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Observations and Modeling of Turbulent Air-Sea Coupling in Coastal and Strongly Forced Conditions

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UNIVERSITY OF MIAMI

A dissertation submitted in partial fulfillment of
the requirements for the degree of
Doctor of Philosophy

OBSERVATIONS AND MODELING OF TURBULENT AIR-SEA COUPLING IN
COASTAL AND STRONGLY FORCED CONDITIONS

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The turbulent fluxes of momentum, mass, and energy across the ocean-atmosphere boundary are fundamental to our understanding of a myriad of geophysical processes, such as wind-wave generation, oceanic circulation, and air-sea gas transfer. In order to better understand these fluxes, empirical relationships were developed to quantify the interfacial exchange rates in terms of easily observed parameters (e.g., wind speed). However, mounting evidence suggests that these empirical formulae are only valid over the relatively narrow parametric space, i.e. open ocean conditions in light to moderate winds. Several near-surface processes have been observed to cause significant variance in the air-sea fluxes not predicted by the conventional functions, such as a heterogeneous surfaces, swell waves, and wave breaking. Further study is needed to fully characterize how these types of processes can modulate the interfacial exchange; in order to achieve this, a broad investigation into air-sea coupling was undertaken. The primary focus of this work was to use a combination of field and laboratory observations and numerical modeling, in regimes where conventional theories would be expected to breakdown, namely: the nearshore and in very high winds. These seemingly disparate environments represent the marine atmospheric boundary layer at its physical limit. In the nearshore, the convergence of land, air, and sea in a depth-limited domain marks the transition
from a marine to a terrestrial boundary layer. Under extreme winds, the physical nature of the boundary layer remains unknown as an intermediate substrate layer, sea spray, develops between the atmosphere and ocean surface. At these ends of the MABL physical spectrum, direct measurements of the near-surface processes were made and directly related to local sources of variance. Our results suggest that the conventional treatment of air-sea fluxes in terms of empirical relationships developed from a relatively narrow set of environmental conditions do not generalize to the coastal and extreme wind environments. This body of work represents a multi-faceted approach to understanding physical air-sea interactions in varied regimes and using a wide array of investigatory methods.
For Tamra M. Suslow
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Chapter 1

Introduction

1.1 Opening Remarks

The atmosphere and ocean are dynamically coupled across a molecules wide interface that covers over two thirds of the Earth’s surface, giving this celestial body its defining idiosyncrasy: blue. At the interface between air and sea, these geophysical fluid bodies exist at human scales of time and space and the ocean surface has been the medium of anthropic culture, commerce, and legend for millennia. The air-sea interface has also been the subject of intense and long-standing scientific investigation. This largely began in the mid 1800’s with investigations into one of the ocean surface’s most emblematic features, free surface gravity waves [Russell, 1845; Stokes, 1847]. Continuing into the early twentieth century, detailed studies were conducted on how waves were generated, through the specific physical interactions between air and water [Jeffreys, 1925; Miles, 1957], and relating the measured wind speed over the water to the physical roughness of the surface [Charnock, 1955]. Deeper understanding was attained with new insights into fluid mechanics and turbulence [Prandtl, 1925; Kolmogorov, 1941; Monin and Obukhov, 1954]. With advancing technology, the observation and numerical modeling of air-sea interaction has significantly progressed into a critical component of the
general understanding of the geophysical system from local to global scales.

Fundamentally, the study of atmosphere-ocean coupling requires an understanding of the four dimensional turbulent interactions at the fluid interface. Theoretically, the full Navier-Stokes equations, i.e. Newton’s Second Law formalized for a continuum, captures these dynamics at all scales. The Navier-Stokes equations prescribe the evolution of fluid momentum density based on the various forces acting on the fluid volume. These forces must be known in time and space and are not necessarily independent of the fluid’s evolution. Therefore, this set of nonlinear and dissipative equations must be solved numerically; and a significant effort has been expended on determining the most robust and efficient means of solving these equations for geophysical problems. Determining the functional form of the forcing terms is critical to modeling the system and so these typically unknown fields are parameterized (or approximated) using sets of known\(^1\) quantities. For air-sea coupled problems, one of the most important forcing terms is the tangential shear stress applied to the ocean surface by the atmosphere. Or, in other words, the vertical transfer of horizontal momentum from the atmosphere to the ocean. As any fluid flows over a surface, the fluid column experiences a viscous shear stress in the plane of the surface, which is due to the physical interactions between fluid particles and that surface. In the case of an infinite, laboratory flume with frictionless walls and rough bed, this tangential shear stress is fairly predictable. In terms of the geophysical atmosphere flowing over a permeable, deformable surface (the upper ocean) the problem becomes quite significant. Therefore, it has become one of the major endeavors of air-sea interaction research to develop a means of representing the tangential shear stress (or wind stress) in terms of parameters that are either universally known or

\(^1\)If necessary, we can substitute known for: easily known, readily known, acquirable, or generally-assumed-to-be known.
easily quantifiable from simple relationships with other known parameters\(^2\). The permeability of the ocean surface expands the motivation for understanding the vertical exchange between air and sea beyond just momentum and characterizing the exchanges of energy and mass are equally important to a wide variety of biogeophysical problems.

Wind stress parameterizations are theoretically based [Prandtl, 1925], but are in large part empirically derived and thus are flow, not fluid, specific. Thus, the observational goal for the past century or so, has been to amass data sets over a variety of oceanic and atmospheric conditions to create statistically reliable functions in terms of bulk, mean environmental parameters, e.g. wind speed, air temperature, air pressure, and humidity. Up until relatively recently, the flow regime which had been the primary focus for observational work was, a light to moderate wind over the open ocean. From an investigator’s perspective, this flow condition represents the majority of the global air-sea interface on an average day. Over decades, methods for data acquisition, processing, and interpretation have significantly developed such that in these conditions the wind stress over the ocean surface is fairly well-known. However, there are clearly geophysical flow regimes that exist outside of this relatively narrow set of environmental conditions. Since these parameterizations are empirically based and flow-dependent, it would not be expected that the relationships developed over the open ocean for certain wind speeds be universally applicable outside that specific regime. Ultimately, the goal of building these empirical relationships is to provide numerical model input; a wind stress parameterization largely based on inherently limited data sets constrains its applicability in models and confines the ability to accurately simulate a wide variety of flows.

The goal of this dissertation is to expand the general understanding of air-sea cou-

\(^2\)It is easy to understand what you know, rather than what you don’t. This is assuming, of course, that you know what you know can be used instead of what you don’t know.
pling to flow regimes that have been relatively under-explored in previous investigations: coastal environments and strongly forced conditions (i.e., very high winds). This work poses two over-arching questions: Do general open ocean wind stress parameterizations breakdown in these two domains? If so, what processes specific to those domain causes this breakdown? Answers to these questions will be developed using a combination of field and modeling work, for the coastal environment, and a laboratory study, for the strongly forced conditions. While both of these domains, coastal environments and high winds, are disparate in nature they provide the opportunity to investigate the role specific processes have on modulating the turbulent exchanges across the air-sea interface. For coastal environments, the fundamental process under investigation is heterogeneity. Over the open ocean, the surface can be considered fairly homogeneous and stationary (within a reasonable time frame), but in the nearshore these assumptions literally reach their limits. In strongly forced conditions, the entrainment of large volumes of sea spray creates an all-together new interface system as an intermediate spray layer develops. Therefore, the notion of an air-sea interface essentially breaks down. Further understanding of air-sea coupling in these two regimes will provide new insights into mechanisms for interfacial exchange and could be significant for observation and modeling studies in these two environments.

1.2 Specific Research Objectives

Presented below are the three specific hypotheses this dissertation aims to address as part of its overall objectives.
1.2.1 Hypothesis 1: Coastal air-sea fluxes are generalizable by bulk, open ocean parameterizations.

This hypothesis seeks to answer these research questions: Do conventional wind stress parameterizations represent the air-sea momentum flux in coastal waters? If not, which coastal processes explain the divergence? What are the relevant temporal and spatial scales to the observed variability? At the on-set, the answer to the first question seems a fairly straightforward, no. However, given that the latter questions have yet to be fully described within the literature, it is necessary to answer this first, albeit simple, question and provide some quantifiable justification. Several field observational data sets were used to evaluate this hypothesis and address these research questions.

Figure 1.1: A 0.6 m resolution panchromatic satellite image of the New River Inlet in North Carolina. A) A zoomed-out view of the entire area; B) a focused frame on the inlet mouth; C) a zoom-in on the box in (B) which is the small research vessel used in the experiment. The vessel is making a sharp turn and it is possible to see the trailing bow wake in the image.

Figure 1.1 is an optical satellite image of the New River Inlet in North Carolina,
a site from which a portion of the field data used in this work was taken. This image highlights the complex interactions that characterize the coastal zone. There is clear evidence of two discrete buoyant plumes outflowing from the inlet. This ebbing flow-meets an incident wave field and there is some evidence of wave-current interaction on the shoals and even propagating up the inlet channel. Depth-limited wave breaking is evidenced on the shoals and in the surf-zone along the shore line. This image also captured long filaments and surface streaks or slicks to the north and south of the inlet. This single frame captures a host of uniquely coastal processes and complex dynamics which could all have a profound effect on the air-sea exchange in the nearshore region. In addressing this research objective we will increase the breadth of the air-sea interaction literature as well as provide insights which have direct relevance to characterizing fundamental coast processes such as surface transport, mixing, and the interfacial gas flux.

1.2.2 Hypothesis 2: The circulation in an operational, process-based model is insensitive to the surface wind forcing.

The research questions under investigation for this hypothesis are: What is the role of the wind forcing on the hydrodynamics at a location like New River Inlet? Does the surface wind forcing contribute significantly to the hydrodynamic momentum balance? With the advent of tides? Or waves? Is there output sensitivity to changing the applied wind forcing parameterization? A series of numerical experiments were conducted in order to test this hypothesis and address these questions in an operational coastal hydrodynamic model. In answering these questions, the generalization of the field observations presented in Chapter 2, that the surface wind forcing exhibits a high degree of spatial and temporal variability in the coastal zone, will be assessed.
In the nearshore, the hydrodynamics are expected to be controlled by the waves, via the radiation stresses, and the tidal flow, while the wind contribution to the flow is expected to be of secondary or minimal importance. A significant amount of effort has been expended in describing the role waves and tides have on typical nearshore processes such as sediment transport, the depth-averaged flow, and mixing. However, relatively little has been done in describing the role the wind forcing on the surface plays in order to confirm or reject the conventional wisdom. A series of numerical experiments were conducted in order to test this hypothesis and address these questions in an operational coastal hydrodynamic model.

1.2.3 Hypothesis 3: *Large droplet spray production is important for spray-mediated fluxes in high winds.*

This hypothesis aims to address the following questions: What is the size-dependent distribution of sea spray above strongly forced wind-waves? Are there discrepancies
between observations and conventional spray models? What is the significance for air-sea interaction in these extreme conditions? The focus of this work will be on laboratory observations of large spray droplets (radius >50 µm), typically referred to as spume.

Figure 1.3: A sample image from the laboratory data set. This image is an optical, shadow image generated from the wave passing between the camera and the light source. This highlights the complex physics directly at the air-sea interface in these extreme conditions. The waves on the bubble sheet are O(0.5 mm) in wavelength.

To-date, the body of ocean spray literature does not have a consensus on the rates of large droplet spray production in very high winds [Andreas et al., 2010; Veron, 2015]. Several hypotheses have been put forward, but major progress in terms of theory, models, and observations has been fairly limited within a relatively narrow set of contributors. This could be largely attributed to two things: 1) The physical complexity of the problem and 2) The dearth of reliable observations. One of the principal aims of the work included in this dissertation was to widen the breadth of available high quality, direct observations of sea spray in wind speeds >30 ms\(^{-1}\). A series of laboratory experiments were conducted which will be used to evaluate the posed hypothesis and help shed light on this mystery topic within the air-sea interaction community.
1.3 Structure of this Dissertation

This dissertation is structured around three chapters, each dealing with a different venue for investigating air-sea coupling: field observations, a modeling study, and laboratory experiments. Each chapter is focused on addressing one of the hypotheses posed in Section 1.2. Due to the disparate nature of these three individual studies, each chapter will contain the relevant introductory and methodological material for understanding how the results were acquired, interpreted, and their significance. Each chapter will also end with a final conclusions where, based on the presented findings, the hypothesis posed for that chapter will be rejected or affirmed. This dissertation will end with a final discussion generally summarizing this entire body of work, as well as a discussion of future efforts.
Chapter 2

Field Observations of Coastal Air-Sea Interaction

Re-statement of Hypothesis

Coastal air-sea fluxes are generalizable by bulk, open ocean parameterizations. This hypothesis seeks to answer these research questions: Do conventional wind stress parameterizations represent the air-sea momentum flux in coastal waters? If not, which coastal processes explain the divergence? What are the relevant temporal and spatial scales to the observed variability? At the on-set, the answer to the first question seems fairly straightforward, no. However, given that the latter questions have yet to be fully described within the literature, it is necessary to answer this first, albeit simple, question and provide some quantifiable justification. Several field observational data sets were used to evaluate this hypothesis and address these research questions.
2.1 Background

The near shore is a dynamic and complex environment where many physical processes converge in relatively shallow depths. In addition to the physical importance, the ocean’s coasts hold significant value for economic and human health-focused environmental management. This motivates the development of sophisticated coastal models to simulate current and wave fields, as well as sediment and tracer transport. For example, Feng et al. [2013] used a physics-based near shore model, coupled with a microbe transport-decay module, to hindcast fecal indicator bacteria levels at a popular South Florida beach. Fujimura et al. [2014] used a biological-physical coupled model to simulate the cross-surf zone transport of larval invertebrates. They found that the transport was significantly dependent on the whether or not there was on-shore wind forcing included in the simulation. Like many near shore modeling platforms, these studies maintained a simple, open-ocean derived wind stress parameterization, which is at best a simplistic, first order approximation [Shabani et al., 2014].

For decades, air-sea momentum flux research has focused on developing a parameterization scheme that can be universally applied over a wide range of wind speeds and sea states. While this has yet to be fully realized, the important role that the wind stress plays in a number of physical processes (e.g. wind-wave generation, surface transport, and mixing), necessitates the continued effort to develop a complete wind forcing model that may be applied from the ocean basins to the coastal shorelines.

Longstanding seminal work done by Charnock [1955], led to the general principle that the wind stress increases with wind speed, where the shear stress is taken as,

\[ \tau = \rho u^* \mathbf{|U_z|}^{2} = \rho C_z |U_z|^2 \]  

(2.1)

here $u_*$ and $U_z$ are the friction velocity (or wind shear velocity) and the wind speed
(referenced to some height, \( z \), above the air-water interface), respectively. The air density is \( \rho_a \) and \( C_z \) is the dimensionless aerodynamic drag coefficient. Decades of study have been dedicated to characterizing this coefficient as some function of wind speed and sea state, but the exact nature of \( C_z \) remains elusive.

Subsequent investigations used a variety of methods to confirm the general nature of the Charnock model, i.e. the drag coefficient increases linearly with wind speed [Smith and Banke, 1975; Garratt, 1977]. This relation was later piece-wise defined for wind speeds generally between 5 m/s and 25 m/s [Smith, 1980; Large and Pond, 1981]. However, significant scatter existed in the literature for low and high wind regimes. Later work revealed a sharp increase in the atmospheric drag for winds approaching 0 m/s [Edson et al., 2013], which Zhu and Furst [2013] demonstrate cannot be explained by Monin-Obukhov Similarity (MOS) Theory [Monin and Obukhov, 1954]. Among limited field observations, Donelan et al. [2004] used laboratory data to show that the drag diverges from linearity at extremely high wind speeds (~ 30 – 35 m/s).

Much of this early work assumed a wind speed dependent roughness formulation,

\[
z_0 = mu^2 / g
\]  

(2.2)

where \( m \) is some dimensionless constant and \( g \) is the gravitational acceleration. However, there has been a significant amount of work done to relate the roughness length, \( z_0 \), to the wave field underneath the atmospheric boundary layer. Since being first proposed by Kitaigorodskii and Volkov [1965], some investigators have sought to relate the atmospheric drag to both the wind- and swell-seas underneath the atmospheric boundary layer [e.g., Donelan, 1990; Anctil and Donelan, 1996; Drennan et al., 1999; Sahlée et al., 2012]. Edson et al. [2013] provides a detailed review of the methods used to parameterize \( C_z \) as some function of wind speed or sea state.
The citations referenced above provide the breadth of techniques and platforms that have been used to measure air-sea momentum fluxes in the field and in the laboratory. Towers, moored and floating buoys, ships, and wind-wave flumes have all been used in the concerted effort to better understand air-sea turbulent exchange in a wide range of ocean and atmospheric conditions. Much of the prior work only considered the open ocean, deep water regime where the seas were near full development [Anctil and Donelan, 1996] and the wind stress was assumed to be aligned with the wind vector [Geernaert, 1988]. However, Zhang et al. [2009] reported that in the presence of strong horizontal current shear, the peak of the wind sea directional spectrum and the wind stress vector were steered off the mean wind direction. Generally, in equation (2.1) the friction velocity is defined as, \( u^2_* = -u_i u_j \), where the only component of the Reynolds' stress considered is the component aligned with the wind direction and off-wind stresses are assumed negligible [Smith, 1980].

There exists a gap in the atmospheric boundary-layer literature detailing the role of shallow water processes on the momentum flux, including current shear [Zhang et al., 2009], intense wave breaking and decreased wave celerity during shoaling [Shabani et al., 2014], and slanting fetch [Ardhuin et al., 2007]. To the authors’ knowledge the suitability of open ocean derived methods for use in coastal waters has yet to be fully tested. Therefore, the aim of this study is to extend our understanding of air-sea momentum fluxes to the currently under-explored coastal environment [Edson et al., 2013]. Large scale processes (e.g., incident wave height and regional circulation) are independent of locally variable wind stresses. However, the near shore wind forcing has been shown to affect littoral transport [Fujimura et al., 2014] and circulation patterns [Wargula et al., 2014]. A more realistic wind stress parameterization could explain the variability in both near shore observations and simulations that generally relied on an
open ocean drag coefficient.

Here, the results from three field observation data sets are presented with the aim of highlighting the complexity of wind stress field in the nearshore zone and providing the coastal processes which contribute to the variability and explain its divergence from the theoretical values. The focus of the analysis will be on the direct mechanic forcing on the ocean surface, i.e. the wind stress vector, but this work fundamentally is an examination of the nature of the sea surface in a coastal regime and bears implications for understanding scalar fluxes such as heat and gas exchange across the interface. This chapter is organized as follows: the observational and analysis methods will be presented, the results from the three independent data sets will be given, and this is followed by a discussion of these findings and their implications for future efforts.

2.2 Methods

2.2.1 Field Data Collection Strategies

For each field data set described within this chapter, a similar measurement philosophy was employed, that being to simultaneously collect a set of atmospheric and oceanographic data that would enable direct measurements of the air-sea fluxes, near-surface currents, surface gravity waves, and other bulk environmental parameters\(^3\). All of this capability was to be deployed on mobile platforms that could navigate the coastal waters from the estuary to the continental shelf. This data collection strategy sought to transfer the measurement capabilities developed over the open ocean, primarily for buoy platforms [e.g. Drennan et al., 1994; Graber et al., 2000; Sahlée et al., 2012; Potter et al., 2015], to small coastal vessels. While stationary platforms provide excellent time se-

\(^3\)Generally, a set of optical and infrared sensors were deployed as part of the measurement package, this will not be the focus of this dissertation, interested readers should see Laxague et al. [2015] for an example. Or Laxague [2016] for even more further details.
ries, there obvious short-coming is a lack of any spatial resolution. With the increase in the capabilities of computational and satellite technologies, there is a growing need for *in situ* observations to provide meaningful validation for these other analytical tools. Numerical models and satellite remote sensing are inherently spatial data sets and therefore field data must match this quality. This is especially important in the coastal zone where strong spatial heterogeneity is expected as hydrodynamics and atmospheric flows converged at the land-air-sea boundary.

