Sediment Transport and Slope Stability in the Northern Gulf of Mexico

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SEDIMENT TRANSPORT AND SLOPE STABILITY IN THE NORTHERN GULF OF MEXICO

A Dissertation

Submitted to the Graduate Faculty of the Louisiana State University and Agricultural and Mechanical College in partial fulfilment of the requirements for the degree of Doctor of Philosophy in

The Department of Oceanography and Coastal Sciences

by
Jeffrey Blake Obelcz
B.Sc., Coastal Carolina University, 2011
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ABSTRACT

Sediment transport and slope stability are fundamental organizing agents of the geological record. These processes have been extensively studied along the northern margin of the Gulf of Mexico basin for both basic and applied purposes, but our knowledge of them is limited by the spatial and temporal sampling capabilities of traditional geologic oceanographic surveying tools such as coring, single-beam echosounders, and sidescan sonar. This dissertation seeks to update the state of knowledge regarding northern Gulf of Mexico sediment transport and slope stability from annual to millennial timescales, primarily using relatively high-resolution acoustic geophysical tools such as swath bathymetric echosounders and swept-frequency subbottom echosounders.

There are three primary findings of this dissertation: (1) the subaqueous Mississippi River Delta Front is a zone of active downslope sediment flux in lieu of major hurricane passage, and the volume of sediment transported downslope during major hurricane and non-major hurricane containing intervals is comparable, (2) mud-capped dredge pits used for coastal restoration projects in Louisiana can be used as proxies for sediment deposition and slope stability along the inner continental shelf, and highlight the important role resuspension and slope failure play in decadal and longer-scale sediment accumulation in this environment, and (3) a drowned forest of age > 40,000 years before present found offshore Gulf Shores, Alabama likely represents a unique or at least fairly localized depositional environment that preserved entire tree stumps during geologic periods that favor destruction of sedimentary fabric, including sea level lowstand and transgression.
CHAPTER 1: INTRODUCTION

A dissertation’s components can be related several ways: spatially, temporally, or thematically. Defining this dissertation in a spatial sense provides the greatest unity; studies are confined to a ~250 km wide swath of the Northern Gulf of Mexico (NGoM) inner continental shelf (ICS) from the Mississippi River Delta Front (MRDF) in the west to Mobile Bay, Alabama in the east. Study sites lay no further seaward than 25 km, which aligns with the research statement and goals of both the LSU Coastal Studies Institute and the Department of Oceanography and Coastal Sciences.

The second unifying feature of this dissertation is methodology. While each of three studies that compose the body of this work are supported by various complementary data sets, including numerical modeling, coring, and in situ observation, the majority of findings are derived from relatively high-resolution acoustic geophysical devices. The most prominent of these devices are swath bathymetric sonars and swept-frequency (Chirp) subbottom sonars. These tools used in tandem can produce a horizontal and vertical sub-meter scale record of large areas of the seafloor and shallow subsurface, yielding a vast amount of information regarding contemporary to millennial-scale sedimentary processes and events.

Investigations that compose the Second and Third Chapters of this dissertation have significantly more commonality than the Fourth. This divergence is especially stark when considering the temporal ranges the studies focus on. Overwhelmingly the Second and Third Chapters are set within the Holocene/Anthropocene epoch and are best conceptualized as studying the interface between contemporary physical oceanographic processes and geologically recent (<1,000 years) deposits. Narrowly confined time-scales result in high-frequency processes
(from a geological perspective) such as tides, ambient oceanographic conditions, and sub-annual storms are heavily emphasized. In the Fourth Chapter, stratigraphy deposited over 10,000 year timescales is examined and therefore the geologic signature of the high-frequency processes mentioned above are subdued or outright destroyed (Jerolmack and Paola, 2010), and longer-wavelength processes such as glacial/interglacial cycles and associated isostatic effects are more prominent.

This dissertation is a sedimentological study and can therefore be placed into a “source-to-sink” context. According to this classification scheme, a study is defined by how far (spatially) it is from an initial sediment source (typically a steep mountain range within the catchment basin) and the final sink before plate subduction (typically the abyssal plain of an oceanic basin) (Helland-Hansen et al., 2016). None of the three studies included in this dissertation dwell outside the source-to-sink pathway’s center; the ICS serves as an intermediary sink before mass failure and resuspension remobilize sediment towards the outer continental shelf (Wright and Nittrouer, 1995). Deposits of sediment analyzed in these studies will either be interred into the long-term stratigraphic record of the continental shelf, or serve as an intermediary source for downslope deposition.

Summarily, this dissertation can be thematically defined by two phenomena: sediment transport and slope stability. Sediment transport is a necessarily broad theme due to the aforementioned dissimilarity between Chapters Two and Three and Chapter Four. Ideologically, sediment transport can be defined as “clastic material moving from one place to another through a medium”; it can be something of a sedimentologically catch-all term. Namely, sediment transport as it applies to Chapters Two and Three of this dissertation refers to river plume deposition, resuspension caused by tides, waves, and currents, and gravitational mass failure.
Sediment transport as it applies to Chapter Four mainly encompasses terrestrial fluvial sedimentation (overbank flooding, crevassing), deltaic progradation, and erosion via marine transgression. Sediment transport is the common thread woven through the three body chapters of this dissertation, but this thread is as ubiquitous in geology as a crawfish boil on a Southern Louisiana Spring Saturday.

Slope stability is the second unifying theme of this dissertation, but predominantly applies to Chapters Two and Three. Slope stability can be defined as a state of resisting forces exceeding destabilizing forces on an angled surface, resulting in no downslope movement. Resisting forces in a subaqueous environment including hydrostatic (water column, pore water) and lithostatic (overlying sediment) pressures. Destabilizing forces as applied to this dissertation are primarily the combination of gravitational attraction and slope angle in addition to excess sediment pore pressure (Bea and Aurora, 1981). When destabilizing forces exceed resisting forces, slope failure occurs; this can be a continuous (creep) or impulse (mass failure) process. Sediment transport and slope stability act on each other; sediment transported to a site provides lithostatic pressure that nominally stabilizes a slope, while a large volume of sediment delivered in short period results in a steep gradient that preconditions failure. Gravitational failure is one of the most important modes of sediment transport globally, since the largest mass failures can transport as much sediment in a day as the discharge of all the world’s rivers (Talling, 2014).

To summarize, this “staple dissertation” is a synthesis of three studies that broadly focus on sediment transport and slope stability. The studies are relatively narrowly geographically confined to the NGoM ICS, but temporally cover timescales from days to tens of thousands of years.
The dissertation is centered on the continental margin segment of the source-to-sink pathway, an area that serves as both an upstream sediment sink and downstream sediment source.

References cited


CHAPTER 2: SUB-DECADAL SUBMARINE LANDSLIDES ARE IMPORTANT DRIVERS OF DELTAIC SEDIMENT FLUX: INSIGHTS FROM THE MISSISSIPPI RIVER DELTA FRONT

2.1 Introduction

Submarine landslides are processes that transport sediment downslope, can dramatically alter seafloor morphology, and are important links in the global source-to-sink sediment system. Large landslides can transport more sediment in a day than a decade of global river discharge (Talling, 2014). Submarine landslides have been the subject of numerous studies because they influence many sedimentary environments worldwide. Such mass transport occurs at a range of scales on subaqueous portions of river deltas, including systems with extensive subaqueous clinothems that are shaped by waves, tides, and gravity (e.g., Amazon, Atchafalaya, Fly, Yangtze) (Denommée et al., 2016), and deltas with more proximal accumulation within river-dominated systems (Fraser, Yellow) (Walsh and Nittrouer, 2009). Submarine mass failures are also well documented on the modern subaqueous Mississippi River Delta Front (MRDF), and can potentially damage human infrastructure such as oil platforms and pipelines (Coleman et al., 1980; Hooper and Suhayda, 2005).

The MRDF is prone to submarine landslides despite its overall gentle gradient (mostly < 1.5° (Abbott et al., 1985)) due to multiple factors, including rapid deposition of relatively impermeable fine-grained sediment, abundant organic material, and subsequent biogenic gas production (Anderson and Bryant, 1990; Goñi et al., 1997). These factors promote oversteepening, hinder sediment consolidation, and can produce elevated pore pressures that precondition seabed failure. The northern Gulf of Mexico is subject to frequent tropical cyclones,
and during their passage, wave heights can exceed 15 m (Wang et al., 2005). Cyclic seafloor loading-unloading conditions can exceed yield strength of underconsolidated, gas-charged sediments, resulting in failures (Prior and Coleman, 1984). Most studies to date document seafloor movement caused by major hurricanes (Landsea, 1993) of category three or greater (Bea et al., 1971; Wang et al., 2005; Walsh et al., 2006; Hitchcock et al., 2008), with few studies addressing sub-decadal triggering events such as river floods, non-major hurricanes, and tropical storms (Son-Hindmarsh et al., 1984; Allison et al., 2005).

Analytical modeling of major tropical storm waves by Henkel (1970) and Bea and Aurora (1981) evaluated potential failures based on the assumption of sinusoidal waves. These studies did not account for effects of nonlinear waveforms, where the transition from wave crest to trough occurs across a shorter horizontal distance than in a linear waveform. To assess the magnitude of sub-decadal mass failures compared with those triggered by decadal-scale events, we use nonlinear wave modeling in combination with data from three MRDF bathymetric surveys collected during a quiescent period (with respect to tropical cyclones: October 2005-June 2014) and during one interval that captures major hurricanes (March 1979-October 2005).

2.2 Geomorphic setting

MRDF geomorphic features were first described by Coleman et al., 1980) as characteristic of river-dominated deltas: mudflow gullies are elongate seafloor depressions tens of kilometers long, hundreds of meters wide, and tens of meters high. Mudflow lobes form at downslope termini of mudflow gullies and have similar dimensions to gullies, but positive relief (Prior and Coleman, 1978).
Mudflow gullies and lobes are typically concentrated in the delta front (0.5–1.5° slope, ~5–80 m water depth); the prodelta is downslope of the delta front and has a smaller gradient and greater distance from the distributary mouth that results in far less seafloor disturbance (Coleman et al., 1980).

2.3 Methods

The study site is a ~55 km² area ~10 km southwest of Southwest Pass (Fig. 2.1). Water depth spans 15–80 m, and covers delta front and prodelta environments. Data used include: single-beam bathymetry of Coleman et al. (1980), multibeam bathymetry collected in October 2005 by Walsh et al. (2006), multibeam bathymetry collected by Fugro Geoservices Inc. in February 2009, and interferometric swath bathymetry collected for this study in June 2014. No hurricanes of category three or greater passed within 100 km of the MRDF during the 9-year period from October 2005 to June 2014 (NOAA, 2016), while the interval between the 1979 and 2005 surveys brackets two major (category 3-5) hurricanes: Ivan (2004) and Katrina (2005).

Bathymetric data were gridded to 25 m² horizontal resolution and subtracted from one another, producing Difference of Depth (DoD) grids. The cut/fill tool in ESRI ArcGIS v. 10.1 was used to assess volumetric flux between surveys.

To evaluate wave forcing on failures, a nonlinear wave model was used to propagate waves over the study area and calculate local wave properties and pressure gradients under the influence of 1-year return-period waves. Results and observations from Guidroz (2009) were used to generate Stokes waves at the marine boundary using the Fenton (1999) approach. Model results were then compared with an earlier linear-wave-model approach (Henkel, 1970). For further details, see Appendix A.
2.4 Results

MRDF sub-decadal scale mass failures transported on the order of $10^6$ m$^3$ of sediment between 2005-2009 in the study area (comparable to sub-decadal failures on other deltas worldwide, Table 1); the volume of sediment transported by major hurricanes in the study area during 1979-2005 is on the order of $10^7$ m$^3$.

