EFFECTS OF WAVES, TIDES, AND VEGETATION ON THE DISTRIBUTION OF BED SHEAR STRESS IN THE GREAT BAY ESTUARY, NH

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EFFECTS OF WAVES, TIDES, AND VEGETATION ON THE DISTRIBUTION OF BED SHEAR STRESS IN THE GREAT BAY ESTUARY, NH

BY

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DISSERTATION

Submitted to the University of New Hampshire
in Partial Fulfillment of
the Requirements for the Degree of

Doctor of Philosophy

in

Oceanography

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“It takes a village to raise a child.”
- ancient African proverb.

It turns out it also takes a village to raise a doctoral student.
I dedicate this work to the village that supported me along the way.
I cannot thank you enough.
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ABSTRACT

Effects of waves, tides, and vegetation on the distribution of bed shear stress in the Great Bay Estuary, NH

by
Salme E. Cook
University of New Hampshire

The goals of this research are to expand our understanding of, and improve predictions of bed shear stress in estuarine environments using both observational datasets and numerical modeling. To accurately predict sediment transport, a good understanding of the bed shear stress that drives the sediment erosion, suspension and deposition is essential. Shear stress is a function of both the hydrodynamics in the system and the characteristics of the sediment that comprise the bed itself. The hydrodynamic forcing is determined by tides, waves, meteorological effects, rivers, or some combination that can change on time scales of a few minutes to a few days. The sediment characteristics are site specific, and often vary spatially within a given estuary. The size, shape, material type, organic content, and time in a given location can determine whether the sediment will move, and what mode of transportation is probable (i.e. bed load or suspended load). The temporal and spatial variability of these factors make it difficult to collect comprehensive observational datasets, and often only represent a small portion of the overall processes of interest. Numerical models become useful tools to predict how the interactions of different hydrodynamic conditions and sediment characteristics can change the bed shear stress on a variety of scales. Consequently, these models require parameterizing sub-grid scale processes, and suppressing noise associated with numerical discretization. A useful model then becomes a balance between capturing the processes of interest within a particular grid scale and the available computational resources. The purpose of this research is to use the observational datasets from both the hydrodynamics and sediment and bed characteristics of a particular estuary, and 1) verify the hydrodynamic model, and 2) use that model to characterize and predict the spatial and temporal variability of bed shear stress and sediment transport under different hydrodynamic conditions (tides, waves, meteorological forcing, etc.) and in the presence/absence of vegetation (eelgrass). Ultimately this knowledge
will useful for more accurate estimates of sediment transport and nutrient fluxes under varying hydrodynamic conditions in the Great Bay estuary (and inform similar estuarine mudflat environments), which has been previously difficult.
CHAPTER 1
INTRODUCTION

1.1 Motivation and Background

Estuaries are dynamic coastal environments where upland freshwater flows mix with tidally and sub-tidally forced coastal marine water to create a highly productive and diverse ecosystem important to local communities and fisheries. Increases in population density and associated anthropogenic impacts have altered the productivity of estuarine environments, resulting in increased nutrient loading and amplified suspended sediment that reduces water quality. Mudflats and salt marshes comprise the intertidal portions of these environments and are the most productive and vulnerable to these stressors. Usually found adjacent to salt marshes and eelgrass beds in sheltered areas, they support commercially harvested populations of shellfish like crabs and clams, as well as provide important habitat for recreational fish and are foraging areas for coastal birds and waterfowl. They also provide an important link to deeper estuarine habitats that support juvenile populations of lobster, bay scallops, winter flounder, and other important commercially fished species. Intertidal areas are primarily comprised of cohesive fine-grained muds and clays, that when disturbed have a longer residence time in the water column than coarser sediment like sands and gravel. Fine-grained sediment also tends to adsorb pollutants and nutrients that can be reintroduced into the environment when disturbed. Therefore, a good understanding of both the sediment characteristics and the hydrodynamic conditions that dictate the transport of sediment, nutrients, and pollutants is imperative for better estuarine management policies and long-term restoration efforts.

Bed shear stress is the physical parameter that has the largest impact on sediment transport and controls how much sediment is suspended. At the water-sediment boundary the magnitude of the bottom shear stress is proportional to the gradient of the overlying fluid velocity. Therefore, the spatial variability of the velocity gradient is very important in estimating the dynamics of the bottom boundary layer, which govern sediment
transport. Tides and waves are the primary forcing mechanisms in these environments, and have different velocity gradients near the sediment bed.

There are several estimation methods that use field observations of water velocity ("law of the wall") or turbulent fluctuations (turbulent kinetic energy method), and require method dependent assumptions. Although these observations exist, they are limited to one specific area, and cannot be used to estimate the bed shear stress across an entire system. It is infeasible to collect enough measurements to determine the spatial and temporal scale of shear stress in these environments; so numerical models are used to make system-wide predictions of shear stress and sediment transport under different hydrodynamic and meteorological forcing conditions. Numerical models are well suited for environments like tidal mudflats, where periodic inundation and drying, and combined wave current flows make it difficult to conduct field studies and collect observations. However these models must be carefully setup and verified in order to represent the physics appropriately. Models also require parameterizing sub-grid scale processes, and suppressing noise associated with numerical discretization. A useful model then becomes a balance between capturing the processes of interest within a particular grid scale and the available computational resources.

Within the Gulf of Maine it is estimated that intertidal mudflats represent over 50% of the total area of estuarine habitat (Roman et al., 2000). The Great Bay Estuary, located in southeastern New Hampshire is comprised of deep tidal channels and expansive mudflats, typical of other estuaries in the Gulf of Maine. The Piscataqua Region Estuaries Partnership (PREP), a part of the Environmental Protection Agency’s National Estuary Program, is charged with protecting the estuarine resources of the Great Bay Estuary. The 2013 PREP report outlined the need for expanded monitoring and additional research to address knowledge gaps in long-term nitrogen/nutrient loading within the Great Bay estuarine system (PREP, 2013). Currently the PREP sampling program includes nitrogen-loading estimates from tributaries, wastewater treatment plants, and atmospheric deposition but does not estimate nutrient loadings from sediments (DES, 2012).

To accurately predict sediment transport, a good understanding of the bed shear stress that drives the sediment erosion, suspension and deposition is essential (Le Hir et
al., 2000). Wengrove, et al. 2015, used field measurements of flows in the bottom boundary layer to estimate the bed shear stress, and compared sediment motion thresholds to erosion chamber experiments. Results indicate a correlation between nutrient release and increasing shear stress and thus a means to estimate nutrient loading in regimes of the bay that are always under water. However, a significant portion of the Great Bay is exposed at low tides creating extensive mud flat areas. Although there is no fluid stress in uncovered areas, at the water’s edge shear stresses can be quite high due to the greater influence of waves on the bottom boundary. There are few field observations estimating combined shear stress from both waves and currents in the very shallow regions near the water’s edge and within the swash in natural estuaries. Therefore we use numerical models to predict the transition from current-dominated to wave-dominated shear stress that in turn will determine the sediment transport patterns both spatially and temporally.

In tidally dominated estuaries like the Great Bay, the tides are the primary forcing mechanism, with meteorological subtidal, river, and wind-induced flows as a secondary driver of the currents. The channels are maintained by tidal and river motions, whereas the sediment transport on the mudflats is governed by both the tide, and wind induced wave motions. In shallow estuarine environments where the tidal range is on the order of the mean depth the hydrodynamics are strongly nonlinear and there are asymmetries in the current strength and duration between the flood and ebb tides. This can have a dramatic effect on the net transport of sediment, nutrients and pollutants.

Small waves in shallow water can have a large effect on sediment transport because they induce shear stresses comparable to current induced shear stresses (Le Hir, 2000). Accurate estimation of shear stress due to waves requires an understanding of the bottom orbital velocities determined by wave height and period and the water depth. Comprehensive observation of bed shear stress over the (often) large spatial extent of estuaries and mudflats would require extensive arrays of instruments, an exceedingly difficult and expensive effort. It is much more feasible to estimate the bed shear stress from numerical models that have been verified with much sparser instrumented field arrays (as in Chapter 3 of this thesis). Use of hydrodynamic, wave model, and coupled
models are needed to deduce the net sediment transport in estuaries due to spatial variations in bed shear stress as a function a both tidal and wave processes.

The purpose of this dissertation is to improve our understanding of the spatial and temporal distribution of shear stress in an estuary using a validated numerical model. Each chapter is written and organized for publication in peer-reviewed journals. Chapter 2, entitled “Modeling nonlinear tidal evolution in an energetic estuary” was accepted in Ocean Modeling on February 14, 2019, outlines the validation process of the numerical model, and co-authored with Thomas C. Lippmann and James D. Irish. Chapter 3 describes the spatial and temporal distribution of numerically modeled shear stress from tidal forcing alone, with comparisons to field studies from 2011 and 2016-2017. This Chapter is presently under internal review and will be submitted to the journal Estuaries, Coastal and Shelf Science. Chapter 4 describes the wave climate of the great bay based on a 2018 field study and meteorological data. Spatial and temporal distribution of shear stress under the influence of modeled waves and tides are described, and compared with analytical estimates of shear stress based on those field studies from spring/summer 2018. Chapter 5 is a conference paper presented at Coastal Dynamics in June 2017 entitled “Tidal energy dissipation in three estuarine environments”, and refers to initial work exploring the tidal evolution of the Great Bay estuary, and compared with two dynamically different estuaries; Hampton-Seabrook estuary, New Hampshire, and New River Inlet in North Carolina.
CHAPTER 2
MODELING NONLINEAR TIDAL EVOLUTION IN AN ENERGETIC ESTUARY

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2.1 Abstract

Three-dimensional numerical simulations of a tidally dominated estuary within the Gulf of Maine are performed using the Regional Ocean Modeling System (ROMS) and validated with observations of sea surface elevation and velocity time series obtained between 1975 and 2016. The model is forced at the ocean boundary with tidal constituents (M2, S2, N2, O1, K1), a time series of observed subtidal elevations and discharge from seven rivers that drain into the estuary. Harmonic analysis is used to determine the tidal dissipation characteristics and generation of overtides within the system. Amplitude decay and phase shift of the dominant semidiurnal (M2) tidal component shows good agreement with observations throughout the main channel of the Piscataqua River and over the channels and mudflats of the Great Bay. The model simulates harmonic growth of the overtides across the spectrum, and indicates a spatial evolution of the tide consistent with a shoaling wave that evolves from a skewed elevation profile with ebb dominance in the lower parts of the estuary, to a more asymmetric, pitched-forward shape consistent with flood dominance. The M4 constituent has spatial variation qualitatively similar to the observations but has magnitudes that are under-predicted in the complex bathymetric region of the Piscataqua River where much of the M2 tidal dissipation occurs. The M6 tidal constituent agrees well with the observations throughout the estuary suggesting that frictional effects on harmonic growth are well modeled. Root-mean-square model-data differences in velocities (~0.05 m/s) and sea surface elevation (~0.1 m) agree to within about 10% of the tidal amplitudes. Differences between model simulations with and without subtidal oscillations in the
estuary are small, suggesting that interactions between the tide and other low frequency (subtidal) mean flows are weak and can be ignored when considering tidal dynamics. Including average fresh water discharge in the model does not affect the behavior of the tidal flows, but can generate high frequency baroclinic velocities potentially important to mixing within the estuary.

2.2 Introduction

The transport and mixing of water, sediment, nutrients and organisms in estuarine and coastal systems is often dominated by astronomical tidal forcing. Of particular interest are the dynamics of shoaling tides induced by nonlinear wave interactions and energy dissipation, and how that process impacts long term coastal planning and environmental conservation efforts. As the tide propagates from the open ocean onto the shelf and into estuaries, it becomes progressively more nonlinear and distorted, leading to growth (shoaling) or decay (dissipation) of tidal amplitudes, shifts in the phase of the tide, and growth of tidal harmonics. Resulting tidal currents are difficult to predict analytically over realistic and complex bathymetry, and require observation or numerical simulation to quantify. Evolution of tidal nonlinearities produces asymmetries in ebb/flood current strength and duration (Boon and Byrne, 1981), that when averaged over a tidal cycle has been used to estimate net sediment transport and circulation patterns (Dronkers, 1986); stronger flood currents drive the movement of coarse sediment and longer slack periods promote the deposition of fine-grained sediment.

Tidal amplitude attenuation in an estuary occurs from energy losses due to turbulent mixing and from frictional affects due to interactions with the bottom and lateral boundaries of the estuary. Energy dissipation of the tidal wave can be described in terms of amplitude decay of the dominant tidal constituent, which for the Gulf of Maine is the semi-diurnal M2 tide that contributes about 90% of the predicted tidal variance. Not all energy is dissipated due to frictional effects, and some is transferred to higher harmonics (overtides; e.g., the M4 and M6 tidal constituents) through nonlinear interactions and frictional effects that create tidal asymmetry (Aubrey and Speer, 1985; Speer and Aubrey, 1985; Parker, 1991). A comparison of the magnitude of the M2
constituent with the first harmonic M4 is a direct measure of nonlinear interactions of the M2 tide, whereas the phase difference qualitatively describes the tidal asymmetries in the system (Friedrichs and Aubrey, 1988). Generation of the M6 component is largely attributed to frictional affects (Parker, 1991).

The dissipation problem is complicated by the highly nonlinear nature of tidal shoaling and propagation, and the need to define representative bottom boundary conditions that characterize the interactions between tidal currents and the seabed. Dissipation in inlets and estuaries leads to development of local phase lags between pressure and velocities that shift slack tide periods up to a quarter of the wave period (90 deg phase shifts between sea surface elevation and along channel velocity), and also impacts the evolution of tidal harmonics that are amplified and phase-shifted relative to open ocean values. This behavior can affect the overall transport in the estuary, thus a good understanding of the spatial and temporal patterns in tidal dissipation can aid in long-term coastal management and planning, for example site selection for tidal renewable energy projects (Neill, et al., 2014).

The tides may also interact nonlinearly with river flow, storm surges and wind driven currents that vary on time scales of hours to months. Often observations from only a few locations are used to describe the overall dynamics of an estuary, and field experiments are limited to one specific area for a discrete amount of time. It is often not feasible to collect enough measurements continuously everywhere to adequately characterize the tides and associated flows; thus, numerical models can be used to produce system-wide predictions of water levels and currents under different hydrodynamic and meteorological forcing conditions (e.g., Warner, et al, 2005a). Quantitative prediction of tidal amplitudes and currents is needed for flooding and inundation studies, mooring and berthing design, safe navigation, interaction with structures, and bottom shear stress prediction for sediment transport, organism transport and nutrient fluxes.

In this study, we discuss the implementation and validation of a three-dimensional high-resolution hydrodynamic model of a tidally dominated well-mixed estuary located within the Gulf of Maine. The Gulf of Maine has a natural resonance close to the semidiurnal (M2) tidal constituent (Garrett, 1972), enhancing the tides throughout the
gulf, including connected estuaries and coastal embayments including the Bay of Fundy. In this study we examine the Piscataqua River - Great Bay estuary located within the Gulf of Maine at the border of New Hampshire and Maine (Figure 1). Tidal forcing for the Great Bay is dominated by the semidiurnal (M2) component of the tide, has a tide range on the order of 2-4 m (depending on the spring-neap cycle), and has variable (but mostly minor) freshwater river discharge. It is home to both the second deepest U.S. naval port, and Portsmouth Harbor, which is home to some of the fastest tidal currents of any commercial port on the U.S. East Coast. The estuary has two tidal regimes: a high dissipative region through the lower Piscataqua River from the mouth to Dover Point, and a low dissipative regime from Dover Pt. through the Little Bay and Great Bay (Brown and Trask, 1980; Swift and Brown, 1983). The former region behaves like a partially progressive wave with concomitant phase shift of the slack tidal period, whereas the latter has phase shifts consistent with standing waves. This behavior causes changes in the timing of tidal currents and the associated net sediment transport throughout the estuary. Previous modeling studies of the Great Bay (Ip, et al., 1998; Erturk, et al., 2002; McLaughlin, 2003) considered depth-integrated, two-dimensional flow fields, with the primary focus of representing the gross tidal behavior to estimate the net transport of water and sediment in the estuary.

The model validation process includes examination of the nonlinear tidal behavior that drives tidal asymmetry and tidal energy dissipation in terms of amplitude decay and phase lags using water level measurements and harmonic analysis. Modeled results are compared with coincident and previous observations, and with results from the literature. This study will form the basis for additional modeling aimed at examining the spatial variation in bottom shear stresses needed for sediment transport calculations, horizontal and vertical mixing within the estuary, and transport of larvae, nutrients and carbon within the estuary.

Section 2.3 describes the field site, observational datasets, the hydrodynamic model and grid development, and the model validation and tidal analysis methodology. Section 2.4 describes model results, and Section 2.5 discusses the model-observation comparison in terms of nonlinear evolution. Section 2.6 presents the conclusions of the study.
2.3 Methods

2.3.1 Site Description

The Great Bay Estuarine system is located along the New Hampshire-Maine border within the Gulf of Maine in the northeastern portion of the United States (Figure 2.1). It is a recessed, drowned river valley connected to the Gulf of Maine via the Piscataqua River (Armstrong, et al., 1976). The tide range is 2-4 m over the spring-neap cycle with tidal currents greater than 2 m/s in the channels at maximum ebb and flood.
tides. At low tide as much as 50% of the Great Bay is exposed as low-lying mudflats, cut with deep tidal channels. The surface area of the estuary is approximately 55 km² measured at mean high water (NHDES, 2007). The volume is 156·106 m³ and 235·106 m³ for low and high tides respectively, with a tidal prism of 79·106 m³ (Swift and Brown, 1983; NHDES, 2007). Seven tributaries contribute fresh water to the system: the Squamscott, Lamprey, Winnicut, Oyster, Bellamy, Cocheco, and Salmon Falls, all feeding the Upper and Lower Piscataqua river that flows into the Gulf of Maine. River fluxes are determined by precipitation and runoff and regulated by dams or weirs that modulate the freshwater volume entering the system. Typically (except during large storms or the spring melt), the freshwater input is relatively small and only contributes 2% of the tidal prism (Short, 1992; NHDES, 2007). The generally small freshwater fluxes and strong tidal mixing results in weak or negligible stratification (except very close to the river mouths) and during periods of little rainfall the salinities at the Great Bay Buoy (Figure 2.3) are nearly equal to the Gulf of Maine indicating that horizontal variation in density due to river fluxes are also weak. As our interests include the ability of the numerical model to represent the vertically varying flow fields, we will include model runs with and without average river discharges to evaluate the influence of baroclinic flows on the tidal behavior.

Ocean waves outside the mouth of the estuary are strongly refracted away from the deep center channel and rapidly attenuate upstream, and thus do not greatly contribute to the velocities or water level fluctuations in the estuary, other studies have shown that waves can have an impact on tidal currents (e.g., Lewis, et al., 2014). Wind-driven surface gravity waves in the large lobe of the Great Bay proper are generally small (5-20 cm significant heights) owing to the limited fetch and strong attenuation by energy loss through interactions with tidal currents and the muddy bottom or shallow aquatic vegetation (eel grass meadows). Although waves on the Great Bay could be important to bottom shear stresses over the mud flats, they do not substantially alter the larger scale circulation, and thus are not considered further in this study. Wind-driven mean currents may be substantial during storm conditions, but are generally much weaker than the tidal currents (Wengrove, et al., 2015) and thus are also not considered in this study.
The bathymetry of the estuary is complex (Figure 2.2), with steep sidewalls in the main channel of the Piscataqua River with water depths ranging 13-26 m. Ocean water flows into mouth of the Piscataqua River through two channels, a main entry point to the north of New Castle Island between New Hampshire and Maine, and a secondary entry point through Little Harbor to the south of New Castle. Tides entering Little Harbor flow through relatively shallow water and around several islands, and join the Piscataqua River between Pierce Island and Portsmouth, NH. Flows through the main channel make a sharp 90 deg turn around New Castle at Fort Point, and then flow around the Portsmouth Naval Shipyard primarily to the south in the deeper channel but also the back bay, a narrow, shallow waterway that reconnects with the Piscataqua River near Pierce Island. The Piscataqua River splits at Dover Pt., with the main flows sharply turning south into Little Bay, and with a smaller portion of the flow heading to the north connecting the lower Piscataqua River with the Upper Piscataqua fed by the Cocheeco, and Salmon Falls rivers to the north, with average summer discharge rates of 8.54 and 15.4 m³/s, respectively (NHDES, 2007).

The channel between the mouth at New Castle Island and Dover Pt. is 12 km long and characterized by a hard, rocky bottom with coarse sediment in the deep channels and steep rocky shorelines for most of the reach. The flows through this part of the estuary are high (exceeding 2 m/s in some locations) on both the flood and ebb tides. Once the flow enters the Little Bay it remains strong through the deep center channels with weaker flows up and over the bordering mud flats. The Oyster and Bellamy rivers that flow into the Little Bay have average summer discharges of 0.94 and 1.32 m³/s, respectively (NHDES, 2007). The Little Bay joins the Great Bay at Furber Strait near Adam’s Pt. The deep center channel gradually shallows and bifurcates into an eastern and western branch flanked by large mud flats that dominate this portion of the estuary. The Squamscott, Lamprey, and Winnicut rivers all flow into this part of the estuary, with average summer discharge rates of 5.3, 10.0, and 0.7 m³/s, respectively (NHDES, 2007). For this study, the tidal analysis focuses on the main channel flows from the mouth of the Piscataqua River to the upper reaches (Squamscott River) of the Great Bay estuary (Figure 2.1).
Figure 2.2: Topographic and bathymetric elevations relative to mean sea level for the Great Bay Estuary. Background image is from Landsat 8.
2.3.2 Observations

Field observations of horizontal currents spanning the water column and sea surface elevation (from bottom pressure and tidal stations) were obtained during several field experiments in 1975, 2007, 2009, 2015, and 2016, and the continuously operating NOAA Tide Gauge station at Fort Point, NH (Station ID: 8423898). Table 2.1 summarizes the dates and durations of the field studies and Figure 3 shows the instrument locations.

Observations of tidal elevations and currents within the estuary were obtained in 1975 by the University of New Hampshire (UNH) in cooperation with the National Ocean Survey (NOS; summarized in Swenson et al., 1977 and Silver and Brown, 1979). Original data were unavailable so tidal analysis estimating M2 tidal amplitudes and phases from Swift and Brown (1983) is used in this study. Observations of bi-directional currents (in 0.5 – 1.0 m range bins) and water levels from the mouth to Adams Pt. were obtained by NOAA in 2007 using six bottom-mounted, upward-looking acoustic Doppler current profilers (ADCPs). The instruments were deployed for between 41 and 45 days, recovered, and then moved to new locations with water depths ranging between 4.3 and 19.3 m. These data are available and described online at https://tidesandcurrents.noaa.gov. Observations of water levels were obtained by UNH in 2009 at four locations in the Great Bay using bottom mounted pressure sensors and an RTK GPS buoy. The instruments were sampled between 30 and 120 s and deployed between 9 and 84 days, and averaged over 6 min intervals following standard NOAA procedures. Observations obtained for 7 – 71 days by UNH in 2015 and 2016 include 1 min averaged bi-directional currents (in 0.25 – 1.0 m range bins) and water levels from six ADCPs deployed across the Great Bay in water depths ranging 3 – 17 m. Bottom pressure was converted to sea surface elevation using observed bottom temperature at the
Figure 2.3: Observational measurement locations. Along-channel distance from the mouth of the estuary is determined using the PIR0701 sensor (most seaward diamond symbol) from the 2007 NOAA Piscataqua River Current Survey.
Table 2.1: Observations used in the study with number of locations and duration of deployments.

<table>
<thead>
<tr>
<th>Year</th>
<th>Program</th>
<th>Data Variable</th>
<th>Number of Locations</th>
<th>Duration</th>
</tr>
</thead>
<tbody>
<tr>
<td>1975</td>
<td>Great Bay Estuary Field Program (Swenson et. al. 1977, Silver and Brown, 1979)</td>
<td>Water Level&lt;sup&gt;a,b&lt;/sup&gt;</td>
<td>10&lt;sup&gt;*&lt;/sup&gt;</td>
<td>21 – 333 days</td>
</tr>
<tr>
<td>2007</td>
<td>Piscataqua River Current Survey (<a href="https://tidesandcurrents.noaa.gov/cdata">https://tidesandcurrents.noaa.gov/cdata</a>)</td>
<td>Water Level and Currents&lt;sup&gt;c,d&lt;/sup&gt;</td>
<td>10</td>
<td>41 – 45 days</td>
</tr>
<tr>
<td>2009</td>
<td>CCOM Great Bay Survey</td>
<td>Water Level&lt;sup&gt;e,f&lt;/sup&gt;</td>
<td>6</td>
<td>9 – 84 days</td>
</tr>
<tr>
<td>2015</td>
<td>Great Bay Field Study</td>
<td>Water Level and Currents&lt;sup&gt;g,h,i,j&lt;/sup&gt;</td>
<td>8&lt;sup&gt;+&lt;/sup&gt;</td>
<td>7 – 35 days</td>
</tr>
<tr>
<td>2016</td>
<td>Great Bay Field Study</td>
<td>Water Level and Currents&lt;sup&gt;d&lt;/sup&gt;</td>
<td>1</td>
<td>71 days</td>
</tr>
<tr>
<td></td>
<td>NOAA Tide Gauge (8423898) at Ft. Point</td>
<td>Water Level&lt;sup&gt;k&lt;/sup&gt;</td>
<td>1</td>
<td>Continuous</td>
</tr>
<tr>
<td>2009-2016</td>
<td>UNH Great Bay Buoy (<a href="http://www.opal.sr.unh.edu/data/buoys/great_bay/index.shtml">http://www.opal.sr.unh.edu/data/buoys/great_bay/index.shtml</a>)</td>
<td>Salinity&lt;sup&gt;l&lt;/sup&gt;</td>
<td>1</td>
<td>Seasonal (~ 9 months)</td>
</tr>
</tbody>
</table>

* Original data unavailable; water levels and current analysis used in this study are provided in Swift and Brown (1983).
+ One instrument was moved to 4 different locations within Great Bay for deployments between 7 and 14 days

a. automatic digital recording (ADR) tide gauge
b. Metritape Inc. Level sensor
c. 600 kHz RDI ADCP
d. 1200 kHz RDI ADCP
e. Aanderaa tide gauge
f. SeaBird Seacat
g. 500 kHz RDI Sentinel V ADCP
h. 1200 kHz RDI Workhorse Sentinel ADCP
i. 3 mHz Sontek Argonaut ADCP
j. 2 mHz Nortek Aquapro ADCP
k. acoustic water level (Next Generation Water Level Measurement System)
l. YSI 6600 Sonde
instrument location and salinity obtained from the Great Bay Coastal Buoy located in the center of the Great Bay Estuary

2.3.3 Hydrodynamic Model

The Regional Ocean Modeling System (ROMS, Haidvogel et al., 2008; Shchepetkin and McWilliams, 2005) is an ortho-curvilinear three-dimensional numerical coastal ocean circulation model that solves finite-difference approximations of the Reynolds-averaged Navier Stokes (RANS) equations using the hydrostatic and Boussinesq assumptions. The objectives of this study focus on the hydrodynamic component to determine the tidal dynamics, which are of first order concern in validating the numerical model. ROMS has been used in both regional (e.g., Zhang, et al. 2009; Yang, et al. 2016) and estuarine modeling studies (e.g., Warner, et al. 2005a; Moriarty, et al., 2014), and implemented into other coupled modeling systems (e.g., Warner, et al., 2008; Warner, et al., 2010).