For each field experiment, the sensor and/or platform design followed these general guidelines (in no particular order):

1. Minimize upstream flow distortion
2. Minimize residual sensor motion
3. Maintain clock consistency
4. Minimize platform impact on measurements
5. Maintain platform mobility

These constraints guided the design of the sensor packages as well as the measurement strategy. Guidelines 1 and 4 are of particular importance for the air-sea fluxes. The eddy covariance technique (section 2.2.3) relies on some statistical assumptions of the eddies advected within the boundary layer and distorted eddies could greatly effect the outcome of the analysis. "Distorted" generally refers to physical interaction between the flow and some structure (a ship railing, guy-wire, etc.) upstream of the measurement volume. A general rule-of-thumb for the minimum distance observations should be made from any non-geophysical vortex generators is 5 diameters (of the vortex gen-
erator). This comes directly from Stokes flow,

\[ \psi = -\frac{1}{2} U_\infty r^2 \left[ 1 - \frac{3}{2} \frac{R}{\sqrt{r^2 + x^2}} + \frac{1}{2} \left( \frac{R}{\sqrt{r^2 + x^2}} \right)^3 \right], \]  

(2.3)

where \( \psi \) is the stream function, \( U_\infty \) is the free-stream velocity, \( R \) is the object radius, and \((r,x)\) are the cylindrical coordinates of the system. While this is used as a measurement guideline, this equation is only valid for uniform free-stream flow and spherical or cylindrical objects. Therefore, while it is a rule-of-thumb, best practices dictate a less quantitative, but possibly more instructive solution: very far away. This consideration must be made for both atmospheric and oceanographic instruments, especially when the object being measured is the flow itself. The platform can also effect measurement quality in other ways, for example large energy sources (i.e. motors or engines) and ferrous material (i.e. magnetic). For wave measurements, consideration has to be made for minimizing the platform effect on the ambient wave forms or from measuring "vessel waves" generated by surface gravity waves interacting with the hull.

### 2.2.2 Field Data Processing

If the above is the "measurement algorithm", the data processing algorithm is of equal importance to the overall quality of the data set. The processing algorithm used for all of the data sets combined a series of quality assessment (QA) and quality control (QC) steps with analytical steps specific to recovering the air-sea fluxes. The general algorithm outline followed flow given in Figure 2.1.
An important step in maintaining data quality is defining the methods used for outlier detection, rejection, and gap-filling. For the wind velocity data a Goring and Nikora [2002] type outlier detection and rejection algorithm was used [Mori et al., 2007]. The results of this method on a created data set is given in Figure 2.2. This method was developed for Acoustic Doppler Velocimeter (ADV) data in noisy conditions, e.g. in the presence of bubbles. In these conditions, spikes appear as clearly defined, usually single-value peaks in the time series. Outlier data points are flagged by fitting an ellipsoid surface to the data cloud generated by the 3D space defined by \([u, \partial u/\partial t, \partial^2 u/\partial t^2]\); all data points outside of the ellipsoid surface are identified as outliers and removed [Goring and Nikora, 2002]. For some sensors, this method was not optimal and so a
simple adaptive Gaussian outlier detection method was used. Gaps left in the record from rejected outliers were generally filled with a piece-wise cubic polynomial. For some data time series, a spline interpolant was used (e.g., waves).

Figure 2.2: Mori et al. [2007] algorithm applied to simple sinusoidal wave with noise and spikes added into the signal.

Determining an appropriate averaging interval for the flux analysis is critically important to the analysis and correctly interpreting the processed data. The Reynolds stress (i.e., the wind forcing on the surface, see next section 2.2.3) is a covariance between the along and off-wind components of the turbulent velocity field and the vertical wind. This operation imposes some discrete time averaging interval on the data set. The amount of time that is deemed "appropriate" varies within the literature, but usually ranges from 15 to 30 minutes. For example, Shabani et al. [2014] used a 15 minute averaging window, Anctil and Donelan [1996] used a 20 minute interval, and Potter [2014] used 30 minutes for flux calculations. The assumption built-in to this approach is that the wind field is ergodic and that the interval is long enough to resolve all of
the turbulent eddies in the atmospheric boundary layer. In the case of a moving platform, this time averaging interval incorporates a spatial averaging interval. Therefore, a balance had to be struck between the constraints of the method and the measurement strategy. For most cases within this data set, an averaging interval of close to 5 minutes was used for the covariances. If another interval was used, it will be given explicitly.

The primary limitation of using such a short averaging interval is that it universally under-samples or misses the low frequency contributions to the total flux (see Figures 2.3 and 2.4). Thus, our flux estimates alone cannot be generalizable as absolute measurements of the air-sea momentum flux. However, our estimates will reliably capture the high frequency signal in the fluxes, which would be expected to carry the bulk of the response to coastal forcing (as opposed to low frequency atmospheric variability that will be driven by larger time scales less sensitive to fine scale, coastal variability). This limitation was mitigated by reporting estimates of the drag coefficient as a ratio between our directly measured drag and a bulk formula representation for the same 5 minute interval.

The motion correction algorithm follows the method developed by Anctil et al. [1994]. This critical step removes the platform linear accelerations and angular velocities from the clean wind velocity signals. These motions are generated from the platform moving across a wavy ocean surface and the corrected wind velocities are transformed into an earth reference frame (i.e. not in a wave-following reference frame). Essentially, the corrected wind velocities are calculated from the raw apparent winds

---

4This study presents the method as applied to the fluxes from a 3-m discus buoy. The method, in general, was developed during the SWADE experiment in order to directly measure eddy covariance fluxes from moving platforms, e.g. buoys and ships. See Drennan et al. [1994] for an example using measurements from a SWATH vessel.
and the simultaneously measured six-degrees of motion:

\[
\mathbf{u} = \mathbf{T} \cdot \mathbf{u}_{\text{raw}} + \mathbf{T} \cdot \int (\mathbf{a} + \mathbf{g}) dt + \Omega \times \mathbf{T} \cdot \mathbf{L},
\]

(2.4)

where \( \mathbf{u} \) is the three-dimensional motion corrected wind velocity observed by the anemometer. From the left-to-right, the RHS terms are: the tilt-corrected apparent wind velocity, the tilt corrected platform velocities, and the angular velocity component as a result of the measurement not being taken at the center of mass and assuming the platform is a rigid body (\( \mathbf{L} \) gives the moment arms of the wind sensor relative to the motion pack, in the vessel coordinate system). \( \mathbf{T} \) is the three-dimensional transformation matrix,

\[
\begin{vmatrix}
\cos \theta \cos \psi & \sin \phi \sin \theta \cos \psi - \cos \phi \sin \psi & \cos \phi \cos \cos \psi + \sin \phi \sin \psi \\
\cos \theta \sin \psi & \sin \phi \sin \theta \sin \psi + \cos \phi \cos \psi & \cos \phi \sin \theta \sin \psi - \sin \phi \cos \psi \\
-\sin \theta & \sin \phi \cos \theta & \cos \phi \cos \theta \\
\end{vmatrix},
\]

the Euler angles, \([\theta, \phi, \psi] \), are the pitch, roll, and yaw measured by the motion pack. The motion pack axes are aligned with the right-handed vessel coordinate system, where \( x \) is the bow-stern line, \( y \) is the port-starboard, and \( z \) is the vertical axis. The angular velocities, \( \Omega \), can also be defined using these angles:

\[
\begin{vmatrix}
-\dot{\theta} \sin \psi + \dot{\phi} \cos \theta \cos \psi \\
\dot{\theta} \cos \psi + \dot{\phi} \cos \theta \sin \psi \\
\dot{\psi} - \dot{\phi} \sin \theta
\end{vmatrix}.
\]

The motion corrected wind velocities are then rotated into the mean vertical plane (a rotation about the \( y \)-axis) and the rotated into the mean horizontal plane defined by the mean azimuthal wind speed (rotation about the \( z \)-axis). This final step, known as the double rotation, ends with the velocities in a wind-centric coordinate system: \( x \) is aligned with the mean wind direction, \( y \) is the across-wind component, and \( z \) is the
vertical. This series of corrections was done for each averaging interval of data. For all the data sets, the translational velocity of the platform was removed using 1 Hz GPS data. Edson et al. [1991] reported that the inertial dissipation method is a viable option for retrieving fluxes from high frequency winds that does not require boat-motion correction. However, it is generally accepted that, if the vessel motion can be quantified to high enough frequencies, the direct method is the preferred technique (see discussion about this method in section 2.2.3).
Figure 2.3: An example of the results of the Anctil et al. [1994] motion correction algorithm applied to a 6 minute segment of vertical wind velocity observed in Monterey Bay on June 13, 2016 at 23:00 UTC. Left) The corrected vertical wind spectra along-side the different motion correction components. $W_e$ is the raw, tilt-correct velocity and the axial and radial components are the second and third terms, respectively, from the RHS of equation 2.4. The omni-directional wave spectrum observed during this segment is also given. Right) The corrected, normalized $S_{ww}$ compared to the corresponding curve from Miyake et al. [1970]. The wave spectrum is also given as a function of normalized frequency. $z$ and $U$ are the anemometer height above the MSL and the mean horizontal wind speed, respectively. $\sigma_w$ is the standard deviation of the vertical velocity. The wave spectrum, $S_{\eta\eta}$ was not scaled, only the frequency.

The goal of the motion correction algorithm is to remove the virtual wind velocity from the observations that were created by the vessel moving across a wavy-surface. An example of how the motion correction algorithm used for this work works on the vertical velocity observed from a sonic anemometer is given in Figure 2.3. From this example, the response of the vessel to the dominant waves is picked up in the axial wind velocity ($T \cdot \int (a + g)dt$) component and is successfully negated from the raw vertical wind component. And so, this signal is not present in the turbulence spectrum
when scaled into boundary layer coordinates following Miyake et al. [1970] (Figure 2.3-right). Similar results have been compiled for the entire data set used in the RIVET I analysis presented here and in Ortiz-Suslow et al. [2015] (Figure 2.4). All of the (co)spectra are given in boundary layer coordinates and compared to their corresponding universal curves from Miyake et al. [1970]. The compiled data represent averages over a 0.02 Hz wide frequency bin, means and variances were calculated independently for each bin.
Figure 2.4: Normalized (co)spectra from the RIVET I experiment, all data used in the analysis were compiled and averaged in equally spaced frequency bins. (Co)Spectra were individually normalized before compilation. The dark green line in all 3 panels gives the mean and the shaded value spans the 95% confidence interval (two times the standard error of the means). The dashed curves give the corresponding curves from Miyake et al. [1970]. The normalized frequencies are given in boundary layer coordinates with $z$ and $U$ being the measurement height and mean wind speed at $z$, respectively. The $-5/3$ slope from the inertial sub-range is also given as a reference. This is the corrected version of Figure 3 in Ortiz-Suslow et al. [2015]. Some notes on the observed cospectra: 1) The cut-off of the low frequency flux components is due to a combination of the averaging interval (5 minutes) and the frequency bin-averaging done to smooth and compute confidence intervals; 2) The $U$ used to transform to boundary layer coordinates was the absolute wind speed relative to ground, this explains the shift of the spectra to the low frequencies.
2.2.3 The Eddy Covariance Technique

Turbulent, incompressible flows can be described using the Reynolds-averaged Navier-Stokes Equation (RANS equation),

\[
\rho \left( \frac{\partial U_i}{\partial t} + U_j \frac{\partial U_i}{\partial x_j} \right) = F_i + \frac{\partial}{\partial x_i} \left[ -P \delta_i + \mu \left( \frac{\partial U_i}{\partial x_j} + \frac{\partial U_j}{\partial x_i} \right) - \rho \overline{u_i u_j} \right]
\]  

(2.5)

where the instantaneous flow has been separated into its mean (U) and fluctuating (u) components. The left hand side of equation (A.6) represents temporal and spatial changes in the momentum, which are balanced by the body forces and the stresses, from left to right: the mean pressure gradient, the mean strain, and the Reynolds stress [Tennekes and Lumley, 1972]. For wind stress analysis, it is generally assumed that over some appropriate time interval the momentum can be considered stationary, that there is no mean advection, and that at some height above the water surface (e.g. 10 m) viscous stresses are negligible. If isotropy is invoked, and body forces are neglected, equation (A.6) reduces to

\[
\vec{\tau}_{Total} = \vec{\tau}_{Re} = \text{const.}
\]  

(2.6)

where the Reynolds stress, \( \vec{\tau}_{Re} \), is

\[
\vec{\tau}_{Re} = -\rho_a \left( \overline{uw\hat{i}} + \overline{vw\hat{j}} \right) = \vec{\tau}
\]  

(2.7)

and \([u, v, w]\) are the turbulent fluctuations of the flow aligned with the local wind direction (i.e. the along, off, and vertical wind components, respectively). The overbar represents a suitable time average. A stress angle, \( \theta \), may be defined as

\[
\tan(\theta) = \overline{vw}/\overline{uw}
\]  

(2.8)

This angle is taken as a rotation about the vertical-wind and is referenced to the mean along-wind velocity (i.e. \( \theta = 0 \) means the wind stress and wind velocity are in-line).
The temporal mean of the vector quantities is taken prior to applying the arctangent. From equation (2.7) and (2.1), the magnitude of the wind shear velocity (or friction velocity) becomes

\[ u_\ast = \left[ \overline{uw^2} + \overline{vw^2} \right]^{1/4} \tag{2.9} \]

Here the off-wind turbulent fluctuations are explicitly considered. Through the dimensional arguments made by Monin and Obukhov [1954], the mean wind speed profile is related to the shear velocity and a dimensionless gradient function,

\[ \frac{dU}{dz} = \frac{u_\ast}{\kappa z} \phi_m(\zeta) \tag{2.10} \]

where \(z\) is the measurement elevation, \(\kappa\) is the Von Kármán constant, and \(\phi_m\) is the nondimensional velocity gradient function for momentum, which is dependent on the stability parameter \(\zeta = z/L\). The Monin-Obukhov Length, \(L\), is defined as

\[ L = -\frac{u_\ast^2 \bar{\theta}_v}{g \kappa \theta'_w} \tag{2.11} \]

Where \(\bar{\theta}_v\) is the mean virtual temperature in air, \(\theta'_v\) is the fluctuating component. By integrating equation (2.10), the mean wind speed as a function of height is found to be,

\[ U(z) = \frac{u_\ast}{\kappa} \left[ \ln \frac{z}{z_0} - \psi_m(\zeta) \right] \tag{2.12} \]

where \(z_0\) is the roughness length scale (see equation 2.2) and \(\psi_m(\zeta)\) is the integrated form of \(\phi_m(\zeta)\). This nondimensional function corrects the logarithmic wind speed profile for non-neutral stratification [Anctil and Donelan, 1996]. Using equation (2.12) with (2.1), the height dependent drag coefficient is [e.g. Donelan, 1990]:

\[ C_z(z) = \left( \frac{u_\ast}{U_z} \right)^2 = \kappa^2 \left[ \ln \frac{z}{z_0} - \psi_m \left( \frac{z}{L} \right) \right]^{-2} \tag{2.13} \]
where the values for $\kappa$ and the non-dimensional buoyancy correction used in this study are those found in Anctil and Donelan [1996]. Conventionally, the drag coefficient and mean wind speed are referenced to 10 m above the air-sea interface, so that $C_z(z = 10) = C_D$ and $U_{z=10}(10) = U_{10}$. Shabani et al. [2014] provides a detailed review of this method for deriving $C_D$ and demonstrates its applicability to wind stresses measured over the surf-zone.

An Alternative: the Inertial Dissipation Method

The eddy covariance technique is a well-established method in air-sea interaction research and was used here to parameterize the atmospheric drag. The inertial dissipation method is another means of extracting drag coefficient estimates from high frequency wind data [e.g., Large and Pond, 1981; Edson et al., 1991; Yelland and Taylor, 1996]. This method relies on Kolmogorov’s hypothesis of the universal shape of the turbulence spectrum of the down-wind component of the velocity over the inertial subrange. From Yelland and Taylor [1996], the turbulent dissipation rate, $\epsilon$, is related to the measured wind velocity, $U$, by,

$$S_{uu}(f) = K \epsilon^{2/3} f^{-5/3} (U/2\pi)$$

(2.14)

where $K$ is the Kolmogorov constant, $f$ is the measurement frequency, and $S_{uu}(f)$ is the frequency-dependent power spectral density converted from wavenumber, $k$, using Taylor’s ”Frozen Turbulence” hypothesis. The inertial dissipation technique has advantages over the eddy covariance method in that it does not require motion correction, is less sensitive to flow distortions, and it is not contingent on a specific temporal averaging interval. The disadvantage of the inertial dissipation method is that it assumes the flux spectrum follows universal scaling (i.e. Miyake et al. [1970]). Therefore, this cannot explicitly take into account processes that would cause similarity theory to break-
down. Furthermore, determining the proper frequency band of the inertial subrange is very difficult and any results from an automatic processor were observed to be highly variable. Although, inertial dissipation has proved useful for a significant amount of air-sea flux research, these limitations discouraged its use as part of these studies.

2.3 Field Observations

Figure 2.5: (a) Google Earth image of the Southeast Coast of the United States. (b) A synthetic aperture radar (SAR) image (taken May 22, 2012) from TerraSAR-X of the inlet mouth, Intracoastal waterway, and southern end of the New River Inlet estuary, © InfoTerra GmbH 2012. (c) A close up of the inlet mouth with an Army Corps of Engineers bathymetric survey overlaid, here color indicates water depth relative to the mean water level. The star marks the local origin, where the shore-normal is X (increasing off-shore), Y and Z follow from the right-hand rule.
Figure 2.6: Time series of bulk meteorological and wave data. From top to bottom: $U_{\text{wind}}$ is 10 minute mean wind speed from an 8 m, land-based tower (location given in Figure 2.5c as "W"); $\theta_{\text{wind}}$ and $\theta_{\text{peak}}$, are the wind direction from the wind tower and the peak incident wave direction from an NDBC waverider buoy moored 6 km off-shore of NRI, respectively; $H_s$ is the significant wave height from the waverider; $T_D$ is the peak period; and finally, $Z$ is the 5 minute mean water surface elevation from a bottom-mounted pressure sensor (location given in Figure 2.5c as "P"). All wave parameters are calculated over 30 minutes. Magenta lines mark start times of SPEC runs (Table 2.1).
2.3.1 RIVET I

The Riverine and Estuarine Transport (RIVET) I experiment took place at NRI from April to June 2012. NRI is a natural, sandy-bottom tidal inlet, free from jetties or breakwaters, on the southeast coast of North Carolina located at 34.528 N and 77.338 W. Prior to the commencement of the experiment, the main inlet channels were dredged by the Army Corps of Engineers (Figure 2.5). The inlet divides North Topsail Beach to the south from Onslow Beach to the north and provides estuary-ocean exchange for the New River Estuary system and the Intracoastal Waterway. The tidal delta extends to a radius of O(1 km) [MacMahan et al., 2014] and is characterized by a large ebb shoal to the south of the main channel (Figure 2.5).

The observed tidal current range during the campaign was ±1 m/s [Clark et al., 2014]. A local significant wave height range of 0.5 m to 2 m and a wind speed range of 0 m/s to 15 m/s was observed, see Figure 2.6 [Wargula et al., 2014]. This presented an opportunity to observe wind stresses in a region with a high degree of spatial and temporal variability in both the wave and current field. Figure 2.5b shows a synthetic aperture radar (SAR) image taken during the RIVET experiment with significant pixel intensity variability in and around the inlet mouth. The brightness in SAR images is directly related to the surface roughness, which in turn is potentially indicative of wind stress variability at NRI.
Table 2.1: SPEC data used for flux analysis.

<table>
<thead>
<tr>
<th>Run</th>
<th>Day</th>
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<th>End</th>
<th>$U$</th>
<th>$\theta$</th>
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<td>10:20</td>
<td>11:35</td>
<td>5.23</td>
<td>213.8</td>
</tr>
</tbody>
</table>

In figures, all runs are used unless explicitly noted
All times are in UTC
Mean wind speed over entire interval ref. to 10 m
Corresponding down-wind direction w.r.t True north
Data Collection

The data presented here were collected aboard the Surface Physics Experimental Catamaran (SPEC). The motivation for developing SPEC was to transfer most of the measurement capabilities of the Graber et al. [2000] Air-Sea Interaction Spar (ASIS) buoy to a low-profile, mobile platform suitable for operation in the inner-shelf and coastal waterways. SPEC is a converted Glacier Bay 2680 Coastal Runner that is 8.2 m long, with a 2.7 m beam and a 0.5 m draft. Some additions to the basic vessel include a 4 m tall meteorological tower and an aluminum framework mounted on the bow for near-field remote sensing and water-side measurements (Figure 2.7).

The meteorological mast was fitted with an ultrasonic anemometer and a probe for fast-sampling of air temperature and relative humidity. The anemometer was mounted on the tower such that the sampling volume was 6.0 m above the mean water level. The bow-package included 8 wave staffs, 2 forward-facing Acoustic Doppler Velocimeters mounted 0.5 m below the mean water line (ADVs), and a downward-facing Ultrasonic Distance Meter (UDM) positioned 0.5 m above the mean water line (MWL). Custom camera mounts were
also secured on the bow for infrared and polarmetric imaging. A downward-looking ADCP was mounted on the starboard-bow 0.5 m below the MWL. SPEC was deployed with two sets of accelerometers and rate gyro packages, these were used to measure the high frequency (up to 10 Hz) vessel motions. This data is critical to recovering the momentum fluxes from the sonic anemometer.