<table>
<thead>
<tr>
<th>Location</th>
<th>Foreset Gradient (degrees)</th>
<th>Presumed Failure Trigger(s)</th>
<th>Time Between Surveys</th>
<th>Magnitude of Volume Transported ($10^6$ m$^3$)</th>
<th>Magnitude of Vertical Change (m)</th>
<th>References</th>
</tr>
</thead>
<tbody>
<tr>
<td>Fraser River Delta</td>
<td>1–3</td>
<td>Current undercutting, cycling loading via earthquake, storm waves leading to liquefaction</td>
<td>~2 months</td>
<td>0.075–1</td>
<td>4–12</td>
<td>(McKenna et al., 1992)</td>
</tr>
<tr>
<td>Squamish Delta</td>
<td>3–6</td>
<td>Summer freshet flood</td>
<td>24 h</td>
<td>&gt;0.02</td>
<td>1–12</td>
<td>(Hughes Clarke et al., 2014; Clare et al., 2016)</td>
</tr>
<tr>
<td>Ogooue Delta</td>
<td>3–8</td>
<td>Oversteepening</td>
<td>1 year</td>
<td>0.02–2</td>
<td>4–15</td>
<td>(Biscara et al., 2012)</td>
</tr>
<tr>
<td>Mississippi River Delta Front</td>
<td>&lt;0.5</td>
<td>River flood, tropical storm or extratropical cyclone passage</td>
<td>4 years</td>
<td>2.2</td>
<td>1–4</td>
<td>This study</td>
</tr>
<tr>
<td>Mississippi River Delta Front</td>
<td>&lt;0.5</td>
<td>Major hurricane passage (2005 Katrina)</td>
<td>24 years</td>
<td>28</td>
<td>3–12</td>
<td>(Walsh et al., 2006), this study</td>
</tr>
</tbody>
</table>

Note: Comparison of worldwide submarine landslides studied via repeat bathymetric surveys. Relevant parameters including foreset gradient, triggering factors, and volume displaced by failures are included.
Normalized for time between surveys, annual transport during a quiescent period ($5.5 \times 10^5 \text{ m}^3/\text{year for 2005–2009}$) is ~50% of that during a period that contained major hurricanes ($1.1 \times 10^6 \text{ m}^3/\text{year for 1979–2005}$). Outside mudflow features, the seafloor showed small positive elevation change (Fig. 2.2), which was generally within the uncertainty range ($\pm 0.5 \text{ m for 2009–2005 DoD}$) and rarely exceeded 1 m. The most drastic deepening (+4 m) occurred in shallow (15–40 m) parts of mudflow gullies (Fig. 2.3A), while the largest shoaling (~3.5 m) was observed in a narrow band at the mudflow lobe downslope terminus. Minimal lateral movement (<200 m) of the gully/lobe “footprint” was observed from 2005 to 2014 (Fig. 2.3B). Seafloor movement during 1979–2005 exceeded 10 m in the gully/lobe complexes and downslope progradation of mudflow lobes exceeded 1 km (Walsh et al., 2006).

**Figure 2.1** Base map showing survey area, Mississippi River Delta front, offshore Louisiana, USA. Bathymetry from Coleman et al. (1980); isobaths are at 25 m intervals. SWP—Southwest Pass; SP—South Pass; PAL—Pass A Loutre. Blue (for 2005), green (for 2009), and red (for 2014) polygons are spatial extents of bathymetric grids used in this study. Pink dashed rectangle demarcates extent of Figure 2.2.
Our modeling shows that smaller nonlinear waves (4.5–6.5 m) with a return period of one year produce pressure differentials comparable to larger hurricane waves (8.5–10.5 m) that were evaluated using the linear theory (Fig. 4b). Local wave heights remained near their boundary value (~6.5 m) and experienced little transformation in depths of >20 m, but reduced to ~6 m in depths of <20 m (Fig. 4A).

2.5 Discussion

This study confirms that appreciable MRDF seafloor movement occurred between October 2005 and June 2014 in the absence of major hurricanes. The snapshot nature of bathymetric surveys does not elucidate whether observed movement was triggered by smaller-scale impulse events, such as extratropical cyclones, tropical storms, or river floods or whether sediment exhibited continuous creep-like motion under the influence of gravity, as has been suggested (Adams and Roberts, 1993). Repeat geophysical surveys can quantify seafloor movement and numerical models can provide a simplified idea of triggering mechanisms, but in situ observation of gully/lobe rheology akin to the Fraser River Delta Observatory (Clare et al., 2016) is necessary to truly understand mudflow kinematics.

The absence of large hurricanes during the 2005–2014 period of our study (significant wave height $H_s << 10$ m) suggests that the seafloor movement observed in bathymetric data was not triggered by major hurricane waves. Simulated peak pressure differentials at the seabed ($\Delta p \approx 35$ kPa) in the study area generated by one-year waves ($H_s \approx 6.5$ m; Fig. 4B) reached similar peak conditions to those produced by simulated hurricane waves, suggesting that movement can be triggered by one-year waves if nonlinear effects are considered.
In depths of 14–50 m, one-year nonlinear waves (~6.5 m, Fig. 4A) produce pressure differentials that exceed pressure differentials of linear waves simulated for hurricanes by more than 15%.

Regardless of the exact triggering factor(s), sub-decadal scale MRDF submarine landslides mobilize volumes of sediment comparable to more catastrophic counterparts, when averaged over multi-decadal timescales. While the calculated volume fluxes are not precise due to large uncertainties and assumptions (see Appendix A), they agree with results from another recent study (Kelner et al., 2016) that indicates smaller, sub-decadal landslides provide major
forcing for shelf-to-slope sediment flux as important as larger, better-studied catastrophic landslides. Submarine landslides are also an important conduit for shelf to deep sea transport of organic carbon, heavy metals, and bioreactive particles (Panieri et al., 2012); the fact that these landslides occur more frequently than previously conceived on margins as disparate as the northern Gulf of Mexico and southern France (Kelner et al., 2016) suggests that the presently accepted flux estimates for these materials are probably incomplete in other locations as well.

Figure 2.3A Profile M-M’ across survey area showing change in gully depth between surveys. Gully depth decreases in westernmost gully, but increases in all other gullies. Figure 2.3B Width of Mississippi River delta front gully plotted against distance from origin along profile N-N’. Note this origin is not head of gully, but shallowest depth common to all three data sets. See Figure 2.2 for locations of profiles M-M’ and N-N’.
Figure 2.4A Relationship between water depth, wave height (blue), and pressure change (black) on sea bottom. Wave heights ($H_s$) plotted against water depth, with results of Henkel (1970) as solid line and this study’s results as dashed. Figure 2.4B Seafloor $\Delta P$ (pressure differential) compared to water depth, with results of Henkel (1970) as solid line and this study’s results as dashed. Evolution of pressure change from water depths between 5 and 70 m appears similar between linear sinusoidal and fully nonlinear waves.

This study documents depth change and substantial ($10^5$ m$^3$/year) volumetric flux within MRDF mudflow gully and lobe complexes during a relatively quiescent period of ocean-wave climate. These failures may be triggered by extratropical cyclones, tropical storm passage, or river floods (Prior and Coleman, 1981); nonlinear wave modeling results presented here demonstrate that storms with return interval of at least 1 year could cause such failures.

Regardless of the forcing mechanism, volumetric analysis of sediment displaced during the 2005–2009 quiescent period indicates “fair weather” subaqueous MRDF movement may be a comparably important driver of sediment transport with the better-studied major hurricane-forced failures. These findings have widespread mass-flux implications not only for the MRDF, but also for margins worldwide because the present understanding for event-scale dispersal of sediment, organic carbon, and bioreactive particles from shelf to deep sea may be skewed towards low-
frequency, high-magnitude events. These results also underscore the need for in situ monitoring programs in tandem with sub-annual repeat surveys in order to elucidate the drivers, periodicity, and magnitude of sediment flux on the MRDF and other deltas worldwide.

2.6 References cited


CHAPTER 3: MUD-CAPPED DREDGE PITS: USING A NEW COASTAL RESTORATION RESOURCE AS A PROXY FOR SEDIMENT TRANSPORT AND SLOPE STABILITY

3.1 Introduction

Anthropogenic modification of the coastal environment has occurred since at least the Bronze Age (Marriner et al., 2006). Early attempts to stabilize dynamic coastal environments came mostly in the form of “hard stabilization,” including structures such as sea walls, jetties, groins, and breakwaters. These structures can stabilize the coastline locally, but can also increase downdrift erosion (Dean, 2002). As an alternate solution, “soft stabilization” projects involve placement of beach-compatible sediment either directly on the shoreface (Park et al., 2009) or upstream of the desired restoration site (de Schipper et al., 2016). This sediment is typically dredged from an offshore depocenter, such as an ebb-tidal delta (Xu et al., 2014), or an abandoned delta distributary mouth bar (Khalil and Finkl, 2009).

The sedimentary environment offshore Louisiana is sand-poor, with little surficial fine and medium sand typically used for beach restoration projects (Buczkowski et al., 2006). The largest surficial sand resources on the Louisiana continental shelf are generally ancient drowned distributary-mouth shoals of the Mississippi River Delta, but their distances > 10 km from many eroding beaches makes such shoals expensive options for nourishment sources. Paleochannels incised during sea level lowstands and subsequently infilled with sand during sea level transgressions are widespread throughout the Northern Gulf of Mexico (NGoM) inner continental shelf (ICS) (Suter and Berryhill, 1985; Blum and Aslan, 2006) and have been targeted as alternative borrow sites (Nairn et al., 2005). The sand within these paleochannels is
typically covered with an overburden of 1-5 m of Holocene mud, which must be removed before sand can be dredged. The resulting mud-capped dredge pits (MCDPs) are typically elongate (length:width ~4:1), deep (~10 m relief) features, following paleo-river morphology.

MCDPs have only been utilized for Louisiana coastal restoration projects since the early 2000s (Nairn et al., 2006), so their post-dredging evolution is not documented or understood as well as more common sand-dominated dredge pits (SDDPs) (Benedet and List, 2008). Relevant parameters include rate of infilling, nature of infill substrate, and erosion outward from initial dredge cuts; this last parameter is relevant to infrastructure risk given the abundance of pipelines and oil rigs in the immediate proximity of MCDP target sites (Nairn et al., 2005). MCDPs also serve as “experiments of opportunity” due to their geometry: their characteristic steep slopes and large relief make them de facto failure-prone sediment traps, which can provide insights into NGoM sediment accumulation and slope stability.

Utilizing several bathymetric datasets acquired from an MCDP dredged proximal to the Mississippi River Delta Front (MRDF), we seek to answer the following questions: (1) how do MCDPs evolve geomorphically; (2) how persistent are MCDPs relative to SDDPs and what is the nature of their infill; (3) are MCDPs oversteepened relative to “natural” NGoM geomorphic features; and (4) what can MCDPs tell us about NGoM sediment dynamics and accumulation?

3.2 Setting

The MCDP analyzed in this study is Sandy Point Southeast dredge pit (hereafter referred to as Sandy Point) and is located 20 km west of the Mississippi Birdsfoot Delta in a water depth of 11 m (Fig. 3.1). A total of $3.7 \times 10^6$ m$^3$ of sand was removed in November 2012, and the muddy overburden was disposed about 1.5 km to the east. Sandy Point is not located directly
proximal to any major Mississippi River distributaries, but is within the intermittently active clockwise gyre of the Louisiana Bight that advects the river plume over Sandy Point (Walker et al., 1996). Oceanographic conditions on the Louisiana ICS are generally mild, with a small (<1 m) semidiurnal or diurnal tidal range and a significant wave height rarely exceeding 3 m (Georgiou et al., 2005).

3.3 Methods

This study primarily utilizes bathymetric data to qualify and quantify the evolution of Sandy Point MCDP post-dredging. Data from five bathymetric surveys were utilized, including two pre-dredging (2003, 2011), and three post-dredging (2012, 2013, 2015). Two bathymetric datasets from the nearby MRDF were also utilized, one collected by Walsh et al. (2006) and one collected by Fugro Geoservices. These data were collected by a range of vessels, sonars, and positioning equipment. The uncertainty inherent in comparing continuous bathymetric datasets (2012 and 2015) acquired using different equipment was quantified utilizing the “fixed reference uncertainty” method described in Schimel et al. (2015). All geomorphic analyses, including Difference of Depth (DoD), volumetric calculations, and gradient analysis, were conducted using ESRI ArcGIS version 10.1; further details of processing and analysis can be found in Appendix B.

