Observations of a Developing Boundary Layer in a Tidally Forced Estuary

Meagan E. Wengrove
University of New Hampshire, Durham

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Observations of a Developing Boundary Layer in a Tidally Forced Estuary

BY

Meagan E. Wengrove
Bachelor of Science, University of New Hampshire, Durham, NH, 2010

THESIS

Submitted to the University of New Hampshire
in Partial Fulfillment of
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in
Civil Engineering

May 2012
This thesis has been examined and approved.

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Date 4 May 2012
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ABSTRACT

Observations of a Developing Boundary Layer in a Tidally Forced Estuary

Meagan E. Wengrove
University of New Hampshire, May, 2012

Field observations of sediment resuspension within a developing tidal boundary layer were collected during two field deployments from summer 2011 in a long straight channel of the Great Bay Estuary of New Hampshire. The first deployment observed boundary layer development during typical tidal forcing, while the second deployment monitored the tidal boundary layer development response to Tropical Storm Irene. During a typical flood tide, the flow field over the flat sandy mud bed at the monitoring location is unidirectional. Bed stress estimations suggest that during typical tidal forcing, the Estuary is in local morphological equilibrium, and the boundary layer supports an observable sublayer in the lowest 0.5 cm of the water column, where viscous effects dominate, suggesting that solute flux from the bed is controlled by molecular diffusion. Additional storm forcing generates bed shear stress that exceeds the threshold of sediment motion (0.15 N/m²), suggesting that turbulence controls solute mixing.
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<td>Molecular diffusivity</td>
<td>$D$</td>
</tr>
<tr>
<td>Median grain size</td>
<td>$d_{50}$</td>
</tr>
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<td>Grain size</td>
<td>$d_s$</td>
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<td>Brinkman Layer thickness</td>
<td>$\delta_0$</td>
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<td>Boundary layer thickness</td>
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<td>$g$</td>
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<td>Mass flux</td>
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<td>Bed roughness</td>
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<td>$\rho$</td>
</tr>
<tr>
<td>Density of sediment</td>
<td>$\rho_s$</td>
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<tr>
<td>Reynolds No.</td>
<td>$Re$</td>
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<td>$Re_*$</td>
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<td>$Re_p$</td>
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<td>Symbol</td>
<td>Description</td>
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<td>Sediment relative density</td>
<td>$s = \rho_s/\rho$</td>
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<td>Schmidt No.</td>
<td>Sc</td>
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<td>Bed shear stress</td>
<td>$\tau_b$</td>
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<td>$\tau_v$</td>
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<td>Total shear stress</td>
<td>$\tau_{total} = \mu \frac{\partial u}{\partial z} - \rho u'w'$</td>
</tr>
<tr>
<td>Velocity alongshore (positive South)</td>
<td>$u$</td>
</tr>
<tr>
<td>Freestream velocity</td>
<td>$U$</td>
</tr>
<tr>
<td>Velocity fluctuations alongshore</td>
<td>$u'$</td>
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<tr>
<td>Friction velocity</td>
<td>$u_*$</td>
</tr>
<tr>
<td>Friction velocity from Indicator function</td>
<td>$u_{*\text{Ind}}$</td>
</tr>
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<td>Friction velocity from log profile</td>
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<td>$u_{*v}$</td>
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<td>Across Channel velocity (positive East)</td>
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</tr>
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<td>Kinematic viscosity</td>
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<td>Vertical velocity (positive up)</td>
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<td>$w'$</td>
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<table>
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<th>Description</th>
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<td>$w_s$</td>
<td>Settling velocity</td>
</tr>
<tr>
<td>$x$</td>
<td>Alongshore direction (positive South)</td>
</tr>
<tr>
<td>$y$</td>
<td>Across channel direction (positive East)</td>
</tr>
<tr>
<td>$z$</td>
<td>Vertical direction (positive up, $z = 0$ at bed)</td>
</tr>
<tr>
<td>$z_0$</td>
<td>Roughness length</td>
</tr>
<tr>
<td>$z^+$ or $Z$ or $y^+$</td>
<td>Direction in wall units</td>
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INTRODUCTION

The Great Bay Estuary of New Hampshire is a tidal estuary that drains into the Piscataqua River and eventually into the Gulf of Maine. The Great Bay Estuary is a shallow estuary that is well mixed and follows a sub-critical flow regime, meaning that there is not a large stratification of flow within the estuary (Swift and Brown, 1983; Bilgili et al., 2005). There are seven rivers that feed into the Great Bay; each provides drinking water, areas for recreation, and a waste water outlet for surrounding New Hampshire communities. These specific land uses provide the Bay with nutrient and trace metal loading that can negatively impact the overall ecosystem health, by degrading its native habitat for fish, birds, seals, oysters, eel grass, and other species (Short, 1992). Within the past 25 years, the concentration of dissolved inorganic nitrogen in the Great Bay Estuary has increased by nearly 60%; this demonstrates effective nutrient overloading (NHEP, 2009). To responsibly manage and minimize further decreases in the ecosystem health of the Great Bay, it is essential to have an accurate assessment of the sources of nutrients to the Bay.

Currently, sources of such nutrients are primarily attributed to local land use practices and waste water treatment plant release through its tributaries. The Piscataqua Region Estuaries Partnership continues to estimate the nutrient loads from tributaries and wastewater treatment plants (Trowbridge, 2009). These in-
fluousential sources of nutrient inflow do not account for all of the nutrient loading within the system. Additional contributions to nutrient loading can occur through sediment porewater solute release by means of molecular diffusion through the near bed diffusive boundary layer or erosion of the sediment bed. The net impact of the benthic fluxes is a function of both hydrodynamic forcing and biological processes. The most relevant previous investigation to this type of contribution was completed by Roseen (2002) where Nitrogen loading due to groundwater influence was characterized using thermal imaging. This study concluded that groundwater did not significantly contribution to Nitrogen loading within the Great Bay, where its average contribution was approximately 2.5%, however this nutrient source has been found to be significant within other estuarine systems (Roseen, 2002).

The Great Bay Estuary is made up of several sediment types. Regions of the Bay with fine-grained sediments are considered to be depositional zones and are at high risk for contamination by excess nutrients and trace metals due to settling of fine-grained particles such as organic matter which releases nutrients when respired, and inorganic matter which carry nutrients and metals (Short, 1992). In coastal environments and rivers, the bottom stress induced by currents, waves, and high flow conditions can suspend deposited particles into the water column. These erosional and depositional events have been shown to mobilize deposited sediments and stored nutrients within the sediment bed (Rivera-Duarte and Flegal, 1997; Kalnejais et al., 2007). This release contributes to the overall nutrient loading within the Bay, but currently, is not estimated as part of the nutrient flux within the Bay by The Piscataqua Region Estuaries Partnership nor in many waterways in gen-
eral (Trowbridge, 2009). Quantitative estimates of the magnitude of the nutrient release stored within the sediment can be made by estimating the contribution of molecular and turbulent diffusion to overall solute mixing within the water column in conjunction with measurements of nutrient concentration.

Nutrient mixing into the bulk water occur through the mechanisms of molecular and turbulent diffusion. When molecular diffusion is the rate-limiting step for nutrient mixing, solutes must molecularly diffuse through a thin layer near the sediment bed called the diffusive boundary layer to be mixed into the bulk water. However, the application of excess shear stress due to turbulent boundary layer development will degrade the diffusive boundary layer and force advection and sediment transport, thus controlling overall nutrient mixing of the bulk water by means of turbulent diffusion. It is evident that turbulent boundary layer development determines the mechanism and rate of nutrient mixing. To begin to completely understand The Great Bay ecosystem and to be able to make informed decisions about land use and water quality standards in the area, it is pertinent to consider nutrient loadings from all potentially large sources.

1.1 Physical Mechanisms of Nutrient Release

Physical mechanisms for nutrient release include both molecular and turbulent diffusion, which depend upon the flow regime present. Laminar flow is characterized by flow moving in horizontal layers, and is not well mixed. Various scales of eddies are characteristic of turbulent flow. These eddies range from the size of the entire water column to a few millimeters in diameter and are responsible for significantly
enhanced energy dissipation. As flow moves over a non-frictionless boundary, a boundary layer is created from a no-slip condition or zero velocity characteristic of the fluid at the boundary interface and ranges in a logarithmic profile to the free-stream velocity.

When a boundary layer is developing from laminar flow to turbulent flow, part of the laminar boundary layer is left behind and this is called the viscous sublayer, the viscous sublayer is a near bed region that is usually only a few grain diameters in thickness, but can range to several millimeters thick depending upon the turbulence present in the water column. The region where viscous effects are significant can include the viscous sublayer and buffer layer. Within the viscous sublayer, nutrient mixing is dominated by molecular diffusion (Jorgensen and Revsbech, 1985; Hondzo, 1998). The thickness of the viscous sublayer, buffer layer, and diffusive boundary layer are not only forced by mean velocity, but also depend upon the magnitude of bottom boundary layer turbulence (Lorke et al., 2003).

On occasion, turbulent mixing can dominate the viscous contributions in the near-bed interface, thus creating an effective excess bed shear stress and incipient motion is induced. At this point, turbulent kinetic mixing and turbulent diffusion are the primary mechanisms for solute release and particle mixing (Hondzo, 1998). Bed shear stress or bed stress ($\tau_b$) is defined to be the stress that is calculated to be acting at the boundary, i.e. the stress which the sediment is subject to. Figure 1-1 demonstrates these three mechanisms of nutrient and particle mixing. The rate of nutrient mixing is controlled by the level of turbulence within the water column.
Figure 1-1: Mechanisms of nutrient mixing: (a) Molecular diffusion through the diffusive boundary layer which is a sublayer of the viscous sublayer, (viscous sublayer indicated by the red dashed line). The diffusive boundary layer and viscous sublayer are near boundary sections of the boundary layer profile created by the no-slip condition when the flow regime is developing from laminar flow to turbulent flow. (b) Sweeping eddy which advects nutrients within the diffusive boundary layer into the bulk water, here indicated by a red eddy. This eddy does not cause excess shear stress, however does turbulently mix nutrients into the bulk water. (c) Incipient motion of sediment, here the threshold of critical shear stress is exceeded and sediment is resuspended into the water column advecting nutrients to be turbulently diffused.
1.1.1 Molecular Diffusion of Nutrients

The diffusive boundary layer creates a partial chemical barrier between the sediment interface and the bulk water, thus for mixing to occur, nutrients must diffuse through this layer (Jorgensen and Revsbech, 1985). The diffusive boundary layer is a portion of the viscous sublayer, and both are controlled by bottom boundary layer turbulence. The thickness of each layer decreases with increased turbulence. In the logarithmic layer, the region above the viscous sublayer and buffer layer, the diffusion of solutes into the water column is controlled by turbulent diffusion, $E$. Upon entering the viscous sublayer, the turbulent diffusivity degrades, and at a particular level it will become equal to molecular diffusivity, $D$, of the solute of interest, this level defines the thickness of the diffusive boundary layer. Molecular diffusivity is characteristic to individual solutes, and therefore the thickness of the diffusive boundary layer is unique to individual nutrients (Boudreau and Jorgensen, 2001). Both the viscous sublayer and diffusive boundary layer are dependent upon hydrodynamic conditions, thus with increased turbulence both layer thicknesses will decrease (Jorgensen and Des Marais, 1990; Hondzo, 1998; O’Connor and Hondzo, 2008).

Molecular diffusion is one of the methods for transport of nutrients into the water column when incipient motion does not occur. Within the diffusive boundary layer nutrients are transported via molecular diffusion, as a flux through the layer, $J$. An approximation to the concentration of nutrients within this layer is represented by Fick’s Law,
\[ J(C, z) = -\frac{D}{\phi^2} \frac{\partial C}{\partial z}, \]  

where the change in \( C \) is the change in concentration from the sediment-water interface to the flow above the diffusive boundary layer, \( z \) is the thickness of the diffusive boundary layer and \( \phi \) is porosity of the sediment (Jorgensen and Des Marais, 1990). The thickness of the diffusive boundary layer can be approximated using the Schmidt number, \( Sc = \nu / D \); the method for approximation varies based upon scaling parameters and an inconsistent assumption that advective transport dominates diffusive transport at the top of the diffusive boundary layer (Dade, 1993; Lorke et al., 2003). It is common to relate the thickness of the viscous sublayer and thickness of the diffusive boundary layer to the Schmidt number by

\[ \frac{\delta_v}{\delta_d} \propto Sc, \]  

where \( \delta_v \) is the thickness of the viscous sublayer and \( \delta_d \) is the thickness of the diffusive boundary layer, although there is some debate whether the Schmidt number should be raised to the 1/3 or 1/2 power. So, \( \delta_d \) can be taken as (Boudreau and Jorgensen, 2001; Lorke et al., 2003; Dade, 1993; Hondzo, 1998),

\[ \delta_d = \delta_v Sc^{1/3}. \]  

There exist several methods for estimating the viscous sublayer thickness. With actual velocity data, the height of the viscous sublayer can be measured using both dimensional and non-dimensional velocity profiles. The general equation is,
Table 1.1: Coefficients that define $\delta_\nu$ according to various sources.

<table>
<thead>
<tr>
<th>Source</th>
<th>Value of constant, $c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Boudreaux and Jorgensen (2001)</td>
<td>7.5</td>
</tr>
<tr>
<td>Nielsen (1992)</td>
<td>11.6</td>
</tr>
<tr>
<td>Kundu and Cohen (2008)</td>
<td>5</td>
</tr>
<tr>
<td>Hondzo (1998)</td>
<td>5 to 11</td>
</tr>
<tr>
<td>Lorke et al. (2003)</td>
<td>11</td>
</tr>
<tr>
<td>Dade (1993)</td>
<td>10</td>
</tr>
</tbody>
</table>

$\delta_\nu = \frac{c\nu}{u_* \nu}$, \hspace{1cm} (1.4)

where $c$ ranges from 5 and 11.6 (see table 1.1 for various estimates and sources).

Equation 1.3 assumes that there are not any sources or sinks of nutrients present and does not account for changes in diffusivity due to permeability or porosity. This can be accounted for by including the thickness of the shear-affected zone within the sediment bed, called the Brinkman layer. The Brinkman layer can be approximated by accounting for the permeability of the sediment, $k$ (McClain et al., 1977);

$\delta_0 \sim |k|^{1/2}$. \hspace{1cm} (1.5)

Figure 1-2 provides an illustration of the segments of the bottom boundary layer discussed.

Upon diffusing through the entire diffusive boundary layer, mixing of nutrients is dominated by eddy diffusivity (turbulent diffusion) and dispersion (Boudreaux and Jorgensen, 2001). Eddy diffusion characterizes the mixing of nutrients into the bulk water due to the motion of eddies or turbulence in the water column. The total diffusive flux can be modeled by Fick's law, similar to equation 1.1 but accounts
Figure 1-2: Plot of the relative heights and shapes of the segments of the bottom boundary layer. From bottom to top, Darcy Flow, Brinkman Layer (which accounts for the porosity of the sediment), Diffusive Boundary Layer, Viscous Sublayer, Logarithmic Layer, and bulk flow.

for eddy diffusion, \( E \), as well:

\[
J(C, z) = -[D + E] \frac{\partial C}{\partial z}.
\]  \hspace{1cm} (1.6)

Where \( E \), also known as eddy viscosity, is generally represented by \( E(z) = \kappa u_* z \), where \( \kappa \) is the von Karman coefficient, taken to be 0.41. This relation was seen earlier when introducing the log-layer. Very close to the bed, empirical approximations based on the dimensionless elevation, \( Z = (zu_*/\nu) \), made by the Van Driest equation and the Reichardt equation are more applicable, and can be found in Boudreau and Jorgensen (2001). Figure 1-3 shows the eddy diffusivity versus the dimensionless elevation above the sediment-water interface, \( Z \), according to the Van Driest and Reichardt equations (Boudreau and Jorgensen, 2001). Also shown
Figure 1-3: Plot of eddy diffusivity, $E(Z)$, with dimensionless height $Z = (zu_*/\nu)$ in the bottom boundary layer. Also plotted are a typical molecular diffusivity, $D$, and kinematic viscosity, $\nu$. The intersections of the curved lines and the vertical lines for $D$, and $\nu$ mark the dimensionless thickness of the diffusive boundary layer and viscous sublayer, respectively (Boudreau and Jorgensen, 2001).

are approximations of the molecular diffusion coefficient, $D$, and the kinematic viscosity, $\nu$. The intersection of the eddy viscosity curve with the molecular diffusion coefficient and with $\nu$ is representative of the dimensionless thickness of the diffusive boundary layer, and viscous sublayer, respectively.

