable Pleistocene surface topography mapped in this study. Fluvial incisions on the Pleistocene surface of the Mississippi Sound have been reported south of the Pascagoula Channel to ~12 m bsl (Curray and Moore, 1963), within the Mobile River delta and across the inner shelf of the Alabama, Mississippi, Louisiana (Greene et al., 2007, Kindinger 1989; Kindinger et al., 1989; Figure 2) and Texas coasts (Morton and Suter, 1996; Rodriguez et al, 2004; Shepard, 1960 a; Wilkinson, 1975).

The thalweg of the largest paleochannel is positioned at ~29.5 m bsl and is flanked by two smaller fluvial incisions to the north and south, positioned at ~20 m bsl (figure 18). The orientation of the channel could not be ascertained from the data used in this study. The channel may strike from northeast to southwest, from northwest to southeast or east to west. The topographic high-grounds to north and south of the channel incisions are interpreted to be either relict shoals formed during previous late Wisconsinan transgression cycles that constrained the path of the active fluvial channel and transgressive deposition within the channel, or the levees of a much larger, pre-Pleistocene channel. In the latter case, the paleochannels seen in the seismic profiles of this study are incising earlier channel fill deposits of the pre-Pleistocene channel.

The conceptual model of barrier island chain and associated delta plain development for southeast Louisiana and western Mississippi Sound proposed by Otvos and Giardino (2004) constrains the chronology of the Mississippi River St. Bernard delta complex's evolution and of the development and evolution of the Pine Island and MS-AL barriers (Otvos, 2005 a; Otvos and Giardino, 2004), but does not provide a compelling mechanism for the formation of Cat Island. Otvos (1970, 1978) proposed that the MS-AL barriers originated as offshore bar-shoal complexes that aggraded *in situ*, emerging from the seafloor, following the De Beaumont (1845) model favored by Johnson (1919) (Otvos, 1970, 1978). As a pre-requisite for application of the De Beaumont (1845) model to the late Holocene Mississippi Sound environment Otvos (1981, 1985 a, 1985 b, 2005 a), Otvos and Giardino (2004) and Otvos and Carter (2008) stipulate existence of a barrier platform as a precondition for bar-shoal aggradation that lessens the water depth, provides a base with a gentle slope and a source of sediment, continually renourished by longshore transport from the east that allows shoal aggradation to outpace sea level rise (Otvos 1981, 1985 a, 1985 b, 2005 a; Otvos and Carter, 2008; Otvos and Giardino, 2004). Otvos (1981, 1985 a, 1985 b, 2005 a), Otvos and Giardino (2004) and Otvos and Carter (2008) hypothesize that a barrier platform formed as tidal currents transported eroding sediment from the Mobile Bay ebb tidal delta that was deposited around the Pleistocene core of Dauphin Island. Sediment then eroded from Dauphin Island and transported by longshore currents and deposited in an elongate, shore-parallel platform over the floor of the Mississippi Sound (Otvos 1981, 1985 a, 1985 b, 2005 a; Otvos and Carter, 2008; Otvos and Giardino, 2004). Otvos (1981, 1985 a, 1985 b, 2005 a), Otvos and Giardino (2004) and Otvos (2005 a) acknowledge that a significant contribution of sediment to construction of the barrier platform was provided by cross-shore transport of inner shelf sediments (Otvos, 2005 a; Otvos and Giardino, 2004). Otvos and Carter (2008) propose that additional sediment was supplied to Cat Island from an unspecified sandy shoal situated to the east of the island that enabled seaward progradation of the beach ridge strandplain (Otvos and Carter, 2008). The principal limitation of all these hypotheses is that they fail to identify an adequate sediment source of sufficient volume proximal to Cat Island to enable construction of multiple beach ridge sets during the preceding ~3800 years.

Shepard (1960 a) questioned whether sediment derived from longshore transport alone would have provided a sufficient volume of sediment to construct the MS-AL barriers (1960 a). Stapor and Stone (2004) question whether wave climate and shelf morphology in the Gulf would have been able to facilitate a net longshore transport of a sufficient volume of sediment eroded from the incipient Lake Pontchartrain headlands 50 km to the east to the site of the Holocene Pine Island barrier complex during the ~1,300 year period of barrier deposition, as proposed by Otvos (1978) (Stapor and Stone, 2004). Stapor and Stone (2004) estimate that a transport rate of

3 million m^3/yr from a sand mass 45 km long, 3.5 km wide and 4.5 m thick would have been required to construct the Pine Island barrier complex, whereas the highest known sediment transport rates in the present-day Gulf occur along less than 5 km sections of Santa Rosa Island, FL at 125,000 m^3/yr (Stapor and Stone, 2004). A net sediment transport model estimating the parameters necessary for construction of the barrier platform underlying the extent of the MS-AL barriers, or for construction of the barriers themselves has either not been attempted or not been published. The scale of a longshore sediment transport system sufficient to construct Cat Island is surmised to be greater than the Lake Pontchartrain-to-Pine Island barrier complex proposed by Stapor and Stone (2004), therefore necessitating discussion of an alternative mechanism of barrier formation for Cat Island.

Proposed Model for the Development and Evolution of Cat Island

A four-stage conceptual model for the evolution of Cat Island employing ridge engulfment and a transgressed paleochannel depositional framework for barrier aggradation and progradation is proposed by this study (time intervals 1-4 [T_1 to T_4] figure 24). The sequence of geologic events presented in this model incorporates elements of ridge engulfment (Hoyt, 1967, 1970; Swift, 1975) and de Beaumont (1845) emergent bar mechanisms of barrier genesis, but the antecedent geologic setting of Cat Island is a determining factor in establishing a local sediment source necessary for barrier emergence and subsequent island evolution. This model proposes that Cat Island aggraded over a landward migrating depocenter of a backfilling fluvial channel as the rate of sea level rise slowed during the mid to late Holocene, ~7000 to ~5000 ybp. Cat Island aggraded at the site along the paleochannel trunk where mid to late Holocene sea level intersected with the topographic highgrounds constraining the backfilling fluvial channel. Infilling of the fluvial channel was greater at the Cat Island site than the seaward trunk of the fluvial channel because landward migration of the depocenter stalled here as sea level rise slowed, allowing re-worked alluvial sediment deposition to overtop the bounding topographic high-grounds and form the extensive sandy barrier platform that provided sediment for beach ridge construction and barrier progradation and vertical aggradation.

