THE ROLE OF MARSH PLATFORM MORPHOLOGY IN THE GEOMORPHIC RESPONSE OF TIDAL INLET SYSTEMS TO SEA LEVEL RISE

By

JESSICA LOREN LOVERING

A DISSERTATION PRESENTED TO THE GRADUATE SCHOOL OF THE UNIVERSITY OF FLORIDA IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY

UNIVERSITY OF FLORIDA

2013
I dedicate this to my loving and supportive husband, Joseph Lovering.
ACKNOWLEDGMENTS

I would like to thank my committee chair, Pete Adams, for always providing support for my research ideas and continuing to push me to strive for greatness in my work. His high expectations of my work helped me to realize my potential and to work hard to achieve my goals. I would also like to thank my committee members, Joe Calantoni, Robert Dean, John Jaeger, Ellen Martin, and Krik Hatfield for all of their thoughtful input throughout my time in the Geological Sciences Department. Without the support of my committee, this work would not have been possible, and for that I am forever grateful.

I would also like to thank the “geomorph team” of students who have become a great research support system and my close friends. I want to thank Katherine Malone for keeping my spirits high and always being my biggest cheerleader even when things got tough. I am thankful for my close friendship with Shaun Kline, my officemate of 6 years. He has always been there to lend an ear and listen to my research ideas and to challenge me with his thoughtful contributions to my work. I would also like to thank Rich MacKenzie for his contagious enthusiasm and passion for the science, as well as being a dependable friend and colleague.

I would like to thank my parents who instilled in me the value of hard work and the joy of education. My mom has always been one of my greatest supporters and I know I would not have been able to achieve this work without her love and support. Lastly, I would like to thank my amazing husband, JD Lovering, who shared in this adventure with me and provided me with the courage to follow my dreams.
TABLE OF CONTENTS

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACKNOWLEDGMENTS</td>
<td>4</td>
</tr>
<tr>
<td>LIST OF TABLES</td>
<td>8</td>
</tr>
<tr>
<td>LIST OF FIGURES</td>
<td>9</td>
</tr>
<tr>
<td>LIST OF ABBREVIATIONS</td>
<td>12</td>
</tr>
<tr>
<td>ABSTRACT</td>
<td>13</td>
</tr>
<tr>
<td>CHAPTER</td>
<td></td>
</tr>
<tr>
<td>1 GENERAL INTRODUCTION</td>
<td>16</td>
</tr>
<tr>
<td>1.1 Problem Statement</td>
<td>16</td>
</tr>
<tr>
<td>1.2 Outline of Presentation</td>
<td>16</td>
</tr>
<tr>
<td>1.3 Tidal Inlet Systems</td>
<td>17</td>
</tr>
<tr>
<td>1.3.1 Morphologic Elements</td>
<td>17</td>
</tr>
<tr>
<td>1.3.1.1 Flood shoal</td>
<td>18</td>
</tr>
<tr>
<td>1.3.1.2 Ebb shoal complex</td>
<td>18</td>
</tr>
<tr>
<td>1.3.2 Inlet Classification</td>
<td>20</td>
</tr>
<tr>
<td>1.3.3 Equilibrium Relationships</td>
<td>21</td>
</tr>
<tr>
<td>1.3.3.1 Inlet channel</td>
<td>22</td>
</tr>
<tr>
<td>1.3.3.2 Ebb shoal complex</td>
<td>24</td>
</tr>
<tr>
<td>1.4 Salt Marshes</td>
<td>26</td>
</tr>
<tr>
<td>1.4.1 Types of Salt Marshes</td>
<td>26</td>
</tr>
<tr>
<td>1.4.2 Salt Marsh Response to Sea Level</td>
<td>26</td>
</tr>
<tr>
<td>2 HOW DOES VERTICAL MARSH ACCRETION CONTROL HYDRODYNAMIC AND MORPHOLOGIC RESPONSES OF A TIDAL INLET TO SEA LEVEL RISE?</td>
<td>37</td>
</tr>
<tr>
<td>2.1 Introduction</td>
<td>38</td>
</tr>
<tr>
<td>2.2 Background</td>
<td>39</td>
</tr>
<tr>
<td>2.3 Methodology</td>
<td>42</td>
</tr>
<tr>
<td>2.3.1 Study Site</td>
<td>42</td>
</tr>
<tr>
<td>2.3.2 Tidal Flow Model and Experimental Design</td>
<td>44</td>
</tr>
<tr>
<td>2.3.3 Empirically Derived Hydrodynamic-Morphologic Relationships</td>
<td>45</td>
</tr>
<tr>
<td>2.3.4 Model Testing and Calibration</td>
<td>46</td>
</tr>
<tr>
<td>2.4 Results</td>
<td>49</td>
</tr>
<tr>
<td>2.4.1 Basin Area</td>
<td>49</td>
</tr>
<tr>
<td>2.4.2 Tidal Prism</td>
<td>49</td>
</tr>
<tr>
<td>2.4.3 Cross-Sectional Area</td>
<td>51</td>
</tr>
<tr>
<td>2.4.4 Ebb Shoal Volume</td>
<td>51</td>
</tr>
<tr>
<td>2.5 Discussion</td>
<td>52</td>
</tr>
</tbody>
</table>
2.5.1 Regional Impacts ............................................. 52
2.5.2 Comparison with Previous Modeling Studies ............. 53
2.6 Conclusions ..................................................... 54

3 ECOGEOMORPHIC FEEDBACKS BETWEEN SEA LEVEL RISE, ESTUARY HYDRODYNAMICS, AND VERTICAL MARSH ACCRETION ......................... 71

3.1 Introduction .................................................... 72
3.2 Background .................................................... 72
3.3 Methods ......................................................... 74
  3.3.1 Study Site ............................................... 74
  3.3.2 Model Setup ............................................. 75
3.4 Results ........................................................ 76
  3.4.1 Spatial Distribution of Tidal Range and Marsh Submergence ........ 76
  3.4.2 Tidal Velocity Asymmetry ................................ 78
3.5 Discussion ...................................................... 80
3.6 Conclusions ..................................................... 82

4 THE ROLES OF MARSH CONFIGURATION AND MARSH MARGIN RETREAT ON TIDAL INLET MORPHOLOGY ............................. 98

4.1 Introduction .................................................... 98
4.2 Model Details .................................................. 100
  4.2.1 Study Sites .............................................. 103
4.3 Results ........................................................ 104
4.4 Discussion ...................................................... 106
  4.4.1 Two Main Mechanisms for Tidal Prism Change ............. 106
  4.4.2 Marsh Loss Spatial Distribution .......................... 107
  4.4.3 Influence on Adjacent Shorelines ......................... 107
  4.4.4 Other Influence on Marsh Loss ......................... 108
  4.4.5 Application to Barrier Systems Worldwide .............. 109
4.5 Conclusions ..................................................... 109

APPENDIX

A DELFT3D MODEL SETUP ............................................. 117
    A.1 Model Physical Processes ................................ 117
    A.2 Delft3D-FLOW Model Setup ............................... 118
      A.2.1 Land Boundary File Generation ...................... 119
      A.2.2 Grid Generation ..................................... 120
      A.2.3 Bathymetry Generation ............................... 121
      A.2.4 Creating an MDF-File ............................... 122
      A.2.5 Executing model run ................................ 126
      A.2.6 Viewing model output ............................... 126

B MODEL INPUT PARAMETERS ........................................ 127
<table>
<thead>
<tr>
<th>Table</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-1</td>
<td>Inlet morphologic variables (from Hubbard et al. (1979))</td>
<td>29</td>
</tr>
<tr>
<td>1-2</td>
<td>Empirical parameters for the tidal prism-inlet area relationships developed by Jarrett (1976) (for metric units)</td>
<td>31</td>
</tr>
<tr>
<td>1-3</td>
<td>Number of inlets and correlations for the tidal prism-inlet area relationships developed by Jarrett (1976) (for metric units)</td>
<td>33</td>
</tr>
<tr>
<td>1-4</td>
<td>Empirical parameters for the tidal prism - ebb shoal volume relationships developed by Walton and Adams (1976) for different wave energy regimes.</td>
<td>34</td>
</tr>
<tr>
<td>4-1</td>
<td>Study sites hydraulic, ecologic, and geomorphic properties</td>
<td>116</td>
</tr>
<tr>
<td>4-2</td>
<td>Inlet morphologic and hydraulic response to wetland loss</td>
<td>116</td>
</tr>
</tbody>
</table>
# LIST OF FIGURES

<table>
<thead>
<tr>
<th>Figure</th>
<th>Description</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>1-1</td>
<td>Tidal inlet elements shown on a typical mixed-energy tidal inlet system.</td>
<td>30</td>
</tr>
<tr>
<td>1-2</td>
<td>Morphology of a typical flood shoal (reproduced from Hayes (1980))</td>
<td>31</td>
</tr>
<tr>
<td>1-3</td>
<td>Conceptual models of sediment bypassing (from FitzGerald et al. (2000)).</td>
<td>32</td>
</tr>
<tr>
<td>1-4</td>
<td>Coastal classification (reproduced from Hayes (1979)).</td>
<td>33</td>
</tr>
<tr>
<td>1-5</td>
<td>Patterns in the number and spacing of tidal inlets as a function of wave versus tidal energy along the Georgia Bight (from FitzGerald (1996)).</td>
<td>34</td>
</tr>
<tr>
<td>1-6</td>
<td>Tidal inlet types in the Georgia embayment (reproduced from Hubbard et al. (1979)).</td>
<td>35</td>
</tr>
<tr>
<td>1-7</td>
<td>Escoffier stability curve (from Escoffier (1940, 1977)).</td>
<td>36</td>
</tr>
<tr>
<td>2-1</td>
<td>Map of the Sea Island Chain with an inset of the Saint Marys River Entrance basin.</td>
<td>56</td>
</tr>
<tr>
<td>2-2</td>
<td>Contour map of the Saint Marys Entrance basin bathymetry.</td>
<td>57</td>
</tr>
<tr>
<td>2-3</td>
<td>The cumulative distribution of the elevations within the St. Marys River basin model domain.</td>
<td>58</td>
</tr>
<tr>
<td>2-4</td>
<td>The temporal distribution of flooded basin area in the Saint Marys River Entrance basin.</td>
<td>59</td>
</tr>
<tr>
<td>2-5</td>
<td>A map of the NOAA Fernandina Beach Currents Project bottom-mounted workhorse ADCP deployment locations and corresponding deployment dates.</td>
<td>60</td>
</tr>
<tr>
<td>2-6</td>
<td>Water elevations at NOAA tide station 8720030 and the co-located Delft3D modeled water levels.</td>
<td>61</td>
</tr>
<tr>
<td>2-7</td>
<td>Time series plots of surface current speeds from ADCP data and co-located model data for the first 3 full days of deployment.</td>
<td>62</td>
</tr>
<tr>
<td>2-8</td>
<td>Scatter plots of surface current U and V components with the 95% confidence ellipse and principal axes from the six NOAA ADCP stations and co-located model data.</td>
<td>63</td>
</tr>
<tr>
<td>2-9</td>
<td>Model comparisons with NOAA ADCP data.</td>
<td>64</td>
</tr>
<tr>
<td>2-10</td>
<td>Model output of water level, instantaneous discharge, and calculated tidal prism.</td>
<td>65</td>
</tr>
<tr>
<td>2-11</td>
<td>The change in the temporal distribution of flooded basin area and the change in the temporally averaged basin area with sea level rise.</td>
<td>66</td>
</tr>
</tbody>
</table>
2-12 The change in temporally averaged flooded basin area due to sea level rise, for a static marsh platform and a marsh platform accreting at the same pace as sea level rise. .......................................................... 67

2-13 The ideal tidal prism, the modeled tidal prism, and the ratio of modeled to ideal tidal prism. ................................................................. 68

2-14 The change in modeled tidal prism .............................................. 69

2-15 Change in cross-sectional area and ebb shoal volume due to sea level rise . . 70

3-1 Map of Saint Marys Entrance with marsh vegetated area highlighted, inset on a map of the Sea Island Chain. ......................................................... 84

3-2 Map of elevations in the Saint Marys Entrance ................................. 85

3-3 The distribution of elevations in the Saint Marys Entrance basin ........... 86

3-4 The temporal distribution of flooded basin area in the Saint Marys Entrance basin for present sea level conditions. ................................. 87

3-5 The spatial distribution of the change in tidal range from present sea level conditions as a result of a sea level rise on an accreting marsh platform. ......................................................... 88

3-6 The spatial distribution of the change in tidal range from present sea level conditions as a result of a sea level rise on a static marsh platform. ........... 89

3-7 The change in the average tidal range over 10 tidal cycles due to sea level rise for seven channel locations. ......................................................... 90

3-8 The spatial distribution of submergence duration for various sea level rise magnitudes on a static marsh platform. ......................................................... 91

3-9 Time series of water depth at three points located on the marsh platform for sea level rise of 0 to 100 cm at 20 cm intervals for a static marsh system. . . 92

3-10 The velocity stage plot and the time series of the water levels and the spatially averaged velocity through the inlet for present sea level and a sea level rise of 60 cm on a static and an accreting marsh platform. ................................. 93

3-11 The change in time at which tidal flow reversals occur due to sea level rise . . 94

3-12 The spatially averaged velocity across the inlet, the cube of the spatially averaged velocity, and the cumulative of the spatially averaged velocity cubed for sea level rise on an accreting marsh platform. ................................. 95

3-13 The spatially averaged velocity across the inlet, the cube of the spatially averaged velocity, and the cumulative of the spatially averaged velocity cubed for sea level rise on a static marsh platform. ................................. 96
3-14 The ratio of the flood to ebb sediment transport proxy .......................... 97
4-1 Conceptual model of tidal inlet geomorphic response to sea level rise. .... 111
4-2 Diagram illustrating how the basins are parameterized into two sections on 
either side of the inlet channel. ................................................................. 111
4-3 Maps of the back-barrier basins examined in this study ......................... 112
4-4 The change in tidal prism due to wetland vegetation loss for PLI and SAI. 113
4-5 The impact of marsh vegetation loss on tidal prism, cross-sectional area of 
inlet channel, and ebb shoal volume. ....................................................... 114
4-6 Schematic Illustration of wetland vegetation loss in two basins with 
ecogeomorphic configurations similar to SAI and PLI ............................ 115
<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ADCP</td>
<td>Acoustic Doppler Current Profiler</td>
</tr>
<tr>
<td>AM</td>
<td>Accreting Marsh</td>
</tr>
<tr>
<td>CIRP</td>
<td>Coastal Inlet Research Program</td>
</tr>
<tr>
<td>HWFR</td>
<td>High Water Flow Reversal</td>
</tr>
<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
</tr>
<tr>
<td>LWFR</td>
<td>Low Water Flow Reversal</td>
</tr>
<tr>
<td>MHW</td>
<td>Mean High Water</td>
</tr>
<tr>
<td>MLW</td>
<td>Mean Low Water</td>
</tr>
<tr>
<td>MSL</td>
<td>Mean Sea Level</td>
</tr>
<tr>
<td>NWI</td>
<td>National Wetlands Inventory</td>
</tr>
<tr>
<td>NGDC</td>
<td>National Geophysical Data Center</td>
</tr>
<tr>
<td>NOAA</td>
<td>National Oceanic and Atmospheric Administration</td>
</tr>
<tr>
<td>PLI</td>
<td>Ponce de Leon Inlet, FL</td>
</tr>
<tr>
<td>SAI</td>
<td>Saint Augustine Inlet, FL</td>
</tr>
<tr>
<td>SM</td>
<td>Static Marsh</td>
</tr>
<tr>
<td>SME</td>
<td>Saint Marys Entrance, FL</td>
</tr>
<tr>
<td>USGS</td>
<td>United States Geological Survey</td>
</tr>
</tbody>
</table>
Abstract of Dissertation Presented to the Graduate School
of the University of Florida in Partial Fulfillment of the
Requirements for the Degree of Doctor of Philosophy

THE ROLE OF MARSH PLATFORM MORPHOLOGY IN THE GEOMORPHIC
RESPONSE OF TIDAL INLET SYSTEMS TO SEA LEVEL RISE

By
Jessica Loren Lovering

May 2013

Chair: Peter N. Adams
Major: Geology

The morphologic evolution of tidal inlets depends on the ecogeomorphic behavior of the back-barrier basin, and exerts strong influence on local shoreline response to sea level rise. In this thesis, I investigate the role of wetland vegetation on tidal inlet response to sea level rise through three approaches, with each approach investigating a specific set of process linkages: (1) Vertical marsh accretion control on changes to the tidal prism, inlet channel cross-sectional area, and ebb shoal volumes due to sea level rise, (2) changes to the sediment availability to the marsh platform by sea level rise induced alterations to the inlet hydrodynamics, and (3) the influence of marsh configuration and edge erosion on the tidal prism, inlet channel cross-sectional area, and ebb shoal volume. Investigating links among changes to wetland vegetation, hydrodynamics of the back-barrier basin, and inlet morphology will be a critical advance in understanding how shorelines with barrier island systems will evolve under sea level rise worldwide.

How does vertical marsh accretion influence the tidal prism, inlet channel cross-sectional area, and ebb shoal volume during sea level rise? Using a numerical hydrodynamic model, paired with empirically derived morphologic relationships, this study explores the behavior of a back-barrier basin under two end-member vertical marsh accretion scenarios: (A) no accretion, wherein marsh islands become submerged, changing both the flooded basin area and the spatial pattern of tidal wave
attenuation, and (B) a marsh accretion rate equal to the rate of sea level rise, wherein marsh tidal channels deepen but maintain their courses and the associated friction reduction leads to a more efficient tidal exchange between the ocean and back-barrier basin. Under both marsh accretion scenarios, a similar increase in tidal prism due to channel deepening and improved exchange efficiency is observed. On a static marsh platform, this increase is amplified by the addition of flooded basin areas contributing to the exchange. Scenario A (no accretion) produced a tidal prism, inlet cross-sectional area, and ebb shoal volume increase double that of the increase calculated for scenario B (pace-keeping accretion). The increase in equilibrium inlet channel cross-sectional area, arising from scour processes, exceeds the increase due to sea level rise alone, illustrating the influential contribution from marsh tidal flow processes.

How does vertical marsh accretion change the hydrodynamic response to sea level rise in the back-barrier basin and the availability of sediment to the marsh platform? Vertical marsh accretion results from allochthonous sediment availability and autochthonous sediment produced in situ by marsh vegetation. Increased duration of marsh vegetation submergence can lead to vegetation waterlogging and drowning, leading to reduced plant production and autochthonous sediment availability on marsh platforms. Changes to the inlet exchange efficiency during a tidal cycle modifies the temporal pattern of tidal velocity asymmetry and the balance of sediment between the ocean and basin, altering the allochthonous sediment availability to the marsh platform. Using a numerical model with the two end-member marsh accretion scenarios described above, results show that, for a static marsh platform, sea level rise causes areas farther from the inlet channel to be submerged for longer periods of time than areas closer to the channel, due to a reduction in drainage efficiency during ebb tidal exchange. The inlet system in this study is ebb dominant under current sea level conditions, but becomes more flood dominant with sea level rise. This response is amplified when the marsh platform is static, transitioning to flood dominance with high magnitudes of sea
level rise (> 70 cm). Marsh platform configuration plays a crucial role in the feedbacks of the system, with the spatial distribution of drainage efficiency throughout the basin and changes in exchange efficiency over a tidal cycle determining the availability of sediment to the marsh platform.

What are the roles of marsh configuration and marsh margin retreat on tidal prism, inlet channel cross-sectional area, and ebb shoal volume? Vegetation loss can change the tidal prism for a tidal inlet system by: (1) increasing the map-view area of open water exchanged between the basin and the ocean during a tidal cycle and (2) reducing the spatial rate of tidal wave attenuation within the basin. In this study, I use a simple conceptual model, as opposed to the numerical hydrodynamic model used in the aforementioned work, to explore the geomorphic response of two Florida inlets, of contrasting wetland configurations, to uniform back-barrier basin vegetation loss that might result from current projections of sea level rise. All results show that vegetation loss causes an increase in tidal prism, inlet cross-sectional area, and ebb shoal volume, but inlets with wetland configurations that strongly increase tidal wave attenuation in the back-barrier basin have an amplified response to vegetation loss. From empirical relationships, it is found that a one percent loss of back-barrier basin vegetated area results in an increase of ebb shoal volume by an amount approximately equivalent to annual to biennial longshore sediment transport rates along the Florida Atlantic coast. The conceptual model developed in this study can be used to examine morphologic response of tidal inlet systems to wetland vegetation loss at similar barrier island complexes worldwide.
CHAPTER 1
GENERAL INTRODUCTION

1.1 Problem Statement

Coastal shoreline response to sea level rise has been studied and debated since the middle of the 20th century, but there have been comparatively few studies that focus on the impact of sea level rise on the morphology and behavior of tidal inlet systems (Dissanayake et al., 2009; FitzGerald et al., 2007; List et al., 1997; Van Goor et al., 2003). A tidal inlet can be considered to be a critical component of a sand sharing system, in which the submerged sand bodies (ebb and flood shoal complexes) and adjacent beaches are interconnected (Dean, 1988). Tidal inlets interrupt the wave current-induced sediment transport alongshore, sequestering sediment in ebb and flood shoals, shallow features that redistribute the pattern of incoming wave energy flux to the shoreline. As a result, shoreline retreat rates near an inlet can be up to two orders of magnitude greater than mean, long-term shoreline retreat rates along uninterrupted portions of sandy coasts (FitzGerald, 1988; Walton and Adams, 1976). If sea level rise increases the tidal prism that inundates the back barrier basin via the inlet, the equilibrium volume of the inlet’s ebb shoal should increase, and the sediment budget for the system will adjust, to accommodate this increase, by removing sediment from adjacent beaches (Dean, 1988). Tidal inlet response to sea level rise is complicated by the relationships between inlet hydrodynamics, morphodynamics, and back barrier basin wetland ecology. This research advances our understanding of the role played by marsh platform accretion on the response of inlet morphology to sea level rise, a critical step in understanding the evolution of sandy coasts, in general, subject to sea level rise.