It is not realistic to assume that all mixing in the water column will take place due to diffusion. As nutrients are introduced into bulk water, turbulent motions and molecular diffusion will work to mix nutrients into the water column (O’Connor and Hondzo, 2008). The horizontal advection-diffusion equation in Cartesian coordinates,

$$\frac{\partial C}{\partial t} + u \frac{\partial C}{\partial x} = -[D + E] \frac{\partial^2 C}{\partial z^2}, \quad (1.7)$$

this equation accounts for mixing due to the velocity, $u$, of the fluid, and does not
1.1.2 Sediment Resuspension and Nutrient Diffusion

Aside from the relatively slow molecular diffusion there are several mechanisms that contribute to the mixing of nutrients into the main water column. If advection contributes to the mixing process, the rate at which solutes are introduced back into the bulk water from the sediment bed is much faster because turbulence is a far more effective agent of diffusion than molecular motion (Tennekes and Lumley, 1972). In the case of mobile sediment beds, there are two methods in which advection can accommodate faster mixing of nutrients: by means of a sweeping eddy, and by means of incipient motion of sediment.

Sweeping Eddies

When a sweeping eddy extends through the viscous sublayer into the diffusive boundary layer it will travel within the diffusive boundary layer for a short period of time and then burst back into the bulk water carrying nutrients with it that then can mix into the bulk water by means of eddy diffusion (O'Connor and Hondzo, 2008). A sweeping eddy is generally assumed to be a turbulent structure that is two dimensional and not fully developed. O'Connor and Hondzo (2008) approximate a similarity solution to the advection-diffusion equation that predicts subsequent concentrations of nutrients in response to sweeping motions.
Incipient Motion

The incipient motion of sediments is the instant when the critical shear stress is just exceeded and sediment particles start to move along the bed. Shields (1936) defined this as the critical shear stress, where the threshold for particle motion is attained from a ratio between driving and stabilizing forces (Fredsøe et al., 1992). Exact formulations for incipient motion due to the variability of the instantaneous shear stress placed upon a sediment particle are variable (Southard, 2006). To understand incipient motion it is helpful to think of it as two probability distributions, that characterize both the instantaneous local shear stress needed to move the bed, and the instantaneous local shear stress acting upon the bed. As these two probability distributions begin to overlap incipient motion is more likely to occur, and if they overlap completely the entire bed could be moving (Southard, 2006).

Figure 1-4 is an example of the Shields’ Diagram taken from the American Society of Civil Engineers used to characterize the threshold for sediment movement (Southard, 2006). In this instance, the critical Shields parameter was determined for the median sediment grain size. If the critical Shields parameter is exceeded then the sediment should be moving. The critical Shields parameter is defined as:

$$\theta_c = \frac{u^2}{\rho(s - 1)gd_{50}},$$  \hspace{1cm} (1.8)

It should be noted that the Shields Diagram was developed for non-cohesive sediments, and when applied to cohesive sediments it is likely that the shear stress needed for incipient motion to occur is greater than that calculated with the Shields
Turbulent boundary layer

Figure 1-4: Shields diagram, with equation for critical shear stress and critical Shields parameter, \( \theta \) presented on the y-axis. The dark line is the critical Shields parameter. Courtesy of the American Society of Civil Engineers (Southard, 2006).

curve because the cohesive forces of the sediment must also be overcome.

Sediment resuspension is a process of advection, once incipient motion occurs the advection-diffusion equation can be used to understand the mixing of nutrients into the bulk water. When there is sediment transport, it is likely that the solutes stored within the sediment will be introduced in to the water column with the sediment. With advection of sediments and nutrients into the water column, the solutes can be turbulently mixed into the bulk water. This type of turbulent mixing is much faster than molecular diffusion, but in order for the nutrients to completely mix into the water column they must diffuse into the bulk water (Jorgensen and Des Marais, 1990).
1.1.3 Fluid Mechanics of the Inner Boundary Layer

The equation of motion characterizes the flow regimes and can be simplified based on characteristics of flow present in the system (Kundu and Cohen, 2008). If turbulent flow is bounded, the presence of viscosity affects the motion near the bed due to the no-slip condition or zero relative velocity at the boundary (Kundu and Cohen, 2008). This no-slip condition is effectively what creates a boundary layer. The boundary layer is a transitional region between inviscid flow and a boundary where the no-slip condition exists and viscous flow dominates. In boundary layer flows, there are both viscous and inertial forces that are non-negligible. In Cartesian coordinates, the incompressible, horizontal momentum equation is

\[
\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + \nu \left( \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial z^2} \right),
\]  

(1.9)

where \( \rho \) is density, \( p \) is pressure, \( \nu \) is kinematic viscosity, and \( u, v, \) and \( w \) are components of velocity for \( x, y, \) and \( z \) directions, respectively. By making assumptions about the flow regime this equation can be used to describe various flow conditions. However, in turbulent flow there is an enormous range of scales to be resolved, and consequently, a direct solution is not presently possibly for most flow conditions. Often, an approximate solution is accomplished through a Reynolds decomposition, which parameterizes mean and turbulent quantities for every velocity and pressure field (i.e. \( u = \bar{u} + u' \) where \( \bar{u} \) is the mean component and \( u' \) is the velocity fluctuation of the vector \( u \)). The resulting mean momentum equation has an additional stress acting in mean turbulent flow called the Reynolds stress, which is an effective
stress exerted by the turbulent fluctuations on the mean flow, or the rate of mean momentum transfer by turbulent fluctuations (Kundu and Cohen, 2008). Following time averaging over the turbulent component which is indicated by an overbar, the mean horizontal momentum equation in Cartesian coordinates becomes,

\[
\frac{\partial \bar{u}}{\partial t} + \frac{\partial \bar{u}^2}{\partial x} + \frac{\partial \bar{u} \bar{w}}{\partial y} + \frac{\partial \bar{u} \bar{w}}{\partial z} = -\frac{1}{\rho} \frac{\partial p}{\partial x} + \nu \left( \frac{\partial^2 \bar{u}}{\partial x^2} + \frac{\partial^2 \bar{u}}{\partial y^2} + \frac{\partial^2 \bar{u}}{\partial z^2} \right) - \left( \frac{\partial \bar{u}^2}{\partial x} + \frac{\partial \bar{u} \bar{v}'}{\partial y} + \frac{\partial \bar{u} \bar{w}'}{\partial z} \right). 
\] (1.10)

The Reynolds stress has nine components, an example of one component is \(-\rho \bar{u}' \bar{w}'\), which is the component which relates how the horizontal velocity fluctuations are related to the vertical velocity fluctuations. This term is a momentum transfer term and characterizes the momentum at the elevation above the boundary at which it is calculated. Very near the boundary the Reynolds stress should be very small due to the no-slip condition, with a mobile boundary the Reynolds stress will increase (Kundu and Cohen, 2008; Sherwood et al., 2006). Within the turbulent boundary layer, momentum transfer, characterized by the Reynolds stress, will increase with distance from the boundary. A characteristic Reynolds stress profile has a positive slope near the boundary due to the no-slip condition and higher probability that there will be positive velocity fluctuations near the boundary due to the no-slip condition; in contrast, it changes direction to have a negative slope upon approaching the free stream for similar reasoning (Davidson, 2004).

Turbulent boundary layers are often characterized by a logarithmic velocity profile due to the no-slip condition at the boundary. If the boundary layer is fully turbulent the logarithmic profile will reach the boundary. A turbulent logarithmic
boundary layer profile is given by the equation,

$$\frac{u_{*\text{log}}}{\kappa} = \frac{\partial u}{\partial ln(z/z_0)}, \quad (1.11)$$

where $u_{*\text{log}}$ is the friction velocity, $\kappa$ is the von Karman coefficient and has been estimated to be 0.41, and $z_0$ is the roughness length. A measure of the bed stress is given by $\tau_b = \rho u_0^2$, which is a function of the friction velocity. The friction velocity from logarithmic boundary layer profile equation is estimated to act at the boundary, and thus an estimate of bed stress can be made when a turbulent logarithmic velocity profile is present.

When there is a no-slip condition, and the boundary layer is not fully turbulent, viscous forces dominate very near the bed. This internal friction introduces a hydrodynamic viscous sublayer which has a linear velocity profile and stress relationship, the viscous stress is given by

$$\tau_{x\nu} = \mu \frac{\partial u}{\partial z}. \quad (1.12)$$

where $\mu$ is the dynamic viscosity as well as the proportionality constant between the linearly related viscous stress ($\tau_{x\nu}$) and strain ($\frac{\partial u}{\partial z}$) of the velocity profile within the viscous sublayer.

From the sediment-water interface through the entire thickness of the viscous sublayer, the shear stress imparted to the water column is dominated by the viscous stress (equation 1.12), which is derived from the momentum equation when molecular viscous forces are the sole mechanism of movement of the fluid (Kundu and Cohen, 2008). It is the internal friction of the water that creates a viscous sublayer.
at the sediment-water interface (Jorgensen and Revsbech, 1985). This is the region adjacent to the bed, where the greatest changes in velocity and greatest resistance to momentum transfer occur (Boudreau and Jorgensen, 2001). The viscous sublayer is present over beds that are smooth, over rough beds an analogous roughness sublayer is present (Boudreau and Jorgensen, 2001). From this shear stress, $u_\ast$, can be calculated by setting equation 1.12 equal to $\tau = \rho u_\ast^2$. Like the Reynolds stress, the viscous stress acts at the elevation above the boundary at which it is calculated, thus the very near bed estimation can be evaluated to be the bed stress. The viscous sublayer is important because it is said to dissipate much of the energy out of the water column, and it is part of the layer that solutes must diffuse to reach the main water column (Caldwell and Chriss, 1979).

There are several methods used to evaluate the presence of the viscous sublayer. One option, is to assess whether the flow near the boundary can be considered hydraulically smooth. This can be accomplished by evaluating the roughness Reynolds number, $Re_\ast$. $Re_\ast$ is equal to $u_\ast k_s/\nu$ where $k_s$ is the roughness length scale, equal to $2.5 d_{50}$ or $d_{90}$ ($k_s$ should be less than $\nu/ u_\ast$, indicating that granular roughness elements do not protrude through the viscous sublayer) (Soulsby, 1997). The roughness Reynolds number should be less than 3.5 for the bed to be considered hydraulically smooth, which increases the likelihood of the presence of a viscous sublayer. The roughness Reynolds number is a ratio between the inertial roughness forces and the viscous forces, the lower the value the more the viscous forces influence the boundary layer (Dade, 1993; Hondzo, 1998; Boudreau and Jorgensen, 2001). An additional option is to assess whether the sublayer thickness is
greater than the granular roughness elements, which is calculated by equation 1.4 to be greater than $3 * k_s$, this is another representation that granular roughness elements are not protruding through the layer (Boudreau and Jorgensen, 2001). A final assessment requires that the bed be immobile, hence the Shield’s parameter should be less than critical Shields parameter for the sediment (equation 1.8). The viscous sublayer traditionally has the following traits:

A. A linear mean velocity profile. The mean velocity profile should follow a linear curve from the sediment to the interface where the buffer layer reaches the main water column and the velocity profile will shift to follow a logarithmic profile. When analyzing the same profile on a logarithmic scale in the vertical direction, the viscous sublayer will appear to have a convex curvature before it reaches a linear logarithmic distribution region (Caldwell and Chriss, 1979). Figure 1-5 demonstrates these characteristic shapes using field data taken off of the Oregon continental shelf by Caldwell and Chriss (1979) in 200 m of water. Caldwell and Chriss (1979) were the first to collect data in a field marine environment that demonstrated the presence of the viscous sublayer. And Figure 1-6 presents the viscous sublayer plotted with dimensionless velocity values $U/u_*$ against a dimensionless height, $Z$ (Boudreau and Jorgensen, 2001).

B. Skewness of the probability density function of the horizontal velocity. In the presence of the viscous sublayer, the probability density function of the horizontal velocity will skew towards the sediment-water interface. Such that for higher elevations above the bed, the probability density function of the
Figure 1-5: To the left, The curved line in the larger figure and the upper line in the inset represent least-squares log-fit, the sloped line in the inset represents a linear fit; this profile shape is characteristic of the viscous sublayer. To the right, velocity profile above sediment-water interface, log-scale shape characteristic of the viscous sublayer (Caldwell and Chriss, 1979).
time series of the horizontal velocity distribution will begin to look more Gaussian distributed. Figure 1-7 is taken from an Eckelmann (1974) data set that analyzes the viscous sublayer in oil as to make the sublayer thickness and characteristic features more pronounced. Within the viscous sublayer the probability density function becomes increasingly skewed with proximity to the wall, however, very near the wall velocity estimates are biased by the no-slip condition, which overwhelms the probability density function with zero velocity estimates. Thus, probability density functions for velocity estimates in figure 1-7 should be considered from a $y^+$ unit of 2.7 and above. Looking from left to right the distance from the wall increases from peak to peak, i.e. the wall is the y-axis. (Brodkey et al., 1974; Kreplin and Eckelmann, 1979; Eckelmann, 1974).

C. Shape of the third and fourth moment profiles. The skewness profile of the
Figure 1-7: Characteristic probability density function of the instantaneous horizontal velocity demonstrating characteristic skewness of the function as the distance to the wall decreases, medium is oil. The x axis is instantaneous velocity, the y axis is frequency of occurrence, and from left to right the position from the wall is increasing, i.e. the wall is the y-axis of the figure. Different symbols are characteristic of distance from the wall (Eckelmann, 1974).
velocity profile will become very large upon reaching the viscous sublayer, and then as approaching the free stream will become negative. The shape of the skewness is a result of the definition of the no-slip condition. It is a measure of the streamwise velocity fluctuations. Near the boundary, because of the no-slip condition, these fluctuations are positive. Kurtosis or flatness is a measure of the probability that a random value in the distribution is likely to fall several standard deviations away from the mean, and is a measure of peakedness and the heaviness of the probability distribution functions tails. A flatness of a standard Gaussian distribution is 3. Near the boundary where the velocity fluctuations are positively skewed, the flatness should become larger than 3. Characteristic shapes for these curves in and adjacent to the viscous sublayer are demonstrated using dimensionless analysis, and an example of each is presented as figure 1-8 (Alfredsson et al., 1988; Klewicki, 1989; Klewicki and R.E., 1989; Floz and Wallace, 2010).

In the viscous sublayer, stress is mainly attributed to the viscous stress, and Reynolds stress is considered to be negligible. Outside the viscous sublayer, stress is mostly due to the Reynolds stress. The region between the logarithmic layer and the viscous sublayer is defined as the buffer layer, where neither viscous stresses nor Reynolds stresses are negligible. By combining the contribution from momentum transfer due to the Reynolds stress and that of the viscous stress, the total stress is defined with $\tau_{Total} = \mu \frac{\partial u}{\partial z} - \rho \overline{u'w'}$ and can be estimated at any elevation within the water column. The Reynolds stress is a momentum transfer term derived from the momentum equation after completing a Reynolds decomposition. It cannot
Figure 1-8: Characteristic shapes of the skewness (top) and flatness (bottom) of velocity measured in and adjacent to the viscous sublayer. Data presented is from laboratory flumes with mediums of water (\(\triangle\)), wind (\(+\)), and oil (\(\square\)) (Alfredsson et al., 1988). \(y^+ = u_\ast y/\nu\) which is equivalent to \(z^+\) and \(Z\) used elsewhere, \(u = u - \bar{u}\) is the alongshore velocity fluctuations, and \(\bar{u}\) is the mean alongshore velocity.
be considered a stress but it does act like a stress, and thus the total stress term is defined (Davidson, 2004). Keeping this in mind, when the viscous stress is negligible, then the total stress results in the Reynolds stress contribution. The total stress very near the boundary can be taken to be the stress that the bed is subject to, making this the bed stress.