Data from most of the sensor systems was routed via multi-conductor cable into the main cabin and wired into three Campbell Scientific CR3000 dataloggers. The camera and ADCP data was captured directly to computers mounted within the SPEC cabin. An internal ethernet network and a Brandywine Network Time Protocol Server (model NTV-100RG) were used to synchronize the various acquisition clocks. The present work will focus on the 10 Hz winds from the anemometer and the 0.8 Hz profiles from the ADCP. The data analyzed in Section 2.3.1 is summarized in Table 2.1 and an overview of conditions at NRI during the SPEC sampling is given in Figure 2.6. The sampling intervals roughly correspond to various satellite overflights, as well as when the flux package and the ADCP were sampling simultaneously.
Figure 2.8: Three different neutral $C_D$ parameterization algorithms using the wind velocities measured by SPEC. Large and Pond [1981] is piece-wise defined and remains constant for all wind speeds below 11 m/s. These algorithms were applied using equation 2.2 for the wind speed dependent roughness length, $z_0$.

**The Aerodynamic Drag Coefficient**

In coastal hydrodynamic modeling, the wind stress magnitude is generally parameterized by the wind speed and a non-dimensional drag coefficient (equation 2.1). Figure 2.8 provides wind stress parameterization schemes from Large and Pond [1981], Smith [1988], and Edson et al. [2013].
While these differ in their algorithm for calculating $C_D$, they remain relatively comparable over a small range of values between 0.0009 and 0.0013, for wind speeds between 0 and 10 m/s. The COARE 3.5 [Edson et al., 2013] is also capable of using a sea state dependent roughness length, however this was not employed for this study. For simplicity, the spatial distributions of the eddy covariance measurements of the atmospheric drag are shown in relation to a Smith [1988] method (Figure 2.9). The metric used to present the results of this analysis is the dimensionless parameter $R_{CD}$, defined as:

$$ R_{CD} \equiv \frac{C_D^{EC}}{C_D^{Smith}} \quad (2.15) $$

where $C_D^{EC}$ is neutral drag coefficient directly calculated using the eddy covariance technique (see equation 11) and $C_D^{Smith}$ is the corresponding bulk estimate using a Smith [1988] algorithm. This ratio between dimensionless coefficients compares the direct estimate of the neutral atmospheric drag to the open-ocean bulk parameterization for any
given wind speed, as both quantities are calculated over the same 5 minute segment of data. Using this ratio can be useful in mapping out regions where observed air-sea momentum fluxes diverge from expected values, however this comes with two significant caveats. Firstly, a 5 minute averaging interval for the eddy covariance fluxes systematically under-estimates the low frequency contributions of the total flux spectrum. Secondly, using this short averaging interval to calculate the mean wind speed used by the bulk parameterization to calculate the drag, increases the per sample variance of each drag estimate due to the gustiness (which increases with decreasing averaging length). Therefore, this short averaging window was done in order to preserve spatial variability, but the trade-off is some effect on the the frequency resolution and quality of individual estimates. Assuming that the second caveat is small at NRI, $R_{CD}$ would be expected to have a mean slightly less than unity.
Figure 2.10: The same data shown in Figure 2.9 filtered by two different criteria: wind direction (left) and wind speed (right). On the left, an on-shore wind (magenta) is defined as a wind coming from anywhere within 45° of the shore-normal line; an along-shore wind (red) was defined as a wind coming from within 45° of the shore-parallel line. An off-shore wind (blue) was defined to be 180° opposite of the on-shore wind criteria. On the right, the data corresponding to wind speeds greater than 5 m/s is shown in red, while data corresponding to wind speeds less than 5 m/s is shown in green.

Figure 2.9 shows a significant degree of scatter in the calculated $R_{CD}$ values. Generally, higher values are observed in and around the inlet mouth (i.e. near $x = 0$), while the observed variability tends to converge toward the bulk estimates in the furthest off-shore samples. A global average taken across all of the data yields a mean $C_D$ 2.8 ± 0.4 times greater than the corresponding bulk estimate. The uncertainties reported here and in the next two paragraphs are the standard error of the mean. To help explain the observed variability in the $R_{CD}$ values, wind direction and speed filters are applied to the across-shore distribution (Figure 2.10).

Along coasts, wind direction has a significant effect on the nature of the atmospheric boundary layer turbulence. In an attempt to account for the complex land-air-sea effects,
Figure 2.11: The cross-shore distribution of the $R_{CD}$ value corresponding to on-shore winds. The data has been separated by wind speeds above 5 m/s (red) and those below 5 m/s (blue).

a three-pronged wind direction filter is utilized [Shabani et al., 2014]. The wind is separated into off-, along-, and on-shore winds using the shore-parallel line as a reference. Figure 2.10 shows the results of this wind direction separation. The highest observed $R_{CD}$ values occur during either off-shore or along-shore winds; while, $R_{CD}$ calculated during on-shore winds do not exceed ten times the bulk drag estimate. The average $R_{CD}$ for both off-shore and along-shore winds are $3.7 \pm 0.6$ and $2.6 \pm 0.5$, respectively. The corresponding mean $R_{CD}$ for on-shore winds is $2.8 \pm 0.4$.

The drag coefficient has been shown to increase sharply at low wind speeds and approaching zero [Zhu and Furst, 2013]. To test this result on the NRI data, the $R_{CD}$ values were filtered wind speeds greater than or less than 5 m/s (Figure 2.10). The mean $R_{CD}$ for the higher wind speeds is $3.1 \pm 0.5$, while the mean for the lower wind speeds is $3.7 \pm 0.5$. Applying this wind speed filter to the on-shore wind data alone shows that for winds greater than 5 m/s the mean $R_{CD}$ is $2.6 \pm 0.5$; and for on-shore winds less
than 5 m/s the mean $R_{CD}$ becomes $4.2 \pm 0.7$ (Figure 2.11). For on-shore winds between 4 and 12 m/s, Shabani et al. [2014] observed drag coefficients roughly twice as high as any give open-ocean result. The results from SPEC for on-shore winds between 5 and 10 m/s fall within 30% of these results and within 10% of the global average reported above.

**Wind Stress Steering**

Zhang et al. [2009] showed that equations 2.8 and 4.15 are necessary in the presence of ocean surface current gradients, meaning the vector nature of the wind stress must be considered. The across-shore distribution of wind stress angles is investigated to discern any significant trend in the stress steering off of the mean wind direction (Figure 2.12). A 400 m wide spatial bin average filter was applied to the 5 minute mean stress angles, which were calculated from the mean across- and along-wind stress components (see equation 2.8). Between 0 and 1800 m across-shore, there was a high propensity for wind stress steering with extreme values exceeding $\pm 70^\circ$. Just off-shore of 2000 m there is significantly less scatter in the data and the wind stress angle generally converges to the mean wind direction. In the furthest across-shore values extrema are observed in the region of $\pm 50^\circ$. 
Figure 2.12: The spatial distribution of observed wind stress angles. The points represent the angle between 5 minute means of the along- and across-wind components of the wind stress. The solid line, with the open circles, gives the data filtered by a 400 m wide bin average. The shaded region spans the 95% confidence calculated from each bin. The red lines mark the ±26° line, which is the cut-off used in Smith [1980] to consider the wind stress in-line with the wind vector.

Wind stress deviations are an inherently noisy parameter [Edson et al., 2013] and in order to assess the significance of the mean spatial trends observed in the across-shore distribution, student’s t-tests were used to compare three different pairs of sub-populations. Each distribution was checked for a Gaussian shape and if necessary, the methods of Niaki and Abbasi [2007] were used to normalize the data in order to satisfy the assumptions of the t-test. Alpha levels were set to 0.05 for all statistical tests. The sub-populations were separated spatially, the scheme used for the statistical comparisons is given below (all distances are in meters).
\[ x \leq 1000 \quad vs. \quad 1000 < x \leq 2000 \]
\[ 1200 < x \leq 2000 \quad vs. \quad 2000 < x \leq 2800 \]
\[ x \leq 2000 \quad vs. \quad 2000 < x \]

The distribution of angles within the tidal radius of 1000 m [MacMahan et al., 2014] was compared to the angles between 1000 m and 2000 m. A significant mean increase in wind stress deviation from the wind direction was found (\( p = 0.003 \)). The second set of test regions was used to assess the transition in the mean wind stress deviation around the 2000 m mark. A significant (\( p = 0.008 \)) 69\% decrease in the mean wind stress angle was observed. The final test was done to investigate the degree of scatter in the furthest off-shore samples, relative to the in-shore data. The comparison done in this test reveals that these two particular sub-populations are not significantly exclusive (\( p = 0.73 \)).

Figure 2.13: A sampling of the ADCP measured current profiles as SPEC crossed the NRI shoals. This data was taken from Run 13. Shown is the two-dimensional along-shore (top) and across-shore (bottom) flow velocity, where the color scale applies to both panels. On the right, the corresponding SPEC GPS track, with the (0,0) marked in yellow.
Figure 2.14: The spatial distribution of the near-surface current direction (top) and depth averaged current direction (bottom) as observed from the ADCP. The 5 minute mean angles are shown as dots and the same spatial filtering done in Figure 2.12 are shown as solid lines. In both panels, the filtered wind stress deviation from the mean wind direction is shown as a black, dashed line.

It has yet to be determined what processes explain the observed variability in the across-shore stress distribution. Current shear has been shown to play a major role in steering the wind stress away from the mean wind direction [Zhang et al., 2009]. A sub-set of the ADCP measured currents, observed as SPEC crossed the NRI shoals, is given in Figure 2.13.

The spatial distribution of the current direction is shown in Figure 2.14, where the directions used are the near surface current (from 1.3 m below the MWL) and the depth averaged current direction. In order to investigate stress steering, Zhang et al. [2009] relied on HF radar to provide two-dimensional maps of the horizontal vorticity in the flow \( \left( \frac{dU}{dy} - \frac{dV}{dx} \right) \). The ADCP aboard SPEC cannot provide this same data, but the across-shore distribution of the 5 minute mean current direction can be used as a proxy for this higher quality measurement. The same spatial averaging scheme used for the
wind stress angle was applied to the ADCP data, comparisons were done between the mean trends in both variables. Qualitatively, the transitions in the spatially meaned wind stress deviation follow the current direction, especially within 2000 m of the inlet mouth; however, across the entire spatial domain there is a weak statistical relationship between the stress angle and current direction, of \( O(r = 0.1) \).

The wind stress angle variability was compared directly to the horizontal current shear and the air-sea heat fluxes, both of which have been shown to steer the wind stress off of the mean wind azimuth [Geernaert, 1988; Zhang et al., 2009]. Wave shoaling is another possible factor, but could not be considered here. The area covered by the moving platform in conjunction with the high spatial variability of the wave field at NRI rendered the wave-stress coupling analysis outside the realm of the present work. The current direction was found to be weakly correlated with the wind stress angle. However, if only data in-shore of 1800 m is considered the correlation becomes stronger (\( r = 0.72 \), for the near-surface current direction). Similar results are found for the depth averaged currents. This region coincides with the strongest current gradients (e.g. Figure 2.13) and studies have shown that the presence of strong, horizontal current gradients contribute to the wind stress steering [Haus, 2007]. This result suggests that the stress-current coupling is inhomogeneous around NRI and that specific spatial scales are important when considering these dynamics.

Another processes associated with wind stress steering is the degree of stratification in the marine boundary layer. Geernaert [1988] found a strong positive, linear relationship between wind stress angles and the heat flux observed from a platform in the North Sea. Following this work, the heat flux, \( < wT' > \), is compared to the stress angles measured by SPEC. Here, \( T' \) is the fluctuating component of the air temperature, measured by the high frequency temperature probe on the mast. The same technique as was
done to compare the stress angles to the current direction, was applied to the observed heat fluxes. A statistical comparison between the stress angles and the heat flux yields a weakly positive correlation \( r = 0.14 \).

Cross-correlation analysis was done to further investigate the spatial relationship between the wind stress and these variables. Again, considering region in-shore of 1800 m, the strong relationship between the stress angle and the near-surface current was confirmed at the zero lag. However, in regards to the heat flux (for the entire across-shore domain) a strong linear relationship was found with the stress angle at the +1 lag, which corresponds to a 400 m interval. Clearly, the horizontal current shear and atmospheric stratification play a significant role in steering the wind stress, however the exact mechanism for this remains obscure. The NRI outflow plume is assumed to be the primary source of horizontal variability in both variables and separating these contributions to the stress steering is beyond the scope this study.

**Summary**

The goal of this study was to evaluate the efficacy of open ocean wind stress parameterization methods in predicting the observed stresses at NRI, to attempt to explain the variability in the in situ data, and to augment the limited collection of air-interaction observations in coastal waters. To first order, the open ocean methods underestimate the atmospheric drag by a factor of 2.6; and the general assumption that the wind stress and wind velocity vector are in-line can only be reliably applied more than 2 km off-shore of NRI. In and around the tidal inlet, observed wind stresses significantly diverge from the previous work, with the 10 m drag exceeding 20 times the bulk method prediction and wind stress deviations from the mean wind azimuth surpassing ±70°. This suggests that uniquely coastal processes, such as strong current gradients, wave shoaling,
and depth-limited breaking, are misrepresented in the open ocean algorithms. The wind
direction (relative to the shoreline) and low wind speed conditions were successfully
used to explain the majority of the variability in the eddy covariance derived drag co-
efficients. The near-surface, horizontal current shear was used to explain a significant
amount of the observed stress angle variance near the inlet mouth. The observed ambi-
guity in the role the heat flux plays in the stress steering highlights the significance of
atmospheric stratification in these boundary layer processes.

2.3.2 RIVET II

RIVET II was the second installment in the RIVET project. The field campaign took
place during May and June of 2013 in the region of the Columbia River Mouth (CRM)
along the border between Oregon and Washington state. This environment provides a
stark contrast to the NRI field experiment. The Columbia River is one of the largest
rivers in western North America and the exchange through the CRM is significant to
dynamics along much of the North American continental shelf [Banas et al., 2009].
Furthermore, the massive flux of fresh water out of CRM on ebbing tides interacts
with large North Pacific swell systems to create a very dynamic plume edge where
wave breaking has been observed to be very intense [Thomson et al., 2014]. Given
the wave climate during the experiment, the strong river discharge from the snow melt,
and physical scale of the CRM, this field experiment presents a unique opportunity to
provide an analytical contrast to some of the observations made during RIVET I.
The Field Experiment

Figure 2.15: Clockwise from top left: Google earth image of the CRM with a SAR image overlay (courtesy of CSTARS), the sharp features just offshore of the North Jetty are the ebb plume edge (running roughly West-East) and the surface signature of a boat wake (running Southwest-Northeast, boat can be seen traveling East about 2 km inshore of the North Jetty), the two features meet just inshore of the jetty; forward deck of R/V Point Sur with meteorological mast (yellow) visible on the bow and the re-purposed ice house on the starboard quarter, the UDM used for data analysis here was mounted on bow-sprit at the base of the tower; image of a flooding tide intrusion front over the Columbia River Bar region; North Pacific swell waves interacting with ebbing tide to create these interesting rough crests and smooth troughs, image was taken inshore of the jetties.

The data collected at RIVET II relevant to this particular study was collected aboard the R/V Point Sur. The data package included a bow-mounted flux tower and a bow-mounted UDM array. Vessel motions were acquired from two motion packs, one
mounted underneath the flux tower and fastened inside a re-purposed ice fishing house on the forward deck (Figure 2.15). Supplementary data came from the R/V Point Sur UDAS, flow-through temperature and salinity, as well as the hull-mounted ADCP (300 kHz). The temporal range for this data set spans May 24 to June 13, 2013 and during that time observations of wind, waves, and currents were made over a very wide spatial domain (see Figure 2.16). Whereas the RIVET I data set at NRI spanned roughly \( \sim 36 \text{ km}^2 \), the CRM data spans \( \sim 3000 \text{ km}^2 \) from the furthest offshore approaching the Astoria Canyon to the inshore extent reaching the upper estuary. The conditions at CRM were fairly typical of spring time in the Pacific Northwest, with winds from either the northwest or southwest and average offshore significant wave heights just under 2 m (Figure 2.17). The waves were generally westerly or northwesterly, with one period early in the experiment with large southerly waves. In general, the winds observed from the R/V Point Sur seem to agree with the NDBC 3-m discus buoy at 20 NM west of the CRM bar region. General agreement is encouraging, but it is not unexpected that over such a large spatial range some differences would be observed between the ship observations and the buoy platform. Unfortunately, due to the experimental objectives of RIVET II the R/V Point Sur ship track was not optimal for collecting air-sea flux data, as can be seen from the relatively patchy data coverage in Figure 2.16. The data points in this map represent 10-minute intervals of data that passed all quality control steps in the flux processing. The largest sources of data loss were from large ship heading deviations within an interval, as well as unfavorable wind directions relative to the ship heading. The main role of the R/V Point Sur during RIVET II was drifter chasing, as well as up-river surveys of river flow. This is an issue in the analysis and interpretation of the results because the air-sea flux data span a variety of environmental conditions (Figure 2.17), but with very few repeat passes in the same location.
Figure 2.16: Map of the R/V Point Sur data coverage used in the flux analysis. Data points represent 10 minute averaging intervals and only data that passed quality control measures in the flux processing. Color of the dots signify time (continuously from yellow to dark brown), start and stop times are provided. The labeled asterisks denote NDBC buoys (including waves) used to gather environmental data during the experimental time frame. The magenta square marks the origin of the local coordinate system, with $X$ increasing eastward and $Y$ increasing northward.
Figure 2.17: From top-to-bottom: Wind velocity components observed from NDBC 46029 (3-m discus meteorological buoy) and the R/V Point Sur; $H_s$ observed from two wave platforms, the 3-m discus buoy and a directional waverider moored near Clatsop spit (NDBC 46243); average wave period from these two buoys; and the corresponding peak wave direction; salinity as observed from an in-water station (NDBC Station JTAW1) near the USCG at Cape Disappointment at Jetty A (the more northern jetty). The vertical magenta dashed lines in the lower 4 panels mark times of 10-minute data processed from the R/V Point Sur observations.
The Aerodynamic Drag Coefficient

**Figure 2.18:** (Upper) $R$ value as a function of $U_{10}$ for all of the observations during the experiment and for only data west (negative in the local coordinate system) of CRM. (Lower) PDF’s of $R$ for three different filtering schemes. Each bar is centered on its averaging interval and spaced every 1 $R$ (the first bar is at 0 and only includes 0 to 0.5 values).

The data from CRM was processed in a very similar way as the RIVET I data set and so the focus of this analysis will be to compare these results to that earlier work. As a first look, the focus will be on the general distribution of the aerodynamic drag coefficient, $C_D$. Similar to the NRI work, the drag coefficient will be compared to a
bulk, open ocean equivalent drag calculated over the same averaging interval used by the eddy covariance method. The ratio of the eddy covariance drag to the bulk drag is again defined as $R$. The bulk formula used for the CRM data analysis was the COARE 3.5 algorithm [Edson et al., 2013], which is a widely used algorithm for calculating flux parameters from bulk observed wind speed, air temperature, and relative humidity. This algorithm can support wave-dependent parameterizations, but for this work a wind-speed only dependent calculation was used.

While the sampling coverage suffers from the issues noted above, it is still of interest to see a general picture of the variance in $R$ over the course of the experiment. It was observed that $R$ tended to converge to unity as wind speed increased beyond $\sim 5$ m/s, this observation held for both data west of CRM as well as inshore of the river mouth (Figure 2.18). Most of the largest $R$ values observed occurred during low winds. Similar to NRI, in a coastal environment wind direction relative to the shoreline would be expected to explain a significant portion of the observed variance. Because the coastline at CRM was roughly inline with the north-south axis, the $R$ data was filtered by the corresponding wind direction quadrant, i.e. westerly, easterly, southerly, and northerly winds (the along-shore quadrants include off-shore and on-shore oriented data). Given the limited sampling, there are very few observations made during westerly wind conditions, when the ship was west of the CRM (see Figure 2.19). However, in general, regardless of the observed wind direction the $R$ value seemed to converge to unity as wind speeds increased beyond 5 m/s. Southerly and easterly winds exhibited the largest variance, but these were also the most common wind directions observed when the ship was offshore of the river mouth.
Figure 2.19: Similar to Figure 2.18, but for $R$ filtered by wind directions (as well as only data west of CRM). The wind quadrants are defined as: westerly is wind coming from compass direction 240 to 330; easterly is the quadrant 60 to 150; northerly 330 to 60; and southerly as 150 to 240. The PDF’s in the lower panel are displayed and calculated similarly to Figure 2.18.