1.2 Outline of Presentation

This chapter provides background information on tidal inlet and back-barrier basin marsh systems. The elements of a tidal inlet system are described and some details concerning the classification of inlets and equilibrium relationships, from the
scientific literature, are reviewed. Salt marsh systems and their response to changes in relative sea level are also discussed. Chapter 2 presents research that uses a numerical hydrodynamic model to explore changes to the tidal prism of an inlet during sea level rise under two end-member, marsh accretion scenarios. The results of the modeling experiment are used to make a statement about the changes expected to occur to the inlet cross-sectional area and ebb shoal volume. In Chapter 3, the same numerical model is used to explore the control of marsh accretion rate on the availability of sediment to the marsh platform, through alterations to the tidal velocity temporal asymmetry and the marsh platform submergence durations. Chapter 4 presents an investigation of the decadal scale impacts of marsh response to sea level rise. Specifically, how does the decrease in marsh areal extent, through platform edge erosion, change the tidal prism, inlet cross-sectional area, and ebb shoal volume? The investigation is applied to two inlet systems of notably different initial marsh configurations.

1.3 Tidal Inlet Systems

1.3.1 Morphologic Elements

A tidal inlet acts as a principal valve for the exchange of water and sediments between the ocean and a back-barrier basin. A tidal inlet system is composed of the inlet channel, ebb and flood shoal complexes, adjacent barrier islands, and the contiguous wetlands. Each component of the tidal inlet is part of a sand sharing system, which is assimilated into the regional sediment budget. The morphology of an inlet is controlled by geologic setting, river inflow, tidal exchange, and waves. As water levels in the ocean increase with the rising tide, water and sediment is transported through the inlet channel and into the basin, as the flow expands in the open basin it slows and deposits sediment on flood shoals (Dean and Walton, 1973). As the water in the ocean lowers with the falling tide, water and sediment from the basin flows back into the ocean and deposits sediment on ebb shoals (Dean and Walton, 1973). Ebb shoals interact
with longshore currents and waves, leading to a sediment transfer between adjacent shorelines and ebb shoals. The flood tidal prism of the system is the total volume of water that flows into the basin with a rising ocean tide, and the ebb tidal prism the volume of water emptied from the basin during the falling ocean tide. Tidal Inlet systems can be divided into different elements, where each element has its own function in the sand sharing system. Figure 1-1 shows these elements on an inlet typical of a mixed energy system.

1.3.1.1 Flood shoal

The characteristic flood delta morphology (depicted in Figure 1-2) consists of a flood ramp and bifurcating flood channels on the seaward side, and ebbs shields, ebb spits, and spillover lobes on the landward side. Flood currents dominate the flow over the flood ramp and flood channels; this can be seen by the existence of flood-oriented sand waves. Smaller bedforms transition between orientations during the tidal cycle (Hayes, 1980). The ebb shield is the topographic high, which acts to protect the inner shoal during ebb tidal exchange. Flood shoal morphology tends toward an equilibrium shape and volume rapidly and then deposition slows as it approaches an equilibrium volume. This is shown in observations made at St. Lucie Inlet, FL, after the inlet was first opened the deposition of sediment close to the inlet channel was rapid and then slowed, at which point the sediment was deposited further into the inlet basin (Dean and Walton, 1973).

1.3.1.2 Ebb shoal complex

The components of the ebb shoal system are shown in Figure 1-1, and are composed of the main ebb channel, channel margin linear bars, the terminal lobe, swash platform, swash bars, and marginal flood channels (FitzGerald and FitzGerald, 1977; Gibeaut and Davis, 1993; Hayes, 1979, 1980). The morphologic features of the ebb shoal complex are determined by the dynamic balance between a net offshore directed sediment flux induced by the ebb dominant currents and a net onshore directed
sediment flux induced by offshore waves (Hayes, 1980). The channel margin linear bars, levee-like bars that flank the main ebb channel, and swash bars, located on the swash platform, are formed by an interaction of tide and wave generated currents (Hayes, 1980). The terminal lobe, the seaward most sediment deposit, has a shallow slope on the basin side of the shoal, transitioning to a steep slope on the ocean side of the shoal (Buonaiuto and Kraus, 2003).

In some mixed-energy coastal systems, the volume of sediment stored in the ebb shoal complex can be comparable to the volume of adjacent barrier islands (FitzGerald, 1988). FitzGerald (1988) stated from observations that the size of the bar complex is proportional to the size of the inlet and the magnitude of longshore transport. Ebb shoals transfer or bypass sediment from the updrift side of the tidal inlet to the downdrift barrier coast according to the state of maturity of the shoal, the magnitude of the tidal prism, and local wave conditions (Bruun and Gerritsen, 1959; FitzGerald, 1988; Kraus, 2000). Natural mechanisms for this bypassing were first described by Bruun and Gerritsen (1959). The mechanism for bypassing is controlled by the ratio of longshore sediment transport ($M_{mean}$ in yd$^3$/yr) to the maximum discharge of the inlet during spring tide ($Q_{max}$ in yd$^3$/sec), using the equation:

$$r = \frac{M_{mean}}{Q_{max}}.$$  \hspace{1cm} (1-1)

Bruun and Gerritsen (1959) described three mechanisms for transport: 1) by wave induced sediment transport along the edge of the terminal lobe, 2) through tidal current induced sediment transport in the channels, and 3) by the migration of tidal channels and sand bars. The study shows that, for a high ratio of longshore transport to tidal currents ($r = 200-300$), the main mechanism for bypassing is by wave action along the terminal lobe. With small ratios ($r = 10-20$), sediment transfer is mainly controlled by tidal flow bypassing and by sand bar migration. They also noted that at many inlets, bypassing occurs as a combination of these processes.
FitzGerald (1982) used field investigations to determine that a large portion of sediment transported at mixed-energy tidal inlet systems occurs through inlet migration, spit breaching, and the formation and landward migration of bar complexes. FitzGerald (1982) presented the existence of downdrift attachment points for sediment bypassing. The location of these points is controlled by the bypassing bar, which is dependent on inlet size, ebb channel orientation, and the relative strength of wave and tidal forces. Gaudiano and Kana (2001) documented the episodic bypassing by cyclic shoal attachment along the coast of South Carolina. Using historical photographs spanning 58 years, they show a cycle of discrete bars becoming detached from ebb shoals and migrating onshore to adjacent shorelines. FitzGerald et al. (2000) constructed a range of sediment bypassing conceptual models built on work done in previous studies by Bruun and Gerritsen (1959), Bruun (1966). and FitzGerald (1982, 1988), and is summarized in Figure 1-3.

1.3.2 Inlet Classification

Davies (1964) classifies depositional shorelines according to tidal range, where microtidal coasts have a tidal range of up to 2 m, mesotidal coasts have a tidal range of 2 to 4 m, and macrotidal coasts have a tidal range of over 4 m. Hayes (1975) was the first to recognize the influence of tidal range on the morphologic features of tidal inlet and barrier island systems. Using the tidal range classification of Davies (1964) and assuming moderate wave energy, Hayes (1975) described the morphologic features of each type of system. Microtidal coasts are typically long and continuous barrier island chains with few tidal inlets, with small or nonexistent ebb tidal shoals and large flood tidal shoals. Mesotidal coasts typically form drumstick type barrier islands, punctuated by frequent tidal inlets. These systems generally have large ebb shoals and flood shoals that are small or nonexistent. Macrotidal coasts typically consist of intertidal flats and salt marshes rather barrier island systems. Hayes (1979) refines the classification system to include the relationship between tidal range and wave energy. The modified
classification is broken into five zones: tide-dominated (high), tide-dominated (low),
mixed-energy (tide-dominated), mixed-energy (wave-dominated), and wave-dominated.
Figure 1-4 illustrates the classification zone ranges for each combination of wave and
tidal range.

Numerous studies discuss the morphology of inlet systems along the Georgia Bight
barrier system, extending between Cape Hatteras, NC and Cape Canaveral, FL (Hayes,
1994; Hubbard et al., 1979; Nummedal et al., 1977). Nummedal et al. (1977) shows
that the wave energy and tidal ranges are controlled by the width of the continental shelf
and inner shelf slope. As the shelf widens, the tidal ranges are increased and the wave
energy is attenuated. The tidal range increases and the wave energy decreases as
the shelf widens between North Carolina and Georgia and the reverse occurs between
Georgia and Florida where the continental shelf narrows. Hayes (1994) quantifies this
trend in order to makes a comparison between the number of inlets per unit length of
coastline, tidal range, and wave energy. Figure 1-5 from FitzGerald (1996) summarizes
the results of these studies and illustrates the trend in shelf width, tidal range, wave
height, and number of inlets along the Georgia Bight.

Hubbard et al. (1979) used observations from the Georgia Bight to classify inlet
morphology as a function of wave versus tide dominance, as these parameters controls
the zones of equilibrium between onshore and offshore sediment and, therefore, the
location of deposition. Figure 1-6 illustrates some of the morphologic features seen
in each of these systems. They describe tide-dominated inlets as characterized by
an ebb dominated deep central channel flanked by extensive channel margin bars.
In contrast, wave dominated inlets are described to be controlled by predominately
landward transport, having small ebb tidal deltas which are often breached by numerous
shallow channels. Transitional inlets have shoals which are contained in the inlet throat.
Hubbard et al. (1979) also notes that inlet systems in the Georgia Bight that have a
larger tidal range have marshes behind the barriers that are extensive and represent a
large portion of the total marshlands along the entire Atlantic shoreline. The morphologic variables associated with each inlet type are described in Table 1-1.

1.3.3 Equilibrium Relationships

Despite the complexity of tidal inlet systems, some predictive relationships have been developed which relate the volume and location of the ebb shoal complex and the inlet cross-sectional area to the tidal prism. Tidal inlet systems tend to remain in a dynamic equilibrium, and variation from the equilibrium state in one of the components will result in sediment exchange between the components (Dean, 1988; Stive et al., 1998). Larger scale permanent changes to the system can cause the entire inlet system to shift towards a new equilibrium state.

1.3.3.1 Inlet channel

LeConte (1905) was the first to look into the relationship between tidal prism and tidal inlet minimum cross-sectional area, based on observations of a small number of inlets on the Pacific coast of the U.S.. The study concludes that “nature requires 33 square feet of mean-tide section for each and every million cubic feet of tidal waters passing in and out at spring tides.” This pioneering work was expanded by O’Brien (1931) to include more inlets. This study determined that the relationship between tidal prism and cross-sectional area is not quite linear, but follow the power law relationship:

\[ A = CP^q \]  \hspace{1cm} (1-2)

where \( A \) (m\(^2\)) is the cross-sectional area, \( P \) (m\(^3\)) is the spring tidal prism, and \( C \) and \( q \) are empirical parameters obtained from observational data. O’Brien (1969) completed a follow-up study with more tidal inlets, including some on the Atlantic coast and one on the Gulf of Mexico. The addition of this data supported the original study completed by LeConte (1905) that the relationship was linear (\( q=1 \)). Physical models of the system also found that the relationship between tidal prism and cross-sectional
area at un-modified inlets could be approximated as liner (Delmonte and Johnson, 1971; Johnson, 1972; Lin, 1969; Nayak, 1971).

Jarrett (1976) compiled data from 108 North American inlets, 59 on the Atlantic coast, 24 on the Gulf of Mexico, and 25 on the Pacific coast. The study separated the inlets into three categories according to engineering modification: (1) all inlets, (2) inlets with no jetties or a single jetty, and (3) inlets with two jetties. Within the categories, inlets were further separated according to the following geographic locations: (a) inlets on all three coasts, (b) inlets on the Atlantic coast, (c) inlets on the Gulf coast, and (d) inlets on the Pacific coast. The empirical parameters, C and q, calculated for each of the twelve combinations, along with the coefficient of determination is shown in Table 1-2 and Table 1-3. Studies investigating the empirical parameters, C and q, for additional inlet data set include van de Kreeke (1992) and Powell et al. (2006) for North American inlets, Shigemura (1980) for Japanese inlets, Diekmann et al. (1988) for inlets in the Wadden Sea, Gerritsen and Louters (1990) and van de Kreeke (1993) for Dutch inlets, and Hume and Herdendorf (1993) for inlets in New Zealand.

Work presented by Escoffier (1940) can be used to explain the theoretical derivation of the relationship between tidal prism and cross-sectional area. The diagram illustrates the hydraulic relationship between the inlet cross-sectional area and the peak mean velocity through the inlet channel. The peak mean velocity is the maximum flow during a tidal cycle spatially averaged through the cross-section of the inlet throat. This relationship produces the so-called “Escoffier stability curve,” the shape of which is dependent on the hydraulic properties of the inlet (i.e., back basin area, tidal range, inlet length, inlet width, and friction through the inlet throat). It was determined that there is a critical value of the velocity, beyond which net scour of the inlet occurs. Field observations have found that this critical velocity is approximately 1 m/s, regardless of sediment characteristics (Bruun et al., 1960; O’Brien, 1969). If the inlet cross-sectional area is less than equilibrium, the velocity exceeds the critical value,
generating sufficiently high bed shear stress to scour until the equilibrium cross-section is reached. If the cross-sectional area of the inlet is greater than the equilibrium value, the velocity decreases, lowering the bed shear stress, leading to deposition in the channel until equilibrium cross-sectional area is attained. Figure 1-7 illustrates an example Escoffier curve for an inlet.

van de Kreeke (2004) used the Escoffier curve to investigate how the reduction of back basin area for the Frisian inlet, located on the North Sea along the Dutch coast, would change the equilibrium cross-sectional area of the inlet. An extrapolation of the observations of the cross-sectional area change during a 17 year period following basin reduction is within the range of values previously predicted using the Escoffier curve for a new equilibrium (van de Kreeke, 1993). Based on the Escoffier methods, the equilibrium cross-sectional area was expected to reduce from 22,000 m$^2$ to 15,500 m$^2$ van de Kreeke (2004). Field measurements of this transformation indicate an adaptation time scale of approximately 30 years (van de Kreeke, 2004).

1.3.3.2 Ebb shoal complex

Tidal exchange through the inlet carries sediment, some transported landward and deposited on flood shoal, and some transported seaward and deposited on ebb shoals. As ebb shoals grow, they become shallower and more susceptible to wave forces. These wave forces tend to transport sediment off the shoal and back into the nearshore system. These two forces act to balance the growth of the shoal. Some studies from inlets along the North American coast have shown that ebb delta volume depends on the tidal prism, inlet geometry, shoreline configuration, offshore bathymetry, wave climate, littoral drift, sediment size, and freshwater runoff (FitzGerald, 1988; Hicks and Hume, 1996; Hubbard et al., 1979; Marino and Mehta, 1987; Walton and Adams, 1976). Walton and Adams (1976) explored this relationship by comparing ebb shoal volumes under the three wave energy regimes: (1) highly exposed, (2) moderately exposed, (3) mildly exposed. The study uses 44 inlets along the Atlantic coast, Pacific coast, and Gulf
of Mexico to performed a linear regression fit to the power law relationship:

\[ V = aP^b, \tag{1-3} \]

where \( V \) is volume of sediment stored in the ebb shoal \( (m^3) \), \( P \) is the Tidal prism \( (m^3) \), and \( a \) and \( b \) are empirical parameters determined from observations. The results of the study are shown in Table 1-4 for each of the wave regimes and for all the inlets. FitzGerald (1988) used the regression curve to calculate that a change in tidal prism of 5% at Kennebec River estuary would lead to an increase in the ebb shoal volume by between \( 0.57 \times 10^8 \) and \( 1.04 \times 10^8 \) m\(^3\). FitzGerald (1988) estimates that the deficit could cause local sediment erosion of the coast, leading to over 100 m of shoreline recession. The response time of the system to reach this new equilibrium is unknown, but is predicted to take tens of years (FitzGerald, 1988).

Marino and Mehta (1987) build upon the Walton and Adams (1976) study by expanding the factors influencing the ebb shoal volume to include the tidal prism, inlet width to depth ratio, the cross-sectional area, and spring tidal amplitude. They estimate ebb shoal volumes from 18 inlets located on the Atlantic coast of Florida, totaling 420 million cubic meters of sediment. The study uses a dimensional analysis approach to determine which parameters most accurately explain the trends found in the ebb shoal volumes off the Florida Atlantic coast. A linear regression is applied to the observational data to fit the same power law relationship as Walton and Adams (1976), finding \( a = 5.59 \times 10^{-4} \) and \( b = 1.39 \), with a correlation coefficient of 0.75. They propose that this is low due to the influence of other factors outside the wave and tidal energy. They discovered that if all other factors were to remain constant, then a greater width to depth ratio of the inlet channel would result in smaller ebb shoals and vice versa.

Hicks and Hume (1996) calculated sand volumes for 17 ebb shoals located at natural inlets along the New Zealand North Island coast. They found that the main control on ebb shoal volume was tidal prism, the angle between the outflow jet and
shoreline ($\Theta$), and the wave climate. Larger tidal prisms, lower energy wave climates, and larger outflow jet angles (more shore normal) tend to increase the ebb shoal volume. They found that the empirical equation:

$$V = 1.37 \times 10^{-3} P^{1.32} (\sin \Theta)^{1.38}$$ (1–4)

accounts for 83% of the variance in the ebb shoal volume. Including the effects of outflow jet angles is of higher importance on active margin coasts, such as those in New Zealand, where rocky shorelines may dictate this angle. They found that high-energy coasts tend to have smaller shoals than on those on low-energy coasts with a similar magnitude tidal prism.

There are links between tidal prism and the extent (both seaward and downdrift) of the ebb shoal complex (Carr-Betts et al., 2012) and the minimum depth over the ebb shoal crest (Buonaiuto and Kraus, 2003). Carr-Betts et al. (2012) found that the seaward and downdrift extent of the ebb shoal remains constant for tidal prisms less that $10^8$ m$^3$ and increase linearly with tidal prism for tidal prisms greater than $10^8$ m$^3$. This relationship between tidal prism and ebb shoal location was highly correlated for mildly or highly wave-exposed inlets, but there was no correlation for moderately exposed coasts. As tidal prism increases, the ebb shoal spreads further offshore and in the downdrift direction, and the shoal occupies a deeper water depth (Buonaiuto and Kraus, 2003; Carr-Betts et al., 2012).

1.4 Salt Marshes

1.4.1 Types of Salt Marshes

Salt marshes inhabit the upper intertidal region between the tidal flats and the upland environment, and are dominated by a variety of halophytic (salt-tolerant) vegetation. Most marsh morphologies can be classified as ramped or platform, depending on the relative amount of supratidal (high marsh) and intertidal (low marsh) vegetation. Marsh systems with large amounts of high marshes plants exhibit a platform
morphology, those with an evenly distributed amount of high and low marshes have a ramped profile. The distribution of low marsh vegetation species are controlled by physical stresses to the system, such as flooding and anoxia, while the distribution of high marsh vegetation is controlled by interspecific plant competition (Donnelly and Bertness, 2001). Net vertical accretion in marshes is dependent on the balance of tidal range, wind-wave climate, sediment supply, relative sea level rise, sediment compaction, subsidence, and vegetation productivity (Reed, 1990). Marsh sediments can have two modes of origins: allochthonous (mineral, fine grain particles brought from flooding water outside the marshes) and autochthonous (dead organic matter produced from the salt marsh vegetation). The relative amount of sediment types depends on the availability of mineral sediments, tidal range, storminess, vegetation type, and vegetation productivity, factors which vary spatially across the marsh (de Groot et al., 2011).