Stage 1(T_0 to T_1)

During early Holocene sea level rise (~7000 - ~5000 ybp; Kindinger, 1989), transgression of the Gulf of Mexico basin forced shoreline retreat and the landward shift of depocenters and mobilized and re-distributed sediment. The fluvial networks of the inner shelf were backfilled with eroded shelf and alluvial sediments (Kindinger, 1989; Rodriguez et al, 2004; Wilkinson, 1975), and the transgressive MAFLA sand sheet was deposited across the inner shelf over the Pleistocene surface (Kindinger, 1989; McBride et al., 2004). As sea level approached the present-day site of Cat Island, re-worked inner shelf and fluvial sediments began to fill the large incised channel situated between the two Pleistocene high-grounds beneath the present-day Cat Island northern and southern nearshore zones. The shoreline began to encroach upon the seaward slope of the southern, gulf-side topographic highground, establishing a shoreface environment. Breaking waves eroded the ridge slope, depositing the eroded sediment in the nearshore zone; an abundant supply of sediment from the adjacent seafloor and the backfilling fluvial channel provided material for cross-shore transport onto the shoreface, enabling maintenance of shoreface equilibrium and depositional progradation during sea level rise (T_0 to T_1, figure 24).

Stage 2 (T_1 to T_2)

As sea level reached the top of the paleochannel levees, positioned at ~10 m bsl, the larger and adjacent smaller incised channels infilled completely. The topographic highgrounds to either side of the paleochannels were still subaerially exposed. Depositional progradation continued on the gulf-side highground shoreface with sediment supplied from offshore, sandy transgressive deposits. Sedimentation occurred in the developing lagoon area landward of the northern, soundside highground, where the sheltered, low-energy lee environment allowed settling of finer-grained material. (T_1 to T_2, figure 24).

Stage 3 ($T_{2\ to}$ T_3)

As sea level approached present-day levels (~5000 ybp) and the rate of sea level rise slowed, the gulfside and soundside topographic highgrounds become completely submerged and the accommodation space between the two ridges is filled with re-worked shelf and fluvial sediments. The recently deposited sediment in trough was held in place between the two ridges instead of being transported and re-distributed landward, creating a compound barrier platform lithosome of mixed Recent and re-worked Pleistocene shelf sediment flanked by Pleistocene highgrounds. Core HC-199 (Brown et al., 1944; figure 10) was likely drilled through the channel fill lithofacies within the trough, accounting for the difficulty of determining the contact between Recent and Pleistocene sedimentary units. (T_2 to T_3, figure 24).

Stage 4 ($T_{3\ to}$ T_4)

With the slowing rate of sea level rise, barrier aggradation and progradation was facilitated by the broad, gently sloping barrier platform, which supplied sediment and created a swash zone environment for construction of inter-tidal bars and subtidal berms. Successive cycles of berm aggradation and welding allowed development of beach ridge sets on the gulfside platform shoreface that aggraded into a subaerial barrier. As the earliest-formed ridges became isolated from the seaward-migrating shoreface environment, eolian transport of sand to the interior ridges allowed vertical accretion of dune deposits over the ridges and vertical aggradation of the subaerial barrier. The highest, and presumably oldest ridge on Cat Island is dated at 3,830 ybp (OXL-1291, Otvos and Giardino, 2004; figure 10), concurrent with the onset of the Mississippi River St. Bernard delta complex progradation. Orientation of the oldest beach ridge strandplains (beach ridge set 1, Rucker and Snowden, 1989; figure 12) on Cat Island indicates that dominant wave approach during formation of the oldest beach ridge set was from the southwest. Progradation of the Mississippi River St. Bernard delta complex from ~3800 to ~2000 ybp (Otvos and Giardino, 2004) blocked wave approach to Cat Island from the southwestern Gulf. The shift in dominant wave approach from the southwest to south of Cat Island during St. Bernard delta complex progradation is reflected by the clockwise angular rotation of beach ridge set orientation with decreasing age (beach ridge sets 2 & 3, Rucker and Snowden, 1989; figure 12). The single ridge on Middle Spit (figure 11) is oriented at a steeper angle relative to the ridges on the main island, indicating that wave angle approach shifted back to the southwest. This would have occurred as the St. Bernard delta complex entered its destructive phase and receded westward, ~2000 to ~1000 ybp. Erosion of the eastern tip of the island and formation of the north-south spit also occurred during this stage, commencing between ~1000 to ~750 ybp (Velardo, 2005). With sea level at its present elevation, the barrier platform has become sequestered beneath the depth of penetration of wave orbitals, and is no longer a source of sediment for beach ridge development or barrier re-nourishment (T_3 to T_4, figure 24).

The Pleistocene geomorphology underlying Cat Island was instrumental in the formation of the island at its current site. At its maximum stage, younger sets of prograded beach ridges likely extended much further gulfward, but have since been eroded. The retrograded clinoforms located in the island's soundside subsurface (figure 23) indicate that the barrier also extended farther into Mississippi Sound than at present. The apparent stability of Cat Island relative to other members of the MS-AL barrier chain (Morton, 2008; McBride et al., 1995) was enabled by a prolonged interval of barrier progradation maintained by an abundant, local sediment supply not present at the other barrier island sites. Sequestration of Cat Island's main source of sediment beneath rising sea level is the primary reason for the island's transition from a constructive to a destructive phase. Initiation of this destructive phase was deferred by the expansion of the St. Bernard delta complex.

Predicted Future Geomorphic Trends for Cat Island

Continued sea-level rise and impacts from intense storms and large-magnitude tropical cyclones pose the greatest threat to Cat Island's survival (Morton, 2008, 2010; Otvos and Carter, 2008). Sea-level rise will increase submergence and erosion of the main island's shoreline and ridges (Morton, 2008; Otvos and Carter, 2008). Sea level rise will also raise the water level of the tidal creeks within the swales enabling expansion of the existing tidal creeks and flooding of the subaerial swales of the central main island, leading to erosion of the beach ridge strandplains from within the island. Persistent wave approach from the southwest, re-established as the Mississippi River St. Bernard delta complex began to recede will continue to erode the island's western margin (Morton, 2008; Otvos and Giardino, 2004; Otvos, 2005 a; Rucker and Snowden, 1989, 1990).