Often, when velocity profile observations near the boundary are not available, the bed stress can be approximated with a quadratic stress law,

\[ \tau_b = \rho C_d \bar{U}^2, \]  

(1.13)

where \( C_d \) is a drag coefficient and \( \bar{U} \) is the free stream velocity. There are several empirical approaches used to estimate the drag coefficient of the boundary and thus the bed stress. These methods relate the roughness length, \( z_0 \), divided by the depth of water, \( h \), to the drag coefficient. Common methods to estimate \( C_d \) include Manning-Strickler, Dawson-Johns, Soulsby, and the full depth log profile, and Colebrook-White (Soulsby, 1997). Figure 1-9 demonstrates each of these method plotted against \( z_0/h \). Estimates of \( z_0 \) when \( u_* \) is not available, is accomplished by approximating \( z_0 \) with grain size as \( z_0 = d_{50}/12 \) for rough turbulent flow (Soulsby, 1997). There are empirical methods used to estimate \( u_* \) which can be used to find \( z_0 \), but none that are explicitly formatted for tidal flows in estuaries.
Figure 1-9: There are several methods used to estimate the drag coefficient, presented here are empirical relations which relate $z_0/h$ to $C_d$ from Manning-Strickler, Dawson-Johns, Soulsby, and the full depth log profile, and Colebrook-White (Soulsby, 1997).
1.2 Project Objectives

In relation to the hydrodynamic component of this research, there were two major objectives:

1. To evaluate the potential for nutrients to re-enter the bulk water from the sediment-water interface by means of molecular diffusion, turbulent diffusion, and/or advection.

2. To evaluate the magnitude of shear stress induced upon the sediment bed in various forcing conditions.

1.3 Hypothesis

During a typical tidal cycle in the Great Bay Estuary of New Hampshire, tidal forcing conditions will induce incipient motion of sediment and thus enhance the release of nutrients into the bulk water column.
CHAPTER 2

OBSERVATIONS

2.1 Field Site

Addressing the scientific goals required to satisfy several geochemical and hydraulic constraints was challenging. The first requirement was that the sediment type be fine enough to store nutrients and metals and be in an area where organic particles were present. However, the sediment needed to be coarse enough to provide a clear delineation between fixed and mobile beds. Thus, it was necessary to find locations in the Great Bay with a sandy mud. Using historical data, as well as recent data a sediment type map of the Great Bay was developed and is presented as Figure 2-1. This comparison between sediment type and local bathymetry provided a basis for assessment of suitable field site locations. Second, analysis of flow conditions is simplified when the flow is unidirectional. Finally, a range of flow fields, and stress magnitudes across the incipient motion threshold was desired.

Two pilot experiments were performed at Wagon Hill Farm and Thomas Point. These sites did not satisfy the above constraints due to either inadequate sediment type or inconsistent large scale eddies within the flow field. Information on particle size distribution for these sites can be found in Appendix A.

The third site evaluated was at Adams Point. The University of New Hampshire
Figure 2-1: The Great Bay of New Hampshire sediment type and bathymetry map, instrument location indicated by pink dot. [Mud (M), sandy mud (sM), gravely mud (gM), muddy sand (mS), sand (S), gravelly sand (gS)] Data courtesy of UNH Prof. Larry Ward (1995), UNH Peter Armstrong (1974), UNH Ata Bilgili (1993, 2000), UNH Prof. Ray Grizzle (2010), NHDES National Coastal Assessment Program (2000-2010), USGS Prof. Lawrence J. Poppe (1986), and compiled by Meagan Wengrove (2011).
Jackson Estuarine Laboratory is located at Adams Point., 2-1 (see pink circle). As seen in figure 2-1, the location of the instrument array is located in a narrow section of the Bay which creates semi-unidirectional flow during the flood tide of the tidal cycle. The sediment type is a sandy-mud and the particle size distribution is presented as figure 2-2. This site was also chosen for its convenient access to power and shelter at the Jackson Estuary Laboratory as the instruments were cabled into the Laboratory.

2.2 Observational Array

To achieve the project objectives observations of velocity profiles very near the boundary and into the free stream are pertinent to understand the hydrodynamic forcing and to estimate the bed stress applied during the tidal cycle. It is also necessary to have an auxiliary method to measure erosion rates of the boundary to verify incipient motion of sediment.

The observational array includes two Nortek Velocimeters, one Nortek Current Profiler, as well as an Imagenex 881a pencil beam sonar. The capabilities of each
instrument along with its use in the field are described below.

- Nortek Vectrino II Acoustic Doppler Profiling Velocimeter (Profiling ADV): This instrument operates at 10 MHz and is capable of sampling up to 100 Hz with a sampling range of 3 cm at a distance of 4 to 7 cm from the transducer with a 1 mm resolution. The evaluation of this instrument is discussed in Appendix B. The Vectrino II was deployed at 6 to 6.5 cm above the sediment-water interface of the Great Bay Estuary, such that the sampling volume extended 0.5 to 1 cm into the bed and 2 to 2.5 cm into the water column. This deployment method allows for characterization of near-bed boundary layer development at a 1 mm resolution.

- Nortek Vector Acoustic Doppler Velocimeter (ADV): This instrument operates at 6 MHz and is capable of sampling a single point measurement at up to 64 Hz. In this experiment, the Vector was used to characterize the mean velocity during deployment in the free stream.

- Nortek Aquadopp High Resolution Acoustic Doppler Current Profiler (ADCP): This instrument is a pulse-to-pulse current profiler which operates at 2 MHz and is capable of sampling over 1 to 2 m with as fine as a 0.7 cm resolution at a rate as fast as 1 Hz for continuous sampling. With instrument configuration there are tradeoffs to accomplish various sampling configurations based upon the experiment objective. For this particular experiment the instrument was sampled over 1.2 m with a 2 cm resolution at 1 Hz continuous sampling for a 6 day autonomous deployment. The Aquadopp HR only sampled during August into September 2011 deployment - storm time data.
• Imagenex 881a with Azimuth Drive: The Imagenex 881a with Azimuth Drive is a variable frequency two-axis pencil beam sonar that is capable of sampling resolution 252 bins over a 1 m to 30 m range length. This instrument can be calibrated based upon sediment type to monitor acoustic backscatter change in the water column based upon sediment resuspension and thus a suspended sediment concentration can be resolved. It can also be used to characterize bottom bathymetry. For this experiment the sampling range was 1 m, where the instrument was 85 cm from the sediment bed and the sampling frequency was 975 kHz. The rotary sonar was used to monitor changes in local bathymetry as well as monitor water column sediment concentrations over a 6 day period. Unfortunately, the rotary head communication failed during the second deployment at Adams Point and other than providing qualifying support for incipient motion, these data will not be discussed.

2.3 Deployment

As mentioned previously, the data obtained at Wagon Hill Farm and Thomas Point sites were discarded primarily due to sediment type or flow conditions that did not fulfill either the chemical nor physical project constraints. The remaining data from approximately 10 days were obtained during two separate deployments at Adams Point out of the UNH Jackson Estuarine Research Laboratory.

Two 6-day deployments of the Vectrino II, Vector, Aquadopp HR (only deployed for second deployment), and Imagenex 881a instruments was completed during the summer of 2011. The first deployment was during early June, and the second
took place during late August into early September. The June 2011 deployment was during calm weather conditions and typical stream discharge rates into the Bay. The August 2011 deployment was during Tropical Storm Irene, which brought wind and rain to the Great Bay area as well as increased the stream discharge rates into the Bay. As seen in Figure 2-1, the orientation of the Bay is such that during flood tide the positive flow direction is South, this will be defined as positive alongshore or positive x-direction. Thus positive across channel or positive y-direction is East of the instrument array, and positive z-direction is oriented from the sediment towards the water surface.

For each deployment there were two separate 3 in. diameter galvanized pipes jetted into the sediment bed. The instruments were fastened to steel and aluminum arms that were attached to collars so that when positioned on the galvanized pipes the instrument can be moved to a desirable height above the sediment boundary and direction into flow. Figure 2-3 shows the locations of the instruments relative to one another and the sediment interface. On the left pipe, the Vectrino II is closest to the bed with the Imagenex 881a directly above it, and on the pipe to the right is the Vector (vertical) and the Aquadopp HR (horizontal). When deployed the Vectrino II and Imagenex 881a were deployed on one pipe, and the Vector and Aquadopp were deployed on a second pipe approximately 8 m North of the first pipe in a location which was approximately the same distance from shore. All instruments were pointed across channel (to the East when looking at figure 2-1). Because the Vectrino II was positioned below the Imagenex 881a and would have been in its acoustic sampling volume without a spatial offset, the direction that
Figure 2-3: The instrument array used during deployment and relative locations of each instrument to one and other as well as the sediment interface. The two instrument poles were located approximately 8 m apart in the alongshore direction, and were not offset from one and other in the across channel direction. Left top, Imagenex 881a sonar. Left bottom, Vectrino II. Right vertical, Vector. Right horizontal, Aquadopp HR. Positive alongshore is $x$ (South), positive across channel is $y$ (East), and positive up is $z$.

the Vectrino II and Imagenex 881a were positioned into the flow was 45 degrees apart in the $x$ - $y$ plane so they did not interfere with one an other. In the vertical direction, the Vectrino II had a sampling volume which was approximately -0.5 cm to 2.5 cm above the bed, the Vector velocity sampling volume was 0.7 m above the bed, and pressure sampling volume was 0.85 m above the bed, the Aquadopp HR sampled -0.1 to 0.94 m from the bed, and the Imagenex was positioned so it sampled from -0.15 to 0.85 m above the bed.
2.4 Field Conditions

New Hampshire is subject to semidiurnal tides, creating a tidal cycle of 12.42 hours. The tidal range at Adams Point in the Great Bay Estuary is approximately 2.5 meters. At the instrument site location, the water depth ranged from 1.3 m to 3.8 m. For all tidal phases and water depths the Froude number, \( Fr = \frac{\bar{u}}{\sqrt{gh}} \), and Reynolds number, \( Re = \frac{\bar{u}h}{\nu} \), were found to be sub-critical and turbulent, respectively, where \( \bar{u} \) is the free stream velocity, \( g \) is the acceleration due to gravity, \( h \) is the water depth at the monitoring location, and \( \nu \) is the kinematic viscosity. These parameters are important to characterize the water column at the monitoring location. During typical tidal forcing the dominant hydrodynamic forcing contribution to the Bay is due to tidal flows (Bilgili et al., 2005). Occasionally, there are also wind or boat waves present. Wind wave generation is duration and fetch limited and does not typically last for enough time to make a significant contribution to mean flow statistics.

As seen in Figure 2-1, just South of the instrument locations, a portion of Adams Point jets out into the main channel of the Bay and becomes an obstruction to unidirectional flow during ebb tide. Consequently, only observations from the flood tide is used to satisfy the research objective of finding the maximum shear stresses applied to the sediment-water interface during a typical tidal cycle, and during storm conditions. A bathymetric and hydrodynamic survey of the across channel at the the instrument location was preformed in late October of 2011 after both deployments were complete and is used to characterize cross channel current. These observations provide insight on overall hydrodynamics at this location, including
current speed and direction throughout the water column. Figure 2-4 shows the bathymetric survey referenced in 2-1, overlaid with the averaged current direction over the first 4 hours of flood tide at the Adams Point site. The current direction is generally in the alongshore direction. Note that the general direction of the current is due South, and the current direction is parallel to the alongshore channel orientation at this location.

Figure 2-5 shows the hourly averaged current magnitude and direction for the second and third hours of flood tide. This data was taken along the transect shown in figure 2-4 using a platform mounted ADCP. The main current is subsurface, and to the West of the main channel, creating faster velocities at the instrument location. Also note that the direction of the current on the West edge of the channel is on average zero radians (near the instrument location), which in this case is positive alongshore, or Southward. These direction profiles demonstrate that the flow through the channel is generally unidirectional during flood tide.

During storm conditions, wind and rain were responsible for significantly altering the hydrodynamics at the field site. Hydrologically, runoff during periods of high rain also increased the drainage into the Bay from the connecting rivers and watersheds, creating increased volume of flow, and flow velocity.

The original hypothesis for this thesis exercised that during a typical tidal cycle, the forcing due to currents should be great enough to surpass shear stress levels attributed to incipient motion, thus resuspending sediments and sequestered nutrients. As will be discussed in this thesis, this was not observed. Thus, it was necessary to complete a subsequent deployment during conditions when incipient
Figure 2-4: Across channel current direction relative to local bathymetry of the Bay. The inset is an overall representation of local bathymetry, where the black line is the transect where channel currents and current direction were surveyed in October of 2011. The main figure highlights the local channel where the field deployment took place. The pink dot is the instrument location, and the black vectors show the average current direction over the first three hours of flood tide. Notice that the current is mainly flowing South. Data was taken with an ADCP courtesy of Prof. Tom Lippmann, Research Faculty, UNH CCOM.
Figure 2-5: (a) Across channel hourly averaged current magnitude and (b) current direction for the second hour of flood tide. (c) Across channel hourly averaged current magnitude and (d) current direction for the third hour of flood tide. A direction of zero radians is alongshore positive or South. Instrument is located at a relative distance along transect of zero meters [pink dot]. Notice that the current is mainly flowing South. Data is taken with an ADCP courtesy of Prof. Tom Lippmann, Research Faculty, UNH CCOM.
Figure 2-6: Non-storm and storm time hydrographs for Oyster and Lamprey Rivers. The left figure is non-storm time hydrographs for both the Lamprey and Oyster Rivers, and the right figure is storm time hydrographs for both River discharges. The yellow highlighted area for each plot recognizes the sampling period for the data set that is analyzed in this thesis. Data is courtesy of the USGS Streamgages.

motion was exceeded, i.e. a storm. Figure 2-6 shows the hygrographs of the Oyster and Lamprey Rivers for a period before and after the September deployment, as well as a highlighted area indicating the time when the detailed analysis is shown. Figure 2-7 displays wind roses for each deployment. The combination of enhanced runoff and high winds were not inconsistent with observed enhanced flows.

The local velocity magnitude during mid-flood tide was 0.28 m/s during the non-storm condition and 0.35 m/s during the storm condition. The critical Shields parameter is $\theta_c = 0.087$ for $d_0 = 113\mu$m. During the non-storm time, $\theta = 0.016$ and during the storm conditions $\theta$ is between 0.2 and 0.37 during mid-flood tide. Providing further support for an immobile bed during the non-storm condition and
Figure 2-7: Wind Rose for non-storm time (left) and storm time (right) for the Great Bay of New Hampshire, wind speed given in m/s. Data courtesy of the Great Bay Coastal Buoy from the UNH Jackson Estuarine Research Laboratory.
a mobile bed during the storm condition. If so, this leads to an observation that diffusion will dominate nutrient release during non-storm times and advection will dominate mixing during storm times.
CHAPTER 3

RESULTS AND DISCUSSION

Figures 3-1 and 3-2 provide a general summary of all the tidal elevation, near free stream velocity, and boundary layer profile observations obtained at Adams Point during non-storm (June 2011, year day 165 through 171), and storm conditions (August and September 2011, year day 239 through 245), respectively. During the storm condition deployment an abundance of eel grass was ripped from the Bay bottom and entangled around the near-bottom Vectrino II instrument, causing usable data to be limited. Sampling periods where eel grass was present on the Vectrino II were correlated to low acoustic correlation and poor amplitude return, and subsequently not considered for further analysis. Consequently, only a subset of the two data sets will be considered. The Vector single point ADV and Vectrino II profiling ADV were available for both deployments, while the Aquadopp HR ADCP was only available for the storm condition deployment.