3.4 Results

3.4.1 Sandy Point infilling

As of May 2015 (approximately 2.75 years post-dredging), Sandy Point MCDP is still a prominent bathymetric basin on the otherwise smooth NGoM seafloor (Fig. 3.1A). The maximum pit depth is approximately 20 m, as compared to the ~11 m depth of the surrounding
seafloor. One-dimensional transects and DoD derived from repeat bathymetric surveys show trends in Sandy Point infilling (Figs. 3.1B, 3.2). The 2003 and 2011 profiles demonstrate that the seafloor was in an apparent equilibrium pre-dredge, with minimal depth changes (Fig. 3.2). Post-dredging, the pit accumulated sediment at an average rate of 54 cm/year, with depositional hotspots of >1 m/year at the deepest parts of the pit (Fig. 3.1B, 3.2) and adjacent to wall failures (Figs. 3.1B, 3.2A).

Post-dredging volumetric analysis indicates that Sandy Point is presently infilling at an average rate of approximately 200,000 m$^3$/year (Table 3.1). The volume of accretion measured between the 2013-2015 surveys is approximately double that of the 2012-2013 survey, so the infilling rate appears to be fairly stable on an annual scale. Dividing the dredged volume (3.74 x $10^6$ m$^3$) by this infill rate yields a projected time to complete filling of ~15 years, if the rate remains constant. Localized volume loss was observed and primarily confined to the pit walls. Total wall volume loss between the 2012 and 2015 surveys was ~55,000 m$^3$, or <10% of infill volume (Table 3.1).

3.4.2 Sandy Point wall gradient change compared to natural features

Between 2012 and 2015 surveys, the floor of Sandy Point MCDP became smoother (Fig. 3.3, Table 3.2) as sharp gradients associated with dredging were mantled with sediment. All pit walls lost steepness in the three years since dredging, although the western wall was initially less steep than the eastern wall (Table 3.2). MCDP gradient changes can be compared with a similar analysis of an MRDF mudflow gully that was surveyed in 2005 (post-Hurricane Katrina) and 2009 (Fig. 3.4). Changes of MRDF gully walls are smaller than Sandy Point, but an opposite trend is apparent in that the gully walls steepened between 2005 and 2009.
Figure 3.1A Basemap with 2015 hill-shaded bathymetry. Contours are in 2 m intervals. Cross sections correspond to Fig. 3.2. Inset shows locations of Sandy Point (red square) and mudflow gully (blue square, Fig. 3.5). Figure 3.1B Difference of Depth between 2012 and 2015 surveys. Red values indicate deepening, while blue values indicate shoaling. Yellow values are within the 2σ range of uncertainty (0.2 m) and are therefore considered no significant change.

Table 3.1 Sandy point volumetric changes quantified via cut/fill analysis. Line spacing was not dense enough in the 2013 survey to measure wall volume losses. Uncertainties based on the “fixed reference uncertainty” method documented in Schimel et al. (2015).

<table>
<thead>
<tr>
<th></th>
<th>Pit infill (m$^3$)</th>
<th>Western Wall Volume Lost (m$^3$)</th>
<th>Eastern Wall Volume Lost (m$^3$)</th>
<th>Pit infill – Wall volume lost (m$^3$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2012-2013</td>
<td>257,630 ± 98,294</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>2013-2015</td>
<td>452,530 ± 112,543</td>
<td>N/A</td>
<td>N/A</td>
<td>N/A</td>
</tr>
<tr>
<td>2012-2015</td>
<td>616,550 ± 87,204</td>
<td>19,140 ± 9,924</td>
<td>29,137 ± 9,433</td>
<td>563,343</td>
</tr>
</tbody>
</table>
Wall slope bulk statistics (Table 3.2) show that MCDP walls are generally steeper than MRDF gully walls (6.9° average of all slopes for Sandy Point vs. 4.3° for MRDF gully). MCDP walls decreased an average of 3.8° in 3 years, while MRDF gully walls increased an average of 0.6° in 4 years. The pit floor decreased in mean gradient by ~0.5° between 2012 and 2015. It should be noted that the standard deviation of all pit wall measurements exceeded 3°, reflecting the large variability of slope values (Figs. 3.3 and 3.4).
Table 3.2 Bulk statistics (mean ± gradient uncertainty, standard deviation) of MCDP wall and floor gradients. Wall boundaries shown as black dashed polygons in Figs. 3.3 and 3.4. See Appendix B for derivation of gradient uncertainty.

<table>
<thead>
<tr>
<th>Location</th>
<th>Mean (Degrees)</th>
<th>Standard Deviation (Degrees)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sandy Point 2012 West</td>
<td>5.7 ± 0.47</td>
<td>4.8</td>
</tr>
<tr>
<td>Sandy Point 2015 West</td>
<td>4.4 ± 0.47</td>
<td>3.1</td>
</tr>
<tr>
<td>Sandy Point 2012 East</td>
<td>9.7 ± 0.47</td>
<td>6.3</td>
</tr>
<tr>
<td>Sandy Point 2015 East</td>
<td>7.9 ± 0.47</td>
<td>6.3</td>
</tr>
<tr>
<td>Sandy Point 2012 Pit Floor</td>
<td>1.38 ± 0.47</td>
<td>1.55</td>
</tr>
<tr>
<td>Sandy Point 2015 Pit Floor</td>
<td>0.89 ± 0.47</td>
<td>1</td>
</tr>
<tr>
<td>MRDF Gully 2005 West</td>
<td>3.8 ± 0.76</td>
<td>3.3</td>
</tr>
<tr>
<td>MRDF Gully 2009 West</td>
<td>5.1 ± 0.79</td>
<td>4.4</td>
</tr>
<tr>
<td>MRDF Gully 2005 East</td>
<td>3.7 ± 0.76</td>
<td>3.2</td>
</tr>
<tr>
<td>MRDF Gully 2009 East</td>
<td>4.1 ± 0.79</td>
<td>3.9</td>
</tr>
</tbody>
</table>

Longitudinally averaged gradient plots (Fig. 3.5) reinforce trends observed in the bulk statistics and also show spatial differences in gradient changes. The largest Sandy Point gradient changes were concentrated in the areas with the steepest slopes (Figs. 3.5A and 3.5B). Conversely, the flattest areas of the MRDF gully walls showed the largest slope increase (Figs. 3.5C and 3.5D). The eastern wall of the MRDF gully showed a larger magnitude of slope increase than the western wall between the 2005 and 2009 surveys.
3.5 Discussion and interpretation

3.5.1 Sandy Point infilling compared to other dredge pits

The amount of sediment removed from Sandy Point and the rate of infill can be compared with published results from dredge pits located in different geographic and geological settings (Fig. 3.6). These dredge pits were SDDPs predominantly excavated in ebb tidal shoals and located along the U.S. eastern seaboard and Gulf coast. The values for other dredge pits were derived from a number of observational and numerical modeling methodologies, the details and uncertainties of which can be found in Appendix B.

Figure 3.3A Gradient map of the Sandy Point dredge pit derived from 2012 bathymetry. Green colors represent flatter surfaces, while red colors indicate steeper surfaces. Dashed boxes are
The time it takes dredge pits to fill level with the seafloor is positively correlated with the initial volume excavated (Fig. 3.6). The volume of sediment excavated from dredge pits is weakly positively correlated ($R^2 = 0.5$) with the average time elapsed for complete infill (Fig. 3.6). The range of time to complete infill spans from half a decade for small ($4 \times 10^5 \text{ m}^3$) volumes to hundreds of years for large excavations ($9 \times 10^6 \text{ m}^3$); the upper end of this range is probably unrealistic due to simplistic modeling parameters in Byrnes (2004a) and Byrnes (2004b).
Sandy Point’s projected time to complete infilling (~15 years) is shorter than predicted assuming a linear correlation between volumes excavated and infilling time; many factors likely control the time to complete infilling but proximity to sediment sources and transport are likely important variables.

Figure 3.5A North-South Latitudinal plot of longitudinally averaged gradient change between 2012 and 2015 surveys, Sandy Point west wall. Figure 3.5B N-S Latitudinal plot of longitudinally averaged gradient change between 2012 and 2015 surveys, Sandy Point east wall. Figure 3.5C N-S Latitudinal plot of longitudinally averaged gradient change between post-Katrina 2005 and 2009 surveys, Mississippi River Delta front mudflow gully east wall. Figure 3.5D N-S Latitudinal plot of longitudinally averaged gradient change between post-Katrina 2005 and 2009 surveys, Mississippi River Delta front mudflow gully west wall.

Pit surface area and orientation relative to net current direction(s) is another important predictor of infilling time (Kennedy, 2010). The geometries of the pits shown in Fig. 3.6 are largely similar (wide in the cross-shore dimension, narrow in the long-shore dimension, (Cialone, 1998)). Sandy Point is predicted to completely fill in ~15 years, while dredge pits with comparable volumes excavated in areas such as Egg Harbor Inlet and Mobile Bay are predicted to exist for decades to hundreds of years due to less sediment supply and movement (Byrnes, 2004a, b). Some of this disparity can be attributed to model oversimplification (for example, the
models used in the Egg Harbor/Mobile Bay studies do not take wall failure contribution to infill into account), but Sandy Point’s faster infilling is likely not solely a product of different study methodologies. A possible contributor to Sandy Point’s relatively rapid infilling is proximity to a massive sediment source in the Mississippi River; all other referenced dredge pits were set in sediment-starved environments without nearby sediment supply from major distributaries. This implies that while initial volume excavated may be the primary controller of dredge pit persistence, proximity to a fluvial sediment source plays an important role as well.

Figure 3.6 Initial volumes excavated from dredge pits versus projected time to complete infilling (based on annual average infilling rate obtained by observational and/or numerical modeling methods; triangles represent hybrid modeling/observation studies, circles are pure observational studies). All dredge pits except Sandy Point were dredged in sand-dominated, sediment-starved conditions. Linear regression fit line is shown both with (solid) and without (dashed) Mobile Bay dredge pit, due to its outlier nature. Inset map shows site locations and is modified from the National Geographic world layer package.

3.5.2 MCDPs as sediment traps

The rapid rate of Sandy Point accretion (~54 cm/year) indicates it functions as a sediment trap on annual timescales. The exact mechanisms that cause Sandy Point to rapidly infill are still being investigated, but the general principle hypothesized in previous dredge pit studies is the increased water depth over the pit 1) decreases current competence due to reduced flow velocity,
and 2) insulates the pit floor from resuspension processes such as storm waves and tidal currents due to deeper depths (Nairn et al., 2005). This second factor was observed by Bokuniewicz (1986), who noted that the salinity within ~10 m deep New York Harbor dredge pits was relatively invariant despite being within the estuarine tidal prism.

Sandy Point accretion can be compared to published short and long-term Louisiana ICS sediment accumulation rates to quantify the efficiency of MCDPs as sediment traps. Beryllium-7 (7Be) radioisotope activity (T1/2 = 53 days) can be used as a proxy for seasonal-scale (<1 year), terrestrially sourced sedimentation; 7Be-derived sedimentation rates proximal to the MRDF range from 11-48 cm/year (Corbett et al., 2004; Keller et al., 2016). 7Be penetration depths observed in a companion Sandy Point coring study documented 12-34 cm of deposition within ~100 days, equivalent to 43.8-124.1 cm/year (O'Connor, 2017). The lower estimate of 7Be-derived infilling agrees well with the repeat bathymetric survey-derived infilling rate of 54 cm/year.