The north-south spit, which impedes westward littoral currents (Otvos and Carter, 2008; Rucker and Snowden, 1989), is depriving the main island shoreface of longshore-transported sediment renourishment, which would likely help extend the life-span of the main island as sea level rise proceeds. The current positions of the north-south spit termini ensure that sediment eroded from the north and south tips of the spit are likely deposited too far offshore or in water too deep to be entrained by nearshore breaking waves to help maintain the main island’s north and south shorefaces. The eastern shoreface of the north-south spit also shelters the main island from erosion by absorbing the impact of westward longshore currents and storm driven and open-marine waves from the southeast (Morton, 2008, 2010; Otvos and Carter, 2008). Subsequently, erosion of Cat Island’s eastern shoreface along the north-south spit has seen the highest shoreline retreat rates on the island during the past ~150 years (Morton, 2008) and erosion of the eastern shoreface is expected to continue in the future. The elevations of the north-south spit dunes and the ridges immediately behind the spit, however, are the highest on the island and prevent storm-driven overwash and spit and barrier breaching (Morton, 2008, 2010). Degradation of the north-south spit will increase vulnerability and exacerbate erosion of the main island.

Summary

Cat Island formed through a confluence of unique, site-specific geologic events involving engulfment of Pleistocene topographic high-grounds that constrained sediment deposited by the landward migrating depocenter of an Oxygen Isotope Stage 3 Phase 1 fluvial channel during mid to late Holocene stillstand. The large volume of sediment that accumulated directly beneath the present-day subaerial barrier enabled barrier progradation during transgression through the construction of successive beach ridge sets. This underlying barrier platform consists of re-worked shelf sediments and alluvium bounded by two Pleistocene topographic highgrounds, overlain by a lithologically similar unit. Sea-level rise sequestered the formerly abundant sediment supply crucial for barrier renourishment, so that the island is currently degrading. The current configuration of Cat Island is the result of the barrier's geomorphic response to the progradation and subsequent recession of the Mississippi River St. Bernard delta complex, which is credited with providing a protective environment for the island and delaying the island's degradation.

Continued rising sea level and storm impacts are expected to drive erosion of Cat Island in the future. Due to the hydrodynamic regime created by the north-south spit, sediment will likely continue to be transported out of the barrier's littoral system. Managers of the island face a paradox: restoration measures such as beach nourishment of the north-south spit's eastern shoreface will preserve the area and height of the north-south spit and help protect the main island from storm driven erosion. This will, however, cause the sediment starvation of the main island to persist, and exacerbate sea level driven erosion.

References

Anderson, J.B., Rodriguez, A., Abdulah, K.C., Fillon, R.H., Banfield, L.A., McKeown, H.A., Wellner, J.S., 2004, Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico margin: A Synthesis, *in* Anderson, J.B. and Fillion, R.H. (eds.), Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico Margin: Society of Economic Paleontologists and Minerologists (SEPM), Special Publication no. 79, pp. 1-23.

Barnhardt, L.B., 2003, Minerologic Study of Sediments from Nearshore Cat Island, Mississippi, Masters Thesis, Mississippi State University, Starkeville, Mississippi, pp. 61.

Boone, P.A., 1973, Depositional Systems of the Alabama, Mississippi and Western Florida Coastal Zone, Gulf Coast Association of Geological Societies, Transactions, vol. 23, p. 266-277.

Boyd, R., Suter, J., Penland, S., 1989, Sequence Stratigraphy of the Mississippi Delta, Gulf Coast Association of Geological Societies, Transactions, vol. 39, p. 331-340.

Brooks, G.R., Kindinger, J.L., Penland, S., Williams, S.J., McBride, R.A., 1995, East Louisiana Continental Shelf Sediments: A Product of Delta Reworking, Journal of Coastal Research, vol. 11, no. 4, p. 1026-1036.

Bruun, P., 1962, Sea Level Rise as a Cause of Shore Erosion, American Society of Civil Engineers, vol. 88, p. 117-130.

Brown, G.F., Foster, V.M., Adams, R.W., Reed, E.W., Padgett, Jr., H.D., 1944, Ground-Water Resources of the Coastal Area in Mississippi, Mississippi State Geological Survey, Bulletin 60, University of Mississippi, University, Mississippi, pp. 232.

Chesapeake Technology, Inc., 2009-2010, http://www.chesapeaketech.com/prod-wizmap.html

Cipriani, L.E. and Stone, G.W., 2001, Net Longshore Transport and Textural Changes in Beach Sediments along the Southwest Alabama and Mississippi Barrier Islands, U.S.A. Journal of Coastal Research, vol. 17, no. 2, p. 443-458.

Corbielle, R.L., 1962, New Orleans Barrier Island, Gulf Coast Association of Geological Societies, Transactions, vol. 12, p. 223-229.

Curray, J.R., 1960, Sediments and History of Holocene Transgression, Continental Shelf, Northwest Gulf of Mexico *in* Shepard, F.P., Phleger, F.B., Van Andel, T.H. (eds.), Recent Sediments, Northwest Gulf of Mexico, American Association of Petroleum Geologists, Special Issue no. 21, p. 221-266.

Curray, J.R. and Moore, 1963, Facies Delineation by Acoustic Reflection: Northern Gulf of Mexico, Sedimentology, vol. 2, p. 130-148.

Dunbar, J.B., Blaes, M.R., Deuitt, S.E., May, J.R., and Stroud, K.W., 1995, Geologic

Investigation of the Mississippi River Deltaic Plain, Vicksburg, Mississippi, U.S. Army Waterways Experiment Station Technical Report GL-84-15 (3 of a series), 36 plates.

EdgeTech, 2010, http://www.edgetech.com/edgetech/gallery/category/sub-bottom-profiling-systems

Field and Duane, 1976, Post-Pleistocene History of the United States Inner Continental Shelf: Significance to Origin of Barrier Islands, Geological Society of America Bulletin, vol. 87, p. 691-702.

Field and Duane, 1977, Post-Pleistocene History of the United States Inner Continental Shelf: Significance to Origin of Barrier Islands: Discussion and Reply, Geological Society of America Bulletin, vol. 88, p. 734-736.

Fisk, H.N., 1947, Geology of the Mississippi Valley Region, Tulsa Geological Society Digest, vol. 15, p. 50-55.

Fisk, H.N., McFarlane Jr., E., Kolb, C.R., Wilbert Jr., L.J., 1954, Sedimentary Framework of the Mississippi Delta, Journal of Sedimentary Petrology, v. 24 no. 2, 76-99.

Flocks, J., Kulp, M., Smith, J., Williams, S.J., 2009, Review of the Geologic History of the Pontchartrain Basin, Northern Gulf of Mexico, *in*, FitzGerald, D. and Reed, D., (eds.), Geological and Environmental Dynamics of the Pontchartrain Basin, Journal of Coastal Research, Special Issue no. 54, p. 12-22.