1.4.2 Salt Marsh Response to Sea Level

Marshes can respond to sea level rise by: (1) actively expanding vertically and laterally if accretion rates are faster than local submergence, (2) maintaining a stable elevation by trapping sediments and accreting vertically at the same rate as local submergence, or (3) the marsh system drowns by not being able to accrete vertically at the same rate of submergence (Orson et al., 1985). When a non-aquatic plant has its roots submerged for prolonged periods of time, the plant is not supplied with sufficient oxygen and will become waterlogged and drown. As the drowning of individual plants occur, the marsh system will lose biomass and the amount of in situ sediment production will become reduced. This further retards the ability of the system to accrete vertically, and the entire marsh system may drown (Orson et al., 1985).

Marsh plants have some negative feedback mechanisms which give them the ability to maintain elevation with moderate levels of relative sea level rise. Morris et al. (2002) completed a study on Spartina alterniflora in South Carolina to investigate how the marsh cordgrass responds to an increase in sea level. They found that the biomass
of this cordgrass increases with depth below high tide (up to a limit). Higher biomass increases aggradation rates on a deeper platform, allowing the plants to equilibrate to a slight increase in the rate of sea level rise. They found that there is an optimal rate of relative sea level rise for plant growth, which also represents the upper limit at which the community is able to sustain the elevation.

Numerous field studies and numerical models have been used to show that the stability of a marsh system is highly dependent on the magnitude of the tidal range of the system (Kirwan and Guntenspergen, 2010; Kirwan et al., 2010; Reed, 1995; Simas et al., 2001). Loss of marsh vegetation can vary greatly between regions, with low tidal range areas (such as the Atlantic coast of North and Central America, the Baltic, and the Mediterranean) being particularly vulnerable (Reed, 1995). Kirwan et al. (2010) compiled the results of five numerical models of marsh evolution to find threshold rates of sea level rise at which marshes would not survive under a variety of suspended sediment concentrations and tidal ranges. They found that for a given suspended sediment concentration, the ability of a macrotidal marsh (TR > 4 m) was able to adapt to a sea level rise rate up to an order of magnitude greater than a microtidal marsh (TR < 2 m). The model results indicate that for suspended sediment concentrations over 20 mg/L and a tidal range over 1 m, the threshold rate of relative sea level would be approximately 10 mm/yr. They predict that this conversion of marsh to subtidal environment would occur about 30 - 40 years after the threshold rate was met.

Donnelly and Bertness (2001) took core samples from a New England marsh system to see the evolution of plant zonal boundaries over time. They found that the lower marshes, dominated by cordgrasses (Spartina alterniflora), slowly moved landward at the expense of higher marsh species, dominated by marsh hay (Spartina patens), spike grass (Distichlis spicata), and black rush (Juncus gerardi). The timing of the onset of this boundary migration was shown to match a regional increase of sea level rise rate (from 2.4 mm/yr to 4.2 mm/yr) recorded by a New York tide gage. The
ability of the cordgrass to oxygenate the substrate allowed the species to survive during
the higher rate of sea level rise. The marsh hay, spike grass, and black rush were not
able to tolerate the relatively high rise rates. The variation in vertical accretion and
tolerance to sea level rise between plants species in difference zones of the marsh
transformed the distribution of plants across the marsh system. Donnelly and Bertness
(2001) predict that the New England marsh will likely become a cordgrass-dominated
system in the future if relative sea level rates remain constant.

Nicholls (2004) used a climate model to show that under an IPCC SRES A1FI
world, that the potential coastal wetland loss due to sea level rise is 5 - 20% by the
2080s. Nicholls (2004) also investigates the socio-economic influence on coastal
wetlands, finding that the wetland loss due to sea level rise is small in comparison to
the potential human-induced direct and indirect influences that increase with population
growth. He finds that coastal wetlands are much more at risk under IPCC scenarios with
lower social environmental consciousness; this factor can greatly influence the marsh
system’s vulnerability to sea level rise. Global coastal wetlands have been declining at a
rate of approximately 1% per year during the late 20th century, primarily due to human
reclamation (Hoozemans et al., 1993).
Table 1-1. Inlet morphologic variables (from *Hubbard et al. (1979)*)

<table>
<thead>
<tr>
<th>Variables</th>
<th>Wave Dominated</th>
<th>Transitional</th>
<th>Tide Dominated</th>
</tr>
</thead>
<tbody>
<tr>
<td>Principal shoal locations</td>
<td>Inside the bay, as a multi-lobate flood delta</td>
<td>In the throat</td>
<td>Seaward of the inlet as long linear channel margin bars</td>
</tr>
<tr>
<td>Ebb-tidal delta</td>
<td>Small and close to the beach</td>
<td>Variable</td>
<td>Large extends far from shore</td>
</tr>
<tr>
<td>Flood-tidal delta</td>
<td>Large; lobate or digitate</td>
<td>Poorly developed or absent</td>
<td>Generally absent</td>
</tr>
<tr>
<td>Channel character</td>
<td>Poorly defined: often multiple</td>
<td>Variable; often one main channel and one or more secondary channels. Unstable in shallower portions 5-10 m depths</td>
<td>Tends toward stability. Depths greater than 10 m.</td>
</tr>
<tr>
<td>Width/depth ratio</td>
<td>Moderate</td>
<td>Very large</td>
<td>Small</td>
</tr>
<tr>
<td>Lagoon</td>
<td>Wide; open</td>
<td>Fringing marsh; marsh-filled</td>
<td>Marsh-filled and channelized</td>
</tr>
<tr>
<td>Swash bars</td>
<td>Poorly developed</td>
<td>Variable</td>
<td>Variable</td>
</tr>
<tr>
<td>Swash platforms</td>
<td>Poorly developed</td>
<td>Variable</td>
<td>Well developed</td>
</tr>
<tr>
<td>Channel margin bars</td>
<td>Absent</td>
<td>Variable</td>
<td>Large</td>
</tr>
<tr>
<td>Sand body character</td>
<td>Tabular</td>
<td>Variable</td>
<td>&quot;Pod&quot;-like; confined to near channels</td>
</tr>
<tr>
<td>Sand by-passing</td>
<td>Bar by-passing</td>
<td>Variable; often in packets; channel abandonment important</td>
<td>Primarily by ebb currents in the main channel and landward transport by waves</td>
</tr>
</tbody>
</table>
Figure 1-1. Tidal inlet elements shown on a typical mixed-energy tidal inlet system.
Figure 1-2. Morphology of a typical flood shoal (reproduced from Hayes (1980)). Arrows indicate dominant direction of tidal currents.

Table 1-2. Empirical parameters for the tidal prism-inlet area relationships developed by Jarrett (1976) (for metric units)

<table>
<thead>
<tr>
<th>Location</th>
<th>All Inlets</th>
<th>No or one Jetty</th>
<th>Two Jetties</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>C</td>
<td>q</td>
<td>C</td>
</tr>
<tr>
<td>All Inlets</td>
<td>$2.41 \times 10^{-4}$</td>
<td>0.93</td>
<td>$3.65 \times 10^{-5}$</td>
</tr>
<tr>
<td>Atlantic Coast</td>
<td>$6.04 \times 10^{-5}$</td>
<td>1.02</td>
<td>$1.98 \times 10^{-5}$</td>
</tr>
<tr>
<td>Gulf Coast</td>
<td>$9.30 \times 10^{-4}$</td>
<td>0.84</td>
<td>$6.94 \times 10^{-4}$</td>
</tr>
<tr>
<td>Pacific Coast</td>
<td>$4.75 \times 10^{-4}$</td>
<td>0.88</td>
<td>$8.83 \times 10^{-6}$</td>
</tr>
</tbody>
</table>
Figure 1-3. Conceptual models of sediment bypassing (from FitzGerald et al. (2000)).
Figure 1-4. Coastal classification (reproduced from Hayes (1979)).

Table 1-3. Number of inlets and correlations for the tidal prism-inlet area relationships developed by Jarrett (1976) (for metric units)

<table>
<thead>
<tr>
<th>Location</th>
<th>All inlets</th>
<th>No or one Jetty</th>
<th>Two Jetties</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>No. of Inlets</td>
<td>r^2</td>
<td>No. of Inlets</td>
</tr>
<tr>
<td>All Inlets</td>
<td>108</td>
<td>0.90</td>
<td>71</td>
</tr>
<tr>
<td>Atlantic Coast</td>
<td>59</td>
<td>0.92</td>
<td>40</td>
</tr>
<tr>
<td>Gulf Coast</td>
<td>24</td>
<td>0.87</td>
<td>21</td>
</tr>
<tr>
<td>Pacific Coast</td>
<td>25</td>
<td>0.92</td>
<td>10</td>
</tr>
</tbody>
</table>
Figure 1-5. Patterns in the number and spacing of tidal inlets as a function of wave versus tidal energy along the Georgia Bight (from FitzGerald (1996)).

Table 1-4. Empirical parameters for the tidal prism-ebb shoal volume relationships developed by Walton and Adams (1976) for different wave energy regimes.

<table>
<thead>
<tr>
<th>Wave Regime</th>
<th>No. of Inlets</th>
<th>a</th>
<th>b</th>
</tr>
</thead>
<tbody>
<tr>
<td>Highly Exposed</td>
<td>7</td>
<td>$8.7 \times 10^{-5}$</td>
<td>1.23</td>
</tr>
<tr>
<td>Moderately Exposed</td>
<td>18</td>
<td>$10.5 \times 10^{-5}$</td>
<td>1.23</td>
</tr>
<tr>
<td>Mildly Exposed</td>
<td>16</td>
<td>$13.8 \times 10^{-5}$</td>
<td>1.23</td>
</tr>
<tr>
<td>All Inlets</td>
<td>44</td>
<td>$10.7 \times 10^{-5}$</td>
<td>1.23</td>
</tr>
</tbody>
</table>
Figure 1-6. Tidal inlet types in the Georgia embayment (reproduced from Hubbard et al. (1979)), including A) tide-dominated inlets B) wave-dominated inlets and C) transitional inlets.
Figure 1-7. Escoffier stability curve (from *Escoffier* (1940, 1977)).
Sedimentary accretion in back-barrier marshes exerts a control on the morphologic response of tidal inlet systems to relative sea level rise. We investigate this phenomenon by applying a numerical hydrodynamic model, paired with empirically derived morphologic relationships, to an inlet system typical of a mixed-energy, tide dominated barrier island system. We consider two end-member, marsh accretion scenarios: (A) no vertical marsh accretion, wherein marsh islands become submerged, changing both the flooded basin area and the spatial pattern of tidal wave attenuation, and (B) a marsh accretion rate equal to the rate of sea level rise, wherein marsh tidal channels deepen but maintain their courses and the associated friction reduction leads to a more efficient tidal exchange between the ocean and back-barrier basin. Model results show a tidal prism increase for both scenarios, leading to increases in channel cross-sectional area and ebb shoal volumes. Under both marsh accretion scenarios, the mechanism of improved tidal exchange efficiency through channel deepening, produces increases of tidal prism that are similar in magnitude. Under conditions with no marsh accretion, an additional mechanism, namely the expansion of flooded basin area, further increases the magnitude of the tidal prism. Scenario A (no accretion) produced a tidal prism, inlet cross-sectional area, and ebb shoal volume increase double that of the increase calculated for scenario B (pace-keeping accretion). The increase in equilibrium inlet channel cross-sectional area, arising from scour processes, exceeds the increase due to sea level rise alone, illustrating the influence of marsh tidal flow processes. Our results indicate that processes that retard back-barrier marsh accretion will have the inadvertent consequences of increasing both inlet channel cross-sectional areas and ebb shoal volumes. Increases in ebb shoal volumes could profoundly impact regional sediment budgets and lead to changes in the erosion-deposition patterns of adjacent shorelines.
2.1 Introduction

A tidal inlet acts as the passageway for exchange of water and sediment between the ocean and a back-barrier basin. The inlet system is composed of the inlet channel, ebb and flood shoal complexes, adjacent barrier islands, and the contiguous wetlands, with each component participating in a sand sharing system, which is assimilated into the regional sediment budget (Dean, 1988; Hayes, 1979). Despite the importance of including the influence of tidal inlets in regional sediment budgets, few studies have investigated the morphologic response of tidal inlet systems to changes in relative sea level (FitzGerald et al., 2008).

The morphology of the inlet system is controlled by local river inflow, tidal exchange, and waves. As water levels in the ocean increase with the rising tide, water and sediment is transported through the inlet channel and into the basin, and upon flow expansion in the basin, transport capacity decreases and sediment is deposited on flood shoals. As ocean water level lowers with the falling tide, water and sediment from the basin flows back out to the ocean and deposits sediment on ebb shoals seaward of the inlet channel. Ebb shoals alter wave transformation and interrupt longshore currents, leading to a sediment exchange between the shoals and adjacent shorelines. Increases to the tidal prism, the total volume of water exchanged during a tidal cycle, can increase the storage of sediment in ebb shoals, which causes the shoals to act as a local sediment sink, altering the wave energy flux, and therefore, the erosional-depositional patterns of the adjacent barrier island shorelines (Dean, 1988; FitzGerald, 1988). Figure 1-1 illustrates some of the components of a typical mixed-energy tidal inlet system.

Two modeling studies exploring inlet response to sea level rise have been undertaken for inlet systems of the West Frisian Islands in the North Sea, which consist of large, un-vegetated tidal flats that connect the open ocean to the Wadden Sea (Dissanayake et al., 2009; Van Goor et al., 2003). Their results predict increases
for inlet channel cross-sectional areas and flood tidal deltas, but decreases in volumes of ebb shoal complexes. These predictions conflict with the changes recorded in field studies performed along the Louisiana coast (FitzGerald et al., 2004, 2007; List et al., 1994, 1997). The inlets in these field studies connect the Gulf of Mexico to Barataria Bay, which has expansive regions of wetlands connected by a complex network of tidal channels. High rates (approximately 10 mm/yr) of relative sea level rise have led to marsh degradation, resulting in an increase in basin area, tidal prism, cross-sectional area, and ebb shoal volumes (FitzGerald et al., 2004, 2007; List et al., 1994, 1997). The loss of more than 1,100 km$^2$ of wetland area has led to an increase in barrier island segmentation and breakup (FitzGerald et al., 2007).

This study explores the role of wetland stability on the hydrodynamic and morphologic response of a tidal inlet system to changes in sea level. We have performed two sets of experiments using the Delft3D numerical model: (A) sea level rise in a basin with no vertical marsh accretion, and (B) sea level rise in a basin where vertical marsh accretion rates are equal to the rate of sea level rise. The first experiment represents a marsh system which is unable to keep pace with sea level rise, whereas the second experiment represents a system that is able. We use these two scenarios to bracket the potential short-term (decadal) response of the system, and to highlight the role of wetland vegetation in this response. We do not investigate marsh edge erosion since it would be expected to occur multiple decades after threshold rates were achieved (Kirwan et al., 2010). The numerical experiments are run for the St. Marys River entrance, on the Florida-Georgia border, a system that is typical of those seen in the Sea Island chain of barrier islands located on the southeast coast of North America. We do not investigate a scenario of marsh expansion for this site because the wetlands cover most of the local undeveloped land and there is little space for lateral expansion.
2.2 Background

Wetland vegetation response to sea level rise is complex, with many factors contributing to the resilience of the system. Wetlands can respond to sea level rise by: (1) actively expanding vertically and laterally if accretion rates are faster than local submergence, (2) maintaining a stable elevation by trapping sediments and accreting vertically at a rate equal to the rate of local submergence, or (3) drowning due to an inability to accrete vertically at the rate of submergence (Orson et al., 1985). If vegetation is submerged, it may become waterlogged and die. This can lead to marsh platform edge erosion, which is expected approximately 30-40 years after threshold rates of sea level rise are reached (Kirwan et al., 2010). The rate of marsh accretion is a balance between storm, ice, and tidal sedimentation, bioproductivity, decomposition, compaction, and subsidence. These processes are controlled by factors that include climate change, sea level rise, and regional tectonics (Argow and FitzGerald, 2006). Numerous field studies and numerical models have been used to show that the resilience of a wetland system to a change in relative sea level is dependent on the magnitude of the tidal range and sediment supply of the system (Kirwan and Guntenspergen, 2010; Kirwan et al., 2010; Reed, 1995; Simas et al., 2001). Marsh ecosystems have the ability to regulate their elevation within a narrow range of the intertidal zone by increasing sediment trapping and bioproductivity when the vegetation is at a lower elevation in the tidal range, leading to feedbacks in the system that promote vertical accretion with increases in sea level (Boorman et al., 2001; Mendelssohn and Morris, 2000; Morris, 2007; Morris et al., 2002). These feedbacks have limitations, given that the optimal rate of plant productivity occurs at a submergence depth equal to the drowning point (Morris et al., 2002).

At most inlets on the U.S. Atlantic coast, wetland vegetation covers a broad areal extent around the back-barrier basin channels. If the wetland system is unable to maintain vertical accretion rates sufficient to keep pace with sea level rise, those broad
areas may become submerged. This would lead to increases in flooded basin area, allowing for an increased tidal prism. The deepening of channels should reduce some of the tidal wave attenuation in the system and lead to a more efficient tidal exchange. Both of these changes would increase the tidal prism and lead to an increase in inlet channel cross-sectional area and ebb shoal volume.

It has been suggested that the size of some morphologic features of a tidal inlet system can be predicted using empirically-derived relationships based on the hydrodynamic characteristics of the system. O'Brien (1931, 1969) compared inlet channel cross-sectional area to tidal prism for stable inlets, which were thought to be in a state of equilibrium, and revealed a power law relationship between the two variables. A larger tidal prism implies that larger flow volumes must be exchanged during a tidal cycle. The cross-sectional area of the channel governs the constriction of flow and the current velocities that occur during the exchange. According to O'Brien (1931, 1969), an equilibrium flow rate exists, at which the current exhibits no net transport in the channel. If the flow velocity exceeds the equilibrium rate, scour will occur until the equilibrium flow rate is attained. If the flow velocity is below the equilibrium rate, channel sedimentation will occur until the cross-sectional area reestablishes the equilibrium flow rate. Jarrett (1976) expanded the investigation performed by O'Brien (1931, 1969) by compiling data from 108 inlets, 59 from the Atlantic coast, 24 from the Gulf of Mexico, and 25 from the Pacific coast, and separated the inlets into three categories: (1) all inlets, (2) inlets with no jetties or a single jetty, and (3) inlets with two jetties. Within each category, they were further separated according to the following geographic locations: (a) inlets on all three coasts, (b) Inlets on the Atlantic coast, (c) inlets on the Gulf coast, and (d) inlets on the Pacific coast. For each of these twelve combinations, power law relationships were fit, and 95% confidence intervals for the power law constants were established.

During tidal exchange, sediment transported landward during the flooding limb is deposited on flood shoal, and sediment transported seaward during the ebbing limb
is deposited on ebb shoal. As ebb shoals grow, the water depths they occupy become shallower and more strongly influenced by the assailing wave field. Wave action entrains sediment and longshore currents transport sediment off the shoal and back to the nearshore system adjacent to the shoal (Kraus, 2000). Therefore, ebb tidal flow and wave action act together to balance the growth of the shoal. Walton and Adams (1976) explored this relationship by comparing ebb shoal volumes under three wave energy regimes defined by wave height and period. They found that their data could be fit to a power law relationship between tidal prism and ebb shoal volume for each of the three wave energy regimes.

2.3 Methodology

2.3.1 Study Site

The Saint Marys Entrance (SME) is a tidal inlet located on the border of Florida and Georgia in southeastern North America that connects the Cumberland Sound to the Atlantic Ocean. The inlet is bordered on the north by Cumberland Island and on the south by Amelia Island. Both Amelia and Cumberland islands are part of the barrier island string known as Sea Islands shown in Figure 2-1. The inlets in this area fall into the mixed-energy tide-dominated range of the classification system developed by Hayes (1979), which categorizes the morphologic features of tidal inlets based on their relative strength of tide and wave forces, a concept further explored by FitzGerald (1996). In general, this island chain characterized by frequently spaced inlets, with well-developed ebb shoal complexes. These barrier islands are separated from the mainland by a complex network of tidal creeks and marsh islands. The banks of the Cumberland Sound and attached channels are mostly bordered by coastal marsh vegetation, covered predominately by the smooth cordgrass, Spartina alterniflora. The area of wetland coverage in the basin, calculated from data available from the NWI, is approximately 197 km², which is approximately 80% of the basin area that lies below the spring high tide water elevation.
The low marsh represents approximately 60% of the total areal extent in Georgia salt marshes, where approximately 80-96% of the net primary production is due to *Spartina alterniflora* (*Frey and Basan, 1985*). Approximately two-thirds of the production of *Spartina alterniflora* occurs as rhizomes below the surface, making it physically inaccessible to grazing herbivores (*Fogel et al., 1989*). Georgia salt marshes exhibit little or no peat due to the extensive rafting of material out of the marsh, the high mean temperatures, intense bacterial degradation, and high levels of bioturbation; a layer of sediment 30 cm thick may remain in the active bioturbation zone for up to several decades (*Fogel et al., 1989*). Average sedimentation rates along the Georgia coast have been estimated as 3 - 5 mm/yr (*Hatton et al., 1983*). Field measurements performed by *Craft* (2007) indicate a wide variation in both the spatial and temporal vertical accretion rate throughout the marsh system, with short-term rates varying between 5.3±0.5 and 9.9±1.1 and long term rates varying between 1.3±0.3 and 3.4±0.6. This variation is likely due to the mixture of sediment sources, both external and internal to the marsh system, which are not supplied at an unsteady pace and/or in a nonuniform spatial pattern (*Nichols and Boon, 1994*). The proportions of sand, silt, and clay vary throughout the marsh system; the creek banks are composed of a mixture of silt- and clay-sized sediment, whereas the high marshes contain mostly sand-sized sediment (*Frey and Basan, 1985*).