A third order upwind advection scheme is used to solve for horizontal advection. A centered-fourth order advection scheme is used to solve for vertical advection. A k-ε generic length scale (GLS) turbulence closure model is used to calculate the horizontal and vertical eddy viscosities (Umlauf and Burchard, 2003; Warner, et al., 2005b) in conjunction with the Kantha and Clayson (1994) stability function. Within ROMS the wetting and drying algorithm (Warner, et al., 2013) is utilized to simulate the inundation and draining of the tide over shallow areas alternatively covered and uncovered by the tide, in which the critical depth (Dcrit) is set to 10 cm. Once the total water depth is less than Dcrit, no flux is allowed out of that cell and it is considered “dry”. Finally, barotropic and baroclinic modes are solved separately in ROMS with the mode-splitting algorithms described in Haidvogel, et al. (2008). Barotropic time steps in model simulations herein are 1/20 of the baroclinic time step.
2.3.3.1 Model Grid

The model domain is defined by a rectilinear Arakawa “C” grid with a constant 30-by-30 m horizontal resolution (Figure 4; downsized by a factor of 33 1/3 in the figure). There are 8 vertical layers in a terrain-following (σ) coordinate system that is adjusted for slightly higher resolution near the surface and bottom boundaries. The domain is rotated 37 deg CCW from true north to align the offshore boundary with the approximate orientation of the shoreline along the New Hampshire-Maine coast. The domain ranges 22.02 by 25.02 km (734 by 834 cells). The grid elevations were defined using bathymetric data obtained from the Center for Coastal and Ocean Mapping (CCOM; http://ccom.unh.edu), and LIDAR data collected by USGS, NOAA, and USACE (https://coast.noaa.gov/dataviewer), and interpolated onto the center of the horizontal grid cells. A hierarchy was defined that weighted the most accurate, recent, and complete topographic and bathymetric data highest, with any gaps filled with more uncertain, older, or less complete data sources. The combined elevation grid (Figure 2.4) was then processed with the MATLAB Easygrid routine (https://www.myroms.org/wiki/easygrid) to create the rectilinear grid and corresponding land mask that was subsequently input into ROMS. During model testing, the grid was smoothed in locations sensitive to numerical instabilities using interpolation methods described in Plant, et al. (2002).
Figure 2.4: Rotated ROMS horizontal grid ($\xi, \eta$) coordinates and model defined boundaries. Displayed gridlines are every 1 km, decimated by a factor of 33 $\frac{1}{3}$ from actual model grid (for display purposes). Cardinal directions of boundaries are relative to the orientation of the rotated grid.
2.3.3.2 Boundary Conditions

At the open ocean boundary (south edge of the rotated domain; Figure 2.4) the model is forced by tidal and subtidal oscillations (see Section 2.3.3.3) using the implicit Chapman (free surface) and Flather (depth averaged velocity) boundary conditions. The Chapman-Flather conditions employ the radiation method at the boundary, assuming all outgoing signals leave at the shallow water wave speed (Flather, 1976; Chapman, 1985). These particular boundary conditions have been shown to be the most suitable for tidal forcing (Palma and Matano, 1998, 2000; Marchesiello, et al., 2001; Carter and Merrifield, 2007). Three-dimensional baroclinic momentum equations were set to radiation and gradient conditions for velocities and tracers. The eastern, northern, and western edges of the domain are closed.

The bottom boundary condition for momentum was parameterized by a simple drag coefficient assuming a logarithmic vertical velocity profile in the bottom vertical cell. The drag coefficient, $C_D$, is represented by

$$C_D = \left(\frac{\kappa}{\ln(z/z_{ob})}\right)^2$$  \hspace{1cm} (1)

where $z$ is the vertical elevation of the mid-point of the bottom cell, $z_{ob}$ is a characteristic bottom roughness (in m), and $\kappa=0.41$ is the von Karman coefficient (Kundu, 1990). A range of bottom roughness values (from 0.015 – 0.030 m) were tested and the best fit was determined iteratively from model-data comparisons of M2 tidal dissipation as a function of distance from the estuary mouth (see Figure 5). Within each run, $z_{ob}$ was assumed to be spatially uniform across the domain. The kinematic bottom stress boundary conditions are given by

$$\tau_b^x = \rho_0 C_D u \sqrt{u^2 + v^2}$$  \hspace{1cm} (2)

$$\tau_b^y = \rho_0 C_D u \sqrt{u^2 + v^2}$$  \hspace{1cm} (3)

where $\tau_b^x$ and $\tau_b^y$ are the bottom stresses in the x and y directions, respectively.
Figure 2.5: M2 tidal energy decay, $E_{\text{norm}}$, (upper panel) and phase evolution (deg Greenwich; middle panel) as a function of distance from the mouth of the estuary for different bottom roughness values ($z_{ob}$) of the logarithmic drag law bottom boundary condition. Observations are included as symbols with error bars based on T_TIDE analysis and following Taylor (1982). Depth profile along the center channel is shown in the lower panel with locations of Fort Point, Dover Point, Adams Point, and the Squamscott River are indicated.
2.3.3.3 Model Initialization and Forcing

Forcing conditions at the open ocean boundary are specified in two ways. The first is with an analytical representation of tidal elevations and velocities considering only the principal semidiurnal (M2, N2, S2) and diurnal (O1, K1) tidal constituents determined by the Oregon State University global Tidal Prediction Software package (OTPS) in conjunction with the United States East Coast regional Tidal Solution (EC2010; Egbert and Erofeeva, 2002). The OTPS provided the necessary tidal amplitude and phases that correspond to the observational datasets for the 2015 field study used in the model-data comparisons of velocities (see Section 2.3.4.2 and Figure 2.6). The amplitudes and phases compared favorably with a harmonic analysis of observed water level fluctuations at Fort Pt. for the 2015 field experiment using T_TIDE (Pawlowicz, et al., 2002).

The second forcing consists of the analytical representation of the tides and including subtidal oscillations associated with atmospheric motions obtained from low-pass filtered (with a 33 hr cut-off period) observed time series of 6-minute averaged water levels at the Fort Pt. tidal station. The subtidal motions can have amplitudes in the Gulf of Maine of 0.10-0.30 m (Brown and Irish, 1992), change the water depth over the shallow mudflats considerably, and although the time scales of the oscillations are generally much longer than the dominant semidiurnal tides, may contribute to the overall water velocities on the flood and ebb. Coastal ocean currents associated with barometric, wind-driven, or other shelf motions at the offshore open boundary are assumed small (consistent with observations of currents from 2007 at the most seaward instrument location, PIR0701) and not considered herein.

In each case (tidal with or without subtidal forcing), time series of water level fluctuations are ramped hyperbolically from rest over a 2-day period. Although tidal currents are included at the open boundary, test simulations in which the boundary currents were set to zero and allowed to evolve with the sea surface fluctuations did not alter the results, suggesting that approximating the forcing by only the pressure gradient at the mouth is reasonable (consistent with Geyer and MacCready, 2014). Time series of at least 32 days are used to force the model so that tidal analysis with T_TIDE produces
amplitudes and phases of the dominant tidal constituents (with confidence intervals). The open ocean boundary is located about 7.5 km from the mouth of the estuary where the Fort Pt. tide station is located. The time for the tide wave to propagate this distance is small, about 7.3-8.1 min based on an average water depth of 30-24 m, and thus has small effect on the phase estimates (about 3.3-3.9 deg) when comparing to coincident observations within the estuary.

Three-dimensional simulations were performed both with and without freshwater flows based on the average summer river discharge (see Section 2.3.1), salinity (varying between 6.93 and 23.54 psu), and water temperature (varying between 19.5 and 25.4 deg. C) for the various rivers for the summer of 2015 was provided by the New Hampshire Department of Environmental Services (https://www.des.nh.gov/organization/divisions/water/wmb/vrap/data.htm). Ocean water temperature (17 deg. C) and salinity (31.5 psu) was assumed constant and given by typical summer values for the Gulf of Maine. Diurnal surface heating and cooling were assumed small in comparison to the tidal mixing and were ignored. Although the precise values of the fluctuating river discharge, temperature, and salinity were not used in the model, the variations in temperature and salinity predicted by the model compare favorably with 2015 observations obtained in the middle of the Great Bay near the surface with the Great Bay Coastal Buoy (http://www.opal.sr.unh.edu/data/buoys/great_bay/index.shtml) and near the bottom with the SeaBird instruments co-located with our ADCP’s deployed in 2015. Modeled and observed fluctuations in temperature and salinity follow tidal cycles and reveal weak vertical gradients in temperature (about 1-2 deg. C) and salinity (about 1-2 psu), consistent with a well-mixed Great Bay environment away from the river mouths during typical summer conditions in New Hampshire.

Model simulations including subtidal oscillations and river fluxes had a very weak effect on the tidal behavior and thus the results presented below will focus on the model simulations for barotropic tides. This is not unexpected for the typical summer conditions examined herein, but might be an important consideration during extreme storms and high runoff periods or in the very shallow depths near the water’s edge over the mudflats. The effect of subtidal oscillations and baroclinic flows is discussed further
in Section 2.5. In the following, tidal analysis from model simulations will be compared with observations within the estuary obtained in different experiments at different time periods (tidal constituents are assumed to be the same throughout). Our model runs focus on the 35 day period spanning the 2015 field experiment, and will be compared with observed velocity and sea surface elevation time series from 2015 and tidal analysis of observations obtained during all experimental periods (Table 2.1).

Time resolution was determined by iteration on grid smoothing and reducing barotropic and/or baroclinic time steps until numerical stability was achieved. For model simulations presented herein, a baroclinic time step of 1.5 s was used, with barotropic time step 1/20 of that value. Computations were performed on a Cray XE6m-200 supercomputer at the Institute of Earth, Ocean, and Space at the University of New Hampshire, and the Blue Waters CRAY XE6 supercomputer located at the University of Illinois-Urbana-Champaign. Output over the whole domain was stored to disk at 30 min average model time intervals, and for 15 min averaged intervals at specific save points corresponding to instrument locations and along a densely sampled line every 100 m along the main transect passing through the entire estuary.

2.3.4 Model Validation Methods

Model validation is accomplished in four ways. The first is by conducting a tidal analysis, and comparing the modeled energy decay and phase shift of the dominant M2 tidal constituent throughout the estuary with similar analysis of observations of sea surface elevation time series. The second is by comparing modeled time series of the vertical variation in currents with observations. The third is with cross-spectral analysis between modeled and observed sea surface elevation, and horizontal velocity components at single locations, and with the evolution of cross-spectral phase at the M2 frequency between sea surface elevation and along-channel velocities. The fourth is by comparing the growth and phase change of M4 and M6 tidal harmonic constituents between modeled and observed time series, and by comparison of the along-estuary evolution of sea surface elevation skewness and asymmetry.
2.3.4.1 Tidal dissipation and phase change

As the tide propagates into shallow coastal regions and interacts with bottom topography and basin geometry, it loses energy through frictional processes that result in tidal amplitude decay and phase changes relative to the open ocean value. Due to phasing of the tide a direct time series comparison is only possible for model runs that coincide with the specific phases of the tide during that particular field study. However, tidal analysis of long (30+ day) time series of sea surface elevation obtained at other times can be compared with non-synchronous model simulations, provided there are no other atmospheric effects that non-linearly interact with the tide and do not substantially change the tidal behavior. Therefore, we conduct a tidal analysis (using T_TIDE; Pawlowicz, et al., 2002) to decompose each time series of sea surface elevation into tidal components and compare the modeled and observed tidal constituent energy from the linear gravity wave relation,

\[ E = \frac{1}{2} \rho g A^2 \]  

(4)

where \( E \) is the total energy per unit surface area, \( A \) is the amplitude of the tidal constituent, and the density \( \rho \) is assumed constant throughout the estuary. In this study the semidiurnal M2 tide dominates, contributing about 88% of the total tidal energy at the mouth of the estuary. The energy at any location within the estuary, \( E_{\text{station}} \), is normalized by the value at the estuary mouth, \( E_{\text{ocean}} \), to represent the fractional energy loss, \( E_{\text{norm}} \), as the tide progresses upstream,

\[ E_{\text{norm}} = \left( \frac{A_{\text{station}}}{A_{\text{ocean}}} \right)^2. \]  

(5)

Assuming the uncertainties in the tidal amplitudes, \( \delta A_{\text{station}} \) and \( \delta A_{\text{ocean}} \), are both independent and random, then the error \( \delta E_{\text{norm}} \) is calculated following Taylor (1982),

\[ \delta E_{\text{norm}} = E_{\text{norm}} \cdot \sqrt{2 \ast \left( \frac{\delta A_{\text{station}}}{A_{\text{station}}} \right)^2 + 2 \ast \left( \frac{\delta A_{\text{ocean}}}{A_{\text{ocean}}} \right)^2}. \]  

(6)

Initial model calibration involves testing different bottom boundary conditions, and iterating to estimate the energy decay as a function of distance from the estuary mouth that best fits the observations.
2.3.4.2 Time series comparison of vertically varying currents

Modeled currents are computed at defined $\sigma$ coordinate levels that range from $\sigma = -1$ at the bottom to $\sigma = 0$ at the surface. The total water depth in the model is given by the elevation of the seabed (relative to the model datum defined) plus the corresponding (fluctuating) sea surface elevation. The observations, on the other hand, are obtained from fixed, upward looking ADCP’s with vertical bin elevations defined in a fixed coordinate system relative to the bottom. The range over which the currents are observed depends on the instrument characteristics (e.g., acoustic frequency and instrument capabilities) and the height of the fluctuating sea surface relative to the bottom. Acoustic interference by side-lobes at the surface limit the range of useable vertical bins to be less than 94% of the total instantaneous water depth (and appropriate filtering methods must be employed to eliminate spurious velocities near the surface). As a consequence, the velocities observed with ADCP’s in the field further from the bottom have bins coming into and out of the water column as the tide rises and falls.

To compare the modeled to observed currents, the modeled currents (in $\sigma$ coordinates) are transformed to the observational coordinate system by linear interpolation over the instantaneous water level at each time step. In this manner, the modeled time series at the transformed upper bins also come into and out of the water surface similar to the observations. Care must also be taken to represent the velocities from the observations at the center of the vertical bins, and the model at the defined location by the $\sigma$ coordinates. A representative example of the time series comparison is shown later (Figure 2.6) and described in Section 2.3.2.

2.3.4.3 Cross-spectral Analysis

A more complete evaluation that includes the overall behavior of the modeled velocities can better be done with cross-spectral analysis that shows the energy density levels for both the model and the data as a function of frequency, and the coherence and phase relationship for each frequency. As our interests lie with the tidal constituents, the frequency resolution of the spectra will necessarily need to be fine enough to resolve the
major constituents, with lowest tidal constituent (the O1 diurnal variation) of about 0.0417 \( \text{hr}^{-1} \). At the same time, the confidence intervals on the spectra, coherence, and phase must be high enough to make reasonable comparisons. For the 30 day time series examined, cross-spectra were computed with 10 degrees of freedom (DOF) by averaging 5 adjacent frequency bands. The frequency bandwidth of the smoothed spectral estimates was 0.0069 \( \text{hr}^{-1} \) with lowest resolved frequency of 0.0035 \( \text{hr}^{-1} \). The 95% confidence intervals are computed for the spectral amplitudes, coherences, and phase. Only those phase estimates for frequencies with coherence greater than the 95% critical value (0.52 for 10 DOF) are shown (phase error bars for incoherent frequencies are meaningless; Bendat and Piersol, 2000). To reduce leakage effects, a Hanning data window is applied to each mean-corrected time series before computing the spectra.

2.3.4.4 Sea surface elevation and along-channel velocity phase difference

Tidal analysis of the sea surface elevation and velocities can be compared to show the relative change in phase as the tide evolves up the estuary. In this case, the observed and modeled bi-directional velocities were rotated to align with the along-channel direction using standard rotary analysis (Gonella, 1972). Ellipse orientations for the dominant M2 tidal frequency defines the angle of the major axis of the rotary ellipse that is used in the rotation to along-channel direction. We conduct a tidal analysis to decompose each time series of the along-channel velocity into amplitudes and phases for each harmonic tidal constituent frequency following the same procedure for the sea surface elevation (see Section 2.3.4.1). The phase difference between the sea surface height \((P)\) and along-channel velocity \((U)\) at the M2 tidal frequency was computed for time series at locations that span the estuary and reported as the P–U phase.

The evolution of the P–U phase for the dominant M2 tidal constituent indicates the nature of the tidal motion throughout the estuary (Figure 2.10; top panel). In a progressive wave, the maximum currents occur at the same time as the maximum height of the wave, and the currents and amplitude are in phase. In a standing wave the
maximum currents occur at mid-tide, half way between the crest and the trough of the wave, and the along-channel currents are 90 deg out of phase with the sea surface height.

2.3.4.5 Tidal harmonic growth and phase difference

The growth of the M4 harmonic relative to the M2 constituent is a measure of the asymmetry and non-linear distortion of the tide (Friedrichs and Aubrey, 1988). Following Speer and Aubrey (1985), the amplitude ratio, $A_{ratio}$, and the phase difference, $\theta_{diff}$, is defined as,

$$A_{ratio} = \frac{A_{M4}}{A_{M2}}$$

$$\theta_{diff} = 2\theta_{M2} - \theta_{M4}$$

where $A_{M4}$ and $A_{M2}$ are the amplitudes of the M4 and M2 tidal constituents, respectively, and $\theta_{M4}$ and $\theta_{M2}$ represent corresponding phase relationships. In general, stronger frictional effects produce larger $A_{ratio}$, and the corresponding $\theta_{diff}$ describes the gross behavior of the tides with phase differences between 0° and 180° (180° and 360°) indicating flood (ebb) dominance (Friedrichs and Aubrey, 1988). Flood dominant systems have characteristically stronger flood currents and longer falling than rising tides, whereas ebb dominant systems have stronger ebb currents and longer rising tides.

The amplitudes and phases of the M2 and M4 tidal constituents are estimated with a tidal harmonic analysis (using T_TIDE) that fits harmonics to the time series and computes error bars on the estimates of amplitude and phases for each constituent, allowing estimates of the uncertainty in $A_{ratio}$ and $\theta_{diff}$ (Taylor, 1982). The error estimates for $\delta A_{ratio}$ and $\delta \theta_{diff}$ are calculated using the following formulations,

$$\delta A_{ratio} = \delta A_{ratio} \cdot \sqrt{(\delta A_{M4}/A_{M4})^2 + (\delta A_{M2}/A_{M2})^2}$$

$$\delta \theta_{diff} = \sqrt{(\delta \theta_{M2})^2 + (\delta \theta_{M4})^2}$$

following Taylor (1982), similar to $\delta E_{norm}$ (Equation 6).
The third moments, skewness and asymmetry, of observed and modeled sea surface elevation time series are computed along the estuary (following Elgar and Guza, 1985). The normalized (by the variance to the 3/2 power) skewness describes the general nonlinear deviation of the wave profile from a sinusoidal shape to a peaked-up waveform symmetrical about the vertical axis through the wave crest. The normalized asymmetry describes the asymmetry about the vertical axis, and can indicate a pitched forward (or pitched backwards) wave form. The nature of the skewness and asymmetry is determined by the phase relationship between the primary frequency and the coupled harmonics. For purely skewed (peaked up, Stokes-like) wave profiles, the asymmetry is zero and the primary and higher harmonics are in-phase. For pitched forward (backward) the asymmetry is nonzero and negative (positive). Sawtooth profiles have high negative asymmetries and phase relationships between the primary and first harmonic up to +/- 90 deg. Evaluation of waveforms for wind-driven surface gravity waves in the ocean and their relationship to third moments can be found in Elgar and Guza (1985).

2.4. Results: Model-Observation Comparison

Results comparing model simulations for barotropic tides with observations are presented here, and follow the methodologies discussed previously. Station data are retained from the model simulations at all the sensor locations, as well as from locations separated by 100 m along a transect down the main channel extending from the first sensor location outside the mouth of the estuary to the upper reaches of the Great Bay by the Squamscott River.

2.4.1 Tidal dissipation and phase change

The observed energy decay and phase change of the M2 tidal constituent relative to the value at the most seaward location along the station transect through the estuary is shown in Figure 2.5. The most seaward observation (1 km from the Ft. Point tidal station) closely matches the predicted tidal amplitude from the OTPS model, and used to
normalize the energy \(E_{norm}\), Eq. 5). Also shown is the variation in the center channel water depth along the transect. Error bars (Eq 5-6) on the energy and phase estimates are based on the T_TIDE analysis. Observations show an increase in tidal energy near the mouth, and then a progressive decrease in energy through the energetic, narrow portion of the lower Piscataqua River. This decay is strong (and somewhat variable) between Portsmouth and Dover Pt., and in general agreement with estimates of dissipation found by Swift and Brown (1983). By the time the tides reach the Little Bay entrance, 45% of the M2 tidal energy has been lost. Over this same reach, the M2 phase has changed 50 deg, significantly larger than for a simple progressive tidal wave propagating upstream (with estimate of about 6 deg phase change based on shallow water wave phase speeds and average water depth of 20 m), and much less than a standing wave with 90 deg phase change.

Interestingly, the tidal amplitudes increase slightly between the entrance to the Little Bay (Dover Pt.) and the upper reaches of the Great Bay (Squamscott Bridge), indicating some amplification as the tide propagates into progressively shallower water. Additionally, the phase continues to evolve (approaching 70 deg) suggesting that the tide here is more reflective. It should be noted that the tidal extent during the flood does not end at the Squamscott Bridge, but continues an additional 8 km inland (as well as up the other rivers; Figure 2.1).

Also shown in Figure 2.5 are model predictions of the M2 tidal decay and phase change for a range of apparent bottom roughness, \(z_{ob}\), from 0.015 – 0.030 m. The best fit to the observation is for \(z_{ob} = 0.02\) m. The model increase in M2 energy across the shallowing Great Bay bathymetry is in general agreement with the observations. In general, the model well predicts the evolution of the tidal phase throughout the estuary.

2.4.2 Time series comparison of vertically varying currents

Comparisons of modeled and observed current time series (for 4 days) from a single location in water depth of about 5.75 m in the Great Bay is shown in Figure 2.6. Both the east-west and north-south velocity comparisons are shown for elevations
(relative to MSL) near the bottom (-4.13 m), mid water column (-2.63 m), and near the surface (-1.13 m). In general, the modeled velocities closely follow the observations including in the upper water column were the “sensor” bins are coming into and out of the water as the tide rises and falls. Root-mean-square (RMS) errors between modeled and observed time series at all elevations above the bottom range 0.035-0.049 m/s and 0.047-0.055 m/s for the east-west and north-south velocity components and 0.095 m for sea surface elevation (each about 10% of the amplitude at that location). In general, the 10% RMS error between model-data time series for all sensors across the Great Bay from the 2015 deployment is quite good, with average RMS errors for sea surface elevations, east-west, and north-south velocities of 0.096 m, 0.054 m/s, and 0.060 m/s, respectively.

2.4.3 Cross-spectral Analysis

Cross-spectra between modeled and observed sea surface elevation, east-west, and north-south currents from a location in the Great Bay are shown in Figure 2.7. Modeled and observed spectral density, \( F \), show similar energy distribution at the tidal constituents, and compare well for the sea surface elevation and both orthogonal components of the velocity. Note that the noise floor associated with the observed spectra is much higher than for the model, a result owing to the sampling uncertainty associated with the pressure sensors and acoustic profiling instruments, as well as the model not considering baroclinic flows (discussed later).

The coherence squared, \( \gamma^2 \), is high (0.99) at the tidal harmonic frequencies, well above the critical value (0.52). The corresponding phase at the energetic M2 frequency is 2.47 deg for the sea surface elevation time series, and 8.48 and 3.98 deg for the east-west and north-south velocities, respectively. The average model-data phase at the M2 frequency for all sensors in the Great Bay during the 2015 deployment for sea surface elevation and the bi-directional velocities was 0.03, 0.34, and 2.32 deg, respectively.

2.4.4 Tidal harmonic growth and phase difference

Modeled and observed power spectra of sea surface elevation, \( F \), from two locations spanning the estuary – one near the mouth at Fort Point and the other in the
Figure 2.6: Modeled (dots) and observed (solid line) time series of east-west (top) and north-south (bottom) velocities from sensor located in 5.75 m water depth in the Great Bay. The vertical elevation relative to mean sea level (in m) of each time series comparison is indicated on the right-hand-side of each panel. The discontinuous time series in the upper three panels are a result of tidal variations in water depth periodically exposing and inundating upper sensor locations near the sea surface. RMS errors range 0.035–0.049 m/s and 0.047–0.055 m/s for the east-west and north-south velocities, respectively.
Figure 2.7: Cross-spectra between modeled and observed sea surface elevation (left panels), east-west depth-averaged velocity (center panels), and north-south depth-averaged velocity (right panels) for sensor location in 5.75 m water depth in the Great Bay. Upper panels show spectral density, $F$, in m$^2$/s for sea surface elevation and m$^2$/s for velocities as a function of frequency ($hr^{-1}$). Spectra were computed with a Hanning data window and 10 DOF. The 95% confidence interval is shown in the upper center panel. Observed spectra have a significantly higher noise floor but still below the energy levels of the harmonics. Center panels show the coherence squared, $\gamma^2$, with 95% significance level as the horizontal dashed line. Lower panels show the phase (deg) with solid circles indicating significant phases with 95% confidence intervals.
Figure 2.8: Modeled (left) and observed (right) spectral density, $F \,(m^2/s)$, of sea surface elevation from two stations, one near the mouth of the estuary (Fort Point; solid line) and one in 5.75 m water depth in the Great Bay (dotted line). Spectra show the growth of the tidal harmonics from the ocean to 20 km up the estuary (the M2, M4, and M6 constituents are indicated). Spectra were computed over a 30 day record and processed with a Hanning data window. Observed spectra have a significantly higher noise floor but still below the energy levels of the harmonics.
Great Bay – are shown in Figure 2.8. The M2 tidal energy decays by about 45% (as shown in Figure 2.5). On the other hand, the spectra show a sharp increase in the energy levels at the tidal harmonics in the Great Bay, evident well beyond the M4 and M6 constituents indicating the strong growth of overtones and nonlinear evolution of the spectra. The growth of the M6 harmonic exceeds that of the M4 harmonic, consistent between the modeled and observed spectra.