Foxworth, R.D., Priddy, R.R., Johnson, W.B., Moore, W.S., 1962, Heavy Mineral Sands from Recent Beaches of the gulf Coast of Mississippi and Associated Islands, Mississippi State Geological Survey, Bulletin 93, University of Mississippi, University, Mississippi, pp. 92.

Frazier, D.E., 1967, Recent Deltaic Deposits of the Mississippi River: Their Development and Chronolgy, Gulf Coast Association of Geological Societies Transactions, vol. 17, p. 287-315.

Greene, D.L., Rodriguez, A.B., Anderson, J.B., 2007, Seaward-Branching Coastal-Plain and Piedmont Incised-Valley Systems Through Multiple Sea-Level Cycles: Late Quaternary Examples from Mobile Bay and Mississippi Sound, U.S.A., Journal of Sedimentary Research, vol. 77, p. 139-158.

Hampson, G.J., Rodriguez, A.B., Storms, J.E.A., Johnson, H.D., Meyer, C.T., 2008, Geomorphology and High Resolution Stratigraphy of Progradational Wave-Dominated Shoreline Deposits: Impact on Reservoir-Scale Facies Architecture, *in*, Hampson, G.J., (ed.), Recent Advances in Siliclastic Shallow Marine Stratigraphy, Society of Economic Paleontologists and Minerologists (SEPM), Special Publication no. 90, p. 117-142.

Hayes, M.O., 1979, Barrier Island Morphology as a Function of Tidal and Wave Regime, *in*, Leatherman, S.P. (ed.), Barrier Islands from the Gulf of St. Lawrence to the Gulf of Mexico: Academic Press, New York, New York, pp. 28.

Heinrich, P.V., 2006, Pleistocene and Holocene Fluvial Systems of the Lower Pearl River, Mississippi and Louisiana, U.S.A., Gulf Coast Association of Geological Societies Transactions, vol. 56, p. 267-278.

Hine, A.C., 1979, Mechanisms of Berm Development and Resulting Beach Growth Along a Barrier Spit Complex, Sedimentology, vol. 26, p. 333-351.

Hoyt, J.A., and Henry, Jr., V.J., 1967, Influence of Island Migration on Barrier-Island Sedimentation, Geological Society of America Bulletin, vol. 78, p. 77-86.

Hoyt, J.A., 1967, Barrier Island Formation, Geological Society of America Bulletin, vol. 78, p. 1125-1136.

Hoyt, J.A., 1967, Influence of Island on Barrier-Island Sedimentation, Geological Society of America Bulletin, vol. 78, p. 77-86.

Hoyt, J.H., 1970, Development and Migration of Barrier Islands, Northern Gulf of Mexico: Discussion, Geological Society of America Bulletin, vol. 81, p. 3779-3782.

Hsu, K.J., 1960, Texture and Mineralogy of the Recent Sands of the Gulf Coast, Journal of Sedimentary Petrology, vol. 3, 380-403.

Isphording, W.C., Imsand, F.D., Flowers, G.C., 1989, Physical Characteristics and Aging of Gulf Coast Estuaries, Gulf Coast Association of Geological Societies Transactions, vol. 39, p. 387-401.

Johnson, D.W., 1919, Shore Processes and Shoreline Development, Hafner Publishing Company, New York, New York.

Kindinger, J.L., 1988, Seismic Stratigraphy of the Mississippi-Alabama Shelf and Upper Continental Slope, Marine Geology, vol. 83, p. 79-94.

Kindinger, J.L., 1989, Depositional History of the Lagniappe Delta, Northern Gulf of Mexico, Geo-Marine Letters, vol. 9, no. 2, p. 59-66.

Kindinger, J.L., Penland, S., Williams, S.J., Suter, J.R., 1989, Inner Shelf Deposits of the Louisiana-Mississippi-Alabama Region, Gulf of Mexico, Gulf Coast Association of Geological Societies Transactions, v. 39, p. 413-420.

Kulp, M.A., Howell, P., Adiau, S., Penalnd, S., Kidinger, J., Williams, J.S., 2002, Late Quaternary Stratigraphic Framework of the Mississippi Delta Region, Transactions: Gulf Coast Association of Geological Societies Transactions, vol. 52, p. 573-582.

Kulp, M.A., Penland, S.P., 2004, Chapter 2: Regional Geology of South Louisiana in Technical Report Series 004-002, Louisiana Gulf Coast Restoration Report, Pontchrtrain Institute of

Environmental Science, University of New Orleans.

Kulp, M., Penalnd, S., Williams, S.J., Jenkins, C., Flocks, J., Kindinger, J., 2005, Geologic Framework, Evolution, and Sediment Resources for Restoration of the Louisiana Coastal Zone, *in*, Finkl, C.W., Khalil, S., (eds.), Saving America's Wetlands: Strategies for Restoration of Louisiana's Wetlands and Barrier Islands, Journal of Coastal Research, Special Issue no. 44, p. 56-71.

Kwon, H.J., 1969, Barrier Islands of the Northern Gulf of Mexico: Sediment Source and Development, Louisiana State University Press, Baton Rouge, Louisiana, 45 p.

LeBlanc, L.R., Mayer, L., Rufino, M., Schock, S.G., King, J., Marine Sediment Classification Using the Chirp Sonar, Journal of the Acoustic Society of America, vol. 91, no. 2, p. 107-115.

Leont'yev, O.K. and Nikiforov, L.G., 1965, Reasons for the Worldwide Occurrence of Barrier Beaches, Oceanology, vol. 5, 61-67.

Lopez, J.A., Penland, S., Williams, J., 1997, Confirmation of Active Geological Faults in Lake Pontchartrain in Southeast Louisiana, Gulf Coast Association of Geological Societies Transactions vol. 47, p 299-303.

McBride, R.A., Byrnes, M.R., Hiland, M.W., 1995, Geomorphic Response-type Model for Barrier Coastlines: A Regional Perspective, Marine Geology, vol. 126, p. 143-159.

McBride, R.A., Moslow, T.F., Roberts, H.H., Diecchio, R.J., 2004, Late Quaternary Geology of the Northeastern Gulf of Mexico Shelf: Depositional History, and Ancient Analogues of a Major Shelf Sand Sheet of the Modern Transgressive Systems Tract, *in* Anderson, J.B. and Fillion, R.H. (eds.), Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico Margin, Society of Economic Paleontologists and Minerologists (SEPM), Special Publication no. 79, p. 55-83.