Decadal to century-scale sedimentation rates can be approximated via Lead-210 (210Pb) radioisotope activity, which is a longer-lived (T1/2 = 22.3 years) radioisotope that adsorbs to fine sediments as they settle through the water column (Muhammad et al., 2008). Sedimentation rates derived from 210Pb proximal to the study area are 1.3-7.9 cm/year (Corbett et al., 2006; Keller et al., 2016). The disparity between seasonal and decadal-scale accumulation rates highlights the importance of sediment resuspension in shallow environments, even in settings without strong oceanographic forcing such as the NGOM. Annual-scale sediment accumulation within Sandy Point MCDP is 1.1-4.8 times and 7-41 times greater than 7Be-derived seasonal and 210Pb-derived decadal Louisiana ICS sediment accumulation, respectively; regardless of mechanism, MCDPs appear to be efficient sediment traps.
3.5.3 Geomorphic model of MCDP evolution

A generalized equation for MCDP infilling can be expressed as:

\[
\frac{\partial z}{\partial t} = (s_{near} + s_{far})^q/t - e_{z90}
\]

where \(\frac{\partial z}{\partial t}\) is the change of pit depth with change in time, \(s_{near}\) is vertical accretion from locally sourced sediment (wall failure), \(s_{far}\) is vertical accretion from far-field sourced sediment (river plume deposition and shelf resuspension), \(q\) is a consolidation coefficient (with consolidation \(q\) growing small as time \(t\) grows large), and \(e_{z90}\) is erosion from currents, waves, and tides when the pit is \(~90\%\) infilled (Nairn et al., 2005). Bathymetric time-series observations cannot discern individual components of the above equation such as consolidation rate and erosion (unless it exceeds deposition, and then only as a minimum estimate), so a conceptual model is presented here to represent the relative impact of these terms on MCDP evolution post-construction (Fig. 3.7). This model builds on the findings of Nairn et al. (2005), who conducted observational and numerical model-based studies of completed and planned MCDPs, including Sandy Point.

The first phase of MCDP evolution occurs immediately after dredging is finished and likely takes place on sub-annual timescales (Fig. 3.7A). During this period, it is difficult to gather observational data because the first post-dredging survey typically does not occur until \(~1\) month after completion. In this phase, locally derived sediment dominates pit infill as steep pit walls slump and fail to achieve short-term equilibrium with seasonal or shorter-timescale forcings (e.g., tidal fluctuations, river floods, and wind-generated waves). After initial rapid failing and stabilization of the pit walls, far-field sedimentation dominates infill volumetrically (Fig. 3.7B); this stage begins months after dredging and persists until the pit is almost entirely
infilled. Sandy Point appears to be currently in the second phase, because over 90% of the infill volume between the 2012 and 2015 surveys was far-field derived. Consolidation is also likely a prevalent factor in pit elevation, as freshly deposited sediments are most prone to compaction (Tornqvist et al., 2008).

The third and final phase of MCDP evolution begins when the pit floor is sufficiently close to the ambient seafloor to be “recoupled” to oceanographic forcings. When this occurs pit infill can be resuspended and transported out of the pit, and if a net current direction exists the downstream pit wall will be eroded (Nairn et al., 2005). These processes retard far-field deposition and increase local infill contribution via wall scouring and undercutting (Fig. 3.7C); this also causes infill rates to slow dramatically as pits reach full capacity (Lu and Nairn, 2011).

3.5.4 MCDP and mudflow gully gradient comparison

High-resolution repeat bathymetric surveys allow fine-scale morphological changes to be quantified. It is instructive to compare gradient changes of MCDPs such as Sandy Point to natural negative relief features found in a similar environment, such as MRDF mudflow gullies. These elongate (>> 10:1 length:width ratio), shallow (relief < 10 m) depressions are a product of rapid sedimentation and gravitational failure and are persistent over decadal timescales (Coleman et al., 1980; Obelcz et al., 2017). Spatial differences and temporal changes in morphology provide valuable information regarding depositional and erosional processes.
Figure 3.7 Conceptual block diagram showing general evolution model of a mud-capped dredge pit situated proximal to a fine-grained sediment source. Model and graphs are dimensionless. Figure 3.7A First phase of evolution, which is hypothesized to take place on the timescale of hours to one year following dredging. In this phase, pit walls rapidly equilibrate with sub-annual scale forcings, including tidal fluctuations, river floods, and storms. This results in the initial infill to be composed largely of wall failure material. Figure 3.7B In the second phase of evolution (1 to 10 years), pit walls are in relative equilibrium with sub-annual recurrence-interval forcings, and are generally much less prone to failure than in the first stage. In this stage, river and inner continental shelf resuspension-derived sediment dominates the pit infill. Figure 3.7C In the final stage (10 to 15+ years), the floor of the dredge pit has infilled to the point it is “recoupled” with oceanographic forcings, and is subject to resuspension. This recoupling also results in downstream pit wall scour, contributing more local sediment to infill. This stage is typified by drastically slower rates of infill, and can potentially last much longer than the first two.
Sandy Point’s eastern wall’s steeper gradient is likely associated with construction geometry (cannot be verified because the 2012 survey did not take place until 1 month after dredging, during which rapid wall changes likely had occurred, Fig. 3.7A); the faster loss of steepness and greater volume loss (1.8° and ~30,000 m³ east, 1.3° and ~20,000 m³ west) is likely associated with steeper slopes equilibrating at a faster rate. The spatial distribution of Sandy Point gradient change reinforces this inference because the steepest sections of wall decrease by the largest margin, while gentler sections better maintain their gradients (Fig. 3.6). Asymmetrical gradient change may be related to initial conditions and oceanographic conditions; Nairn et al. (2005) conceptualized that the downstream pit wall will experience stronger currents due to flow constriction associated with the rapid change of water depth, which may promote undercutting and wall failure (Fig. 3.7C).

Naturally formed MRDF gully walls do not show an apparent east-west gradient asymmetry, which reinforces the notion that asymmetric Sandy Point walls are a product of initial dredging geometry. MRDF gully wall gradients are also universally gentler than MCDP gradients (4.2° and 7° averaged across all walls and surveys, respectively). The largest apparent difference between MRDF and MCDP walls in a 3-4 year time window is the opposite temporal trends-MRDF gully walls appear to steepen with time, while MCDP walls become gentler (Fig. 3.5). Furthermore, it appears the gentlest sections of walls in the initial 2005 survey steepen to the greatest extent (Figs. 3.5C, D). It should be noted that both of these trends only apply to small spatial and temporal samples, and may not reflect longer-term (decadal-scale) gradient trends.
Differences between artificial MCDPs and natural MRDF gullies in terms of geology, oceanography, and surveying intervals must be recognized before interpretations can be made. The MRDF seaward of distributaries is predominantly composed of muddy substrate derived from plume settling and downslope mass failures (Keller et al., 2016), while MCDPs have a bipartite composition of a 1-5 m mud cap overlying primarily Pleistocene paleochannel sand (Suter and Berryhill, 1985). This lithological difference may affect slope stability, if the cohesive sediment cap on top of noncohesive MCDP sand does not have the same mechanical properties as relatively homogenous MRDF mud. The mudflow gully surveyed is also directly offshore of Southwest Pass of the Mississippi Delta, which is currently the most active distributary of the Mississippi River in terms of discharge (Allison et al., 2012).

Sandy Point apparently receives an appreciable quantity of Mississippi-derived sediment, but the greater sedimentation rates directly offshore Southwest Pass may result in constantly oversteepening and failing slopes which are not found outside the direct zone of greatest sediment input (Obelcz et al., 2017). Finally, the survey time windows were different (post-Hurricane Katrina 2005 to 2009 for MRDF, 2012-2015 for Sandy Point), but the generally quiescent NGOM hurricane history since Katrina means oceanographic conditions should have been relatively similar between surveys (NOAA, 2016).

If the gradient disparity between Sandy Point and MRDF mudflow gully is not entirely a product of the various differences described above, the convergence of gully and MCDP gradients towards a common value (4-5°) could represent an equilibrium condition between slope and annual recurrence interval events, such as tidal fluctuations, river floods, cold fronts, and tropical storms. A slope will remain stable under wave loading as long as sediment shear strength exceeds shear stress, which Seed and Rahman (1978) express as:
\[
\frac{\tau_w}{\gamma' h} = \sin \alpha + \frac{\pi \gamma_w H}{\gamma' L \cosh \frac{2\pi d}{L}}
\]

where \(\tau_w\) is wave-induced shear stress, \(\alpha\) is slope angle, \(\gamma_w\) is the unit weight of water, \(H\) is wave height, \(L\) is wavelength, and \(d\) is water depth. Of these variables, the one that most directly affects slope stability (and is shown to differ between Sandy Point and the gullies, Table 3.2) is angle \(\alpha\).

This hypothesis could be tested through further repeat surveying of Sandy Point, assuming the pit is not completely infilled before the walls reach an equilibrium slope. If the average rate of slope decrease observed (~1.4°/year) remains constant, equilibrium (no change in slope) should occur before the pit is completely infilled (~15 years). Conversely, if this hypothesis is valid, MRDF mudflow gully gradients should remain relatively constant or decrease after exceeding a threshold value. Knowledge regarding an empirical “angle of repose” in this substrate and setting has a value to future basic and applied research, in that it will inform morphological evolution models and provide a quantitative metric of geohazard risk.

3.6 References cited


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CHAPTER 4: SHALLOW STRATIGRAPHIC AND SURFICIAL ANALYSIS OF A WELL-PRESERVED PLEISTOCENE DROWNED FOREST: HOW DOES IT FIT INTO THE NORTHERN GULF OF MEXICO GEOLOGIC CONTEXT AND HOW WAS IT PRESERVED?

4.1 Introduction

4.1.1 Geologic setting

The Northern Gulf of Mexico (NGoM) passive continental margin is often characterized from a depositional perspective, in which sediment sources (rivers) define the lateral boundaries between provenances (Anderson et al., 2004). The second-easternmost of these provenances is known as the Mississippi-Alabama-Florida (MAFLA) sand sheet, and is bounded by the St. Bernard lobe of the Mississippi River Delta to the west and the carbonate-ramp platform of the Florida Peninsula to the east (Doyle and Sparks, 1980). The MAFLA region’s Quaternary evolution is largely tied to retention of sandy sediment discharged by small rivers, including the Pearl, Mobile, and Apalachicola (Anderson et al., 2004). During sea level highstands these rivers’ sediment discharge is mostly confined to bays and the inner continental shelf, but during glacial periods the depocenters shift seaward with falling sea level (Van Wagoner et al., 1988).

Offshore of Alabama, the most prominent geological features are the paleovalleys incised by the Pascagoula (west) and Mobile (east) rivers during glacial intervals (Fig. 4.1). These valleys were initially carved during Marine Isotope Stage (MIS) six or eight, and were reoccupied by their respective rivers during the Last Glacial Maximum (LGM) approximately 21,000 years before present (ybp) (Bartek et al., 2004). There is no surficial expression of the paleovalleys since they were infilled during and possibly after sea level transgression with a combination of deltaic, estuarine, and marine sediments (Figs. 4.1, 4.2; Kindinger et al. (1994)).
4.1.2 Drowned forest

This study focuses on a well-preserved drowned forest discovered in 2005 approximately 12 km offshore Gulf Shores, AL shortly after the passage of 2004 Hurricane Ivan. The eye of this category three hurricane passed within 10 km of the forest, so it is likely that intense storm-
generated waves scoured the overburden that had preserved the forest, exposing it to the water column (Zambon et al., 2014). The site became known to academic researchers around 2010, after which reconnaissance coring and multibeam bathymetric surveying was conducted (Raines, 2012). In 2015 and 2016, a full campaign of vibracoring and shallow (penetration < 30 m, Fig. 4.2), high-resolution (vertical resolution ~ 30 cm) geophysical surveying was conducted at and around the exposed forest (Fig. 4.3).

Figure 4.2A Uninterpreted boomer seismic profile from Kindinger et al. (1994). Location corresponds to green line in Fig. 4.1. Figure 4.2B Interpreted boomer seismic profile modified from Kindinger et al., (1994). Features that correspond with reflectors observed in 2015-16 Chirp data (blue dashed line indicates approximate maximum Chirp subbottom penetration) include water bottom with ridge-and-trough morphology (yellow line), base of Holocene sand sheet (green line), and late Pleistocene reflectors (red lines).

The drowned forest is subaqueously exposed over ~30,000 m² of the seafloor and is predominantly composed of bald cypress (*Taxodium distichum*). These species are intolerant to salt intrusion and thrive in subtropical environments similar to the contemporary NGoM coast (Conner and Toliver, 1990). The site straddles the lateral boundary between the Mobile
paleovalley and its western interfluve (Fig. 4.1). Radiocarbon age estimates from forest wood indicate the trees grew approximately 41,830±880 ybp (Gonzalez, 2016), but since this age is close to the upper limit of radiocarbon detection and other samples were radiocarbon dead it should be treated as a minimum estimate (Taylor, 1997).

