Holocene Formation and Evolution of Horn Island, Mississippi, USA

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*University of Southern Mississippi*

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HOLOCENE FORMATION AND EVOLUTION OF HORN ISLAND, MISSISSIPPI, USA

by

Nina Sabine Margarete Gal

A Thesis
Submitted to the Graduate School, the College of Arts and Sciences and the School of Ocean Science and Engineering at The University of Southern Mississippi in Partial Fulfillment of the Requirements for the Degree of Master of Science

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ABSTRACT

Horn Island, one of the most stable barriers along the Mississippi-Alabama chain, provides critical habitat, helps regulate estuarine conditions in the Mississippi Sound, and reduces wave energy and storm surge for the mainland. This study integrates 2,200 km of high-resolution geophysics, 35 sediment cores, and 15 radiocarbon ages to better understand the formation and evolution of the island in response to sea-level rise, storms, and antecedent geology. The Biloxi and Pascagoula incised valleys converge at Horn Island and have played a profound role in the evolution of the system. Within the incised valleys, numerous shallow paleochannels between 4 and 9 meters below sea level exist only on the landward side of the island, indicating seaward transgressive and tidal ravinement. Sand released due to ravinement processes thus contributed to the formation of Horn Island. Based on radiocarbon ages, an ancestral island existed 8,000 years BP that was ephemeral, frequently overwashed, and unable to build a sandy shoreface. This time period occurs during known rapid rates of relative sea-level rise of 4 mm/yr. Approximately 4,500 years BP coinciding with a deceleration in sea-level rise to about 1 mm/yr, radiocarbon ages associated with Horn Island’s barrier complex and lower shoreface indicate progradation and island formation, in addition to lateral migration in a westward direction that continues to present day. Past sensitivity to rates of sea-level rise coupled with an exhausted sediment supply make the future of Horn Island uncertain.
ACKNOWLEDGMENTS

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Numerous scientific discussions with my lab mates Robert Hollis, Clayton Dike and Shara Gremillion have been invaluable in understanding and interpreting data. Their assistance during fieldwork made data collection possible. I would also like to acknowledge Eve Eisemann, who taught me how to analyze a LiDAR dataset.

James Flocks and Kyle Kelso from the USGS in St. Petersburg kindly allowed me to analyze sediment cores they collected in their laboratory, the results of which have led to important research results that represent a large part of my thesis.
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<table>
<thead>
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<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td><strong>USM</strong></td>
<td>The University of Southern Mississippi</td>
</tr>
<tr>
<td><strong>MIS</strong></td>
<td>Marine Isotope Stage</td>
</tr>
<tr>
<td><strong>BP</strong></td>
<td>before present</td>
</tr>
<tr>
<td><strong>mm/yr</strong></td>
<td>millimeter per year</td>
</tr>
<tr>
<td><strong>MS-AL</strong></td>
<td>Mississippi-Alabama</td>
</tr>
<tr>
<td><strong>GPS</strong></td>
<td>global positioning system</td>
</tr>
<tr>
<td><strong>LiDAR</strong></td>
<td>Light Detection and Ranging</td>
</tr>
<tr>
<td><strong>NOAA</strong></td>
<td>National Oceanic and Atmospheric Administration</td>
</tr>
<tr>
<td><strong>USGS</strong></td>
<td>United States Geological Survey</td>
</tr>
<tr>
<td><strong>LLC</strong></td>
<td>Limited liability company</td>
</tr>
<tr>
<td><strong>SB</strong></td>
<td>sub-bottom</td>
</tr>
<tr>
<td><strong>kHz</strong></td>
<td>kilohertz</td>
</tr>
<tr>
<td><strong>ms</strong></td>
<td>milliseconds</td>
</tr>
<tr>
<td><strong>Z’</strong></td>
<td>Depth</td>
</tr>
<tr>
<td><strong>NAVD88</strong></td>
<td>North American Vertical Datum of 1988</td>
</tr>
<tr>
<td><strong>USACE</strong></td>
<td>United States Army Corps of Engineers</td>
</tr>
<tr>
<td><strong>NCMP</strong></td>
<td>National Coastal Mapping Project</td>
</tr>
<tr>
<td><strong>μm</strong></td>
<td>micrometer</td>
</tr>
<tr>
<td><strong>mbsl</strong></td>
<td>meters below sea level</td>
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1. INTRODUCTION

Barrier islands occur and are abundant on all continents of the world except for Antarctica, and are geologically young, often forming during the late Holocene (Davis, 2014). Covering 15% of the planet’s coastlines, barrier islands not only reduce ocean energy but their bays also provide highly productive ecosystems (Davis & Fitzgerald, 2010). Barrier islands cover approximately half of the United States coastline, are almost uninterrupted along the East and Gulf Coasts and are also common around the world, especially in arctic Alaska, Holland, west-central Africa, southern Brazil, and southeastern Australia (Shepard, 1973). Complex barrier island dynamics are largely controlled by storm impacts, sediment supply variations, antecedent geology, and relative sea-level rise. These variables influence barrier island stability and geomorphology (Otvos & Carter, 2013). Accelerated sea level rise alone has produced deleterious impacts to barrier islands by disrupting the balance between sediment supply and the creation of accommodation space (Bird, 1985, 1996). Coupled with a likely increase in the frequency of intense hurricanes (Emanuel, 2013), shorelines and coastal features have largely become unstable and are eroding (Woodruff et al., 2013). Humans have also significantly decreased sediment delivery in many systems around the world (Syvitski et al., 2005). As a result of these issues, many barriers on all continents are eroding at unprecedented rates.

Because detailed historic records of barrier island morphodynamics are short and fragmentary, studies must utilize the geologic record. Many of these studies have sought to quantify the response of coastal systems to these coastal change forcing mechanisms, in particular as it relates to vulnerable, eroding systems. A common response to the
combination of sea level rise and storm impacts for barrier islands is rollover, which can diminish the islands’ sediment volume and ability to act as storm buffers for the mainland (Penland & Ramsey, 1990; Odezulu et al., 2018). Overwash and island breaching are becoming more frequent as ocean waters heat up in areas where tropical storms are prevalent (Webster et al., 2005; FitzGerald et al., 2008), which causes a weakening in island resistance to erosion and often retrogradation. Some barriers around the world are relatively stable and even progradational, as is the case for drumstick barrier islands in west-central Florida (Davis Jr. et al., 2003), beach ridges in Guichen Bay, Australia (Murray-Wallace et al., 2002), and central and upper Texas barriers (Rodriguez et al., 2001; Anderson et al., 2014). This is a rare occurrence due to storm impacts and accelerated rates of sea level rise, which is causing increased erosion (Wallace & Anderson, 2013). Understanding long-term barrier formation and evolutionary processes leading to stability is critical towards predicting future responses to accelerated relative sea-level rise, sediment supply variations, and climate change. Yet, there is a paucity of studies worldwide that can shed light on the issue.