Morton, R.A., Paine, J.G., Blum, M.D., 2000, Responses of Stable Bay-Margin and Barrier-Island Systems to Holocene Sea-Level Highstands, Western Gulf of Mexico, Journal of Sedimentary Research, vol. 70, no. 3, p. 478-490.

Morton, R.A., 2008, Historical changes in the Mississippi-Alabama Barrier Island Chain and the Roles of Extreme Storms, Sea Level, and Human Activities, Journal of Coastal Research, vol. 24, no. 6, p. 1587-1600.

Morton, R.A., 2010, First-Order Controls of Extreme Storm Impacts on the Mississippi-Alabama Barrier Island Chain, Journal of Coastal Research, vol. 26, no. 4, p. 635-648.

Morton, R.A and Rodgers, B.E., 2010, Geomorphology and Depositional Sub-Environments of Gulf Islands National Seashore, Mississippi, USGS Open File Report 2009-1250, http://pubs.usgs.gov/of/2009/1250/

Morton, R.A. and Suter, J.R., 1996, Sequence Stratigraphy and Composition of Late Quaternary Shelf-Margin Deltas, Northern Gulf of Mexico, American Association of Petroleum Geologists Bulletin, vol. 80, no. 4, p. 505-530.

Knabb, R.D., Rhome, J.R., Brown, D.P., 2006, Tropical Cyclone Report, Hurricane Katrina, National Hurricane Center, http://www.nhc.noaa.gov/pdf/TCR-AL122005_Katrina.pdf

National Park Service, 2009, http://www.nps.gov/guis

Oivanki, S.M., 1998, Geogolgy and Geomorphology of the Mississippi Gulf Coast in Marine Resources and History of the Mississippi Gulf Coast, Volume 2, Klein, L.A., Landry, M., Seward, J.E., (ed.), Mississippi Department of Marine Resources, 271 p.

Otvos, E.G., 1970, Development and Migration of Barrier Islands, Northern Gulf of Mexico, Geological Society of America Bulletin, vol. 81, p. 241-246.

Otvos, E.G., 1977, Post-Pleistocene History of the United States Inner Continental Shelf: Significance to Origin of Barrier Islands: Discussion and Reply, Geological Society of America Bulletin, v. 88, p. 734-736.

Otvos, E.G., 1978, New Orleans-South Hancock Holocene Barrier Trends and Origins of Lake Pontchartrain, Gulf Coast Association of Geological Societies Transactions, vol. 28, p. 337-355.

Otvos, E.G., 1981, Barrier island formation through nearshore aggradation-stratigraphic and field evidence, Marine Geology, vol. 43, 1981, p. 195-243.

Otvos, E.G., 1985 a, Barrier Platforms: Northern Gulf of Mexico, Marine Geology, v. 63, 1985, p. 285-305.

Otvos, E.G., 1985 b, A New Stratigraphic System, Geological Evolution and Potential Economic Sand Resources in the Mississippi Sound Area, Mississippi-Alabama, Open-File Report 85-6F, Mississippi Mineral Resources Institute, University of Mississippi, University, Mississippi, 71 p.

Otvos, E.G., 1986 a, Island Evolution and "Stepwise Retreat": Late Holocene Transgressive Barriers, Mississippi Delta Coast, Marine Geology, vol. 72, p. 325-340.

Otvos, E.G., 1986 b, Stratigraphy and Potential Economic Sand Resources of the Mississippi-Alabama Barrier island System and Adjacent Offshore Areas, Open-File Report 86-1F, Mississippi Minerals Resource Institute, University of Mississippi, University, Mississipi, 67 p.

Otvos, E.G., 2000, Beach Ridges- Definitions and Significance, Geomorphology, vol. 32, p. 83-108.

Otvos, E.G., 2005 a, Coastal Barriers, Gulf of Mexico: Holocene Evolution and Chronology, Journal of Coastal Research, *in*, Finkl, C.W., The Sun, Earth and Moon: a Tribute to Rhodes W. Fairbridge, Special Issue no. 42, p. 141-163.

Otvos, E.G., 2005 b, Numerical Chronology of Pleistocene Coastal Plain and Valley Development: Extensive Aggradation During Glacial Low Sea Levels, Quaternary International, v. 135, p. 91-113.

Otvos and Carter, 2008, Hurricane Degradation- Barrier Development Cycles, Northeastern Gulf of Mexico: Landform Evolution and Island Chain History, Journal of Coastal Research, vol. 24, no. 2, p. 463-478.

Otvos, E.G. and Giardano, M.J., 2004, Interlinked Barrier Chain and Delta Lobe Development, Northern Gulf of Mexico, Sedimentary Geology, vol. 169, p. 47-73.

Penland, S. and Suter, J.R., 1983. Transgressive Coastal Facies Preserved in Barrier Island Arc Retreat Paths in the Mississippi River Delta Plain, Gulf Coast Association of Geological Societies Transactions, v. 33, p. 367-381.

Penland, S. and Boyd, R., 1981, Shoreline Changes on the Louisiana Barrier Coast. Institute of Electrical and Electronic Engineers, Oceans, vol. 81, p. 209-219.

Penland, S., Suter, J.R., McBride, R.A., Williams, S.J., Kindinger, J.L., Boyd, R., 1989, Holocene Sand Shoals off the Mississippi River Delta Plain, Gulf Coast Association of Geological Societies Transactions, vol. 39, p. 471-480.

Roberts, H.H., 1997, Dynamic Changes of the Holocene Mississippi River Delta Plain: The Delta Cycle, Journal of Coastal Research, vol. 13, no. 3, p. 605-627.

Rodriguez, A.B., Anderson, J.B., Siringan, F.P., Taviani, M., 2004, Holocene Evolution of the East Texas Coast and Inner Continental Shelf: Along Strike Variability in Coastal Retreat Rates, Journal of Sedimentary Research, vol. 74, no. 3, p. 405-421.

Rodriguez, A.B., Meyer, C.T., 2006, Sea-Level Variation During the Holocene Deduced from the Morphological and Stratigraphic Evolution of Morgan Peninsula, Alabama, U.S.A., Journal of Sedimentary Research, v. 76, p. 257-269.

Rosati, J.D., Dean, R.G., Stone, G.W., 2010, A Cross-Shore Model of Barrier Island Migration Over a Compressible Substrate, Marine Geology, vol. 271, p. 1-16.