To examine the spatial nonlinear evolution of the tidal spectra, the M2, M4, and M6 tidal constituents (as determined by T_TIDE analysis) along the center channel from the mouth to the upper reaches of the Great Bay is shown in Figure 2.9 (along with the depth variation along the transect). The M2 tidal amplitude decays as expected. Modeled M4 and M6 harmonics increase from 2% to 7% of the M2 amplitude, consistent with the observations. Interestingly, the M4 amplitude first grows through the first 8 km of the Piscataqua River, then decays to very small value at Dover Pt., and then grows again in the upper reaches (last 3 km) of the Great Bay over the mudflats. The spatial evolution of the M4 tidal constituent is qualitatively similar to the observations but underestimates the magnitude by about a factor of 2 in the narrows of the lower Piscataqua River, and overestimates in the upper reaches of the Great Bay. Similar results are obtained if we include baroclinic or subtidal flows. We do not fully understand why this is occurring, but may arise from complexities in the bathymetry and sidewalls in this part of the estuary not well resolved in the model, or from viscous or turbulent effects assumed constant throughout the model domain. Moreover, it has been shown that locally high values of the M4 tide can occur near headlands as a result of the centrifugal component of the advection of M2 momentum (Parker 1984). Further investigation will need to address the role of bathymetric resolution and topography in the local generation of the M4 tide. The M6 tidal amplitude shows a steady increase throughout the estuary, leveling off (and even decaying near the Squamscott) over the final 3 km in the Great Bay. The M6 tidal constituent, driven primarily by frictional effects (Parker, 1991), appears to be well modeled throughout the estuary.
Figure 2.9: Modeled (lines) and observed (symbols) amplitude evolution of the M2 (top), M4 (2\textsuperscript{nd} from top), and M6 (3\textsuperscript{rd} from top) tidal constituents from Fort Point, near the mouth of the estuary, to the Great Bay. Amplitudes were determined with T\_TIDE analysis of 30+ day records (or for the 1975 data from the literature of which no error bars are available). Model results for a range of bottom roughness, $z_{ob}$, are indicated in the legend. The depth profile along the center channel is shown in the lower panel.
The phase evolution across the estuary is shown in Figure 2.10 (top panel) for the M2 tidal frequency at all observation stations where time series are available (Table 2.1). The modeled evolution of the P-U phase closely follows that of the observations. The P-U phase relationship in the first 12 km of the estuary is consistently about 45 degrees indicating a partially progressive and standing wave motion. However, 12 km upstream the P-U phase abruptly changes to +90 deg, consistent with a standing wave from Dover Pt. through the Great Bay Estuary. This change in P-U relationship is consistent with the observed tidal dissipation and relative phase change of the M2 tidal constituent (Figure 2.5).

Also shown in Figure 2.10 is the evolution of the growth of the M4 relative to the M2 constituent ($A_{ratio}$; Eq. 7). The modeled growth of the M4 harmonic increases through the first half of the lower Piscataqua River, decreasing at Dover Pt., and then increasing again through the upper reaches of the Great Bay (to about 8% of the M2 amplitude) where the depth shallows significantly over the mudflats. The evolution of the tide depends strongly on the water depth, consistent with a nonlinearly shoaling tidal wave. This spatial behavior is qualitatively consistent with the observations that show about twice as much harmonic growth as the model in the lower Piscataqua.

Also shown in Figure 2.10 is $\theta_{diff}$ (Eq. 8), an indication of the relative importance of the ebb and flood tide to the circulation (following Friedrichs and Aubrey, 1988). Although the model under-predicts the growth of the M4 constituent, the phase differential is qualitatively consistent with the observations. The lower reaches of the estuary in the Piscataqua River show ebb dominance, consistent with a stronger receding tide as the estuary drains. The Great Bay (beyond Adam’s Pt.), on the other hand, shows a strong flood dominance, indicating the flows into the bay and over the mudflats are greater than that produced by the ebb tide. This behavior is consistent with the evolution of the sea surface elevation skewness and asymmetry (Figure 2.10). The skewness shows similar trend to $A_{ratio}$ and $\theta_{diff}$, and is relatively low through the Piscataqua river, growing in the Little Bay and Great Bay suggesting a strong nonlinear evolution to the shoaling tide wave with asymmetrical form about the horizontal (along-channel) axis. The asymmetry increases in magnitude sharply in the Great Bay, indicating a pitched
Figure 2.10: Modeled (lines) and observed (symbols) along-channel evolution of the $P-U$ phase ($\text{deg}$; top panel), $A_{\text{ratio}}$ (2nd from top), $\theta_{\text{diff}}$ (3rd from top; showing flood and ebb dominance), normalized skewness (4th from top), and normalized asymmetry (5th from top) of 30 day sea surface elevation time series from the ocean to the upper reaches of the Great Bay. The nonlinear evolution of the tide is clearly evident with the sea surface profile evolving from a partially progressive nearly sinusoidal form and ebb dominance between Fort Pt. and Dover Pt., to a nearly standing wave with highly skewed and pitched-forward shape and flood dominance in the Great Bay. Model results for a range of bottom roughness, $z_0$, are indicated in the legend. The depth profile along the center channel is shown in the lower panel.
forward wave profile that has shorter duration but stronger flood currents and longer
duration but weaker ebb currents, consistent with the flood dominance estimated from
\( \theta_{diff} \).

2.5. Discussion

The tidal dissipation and phase evolution in the model is modified by the choice
of apparent bottom roughness, \( z_{ob} \). A range of values for \( z_{ob} \) were introduced in model
simulations and the best fit of the model tidal analysis to the observed M2 energy and
phase evolution used to determine the most appropriate value. Our best estimate, \( z_{ob} = 0.02 \ m \), is consistent with Swift and Brown’s (1983) estimates based on the 1975
observations. In their work, they find a range of frictional coefficients from 0.015 to
0.054. They also note that the dissipation was highest in regions where the flows were
larger, generally occurring in parts of the estuary where there are constrictions in the flow
owing to a narrowing of the river channel. Our model results show that ranges of \( z_{ob} \)
from 0.015 to 0.030 \( m \) give reasonable results throughout the estuary, and suggest that the
dissipation is well represented with a single value. This is somewhat surprising in that
the character of the seafloor (ranging from rocky and coarse sediments in the channels to
fine sands and muds on the flats) changes significantly over the estuary. On the other
hand, the flows also change similarly. That is, where the flows are highest, the more
rocky the bottom and more coarse the sediments (i.e., the fine material is washed away),
and where the flows are weak, the more fine-grained the sediments and the nature of the
bottom changes (i.e., with tidal channels cut through the mud and vegetation).

Model simulations that include and exclude subtidal forcing show that the tidal
dissipation (based on tidal analysis and considering only the M2 tidal constituent) does
not change significantly (Figure 2.11). This suggests that for the conditions examined
with subtidal amplitudes ranging 0.10-0.30 \( m \) over the 30-day model runs and
observation periods, the nonlinear interaction with the tides is weak. This also suggests
that tidal dissipation and phase change produced from the model simulations conducted
with 2015 forcing conditions can be compared with observations taken at other times (for example, from all the other experiments; Table 2.1).

The freshwater input to the Great Bay estuarine system is relatively small and during non-storm conditions contributes about 2% of the tidal prism (Short, 1992; NHDES, 2007). Baroclinic model simulations with average river discharge and average salinity and temperatures had a negligible effect on the tidal constituent amplitudes and phases, and can generally be ignored for the Great Bay when considering the tidal dynamics. However, comparisons of modeled time series and spectra with observations suggest that baroclinic flows are present. RMS velocity comparisons between barotropic and baroclinic model simulations away from the rivers but within the Great Bay are quite similar, and agree to within about 0.01-0.02 m/s. However, in the deep channel of the Little Bay where the flow field is high and has strong lateral shear, the baroclinic model velocities deviate from the barotropic velocities by about 0.05-0.10 m/s. Moreover, spectral comparisons show that, although the energetic tidal frequencies are not strongly affected, the high frequencies and the noise floor between the tidal harmonics increases for the baroclinic flows. This suggests that if higher frequency flows are of interest, then baroclinic models should be considered, but that tidal dynamics are well modeled with barotropic approximations.

In this work, we have not considered the effects of waves or winds on the tidal circulation and dissipation. In hindsight, this appears to be a reasonable assumption, at least for the conditions that occurred during the various field experiments. As noted by Wengrove, et al. (2015), wind-generated currents during a large storm can enhance the tidal flows when the winds are in the same direction as the current. Considering that the tides reverse every 12.4 hr in the Great Bay, this direct wind-driven flow might have an asymmetric effect on the overall current speeds and directions, sometimes in the direction of the flow and other times opposing or acting at an angle. In any case, the effect appears to be small even for the large wind event examined in Wengrove, et al. (2015), and does not likely change the overall character of the tidal currents owing to the order of magnitude difference between the wind-induced flows (of order 0.1 m/s) and the tides (of order 1-2 m/s). This may not be true closer to shore where the tidal flows are weaker and the wind-induced currents may be proportionally larger.
Figure 2.11. Modeled amplitude evolution for tidal only (solid lines) and tidal plus subtidal forcing (symbols) for the M2 tidal constituents from Fort Point, near the mouth of the estuary, to the Great Bay. Amplitudes were determined with T_TIDE analysis of 30+ day records. Model results for a range of bottom roughness, $z_{ob}$, are indicated in the legend. The depth profile along the center channel is shown in the lower panel.
The model-data comparisons show that the ROMS model reasonably well simulates the tidal dissipation and nonlinear evolution throughout the Great Bay Estuarine system. Ignoring baroclinic flow and subtidal oscillations does not strongly affect the tidal dynamics, at least for typical non-storm conditions for the Great Bay region. The model makes the hydrostatic approximation, and solves the RANS equations in three-dimensions following rectilinear horizontal grid and a vertical terrain-following $\sigma$ coordinate system. Many other models (such as ADCIRC, Westerink, et al., 1992; FVCOM, Chen, et al., 2003; Delft3D, Lesser, et al., 2004) also solve the same equations with similar approximations for rectilinear or unstructured grids and would likely also produce similar results. The good agreement between modeled and observed velocities across the estuary tidal channels and over the mud flats suggests that modeled currents from these fully nonlinear models would produce a good representation of the flow fields useful for sediment transport and nutrient flux studies (the subject of ongoing work).

2.6. Conclusions

A high-resolution three-dimensional hydrodynamic model (ROMS) was implemented for the Piscataqua River - Great Bay estuary using observed bathymetry and validated with several observational datasets spanning the estuary. The model was able to reproduce the observed tidal dissipation characteristics including dominant semidiurnal M2 tidal amplitude decay and phase changes, as well as the nonlinear growth of the M4 and M6 harmonics. The model underestimates the spatial evolution of the M4 magnitude by about a factor of 2 in the narrows of the lower Piscataqua River, and overestimates the values in the upper reaches of the Great Bay toward the Squamscott River. This could be due to complexities in the bathymetry and sidewalls in this part of the estuary not considered in the model, or from viscous or turbulent effects assumed constant throughout the model domain, and should be the topic of further investigation. The modeled behavior reproduces a highly dissipative, partially progressive wave in the lower 12 km of the Piscataqua River (with 45% tidal energy loss by Dover Pt., consistent with previous observational studies; Swift and Brown, 1983), and a (nearly) standing wave in the low dissipative region between Dover Pt. and the upper reaches of the Great Bay.
The spatial evolution from the mouth upstream in the estuary of the tidal harmonics, sea surface elevation skewness and asymmetry, and phase relationship between the along-channel velocity and sea surface time series, indicates a strong nonlinear tidal evolution consistent with an ebb dominant flow in the lower Piscataqua, and a flood dominant flow in the Great Bay. The good comparisons with observations suggest that the model well represents the nonlinear behavior of the tide, and accurately simulates the velocity and sea surface elevation time series throughout the estuary. Differences between model simulations with and without subtidal oscillations or river fluxes for the Great Bay are small, suggesting that interactions between the tide and other low frequency (subtidal) or baroclinic flows are weak and can be ignored when considering tidal dynamics.
CHAPTER 3

THE INFLUENCE OF SUBMERGED AQUATIC VEGETATION ON
MODELED ESTIMATES OF BED SHEAR STRESS AND NUTRIENT FLUXES
IN A TIDALLY DOMINANT ESTUARY

3.1 Abstract

Bed shear stress and contribution to the net nutrient flux by sediment sources in the New Hampshire Great Bay Estuary are estimated using the Regional Ocean Modeling System (ROMS), a three-dimensional numerical hydrodynamic model within the Coupled-Ocean-Atmosphere-Wave-Sediment Transport (COAWST) modeling system that incorporates wetting and drying over shallow mudflats that includes aquatic vegetation. The estuarine system is tidally dominated with mudflats and eelgrass meadows cut by several tidal channels and with over 50% of the surface area exposed at the lowest tides. In the Great Bay the spatial densities of eelgrass meadows ranged from 30-100% between 2011 and 2016 (Short, 2015). Model simulations for both 10 m and 30 m rectilinear grids were considered, with water depths determined from bathymetric datasets obtained in 2009, 2013 and 2015. The model is forced with tides offshore the mouth of the estuary, about 20 km downstream from Adam’s Pt., the entrance to the main lobe of Great Bay. Model estimates of bed shear stress compare well with field observations of near-bed velocity profiles obtained in 2011 (Wengrove, et al., 2015) and 2016 (Koetje, 2018) at two different shallow locations on the mud flats. Basin wide estimates of bed shear stresses are used to estimate total nutrient flux from sediments over month-long periods as well as over single spring and neap tidal cycles, both with and without the presence of eelgrass meadows. The presence of subaquatic vegetation was shown to decrease shear stress and lower bay nutrient flux estimates by 18% and 8.9% for meadows with 100% and 30% density, respectively. Total modeled nutrient loading over month-long periods from sediment sources was 3-20% of the average
estimated flux from river sources (Oczkowski, 2002) for higher discharge winter and spring periods, but much higher, about 52-60% during typically low discharge summer and fall periods, suggesting that nitrogen loading from sediments is a significant source to the estuary during half of the year. Simulated shear stress for 10 and 30 m grids were similar and had less than 10% impact on the monthly nutrient fluxes suggesting that coarser grids are adequate for this application.

3.2. Introduction

Estuaries are integral parts of coastal regions, providing a highly productive and critical habitat for many species, a valuable recreational and tourist destination, and are often the centers for transportation commercial trade activities. They are often described as traps (Odum, 1971, Biggs and Howell, 1984) or sinks, as well as buffers (Schubel and Kennedy, 1984) and exporters (Odum, 1971) for nutrients, sediments and pollutants to the world’s coastal and ocean waters. Healthy ecosystems require nutrients, primarily nitrogen and phosphorus, to foster plant growth; however, excess loading can lead to harmful algal growth and eutrophication (e.g., Cloern, 2001; Short and Wyllie-Echeverria, 2009). In recent decades, increased human interaction and modification of coastal wetlands and surrounding areas has dramatically altered transport pathways and increased the flow of nutrients into these systems (Nixon, 1995; NRC 2000). Coupled with sea level rise, and possible increased frequency and intensity of storm systems associated with global climate change (Emanuel, 2007; Emanuel, 2013; Knutson et al., 2013), increased terrestrial runoff and sediment resuspension will only exacerbate the problem.

Nutrient pollution and accompanying decline in water quality and clarity is one of the largest and longest standing problems facing coastal waters in the United States (Howarth et al., 2000), and continues to threaten economic and ecological sustainability. The presence of submerged aquatic vegetation (SAV) is a sensitive indicator of estuarine water and sediment quality and overall ecosystem health (Kenworthy and Haunert, 1991; Dennison, 1993; Bricker et al., 2008). Seagrass beds provide essential food for waterfowl and critical habitat for shellfish and finfish, net removal of nutrients from the water
column, and lower water turbidity by stabilizing bottom sediments through their root structures and promoting deposition of suspended sediment by damping currents. The greatest threat to SAV is decreased light attenuation (Dennison et al., 1993; Ralph et al., 2007) from high levels of suspended sediment due to increased loading from rivers or internal resuspension that severely restricts light penetration through the water column (Zimmerman et al., 1995; Lawson et al., 2007).

Quantifying nutrient budgets and loads is useful for determining impacts from point and nonpoint nutrient sources, providing a basis for comparing one estuary to another, and informs local practitioners that can lead to better management practices. Point source inputs of nutrients from wastewater treatment plants and rivers are relatively well understood and quantified (e.g., Loder and Gilbert, 1980; Jaworski, 1981; Nixon et al., 1982; Boynton and Kemp, 2000); however, nonpoint sources from groundwater seepage, surface runoff, precipitation and sediments are complex and difficult to measure, but often are much larger than point sources (Valiela et al., 1990; Howarth et al., 1996). In estuaries with large watersheds for example, nonpoint source loading of nitrogen and phosphorus is greater than wastewater inputs (NRC 1993a). Nutrient release from seafloor sediments and sediment resuspension has been shown to be important (Boynton and Kemp, 1985; Cowan et al., 1996); however, few studies have estimated the internal nutrient load from resuspended estuarine sediments (e.g., Percuoco et al., 2015; Wengrove et al., 2015).

Biogeochemical feedbacks in bottom sediments have been shown to increase the availability of nitrogen and phosphorus (Conley et al., 2007; Vahtera et al., 2007; Kemp et al., 2009). Nutrient flux from sediments occurs during sediment resuspension and subsequent release of interstitial pore waters. To accurately predict sediment transport processes, a good understanding of the bed shear stress that drives the sediment erosion, suspension and deposition is essential. Lack of quantifiable shear stress estimates over intertidal areas limits our overall estimation of net nutrient fluxes from sediments.

Wengrove, et al. (2015) used field measurements of flows in the bottom boundary layer to estimate the bed shear stress, and compared sediment motion thresholds to erosion chamber experiments in the Great Bay Estuary, New Hampshire, USA. Percuoco, et al. (2015) performed these erosion experiments on sediment cores collected from
several muddy locations in the Great Bay and determined that nutrients are released after a shear stress threshold is exceeded. Results from these studies indicate a correlation between nutrient release and increasing shear stress and thus a means to estimate nutrient loading in regimes of the bay that are always under water. However, at low stands of the tide almost 50% of the bay is (nearly) exposed as intertidal mudflats which are difficult to study owing to the temporally varying and spatially extensive region where the tides inundate.

To accurately estimate total loads across the bay, estimates of bed shear stress over tidal cycles from the water’s edge to the deep channels are required. Acquisition of field observations is not practical owing to spatially large mudflat areas and high spatial and temporal variability in currents, bed characteristics, and presence or absence of aquatic vegetation. To overcome this limitation, verified numerical hydrodynamic models are required. Previous models of the Great Bay Estuary have estimated shear stress (Ip et al., 1998; Erturk et al., 2002); however, model performance was somewhat limited by the available bathymetry and grid resolution, inability to include wetting and drying, and lack of verification with observational based estimates of shear stress. More recent models, such as the Regional Ocean Modeling System (ROMS; Haidvogel et al., 2008; Shchepetkin and McWilliams, 2005), includes a wetting and drying scheme (Warner et al., 2013) that simulates the rising and falling tide over the extensive mudflats and allows for flooding and draining across low-lying topographic areas. ROMS has been widely used to simulate flows, sediment transport, water quality, and morphological change, and recently formulations for the presence of vegetation and cohesive sediment have been incorporated (Beudin et al., 2017; Sherwood et al., 2018). ROMS has been implemented for the Great Bay Estuary, and verified with observations of currents and water levels spanning the estuary (Cook et al., 2019).

In this study we use the verified ROMS model (from Cook et al., 2019) to estimate the spatial and temporal distribution of shear stress across the Great Bay Estuary over 30 day periods with and without the presence of eelgrass meadows, for both 10 and 30 m grids. Modeled bed shear stress are compared with observations obtained in 2011 (Wengrove et al., 2015) and 2016 (Koetje, 2018). Spatial maps of the surficial mud fraction distribution across the bay (Wengrove et al., 2015), coupled with laboratory
estimates of the critical bed stress threshold and nutrient release (Percuoco et al., 2015), are combined with modeled bed shear stresses to produce bay-wide estimates of nutrient flux from sediment sources, with and without the presence of SAV, over monthly and typical neap and spring tidal cycles. Section 3.2 below describes the methodology behind the numerical model and observational approaches to estimating shear stress, including the nutrient flux estimates with and without eelgrass meadows. Results of model-data comparisons are presented in section 3.3 and discussed in section 3.4 in terms of nutrient fluxes and model limitations. Section 3.5 summarizes the findings.

3.3 Materials and methods

3.3.1 Study Site: Great Bay estuary, New Hampshire, USA

The Great Bay is a well-mixed, tidally dominated estuary located along the New Hampshire-Maine border (Figure 3.1). It is a recessed, drowned river valley characterized by deep (~10 m) tidal channels, large fringing mudflats, and is connected to the Gulf of Maine via the tidally energetic Piscataqua River. The surface area of the estuary is approximately 55 km² measured at mean high water (NHDES, 2007). The volume is approximately 156·10⁶ m³ and 235·10⁶ m³ for low and high tides, respectively, with a tidal prism of 79·10⁶ m³ calculated by Swift (1983). The seven major tributaries in this system are the Winnicut, Squamscott, Lamprey, Oyster, Bellamy, Cocheco, and Salmon Falls Rivers that flow into the Great Bay and Piscataqua River (Figure 3.1). Tidal excursion up these rivers is blocked by dams, which somewhat regulate the freshwater input into the system. Overall, the freshwater discharge is relatively small and only 2% of the tidal prism, shown by Short (1992) and confirmed by (NHDES, 2007). The tide range is 2-4 m over the spring-neap cycle with tidal currents up to 2 m/s in the channels at maximum ebb and flood tides (Cook et al., 2019). At low stands of the tide as much as 50% of the much Great Bay is exposed as low-lying mudflats, cut with deep tidal channels. There are significant areas of the mudflats that contain eelgrass meadows that change in their spatial extent annually and throughout each growing season (Short et al., 2006a, 2014). There are also areas of macroalgae within the Great Bay, but the spatial
Figure 3.1. The Great Bay Estuary and model bathymetry. Also shown are the major tributaries that flow into the system (inset) Location within the Gulf of Maine.
extent is significantly less than the eel grass meadows (Pe’eri et al., 2016) and are not considered herein.

The Great Bay estuarine system extends 25 km inland and defines the border between southern New Hampshire and Maine. It provides important economic, recreational and ecological benefits to this coastal region, and is often referred to as New Hampshire’s “hidden coast” (Figure 3.1). Because of its regional importance, it is designated as one of 28 estuaries of national significance by the U.S. Environmental Protection Agency’s (USEPA) National Estuary Program (NEP), established by Congress and authorized by the 1987 Clean Water Act. Through this program, the Piscataqua Region Estuaries Partnership (PREP) was established to preserve the health of estuarine resources in the region in part through regular assessments and monitoring programs. The 2018 PREP report outlined the need for expanded monitoring and additional research to address knowledge gaps in long-term nutrient loading to the Great Bay (PREP 2013, 2018), particularly nitrogen. Increases in population density and associated anthropogenic impacts have altered the estuarine productivity, resulting in increased nutrient loading and amplified suspended sediment that reduces water quality (PREP 2013, 2018). Currently the PREP sampling program includes nitrogen-loading estimates from tributaries, wastewater treatment plants, and atmospheric deposition but does not estimate nutrient loadings from sediments (PREP 2013, 2018).

### 3.3.2 Hydrodynamic Model

The Regional Ocean Modeling System (ROMS, version 3.7) is a community developed numerical hydrodynamic model that solves the Reynolds-averaged Navier Stokes (RANS) equations using finite difference approximations, the hydrostatic and Boussinesq assumptions and a split explicit time stepping algorithm using depth integrated momentum equations (Shchepetkin and McWilliams, 2005; Haidvogel et al., 2008). ROMS was implemented within the Coupled Ocean Atmosphere Wave and Sediment Transport (COAWST, version 3.3.1, r1270) numerical modeling system (Warner et al., 2010), to take advantage of the recently incorporated wave-flow-vegetation module.
The momentum, scalar advection and diffusive processes are characterized using transport equations. A third order upwind advection scheme is used to solve for horizontal advection. A centered-fourth order advection scheme is used to solve for vertical advection. In order to resolve sub-grid scale turbulent processes for the vertical mixing of momentum and mass, an eddy viscosity \( K_M \) is parameterized using a two equation, generic length scale (GLS) method with the \((k-\varepsilon)\) turbulence closure model (Umlauf and Burchard, 2003; Warner, et al., 2005) in conjunction with the Kantha and Clayson (1994) stability function. A summary the GLS method and numerical implementation in ROMS can be found in Warner et al. (2005). The wetting and drying algorithm within ROMS (Warner, et al., 2013) is utilized to represent the flooding and drying of the tide over shallow areas with a critical depth \( D_{crit} \) is set to 10 cm. Model verification and implementation for the Great Bay Estuary is discussed in detail in Cook et al. (2019).

### 3.3.2.1 The flow-vegetation module

The vegetation module (Beudin et al., 2017) incorporates a spatially averaged drag force that is approximated using a quadratic drag law,

\[
F_{d,\text{veg},u} = \frac{1}{2} \rho C_D b_v n_v u \sqrt{u^2 + v^2}
\]

\[
F_{d,\text{veg},v} = \frac{1}{2} \rho C_D b_v n_v v \sqrt{u^2 + v^2}
\]

where \( \rho \) is the density of seawater, \( C_D \) is the plant drag coefficient, \( b_v \) is the width of the individual plants, \( n_v \) is the number of plants per unit area, and \( u \) and \( v \) are the horizontal components of velocity at each vertical level in the canopy height, \( l_v \). A limiter is imposed to prevent large vegetation drag forces from reversing the flow (Warner et al., 2013).

The turbulent production due to vegetation (Uittenbogaard, 2003) is defined as

\[
P_{\text{veg}} = \sqrt{\left(F_{d,\text{veg},u} \cdot u\right)^2 + \left(F_{d,\text{veg},u} \cdot v\right)^2}
\]
and dissipation due to vegetation is

\[ D_{\text{veg}} = c_2 \frac{P_{\text{veg}}}{\tau_{\text{eff}}} \]  

(4)

where \( c_2 \) is a coefficient from the GLS turbulence model (Warner et al., 2005) and \( \tau_{\text{eff}} \) is the minimum between the dissipation time scale of free turbulence and dissipation time scale of eddies between plants. The bottom stress is calculated assuming a logarithmic profile in the bottom cell (Warner et al., 2008a, 2008b). The dissipation due to vegetation acts as an added dissipation term in the 2-equation turbulence model.

### 3.3.2.2 Model Domain and Grid Development

The model domain ranges 22.02 by 25.02 km and is rotated 37 deg CCW from true north to align the offshore boundary with the approximate orientation of the shoreline along the New Hampshire-Maine coast. The grid is a rectilinear Arakawa C grid, with a constant horizontal grid spacing of 10 m and 30 m (Figure 3.2). The grid is discretized into 8 vertical layers in a stretched terrain-following (\( \sigma \)) coordinate system adjusted for higher resolution near the surface and bottom boundaries. Details on grid development can be found in Cook et al. (2019).