The St. Marys, Cumberland, Crooked, North, Jolly, Bell and Amelia Rivers flow into the estuary basin, with the St. Marys River being the primary contributor of fresh water. The river discharge at USGS station 02231254 (located 30.7439 N, 81.6544 W), approximately 20 km inland from the inlet entrance, is generally tidally dominated, punctuated with rain events, which temporarily increase discharge values. Two harbors are located in the estuary, which serve as a base for substantial local commercial and fishing fleets. The Kings Bay Naval base, located approximately 14 km north of the inlet in the Cumberland Sound, serves as a port for naval submarines. In support of the
defense, commercial, and recreational navigation, the inlet is stabilized with two jetties and is regularly dredged to maintain a depth of approximately 14 m (Parchure, 1982). Channels between the inlet and the naval base (north of the inlet), as well as between the inlet and Fernandina Beach Marina (south of the inlet), are dredged to maintain navigation. Bathymetry obtained from the National Geophysical Data Center (NGDC) 3 arc-second Coastal Relief Model is displayed in Figure 2-2.

The tide is predominately semi-diurnal with a mean range of 1.75 m and spring range of 2.5 m. Tidal prism measurement for this inlet have been reported as $1.75 \times 10^8$ m$^3$, $1.58 \times 10^8$ m$^3$, $1.70 \times 10^8$ m$^3$, and $3.3 \times 10^8$ m$^3$ by Bruun et al. (1978), O’Brien and Clark (1974), Environmental Science and Engineering, Inc. (1980), and Powell et al. (2006), respectively. Figure 2-3 shows the basin hypsometric curve, the distribution of basin elevation, illustrating that 25% of the basin resides within the spring tidal range. Using the distribution of basin elevations, and the temporal observations of water levels (taken over 40 tidal cycles), an estimate of the flooded portion of basin area can be calculated. Figure 2-4 displays the temporal distribution of flooded basin area over the range of tidal elevations (shaded area) and the temporally averaged basin area (blue line). The distribution of elevations is bimodal representing the marsh channels and the marsh platforms. During the transition from ebb to flood tide, the marsh channels are the first to experience inundation until the marsh platform edge is overtopped (Zone A), after which time the flooded basin area rapidly expands while the marsh platform (Zone B) continues to be inundated. During the higher portion of the rising tide limb, the region of flooded marsh platform enlarges until maximum tidal elevation is reached (Zone C). The temporally average flooded basin size is 178 km$^2$, but that precise amount of basin area is submerged for only a short duration.

2.3.2 Tidal Flow Model and Experimental Design

Delft3D is a modeling suite developed by Deltares and is capable of simulating three-dimensional unsteady flow fields resulting from waves, tides, rivers, winds, and
other coastal currents. The system of equations consists of the horizontal momentum equation, the continuity equation, the transport equation, and a turbulence closure model, described in detail by Lesser et al. (2004). The model uses curvilinear, boundary-fitted grids, and can be used to provide flow predictions for areas with complex bathymetry. In our simulations, the model is forced with water levels proscribed by a set of astronomical tidal constituents derived from NOAA tidal stations. The model does not incorporate freshwater inflow from local river sources since the net discharge through the inlet channel due to river input into the system is less than 5% of the total inlet discharge attributable to the tides (Parchure, 1982).

The model domain is a curvilinear set of 419 x 319 x 8 grid cells, with an average horizontal grid spacing of 100 m x 100 m in the estuary, increasing in the x-direction towards deeper ocean cells. The model uses a sigma coordinate transformation in the vertical direction, dividing the depth into an evenly spaced, constant number of vertical slices, varying in thickness with changes in water level. This results in a smooth representation of the local bathymetry and high computing efficiency.

Water surface elevations and flow velocities from one model run, simulating a 52-day period was compared with available in situ data, in order to conduct a model validation study. Subsequently, two series of model runs were conducted for each of the wetland accretion scenarios. The first experimental series, with no marsh accretion, was executed by lowering the base level of the bathymetry evenly throughout the model domain. In the second experimental set, with wetland accretion equal to sea level rise, the base level was lowered in areas of the bathymetry that are not classified by the National Wetland Inventory (NWI) as vegetated area. Each series of model runs ran through the following set of sea level rise magnitudes: 5, 10, 20, 30, 40, 50, 60, 70, 80, 90, and 100 cm. All of the simulations were run using tidal elevation conditions during the same six day interval, from November 08, 2011 to November 14, 2011, which comprised 10 full tidal cycles. This model does not incorporate sediment transport into
the system because there is not sufficient data used for calibration to confidently model morphology change to the system. In order to accurately model sediment transport at this location, detailed information about the sediment distribution, longshore transport, and riverine input would be required. Since the model does not incorporate sediment transport, the magnitude, rather than the rate of sea level rise, is used in this study. The rate of sea level rise dictates the time to which each sea level rise magnitude would be reached. For example, a sea level rise rate of 2 mm/yr would require 250 years to reach 50 cm of sea level rise, while a rate of 10 mm/yr would require 50 years to reach this magnitude.

2.3.3 Empirically Derived Hydrodynamic-Morphologic Relationships

In this study, we use the model output of the instantaneous discharge through the inlet channel to calculate a mean tidal prism over the 10 modeled tidal cycles, in order to make estimates about changes to the inlet cross-sectional area and ebb shoal volume. The relationship between the tidal prism and the inlet channel cross-sectional area that we apply in this study was established by Jarrett (1976) for U.S. Atlantic coast inlets with two jetties:

\[ A = 1.5840 \times 10^{-4} P^{0.95} \]  \hspace{1cm} (2.1)

where \( P \) is the tidal Prism (m\(^3\)) and \( A \) is the minimum cross-sectional area (m\(^2\)). Similarly, we use an empirical relationship to investigate changes to the equilibrium ebb shoal volume using a relationship published by Walton and Adams (1976). We chose the relationship developed for moderately exposed coasts, based on the wave energy regime in the area of the Saint Marys Entrance:

\[ V = 6.6 \times 10^{-3} P^{1.23} \]  \hspace{1cm} (2.2)

where \( V \) is the ebb shoal volume (m\(^3\)). We apply these formulas to the tidal prism model calculations for both marsh platform accretion scenarios and all sea level rise
magnitudes in order to understand the control of marsh processes on the response of the inlet morphology to changes in relative sea level.

2.3.4 Model Testing and Calibration

Prior to executing the full suite of numerical experiments, the model was run for purposes of testing and calibration, to simulate tidal conditions for a 52-day period from November 08, 2011 to December 30, 2011. Comparison sites included points closest to the NOAA tidal station and ADCP locations. The NOAA Fernandina Beach Currents Project bottom-mounted workhorse ADCP locations and deployment periods are shown in Figure 2-5.

Water surface elevation comparison with tide station 8720030 is shown in Figure 2-6, with the length of the record broken into four panels. Displayed are (1) the tidal station data (red line), (2) the Delft3D model results (blue), and (3) the difference (green) between observation and model results - underpredictions have positive values and overpredictions have negative values. The phase of the signal appears to be accurately predicted, with most of the differences arising from discrepancies in magnitude. The water elevation differences for this record passed a t-test of the null hypothesis that they follow a normal distribution with a mean of -2.1 cm and a standard deviation of 19.7 cm. The standard deviation is approximately 8% of the range of tidal elevations in this record. Some of this difference is likely due to the lack of wind induced set-up and set-down incorporated into the model.

Time series plots of observed and modeled surface current speeds, for each of the six stations, are shown in Figure 2-7. The data duration displayed is for the first three full days of deployment at each station in order to explore a meaningful comparison. A plot of the surface currents for the entire deployment of each station are shown in Figure 2-8 as U-V scatter plots along with the 95% confidence ellipses of the ADCP current measurements and the co-located modeled currents. The U and V velocities are the easting and northing components of velocity, respectively. The axes are stretched
to show variability (axis exaggerations indicated on plots). For both plots, the colored data represent the ADCP observation data, colored by station, and the black data points are the co-located model results. Figure 2-9 shows the calculated mean speed, speed standard deviation, and principal axis calculated over the deployment period at each of the six stations.

The majority of the model stations have a phase similar to the reported ADCP currents, with the exception of FEB1108, which is located close to the southern jetty. This location may have some modeling inaccuracies due to the proximity of the hard structure (jetty). This location may also be influenced by a combination of channel and alongshore coastal flow; the model output exhibits greater channel directed flow and smaller magnitudes than the ADCP observations at FEB1108. Longshore currents are not included in the model calculations, making this a plausible source of the discrepancy.

The model tends to underpredict current speeds during flood tide at station FEB1101, which is located in the center of the jetty structures on the ocean side of the inlet. This may be due to the influence of the jetties, which constrict flow and increase current velocity through the inlet. The jetties are included in the model as shallow regions, rather than thin dams, because the jetties do allow some passage (overtopping) of water during high tide. Experimental runs using thin dam jetties reduced the flow through the channel and the currents in the basin were significantly underpredicted.

The model overpredicts current speeds at station FEB1104, a point in a narrowing section of the Amelia River, 2.5 km south of the inlet channel, by approximately 20 cm/s. There also exists a depth discrepancy between the NOAA ADCP data and the NGDC Coastal Relief Model data used to construct the model bathymetry at that location; the ADCP depth is reported at 12 m, while the NGDC depth is 8.3 m. This area is in the Quarantine Reach, a dredged section of the basin, maintained at approximately 11 meters. The dates corresponding to the bathymetric survey and ADCP deployments
may be separated by a dredging event leading to this depth discrepancy, which would certainly influence current velocities.

The two ADCPs in the channel throat (FEB1102 and FEB1103) are located on either side of where the model tidal prism calculations are made. This minimum cross-section of the inlet channel was chosen because the total volume flux through this plane over each tidal cycle is equivalent to the tidal prism. The model makes the most accurate velocity predictions in this area; the co-located model outputs have mean speeds within 2.2 cm/s, standard deviations within 4.4 cm/s, and direction within 2.8 degrees of the ADCP data. Model calculations of tidal prisms during the 52-day simulation range from $1.3 \times 10^8$ m$^3$ to $2.7 \times 10^8$ m$^3$. Modeled inlet channel water elevations, modeled instantaneous discharges through the inlet, and modeled tidal prism values, for each tidal cycle, are shown in Figure 2-10. The mean of the tidal prism during this model simulation was $1.87 \times 10^8$ m$^3$, comparing well in the range of reported values of $1.75 \times 10^8$ m$^3$, $1.58 \times 10^8$ m$^3$, $1.70 \times 10^8$ m$^3$, and $3.3 \times 10^8$ m$^3$ published in Bruun et al. (1978), O’Brien and Clark (1974), Environmental Science and Engineering, Inc. (1980), and Powell et al. (2006), respectively. These values are shown, for comparison purposes, in the bottom panel of Figure 2-10 as the dashed horizontal lines.

2.4 Results

Below, we present the modeling results of simulated sea level rise and the morphologic effects organized into four sections corresponding to the variables of interest. We describe the simulated changes in basin area, tidal prism, inlet cross-sectional area, and ebb shoal volume for the incremented sea level rise scenarios in the two, end-member marsh accretion scenarios described above: no marsh accretion and marsh accretion at a rate equal to that of sea level rise.

2.4.1 Basin Area

The change in temporally averaged flooded basin area was calculated for each sea level rise magnitude under the two end member marsh accretion scenarios. Figure 2-11
displays the change in the temporal distribution of flooded basin area and the change in the temporally averaged basin area (vertical lines) with sea level rise for each marsh accretion scenario. The black lines show initial basin area distribution, the blue lines are 10 cm incremental sea level rise steps between zero and 100 cm, with a sea level rise of 50 cm and 100 cm highlighted as green and red lines, respectively. In the case of no marsh accretion, both the channels and marsh platform are flooded with increased frequency and duration. In the case of sea level rise-paced marsh accretion, the channel edges become flooded more frequently with sea level rise, but the platform is flooded with the same frequency and duration. Eventually, the channel sides become steep and are continuously submerged. Figure 2-12 displays the relationship between sea level rise magnitude and temporally averaged submerged basin area for both of the marsh accretion scenarios.

2.4.2 Tidal Prism

The ideal tidal prism, considered to be that which would occur if tidal exchange were instantaneous and uniform throughout the basin, is calculated as the temporally averaged submerged basin area multiplied by the mean tidal range. The modeled tidal prism, averaged for the ten tidal cycles simulated, was calculated from the instantaneous discharge through the inlet channel for both of the marsh accretion scenarios and for each of the sea level rise magnitudes. The upper panel in Figure 2-13 displays the calculated ideal tidal prisms (circles) and modeled tidal prism (squares) for the scenario of zero marsh accretion (red) and marsh accretion equal to sea level rise (blue). In all cases, the ideal and modeled tidal prisms exhibit an increase with rising sea level. For the static marsh scenario, the tidal prism increases at a greater rate than for the pace-keeping marsh accretion scenario, because the basin area increases at a greater rate with rising sea level.

The tidal exchange efficiency was calculated as the ratio of modeled to ideal tidal prism, and is shown in the lower plot in Figure 2-13 over the range of investigated
sea level rise values. The increase in this ratio indicates that the tidal prism exchange becomes more efficient with increases in sea level due to a reduction in tidal wave attenuation. The increase is slightly greater in the marsh accretion scenario because the only additional basin area that is becoming flooded is in the channels. In the case of a static landscape, shallow areas that were not previously within the tidal range are now accessible; these shallow areas retard exchange flow more severely than do the basin channels.

The tidal prism increases due to a combination of the increased basin area and increased tidal exchange efficiency. In Figure 2-14 we separate the increased modeled tidal prism into these two, aforementioned, components for both the static marsh (red) and pace-keeping marsh accretion (blue) scenarios. The total modeled tidal prism increases are shown as solid lines, and the components, from the basin area increase and from more efficient exchange, are shown as circles and dashed lines, respectively. The increase in tidal exchange efficiency contributes a similar amount to each of the systems, regardless of marsh accretion rates, due to channel deepening. The increased efficiency is responsible for approximately two-thirds the increase in tidal prism in the marsh accretion scenario, whereas it is only one-third in the case of a static marsh system.

2.4.3 Cross-Sectional Area

Sea level rise increases the cross-sectional area of the inlet channel due to increasing water level and flooding of adjacent shoreline. The equilibrium value of the channel cross-sectional area, the area necessary to convey tidal flows, also increases with sea level rise due to the scour associated with the increase in tidal prism. The empirical relationship between tidal prism and cross-sectional area developed by Jarrett (1976) is used to calculate equilibrium cross-sectional area for each sea level rise scenario. The upper panel of Figure 2-15 shows the change in cross-sectional area due to sea level rise in the channel (green) and the change in equilibrium cross-sectional
area due to scour from tidal prism increase for a static landscape (red) and for a basin with pace-keeping marsh accretion (blue). The approximate rate of increase is 61 m\(^2\) and 28 m\(^2\) per 1 cm of sea level rise for static marshes and accreting marshes, respectively. At present-day sea level, the measured cross-sectional area is greater than the equilibrium prediction, likely due to channel dredging regularly performed at this site (Johnston et al., 2002). As sea level rises, the equilibrium area increases at a faster pace than the increase in the cross-sectional area due to sea level rise in the channel. For sea level rise magnitudes at which the equilibrium cross-sectional area exceeds the cross-sectional area created by sea level rise in the channel, scour should occur to enlarge the channel by widening or deepening, until sufficient area is achieved to accommodate the increased tidal prism.

### 2.4.4 Ebb Shoal Volume

Using an empirically derived equilibrium relationship developed by Walton and Adams (1976), we explore the impact of sea level rise on the ebb shoal volume of the inlet system. A plot of the change in ebb shoal volume over the range of sea level rise magnitudes for static marshes (red) and pace-keeping marshes (blue) is provided in the lower panel of Figure 2-15. This figure displays the sensitivity of ebb shoals to changes in marsh accretion rates; static marsh environments show an increase in ebb shoal volume more than double the amount calculated when marshes accrete at the same rate as sea level rise. For every 1 cm of sea level rise, increases in the equilibrium volume of sediment stored in ebb shoals of 0.66 km\(^2\) and 0.30 km\(^2\) are predicted to occur for static and pace-keeping accreting marshes, respectively.

### 2.5 Discussion

In this section, we consider the potential impacts of sea level rise on ebb shoal volume and the influence that such a change might exert on local longshore transport patterns. We compare our modeling results with previous studies focused on tidal inlet response to sea level rise and speculate on why there is disagreement with modeling
studies of inlets along the West Frisian Islands in the North Sea. Finally, we discuss the qualitative agreement of our model results with field studies from the Louisiana (U.S.) Gulf coast.

2.5.1 Regional Impacts

The current rate of sea level rise at NOAA tidal station 8720030 (location shown in Figure 3-1), calculated using data over a 110 year record between 1897 and 2006, is $2.02 \pm 0.2$ mm/yr (Zervas, 2009). Assuming a steady rate, the magnitude of sea level rise should be 5 cm in 22.7 years. Results of this modeling study predict an increase in the ebb shoal volume at this site of between $3.6 \times 10^6$ and $6.83 \times 10^6$ m$^3$ (2.23% to 4.24% increase), for the pace-keeping marsh accretion and static marsh scenarios, respectively; nearly a doubling of the ebb shoal volume if accretion halts. The local longshore sediment transport rate has been reported to be approximately $3.8 \times 10^5$ m$^3$/yr (Dean, 1988), totaling $8.63 \times 10^6$ m$^3$ over the 22.7 years. The predicted increase in equilibrium ebb shoal volume is between 41 and 79% of this longshore transport. As tidal prism increases, the ebb shoal spreads further offshore and in the downdrift direction, and the shoal occupies a deeper water depth (Buonaiuto and Kraus, 2003; Carr-Betts et al., 2012). This change in location reduces the impact of wave-induced transport off of the shoals.

If all 16 inlets in the sea island chain were to experience similar increases in ebb shoal volume, the longshore transport alone could not provide adequate sediment to sustain this growth. This deficiency in the sediment budget could trigger shoreline retreat and beach volume loss from the barrier islands within the Sea Island Chain. Although this model was set up for a particular site, the St. Marys Entrance, the results suggest that the relationships and consequences of these interrelated processes are applicable to barrier island systems worldwide.
2.5.2 Comparison with Previous Modeling Studies

There have been two, previously published, modeling studies, which investigate the influence of sea level rise on tidal inlets located along the West Frisian Islands in the North Sea. The first study, by Van Goor et al. (2003) uses a three-element semi-empirical equilibrium model and the second study, by Dissanayake et al. (2009), uses Delft3D, the same model used in our study. Both previous modeling studies conclude that an increase in sea level would drive an increase in cross-sectional area and a decrease in ebb shoal volume. Dissanayake et al. (2009) hypothesized that the decrease in ebb shoal volume is most likely due to the decreased bed friction from inlet channel deepening, allowing for higher flood current velocities through the inlet. The inlets in this area have expansive regions of unvegetated tidal flats, with different basin hypsometries to those seen on the U.S. Atlantic coast. This may account for the difference in ebb shoal change predictions between studies at that location and the research conducted in this study.

Our results are in agreement with field observations reported by List et al. (1994, 1997) and FitzGerald et al. (2004, 2007) at sites located within Barataria Bay, Louisiana, which found an increase in both inlet channel cross-sectional area and ebb shoal volume. Studies of the Barataria tidal inlets have been performed to show how a multiple tidal inlet system evolves with high rates of relative sea level rise (approx. 10 mm/yr) mixed with substantial marsh loss (FitzGerald et al., 2004, 2007; List et al., 1994, 1997). The Barataria Bay is connected to the Gulf of Mexico by four tidal inlets: Abel, Barataria, Caminada, and Quatre Bayou. List et al. (1994, 1997) performed a comparative study of the bathymetric evolution of a 157 km reach of this barrier islands chain, located off the Louisiana coast west of the Mississippi River delta, over three survey periods (1880s, 1930s, and 1980s), with depth soundings between 7 km landward of the islands to an offshore depth of 12 m (3000 km$^2$ total area). Their assessment revealed increases in ebb shoal volume and tidal prism that followed the relationships developed by Walton...
and Adams (1976). They concluded that the changes in barrier islands and tidal inlet systems cannot be considered independent of the hydrodynamic and morphologic behavior of the back-barrier wetlands.