Figure 4.3 Three-dimensional digital elevation model (DEM) of 2015 and 2016 bathymetric data collected at the drowned forest site. Red dashed outline shows the seafloor depression where trees are exposed to the water column, and black dashed outline indicates an artifact from merging the 2015 and 2016 datasets.

Flora that have specific temperature, precipitation, and salinity requirements are useful as both biostratigraphic and paleoclimatic indices (Hautevelle et al., 2006). This information is usually only conveyed through seeds or plant debris, because some combination of aerobic/anaerobic decomposition, early diagenesis, subaerial erosion, and sea level transgression physically and chemically degrades specimens over the course of eustatic sea level fluctuation (Damste et al., 2002). The aforementioned drowned forest gives a rare opportunity to assess an
entire paleoenvironment more or less in situ, and we hope to use this opportunity to: 1) place the forest within the established geological context of the MAFLA provenance, 2) use a high-resolution geophysical dataset nested within existing lower-resolution regional surveys (Kindinger et al., 1994; Bartek et al., 2004) to expand the Late Pleistocene and Holocene MAFLA state of knowledge, and 3) develop a working hypothesis of forest growth, burial, preservation, and exposure that can be utilized to identify potential analogous sites.

4.2 Methods

This study largely utilizes the geophysical data obtained during the 2015-2016 surveys. Two sonars were deployed: an EdgeTech 512i (Chirp subbottom, ~0.3 m vertical resolution, 1-2 m horizontal resolution) and an EdgeTech 4600 (coregistered swath bathymetry and sidescan sonar, ~0.25 cm vertical resolution, 1 m² horizontal resolution). Line spacing was 75 and 100 m in strike and dip orientation, respectively, yielding continuous bathymetric digital elevation models (DEMs) and backscatter maps, and a dense grid of 2-D seismic reflection profiles imaging the top 10-20 m of the subsurface. Surficial data (bathymetry, backscatter) were processed with Caris HIPS and SIPS v. 9.0 and imported into ESRI ArcGIS for analysis, while subbottom data were processed using SIOSEIS v. 2014.2.1 and imported into IHS Kingdom Suite and QPS Fledermaus for analysis and interpretation.

Interpretation of the geophysical data described above was aided by vintage seismic data (Fig. 4.2) collected in the 1990s by the U.S. Geological Survey and preliminary interpretation of vibracores collected during the 2015-2016 field campaigns.
These cores ranged in length from 0.5 – 5 m, and were analyzed for gamma density, lithology, grain size, and loss on ignition (Gonzalez, 2016). The cores have also been prepared for future optically stimulated luminescence (OSL) dating, which should provide reliable dates to constrain forest chronology.

4.3 Results

4.3.1 Seafloor and uppermost stratigraphy

Bathymetry around the exposed forest region is similar to that described in prior regional studies (Dufrene et al., 2003); the most prominent features are a series of northwest-southeast trending ridges and troughs with 2-5 m vertical relief and ~0.5 km wavelength (Fig. 4.3). The drowned forest is located within a hole (~2 m relief, ~100 m wide, <10° walls) nested in a trough (Fig. 4.3). Seafloor backscatter is uniformly lower within the troughs compared to the ridges (Fig. 4.4A), and particularly low within the hole where tree stumps are exposed. Individual tree stumps can be identified via sidescan sonar, and appear to have lower reflectance than the surrounding seafloor (Fig. 4.4B).

The uppermost stratigraphic layer present across the entire survey site except where the forest is exposed is a sand sheet. Lithology indicates this layer is predominantly well-sorted medium to fine sand and shell hash (Gonzalez, 2016). The sand varies in thickness between 0 and 5 m, and is thickest on ridges and thinnest or absent in troughs (Fig. 4.5). Bathymetry and sand sheet thickness generally correlate very well, wherein shoalest areas have the thickest deposits and deepest areas have the thinnest. This indicates strata underneath the sand sheet are relatively flat-lying, and sand distribution is not controlled by precedent stratigraphy.
Figure 4A Sidescan sonar mosaic illustrating seafloor sediment texture variation at drowned forest survey site. Light and dark colors indicate higher (generally coarser sediment) and lower (generally finer sediment) backscatter, respectively. Backscatter is lowest between ridges (dashed black lines) and particularly where trees and particularly where Pleistocene swamp sediments are exposed to the seafloor (purple box, extent of Fig. 4B). Histograms quantitatively demonstrate backscatter difference between ridges (top left) and troughs (bottom right). Figure 4B Sidescan sonographs of trees exposed to the seafloor (purple boxes). Also apparent is the bathymetric depression the exposed trees lie within and around (low backscatter dark tones). Left and right panels are from the same approximate area (purple box on Fig. 4A) but acquired on adjacent tracklines.
4.3.2 Lower stratigraphy

The dense Chirp subbottom grid can be projected into 3-D space and viewed as a “fence diagram” to aid the analysis and interpretation (Fig. 4.6). The Chirp data in general provides higher vertical resolution than previous surveys, but do not resolve strata > 30 m below the subsurface, likely due to the acoustically reflective surficial sand sheet (Fig. 4.4). Based on seismic facies and stratal geometry, the subsurface is divided into two units, unit 1 and unit 2, from deepest to shallowest.

![Figure 4.5A Isopach map of Holocene sand sheet thickness, derived from Chirp seismic data. Blue indicates thickest sand deposits, while red indicates thin or absent cover.](image1)

![Figure 4.5B 2015-16 bathymetry, which demonstrates that the Holocene sand sheet is not largely influenced by precedent stratigraphy.](image2)

Unit 1 (red annotation, upper bounded by green horizon in Figs. 4.6 and 4.7) extends from the deepest resolvable reflectors to 0-5 m below the seafloor, and is characterized by concordant, subparallel reflectors that dip S-SW (Fig. 4.6). The dip angle of these reflectors
generally decrease from east to west (particularly flat-lying reflectors underlay the exposed drowned forest region), with deepest resolvable reflectors in the east displaying clinoform geometry (Fig. 4.7). Localized cross-cutting is apparent within unit 1 (blue annotation in Fig. 4.6), and regional but discontinuous unconformities truncate dipping strata from overlying flat-lying reflectors (Fig. 4.7). Unit 2 (bounded below by green horizon and above by yellow horizon in Figs. 4.6 and 4.7) varies in thickness between 0-5 m, and overlies a strong and laterally continuous reflector. Unit 2 has no internal reflectors and is acoustically homogeneous.

Figure 4.6 Three-dimensional fence diagram of interpreted Chirp data collected at the drowned forest site in 2015-16. Aspect is looking north, vertical exaggeration 10x. Gold ellipse represents area where trees are subaqueously exposed. Annotated reflectors are: water bottom (yellow), base of Holocene sand sheet (green), paleochannels (blue), and Pleistocene surfaces (red). Pleistocene strata generally dip to the south/southeast, with dip angle decreasing both east-west (reflectors in the forest area are relatively flat-lying) and deep-shallow. Pleistocene reflectors are incised by paleochannels and truncated by base of Holocene sand sheet.

4.4 Discussion

4.4.1 Facies identification via lithology and seismic interpretation

Lithology, gamma density, and organic matter analysis is described in Gonzalez et al. (2016). Cores displayed four different lithofacies: (1) fine-grained muddy sediments with abundant organic matter and wood chips (swamp), (2) highly weathered and oxidized,
compacted fine-grained sediments with shell fragments (paleosol), (3) interbedded fine-grained and sandy sediments (transitional), and (4) fine to medium sand with shell hash (transgressive/Holocene sand sheet). Swamp and paleosol are generally deepest in cores, transitional is an intermediate depth layer (when present), and sand sheet is a surficial layer.

Figure 4.7 Three-dimensional fence diagram of interpreted Chirp data collected at the drowned forest site in 2015-16. Aspect is looking south, vertical exaggeration 10x. Gold ellipse represents area where trees are subaqueously exposed. Annotated reflectors are: water bottom (yellow), base of Holocene sand sheet (green), paleochannels (blue), and Pleistocene surfaces (red). Note steeply dipping clinoform geometry of Pleistocene reflectors in the northeast quadrant of the survey area, as compared to relatively flat-lying western reflectors in the exposed forest area.

Based on previous research and new Chirp data (Figs. 4.2 and 4.7; (Kindinger, 1988; Bartek et al., 2004)), the dipping reflectors underlying the entire survey area (blue shaded regions in Figs. 4.8 and 4.9) are interpreted to be bay-head delta deposits (although the geometry could also indicate fluvial lateral accretion deposits, Plink-Björkland, 2005). This interpretation cannot be ground truthed because no cores penetrated this sequence, but is based on sigmoidal reflector geometry (Figs. 4.7 and 4.9) and overlying terrestrial facies. These deposits are associated with sea level transgression, wherein valleys incised across the shelf during sea level lowstands become deltaic depocenters as the coastal zone (and locus of deposition) moves landward with rising sea level.
(Mars et al., 1992). The steeper-dipping reflectors to the east are logical given the site’s location on the western edge of the Mobile paleovalley; steeper gradients would be expected as the axis of the paleovalley (greatest accommodation space) is approached.

Figure 4.8 Uninterpreted (top) and interpreted (bottom) chirp subbottom profile dip-oriented across the subaqueously exposed drowned forest (see Fig. 4.3 for location). Holocene sand sheet (red) is thin or absent near the forest (green); the forest area presents as a negative amplitude anomaly, but individual trees cannot be identified likely due to being at or below the horizontal resolution threshold of the system. Below the sand sheet is undifferentiated Pleistocene terrestrial deposits (which may include swamp, paleosol, or floodplain facies, yellow), and below that bayhead delta facies with clinoform geometry (solid black lines). Gamma density profile of core acquired in the vicinity of chirp profile shown to right; tan indicates sand sheet facies and green indicates swamp facies.
Bay-head facies are overlaid by terrestrial facies. These include the swamp facies in which the drowned forest is situated (Fig. 4.8), and paleosol in the eastern survey area (Fig. 4.9). There is an erosive unconformity (which forms a reflector of variable amplitude) at the base of the swamp/paleosol strata that truncates the bay-head facies; this is interpreted as the transition between marine/estuarine (bay-head delta) and terrestrial (swamp/paleosol) environments. The swamp and paleosol lithofacies are stratigraphically coeval (Figs. 4.8 and 4.9); the lateral boundary between them is not apparent in the seismic data.

The penultimate stratigraphic layer is transitional facies, which is not ubiquitous throughout the survey area. Where it is present, it lies in between swamp/paleosol and sand sheet facies, and is lithologically a mix of interbedded muds and sands. The transitional layer is seismically indistinguishable from the underlying terrestrial facies. This layer is interpreted to be a transition between terrestrial or potentially estuarine and marine environments, hence the nomenclature. The bay head, swamp, paleosol, and transitional facies combine to compose seismic unit 1 (Figs. 4.6 and 4.7).

The uppermost layer is the transgressive/marine MAFLA sand sheet, or seismic unit 2. Previous studies have speculated whether the MAFLA sand sheet was entirely formed during and after sea level transgression, or whether bathymetric ridges (Fig. 4.3) are partially or entirely relict Pleistocene barrier island complexes (McBride et al., 1999). Due to the lack of internal stratigraphy (Fig. 4.9) characteristic of barrier islands (washover fans, seaward progradation, Aksu et al. (1999)), we interpret the sand sheet in this region to be a syn or post-transgressive feature.
Figure 4.9 Uninterpreted (top) and interpreted (bottom) strike-oriented chirp subbottom profile taken across the largest sand ridge in the survey area (see Fig. 4.3 for location). Note lack of internal stratigraphy within sand sheet (red), which indicates these features are entirely reworked during or after transgression, as opposed to relict Pleistocene beach ridges or barrier islands. Yellow layer is undifferentiated terrestrial Pleistocene deposits, which may include swamp, paleosol, or floodplain facies. Below that are bay-head delta facies (blue) with diagnostic clinoform geometry. Red line ‘M’ is a seafloor multiple. Gamma density profile of core acquired in the vicinity of chirp profile shown to right; tan indicates sand sheet facies, gray indicates transitional facies, and green indicates swamp facies.
4.4.2 Working hypothesis of forest growth, burial, preservation, and exposure

The lack of absolute chronology prevents the unequivocal placement of the drowned forest within the established NGoM geological history, but a working hypothesis has been developed for forest growth, burial, preservation, and exposure. This hypothesis is based on the observations described in section 4.3, and the stratigraphy/lithology to facies interpretation in section 4.4.1. Future dating via Uranium-Thorium radiochemistry and OSL may revise our working hypothesis and chronology.