The Mississippi/Alabama (MS-AL) barrier island chain along the northern Gulf of Mexico is a highly vulnerable section of United States coastline (Pendleton et al., 2010). As one of the costliest and deadliest storms in the United States, Hurricane Katrina in 2005 inundated all of the Mississippi barrier islands, and the recorded storm surge on many of the islands exceeded 6 meters (Fritz et al., 2007) causing severe erosion (Eisemann et al., 2018). Even before Katrina’s impact, these islands had been demonstrated to be extremely dynamic and largely erosional. The lone exception is the centrally located Horn Island, which remains among the most stable in the chain (Morton,
2008). The barriers flanking Horn Island are eroding at a faster rate than Horn Island (Byrnes et al., 2013), but the reason for this occurrence still needs to be resolved. One feature that is important in keeping Horn Island stable is its maritime forest that helps the sediments stay fixed in place. The entire chain acts as an incredibly valuable natural laboratory, as the protected status as national seashores ensures they are almost entirely natural. This study targets Horn Island for two primary reasons. First, a large amount of previously collected geophysical and sediment core data exist for the system, but this information remains to be holistically integrated. Supplementing this prior information with new data allow for an unparalleled opportunity to understand a large barrier system. Coastal resource management aimed at protecting citizens living along the coast is critical, and science should inform public policy and management decisions (Dolan & Wallace, 2012). Secondly, Horn Island would be a valuable addition to the global database of barrier island studies as it is stable, yet surrounded by erosional islands responding to barrier change forcing mechanisms. Thus, it could provide an important contribution towards understanding stable barrier systems in the context of general global transgression.
2. REGIONAL SETTING

2.1 Coastal and physical setting

As part of the Mississippi-Alabama barrier islands, Horn Island is about 19 km long and on average 12 km seaward of the mainland. From west to east, the islands include Cat, Ship, Horn, Petit Bois and Dauphin, respectively (Figure 1). Tidal inlets flank either side of Horn Island, Dog Keys Pass to the west and Horn Island Pass to the east. Horn Island Pass incorporates a dredged shipping channel that disrupts natural sediment movement to the west. Dog Keys Pass is a natural, pristine tidal inlet that contains two deep tidal channels, which have been deepening since 1847 (Buster & Morton, 2011). In close proximity to the Mississippi-Alabama barrier island chain are the Chandeleur Islands chain in Louisiana that represent reworked Mississippi River delta lobes, which do not exhibit strandplain surfaces as the MS-AL barrier islands (Otvos, 1981). Otvos & Giardino (2004) hypothesize that Horn Island’s formation started at Dauphin Island, which they concluded to be the oldest island of the barrier chain, that proceeded to transmit sand westward about 4,500 years ago, forming a large, sandy shoal platform that the other barrier islands used as a foundation to build on. This is largely based on a single existing radiocarbon age taken by (Otvos, 1981) from a shell found at the transition from mud to sandy mud at 11.1 m depth, suggesting that Horn Island formed 4,857 years BP.
A strong westward longshore current restricts the flow of sediments coming eastward from the Mississippi River Delta (Davies & Moore, 1970). The loop current is able to bring winnowed clays from the Mississippi River towards the eastern shelf where Horn Island is located (Doyle & Sparks, 1980). Wind-driven currents drive the circulation of the northern Gulf of Mexico shelf, which dictate a westward-driven plume at the Mississippi River delta (Schiller et al., 2011) and general east to west longshore transport occurs along the MS-AL barrier chain. During cold fronts, currents tend to be reversed, creating significant transport of freshwater to the MS-AL barrier islands (Walker et al., 2005). More southerly winds in the summer can also reverse the general western circulation and bring freshwater to the east (Morey et al., 2003).

Morton (2008) utilized historical U.S. Coast and Geodetic Survey Topographic Sheets, aerial photographs, the global positioning system (GPS) and LiDAR surveys to reveal that Horn Island has been laterally accreting for the past 150 years, moving
westward. The subaerial portion of the barrier’s characteristic geomorphic turn likely formed at the beginning of the 20th century, making the rapidly changing strike of the island a recent geomorphological feature (Morton, 2008). Horn Island has experienced a significantly higher relative resilience to erosion from increased storms and sea level rise compared to other islands in the MS-AL barrier island chain for the past 100 to 150 years (Morton, 2008; Byrnes et al., 2013). However, the reason remains largely unknown.

2.2 Sea level rise

Horn Island shares an age with many other barrier islands that formed along the U.S. Atlantic and Gulf coasts in the mid-Holocene (Anderson et al., 2016), making its response to accelerated sea level rise, storm impacts, and sediment supply variations an important possible connection with other barriers. Indeed, the oldest barriers that currently exist globally are from reworking in the late Holocene as sea level started to rise after the last glacial maximum (LGM) (Swift, 1975; Woodruff et al., 2013). Around the world, barrier island ages range from early-, mid- to late Holocene (from Mellet & Plater, 2018), indicating that barrier systems formed due to regional factors beyond decreasing global sea level rise rates. Examples include barrier islands such as Sylt in northern Germany that formed 6,000 years ago (Jessel, 2001). In Mozambique, similar ages for barrier island formation were reported that formed as a result of Pleistocene spit progradation (Cooper & Pilkey, 2002). In the Gulf of Mexico, most barrier islands in Texas began forming 4,000-5,000 years ago, during a time when the rate of sea level rise ranged from 0.4-0.6 mm/yr (Anderson et al., 2014, 2016). Decelerating sea level rise in the Holocene created optimal conditions for barrier island formation in the Gulf of Mexico during this time (Figure 2; Otvos & Giardino, 2004; Anderson et al., 2014).
current rate of relative sea level rise near Horn Island ranges from approximately $3.6 \pm 0.59$ mm/yr based on a NOAA water level station in Dauphin Island, AL that has collected data since 1966 to $4.5 \pm 0.86$ mm/yr based on a NOAA water level station in Bay St. Louis, MS that has collected data since 1978 (NOAA, 2018b), which is over an order magnitude higher than the rates of sea-level rise under which many Gulf barriers formed and were progradational (Anderson et al., 2016).

To understand the long-term record of sea level rise, the ratio between oxygen isotopes $^{16}O$ and $^{18}O$ are compared, which reflect the expanse of ice sheets on Earth, relating these to the elevation of sea level. This record is referred to as marine isotope stages (MIS). The boundary between Pleistocene and Holocene roughly corresponds to the change from MIS 2 to MIS 1. MIS 2 ranges from approximately 29,000 years to 14,000 years ago, encompassing the last glacial maximum and the initiation of the current interglacial. MIS 1, from approximately 14,000 years ago to present, reflects sea level rise from 60 meters below current, pre-industrial sea level. In this study, the MIS 2 flooding surface corresponds to the boundary between Pleistocene and Holocene sediments.
Figure 2: Hypothetical flooding of a lowstand delta and associated channel. Cross section shows the subsequent subsurface stratigraphy. Panel A depicts the sedimentological differences between Holocene and Pleistocene sediments found in the study area and shows how the different types of sediments are formed due to sea level rise. Two representative cores are shown from USGS sediment cores in the study area, collected by Kelso and Flocks (2015). Panel B is a cross section of a partially ravined paleochannel, depicting a ravinement surface, reworked and preserved sediments, as well as the Holocene-Pleistocene/MIS 2 boundary.
2.3 Hurricanes

Storms transport material to the back-barrier via overwash and breaching, or can move beach sediments offshore (Houser et al., 2008). Repeated episodes of breaching have shown to cause increasing segmentation and even submergence (Sallenger Jr., 2000; FitzGerald et al., 2008). Shoreline and barrier island ecosystems are able to recover from hurricane impacts (Conner et al., 1989), but more frequent high-intensity hurricanes due to climate change may disrupt the balance between sediment supply and erosion (Bender et al., 2010). Hurricanes are also the cause for the decrease and mortality of plant species on barrier islands, which often aid in mitigating areal and volume loss during the storm, as well as recovery after the event (Carter et al., 2018).