Rucker and Snowden, 1988, Recent Morphologic Changes at Dog Keys Pass, Mississippi: the Formation and Disappearance of the Isle of Caprice, Gulf Coast Association of Geological Societies, vol. 38, p. 343-349.

Rucker, J.B. and Snowden, J.O., 1989, Relict Progradational Beach Ridge Complex on Cat Island in Mississippi Sound, Gulf Coast Association of Geological Societies Transactions, vol. 39, p. 531-539.

Rucker, J.B. and Snowden, J.O., 1990, Barrier Island Evolution and Reworking by Inlet

Migration along the Mississippi-Alabama Gulf Coast, Gulf Coast Association of Geological Societies Transactions, vol. 50, p 745-753.

Saucier, R.T., 1963, Recent Geomorphic History of the Pontchartrain Basin, Louisiana State University Coastal Studies Series no. 9, Baton Rouge, Louisiana, pp. 114.

Schmidt, K., 2001, Cat Island Evolution, Morphology, and Hurricane Response- 1995 to 2000, Mississipp Department of Environmental Quality Office of Geology, Open File Report 132, pp. 29.

Schwartz, M.L., 1971, Multiple Causality of Barrier Islands, Journal of Geology, vol. 79 no. 1, p 91-94.

Scruton, 1960, Delta Building and the Deltaic Sequence, *in*, Shepard, F.P., Phleger, F.B., Van Andel, T.H. (eds.), Recent Sediments of the Northwestern Gulf of Mexico, American Association of Petroleum Geologists, Special Issue no. 21, p. 82-102.

Shepard, F.P., 1960 a, Gulf Coast Barriers *in*, Shepard, F.P., Phleger, F.B., Van Andel, T.H. (eds.), Recent Sediments of the Northwestern Gulf of Mexico, American Association of Petroleum Geologists, Special Issue no. 21, p. 197-219.

Shepard, F.P., 1960 b, Rise of Sea Level Along Northwest Gulf of Mexico, *in*, Shepard, F.P., Phleger, F.B., Van Andel, T.H. (eds.), Recent Sediments of the Northwestern Gulf of Mexico, American Association of Petroleum Geologists, Special Issue no. 21, p. 338-344.

Stapor and Stone, 2004, A New Depositional Model for the Buried 4000 yr BP New Orleans Barrier: Implications for Sea Level Fluctuations and Onshore Transport from a Nearby Shelf Source, Marine Geology, vol. 204, p. 215-234.

Stanley, D.J., 1997, Holocene Sea Level Change and Early Human Utilization of Deltas, GSA Today, vol. 7, no. 2, p. 1-6.

Swift, D.J.P., 1975, Barrier Island Genesis: Evidence from the Central Atlantic Shelf, Eastern U.S.A., Sedimentary Geology, vol. 14, p. 1-43.

Stutz, M.L. and Pilkey, O.H., 2002, Global Distribution and Morphology of Deltaic Barrier Island Systems, Journal of Coastal Research, Special Issue 36, p 694-707.

Tanner, W.F., 1990, Origin of Barrier Islands on Sandy Coasts. Gulf Coast Association of Geological Societies Transactions, vol. 50, p. 819-373.

Tanner, W.F., 1995, Origin of Beach Ridges and Swales, Marine Geology, vol. 129, p. 149-161.

Taylor, M. and Stone, G.W., 1996, Beach Ridges: A Review, Journal of Coastal Research, vol. 12, no. 3, p. 612-621.

Timmons, E.A., Harden, E.L., Freeman, C., Rudolph, G., Rodriguez, A.B., 2008, Underlying Paleochannels Facilitate Barrier Progradation: Bogue Banks, North Carolina. Abstract from the 2008 Joint Annual Meeting of the Geological Society of America, Soil Science Society of America, American Society of Agronomy, Crop Science Society of America, Gulf Coast Asociation of Geological Societies and the Houston Geological Society, http://a-c-s.confex.com/crops/2008am/webprogram/Paper50304.html

Treadwell, R.C., 1955, Sedimentology and Ecology in Southeast Coastal Louisiana, Louisiana State University Technical Report 6, Baton Rouge, Louisiana, 78 p.

The Trust for Public Land, 2002, www.tpl.org/tier3_cd.cfm?content_item_id=7889&folder_id=671

U.S. Army Corps of Engineers, Mobile District, 2009, Mississippi Coastal Improvements Program (MsCIP), Hancock, Harrison, and Jackson Counties, Mississippi, Comprehensive Plan and Integrated Programmatic Environmental Impact Statement, Volume 1- Main Report: http://www.sam.usace.army.mil/mscip/MsCIP_Comprehensive_Plan.htm

Velardo, B., 2001, Detailed Geochronology of the Mississippi Sound During the Late Holocene. Masters Thesis, Louisiana State University and Agricultural and Mechanical College, Baton Rouge, Louisiana, 90 p.

Van Andel, 1960, Sources and Dispersion of Holocene Sediments, Northern Gulf of Mexico. *in*, Shepard, F.P., Phleger, F.B., Van Andel, T.H. (eds.), Recent Sediments of the Northwestern Gulf of Mexico, American Association of Petroleum Geologists, Special Issue no. 21, p. 34-55.

Wilkinson, B.H., 1975, Matagorda Island, Texas: The Evolution of a Gulf Coast Barrier Complex. Geological Society of America Bulletin, vol. 86, p. 959-967.