### 3.3.2.3 Boundary Conditions

The bottom boundary condition for momentum is parameterized with a simple drag coefficient, assuming a logarithmic vertical velocity profile in the bottom vertical cell (discussed later). The drag coefficient is represented by

\[ C_D = (\kappa/\ln(z/z_o))^2 \]  

(5)

where \( \kappa = 0.41 \) is the von Karman coefficient (Kundu, 1990), \( z \) is the elevation above the bottom (the vertical mid-point of the bottom cell) and \( z_o \) is a characteristic bottom roughness (in m). A bottom roughness of 0.02 m was found to give consistent comparisons between observed and modeled tidal flows and water levels for the entire estuary (Cook et al., 2019). The equation for the kinematic bottom stress is given by
Figure 3.2. Model Domain and observational locations (30 m grid). (left) Grid lines are decimated by a factor of 9, and do not reflect the actual grid size. Insets (right) show details of the bathymetry near the location of instruments. Green and red dots indicate the instrument locations from 2011 and 2016, respectively.
\[ \tau_{b, log, x} = -\frac{1}{2} C_D \cdot u \sqrt{u^2 + v^2} \quad (6) \]
\[ \tau_{b, log, y} = -\frac{1}{2} C_D \cdot v \sqrt{u^2 + v^2} \quad (7) \]

for the \( x \) and \( y \) directions, respectively corresponding to the \( \xi - \eta \) coordinate system in ROMS.

The east, west, and northern lateral boundary conditions were closed, and the southern edge was defined by an open boundary condition for sea surface height, barotropic and baroclinic velocities, and tracers. To account for the tides at the open boundary, an implicit Chapman condition (Chapman, 1985) was applied to sea surface height in conjunction with a Flather condition (Flather, 1976) for 2D momentum. These conditions assume all outgoing signals leave at the shallow water wave speed. In order to eliminate reflections at the boundary, deviations of the normal component of velocity from the exterior value are allowed to radiate out of the domain at the speed of external gravity waves. 3D momentum and tracers were set to radiation and gradient boundary conditions, respectively (Marchesiello et al., 2001).

### 3.3.2.4 Model Initialization and Forcing

The 10 and 30 \( m \) grid models are forced with three semidiurnal (M2, S2, N2) and two diurnal (O1, K1) tidal constituents using the Oregon State University Tidal Prediction Software package (OTPS) with the United States East Coast Regional Tidal Solution (EC2010; Egbert and Erofeeva, 2002). All model runs were ramped with a hyperbolic tangent to full amplitude over two days. Temperature and salinity were held constant at 17 \( \text{deg C} \) and 30 \( \text{psu} \) for all runs. The time steps for the 10 \( m \) and 30 \( m \) were set to 1 \( s \) and 1.5 \( s \), respectively. The model was run for 10 days coinciding with a field deployment in July 2016 (see Section 3.3.3.2). River discharges were not included in these runs, as both field measurement campaigns (see Section 3.3.3) were conducted during times of low river input.
3.3.3 Field Measurements

Two field data sets were used to compare modeled velocities and bed shear stresses with observations. The first was obtained in the summer of 2011 by Wengrove et al. (2015); the second in the fall 2016 by Koetje (2018). Details of the experiments can be found in therein, and are briefly summarized below.

3.3.3.1 2011 Field Deployment (Wengrove et al., 2015)

The sampling array was located in the shallow western portion of the main channel of the Great Bay, just north of Adam’s Point (Figure 3.2) in water depths ranging 1.3 to 3.8 m. The instrument array (see Table 3.1) included a single point acoustic Doppler Velocimeter (Nortek Vector ADV) that sampled three components of velocity \(u, v, w\) at 64 Hz, in a single 2 cm bin 80 cm above the bed. A high resolution profiling ADV (Nortek Vectrino Profiler) sampled the horizontal \(u, v\) velocity profiles at 64 Hz with 1 mm resolution over a 3 cm range intersecting the bed. An acoustic Doppler current profiler (2 MHz Nortek Aquadopp High Resolution ADCP) was used to obtain high-resolution velocity estimates at 1 Hz over a 1.2 m profile intersecting the bed.

The 14-20 June 2011 deployment was conducted under neap tide conditions with typical low river discharge rates and light winds (1-4 m/s). During neap-tide conditions the viscous sublayer was potentially visible (Wengrove and Foster, 2014), and the friction velocity, \(u_\ast\), is calculated using the viscous and the log-layer methods (indicator function; described later in section 2.4). The 6 hr flood phase of 20 June 2011 was used for the study when the bed elevation was found to be undisturbed.

Enhanced flows were observed during spring tide conditions (27 August – 2 September 2011) with elevated winds (6-8 m/s) and precipitation, with increased river discharge and terrestrial runoff. For these data \(u_\ast\) is calculated using the log-layer method (indicator function; describe later in section 2.4). During the deployment eelgrass was observed on the instrument array, limiting the amount of usable data. Hourly averaged velocity profiles for the alongshore and across channel directions of flow were calculated and the direction in relation to the flood was also evaluated. Change in bed elevation was found to be small, about 1 mm.
Flooding flow (along channel; southward directed at the site) was found to be unidirectional and parallel to the channel. However, ebbing flows were found to be sheltered from a small peninsula just south of the sampling site, creating flows away from the mudflats and toward the deeper channel. During the ebbing tide the observed velocities show little change, most likely owing to local topographic effects. Thus, results from this deployment are focused on the flood phase of the tide. During the rising tide the maximum velocity magnitudes occurred at about mid-tide; ranging 0.28 m/s and 0.35 m/s. In both cases, surface gravity wave energy was small and velocities were dominated by tidal currents.

3.3.3.2 2016 Field Deployment (Koetje, 2018)

During this field experiment the sampling array was located on a mudflat south of Adams Pt. in the Great Bay with tidally modulated water depths ranging 0.3 to 2.8 m and located horizontally about 30 m from ~2 m deep tidal channel (Fig. 3.2). The instrument array (see Table 3.1) included the same (as in 2011) downward facing ADCP, and was used again to sample horizontal \((u, v)\) velocity profiles continuously at 1 Hz over a 1 m range intersecting the bed with profiling bins at 4 cm resolution. The same (as in 2011) single point ADV sampled \((u, v, w)\) velocities at 64 Hz, 50 cm above the bed, and was used to validate flow velocities measured by the ADCP.

Flow velocity components were reported in Koetje (2018) in east-west and north-south directions, as well as velocity magnitude and direction over a four-day deployment beginning 7 July 2016. Velocities observed 40 cm above the bed were ensemble averaged over 20 min intervals with velocity outliers removed using three passes of a 3 standard deviation filter applied to all data that exceeded a 90% correlation threshold. Pressure data at the site were unavailable, and water levels were approximated using data obtained from the NOAA Tidal Station (ID 8423898) located at the mouth of the estuary near Fort Point, NH, and adjusted for tidal energy dissipation and travel time to the site to determine the flood/ebb tide (Koetje, 2018). East-west directed flows were dominant, with flooding currents stronger than the ebb and approximately steady for several hours.
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<td>July 2016</td>
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<td>1 m profile within water column</td>
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<td>0.02 m bin resolution</td>
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<td>6-6.5 cm (1 mm) bin resolution</td>
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* only deployed during the August and September 2011 deployments
3.3.4 Bed shear stress

The method for estimating the shear stress is dependent on the hydrodynamics of the system. Since bed shear stress were not measured directly, proxy methods are used to estimate shear stress in tidal boundary layers (Gross and Nowell, 1983, Kim et al., 2010).

Shear stress estimates during neap tide conditions of the June 2011 deployment were calculated assuming smooth turbulent flow with a viscous sublayer (Wengrove and Foster, 2014) and no sediment resuspension. In these conditions the shear stress is equal to the viscous stress, and estimated by

\[ \tau_{b,viscous} = \mu \frac{\partial u}{\partial z} \]  

(8)

where \( \mu \) is the dynamic viscosity and the velocity \( u \) taken in the dominant flow direction. In the viscous sublayer the bed stress is the result of molecular momentum transfer (Boudreau and Jorgensen, 2011). The viscous stress was estimated using data from the profiling ADV (Wengrove and Foster, 2014). The typical estimate for the critical shear stress for incipient motion of sediment of 0.1 N/m² (Wengrove et al., 2015) was not exceeded during this deployment, and the bed elevation did not change. The numerical model simulations are incapable of resolving the viscous sublayer at the grid resolution, and therefore direct comparison with observed shear estimates is done using the log-layer approach described next.

Shear stress estimates during spring tide conditions of the August-September 2011 deployments, and during the July 2016 study were calculated using a log-layer approach. The boundary conditions during these studies were assumed to be rough turbulent and therefore the viscous forces were assumed negligible (Wengrove et al., 2015; Koetje, 2018). Two methods were used to estimate shear stress. In the first, a quadratic drag law was assumed and given by

\[ \tau_{b,quad} = \rho C_D \bar{U}^2 \]  

(9)

where \( \rho \) is the density of the fluid, \( C_d \) is a drag coefficient, and \( \bar{U} \) is the free stream velocity (Soulsby, 1997). The free stream velocity was taken as the 20 min average.
horizontal velocity measured by the ADV 70 cm above the bed. The drag coefficient was set to 0.001, using an average of Dawson-Johns, Soulsby, full depth log profile, and Cole-brook White methods (Soulsby, 1997). This method tends to be used when observations near the boundary are unavailable. This approach is limited in that the estimation is based on a single point measurement of velocity and the drag coefficient is based on empirical approaches.

The second approach assumes a logarithmic velocity profile from the free stream to the bottom boundary,

\[ u(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right) \]  

(10)

where the velocity is taken in the direction of dominant flow (Tennekes and Lumley, 1972). The velocity profile is determined with the profiling ADCP. The shear velocity is estimated using the indicator function by taking the derivative of (10) with respect to \( z \) (Orlu et al., 2010), given by

\[ u_* = \frac{\partial u}{\partial z} \kappa. \]

(11)

The shear stress, \( \tau_{b,\log} \), is then determined assuming a quadratic bed shear stress,

\[ \tau_{b,\log} = \rho u_*^2. \]  

(12)

The log-layer method provides two estimates of bed stress using the profiling instruments (given in Table 3.1). This method has been shown to provide reasonable estimates of shear velocity (Tennekes and Lumley, 1972; Orlu et al., 2010; Wengrove and Foster, 2014).

### 3.3.5 Sediment mud fraction distribution

To estimate nutrient fluxes from sediments, Wengrove et al. (2015) assumed that nutrients were released only in regions with at least 50% surficial mud fraction. Regions of the Great Bay with at least 50% mud fraction were estimated using observed grain size distributions obtained from both sediment grab samples and a logarithmic depth model.
based on comparison of water depths and historical sediment data from the region (Lippmann, 2013; Wengrove et al., 2015). The same data is used herein to compare with estimates obtained by Wengrove et al. (2015).

### 3.3.6 Eelgrass Distribution and Model Input

Trends in eelgrass (*Zostera marina*) distribution in the Great Bay have been aerially surveyed and mapped yearly since 1986 (Short and Burdick, 1996). The eelgrass distribution map for 2016 (Figure 3.3) was created from aerial photography collected on 5 August 2016 (Barker, 2017) and downloaded as an ESRI shape file from the New Hampshire Geographically Referenced Analysis and Information Transfer System (UNH GRANIT) Data Recovery Tool (http://www.granit.unh.edu/data/search). Ground-truth observations are made at low tide and assess 10-20% of the aerially mapped eelgrass. In 2007, a global monitoring program designed to document seagrass habitat change (SeagrassNet; Short et al., 2006a, 2014) set up a permanent site in the Great Bay. Quarterly sampling parameters include photographic records, percent cover, canopy height, biomass, shoot density, and flowering shoots (sexual production). In 2016 sampling was conducted on 26-27 April, 4-6 July, and 24-26 October. Eelgrass meadows in New Hampshire and southern Maine typically develop maximum canopy height in the late summer and fall (Gaeckle and Short, 2003) and greatest shoot density in July (Ochieng et al., 2010). However, results for July revealed mean percent cover at approximately 30%, much lower than the nearly 100% reported in previous years (going back to 2007; Short 2016).

The vegetation model inputs follow those in Beudin et al., 2017, which designed parameters to resemble eelgrass (*Zostera marina*). The modeled plants are 30 cm high, 0.3 cm wide, 0.3 mm thick (corresponding to flexible stems), mass density of 700 kg/m$^3$, elastic modulus to 1 GPa (Luhar and Nepf, 2011), and drag coefficient, $C_d$, in the flow model is set to 1. To test the sensitivity of model flows and shear stress estimates to canopy density, the plant density is set to a dense canopy of 2500 stems/m$^2$ (Ghisalberti and Nepf, 2004; Nepf 2012), and 30% of that value to mimic conditions in July 2016. The vegetation patches are ramped at the edges from no vegetation to test case canopy
density to avoid numerical perturbations of the corners of a vegetation patch (Beudin et al., 2017).

### 3.3.7 Nutrient Load Calculation

The calculation of bay wide nutrient release is based on the same approach as Wengrove, et al. (2015), using a value $0.35 \text{ N/m}^2$ as the shear stress threshold for nutrient release obtained from Eromes chamber laboratory experiments on undisturbed sediment cores taken from the field (Percuoco et al., 2015). In their estimates, average nutrient release occurs when bed shear stresses exceed the threshold value for at least 15 minutes in regions with at least 50% surficial mud fraction. Using the same approach as Wengrove et al. (2015), nutrient release can be estimated as

$$\text{Nutrient Release (kg)} = \frac{N_{SS}}{1000} A_{sed} \frac{AW}{1000} \quad (13)$$

where $N_{SS}$ is the average nutrient release (Eromes chamber) for all shear stress up to and including the shear stress threshold of $0.35 \text{ N/m}^2$ in this study (Percuoco, 2015), $A_{sed}$ is the area with 50% mud fraction in the surface seafloor sediments of the Great Bay ($\sim 12 \text{ km}^2$), and $AW$ is the atomic weight of nitrogen or phosphorus (in g/mol).

### 3.4. Results

Model-observation comparisons are presented with and without subaquatic vegetation (eelgrass) incorporated into the model. The distribution of depth averaged velocity and bed shear stress are shown for different stages of the tide. A time series of the modeled depth averaged velocity and shear stress is compared with observations from 2011 and 2016 field deployments.
Figure 3.3: Eelgrass coverage from 2016 (Barker, 2017). Inset shows close up of 2016 instrument deployment relative to eelgrass meadows.
3.4.1 Distribution of bed shear stress

Examples of the modeled spatial distribution of depth-averaged velocities and bed shear stress are shown in Figure 3.4 for four discrete stages of the tide: low, rising (flood), high, and falling (ebb) tides. The flooding tide corresponds to 1445 hr on 8 July 2016, coinciding with the 2016 field deployment. At low tide (Figure 3.4, left column), currents are not slack and strong bed stress occurs in the channels and localized areas of the mudflats. At flooding tides (Figure 3.4, middle left column), shear stresses are strongest in the channels and extend to the edges of the mudflats. At high tide (Figure 3.4, middle right column), also close to slack tide, shear stresses are minimum across the bay. On ebbing tides (Figure 3.4 right column), high shear stresses are found mostly in the deeper channels and away from the mudflats. In general, bed shear stress tends to be strongest in the channels where the currents are strongest and the bottom substrate is mostly composed of coarse sand and gravel (Wengrove et al., 2015). Shear stress tends to be lower on the mudflats, yet high enough to exceed the critical threshold for sediment incipient motion (0.1 N/m²), and where the bottom substrate is composed of finer muddy sediment (and in some areas aquatic vegetation).

3.4.2 Time series comparison

Time series of modeled sea surface height, depth averaged velocity, and shear stress are shown in Figure 3.5 corresponding to the 2011 field data, and in Figures 3.6-3.7 corresponding to the 2016 field deployment, each discussed below. Model simulations are performed for 30 days using forcing conditions from 2016 for both 10 and 30 m grids. Model-data comparisons are performed with the 10 and 30 m model grid point closest to the instrument locations in each respective field deployment.

2011 Model-Data Comparison

Although the model was run with hydrodynamic forcing conditions from 4-13 July 2016, the tidal cycles are very nearly the same and we can qualitatively compare with observations obtained during the neap (June 2011) and spring (August 2011; September 2011) periods. The modeled neap conditions were chosen to correspond to the
neap-like tidal cycle of 13 July 2016 when the sea surface elevation at the site closely matched the observed water level time series (top left panel of Figure 3.5) and the modeled depth-averaged velocities from the 30 m grid closely matched the observations (middle left panel of Figure 3.5). The modeled spring tide conditions were chosen to correspond to two tidal cycles during the more energetic spring-like conditions of 6 July 2016, with modeled sea surface elevation and 30 m depth-averaged velocities closely matching the observations (top right and middle right panels, respectively, of Figure 3.5).

Observational based estimates of shear stress using the three methods (viscous, quadratic, and logarithmic) described in Section 3.3.4 are shown in the bottom panels of Figure 3.5. The best estimate (average of the various methods) of bed shear stress according to Wengrove et al. (2015) is shown with the solid black dots and all estimates and methods are reported in Wengrove (2012) and shown on Figure 3.5 (and tabulated in Appendix A3.1 at the end of this chapter). Overall the simulations from the 30 m grid well reproduces the magnitude of the shear stress corresponding to both the neap and spring tide conditions. Root-mean-square errors (RMSE) between model estimates and observations for the June, August, and September conditions are 0.0525, 0.0345, and 0.0874 N/m², respectively. Shear stress estimates from the 10 m resolution model were all below 0.1 N/m², including the 2 grid cells adjacent to the instrument location. The 10 m grid results are underestimating the shear stress compared to the 30 m grid in the same locations. A possible explanation for this are boundary effects from the coast and eddy formation from the interaction with the tidal currents and the coastline aren’t resolved in the 30 m grid, but those effects are modeled in the 10 m grid.

2016 Model-Data Comparison

The model simulations shown in Figures 3.6 and 3.7 were forced with tidal conditions concurrent to the July 2016 field experiment. Observational based estimates of shear stress for this experiment are described in Section 3.3.3. Observed depth-averaged velocities and shear stresses are compared with model estimates in Figure 3.6 for the 30 m grid with 0, 30, and 100% vegetation density. Model simulations show that the presence of eelgrass vegetation retards the flow, particularly during ebb tidal phases.
Figure 3.4: Example model results from the 30 m grid showing the spatial variability in velocity (upper panels) and shear stress (lower panels) for low (A), rising (B), high (C), and ebbing (D) tide.
Figure 3.5: Sea surface height (upper panels), depth-averaged velocity (middle panels) from the 30 m (solid line) and 10 m (dashed line) model runs and bed shear stress comparisons with observations (symbols indicated in the legend) from 2011. Left panels show the spring tide case; right panels show the neap tide case. Open circles in the upper two panels indicate the tide stage and modeled currents during the simulations and observations.
when the model compares quite well with the observations when characteristics and presence of the vegetation are included. However, the flood tide remains under-predicted possibly owing to channelization at a scale smaller than is resolved with the coarse grid. Comparisons between simulations from the higher resolution 10 and 30 m grids show little variation (Figure 3.7), indicating that the finer scale 10 m grid at this particular location does not provide a better estimate of flow or shear stress. RMSE values are calculated for both the flood and ebb tide, to account for the tide asymmetry. Flood and RMSE values between the model and observations for the 30 m grid, 100% dense vegetation, 30% vegetation, and 10 m resolution are summarized in Table 3.2.

<table>
<thead>
<tr>
<th>Model Run</th>
<th>June 2011 RMSE (N/m²) (no. obs)</th>
<th>July 2016 RMSE (N/m²) (no. obs)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Calm</td>
<td>Storm</td>
</tr>
<tr>
<td>30-m grid</td>
<td>0.0525</td>
<td>0.0345</td>
</tr>
<tr>
<td>Dense vegetation (30-m grid)</td>
<td>0.0521</td>
<td>0.0333</td>
</tr>
<tr>
<td>30% vegetation (30-m grid)</td>
<td>0.0524</td>
<td>0.0325</td>
</tr>
<tr>
<td>10-m grid</td>
<td>0.0314*</td>
<td>0.164</td>
</tr>
</tbody>
</table>

* model used logarithmic log law, while observations used viscous stress
Figure 3.6: Sea surface height (upper panel), depth-averaged velocity (middle panel) and bed shear stress (lower panel) for model (lines) and observations (symbols) from 2016. Simulations are performed on the 30 m grid with and without eelgrass vegetation (symbols indicated in the legend).
Figure 3.7: Sea surface height (upper panel), depth-averaged velocity (middle panel) and bed shear stress (lower panel) comparisons between model (lines) and observations (symbols). Simulations are performed for the 10 m and 30 m grid without eelgrass vegetation.
3.5. Discussion

3.5.1 Validation of modeled estimates of bed shear stress

The study presents an estimate of the spatial and temporal distribution of bed shear stress for a tidally dominant estuary using a verified hydrodynamic model. Time series model output are compared with observational based estimates at two locations in the Great Bay Estuary. The model well represents the estimates of shear stress at both sites, although under-predicts the magnitude of the flooding-tide depth-averaged velocity (Figures 3.6 and 3.7). There is a clear flood-ebb asymmetry at the 2016 field site and including the effects of vegetation clearly improves the estimates of both depth averaged velocity and shear stress on the ebb tide, but the results for the flood tide are less clear. There is also asymmetry in the flood-ebb tides at the 2011 field site (Figure 3.5), which we believe are due to the sheltering effects of Adam’s point (Figure 3.2) as pointed out in Wengrove et al. (2015). The modeled velocity used for shear stress estimates is from the center of the bottom vertical bin (Warner et al. 2008a, 2008b), and not the depth-averaged velocity shown in Figures 3.6 and 3.7.

Koetje (2018) shows that a transitional log layer is present in the vertical profiles from the 2016 data. Here we compare stress estimates with the “upper” log layer, as the model does not resolve the dynamics very close to the bed. There is a difference in the depth-averaged velocity during the flooding tide, on average about 5 cm/s (similar in magnitude found by Cook et al., 2019). This is seen in all the model runs including the vegetation and 10 m higher-resolution run (Figures 3.6 and 3.7). This difference might be due to channelized flow at a sub-grid length scale (less than 10 m). The results are only marginally better by reducing the grid resolution from 30 to 10 m suggesting that the model is able to capture the bulk flow and patterns of shear stress using the coarse, 30 m grid.

This is a convenient result for future studies requiring accurate shear stress estimates (e.g., sediment transport studies), as the computational savings and data generated is greatly reduced. Currently the 30 m grid is performed on Trillian, a Cray XE6m-200 supercomputer at the Institute of Earth Oceans and Space, University of New
Hampshire, using 20 nodes with 32 processors and takes ~39 hours to complete a 10 day run, and generates ~500 GB of output data. The 10 m grid is run on the Blue Waters Cray XE6 supercomputer at the University of Illinois using 150 nodes with 32 processors and takes ~44 hours to complete a 10-day run, and generates ~2 TB of output data. These are relative comparisons, but demonstrate the practical computational savings of using the coarse grid.

3.5.2 Effects of vegetation on distribution of shear stress

The flow-vegetation module is a relatively recent addition to the COAWST modeling system with several applications to estuarine and coastal environments (Donatelli et al., 2018). The flow-vegetation model has been validated for two locations within the Great Bay, one at the entrance to the Great Bay proper near the main channel (Figure 3.5) and a second adjacent to an eelgrass bed (Figures 3.6 and 3.7). The spatial distribution of depth averaged velocity (Figure 3.8) and shear stress (Figure 3.9) is shown across the Great Bay for a typical flood and ebb tide. Including vegetation in the model, particularly in the mudflat regions of the Great Bay clearly changes the spatial patterns of the flow around the eelgrass beds (Figure 3.8, panels B and C). Unsurprisingly, the dense vegetation (Figure 3.8; panel B) affects the flow more than the sparse case (Figure 3.8; panel C). However the sparse case better matched the conditions of the 2016 eelgrass meadow (Short 2016; Barker, 2017), suggesting that vegetation density is important for accurate estimates of the spatial distribution of shear stress.

The presence of vegetation also has clear effect on the spatial distribution of bed shear stress (Figure 3.9). Depth averaged velocities and bed shear stress is decreased within and around the vegetated areas, although the sparse vegetated case shows increased shear stress in areas directly adjacent to the eelgrass meadows (Figure 3.9, Panels B and C).

One model limitation is that the total shear stress does not include the contribution of canopy shear stress and wake generated turbulence from eelgrass meadows to the water column. Also, the current assumption of a logarithmic boundary layer for calculating shear stress at the sediment bed within the vegetation canopy (Beudin et al.,
2017) might not appropriately capture the boundary layer dynamics. Lastly, in the present modeling formulation, there is a lack of enhanced bed shear stress and sandification that is observed in sparsely vegetated canopies (Van Katwijk, et al., 2010; Nepf 2012), like the sparse (30% coverage) case considered here. Although this study does not incorporate sediment transport directly, understanding these limitations will help guide future research.

3.5.3 Estimates of nutrient loads from sediment resuspension

Estimating nutrient loads from sediment resuspension is very complex and generally not well understood (Couciero et al., 2013). The primary source of information is derived from in situ or laboratory based Eromes chamber experiments (Kalnejais et al., 2010; Couceiro et al., 2013; Kleeburg and Herzog, 2012). Nutrient release data obtained in the summer 2011 Eromes erosion chamber experiments is used in conjunction with model output to make estimates of nutrient loads from sediment resuspension in the Great Bay estuary (Eq. 13). The model output is time averaged over 15 minutes, and any values in areas where the mud fraction is greater than 50% and the shear stress is greater than or equal to the nutrient release threshold of 0.35 N/m² are considered to have released 1.3 mmol/m² and 0.21 mmol/m² of dissolved inorganic nitrogen (DIN) and phosphate, respectively. Estimates of nutrient loads from sediments are made for a month (30 days) and an event (1 average, neap, or spring tidal cycle). The month-long simulation is compared with estimates from riverine inputs (Oczkowski, 2002) and the single tidal cycles are compared with an estimate of nutrient loading from Wengrove et al. (2015) with results summarized in Table 3.3.

Our results suggest that nutrient loading from sediment resuspension is an important internal and nonpoint source to the overlying water column and across the entire estuary. During higher discharge winter and spring periods, nitrogen fluxes from sediment sources are about 3-20% of the river inputs. On the other hand, during low flow summer and fall months, the sediments contribute about 52-60% of the river inputs, a significant contribution to the net flux. The role of vegetation clearly lowers these estimates, yet only by 9-18% for sparse (30%) and dense (100%) canopies.
Figure 3.8: Depth averaged velocity on the flooding and ebbing tide for the western portion of the Great Bay and inclusive of the 2016 field site (white triangle) for the 30 m grid with (A), dense vegetation (B), 30% vegetation (C), and 10 m (D) model runs (no vegetation).
Figure 3.9: Bed shear stress on the flooding and ebbing tide for the western portion of the Great Bay and inclusive of the 2016 field site (white triangle) for the 30-meter grid (A), dense vegetation (B), 30% vegetation (C), and 10-meter (D) model runs (no vegetation).
Between 2012-2016, nonpoint source loads accounted for 67% of the nitrogen load to the Great Bay (PREP, 2017). These estimates include nonpoint sources in watersheds, groundwater discharge, and atmospheric deposition. These methodologies do not consider the internal cycling of nutrients from sediment resuspension, and therefore the contribution of nonpoint sources to the overall nutrient load to the estuary is underestimated. Although there is large variability in river flow and surface runoff, and concurrent variability in estimated flux of nutrients, tidal forcing and associated currents are consistent. Our results suggest that even an average tide contributes about 25 kg DIN across the estuary per tidal cycle and about 700 kg per month.