FitzGerald et al. (2007) looked at the evolution of the cross-sectional area of the four tidal inlets connected to Barataria Bay and found consistent increases over time due to the increased tidal prism, likely driven by local wetland loss. Notably, the total cross-sectional area of the inlets has quadrupled since 1880. ADCPs were used to calculate the tidal prism for each of the inlets, and the resulting comparisons with the Jarrett (1976) equation, which relates tidal prism and inlet cross-sectional area, showed that the measured tidal prisms of 3 (out of 4) inlets fell within the 95% confidence interval published by Jarrett (1976). The one inlet that did not match the predicted relationship (Barataria Pass) has a smaller cross-sectional area than predicted, which, the authors postulate, may be due to the stratigraphy of the area; the inlet may be trying to cut into consolidated clays that resist erosion. Measurements from the field studies, described above, follow a similar qualitative trend to the modeling results presented in this paper.

### 2.6 Conclusions

Because of their critical role in the coastal landscape, the response of tidal inlet systems to sea level rise is important to consider over the course of the next century. In this paper, we have demonstrated that the accretionary behavior of back-barrier wetlands exert significant control on the geomorphic response of the inlet systems and their morphologic characteristics (e.g. inlet channel cross sections and ebb shoals). As sea level rises, marshes must accrete vertically at a comparable rate in order to counteract submergence. If marshes do not maintain sufficient vertical accretion, the submerged portion of the back-barrier basin will increase in areal extent, which leads to an increase in the tidal prism. If the marsh vegetation accretes at a pace sufficient to keep up with sea level rise, then only the marsh channels will deepen and unvegetated
coastline will become flooded. We used the two end-member scenarios of zero marsh accretion and pace-keeping marsh accretion to bracket the range of potential inlet system responses.

After satisfactory calibration, the modeling procedure used in this study calculates that sea level rise, in both the non-accreting and pace-keeping marsh cases, drives an increase in tidal prism through two mechanisms: (1) an increase in submerged basin area and (2) an increase in the tidal exchange efficiency from channel deepening. Sea level rise, affecting a landscape with non-accreting marshes, increases the tidal prism approximately twice as much as a landscape with marshes maintaining an accretionary pace comparable to rising sea level.

The morphologic consequences of this increased tidal prism include changes to tidal inlet channel cross-sectional area and ebb shoal volume. Equilibrium inlet channel cross-sectional area increases in both the non-accreting and pace-keeping marsh experiments. In a system experiencing marsh accretion equal to sea level rise, the increase in equilibrium channel area is approximately 1.5 times the increase due to channel flooding alone (i.e. elevated water levels in the channel due to sea level rise alone). In a non-accreting marsh landscape, the increase in equilibrium channel area is approximately 3 times the increase due to flooding alone. Ebb shoal volumes increase due to sea level rise in both non-accreting and accreting marshes. The ebb shoal volume approximately doubles for a system with no marsh accretion compared to a system of pace-keeping marsh accretion. Increases in ebb shoal volumes could profoundly impact the regional sediment budget and lead to changes within the erosional-depositional patterns of adjacent shorelines.
Figure 2-1. Map of the Sea Island Chain with an inset of the Saint Marys River Entrance basin. This map displays vegetated area, NOAA tidal station 8720030, and ADCP deployment locations.
Figure 2-2. Contour map of the Saint Marys Entrance basin bathymetry from data available from the National Geophysical Data Center (NGDC) Coastal Relief Model at 3-arc second resolution.
Figure 2-3. The cumulative distribution of the elevations within the St. Marys River basin model domain (green shaded area). Bars on the left represent the percentage of surface area within each 1 m interval, showing a peak around MSL. The yellow area highlights the elevations that are within the spring tidal range.
Figure 2-4. The temporal distribution of flooded basin area in the Saint Marys River Entrance basin (shaded area) and the temporally averaged flooded basin area (blue line). There are two distinct regions of elevations: the marsh channels and the marsh platform. From low to high tide, the marsh channels expand until the marsh platform edge is reached (Zone A). As the marsh platform is overtopped, the basin area rapidly expands (Zone B). Slowly the flooded area of marsh platform expands until maximum tidal elevation is reached (Zone C).
Figure 2-5. A map of the NOAA Fernandina Beach Currents Project bottom-mounted workhorse ADCP deployment locations and corresponding deployment dates. The colored location dots match the colored bars for each location. Names were retained from the NOAA project, beginning with FEB followed by the two digit year 11 and the project station number (01-04, 07, and 08).
Figure 2-6. Water elevations at NOAA tide station 8720030 and the co-located Delft3D modeled water levels. The model data is shown in black, the tidal observations are red, and the difference is shown in green. Underpredictions have positive values and overpredictions have negative values.
Figure 2-7. Time series plots of surface current speeds from ADCP data and co-located model data for the first 3 full days of deployment at each station. The ADCP data is colored according to the station and the model data is black.
Figure 2-8. Scatter plots of surface current U and V components with the 95% confidence ellipse and principal axes from the six NOAA ADCP stations and co-located model data. The ADCP data is colored according to station, the co-located model data is black, and the principal axes are dashed lines. Note unequal size axes to highlight differences in ellipses.
Figure 2-9. Model comparisons with NOAA ADCP data. A) The mean surface current speed, B) the standard deviation of the surface current speed, and C) the principal axis direction for the six NOAA ADCP stations (red) and the co-located model data (blue).
Figure 2-10. Model output of water level, instantaneous discharge, and calculated tidal prism. A) The water elevations in the inlet channel, B) the model output of instantaneous discharge through the inlet, and C) the calculated tidal prism for each tidal cycle during a 52 day simulation. Tidal prism calculations are between $1.3 \times 10^8$ m$^3$ and $2.7 \times 10^8$ m$^3$ for this simulation, with a mean of $1.87 \times 10^8$ m$^3$. Reported values byBruun et al. (1978), O’Brien and Clark (1974), Environmental Science and Engineering, Inc. (1980), and Powell et al. (2006) are plotted in (C) as horizontal dashed lines.
Figure 2-11. The change in the temporal distribution of flooded basin area and the change in the temporally averaged basin area (vertical lines) with sea level rise. The black lines show initial basin area distribution, the blue lines are 10 cm incremental sea level rise steps between zero and 100 cm. A sea level rise of 50 cm is shown in green, and 100 cm is shown in red. The changes in basin area with sea level in the accreting marsh platform scenario are limited to channel edges. In the model scenarios with a static marsh platform, all areas of the basin that are flooded become flooded for a longer portion of the tidal cycle.
Figure 2-12. The change in temporally averaged flooded basin area due to sea level rise, for a static marsh platform and a marsh platform accreting at the same pace as sea level rise.
Figure 2-13. The ideal tidal prism, the modeled tidal prism, and the ratio of modeled to ideal tidal prism. A) The ideal tidal prism (circles) and modeled tidal prism (squares) for static marshes (red) and accreting marshes (blue), and B) the ratio of modeled to ideal tidal prism. The increase in this ratio indicates that the tidal prism exchange becomes more efficient with increases in sea level due to a reduction in tidal wave attenuation.
Figure 2-14. The change in modeled tidal prism (solid lines) from the basin area increase (circles) and from more efficient tidal exchange (dashed), for static marshes (red) and accreting marshes (blue).
Figure 2-15. Change in cross-sectional area and ebb shoal volume due to sea level rise. The change in cross-sectional area due to channel flooding (green) and the change in the equilibrium cross-sectional area for static marshes (red) and accreting marshes (blue) is shown in (A). The changes in equilibrium ebb shoal volume with sea level rise for static marshes (red) and accreting marshes (blue) is shown in (B). The grey areas highlight the range of predictive changes if marsh vertical accretion were between the end-member scenarios.
CHAPTER 3
ECOGEOMORPHIC FEEDBACKS BETWEEN SEA LEVEL RISE, ESTUARY HYDRODYNAMICS, AND VERTICAL MARSH ACCRETION

The morphologic response of a tidal inlet to sea level rise is dependent on the marsh platform response in the back-barrier basin. The vertical accretion rate of the marsh system controls the flooded basin area and the tidal wave attenuation in the estuary, regulating the tidal prism, inlet cross-sectional area, and ebb shoal volumes. Alterations of the hydrodynamics within the estuary lead to changes in the spatial distribution of marsh submergence. Submergence duration impacts the access of oxygen by halophytic vegetation and, therefore, the stability of the marsh platform. Deviation of temporal symmetry of the tidal current velocity through the inlet channel results in a difference in net transport between the basin and ocean. Net transport influences the availability of sediment for the marsh platform, a critical component of marsh vertical accretion. In this study we used a numerical model to investigate the role of marsh vertical accretion on the hydrodynamic response in the back-barrier basin to changes in sea level rise by investigating two marsh accretion end-member scenarios: (A) no vertical marsh accretion, and (B) marsh accretion at a rate equal to the rate of sea level rise. We ran simulations for a suite of sea level rise scenarios between 0 and 100 cm at 10 cm increment for the Saint Marys Entrance, an inlet representative of a mixed-energy, tide dominated system. Model results indicate for a static marsh platform (scenario A), sea level rise causes areas far from the inlet channel to be submerged for longer periods of time than areas close to the inlet channel, due to a reduction in drainage efficiency during the ebb tidal exchange. The tidal asymmetry for this system becomes more flood dominant with sea level rise because the difference in storage capacity between low and high tide is reduced. The inlet system transitions from ebb to flood dominance at high levels of sea level rise (>70 cm). When the platform is able to accrete vertically at the same rate as sea level rise (Scenario B), the system becomes slightly less ebb dominant, with more gross sediment transport through the inlet,
but similar net transport. Sea level rise alters the back-barrier basin hydrodynamics, changing the distribution of allochthonous sediments and the submergence of the marsh platform, leading to feedbacks between marsh accretion, hydrodynamics, and inlet morphology.

3.1 Introduction

The response of barrier island coastal systems to sea level rise is critically influenced by tidal inlet morphology and hydrodynamics. Inlets act as part of a sand sharing system, in which the inlet channel, ebb and flood shoal complexes, and adjacent barrier islands are interconnected (Dean, 1988; Hayes, 1979). Chapter 2 explores the relationship between marsh vertical accretion in the back-barrier basin and the tidal prism, cross-sectional area, and ebb shoal volume for a range of relative sea level rise scenarios. Results of that study show that marsh platforms with no vertical accretion lead to an amplified increase in tidal prism, inlet cross-sectional area, and equilibrium ebb shoal volume with sea level rise. Increases in the equilibrium ebb shoal volume may provide a sink for longshore sediment transport in the vicinity of an inlet, leading retreat of adjacent shorelines. Sea level rise induced changes in the hydrodynamic characteristics of an inlet system can change the resilience of a marsh platform by altering the availability of sediment and the duration of vegetation submergence. In this study, we used a numerical model to investigate how sea level and marsh platform elevation influence the spatial distribution of tidal range and marsh submergence durations, as well as the tidal inlet velocity asymmetry, in order to understand how changes in sea level would influence factors that control marsh platform resilience.

3.2 Background

Marsh vertical accretion rate is a function of tidal range, sediment availability, and vegetation productivity since the marsh platform accretes through the accumulation of autochthonous plant material and allochthonous sediment (Argow and FitzGerald, 2006; D’Alpaos et al., 2011; Kirwan and Guntenspergen, 2010; Kirwan et al., 2010;
Changes in the hydrodynamics of an inlet system due to sea level rise may impact these factors, leading to feedbacks between the marsh vegetation, hydrodynamics, and morphodynamics. Assuming no net sediment accumulation due to a reduction in flow velocity, increased water depth decreases the friction in the inlet and basin channels, modifying the spatial distribution of flow throughout the basin, the tidal current velocity asymmetry, and the duration of submergence of the marsh platform (Seelig and Sorensen, 1978; Speer and Aubrey, 1985). Changes in the tidal velocity asymmetry alter the net flow of allochthonous sediments into and out of the basin, as well as the availability of sediment for accretion of the marsh platform (Gardiner et al., 2011).

There are two principal controls on tidal inlet velocity asymmetry altered by changes in relative sea level: (1) frictional interaction between the tidal current and channel bottoms, and (2) the intertidal water storage. High ratios of tidal amplitude to channel depth correspond to short ebb durations with high flood velocities and flood dominant sediment transport (Friedrichs and Aubrey, 1988; Seelig and Sorensen, 1978; Speer and Aubrey, 1985). Channel deepening (and therefore friction reduction) should make a system less flood dominant (more ebb dominant).

A high ratio of intertidal storage volume to basin channel volume prolongs flood duration with higher ebb velocities and ebb dominant sediment transport (Boon and Byrne, 1981; DiLorenzo, 1988; Friedrichs and Aubrey, 1988; Nummedal and Humphries, 1978; Speer and Aubrey, 1985). At high tide, the larger flooded surface area does not drain efficiently through the inlet, leading to a substantial lag time between the ocean and basin tidal peaks. During low tide, the surface area is smaller and there is a shorter lag time between the ocean and basin tidal peaks. These differences in lag times result in prolonged flood durations and higher ebb flow velocities/shear stresses (FitzGerald and Nummedal, 1983).
If marsh accretion rate is not sufficient to keep pace with sea level rise, the marsh platform will become submerged for prolonged periods of times and the vegetation will be denied adequate oxygen, leading to plant waterlogging and drowning (D’Alpaos et al., 2011; Kirwan et al., 2010; Mariotti and Fagherazzi, 2010; Morris et al., 2002; Orson et al., 1985). Local plant drowning alters the productivity of the marsh platform system and reduces the availability of autochthonous sediments, making the entire system more vulnerable to submergence with sea level rise (Orson et al., 1985).

3.3 Methods

3.3.1 Study Site

The study site for this research is the Saint Marys Entrance (SME, Figure 3-1), an inlet located on southeastern Atlantic coast of the North America, at the southern end of a region referred to as the Sea Island Chain, which extends from southern North Carolina to northern Florida. The SME is located on the state border between Florida and Georgia, at the south end of Cumberland Island and the north end of Amelia Island. The inlet has characteristics that are typical of a mixed-energy tide dominated system (Hayes, 1979), with a large ebb shoal complex and an expansive region (197 km$^2$) of wetlands in the back-barrier basin.

The back-barrier basin consists of a large, flat marsh platform dissected by tidal channels. Figure 3-2 shows the 3 arc-second resolution bathymetry of the basin derived from the freely-available, National Geophysical Data Center (NGDC) Coastal Relief Model, with the mean high water (MHW) and mean low water (MLW), at 0.88 m and -0.95 m relative to mean sea level (MSL), shown as thin and thick black contour lines, respectively. The cumulative distribution of elevation within the basin is shown in Figure 3-3 as the green shaded region. The elevations are divided into 1 m intervals, and the percentage of the basin that resides within each of those intervals in shown in the bar graph on the left. The proportion of basin area that lies within the spring tidal range is shown in yellow.
Given that a substantial fraction of the basin has an elevation within the spring tidal range, the size of the flooded basin area more than doubles during a tidal cycle. The temporal distribution of the flooded basin area calculated using model simulations described in section 3.3.2 is shown, as the shaded area, in Figure 3-4. As the basin fills during the rising limb of the tidal cycle, the sloped channel edges become submerged, then, as the marsh platform edge is overtopped, the fraction of the basin submerged quickly increases drastically in size. As the tide continues to rise, the flooded area of the platform continues to expand, until the maximum elevation is reached. The reverse pattern is witnessed during the ebbing portion of the tidal cycle.

3.3.2 Model Setup

The three-dimensional hydrodynamic numerical modeling suite, Delft3D, was used to investigate the role of vertical marsh accretion on the hydrodynamic response of the back-barrier basin to increases in sea level rise. We applied two end-member responses of marsh accretion to the model: (A) no vertical marsh accretion, and (B) vertical marsh accretion that keeps pace with sea level rise. Areas of marsh coverage were determined using vegetation coverage information available from the National Wetlands Inventory (NWI) provided by the U.S. Fish and Wildlife Service. In model simulations using an accreting marsh platform (scenario B), only the basin channels, unvegetated shorelines, and ocean deepened with the rising sea level. In simulations using a static platform (scenario A), increases to mean sea level were applied evenly throughout the entire model domain (i.e. ocean, basin channels, marsh platforms).

A series of model simulations, covering a range of sea level rise magnitudes, between 0 and 100 cm at 10 cm increments, were conducted for both marsh accretion scenarios. The model grid is curvilinear, consisting of a horizontal grid of 419 x 319 cells, which is further expanded vertically into 8 layers. The average grid spacing in the basin is 100 m x 100 m, increasing in the x-direction (eastward) into the deepening ocean domain. We utilized a sigma coordinate transformation in the vertical direction,
which divides water elevation into evenly spaced slices, varying in thickness with changes in water elevation.

In these model simulations, we do not incorporate sediment transport or morphologic changes, because of a lack of sufficient *in situ* observations to adequately calibrate simulations of sediment transport within the system. A calibration and validation study was performed using NOAA ADCP data, tidal station water elevations, and previously documented tidal prism measurements. Detailed information and results of this study can be found in Chapter 2.

### 3.4 Results

#### 3.4.1 Spatial Distribution of Tidal Range and Marsh Submergence

As mentioned above, for simulations in which the marsh platform is able to maintain vertical accretion at the same pace as sea level rise, only the basin channels and their adjacent unvegetated shoreline have increased depth. Deeper basin channels reduce the degree of tidal wave attenuation within the basin, leading to an increase in the tidal range in areas far (10 - 20 km) from the inlet channel. This can be seen in the Figure 3-5, where the change in tidal range, with respect to its value for present sea level, is shown for 10, 20, 40, 60, 80, and 100 cm of sea level rise. Channel locations close to the throat show small increases in tidal range (e.g. $\approx 1\text{-}5 \text{ cm per meter of sea level rise}$ for sites within a 10 km along-channel path of the inlet throat), but the tidal range increase grows with greater along-channel distances. The increase in tidal range at distal regions of the basin raises the effective basin area contributing to the tidal prism. An increase in tidal range also occurs on unvegetated sections of the channel edges as sea level rise causes those areas to reside lower within the tidal range.

For simulations in which marsh platform elevation is static, sea level rise causes a decrease in the tidal range in channel reaches within approximately 15 km of the inlet throat because water is redistributed onto the marsh platform. Once the flow is allowed to expand horizontally, the vertical change in water elevation is reduced. Figure 3-6
shows the change in tidal range throughout the basin on a static marsh platform. Tidal ranges for all areas of the marsh platform increase, with the greatest increase for locations close to the inlet channel, probably because these are the locations where tidal exchange is most efficient.

Figure 3-7 shows the change in tidal range for seven locations within basin channels in order to highlight the difference in tidal range change between the marsh accretion scenarios. The four points within 15 km from the inlet channel experience a slight increase in tidal range with sea level rise when the marsh platform is accreting vertically at the same pace as sea level rise. Points in more distal regions, with longer along-channel travel distances, experience more dramatic increases in tidal range with sea level rise. A similar trend is seen in the case of a static marsh platform, where these distal points have increased tidal ranges due to a reduction in friction in the channels allowing more water to access this portion of the basin. When the marsh platform is static, the closer points decrease in tidal range due to the distribution of water onto the adjacent marsh platform.

When the marsh vegetation is able to accrete vertically at the same pace as sea level rise, little change in submergence duration should occur since the platform maintains a steady elevation, relative to sea level rise, within the tidal range. Under present sea level conditions, the marsh platform is submerged approximately 50% of the time throughout the basin. Model results show that when marsh platform elevation is static (non-accreting), the submergence duration increases non-uniformly, in spite of a nearly uniform marsh platform elevation throughout the back-barrier basin. Figure 3-8 shows the spatial distribution of submergence frequency for sea level rise magnitudes of 0, 20, 40, 60, 80, and 100 cm on a static marsh platform. Distal marsh platform areas, located far from the inlet channel, experience longer durations of submergence, for a given sea level rise, as compared to proximal areas.
Figure 3-9 displays the change in the water depth at three points (Site A, Site B, and Site C shown in the figure from top to bottom) on the marsh platform for sea level rise conditions of 0, 20, 40, 60, 80, and 100 cm, on a static marsh platform. Sites A, B, and C are approximately 18, 13, and 9 km along-channel from the inlet throat, respectively. While there are increasing maximum depths during high tide with sea level rise at all three locations, Site A shows increasing low tide levels because the water does not fully draining from the platform during low tide. These plots illustrate the non-uniform spatial distribution of drainage efficiency. Areas of the marsh platforms that are farther from the inlet channel do not fully drain during a tidal exchange and, therefore, have a slightly lower increase in tidal range and maintain a layer of water for a larger proportion of the tidal cycle.