Appendix

Table 1. Survey line numbers with corresponding seismic file numbers

Survey Line #	SB-216S File #	SB-424 File #
N-S LINES		
1		09c692.sgy, 09c640.sgy, 09c662.sgy
2		09c639.sgy
3		09c630.sgy, 09c637.sgy
4		09c664.jsf, 09c636.sgy
5		09c634.sgy
6		09c686.sgy, 09c632.sgy
7	7.jsf, 7.001.jsf, 7.002.jsf, 8.5B.sgy	09c682.sgy, 09c656.sgy
8	8.jsf, 8.001.jsf, 8.5B.001.sgy, 8.5B.002.sgy, 8.5B.003.sgy	
9	9.jsf, 9.001.jsf	09c680.sgy
10	10.001.jsf, 10.002.jsf	
11	11.jsf, 11.001.jsf, 11B.sgy, 11B.001.sgy, 11B.002.sgy	09c678.sgy, 09c652.sgy
12	12.jsf, 12.001.jsf, 12B.sgy, 12B.001.sgy	
13	13.jsf, 13.001.jsf	09c676.001.sgy, 09c651.sgy
14	14.001.jsf, 14.004.jsf	
15		09c674.sgy, 09c649.sgy
16	16B.sgy, 16B.001.sgy, 16B.002.sgy, 16B.003.sgy, 16B.004.sgy, 16B.005.sgy	
17	17.jsf, 17.001.jsf	09c672.sgy, 09c647.sgy
18	18.jsf, 18.001.jsf	
19	19_A.001.jsf, 19_A.002.jsf, 19_A.003.jsf, 19_A.jsf	09c684.sgy, 09c645.sgy
21	test.001.sgy, test.002.sgy, test.003.sgy, test.004.sgy, test.005.sgy, test.006.sgy, test.007.sgy	09c643.sgy, 09c660.sgy
E-W LINES		
110		09c629.sgy
111		09c687.sgy
112	11.002.jsf, 13.001.jsf, 19A.004.jsf	09c677.sgy, 09c681.sgy
113		09c685.sgy, 09c673.sgy, 09c638.sgy
114		09c683.sgy, 09c689.sgy
115		09c691.sgy
116		09c642.sgy, 09c641.sgy, 09c693.sgy, 09c646.sgy, 09c650.sgy, 09c635.sgy
117	171.sgy, 171.001.sgy, 171.002.sgy, 171.003.sgy, 171.004.sgy, 171.005.sgy, 171.006.sgy, 171.007.sgy, 171.008.sgy, 171.009.sgy, 171.010.sgy, 171.011.sgy, 171.012.sgy, 171.013.sgy, 171.014.sgy, 171.015.sgy, 171.016.sgy, 171.017.sgy	09c695.sgy, 09c658.sgy, 09c659a.sgy
117.5	171.5.sgy, 171.5.001.sgy	09c659.sgy
118	181.001.sgy, 181.002.sgy, 181.003.sgy, 181.004.sgy, 181.005.sgy, 181.006.sgy, 181.007.sgy, 181.008.sgy, 181.009.sgy, 181.010.sgy, 181.011.sgy, 181.012.sgy, 181.013.sgy, 181.014.sgy, 181.015.sgy	09c661.sgy
118.5	181.5.jsf, 181.5.001.jsf, 181.5.002.jsf, 181.5.003.jsf	

(table 1. continued)

Survey Line #	SB-216S File #	SB-424 File #
119	191.jsf, 191.001.jsf, 191.002.jsf, 191.003.jsf, 191.004.jsf, 191.005.jsf	09c644.sgy, 09c648.sgy, 09c663.sgy
OBLIQUE LINES		
100		09c627.sgy, 09c631.sgy, 09c690.sgy, 09c694.001.sgy
101		09c628.sgy
102		09c665.sgy
103	19.jsf, 12.002.jsf, 10.003.jsf,	09c666.sgy, 09c671.sgy, 09c675.sgy, 09c679.sgy
DREDGED CANAL LINES		
104	00000000.003.jsf, 00000000.004.jsf	
105	00000000.001.jsf, 00000000.002.jsf	

Table 2. Characteristics of sediments recovered from well HC-199 (from Brown, et al., 1944)

Harrison County Well 199
Owner: U.S. Army
Driller: Layne Central Company
Drill Date: 1943
Location: "200 ft N, 200 ft E of north end of pier at mouth Little Bay, Cat Is."
Well diameter: 6 inches
Well depth: 971 ft (296 m)
Altitude: +/- 10 ft (3 m)
USACE "Distribution of Deltaic and Marine Deposits" interprets Pleistocene surface at -3.4 m

Facies Thickness	Facies Depth	Geologic Interpretation	Minerology	Fauna
86 ft (26 m)	86 ft (26 m)	Recent and Pamlico sands Overlie the Prairie Formation Pamlico sand overlain with Recent beach deposits, difficult to define contact	Sand: magnetite, kyanite, clear zircon, staurolite, rutile, tourmaline, colored zircon, trace of a blue-green mineral; central interval- trace of epidote; lower interval- traces of horneblende and garnet; Feldspars uncommon.	Bolivinia sp., Cibicides concentricus, Elphidium gunteri var. galvestonense, Nonion depressula var. matagordana, Quinqueloculina sp., Rotalia becarii var. tepida
42 ft (12.8 m)	128 ft (39 m)	Citronelle Formation Downwarped west of southeast trending structural flexure that extends from Bogalusa, LA to Bay St. Louis and the western end of Ship Island caused by deltaic loading from Mississippi River	Sand and thin clay strata: magnetite, kyanite, zircon, staurolite, rutile, tourmaline, traces of hornblende and epidote from 93 to 117 ft, no feldspar	Bolivinia sp., Buliminella curta, Buliminella elegantissima, Cibicides americanus, Cibicides concentricus, Elphidium gunteri var. galvestonense, Nonion depressula var. matagordana, Quinqueloculina cf. lamarckiana, Quinqueloculina sp., Rotalia becarii var. tepida, Textutularia mayori, Virgulina sp.
88 ft (26.8 m)	216 ft (65.8 m)	Citronelle Formation (continued)	Sand; magnetite, kyanite, zircon, staurolite, rutile, tourmaline; pyrite and opaque light gray grains at 117 to 136 ft; traces of hornblende, a lavender mineral, ceylonite (?), pyrite at 136 to 160 ft; trace of garnet at 160 to 192 ft; lower 44 ft contain light colored minerals; no feldspar	Bolivinia sp., Buliminella curta, Buliminella elegantissima, Cibicides americanus, Cibicides concentricus, Cibicides cf. pseudoungerianus, Elphidium Elphidium gunteri var. galvestonense, Nonion depressula var. matagordana, Quinqueloculina sp. Virgulina punctata
30 ft (9.1 m)	971 ft (296 m)	Graham Ferry Formation Series of deltaic sediments above the Pascagoula and below the Citronelle Fm. Disconformable contact with the Citronelle Fm. Heterogeneous sediments, continental and marine beds		

Table 3. Characteristics of sediments recovered from well HC-200 (from Brown, et al., 1944)