Estimates of phosphate loads are more complex than dissolved inorganic nitrogen in these environments. Depending on the presence of phosphorus-sorptive molecules like iron (III) hydroxides and oxides and calcium carbonate material, estuarine sediments may or may not retain phosphorus (Sundby et al., 1986; Howarth, 1995). Percuoco et al. (2015) found that spatial variability in the solid phase iron in the surface sediments of two cores may have contributed to the suppression of phosphate fluxes. They concluded that phosphate release could be influenced by shear stress duration and the concentration of suspended material. As a result, our estimates of phosphorus loading are likely to well over-predict the actual flux, but could be improved with better understanding of the geochemical processes at the sediment-water interface.

3.6. Conclusions

Sediment transport formulations rely on accurate estimates of bed shear stress; however, this parameter is not directly measurable. Estimation methods are dependent on hydrodynamic conditions present and rely on observations of velocity close to the sediment bed, which are difficult to obtain in shallow coastal environments like the Great Bay. In this study we employ a validated (Cook et al., 2019) three-dimensional hydrodynamic model with a vegetation module to study the effects of vegetation on the distribution of shear stress across the estuary. The model results in this study compare
### Table 3.3: Nutrient Load Estimates

**Monthly nutrient loads**

<table>
<thead>
<tr>
<th>Source</th>
<th>Dissolved Inorganic Nitrogen load (kg/month)</th>
<th>Phosphate load* (kg/month)</th>
<th>Data</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rivers</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Winter (Dec-Feb)(^A)</td>
<td>3,700</td>
<td>92</td>
<td>2000-2001</td>
</tr>
<tr>
<td>Spring (Mar-May)(^A)</td>
<td>17,000</td>
<td>720</td>
<td>2000-2001</td>
</tr>
<tr>
<td>Summer (June-Aug)(^A)</td>
<td>1,300</td>
<td>120</td>
<td>2000-2001</td>
</tr>
<tr>
<td>Fall (Sep-Nov)(^A)</td>
<td>1,200</td>
<td>70</td>
<td>2000-2001</td>
</tr>
<tr>
<td>Sediments – Tidal Forcing</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>No vegetation – 30m grid</td>
<td>747</td>
<td>267</td>
<td>2019</td>
</tr>
<tr>
<td>Dense vegetation</td>
<td>614</td>
<td>219</td>
<td>2019</td>
</tr>
<tr>
<td>30% vegetation</td>
<td>680</td>
<td>243</td>
<td>2019</td>
</tr>
<tr>
<td>No vegetation – 10m grid</td>
<td>719</td>
<td>257</td>
<td>2019</td>
</tr>
</tbody>
</table>

**Event based nutrient loads**

<table>
<thead>
<tr>
<th>Source</th>
<th>Dissolved Inorganic Nitrogen load (kg/event)</th>
<th>Phosphate load* (kg/event)</th>
<th>Date</th>
</tr>
</thead>
<tbody>
<tr>
<td>Storm (Tropical Storm Irene)(^B)</td>
<td>220</td>
<td>80</td>
<td>2011</td>
</tr>
<tr>
<td>Tidal Cycle (average)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>30m grid</td>
<td>25</td>
<td>9</td>
<td>2019</td>
</tr>
<tr>
<td>Dense vegetation</td>
<td>20</td>
<td>7</td>
<td>2019</td>
</tr>
<tr>
<td>30% vegetation</td>
<td>23</td>
<td>8</td>
<td>2019</td>
</tr>
<tr>
<td>10m grid</td>
<td>24</td>
<td>9</td>
<td>2019</td>
</tr>
<tr>
<td>Spring Tidal Cycle (maximum)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>30m grid</td>
<td>91</td>
<td>33</td>
<td>2019</td>
</tr>
<tr>
<td>Dense vegetation</td>
<td>56</td>
<td>20</td>
<td>2019</td>
</tr>
<tr>
<td>30% vegetation</td>
<td>66</td>
<td>24</td>
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<tr>
<td>10m grid</td>
<td>80</td>
<td>28</td>
<td>2019</td>
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<tr>
<td>Neap Tidal Cycle (minimum)</td>
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<td></td>
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<tr>
<td>30m grid</td>
<td>13</td>
<td>5</td>
<td>2019</td>
</tr>
<tr>
<td>Dense vegetation</td>
<td>10</td>
<td>4</td>
<td>2019</td>
</tr>
<tr>
<td>30% vegetation</td>
<td>10</td>
<td>4</td>
<td>2019</td>
</tr>
<tr>
<td>10m grid</td>
<td>9</td>
<td>3</td>
<td>2019</td>
</tr>
</tbody>
</table>

A. Oczkowski (2002)
B. Based on results from Percuoco (2013)
* Uptake not considered for Phosphate
well with observed depth average velocities and shear stress estimates in two locations in the Great Bay estuary.

The study presents an estimate of spatial distribution of shear stress in a tidally dominated estuary using a verified numerical model and compared with stress estimates from observed currents at several locations. The model runs included the effects of vegetation and resolution and found that incorporating vegetation was an important improvement to the model, whereas higher grid resolution had little effect on estimates. The distributions of shear stress were used to estimate internal nutrient loading from sediments for a typical tidal cycle, spring and neap cycle, and averaged over a month. Model results demonstrate that a typical spring tidal cycle compared in magnitude with estimates of DIN and phosphorus, suggesting that when considering tides alone, hydrodynamic conditions favor nutrient release from sediments at least once a month. When compared with rivers, model results suggest that internal sources of nutrient loads from sediment were shown to be about 52-60% of that contributed by rivers for at least half of the year (during low discharge summer and fall periods). Including eelgrass in the model lowers the estimates of nutrient loading by 18% and 8.9% for dense vegetation and 30% vegetation, respectively. These results indicate that when eelgrass meadows are healthy and abundant, the lower the availability of sediment for resuspension and subsequent release of nutrients. This study demonstrates that a coupled hydrodynamic-vegetation model is capable of reasonably estimating the distribution of shear stress for a tidally dominant estuary.

Future modeling studies should incorporate small amplitude waves that can be potentially induce strong shear stresses when the amplitude is on the order of the water depth, which is true for most mudflat and shallow estuarine environments. This can be accomplished through a ROMS-SWAN coupling study within the COAWST modeling system. There is a need for future observational studies to validate waves in these environments as well as continued observational based estimates of shear stress under various hydrodynamic conditions. Future studies in the field of nutrient regeneration in muddy sediments would help constrain the amount of material available on every tidal cycle.
### Table A3.1: Shear Stress calculations: Model vs. Observations (adapted from Wengrove, 2012)

<table>
<thead>
<tr>
<th>2011 Study (Wengrove, 2012)</th>
<th>Shear Stress (N/m²)</th>
<th>Observations</th>
<th>Model</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Viscous</td>
<td>Log law</td>
<td>Quadratic</td>
</tr>
<tr>
<td>Calm June 20th, 2011</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hour 1</td>
<td>0.0033</td>
<td>0.0012</td>
<td>0.0004</td>
</tr>
<tr>
<td>Hour 2</td>
<td>0.0083</td>
<td>0.012</td>
<td>0.0795</td>
</tr>
<tr>
<td>Hour 3</td>
<td>0.0448</td>
<td>0.0117</td>
<td>0.0823</td>
</tr>
<tr>
<td>Hour 4</td>
<td>0.0666</td>
<td>0.0906</td>
<td>0.1075</td>
</tr>
<tr>
<td>Hour 5</td>
<td>0.0589</td>
<td>0.0075</td>
<td>0.0904</td>
</tr>
<tr>
<td>Hour 6</td>
<td>0.0197</td>
<td>0.0042</td>
<td>0.0383</td>
</tr>
<tr>
<td>Peak Storm</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hour 1</td>
<td>0.0119</td>
<td>0.0025</td>
<td>0.1339</td>
</tr>
<tr>
<td>Hour 2</td>
<td>0.0016</td>
<td>0.1677</td>
<td>0.1318</td>
</tr>
<tr>
<td>Hour 3</td>
<td>0.0119</td>
<td>0.1509</td>
<td>0.1410</td>
</tr>
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<td>Hour 4</td>
<td>0.0119</td>
<td>0.1633</td>
<td>0.1604</td>
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<tr>
<td>Hour 5</td>
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<td>0.1242</td>
<td>0.1491</td>
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<tr>
<td>Hour 6</td>
<td>0.0016</td>
<td>0.0163</td>
<td>0.0753</td>
</tr>
<tr>
<td>Waning storm</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Hour 1</td>
<td>0.0025</td>
<td>0.0001</td>
<td>0.0054</td>
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<tr>
<td>Hour 2</td>
<td>0.0003</td>
<td>0.3180</td>
<td>0.0809</td>
</tr>
<tr>
<td>Hour 3</td>
<td>0.0030</td>
<td>0.3398</td>
<td>0.1218</td>
</tr>
<tr>
<td>Hour 4</td>
<td>0.0068</td>
<td>0.3004</td>
<td>0.1666</td>
</tr>
<tr>
<td>Hour 5</td>
<td>0.0186</td>
<td>0.1677</td>
<td>0.1255</td>
</tr>
<tr>
<td>Hour 6</td>
<td>0.0179</td>
<td>0.1205</td>
<td>0.0779</td>
</tr>
</tbody>
</table>

a, viscous stress method
b, average of quadratic and logarithmic law method
CHAPTER 4
SPATIAL AND TEMPORAL DISTRIBUTION OF SHEAR STRESS DUE TO WAVES AND TIDES IN GREAT BAY ESTUARY

4.1 Abstract

Observations of wind-generated waves obtained in the meso-tidal Great Bay Estuary are compared with analytical formulations that predict wave properties based on wind speeds and fetch length, and a coupled numerical model that predicts waves, currents, and bed shear stress spanning the estuary. Observations show that the (typical) summer wind field in 2018 was dominated by oscillating land and sea breezes oriented in opposite directions that spin up and down on regular diurnal cycles. Significant wave heights are on average between 5.5 – 7.3 cm with spectral peak periods of about 1.52 – 1.66 sec. Maximum significant wave heights observed were 32 cm, but generally only reach about 14-17 cm during typical wind conditions. Analytical formulations for significant wave heights and peak wave periods based on fetch length and wind speeds agree well with observations in the center of the estuary suggesting that simple formulations provide reasonable estimates of the wave field. Numerical model results using the SWAN wave model coupled to a hydrodynamic model (ROMS) with the COAWST modeling system show that wave generation and decay are strongly modified by the tides that change the water depth and fetch length, and further depend on the timing of diurnal wind events with the phase and direction of the flooding or ebbing currents. Bed stress from surface waves is weak during lower stands of the tide when the fetch is short and the ebb currents are strong, limiting wave growth, and also during high stands of the tide when orbital velocities do not reach the seafloor and when weaker flooding tidal currents have lesser effects on wave attenuation. Wave-induced bed shear stress is maximum during mid tidal periods when the fetch is large and the water depth shallow allowing waves to interact with the bottom. However, even in those cases the bed stress is dominated by mean currents over most of the estuary, except near the very shallow fringing areas where the currents diminish with progressively shallower depth. Results suggest that during typical
summer conditions sediment transport (when incipient motion critical bed stress is exceeded) is dominated by the strong tidal currents and only weakly affected by waves during mid tidal periods.

4.2 Introduction

Ocean sea and swell dominates wave action at the mouths of rivers and estuaries; however, waves are often dissipated and refracted before traveling upstream through the inlet or into the back bay. Waves in shallow back bays and on tidal flats are mostly locally wind generated (Boon et al., 1996, Lin et al., 2002) and are generally damped when their orbital velocities reach the bottom and interact with seafloor substrate (Wells and Kemp, 1986; Li and Mehta, 1997). However, over shallow regions of estuaries and mudflats very small amplitude, short-period waves associated with light winds have been shown to resuspend sediments (Anderson, 1972; Uncles and Stephens, 2000; Green, 2011, Shi et al., 2017). In these fringing regions wind waves can have a disproportionately large effect on sediment transport when their bed shear stresses are comparable to current induced shear stresses (Le Hir, 2000), and can be strong enough to exceed the critical threshold for sediment resuspension. As mean currents tend to decrease toward shallow areas of the estuary, progressively more fine grained sediments (muds and clays) are deposited over intertidal flats. Sediment transport in these areas is generally a result of the combined effects of wave action and tidally driven flows.

Anderson (1970) found that wave-induced resuspension of fine grained sediments is correlated with significant daily patterns of suspended material in estuaries. Anderson (1972) investigated the influence of small amplitude waves on resuspending sediment over a sheltered, recessed portion of a tidal mudflat along the fringes of the Great Bay Estuary, New Hampshire, USA. They found that average wave heights were small, between 3 – 6 cm, yet capable of suspending sediment during both flood and ebb tides. The maximum suspended sediment concentrations were found during the middle of the ebb tide as the water rushed off the mudflat following a wind pulse and increase in wave activity. During flood tide suspended sediment concentration was linearly correlated with water depth and small amplitude waves. Uncles and Stephens (2000) observed
sediment resuspension associated with relatively light winds in a subestuary of the Tamar River Estuary, UK. Observations of small breaking waves on either advancing (flooding) or retreating (ebbing) flows, combined with the peak flood and ebb currents that occurred shortly after inundation or shortly before drying, were found to enhance the turbidity along the fringe of the estuary.

Other studies have investigated the role of small amplitude waves (less than 20 cm) on the role of sediment resuspension in estuaries. Green (2011) studied the role of small amplitude wind waves (heights less than 10 cm and periods between 1.0-1.8 sec) in sediment suspension and the geomorphological evolution of an estuarine intertidal mudflat in the Tamaki Estuary, New Zealand. Moderate winds (~5 m/s) dominated the resuspension integrated across the estuarine mudflat; however, the primary resuspension was due to waves towards the higher elevations of the mudflat. Sediment resuspension was reduced by wave dissipation as well as attenuation of orbital motions in greater water depths. They found that the location of maximum wave driven sediment resuspension was not the same as the location of maximum duration of resuspension. Sediment concentrations were highest around low tide when attenuation of orbital motions due to depth was reduced, even though the waves were smaller than during higher stands of the tide owing to smaller fetch.

In shallow bays and estuaries, modulations in water depth are determined by tidal dynamics, often considered the primary governor of the spatial and temporal distribution of wave induced bed shear stress. The large spatial extent of tidal flats leads to greater fetch during flooding tides when the intertidal flats are covered and the surface area increases, and oppositely progressively lower fetch during ebbing tides. Low sloping tidal mudflats allow for waves to propagate long distances (encompassing many wave wavelengths) while still interacting with the sea bed. Tidal currents in these areas are generally smaller as the depth decreases and rarely exceed the critical threshold for incipient motion of sediment (0.1 N/m²; Wengrove et al., 2015; Chapter 3 of this thesis). However, by simply changing the water depth the tides act as an important moderator of wave action because wind generated wave heights and periods depend on the fetch length.
Wave climate has been shown to be important to the spatial and temporal variability of wave-induced shear stress determining the long-term stability and evolution (equilibrium) of tidal flats (Fagherazzi et al., 2007). Fagherazzi and Wiberg (2009) found that between mean sea level (MSL) and mean higher high water (MHHW) the height of the waves increases more strongly than the depth resulting in highest bottom stress over any given tidal cycle, a consequence of the balance between increasing fetch length and wave attenuation near the shallow fringing areas. They concluded that the morphology at elevations greater than MHHW are primarily controlled by water depth.

Accurate estimation of shear stress due to waves requires an understanding of the bottom orbital velocities determined by wave height and period and the water depth. Field observations of wave orbital velocities in estuaries have been shown to exceed 10-30 cm/s, enough to induce suspended sediment of fine sediment (MacVean and Lacy, 2014; Chapter 3). Green et al. (1997) found that the maximum orbital speed at the bed occurred around mid-tide on both the flooding and ebbing tides, similar to Christie et al. (1999) who showed that maximum current speeds under waves occurred around mid-tide and minimum currents during high (slack) tide. Previous observational studies estimated bed shear stress on estuarine mudflats from single point acoustic observations (Verney et al., 2007), but were unable to describe the spatial variability of the hydrodynamic features. Comprehensive observation of bed shear stress over the (often) large spatial extent of estuaries and mudflats would require extensive arrays of instruments, an exceedingly difficult and expensive effort. It is much more feasible to estimate the bed shear stress from numerical models that have been verified with much sparser instrumented field arrays (as in Chapter 3 of this thesis). Use of hydrodynamic, wave model, and coupled models are needed to deduce the net sediment transport in estuaries due to spatial variations in bed shear stress as a function of both tidal and wave processes.

Numerical modeling of surface wind-generated waves has been successfully applied to shallow water estuaries (e.g., Umgiesser et al., 2004; Chen et al., 2005; Lettmann et al., 2009) using the Simulating Waves in the Nearshore model (SWAN; Booij et al. 1999). SWAN uses a phase-averaging approach to model wind waves and has been successful at accounting for wave generation and decay due to bottom friction and depth-induced breaking in shallow enclosed basins (Carniello et al., 2005). In an
observational-modeling study of the Chesapeake Bay, USA (Lin et al., 2002), modeled (SWAN) significant wave heights and peak periods compared well with observations under slowly varying atmospheric conditions; however, sudden changes in the wind field led to over-predictions in the wave height and under predictions in the peak wave period. Lin et al. (2002) also found that wave heights were correlated with wind speed and that waves in the middle of the estuary were fetch limited. A similar study by Chen et al. (2005) at Mobile Bay, AL, USA, compared modeled significant wave height and peak wave period with observations and found that the model was sensitive to changes in unsteady forcing and ambient currents and water levels near the shoreline. They concluded that good spatial resolution is needed to characterize the wave field across large bathymetric gradients; for example, from a deep tidal channel to a shallow intertidal mudflats that characterizes many estuaries worldwide.

Carniello et al. (2005) used a coupled hydrodynamic-wave model to predict wind wave generation and determined bottom shear stress patterns in the micro-tidal Venice Lagoon, Italy. Their simulations compared moderate winds (10-16 m/s) with a no-wind case and found that on tidal flats the wind waves wave a high influence on the bed shear stress through both wave induced shear stress and introducing residual currents that increased local velocities. In order to realistically simulate wave breaking at the banks of the deep channels, the dissipation of wave energy was split into a boundary component near major topographical changes and a uniform dissipation component a few model elements away from the channel. Model results showed maximum bed shear stress due to currents during maximum flood and ebb currents in the channels, whereas maximum shear stress on the tidal flats were due to wind waves at average depths. They concluded that without incorporating wind waves into bottom shear stress formulations, tidally-induced bed shear stress was incapable of suspending sediment on tidal flats, a result clearly seen in their observations. This study used moderate wind conditions (10-16 m/s), whereas our study is focusing on the ubiquitous relatively light winds (2-6 m/s) that dominate the daily summer conditions.

In this study, we use a coupled hydrodynamic and wave modeling system to predict the spatial and temporal distribution of bed shear stress due to waves, currents (tidal and wind-driven), and combined wave-current flow in the Great Bay Estuary
located in southeastern New Hampshire, USA under typical (mild) wind conditions that prevail for much of the year. We compare wave predictions (significant wave height, peak spectral period) with wave observations obtained in the estuary for the first time. Model simulations are used to determine the transition regions from current to wave dominated bed shear stress. The variation of the transition region is examined under different hydrodynamic conditions (e.g. wind or tidally dominated motions) using the verified hydrodynamic model. The relative importance of waves and currents and the spatial distribution of bed shear stress (and role in sediment motion and nutrient flux as a function of wind conditions and tidal stage are examined. The wave climate in the Great Bay is observed for the first time and used to drive a coupled numerical model for waves and currents. Observations and model results are compared to simple analytical approaches that use antecedent wind conditions and fetch length to predict wave heights.

In the following we review theoretical seafloor boundary layer formulations that can be used in analytical calculations of bed shear stress and the basis from which numerical models incorporate hydrodynamics and wave motions to compute bed stress. We then describe the field site and the observations of waves, currents, and winds, followed by a brief description of the numerical model that incorporates tidal forcing of currents and water levels, wind forcing of waves and currents, wave-current interactions, and wave dissipation from both bottom boundary layer friction and wave breaking. Results are compared with analytical calculations based on the typical observed wave field.

4.3 Theoretical bed shear stress

The bottom shear stress due to tidal currents can be calculated assuming a turbulent boundary layer with logarithmic mean velocity profile extending from the bottom boundary to the free stream,

$$\bar{u}(z) = \frac{u_*}{\kappa} \ln \left( \frac{z}{z_0} \right)$$

(1)
where $\bar{u}(z)$ is the vertical profile of the mean horizontal velocity, $u_*$ is the shear velocity, $\kappa = 0.4$ is the von Karman coefficient, $z$ is the distance from the bed, and $z_0$ is the roughness length scale. The shear velocity can be estimated without consideration of $z_0$ by taking the derivative of equation (1) with respect to $z$ (Orlu et al., 2010). The so-called indicator function is given by

$$
\tau_{b, currents} = \rho u_*^2
$$

(3)

where $\rho$ is the density of the water.

The bed shear stress due to currents assuming a logarithmic drag law is then calculated as,

$$
\tau_{b, \text{currents}} = \rho u_*^2
$$

(3)

where $\rho$ is the density of the water.

The bed shear stress due solely to waves is often calculated using a quadratic bottom friction relationship,

$$
\tau_{b, waves} = 0.5 f_w \rho u_{b,w}^2
$$

(4)

where $f_w$ is a wave friction factor and $u_{b,w}$ is the maximum wave orbital speed at the bed. Assuming a hydraulically rough turbulent boundary layer, the wave friction factor can be calculated (following Soulsby 1997) by

$$
f_w = 1.39 \left( \frac{A_b}{k_b} \right)^{-0.52}
$$

(5)

where $A_b = u_{b,w} T$ is the wave orbital excursion with $T$ the wave period evaluated at the bed, and bed roughness, $k_b$, calculated using
where $D_{50}$ is the median grain size.

Following linear wave theory, given a wave period, $T$, and wave height, $H$, the wave orbital speed at the bed can be expressed as

$$u_{b,w} = \frac{\pi H}{T \sinh (kh)}$$

(7)

where $h$ is the water depth, and $k = 2\pi/L$ is the wavenumber with $L$ the wavelength. The dispersion relation, $\sigma^2 = g k \tanh (kh)$ with $g$ being acceleration due to gravity, is used to estimate $k$ from the radian frequency $\sigma = 2\pi / T$ following iterative methods described in Wiberg and Sherwood (2008) using the Newton-Raphson method (Soulsby 2006). In shallow water, the bottom orbital velocity becomes

$$u_{b,w} = \frac{H}{2} \frac{\sqrt{g}}{\sqrt{h}}$$

(8)

and is not dependent on $k$.

The combined shear stress due to waves and mean currents is the vector sum

$$\tau_{b,combined} = (\tau_{b,waves} + \tau_{b, currents} \cos \varphi)^2 + \tau_{b, currents}^2 \sin \varphi^{1/2}$$

(9)

where $\varphi$ is the angle between currents and wave propagation (Nielsen, 1992). The combined friction velocity is

$$u_{*,combined} = (\tau_{b,waves} + \tau_{b, currents} \cos \varphi)^2 + \tau_{b, currents}^2 \sin \varphi^{1/2}.$$  

(10)
4.4 Field Site: Great Bay Estuary

The Great Bay Estuary is located within the Gulf of Maine in the southeastern part of New Hampshire (Figure 4.1). It is a recessed drowned river valley, connected to the ocean via the Little Bay, an adjacent deep bay located between Adam’s Point and Dover Point and the main Piscataqua River characterized by strong currents in a deep (17-25 m), narrow, rocky channel extending 10 km to the mouth at Portsmouth, NH. Tidal currents at the constrictions can exceed to 2 m/s, with average currents in the Great Bay proper south of Adam’s Point about 0.50 m/s (Cook et al., 2019). Ocean waves are generally refracted towards shore at the mouth of the estuary and do not propagate up the Piscataqua River. Until now, the wave climate of the Great Bay has only been qualitatively characterized. The Great Bay south of Adam’s Pt. is comprised of primarily mudflats (greater than 50% of the spatial area) and deep channels that are maintained by both tidal and river flows. The average depth of the bay proper is 3.2 m (NHDES, 2007), with a 15 m main channel leading to head of the estuary at the mouth of the Lamprey and Squamscott Rivers. There are 7 principal rivers that feed into the Great Bay Estuary (see Cook et al., 2019); however, the discharge is weak for most of the year and freshwater makes up 2% of the tidal prism (NHDES, 2007; Short, 1992). The morphology of the estuary is relatively stable, expressing a convex shape suggesting that this environment is depositional, tidally dominated, and wave energy is locally generated (Friedrichs, 2011).
Figure 4.1: Site location of the Great Bay Estuary and Piscataqua River in southeastern New Hampshire, USA, within the Gulf of Maine (inset).
4.5 Observations

In the summer of 2018, four directional wave buoys (Spoondrift Spotters; Raghukumar et al., 2019) were deployed in the Great Bay (Figure 4.2). The wave buoys were deployed in the center of the bay in the main channel (Spotter 0080), in the minor eastern channel towards the Winnicut River (Spotter 0074), in the western channel (Spotter 0073), and in the upper reach of the western channel near the mouths of the Lamprey and Squamscott Rivers (Spotter 0071). Each GPS-based buoy measured surface displacements in three orthogonal directions at 2.5 Hz and computes directional spectra over the frequency range 0.05 to 1.2 Hz (Raghukumar et al., 2019). Buoy 0080 was deployed from 10 April through 05 August, buoys 0071 and 0074 from 11 July through 20 August, and buoy 0073 from 11-16 July.

The wave energy spectra (based on the vertical displacement time series) with 500 degrees of freedom over the entire deployment periods for each buoy are shown in Figure 4.3. Red noise was apparent in the buoy measurements owing to the relatively low amplitude, high frequency waves in the estuary (compared with typical open ocean values), restricting the useful range of frequencies to greater than about 0.4 Hz (similar to observed wave fields in Green, 2011).