Some interesting variability was observed when viewing $R$ as a function of distance west of CRM (Figure 2.20). Within 10 km of CRM, there seems to be a consistent increase in $R$, which appears relatively insensitive to wind direction. Immediately west of 10 km, the observations show that for easterly winds $R$ is consistently less than unity, while for southerly winds $R$ is spread about 1. Further west out to continental shelf
edge and Astoria Canyon (see marker buoy in Figure 2.16), $R$ is consistently spread about 1, regardless of wind speed and the fact that winds happen to be coming from the east (presumably against the waves). Further variance in $R$ as a function of space can be explained if the wind direction filter is changed to be relative to the peak wave direction (Figure 2.21). For wind traveling obliquely to the peak waves, $R$ is generally greater than unity and insensitive to distance westward of CRM; while for winds traveling against the waves $R$ was observed to have a strong spatial signal. Near CRM, wind opposing the waves coincided generally with $R$ values greater than 1, while far offshore $R$ corresponding to this wind condition was more centered on 1. Interestingly, the vast majority of the significantly lower than 1 $R$ values were observed during conditions of perpendicular wind flow relative to the peak waves. In fact, during these conditions almost no $R$ values exceeded unity (see Figure 2.21). Perhaps somewhat counter-intuitively, the very low values of $R$ tended to be observed during peak wave periods less than 10 seconds (Figure 2.22). In fact, only considering wind perpendicular to the peak waves, a $R$ exhibited a linearly increasing trend with increasing wave period ($r^2 = 0.35$). This trend was less significant in the other wind direction regimes, suggesting a strong flow-relative-wave crest dependence in the air-sea momentum flux coupling with the peak waves.
Figure 2.20: Cross-shore spatial distribution of $R$ for different filtering schemes: westerly and easterly winds (top), northerly and southerly winds (middle), and for winds exceeding 5 m/s (bottom). In all panels, all of the offshore data (x’s) is provided as a reference.
Figure 2.21: Cross-shore spatial distribution for $R$ observations made west of CRM filtered by different wind directions relative to the dominant wave period reported by the NDBC waverider at Clatsop Spit. With, against, and perpendicular filters were determined as ±30 degrees about the 0, ±180, and ±90 degree axes, respectively. Oblique was taken as any left over data, which includes oblique “with” and “against”.
Figure 2.22: $R$ as a function of peak wave period reported from the Clatsop Spit directional waverider, only for the data west of CRM and winds exceeding 5 m/s. Data have been filtered by their wind direction relative to the peak wave direction.

Summary

The RIVET II air-sea coupling data was very limited, primarily due to the sampling strategy of the R/V Point Sur. However, some interesting observations can be made from this sparse data set. In general, the order of variance in the eddy covariance drag coefficient relative to the open ocean equivalent was comparable at CRM relative to the observations made at NRI. This is interesting to note due to the vastly different environmental conditions and physical scales of these two estuary systems. In general, regardless of location in the CRM region, the drag coefficient was observed to converge to unity as wind speed increased beyond 5 m/s, this convergence became more consistent beyond 10 m/s (though this coincided with fewer data points). Interestingly,
the CRM data did not exhibit as clear of a spatial dependence on distance from inlet mouth, as was observed at NRI. This was most likely due to complications from the presence of swell waves throughout the observational record. $R$ was found to depend fairly strongly on the wind direction relative to the peak wave direction, as opposed to the wind relative to the coastline. Furthermore, it was observed that significantly lower than unity $R$ values tended to be observed during wind flowing perpendicular to the waves. Within this subset of observations, a statistically significant positive trend was observed between $R$ and the peak wave period. All of these particular observations occurred within 20 km of the CRM and at locations where significantly heightened $R$ values were observed in other flow-relative-to-wave conditions. One potential explanation for this observation is that these short period, steeper waves moving perpendicular to the flow direction suppress short wave development in the wind direction. Thus the surface in the wind direction would be less rough, decreasing the observed drag relative to the expected drag coefficient given the wind speed alone. Unfortunately, given the limits of the data set this hypothesis cannot be confirmed, but it does suggest further investigations into how wave-wave interactions can influence the turbulent transfer of momentum across the air-sea interface. RIVET II potentially could have provided a very interesting contrast to NRI and specifically given the very dynamic wave field at the CRM, it could have been a very fruitful observational data set. Unavoidable data sampling strategies (for our purposes) prevented the full realization of this and so further work, in similar environments, would be needed to develop complete answers to the questions the CRM data set posed.
2.3.3 CLASI

The outcomes of the RIVET project motivated a further effort to understand the nature of the atmospheric boundary layer in the transition region between the land and sea surfaces. The observations made at NRI revealed a complexity in the wind stress field over a coastal inlet system that had not been explicitly investigated in the literature. However, these findings were made for a fairly specialized environment and during a relatively narrow set of conditions. Further effort is needed to understand the coastal processes driving the observed variability in the air-sea fluxes in the nearshore and their spatial and temporal scales. Of particular interest was developing a better understanding of how land-air-sea and orographic effects could translate to variance in the ocean surface roughness. While this has implications for the hydrodynamics and air-sea gas fluxes in the nearshore, these dynamics also coincide with challenging environments for land-air-sea coupled models. In addition, this region poses a difficult environment for satellite remote sensing data interpretation. Further observations are needed in a variety of conditions in order to provide validation for model simulations and these remotely sensed observations.

In order to address this need, the Coastal Land-Air-Sea Interaction (CLASI) project was initiated. The first field effort took place in and around Monterey Bay, California using a combination of in situ observations, numerical models, and satellite remote sensing. Some preliminary findings from CLASI will be presented here; analysis for this project is currently on-going.
Monterey Bay (herein MB) provides an excellent test site for achieving the aims of the CLASI project. This large bay is located on the West Coast of North America just south of the San Francisco Bay (Figure 2.23). MB is well-exposed to Pacific swell systems.
because of its wide, west-facing opening. Also, atmospheric systems pass through MB regularly, driving strong west-northwesterly winds in the northern half of the bay and setting up fairly strong local wind seas. This combination of large Pacific swell and local wind seas creates a rich wave climate and strong signal for the air-sea fluxes. Morphologically, the eastern shore of MB is almost a continuous planar, sandy beach running from Aptos in the northeast corner to the city of Monterey in the south. Both the north and southern coastlines, are oriented almost due west and characterized by rock reef, kelp forests, and steep bluffs on the shorelines. The Santa Cruz Mountains above the northern coastline and the mountainous southern Monterey Peninsula can create significant orographic effects on the in-coming air masses from the open Pacific Ocean.

The in situ data for CLASI were collected from a variety of platforms both over-the-water and land-based, however only observations made from the smaller research vessel will be presented here. This vessel was a ∼8 m rigid-hulled inflatable boat (RHIB), which was outfitted with a suite of atmospheric and oceanographic sensors (Figure 2.24). All sensor data was logged in real-time on-board using Campbell Scientific data loggers. The loggers were continuously synced using a network time protocol server equipped with a GPS link. This system was inspired by the design and instrumentation of the SPEC deployed in RIVET I.
Figure 2.24: The RHIB used for making some of the *in situ* observations for CLASI. Some of the instrumentation is highlighted in the image. The ADCP and CTD were mounted to the platform shown, but the instruments are not visible in the image. Also, in addition to the IRGASON, two additional 3D-sonic anemometers were mounted on the RHIB frame on separate staffs. The vessel was motion was recorded by a 6-axis IMU fastened to the ship hull in the stern of the boat, just under the aft-most person in the photograph.
Case Study 1: the Southern Surge

21:30 June 12, 2016
Figure 2.25: A composite overlay of a COSMO-SkyMED™ (large) synthetic aperture radar (SAR) and a panchromatic ImageSat (inset in green) acquisition. The pixel resolution of the inset is \( \sim 0.7 \) m and the SAR has resolution of several meters. The arrow in the upper indicates north. The time stamp is referenced to the inset, the SAR image was taken several hours earlier. While the SAR reveals the incident wave field into MB and the presence of slicks in the northern half of the bay, the optical image nicely captures atmospheric gravity waves propagating through some low-level clouds in the central portion of MB. The waves are coming from the southwest and most likely generated through interaction with the Monterey Peninsula. The transect line used for analysis in this case study is also given with yellow (red) being the start (stop). Both satellite images were downlinked and processed by CSTARS.

On June 12, a southerly atmospheric system surged up-and-over the Monterey Peninsula and into the southern half of MB. The system had been generated off the coast off of Pt. Conception (\( \sim 250 \) miles south) and was driven northward by a larger scale atmospheric system traveling eastward across the western US. This relatively rare low-level, southerly jet created an excellent opportunity to observe complex land-air-sea interactions driven by offshore flow in MB, which is characterized by diurnal on-shore winds. On this day, the goal of the RHIB was to do a westward transect in the southern corner of MB and capture the transition of the air-sea interface along the Monterey-Pacific Grove coastline (Figure 2.25). The transect began off of Del Monte Beach (yellow circle in Figure 2.25) around 21:40 UTC and ended an hour later just north of Pt. Piños on the eastern edge of MB (red circle in same figure). The total transect length was just over 6.5 km. At the beginning of the transect, the southerly jet had weakened and the winds had shifted to the WSW. A strong horizontal roughness front, or wind line, was visually observed (Figure 2.26) about halfway along the transect and the approximate location in time and space of this feature was determined using the time stamp of the note in the field log. This feature will be used in the following figures as a reference for the along-transect data.
Figure 2.26: A photograph taken from the RHIB during the westward transect used for this case study. Directly to the left is the Monterey coastal bluffs (Pacific Grove is in the background) and the point where the land meets the water in the background is Pt. Piños. The border between the light and dark blue water in the background is the wind line noted in the field log. The thick scum lines to the left near the coast are over a kelp forest.

Data were processed in a 1-minute time bins, which based on an over-ground vessel speed of 1 m/s corresponds to about a 100 m wide horizontal footprint. The mean horizontal wind vectors along this transect are given in Figure 2.27. Along the transect a noticeable shift in the wind direction and speed was observed. In the southeast corner of MB, the winds are light and the directions confused. At about the half-way mark, the winds increase and the direction is consistently WSW-W out to Pt. Piños at the transect terminus. This coincides with a transition in the land topography just west-southwest of the transect line. Also, this change in the mean wind vectors is co-located with the location of the wind line observed in the field log (Figure 2.26).
This transition in the horizontal wind field also coincided with a noticeable transition in the atmospheric boundary layer stability (middle panel Figure 2.28). East of the transition zone, the boundary layer is slightly unstable with water surface temperature about a degree higher than the air temperature as observed on the RHIB met-mast. Within a 1 km transition zone, the boundary layer flips stability and a nearly two degree difference in air-sea temperature was observed. This was consistent out to the terminus of the transect, but with both air-water showing a mean decrease in temperature. Near this transition zone, some significant horizontal gradients in water surface temperature were observed with steps of 1.5° over 500 m. Interestingly, the wind and stability transitions seem to correspond to changes in the observed normalized acoustic backscatter (ABS) near the water surface. The layer of high ABS seems to thicken and the response very close to the surface also seems to increase in magnitude. This is most pronounced
within 2 km of the transition zone. The normalized ABS was estimated following the algorithm of Wall et al. [2006] and derives directly from the acoustic echo amplitude observed from the ADCP mounted on the RHIB.

**Figure 2.28:** Mean wind speed and direction (top), air and water surface temperature (middle), and normalized ABS. The air and water temperature come from the shielded probe on the tall met-mast and a CTD mounted 0.3 m below the surface, respectively. The vertical dashed line marks the estimate of the wind line derived from the photographs and notes in the field log, this is the case in all of the following figures. The strong near-surface signal in the ABS profiles, at about 2.25 km, is due to the RHIB crossing the bow wake of a whale sight-seeing ship going back to Monterey harbor—this was noted in the field logs.
Figure 2.29: $R_{CD}$ (top) and Off-wind stress angle (bottom) along the transect. The color scale denotes $U_{10}$ in m/s.

The transition noted in the mean wind and air-sea temperature difference corresponds to marked changes in the wind stress field (Figure 2.29). For $R_{CD}$ (ratio of eddy covariance $C_D$ to bulk algorithm $C_D$) a dramatic collapse to 1 was observed within a 500 m region of the transition zone. East of the transition zone, higher and more variance is observed in the directly observed drag coefficient with respect to the open-ocean equivalent. West of the transition, the drag converges to the open ocean value and the variance drastically decreases. Curiously, 5 km along the transect there is consistent increase in the drag, which corresponds to $\sim 2$ m/s drop in the mean wind speed (upper panel Figure 2.29). This signal was only observed for about 1 km footprint, centered on the 5 km mark. This could be due to some topographic feature upwind, but it is not easy to discern the source of the deviation given the data and field notes. The consistency of this signal and the rest of the data lends some credence to this being a real, physical response in the wind stresses.

The off-wind stress angle also shows some consistent trends along this transect line.
The stress angle was calculated using both components of the Reynolds stress over the 1-minute average and then a 300 m bin-average was applied to reveal the overall trend in the observations. East of the transition zone, a consistent shift in the wind stress angle from $+45^\circ$ to $-45^\circ$ was observed across the entire 3 km stretch (lower panel Figure 2.29). The sign indicates a shift from the left-to-the-right of the mean azimuthal wind direction. West of the transition line, the stress angle exhibits larger variance, but with a mean centered on zero. At the 5 km west mark, unlike with $R_{CD}$, no consistent signal in the stress angle response was observed.

**Figure 2.30:** Omni-directional wave spectra from one of the acoustic wave gauges mounted on the RHIB. The spectra are ensemble averages of three consecutive 1-minute spectra and the 95% CI of this ensemble mean (for each frequency band, df) is given as the shaded region. The location along the transect in km from the beach for each spectra is also given. Note: the wind line corresponds to $x = 3.1030$.

The RHIB was also outfitted with 3 acoustic wave gauges, which, in a downward-looking orientation, use the time-of-flight of an acoustic pulse to estimate the distance from the transducer to the water surface.
Figure 2.31: Directional wave spectra from NDBC #46042, which is located N 36.79, W 122.45 (this is ~27 nautical miles west of Moss Landing). Frequency for each radial is given in Hz and the azimuthal angles are in a meteorological (aka nautical) convention. The energies have been normalized by the peak of the omni-directional spectra.

From these gauges, the water surface elevation was observed and the corresponding omni-directional wave spectra were estimated (Figure 2.30). The dominant peak was consistently observed to be between 0.066 and 0.1 Hz (15-10 seconds) along the entire transect. However, the waves were largest and most broad-banded near the western terminus of the transect. Only energy up to 0.3 Hz is shown here because these spectra have not been corrected for the Doppler effect due to the vessel translation relative to the wave propagation.

Wave direction spectra from a nearby NDBC buoy (#46042) show that for the entire time period of this transect, roughly 21:30 to 22:30, the dominant incident waves into MB were broadly from WNW to WSW between 10 to 8 seconds (Figure 2.31). The buoy also observed a southerly swell at ~17 seconds. For the most part, these buoy observations were not reflected in the RHIB wave observations, except for the furthest westward portion of the transect (see Figure 2.30).
Figure 2.32: Current profiles, speed (top) and direction (bottom), observed from the ship-mounted ADCP. Profiles have been smoothed using a 30 second wide running average.

The 1-minute current profiles observed from the ADCP also reveal a sharp transition coinciding with the observed wind line and shift in the mean atmospheric variables (Figure 2.32). East of the transition, current were observed to be <0.15 m/s and fairly depth-uniform in magnitude. A seemingly wind-driven surface layer was observed by examining the current direction profiles. Directly west of the transition zone, a water mass moving 3 to 4 times faster than the water mass east of the transition was observed. This water body is also moving consistently southwestward. The ADCP profiles revealed the cross-section of this water mass as a wedge with distinct undulations along the sloping face of the front, suggesting a transition to baroclinicity west of the wind-line. The water levels from the NDBC tidal station #9413450 in Monterey Harbor, indicate the low-low tide of $-0.56$ m occurred at about 18:30 UTC, which would be nearly 3 hours before the beginning of this transect.
Summary

The observations from Figures 2.27 to 2.32 revealed strong, horizontal spatial gradients in the wind, wave, and current fields along a single transect running parallel to the Monterey-Pacific Grove coastline. Of particular note, is the apparent coupling between the variability in the wind velocity and stress fields with the underlying water mass. This also coincided with the spatial gradients in the topography just upwind of the transect line. Using a back-of-the-envelope estimate based on field log notes and pictures taken during the transect, these quantitative measures match up well in space with an observed horizontal transition in the apparent surface roughness (Figure 2.26).

As of yet, the precise nature of the land-air-sea coupling has not been described given the challenges in untangling all of the sources of variance observed. For example, it is difficult, with a single transect, to untangle the separate contributions tidal currents and orographic wind effects have on the observed spatial wind stress variability. Also, the spatial variance in the wave field does not match the wind and currents. Since the short waves are the most effected by the Doppler shift and these waves are the most responsive to short-fetch winds, this correction needs to be done before definitive conclusions can be made about the wave field. This may be contributing to some of the differences observed between the RHIB wave data and the NDBC buoy at the mouth of MB. However, these differences may also be a result of wave shoaling over the Monterey Canyon, which splits MB along its E-W axis. In order to confirm that this is the result of some physical process and not instrument error, a comparison will have to be made when the RHIB is closer to the NDBC buoy, thus removing the potential canyon effects. While this analysis is incomplete and is currently on-going, the dynamics captured in this roughly 60 minute segment highlight the objective of the CLASI project
and suggests that further investigations would be fruitful. Future efforts will focus on more detailed analysis of this and other case studies as well as direct comparison to models and satellite remote sensing observations.

### 2.4 Conclusions

![Aerial photo taken at the New River Inlet, NC during the RIVET I experiment. Courtesy of G. Farquharson and APL-UW.](image)

**Figure 2.33:** Aerial photo taken at the New River Inlet, NC during the RIVET I experiment. Courtesy of G. Farquharson and APL-UW.

The hypothesis posed at the beginning of this chapter was: **Coastal air-sea fluxes are generalizable by bulk, open ocean parameterizations.** From analyzing several field observational data sets it seems clear that the air-sea fluxes in the coastal zone are not generalizable by parameterizations developed from open ocean measurements. This failure of open ocean parameterizations can be linked to sources of variance in the
observed coastal air-sea fluxes that are unique to the coastal environment. These sources include a depth-limited domain, wave shoaling and depth-limited breaking, horizontal heterogeneity of the surface, and strong current gradients. While it is possible to witness some of these processes (i.e., the latter two) the coastal environment tends to amplify those signals beyond what is typically observed over the open ocean. This study has demonstrated that the relevant spatial and temporal scales of variability are strongly a function of the particular coastal domain as well as the local environmental conditions. The local wave climate has been shown to play a critical role in the observed variability as coastal environments with and without strong swell conditions exhibit different scales of air-sea flux variability. The focus of this study has been on the air-sea momentum flux because this particular parameter to the dynamics the air-sea coupled system. The momentum flux drives currents, is critical to wave development, and facilitates material transport. One of the most fundamental observations made throughout these analyzed data sets was that the assumption that the wind vector and wind stress vector are inline regularly breaks down in coastal environments. This has significant implications for observations and modeling studies, which rely on wind velocity as a proxy for wind stress—the actual force applied from the atmosphere to the ocean. Other air-sea fluxes, e.g. heat and mass, may be important to consider in the coastal zone for a variety of processes; and while this study does not consider these processes in the analysis, the findings may suggest similar divergence in the other flux parameters from the expected open ocean values.
Chapter 3

Wind-Current Coupling in a Coastal Model of a Tidal Inlet System

Re-Statement of Hypothesis

The circulation in an operational, process-based model is insensitive to the surface wind forcing. The research questions under investigation for this hypothesis are: What is the role of the wind forcing on the hydrodynamics at a location like New River Inlet? Does the surface wind forcing contribute significantly to the hydrodynamic momentum balance? With the advent of tides? Or waves? Is there output sensitivity to changing the applied wind forcing parameterization? A series of numerical experiments were conducted in order to test this hypothesis and address these questions in an operational coastal hydrodynamic model. In answering these questions, the generalization of the field observations presented in Chapter 2, that the surface wind forcing exhibits a high degree of spatial and temporal variability in the coastal zone, will be assessed.
3.1 Background

3.1.1 A Conventional Definition of the Atmospheric Forcing

The atmosphere directly forces the surface of the ocean through the wind shear stress, $\tau$. Typically, $\tau$ is represented using a straightforward drag relationship:

$$\tau = \rho_{air} C_D U_z^2,$$  \hspace{1cm} (3.1)

where $\rho_{air}$ is the air density, $U_z$ is the mean azimuthal wind speed observed at some height $z$ above the surface, and $C_D$ is the aerodynamic drag coefficient, which summarizes the entirety of the turbulent interactions between the atmosphere and surface. The vector nature of the stress is generally ignored because it is assumed to always be in-line with the mean azimuthal wind direction. For a typical atmosphere-circulation coupled model, the wind shear stress (in the form of Equation 3.1) is applied as a source (or sink) of momentum and prescribes the dynamic surface boundary condition of the water column. If waves are introduced, this term acts as a source/sink of wave energy in the wave action balance equations.