During the instrumental storm period (from 1852 to present), 14 hurricanes of ≥ Category 1 passed within 50 km of Horn Island (NOAA, 2018a). Hurricane Katrina (2005), with a recorded storm surge of ~ 6 meters, caused significant erosion to the eastern tip of Horn Island and forced tidal inlets to widen between the islands (Fritz et al., 2007). Between the 1950’s to 2005, Horn Island lost a total amount of 2.5 hectares/year, with this land loss steadily increasing every year (Morton, 2007). The increasing strength of hurricanes throughout the 20th century contributed to significant erosion and overwash on Horn Island and the other barrier islands (Morton, 2007).

2.4 Antecedent Geology

Incised valleys represent the buried floodplain of ancient rivers and can profoundly influence the location and stability of barrier islands (Anderson et al., 2014, 2016). Incised valleys form during lowstands and regression, when rivers are able to downcut to the shelf edge, and are subsequently filled during transgression (Greene,
During Holocene sea-level rise, sand-rich paleochannels and deltas located within valleys can be eroded and reworked introducing new sand to the active coastal system as it migrates landward. This process, known as transgressive ravinement, has been demonstrated to be the primary factor in delivering sediment to the Texas coast during Holocene sea-level rise (Wilkinson & Basse, 1978). Furthermore, transgressive ravinement of incised valleys can provide enough sediment for barrier islands to exist offshore where they are out of equilibrium with sea level (Rodriguez et al., 2004; Timmons et al., 2010). The contact between reworked and preserved sediments is referred to as the ravinement surface (Figure 2). Incised valleys also provide higher preservation potential during shoreface ravinement, as they offer ample accommodation space due to higher rates of compaction-related subsidence, and are situated at a lower elevation relative to contemporaneous deposits (Anderson et al., 2014, 2016). Tidal ravinement, known as the erosion of sediments as tidal inlets migrate laterally with a barrier island (Miner et al., 2007), are also able to scour deeply into incised valley fill and offer high preservation potential. The interfluvial areas between valleys can also exert an important control, as the inundated Pleistocene deposits that were subaerially exposed are more resistant to erosion. As a result, incised valleys represent antecedent topography that influences the location and sediment availability of coastal features such as barrier islands (Rodriguez et al., 2004).

Potential modern analogs for the incised valleys found intersecting Horn Island are the Pascagoula River to the east and the Biloxi River to the west. The Pascagoula River has a basin area of 24,599 km² and a mean discharge of 432 m³/s (Benke & Cushing, 2011), and the Biloxi River has a watershed of 1,804 km² (Poppenga &
Worstell, 2008). Discharge information for the Biloxi River is not available. Although the Biloxi watershed is more than an order of magnitude smaller in area than the Pascagoula River, satellite imagery and field reconnaissance suggest that it is a sandy system with visibly prominent sandy point bars, compared to the Pascagoula River, which appears to be muddier. Prominent point bars of the Pascagoula River exist farther north than those present in the Biloxi, which is related to the size of its watershed and geology of the drainage basin.

Incised valleys coinciding with abandoned Mobile River, Fowl and La Batre systems lie just east of Horn Island (Greene et al., 2007; Bartek et al., 2004) and may have had a significant impact on Horn Island’s evolution (Figure 3). While these systems have been mapped using seismic data and sediment cores on the inner to outer shelf, no current information exists concerning where the valleys intersect Horn Island, as well as their connections with modern rivers through Mississippi Sound.
Figure 3: Synthesis of studies mapping paleovalleys, which are shown in light green (modified from Greene et al., 2007). The study area is outlined in red. Note the converging incised valleys south of Horn Island and the lack of data to the west.
3. METHODS

The main tool for analyzing Horn Island’s subsurface are two types of geophysical data: seismic and chirp subbottom profilers. Pre-existing datasets offshore and in the MS Sound provide sufficient data to locate paleochannels, the MIS 2 flooding surface, and incised valleys using the software Sonarwiz by Chesapeake Technology Inc. Features interpreted from the geophysical data profiles are then exported, stored and visualized in ArcGIS. The program Surfer by Golden Software LLC aids in gridding surfaces. Pre-existing sediment cores were used to ground truth the MIS 2 flooding surface reflector found in seismic and chirp data. Further, the cores were also used to investigate barrier island facies such as shoreface sediments, washover fans and the Pleistocene surface, as well as provide radiocarbon age constraints.

3.1 Geophysical analyses

High-resolution, shallow penetration geophysical data (0.5-24 kHz chirp) and 1.5 kHz geophysical data (boomer) are used to analyze the subsurface stratigraphy surrounding Horn Island. For the chirp data, the USGS utilized EdgeTech SB-512i and SB-424 Chirp systems, which use a swept frequency signal, but can only penetrate up to 50 ms.

The towfish was towed 1-2 m below the sea surface, and this draft was corrected for in post-processing to provide accurate depths below sea-surface. Seismic data were obtained using a boomer that sounds and geophones that receive signals. The signal sounded by the towfish is reflected at different density interfaces in the subsurface and then detected and recorded digitally by a PC system. A two-dimensional vertical image is produced from the process that records sound reflectors at timed intervals. To synthesize
core and seismic/chirp interpretations into one program, they are uploaded and interpreted in Sonarwiz and gridded in Golden Software Surfer.

Relevant Chirp lines were collected by the USGS in 2008, 2009 and 2010 (08CCT01, 09CCT03, 09CCT04, 10CCT02, 10CCT03; 2010-012-FA; Forde et al., 2011a, 2011b, 2011c; Pendleton et al., 2011) around Horn Island (Figure 4). Relevant boomer lines were collected by the USGS in 1990, 1991 and 1992 (Erda 90-1, Erda 91-3, 92ER2, 02ER4 and 91KI2; Bosse et al., 2017, 2018; Sandford et al., 1991) (Figure 4). Strategically collected new boomer data were collected to fill in gaps where no pre-existing data has been collected or where data quality was poor (Figure 4).
Figure 4: Data coverage of previously collected seismic and chirp datasets (Bosse et al., 2017, 2018; Sandford et al., 1991; Forde et al., 2011a, 2011b, 2011c; Pendleton et al., 2011) in addition to new data. Dense lines are chirp and more sparsely spaced lines are seismic, which are mostly located in the Mississippi Sound. Data quality from chirp lines in the eastern section of the plot is poor and offshore lines also tend to be of worse quality than back-barrier lines.

The MIS 2 flooding surface marks the boundary between the Holocene and Pleistocene periods, which represent two very different sediment types. The Holocene consists of generally soft, compressible sediments of lower density, and the Pleistocene is an exposure surface which is dense, dewatered, and oxidized (Figure 2). Incised valleys tend to be large-scale features in seismic data that have deeply eroded into the MIS 2 flooding surface. To be able to broadly interpret incised valley systems, the MIS 2 flooding surface needs to be verified in sediment cores and seismic data. In seismic data, the MIS 2 flooding surface is typically manifest as a hard and thick reflector in the
subsurface due to the high impedance contrast. The MIS 2 flooding surface around Horn Island is created by interpreting seismic and chirp lines in Sonarwiz. Xyz (x and y are coordinates, and z represents depth) points of the MIS 2 flooding surface reflector were exported and imported into Surfer, a gridding tool of Golden Software LLC. Using the kriging method of interpolation, a gridded surface was created from the xyz data.

3.2 Sedimentology

Coastal environments from 12 sediment cores taken by the USGS in 2010 were previously described (Kelso & Flocks, 2015). Five cores were collected on the Gulf-side of Horn Island, five on the Sound-side of Horn Island, one in Dog Keys Pass and one in Horn Island Pass. The interpretations are mainly based on previous grain size analyses conducted by the USGS, as well as macro-paleontological samples that were extracted from the cores during our sampling.