Significant wave heights, $H_s$, for each buoy were calculated over the high frequency portion of the wave spectrum

$$H_s = 4 \left[ \int_{0.4 \text{ Hz}}^{1.2 \text{ Hz}} S_\eta(f) df \right]^{1/2} \tag{11}$$

The mean wave period, $T_{Avg}$, is calculated by integrating $S_\eta$ using

$$T_{Avg} = \left[ \frac{\int_{0.4 \text{ Hz}}^{1.2 \text{ Hz}} S_\eta(f) df}{\int_{0.4 \text{ Hz}}^{1.2 \text{ Hz}} f^2 S_\eta(f) df} \right]^{1/2} \tag{12}$$

and the peak period, $T_p$, is the inverse of the peak frequency, $f_p$, of $S_{\eta(t)}$

$$T_p = \frac{1}{f_p} \tag{13}$$
Figure 4.2: Location of 4 Spoondrift wave buoys (colored circles), Nortek AWAC (yellow triangle), and Greenland meteorological station (left pointing magenta triangle).
Figure 4.3: Wave spectra from each wave buoy over the deployment period. All spectra are computed with 500 degrees of freedom.
The time series of vertical displacements, $\eta(t)$, were used to compute sea surface elevation spectra, $S_\eta(f)$, over 30 min intervals, where $f$ is frequency and $t$ is time. Waves, bottom pressure, and mean currents were measured with a Nortek AWAC wave and current profiler deployed in 6.5 m water depth in the eastern channel of the Great Bay (yellow triangle in Figure 4.2). Waves were sampled at 2 Hz for 17.1 min every hour. Three-dimensional current velocity profiles spanning the water column in 0.5 m bins were averaged at 5 min intervals between wave sampling periods, from which current magnitude, $U_m$, and direction, $\theta_m$, were computed. Water levels relative to MSL, $\eta_L$, were estimated at the same 5 min averaging intervals using the AWAC pressure measurements corrected for the average density in the estuary over the time period of the deployments (estimated from AWAC temperature and salinity measurements obtained at the Great Bay buoy; http://www.opal.sr.unh.edu/data/buoys/great_bay/index.shtml). The wave field was generally too small and too high frequency for the AWAC to resolve directional spectra or wave heights, and thus only the water levels and current profiles are used herein to qualitatively assess the phase of the tides and currents in the Great Bay proper.

Wind speed ($U_w$), wind direction ($\theta_w$), and photosynthetically active radiation ($PAR$) were obtained from a nearby meteorological station located in Greenland, NH (left pointing triangle, Figure 4.2) maintained by the Great Bay National Estuarine Research Reserve (GBNERR; data available from the NOAA Centralized Data Management Office website http://edmo.baruch.sc.edu). The wind speed sensor is located 6 m above the ground and averaged over 15 min. The wind speed is converted to values corresponding to 10 m above sea level in accordance with meteorological conventions, and the wind field is assumed to be homogenous over the entire estuary. $PAR$ was used as a proxy for identifying day and nighttime periods.

Time series of observed significant wave height, $H_s$, from each buoy are shown in Figure 4.4 for the time period spanning 11 July through 20 August 2019 (a shorter 9-day record of the time series is shown later with higher resolution in Figure 4.8). Significant wave heights in the Great Bay during the summer months varied between 1 – 17 cm. Time series of wind velocity speed, $U_w$, and direction, $\theta_w$, and $PAR$ (also shown in Figure
4.4) show that the wave heights have strong variation associated with diurnal wind fields driven by differential atmospheric heating and cooling in the coastal New Hampshire region. The wave fields spin up and spin down rapidly, with magnitudes that depend on the wind direction as well as wind speeds. The magnitudes of the waves also depend on the phase of tidal water levels, $\eta_L$, during the wind events that change the fetch length, water depths, and direction and magnitude of the currents, $C_m$ (also shown in Figure 4.4).

The probability density function (pdf) for $H_s$ for each buoy is shown in Figure 4.5 indicating the distribution of the wave field observed at each location in the estuary. The pdf’s are similar for all 4 buoys. It should be noted that very small amplitude waves are not likely accurately measured below $H_s$ of about $2 - 3$ cm owing to the response dynamics of the Spotter buoys. The most common (mode) wave heights at all buoy locations is the no wave condition ($H_s < 2$ cm). Mean $H_s$ ranged 0.055 – 0.073 m, with maximum significant wave heights, $H_{max}$, of 0.32 m observed at buoy 0080 (in the approximate center of the Great Bay during a strong wind event in mid May; time period not shown in Figure 4.4) and between 0.14 and 0.17 m at the other three buoys.

Time series of the spectral peak period, $T_p$, are shown in Figure 4.4 only for times when $H_s > 3$ cm (with values for the lesser wave heights deemed unreliable owing the Spotter response). The pdf’s of $T_p$ and $T_{avg}$ (Figure 4.5) are similar at each buoy location, with the exception of the bi-modal $T_p$ distribution for buoy 0073 likely due to its relatively short record length. Mean $T_p$ and $T_{avg}$ ranged 1.52 – 1.66 sec and 1.44 – 1.53 sec, respectively. Higher observed $T_p$ greater than about 2 sec is likely due to red noise of the spectra appearing at higher frequencies during low wave periods (Figure 4.3).
Figure 4.4: Hydrodynamic, wind and wave climate in the Great Bay (a) Sea surface height (m) and depth-averaged velocity magnitude (m/s) from AWAC, (b) wind speed (m/s) and photosynthetically active radiation (PAR), (c) wind direction (0 deg = True North; +CW), (d) significant wave height (m), (e) peak wave period (sec) measured from four Spotter wave buoys. Wind data from the Greenland meteorological station (GBNERR; https://cdmo.baruch.sc.edu).
Figure 4.5: Pdf’s of observed Hs, Tp, and Tavg from each wave buoy. Tp and Tavg only for Hs > 3 cm.
Directional estimates of the high frequency waves obtained with the Spotter buoys (deduced from the recorded first 4 directional Fourier coefficients over the frequency range 0.4 to 1.2 Hz) but were not reliably estimated, especially for the very small waves. There is a bias in the horizontal motions of the buoy induced by the effects of the relatively large mean currents (relative to the orbital velocities of the waves) acting on the buoy. This bias does not likely effect the vertical displacements used to estimate $H_s$, $T_p$, and $T_{avg}$. Qualitatively, the direction of wave propagation of the high frequency waves was visually observed to be in the general direction of the wind fields.

The winds are relatively mild (2 – 3 m/s) and variable, with no notable wind events (except for a brief storm in May; time series not shown in Figure 4.4). The wind rose for the Greenland meteorological station is shown in Figure 4.6. The winds are primarily out of the ESE (~115-130 deg) and WNW (~270-290 deg), following a land-sea breeze pattern associated with the differential diurnal heating and cooling of the atmosphere in the coastal New Hampshire region (note that the Greenland met station is located about 5 miles from the coast).

There is a clear spring-neap modulation to the tidal flows in the Great Bay. Spring tide current velocities in the vicinity of the AWAC averaged 50 cm/s and 35 cm/s on the flood and ebb tide, respectively. Neap average current velocities were lower, about 45 cm/s on the flood and 25 cm/s on the ebb. Currents in the deeper channel are in general stronger (Cook et al., 2019), and can have a substantial effect on the refraction and dissipation of the wave field. It would be expected that any wave model would need to take the wave-current interactions into account.
Figure 4.6: Wind rose from the Greenland, NH meteorological station applied to [a] July and August and [b] subset data from July 12-22nd 2018.
4.6 Numerical Model

The numerical model used is the Coupled Ocean Atmosphere Wave and Sediment Transport (COAWST) model (Warner et al., 2008ab, 2010; v. 3.4 svn. 1401). In this work we utilize the hydrodynamic model Regional Ocean Modeling System (ROMS v. 3.7; http://www.myroms.org; Haidvogel et al., 2000; Shchepetkin and McWilliams, 2005; Haidvogel et al., 2008) and the wave model Simulating Waves in the Nearshore (SWAN v. 41.20; Booij et al. 1999; http://www.swan.tudelft.nl).

ROMS is a regional ocean circulation model that has been successfully applied to many coastal and estuarine systems, including the Great Bay (Cook et al., 2019). It is a fully three dimensional, free-surface, terrain following numerical model that solves the Reynolds-averaged Navier-Stokes (RANS) equations with finite difference approximations and discretizes the model domain in an Arakawa “C” grid structure. The Generic Length Scale (GLS) method is used for turbulence closure (Warner et al., 2005). The baroclinic and barotropic time steps in ROMS are 1.5 sec and 0.05 sec, respectively (following Cook et al., 2019).

The ROMS bottom boundary condition follows the Sherwood-Signell-Warner bottom boundary layer parameterization (SSW_BBL; described in Warner et al., 2008b). The bottom boundary is divided into two layers: a combined wave-current boundary layer, and a current-only boundary layer (described earlier). Current-only and wave-only shear stress estimates are made separately first to provide consistent eddy viscosity and velocity profiles between a defined bottom roughness length scale, $z_M$, and reference elevation, $z_V$, following Styles and Glenn (2000).

The initial current-only shear stress estimate is determined assuming a logarithmic velocity profile by combining equations 1-3, and representing the velocity as the vector sum of orthogonal velocity components, $u$ and $v$, where $|\bar{u}| = (u^2 + v^2)^{1/2}$, such that

$$\tau_c = \frac{\rho_0 (u^2 + v^2) \kappa^2}{\ln(z/z_0)}$$

(14)

and wave-only case given by equation 4.
The friction factor, $f_w$, depends on the ratio of the wave-orbital excursion amplitude to the bottom roughness length, $A_b/k_b$, as in equation 5, with $A_b$ and $k_b$ approximated by Madsen (1994) as

$$A_b = u_{br}T/2\pi$$

$$k_b = 30z_0$$

such that

$$f_w = 0.3, \quad A_b/k_b \leq 0.2$$

$$f_w = \exp(-8.82 + 7.02(A_b/k_b)^{-0.078}), \quad 0.2 < A_b/k_b \leq 100$$

$$f_w = \exp(-7.30 + 5.61(A_b/k_b)^{-0.109}), \quad A_b/k_b > 100.$$
\[ N(\sigma, \theta) = E(\sigma, \theta)/\sigma \]  

where \( \theta \) is the wave direction, and \( c_x, c_y, c_\sigma, \) and \( c_\theta \) are the wave group velocities corresponding to \( x, y, \sigma, \) and \( \theta \), respectively. \( S_{in} \) represents wave generation through wind input, \( S_{ds} \) represents wave dissipation (through white-capping, bottom friction and depth-induced wave breaking), and \( S_{nt} \) represents the sources and sinks of wave energy through wave-wave interactions (Tolman et al., 2002).

Wave growth due to wind is described as a sum of a linear resonance term (A; Cavaleri and Malanotte-Rizzoli, 1981) and the exponential feedback growth term \( (BE(\sigma, \theta); Komen et al., 1984) \) through

\[ S_{in}(\sigma, \theta) = A + BE(\sigma, \theta). \]  

The default formulation is based on the work of Snyder et al. (1981) and given in Komen et al. (1984) by

\[ B = \max \left[ 0, 0.25 \frac{\rho_d}{\rho} \left( 28 \frac{U_*}{c_{ph}} \cos(\theta - \theta_w) - 1 \right) \right] \sigma. \]  

where \( \rho_a \) is density of air, \( U_* \) and \( c_{ph} \) are the wind friction velocity and wave phase speed, respectively, and \( \theta_w \) is the mean wind direction. \( U_* \) is calculated from the wind speed at 10 m elevation

\[ U_*^2 = C_d U_{10}^2. \]  

where \( C_d \) is the wind drag coefficient calculated following Wu (1982),

\[ C_d = 1.2875 \times 10^{-3} \quad \text{for} \quad U_{10} < 7.5 \text{ m s}^{-1} \]  

\[ C_d = (0.8 + 0.065 U_{10}) \times 10^{-3} \quad \text{for} \quad U_{10} \geq 7.5 \text{ m s}^{-1}. \]  

Wave dissipation due to depth-induced wave breaking follows Battjes and Janssen (1978). Maximum wave height, \( H_{max} \), is determined with a breaker saturation parameter \( \gamma \), and total water depth, \( h \), through
\[ H_{\text{max}} = \gamma h. \]

A constant value of \( \gamma = 0.73 \) is used following Battjes and Stive (1985). Wave dissipation due to white capping is based on Hasselmann (1974) and Komen et al. (1984). Wave dissipation due to bottom friction follows an eddy viscosity model (Madsen et al., 1988) that depends on a friction coefficient, \( C_b \), taken as 0.05.

Four-wave (quadruplet) and three-wave (triad) interactions are solved using the DIA method (Hasselmann et al., 1985). Triad wave interactions are computed by the lumped triad approximation (Eldeberky, 1996).

Both the hydrodynamic and wave model were run on the same model grid with 30 m grid resolution and 8 vertical sigma levels. Bathymetric depths were obtained from several sources. A more detailed description of the grid and bathymetric data sets used can be found in Cook et al. (2019). Often, wave model grids are coarser than for the hydrodynamic model; however, using a higher resolution grid for the wave model better accounted for sharp topographic changes, particularly near transition regions from the deep channels to the shallow mudflats that can strongly influence wave characteristics (i.e., wave height and period).

The coupled COAWST model was forced with tides at the ocean boundary seaward of the mouth of the Piscataqua River and winds uniformly distributed over the domain. Tidal forcing was applied in ROMS using tidal amplitudes and currents defined by the Oregon State University Tidal Prediction Software package (OTPS) with the United States East Coast Regional Tidal Solution (EC2010; Egbert and Erofeeva, 2002). Wind forcing was derived from meteorological station data located in Greenland, NH. Wind direction was converted from meteorological convention (0 deg N; +CW) to Cartesian coordinate system (0 deg E; +CCW). Wind speed observed at 6 m elevation, \( U_6 \), was adjusted to 10 m elevation, \( U_{10} \), following Large and Pond (1981),

\[ u_{10} = u_6 \left( 1 + \frac{\sqrt{C_p}}{\kappa} \ln \left( \frac{6}{10} \right) \right) \]
where $C_d$ is an atmospheric drag coefficient assuming a bulk aerodynamic formula with neutrally stable boundary layer

\[ C_d = 1.2 \left(10^{-3}\right) \quad \text{for} \quad 4 \leq U_{10} < 11 \, m \, s^{-1} \quad (26) \]

\[ C_d = 10^{-3} (0.49 + 0.065U_{10}) \quad \text{for} \quad 11 \leq U_{10} \lesssim 25 \, m \, s^{-1}. \]

The wind speed is transformed into vector form $(u_w, v_w)$, and converted to orthogonal surface stresses by applying

\[ \tau_{xx} = \rho_a C_d u_w^2 \quad \quad (27) \]

\[ \tau_{xy} = \rho_a C_d v_w^2 \quad \quad (28) \]

where $\rho_a$ is given by $1.225 \, kg/m^3$. Kinematic surface stresses are given by

\[ \tau_{xx} = \rho_a C_d u_w^2 / \rho \quad \quad (29) \]

\[ \tau_{xy} = \rho_a C_d v_w^2 / \rho. \quad \quad (30) \]

with $\rho = 1025 \, kg/m^3$, the approximate average ocean density in the Great Bay.

SWAN was run in non-stationary mode with the wave spectrum discretized logarithmically into 45 frequency bins between $0.04 - 3.0 \, Hz$, and wave direction distributed into 36 bins with constant bandwidth of $10 \, deg$. Iterations were set at 5 to ensure the model to converged on each of the runs. SWAN was forced with variable wind speed from the meteorological station observations.

4.7 Results

**Analytical wave induced shear stress distribution**

Near bed wave induced shear stresses were estimated using observations of significant wave height and peak period from the Spotter buoys. Calm conditions were represented by significant wave height of $10 \, cm$ with peak period of $1.75 \, sec$ from the July 12th-July 22nd deployment. A significant wave height of $20 \, cm$ is chosen to represent moderate wind conditions with $10 \, m/s$ winds with the same peak period. Using analytical formulations (Eqs. 4-7) the resulting spatial distribution of shear stress during a
wave height of 10 cm (mild; \( \sim 3 \) m/s winds) or 20 cm (moderate; \( \sim 10 \) m/s winds) with a wave period of 1.75 sec is shown in Figure 4.7.

The analytical approach uses a representative wave height, wave period, and median grain size across the entire domain allowing for basin-wide estimates of wave-induced shear stress for the Great Bay as a function of tidal stage. The maximum shear stress occurs at low and mid tides along the fringe of the estuary and across a large portion of the mudflats. The bed shear stress in these areas is greater than 0.1 N/m\(^2\), the critical threshold for incipient sediment grain motion (Wengrove et al., 2015; Shields, 1936). This approach, however, it does not incorporate the effects of tidal currents on wave growth and propagation, or wave damping due to wave-bottom interactions in shallow water. Waves are damped in shallow water, decreasing their wave height resulting in lower their orbital velocities and associated bed shear stress.

**Modeled waves**

A subset of the data was chosen for the modeling study based on representative wind and tidal characteristics. Time series of observed wave, wind, and current properties are shown in Figure 4.8. Nine wave events are delineated with the gray regions and summarized in Table 4.1. Two periods outlined in red were chosen to represent two different wind states that characterize the summer wind climate in the Great Bay (Figure 4.6). Model A is represented by a typical “sea breeze” represented by 2 – 3 m/s winds from the ESE (115-130 deg) and Model F is represented by a typical “land breeze” with 2 – 3 m/s winds from the WNW (270-290 deg).

The sea breeze (Model A) model-observation comparison of \( H_s \) and \( T_p \) at each Spotter buoy location is shown in Figure 4.9. Also shown are the modeled wave direction, observed wind direction, modeled water level and modeled depth averaged tidal currents from each location. The land breeze (Model F) model-observation comparison is shown in Figure 4.10 (note that Spotter buoy 0073 data is not available during this period).
Figure 4.7: Wave induced bottom shear stress for (top row) mild winds (~3 m/s) and 10 cm Hs, and (bottom row) moderate winds (~10 m/s) and 20 cm Hs. Tide stage for model runs: (left panels) low tide, (right panels) mid tide (~ MSL), (right panels) high tide.
Figure 4.8: (a) Sea surface height (m) and (b) depth-averaged velocity from AWAC. (c) Wind Speed (m/s) and PAR. (d) wind Direction (deg; True North = 0). (e) Observed significant wave height (m) and (f) peak period from each Spotter buoy. Grey panels are “wave events” encompassing with Hs > 3 cm. Red bold outlines delineate the two events that were modeled (A) sea breeze case; (F) land breeze case.
Table 4.1: Model Time Periods

<table>
<thead>
<tr>
<th>Time period</th>
<th>Tide (stage)</th>
<th>Wind Speed (Dir) (m/s)</th>
<th>Max Significant Wave Height (cm)</th>
<th>Max Peak Wave Period (sec)</th>
</tr>
</thead>
<tbody>
<tr>
<td>A*</td>
<td>Falling</td>
<td>3 (115 deg ESE)</td>
<td>10</td>
<td>1.92</td>
</tr>
<tr>
<td>B</td>
<td>Falling</td>
<td>2 (115 deg ESE)</td>
<td>8</td>
<td>1.89</td>
</tr>
<tr>
<td>C</td>
<td>Falling</td>
<td>1.5 (115 deg ESE)</td>
<td>4</td>
<td>1.95</td>
</tr>
<tr>
<td>D</td>
<td>Falling</td>
<td>1.5 m/s (115 deg ESE)</td>
<td>5</td>
<td>1.93</td>
</tr>
<tr>
<td>E</td>
<td>Rising</td>
<td>3 m/s (WSW)</td>
<td>9</td>
<td>1.98</td>
</tr>
<tr>
<td>F*</td>
<td>Rising</td>
<td>2 m/s (WNW)</td>
<td>10</td>
<td>1.78</td>
</tr>
<tr>
<td>G</td>
<td>Rising</td>
<td>2 m/s (WNW)</td>
<td>7</td>
<td>1.88</td>
</tr>
<tr>
<td>H</td>
<td>Rising</td>
<td>2 m/s (ESE)</td>
<td>10</td>
<td>1.95</td>
</tr>
<tr>
<td>I</td>
<td>Rising</td>
<td>2 m/s (ESE)</td>
<td>6</td>
<td>2.02</td>
</tr>
</tbody>
</table>

[1] Tide stage from AWAC pressure data
[2] Wind speed and direction from Greenland, NH meteorological data
[3] Maximum significant wave height and peak period from the distribution observed at all wave bouys.
(*) used for the modeling study

For the sea breeze case (Figure 4.9) the modeled waves follow the same pattern as the observed wave heights although are underestimated between 20-50% (about 3-8 cm). The peak wave period is underestimated by about 0.5 sec. Assuming the wave direction follows the wind direction, then the model represents the wave direction well. In all cases a jump in wave direction is seen at the turn of the tide around low tide (which includes a change in the current direction) implying a strong wave-current interaction. Similar behavior is observed at all buoy locations.

For the land breeze case (Model F; Figure 4.10) the modeled waves follow the same pattern as the observed wave heights for buoys 0071 and 0080, but do not predict the observed 10 cm waves at buoy 0074. The modeled $T_p$ is highly variable, and does not appear to depend on wind direction. It could be that wave growth from winds out of the WNW are acting against the ebbing tide and the model isn’t properly accounting for wave-current interactions at this scale. On the other hand, during flooding tide with the winds in the direction of the currents, the model is able to reasonably well represent the
wave height for each of the buoys. In general, the peak wave periods better agree on the flooding tide than the ebbing tide.

**Numerical Shear Stress distribution**

Figures 4.11 and 4.12 shows the spatial distribution of modeled significant wave height, tidal currents, combined wave-current bed shear stress, and the ratio of wave bed stress to current bed stress for the model A (sea breeze) and model F (land breeze) time periods. For each case model output were extracted at mid tide. A contour line for 0.1 $N/m^2$ is added to the combined wave-current shear stress panels indicating the threshold for incipient motion (Wengrove et al. 2015).

Model A (Figure 4.11; sea breeze) shows 5-10 cm waves across the western portion of the bay during the middle of the ebbing tide. Winds from the ESE result in greater wave growth on the downwind (northwest) side of the estuary than the upwind (southeastern) portion of the estuary. At the mid-tide stage the mean tidal currents in the channel act to shelter the downwind side of the estuary presumably through wave dissipation, dispersive, and refraction processes. It is generally apparent in the model runs that during average water depths wave impacts on the seabed are at their maximum. The combined bed shear stress is strongest in the channels and on parts of the mudflat where waves are present; some areas of the northwestern mudflat have wave induced shear stress greater than 0.1 $N/m^2$. The ratio of the wave-induced to current-induced shear stress shows waves to be important along the fringing mudflat on the western and northwestern side of the estuary. Although during ebbing tide the currents are not in the same direction as the waves on the northwestern portion of the bay, the contribution of wave-induced bed stress is still relatively strong, approximately 50-100% of the current-induced bed stress (Figure 4.11, bottom right panel).
Figure 4.9: Sea breeze case. Modeled (solid lines) and observed (dotted lines) significant wave heights and peak periods (open circles) for each of the 4 buoys (a) Spotter 0071 (b) Spotter 0073 (c) Spotter 0074 (d) Spotter 0080. Modeled wave direction is compared with wind directions as the buoy directions were not reliable for the small waves and large currents in the estuary. The bottom panels of (a)-(d) show modeled tidal currents and water levels for each location.
Figure 4.10: Land breeze case. Same format as for Figure 4.9. (a) Spotter 0071 (b) Spotter 0074 (c) Spotter 0080. Spotter 0073 was inactive during this period.
Model F (Figure 4.12; land breeze) shows 2-6 cm waves across the southeastern portion of the bay during the middle of the flooding tide. Winds from the WNW result in greater wave growth on the downwind (southeastern) side of the estuary than the upwind (northwestern) portion of the estuary. The fetch in this case is smaller than for Model A, with smaller winds (~1-2 m/s), resulting in smaller waves. During the ebbing tide (from high to low water the model) the model did not generate the observed 8-10 cm waves at buoy 0074 (Figure 4.10). In this part of the bay, winds and waves are acting at an angle to the flooding tide when the tidal currents are directed down the western channel of the Great Bay to the Winnicut River. The combined wave-current shear stress shows a smaller bounded area where bed stress is greater than 0.1 N/m², suggesting there would be less sediment transport than for Model A. The ratio of wave-to-current induced shear stress shows stronger wave contributions on the southeastern lobe of the Great Bay in line with the wind and wave direction, and consistent with modeled tidal currents that diminish as depths shallow over the mudflats towards the estuary banks (Figure 4.12; top right panel).

4.8 Discussion

Observed Wave Climate

The summer wave climate in the Great Bay estuary is characterized by light winds 1-4 m/s from dominant ESE and WNW directions (Figure 4.6) characterized by strong land and sea breezes determined by the passage of atmospheric fronts. Differential atmospheric heating and cooling produces strong diurnal oscillations in wind speeds with winds spinning up at dawn and down at dusk, and calm winds at night (Figure 4.4). The summer surface wind-wave climate of the Great Bay Estuary is observed with 4 wave buoys for the first time. Daily wind patterns produce (typically) small amplitude waves
Figure 4.11: Model case A (sea breeze): Modeled wave heights (middle left), depth-averaged currents (middle right), combined wave-current induced shear stress (bottom left), and the ratio of wave induced shear stress to current induced shear stress (bottom right). The top panel shows modeled sea surface elevation at Spotter Buoy 0080 (station location shown with red circle in Hs panel) and wind vectors from the Greenland meteorological station. Red contour line for the combined wave-current induced shear stress are for areas greater than 0.1 N/m^2.
Figure 4.12: Model case F (land breeze): Same format as for Figure 4.11.
with significant heights of 3–10 cm with average peak periods between 1.5–2 sec (Figures 4.4, and 4.5). Maximum $H_s$ was about 0.32 m occurring during a spring storm.

Wave spectra show broad distribution of peak periods between 1.25-1.75 sec, except for buoy 0073 which shows a bi-modal distribution that was likely due to the limited 4 day record. Directional data from the Spotter buoys was likely biased by the presence of relative strong currents relative to the orbital velocities of the wave field and the dynamics of the buoys, and were not reliably estimated. Qualitatively, the waves always were propagating downwind in the approximate direction of the wind vectors (in agreement with modeled wave directions).

Observations of winds (> 2 m/s) from the ESE (115-155 deg) show that waves develop on the southeastern portion of the bay first (buoys 0074 and 0080; Figure 4.7) and then later in the western portion of the bay (Buoys 0071 and 0073; Figure 4.9), as expected due to the growth across the fetch length. However, when winds are out of the WNW (270-290 deg) a different pattern arises (Figure 4.10). Not surprisingly wave heights in the southeastern corner of the bay, at buoy 0074 are larger than the center of the bay at buoy 0080. This spatial and temporal variability in wave heights and directions driven by the dominant, persistent oscillating diurnal land/sea breeze pattern results in asymmetric distribution of wave-induced bed shear stress in the southeastern and western regions along the downwind fringes of the bay.

The growth of waves are modulated by the phase of the tidal water levels and currents which govern the surface area of the bay and determines the fetch length and the dissipation induced by wave-current interactions. Regions of the estuary with large fetch in line with the wind directions are more likely to be affected by waves. Tidal stage also modulates the water depth and in concert with the wave periods (and hence wavelengths) determines where the orbital velocities of the waves interact with the sea bed and generate bed stresses. Waves are largest at mid-tidal stages, during either the rising or falling tides (Figure 4.8), a consequence of fetch length that increases as water levels are higher, wave-current interactions that decrease waves (especially when wave direction opposes the current), and bottom depth that increases wave attenuation in shallow water more strongly at lower tide levels. Waves are observed to increase more strongly on
rising (flooding) tides, and decrease during falling (ebbing) tides, consistent with observed small amplitude waves in other estuaries (Green, 2011; Green and Coco 2014).