The relevant horizontal momentum equation can be represented following Lesser et al. [2004],

$$\frac{\partial U}{\partial t} + U \frac{\partial U}{\partial x} + V \frac{\partial U}{\partial y} + w \frac{\partial U}{\partial \sigma} - fV = \frac{1}{\rho_0} P_x + \nu_H \nabla^2 U + D_u + M_x \quad \text{(3.2)}$$

$$\frac{\partial V}{\partial t} + U \frac{\partial V}{\partial x} + V \frac{\partial V}{\partial y} + w \frac{\partial V}{\partial \sigma} + fU = \frac{1}{\rho_0} P_y + \nu_H \nabla^2 V + D_y + M_y, \quad \text{(3.3)}$$

where $x$ ($U$) and $y$ ($V$) are the local coordinate (velocity) system. The fourth term on the LHS are the vertical advection term with $w$ as the vertical velocity in the $\sigma$ coordinate system—this will be explained below. The last term on the LHS is the Coriolis term. For the RHS from left-to-right, the hydrostatic pressure gradient where
$P_{xy}$ are the horizontal, Boussinesq pressures, the horizontal viscosity term, vertical diffusion, and $M_{xy}$ representing external sources or sinks. The wind stress would fall into this term as $\tau/\rho_0$ [Wargula et al., 2014] with $\tau$ being defined as in equation 3.1. If considering waves, the horizontal velocities $U$ and $V$ are defined as the Generalized Lagrangian Mean [Lesser et al., 2004],

$$U = u + u_{st}$$

$$V = v + v_{st},$$

where $u$ and $u_{st}$ are the Eulerian and Stokes’ drift velocity components, respectively (the same holds for $V$). Also, if considering waves, in $M_{xy}$ there would be included radiation stress gradients, which are functions of the phase-averaged wave field [e.g. Longuet-Higgins and Stewart, 1964]. Therefore, in the momentum balance the wind stress is included both explicitly, as a mechanical source of momentum, and implicitly, through the Stokes’ drift and radiation stresses as a source of wave action.

### 3.1.2 Literature Review

The focus of this review will be on the role the atmospheric forcing plays in the coastal ocean. For a detailed review of the wind shear stress itself, the reader is directed to Chapter 2 and to Appendix A.

The coastal ocean is the margin of the global ocean spanning from the continental shelf to the shoreline. From a hydrodynamics perspective, this marginal region can be separated into specific zones characterized by the relative importance of different dynamical forcing mechanisms. Understanding how these zones differ and their relevant spatial-temporal scales of variability, is critical to characterizing the general circulation
is this complex region. The outer continental shelf\(^5\) marks the transition from the generally approximated “infinite” water column of the deep ocean to a water column where surface forcing mechanism (i.e. wind, heat flux) influence a greater proportion of the entire depth [Csanady, 1982]. The inner shelf is the zone where surface and bottom boundary layers directly interact with each other [Fewings et al., 2008]. The surfzone is the portion of the coastal ocean characterized by depth-limited wave breaking and extends all the way to the beach face [Hally-Rosendahl et al., 2014]. The surfzone itself can be divided into two regions: an outer surfzone, where waves are actively breaking, and an inner surfzone, where the broken waves propagate towards the shore as bores. The horizontal scales of these zones within the coastal ocean are constrained by the local bathymetry, but can be highly dependent on local wind and wave conditions. Generally, the hydrodynamics in the surfzone are dominated by surface gravity wave forcing, while over the shelf the flows are largely wind-driven (with contributions from tides, buoyancy forces, and non-breaking waves [Lentz and Fewings, 2012]). The border between the inner shelf and the surfzone represents a stark dynamical boundary and it remains a significant challenge to separate the independent contributions from along- and cross-shelf wind stress and surface gravity wave forcing in the nearshore\(^6\) [Lentz and Fewings, 2012].

Using a multi-year time series, Fewings et al. [2008] isolated the roles along- and cross-shelf wind stress play in driving cross-shelf circulation over the inner shelf. Over the middle and outer continental shelf, exchange across the shelf is driven by coastal upwelling and downwelling events, which are caused by along-shelf wind stress [Lentz

\(^5\)In the discussion of this topic, a conventional continental shelf coordinate system will be adopted: the cross-shelf direction is in-line with the shore-normal direction and along-shelf is oriented perpendicular to this in the horizontal plane (the third dimension is always vertical). Different investigators use different conventions for the sign, but for the purposes of a general discussion this is not critical. When necessary, the orientation of the coordinate system will be explicitly noted.

\(^6\)Herein, we consider the nearshore as anything landward of the outer-inner shelf boundary.
and Fewings, 2012; Ekman, 1905]. However, in water depths O(10 m) along-shelf wind forcing does not drive cross-shelf exchange, but rather the relative importance of the Coriolis force diminishes and cross-shelf transport is driven by cross-shelf wind stress [Fewings et al., 2008]. This difference between the inner and mid-outer shelves is important because it effects how material is exchanged between the inner shelf and the surfzone. The conclusions of the Fewings et al. [2008] study could only be reached because of the length of the time series dataset used in the analysis. While this study contained excellent temporal resolution in order to parse the effects of along- and cross-shelf wind stress and wave forcing, it lacked any substantial spatial resolution (being from the Martha’s Vineyard Coastal Observatory [http://www.whoi.edu/mvco]).

Hally-Rosendahl et al. [2014] focused on the exchange of material between the surfzone and inner shelf using rhodamine dye and a suite of stationary and mobile sampling platforms. The authors highlight one particular case study dominated by swell wave forcing to demonstrate that cross-shelf exchange from the surfzone to the inner shelf was driven by transient rip current ejection events. These events occurred periodically in space along a wide, planar sandy beach in Southern California during a day of relatively low winds. Using a variety of spatially-distributed observation stations, Hally-Rosendahl et al. [2014] captured the evolution of the surf zone-inner shelf coupled system. However, drawing general conclusions from this case study remains challenging given its limited duration (~12 hours of observations). Especially given that the Hally-Rosendahl et al. [2014] do not include in their analysis the role of the wind because it was not directly observed at the study site during the observation period (though the authors do reference an unnamed inland station as reporting <5 m/s winds on that day). In the context of this dissertation, this motivates the question: would these observations hold in stronger winds?
Both Fewings et al. [2008] and Hally-Rosendahl et al. [2014] conducted studies at sites with gently sloping, planar shorelines. In other coastal domains the dynamics are dramatically different and this changes the relative importance of the wind stress in the momentum balance equations. In the general sense, the effect of wind forcing on the tidal exchange at an estuarine or river mouth is partitioned into remote and local forcing [Wong and Valle-Levinson, 2002]. The former being the result of winds over the continental shelf inducing water level fluctuations (set-up or set-down) at the shoreline [Wong and Valle-Levinson, 2002; Lentz and Fewings, 2012]. This cross-shore pressure gradient can drive uni-directional flow governed by the direction of the sea level, outflow for set-down and inflow for set-up [Wong, 1994]. The local wind effects, driven by winds at the estuary or river mouth, may contribute significantly to the subtidal flows [Wong and Valle-Levinson, 2002], but on different spatial-temporal scales [Wong and Moses-Hall, 1998]. These locally wind-driven flows are more three-dimensional in nature and this variability may be lost by large scale spatial-temporal averages [Wong and Moses-Hall, 1998]. The influence of local wind forcing is also highly dependent on the topographic structure of the estuary [Geyer, 1997; Wong, 1994] and degree of stratification in the water column, which may vary seasonally [Wong and Valle-Levinson, 2002].

At the New River Inlet (the same site used in this and the previous chapters), War- gula et al. [2014] used an array of observation stations to analyze the spatial-temporal variability of the various terms in the momentum balance. They compared observations from the main tidal channel and on the ebb shoal directly off-shore of the inlet and found that waves enhance the along-channel momentum flux in the inlet and over

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7The cross-shore direction is in-line with the cross-shelf direction. "Shelf" is replaced by "shore" landward of the inner shelf, though the exact border may depend on which group of investigators is being considered.
shoals. However, they also observed discrepancies of up to 50% between source/sink terms in the momentum balance when the wind forcing term was not included (see Figures 8b and 9b of that paper). Further analysis of this data set suggests that the importance of the wind term has spatial variability and is dependent on the wind direction relative to the tidal channel (A. Wargula personal communication, 2015). The Wargula et al. [2014] study focused on the mouth of a tidal inlet, which is where wind effects, remote or local, would not be expected to play a significant role [Wong and Moses-Hall, 1998]. Somewhat similar observations were made by Muscarella et al. [2011] at the mouth of the Delaware Bay, a much larger estuary-ocean system than New River Inlet. Using high-resolution surface wind model output and HF radar-derived surface current maps, they attempted to track wind-current coupling over an eight month period. There results show a strong linear relationship between the surface currents and local wind stress response; however, wind events were not persistent enough to set up the classical Ekman flow expected over the continental shelf [Ekman, 1905; Lentz and Fewings, 2012].

3.1.3 Study Objective

This brief literature review has highlighted that some significant gaps remain in the general understanding of the role atmospheric surface forcing plays in the nearshore hydrodynamics. Especially, in relation to the more visibly energetic forcing mechanisms of tides and surface gravity waves. Many previous observational studies were constrained either in space [Fewings et al., 2008] or in time [Hally-Rosendahl et al., 2014] and so drawing general conclusions from these studies is challenging. Field studies that surmounted both of these constraints [Wargula et al., 2014; Muscarella et al., 2011] become hampered by their ability to precisely separate the different forcing mechanisms.
This present work attempts to tackle all of these constraints by using hindcasts from an operational coastal model, which incorporates \textit{in situ} observations, in order to assess the effect of the surface wind forcing on the local circulation of a nearshore region, namely the New River Inlet in North Carolina. The over-arching goal of addressing this problem and testing the hypothesis under consideration in this study is to motivate the development of physically realistic air-sea coupling in coastal models. While numerical methods have their own constraints and limitations, the advantages in spatial and temporal resolution are critical to properly characterizing the independent contributions of tidal, wind, and wave forcing on the momentum balance in the nearshore. Models also have the distinct advantage of allowing for complete control of the physical drivers of the system.

\section*{3.2 Methods}

For a more details about the study site New River Inlet (herein, NRI) in North Carolina, the reader is directed to Chapter 2 and Wargula et al. [2014] and MacMahan et al. [2014]. Below will focus on the design of the numerical experiments and the methods used for answering the posed hypothesis.

The model used to test the posed hypothesis uses the Delft 3D package developed by WL—Delft Hydraulics and Technical University of Delft [Lesser et al., 2004]. The Delft 3D package is comprised of a variety of components, such as circulation, waves (using SWAN, [Booij et al., 1996]), sediment transport, water quality, etc. and is an open-sourced, operational platform used by a wide community of individual researchers and official organizations. This makes it a good candidate for this type of test because the model has been subject to rigorous numerical evaluation and, if properly set-up, can generate plausible output. In addition, this particular model of the NRI used here was
initialized and calibrated to local conditions using \textit{in situ} observations as part of RIVET I [Rynne, 2016]. Therefore, some initial investment has been spent on assessing the skill of this model at this particular inlet and this will be discussed below.

3.2.1 Experiment Scheme

A series of numerical experiments was designed in order to test the posed hypothesis. These experiments were designed to isolate the effects of wind, tidal, and wave forcing on the simulated NRI system. Table 3.1 provides a layout of all of the simulation runs used in the analysis. The aim was to run the model and systematically turn on/off different forcing terms in the momentum balance. For example, a simulation was run with tidal forcing being the only external driver of momentum. This can be thought of as the O(1) simulation a modeler would do to get a sense of the hydrodynamics at NRI.

Four different wind fields were used for these model experiments. \( U(t) \) is a spatially uniform, time varying wind field imposed on the model grid. The difference between 0 and 1 wind fields: 0 runs are done with the default wind stress parameterization used in Delft 3D [Smith and Banke, 1975] and 1 runs have a universally doubled wind stress value (i.e. the default value, for a given wind speed, \( \times 2 \)). A spatially varying wind field in both default and this ”enhanced” mode were also run. Therefore, the axes of Table 3.1 can be thought of as the balance between the complexity of the tidal inlet versus the reality of the imposed wind field. Therefore, the bottom-right corner of Table 3.1 is a simulation run with the most\(^8\) complex tidal inlet and the most realistic wind field. A description of the wind stress parameterization used in Delft 3D is given further down. In total, all of the simulations took a little more than 2,000 CPU hours.

In order to confirm or reject the hypothesis under evaluation, the analysis will have to show that there is a statistically significant change in the direction and/or magnitude

\(^8\)“Most” relative to the other numerical experiments done in this study.
Table 3.1: Numerical experiment plan.

<table>
<thead>
<tr>
<th>Processes</th>
<th>( \tau_0 )</th>
<th>( \tau_1 )</th>
<th>( \tau_2 )</th>
</tr>
</thead>
<tbody>
<tr>
<td>Wind N, Tides Y, Waves N</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Wind Y, Tides N, Waves N</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Wind Y, Tides Y, Waves N</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Wind Y, Tides Y, Waves Local</td>
<td>X</td>
<td>X</td>
<td>X</td>
</tr>
<tr>
<td>Wind Y, Tides Y, Waves All</td>
<td>X</td>
<td>O</td>
<td>O</td>
</tr>
</tbody>
</table>

X/O refers to that process being activated or not during the simulation. Waves Local refers to wave-coupled runs where the only waves generated were within the domain by the local wind. Waves All refers to runs with local generation and incident waves from the boundary. The columns denote different wind stress modes: 0 is nominal (default Delft 3D wind stress), 1 refers to global enhancement of wind stress forcing, and 2 refers to a spatially varying wind stress. More details about this are given below.

of the circulation in and around NRI that can be directly linked to the wind stress applied to the water surface. And this effect persists with the advent of tidal and wave forcing. This would demonstrate the wind forcing is critical to accurately representing the hydrodynamics at the NRI and therefore, the hypothesis statement can be rejected. If no statistically significant change in the NRI circulation can be attributed to the wind, while tides and/or waves are turned on in the model, then the hypothesis cannot be rejected with any significant confidence. This outcome would either suggest that the wind stress plays a minor or negligible role in the momentum balance. Or that the numerical implementation of the atmospheric surface forcing is misrepresenting the physical air-water interaction.

Based on field observations made at NRI and other studies at similar locations, it is possible to develop an *a priori* sense of where the wind stress would play a significant role and where it is expected to be of negligible importance. Over the shelf, the wind forcing is expected to be an important, if not the dominant forcing on the system
(Region I of Figure 3.1). Over the shoals and in the nearshore (surfzone plus inner shelf), there is more uncertainty in the degree to which atmospheric forcing is a driver of the flow (Region II). In the tidal channel and the Intracoastal Waterway (ICW), Region III, the wind stress would be expected to play a minor, if not negligible, role in the hydrodynamics. The shallow estuary represented by Region IV in Figure 3.1 would be expected to be heavily influenced by local wind effects (see Geyer [1997] for an example of wind-driven flows in a shallow estuary).

![Figure 3.1: The major dynamic regions within the NRI numerical domain. The highlighted portions are meant as qualitative representations.](image)

**3.2.2 Model Set-Up**

Delft 3D was set-up to run in three-dimensions with online (direct) coupling between the FLOW module (currents) and SWAN (waves). The wind field fed into Delft 3D was applied to both the current and wave simulations. The main focus of this study is on the circulation and so in order to save computational effort, the simulations were run without sediment transport or heat fluxes.
Domain

Figure 3.2: An overview of the Delft 3D model domains used for this study. The outer domain is enclosed in the dashed red box and the nested domain is shown in cyan. Depth contours are given for the outer domain at 5 m intervals and corresponding to the colorbar with units given in m’s.

The model domain encompasses the New River Inlet and Estuary in North Carolina, which is south of Cape Hatteras, and the surrounding coastal region (Figure 3.2). Delft 3D employs a one-way nesting scheme and when using wave-current coupling a larger, wave-dedicated domain (red-dashed box in Figure 3.2) is necessary to propagate wave energy down to the boundaries of the nested grid [Rynne, 2016]. This larger grid extends offshore 60 km and 60 km in either along-shore direction with a variable reso-
olution of O(1 km) at the offshore extent and O(100 m) in the surfzone and inlet mouth. The fine scale nested, curvilinear grid (cyan in Figure 3.2) is used for both the hydrodynamic and wave models. This grid includes the upper New River Estuary at its inshore limit to ~9 km offshore of NRI and extends 9 km in both along-shore directions. This nested grid also includes a portion of the ICW, which generally runs parallel to the coast connecting NRI to other tidal inlets in the region.

The nested grid resolution varies between O(10 m) in the center of the inlet mouth up to O(100 m) at the offshore boundary. The bathymetry data used in both wave-only and current-wave coupled nested grid was derived from several datasets. The 30 m resolution GEODASS global relief nearshore survey was used for the lower estuary out to the mouth of NRI; the upper estuary, beaches, inlet channels, and ICW were survey using a combination of on-foot, jetski, and LARC amphibious vehicle surveys [Rynne, 2016]. Surveys were conducted using survey grade GPS systems and acoustic altimeters. The bathymetry of the nested grid is given in Figure 3.4.

Delft 3D was set-up as fully three-dimensional using a $\sigma$ coordinate system (contour following). The relationship between this system and a cartesian grid is,

$$\sigma = \frac{z - \zeta}{H},$$

\[(3.6)\]
where \( z \) is the cartesian vertical position, \( \zeta \) is the free surface elevation above the reference plane \( (z = 0) \), and \( H \) is the total water depth (Delft 3D FLOW User’s Manual). The advantage of the \( \sigma \) grid is that the flow mimics the contours of the bed, but with the cost of not being able to specify a fixed vertical resolution. For all of the simulations, nine \( \sigma \) layers were used with a non-uniform distribution in the vertical in order to have finer resolution at the surface and bed (see Figure 3.3).

Figure 3.4: Bathymetric map used in the nested hydrodynamic-wave coupled grid. These data were compiled from available data sets and \textit{in situ} surveys done as part of the RIVET I field campaign. The color scale is given in units of m’s. Also labeled, the boundary condition types for the hydrodynamic grid.

**Boundary Conditions**

The wave-only grid was driven by wave observations made from a directional waverider buoy \( \sim 6 \) km directly offshore of NRI (NDBC 41109, not shown). These observations were used as input into SWAN assuming that wave conditions are fairly homogeneous.
over the shelf [Rynne, 2016]. The sea state developed on the large wave-only grid is used to force the wave conditions on the boundary of the nested grid. The hydrodynamic boundaries are forced by water level on the seaward edge (see Figure 3.4) and Neumann conditions for the cross-shore boundaries both for the ocean-side and ICW (following Lesser et al. [2004]). The offshore water level was determined from the waverider buoy assuming along-shelf homogeneity in water level. The inland boundary is forced by river discharge from the USGS station in Jacksonville, North Carolina. At the bed, a free slip boundary condition was used and a uniform Manning coefficient of 0.028 was applied [Rynne, 2016]. For the waves, depth-limited breaking was prescribed following Battjes and Janssen [1978] with an \( \alpha \) of 1 and \( \gamma \) of 0.78. Zippel and Thomson [2015] used observations from NRI to show that depth-limited breaking models that do not explicitly incorporate currents do not fully capture wave breaking at the inlet mouth due to strong wave-current interactions. These new observations were not incorporated into this study because Delft 3D is yet unable to apply such a scheme.

Bottom friction for waves was applied using a JONSWAP scheme with a friction coefficient of 0.067 m\(^2\)s\(^{-3}\). For the turbulence closure, the Horizontal Large Eddy Simulation (HLES) approximation was employed with a Prandlt-Schmidt number of 0.7 (suitable if not considering suspended sediment transport: Delft 3D FLOW User’s Manual) and a relaxation time of 30 minutes. The data used for initializing the model and prescribing the boundary conditions are given in Figure 3.5.