Figures 3-1 and 3-2 show that during flood tide the velocity magnitudes increase from low tide to mid-tide, where the velocity magnitudes are at a maximum, and then decrease as slack tide at high tide is approached. However, during ebb tide, this effect is not observed. The velocity magnitudes for all three instruments show a little relative change over the ebbing tides (please note, this is easiest to see when looking at the Vector [blue] and Aquadopp HR [green] data during storm conditions). This
Figure 3-1: Overview of all non-storm condition data recorded at Adams Point recorded in June 2011, year day 165 through 171. (a) Predicted tidal heights [black] from the University of South Carolina Tide Predictor (Pentcheff, 2012), and the recorded water depths from the Vector [blue], yellow bands represent usable data records. (b) Relative magnitude of velocity averaged hourly, mean and standard deviation of velocity data from the Vector single point ADV [blue], which had a sampling volume of 0.7 m from the bed; and Vectrino II profiling ADV [red] with sampling volume within 9 cm of the bed. The cyan box highlights the data sets that are discussed in this thesis.
Figure 3-2: Overview of all storm condition data recorded at Adams Point during Tropical Storm Irene. (a) Predicted tidal heights [black], and the recorded water depths from the Vector [blue], yellow bands represent usable data records. (b) Relative magnitude of velocity averaged hourly, mean and standard deviation of velocity data from the Vector [blue], which had a sampling volume of 0.7 m from the bed, and Vectrino II [red] with sampling volume within 3 cm of the bed. (c) 1 m alongshore velocity profiles taken with the Aquadopp HR ADCP [green], this instrument was only available during the storm condition deployment. From these plots, notice the relative magnitude of velocity from the flood phase compared to the ebb phase of the tide. The cyan boxes highlight the data sets considered in a more in-depth manner in this thesis.
effect is likely due to the local topography at Adams Point as discussed in Chapter 2. During ebb tide, the flow field is directed into the channel and away from the instrument site. This asymmetry in the tidal flows could additionally be affected by the baroclinic and tidal barotropic pressure gradients acting together during the flood phase, but in opposition during the ebb phase. This causes the flood phase boundary layer to become more fully turbulent and yield higher velocities than the ebb phase (Jay and Musiak, 1996). However, since the Great Bay Estuary has been found to be fully mixed (Bilgili et al., 2005), this may not be a contributing factor to the discontinuity between the Adams Point flood and ebb tides.

It is also evident from the standard deviation of the velocity profiles that the only apparent waves occurred at low tide, at the beginning of flood tide (please see Appendix C, which demonstrates spectra in support of this observation).

This thesis will be explicitly focussed on the three flood tidal cycles taken during year days 171 and 240 of 2011 (please see the cyan boxes in figures 3-1 and 3-2). Results of non-storm time data and storm time data will be presented simultaneously to allow for visual comparison of non-storm time and storm time effects. In each case, the Vectrino II was positioned with its sampling volume overlapping the bed. Results from an additional flood tidal cycle taken on year day 167, where Vectrino II sampling volume was 5.5 to 8.5 cm above the bed along with the year day 171 data set where the sampling volume was -0.5 to 2.5 cm above the bed were used to evaluate the Vectrino II performance and are shown in Appendix B.
3.1 Overview of non-Storm and Storm Statistics

Figure 3-3 shows the non-storm water depth, velocity at \( z = 0.7 \text{ m} \) [blue] and \( z = 1.4 \text{ cm} \) [red], and change in bed elevation for the flood tide during year day 171 (highlighted in cyan in figure 3-1). The mean horizontal velocity ranges from less than 1.5 cm/s at low tide to 40 cm/s at mid-tide. The standard deviation over each of the hourly averaged records ranges from 1 cm/s at low tide to 3.1 cm/s at mid-tide, large variability in standard deviation is a measure of waves, turbulence, and trends. Change in bed elevation was calculated using the bottom check feature of the Vectrino II, which uses signal amplitude return to evaluate where the bed is located in relation to the instrument transducer. The change in bed height was calculated by subtracting the mean of the first twenty minutes of bottom check data from the record in order to show the relative change in bed elevation. The bed elevation remains undisturbed over the 6 hour flood phase of the tide for the non-storm condition.

Figure 3-4 shows the water depth, velocity and change in bed elevation for the storm conditions of year day 240. In this figure, the Aquadopp HR profiler data is also available, and therefore a velocity profile of the lower 1 m of the water column is shown (see additional panels in figure 3-4). The hourly averaged velocity profiles for the alongshore and across channel directions of flow, and the direction of the current in relation to alongshore velocity during flood tide (positive alongshore is South) are also given. If the current is moving in the positive alongshore direction the plot color is green (zero radians, or South). For every degree the current rotates East of South at the instrument location, the current magnitude will rotate to the
Figure 3-3: Overview of year day 171 data recorded at Adams Point non-storm conditions. (a) Predicted tidal water depths [black], and the recorded water depths from the Vector pressure gage [blue]. (b) Relative magnitude of velocity averaged hourly, mean and standard deviation velocity data from the Vector [blue], with sampling volume of 0.7 m from the bed and Vectrino II [red] with sampling volume between -0.5 and 2.5 cm above the bed. (c) Change in bed elevation over the data collection period.
red spectrum, for every degree West of South the current will migrate to the blue spectrum, where $-\pi$ and $\pi$ indicate North or negative alongshore. During flood tide most of the flow over the profile is moving in the alongshore direction as apparent by the zero radian current direction between the alongshore and across channel velocities. These local observations are comparable to the direction of the current through the across channel transect as presented in figures 2-4 and 2-5. The direction of the flow at this location suggests that the flow is primarily unidirectional in the alongshore direction. During flood tide (green) the flow is moving Southward, and during ebb tide (red or blue), the flow is moving Northward but very low in magnitude because of the geometry of the Bay and low flow conditions. During the periods of flood tide there is a reversal in flow direction as indicated by the blue banding, that could be attributed to large scale eddies within the current. This could be attributed to a large vertical sweeping eddy, however, the length scale of the eddy would need to be rather large for the period and duration of the eddy to persist, making a channel standing wave motion or a large eddy which shed from the sharp bend in flow direction further upstream of the instrument location other possibilities. Finally, it is apparent from the change in bed height that during storm conditions the bed eroded over 1 mm (year day 240). Change in bed elevation for the storm condition was calculated using the same method presented for the non-storm condition. Between the two storm time conditions, the later flood tide storm condition for year day 240 is more mild than the earlier flood tide storm condition as evident through observations of lower velocity magnitudes and sediment deposition during the later flood tide storm condition. Therefore, the later flood tide storm
condition analyzed will be referred to as the waning storm condition, while the earlier, more energetic, flood tide storm condition analyzed will be referred to as the peak storm condition.

Figure 3-5 and 3-6 show the linear and logarithmic near-bed alongshore velocity profiles, respectively, from the Vectrino II. These are six hourly averaged records over flood tide for the near bed data sets (year day 171 and 240), non-storm and storm conditions. At the top of each figure is a representative color coded flood tide chart, to relate the phase of the tide to each profile. As the tide floods the Bay, it is evident that there is a lag time during a portion of the first hour of flood tide, where the flow field is gaining momentum, and the boundary layer begins to develop. During the second hour of flood tide, the velocity field is increasing in magnitude. The mean velocity continues to increase until the record before high tide (blue symbols). From these velocity profiles, the storm conditions yield a more hydraulically well-mixed water column with a logarithmic velocity profile, especially evident in the peak storm (yd 240) flood tide profiles during peak storm conditions. In the peak storm condition, there is a bend in the upper bins of the velocity profile, this is potentially attributed to acoustic error, and is not considered to be characteristic of the hydrodynamics of the record period. The non-storm conditions (yd 171) suggest a sublayer consistent with that of Caldwell and Chriss (1979) during most of the phases of the flooding tide. The waning storm condition observations (yd 240) also show a sublayer for the second two hourly tidal phases (see figure 3-5 and 3-6 (i) and (j)). Once the the green and blue profiles (fourth and fifth hours (k) and (l)) are reached, the boundary layer takes on a characteristic
Figure 3-4: Overview of year day 240 data recorded at Adams Point storm conditions. (a) Predicted tidal water depths [black], and the recorded water depths from the Vector pressure gage [blue]. (b) Relative magnitude of velocity averaged hourly, mean and standard deviation velocity data from the Vector [blue], with sampling volume of 0.7 m from the bed and Vectrino II [red] with sampling volume between -0.2 and 2.7 cm above the bed. (c) and (d) Alongshore and across channel profiles from the Aquadopp HR [green], with velocity comparisons from the Vector [blue] plotted on top of the profile. (e) Profile of the current angle taken from the Aquadopp HR data. (f) Change in bed elevation over the data collection period.
Figure 3-5: Linear hourly-averaged velocity profiles for year days 171 and 240, non-storm and storm conditions matched with change in tidal phase by color. (a) Typical [black] and observed water depth. (b - g) Non-storm condition (year day 171) (□), (h-m) Waning storm condition (year day 240) (○), (n-s) Peak storm condition (year day 240) (▽).

logarithmic shape indicating a fully turbulent boundary layer. During the waning storm condition, as seen in figure 3-4, there was possible deposition occurring over the first half of the flood tide before tidal forcing accelerated the flow field. This event may have changed the density of the water in this near bed region, mimicking a sublayer in the velocity profiles. Characteristic profile shapes will be discussed more in the Boundary Layer and Viscous Sublayer section.

To put these data into perspective, figure 3-7 relates the phase of the tide to
Figure 3-6: Representation of velocity profiles for year days 171 and 240, non-storm and storm conditions in logarithmic scale. Here the hourly averaged Vector velocities (near free stream) are also presented with a blue outline, in contrast to the black outline used to present Vectrino II data. (a) Typical [black] and observed water depth. (b - g) Non-storm condition (year day 171) (☐), (h-m) Waning storm condition (year day 240) (○), (n-s) Peak storm condition (year day 240) (▽)
statistics for potential of sediment movement. The average boundary layer thickness is shown for each hourly averaged phase of the tide, using $\delta_{99}$ from the Aquadopp HR profile data, $\delta_{99}$ is calculated by finding the boundary layer elevation that has a velocity that is 99% of the freestream velocity, $\delta_{99}$ values presented are an average for all flood tide records during the second storm condition deployment. A boundary layer can be considered turbulent when the shear velocity Reynolds number is greater than 1, $Re = \frac{u_*^2}{\nu}$ (Kim et al., 1987). The $u_*$ to calculate Re is taken from within log-fit for the Vectrino II for all conditions. This shows that for all data discussed, the near-bed boundary layer is fully turbulent at the instrument location. The Shields parameter, as outlined in Chapter 1, is presented to characterize incipient motion of sediment, where the critical Shields parameter at Adams Point is $\theta = 0.087$. During both the peak and waning tidal cycles of the storm, the Shields parameter surpasses critical Shields, suggesting at least incipient motion of the sediment bed. However, please recall that the Shields parameter was developed for non-cohesive sediments, and the sediment at the Adams Point site is cohesive. With this stated, the critical Shields may be higher than calculated. This fact will not change the prediction that the sediment during storm conditions will move because the calculated Shields parameter for these conditions (between 0.2 and 0.37) far exceeds critical Shields, thus incipient motion is reached, this is consistent with the observation that the bed was eroding at this site. Also notice that the Shields parameter for the peak storm is less than that for the waning storm, this could be due to the no-slip condition being exceeded during the peak storm condition, which effectively reduces the friction velocity estimate within a
Figure 3-7: (a) Typical [black] and observed water depth. (b) \( \delta_{99} \) boundary layer thickness estimates indicated as *. (c) Shear velocity Reynolds number for the water column at the instrument location. (d) Shields parameter for each hourly averaged record, where the red dashed line in the lower panel indicates critical Shields of \( \theta = 0.087 \). The (\( \square \)) is for non-storm condition (year day 171), the waning storm conditions (\( \circ \)), and the peak storm condition data, (\( \nabla \)) (year day 240).

Sediment characteristics can offer further insight on hydraulic conditions. The particle Reynolds number is a measure of the inertial effects of a settling grain related to the viscous effects of the surrounding fluid and is given by \( Re_p = \frac{w_s d_5}{\nu} \), where \( w_s \) is settling velocity. For the Adams Point \( d_{50} \) of 113 \( \mu \)m, the setting velocity was found to be 11.3 mm/s resulting in \( Re_p = 1.21 \). The Rouse Number \( (Ro = \frac{w_s}{\kappa u_*}) \), characterizes the relative balance between the particle settling and the turbulence present to keep the particle in suspension. The higher the Rouse Number, the less suspended sediment in the water column. For the non-storm condition this number ranges from 6.2 to 11.5 and for storm condition it ranges...
from 1.1 to 3.2. Udo and Mano (2011) suggest that a Rouse Number of 5 indicates bed load, 2 indicates suspension, and 0.1 indicates wash load. This would suggest that even if the sediment bed were in motion in the non-storm condition, there would not be enough water column turbulence to maintain suspension. However during the storm condition, when the incipient motion threshold was exceeded, there was enough water column turbulence for suspension to be maintained.

3.2 The Boundary Layer and Viscous Sublayer

As the current accelerates past Adams Point, the friction of the boundary satisfies the no-slip condition and retards the flow field. During the non-storm condition, this no-slip condition creates the effect of a sublayer with significant influence of viscous effects.

(Caldwell and Chriss, 1979) found the viscous sublayer and buffer layer to be present in the bottom 6 mm off of the Oregon Continental Shelf. This relationship is modeled by the viscous stress equation, equation 1.12, where the bed stress is linearly proportional to the strain, $du/dz$, by the dynamic viscosity, $\mu$. Upon reaching the top of the viscous sublayer, the profile starts to take on a logarithmic shape, and stress in this region can be modeled by the log-layer equation, equation 1.11, paired with the bed stress equation, $\tau_b = \rho u^2$. When comparing the shape of the boundary layer with linear and log scales from figures 3-5 and 3-6 to the Caldwell and Chriss (1979) viscous sublayer characteristic profiles in figure 1-5, evidence of the visible viscous sublayer is clear. In the viscous sublayer, there is a linear relationship between velocity and distance from the boundary. There is
visual support for a linear profile from the boundary to 3 to 5 mm above the bed for the non-storm condition profiles (yd 171). This region may include the buffer layer, which still is subject to the viscous effects as it transitions into bulk flow where viscous forces are negligible. In the non-storm condition this linear, viscous region is present throughout the flood tide, however, it is difficult to determine whether this region is present during low tide because of the very low flow rates during this tidal phase.

The magnitude of the velocity profile between non-storm and storm conditions is of comparable magnitude, but during storm conditions the boundary layer takes on a characteristic logarithmic shape for the peak storm condition and the later half of the flood tide for the waning storm condition. This can be observed more readily on the logarithmic scale plot, figure 3-6, because in logarithmic scale the log-layer will look linear. The storm-time profiles have a logarithmic shape for the peak storm condition, showing evidence of the boundary layer developing into completely turbulent flow. The viscous sublayer, when observable, shows a distinct convect curvature on logarithmic scale, which is seen in the non-storm time condition and during the first three hourly averaged phases of the flood tide during the waning storm condition for year day 240, comparable to that of Caldwell and Chriss (figure 1-5). In the logarithmic profiles, the free stream velocity measured from the Vector, is also presented. The free stream velocity was larger in magnitude during the storm condition when compared with the non-storm condition. This storm was fairly mild when comparing the external forcing conditions associated with it to other storms within the area over the past year. As suggested, during typical tidal forcing,
sediment at Adams Point is at the brink of incipient motion, and even a mild storm can force sediment resuspension.

Another way to examine the velocity profile is in non-dimensional wall units, or \( Z \) units (or \( z^+ \) units). The characteristic velocity profile for the viscous sublayer in \( Z \) units is presented as in figure 1-6. In this unit style, all velocity profiles should collapse upon one another, the calculated \( u_* \) shifts the entire profile vertically and horizontally based upon how it scales with the profile velocity and distance from the boundary, respectively. The non-dimensional wall unit velocity profile for the non-storm condition (yd 171) and waning and peak storm condition (yd 240) are shown in figure 3-8. For each data set the low tide velocity profile (magenta) is not plotted because of the exceptionally low flow rates.