3.4.2 Tidal Velocity Asymmetry

Figure 3-10 (A) shows a velocity-stage plot for results from three numerical model simulations: (i) for the present sea level condition, (ii) for a sea level rise of 60 cm on a static marsh platform, and (iii) for 60 cm on an accreting marsh platform. The plot illustrates the flow speed at different times during the tidal cycle. The time series of water elevation in the channel and flow speeds are shown in Figure 3-10 (B). The velocity-stage plot demonstrates that there is a reversal from flooding to ebbing flow (negative to positive velocity values), referred herein as high water flow reversal (HWFR), that occurs nearly two hours after high tide, during the falling limb of the tidal cycle. This delay is due to the phase shift between the tide in the basin and ocean caused by the inertial effect of the water in the inlet. HWFR is observed to occur at two discrete water levels. The flow also reverses from ebbing to flooding (positive to negative velocity values), referred to herein as low water flow reversal (LWFR), during the rising limb of the tidal cycle, also at two water levels. The two water levels, occurring during each flow reversal, are caused by diurnal inequality produced by the effect of the moon’s declination, which contribute to changing the water level of high and low tide occurrence.
For present sea level conditions, flow into the basin (tidal flooding) occurs for a longer portion of the tidal cycle than flow out of the basin (tidal ebbing); this causes the peak flow speeds to be greater during the ebb exchange than during the flood exchange. This difference in flow durations is created by the interaction of the tidal harmonic constituents of the forcing tides. Sea level rise causes peak current velocities, during both ebb and flood exchange, to increase in both the static and accreting marsh scenarios, though the effect is demonstrably greater in model simulations for the static marsh platform. The position within the tidal cycle at which the flow reverses from flooding to ebbing does not change significantly with sea level rise in either of the marsh accretion scenarios. The reversal from ebb to flood tidal exchange is delayed in the simulations with a static marsh platform, causing the ebbing flow to operate over an increased portion of the tidal cycle.

The instant at which the water elevation on the basin side of the inlet channel is equal to the elevation on the ocean side of the inlet channel represents the time at which there is no tidal flow through the inlet channel - identified above as the reversals (HWFR and LWFR). The change in the timing of HWFR and LWFR as a function of sea level rise is shown in Figure 3-11. Figure 3-11 (A) shows this change for an accreting marsh platform and Figure 3-11 (B) shows this change on a static marsh platform. An increase in the lag during HWFR should prolong flood tidal exchange and an increase in the lag during LWFR should prolong ebb tidal exchange. The relative durations of each lag change determines if the system is in transition to a more flood or ebb dominant system. On a marsh platform that keeps pace with sea level rise, the LWFR lag increases and the HWFR lag decreases, leading to an overall increase in the duration of the ebb tidal exchange. Sea level rise on a static marsh platform increases the lag of both the LWFR and HWFR. In this case, the LWFR lag is greater than the HWFR lag, and therefore, the system has increased ebb tidal exchange durations.
At present sea level, the inlet is classified as an ebb dominant system because of the higher ebb directed current velocities. Bedload sediment transport is proportional to the cube of the spatially-averaged velocity across the inlet channel resulting in a net seaward sediment transport in ebb dominant systems (Bagnold, 1963; Fry and Aubrey, 1990). Figure 3-12 and 3-13 show model results of the spatially averaged velocity, the cube of the spatially averaged velocity, and cumulative sum of the cube of the spatially averaged velocity for the accreting marsh platform and the static marsh platform, for sea level rise magnitudes ranging between 0 and 100 cm at 10 cm intervals. Positive values represent flow in the ebb direction. For both marsh accretion scenarios, the magnitude of the spatially averaged velocity increases in the ebb and flood direction with sea level rise. This increase is amplified when the marsh platform is static. The velocity cubed is used as an estimate of net transport in order to estimate dominant direction. Results for the accreting marsh platform show an overall slight increase in the gross transport through the inlet, and a slight shift toward the balance of transport in the flood dominant direction. For a static marsh platform, the gross transport through the inlet increases, and the net transport transitions from being dominantly in the ebb direction to the flood direction with increases in sea level.

Figure 3-14 displays model results of the ratio of flood to ebb transport, estimated as the spatially-averaged velocity cubed for the range of sea level magnitudes investigated on an accreting and a static marsh platform. On an accreting marsh platform, the ratio of flood to ebb transport slightly increases with higher magnitudes of sea level rise, implying that the system becomes slightly less ebb dominant. When the platform is static, the ratio of flood to ebb transport increases, becoming a flood dominant system at high levels of sea level rise (>70 cm). Once the transition to flood dominance occurs (ratio of flood to ebb sediment transport greater than 1), the ratio of flood to ebb sediment exchange becomes less sensitive to sea level rise.
3.5 Discussion

In our model results, we reveal non-uniform increases to the vegetation (marsh platform) submergence duration when the marsh was static (non-accreting), despite the relatively flat topography on the platforms. It is notable that, holding other factors constant (sediment distribution, vegetation types, etc.), marsh vegetation at greater distances from the inlet channel may, surprisingly, be more vulnerable to drowning from an increase in local sea level. This phenomenon arises because of the decreased drainage efficiency at distal regions of the basin. When an area of vegetation dies off, the plants cease to contribute to the allochthonous sediments and the marsh may become less efficient at vertical accreting. Vertical accretion at a rate slower than the rate of sea level rise will lead to an increase in the submergence durations that the system experiences during increases in local sea level, leading to a feedback that puts the entire marsh system at risk of submergence and die-off.

The SME back-barrier basin has a large portion of marsh platform at an elevation close to present mean sea level and is dissected by numerous tidal channels. The flooded area of the basin approximately doubles in size when comparing between low and high tide. This type of basin hypsometry is in line with those described to be ebb dominant, which is seen in model results by the prolonged flood duration and higher quantities of net sediment transport out of the basin, seaward. Both the friction in the channels and the drainage efficiency of the system are altered by changes to local sea level and by the rate of vertical marsh accretion. In model experiments with accreting marshes, the basin channels deepen, thereby reducing the frictional drag within the system, which would be expected to lead to an increase in the ebb dominance of the system (Friedrichs and Aubrey, 1988; Seelig and Sorensen, 1978; Speer and Aubrey, 1985). This is not seen in our results, possibly due to the flooding of unvegeted areas adjacent to the basin channels imposing a greater influence on the system. There is
only a slight change in the tidal velocity asymmetry in this system (approximately 1% increase in the ratio per 10 cm of sea level rise).

In model simulations with a static marsh platform, the flooded basin area increases during lower tidal stages and the difference in the water storage between high and low tide is reduced. This reduces the difference in the lag time between the ocean and basin high and low tide occurrence times, decreasing the ebb dominance of the system. This change is illustrated by the increase in the LWFR lag shown in figure 3-11. At sea level rise over 70 cm, the change in storage capacity due to sea level rise is reduced because the marsh platform is flooded for the majority of the tidal cycle. At that point frictional changes become a more significant force for change in the system’s behavior.

Tidal asymmetry has been found to be an important factor in the transport and accumulation of sediment in an inlet system. When the marsh platform is static, the net sediment transport transitions from ebb to flood dominance. This increase in transport into the basin provides more allochthonous sediment to the marsh platform, increasing the ability of the platform to accrete vertically. A correlation between flood tidal dominant systems and high rates of marsh accretion has been documented (Gardiner et al., 2011).

This feedback could help to protect the marsh system by allowing it to maintain vertical accretion rates equal to sea level rise. Increases in sediment transport into the basin may have implications for longshore sediment movement and accumulation on adjacent shorelines. These changes to the balance of the inlet system could alter the alongshore erosional-depositional pattern and lead to local shoreline erosion. This impact may be amplified by changes to the equilibrium ebb shoal volume, leading to a local sediment sink in both the marsh platform and the ebb shoal (Lovering and Adams, in review).
3.6 Conclusions

In this research, we aim to understand the role of vertical marsh accretion and sea level rise on some of the sediment feedback within the estuary, specifically submergence duration and tidal velocity asymmetry. Submergence duration is a critical factor in the vegetation’s ability to maintain adequate access to oxygen and the plant’s susceptibility to drowning. If the vegetation drowns then there is decreased in situ sediment production and the entire marsh system will have less autochthonous sediment for vertical accretion. The tidal velocity asymmetry in the inlet system controls the net transport of sediment into and out of the estuary, impacting allochthonous sediment delivery to the marsh platform. Using the results of our numerical modeling study, we conclude:

- **Submergence duration** increases non-uniformly throughout the basin with sea level rise, despite the relatively flat marsh platform topography, due to a decrease in drainage efficiency throughout the basin and inlet channels. Sea level rise causes distal reaches of the marsh platform to maintain a layer of water cover for a longer portion of the tidal cycle than regions more proximal to the inlet channel, making these areas more susceptible to waterlogging and drowning.

- **Tidal velocity asymmetry** shifts in the flood dominant direction with sea level rise due the decrease in the difference in storage capacity between high and low tide for both vertical marsh accretion scenarios. This effect is amplified when the marsh platform is static.

- **A shift in the tidal velocity asymmetry** in the flood dominant direction causes an increase in the net sediment transport in the basin direction leading to basin infilling and more sediment availability to the marsh platform for vertical accretion. The sediment delivered to the marsh platform is at the expense of other morphologic elements of the inlet, including the ebb shoal complex and adjacent barrier islands, which could lead to erosion of the local shoreline.
Figure 3-1. Map of Saint Mary’s Entrance with marsh vegetated area highlighted, inset on a map of the Sea Island Chain. SME is the southernmost inlet in the Sea Island chain located on the Atlantic coast of the US.
Figure 3-2. Map of elevations in the Saint Marys Entrance referenced to MSL, with the MLW and MHW elevations contours shown. Elevations are from the National Geophysical Data Center (NGDC) Coastal Relief Model at 3-arc second resolution.
Figure 3-3. The distribution of elevations in the Saint Marys Entrance basin is shown in the bar graph. The elevations that lie within the spring tidal range are highlighted in the yellow zone. The grey shaded regions is the cumulative distribution of elevations, indicating the percentage of the basin that is above each elevation. There is a peak in the basin area is around MSL, with approximately 20% of the basin residing within 0.5 m of MSL and approximately 25% within the spring tidal range.
Figure 3-4. The temporal distribution of flooded basin area in the Saint Marys Entrance basin for present sea level conditions. The flooded area of the basin is always 100 km$^2$ or larger. As the tide rises from low tide, the channel banks are submerged and the flooded area increases to approximately 150 km$^2$. As the marsh platform is overtopped, the flooded area grows quickly to nearly 225 km$^2$. At that point, the tide rises to its maximum elevation and the flooded basin area continues to expand until it reaches a maximum of 240 km$^2$. 
Figure 3-5. The spatial distribution of the change in tidal range from present sea level conditions as a result of a sea level rise on an accreting marsh platform. Increases in tidal range occur in areas of the basin channels that are farther from in the inlet and along non-vegetated areas of the channel edges. Little change occurs on the marsh platform or in the basin channels that are close to the inlet.
Figure 3-6. The spatial distribution of the change in tidal range from present sea level conditions as a result of a sea level rise on a static marsh platform. Increases in the tidal range on the marsh platform occur over the entire platform, but there is an amplified increase in areas that are closer to the inlet. Tidal range in the channels decrease, likely due to the redistribution of water onto the adjacent platforms.
Figure 3-7. The change in the average tidal range over 10 tidal cycles due to sea level rise for seven channel locations, indicated on the map in (A). The change is shown for an accreting (B) and a static (C) marsh platform. When the marsh platform accretes vertically, the tidal range in all of the channel locations increase due to a reduction in tidal wave attenuation with channel deepening. In the static marsh platform, more proximal locations decrease in tidal range because the flow is distributed between the channels and the marsh platform.
Figure 3-8. The spatial distribution of the submergence duration due to sea level rise on a static marsh platform. The duration is shown as the percentage of time during a 10 tidal cycle period. Initially, the marsh platform is submerged approximately 50% of the time. The submergence durations increase throughout the basin as sea level increases, with more dramatic increases at more distal areas of the basin.
Figure 3-9. Time series of water depth at three points located on the marsh platform for sea level rise of 0 to 100 cm at 20 cm intervals for a static marsh system. The location of the points are indicated on the map, labeled as site A, B, and C. They are at approximately 18, 13, and 9 km along-channel distances from the inlet channel, respectively. The two closer sites (B and C) tend to drain more efficiently, with a more rapid decreases in water level, reaching zero at low tide for sea level rise up to 60 cm.
Figure 3-10. The velocity stage plot and the time series of the water levels and the spatially averaged velocity through the inlet for present sea level and a sea level rise of 60 cm on a static and an accreting marsh platform. The top plot shows the velocity stage plot for the spatially averaged current across the inlet cross-section for present sea level (black) and for a sea level increases of 60 cm on an accreting (blue) and static (red) marsh platform for the tidal signal shown in the lower plot. On the lower plot, the green line is the time series of water elevations and the black, blue, and red lines correspond with the top plot. Positive values of current speed represent flow out of the basin (ebb currents) and negative values represent flow into the basin (flood current). The change from ebb to flood and from flood to ebb are indicated as the low water flow reversal (LWFR) and high water flow reversal (HWFR), respectively.
Figure 3-11. The change in time at which tidal flow reversals occur due to sea level rise for A) an accreting and B) a static marsh platform. This is shown for the high water flow reversal (HWFR) and the low water flow reversal (LWFR). The time change of the reversals indicate the change in duration of the flood and ebb tidal exchange, with an increase in the HWFR indicating longer flood durations, and an increase in the LWFR indicating longer ebb durations.
Figure 3-12. The spatially averaged velocity across the inlet, the cube of the spatially averaged velocity, and the cumulative of the spatially averaged velocity cubed for sea level rise on an accreting marsh platform. Positive flow values indicate flow out of the basin (ebb current) and negative values indicate flow into the basin (flood current). The cube of the velocity is used as a proxy for the net bedload sediment transport, with the cumulative of the cube indicating the dominant transport direction.
Figure 3-13. (The spatially averaged velocity across the inlet, the cube of the spatially averaged velocity, and the cumulative of the spatially averaged velocity cubed for sea level rise on a static marsh platform. Positive flow values indicate flow out of the basin (ebb current) and negative values indicate flow into the basin (flood current). The cube of the velocity is used as a proxy for the net bedload sediment transport, with the cumulative of the cube indicating the dominant transport direction.)
Figure 3-14. The ratio of the flood to ebb sediment transport proxy is shown for an accreting marsh platform (blue) and a static marsh platform (red) for an increase of sea level between 0 and 100 cm. The spatially averaged velocity cubed is used as a proxy for sediment transport. A value of one represents equal flood and ebb transport, with values less than one indicating ebb dominant transport, and greater than one representing flood dominant transport.
CHAPTER 4
THE ROLES OF MARSH CONFIGURATION AND MARSH MARGIN RETREAT ON TIDAL INLET MORPHOLOGY

Morphologic evolution of tidal inlets depends on the ecogeomorphic behavior of the back-barrier basin, and exerts a strong influence on local shoreline response to sea level rise. A tidal inlet channel acts as the principal valve for water and sediment exchange in a barrier system, but changes to back-barrier basin ecology, hypsometry, and flow network configuration can alter discharge conveyed through the inlet channel. Wetland vegetation loss can change the tidal prism for a particular inlet by: (1) increasing the map-view area of open water exchanged between the back-barrier basin and the ocean during a tidal cycle and (2) reducing the spatial rate of tidal wave attenuation within the basin. We use a simple conceptual model to explore the geomorphic response of two Florida inlets, of contrasting wetland configurations, to uniform back basin vegetation loss that might result from current projections of sea level rise. All results show that vegetation loss causes an increase in tidal prism, inlet cross-sectional area, and ebb shoal volume, but inlets with wetland configurations that strongly increase tidal wave attenuation in the back-barrier basin have an amplified response to vegetation loss. Using empirical relationships, we find that a one percent loss of back-barrier basin vegetated area results in an increase of ebb shoal volume by an amount approximately equivalent to annual to biennial longshore sediment transport rates along the Florida Atlantic coast. The conceptual model developed in this study can be used to examine morphologic response of tidal inlet systems to wetland vegetation loss at similar barrier island complexes worldwide.

4.1 Introduction

The response of a tidal inlet to sea level rise has been shown to be highly dependent on the ecogeomorphic response of vegetation in the adjoining back-barrier basin. The Barataria Bay, located in southeastern Louisiana, has experienced marsh loss due to sea level rise and high energy storm waves (Barras, 2006; Barras et al.,
1994; Britsch and Dunbar, 1993), which has led to an increase in the tidal prism of the interconnected four inlet complex, an increase in the cross-sectional area of the inlet channels, and an increase in ebb shoal volumes (FitzGerald et al., 2004, 2007; List et al., 1994, 1997). Studies of the area have shown that during the past 100 years there has been a rapid increase in the number and size of inlets (Levin, 1993; McBride et al., 1992), which may be a direct response to increases in tidal prism volumes. The increase in ebb shoal volume resulting from increased tidal prism is in agreement with the predictive relationship developed by Walton and Adams (1976) (List et al., 1997). FitzGerald et al. (2007) found that the four tidal inlets have quadrupled in size since 1880. Three of the four inlets fell within the 95% confidence interval of the tidal prism and cross-sectional area relationship developed by Jarrett (1976). Relationships between tidal prism and morphology have been extensively studied and quantified, however, the process by which wetland degradation and conversion to open water exerts control on the tidal prism of a system must be further explored. Studies using the HadCM3 climate model predict that, under an IPCC SRES A1FI world, the potential worldwide coastal wetland loss due to sea level rise is estimated to be 5 - 20% by the 2080s (Nicholls, 2004).

This study examines the influence of back-barrier basin vegetation loss on transformation of the tidal prism and, subsequently, cross-sectional area and ebb shoal volume. We investigate two mechanisms for tidal prism change resulting from wetland area conversion to open water: (1) the contribution of increased back-barrier basin open water area to tidal prism volume, and (2) reduction in tidal wave attenuation within the back-barrier basin. We investigate two inlets with similar tidal ranges and wave climates but with strongly contrasting initial ecogeomorphic conditions in their accompanying back-barrier basins, in order to highlight the control of the initial wetland configuration on the evolution of the tidal inlet due to vegetation deterioration. We conclude with a discussion of the implications of continued sea level rise on the morphologic evolution
of tidal inlets, ebb shoal, and adjacent beach/dune complexes and comment on the
application of this concept to barrier systems worldwide.

4.2 Model Details

We have developed a conceptual model to explore the process linkages among
back-barrier basin wetland vegetation, hydraulics of a tidal inlet, and the geomorphic
evolution of the inlet system under a scenario of wetland conversion to a subtidal
environment. Figure 4-1 organizes the processes and linkages, which are described
in some detail, below. If the rate of sea level rise is rapid enough to outpace vertical
accretion of the vegetation, non-aquatic plants may experience root submergence for
prolonged periods, preventing access to oxygen, which will promote waterlogging and
subsequent drowning (Figure 4-1, Process A). As plant drowning occurs, the wetland
system loses biomass, decreasing the amount of in situ sediment production. This
inhibits the basin’s ability to accrete vertically, and the back-barrier ecosystem may
become more vulnerable to inundation from sea level rise (Mudd et al., 2009; Orson
et al., 1985). The susceptibility of salt marshes to inundation has been shown to be
dependent on the sediment supply and tidal range of the system; systems with greater
sediment availability and higher tidal range are able to accrete vertically at a more rapid
pace (D’Alpaos et al., 2011; Kirwan and Guntenspergen, 2010; Kirwan et al., 2010;
Reed, 1995; Simas et al., 2001). There exists a threshold rate of sea level rise for which
marshes fail to maintain a sufficient vertical accretion rate; if this threshold value is
reached, then degradation can be expected to commence approximately 30-40 years
later (Kirwan et al., 2010). Mangrove forests respond similarly to sea level rise and are
sensitive to the availability of allochthonous sediments (Ellison and Stoddart, 1991).
The relationship between sea level rise and vegetation degradation is not quantitatively
addressed in this study, but is a key component in the sequence of back-barrier basin
expansion. Sea level rise could lead to submergence of low-lying, unvegetated land,
generating increased basin area and tidal prism (Figure 4-1, Process B). In this
study, this is considered a negligible increase since coastal structures border most unvegetated sections of the basin boundaries in our study sites, but this should be addressed in undeveloped locations. Sea level rise increases depth in the back-barrier channels thereby reducing the spatial rate of tidal wave attenuation (Figure 4-1, Process C), a component which is significantly smaller than the effects of vegetation on tidal wave attenuation (Möller and Spencer, 2002) and is ignored in this study.