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9This station is not shown in Figure 3.4, but it would be at most western extent of the nested domain almost exactly at the terminus of the "Discharge" pointer.
Figure 3.5: Input data used to define boundary conditions. From top-to-bottom: significant wave height, $H_s$, average ($A$) and peak ($P$) wave period, mean wave direction (MWD) relative to the shore-normal (positive: from left-of-normal), 10 m wind speed in cross- and along-shore components $U$ and $V$ respectively, and water elevation, $\eta$, at the offshore boundary. The space between the vertical red-dashed lines represents the period simulated in all model runs used for this study.

Wind Stress Parameterization

The wind stress on the water surface in the hydrodynamic and SWAN models is parameterized using the aerodynamic drag formulation (Equation 3.1). It is applied assuming that the wind stress vector, $\tau$, is always parallel to the wind velocity vector, $U$. For the Delft 3D package, the wind stress parameterization scheme is a simple piece-wise defined drag coefficient about three "break-points" delimited by wind speed, which is assumed to be the neutral 10 m equivalent. In this fashion, one can input the wind drag
coefficient at three wind speeds and the model will create a discontinuous function of the wind stress at each time step and grid cell. This is unlike the bottom friction formulations for the hydrodynamic and wave models, which provide a variety of option from the literature with pre-determined behavior and tuning parameters made available to the user. Following Smith and Banke [1975], Delft 3D prescribes a default wind stress parameterization with a constant wind drag of 0.0011 from 0 to 10 m/s and then increasing monotonically to 0.00723 at 100 m/s, from there it holds constant to infinity. The max wind speed observed at NRI during the experiment and applied in the model was <15 m/s and so the behavior of the parameterization at these upper extremes is inconsequential to the study.

A comparison between this default scheme and three conventional drag parameterizations is given in Figure 3.6. The COARE 3.5 curve is derived from the eponymous algorithm presented as an update to COARE 3.0 in Edson et al. [2013]. The wind speed-drag relationship used by COARE 3.5 is derived from reanalyzing various field observation data sets. At both the lower wind regime and with increasing wind speed there is significant divergence between COARE and the wind stress scheme employed for this study. However, it was important for this study to use the default set-up in Delft 3D for the wind drag coefficient. The perspective of the posed hypothesis and this study as a whole is of a modeler using Delft 3D to hindcast the flows in a coastal area, such as NRI. This study is explicitly designed to not be a study by an atmospheric scientist or an investigator expressly interested in air-sea interaction. Delft 3D is widely used in nearshore studies by users who do not change the default wind stress parameterization.
Numerically, the hydrodynamic model uses the input wind field and the wind-drag relationship defined by the piece-wise function (and the density of air) to calculate the wind shear stress using Equation 3.1. This stress is used as an external force in the momentum balance and as the dynamic surface boundary condition for the water column. In the wave model, the wind stress is a source of wave action balance using an input term with both resonant interaction [Phillips, 2017] and feedback [Miles, 1957]. There are two complicated parameters $A$ and $B$ (see Delft 3D WAVE User’s Manual and/or [Booij et al., 1996]) that are functions of the sea state, wind speed, and wind direction relative to the water surface. However, fundamentally the process is the same as the hydrodynamic module where a wind-drag relationship is used to turn wind speed input by the user into wind shear stress through Equation 3.1. In SWAN, the wind shear velocity, $u_*$ is used instead of $\tau$, but the two can be related directly to the wind drag,

$$\tau = \rho u_*^2 = \rho C_D U_{10}^2.$$  \hfill (3.7)

### 3.3 Results

A fully wave-current coupled, three-dimensional model was used to hindcast the conditions at the NRI from May 1 to 20, 2012. The model was run following an experiment scheme designed to assess the role of the atmospheric surface forcing on the
three-dimensional circulation in the nearshore region of the domain (Table 3.1). This particular model set-up has been used in other studies as part of the RIVET I experiment and the ability of the model to accurately represent the dynamics was assessed by Rynne [2016]. These model runs differ from the present study in that they were run in a depth-averaged, or two-dimensional, mode. Model skill was evaluated using the relative average error between modeled and observed quantities [Willmott et al., 1985],

$$
\chi = 1 - \frac{\langle |x_m - x_o|^2 \rangle}{\langle (|x_m - \bar{x}_o| + |x_o - \bar{x}_o|)^2 \rangle},
$$

(3.8)

where $\langle \rangle$ are averages over the entire simulation period, $m$ and $o$ refer to modeled and observed quantities, respectively. This statistic is known as the Willmott Skill Score [Chen et al., 2015] and is a measure of the agreement between any model output quantity $x_m$ and the corresponding observed quantity $x_o$ [Willmott et al., 1985]. A score of 1 means perfect agreement and 0 means no agreement is observed. Table 3.2 gives skill scores calculated at various observation stations deployed during the RIVET experiment. The locations of the stations are given in Figure 3.7. The $\chi$ values for along- and cross-shore water velocity were calculated using depth-averaged values of the observed water velocity profiles. Using depth-averaged values may be misrepresenting the effect of local wind-driven flows, which can be highly three-dimensional [Wong and Valle-Levinson, 2002]. In general, the model agrees with observed water level well ($\tilde{\chi} = 0.9115$). For the depth-averaged circulation, there is generally better skill for the cross-shore currents ($\tilde{\chi} = 0.8306$) than for the along-shore component ($\tilde{\chi} = 0.66$). The lower skill in the along-shore would be more reasonable in the tidal channels where the flow is polarized by the tidal currents and cross-stream flows would tend to be small and difficult to capture in both the model and observations. This is evident by comparing

10The skills reported here were calculated using model runs from Rynne [2016]. The time period of those runs extend to May 22, 2012.
skill of $U$ and $V$ (cross- and along-shore, respectively) at stations C1 and 52. For the latter, the tidal flow is oriented cross-shore and along-shore for the former; and there is a corresponding switch in model-observation agreement from $U$ to $V$. While the focus of the analysis for this study is not on model-observation comparisons, it is useful to assess the ability of the model to represent the physical conditions at NRI.

Table 3.2: Table of skill scores at select observation station locations for water elevation ($\eta$), cross-shore velocity ($U$), and along-shore velocity ($V$). Entries with "-" means no observation data was available for comparison. Adapted with permission from Rynne [2016].

<table>
<thead>
<tr>
<th>Station</th>
<th>$\eta$</th>
<th>$U$</th>
<th>$V$</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.91</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>2</td>
<td>0.83</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>3</td>
<td>0.98</td>
<td>0.98</td>
<td>0.96</td>
</tr>
<tr>
<td>4</td>
<td>0.84</td>
<td>0.96</td>
<td>0.84</td>
</tr>
<tr>
<td>5</td>
<td>0.97</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>6</td>
<td>0.64</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>7</td>
<td>0.90</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>8</td>
<td>0.97</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>9</td>
<td>0.92</td>
<td>0.70</td>
<td>0.41</td>
</tr>
<tr>
<td>15</td>
<td>0.99</td>
<td>0.46</td>
<td>0.70</td>
</tr>
<tr>
<td>26</td>
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Figure 3.7: A portion of the NRI domain with the observation stations used for the skill assessment, Table 3.2, shown. The station locations correspond to actual locations of *in situ* observations made during the simulation period.
Figure 3.8: Cross-sectional view of the stations 3 to 9. Axes are given in local coordinate system, where $X$ is the distance across-shore and $Z$ is defined as positive downward. Note the bed slopes are exaggerated by the axis scaling and the physical slopes on the left and right of the shoal are both $\sim 0.006$ and $\sim 0.01$, respectively. The contours show the deformation of the grid cells due to the $\sigma$ coordinate system ($\Delta X$ reduced for illustration purposes).

3.3.1 Control Simulation: Tide Only

Time series of surface flows and water levels at select grid cells (red squares from Figure 3.7; see Figure 3.8 for cross-sectional view) along a cross-shore transect in the NRI mouth are given in Figure 3.9. These are only given for the Tide Only (TO) simulation, which will be considered as the control test for the time series analysis. These currents have been band-passed filtered with cut-off periods at 200 and 3 hours, the total simulation period for all model runs was 432 hours (18 days). This filtering was applied to all currents output in order to remove ultra-low frequency signals that may cause spurious correlation when different time series are compared directly.
During the entire simulation period, a spring-neap tidal cycle was captured with spring surface currents ranging from 1.3 m/s to 1.5 m/s on the ebbing and flooding tide, respectively (at station 4). This net-flood discharge through the inlet mouth was also observed in the along-channel momentum from in situ observations made during this period and at a similar location in the inlet channel [Wargula et al., 2014]. The strong
along-shore flows at station 3 are most likely due to its proximity to the S-Bend (see Figure 3.7) just up-stream—this corresponds to lower across-shore velocities at this station relative to station 4. A significant drop in both ebb/flood surface currents was observed in the model output along the cross-shore transect, which spans the ebb shoal region of NRI. Station 9 observed maximum (ebb or flood) currents that were nearly five times lower than station 4, which observed the highest currents. Cross-correlation analysis of the across-shore surface currents showed that station 9 exhibited a 1 hour lag relative to station 4, but no appreciable lag is observed in any of the other stations with the respect to station 4 time series (Figure 3.10). While a steady drop in max correlation from stations 4 to 8 was observed, a far more dramatic decrease in the max correlation from station 8 to 9 \((r = 0.9246\) and \(0.7242\), respectively) was observed. Except for station 3, all other stations in this transect observed relatively weak along-shore currents with a high degree of station-to-station variability.

Spectral decompositions of the across- and along-shore surface currents are given in Figure 3.11 and the spectral amplitudes have been normalized by their respective maximum. For the across-shore flows, the dominant signal is from the semi-diurnal, \(M_2\), tidal frequency band with some energy at the diurnal band, but this was observed to be only 3% of the energy in the semi-diurnal band. All stations also observed a peak at the quarter diurnal band, but this was only significant

![Figure 3.10: Cross-correlation analysis for \(U(x,z=0)\) over the entire simulation period. The linear correlation coefficient is given on the vertical axis. For the cross-correlation, station 4 was used as the reference. Color scheme is same as Figure 3.9.](image)
relative to the semi-diurnal peak for station 9. For the along-shore spectra, there is much more inter-station variability and the energy is distributed over a much wider range of frequencies. However, in absolute terms the energy in the along-shore spectra tends to be 1 to 3 orders of magnitude lower than in the across-shore spectra, for all stations. The exceptions would be stations 1 and 9 where there was less difference between across- and along-shore currents (see Figure 3.9).

Figure 3.11: Across- (left) and along-shore (right) power spectral densities of the autovariance functions for surface currents observed at stations 3 through 9. The spectral amplitudes have all be normalized to the peak.
3.3.2 Delft 3D Sensitivity to Nominal Wind Forcing

The first step in addressing the hypothesis posed in this chapter is to quantify the net effect of the nominal wind forcing on the modeled circulation. Here, nominal means time varying, spatially uniform wind forcing applied using the default wind stress parameterization (see Figure 3.6). The Tide+Wind (TW) simulation (Figure 3.12) was run with tidal and wind forcing being the only primary drivers of the NRI system in the model. Qualitatively, there is a noticeable change in the surface flows when wind forcing is applied to the model, especially for the along-shore velocity component. These effects are most noticeable in stations 8 and 9, which are the furthest offshore in stations this sample transect.
Figure 3.12: Wind velocity (top) and across-(middle) and along-shore (bottom) surface currents time series for the TW simulation. Stations 3 to 9 are colored the same as previous figures.

Spectral decomposition of TW reveals that the main sources of wind variance are in the low frequencies, between the diurnal and synoptic scales (\(\sim 160\) hours; Figure 3.13). For both wind components, the energy in the diurnal band is close to 20\% of the peak amplitude. For the surface flows, the autovariance spectra showed that the across-shore variability was dominated by the semi-diurnal tide. Station 9 did show some relatively strong low frequency signals, which could be attributed to the wind, but the peak in the autospectra remained undeniably tidal. Far more station-to-station variability was
observed for the along-shore components and at Station 9 the spectral peak was clearly correlated to the synoptic variability in the along-shore winds.

**Figure 3.13:** Across-(left) and along-shore (right) power spectral densities of the autovariance for wind velocity (top) and surface currents (bottom). The spectral amplitudes have all been normalized to their respective peak amplitudes.

Time series of surface currents for TO and TW simulations for select stations along this example transect are given in Figures 3.14 and 3.15, for the across- and along-shore components, respectively. For the across-shore surface currents, through the entire simulation the addition of wind forcing has little effect on the flow. Probability density
functions of difference between the across-shore surface flows of TO and TW show that $\sim90\%$ of the PDF is contained within $\pm0.1$ m/s. The only exception is for Station 9, which is 1.33 km offshore of NRI and just offshore of the ebb shoal break. For this station, the surface flows are dramatically reduced: from ebb currents of $O(1)$ m/s at station 7 (0.36 km from NRI), to $O(0.25)$ m/s ebb flows. However, this decrease in the across-shore tidal strength corresponds with an increasing effect of the wind forcing, which appeared to be largely episodic in nature. The largest differences between TO and TW at Station 9 tended to occur during offshore-directed flow conditions, suggesting that the effect of the wind forcing is strongest on the ebbing tide.

**Figure 3.14:** Left) Time series of across-shore surface currents from Stations 3 (top), 7 (middle), and 9 (bottom) for TO (black) and TW (red) simulations. The corresponding across-shore distance in the local coordinates are provided. Right) The corresponding PDF of the difference between TO and TW. For Station 9, the positive skewness of the PDF reflects that the wind appeared to have a larger effect on ebbs as opposed to floods. This can be seen as the many of the peak ebb currents in the TO simulation being reduced during the TW model run. This suggests on-shore winds arresting the ebb jet as it come off of the ebb shoals.
For the along-shore currents, the difference between TO and TW are larger and occur closer to NRI (Figure 3.15). At Station 7, the surface currents were largely effected by the addition of the wind, not only in the magnitude and direction of the flow, but also the wind provides a new source of variability that was not present in the TO simulation. This effect is larger at Station 9, where the tidal signal is all-but removed from the surface currents and the flows at this location are almost entirely wind-driven.

**Figure 3.15:** Same as previous figure, but for the along-shore surface currents. Note the difference in vertical scale for Station 7 and 9, relative to Station 3, for the time series.

In order to quantify the net contribution of the wind forcing to the surface currents along the transect from Stations 3 to 9, the Fraction of Coherent Power (FCP) will be defined as,

$$
FCP = \frac{\sum_{i=a}^{b} S_{UU}(\gamma_i \geq \chi^2, \phi_i \geq 0) df_i}{\sum_{i=1}^{N} S_{UU} df_i},
$$

(3.9)

which is the fraction of total power in the autovariance function $S_{UU}$ that can be directly correlated to the wind velocity spectrum through the coherence, $\gamma$, and the
phase $\phi$. In other words, FCP quantifies the direct contribution of the wind to the observed variance in the currents. These parameters, $\gamma$ and $\phi$, are derived from the covariance spectrum ($Q$) between the wind ($u$) and current ($U$) time series,

$$\gamma = \frac{|Q_{uU}|}{|S_{uu}| |S_{UU}|},$$

(3.10)

$$\phi = \arctan \left( \frac{-\text{Im}(Q_{uU})}{\text{Re}(Q_{uU})} \right),$$

(3.11)

this was done for each station and both velocity components. In equation 3.9, $df$ is the frequency bandwidth corresponding to the spectral amplitude $i$, for these calculations $df$ was constant for all frequencies $N$. Only the across-shore components are shown here. The summation limits $a$ and $b$ are related to the total number of frequencies by:

$$1 \leq a \leq b \leq N.$$ 

The coherence cut-off, $\chi^2$ was calculated from,

$$\chi^2 = \frac{2F_\alpha}{\epsilon - 2 + 2F_\alpha},$$

(3.12)

where $\epsilon$ is the equivalent degrees of freedom, $F_\alpha$ is the value of the F-distribution at the confidence level $\alpha$ (unless explicitly noted, $\alpha = 0.05$). The phase and coherence of the cospectrum, $Q$, describe the relationship between the input and output signal, the wind and surface currents, respectively for all frequency bands. Fundamentally, spectral coherence-phase analysis seeks to quantify the transfer of energy from the input to the output. Therefore, if the two time series are stationary and the relationship between input and output is linear, there will be a strong coherence with a corresponding phase difference (i.e., the delay between input and output). If the transfer function between input and output signals is nonlinear, the coherence (i.e., spectral correlation) may not fully capture the relationship. Furthermore, the assumption is that the output response is solely due to the input, but if some other process contributes significantly to the output signal variance, this could result in a spuriously high coherence. These limitations
to this analysis technique were mitigated by bandpass filtering the time series prior to spectral analysis, to insure stationarity, and imposing coherence-phase conditions on acceptable spectral densities. Thus, the FCP value only captures the linear response between wind and currents and only those responses that could be reasonably considered as wind-driven. The issue of considering higher order processes was at least partially mitigated by running the analysis on simulation runs with different physical processes activated. The only obvious source of variability not isolated from these numerical experiments was bathymetry, since no model runs were run with a modified depth map.

Figure 3.16: Across-shore distributions of FCP for both the across-shore (left) and along-shore (right) currents. Here, the control model run is represented as a simulation with Wind Only (WO). FCP was calculated for surface (solid w/ marker) and bed ($z = -H$, dashed) flows. For the surface currents, an FCP calculated using a lower confidence interval to determine the cut-off $\gamma$ is also provided (thin solid).

The FCP calculated at each Station from 3 to 9 revealed a clear asymmetry in the role of the wind forcing on the currents (Figure 3.16). For the TW simulation, across-shore currents appear to be largely insensitive to the wind forcing, except for Station
9 where about 15% of the total variance could be directly attributed to the wind. This contrasts the along-shore components, which exhibit a stronger dependence on the wind forcing. Moving across the ebb shoal, FCP is relatively low, but steadily increasing until about 0.5 km offshore of NRI, where there is a dramatic increase in FCP to 0.5. From there, FCP increases steadily to between 0.8 and 0.9. For both components of the currents, the trend in FCP was reflected in the surface and bed flows, suggesting that the effect (or lack-there-of) of the wind forcing is consistent throughout the modeled water column.

### 3.3.3 Delft 3D Sensitivity to Enhanced Wind Forcing

Thus far, only Delft 3D simulations with default wind stress parameterizations have been presented. This was useful to establish the baseline sensitivity of the model to the surface wind forcing, but from the observations made in Ortiz-Suslow et al. [2015] (Chapter 2 of this dissertation) it was clear that the wind stresses in and around NRI are highly variable and not well-captured by a simple piecewise linear drag coefficient (Figure 3.6). Two enhanced wind stress parameterizations were used for this study: spatially uniform and spatially varying enhanced wind stress. The former was done simply by doubling the drag coefficient, for a given wind speed, at every grid cell in the model. This effectively doubles the nominal wind stress in the model everywhere and for all times. The spatially varying wind stress scheme was done as an O(1) recreation of the generally observed stress distribution from Ortiz-Suslow et al. [2015]. The goal was to build a cross-shore distribution of stress that was increased over land and then decreased to the nominal stress value (for a given wind speed) at the offshore boundary of the domain. Between these two states, there would be a 2 km wide transition zone. This cross-shore distribution was fixed in space, but could vary in time to match the
time varying wind forcing used in the nominal wind simulations. The highest stress value over land was fixed to three times the nominal value. An example of this spatial distribution for one particular wind speed is given in Figure 3.17. This method of creating a spatially varying wind stress field is a limited approach for two reasons: 1) the spatial distribution is wind speed independent, and 2) it does not take into account wind direction. Ortiz-Suslow et al. [2015] showed that both wind speed and direction play a role in spatial distribution of the stress field near NRI. However, there is currently no way of explicitly taking this into account in the model and any sort of scheme that attempts to account for these two limitations would essentially be speculative. While the approach used here is limited, it may still be instructive in providing some insights into the effects of a spatially varying wind stress field has on the model output.

![Figure 3.17](image)

**Figure 3.17:** An example of the resulting spatially varying wind stress field (magenta mesh) when the nominal wind speed $U$, here 11.1 m/s, was enhanced using the multiplicative factor $M$. For this field $\tau_0 = 0.164 \text{ N/m}^2$ and the maximum stress over land is three times this nominal value. The local topography is given underneath the mesh where color refers to meters above sea level.