Figure 3-8 demonstrates that these three data sets collapse fairly well into the characteristic dimensionless velocity profile as presented in 1-6. The collapse is not perfect, but not unexpected given variations in field conditions and the approximate error associated with the calculation of the friction velocity. When the viscous sublayer was observable, the data shows the best collapse and supports estimates of \( u_* \). The linear portion above 20 wall units on the characteristic curve is also visible in figure 3-8. This linear portion is the logarithmic region of the velocity profile. Figure 3-8 also demonstrates the presence of the viscous sublayer below 10 wall units, as evident by the curvature of the profile, and is also comparable to the characteristic profile. This type of chart is called a Clauser chart. The Clauser chart follows a universal logarithmic form in the overlap region of the boundary layer and it was also originally based upon smooth wall, laboratory data (Wei et al., 2005).
Figure 3-8: Velocity profile for each tidal phase (with the exception of low tide) presented as a collapsed velocity profile in non-dimensional wall units. Non-storm conditions for year day 171 (□), the waning storm conditions (○), and the peak storm condition data, (∇) (year day 240). The color of each hourly averaged profile corresponds to previously presented tidal phase plots. And the solid black line is the typical collapsed velocity profile as seen in figure 1-6.
With this caveat, the fit which this field data portrays compared to the accepted collapsed form is reasonable.

To calculate non-dimensional velocity profiles, the friction velocity must be found. For the non-storm profile the friction velocity found from the viscous stress was used where \( u_* = \sqrt{\mu \frac{du}{dz}} \), and for the storm profiles the friction velocity found from the log-layer was used where \( u_{*\log} = \frac{kdu}{dln(z/z_0)} \). \( u_* \) is a function of the velocity gradient in the viscous sublayer. To find the \( u_* \) associated with the log-layer there is the traditional method presented as equation 1.11 or the indicator function can be used by plotting \( u_{*\text{ind}} = \frac{du}{dz} z \kappa \) against distance from the bed, \( z \). Figure 3-9 show the Indicator function for the non-storm condition and storm condition data. \( u_{*\text{ind}} \) is found by finding the flat portion of the indicator function after the maxima and averaging the flat portion as a function of distance from the bed, this value is \( u_{*\text{ind}} \). It compares very closely to that of the log-layer \( u_{*\log} \) since the indicator function is the derivative of the log-layer equation. Using the indicator function, the value of the friction velocity, \( u_* \), can accurately be found graphically without dependence upon \( z_0 \) (Orlu et al., 2010).

The friction velocity is used to calculate several parameters which can further suggest the presence of the viscous sublayer. Aside from the shape of the velocity profile, the consequences of viscous effects are seen within the roughness Reynolds number, the estimated thickness of the sublayer, and values of \( u_* \). Each of these parameters are calculated using \( u_* \) from either the log-fit, indicator function, or if applicable, the viscous stress.

The presence of the viscous sublayer is characterized by the hydraulic smooth-
Figure 3-9: (a) Indicator function for non-storm condition year day 171. (b) Indicator function of the waning storm condition year day 240. (c) Indicator function for the peak storm condition year day 240. Find flat portion of graph and continue over to y-axis to find value of $u_{\text{ind}}$ for each tidal phase. Color of lines indicate phase of tide as seen in previous figures.
ness of the boundary layer. The roughness Reynolds number characterizes the hydraulic smoothness of the boundary layer. For the viscous sublayer to be present the roughness Reynolds number must be less than 3.5 \( (Re_* = u_* k_s/\nu) \). Figure 3-10 shows that the roughness Reynolds number is less than 3.5 for all conditions, which introduces the possibility of a visible viscous sublayer. An observable viscous sublayer also requires that the sublayer thickness be greater than the physical roughness of the bed elements. Figure 3-10 demonstrates that during the non-storm condition (yd 171), \( \delta_\nu \) is greater than \( 3 \times k_s \), and during the storm conditions (yd 240), all viscous sublayer thicknesses calculated are less than \( 3 \times k_s \), suggesting that the granular roughness elements of the bed are protruding beyond the viscous sublayer. Please note, the Nikuradse roughness, \( k_s \) is used to characterize the scale of the bed roughness by \( 2.5d_{50} \) or \( d_{90} \) which both happen to be approximately 285 \( \mu \text{m} \).

In this figure, \( \delta_\nu \) for the non-storm condition was calculated as the average value between \( 11.6\nu/u_* \) (high) and \( 5\nu/u_* \) (low), where range bars are presented from the high value to the low value. The calculated thickness of the viscous sublayer for non-storm condition is between 1 and 2 mm, this estimate does not include the added thickness which the buffer layer contributes. For the storm conditions, \( \delta_\nu \) was calculated as \( 11.6\nu/u_* \) to show that even with the highest estimate of sublayer thickness, the storm time conditions have a sublayer thickness that is less than \( 3 \times k_s \), indicating that the granular roughness elements protrude through the viscous sublayer for the entire flood tide during storm conditions.

Finally, values of calculated \( u_* \) are compared between the three data sets. For year day 171, non-storm condition, when the viscous sublayer is potentially vis-
ible, there is a $u_*$ value calculated from the viscous method (hollow □) and the log-layer method (opaque □). Both methods used to calculate $u_*$ in the non-storm time are comparable, especially for the first three phases of the tide, this is another indication that the viscous sublayer is present. The calculated $u_*$ values for year day 240, storm condition, are calculated with the log-layer or indicator function fit. It is apparent that $u_*$ values during storm times are larger than those during the non-storm condition. $u_*$ values for the peak storm condition could be of smaller magnitude than those for the waning storm condition because the no-slip condition is exceeded during the peak storm condition. (Please note, values of $u_*$ for non-storm conditions are from the viscous fit, while $u_*$ for storm conditions are attributed to the log fit for all calculations of $Re_*$ and $\delta_v$.)

Effects of a viscous region are also evident when evaluating the higher order statistics. The probability density function of the velocity as a function of distance from the boundary will portray positive skewness as approaching the boundary if the viscous sublayer is present. Figure 1-7 is an example of this trait. Figure 3-11 shows the probability density function for the lower eight velocity profile bins just above the boundary for each condition at mid-tide. The non-storm condition probability density function shows the most visual skewness as the boundary is approached. The actual values of skewness for each bin are presented in a table within each figure. The storm condition probability density function distributions also show some skewness as the boundary is approached, but not as much as the non-storm condition. For the flood tide during the waning storm condition the probability density function is flattened as the bed is approached. This could be
3.5
Re = \frac{u \cdot L}{v}

- Non-Storm 171 (Near bed)
- Waning 240 (Near bed)
- Peak 240 (Near bed)

Figure 3-10: (a) Typical [black] and observed water depth. (b) \(Re_\ast\) for non-storm and storm conditions, the red dashed line indicates an \(Re_\ast\) of 3.5. (c) \(\delta_v\) for non-storm and storm conditions, where the red dashed line indicates \(3k_s\) or the height of the granular roughness elements for this boundary. (d) \(u_\ast\) for non-storm and storm conditions. Non-storm year day 171 \(u_\ast\) value calculated from the viscous method (hollow □) and the log-layer method (opaque □). Waning storm conditions (○), and the peak storm condition data, (▽) (year day 240).
indicative of a region of higher density since deposition was evident during the
waning storm condition as seen in figure 3-4. It is important to note that the
Vectrino II does have limitations in very near boundaries, therefore the probability
density functions within less than 1 mm of the bed can be biased by zero velocities.
For this analysis, the skewness was taken into consideration at distance further
than 1 mm from the bed. The skewness of the non-storm profile is approximately
2 times greater than that of the storm profiles when approaching the bed. A high
skewness of the velocity profile near the bed indicates the presence of the viscous
sublayer.

As portrayed through the probability density function, skewness is an impor-
tant parameter to evaluate the presence of the viscous sublayer. Recall, that the
skewness is a measure of the asymmetry of the probability distribution about its
mean value. The streamwise velocity fluctuations become positively skewed in the
vicinity of the wall because they are limited in amplitude on the negative side of the
mean. This is the result of the no-slip condition. There is a much higher probability
of generating large amplitude positive (greater than the mean) fluctuations, while
most lower amplitude fluctuations are negative. Near the freestream the skewness
becomes negative for this reasoning.

Kurtosis (or flatness) is a secondary, higher order metric which when applied
to the velocity distribution will take on a distinct shape in the presence of viscous
effects. The flatness is the fourth moment, it is a measure of the probability that a
random value in a distribution will fall several standard deviations away from the
mean. In the presence of the viscous sublayer, the flatness will increase above 3,
Figure 3-11: Probability density function of velocity as a function of distance from the boundary for (a) non-storm year day 171, (b) waning storm condition year day 240, and (c) peak storm condition year day 240. As the boundary is approached the color of the probability density function will change from brown to light orange. If viscous sublayer is present the skewness of the probability density function of the velocity will become more positive as the boundary is approached.
which is the flatness of a Gaussian distribution. This is again because the velocity fluctuations near the boundary are more likely to fluctuate widely around the mean in the positive direction.

Figure 1-8 is a representation of what the skewness and flatness should be in the presence of the viscous sublayer. Figure 3-12 shows the skewness and flatness plotted against the non-dimensional wall units, $Z$. The skewness plot does not include the low tide (magenta) profiles. The skewness within 1 mm the bed is not included in this plot because it is biased by instrument limitations near the boundary. A $Z$ value of 10 is approximately equal to 2 mm from the bed. The skewness of the velocity profile near the bed for non-storm conditions (☐) increases in magnitude as the boundary is approached, this characteristic is also seen in figure 1-8, and a $Z$ of 15 is considered to be within the region where viscous affects are important. The storm conditions (o and ▽) do not show skewness in this region. The migration to a negative skewness in the velocity profile is not seen here because analysis is very close to the boundary. The flatness plot is only presented for the non-storm condition, for the middle four hours of the tide. Near the bed the flatness becomes greater than 3, and upon exiting the viscous sublayer it levels off around 3, as expected. The flatness profile is clearly affected by viscous effects, even though the data presented is not within the sublayer, but instead within the buffer layer, where the fluid is still subject to the effects of the viscous sublayer.
Figure 3-12: (a) Skewness of the velocity profile for each tidal phase and condition (with the exception of low tide) presented in non-dimensional wall units. (b) Flatness of the velocity profile for each tidal phase of the non-storm condition (with the exception of low tide and high tide) presented in non-dimensional wall units. Non-storm condition for year day 171 (□), the waning storm condition (○), and the peak storm condition (▽) (year day 240). The color of each hourly averaged profile corresponds to previously presented tidal phase plots.
3.3 Evaluation of Shear Stress

As highlighted in the Introduction, there are various methods used to estimate the shear stress imparted to the sediment bed due to hydrodynamic forcing. One method is to estimate a friction velocity, $u_*$, from the log-layer (equation 1.11) or Indicator function ($u_{*ind} = \frac{du}{dz}z\kappa$) and use this to solve for bed stress, $\tau_b = \rho u_*^2$. For the Vectrino II and Aquadopp HR the Indicator function was used to find the friction velocity and used to solve for bed stress. The log-layer method provides two estimates of bed stress using the profiling instruments. The log-layer method assumes that the velocity profile follows a logarithmic shape all the way to the boundary. This assumption is not valid for all data presented for analysis.

An additional method of estimating bed stress is by estimation of the Reynolds stress, which is a momentum transfer term, and combining this with the viscous stress to estimate the total stress. Very near the bed the Reynolds stress should be equal to zero because of the no-slip condition. Higher in the water column the viscous stress should be equal to zero because the fluid is no longer dominated by viscous forces, where $\tau_{Total} = \mu \frac{\partial u}{\partial z} - \rho \overline{u'w'}$. The total stress estimated at any location in the water column are a function of the momentum imparted to the water at that elevation above the bed. Total stress estimates were evaluated for the Vectrino II, where Reynolds stress and viscous stresses were calculated for the entire profile. When moving higher into the water column the viscous stress does in fact result in zero stress, thus the stress the water column is subject to at that elevation is due to the Reynolds stress. The Reynolds stress was also estimated for the Vector and since the Vector is approximately in the free stream, the viscous stress is assumed.
to be zero. This method provides an additional estimate of bed stress, by taking the closest estimate of total stress to the boundary from the Vectrino II and assuming that this is representative of bed stress.

Another common approach to estimating shear stress is by assuming a drag coefficient and using equation 1.13 to calculate the shear stress. For the Vector, a roughness length, $z_0$, is assumed based off of $d_{50}$ and then paired with water depth measurements to calculate $z_0/h$ which can be used to find an average $C_d$ from figure 1-9. Shear stress with the single point velocity measurement from the Vector is then found using the estimated $C_d$. The Vector only has a single point measurement of the velocity at 0.7 m above the bed, therefore it is not appropriate to fit a log profile to this data because of probable inaccuracies.

Figure 3-13 demonstrates all stress estimates made for each condition during each phase of the flood tide. Each stress estimate is positioned where it is acting, for instance if the stress is acting at the bed it is positioned at $z = 0$. Momentum transfer terms that are acting higher in the water column, are positioned at that elevation. To compare estimates of stress imparted upon the sediment bed numerically, table 3.1 compiles an estimate of bed stress for each method over each tidal phase for each condition.

The log-layer estimate of stress was the first estimation used to make inferences about the hydrodynamics of the area. This estimate of stress is consistent during storm conditions but can be inconsistent during non-storm conditions for Vectrino II profiles (o). The log-layer fit is potentially more accurate during storm conditions when the assumption of a log-layer fit actually matches the hydrodynamics for most
Figure 3-13: Plot of bed stress estimates for each phase of the tide and forcing condition. (a) Predicted and observed water depth. (b - g) Non-storm condition (year day 171). (h - m) Waning storm condition (year day 240). (n - s) Peak storm condition, (year day 240). Symbology: Log-Layer method for Vectrino II (o), viscous stress method for Vectrino II (+), Reynolds stress for Vectrino II (•), $C_d$ method for Vector (opaque ∘), Reynolds stress method for Vector (hollow ∘), Log-Layer method for Aquadopp HR (∗). The Vector Reynolds stress is plotted at 3 cm from the bed to allow for comparison with other stress estimates, although it is actually acting 0.7 m from the boundary.
Table 3.1: Estimates of bed stress ($N/m^2$).

<table>
<thead>
<tr>
<th>Phase of Tide</th>
<th>Bed Stress Method</th>
<th>Non-Storm (yd 171)</th>
<th>Waning Storm (yd 240)</th>
<th>Peak Storm (yd 240)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>ṭ_log</td>
<td>ṭ_Cd</td>
<td>ṭ_v</td>
<td>ṭ_total</td>
</tr>
<tr>
<td></td>
<td>ṭ_log</td>
<td>ṭ_Cd</td>
<td>ṭ_v</td>
<td>ṭ_total</td>
</tr>
<tr>
<td></td>
<td>ṭ_log</td>
<td>ṭ_Cd</td>
<td>ṭ_v</td>
<td>ṭ_total</td>
</tr>
<tr>
<td>low tide</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
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<td>0.0004</td>
<td>0.0090</td>
<td>0.0025</td>
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<tr>
<td></td>
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<td>0.0795</td>
<td>0.0889</td>
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<tr>
<td></td>
<td>0.0083</td>
<td>0.0003</td>
<td>0.0016</td>
<td>0.1539</td>
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<td>0.0120</td>
<td>0.0091</td>
<td>0.1524</td>
<td>0.3554</td>
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<tr>
<td>hour 2</td>
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<td>0.1075</td>
<td>0.1666</td>
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<tr>
<td></td>
<td>0.0666</td>
<td>0.0068</td>
<td>0.0019</td>
<td>0.4554</td>
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<tr>
<td></td>
<td>0.0075</td>
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<td>0.1491</td>
</tr>
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<td>hour 3</td>
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<td></td>
<td>0.0197</td>
<td>0.0179</td>
<td>0.0016</td>
<td>0.0753</td>
</tr>
<tr>
<td>high tide</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>hour 6</td>
<td></td>
<td>0.1205</td>
<td>0.0779</td>
<td>0.0753</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.0179</td>
<td>0.0016</td>
<td>0.0753</td>
</tr>
<tr>
<td></td>
<td></td>
<td>0.1040</td>
<td>0.1543</td>
<td></td>
</tr>
</tbody>
</table>
of the data records as viewed in figure 3-6. During non-storm conditions when the viscous sublayer was present, the log-layer approximation to estimate $u_{*\log}$ was fit to the data above the sublayer. This estimate was then used to calculate the stress imparted to the sediment bed, even though this portion of the curve did not reach the boundary, which could account for the very low log-layer estimates. During the storm time conditions, the Aquadopp HR was available to estimate the stress from the log-layer fit ($*$). This estimate is usually somewhat higher than that of the Vectrino II log-layer fit. Between these two instruments, it is difficult to determine which estimate is closer to the actual $\tau_b$ because both instruments have a high resolution profile. The benefit of the Vectrino II estimate is that it is very near the boundary, however, theoretically the log-layer does approach the free stream, and the Aquadopp HR estimate has the ability to fit a log curve velocity data near the bed and at the free stream.