Before Hurricane Katrina destroyed them in 2005, the Department of Environmental Quality housed 10 Horn Island drill cores at Ocean Springs, MS. While these cores can no longer be sampled, the information that exists is still valuable to understand Horn Island’s evolution. Detailed handwritten corelogs are available that detail grain size, foraminiferal information and sparse radiocarbon ages. Grain size classifications are based on phi sizes and include categories such as sand, sandy mud, muddy sand, silt, and clay. Although mud and clay have similar low sand percentages, the two categories are divided based on having high and low organic content, respectively.

The surficial presence of unidentified shell hash and macro-paleontological samples were described by the USGS, but the majority of shells used for radiocarbon
ages were extracted from the core and is not included in the USGS lithological
descriptions.

3.3 Radiocarbon dating

$^{14}$C found in an organic sample in a sediment core can be measured, compared
and calibrated to a standard curve, providing an age estimate with an associated error
(Törnqvist et al., 2015a). Fifteen carbonate macrofossils were analyzed at the
National Ocean Sciences Accelerator Mass Spectrometry (NOSAMS) facility at the
Woods Hole Oceanographic Institution. Each shell sample was also identified. Of the 15,
8 were pristine, small articulated bivalves, indicating they are in situ.

Cores obtained by the USGS in 2010 (Kelso & Flocks, 2015; Forde et al., 2011c)
were stratigraphically analyzed and radiocarbon dated in this study. Radiocarbon dates
were taken from cores 10CCT05-48, 10CCT05-13, 10CCT05-09, 10CCT05-14B,
10CCT05-40, 10CCT05-41, 10CCT05-42 and 10CCT05-43, and were calibrated using
Marine13 (Reimer et al., 2013) with the standard global marine reservoir correction. Due
to samples originating in the shoreface and from the offshore environment, there is likely
no estuarine source that might cause a differential reservoir effect.
4. RESULTS

4.1 Barrier chronostratigraphy

4.1.1 Unit lithology and seismic characteristics

Figure 5 illustrates the barrier and coastal environments present in the interpreted USGS cores, which will be further explained below. Grain size and sedimentology were used to delineate unit lithologies and radiocarbon dates reflect the age of these environments. All grain size measurements were done by the USGS (Kelso & Flocks, 2015).

In addition to dating the shell material shown in figure 6, each macro specimen was also identified and associated with a depositional, if possible (table 1). In general, the shell species that were dated have an associated environment that corresponds to the location they were sampled (table 1) (Andrews, 1981).
Figure 5: Holocene sedimentological environments (modified from Kelso & Flocks, 2015).
Figure 6: Depiction of 10CCT05 cores surrounding Horn Island (modified from Kelso & Flocks, 2015), including radiocarbon dated sections. The elevation of these cores is plotted with respect to sea-level, which is shown in the center of the figure. USGS grain size information was obtained by Kelso & Flocks (2015). LiDAR data were collected in 2010 (USACE NCMP Lidar: Gulf Coast - LA, MS), and bathymetry data were collected in 2007 (Taylor et al., 2008). Note the radiocarbon ages for various cores quantifying overwash deposition and shoreface migration. Pleistocene depth is based on core and seismic interpretations.
Table 1

Information on location, species and environment of radiocarbon dated shells.

Environmental information was taken from (Andrews, 1981).

<table>
<thead>
<tr>
<th>Core</th>
<th>Depth relative to core (cm)</th>
<th>Calibrated age BP (2 sigma)</th>
<th>Taxa</th>
<th>Environment (from Andrews, 1981)</th>
</tr>
</thead>
<tbody>
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<td>354</td>
<td>5844-6836</td>
<td>Raeta plicatella</td>
<td>Lives on sandy bottom of outer surf zone; epifaunal</td>
</tr>
<tr>
<td>10CCT05-13</td>
<td>106</td>
<td>113-671</td>
<td>Cirripedia: Balanomorpha (Barnacle)</td>
<td>N/A</td>
</tr>
<tr>
<td>10CCT05-9</td>
<td>164-169</td>
<td>7300-8177</td>
<td>Strombus alatus</td>
<td>Intertidal to nearshore; epifaunal</td>
</tr>
<tr>
<td>10CCT05-9</td>
<td>260</td>
<td>6678-7623</td>
<td>Mulinia lateralis</td>
<td>Clayey sediments; infaunal</td>
</tr>
<tr>
<td>10CCT05-14B</td>
<td>57</td>
<td>887-1552</td>
<td>Tellina (Eurytellina) alternata</td>
<td>Inlet environment</td>
</tr>
<tr>
<td>10CCT05-40</td>
<td>363-369</td>
<td>2294-3324</td>
<td>Raeta plicatella</td>
<td>Lives on sandy bottom of outer surf zone; epifaunal</td>
</tr>
<tr>
<td>10CCT05-40</td>
<td>393</td>
<td>2814-3852</td>
<td>Mulinia lateralis</td>
<td>Clayey sediments; infaunal</td>
</tr>
<tr>
<td>10CCT05-40</td>
<td>460</td>
<td>3097-4248</td>
<td>Parvilucina multilineata</td>
<td>Offshore and inlet-influenced areas; infaunal</td>
</tr>
<tr>
<td>10CCT05-41</td>
<td>345</td>
<td>800-1471</td>
<td>Gastropod</td>
<td>N/A</td>
</tr>
<tr>
<td>10CCT05-42</td>
<td>489-490</td>
<td>3922-4599</td>
<td>Diplodonta (Phlyctiderma) cf. D. soror</td>
<td>Inlet-influenced areas; infaunal</td>
</tr>
<tr>
<td>10CCT05-42</td>
<td>505</td>
<td>3819-4410</td>
<td>Oliva (Ispidula) sayana</td>
<td>Inlets and offshore; infaunal</td>
</tr>
<tr>
<td>10CCT05-42</td>
<td>510</td>
<td>4817-5651</td>
<td>Diplodonta (Phlyctiderma) semiaspera</td>
<td>Open-bay centers, jetties, inlet-influenced areas; infaunal</td>
</tr>
<tr>
<td>10CCT05-43</td>
<td>80</td>
<td>3137-3692</td>
<td>Oliva (Ispidula) sayana</td>
<td>Inlets and offshore; infaunal</td>
</tr>
<tr>
<td>10CCT05-43</td>
<td>145</td>
<td>4722-5302</td>
<td>Strombus alatus</td>
<td>Intertidal to nearshore; epifaunal</td>
</tr>
<tr>
<td>10CCT05-43</td>
<td>276</td>
<td>4276-4836</td>
<td>Oliva (Ispidula) sayana</td>
<td>Inlets and offshore; infaunal</td>
</tr>
</tbody>
</table>
Table 2

Grain size information of different lithofacies according to the Folk and Ward (1957) Method done by Kelso & Flocks (2015).