Delft 3D does not currently allow for prescribing spatially varying stress fields. Therefore, in order to create the distributions of stress the nominal wind time series was
artificially enhanced using a spatially and time varying factor, $M$. The nominal wind stress, $\tau_0$, is calculated in Delft 3D using this piecewise relationship:

$$
\tau_0 = \begin{cases} 
\rho C_A^D U^2 & U > U_A \\
\rho \left[ C_A^D + \left( \frac{C_B^D - C_A^D}{U_B - U_A} \right) (U - U_A) \right] U^2 & U_A < U < U_B \\
\rho C_B^D U^2 & U > U_B.
\end{cases}
$$

(3.13)

Again, the user defines $C_A^D$, $U_A$, $C_B^D$, and $U_B$. The enhanced wind stress was calculated by substituting $U \rightarrow M \times U$. For $U < U_A$ and for $U > U_B$, $M = \sqrt{3}$ to yield a maximum enhanced stress of three times the nominal wind stress. For the middle equation, substituting in $M \times U$ yields a relationship that is quadratic in $M$, which can be readily solved numerically. The transition between a maximum $M$ and an $M = 1$ was 2 km wide and it was assumed that $M$ decreased linearly in $X$. The land-edge for the cross-shore distribution of $M$ was determined by searching from inshore to offshore and finding the last grid cell with an elevation greater than 0. Manipulating the nominal wind speed in this manner does not have any other effects on the model, since the primary purpose of inputting a wind field is so that Delft 3D can calculate a surface stress.
Figure 3.18: Similar to Figure 3.16, but with the results of the enhanced wind stress parameterizations included. All FCP values reflect surface currents and confidence levels with $\alpha = 0.05$. Enhanced 1 and 2 are the spatially uniform and spatially varying wind stress parameterization schemes, respectively.

The effects of universally and spatially varying increased wind stress does not have a drastically different outcome for FCP calculated along the test transect, relative to the nominal wind simulations for both along-shore and across-shore flows (see Figure 3.18). However, some slight differences were observed. For example, in the across-shore surface currents, the amount of variance driven by the wind almost doubles for Station 9 in both enhanced wind stress model runs, relative to the nominal wind simulation. The stations in the channel and over the majority of the ebb shoal report the same outcome. For the along-shore flows, the FCP value estimated at each station generally increases across the entire transect, but this effect is relatively small for simulations with tidal forcing activated. Typically, for both components of the circulation, the spatially
uniform enhancement resulted in slightly higher FCP relative to the spatially varying wind stress. In general, these results would suggest that changing the wind stress parameterization in the model does not create any new sources of wind-driven variance in the currents near NRI. Instead, the effects of the wind forcing are simply amplified.

Figure 3.19: Surface current vector map after 1 day of simulation for three different model simulations ($\tau_2$ refers to spatially varying wind stress enhancement). The dashed lines mark bathymetry contours as 4 m intervals. The green arrow is the nominal wind vector (5.7 m/s). The tide is near peak ebbing flow. All vectors are in a "going towards" convention. Every tenth vector is shown for clarity.

The FCP calculation is an integrated look at the effects of wind of the modeled flows, but further information can be gained by simply looking at snapshots of surface currents in different wind and tide conditions. Figures 3.19 to 3.21 provide three examples of surface current maps comparing the results for the TO, TW nominal, and TW enhanced model simulations. In broad terms, the surface current maps reveal a similar pattern as was determined from looking at the cross-shore distribution of FCP along the sample
transect. The wind has little effect at the inlet mouth and up in the channel, but this changes on the offshore edge of the ebb shoal and in the inner shelf region. This is general consistent between nominal and enhanced wind stress runs. However, closer inspection reveals some significant differences between the three simulations that are not very well-captured by the cumulative nature of FCP.

Figure 3.20: Same as previous, but after 4.8 simulation days. Nominal wind speed is 4.19 m/s.

Figure 3.19 provides a case of ebbing tide with offshore directed wind. In this scenario, just off of the shallowest part of the ebb shoal, the surface currents from the TW simulation have turned \( \sim 90^\circ \) relative to the TO currents. This steering of the currents off of the main tidal flow occurs earlier on the ebb shoal and the strength of the steered currents are stronger for the enhanced wind stress run versus the nominal simulation. During a flooding tide and onshore winds (Figure 3.20), the differences between the two TW simulations are small, relative the effect of adding wind to the
model. Also, during this scenario significant changes in surface current magnitude and direction, due to the wind, were evident to within a few hundred meters of the NRI mouth. The final example provided, gives an ebbing tide with an onshore wind (Figure 3.21). In this case, there is evidence for the modeled ebb jet being arrested by the opposing wind as soon as the flow moves offshore of the ebb shoal break. This retarding of the flow was more pronounced for the enhanced wind stress simulation versus the nominal run. The differences between the TO and TW simulations for each of three scenarios are summarized in Figure 3.22.

**Figure 3.21:** Same as previous, but after 16 simulation days. Nominal wind speed is 4.75 m/s.
Figure 3.22: Scatter plots of across-shore (right) and along-shore (left) surface currents for TO versus TW simulations for the three example vector maps given in Figures 3.19 to 3.21 (top-to-bottom in sequential order). In most panels, the clump of data close to zero that seems to diverge from the linear trend corresponds to data offshore of NRI, i.e. the region most effected by the wind. The highest values occur in the tidal channels upstream of the NRI.
3.3.4 The Effect of Waves

![Figure 3.23](image)

**Figure 3.23:** Time series of significant wave height (top) and average wave period (bottom) outputted from SWAN during wave-coupled simulations. The colored values come from Station 8 (the black lines reflect Station 9 nominal wind forcing). The output for each different wind forcing scheme is shown.

Until now, wave effects have been largely ignored. The wave model output was found to be largely insensitive to the wind stress parameterization scheme used, whether nominal or enhanced (Figure 3.23). This is not unexpected for two reasons: the fetch in the nested hydrodynamic grid and the nature of air-water coupling in the Delft 3D package. The maximum across-shore fetch (for an onshore wind) in the hydrodynamic grid is \( \sim 6 \text{ km} \), this is fairly short given the relatively mild winds forcing the model. Also, for the enhanced wind stress runs, these manipulations have been applied in the FLOW module. When waves are on, FLOW couples with SWAN for the wave current interaction, but it does not use the same wind stress formulation applied to FLOW. Therefore, the only effects the waves would feel from the enhanced wind stresses would be communicated through changes in circulation. An exception is for the spatially
varying wind stress simulations, because this was done by manipulating the nominal wind speed input this same wind input is felt in SWAN.

**Figure 3.24:** Similar to Figures 3.16 and 3.18, but showing simulations with wave coupling (and tides) enabled. The TW and WO simulation are provided as a reference. Enhanced 1 and 2 refer to wind forcing enhancement that was spatially uniform and spatially varying, respectively. Here, only locally generated (by the wind in the domain) waves are included.

Doing similar analysis with the wave simulations for the across- and along-shore surface currents, as was done for TW simulations, reveals that including waves does effect the FCP value for across-shore currents at the offshore edge of the ebb shoal (Figure 3.24). In fact, three times more variance can be directly linked to the wind at Station 9 when waves are turned on versus when they are off (compare Black and Green lines in Figure 3.24). An additions 10-15% of variance can be attributed to the wind when an enhanced wind forcing scheme is used; however, there is little difference between the enhanced wind stress schemes. This confirms that the increased wind stress in the
FLOW-WAVE coupled runs of Delft 3D does not actually communicate the increased wind stress from FLOW to WAVE and both modules separately handle physical air-sea coupling\textsuperscript{11}. It is somewhat difficult to interpret the observed model sensitivity of FCP when waves are activated. These are signals in the surface flows, which are coherent to the 95\% confidence interval with the wind and they occur as a response to the wind input ($\phi > 0$). This wave-added wind-coherent variability in the surface currents can be considered as a secondary (or indirect) effect of including the wind in the model—an effect which would be missed if wind were not included in the simulation. In the along-shore direction, the inclusion of waves has little effect on the FCP and one can see the change in FCP across-shore is mainly sensitive to the tidal and wind forcing.

3.4 Conclusions

The hypothesis posed at the beginning of this chapter was: \textbf{The circulation in an operational, process-based model is insensitive to the surface wind forcing.} The results of this modeling study indicate that this statement can be rejected. The model circulation demonstrated a clear sensitivity to the surface wind forcing even within portions of the domain where other forcing mechanisms (e.g., tidal) were shown to have significant influence. However, the degree of sensitivity was found to vary significantly from the wind having almost no effect on the circulation to the wind being the dominant source of variance. The influence of the wind forcing on the modeled circulation was isolated by running a series of models with different forcing terms included, as well as using statistical methods to determine variance in the currents that could plausibly be linked to wind input. The results revealed that the circulation dependence on the winds was

\textsuperscript{11}This is actually an entirely separate issue, which makes doing this type of analysis in wave-coupled runs difficult. van Nieuwkoop et al. [2015] actually observed model output sensitivity to making sure the air-sea coupling was consistent, using a different wind-wave-flow coupled model. For this reason, this wave analysis is treated as more of a \textit{caveate}.}
asymmetric between the across-shore and along-shore momentum and the wind was found to have a significantly larger influence on the along-shore flows versus across-shore flows. Further tests revealed that enhancing the wind forcing applied to surface did not dramatically change the overall contribution of variance the wind provided to the flow, but it did have significant local influence on both the across- and along-shore momentum. These local effects of the enhanced wind forcing were strongly time and space dependent. Therefore, it can be concluded that the wind influence on the modeled hydrodynamics exists in both a global effect (simply including wind) and a local wind influence, which was found to be sensitive to local changes in the wind forcing magnitude. These findings help to satisfy one of the original motivations of conducting this study, that being to determine if the local variability in the air-sea momentum flux observed at NRI, during RIVET I, translated to a net, local hydrodynamic effect.

A potential short-coming of this study is that it was done as an isolated numerical experiment and no significant attempt was made to compare these findings to in situ observations made at this coastal system. However, this study was designed to assess the reasoning behind a hypothetical coastal modeler’s assumption: that to first order modeling the circulation in a tidal system like NRI does not need to include wind forcing. The sensitivity analysis done here clearly shows that the wind forcing can be a significant, if not dominant, source of momentum even in portions of the domain where one would expect tidal forcing to be the primary driver of the flow. These findings suggest that accurately representing the local wind forcing in a coastal domain can have significant effects on the local circulation. This bears implications for coastal modeling, as well as in situ observational studies of coastal flows that may under-appreciate the role the of the wind in the local dynamics. While the focus of this study has been on the circulation and characterizing the role the wind plays in a statistical sense, these results
are also significant for other coastal processes, such as sediment and material transport, wave-current interaction, and mixing. Essentially, if the wind has a role in determining the circulation, there are a myriad of downstream (no pun intended) effects that would result from accurately accounting for variance in the local flows.
Chapter 4

Sea Spray Production in Very High Winds

Re-statement of Hypothesis

Large droplet spray production is important for spray-mediated fluxes in high winds. This hypothesis aims to address the following questions: What is the size-dependent distribution of sea spray above strongly forced wind-waves? Are there discrepancies between observations and conventional spray models? What is the significance of these droplets for air-sea interaction in these extreme conditions? The focus of this work will be on laboratory observations of large spray droplets (radius $>50$ $\mu$m), typically referred to as spume.
4.1 Background

4.1.1 Motivation

The balance between moist enthalpy input and wind energy dissipation at the air-sea interface is thought to be critical to tropical storm development [Emanuel, 1986]. Over the last several decades a respectable body of knowledge has been developed on the processes of mass, momentum, and energy fluxes at the ocean surface in light and moderate winds. However, our direct knowledge of these fluxes in high winds remains sparse. At wind speeds (referred to 10-m height) above 30 m/s, there is generally a significant quantity of spray entrained in the atmospheric boundary layer [Fairall et al., 1994]. This transition zone from the spray-free to a spray-laden boundary layer seems to correlate with the apparent saturation of the drag coefficient which occurs in the wind speed range of 30 to 40 m/s as observed in both laboratory [Donelan et al., 2004] and field [e.g., Powell et al., 2003; Potter et al., 2015] studies. One potential explanation for this is that the presence of spray in the boundary layer alters the vertical wind profile [Pielke and Lee, 1991; Barenblatt et al., 2005], which can be shown to change the theoretical logarithmic relationships used to derive the aerodynamic roughness [Lykossov, 2001]. Kepert et al. [1999] demonstrated that including spray affects in a coupled atmosphere-ocean model can have a dramatic affect on the air-sea fluxes during a simulated tropical cyclone. However, direct observations of this remain elusive. The impact that the development of this spray layer has on the exchange of heat and momentum remains relatively unknown [Haus et al., 2010; Jeong et al., 2012]. Determining this relationship holds implications for a number of oceanographic and atmospheric processes, including the rate of intensification of tropical cyclones [Soloviev et al., 2014].
The effects of sea spray on the air-sea interface is highly size-dependent since particles are generated over a wide distribution of sizes (nominally 1 μm to 1 mm). Droplets with radii less than about 25 μm are likely produced by bubble bursting [e.g., Monahan et al., 1986; Clarke et al., 2006] and have a minimal effect on momentum and heat fluxes. This generation mechanism is likely only a minor source for particles larger than 25 μm [Lewis and Schwartz, 2004]. Jones and Andreas [2012] summarized many oceanic spray observations and found that for $U_{10}$ greater than 16 m/s, spume (denoting spray generated by the mechanism of wave breaking) droplets greater than 50 μm may be generated. In the case of even higher winds, it becomes evident that large spume particles (300 - 600 μm) can be directly torn from the wave crests [Anguelova et al., 1999] or the destabilized interface [e.g. Marmottant and Villermaux, 2004; Soloviev et al., 2014].

### 4.1.2 Spray Effects on Air-Sea Momentum Exchange

There are a number of alternative explanations for the potential role of spray in altering the air-sea momentum exchange: (1) The return of spray droplets to the water surface suppresses the short waves that carry much of the stress [Andreas, 2004]. (2) The existence of a turbulence-suppressing, spray-laden layer above the air-sea interface inhibits the direct physical interactions between the sea surface and the atmosphere [Barenblatt et al., 2005; Bye and Jenkins, 2006]. (3) The onset of Kelvin-Helmholz instabilities leading to the production of spray and spume and the stabilization of the surface roughness as the crests of steep waves are blown off [Soloviev et al., 2014].

Early research suggested that the primary effect of spray would be to enhance air-sea momentum exchange because of the momentum required to accelerate spray droplets to the wind speed, which would then be transferred to the water upon re-entry [Munk,
1955; Pielke and Lee, 1991]. Subsequent studies based primarily on low-moderate wind regime measurements found this effect to be insignificant [Wu, 1972]. Fairall et al. [1994] calculated that the spray stress (wind to water) would have only a small influence (10% of total stress) for winds up to $U_{10} = 50$ m/s. However, this result may be primarily relevant to spray, not spume, particles given assumptions of the initial horizontal velocity of the droplets considered in that work. A confounding factor in these estimates is the difficulty in defining a realistic spray generation function in high winds. Andreas [2004] suggested that spray’s main role is to redistribute the wind’s momentum in the near-surface layer, with the spray acting to slow the near-surface wind speed by roughly 10% (for winds $\sim 30$ m/s). As a result, although the total surface stress may be the same as in the absence of spray, the spray contribution to that stress increases with wind speed while the interfacial contribution decreases. The author recalculated the spray contribution by separating the stress into an interfacial and spray contribution, and found that spray could have a much stronger role than estimated by the earlier works. Furthermore, he speculates that the spray droplets returning to the sea surface would suppress short waves and thereby lead to a reduction in the drag coefficient. The existence of a droplet evaporation layer [Andreas et al., 1995] close to the surface would further suppress turbulence close to the interface [Andreas et al., 2008] and lead to a reduction in interfacial transfers.

Alternatively, a theoretical model developed by Lighthill [1999], and expanded upon by Barenblatt et al. [2005], suggests a mechanism that would produce a sharp reduction in the atmospheric drag coefficient due to spray loading. This so-called “sandwich” model postulates that once the spray droplet size and concentration in the boundary layer reaches a threshold concentration the dynamics will evolve into a stably stratified multi-phase flow, where the effective density of each layer is determined by spray
concentration (akin to thermodynamic stratification with spray instead of temperature). This model results in a significant decrease in the stress that is supported across the ocean-atmosphere interface—now the ocean-spray-atmosphere interface. Their spray layer thickness and the boundary layer velocity profile were found to be strongly dependent upon the spray droplet size, with the larger particles being the most significant to the stress reduction.

Bye and Jenkins [2006] and Bye and Wolff [2007] presented a unified boundary layer model that incorporated some of the basic elements of the Lighthill [1999] approach. The model is unified in the sense that it combines a wave model, a spray generation function based on breaking waves, and a boundary layer turbulence model. The unified model predicts a leveling off of $C_D$ due to the suppression of turbulence by spray for winds greater than $\sim 40 \text{ m/s}$, with a maximum value at $42 \text{ m/s}$. Their results also show a flattening of the short wave field due to impinging spray, leading to reduced stress and a transfer of wave energy to lower wavenumbers. This sandwich model, along with alternatives presented, hinge on the vertical distribution of stress-carrying particles in the spray-laden boundary layer. Observations in very high winds where this is expected to be significant are generally lacking and these theories require further validation.

### 4.1.3 Spray Effects on the Enthalpy Flux

Latent and sensible heat fluxes both contribute to the total moist enthalpy flux. At winds high enough to generate spray, these fluxes are generated through two distinct mechanisms: (1) molecular scale interfacial fluxes that occur at the air-sea interface; and (2) spray fluxes produced by droplets ejected from the water surface into the air [Andreas et al., 2008]. Many authors have suggested [see Wu, 1979; Andreas and Decosmo,
1999, 2002; Andreas et al., 2008; Emanuel, 2003] that the spray contribution becomes important at wind speeds above $\sim 12$ m/s thereby increasing the overall moist enthalpy exchange coefficient. Models have shown that for wind speeds greater than approximately 20 m/s the spray sensible and latent heat fluxes are as large as the interfacial fluxes [Andreas, 1992]. Dropsonde data reported in Richter and Stern [2014] supports the Andreas [1992] findings for winds up to 70 m/s. However, in the laboratory, Jeong et al. [2012] found that the total moist enthalpy transfer was essentially constant with increasing wind speed up to 40 m/s. In a response to this article by Andreas and Mahrt [2015], the authors argued that Jeong et al. [2012] only captured the interfacial transfers, pointing to the need for additional spray-mediated enthalpy flux observations.

It has been shown by Andreas and Emanuel [2001] that spray can only cool the water volume by being ejected from the warmer water, cooling in the air, and then returning to water. To quantify this cooling mechanism, there are three key time constants (following Andreas et al. 1992, 2005, 2010) that must be considered: the duration of suspension of the spray droplet in the airflow $t_f$, the temperature evolution time or e-folding scale for droplet cooling $t_t$, and the droplet radius evolution time scale $t_r$. In the laboratory studies of [Haus et al., 2010] and [Jeong et al., 2012], the time $t_L$ before a droplet is advected out of the wind-wave tank test section must also be considered. The $t_L$ scale comes directly from the wind speed in the control volume, the location of spray generation, and the length of the tank.

The $t_t$ and $t_r$ scales have been estimated by Andreas [2005] for a range of salinities, however the duration of suspension remains uncertain. Andreas et al. [2010], assumed that spray was generated at the elevation of the significant wave height ($H_s$) and then would fall back to the mean water surface. Jeong et al. [2012] argued that this conceptual framework does not apply in the case of strongly forced waves. In previous studies
$H_s$ was usually chosen for the suspension height, although considerable lofting above this level clearly can occur due to nonzero vertical droplet velocities at the wave crest [e.g. Fairall et al., 2009]. Jeong et al. [2012] showed that the distance over which the droplet must fall before reentering the water depends on the downwind surface elevation in addition to the generation height. Andreas et al. [2010] used $H_s/2$ (i.e. significant wave amplitude) as the vertical length scale over which the droplet needed to fall, but in strongly forced conditions the droplet will likely impact a downwind wave at a level greater than the suspension height. It is likely that in wave tank studies this effect is more pronounced than in the open ocean as the waves grow in height over relatively short downwind length scales. The relevant time scale in this case is then related to the speed of the droplet relative to the wave phase speed and the wavelength. Mueller and Veron [2014] recognized this and developed a stochastic particle model to investigate residence times for a range of particle radii and wind speeds. They found that the droplet residence time is not represented by the simple free-fall concept and that there is a far more complicated relationship between the vertical particle distribution and the near-surface turbulent flow.