For the non-storm condition, year day 171, when the viscous sublayer is present it is more accurate to estimate the stress imparted to the bed due to the viscous stress. The viscous stress assumes a linear relation between $u$ and $z$ within this layer, and this assumption visually is accurate to the hydrodynamics of the profile within the bottom 3 to 5 mm (see figure 3-5). However, the total stress is the viscous stress in addition to the Reynolds stress. In this region during the non-storm condition, the Reynolds stress within the lower 3 to 5 mm is small. Moving up through the Vectrino II profile, for non-storm and storm conditions, the Reynolds stress ($\cdot$) grows, while the viscous stress ($+$) is reduced to approximately zero, as expected. During the non-storm condition, at higher elevations in the profile it seems as
it seems as though the viscous stress is non-zero, this is potentially attributed to acoustic error associated with the instrument. In the storm time profiles, the viscous stress is very close to zero, if not zero for the entire profile, which indicates that the viscous sublayer is not present, even for waning storm condition cyan and red profiles (second and third phases), which demonstrate a sublayer in their velocity profiles. There is also a Reynolds stress momentum transfer term calculated for the Vector (hollow o). This estimate can be assumed to be the total stress imparted to the water at the Vector elevation because the viscous stress at this distance from the boundary is assumed to be zero. The Reynolds stress associated with the Vectrino II and the Reynolds stress associated with the Vector create a stress profile, which is visually consistent with theoretical observations of the Reynolds stress profile, where very near the boundary $\rho u'w'$ has a positive slope and as distance from the boundary increases the Reynolds stress profile slope becomes negative.

From a single point velocity measurement, water depth data, and an estimated $z_0$ based upon grain size, the shear stress imparted to the sediment bed can be approximated by means of the drag coefficient for Vector data (opaque o). This estimate of shear stress seemed to be slightly lower than the estimates made by other methods, but is reasonable based upon all of the assumptions made about the site to deduce a $C_d$. The danger in using a drag coefficient to estimate bed stress is the lack of data to support such a conclusion. In these cases, the Vector estimate of shear stress using the drag coefficient, portrays the sediment right at the brink of incipient motion during storm conditions, even though it is known from visual observation and other approximations of stress that threshold of incipient motion
during the mid-hours of the flood tide for storm conditions were surpassed.

After estimating bed stress, in several ways it is appropriate to make a conclusion about what the bed stress is for each condition. Table 3.2 concludes an approximate shear stress that the bed is subject to during non-storm and storm conditions for each phase of the tide. These stresses were approximated by considering characteristics of the flow condition, and giving weight to the stress estimate which is most relevant to the hydrodynamics. For example if the viscous sublayer was present, the total stress estimate was used over the logarithmic stress estimates because one of the assumptions of the logarithmic profile is not relevant to the velocity field at the boundary. However, when the viscous sublayer was not present the stress estimations from the logarithmic methods and the drag coefficient method were effectively averaged to find the approximate bed stress. From $\theta_c$ the $\tau_c$ is found as approximately $0.10 \, N/m^2$. As seen in table 3.2, during non-storm conditions $\tau_b$ is at this limit, but from acoustic observations sediment transport did not occur, this is most likely due to the cohesiveness providing additional stabilization of the bed. However, during storm conditions this critical value is far surpassed, thus resuspension of sediment and nutrients occur.

Figure 3-14 presents calculated values of $C_d$ from final approximations of bed stress made for each tidal phase and condition (table 3.2) plotted on top of figure 1-9, which shows the empirical relations for $C_d$. It is evident that values for $C_d$ are larger than what the empirical relations suggest. For the Vector data, $z_0$ was estimated by $d_{50}/12$ which gave $z_0/h$ values ranging from $2 \times 10^{-6}$ to $7 \times 10^{-6}$. However, when using the estimated $z_0$ from the logarithmic fit to the Vectrino II
velocity profile, the $z_0$ values ranged more widely. $C_d$ estimations for the Vector were an average of the Dawson-Johns, Soulsby, Log-Profile, and Colebrook-White relations, thus fell within the extents of these four curves within this $z_0/h$ range (Manning-Strickler was not used because it is meant for values of $z_0/h > 10^{-4}$).

These methods for estimating a drag coefficient based upon a velocity measurement taken at 0.7 m above the boundary are consistent with the methodology discussed in Bricker et al. (2005), where $\bar{U}$ is taken at some elevation, $z$, above the boundary and is accounted for within the estimate of $C_d$ based upon a logarithmic velocity profile. These estimates of stress from finding a drag coefficient are usually somewhat lower than estimates made by other methods during storm conditions, but are not unreasonable for all observations. However, during non-storm conditions, the drag coefficient method and viscous method compare fairly well.

Figures 3-15 and 3-16 demonstrate the stress field over time with comparison to water depth, velocity, and bed elevation change for the non-storm, and storm conditions, respectively. From these figures it is apparent that bed stress during the non-storm condition was much lower than that of the storm condition and flood

Table 3.2: Approximations of bed stress for each data set analyzed based upon various estimates ($N/m^2$).

<table>
<thead>
<tr>
<th>Phase of Tide</th>
<th>Non-Storm (yd 171)</th>
<th>Waning Storm (yd 240)</th>
<th>Peak Storm (yd 240)</th>
</tr>
</thead>
<tbody>
<tr>
<td>low tide</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>hour 1</td>
<td>0.0033</td>
<td>0.0048</td>
<td>0.0458</td>
</tr>
<tr>
<td>hour 2</td>
<td>0.0083</td>
<td>0.3007</td>
<td>0.1925</td>
</tr>
<tr>
<td>hour 3</td>
<td>0.0448</td>
<td>0.3142</td>
<td>0.2158</td>
</tr>
<tr>
<td>hour 4</td>
<td>0.0666</td>
<td>0.3494</td>
<td>0.2657</td>
</tr>
<tr>
<td>hour 5</td>
<td>0.0589</td>
<td>0.2569</td>
<td>0.1991</td>
</tr>
<tr>
<td>high tide</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>hour 6</td>
<td>0.0197</td>
<td>0.1328</td>
<td>0.0971</td>
</tr>
</tbody>
</table>
Figure 3-14: Comparison of empirical relations for $C_d$ to estimations made from $\tau_b$ found for the Adams Point field site. The (□) is for non-storm condition (year day 171), the waning storm conditions (○), and the peak storm condition data, (▽) (year day 240), and the color of each hourly averaged profile is consistent with previously presented tidal phase plots. Empirical relations are Manning-Strickler, Dawson-Johns, Soulsby, and the full depth log profile, and Colebrook-White (Soulsby, 1997).
Figure 3-15: Non-storm condition water depth, velocity, and change in bed elevation compared to bed stress. (a) Predicted tidal water depths [black], and the recorded water depths from the Vector pressure gage [blue]. (b) Relative magnitude of velocity averaged hourly, mean and standard deviation velocity data from the Vector [blue], with sampling volume of 0.7 m from the bed and Vectrino II [red] with sampling volume between -0.5 and 2.5 cm above the bed. (c) Bed stress estimates, viscous stress from Vectrino II [red], quadratic stress law from Vector [blue], and critical stress is indicated by the black dashed line. (d) Change in bed elevation over the data collection period.

tide greatly influences the bed stress even during storm conditions. These temporal bed stress estimates were a result of a 10 minute velocity record average. For the non-storm condition, the viscous stress for the Vectrino II [red] and the quadratic stress law resultant for the Vector [blue] are compared to the critical stress. For the storm conditions, the quadratic stress law resultant for the Vector [blue] and the logarithmic stress from the AquadoppHR [green] are compared to the critical stress.
Figure 3-16: Non-storm condition water depth, velocity, and change in bed elevation compared to bed stress. (a) Predicted tidal water depths [black], and the recorded water depths from the Vector pressure gage [blue]. (b) Relative magnitude of velocity averaged hourly, mean and standard deviation velocity data from the Vector [blue], with sampling volume of 0.7 m from the bed and Vectrino II [red] with sampling volume between -0.5 and 2.5 cm above the bed. (c) AquadoppHR velocity profiles [green], Vector velocity comparisons [blue] (d) Bed stress estimates, quadratic stress law from Vector [blue], logarithmic stress from AquadoppHR [green], and critical stress is indicated by the black dashed line. (e) Change in bed elevation over the data collection period.
There is evidence that river inflow only contributes to approximately 2% of the Great Bay volume on average (Roseen, 2002; Bilgili et al., 2005). This leads to an assumption that excess shear stress in the Bay is due to wind. Taking this assumption to be accurate, figure 3-17 is an analysis for how wind affects bed stress for the temporal stress curves shown in figures 3-15 and 3-16. The cumulative distribution of wind speed is taken from the Great Bay Tidal Buoy data for the summer and fall months of 2011. This distribution shows at what percentage of time various wind speeds are reached within the Great Bay, a more accurate representation would be to include a cumulative distribution of wind speed over the past 5 years. The scatter plot of bed stress vs. wind speed indicates the approximate wind speed to induce excess shear stress at the boundary. As seen in previous figures, bed stress is also dependent upon the phase of the tide, here the phase of the tide and wind speed are not differentiated, thus some of the smaller bed stress estimates at higher winds speeds probably correlate to ebb tide flows at Adams Point. Finally, the cumulative distribution of bed stress shows that approximately 60 to 75% of the time excess shear stress (of 0.10 N/m²) is not reached from the data sets analyzed. It is probable that with more data this percentage of non-exceedance would increase. When these three plots are considered together an overall representation of how wind speed affects bed stress within the Bay at locations with a sediment type of mud and sandy mud begins to emerge. This analysis does support the conclusion that wind velocity does affect the magnitude of bed stress within the Bay. From figure 2-1 it can be approximated that mud and sandy mud make up 60% of the submarine surfaces, thus this bed stress in relation to wind speed analysis can po-
Figure 3-17: Comparison of bed stress and associated wind speed for the non-storm and storm condition data sets. (a) Cumulative distribution of wind speed for the summer and winter months of 2011, data courtesy of the Great Bay Buoy. (b) Scatter plot of bed stress and wind speed. Viscous stress from Vectrino II for non-storm condition [blue], log stress for AquadoppHR for storm condition [red]. (c) Cumulative distribution of bed stress for the non-storm and storm condition datasets combined, critical stress is indicated by the vertical dashed line.

tentially apply to areas within the Bay that are composed of these sediment types. However, it is important to consider the individual hydrodynamic conditions at each site, as they may not correspond to the unidirectional flow seen at Adams Point.

3.4 Hydrodynamics Effecting Nutrient Diffusion

The Shields parameter and calculated bed stress for the non-storm condition show that hydrodynamic forcing pushes the sediment to the brink of incipient motion.
in the non-storm condition, yet stresses imparted to the bed and bed elevation change over the storm indicate that incipient motion does occur. This characteristic of the Bay during typical tidal forcing indicates that nutrients must diffuse through the diffusive boundary layer, following Fick’s Law, with constant of molecular diffusion, $D$, characteristic to each nutrient. The thickness of the diffusive boundary layer can be found for various molecular diffusivities as a relation to the thickness of the viscous sublayer. In this instance, because the thickness of the viscous sublayer is known for each phase of the tidal cycle, the thickness of the diffusive boundary layer can be found based upon the molecular diffusivity of the nutrient of interest. This relation comes from equation 1.3, and is equal to,

$$\delta_d = \delta_u \left( \frac{\nu}{D} \right)^{1/3}, \quad (3.1)$$

where the thickness of the diffusive boundary layer will change with each phase of the tide as hydrodynamic conditions change. Since $D$ is characteristic to each nutrient, and $\nu$ is characteristic to the fluid, the only unknown in this equation is $\delta_d$.

After nutrients diffuse through the diffusive boundary layer, the effective diffusion constant will change from molecular diffusion to eddy diffusion, $E$. Figure 1-3 shows two empirical representations from Van Driest and Reichardt, which are based off of the Prandtl mixing length and eddy viscosity of, $E(z) = \kappa u_z z$, as discussed. These empirical equations also characterize the thickness of the viscous sublayer and diffusive boundary layer graphically. Van Driest and Reichardt equations are used to estimate the eddy diffusivity as distance from the bed increases.
The eddy diffusivity is a function of distance from the boundary because the length of eddy responsible for momentum transfer at any location is proportional to the distance from the boundary as stated by the Prandtl mixing length hypothesis.

Figure 3-18 shows the Prandtl, Van Dierst, and Reichardt curves based upon characteristics of the velocity profiles taken from the Vectrino II non-storm data on year day 171. The vertical black dashed line indicates the kinematic viscosity of sea water, and the red dashed lines bracket the calculated minimum and maximum thickness of the viscous sublayer for each phase of the tide. The black dashed line ($\nu$) should cross the Van Dierst and Reichardt empirical relationships at the height of the viscous sublayer ($\delta_v$). For each phase of the tide the eddy diffusivity curves do predict the height of the viscous sublayer fairly well when for the maximum calculated sublayer thickness or the higher red dashed line indicating $\delta_v = 11.6\nu/\nu_\ast$. Another method to calculate the thickness of the diffusive boundary layer is to take the elevation from the bed where the molecular diffusivity, $D$, vertically crosses the empirical eddy diffusivity relationships. $D$ is variable based upon the nutrient of interest, so it is not represented here. These empirical relationships to determine the height of the viscous sublayer are fairly accurate based upon field observations of the viscous sublayer.

During storm conditions, the eddy diffusivity is still important, however, there is the added condition of nutrient mixing through advection and resuspension of sediments, which is much more effective than molecular diffusion. With increasing Reynolds stress, it is probable that the rate at which nutrients mix into the water column will increase as well. This is attributed to the Reynolds stress being a
Figure 3-18: The Prandtl, Van Driest, and Reichardt equations to estimate eddy diffusivity, \( E \), as a function of \( z \). Eddy diffusivity can be related to molecular diffusivity and the thickness of the viscous sublayer through these empirical relationships. Prandtl [blue], Van Dierst [green], and Reichardt [magenta], the vertical black dashed line indicates the kinematic viscosity of sea water, and the red dashed lines bracket the minimum and maximum thickness of the viscous sublayer for each phase of the tide. All data presented for non-storm condition, year day 171, tidal phase is indicated on far right for each panel. The left column of plots is the full eddy diffusivity curve calculated for Prandtl, Van Dierst, and Reichardt, and to the right is a zoomed in section of each plot to show where the dashed lines intersect with the eddy diffusivity curves.
column will increase as well. This is attributed to the Reynolds stress being a momentum transfer term. Overall, during non-storm conditions nutrient mixing is dominated by molecular diffusion through the diffusive sublayer until it reaches the bulk water where the nutrients can turbulently mix in the Great Bay. However, during mild storm conditions nutrient mixing will be dominated by advection due to resuspension of nutrients into the water column due to incipient motion of sediment in the Great Bay. These observations are made for flood tide. During ebb tide, it is probable that nutrient mixing is dominated by molecular diffusion even during mild storm conditions due to subdued hydrodynamic tidal forcing.