The actual tidal prism of an inlet ($P$) can be estimated as the ideal tidal prism, where the tidal exchange is instantaneous and uniform throughout the basin, less the loss in tidal prism due to tidal wave attenuation. At each of our study sites, the inlet connects a long, narrow, shore-parallel basin to the ocean. We parameterize basin geometries, by defining two sections of basin separated at the location of the inlet channel, and by calculating the lengths ($L_1$ and $L_2$) and mean widths ($w_1$ and $w_2$) of each section (Figure 4-2). For more complex basin geometries, the basin may be subdivided into more sections. We assume that the tidal wave height attenuates at a spatially-uniform rate ($R$) within the basin. Under these assumptions, the tidal prism can be calculated as

$$P = H(L_1w_1 + L_2w_2) - \frac{1}{2}R(L_1^2w_1 + L_2^2w_2), \quad (4-1)$$

where $H$ is the tidal range. It is expected that the vegetation loss would occur from marsh edge erosion (Allen, 1997; D’Alpaos et al., 1993; Fitzgerald et al., 2007; Kirwan et al., 2008) or seaward edge retreat of mangroves (Ellison, 1991). A reduction in wetland area would, therefore, increase channel width and reduce the tortuosity of the system; wetland vegetation loss should reduce tidal wave attenuation in the back-barrier basin (Figure 4-1, Process D). Decreased tidal wave attenuation increases the actual tidal prism, $P$, or the volume of water moving through the basin during a tidal cycle (Figure 4-1, Process E). Initial mean attenuation rates can be estimated using equation 4–1 if basin geometry and tidal prism are known. We assume that tidal
wave attenuation rate varies linearly between initial mean attenuation rate and the rate calculated within an unvegetated end-member version of the basin, as wetland vegetation loss varies from 0 to 100%.

A reduction in vegetated wetland area would likely increase the open-water back-barrier basin area, resulting in an increased tidal prism (Figure 4-1, Process F). In practice, this change in vegetation area can be calculated using information from the U. S. Fish and Wildlife Services National Wetlands Inventory (NWI), for example. Wetland areas can be divided into two categories: high vegetation zone and low vegetation zone, classified through the NWI as areas that are irregularly flooded and regularly flooded, respectively. In this study, we estimate that low vegetation zones, where the tidal waters alternately flood and expose land surfaces at least once daily, effectively contribute their area to the tidal prism during half of the tidal cycle. High vegetation zones flood less frequently and, therefore, do not significantly contribute to the tidal prism. In this study, we estimate that vegetation conversion to a subtidal environment occurs at a rate proportional to the initial coverage by each of the two zones. This is not always the case, however, as evidenced by sites in southern New England where replacement of high marsh vegetation with low marsh vegetation has occurred because of the ability possessed by some cordgrass plant species (in low marsh areas) to withstand high rates of sea level rise (Donnelly and Bertness, 2001). It is assumed that the loss occurs uniformly throughout the basin, but we note that vegetation loss occurring close to the inlet should increase the tidal prism more significantly than loss occurring at distal sites within the basin, where the influence of tidal wave attenuation is low.

The cross-sectional area of the inlet channel \( A_c \) increases with tidal prism in order to convey a larger volume of water during a tidal cycle (D’Alpaos et al., 2010; Jarrett, 1976; O’Brien, 1931; Powell et al., 2006) (Figure 4-1, Process G). In this study, we employ the relationship developed by Powell et al. (2006), which was derived from
In addition to the influence on inlet morphology, tidal prism has been shown to correlate with gross and net sediment transport through the inlet and with deposition volume on ebb shoals (Figure 4-1, Process H) (FitzGerald, 1988; Marino and Mehta, 1987; Walton and Adams, 1976). In this study, we estimate ebb shoal volume ($V_E$) from an empirically derived, power-law relationship by Marino and Mehta (1987), that depends solely upon tidal prism size:

$$V_E = 5.59 \times 10^{-4} P^{1.39}.$$  

### 4.2.1 Study Sites

We chose two sites from the Florida Atlantic coast to which we apply the conceptual model. Saint Augustine Inlet (SAI) and Ponce de Leon Inlet (PLI) are separated by approximately 100 km and are both subject to semi-diurnal, micro-tidal (range <2 m) conditions. Some of the hydraulic, morphologic, and ecogeomorphic properties of the inlet systems are shown in Table 4-1. In this study, we use the mean of the measured data, derived from Walton and Adams (1976) and Powell et al. (2006) for tidal prisms, inlet cross-sectional areas, and ebb shoal volumes. The back-barrier basin of SAI is an order of magnitude smaller in spatial extent than that of PLI, and the configurations of these basins are illustrated in Figure 4-3. Despite the significantly larger basin area, PLI has a smaller measured tidal prism, which is likely due to the presence of an extensive network of channels separated by islands of wetland vegetation between the inlet and Mosquito Lagoon. SAI has a distinct main channel bordered by marsh vegetation, whereas PLI has a mixture of marsh vegetation and mangrove forest occupying the sides of the basin as well as the basin islands, which characterize the anastomosed network of small channels. The vegetated islands attenuate the tidal wave as it moves through the basin, reducing the tidal prism of the system.
The measured tidal prism at PLI is approximately 11% of the ideal tidal prism, which is calculated by multiplying tidal range by open-water basin area. Using equation 4–1, a mean attenuation rate of 4.7 cm/km throughout the basin is required in order to cause this degree of tidal wave attenuation. For comparison, studies of water levels through an unimpeded section of the Indian River close to the inlet at PLI determined an attenuation rate of approximately 1.1 cm/km (Militello and Zarillo, 2000). SAI tidal prism measurements yield values that are closer to the ideal (88%), requiring a mean rate of only 1.6 cm/km throughout the basin to account for the difference from ideal.

PLI has a smaller cross-sectional area and ebb shoal volume than SAI, as would be expected to accompany the smaller observed tidal prism. At both study sites, the predictive relationship developed by Powell et al. (2006) (equation 4–2) and Marino and Mehta (1987) (equation 4–3) underpredicts the cross-sectional area and ebb shoal volumes measurements presented in Powell et al. (2006) and Walton and Adams (1976). This is shown in Figure 4-5 as the green (SAI) and blue (PLI) stars. At PLI, the cross-sectional area and ebb shoal volumes are 23% and 158% larger than predicted using the tidal prism. At SAI, the cross-sectional area and ebb shoal volumes are 82% and 320% larger than predicted using the tidal prism. The underprediction of the cross-sectional area at these study sites may be due to the dynamic nature of these inlets; until SAI was jettied in the 1940s and PLI was jettied in the 1960s, they showed high variability in inlet width. Aerial photographs available from the U.S. Army Corp of Engineers Coastal Inlet Research Program (CIRP) show that the width of SAI has varied between 450 m and 250 m over the 24 year interval from 1975 to 1999 fluctuating by as much as 50 m within a 6 month period. While fewer aerial photographs are available for PLI, they show a similar range of variability, with a 65 m fluctuation in width between photographs taken over a two-year interval. The difference between actual and predicted values for ebb shoal volume at the study sites may be due to the difficulty in measurement (reported values vary greatly between studies, as shown in Table 4-1), or
the influence of other factors such as, channel depth to width ratio, inlet cross-sectional
area, or longshore component of wave energy flux, all of which have been shown
to exert influence on ebb shoal volume (Marino and Mehta, 1987). We use these
relationships to investigate trends that are seen in nature between the morphology of an
inlet and its hydrodynamic properties, in order to show order of magnitude changes from
vegetation loss - not to make predictions for a given site.

4.3 Results

We use equation 4–1 to calculate changes to tidal prism, which should arise due
to a conversion of wetland vegetation to open water in the back-barrier basin, then we
use equations 4–2 and 4–3 to calculate the corresponding changes to cross-sectional
area of inlet channel, and ebb shoal volume. This conversion is considered to represent
a situation in which basin vegetation cannot maintain a sufficiently high accretion rate
to keep up with sea level rise. Over the range of calculations, it is evident that wetland
loss leads to an overall increase in tidal prism for both study sites (Figure 4-4). Table 2
presents the change in vegetated area, tidal prism, cross-sectional area, and ebb shoal
volume for each inlet based on the 5-20% wetland loss predicted by Nicholls (2004)
under IPCC SRES A1FI scenario by the year 2080. Despite PLI having a similar amount
of wetland area to SAI, it is more sensitive to changes in wetland area percentage. We
calculate that these inlet systems should experience an increase in tidal prism of 3.02
m$^3$ and 1.47 m$^3$ for every 1 m$^2$ of vegetation loss at PLI and SAI, respectively.

Figure 4-4 illustrates the relative influence of the increase in basin area (first term in
equation 4–1) and the reduction in tidal wave attenuation (second term in equation 4–1)
on the tidal prism of each inlet. Dashed lines represent PLI and solid lines with circular
markers represent SAI. The black lines are the tidal prism (P), which is the ideal tidal
prism (green line), dependent solely on basin area, less the attenuated tidal prism (blue
line), dependent on basin area and attenuation rate. PLI showed an increase of 1.24
m$^3$ per 1 m$^2$ loss due to an increase in basin area, and 1.77 m$^3$ per 1 m$^2$ loss due to
a decrease in the amount of tidal prism reduction from tidal wave attenuation. At SAI, the basin area increase causes a tidal prism increase of 1.51 m$^3$ per 1 m$^2$ loss. Given that the mean tidal wave attenuation rate is similar to the rate in un-vegetated areas, the increase in open-water area (over which the tidal wave attenuates) causes a decrease in the tidal prism by 0.4 m$^3$ per 1 m$^2$. Figure 4-4 highlights the significance of the decrease in tidal wave attenuation with wetland loss for some inlet systems, such as PLI, which have wetland configurations that provide a high degree of tidal wave attenuation. This is not the case at all inlets, as seen in the SAI example, where changes to the tidal wave attenuation rate have relatively little influence on how the tidal prism responds to changes in marsh area.

Initially, an increased tidal prism should increase the mean gross discharge through the inlet channel during each tidal cycle, increasing the flow velocity, shear stress, and sediment transport capacity through the throat of the inlet. This has the effect of scouring the channel until the cross-sectional area of the throat accommodates the decreased flow such that there is no net transport through the channel; a new equilibrium cross-sectional area is achieved at that point (O'Brien, 1931). Using the relationship presented by Powell et al. (2006) (equation 4–2), we find that the increase in cross-sectional area for every 1 m$^2$ of vegetation loss is 1.78 cm$^2$ and 0.92 cm$^2$, at PLI and SAI, respectively (Figure 4-5). The predictive relationship of Marino and Mehta (1987) reveals that 1 m$^2$ of vegetation loss should result in an increased ebb shoal volume of 2.98 m$^3$ for PLI, and 1.29 m$^3$ for SAI (Figure 4-5).

4.4 Discussion

4.4.1 Two Main Mechanisms for Tidal Prism Change

This study examines the influence of changes in configuration of vegetated wetlands, which may arise from sea level change, on the morphologic evolution of a tidal inlet system. The results are not intended to be used as a predictive tool for either of the example tidal inlet systems presented, but rather to identify and explore
the two main factors governing tidal prism response to changes in wetland vegetation area: (1) map-view area of the back-barrier basin, and (2) the spatial rate of tidal wave attenuation within the basin. Herein, we investigated a simple scenario of uniform wetland vegetation retreat within the back-barrier basin in order to highlight the relative importance of the two factors at PLI and SAI, stemming from their differences in initial wetland configuration. While the two study sites have similar total wetland area in their back-barrier basins, the configurations of these wetland systems differ. SAI has fringing wetlands along the sides of the basin, creating one main basin channel, whereas PLI has a combination of fringing wetlands and a network of wetland islands which creates a complex web of highly sinuous basin channels. The intricate channel network at PLI reduces the tidal wave as it travels through the basin by increasing the flow resistance by wetland vegetation and increasing the path distance through which the tidal wave travels. A reduction in wetland area at PLI would increase the map view open water area and decrease the spatial rate of tidal wave attenuation. Such a change would cause the tidal prism at PLI to respond (increase) more sensitively than would be expected to occur at SAI, where wetland loss would have a negligible impact on tidal wave attenuation rate. These results indicate that an inlet whose back-barrier basin has a higher initial spatial rate of tidal wave attenuation holds greater potential for tidal prism change due to vegetation loss.

4.4.2 Marsh Loss Spatial Distribution

In our conceptual model, we assume that, as wetland loss occurs, the wave attenuation rate changes linearly between the initial estimated average attenuation in the basin and the measured attenuation that occurs in an unvegetated basin. If wetland loss occurs in areas far from the inlet channel, it would not have the same impact on wave attenuation as loss of a wetland island close to the inlet channel. This is due to the fact that at distal basin regions, the tidal wave is small in comparison to areas proximal to the inlet channel. The assumption of linear rate change with wetland loss will overpredict
tidal prism change where wetland loss occurs at a fringing marsh in the back of the basin, and will underpredict in the case of near-channel wetland loss. Given that wetland loss patterns cannot be reliably predicted, we opt for a conservative method (linear) to estimate wave attenuation change as a function of wetland loss.

4.4.3 Influence on Adjacent Shorelines

Ebb shoal volume changes can impact adjacent environments, such as beach and dune complexes on either side of the inlet. Shoreline change is particularly sensitive to inlet system morphology as ebb shoals interrupt longshore sediment transport (LST) and gradients in LST lead to beach erosion and accretion. For inlet systems comparable to those presented in the study, a vegetation loss of 1%, should increase ebb shoal volume by $6 \times 10^5$ m$^3$, by incorporating a fraction of the steady flow of LST passing the inlet. In the sand-sharing concept of Dean (1988) and Kraus (2000), the ebb shoal complex consists of a main ebb tidal shoal, a series of bypassing bars, and attachment bars. If the ebb shoal complex is out of equilibrium due to a change in hydrodynamic properties, such as a shift in the tidal prism, it will adjust its volume by sequestering sand from the passing flow of LST until a new equilibrium volume is achieved. In this scenario, the ebb shoal acts as a local sediment sink and interrupts the flow of LST past the inlet channel. Along the coastal reach in the vicinity of the two example sites presented herein, the LST rates are approximately $3.8 \times 10^5$ m$^3$/yr (Dean, 1988), implying that such a change in ebb shoal would be equal to approximately 19 months of LST in the area. There are links between tidal prism and the extent (both seaward and downdrift) of the ebb shoal complex (Carr-Betts et al., 2012) and the minimum depth over the ebb shoal crest (Buonaiuto and Kraus, 2003). As tidal prism increases, the ebb shoal spreads further offshore and in the downdrift direction, and the shoal occupies a deeper water depth.
4.4.4 Other Influence on Marsh Loss

Although we have chosen to identify sea level rise as the most likely driver of vegetation loss in back-barrier basins, anthropogenic disturbances can play a major role in exacerbating sea level rise effects (Mudd, 2011; Nicholls, 2004). Natural and anthropogenic processes, such as invasive species establishment, overfishing, nitrogen eutrophication, rising water temperatures, increased atmospheric carbon dioxide, altered hydraulic and sedimentation regimes, drainage, reclamation, and shoreline development, might also contribute to wetland loss and therefore influence morphologic changes in these systems (Silliman et al., 2009). Further research into the relationships between wetland vegetation and tidal inlet morphology will provide valuable progress toward our understanding of the long-term morphological evolution of sandy coastal systems globally.

4.4.5 Application to Barrier Systems Worldwide

While the focus of this study was on contrasting wetland configurations in the basins of two Florida inlets, the conceptual model can be applied to other barrier island systems. For the two study sites used here, the basin geometries could be parameterized into two main sections, which may need to be modified for more complex basin systems. We were able to neglect the change in basin area due to sea level rise flooding adjacent shorelines because of the highly urbanized and modified coastal area surrounded these basins. We note that basin area change at the margins should not be neglected in cases with basin edges are gently sloped and unmodified by seawalls. The relationships between tidal prism and cross-sectional area was chosen from a study of Florida inlets for our sites, but could be modified to more general relationship, such as those presented by Jarrett (1976), for other sites. Our conceptual model, which includes the influence of the wetland vegetation changes on tidal inlet response to sea level rise, can provide insight into the morphologic response of other inlet systems, including
changes to the channel cross-sectional area and the volume and position of the ebb shoal complex.

4.5 Conclusions

In this paper, we show progress toward understanding how the ecogeomorphic properties of a tidal inlet system control the evolution of the inlet and adjacent shoreline morphology due to loss in wetland area. Using our simple conceptual model, we conclude:

- At both study sites, wetland vegetation loss that leads to bank destabilization and erosion creates an increase in tidal prism, increase in cross-sectional area, and increase in ebb shoal volume.
- Initial wetland configuration controls the response of the tidal inlet morphology to wetland loss; inlets with wetland configurations that impose a higher spatial rate of wave attenuation have greater potential for tidal prism change due to vegetation loss that leads to bank erosion.
- For the inlets in this study, small amounts of marsh platform loss (1%) can increase the equilibrium volume of the associated ebb shoal complex by an amount equivalent to one to two years of accumulated net longshore transport in the area.

Figure 4-6 shows the changes to the tidal inlet morphology that would occur due to a loss in wetland vegetation, including: (1) an increase in cross sectional area of the inlet, (2) an increase in the volume of the ebb shoal complex, and (3) an increase in the downdrift and seaward extent of the main ebb shoal. Plot A and B of Figure 4-6 illustrates a uniform wetland vegetation loss that could occur in basins similar to SAI and PLI, respectively.
Figure 4-1. Conceptual model of tidal inlet geomorphic response to sea level rise. Letters represent specific processes linking key variables (shown in blue boxes) as described in the text.

Figure 4-2. Diagram illustrating how the basins are parameterized into two sections on either side of the inlet channel.
Figure 4-3. Maps of the back-barrier basins examined in this study: (A) the basin associated with Saint Augustine Inlet (SAI), and the (B) north and (C) south sections of basin associated with Ponce de Leon Inlet (PLI). Both sites are located along the North Florida Atlantic Ocean coast, within approximately 100 km of one another. This location is shown as the red box in the inset map of Florida. The scale bar and north arrow apply to all three basin maps.
Figure 4-4. The change in tidal prism due to wetland vegetation loss for PLI and SAI. The black lines are the total change in tidal prism, which is calculated as the ideal tidal prism (green lines) less the loss in tidal prism due to tidal wave attenuation in the basin (blue lines). The percent loss for each inlet in shown below the figure for each inlet and the grey highlighted areas correspond to the extent of worldwide wetland loss, as predicted by Nicholls (2004) under IPCC SRES A1FI scenario by 2080. This figure highlights the relative importance of the reduction in tidal wave attenuation and increase in basin area on the hydrodynamics response of the inlet systems to changes in wetland area.
Figure 4-5. The impact of marsh vegetation loss on tidal prism, cross-sectional area of inlet channel, and ebb shoal volume. Saint Augustine Inlet (SAI) is shown in green, and Ponce de Leon Inlet (PLI) is shown in blue. Stars indicate the average measurements of cross-sectional area and ebb shoal volume published in Powell et al. (2006) and Walton and Adams (1976). Gray shading shows extent of worldwide wetland loss, as predicted by Nicholls (2004) under IPCC SRES A1FI scenario by 2080.
Figure 4-6. Schematic Illustration of wetland vegetation loss in two basins with ecogeomorphic configurations similar to SAI and PLI. Vegetation loss within each basin increases channel cross-sectional area, ebb shoal volume, and the downdrift and seaward extent of the main ebb shoal. Decrease in the tidal wave attenuation in basin (B) leads to a higher change in tidal prism of the system and an intensification of these effects.
Table 4-1. Study sites hydraulic, ecologic, and geomorphic properties

<table>
<thead>
<tr>
<th></th>
<th>Saint Augustine Inlet, FL</th>
<th>Ponce de Leon Inlet, FL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Tidal Range (m)</td>
<td>1.6</td>
<td>1.3</td>
</tr>
<tr>
<td>Meas. Tidal Prism (m³)</td>
<td>2.5 x 10⁷</td>
<td>1.7 x 10⁷</td>
</tr>
<tr>
<td></td>
<td>3.71 x 10⁷</td>
<td>1.63 x 10⁷</td>
</tr>
<tr>
<td>Ideal Tidal Prism (m³)</td>
<td>3.52 x 10⁷</td>
<td>1.56 x 10⁸</td>
</tr>
<tr>
<td>Cross-Sectional Area (m²)</td>
<td>4600</td>
<td>1500</td>
</tr>
<tr>
<td></td>
<td>2461</td>
<td>1068</td>
</tr>
<tr>
<td>Ebb Shoal Volume (m³)</td>
<td>4.3 x 10⁷</td>
<td>1.7 x 10⁷</td>
</tr>
<tr>
<td></td>
<td>8.1 x 10⁷</td>
<td>1.45 x 10⁷</td>
</tr>
<tr>
<td>Basin Area (m²)</td>
<td>2.2 x 10⁷</td>
<td>1.2 x 10⁸</td>
</tr>
<tr>
<td>High Vegetation Area (m²)</td>
<td>4.69 x 10⁷</td>
<td>5.02 x 10⁷</td>
</tr>
<tr>
<td>Low Vegetation Area (m²)</td>
<td>5.82 x 10⁶</td>
<td>4.38 x 10⁶</td>
</tr>
<tr>
<td>N. Basin Length, L₁ (km)</td>
<td>26</td>
<td>37</td>
</tr>
<tr>
<td>S. Basin Length, L₂ (km)</td>
<td>18</td>
<td>52</td>
</tr>
<tr>
<td>N. Basin Avg. Width, w₁ (m)</td>
<td>495</td>
<td>550</td>
</tr>
<tr>
<td>N. Basin Avg. Width, w₂ (m)</td>
<td>500</td>
<td>1860</td>
</tr>
<tr>
<td>Avg. Att. Rate (cm/km)</td>
<td>1.6</td>
<td>4.7</td>
</tr>
</tbody>
</table>

First values of measured tidal prism, cross-sectional area, and ebb shoal volumes are from published data in *Powell et al. (2006)* and the second listed values are published in *Walton and Adams (1976)*. Wetland area values calculated based on NWI estuarine environment data. The basin area is classified as subtidal. High and low vegetation area are classified as intertidal with emergent vegetation or scrub-shrub, and irregularly and regularly flooded water regimes, respectively.