Figure 4.10 Working hypothesis of drowned forest growth conditions. The approximate time period of forest growth is hypothesized to be 45 kybp, based on a combination of eustatic sea level (modified from Rittenour et al., 2007), cypress growth requirements, and stratigraphy derived from cores and seismic data. In this diagram, D-D’ represents a vertical cross-section of arbitrary shallow depth, and a plan-view of lateral facies distribution is shown below. The forest is hypothesized to have grown in a relatively low-lying swamp in the floodplain of a Mobile valley tributary river; all of the topmost facies shown are hypothesized to have been deposited on top of bay-head facies formed during an older transgressive phase.
Uncertainty regarding age and elevation relative to sea level of the forest while it was alive is largely due to two factors: the minimum nature of the radiocarbon sample ($\leq 45,000$ ybp), and whether the forest’s elevation has remained stable throughout time. The radiocarbon age will be assumed to be accurate, due to a lack of alternatives. The eastern MAFLA has been interpreted as tectonically stable (~2.3 m subsidence/10 ky, Bartek et al. (2004)), but recent studies indicate the entire NGoM margin may deviate from the eustatic sea level curve due to continental levering associated with the Laurentide ice sheet (Love et al., 2016). These uncertainties will have to be budgeted once absolute dates are acquired, but for the time being are assumed to be minimal.

If it is assumed the forest has maintained its absolute elevation since growth ~45,000 ybp and a eustatic sea level curve applies, the forest would have grown ~40-60 m above sea level (Waelbroeck et al., 2002). Based on analogous contemporary environments and core lithology, the forest likely grew in a freshwater swamp within a low-lying river floodplain (Fig. 4.10; Conner and Toliver (1990)). This floodplain could have been adjacent to a tributary river feeding into the main paleo-Mobile River, or the Mobile River itself. Either way, the relatively low elevation of a floodplain and a proximal fluvial source would provide the accommodation space and sediment supply, respectively, necessary for the observed forest preservation state (Damste et al., 2002).
Figure 4.11 Working hypothesis of drowned forest burial conditions. The approximate time period of forest burial is hypothesized to be 40 kybp, based on a combination of eustatic sea level (modified from Rittenouer et al., 2007), cypress growth requirements, and stratigraphy derived from cores and seismic data. In this diagram, E-E’ represents a vertical cross-section of arbitrary shallow depth, and a plan-view of lateral facies distribution is shown below. The forest is hypothesized to have been rapidly buried by floodplain aggradation associated with the short sea level rise interval around 40 kybp.

Assuming the postulated growth setting (Fig. 4.10), forest burial could be associated with autogenic or allogenic events. An example of autogenic burial would be a river crevasse splay that rapidly deposited a large mass of sediment on top of the forest (Shen et al., 2015), preventing degradation via aerobic respiration and/or exposure via erosion. Allogenic burial would involve external forcing, such as rapid sea level rise raising upstream river stage and promoting overbanking and associated floodplain aggradation (Shen et al., 2012). The relatively rapid sea level rise of ~15 m over ~3000 years (~5 mm/yr) between 43,000-40,000 ybp is a
plausible driver of allogenic burial, and is therefore favored in the working hypothesis.

Regardless of mechanism, the net outcome is rapid and deep burial of the forest beneath a relatively impermeable sediment cap (Fig. 4.11).

![Diagram of forest preservation conditions during the LGM.](image)

Figure 4.12 Working hypothesis of drowned forest preservation conditions. The forest was preserved during the LGM, when sea level was approximately 120 m below current level and the entire continental shelf was subaerially exposed to cooler, drier conditions (modified from Rittenouer et al., 2007). In this diagram, F-F’ represents a vertical cross-section of arbitrary shallow depth, and a plan-view of lateral facies distribution is shown below. The forest is hypothesized to have been sufficiently buried so as not to be subaerially exposed during lowstand erosion and paleosol formation.

The forest survived the falling and lowstand sequence stratigraphic stages that favor destruction of preexisting sedimentary strata (Van Wagoner et al., 1988). Assuming the forest was buried ~40 kybp, the site would have been subaerially exposed for another ~30 ky. This includes the Last Glacial Maximum (LGM), during which the climate was cooler and drier, i.e. conducive to eolian erosion and paleosol formation (Fig. 4.12; Rutter et al. (2017)).
Figure 4.13 Working hypothesis of drowned forest current conditions. The forest survived marine transgression and was subaqueously exposed by 2004 Hurricane Ivan (modified from Rittenour et al., 2007). In this diagram, G-G’ represents a vertical cross-section of arbitrary shallow depth, and a plan-view of lateral facies distribution is shown below. The modern facies distribution is almost uniform sand sheet, with a small forest section exposed to the water column.

After the LGM, deglaciation drove eustatic sea level rise and marine transgression. Sea level fluctuation is rarely linear, and high-frequency rises and fall in sea level are often superimposed on a larger-scale rising and falling trend (Siddall et al., 2003). This “transgressive belt sander” (D. Swift, pers. comm. 2017) moves the high-energy shoreline environment back and forth across preexisting strata. The MAFLA sand sheet is formed largely by transgressive processes (McBride et al., 1999), and the drowned forest must have been sufficiently buried and/or rapidly bypassed by the nearshore environment to avoid being integrated into the MAFLA sheet (Fig. 4.13). After transitioning to a full marine environment ~8000 ybp, the forest was
evidently unperturbed until hypothesized exposure around 2005 coincident with the passage of Hurricane Ivan (Figs. 4.1 and 4.13). Now that the forest is exposed to the water column, degradation is apparent and preservation is not expected to exceed 100 years. A synthesis of seismic and lithological data with the forest preservation hypothesis is presented in Fig. 4.14.

4.4.3 Implications of forest model for analogous sites and Gulf Coast paleoclimate and sea level history

Partial preservation of forests following sea level transgression via wood debris, seeds, and pollen is fairly common, and can provide paleoenvironmental and climatic information including species assemblage, aridity, temperature, and radiocarbon age (Hautevelle et al., 2006). However, preservation of an entire forest in growth position post-transgression is rare and provides information fragments, seeds, and pollen cannot, such as dendrochronological records of interseasonal climate variation (Schongart et al., 2006). The hypothesis of forest growth setting, burial, and preservation conditions can serve as a model for discovering analogous environments, both along the NGOM and on other passive margins around the world.

The conjecture that the forest grew in a floodplain setting proximal to a river is based on analogous contemporary environments (Conner and Toliver, 1990). Major river paleochannels are resolvable in seismic data and well-mapped for many passive continental margins (Shaw and Courtney, 1997; Schwab et al., 2014; van Heteren et al., 2014), so potential forest growth locations can be identified in many cases with preexisting maps. The other two major factors expected to dictate the abundance of analogous preserved forests are burial and post-burial erosion and disturbance.
If the forest detailed in this study was buried by widespread overbank flooding associated with eustatic sea level rise, other still-buried preserved forests may exist as a “bathtub ring” proximal to paleochannels. If the forest was buried by an autogenic process with smaller spatial extent such as a river crevasse splay, the number of analogous sites may be much lower.

Fig. 14: Long strike profile synthesizing chirp and core data with working hypothesis of drowned forest facies and stratigraphy. A-A’ corresponds to basemap location in Figs. 4.10-4.13. Chirp profile is shaded to correspond with major conceptual diagram facies groups: pink for Holocene sand sheet, green for Pleistocene swamp, Yellow for all other Pleistocene facies, and blue for bay head delta.

Regardless of autogenic or allogenic forcing leading to forest burial, the existence of similar sites is also heavily dependent on surviving conditions unfavorable for stratigraphic preservation, including sea level lowstand and transgression (Figs. 4.12 and 4.13). The proximity of the drowned forest to the current shoreline may have aided its preservation, since sea level rise rapidly accelerated 11,000 years before present (Love et al., 2016), reducing the amount of time the forest area was exposed to the “transgressive band saw”. For this reason, high-resolution constraints on sea level fluctuation history should assist in identifying areas conducive to forest preservation.
4.5 Conclusions

New surficial and shallow subsurface geophysical data collected at a drowned forest offshore of Gulf Shores, Alabama provides geologic and stratigraphic context in resolution higher than previous regional surveys. The deepest resolvable strata are tentatively identified as bay-head deltaic facies (but may also represent lateral fluvial point bar accretions), and have clinoform reflectors with increasing dip angle approaching the main axis of the Mobile paleovalley. Above the deltaic facies are terrestrial facies, including paleosol to the east and swamp to the west. The surficial layer is the ridge-and-trough morphology of the MAFLA sand sheet; backscatter is higher on ridges than troughs and lowest where the forest is exposed. Bathymetric variation is controlled almost entirely by MAFLA sand sheet thickness, and lack of internal stratigraphy indicates the ridges formed either during or after sea level transgression, as opposed to being Pleistocene relict features. This sequence of facies documents a marine-terrestrial-marine succession, but a lack of age controls prevents absolute chronology of this sequence from being determined.

Integration of vintage and newly acquired geophysical data, in tandem with preliminary radiocarbon ages and core lithology has yielded a working hypothesis of forest growth, burial, preservation, and exposure. The forest is hypothesized to have grown ~45,000 ybp in the floodplain of either the paleo-Mobile River or one of its tributaries. Rapid fine-grained sediment buried the forest, either through allogenic floodplain aggradation associated with 40,000 ybp sea level rise or autogenic levee crevassing/overbank flooding. This rapid burial preserved the forest during periods of sediment bypass and/or erosion, including the LGM and sea level transgression leading to the current highstand. Exposure of the forest is hypothesized to have been caused by 2004 Hurricane Ivan, which passed within 10 km of the site as a category 3 hurricane. The forest
evolution hypothesis can be used as a model to predict the location of analogous sites, both along the NGoM margin and on other passive margins. Using this model, floodplains proximal to paleochannels in areas known to have been transgressed over rapidly have the highest probability of harboring preserved drowned forests.

4.6 References cited


Gonzalez, S., 2016, Facies reconstruction and stratigraphy of a Late Pleistocene bald cypress forest discovered on the Northern Gulf of Mexico continental shelf, Undergraduate senior thesis, Louisiana State University, 48 p.


CHAPTER 5: SUMMARY

The goal of a dissertation is to demonstrate the PhD candidate is capable of making original scientific contributions that advance the state of knowledge in their chosen field. This dissertation focuses on submarine sediment transport and slope stability, and improves our general understanding of these processes through case studies in the NGoM. Chapter Two concludes that mudflow activity on the MRDF occurs more frequently, and transports more sediment (and with it organic carbon and adsorptive particulates including heavy metals) downslope, than previously conceptualized.

This finding can be transitorily applied to other margins, as it corroborates the conclusions of recent studies with similar goals and motivations. It was also noted that despite appreciable volumetric flux within the geomorphic confines of the mudflow zones occurred, the “footprint” of the mudflow zones showed little change over a nine year “fair weather” period. This is an important finding since lateral movement of the mudflow zone (by gully wall retrogressive failure and lobe progradation) is a characteristic feature of submarine landslides triggered by major hurricane passage. This lateral movement disparity implies there is a threshold between mudflow-confined failures and mudflow-expanding failures, which will need to be quantified in future studies.