<table>
<tr><td colspan="4">Harrison County Well 200
Owner: U.S. Army (old artesian well)
Driller: Sutter Well Works
Drill Date: 1929
Location: “At head of Spit Cove, on Cat Island”
Well diameter: 3inches
Well depth: 530 ft (161.5 m)
Altitude: none given
USACE “Distribution of Deltaic and Marine Deposits” interprets Pleistocene surface at -9.1 m</td></tr>
<tr><th>Facies Thickness</th><th>Facies Depth</th><th>Geologic Interpretation</th><th>Description</th></tr>
<tr><td>6 ft
(1.8 m)</td><td>6 ft
(1.8 m)</td><td rowspan="2">Recent deposits</td><td>Sand, white</td></tr>
<tr><td>19 ft
(5.8 m)</td><td>25 ft
(7.6 m)</td><td>Marsh mud or blue clay</td></tr>
<tr><td>65 ft
(19.8 m)</td><td>90 ft
(27.5 m)</td><td>Pamlico sand</td><td>Sand, fine gray</td></tr>
<tr><td>55 ft
(16.7 m)</td><td>145 ft
(44 m)</td><td rowspan="4">Citronelle Formation</td><td>Clay, soft blue</td></tr>
<tr><td>70 ft
(21.3 m)</td><td>215 ft
(65.5 m)</td><td>Sand, coarse and fine gravel</td></tr>
<tr><td>45 ft
(13.7 m)</td><td>260 ft
(79.2 m)</td><td>Clay, soft blue</td></tr>
<tr><td>30 ft
(9 m)</td><td>290 ft
(6 m)</td><td>Sand, coarse and gravel</td></tr>
<tr><td>180 ft
(55 m)</td><td>470 ft
(143.2 m)</td><td rowspan="2">Graham Ferry Formation</td><td>Clay, mixed blue and fine gray sand</td></tr>
<tr><td>60 ft
(18.2 m)</td><td>530 ft
(161.5 m)</td><td>Green sand and gravel, water-bearing</td></tr>
</table>

Table 4. Gross descriptions of MMRI Vibracore Series A and C (Cat Island, MS vicinity)

Core Number	Depth to seafloor	Unit Depth	Unit Thickness	Facies Description
A5/23/1	-14.0 ft (4.25 m)	-14.0 ft to -25.8 ft (-4.25 m to -7.86 m)	3.6 m	Very brackish, very fine and fine sandy clays; clays and clayey fine sand
A5/23/3	-10.0 ft (-3.1 m)	-10.0 ft to -20.0 ft (-3.1 m to -6.1 m)	3.0 m	Very brackish, muddy fine sand, fine sandy mud, clay
		-20.0 ft to -23.3 ft (-6.1 m to -7.1 m)	1.0 m	Moderately brackish, clayey fine sand
		-23.3 ft to -27.4 ft (-7.1 m to -8.4 m)	1.3 m	Very brackish, clay, fine sandy mud
		Sand Units (at least 80% sand)		
		-15.5 ft to -17.3 ft (-4.7 m to 5.3 m)	1.8 m	Fine sand, poorly sorted
		-21.5 ft to -23.3 ft (6.6 m to -7.1 m)	0.5 m	Clayey fine sand, poorly sorted
A5/23/4	-10.1 ft (-3.1 m)	-10.1 ft to -15.5 ft (-3.1 m to -4.7 m)	1.6 m	Very brackish, muddy fine sands, fine sandy mud; clay and mud
		Shell content		
		-10.1 ft to -12.5 ft (-3.1m to -3.8 m)	0.7 m	46% shell content (Crassostrea virginica)
		-13.6 ft to -15.5 ft (-4.1 m to -4.7 m)	0.6 m	42 % shell content (Crassostrea virginica)
A5/23/6	-11.0 ft (-3.4 m)	-11.0 ft to -12.0 ft (-3.4 m to -3.7 m)	0.3 m	Very brackish, muddy medium sand
		-12.0 ft to -19.0 ft (-3.7 m to -5.8 m)	2.1 m	Moderately brackish, fine sand, muddy fine sand, very fine sandy mud
		-19.0 ft to -27.0 ft (-5.8 m to -8.2 m)	2.4 m	Very brackish, clay
		-27.0 ft to -31.0 ft (-8.2 m to -9.4 m)	1.2 m	Moderately brackish, clay
		Sand Units (at least 80% sand)		
		-11.0 ft to -13.5 ft (-3.4 m to -4.1 m)	0.7 m	Muddy medium sand, fine sand, poorly sorted
A5/23/7	11.0 ft (-3.4 m)	-11.0 ft to -14.8 ft (-3.4 m to -4.5 m)	1.1 m	Brackish, very fine sandy mud
		-14.8 ft to -18.7 ft (-4.5 m to -5.7 m)	1.2 m	Marine, muddy fine sand

(table 4 continued)

C1	-12.0 ft (-3.7 m)	-12.0 ft to -13.5 ft (-3.7 m to -4.1 m)	0.4 m	Biotype- too few to identify, medium sand
		-13.5 ft to -14.0 ft (-4.1 m to -4.3 m)	0.2 m	Brackish, fine sand
		-14.0 ft to -20.0 ft (-4.3 m to -6.1 m)	1.8 m	
		Sand Units		
		-12.0 ft to -20.0 ft (-3.7 m to -6.1 m)	5.4 m	Medium to fine sand, well to poorly sorted

Table 5. Gross descriptions of GRCL drill cores (Cat Island, MS vicinity)

GCRL Drillcore- Holocene Sediments				
CE	-4.5 ft (-1.4 m)	-4.5 ft to -6.0 ft (-1.4 m to -1.8 m)	0.4 m	Very brackish, fine sand
		-6.0 ft to -19.5 ft (-1.8 m to -6.0 m)	4.2 m	Moderately brackish, fine sand
		-19.5 ft to -46.0 ft (-6.0 m to -14.0 m)	8.0 m	Marine, medium sand, silty and muddy very fine sand
OC	-4.5 ft (-1.4 m)	-4.5 ft to -6.0 ft (-1.4 m to -1.8 m)	0.4 m	Very brackish, fine sand
		-6.0 ft to -17 ft (-1.8 m to -5.1 m)	3.3 m	Brackish, fine sand
		-17.0 ft to -42.0 ft (-5.1 m to -12.8 m)	7.7 m	Marine, fine sand
		-42.0 ft to -45.0 ft (-12.8 m to -13.7m)	0.9 m	Very brackish, muddy and silty fine sand

Vita

The author was born in London, England and grew up in Singapore, Virginia and Michigan. As an undergraduate student, she attended Virginia Commonwealth University in the Biology Department and completed a Bachelor of General Studies degree at the University of New Orleans in 2007. She was accepted to the Earth and Environmental Studies Masters Program at University of New Orleans in 2008 and worked in the U.S. Geological Survey Gulf Coast Science Coordination Office while completing her studies.