<table>
<thead>
<tr>
<th>Unit</th>
<th>Core</th>
<th>Sediment description</th>
<th>Mean grain size (μm) – (number of samples)</th>
<th>Standard Deviation associated with grain size</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper shoreface</td>
<td>10CCT05-41</td>
<td>Medium sand</td>
<td>296.5 – (5)</td>
<td>2.3</td>
</tr>
<tr>
<td>Upper shoreface</td>
<td>10CCT05-42</td>
<td>Muddy sand</td>
<td>136.5 – (3)</td>
<td>2.9</td>
</tr>
<tr>
<td>Upper shoreface</td>
<td>10CCT05-44B</td>
<td>Medium sand</td>
<td>266 – (4)</td>
<td>7.7</td>
</tr>
<tr>
<td>Lower shoreface</td>
<td>10CCT05-40</td>
<td>Muddy sand</td>
<td>190.1 – (2)</td>
<td>4.6</td>
</tr>
<tr>
<td>Lower shoreface</td>
<td>10CCT05-42</td>
<td>Sandy mud</td>
<td>109.5 – (2)</td>
<td>3.6</td>
</tr>
<tr>
<td>Lower shoreface</td>
<td>10CCT05-43</td>
<td>Muddy sand</td>
<td>149.7 – (2)</td>
<td>10.4</td>
</tr>
<tr>
<td>Lower shoreface</td>
<td>10CCT05-40</td>
<td>Sandy mud</td>
<td>57.8 – (4)</td>
<td>4.0</td>
</tr>
<tr>
<td>Lower shoreface</td>
<td>10CCT05-43</td>
<td>Sandy/Silty clay</td>
<td>48.0 – (3)</td>
<td>3.1</td>
</tr>
<tr>
<td>Bottom of lower shoreface</td>
<td>10CCT05-40</td>
<td>Sandy/Silty clay</td>
<td>51.1 – (1)</td>
<td>8.6</td>
</tr>
<tr>
<td>Overwash</td>
<td>10CCT05-48</td>
<td>Muddy sand (360 cm)</td>
<td>141.9 – (1)</td>
<td>16.8</td>
</tr>
<tr>
<td>Overwash</td>
<td>10CCT05-9  (170 cm)</td>
<td>Medium sand</td>
<td>335 – (2)</td>
<td>20.4</td>
</tr>
<tr>
<td>Overwash</td>
<td>10CCT05-9 (230 cm)</td>
<td>Muddy sand</td>
<td>233.3 – (1)</td>
<td>7.1</td>
</tr>
<tr>
<td>Back-barrier sands</td>
<td>10CT05-13</td>
<td>Mud</td>
<td>9.3 – (1)</td>
<td>0.2</td>
</tr>
<tr>
<td>Tidal delta deposits</td>
<td>10CCT05-49</td>
<td>N/A (Wide range)</td>
<td>3.4-242 – (6)</td>
<td>0.7</td>
</tr>
<tr>
<td>Tidal delta deposits</td>
<td>10CCT05-47B</td>
<td>N/A (Wide range)</td>
<td>7.9-216 – (4)</td>
<td>0.2-0.9</td>
</tr>
<tr>
<td>Tidal delta deposits</td>
<td>10CCT05-14B</td>
<td>N/A (Wide range)</td>
<td>332.4-539.2 – (3)</td>
<td>7.0-17.0</td>
</tr>
</tbody>
</table>
Table 2 (continued)

<table>
<thead>
<tr>
<th>Barrier platform</th>
<th>Average from DEQ cores</th>
<th>Mud</th>
<th>9 – (16)</th>
<th>N/A</th>
</tr>
</thead>
<tbody>
<tr>
<td>Barrier platform</td>
<td>Average from DEQ cores</td>
<td>Sandy mud</td>
<td>44.2 – (4)</td>
<td>N/A</td>
</tr>
<tr>
<td>Barrier platform</td>
<td>Average from DEQ cores</td>
<td>Sandy silt</td>
<td>38.5 – (12)</td>
<td>N/A</td>
</tr>
<tr>
<td>Barrier platform</td>
<td>Average from DEQ cores</td>
<td>Muddy sand</td>
<td>203.0 – (45)</td>
<td>8</td>
</tr>
<tr>
<td>Barrier platform</td>
<td>Average from DEQ cores</td>
<td>Medium sand</td>
<td>353.6 – (43)</td>
<td>0.5</td>
</tr>
</tbody>
</table>

4.1.1.1 Unit I: Upper and lower shoreface

The shoreface represents a sensitive transition zone between the foreshore and inner shelf, where the existence, response and geometry of barrier island facies can be altered through sea-level changes, sediment supply and hydrodynamic processes (Swift, 1975; Rodriguez et al., 2001). At the Texas-Louisiana border, upper shoreface deposits extend to 3 kilometers offshore, while the entire shoreface can range as far as 7 kilometers offshore (Rodriguez et al., 2001). In Galveston Island, Texas, the shoreface extends ~ 5 km offshore (Wallace et al., 2010). Using sedimentological and bathymetric data, the shoreface of Horn Island can be measured and examined. Shoreface deposits generally range from fine to medium sand, and sand grain size decreases with distance from the barrier platform (Rodriguez et al. 2001; Timmons et al., 2010; Raff et al., 2018). The upper shoreface and barrier platform sediments are very similar in grain size. Shells that characterize the shoreface environment in the northern Gulf of Mexico include *Mulinia lateralis, Anadara transversa, Anadara ovalis,* and *Oliva sayana* (Parker, 1960; Rodriguez et al., 2004). The shoreface environment of Horn Island was subdivided into the upper and lower shoreface in this study. This distinction was made based on the mud
fraction of each facies, as the lower shoreface contains thick layers of mud and only sparse, thinly-bedded layers of sand (Rodriguez et al., 2001).

Several cores contain massive, fine-grained sand facies and are located within one kilometer of Horn Island’s Gulf-side, indicating they are associated with the upper shoreface (Table 2). These include core 10CCT05-41, the upper portion of core 10CCT05-42 and core 10CCT05-44B. One radiocarbon age from a gastropod in core 10CCT05-41 produces a calibrated 2 sigma age range of 800-1471 years BP (core depth: 345 cm) (Figure 6).

Cores 10CCT05-40, the lower portion of core 10CCT05-42, and core 10CCT05-43 contain sections of muddy sand, sandy mud, and silty/sandy clay (Table 2). All have a significant decrease in sand percentage compared to the three core sections from the upper shoreface, indicating these environments represent the lower shoreface. The bottom of the lower shoreface coincides with the transition from the above-mentioned facies to clay and silty/sandy clay that contain almost zero percent sand. In addition to being approximately two to four kilometers offshore from the island’s Gulf-side, the sedimentology and macrofossils indicate that these facies are also associated with the lower shoreface. Six of the nine ages from the lower shoreface originated from offshore/inlet-influenced areas (Table 1). Ages corresponding with each core can be found in figure 6, and all date the bottom of the lower shoreface at different locations. The trend of ages become younger in a westward direction. These ages, moving westwards between cores, are 4276-4836 years BP (core depth: 276 cm), 4722-5302 years BP (core depth: 145 cm), and 3137-3609 years BP (core depth: 80 cm) from core 10CCT05-43, 4817-5651 years BP (core depth: 510 cm), 3819-4410 years BP (core
depth: 505 cm), and 3922-4599 years BP (core depth: 489 – 490 cm) from core 10CCT05-42, and 3097-4248 years BP (core depth: 460 cm), 2814-3852 years BP (core depth: 393 cm) and 2294-3324 years BP (core depth: 363-369 cm) from core 10CCT05-40.