Table 4-2. Inlet morphologic and hydraulic response to wetland loss

<table>
<thead>
<tr>
<th></th>
<th>Saint Augustine Inlet, FL</th>
<th>Ponce de Leon Inlet, FL</th>
</tr>
</thead>
<tbody>
<tr>
<td>Vegetation Loss (km²)</td>
<td>2.64 - 10.54</td>
<td>2.73 - 10.92</td>
</tr>
<tr>
<td>ΔP (x10⁷ m³)</td>
<td>0.38 - 1.53</td>
<td>0.72 - 2.93</td>
</tr>
<tr>
<td>ΔA_C (m²)</td>
<td>238 - 955</td>
<td>448 - 1834</td>
</tr>
<tr>
<td>ΔV_E (x10⁷ m³)</td>
<td>0.25 - 1.07</td>
<td>0.39 - 1.89</td>
</tr>
</tbody>
</table>
APPENDIX A
DELT3D MODEL SETUP

This appendix is intended to act as a guide for the setup and execution of the Delft3D-FLOW module within the Delft3D modeling suite for simulations similar to those presented in Chapter 2 and Chapter 3. These model runs used Delft3D utilities and the Delft3D-FLOW module without wind or wave inputs and did not calculate sediment, temperature, or salinity transport. This appendix will give a brief description of the physical processes that can be modeled using Delft3D-FLOW and to act as a guide for the general steps of model setup. More detailed information is provided in the Delft3D-FLOW user manual (Deltares, 2009).

A.1 Model Physical Processes

The Delft3D-FLOW model is capable of simulating two-dimensional (depth averaged) or three-dimensional flow using the unsteady shallow water equations. The equations are derived from the three-dimensional Navier Stokes equations for incompressible free surface flow under the shallow water and Boussinesq assumptions. Detailed hydrodynamic equations are presented in Deltares (2009). The model is forced at the open boundaries by tides, at the free surface by wind stress, and by pressure gradients due to gradient in the free surface elevation and density. Delft3D-FLOW is capable of simulation the following physical phenomena:

- Free surface gradients (barotropic effects).
- The effect of the Earth’s rotation (Coriolis force).
- Water with variable density (equation of state).
- Horizontal density gradients in the pressure (baroclinic effects).
- Turbulence induced mass and momentum fluxes (turbulence closure models).
- Transport of salt, heat and other conservative constituents.
- Tidal forcing at the open boundaries.
- Space and time varying wind shear-stress at the water surface.
• Space varying shear-stress at the bottom.
• Space and time varying atmospheric pressure on the water surface.
• Time varying sources and sinks (e.g. river discharges).
• Drying and flooding of tidal flats.
• Heat exchange through the free surface.
• Evaporation and precipitation.
• Tide generating forces.
• Effect of secondary flow on depth-averaged momentum equations.
• Lateral shear-stress at wall.
• Vertical exchange of momentum due to internal waves.
• Influence of waves on the bed shear-stress (2D and 3D).
• Wave induced stresses (radiation stress) and mass fluxes.
• Flow through hydraulic structures.
• Wind driven flows including tropical cyclone winds.

A.2 Delft3D-FLOW Model Setup

Before beginning model setup, the user should select the working directory for the model simulation. This can be done by selecting the ‘Select working directory’ button on the bottom of the main Delft3D menu. It is important to note that all user input files and attribute files for a model simulation should be stored in the same directory.

The general steps that will be presented in this appendix for the setup and execution of the Delft3D-FLOW module, are:

1. Creation of a land boundary file
2. Generation of model grid and enclosure files
3. Generation of a bathymetry file using sample files
4. Creation of an MDF-file which specifies:
• Model domain
• Time frame
• Processes
• Initial conditions
• Boundary conditions
• Physical parameters
• Numerical parameters
• Operations
• Monitoring
• Output

5. Execution of model simulations

6. Viewing model output

A.2.1 Land Boundary File Generation

The land boundary file consists of closed polygons which represent the land-water interface. These files can be used for grid generation and in the creation of output images. The files must be created by the user in a utility outside of the Delft3D model, such as MATLAB. The files are ascii text files with the extension ‘.ldb.’ In these files, each polygon has a header with the polygon label on the first line, followed on the next line by the number of polygon points and the number of coordinates specified (usually 2 for X,Y coordinates), followed on the lines below as a list of X and Y points. This is repeated for each polygon that makes up the entire land boundary of interest. Each polygon must have the same first and last set of X,Y coordinates. Example of the file format using a spherical coordinate system for a land boundary is shown below:

```
L001
6 2
-81.492310 30.647717
-81.492653 30.647631
-81.493416 30.647675
```

121
A.2.2 Grid Generation

The model grid can be created using the built-in utility, Delft3D-RGFGRID, developed by Deltares. This program can be launched through the main Delft3D menu, by clicking on the ‘Grid’ button and then the ‘RGFGRID’ button. This application is capable of generating curvilinear grids using either a spherical (in decimal degrees) or cartesian (in meters) coordinate system. A simple rectangular grid can be created by inputting the X and Y origins, grid cell spacing, and the number of grid cells in the X and Y direction. More complex grids can be created by importing a land boundary file to use
as a reference for grid boundary locations. Individual grid cells can be manually moved or deleted, grid lines can be snapped to land boundary locations, and grid sections can be smoothed, allowing for complete user control over each grid cell location. Grid lines should be smoothed along land boundaries, reducing the “stair case” boundaries and reducing artificial diffusion. The output of this program is a grid file with the extension `.grd' and a grid enclosure file with the extension `.enc.' The enclosure file is generated automatically when the grid file is exported.

A.2.3 Bathymetry Generation

Once a model grid has been created using the Delft3D-RGFGRID utility, a bathymetry may be interpolated onto the grid using the Delft3D-QUICKIN utility. This utility can be launched through the main Delft3D menu, by clicking on the 'Grid' button and then the ‘QUICKIN’ button. Once the program is opened, the grid must be imported into the program by selecting File → Import → Grid. A bathymetry can be created using a sample file, an ascii text file, with a list of (X,Y,Z) coordinates. The set of coordinates for each sample point are on a line, and are separated by a space or tab. In spherical coordinates, the X,Y coordinates are in decimal degrees and the Z coordinates are in meters. In cartesian coordinates, all coordinates are in meters. Depths in Delft3D are positive values. These sample points do not need to be evenly spaced, and multiple files can be imported. The sample files can be imported in the program by selecting File → Attributes → Open Samples.

Once the samples are imported into the program, a bathymetry can be created to correspond with the imported grid by either triangular interpolation, grid cell averaging, or a combination of the two. Grid cell averaging requires higher resolution of sample points than triangular interpolation, and may not be an options for fine grid resolutions relative to sample resolution. These two methods are available under the Operations tab. Polygons can be created and used to isolated individual grid areas to apply these interpolations, which is useful for large grids or grids with many sample input files.
with different resolutions. Internal diffusion can be used to populate grid cells that are on the edges of the grid which may not have enough samples surrounding the cell to estimate a value for that location. This diffusion will assign the depth value of the closest cell to all the cells with no depth.

The depth of individual grid cells can be manually altered in this program and large areas can be smoothed. The parameters for smoothing and interpolation of samples to the grid can be changed in the Settings → General Parameters. Once a desired depth value has been found for each grid cell, the bathymetry file should be exported in File → Export → Depth. It will be saved as a ‘.dep’ and must be used in conjunction with the grid file from which it was created because it does not contain horizontal coordinate data.

A.2.4 Creating an MDF-File

The Master Definition Flow file, or MDF-File, is the main input file used to run the Delft3D-FLOW module. It can be created by launching a user interface through the Delft3D main menu, by selecting ‘Flow’ and then selecting ‘Flow input.’ A user interface launches with tabs on the left which open a difference set of input options on the right. The tabs should be selected from top to bottom and the appropriate information for each tab should be completed before continuing onto the next section. Below are basic instructions for each tab. For detailed instruction see the Delft3D-FLOW user manual (Deltares, 2009)

Description. The description input is used only for the reference of the user and does not influence the model run.

Domain. In this section the user should select ‘Grid parameters’ and then open the grid and enclosure files created in the previous steps. The co-ordinate system and number of grid cells should be read from the grid file. The user then needs to specify the number of layers in the z-direction. A value of one indicates a two-dimensional, depth averaged model run. For a three-dimensional model simulation, a value greater than...
one should be selected. An option for layer thickness will populate the user interface and
the user can specify the layer thickness (as a percentage of the water column) for each
layer. The default is for evenly spaced layers.

The user should select the ‘Bathymetry’ button at the top and open the bathymetry
file created in the previous steps. Dry points and thin dams may be added in this section
to represent jetties and areas with no flow.

**Time frame.** Here the model simulation date range and time step should be
specified. Generally, the reference date is the same as the simulation start date. The
default time zone is GMT, but can be modified in this section.

**Processes.** This tab allows the user to specify which processes to include in the
model simulation. For this example, none of the constituents or physical processes
are selected. If these are selected then more input information in the follow sections
become available. If no options are selected, as in this example, only the hydrodynamic
properties are calculated without wind or wave forces.

**Initial conditions.** This section allows users to specify a uniform water level
throughout the domain as the initial condition, to include an initial conditions file, to
include a restart file, or to include a map file. Using initial conditions or restart files
can reduce the “spin-up” time of the model, but are not necessary. The default for this
section is a uniform water level of zero meters throughout the domain.

**Boundary Conditions.** A set of initial and boundary conditions for water levels
and horizontal velocities must be specified. Initial vertical velocities are not necessary
to specify, as they are computed through the continuity equation. Boundaries of the
model are classified as opened or closed boundaries. Closed boundaries are locations
without flow, such as “land-water” lines (river banks, coastlines). Open boundaries
are always “water-water” boundaries that intersect the flow field. These boundaries
should always be situated as far from the area of interest as possible. Reflection at open
boundaries should be minimal since the open boundaries should not hamper long wave propagation.

To specify the boundary conditions, first individual boundary segments must be defined by clicking the ‘add’ button. Each segment must have an unique name and the user must specify the indices (m,n) of each end point. The visualization area can be used to help with this process. It can be opened through the top menu by selecting View → Visualization area. In this window, boundary segments can be added manually. The boundaries can be broken into numerous segments, and should depend on the domain size. All areas of open water at the border of the grid should be defined as a boundary. Closed boundaries, which border land, do not need to be specified as a boundary here.

Once each boundary is defined in space, the boundary conditions must be specified. The type of boundaries that can be specified, include: Water level, Current, Neumann, Total discharge, Discharge per cell, or Riemann. For this example we use water levels, but details of each of these types of boundaries can be found in Deltares (2009). The forcing type for water levels can be astronomic, harmonic, QH-relation, or time series. For this example we chose astronomic, which is convenient if tidal constituents can be derived from local tidal gauges.

Once the appropriate selections have been made, the user needs to edit the details of the boundary by selecting the ‘Edit flow conditions’ button. A new box will open where the details of the boundary can be input. Different sets of conditions can be specified for each end of each of the boundaries. These sets of conditions can be specified by selecting the ‘Add’ button and inputting each set of constituents. The component set is linked to each end of each boundary segment by the dropdown menus on the right side.

Once the boundary details have been input, the user closes the box by clicking the ‘close’ button in the bottom right corner. Once this process has been completed for each boundary segment, the files need to be saved by selecting the ‘Open/Save’ button. The boundary definitions file (with the extension ‘.bnd’) contains the coordinates of each
boundary segment and the astronomic flow conditions file (with the extension ‘.bca’) contains the constituents for each end of each boundary segment.

**Physical parameters.** In this section the user specifies the constants for gravity and water density, the bottom and wall roughness, the viscosity/diffusivity, and three-dimensional turbulence closure model to be used in the simulation. The roughness and viscosity/diffusivity can be specified using a uniform value throughout the domain or as a file that specifies a value at each grid cell location. For more information on these values and how to create these files, refer to *Deltares (2009)*.

**Numerical parameters.** In this section the user can specify parameters related to drying and flooding and some other advanced options for numerical approximations. The smoothing time is the time interval at the start of the model simulation that is used to create a smooth transition between the initial and boundary conditions. Longer smoothing time periods create smoother results in the beginning of the simulation but increase the computation time period.

**Operations.** Discharge values can be specified here to represent flow into the system by a source such as a river. The discharge at each grid cell location must be specified, so flow across a river must be divided into segments.

**Monitoring.** Different observation types for output can be specified in this section. Observations are individual points, drogues monitor particle paths, and cross-sections are shown for all depths along an m- or n-grid index segment. More than one specific observation point or cross-section must be specified or the model will not run (i.e. there can be one observation as long as there is also one cross-section specified).

**Additional parameters.** This section provides access to additional functions that are not supported by the FLOW-GUI, allowing for more flexibility without altering the FLOW-GUI. Unless the user has access to additional options, it should remain blank.

**Output.** The time interview for model output and the details of the output can be specified in this section. The map results are the snap shots of the computed quantities
for the entire model domain, the history file stores the results for the specified monitoring observation points, drogues, and cross-sections, and the communication file stores data required for other Delft3D modules. The start time and end time must be in the range of the model time frame, but do not have to cover the entire range. The interval sets the time step of output for the map file and the history interval sets the time step for output of the history file. Both of these values must be a multiple of the model time step. Unless other modules are being used, the interval for the communication file can be set to zero. Online visualization should generally be unchecked unless it is needed for troubleshooting. By default all model output details (under the ‘Details’ button) are selected. Some of these may be unselected in order to reduce output file size.

A.2.5 Executing model run

After all the details of the run have been specified through the FLOW-GUI, the MDF-file should be saved by selecting File → Save MDF. This creates an ascii text file with the extension ‘.mdf’ that contains input information and the names of files to reference during the model simulation. All files for the simulation need to be saved in the same folder. Once the file has been saved, the FLOW-GUI can be closed. When the model is ready to be run, select the ‘Start’ button from the main Delft3D menu and select the MDF-file that has been created and select OK. A box should appear to show that the model is running.

A.2.6 Viewing model output

The QUICKPLOT utility can be used to open, view, and export model output. It reads the map output file (trim-name.dat) and the history file (trih-name.dat). Once these files have been loaded into the program, specific date ranges and model data can be exported in a variety of formats for manipulation outside of the program, including MATLAB mat-files.
APPENDIX B
MODEL INPUT PARAMETERS

This appendix includes the model input parameters for the 52-day model simulation described in Chapter 2. This simulation was used for model calibration and comparison with available ADCP data. The lists are broken into groups which correspond to the tabs in the FLOW-GUI described in Appendix A.

Domain.

- Coordinate System: Spherical
- Grid points in the M-direction: 420
- Grid points in the N-direction: 302
- Number of layers: 8 (12.5% of depth each)

Time Frame.

- Reference date: 08 11 2011
- Simulation start time: 08 11 2011 00 00 00
- Simulation stop time: 30 12 2011 00 00 00
- Time step: 0.5 min
- Local time zone: 0 +GMT

Processes.

- No constituent or physical processes selected (hydrodynamics only)

Initial Conditions.

- Uniform values
- Water level: 0 m

Boundaries.

- 3 boundary segments: North, East, and South
- Type of open boundary: Water level
• Forcing type: Astronomic
• Reflection parameter alpha: 0 s^2
• Tidal constituents Included (Name, Amplitude (m), Phase (deg)):
  – M2, 8.0269704e-001, 1.5706319e+001
  – S2, 1.5658684e-001, 1.3463423e+001
  – N2, 1.6423517e-001, 3.5913640e+002
  – K2, 2.5928995e-002, 8.6506397e+000
  – K1, 1.0109143e-001, 1.9361364e+002
  – O1, 7.7357689e-002, 2.0546273e+002
  – P1, 3.7856681e-002, 1.9961633e+002
  – Q1, 1.7017450e-002, 2.0305634e+002
  – MF, 7.5714709e-003, 3.4363395e+002
  – MM, 2.9204701e-003, 3.2804466e+002
  – M4, 7.1590335e-003, 3.4327001e+002
  – MS4, 4.6285343e-003, 2.4064482e+002
  – MN4, 1.8140395e-003, 1.3922591e+002

Physical Parameters.
• Gravity: 9.81 m/s^2
• Water density: 1025 kg/m^3
• Roughness formula: Chezy, Uniform U = 65, V = 65
• Wall roughness slip condition: Free
• Background horizontal viscosity/diffusivity: Uniform 1 m^2 horizontal eddy viscosity
• Background vertical viscosity/diffusivity: Uniform 0 m^2 vertical eddy viscosity
• Turbulence model for 3D: k-Epsilon

Numerical Parameters.
• Drying and flooding check at: grid cell centres and and faces
• Depth specified at: Grid cell corners
• Depth a grid cell centres: Max
• Depth at grid cell faces: Mean
• Threshold depth: 0.1 m
• Marginal depth: -999 m
• Smoothing time: 60 min
• Advection scheme for momentum: Cyclic

**Operations.**
• No discharges included

**Monitoring.**
• Observations: one at tidal station 8720030 and one in inlet channel
• Drogues: none specified
• Cross-section: two through inlet channel, one at minimum cross-section

**Additional Parameters.**
No additional parameters included

**Output.**
• Map storage start time: 08 11 2011 00 00 00
• Map storage stop time: 30 12 2011 00 00 00
• Map storage interval: 30 min
• History interval: 6 min
• Communication interval: 0 min
• Restart interval: 0 min
REFERENCES


Bruun, P. (1966), Tidal inlets and littoral drift, Universitetsforlaget, Norway.

Bruun, P., and F. Gerritsen (1959), Natural bypassing of sand at coastal inlets, J. Waterways and Harbors Division, 85(4), 75–107.


Environmental Science and Engineering, Inc. (1980), Draft supplement to the environmental impact statement for preferred alternative location for a fleet ballistics missile submarine support base at Kings Bay, Georgia, *Tech. rep.*


FitzGerald, D. M. (1982), Sediment bypassing at mixed energy tidal inlets, in *18th Coastal Engr. Conf.*, ASCE.


Hayes, M. O. (1979), Barrier island morphology as a function of tidal and wave regime, in *Barrier Islands from the Gulf of St. Lawrence to the Gulf of Mexico*, edited by S. Leatherman, pp. 1–27, Academic Press, New York.


LeConte, L. J. (1905), Discussion on river and harbour outlets, *Transactions, American Society of Civil Engineers, Paper No. 1009*, 306–308.


Mudd, S., S. Howell, , and J. Morris (2009), Impact of dynamic feedbacks between sedimentation, sea level rise, and biomass production on near-surface marsh


van de Kreeke, J. (1992), Stability of tidal inlets; Escoffier’s analysis, *Shore and Beach, 60*(1), 9–12.


BIOGRAPHICAL SKETCH

Jessica Lovering was born in Maryland and moved to the Florida Keys in late elementary school. She lived on Big Pine Key and graduated from Key West High School in 2001. She began her undergraduate career at the University of Maryland and completed her degree in civil engineering with a specialization in geotechnical engineering at the University of Florida. She went on to complete a master’s degree in coastal and oceanographic engineering. She remained at the University of Florida in order to pursue her Ph.D. in the Geological Sciences Department with Peter Adams. She was awarded the SMART (Science Math and Research for Transformation) fellowship and will be working for the Department of Defense after graduation. She will begin her career as a physical oceanographer at the Naval Oceanographic Offices at the Stennis Space Center in Mississippi.