Waves rarely break due to depth limited conditions in the Great Bay (except very near the edges of the estuary), and energy dissipation is primarily due to bottom friction and wave-current interactions. The tide provides a strong influence on fetch length by establishing the water depth and spatial area of the bay surface, and therefore exerts an order one control over the potential wave heights across the estuary. We can compare the observed significant wave height to the JONSWAP fetch-limited prediction (Green, 2011; Massel, 1996), where

\[
H_{s-j} = \frac{U_{\text{met}}^2}{g} \times 0.0016 \left( \frac{g F_x}{U_{\text{met}}} \right)^{1/2}
\]  

(31)

where \(U_{\text{met}}\) is the meteorological wind speed (given in m/s) along the fetch, \(F_x\) (given in m). The peak wave period can be compared with the fetch-limited JONSWAP formulation,

\[
T_{p-j} = 0.286 \frac{U_{\text{met}}}{g} \left( \frac{g F_x}{U_{\text{met}}^2} \right)^{1/3}
\]  

(32)

which also depends on \(U_{\text{met}}\) and \(F_x\). Using numerical sea surface elevation from the verified hydrodynamic model of the Great Bay (Cook et al., 2019), the fetch can be calculated using observations of wind direction at any point in the estuary.

Sea surface height, wind speed and direction, calculated and observed \(H_s\), calculated \(F_x\), and calculated \(T_{p-j}\) and observed \(T_p\), are shown in Figure 4.13. Observed sea surface elevation values are taken from the AWAC and compared with the numerical model. Calculated \(H_{s-j}\) are compared with observed \(H_s\) from Spotter buoy 0080. Predicted wave heights agree well the observations, with slight overestimation (generally) during larger wind events, likely owing to wave dissipation processes not accounted from in the simple analytical formulation (equation 31). Analytical estimates of peak wave period, \(T_{p-j}\) (equation 32) agrees reasonably well with observed \(T_p\) from buoy 0080 in the center of the bay (Figure 4.13) for times when \(H_s\) exceeds 3 cm (to
avoid possible influence of red noise in the wave spectra; Figure 4.3). Overall, using fetch length and wind speed is considered a good estimator of peak wave period even though this formulation does not incorporate the effect of wave-current interactions on wave period.

Previous studies have applied a characteristic wave height and peak period across the estuary to estimate a bottom orbital velocity and shear stress due to waves alone. This method is useful when making a first guess of the orbital velocities and wave induced bed shear stress (following Wiberg and Sherwood, 2008), however with some limitations owing to the assumption that waves have not evolved through bottom friction, wave breaking, dispersion, or wave-current interaction. The simple analytical method uses one (or a few) estimate(s) and applies the results across the entire domain irrespective of the spatial variation in water depth or mean currents, yet could be considered a reasonable approach to quickly estimate the effects of wind-generated waves on bed stress.

**Numerical estimate**

In principle, incorporating a numerical model should be able to provide a better estimate wave statistics than the simple analytical formulations, and improve the estimate of the bed shear stress across the estuary. The COAWST coupled ocean-wave numerical model includes the effects of tidal currents and wave-current interactions that are neglected with simple parameterizations. Of interest is that model-observation comparison shows a temporal lag in the model’s ability to produce waves in both the sea breeze (Model A) and land breeze (Model F) model runs (Figures 4.9 and 4.10). Chen et al. (2005) showed that in the presence of ambient currents or shallow water depths, model predictions of wave height and wave period are reduced by 50% and 30%, respectively. They found that having high model grid resolution in the presence of strong geometric and bathymetric gradients is important for considering the effects of ambient currents on wave predictions. Further modeling studies are needed to examine the sensitivity of wave-current formulations in COAWST to the prediction of wave height and period during different hydrodynamic and meteorological conditions.
Figure 4.13: (a) Observed (red) and modeled (black) sea surface height (m) at the AWAC. (b) Observed wind speed (m/s; blue) and wind direction (0 deg N, CW+; green)). (c) Fetch length (km; black) based on modeled water levels and observed wind direction, and observed (blue) and analytical predictions (Massel, 1996; red) significant wave height (m) at Spotter buoy 0080. Observed (blue circles) and analytical predictions (Massel, 1996; red) peak wave period (sec) from Spotter buoy 0080. Observed peak periods only for Hs > 3 cm.
At times, the coupled ocean-wave numerical model was unable to reproduce the wave height and periods observed in the Great Bay, and as a consequence would under predict the wave-induced bed shear stress. In both cases the wave-only bed shear stress exceeded the critical threshold for incipient motion (0.1 \(N/m^2\); Shields, 1936; Wengrove et al., 2015) along the downwind fringes of the estuary. Sediment resuspension from wave-induced bed stress predicted to occur when bed stresses exceed 0.35 \(N/m^2\) (Wengrove et al., 2015) was not widespread for the conditions observed during the summer months. These results are consistent with previous studies with small amplitude waves and suspended sediment across mudflats (Anderson, 1970; Anderson, 1972; Green, 2011). Results suggest that, except for the fringes of the estuary, bed shear stress is dominated by mean tidal currents.

Fagherazzi and Wiberg (2009) found that wind waves produced four bottom shear stress regimes depending on water elevation. Between MSL and MHHW the height of the waves increases more than the depth and result in a peak bottom stress, and that these elevations in the estuary are controlled by fetch. Between MSL and MLLW – regions where tidal flats exist – they found that the increase in water depth reduces bottom stress, consistent with theoretical formulations and our model results. They also found that wave height, orbital velocities, and bed shear stress grow monotonically with wind speed. Our model results support this finding; however, there is a lag between modeled wave generation and the observations.

4.9 Conclusion

Observations of surface wind waves characterize the wave climate of the Great Bay Estuary for the first time, and are used to verify model simulations from the coupled ocean-wave numerical modeling system COAWST that utilizes the ROMS hydrodynamic and SWAN wave models. Waves are driven by a dominant diurnal sea-land-breeze pattern aligned narrowly (in direction) across the estuary. Wave growth is dependent on the fetch length determined by the tide stage that changes the water depths and surface area of the bay. The numerical model reproduces the patterns of wave heights and periods observed from 4 wave buoys, but lag the observations in all cases, especially
when the wind and tide are in opposition. The spatial and temporal distribution of wave-induced bed shear stress was described with analytical relationships and compared with numerical model formulations. Results suggest that mean currents dominate the bed shear stress across the estuary, largely owing to the depth attenuation of small, short period wind waves that do not interact with the bottom except at lower stages of the tide or are diminished by wave current interactions or wave dispersion.

Analytical formulations and model simulations of wave-only induced bed shear stress show that in the shallow portions of the estuary near the fringes of the mudflats, waves are strong enough to induce sediment mobilization into the overlying water column by exceeding a critical threshold for insipient motion (0.10 N/m²; Wengrove et al. 2015). However, the critical threshold for sediment resuspension (0.35 N/m²; Wengrove et al., 2015) was not widely exceeded for wave-induced bed stress, indicating that nutrient fluxes from sediment sources are largely owing to mean currents. Overall patterns of wave growth and spatial distribution of bed-shear stress show that shallow areas are more affected by tidal currents than by wave-induced bed stresses except along the downwind edges of the bay. Analytical formulations based on fetch length and wind velocities reasonably well reproduces the observed wave heights and periods in the middle of the estuary, suggesting that simple applications can provide a good first order estimate of the local wind-generated wave field.
CHAPTER 5
TIDAL ENERGY DISSIPATION IN THREE ESTUARINE ENVIRONMENTS

COASTAL DYNAMICS 2017
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5.1 Abstract

Tidal energy dissipation was examined in three estuaries using a three-dimensional hydrodynamic model (COAWST; Warner, et al., 2010). The modeled M2 tidal amplitude decay and phase lag were estimated at specific locations along transects from the mouths of the estuaries to the furthest inland extent and compared to observations where available. Nonlinear evolution of the tides was qualitatively examined with the spatial evolution of the skewness and asymmetry, and the growth of harmonic constituents. Harmonic constituents and over-tides were estimated from modeled time series of water levels and three-dimensional currents with T_TIDE (Pawlowicz, 2002). Observed evolution of tidal dissipation, harmonic growth, and nonlinear statistics are also well modeled, indicating that the nonlinear evolution of the tides is well represented in COAWST.

Key words: estuarine hydrodynamics, tidal asymmetry, numerical modeling

5.2 Introduction

Astronomical tides are the dominant force in most coastal and estuarine environments, driving the transport of water, sediment, nutrients, and organisms between terrestrial and marine ecosystems. As tides propagate across the ocean and into shallow inlets and bays, they interact with the bottom and become distorted, leading to asymmetries in the duration of the tide and magnitude of the tidal currents, and growth of the tidal harmonics. The asymmetries arise from the inherently nonlinear nature of the tidal shoaling process, leading to the development of local phase lags between pressure and velocities that shift slack tide periods up to 90 degrees (or ¼ wave period). When
averaged over a tidal cycle this asymmetrical behavior determines net sediment transport and circulation patterns (Dronkers, 1986). This behavior has important implications for sediment transport where stronger flood currents drive the movement of coarse sediment and longer slack periods lend themselves to the deposition of fine-grained sediment.

Tidal propagation also leads to amplitude attenuation from energy losses due to frictional interaction with the bottom and geometry of the estuary. Energy dissipation of the tidal wave can be described in terms of amplitude decay of the dominant tidal constituent; the semi-diurnal M2 tide in this study. Not all energy is dissipated due to frictional effects, and some is transferred to higher harmonics (overtides, e.g. M4 and M6) through nonlinear interactions that create tidal asymmetry (Aubrey and Speer, 1985, Speer and Aubrey, 1985). A comparison of the magnitude and phase differences of the M2 with the first harmonic M4 qualitatively describes the tidal asymmetries in the system (Friedrichs and Aubrey, 1988).

In this work, we model the tidal motion in three distinct estuaries with a three-dimensional, high-resolution hydrodynamic model and examine the nonlinear evolution as the tides propagate upstream. The model used is the Coupled Ocean-Atmosphere-Wave-Sediment Transport modeling system (COAWST; Warner, et al., 2008). COAWST includes state-of-the-art atmospheric (WRF) and wave (SWAN) models coupled with the Regional Ocean Modeling System (ROMS), a three-dimensional fully nonlinear hydrodynamic, ortho-curvedinear primitive equation model (Shchepetkin and McWilliams, 2005; Haidvogel et al. 2008). Previous hydrodynamic modeling and observational studies of tidal inlets show that nonlinear advection, nonlinear channel friction and tidal interaction with coastal geometry drive tidal distortion (Dronkers, 1986, Aubrey and Speer, 1985, Speer and Aubrey, 1985), and that evolution of the tides is strongly dependent on the bottom boundary conditions (e.g., MacMahan, et al., 2014).

We will examine the nonlinear tidal behavior that drive tidal asymmetry and energy decay in each estuary, and discuss modeled tidal dissipation characteristics in terms of tidal amplitude decay and phase lags determined from harmonic analysis of tidal constituents. Modeled results are compared with observations where available or with previous results from the literature. Section 5.3 provides site background and model grid development of each estuary, Section 5.4 describes the model and tidal analysis
methodology, Section 5.5 discusses the results in terms of nonlinear evolution, and Section 5.6 summarizes the conclusions of the study.

5.3 Site background and model grid development

Each of the three estuaries in this study are well mixed and tidally driven, and river influences are considered negligible. New River Inlet, NC (Figure 5.1, 1A; discussed in Section 5.2.1) demonstrates progressive wave characteristics and a highly dissipative environment. Hampton Inlet, NH (Figure 5.1, 1B; discussed in Section 5.2.2) demonstrates very different dynamics than New River Inlet, as the tide acts like a standing wave as it propagates inshore on two branches of the estuary, and like a progressive wave on the third. The Piscataqua River-Great Bay system, NH (Figure 5.1, 1C; discussed in Section 5.2.3) is characterized by both progressive and standing wave characteristics, depending upon the section of the main branch of the estuary. In order to characterize the tidal wave properties, time series were extracted from stations within each estuary (Figure 5.1).

Figure 5.1: Model Grid Bathymetry and Station Locations, 1A) New River Inlet, North Carolina, 1B) Hampton Inlet, New Hampshire, and 1C) Piscataqua River-Great Bay estuary, New Hampshire.
5.3.1 New River Inlet, North Carolina

New River Inlet is a part of the White Oak River Basin located in the Carolina Cape region of North Carolina. It serves as an important habitat for birds and fish, an economic resource for commercial fisherman, as well as a strategic location for the Camp Lejeune Marine Corps military base. The New River estuary is a coastal plain system, influenced by barrier islands and the Atlantic Intracoastal Waterway. For decades the inlet has been dredged in order to maintain a navigable channel. The inlet is 1 km wide at the mouth, and has a mean tide range of 1.31 m. Water depths range from 1-3 m at the mouth, over 10 m in the main channel, and again ranging 1-3 meters over an extensive estuarine back bay. The watershed drains an area of 1197 km², and has a surface area of 88 km² (NOAA, 1999). Historically it has been described as eutrophic, with excessive nutrient loading from local wastewater treatment facilities and historical hog waste dumping, (Burkholder et al. 1997; Mallin et al. 1997). Despite recent improvements in both treatment facilities and non-point source loading, eutrophication still persists because New River is a shallow and poorly flushed system.

Observations show that the principal semi-diurnal amplitude decays by about 87% from the mouth to 10 km upstream, consistent with a strongly tidally-choked system (MacMahan, et al., 2014). A simple model balancing the pressure gradient by a quadratic bottom friction formulation suggests that nonlinear interactions induced by the bottom drag modify the amplitude and phase changes of the tide as it propagates upstream. 

**New River Inlet: Model Grid:** For the implementation of COAWST at New River Inlet, we generated a structured orthonormal grid with constant 30 m horizontal resolution and 8 vertical sigma levels. The bathymetric dataset used for this grid was created using a 10-m digital elevation model (DEM) that includes several sources of topography (Lidar) and bathymetry (hydrographic surveys), interpolated using Fledermaus software, and compiled by the researchers at the United States Army Corps of Engineers (USACE) Field Research Facility (FRF).

**New River Inlet: Model Setup:** The model is forced with five tidal constituents (M2, N2, S2, O1, K1). The harmonic constituents were obtained using values from Wrightsville Beach, NC, harmonic station (#8658163), from the NOAA Tide Prediction
service (http://tidesandcurrents.noaa.gov). Tides were forced on the open boundary at the ocean about 2 km from shore. The lateral boundaries within the inlet at the junctions with the Atlantic Intracoastal Waterway (ICW) were left open. For the purposes of this study, no mean flow was forced through the ICW, nor was any river discharge (typically small) from any tributary flowing into the inlet considered. A total of 14 model stations were placed from the open ocean to the head of the estuary, near the town of Jacksonville, NC (Figure 5.1, 1A).

5.3.2 Hampton Inlet, New Hampshire

The Taylor River, and Hampton Falls River feed Hampton River to the north and Blackwater River to the south that drain through Hampton Inlet, a barrier beach system located in southeastern New Hampshire in the Gulf of Maine. With direct and easy access to the Atlantic Ocean, the harbor is home to several commercial fishing and recreation charter boating businesses. The inlet is maintained through regular dredging conducted by the United States Army Corps of Engineers (USACE) and is stabilized by two jetties on either side of the inlet. Hampton Beach is located directly north of the inlet and through predominantly southern alongshore transport, sedimentation builds shoals on the north side of the inlet and form a spit to the south. Tidal currents on flood and ebb tides can exceed 2 meters per second respectively (Mckenna, 2013). Strong currents and active shoaling lead to potentially hazardous navigational conditions. Extensive salt marshes characterize the backbay with several flats used for recreational shellfishing.

Hampton Inlet: Model Grid: The horizontal grid used at Hampton Inlet was similar to New River, with horizontal resolution of 30 m, with 8 vertical sigma levels. The bathymetric grid is compiled from seven different Lidar and hydrographic sources, ranging in years from 1999 to 2015 and acquired from NOAA’s coast dataviewer database and the Western Gulf of Maine (WGOM) dataset compiled at the Center for Coastal and Ocean Mapping-Joint Hydrographic Survey (CCOM/JHC) at the University of New Hampshire.
Hampton Inlet: Model Setup: The model again is forced with five tidal constituents (M2, N2, S2, O1, K1) at the offshore open boundary. The harmonic constituents were obtained using Oregon State University’s global Tidal Prediction Software Package (OTPS) in conjunction with the United States East Coast Regional Tidal Solution (EC2010) (Egbert, 2002). This software package provided the necessary tidal amplitude and phases that correspond to 19 September 2011, thereby coinciding with some of the observational datasets for future model-data comparisons. A total of 21 model stations were placed from the open ocean to the head of each branch of the estuary (North, Middle, and South; Figure 5.1, 1B).

5.3.3 Piscataqua-River Great Bay Estuary, New Hampshire

The Great Bay/Piscataqua River estuarine system is located along the New Hampshire-Maine border. The Little Bay-Great Bay estuary is a recessed, drowned river valley connected to the Gulf of Maine via the Piscataqua River. There are seven major tributaries in this system, including the Squamscott, Lamprey, Oyster, Bellamy, Cocheco, Salmon Falls, and Winnicut rivers. Tidal excursion up these rivers is blocked by dams, which regulate the freshwater input into the system. Overall, the freshwater input is relatively small and only 2% of the tidal prism (Short, 1992; Trowbridge, 2007). The tide range is 2-4 m over the spring-neap cycle with tidal currents exceeding 2 m/s in the channels at maximum ebb and flood tides. At low stands of the tide as much as 50% of the Great Bay is exposed as low-lying mudflats, incised by deeper tidal channels.

Piscataqua-River Great Bay estuary: Model Grid: Again, the horizontal grid resolution is 30 m, with 8 vertical sigma levels. Bathymetric data from several sources (CCOM/JHC, USGS, NOAA, and USACE) were compiled, weighted based on coverage and resolution, and then interpolated to create a composite DEM. The combined elevation data were used with the Easygrid routine to create the model grid (available at https://www.myroms.org/wiki/easygrid). Unlike the other two case studies, this grid required extensive smoothing in areas with steep bathymetry gradients in order to obtain numerical stability.
**Piscataqua-River Great Bay estuary: Model Setup:** The model is forced with five tidal constituents (M2, N2, S2, O1, K1) at the open boundary offshore. The amplitudes and phases for these constituents were obtained through harmonic analysis using T_TIDE (Pawlowicz, 2002) and a surface elevation dataset from 28 March 2006 as a part of the NOAA Marine Aquaculture Program. A total of 21 model stations (save points) were placed from the open ocean to the head of the estuary (Figure 5.1, 1C).

**Piscataqua-River Great Bay estuary: Observations:** Field observations of horizontal currents spanning the water column, sea surface elevation (from bottom pressure and tide gauge), water temperature, and salinity were obtained during field experiments in 2007 and 2015, the long term Great Bay Buoy (http://www.opal.sr.unh.edu/data/buoys/great_bay), and the NOAA Tide Gauge station at Fort Point, NH (Station ID 8423898). Between May and September 2007 bottom-mounted, upward-looking acoustic Doppler current profilers (ADCPs) were deployed in the tidal channel from the mouth of the Piscataqua River to Furber Strait within the Great Bay in water depths ranging between 4.3 m and 19.3 m. A total of 11 different deployments collected current, temperature, and conductivity data for record lengths between 41-45 days (data available at https://tidesandcurrents.noaa.gov). Between August and September 2015 four ADCPs were used in 8 different deployments in Great Bay proper, in water depths ranging between 3 m on the mudflats and 17 meters in the main channel. Currents measurements were sampled continuously, for record lengths between 8-35 days.

5.4 Methods

5.4.1 Hydrodynamic Model

The hydrodynamic model used is the Regional Ocean Modeling System (ROMS) within the Coupled Ocean Atmosphere Wave and Sediment Transport (COAWST) modeling system. ROMS is a three-dimensional finite difference model that solves the Reynolds-averaged Navier-Stokes equations using the hydrostatic and Boussinesq assumptions (Shchepetkin and McWilliams, 2005). Measured bathymetric data were used
to define the model grid (discussed in Section 2). The tidal forcing are ramped up hyperbolically over a 2-day period. The bottom boundary conditions are based on a logarithmic drag law, derived from a characteristic bottom roughness element. A k-ε generic length scale (GLS) turbulence closure model is used to calculate the horizontal and vertical eddy viscosities (Umlauf and Burchard, 2003; Warner et. al., 2005). Each model is run for 30 days with output of averaged data over the whole domain at 30-minute intervals and at specific station locations at 5-minute intervals. Within ROMS the wetting and drying algorithm (Warner, et. al., 2013) is utilized to simulate the inundation and exposure of the mudflats by the tide in shallow areas. The critical depth, Dcrit is set to 10 cm; when the total water depth is less than Dcrit, no flux is allowed in or out of that cell and it is considered “dry”.

5.4.2 Tidal Dissipation Analysis

As the tide propagates into inlets, bays, and other coastal regions, it interacts with the bottom boundary and basin geometry, and loses energy in the form of small turbulent motions. Energy dissipation results in tidal amplitude decay and phase changes that modify the asymmetries of the waveform and flow field. Figure 5.2 shows modeled time series for the sea surface elevation change in the Piscataqua River-Great Bay model case from.

There is a noticeable change between the station closest to the ocean (Fort Point, NH – mouth of the Piscataqua River) to the back-bay area of the Great Bay proper near the Squamscott railroad bridge. Results for New River and Hampton are discussed in Section 4.

The total energy per unit surface area of any tidal constituent is,

\[ E = \frac{1}{2} \rho g A^2 \]  

where \( \rho \) is the density of water, \( g \) is the acceleration due to gravity, and \( A \) is the amplitude of that constituent (dominated by the M2 tide in our estuaries). The amplitude at any location within the estuary, \( A_{\text{station}} \), can be normalized by the ocean amplitude, \( A_{\text{ocean}} \), and squared to represent the fractional energy loss, \( E \), as the tide propagates inland,
Figure 5.2: Modeled time series of the tidal amplitude decay for the Piscataqua River-Great Bay model case.
Figure 5.3: Normalized energy decay and phase change of the M$_2$ tidal constituent as a function of distance from the ocean (inlet) for the Piscataqua River-Great Bay model case.
The normalized energy decay of the $M_2$ tide is shown in Figure 5.3 and demonstrates that about 40% of the $M_2$ tidal signal is dissipated through the narrow, high flow Piscataqua River at a distance about 12 km into the estuary near the entrance to the Little Bay. Further inland over the next 13 km very little energy is lost. Coincident with the energy decay is a nearly linear change in phase to about 45 degrees at the 12 km point, and then nearly constant phase within the Great Bay beyond 18 km in the expansive and relatively shallow Great Bay region with extensive mud flats.

A comparison of this phase change to the higher harmonics, $M_4$ and $M_6$, is an indication of the flood or ebb dominance and asymmetry in the system (see Section 5.3.4). The corresponding results for New River Inlet, NC and Hampton Inlet, NH are shown in Figure 5.6 and discussed in Section 5.4.

5.4.3 Tidal Dissipation Analysis

Time series of sea surface height and depth averaged currents in the $u$ (east-west) and $v$ (north-south) directions were retained at each station. Velocities were rotated to align with the along-channel direction using standard rotary analysis. Figure 5.4 shows a time series of sea surface height (red line), and the along-channel velocity component (black line), for five locations in the Piscataqua River-Great Bay estuary. In the top panel, the velocity lags the elevation time series by approximately 50 degrees, indicating a mix of shoreward progressive and seaward-reflected wave components. The middle panels indicate a shift in phase differences from velocity leading to velocity lagging in the vicinity of entrance to the Little Bay (at the General Sullivan Bridge). The bottom panel, from the upstream extent of the Great Bay near the Squamscott railroad bridge, shows an asymmetric, pitched forward sea surface height profile that leads the upstream directed velocity maxima by almost 90 degrees, consistent with a standing tidal waveform. Estimates of the phase difference between sea surface height and along-channel velocities are shown by the cross spectra in Figure 5.5 for the station nearest the ocean boundary and the most upstream station.
Figure 5.4: Sea surface elevation and along-channel velocity time series from the Piscataqua River-Great Bay model case. Top panel is from the station closest to the ocean, and each subsequent panel is closer to the back bay.
Figure 5.5: Spectral analysis of the Ocean station [Left] and Great Bay station [Right]. Top panel is the spectra of the sea surface elevation (blue) and the along-channel velocity (red), coherence squared, and phase. Significant phases are filled and include confidence intervals and for most of the significant phases the confidence intervals are smaller than the marker size. For the M2 frequency, the phase at the ocean station is about 50 degrees out of phase, and 96 degrees out of phase at the Great Bay station location. Spectra were computed with a Hanning data window and 10 degrees-of-freedom.
5.4.4 Nonlinear Harmonic Growth

The growth of the M4 harmonic relative to the M2 constituent is a measure of the asymmetry and non-linear distortion of the tide (Friedrichs and Aubrey, 1988). Spectra of sea surface elevation time series from three stations spanning the estuary show the growth of the M4 and M6 harmonics (in particular) as the tide shoals upstream (Figure 5.6). Following Speer and Aubrey (1985), the amplitude ratio and the phase difference defined as,

\[ A_{diff} = \frac{A_{M4}}{A_{M2}} \quad (3) \]

\[ \theta_{diff} = 2\theta_{M2} - \theta_{M4} \quad (4) \]

where \(A_{M4}\) and \(A_{M2}\) are the amplitudes of the M4 and M2 sea surface elevation or velocity, respectively, and \(\theta_{M4}\) and \(\theta_{M2}\) represent corresponding phase relationships between the constituents.

In general stronger frictional effects produce larger \(M_4/M_2\) ratios and the phase differences describe flood or ebb dominance. Phase differences between \(0^\circ\) and \(180^\circ\) indicate flood-dominance, and between \(180^\circ\) and \(360^\circ\) ebb dominance. Flood dominant systems have characteristically longer falling than rising tides, and ebb dominant systems have characteristically longer rising tides. Harmonic analysis of the station data provides the phases and amplitudes for these components (discussed in Section 5.5).

5.5 Discussion

5.5.1 Tidal Dissipation

Energy decay of the M2 tidal constituent relative to the value at the entrance is shown in Figure 5.7 as a function of distance from the mouth of each respective estuary. Each region has markedly different dissipation characteristics, with the 92% and 40% energy loss at New River and Great Bay, respectively, whereas energy variation in Hampton shows very little dissipation in the north and middle branches, but 80% loss in the south channel.
Figure 5.6: Spectral analysis of three stations show the growth of the M4 and M6 higher harmonics from the ocean to the bay. Spectra were computed with a Hanning data window and 10 degrees-of-freedom.
Figure 5.7: Relative energy decay and phase change in the M2 tidal signal for all of the model cases.
Relative phase change as a function of distance up the estuary shown in Figure 5.7, corresponds to the relative dissipation differences in each estuary. New River Inlet shows the greatest phase change (~145) whereas Hampton Inlet shows the lowest (<10). New River Inlet acts as a progressive wave, with a sea surface elevation-along-channel velocity phase closer to 0° than 90° (Figure 5.8, top panel), and energy dissipation is high. The north and middle channels of Hampton Inlet, however are more reflective in nature, and show a sea surface elevation-along-channel velocity phase of almost 90°, and corresponds with low energy dissipation. The south channel of Hampton Inlet is similar in nature to New River Inlet, with large energy losses and phase changes in the M2 tide, however sea surface elevation-along-channel velocity phase is more reflective in nature. Further work is needed to determine the dynamics in this channel. The Piscataqua River-Great Bay system, NH lies somewhere in the middle and demonstrates both areas of progressive wave and high dissipation, as well as standing wave, low energy loss dynamics. The transition seems to occur in the Little Bay region, which connects the Great Bay to the Piscataqua River. Further observation and modeling studies are needed to determine the nature of this transition.