Chirp and seismic data added a spatial understanding to the architecture and extent of the upper and lower shoreface. Core 10CCT05-43 intersects a seismic line in dataset 10CCT02 and was thus used to interpret reflectors. Figure 7 depicts the 7-meter-thick lower shoreface of Horn Island as well as a partially ravined paleochannel. Shoreface deposits generally have medium frequency and moderate to low reflectors (Bartek et al., 2004). The upper shoreface is characterized by chaotic reflectors, which are often associated with transgressive shoreface reworking (Bartek et al., 2004) and muddy sands, which are seen in core 10CCT05-43. In contrast, parallel reflectors are present in the lower shoreface (Figure 7). The internal structure is not identifiable, but the ravinement surface reflector is distinctive. The paleochannel in figure 7 has a distinctive channel-like external geometry and is capped by an erosional ravinement surface. The fill reflectors are of moderate amplitude, moderate to high frequency, and moderate continuity.
Figure 7: A Chirp line from USGS cruise 10CCT02 (modified from Forde et al., 2011c) includes core 10CCT05-43, showing the extent of the lower shoreface, the bottom and fill of a paleochannel and the ravinement surface of this channel. Chaotic reflectors suggesting sandy sediments characterize the shoreface. Partial erosion of a paleochannel is indicated by a ravinement surface.
4.1.1.2 Unit II: Storm overwash deposits

Overwash deposits in the north-central and north-western Gulf of Mexico often contain sand that is sometimes interbedded with mud (Raff et al., 2018; Odezulu et al., 2018). Cores 10CCT05-48 and 10CCT05-9 are within the clear boundaries of the back-barrier environment of Horn Island. The oldest ages presented in this study are likely sandy overwash deposits interfingered with lagoonal muds north of the island (Figure 6). These deposits may also indicate tidal delta deposition, but both cases suggest the presence of a barrier island. Core 10CCT05-48 shows a calibrated age range of 5844-6836 years BP (core depth: 354 cm) and core 10CCT05-9 has calibrated ages of 6678-7623 years BP (core depth: 260 cm) and 7300-8177 years BP (core depth: 164 – 169 cm). Each core contains muddy sand, sand or interbedded mud and sand facies that have a sand content between 70 and 100 percent (Table 2). One of the shells extracted from core 10CCT05-9 was an articulated Mulinia lateralis, which lives in salinities ranging from 10-30 ppt and is indicative of bay environments, but can also be associated with shoreface sediments, as explained above (Lippson & Lippson, 2006; Rodriguez et al., 2004). The thin nature of overwash deposits and the sparse as well as moderate to poor quality of seismic and Chirp data available makes it difficult to identify overwash using geophysics.

4.1.1.3 Unit III: Back-barrier muds

Back-barrier marine muds are also present in the lagoonal environment of Horn Island and accumulated during quiescent weather patterns. The sand content of back-barrier muds is below 10 percent and shell content is minimal (Table 2). When in contact with overwash deposits, back-barrier muds can be interbedded with sand. They represent
material that naturally accumulates by settling of fine material in the water column. One radiocarbon age obtained from a barnacle, located in backbarrier muds found in core 10CCT05-13 has a two sigma of 113-671 years BP (core depth: 106 cm) shows that the marine muds have a rate of deposition of approximately 1 meter per 400 years. Grain size information is available at a depth of 50 cm in the core, which still corresponds to the back-barrier mud facies.

Back-barrier muds imaged in seismic and Chirp data show up as thin, high-amplitude, parallel reflectors with a sheet-like external geometry.

4.1.1.4 Unit IV: Tidal delta deposits

Tidal delta deposits are characterized by interbedded sands and muds (Miner et al., 2007), and are present in cores west of the characteristic geomorphic turn, as well as cores from both tidal inlets, including cores 10CCT05-49, 10CCT05-47B and 10CCT05-14B (Table 2). A single radiocarbon date exists in core 10CCT05-14B, that indicates the age of Horn Island Pass is at least 887-1552 years BP (core depth: 57 cm). The shell extracted from this core is associated with tidal inlet environments, making this a representative age estimate for the inlet (Andrews, 1981). In the Chirp dataset 08CCT01, and seismic lines collected by USM, there are clear spit progradation and tidal channel features present just northwest of Horn Island’s geomorphic turn. Spit accretion reflections are high-amplitude and wavy-parallel and contain a tidal channel between reflectors, which has a channel-like external geometry with chaotic fill.

4.1.1.5 Unit V: Barrier platform

Barrier platform sediments are thick deposits that extend to the upper shoreface, where they are then modified (Rodriguez et al., 2001). Figure 8 depicts drill cores from
Horn Island’s barrier platform. All cores have a coarsening upwards sequence during Holocene deposition. Grain size information is published online with the Mississippi Department of Environmental Quality. The sequence begins with mud, sandy mud and sandy silt, and continues to muddy sand (Table 2). The coarsest facies is medium sand. Dates in some of the cores provide information about the age of the facies, which indicate that the very fine-grained, muddy deposits in each core represent the start of Holocene deposition, including 4523-5192 years BP (core depth: 12.8 m) in core BI-7 and 8303-8687 years BP (core depth: 18.7 m) in core BI-8 (Otvos et al., 1981). One Pleistocene age in core BI-13 dates to 35042-42141 years BP (core depth: 21.3 m) (Otvos, 1981).
Figure 8: Strike-oriented litho-stratigraphic cross-sectional transect along Horn Island based on grain size interpretations and radiocarbon ages done by Otvos (1981). Note the coarsening upwards sequence in the Holocene showing increased barrier upbuilding and stability.
4.2 Geophysics

While the Pascagoula incised valley extends the current river valley by continuing in a north to south direction, the Biloxi incised valley forms anastomosing channels in the Mississippi Sound (Figure 9). Three different branches of the Biloxi incised valley intersect Horn Island. In the southwest corner of the MIS 2 surface smaller, more distinct paleochannels can be identified due to the high-resolution, dense Chirp coverage available. In the Mississippi Sound, multiple shallow Pleistocene highs were mapped. They represent interfluvial areas that were not incised and eroded by the Biloxi and Pascagoula river systems. Horn Island lies just south of these Pleistocene highs. The low elevation sections of the DEM overlaying the MIS 2 surface show potential modern continuations of incised valleys.

The depths of the tops of the Biloxi and Pascagoula incised valleys range from 13 to 15 meters below sea level, respectively. The bases of both valleys are at least 26 meters deep, but could not be imaged entirely. In the Mississippi Sound, the width of the Biloxi incised valley ranges from 1.5 km to 5 km and increases to approximately 7 km in the Gulf of Mexico. In comparison, the width of the Pascagoula incised valley is approximately 5 km in the Mississippi Sound and widens to about 8 km in the Gulf of Mexico. The length of the Biloxi incised valley that was mapped is about 90 km, including smaller, distinctive lowstand channels in the southwestern quadrant of figure 9. The length mapped of the Pascagoula incised valley is approximately 25 km.

Figure 10 depicts a section of the Biloxi incised valley. The fill of the valley is at least 15 meters thick and it is unclear whether the bottom of the valley was imaged, yet valley walls and paleochannels were imaged at the surface of the incised valley.
Figure 9: MIS 2 flooding surface with 2007 bathymetry as background (NOAA, 2007). This figure depicts the MIS 2 flooding surface around Horn Island that is overlaid by a digital elevation model. The orange/red color in the MIS 2 surface represents shallow Pleistocene deposits and the green and blue colors depict deeper Pleistocene deposits. The continuation of the lowstand incised valleys of the Pascagoula and Biloxi Rivers can be seen in green and blue. Note the Pleistocene highs in the Mississippi Sound, the smaller tributary channels in the southwestern quadrant and the varying geometries of the Biloxi and Pascagoula River incised valleys. In addition, the data shows a potential convergence of the Biloxi and Pascagoula valleys.
Figure 10: Seismic line from USM cruise imaging the Biloxi incised valley. Note valley wall and prominent horizontal reflectors representing valley fill.
4.3 Paleochannels

Within the incised valleys above the MIS2 sequence boundary, chirp and seismic lines show paleochannel features that were ground-truthed by the USGS in 2010 using vibracores. Figure 11 depicts the locations of paleochannels with respect to incised valleys found in the MIS 2 surface. One notable observation is that paleochannels on the Sound-side of Horn Island are found at varying depths ranging from deep to shallow, while on the Gulf-side, existing paleochannels only have a deep expression.