The flux of nutrients within the Great Bay system as a whole can be modeled by a mass balance. Potential sources for nutrients are from river inflow, groundwater inflow, tidal inflow and outflow, waste water treatment plant inflow, and solute flux from the sediment bed. Each of these potential sources have an effective concentration per volume of water to contribute to the mass balance, however solutes within the sediment bed can be released into the bulk water through a flux and can be modeled in two ways: by a thin-film model, where nutrients must diffuse across the diffusive boundary layer during times where excess shear stress is not reached, and by advective flux when incipient motion of sediment occurs. This flux can be associated with the 60% of the Bay which is made up of mud or sandy mud. Because the Great Bay is a tidal estuary, the bulk water is partially renewed during each tidal cycle, which can dilute any nutrient base within the Bay. This mass balance should be completed for one tidal cycle, which includes a flood and an ebb tide (12.42 hours) for both a non-storm and storm forcing scenario. Then using a
loading of the Bay can be calculated for a period of time. On average there are 14 storms per year within the Great Bay, this was found by considering both elevated wind and hydrologic runoff conditions from the Pease Tradeport wind data, and the USGS stream gaging statistics, respectively, over the past 5 years. These loads can then be compared to effective water quality standards within New Hampshire as published by the New Hampshire Department of Environmental Safety (NHDES).
CHAPTER 4

CONCLUSIONS

Two separate field deployments were completed in the summer of 2011 to evaluate the potential for nutrients, stored within fine grained sediments, to re-enter the bulk water through the sediment-water interface in the Great Bay Estuary of New Hampshire. The project objective was to characterize the flow field and magnitude of bed stress within a representative location within the Great Bay. After two pilot studies, Adams Point was chosen as the representative site because of its sandy-mud sediment properties and unidirectional flow characteristics. This site allowed for flow field and stress characterization over a range of velocities within a semi-controlled field environment.

The first deployment occurred during typical tidal forcing conditions on year day 171 of 2011 (mid June). During this deployment, the threshold for incipient motion of sediment was approached but not surpassed, and the change in bed elevation over the non-storm records was approximately 0 mm. Furthermore, there is substantial evidence that supports the presence of an observable viscous sublayer and buffer layer during this time. Estimates of the bed shear stress show that during a typical flood phase tidal forcing, critical shear stress ($\tau_c = 0.15 N/m^2$) is approached but not exceeded, most likely due to the cohesiveness of the sediment. These observations, suggest that during typical tidal forcing, Adams Point is at
local morphologic equilibrium, thus nutrient mixing is primarily due to nutrient diffusion.

Given that incipient motion of sediment did not occur during typical tidal forcing within this area, the question arises regarding under what hydrodynamic condition will the critical threshold be exceeded, and thereby resuspending sediment and mixing nutrients by means of advection. Thus, a second field deployment targeted a storm condition during Tropical Storm Irene, which was preformed in late August of 2011, with data analysis presented for year day 240. Tropical Storm Irene was a mild to moderate storm for the Great Bay region. During the time of the strongest external forcing due to wind and hydrologic runoff, the sediment bed elevation only eroded just over 1 mm in 6 hours. Although a relatively small amount of erosion, the event demonstrated that incipient motion of sediment resulted when the stress exceeded the critical threshold. There were two data sets evaluated during this event. The earlier year day 240 flood tide cycle exhibits the largest flows of the two, with turbulent velocity profiles, high estimates of bed shear stress, and observations of bed erosion, this is referred to as the peak storm condition. However, at the later year day 240 flood tide cycle, the storm was weakening, this is referred to as the waning storm condition. Estimates of stress during the most energetic phases of flood tide still exceeded critical shear stress, because as stated, during typical tidal forcing Adams Point is at the brink of incipient motion. However, the near bed boundary layer shape suggest a significant viscous contribution during the waning storm condition. Also, when considering the change in bed elevation, there is evidence of a potential depositional event during this cycle, suggestive of
a weakening flow field even though the Shields parameter and shear stress values were estimated to exceed the critical threshold.

Boundary layer development is crucial to characterizing the type of flow condition present. As suggested through this thesis, during the typical flood tidal forcing the boundary layer develops through low to high tide, and there is significant evidence of the presence of a region where viscous forces dominate the flow regime. The boundary layer at Adams Point develops fairly quickly during flood tide, and is fairly consistent during the mid-four hours of the flood phase. Within these four hours during the non-storm forcing, an observable viscous sublayer is present. Hydrodynamically, the boundary layer is considered hydraulically smooth parameterized by the roughness Reynolds number. The shape of the velocity profiles are consistent with the characteristic velocity profile shape when the viscous sublayer is present in the Caldwell and Chriss (1979) observations and also in the non-dimensional wall units profile. Unlike during the storm conditions, the calculated thickness of the non-storm viscous sublayer is greater than the height of the granular roughness elements when defined by the Nikuradse roughness, $k_s$. Also, the critical Shields parameter and stress are not exceeded during the non-storm condition, while these parameters indicate a mobile bed during the storm condition. Higher order moments of the velocity profile provide for skewness and kurtosis provide further support for an observable viscous sublayer during the non-storm condition. Finally, within the viscous sublayer and buffer layer during the non-storm condition there is a significant stress contribution from the viscous stress, while during storm conditions the viscous stress does not appreciably contribute to
the total stress estimate, even in the very near bed region. Each of these parameters produces independent evidence for the presence of an observable viscous sublayer during the flood phase of the tide at Adams Point during typical tidal forcing.

The second and third hourly averaged velocity profiles within the flood tide of the waning storm condition also exhibit the characteristic shape for a boundary layer with significant influence from viscous forces. However, these profiles did not satisfy any of the other parameters in support of an observable viscous sublayer. Moreover, the viscous stress in the near bed region was significantly smaller than the Reynolds stress. A possible explanation for this discrepancy is particle settling and increased drag due to sediment resuspension. A region of altered density near the bed could affect the dynamics in a way that is not considered here. If this is in fact reality, the region of altered density would effectively mimic a viscous sublayer shape due to altered acoustic properties of the fluid.

This consideration suggests that the presence of the viscous sublayer and buffer layer during the non-storm condition could be affected by a region of altered density very close to the bed. Without further observation this caveat cannot be completely dismissed. However, if the suggestion that the shape of the velocity profile during the waning storm condition is due to a density change in the near bed region, and this region of altered density did not bias the viscous stress estimates nor the higher order moments and other indicative parameters of the presence of the viscous sublayer, the validity of the suggestion that the viscous sublayer is in fact observable during typical tidal forcing at Adams Point is appreciably more compelling. Thus, observations and analysis assert that an observable viscous sub-
layer is present at the Adams Point location during typical tidal forcing conditions. This is exciting because this observation of the viscous sublayer in a marine field environment is only the second within the past 30 years.

This analysis suggests that during non-storm times molecular diffusion through the diffusive boundary layer should dominate nutrient mixing into the bulk water. The diffusive boundary layer thickness and viscous sublayer thickness will change based upon local turbulence and hydrodynamics. During mild storm conditions, the mechanism of nutrient mixing becomes advection. The hypothesis that typical tidal forcing conditions would induce incipient motion of sediment is not valid, and consistent with local morphologic equilibrium at Adams Point. Results demonstrate that even mild external forcing conditions will result in excess shear stress and thus resuspension at the Adams Point site. This information is important to begin to better understand and characterize the nutrient loading of the Great Bay which supports many species, habitats, and recreational uses.

On average, this region is subject to approximately 14 storms per year which match or surpass the external wind and runoff conditions of Tropical Storm Irene. Turbulent diffusion of solutes from the sediment will dominate nutrient mixing during these times of excess wind and hydrologic runoff, while mixing during the rest of the time is dominated by molecular diffusion and biological activity. The Great Bay is made up of approximately 60% of mud or sandy mud, which has a similar particle size distribution to the sediment at Adams Point and is capable of storing nutrients in its porewater, and thus has the potential to contribute to the overall nutrient flux within the Bay. This analysis, paired with an analysis
of the nutrient geochemistry of the sediment bed can inform municipalities and
government agencies whether the Great Bay is in danger of not meeting effective
total maximum daily loads (TMDL), these regulations are set to consider the overall
health of the ecosystem for wildlife as well as human contact and consumption
by the New Hampshire Department of Environmental Services (NHDES) and the
United States Environmental Protection Agency (USEPA).

Further investigations could combine the capabilities present for characteriza-
tion of local hydrodynamics to a study which can resolve the diffusive sublayer by
means of dissolved oxygen content. Since it is probable that the viscous sublayer
is observable in a field environment at Adams Point, this would be an excellent
site to perform this type of collaborative analysis. It is also relevant to perform a
third field experiment to verify findings of an observable viscous sublayer at Adams
Point, with an auxiliary technique to verify the state of the boundary. There is
also potential for subsequent investigation to characterize the effects of a near bed
density differential in relation to boundary layer development. Finally, this study
can be extended to other locations within the Great Bay and within other more in-
tense flow conditions. With additional data during normal forcing conditions, and
during various external forcing conditions, a cumulative stress distribution can be
created to show how often and at what hydrodynamic conditions will cause excess
shear stress, and thus a more accurate nutrient budget from the sediments can be
accounted for.
Two pilot deployments for this project were completed in August of 2010.

The first was at Wagon Hill Farm and the second was at Thomas Point, before deciding on a final project site at Adams Point. The sediment at Wagon Hill Farm was a muddy sand. The hydrodynamics at Wagon Hill Farm yielded maximum velocities of 40 cm/s during normal tidal forcing conditions, with the added outflow of the Oyster River. This site had unidirectional flow, however the sediment type was too coarse to hold nutrients for the nutrient chemistry portion of the project.

The second deployment was at Thomas Point. As seen, Thomas Point is a peninsula that reaches into the channel of the Bay. The sediment type at Thomas Point was a mud, however the hydrodynamics at this site were not ideal for characterizing stress. Just off-shore of the instrument location, there was a steep drop off into the 19 m channel, it was concluded that the main flow path at this site was through the channel. It was also found that due to the geometry of Thomas Point, it is subject to large recirculating eddies, causing the flow to change directions throughout the entire tidal cycle, whether it is flood or ebb tide. These eddies yielded low velocity magnitudes, of only 20 cm/s at maximum.
The Great Bay of NH, Sediment Type, Channel Bathymetry, and Field Site Location

Figure A-1: The Great Bay of New Hampshire sediment type and bathymetry map. Two pilot deployments were completed in August of 2010 at Waggon Hill Farm, on the Oyster River, and at Thomas Point as indicated by the labeled pink dots. The Final deployments were made at Adams Point, also shown.
Table A.1: Percent finer grain size analysis for pilot sites and for the Adams Point site.

<table>
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<th>Sieve Size</th>
<th>Location:</th>
<th>Oyster River</th>
<th>Thomas Pt. 1</th>
<th>Thomas Pt. 2</th>
<th>Thomas Pt. 4</th>
<th>Adams Pt.</th>
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<td>Many big rocks</td>
<td>Few Rocks</td>
<td>No big rocks</td>
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<td>100</td>
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<td>97</td>
<td>75</td>
<td>82</td>
<td>99.3</td>
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<td>92</td>
<td>66</td>
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<td>96.9</td>
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<td>49</td>
<td>52</td>
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<td>error 2.7%</td>
<td>error 3.5%</td>
<td>Error 2.1%</td>
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APPENDIX B

Evaluation of the Nortek Vectrino II Acoustic Doppler Profiler

The Nortek Vectrino II is a profiling acoustic Doppler velocimeter. This instrument was released to the public in August of 2011, and was beta tested by several clients over the course of June 2010 through August 2011. The Vectrino II is a profiling bistatic system, which means that it simultaneously samples closely spaced volumes, and samples multiple volumes per measurement cycle, thereby providing full three dimensional measurements of a range of velocities (Craig et al., 2010). Figure B-1 shows the Vectrino II as it was mounted during the Adams Point field site deployment. The probe is made up of, four passive transducers, angled at 30° towards the center surrounding the central active transducer producing an intersection point 50 mm below the central transducer. This provides a usable profiling region approximately 40-80 mm in height away from the central transducer. The unit generates a continuous train of transmit pulses at 10MHz organized into ensembles, the number of pings per ensemble is determined by the sampling rate (up to 100Hz) and other configuration parameters (Craig et al., 2010).

One of the biggest problems with pulse to pulse coherent Doppler instruments is acoustic interference from the previous pulse. Acoustic reflections depend upon boundary type, as well as an array of other parameters. With single point ADVs
Figure B-1: The Vectrino II is a bi-static, 10 MHz acoustic Doppler profiling velocimeter. Here it is shown as it was deployed during the Adams Point experiment. The probe is to the left and the hardware pressure housing is to the right, mounted on the pole.

This can be a problem, but it is easily resolved by moving the instrument’s relative location to the boundary, or changing one of several sampling parameters so that the reflection does not interfere with the sampling volume. With a profiling ADV this is more difficult. The profiling attribute is very useful for resolution of near boundary flows. This characteristic introduces problems, because when near the boundary an acoustic reflection is difficult to avoid, and hard to move out of the sampling range. The Vectrino II does have the ability for adaptive pinging, but even this does not completely resolve reflection problems, especially for hard boundary surfaces. For more background information about the Vectrino II as well as information about Nortek’s position on limiting acoustic reflections, refer to Craig, Loadman, Clement, Rusello, and Seigel (2010).

In the Great Bay, the boundary was soft and acoustically absorbent, so acoustic interference due to bottom conditions was easily avoided. The instrument evaluation presented here is for two data sets during mid-tide, where the flow conditions
are fairly uniform throughout the period of data collection. The first data set is for a deployment where the profiling sample volume was from 5.5 to 8.5 cm above the bed (year day 167), and the second data set is from year day 171 where the sampling volume was -0.5 to 2.5 cm above the bed. These two data sets allow for comparison between instrument performance with a sampling volume that was and was not influenced by near boundary effects. For both data sets the range used was from 40 to 70 mm from the instrument, this is because the "sweet spot" for the Vectrino II is at approximately 50 mm from the transducer. Extending past approximately 70 cm introduces more error into corrected data based upon the instrument beam pattern, and above 40 mm introduces fluid disturbance from the Vectrino II probe.

An acoustic model of the 10 MHz Vectrino II beam pattern from transmitter (center) and one receiver over the nominal Vectrino II measurement region of 30 to 75 mm from the transmitter is shown if figure B-2. The theoretical sampling zone is identified with black dots. The cyan dots indicate the modeled average region of maximum power and sensitivity that determines the sample location of each measurement cell. The red bands are the actual passive transducer beam patterns. Notice that the "sweet spot" is where these bands intersect. Above and below this intersection, internal post processing must account for the spacial difference between the beam patterns to resolve a velocity that is centrally located. Also notice the ellipsoid bands, this is the pattern which the profiler resolves its velocities around. In the observations presented here, the sampling zone, as determined by the beam intersection over 31 mm zone, (please note, in figure B-2 shows 50 mm
Figure B-2: The Vectrino II is a bi-static, 10 MHz acoustic Doppler profiling velocimeter. This model shows the beam pattern (red), modeled average region of maximum power (cyan), and theoretical sampling zone (black). The region where the red bands intersect is the "sweet spot" for the bi-static instrument. Model results courtesy of Len Zedel, Memorial University.

sampling zone, which is not accurate to Vectrino II sampling volume presented in this analysis) (Nortek). Transformation between beam coordinates and XYZ coordinates for the Vectrino II is based upon a regression used for the single point Vectrino ADV. This regression may not be the best fit to resolve velocities near the upper and lower extents of the profile range (Craig et al., 2010).

### B.1 Velocity, Amplitude return, and Correlation Profiles

The first method of analysis is considering velocity profiles and how much variability is associated with them, when the velocity magnitude is not changing. This can be
accomplished by plotting mean velocity with plus or minus one standard deviation for a 10 minute velocity record as in figure B-3 for the \( u \) and \( v \) components, and figure B-4 for \( w_1 \) and \( w_2 \) components (the Vectrino II has a \( w \) component associated with each \( u \) and \( v \) because the instrument is basically two 2D ADVs put together to create a 3D ADV). Figures B-5, and B-6 present the beam amplitude and correlation signals for both data sets, respectively. Another way to look at the spread of velocity contributions over the record is to plot an array of probability density functions of velocity for each elevation above the bed. Figure B-7 shows the probability density function for the \( u \) component of alongshore velocity time series. Figure B-3 through B-7 plot year day 167 and year day 171 on the same plot at the elevation above the bed that data was recorded. These data sets were taken at the same phase of the tide, but were not taken during the same tidal cycle, thus please note that a perfect velocity profile when combining the two data sets may not exist.