5.5.2 Time Series and Harmonic Growth

Upstream evolution of the pressure-velocity phase relationships at the M2 tidal frequency is shown in the top panel of Figure 5.8. Both New River and Great Bay phase relationships show transition from a dominantly progressive wave motion at the mouth (with slack water occurring about 1.7-2.6 hours after high tide near the mouths) to a more standing wave motion (only 0.3 hours difference in the upper parts of the estuary). Conversely at Hampton, slack tides at the mouth occur about 0.2-0.3 hours after high tide, and progressively later as the tide propagates up the estuary (reaching about 1 hour delay far into the south channel).
Figure 5.8: Relative energy decay and phase change in the M2 tidal signal for all of the model cases.
The ratio of the M4 to M2 amplitudes are a measure of the non-linear distortion of the tide (Friedrichs and Aubrey, 1988) and depends strongly on the geometry and frictional features in the tidal channels and mudflats. Both New River Inlet and the south channel of Hampton Inlet show large ratios, and therefore large tidal distortion. In New River, within the first 7 km of the estuary exists the greatest growth of the M4 overtide, corresponding to the greatest energy dissipation in the system. As before, similar dynamics are observed in the south channel of Hampton Inlet. Both systems show low harmonic phase differences and exhibit flood-dominant characteristics.

The growth of the M4/M2 ratio around Great Bay (~20 km in Figure 5.8) aligns with earlier observations of tidal asymmetry growing with distance into the estuary (shown in Figure 5.4). The corresponding phase difference shows a transition towards flood-dominance in the same region as the ratio, possibly due to geometry changes in the estuary. The Piscataqua River is a relatively deep channel with strong tidal currents, and transitioning to a large spatial region of mudflats in Great Bay, may dominate the tidal distortion in this case (Friedrichs and Aubrey, 1988).

5.6 Conclusion

Tidal energy dissipation was examined through tidal amplitude and phase changes as well as harmonic growth in three estuaries using a three-dimensional fully nonlinear hydrodynamic model. Nonlinear evolution of the tides was qualitatively examined with the spatial evolution of the skewness and asymmetry, and the growth of harmonic constituents. Previous hydrodynamic modeling and observational studies of tidal inlets show that the nonlinear evolution of the tides is strongly dependent on the bottom boundary conditions (e.g., MacMahan, et al., 2014).

Modeled tidal behavior in New River is calibrated with previous results based on force balance between pressure gradients and bottom drag and verified with observed elevation time series (MacMahon, 2014). Strong tidal dissipation is evident in the energy decay and phase change in the M2 tidal component as a function of distance from the inlet, shown in Figure 5.7. The high ratio of the M2/M4 shown in Figure 5.8 tide also
corresponds to higher energy losses than the other two estuaries. The M2-M4 phase difference suggests flood-dominance.

In the Piscataqua River/Great Bay Estuary, model bottom roughness (assumed constant over the domain) was calibrated with observations of surface elevation and current time series obtained throughout the estuary. The modeled behavior reproduces a highly dissipative progressive wave in the Piscataqua River with 45% tidal energy decay, and a standing wave low dissipative region in the Great Bay. This is similar to results shown in the literature (~52%, Swift and Brown, 1979) and observations. Future work is needed to determine the spatial variability of the bottom roughness in the model, and how that relates to the spatial variability in the energy dissipation.

Modeled tidal behavior in Hampton shows marked differences in tidal dissipation between channels, confirming previous estimates of limited energy loss in two channels (Ward and Irish, 2014), but with significant energy loss (> 80%) in the third Hampton channel, similar to New River Inlet. Strong spatial variation in the nonlinear evolution of the higher harmonics at Hampton reveals complex tidal shoaling within the shallow back-bay area. The differences in nonlinear tidal evolution between estuaries and channels are attributed to the integrated amount of energy dissipated along different path lengths of the various estuarine branches. Observed evolution of tidal dissipation, harmonic growth, and nonlinear statistics are also well modeled, indicating that the nonlinear evolution of the tides is well represented in COAWST.
CHAPTER 6
CONCLUSIONS AND FUTURE WORK

The goals of this research were to expand our understanding of, and improve predictions of bed shear stress in estuarine environments using both observational datasets and numerical modeling. This was accomplished through an in-depth modeling study of the Great Bay Estuary, NH, using historical and current observational studies.

Chapter 2 described the high-resolution three-dimensional hydrodynamic model (ROMS) that was implemented for the Piscataqua River-Great Bay estuary using observed bathymetry and validated with several observational datasets spanning the estuary. The model was able to reproduce the observed tidal dissipation characteristics including dominant semidiurnal M2 tidal amplitude decay and phase changes, as well as the nonlinear growth of the M4 and M6 harmonics. The modeled behavior reproduces a highly dissipative, partially progressive wave in the lower 12 km of the Piscataqua River (with 45% tidal energy loss by Dover Pt., consistent with previous observational studies;Swift and Brown, 1983), and a (nearly) standing wave in the low dissipative region between Dover Pt. and the upper reaches of the Great Bay. Differences between model simulations with and without subtidal oscillations or river fluxes for the Great Bay are small, suggesting that interactions between the tide and other low frequency (subtidal) or baroclinic flows are weak and can be ignored when considering tidal dynamics.

These results were contrasted with two other estuaries, New River, NC and Hampton-Seabrook, NH. Modeled tidal behavior in New River shows strong tidal dissipation is evident in the energy decay and phase change in the M2 tidal component as a function of distance from the inlet. Modeled tidal behavior in Hampton shows marked differences in tidal dissipation between channels, confirming previous estimates of limited energy loss in two channels (Ward and Irish, 2014), but with significant energy loss (> 80%) in the third Hampton channel, similar to New River Inlet. Strong spatial variation in the nonlinear evolution of the higher harmonics at Hampton reveals complex
tidal shoaling within the shallow back-bay area. Future work is needed to determine the spatial variability of the bottom roughness in the model of the Great Bay Estuary, and how that relates to the spatial variability in the energy dissipation.

Once the hydrodynamic model of the Great Bay Estuary was validated (Cook et al, 2019), the model was then used to understand the spatial variability of bed shear stress in the presence of vegetation in Chapter 3 and with the inclusion of wind drive waves in Chapter 4. Chapter 3 incorporated a vegetation module within the COAWST modeling system to study the effects of vegetation on the distribution of shear stress across the estuary. Model output compares well with observed depth average velocities and shear stress estimates in two locations in the Great Bay estuary. Results demonstrate that incorporating vegetation was an important improvement to the model, whereas higher grid resolution had little effect on estimates, allowing future studies to save computational expense and progress with the coarse 30 m grid. When compared with rivers, model results suggest that internal sources of nutrient loads from sediment were shown to be about 52-60% of that contributed by rivers for at least half of the year (during low discharge summer and fall periods). Including eelgrass in the model lowers the estimates of nutrient loading by 18% and 8.9% for dense vegetation and 30% vegetation, respectively. This study demonstrates that a coupled hydrodynamic-vegetation model is capable of reasonably estimating the distribution of shear stress for a tidally dominant estuary. Future studies in the field of nutrient regeneration in muddy sediments would help constrain the amount of material available on every tidal cycle.

Chapter 4 described observations of surface wind waves characterize the wave climate of the Great Bay Estuary for the first time, and were used to verify model simulations from the coupled ocean-wave numerical modeling system COAWST. The same hydrodynamic parameters from Chapter 2, 3, and 5 were coupled with the wave-generation and propagation model, SWAN. Observations show that the (typical) summer wind field in 2018 was dominated by oscillating land and sea breezes, with significant wave heights are on average between 5.5 – 7.3 cm with spectral peak periods of about 1.52 – 1.66 sec. Analytical formulations based on fetch length and wind velocities
reasonably well reproduces the observed wave heights and periods in the middle of the estuary, suggesting that simple applications can provide a good first order estimate of the local wind-generated wave field. The numerical model reproduces the patterns of wave heights and periods observed from 4 wave buoys, but lag the observations in all cases, especially when the wind and tide are in opposition. Results suggest that during typical summer conditions sediment transport (when incipient motion critical bed stress is exceeded) is dominated by the strong tidal currents and only weakly affected by waves during mid tidal periods, consistent with observational and modeling studies conducted in other estuaries. However, analytical formulations and model simulations of wave-only induced bed shear stress do show that in the shallow portions of the estuary near the fringes of the mudflats, waves are strong enough to induce sediment mobilization into the overlying water column by exceeding a critical threshold for insipient motion (0.10 N/m$^2$; Wengrove et al. 2015). However, the critical threshold for sediment resuspension (0.35 N/m$^2$; Wengrove et al., 2015) was not widely exceeded for wave-induced bed stress, indicating that nutrient fluxes from sediment sources are largely owing to mean currents. There is a need for future observational studies to validate waves in these environments as well as continued observational based estimates of shear stress under various hydrodynamic conditions.

This validated model has been shown to be useful in understanding the spatial and temporal patterns of shear stress in the great bay under various forcing conditions (tides, subtidal, rivers, waves) and incorporated the effects of vegetation which is ubiquitous in estuarine environments. It is my hope that this tool will be useful for future work involving deeper scientific studies into the influences of vegetation and waves with respect to the transport of sediment and nutrients within the Great Bay Estuary.
BIBLIOGRAPHY


APPENDIX A: MODEL SETUP

Model Forcing: Tides
The tidal forcing for the model runs is provided using the Oregon State University Tidal Prediction Software (OTPS; http://volkov.oce.orst.edu/tides/; Egbert and Erofeeva, 2002). This software predicts the tides based on global and regional barotropic inverse tidal solutions using the OSU Tidal Inversion Software (OTIS). All tidal elevations and tidal currents are given relative to mean sea level. This work uses the East Coast of America 1/30 degree regional tidal solution.

Model Forcing: Winds (Surface Stress)
Data was collected from the National Estuarine Research Reserve (NERRS) Centralized Data Management Office (CDMO) online repository (http://cdmo.baruch.sc.edu). The weather station used was the Great Bay Greenland meteorological station (GRBGLMET), located at 43.058768 N and 70.830383 W, and located about 0.5 km from the eastern shore of the Great Bay estuary. The CR1000 logger and all probes (less the rain gauge) are located on a 6.1 m tower, with the rain gauge located on a separate installation nearby. The tower is located approximately 3 m above mean high water.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Units: meter per second (m/s)</th>
<th>Location</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind speed</td>
<td>Sensor type: 18 cm diameter 4-blade helicoids propeller molded of polypropylene</td>
<td>6 m above ground on top of tower</td>
</tr>
<tr>
<td></td>
<td>Model #: R.M. Young 05103 Wind Monitor</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Range: 0-60 m/s (134 mph); gust survival 100 m/s (220 mph)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Accuracy: +/- 0.3 m/s</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Serial Number: WM60353</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Date of Last Calibration: 9/5/2017</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Dates of Sensor Use: 3/7/2018 to present (10/19/2018)</td>
<td></td>
</tr>
<tr>
<td></td>
<td>Serial Number: WM87987</td>
<td></td>
</tr>
</tbody>
</table>
| **Wind Direction** | **Units:** degrees  
**Sensor type:** balanced vane,  
**Model #:** R.M. Young 05103 Wind Sentry  
**Range:** 360° mechanical, 355° electrical (5° open)  
**Accuracy:** +/- 5%  
**Serial Number:** WM60353  
**Date of Last Calibration:** 9/5/2017  
**Dates of Sensor Use:** 3/7/2018 to present (10/19/2018)  
**Serial Number:** WM87987  
**Date of Last Calibration:** 2/20/2015  
**Dates of Sensor Use:** 2/29/2016 to 3/7/2018 | 6 m above ground on top of tower |
| **Photosynthetically Active Radiation (PAR)** | **Apogee Quantum Sensor**  
**Units:** mmoles m-2 (total flux)  
**Sensor type:** High stability silicon photovoltaic detector (blue enhanced)  
**Model #:** SQ-110  
**Light spectrum waveband:** 400 to 700 nm  
**Temperature dependence:** 0.06 ± 0.06% per °C  
**Stability:** <= 2% change over 1 yr  
**Operating Temperature:** -40°C to 65°C; Humidity: 0 to 100%  
**Sensitivity:** 0.2 mV per µmoles m-2 s-1  
**APOGEE**  
**Serial Number:** 11887  
**Date of Calibration:** 8/26/2016  
**Dates of Sensor Use:** 10/20/2016 to present (10/19/2018)  
**Multiplier 0.025** | 6.1 m above ground on southernmost location to avoid shading |
| **Data Logger** | **CR1000 Serial Number:** 5243  
**Date CR1000 Installed:** July 2006  
new; reinstalled: January 6, 2015  
**Date CR1000 Calibrated:** December 23, 2014  
**CR1000 Firmware Version(s):**  
OS 27.05  
1/1/2017 to present (10/19/2018)  
**CR1000 Program Version(s):**  
GRBGLMET_6.5L_04_04_17b.CR1  
4/11/2017 to present (10/19/2018) | 1.8 meters above ground |
Convert wind speed and direction into model coordinates

In order to convert wind speed and direction into u- and v- components of the wind vector we need to first convert the meteorological direction (North – 0 deg, + CW) to a Cartesian coordinate system (East – 0 deg, + CCW)

\[ \theta_{\text{cart.}} = 270 - \theta_{\text{met.}} \]

Now we convert wind speed and direction using

\[ u_x = w_{\text{speed}} \cos(\theta_{\text{cart.}}) \]
\[ u_y = w_{\text{speed}} \sin(\theta_{\text{cart.}}) \]

And finally convert onto our ROMS (xi, eta) grid using

\[ u_{x-\text{ROMS}} = u_x \cos(-\theta_{\text{ROMS}}) + u_y \sin(-\theta_{\text{ROMS}}) \]
\[ u_{y-\text{ROMS}} = u_x \sin(-\theta_{\text{ROMS}}) - u_y \cos(-\theta_{\text{ROMS}}) \]

and to take these wind speed values and convert to a surface stress by using

\[ \tau_{sx} = \rho_{\text{air}} C_d u_{x-\text{ROMS}}^2 \]
\[ \tau_{sy} = \rho_{\text{air}} C_d u_{y-\text{ROMS}}^2 \]

where \( \rho_{\text{air}} \) is the density of air (1.225 kg/m\(^3\)), \( C_d \) is a dimensionless drag coefficient, \( U_{10} \) is the wind speed 10m above the sea surface. The drag coefficient, \( C_d \) is typically found using the following formula from Large and Pond (1981 – J. Phys. Oce. 11, 324-336) assuming a bulk aerodynamic formula based on the assumption of neutrally stable boundary layer
\[
\begin{bmatrix}
C_{DN} = 1.2 \times 10^{-3} & \text{for } 4 \leq U_{10} < 11 \text{ m s}^{-1} \\
C_{DN} = 10^{-3} (0.49 + 0.065U_{10}) & \text{for } 11 \leq U_{10} \leq 25 \text{ m s}^{-1}
\end{bmatrix}
\]

Because the GRBGLMET values of wind speed and direction are 6m above the ground, we need to convert to 10m and remove the dependence on \(z_0\) based on the following equation (Long and Pond, 1981):

\[
u_{10m} = u_{6m} \left(1 + \frac{\sqrt{C_{DN}}}{k} \left\{ln \left(\frac{6m}{10m}\right)\right\}\right).
\]

To convert to a kinematic surface stress we then must take those shear stress values and divide by a reference density, \(\rho_{ocean}\) is the density of ocean water (1025 kg/m³),

\[
\tau_{sx} = \rho_{air}C_d u_{x-ROMS}^2/\rho_{ocean}
\]

\[
\tau_{sy} = \rho_{air}C_d u_{y-ROMS}^2/\rho_{ocean}
\]

**Model Forcing: Wind driven waves in SWAN**

The source of action density due to wind \((S_{in}(\sigma, \theta))\) is described in terms of the sum of the initial linear and exponential growth terms:

\[
S_{in}(\sigma, \theta) = A + BE(\sigma, \theta)
\]

where \(\sigma\) is the frequency, \(\theta\) is the wave direction (CCW from geographical east), \(A\) and \(B\) depend on wave frequency and direction and wind speed and direction. \(A\) is due to Cavalieri and Malanotte-Rizzoli (1981) with a filter to avoid low frequency growth.

\[
A = \frac{1.5(10^{-3})}{2\pi g^2} (U_\ast \max [0, \cos (\theta - \theta_w)]^4 H, H = \exp \left\{-\left(\frac{\sigma}{\sigma_{PM}^\ast}\right)^4\right\}, \sigma_{PM}^\ast = \frac{0.13g}{28U_\ast} 2\pi
\]
where $\theta_w$ is the wind direction, $H$ is the filter and $\sigma_{pM}$ is the peak frequency of the fully developed sea state (Pierson and Moskowitz (1964). $B$ is determined in one of two ways. The first option is through Snyder et al 1981, with rescaled terms of friction velocity $U_*$ by Komen et al 1984. This method relates the drag coefficient to relate to $U*$ to drive wind speed at the 10 m elevation (Wu, 1982 or Zijlema et al 2012) and is represented by:

$$B = \max\left[0, 0.25 \frac{\rho_a}{\rho_w} \left(28 \frac{U_*}{c_{ph}} \cos (\theta - \theta_w - 1) \right) \right] \sigma$$

The second method is from Janssen 1991a and explicitly accounts for the interaction between wind and waves by considering atmospheric boundary layer effects. $BE$ is the exponential growth term following Janssen (1989, 1991a) and based on quasi-linear wave theory

$$B = \beta \left( \frac{U_*}{c_{ph}} \right)^2 \max [0, \cos (\theta - \theta_w)]^2 \sigma$$

where $\beta$ is the Miles constant and is estimated using a dimensional critical height $\lambda$

$$\begin{cases} 
\beta = \frac{1.2}{\kappa^2} \lambda \ln^4 \lambda & \lambda \leq 1 \\
\lambda = \frac{g z_e}{c_{ph}^2} e^r & r = \kappa c/[U_* \cos (\theta - \theta_w)] 
\end{cases}$$

where $\kappa$ is the Von Karman constant and $z_e$ is the effective surface roughness. If $\lambda \geq 1$ then $\beta = 0$. Following Janssen 1991a, the wind profile looks like

$$U(z) = \frac{U_*}{\kappa} \ln \left[ \frac{z + z_e + z_0}{z_e} \right]$$
where $U(z)$ is the wind speed at height $z$ (10m in SWAN) above the mean water level, and $z_0$ is the roughness length. The effective roughness $z_e$ depends on $z_0$ and the sea state through a wave induced stress

$$z_e = \frac{z_0}{\sqrt{1 - \left| \frac{\tau_w}{\bar{\tau}} \right| \bar{\tau}}}$$

$$z_0 = \hat{\alpha} \frac{U^2}{g}$$

where the total surface stress is $\bar{\tau} = \rho_a |\overline{U_*}| \overline{U_*}$, $\hat{\alpha}$ is a constant equal to 0.01 (similar to Charnock style relationship). The wave stress is found using,

$$\overline{\tau_w} = \rho_w \int_0^{2\pi} \int_0^\infty \sigma \sigma E(\sigma, \theta) \frac{k}{k'} d\sigma d\theta$$

where the value of $U^*$ is determined for a given wind speed $U_{10}$ and given wave spectrum $E(\sigma, \theta)$. In SWAN the iterative procedure is used (Mastenbroek et al 1993).

The value of $U_*$ is found from using the wind input $U_{10}$ in the following way:

$$U_*^2 = C_D U_{10}^2$$

Where $C_D$ is found using Wu (1982)

$$C_D(U_{10}) = \begin{cases} 1.2875 \times 10^{-3}, & \text{for } U_{10} < 7.5 \text{ m/s} \\ (0.8 + 0.065 \frac{s}{m} U_{10}) \times 10^{-3}, & \text{for } U_{10} \geq 7.5 \text{ m/s} \end{cases}$$

This formulation for the drag coefficient is reported to work well for wind speeds less than 20 m/s, which are typical wind speeds for this study.
APPENDIX B: MODEL GRID DEVELOPMENT

Grid Development
Bathymetric and topographic data were obtained from several sources; including hydrographic surveys conducted by UNH, USGS, NOAA, and USACE and LIDAR by the USGS and USACE. The most recent and accurate bathymetry was loaded first, gaps were filled with the remaining data, and then the LIDAR was loaded. Gaps were linearly interpolated. The Upper Piscataqua lacked historical data, so the bathymetry in this area was approximated manually. Future hydrographic surveys are planned for this area to include in more accurate grid development. The combined data was linearly interpolated into 10 by 10 meter rectilinear grid cells with 8 vertical layers.

Datasets

Western Gulf of Maine
Western Gulf of Maine: The bathymetric map of the western Gulf of Maine is based on composite of high resolution multibeam echo sounder (MBES) surveys conducted by several government agencies and private organizations. The composite grid used in this grid development was the 8 meter grid. Original gridding and survey sources are included in the metadata for each survey and available at http://ccom.unh.edu/project/wgom-bathbackscatter. Development of the bathymetry and backscatter syntheses were supported primarily by UNH/NOAA Joint Hydrographic Center Award NA10NOS4000073. Partial support was provided by the Bureau of Ocean and Energy Management (BOEM) Marine Minerals Program Award M14AC00010. Larry Ward, Zachary McAvoy, and Maxlimer Vallee-Anziani aided with the analysis and development of the syntheses.
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<tr>
<th>Location</th>
<th>Year</th>
<th>Agency</th>
<th>Vertical</th>
<th>Horizontal</th>
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<tbody>
<tr>
<td>York ME (Job#285186)</td>
<td>2006</td>
<td>FEMA</td>
<td>iElevRef = 7</td>
<td>WGS 1984</td>
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<tr>
<td>York ME (Job#285191)</td>
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<td>FEMA</td>
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<td>NAD83</td>
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<td>Lidar: New Hampshire (Job#285188)</td>
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<tr>
<td>Lidar: New Hampshire (Job#285190)</td>
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<tr>
<td>Piscataqua River</td>
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<td>NOAA</td>
<td>MLLW (meters)</td>
<td>NAD83</td>
</tr>
</tbody>
</table>

Seacoast  

**UNH-NOAA-CCOM-JHC**

*Lippmann Lab survey vessel – Great bay NH, 2015*

In the winter of 2015-2016 the Great Bay was surveyed with the Galen J. The track lines were based on a previous survey conducted in 2009. All instrumentation and specifications were chosen based upon the 2009 survey.
## APPENDIX C: TABLE OF MODEL RUNS

### Chapter 1 (Paper 1)

<table>
<thead>
<tr>
<th>Model Type</th>
<th>Run Files</th>
<th>Tidal forcing</th>
<th>Z0 values</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>TIDES</strong></td>
<td>Ocean_sta_run03v7C_v11zob15.nc, Ocean_sta_run03v7C_v11zob20.nc, Ocean_sta_run03v7C_v11zob25.nc, Ocean_sta_run03v7C_v11zob30.nc</td>
<td>tide_Combined2016_v11.nc</td>
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<td>z0=0.020</td>
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<tr>
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<td></td>
<td></td>
<td>z0=0.025</td>
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<tr>
<td></td>
<td></td>
<td></td>
<td>z0=0.030</td>
</tr>
<tr>
<td><strong>TIDES+SUBTIDAL</strong></td>
<td>Ocean_sta_run03v7C_v11zob15_sub.nc, Ocean_sta_run03v7C_v11zob20_sub.nc, Ocean_sta_run03v7C_v11zob25_sub.nc, Ocean_sta_run03v7C_v11zob30_sub.nc</td>
<td>tide_Combined2016_v11.nc</td>
<td>z0=0.015</td>
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<tr>
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<td>Grid: Combined2016_v11_grd.nc</td>
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<td>z0=0.025</td>
</tr>
<tr>
<td></td>
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<td></td>
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<tr>
<td><strong>TIDES+RIVERS</strong></td>
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<td>riversv3_run03v9A.nc</td>
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### Chapter 2 (Paper 2)

<table>
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<th>Run Files</th>
<th>Tidal forcing</th>
<th>Z0 values</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>TIDES (30 m grid)</strong></td>
<td>Ocean_sta_run03v13test.nc, Ocean_his_run03v13test.nc, Ocean_avg_run03v13test.nc</td>
<td>tide_Combined2016_v11_July2016.nc</td>
<td>z0=0.020</td>
</tr>
<tr>
<td><strong>TIDES+ 30% vegetation</strong></td>
<td>Ocean_sta_run03v13test_vt30p.nc, Ocean_his_run03v13test_vt30p.nc, Ocean_avg_run03v13test_vt30p.nc</td>
<td>tide_Combined2016_v11_July2016.nc</td>
<td>z0=0.020</td>
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</table>
| **TIDES+100% vegetation** | Ocean_{sta}_run03v13\text{test}_vt.nc  
Ocean_{his}_run03v13\text{test}_vt.nc  
Ocean_{avg}_run03v13\text{test}_vt.nc  
Grid: Combined2016\_v11\_grd.nc  
Tidal forcing: tide\_Combined2016\_v11\_July2016.nc  
Input files: gb\_vegetation.in  
GB\_30m\_July2016\_init.nc  

\text{z}_0=0.020 |
|---|---|
| **TIDES (10 m grid)** | Ocean_{sta}_run04H\_bw.nc  
Ocean_{his}_run04H\_bw.nc  
Ocean_{avg}_run04H\_bw.nc  
Grid: Combined2016\_10mv8\_grd.nc  
Tidal forcing: tide\_Combined2016\_July2016\_10mgrid.nc  

\text{z}_0=0.020 |

| **Chapter 3 (Paper 3)** |
|---|---|
| **TIDES+WAVES (Model A)** | Ocean_{sta}_run03v11\_PP3A\_nc  
Ocean_{his}_run03v11\_PP3A\_nc  
Ocean_{avg}_run03v11\_PP3A\_nc  
Grid: Combined2016\_v11\_grd.nc  
Boundary forcing: tides: run03v11\_PP3A\_30m\_bry.nc  
Wind Stress file: frc\_uvstress\_PP3A\_nc  

\text{z}_0=0.020 |
| **TIDES+WAVES (Model F)** | Ocean_{sta}_run03v11\_PP3F\_nc  
Ocean_{his}_run03v11\_PP3F\_nc  
Ocean_{avg}_run03v11\_PP3F\_nc  
Grid: Combined2016\_v11\_grd.nc  
Boundary forcing: tides: run03v11\_PP3F\_30m\_bry.nc  
Wind Stress file: frc\_uvstress\_PP3F\_nc  

\text{z}_0=0.020 |
### APPENDIX D: ROMS MODEL PARAMETERS

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<th>Parameter</th>
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<td>R0</td>
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