The discrete number of paleochannels can only be estimated in the southwest quadrant of figure 11, as data are too sparse anywhere else. Roughly ten paleochannels between 10 and 25 meters below sea level can be identified and associated with the Biloxi River incised valley system. Their fill is the same sediment that the Biloxi River carries, and would therefore be fine to medium sand.

Twichell et al. (2011) also imaged the southwestern quadrant of the study area and found similar paleochannels, tidal and lowstand channel sands. Paleochannels are more abundant where there is a higher density of seismic and chirp lines (Figure 11).
Figure 11: Paleochannel features mapped using geophysics. The MIS 2 Surface color scheme is identical to figure 10. The different colors of the points associated with paleochannels represent depths, with light blue being the shallowest and pink being the deepest.
5. DISCUSSION

5.1 Lowstand system

Antecedent geology is a dominant factor in controlling barrier island evolution in passive margin regions that have only limited sediment supply, as is the case for the northern Gulf of Mexico and the east coast of the United States (Riggs et al., 1995). During the falling stage of sea level from the past sea level cycle, the Biloxi and Pascagoula valleys either reoccupied valleys from previous sea-level cycles or incised into Pleistocene sediments. Both valleys exhibit widening in a seaward direction, which is a common scenario, as incised valleys tend to widen in a seaward direction due to an increase in slope steepness of the continental shelf (Schumm, 1993; Anderson et al., 2014). Similarly, the Rio Grande incised valley in Texas widened seaward and even deepened into a canyon head (Anderson et al., 2016). The convergent valley of the Sabine and Trinity River incised valley also widens on the outer shelf (Simms et al., 2007), indicating that valleys behave similarly in Texas and Mississippi.

The potential convergence of the Pascagoula and Biloxi River incised valleys is suggested in the synthesis study done by Greene et al. (2007) (Figure 3 & 9). Comparing these valleys to northeastern Texas, the converging incised valleys vary widely in geometry, with the Biloxi and Sabine incised valleys exhibiting narrow cross-sectional profiles and the Trinity and Pascagoula incised valleys being broad and terraced (Anderson et al., 2014). This indicates that the geomorphologically and sedimentologically differing Biloxi and Pascagoula river valleys likely nourished an ancient delta, similar to other systems around the Gulf of Mexico. This delta is likely the Lagniappe delta (Greene et al., 2007). The buried lowstand continuation of the Biloxi and
Pascagoula Rivers keep the sedimentological characteristics of their modern analogs due to the nature of drainage basin sediment production (Anderson et al., 2016).

The incision of paleovalleys into the MIS 2 flooding surface is an important framework for modern coastal features to develop (Mallinson et al., 2010). The Pleistocene ridge at Dauphin Island was the site of initial shoal accretion that ultimately incorporated Horn Island (Otvos, 1981), just as Cape Hatteras on the US east coast is perched on an interfluvial Pleistocene high (Mallinson et al., 2010). As shown in previous studies from Texas, the MIS 2 exposure surface created by the incision of river valleys provides the framework for modern coastal features to develop (Anderson et al., 2014). This is seen at Galveston Island, where the barrier varies greatly in thickness due to its location above the Trinity incised valley, which was likely the source of sand for the barrier (Anderson et al., 2014). Furthermore, Mustang Island’s location over the Nueces incised valley depths to the MIS 2 surface can be up to 38 m (Simms et al., 2006). Even though this valley is smaller than other incised valleys in this area, it can still withstand numerous sea-level cycles and downcut deeply into the Pleistocene (Anderson et al., 2014).

Horn Island is situated above two incised valleys and an MIS 2 surface up to 26 meters deep. Similar to the Nueces, the Pascagoula and Biloxi are smaller rivers compared to the surrounding area that includes the Pearl and Mobile Rivers, but the incised valley resulting from the Biloxi and Pascagoula are likely reoccupied during multiple sea-level cycles and replenish their fluvial sands within the system.
5.2 Ravinement and initial island formation

The three branches of the Biloxi incised valley that intersect Horn Island bring sandy paleochannels with them that can be ravined, providing ample sand. This is evidenced through paleochannels in seismic and chirp exhibiting chaotic reflectors associated with sand (Bartek et al., 2004). Although the Pascagoula incised valley is muddier, some sand can also likely be ravined from it. Ephemeral paleochannels are located in these incised valleys and represent meanders within the incised valley floodplain (Figure 11). As sea level rose during the Holocene, transgressive ravinement eroded much of the sand from the channels. Paleochannels can act as local sediment sources, where ravinement is more effective than in interfluvial areas where Pleistocene sediments are shallow, providing sediment to barrier islands in close proximity (Timmons et al., 2010). The shallow paleochannels that were eroded on the front side of Horn Island likely released a significant amount of sand into the coastal system as it migrated landward across the shelf (Figure 10). The proximity of paleochannel sands can aid in the building and preservation of large dunes and may help in the stability of an island in response to storms (Wernette et al., 2018). Although some sand is sourced from the east via Dauphin Island (Otvos, 1981), Horn Island’s relative stability compared to adjacent barrier islands can likely be attributed to its more local source of sediment from underlying and adjacent paleochannels from the Pascagoula and especially the sand-rich Biloxi River. Similarly, the barrier Bogue Banks in North Carolina is wider, more stable and less erodible where paleochannels intersect the island (Timmons et al., 2010).

One explanation for the potential relict overwash deposits (7,150 and 7,738 years BP) found in the backbarrier environment of Horn Island is the existence of an ephemeral
ancestral transgressive island in near modern-day position. When sea level rates were much higher at ~4 mm/yr, approximately 8,000 years BP, barrier islands were not able to maintain pace with sea level rise even though there was ample sediment supply (Anderson et al., 2016). The ancestral transgressive Horn Island likely scoured sand during ravinement but was continuously overwashed by sea level rise and storms. Relict overwash deposits in the back-barrier correspond with older ages in cores 10CCT05-48 and 10CCT05-9. Ancestral barrier islands associated with Sabine and Heald Banks existed from approximately 8.4 to 7.5 ka (Anderson et al., 2016). This indicates that around this time, frequently overwashed, ephemeral islands existed in numerous places in the northern Gulf of Mexico. As the ephemeral island became more stable with lower sea-level rise rates, shoreface ravinement was much more effective on the Gulf-side of Horn island because a high amount of wave energy is able to erode much more sediment there without the protection of the island. Furthermore, lowering sea level rise rates allowed the island to transition from being transgressive to prograding seaward, incorporating sand from paleochannels via shoreface ravinement that helped establish a stable shoreface. After the continuous presence of a stable Horn Island, wave energy in the Mississippi Sound was reduced and erosion in the back-barrier was limited to shallower bay ravinement. Tidal ravinement also releases sand from paleochannels as tidal inlets and barriers migrate laterally. Tidal inlets are also more stable in preserved channel-like features due to enhanced erodibility of sediments (Kulp et al., 2007). The ephemeral Horn Island formed near its modern-day position ~8,000 years ago, meaning it has not retrograded, and has had a progradational evolution (Figure 6). If Horn Island
experienced retrogradation, shoreface ravinement would have reworked the shallow paleochannels currently existing on the Sound side of the island.