From figures B-3 and B-4 it is apparent that there is high variability of the velocity signal for a ten minute record. The mean profile does look uniform. However the standard deviation of the filtered time series is large. This could be attributed to turbulence in the water column, but also could be due to instrument calibration, and the regression used to resolve the XYZ component of velocity. The \( w_1 \) and \( w_2 \) signals are almost perfectly aligned. The associated error between \( w \) components close to the bottom of the profiling range is the largest. In the away from bed profile it looks as though both \( w_1 \) and \( w_2 \) velocity signals reach zero velocity at the "sweet spot", this characteristic is concerning being that at mid-tide there probably was low vertical movement through the water column, but zero vertical movement is
Figure B-3: Velocity profiles with one standard deviation from mean for $u$ and $v$ components of near-bed (yd 171) and away from bed (yd 167) data sets at mid-tide.

Figure B-4: Velocity profiles with one standard deviation from mean for $w_1$ and $w_2$ components of near-bed (yd 171) and away from bed (yd 167) data sets at mid-tide. $w_1$ is associated with $u$ and $w_2$ is associated with $v$. If instrument is not tilted and velocity transformation is accurate, $w_1$ and $w_2$ should be almost the same.
Figure B-5: Amplitude return profile for beam coordinates, with one standard deviation from mean for near-bed (yd 171) and away from bed (yd 167) data sets at mid-tide.

Figure B-6: Beam correlation profile, with one standard deviation from mean for near-bed (yd 171) and away from bed (yd 167) data sets at mid-tide.
Figure B-7: Probability density function of $u$ component of velocity for both data sets. This plot shows the frequency of return for each velocity signal over the hour long record for near-bed (yd 171) and away from bed (yd 167) data sets at mid-tide.
not probable. In the near-bed profile, this same effect is not seen. The beam coordinate amplitude and correlation profiles the standard deviation is relatively lower than the standard deviation of the of the velocity profiles. The amplitude should be below -20 dB to be considered quality data. This is true except for the bins within the boundary, which is reasonable. The correlation of signal between each of the four beams is above 95 percent for each beam, except within the boundary, which is reasonable.

The probability density function of the $u$ component of velocity is shows that the velocity within the boundary is zero, this is known to be accurate with an immobile boundary, thus it should be highlighted that this acoustic devise is capable of resolving these values. The profile shows the similar characteristics to figure B-3 in terms of the spread of data. Figure B-7 is another method to view characteristics of the velocity distribution.

### B.2 Geometry and Stress Calculations Along the Profile

When considering figure B-2 it is apparent that when resolving the velocity signal or the Reynolds stress, as presented within the thesis main body, it is important to consider the angle change between the passive and active transducers at elevations along the profile. The 30° angle which the passive transducers are directed toward the active transducer is actually variable depending on position within the profile. Figure B-8 demonstrates the degree with which the bi-static angle changes along
Figure B-8: Calculated bi-static angle for each profile bin for the Vectrino II profiling ADV. Bi-static angle is a function of geometry of the probe, and distance from the transducer.

The bi-static angle is the angle that bisects the angle between the transducer and receiver paths. This angle can be used to more accurately calculate the Reynolds stress through a method presented by Alex Hay of University of Dalhousie (refer to Hay et al. (2012) for more information on this method). In this thesis it was found that for the Vectrino II the method to calculate Reynolds stress accounting for the change in bi-static angle does not make a significant contribution to stress estimate. Reynolds stress estimates accounting for the bi-static angle difference and not accounting for this difference are different by approximately 0.01% at the furthest profile bin from the "sweet spot".
B.3 Spectra

The power spectral density (PSD) of the two alongshore ($u, v$) and vertical ($w_1, w_2$) velocities at four elevations from the bed are shown in figure B-9, B-10, B-11, and B-12. The spectra were calculated over a 10 minute record at mid-tide with 80 degrees of freedom. All of the spectra are red with decreasing energy as a function of frequency. None of the signals show evidence of free surface gravity waves. The horizontal velocities reach a $-5/3$ slope at frequency higher than 5 Hz for each spectra excluding the spectra within 2 mm of the bed. The PSD for the vertical velocities is consistently lower than that of the horizontal velocities. Moreover, there are no significant differences between the two vertical velocity channels estimated from $u$ and $v$ components. In general, the PSD for the vertical velocities show a larger frequency band that is consistent with a $-5/3$ slope.

The magnitude of the PSD as the distance to the bed decreases, also decreases. This is consistent with intuition and existing boundary layer theories. At 1.6 mm from the bed (the lowest panel in B-12) when the sampling volume is within the viscous sublayer, the PSD shows the largest deviation from the $-5/3$ slope.

Figure B-12 is a single spectra taken from the lowest flow conditions at low tide on year day 167. In this record, the average alongshore velocity at an elevation of 6.2 cm from the boundary is 1.9 cm/s. The tapering of the PSD at frequencies beyond 15 Hz provides evidence that the noise floor of the instrument not observed in the other PSD. See Appendix C for more PSD plots.
Figure B-9: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 167, non-storm condition for mid-tide, Spectra has 80 degrees of freedom. And is located 7.3 cm from boundary.

Figure B-10: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 167, non-storm condition for mid-tide, Spectra has 80 degrees of freedom. And is located 6.1 cm from boundary.
Figure B-11: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 171, non-storm condition for mid-tide. Spectra has 80 degrees of freedom. And is located 1.0 cm from boundary.

Figure B-12: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 171, non-storm condition for mid-tide. Spectra has 80 degrees of freedom. And is located 0.2 cm from boundary.
Figure B-13: Spectra presented for $u$, $v$, $w_1$, $w_2$ components of velocity 6.2 cm above the boundary taken at low energy, low tide when velocity was 1.9 cm/s from year day 167. This plot shows indication of the noise floor after 15 Hz.

B.4 Coherence and Phase

The coherence is used to show how two signals are related. Figure B-14 relates the coherence of the "sweet spot" to the coherence at every other profile elevation for year day 167 (away from the boundary) and year day 171 (near the boundary). Any area within the plot that has a coherence that is less than the 95% significance of 0.1173 is white. For both near bed and away from bed profiles the "sweet spot" (bin 10) has a coherence of 1, as expected. The coherence drops as distance from bin 10 increases and as frequency increases. Beyond 5 Hz, even neighboring bins are incoherent. The coherence is stronger for year day 167 when the probe was located higher in the water column and the signals were more energetic. The coherence signal indicates that at the sweet spot, neighboring bins are sampling independently. Plotting the coherence of every $u$ signal compared to the $u$ signal
of any bin, moving further away from the sweet spot, a noticeable trend of more coherent signal is seen between bins, this indicates that at further distances from the sweet spot, the measurements of velocity become less independent, especially at low frequencies. Figure B-15 shows the coherence at bin 17 compared to the rest of the data.

The coherence between alongshore velocity ($u$) and vertical velocity ($w_1$) is shown in figure B-16 for year day 167 and 171. Near the center of the sampling volume, the coherence between $u$ and $w$ is insignificant. However, at the upper and lower ends of the sampling profile, the coherence increases. An explanation for this results is that the beams are not directly overlapping in this region and some of the horizontal velocity signal is pushed into the vertical signal and visa versa within the transformation matrix from beam coordinates to XYZ signal. This suggests that careful interpretation is needed before using the Reynolds stress estimates. In theory, the $u$ and $w$ signals should be coherent at the sweet spot, however this is not true. This suggests that the sweet spot may have more noise associated with it because noise is not coherent. This could be due to less averaging between beams to find the $u$ and $w$ components of velocity at the sweet spot.

Figure B-17 presents the phase spectra for $u$ and $u10$ at the "sweet spot". Both plots show for the away from boundary data set (yd 167) and the near boundary data set (yd 171) that all $u$ signal is in phase with the sweet spot $u$ signal at low frequencies, and higher frequencies for data that is closer to the sweet spot. Figure B-18 shows the phase spectra for the $u$ and $w_1$ components. White data is not significant in the coherence plots. This spectra shows that $u$ and $w_1$ are 90° out
Figure B-14: Spectral coherence between the $u$, alongshore velocity signal, and the $u10$ alongshore velocity signal which is the location of the sweet spot for the Vectrino II. Year day 167 (top) is away from the bed, and year day 171 (bottom) is near the bed. 95% significance level is 0.1173, white data is below this level, not significant.
Figure B-15: Spectral coherence between the $u_1$ alongshore velocity signal, and the $u_{17}$ alongshore velocity signal. Year day 167 (top) is away from the bed, and year day 171 (bottom) is near the bed. 95% significance level is 0.1173, white data is below this level, not significant.
Figure B-16: Spectral coherence between the $u$, alongshore velocity signal, and the $w$ vertical velocity signal. Year day 167 (top) is away from the bed, and year day 171 (bottom) is near the bed. 95% significant level is 0.1173, white data is below this level, not significant.
of phase in the colored areas, this does not affect the Reynolds stress estimates because of the nature of Reynolds stress, however it suggests that there could be a better technique for transformation between beam and XYZ coordinates for this instrument.

B.5 Conclusion

Field observations of a tidal boundary layer evolution were obtained with a new Vectrino II profiling acoustic Doppler velocimeter. While designed for laboratory environments, the probe surveyed field conditions. The mean velocities show a typical boundary layer profile with the velocity increasing with distance from the bed. In the very near bed region, the velocity variance becomes small and the mean profile is consistent with a viscous sublayer of 3 mm, suggesting the probe is applicable to near wall studies.

The power spectral density of various range bins show a decreasing amount of energy with increasing frequency. The spectral slopes approach a -5/3 slope for part of the turbulent band. The noise floor is only evident during very low flow conditions. The coherence between the horizontal velocities over the profile is reasonable, given that the signals are only coherent for low frequencies and neighboring bins. This would suggest that the bins are sampling independently. The coherence between horizontal and vertical velocity fields suggest that divergence of the beam at outer portion of the sampling profiles may result in contamination of horizontal velocity signal with the vertical velocity signal with transformation from beam to XYZ coordinates. Theoretically, correlation at the sweet spot between $u$ and
Figure B-17: Spectral phase between the $u$, alongshore velocity signal, and the $u_{10}$ at the sweet spot. Year day 167 (top) is away from the bed, and year day 171 (bottom) is near the bed. 95% significant level is 0.1173 for Coherence, white data is below this level, not significant.
Figure B-18: Spectral phase between the $u$, alongshore velocity signal, and the $w$ vertical velocity signal. Year day 167 (top) is away from the bed, and year day 171 (bottom) is near the bed. 95% significant level is 0.1173 for Coherence, white data is below this level, not significant.
$w$ signal should be the best, but for this instrument that is not the case. These characteristics of the correlation profile and the phase profile suggest that there are still certain aspects about this new instrument that are not understood. With this said, most data which is produced does look reasonable, and it is important to acknowledge that this product is one of the first commercially produced of its kind. The Vectrino II is an exciting instrument, and can only be improved upon.
APPENDIX C

Spectra for Presented Non-Storm and Storm Data Sets

This appendix is organized by data set. Year day 171, non-storm condition is presented first with spectra from low-tide (low velocity condition), and mid-tide (maximum velocity condition) for various range bins (1, 7, 10, and 22) along the profile. Spectra for year day 240 during the earlier and later flood tidal cycle storm condition are shown in this manner as well but for bins (1, 10, and 22). Please note that the spectra for the four hourly averaged phases of the tide not presented here, however, are similar to the spectra presented for mid-tide. Also note that for year day 171 the sweet spot for the data set occurs at bin 7 because the this data set was corrected for a time jump problem causing it to have only 28 bins. The sweet spot for year day 240 occurs at bin 10.
Figure C-1: Spectra for \( u, v, w_1, \) and \( w_2 \) year day 171, non-storm condition for low tide, low energy signal, notice the wave signal at 0.5 Hz indicating a 2 second wave signal. Noise floor is present near 30 Hz. Taken for range bin 1 (profile bin highest in water column), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-2: Spectra for \( u, v, w_1, \) and \( w_2 \) year day 171, non-storm condition for low tide, low energy signal, notice the wave signal at 0.5 Hz indicating a 2 second wave signal. Noise floor is present near 30 Hz. Taken for range bin 10 (middle profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-3: Spectra for $u$, $v$, $w$, and $w_2$ year day 171, non-storm condition for low tide, low energy signal, notice the wave signal at 0.5 Hz indicating a 2 second wave signal. Taken for range bin 22 (near bed profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-4: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 171, non-storm condition for mid-tide, high energy signal, slope follows turbulence signal slope, no significant waves signals. Taken for range bin 1 (profile bin highest in water column), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-5: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 171, non-storm condition for mid-tide, high energy signal, slope follows turbulence signal slope, no significant waves signals. Taken for range bin 7 (this is at sweet spot where beams intersect), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-6: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 171, non-storm condition for mid-tide, high energy signal, slope follows turbulence signal slope, no significant waves signals. Taken for range bin 10 (middle profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-7: Spectra for $u, v, w_1$, and $w_2$ year day 171, non-storm condition for mid-tide, high energy signal, slope follows turbulence signal slope, no significant waves signals. Taken for range bin 22 (near bed profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-8: Spectra for $u, v, w_1$, and $w_2$ year day 240, late time storm condition for low tide, low energy signal. Noise floor is present near 30 Hz. Taken for range bin 1 (profile bin highest in water column), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-9: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 240, late time storm condition for low tide, low energy signal, notice the wave signal at 0.7 Hz indicating a 1.5 second wave signal. Noise floor is present near 30 Hz in $w$ signal. Taken for range bin 10 (middle profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-10: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 240, late time storm condition for low tide, low energy signal, notice the wave signal at 0.5 Hz indicating a 2 second wave signal. Taken for range bin 22 (near bed profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-11: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 240, late time storm condition for mid-tide, high energy signal, slope follows turbulence signal slope. Taken for range bin 1 (profile bin highest in water column), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-12: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 240, late time storm condition for mid-tide, high energy signal, slope follows turbulence signal slope. Taken for range bin 10 (sweet spot of the profile), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-13: Spectra for \( u, v, w_1, \) and \( w_2 \) year day 240, non-storm condition for mid-tide, high energy signal, slope follows turbulence signal slope, no significant waves signals. Taken for range bin 22 (near bed profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-14: Spectra for \( u, v, w_1, \) and \( w_2 \) year day 240, early time storm condition for low tide, low energy signal, notice the wave signal at 0.7 Hz indicating a 1.5 second wave signal. Noise floor is present near 30 Hz. Taken for range bin 1 (profile bin highest in water column), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-15: Spectra for \( u, v, w_1 \), and \( w_2 \) year day 240, early time storm condition for low tide, low energy signal, notice the wave signal at 0.7 Hz indicating a 1.5 second wave signal. Noise floor is present near 30 Hz in \( w \) signal. Taken for range bin 10 (middle profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-16: Spectra for \( u, v, w_1 \), and \( w_2 \) year day 240, early time storm condition for low tide, low energy signal, notice the wave signal at 0.5 Hz indicating a 2 second wave signal. Taken for range bin 22 (near bed profile bin), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and -5/3 slope are also plotted for reference. Spectra has 80 degrees of freedom.
Figure C-17: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 240, early time storm condition for mid-tide, high energy signal, slope follows turbulence signal slope. Taken for range bin 1 (profile bin highest in water column), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and $-5/3$ slope are also plotted for reference. Spectra has 80 degrees of freedom.

Figure C-18: Spectra for $u$, $v$, $w_1$, and $w_2$ year day 240, early time storm condition for mid-tide, high energy signal, slope follows turbulence signal slope. Taken for range bin 10 (sweet spot of the profile), relative velocity and distance from the boundary are shown in the lower left hand corner, confidence interval and $-5/3$ slope are also plotted for reference. Spectra has 80 degrees of freedom.
LIST OF REFERENCES