Chapter Three used an MCDP as a de facto sediment trap and slope stability experiment. Several basic research conclusions regarding both sediment transport and slope stability in the Louisiana ICS region came from this study, first and foremost the important role resuspension plays in the regional short and long-term sediment accumulation rate (SAR). When sheltered from ambient oceanographic conditions (and in lieu of major hurricane passage), Sandy Point
MCDP vertically accreted ~100-500% faster than previously documented ICS SAR, and an order of magnitude faster than centennial-timescale ICS SAR. SAR within the MCDP was consistent between the first and second/third years following initial dredging, but the proposed model for pit evolution predicts infill will slow drastically once the pit floor is sufficiently close to the seafloor to be “recoupled” to ambient oceanographic conditions. MCDP walls are on average steeper than MRDF mudflow gully walls (6.9º and 4.3º, respectively), but lost steepness at an average rate of 0.55º/year, while MRDF gully walls steepened at a comparable rate (0.44º/year). The gradient convergence of unnaturally oversteepened MCDP walls and natural mudflow gully walls may represent an angle of repose for this region/substrate in lieu of extreme events, which may provide an empirical measure of slope stability in this region.

Chapter Four is based on a research project in an earlier stage than Chapters Two and Three, so the conclusions are more speculative. Regardless, the state of knowledge regarding the drowned forest offshore Gulf Shores, AL and the MAFLA sand sheet it is surrounded by has been advanced by preliminary findings reported in this dissertation. The lack of internal stratigraphy (washover fans, prograding wedges) diagnostic of barrier island complexes indicates the characteristic MAFLA surficial ridges, at least in the region ~15 km offshore Mobile Bay, are completely reworked (post-transgressive) features. Below the MAFLA sand sheet, seismic stratigraphic analysis, in tandem with a companion coring study, document terrestrial strata, including the swamp facies the drowned forest is set within and a paleosol to the east. Below the terrestrial layer, all resolvable strata (maximum Chirp penetration depth < 30 m below seafloor) took the form of south-southeast dipping clinoforms which dipped at steeper angles to the east. These clinoforms are interpreted as bay-head delta facies deposited during a phase of sea level backstepping, infilling incised valleys carved during sea level lowstand.
The marine-terrestrial-marine sequence documented above is combined with preliminary forest radiocarbon ages (minimum age estimate 45 kybp) to construct a working hypothesis of forest evolution. Based on location proximal to the Mobile River paleovalley and contemporary analogues, it is likely the forest grew in a low-lying river floodplain. Forest burial and subsequent preservation requires accommodation space and rapid fine-grained burial; the sediment the forest was buried in could have been sourced from autogenic (seasonal overbank flooding or levee crevassing) or allogenic (similar processes, but tied to boundary condition changes such as sea level rise) processes. The forest remained burial through maximum subaerial exposure during the LGM and sea level transgression, and is hypothesized to have been exposed in 2004 by the nearby passage of Hurricane Ivan. The working hypothesis of forest evolution can be used as a predictive model to identify analogous sites with still-buried forests; using this model, river floodplains that were known to be rapidly transgressed over during sea level rise (reducing sedimentary fabric destruction) have the highest likelihood of harboring preserved buried forests.
APPENDIX A: CHAPTER 2 SUPPLEMENTAL MATERIALS

A.1 Data acquisition and processing

Table A1 shows the equipment used in bathymetric data acquisition during 1977-1979, October 2005, February 2009 and June 2014 bathymetric surveys.

Table A1 Details of bathymetric data acquisition. Reson sonars are multibeam units, Edgetech 4600 is interferometric. ECU = East Carolina University; OSU = Oregon State University; BLM = Bureau of Land Management, USGS = United States Geological Survey; LSU = Louisiana State University; UNO = University of New Orleans; BOEM = Bureau of Ocean Energy Management; SBES=singlebeam echosounder, GPS= Global Positioning System; CTD = Conductivity, Temperature, and Depth; SVP= Sound Velocity Profiler.

<table>
<thead>
<tr>
<th>Survey</th>
<th>Vessel</th>
<th>Operator</th>
<th>Sonar</th>
<th>Positioning System</th>
<th>Heave/pitch/roll compensation</th>
<th>CTD/SVP</th>
<th>Area (km²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>October 1977-</td>
<td>Various</td>
<td>BLM, USGS</td>
<td>110 kHz SBES</td>
<td>Loran C</td>
<td>N/A</td>
<td>N/A</td>
<td>775</td>
</tr>
<tr>
<td>March 1979</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>October 2005</td>
<td>R/V Cape Hatteras</td>
<td>ECU, OSU, USGS</td>
<td>Reson 8101</td>
<td>Furuno GP-90</td>
<td>TSS MAHRS</td>
<td>Sea-Bird SBE 9</td>
<td>70</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>SeaBat 7125</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>June 2014</td>
<td>R/V Coastal Profiler</td>
<td>LSU, UNO, BOEM</td>
<td>Edgetech 4600</td>
<td>Hemisphere VS111</td>
<td>SMC IMU-108</td>
<td>Valeport MiniSVP</td>
<td>55</td>
</tr>
</tbody>
</table>

Bathymetric data from all three modern surveys were processed using Caris HIPS and SIPS. Processing began with automatic filtering (including swath width and depth filters) to remove the majority of bad data. Manual cleaning was then used to further remove spurious data. Modern datasets were corrected for sound velocity artifacts (utilizing the closest CTD casts in time and space), and tidal corrections and vertical referencing were derived from mean sea level of the NOAA Southwest Pass tidal gauge (station ID 8760959). Digital Elevation Models
(DEM) were then constructed from point clouds using the Combined Uncertainty and Bathymetry Estimator (CUBE) algorithm (Fig. A1). DEMs were exported from Caris and imported into ESRI ArcGIS for analysis and interpretation. All morphometric analysis (cross sections, measurements, surface differencing) was done in ArcGIS. Two DoDs were produced, 2014-2009 and 2009-2005 (Fig. 2.2). The DoD shown in Fig. 2.2 was regridded using Kriging interpolation method to 25 m² cell size using Golden Surfer. The 1977-1979 data was received as a digitized version of the original hand-contoured maps; the raster was gridded to 100 m² resolution.

A.2 Volumetric calculations

The overall goal of the volumetric calculations was to assess the magnitude of major hurricane-induced failures to failures that occurred without major hurricane forcing. In order to calculate volumetric changes between surveys, the ESRI ArcGIS Cut Fill tool was used. More information on Cut Fill can be found here. No single gully/lobe complex was covered by the
three bathymetric datasets used in volumetric calculations (1977-1979, 2005, 2009, Fig. 2.1), so several approximations and assumptions had to be made to obtain meaningful comparisons.

These assumptions create large uncertainties (discussed in detail in the next section), so the volumetric calculations are only intended to provide an order of magnitude sense of comparison.

The volumetric changes are presented in Table A2 below:

Table A2 Mississippi River Delta Front volumetric changes

<table>
<thead>
<tr>
<th>Time Interval</th>
<th>Subset of Study Area</th>
<th>Area (m²)</th>
<th>Bulk Volumetric Change (m³)</th>
<th>Net Volumetric Change (m³)</th>
<th>Uncertainty (+/− m³)</th>
<th>Annual Volume Transported (m³/y⁻¹)</th>
</tr>
</thead>
<tbody>
<tr>
<td>March 1979-October 2005</td>
<td>Mudflow Lobes</td>
<td>1.1 x 10⁷</td>
<td>1.2 x 10⁷</td>
<td>2.8 x 10⁷</td>
<td>3.6 x 10⁶</td>
<td>1.1 x 10⁶ ± 1.5 x 10⁵</td>
</tr>
<tr>
<td>October 2005-February 2009</td>
<td>Mudflow Gullies</td>
<td>1.1 x 10⁷</td>
<td>5.5 x 10⁶</td>
<td>2.2 x 10⁶</td>
<td>2.9 x 10⁵</td>
<td>7.3 x 10⁵ ± 7.3 x 10⁴</td>
</tr>
</tbody>
</table>

To focus on major storm-driven changes as exclusively as possible, volumetric calculations were done for the 1979-2005 time interval on the mudflow lobe area only (Fig. A2). The lobe area was defined and digitized on the 2005 Walsh DEM using several criteria, including 1) a positive relief above the surrounding seafloor demarcated by a sharp gradient (5-30°), 2) a hummocky surficial appearance, 3) location directly downslope of mudflow gully(s). The rationale for choosing the mudflow lobe zone to document major hurricane-induced failure is based on two observations: previous authors (including Bea et al. (1983) and Hitchcock et al. (1996)) have noted that major hurricane passage triggers large-scale (<1 km) downslope movement of mudflow lobes, and our assessment that the mudflow lobes are relatively laterally immobile during the 10-year quiescent observation period (Fig. S3). It is inevitably
oversimplifying to assume all movement observed in the lobe zone is hurricane-induced, but volumetric changes in the form of mudflow lobe nose downslope movement are likely to be hurricane-driven changes. The mudflow lobes were assumed to be purely depositional systems, so only volume gains were calculated.

Figure A2 Digital elevation map from October 2005 with mudflow gully (blue) and mudflow lobe (red) areas used for volumetric calculations digitized. The extents of Figure A3 are shown as the dashed black box.

It is probably not a valid assumption to assume that the entire study area has remained static outside the gully/lobe zones. In order to account for change that had occurred throughout the entire study area, a small (0.05 km²) area of the prodelta (pink polygons in Fig. A1) presumably outside the reach of gully/lobe activity was also assessed for changes using surface differencing, and found to have undergone an average of 1.51 m of erosion between the 1970s
and 2005 surveys. This means that the lobe accretion estimates are conservative (bulk volumetric change in Table A2), assuming the erosion observed in the prodelta reference area is representative of the entire survey area. If the volume lost by regional erosion is added to the accretion estimates in a simple manner (average erosion of reference area multiplied by surface area of lobes), the volume accreted across the mudflow lobe area increases from $1.6 \times 10^7$ m$^3$ to $2.8 \times 10^7$ m$^3$ (net volumetric change in Table A2).

Figure A3 Gradient map of a mudflow lobe showing the relative lack of lateral change between the 2005, 2009, and 2014 surveys. The lack of lateral change stands in contrast to surveys bracketing major hurricanes, where mudflow lobes move hundreds of meters to kilometers downslope.

The same area used for volumetric analysis of the 1979–2005 period could not be used for the 2005–2009 period, because of insufficient survey overlap of the mudflow lobe zones (Fig. 2.1). Therefore, changes were calculated in the gullies instead; the extent of the gullies was digitized in a manner similar to that described above for the lobes, but negative relief was used for discrimination criteria instead of positive relief. Since the assumption was made that
mudflow lobes are purely depositional areas, only negative volumetric changes were tallied for the gullies, simplifying to assume they are purely erosional systems. The total area of the gully zone used in the 2005-2009 calculations and the lobe zone used in the 1979-2005 calculations was approximately equal (both ~1.1 x 10^7 m^2). There were also changes in the reference area in between 2005 and 2009; the mean vertical change was 0.3 m depth increase, which means the volume of material eroded from the gullies is a maximum estimate. When the “ambient” vertical change is accounted for, the volume of sediment removed from the gullies between 2005 and 2009 drops to 2.2 x 10^6 m^3.

To put these numbers into a sediment budget context, the volumetric estimates were compared with the total suspended load discharge out of Southwest Pass, as calculated by (Allison et al., 2012). The bed (sand) load was disregarded because the majority of it is presumably deposited immediately proximal to the distributary mouth. The suspended load value was expressed as mass, so to convert to volume a bulk density value (1.5 g/cm^3) derived from gravity and multicores obtained from the study area was used (Keller et al., 2016). Multiplying the bulk density by the total suspended load (20.8 million tons/year) yielded an annual volumetric flux of 1.4 x 10^7 m^3/year out of Southwest Pass, which means averaged annually ~8% and ~5% of the suspended sediment that arrives at Southwest Pass is mobilized by mass failures in the survey area during a given “major storm” and “quiescent” year, respectively.

A.3 Estimation of uncertainty

In order to acquire an estimate of uncertainty for the DoDs, the “fixed reference uncertainty” was calculated, following the methods detailed in Schimel et al. (2015). The general premise behind this method is to use a “reference area” that is assumed to be relatively stable
between two surveys to acquire a statistical estimate of the DEMs’ vertical uncertainty. A 0.05 km² area of the prodelta (~70 m water depth, four pink polygons in Fig. A1) was selected as the reference area because it is downslope of mudflow activity and relatively distant from the sediment plume of Southwest Pass. Values were extracted from the reference area, and statistical parameters were calculated (Fig. A4). The mean values for the DoDs were within 0.3 m of zero change, validating the assumption that this area remained relatively unchanged between surveys. A 2σ (95% confidence interval for normally distributed data) value was used as the uncertainty range; i.e. values within 2 standard deviations of 0 m were considered within the range of uncertainty and therefore not interpreted as actual change.