Lithologic information from cores (Figure 8) is another important tool to understand the formation of Horn Island. The transition from mud to sand in all cores in figure 8 represents initial subaqueous formation of Horn Island, which occurred around 8,000 years ago. Subsequently, the transition from muddy sand to sand shows the initial subaerial exposure and growth of the island. Sandy silt and sandy mud sections below the muddy sand facies could represent an even earlier initiation of the upbuilding of Horn Island. Clay facies towards the bottom of the cores likely represent oxidized Pleistocene material.

Once the island was fully established, coarser grained material began to accrete ~4.5 ka (Otvos, 1981). Due to the nature of the island, the barrier platform is characterized by a coarsening upwards sequence. This relationship is similar in the shoreface, described below.

5.3 Horn Island Evolution

Once sea level rise rates slowed during the Holocene, the ephemeral island transitioned to a more stable system and kept pace with sea level rise. The single radiocarbon age taken by Otvos (1981) does not allow for much chronological understanding of system evolution. Unfortunately, drilling through these thick barrier sands to obtain more ages is no longer an option due to the protected status of a National Seashore. However, the shoreface is a continuation of barrier island facies that thin offshore of the island (Rodriguez et al., 2004). For example, a dip-oriented transect through Galveston Island, Texas shows that barrier island facies thin towards the toe of
the shoreface, but that all sediments present in the center of the island are still preserved there due to isochronous surfaces extending to the shoreface (Rodriguez et al., 2004). In addition, shoreface deposits can show retrogradation and progradation of the barrier by onlapping or offlapping of sediments with respect to offshore marine muds (Rodriguez et al., 2001). Similar ages in the shoreface and backbarrier can further be used to constrain island evolution (Odezulu et al., 2018).

Sampling of in situ macro fossils shown in figure 12 reveal the timing and processes of barrier evolution. The shoreface shows slightly older ages than previously reported (Otvos, 1981), indicating the first signs of island progradation.

Cores 10CCT05-40, 10CCT05-42 and 10CCT05-43 dated the shoreface of Horn Island at various locations. Each core has three dates clustered at the lower shoreface, and the lower shoreface becomes younger in a westward direction, supporting the argument for the westward lateral movement of the island. Cores 10CCT05-43 and 10CCT05-42 date the bottom of the lower shoreface to ~4,500 years BP, and core 10CCT05-40 dates the shoreface to ~3,500 years BP. As Horn Island laterally accretes westwards, tidal inlets move accordingly. As a result, the shells associated with the bottom of the lower shoreface are also associated with offshore/tidal inlet habitat. Six of the nine macrofossils sampled at the bottom of the lower shoreface are associated with tidal inlets and offshore locations (Andrews, 1981), corresponding to the expected macrofossil assemblage.

Tidal deposits are only present in backbarrier cores on the western side of Horn Island that also exhibit a lack of overwash deposits (Figure 6). This indicates that Horn Island was not subaerially exposed for overwash to deposit on its western side, and that a
tidal inlet was present at this location. Historical aerial maps of Horn Island support this claim, as the part of Horn Island west of its geomorphic turn only started being subaerially exposed at the turn of the 20th century. Figure 12 incorporates barrier platform cores shown in figure 8 with offshore and back-barrier cores from figure 6, producing corresponding up-dip cross sections. Where radiocarbon dates from this study are available, they are drawn into figure 12 to illustrate chronostratigraphic horizons. Overwash deposits are only seen on the eastern side of the island and tidal deposits only on the western part, indicating an initial barrier nucleation point in the east. Incised valleys intersect much of the island, potentially providing sand for the island during ravinement. Where valleys are not present, stiff Pleistocene deposits are much shallower. The various facies of Horn Island differ significantly in terms of age, ranging between 400 years BP and 8,000 years BP. The ~400 years BP age taken from back-barrier muds shows recent, fine-grained deposition in the Mississippi Sound, covering and preserving much older overwash deposits. One upper shoreface age of ~1,000 years BP west of the geomorphic turn indicates that while Horn Island has not been subaerially exposed until the turn of the 20th century, the island has been building out its shoreface there for a long time before it was visible above water. The 3,000 years BP age in the lower shoreface of the westernmost cross-section illustrates the same point. The 4,500 years BP age at the bottom of the lower shoreface in central Horn Island is a good upper estimate for the age of the island.
Figure 12: Correlations between dip-oriented cross sections from sediment core information showing chronostratigraphic horizons. Note the lower shoreface age on the youngest, western part of the island, as well as the locations of tidal and overwash deposits.
Interestingly, the antecedent geology is still influencing the modern morphodynamics of the system. The geomorphic anomalies of Horn Island encompass its characteristic turn and the fact that it is far from the mainland. These can be explained by the location of incised valleys in the Mississippi Sound (Figure 12). Horn Island is preferentially located where the MIS 2 flooding surface is deep, coinciding with locations of incised valleys. Numerous Pleistocene highs currently prevent major retrogradation of the island, which would be the natural path of the island as sea level is rising. Instead the island has historically accreted westwards (Morton, 2008), ravening incised valleys filled with sand, a process that has remained unchanged since the island’s formation approximately 4,500 years BP. Horn Island’s characteristic geomorphic southwest turn is similarly nestled alongside a Pleistocene high, whose stiff and hard-to-erode sediments prevent the island from evolving into a straight shoreline. Spit accretion is apparent just north of the geomorphic turn from Chirp data, indicating that at one point, the island was attempting to laterally accrete in this direction. However, spit accretion was unsuccessful, eroding only little of the Pleistocene sediments sitting below.

Anthropogenic effects on global climate change are causing an increase in the rate of sea level rise. Rates are approaching when Horn Island was ephemeral at 4 mm/yr and not able to build out a shoreface due to high rates of sea level rise approximately 8,000 years BP. This means that the island will likely become unstable and sediment supply via ravinement will not be able to keep pace with sea level rise. Moreover, the shifting of the loop current closer to the northern Gulf of Mexico, as it has done in the geologic past
(Bregy et al., 2018), could increase intense hurricane activity near Horn Island and further alter sediment dynamics.
6. -- CONCLUSIONS

Horn Island provides a unique opportunity to examine barrier island response to sea level rise, storms, antecedent topography, ravinement processes and sediment supply variations. Using geophysical datasets and sediment cores, this study shows that sediment supply was not only sourced from the east via longshore processes (Otvos, 1981), but also from transgressive and tidal ravinement of previously deposited coastal and fluvial lithosomes that contributed large volumes of sand towards island formation. The island has been experiencing westward lateral migration since its formation ~4,500 years BP. However, an ephemeral island existed ~8,000 years BP, based on overwash deposits in the back-barrier environment. Due to high rates of sea level rise of ~4 mm/yr around this time (Anderson et al., 2016), this barrier was not able to nourish a shoreface. Once sea level rise rates slowed down around 4,500 years ago, Horn Island began to establish itself. The Biloxi and Pascagoula incised valleys merge north of Horn Island and intersect the barrier. The tidal and shoreface ravinement of Holocene paleochannels from these incised valleys was a significant contributor of sediment. The Biloxi River is relatively sand-rich compared to the Pascagoula River, so ravinement from associated paleochannels were an important contributor to island stabilization. Antecedent topography in the form of a Pleistocene high representing a valley wall restricts ravinement due to the stiffness of Pleistocene sediments, causing the geomorphic turn of Horn Island. Barrier movement and ravinement continues further south, where an incised valley is present and provides more erodible sediment in the form of thick Holocene sediment packages. Future rates of relative sea-level rise for the Gulf of Mexico will near the rates of sea-level rise ~8,000 years BP, a time when the island was ephemeral,
frequently overwashed, and not able to build out a sandy shoreface. Therefore, the future of the Island system could be approaching unstable conditions.
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