Figure A4 Histogram showing distribution of depth change within the fixed reference area for the 2009-2005 DoD. Values are shown as a percent of the total cells included within the reference area (shown as four pink polygons in Fig. A1). These data are roughly normally distributed around -0.25 m, and 2 times the standard deviation (0.5 m, 95% confidence interval) were chosen as the uncertainty range.

A similar method was used for estimation of volumetric uncertainty, with the necessary extra step of scaling into three dimensions. Cut Fill was performed on the same reference area as described above and central tendencies were calculated, in addition to the standard deviation. The standard deviation was multiplied by the surface area of the zones (mudflow lobes for
1979-2005, mudflow gullies for 2005-2009), yielding an empirical estimate of volumetric uncertainty. In both cases, the uncertainty was approximately an order of magnitude less than the measured volumetric change.

A.4 Simulation of non-linear waves

To generate and propagate higher order waves (non-linear) on the MRDF, we selected computational fluid dynamics software with built-in capabilities to match requirements specific to our experiments. The FLOW-3D model was selected to perform this analysis, among other available research and commercial codes with varying degrees of strengths and limitations. The FLOW-3D model was selected due to its capability to simulate free-surface flows accurate, using a novel approach, its ability to generate higher order wave theories near the model domain boundaries, and for solving fully three-dimensional flows, without the shallow water approximation.

FLOW-3D is a three dimensional model where fluid motion is described with non-linear transient, second-order differential Navier-Stokes equations. The numerical algorithm used in FLOW-3D is based on both finite difference and finite volume methods applied to a structured computational grid. Structured grids are known for their computational efficiency and ease of discretizing the flow domain. The ability of the model to maintain a sharp interface (air-water) helped retain the non-linear waveform as waves were advancing across the MRDF, and provided for more accurate pressure fields. The finite volume method used in FLOW-3D derives directly from the integral form of the conservation laws for fluid motion, and therefore, retains the conservation properties (FLOW-3D, 2010; Meselhe et al., 2012). FLOW-3D is also capable of capturing the water free-surface accurately, using the so called true Volume Of Fluid – TrueVOF
(Barkudarov, 2004). This approach computes the advection of fluid to all neighboring cells according to the orientation of the fluid within the cell, and using pressure and velocity boundary conditions it computes the sharp free surface interface. This method is ideal for propagating non-linear waves on the delta front while preserving the non-linear waveform.

The governing equations used in FLOW-3D can be found in (FLOW-3D, 2010). FLOW-3D includes several turbulence closure models, namely Prandtl mixing length, One-equation transport, two-equation k–e transport, Renormalized group theory (RNG), and Large Eddy Simulation (LES) models. The two-equation turbulent closure models are widely used due to their relative computational efficiency and adequate performance for wide range of practical applications (e.g. (Muste et al., 2001)). For the simulations performed here, the Renormalization-Group (RNG) method (Yakhot and Orszag, 1986; Yakhot and Smith, 1992) was used. The RNG model applies statistical methods to the derivation of turbulent kinetic energy and its dissipation rate, and appears to have wider applicability than the standard k–e model when dealing with applications with strong shear regions (e.g., velocity gradients from crest to trough along a waveform).

A.5 Model domain and initial conditions

The computational domain for the model included a 2,000 m long, 5 m wide, and 100 m high rectangular basin. The model resolution was constant (but different) in each x and z dimensions. Horizontal resolution was ~ 0.6 m, vertical resolution ~ 0.25 m, and a time-step of <0.01 s. Model experiments were performed using a flat slope (to eliminate the effect of shoaling) and varied water depths (5-70 m) were instead used to establish pressure differential fields across the study area. All simulations were initialized with a fluid at rest for the required
mean fluid depth for each simulation experiment. The fluid density was equal throughout set to seawater density. To evaluate only the effect of depth and reduce further wave transformations once waves were applied at the boundary, a flat slope bed was selected. This is also a conservative approach, as a sloped delta front would promote shoaling and other forms of dissipation that could render the waves, and thus the seabed pressure differential to be higher. Friction was only applied at the seabed (partial slip) at a value that approximates that of a muddy seabed. Lateral friction was eliminated (full slip) to avoid lateral friction of the waves due to the narrow basin (~5 m).

A.6 Model boundary conditions

*FLOW-3D* possesses the capabilities to simulate regular linear and nonlinear propagating surface waves as well as irregular waves. A linear wave has a sinusoidal surface profile with small amplitude and steepness, while a nonlinear wave has larger amplitude (finite-amplitude), sharper crests and flatter troughs than the linear wave. The nonlinear waves can be categorized into Stokes, cnoidal and solitary waves, according to the wave characters and the mathematical methods used to obtain their solutions (*FLOW-3D*, 2010). Although the linear wave theory (Airy, 1845) has been used in many applications, the nonlinear wave theories often provide significant improvement in accuracy over the linear wave theory when the wave amplitude is not small. In *FLOW-3D*, three nonlinear wave theories are used for nonlinear wave generation: the fifth-order Stokes wave theory (Fenton, 1985), the Fourier series method for Stokes and cnoidal waves (Fenton, 1999), and McCowan’s theory for solitary wave (McCowan, 1891; Munk, 1949). Among them, Fenton’s Fourier series method is generally valid for all kinds of periodic propagating waves in deep water, transitional water and shallow water, including linear, Stokes
and cnoidal waves, it possesses higher order of accuracy and was the method used for all non-linear waves in this study. An example of wave propagation is shown in Fig. A5. Each simulation was run until the entire domain was filled with waves.

Figure A5 Example pressure distribution and velocity variation along a non-linear wave using FLOW-3D; the approximate wave at the boundary (left – outside the frame) is $H_s = 6.5$ m, $T = 9$ seconds). At the right boundary, outflow boundary was selected, to radiate the entire wave outside the domain and avoid wave reflections back into the domain.

A.7 Evaluating pressure change near the seabed

Simulation results were post-processed and visualized using Tecplot360® (Tecplot Inc.). Results included instantaneous three-dimensional velocities, pressures, position of the free-surface, and other hydraulic information such as flow depth, Froude number etc. At approximately the middle of the model domain, away from boundaries, pressures were extracted at the crest and trough of each wave and were differenced to calculate the pressure change. Wave height, length and period were also extracted to ensure that the wave retained (albeit some frictional dissipation) the wave characteristics applied at the boundary. All simulation results were then plotted against those in reported by Henkel (1970). All results (Henkel, 1970) and this study) were converted to SI units (see Fig. 2.4).
The selection of flat slope in the model may overestimate wave height and thus pressure, and so can the treatment of a rigid bed. (Bea and Aurora, 1981) reported that a deformable bed could attenuate waves and hence produce lower pressure differentials resulting in lower shear delivered to MRDF sediments. However, even without attenuation (which at these depths is not expected to be large) the results show that pressure differentials simulated here ($\Delta p \sim 7-35 \text{ kPa}$) exceed a range of values reported by Henkel (1970) and other authors required to produce failure in the study area (Fig. 2.4).

A.8 Supplemental references

Airy, G. B., 1845, Tides and Waves, Encyclopedia Metropolitan.


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A.9 Permission for reproduction

Chapter Two and Appendix A have been previously published in the Geological Society of America (GSA) peer-reviewed journal Geology. The published version can be accessed at the following URL: http://geology.gsapubs.org/content/early/2017/05/19/G38688.1.abstract?papetoc

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APPENDIX B: CHAPTER 3 SUPPLEMENTAL MATERIALS

B.1 Bathymetric data processing
Swath bathymetric data from the 2015 survey were processed using Caris HIPS and SIPS, while Golden Surfer was used to generate the 2012 surface from single-beam echosounder data. Swath processing began with automatic filtering (including swath width and depth filters) to remove the majority of bad data. Manual cleaning was then used to further remove spurious data. All datasets were corrected for sound velocity artifacts (utilizing the closest CTD casts in time and space), and tidal corrections were applied using measured tides from the NOAA Southwest Pass tidal gauge (station ID 8760959). Digital Elevation Models (DEMs) were then constructed from point clouds using swath angle weighting. The DEM was referenced to mean water level of the NAVD88 vertical datum.

B.2 Dredge pit volume change quantification
In order to calculate volumetric changes between surveys with sufficiently dense soundings to generate continuous surfaces (2012 and 2015), the ESRI ArcGIS Cut Fill tool was used. More information on Cut Fill can be found here. A 2-D polygon encompassing the entire pit area (as of the 2015 survey) was first digitized in ArcMap. The Cut Fill tool was then applied to the area within this polygon, yielding volumetric gains and losses within the dredge pit between the two surveys.

B.3 Wall slope analysis
To spatially quantify gradient change, eastern and western pit/gully walls and the pit floor were digitized (black dashed polygons in Figs. 3.4 and 3.5) and the mean value of adjacent horizontal bins was calculated, resulting in one average slope value for each horizontal bin grouping along a
north-south wall transect. The walls were defined as the sharpest gradient breaks between the ambient seafloor and the pit/gully floor. In Sandy Point, this gradient break tended to be laterally small, while the mudflow gully’s stair-stepped wall morphology created a wider wall area in places (Fig. 3.5).

Uncertainty of slope derived from a DEM is largely a function of terrain complexity and DEM resolution. An empirically derived formula from Tang et al. (2003) was used to quantify the slope uncertainty of each DEM:

\[
0.0015S^2 + 0.031S - 0.0325X - 0.0045S^2 - 0.155S + 0.1625
\]

where \( S \), stream density (m/m\(^2\)) is used as a proxy for terrain complexity and \( X \) is the DEM resolution in m\(^2\). All slope uncertainties were < 1º (Table 3.2).

**B.4 Quantifying bathymetric and volumetric uncertainty**

In order to acquire an estimate of uncertainty for the DoDs, the “fixed reference uncertainty” was calculated, following the methods detailed in Schimel et al. (2015). The general premise behind this method is to use a “reference area” that is assumed to be relatively stable between two surveys to acquire a statistical estimate of the DEMs’ vertical uncertainty. For this study, the seafloor > 100 m away from the pit was used as a reference area. Values were extracted from the reference area, and statistical parameters were calculated. A 2σ (95% confidence interval for normally distributed data) value was used as the uncertainty range (0.2 m); i.e. values within 2 standard deviations of 0 m were considered within the range of uncertainty and therefore not interpreted as actual change.
A similar method was used for estimation of volumetric uncertainty, with the necessary extra step of scaling into three dimensions. Cut Fill was performed on the same reference area as described above and central tendencies were calculated, in addition to the standard deviation. The standard deviation was multiplied by the surface area of the reference zone, yielding an empirical estimate of volumetric uncertainty.

B.5 Dredge pit literature infilling comparison

Infilling rates from published literature were compared with Sandy Point infilling rates to place our study into a broader context. Details of these studies could be found within publications are described in Table B1. Volumetric estimates removed from pits were provided in literature and derived from volume of sediment emplaced on restoration target beaches. Infilling rates were given as an average of repeated surveys or modeling results.
Table B1 Relevant parameters of dredge pit studies reported in literature.

<table>
<thead>
<tr>
<th>Pit Location</th>
<th>Reference</th>
<th>Date(s) dredged</th>
<th>Method for post-dredge monitoring</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boca Raton, FL</td>
<td>Cialone and Stauble, 1998</td>
<td>1985</td>
<td>Annual hydrographic surveys</td>
</tr>
<tr>
<td>Mobile Bay, AL</td>
<td>Byrnes et al., 2004a</td>
<td>None (proposed sites)</td>
<td>Numerical modeling combining wave, current, and sediment transport historical data to predict infilling rates</td>
</tr>
<tr>
<td>Egg Harbor Inlet, NJ</td>
<td>Byrnes et al., 2004b</td>
<td>None (proposed sites)</td>
<td>Numerical modeling combining wave, current, and sediment transport historical data to predict infilling rates</td>
</tr>
</tbody>
</table>
VITA

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