4.1.4 Spray Production

The evolution, development, and ultimate impact of entrained spray and spume in the atmospheric boundary layer all depend on the rate at which spray is produced and the size-dependent vertical distribution of these particles above the ocean surface. This foundational knowledge has yet to be fully described in the literature for the full spectrum of particle sizes. Particularly at wind speeds when spray is expected to be most significant for heat and momentum exchange across the air-sea interface. Spray generation source functions derived from limited field observations exhibit a very wide range
of values and none of these studies have achieved reliable measurements in hurricane conditions. The extreme conditions necessary to produce the spray volumes of interest make meaningful field observations an arduous, if not impossible, task [Melville, 1996]. Although the laboratory is a much simpler environment in which to make spray observations than in the field, it is still a complicated undertaking; consequently, there have been only limited observations of the spray distributions above breaking waves for either fresh or salt water [e.g. Fairall et al., 2009] in the laboratory. Veron et al. [2012] made observations in high wind speeds that show orders of magnitude divergence from the Fairall et al. [2009] production rate. More observational work is necessary to understand these results and to fully characterize the distribution of spume droplets above the wavy surface.

The present study addresses some of the gaps in the literature through laboratory observations in salt water. Direct measurements of spume concentration profiles were made in hurricane force winds above actively breaking waves. Experiments were done using filtered sea water and were conducted in 10-m equivalent wind speeds ranging from 36 m/s to 54 m/s. An unobtrusive optical technique was used which minimized the flow distortion and particle disruption. From these observed profiles, a bulk parameterization was used to estimate the corresponding source functions, these were compared to previous observations and typical source function models. Imaged particles ranged in radius from 80 µm to just over 1400 µm, which lie in the spume regime of the production spectrum. While laboratory experiments in these conditions are challenging, having controllable, repeatable experimental conditions makes this approach an attractive alternative to similarly-aimed field campaigns.
4.2 Methods

4.2.1 Quantifying Sea Spray Production

The following is a condensed treatment of this material. For a comprehensive review of ocean spray research to-date the reader is directed to Veron [2015].

The cumulative size distribution of spray particles above the air-sea interface is the integrated spray concentration of droplets with radius less than some radius, \( r \),

\[
N(r, x, t) = \int_0^r n(r', x, t) \, dr',
\]

where \( n(r, x, t) \) is the total number concentration of spray drops per unit volume of air per discrete particle radius increment from \( r' \) to \( r' + dr' \) [Veron, 2015]. The quantity \( n(r, x, t) \) is sometimes referred to as the concentration function [Veron et al., 2012]. The particle concentration is a function of droplet radius, height above the surface, and time. The temporal evolution of \( n(r, x, t) \) is given as,

\[
\frac{\partial n}{\partial t} = - \nabla \cdot (u_p n - k_p \nabla n) - \frac{\partial}{\partial r} \left( n \frac{\partial r}{\partial t} \right) + Q_n,
\]

(4.2)

here the explicit dependencies of \( n \) have been dropped. Equation (4.2) describes the time dependence of \( n(r, x, t) \) in terms of (from left to right) the three dimensional particle velocity \( u_p \), the radius-dependent particle diffusivity \( k_p \), the rate of change of the droplet radius, and a source-sink function \( Q_n \).

The first step towards a tractable solution to equation (4.2) is applying a standard Reynolds decomposition, i.e. \( u_p = \overline{u}_p + u'_p \), where overbars represent mean quantities and primed values are fluctuating values (i.e. \( \overline{u}_p' = 0 \)). This decomposition is also applied to \( n \) and \( r \) and equation (4.2) becomes,

\[
\frac{\partial \overline{n}}{\partial t} = - \nabla \cdot (\overline{u}_p \overline{n} + \overline{u}_p' n' - k_p \nabla \overline{n} + \overline{S}_n) - \frac{\partial}{\partial r} \left( \overline{n} \frac{\partial r}{\partial t} + n' \frac{\partial r'}{\partial t} \right),
\]

(4.3)
The last term on the RHS can be considered negligible for the particles sizes considered here, i.e., the relaxation time scale for the droplet radius is sufficiently large [Andreas, 1992]. Also, the source-sink term, $Q_n$, is brought into the parenthesis on the RHS and becomes the mean, size-dependent flux function, $\mathbf{S}_n = d\mathbf{F}/dr$. This term has units of number of particles per unit water surface area per unit time per discrete radius increment $dr$. The deposition velocity or settling velocity, $V_d$, is defined as the mean particle velocity relative to the mean flow velocity, $\mathbf{u}_p - \mathbf{u}$. Substituting this into equation (4.3) and removing the $\partial r / \partial t$ terms,

$$\frac{\partial \bar{n}}{\partial t} = -\nabla \cdot (V_d \bar{n} + \mathbf{u} \bar{n} + \mathbf{u}_p' n' - k_p \nabla \bar{n} + \mathbf{S}_n)$$

(4.4)

Equation (4.4) can be further simplified by moving the advective term, $(\nabla \cdot \mathbf{u} \bar{n})$, to the LHS and re-writing in terms of the total or material derivative of $\bar{n}$,

$$\frac{D \bar{n}}{Dt} = -\nabla \cdot (V_d \bar{n} + \mathbf{u}_p' n' - k_p \nabla \bar{n} + \mathbf{S}_n)$$

(4.5)

This flux equation further simplifies if assuming horizontal homogeneity, steady state conditions, and that particle diffusion is negligible. Invoking these conditions and re-arranging to get an expression for $\mathbf{S}_{nw}$, the vertical droplet flux,

$$\mathbf{S}_{nw} = V_d \bar{n} - \bar{w}_p n'$$

(4.6)

where the quantity $V_d = -|V_d|$ and $w$ represents the vertical velocity component. This states that the vertical flux of spume droplets can be described as a simple balance between particle deposition and vertical particle transport. Eddy covariance techniques can be used to measure this directly [Norris et al., 2012]. However, parameterizing this in terms of an eddy diffusivity model may be a necessary alternative [e.g. Rouault et al., 1991; Fairall et al., 2009]. The latter method is used here to infer the source
function because the covariance term, $w' \bar{n}'$, could not be directly observed during the experiments. Then equation (4.6) becomes,

$$\overline{S}_{nw} = V_d \bar{n} + K_p^t(z) \frac{\partial \bar{n}}{\partial z},$$

(4.7)

where $K_p^t(z)$ is a turbulent particle diffusion coefficient,

$$K_p^t(z) = \frac{\kappa u^* z f_s}{Sc},$$

(4.8)

here $\kappa$ is the von Karman constant (taken as 0.4), $u^*$ is the wind shear velocity, $z$ is the height above the free surface, $f_s$ is a slip factor, and $Sc$ is the spray droplet turbulent Schmidt number from [Rouault et al., 1991]. The slip factor attempts to parameterize the inertial diffusion of particles. The functional form of $f_s$ comes from Rouault et al. [1991] and is defined as:

$$f_s = \frac{1}{1 + C \left( \frac{V^2}{w'^2} \right)},$$

(4.9)

where $C \approx 2$ and $w'$ is the vertical wind variance. Fairall et al. [2009] assumed that $f_s$ was O(1) for their entire size spectrum (radii from 10 µm to 600 µm). For the purpose of this study $f_s$ will also be taken as $\sim 1$, but sensitivity analysis of this assumption was done and is provided later in this article.

Directly from equations (4.7) and (4.8), $\overline{S}_{nw}$ can be estimated given a measure of the mean vertical droplet concentration, $\bar{n}$. Conceptually, all of the spray observed in the boundary layer is assumed to be generated within some region above the breaking waves. Above this layer no spray is generated and therefore $Q_n$ is typically piece-wise defined as some delta function, $\bar{Q}_n = \bar{S}_0(r) \delta(z - h)$ [Veron, 2015; Fairall et al., 2009]. In this conceptual model $h$ is taken as the height above the surface that defines the generation layer, it is generally referred to as the theoretical source height and is
taken as the significant wave height, $H_s$. Following this framework, equation (4.7) then becomes a system of equations,

$$0 = V_d \bar{n} + K'_p(z) \frac{\partial \bar{n}}{\partial z}, \quad z > H_s \tag{4.10}$$

$$\bar{S}_0(r) = V_d \bar{n} + K'_p(z) \frac{\partial \bar{n}}{\partial z}, \quad z < H_s, \tag{4.11}$$

This is simplified if the concentration gradient within the spray generation layer is negligible [Fairall et al., 2009]. These two equations must match at the boundary $z = H_s$,

$$\bar{S}_0(r) = V_d \bar{n}(r, z = H_s) = V_d \bar{n}(r, z) \left( \frac{z}{H_s} \right)^{\frac{V_d}{H_s}}, \tag{4.12}$$

and the spray flux can be estimated from a mean size-dependent droplet concentration measured at some height $z$ above the reference level $H_s$. From Veron [2015], $\bar{S}_0(r)$ is known as the size-dependent spray generation function.

A critical problem in this formulation is how to represent $V_d$. Andreas et al. [2010] provides a detailed characterization of the vertical settling velocity for particles ranging from 0.5 to 300 $\mu$m. This velocity scale is both a function of height and particle radius,

$$V_d(r_0, z) = -\frac{V_g + V_a}{1 + \frac{V_a}{V_g}(1 - f_{\delta z})}, \tag{4.13}$$

here $r_0$ is the observed particle radius, $V_g$ is the size-dependent gravitational settling velocity, $V_a$ is a molecular sublayer transfer velocity, and $f_{\delta z} = \left( \frac{z}{\delta} \right)^{\frac{V_g}{\kappa u^*}}$, where $\delta$ is the sublayer thickness; the reader is directed to Andreas et al. [2010] (and the appendix therein) for details regarding the origin of these parameters. For this study, $V_d$ was calculated following this method and ranged from 0.6 to 10 m/s for the smallest to largest drops, respectively. For the size of the particles considered here (radius 80 $\mu$m to 1400 $\mu$m), the height dependence was found to be negligible. This comes directly from the the $V_g/\kappa u^*$ quantity, which was observed to be $>1$ for all radii and wind speeds considered here [see Andreas et al., 2010].
4.2.2 The Laboratory Facility

![Diagram of the ASIST facility with labels like Fan, Spray Obs., Wind Obs., Wave Obs., Open Circuit Inlet, Tail Tank, Beach, Flow Direction, Pump, Head Tank, Flow Straightener, and Settling Chamber.]

**Figure 4.1:** Upper panel: a diagram of the ASIST facility; lower left: the light source used equipped with telecentric lens and liquid light guide, this was mounted on a frame opposite the camera used for data collection (lower right picture). The green tinge in the water is from a fluorescent dye used as part of a completely different experiment, the observations presented here were done using non-dyed, filtered sea water.

The experiments were carried out in the University of Miami Air-Sea Interaction Saltwater Tank (ASIST), which has a 15 x 1 x 1 m acrylic test section (Figure 4.1). This facility is capable of generating both wind waves with a single turbine recirculating fan, as well as one-dimensional mechanical waves. Spray images were collected 11.05 m downwind from the wind inlet and the maximum sustained winds in the tank reached
54 m/s $U_{10}$. This was observed via sonic anemometer 2.35 m upwind of the imaged volume and the sampling volume of the anemometer was 20 cm above the still water line. The winds were referenced to 10-m following previous work done in ASIST [Donelan et al., 2004; Haus et al., 2010]. All of the experiments presented here were done with an initial water depth of 0.42 m and using 10 µm filtered sea water pumped in from a nearby tidal inlet (Bear Cut). There are no nearby freshwater sources and before being filtered the sea water was settled in a large basin to remove large particles. For all of the experiments, the salinity of the sea water was $\sim$33 psu.

An important physical reference in spray observations is the wave height, typically generalized via the significant wave height, $H_s$. The wave heights in ASIST were sampled at 10 Hz using a downward-looking Ultrasonic Distance Meter (UDM) mounted 5.7 m downwind of the inlet on the roof of the tank. The wave sampling was done in discrete 300 second blocks for separate wind speed regimes inside the tank; the conditions in the tank were allowed 120 seconds to become stationary prior to sampling. The $H_s$ used in this study was calculated spectrally,

$$H_s = 4\sqrt{m_0},$$  \hspace{1cm} (4.14)

where $m_0$ is the 0th moment of the elevation variance spectrum. The particle concentration profiles were adjusted to account for the change in MWL due to spray exiting the test section of the tank using the low frequency trend observed by the UDM for each respective wind trial. This trend in the water surface elevations was removed prior to the spectral analysis. This can be considered a conservative correction to the profile height because all of the spray imaging was done in less than 200 seconds of stationary wind forcing and the UDM-sensed water levels were sampled for 300 seconds.

Equation (4.8) requires some knowledge of the wind forcing on the water surface in
order to determine the $\bar{S}_0$. For a given $U_{10}$, the $u_*$ can be calculated following:

$$\frac{\tau}{\rho} = C_D U_{10}^2 = u_*^2. \quad (4.15)$$

The drag parameter used in the present study comes from work done in ASIST by Donelan et al. [2004], where three independent methods were used to estimate this parameter as a function of wind speed. This laboratory study compared the results of the eddy correlation technique [e.g. Edson et al., 2013] with those from a profiling and momentum budget method, the latter being a technique that uses mass conservation in the flume and the pressure-slope relation. This provides a robust estimation of the aerodynamic drag as a function of wind speed for this facility. The observed $H_s$ and $u_*$ in ASIST for a given wind speed are provided in Table 4.1.

### 4.2.3 Image Collection and Processing

Droplet concentrations were estimated using an optical technique and a Dantec Dynamics PIV data acquisition system was used to collect the spray imagery. The PIV system was modified for spray observations by re-routing the laser sheet through a liquid light guide to a strobe (Dantec Shadow Strobe). The strobe produces a collimated beam that eliminated size distortion based on the distance from the light source. The beam illuminated a small section of a diffuser screen mounted on the outside of the ASIST wall and a camera (JAI CV-MSCL, 1.9 MP, 30fps) was oriented directly opposite in order to image this region. Spray droplets being advected through the volume between the camera and screen were imaged as a shadow projection (see Figure 4.2). The camera-strobe system was mounted outside the acrylic tank and enabled undisturbed observations of vertical spray concentration profiles.
Figure 4.2: Examples of acquired images. The U (L) signify the Upper (Lower) acquisition levels and the number refers to 10-m equivalent wind speed. The red boxes and yellow circles represent particles identified and contoured by the automatic processing algorithm. Some identified spray have been circled, clockwise from upper left these drops have area equivalent radii of 92.5, 303.5, 181.5, and 396 µm.

Sampling was done at two reference levels centered at 95 mm (Lower) and 145 mm (Upper) above the MWL in order to reconstruct π profiles of sufficient length to capture the vertical variability in the droplet distribution. The acquisition was done in a “double frame mode” where images were collected in pairs separated by 500 µs. A total of 250 such pairs per collection were acquired at 15 Hz for five different wind speed regimes and for both the Lower and Upper reference levels. The pair sampling was limited by the laser flash rate, not the camera acquisition rate (30 Hz). The timing of the PIV system is precisely controlled and a collection of 250 paired images took 16.667 seconds to complete. Multiple collections of these sets of 250 double frame images were carried out continuously under stationary wind forcing conditions, with maximum acquisition time not exceeding 175 seconds. For each wind speed regime, it is possible that a variable number of collections of 250 images were acquired during the sampling (details in Figure 4.3). Though image pairs were acquired, only one frame from each pair was used for determining the size-dependent droplet number concentrations. Unfortunately, it was determined during the image analysis that the experimental conditions were not optimal for using these double-frames to extract the particle trajectories; however, single images are sufficient for quantifying the height- and size-dependent spray concentrations.
Figure 4.3: The sampling strategy used in this study. The far right column, for both levels, provides the total number of image collections, the number of images analyzed per collection, and the visually verified percent success rate of the counting algorithm. The variable total number of images per wind speed regime (middle column) was taken into account when computing the mean particle concentrations. Each set of images for a given wind speed was collected independently with laboratory conditions reset before starting another trial.

Image processing was done in two steps using the Dantec Dynamics shadow imaging software package. Each raw image was balanced initially to correct for irregularities in the image light sheet by taking the mean intensity of the set of 250 images. The raw frames were then normalized by this mean image, which resulted in higher contrast and easier particle detection. Droplet characterization was then carried out using an automatic shadow sizing routine. The detection algorithm was trained first by selecting a particle in one image in order to get a baseline for gray level contrast and edge gradients. The steepness of the edge gradient determines how in and out of focus detected
particles are—only gradients above a threshold steepness are considered in the plane of focus and counted. After the detection parameters were determined based on this “training”, the detection algorithm was automatically applied to every set of shadow images (>17,500 individual frames). The end result provided droplet centroid location and surface area and the radius reported in this study was calculated assuming spherical drops. This clearly begins to stop being valid for particles with radii exceeding 500 \( \mu m \) (see Figure 4.3). This is an issue for the radius-dependent spray generation function which relies on direct measurement of the particle radius at formation, \( r_0 \). For the observations in ASIST, some large particles are ellipsoids and \( r_0 \) may in fact be unstable or at least ill-defined, however little observational data exists that can quantify the significance of this distinction with regards to sea spray research. Mueller and Veron [2009] present a numerical model that does take into account these affects (as well as other processes related to particle deformation), which is partially based on previous laboratory experiments that investigated solid particle dynamics in turbulent flows [e.g., Clift and Gauvin, 1971]. For the purposes of this work, a simplified approach was used and the particle radius in question represents an area equivalent estimate.

Applying the identification criteria universally does not maximize detections for a particular set of 250 images, but it is a standardized means of spray detection and sizing, enabling direct comparison between respective data collections. This method is not only more expeditious than employing visual identification criteria, but it also minimizes the experimenter biases from the data set. It is expected that these biases would have been significant given the difficulty of consistently analyzing thousands of images. The success rate of the automatic processing algorithm was tested against visual inspections of the sets of 250 images. The results of the user-observed particle detection and the automatic processor were compared for every 25th frame of the 250
image sets. In general, the algorithm successfully detected 75-90% of the droplets. The lowest wind speed trial tended to be under-detected by the algorithm with a success rate just above 60%. Each set of 250 images was collected independently and it is possible that this may be a result of low image contrast in this particular trial, which was observed when the images were inspected and compared to other trials. For all of the experiments a radius dependence was observed in the algorithm’s success rate, in that smaller particles (10 pixels or less in diameter) were more likely to be missed than larger particles.

The camera used to acquire the droplet images was a medium telephoto lens with a 23.3° field of view. This is a non-telecentric lens and thus the image magnification exhibits a dependence on an object’s location within the depth of field; however, experimentally this source of error in the reported particle sizes is small. The focal plane of the camera was calibrated 0.59 m away from the lens (center of the air space in ASIST). This yields a depth of field of order 3 mm and results in a magnification error in the particle sizing around 1%. This is much less than the uncertainty associated with the automatic detection and sizing algorithm used to process the imagery. In the plane of the ASIST center line, the sampling volume represented by the camera frame was 55 mm x 75 mm (Figure 4.2). The across-tank dimension was determined in post-calibration to be 70 mm. This was determined to be the operational depth of field and was quantified using a target with standardized circles of known diameters (and separation) between 1 and 2.5 mm. The edge-detection algorithm was unable to discriminate between in and out-of-focus droplets to the precision of the camera lens’ depth of field (i.e., ± 1.5 mm). The pixel resolution of each frame was 42 µm, which was determined using the standardized calibration target.
4.3 Results

4.3.1 The Observed Droplet Distribution

Figure 4.4: Total mass concentration observed for each wind speed (color) and at the Upper (circles) and Lower (no symbols) collection levels, respectively.

Raw spray counts were expressed as the total mass concentration observed at each acquisition level and radius class normalized over the entire imaged air volume and for all of the collected images (Figure 4.4). These data indicate that the amount of observed spray decreased with height above the MWL and increased with wind forcing. There is a discernible peak in the mass concentrations spectra for both acquisition levels in the radius range between 500 and 800 \( \mu \text{m} \). The slope on the large radius side of the peak tends to be steeper than the small droplet side. When integrated across all radii, about 150\% more water mass per unit air volume is observed in the Lower frame versus the Upper frame, with this difference increasing slightly with increasing wind speed. The vertical difference is most pronounced for the larger particles. The mass concentration observed in the Upper frame saturates between 49.5 m/s and 54 m/s \( U_{10} \), but this does
not occur at these wind speeds in the Lower acquisition level, except at the largest particle sizes.

The raw image sets taken at the Lower and Upper levels were reconstructed into continuous profiles with 3 mm vertical resolution, i.e. horizontal slices through the entire imaged air volume that are 3 mm thick. The droplet concentration observed across each vertical slice and for each radius class (50 µm wide) was calculated using,

\[ n(r_i, z_j) = \frac{C_{ij}}{\Delta V N_I dr}, \]

(4.16)

where \( C_{ij} \) is the total number of observed particles in the \( i^{th} \) radius class and \( j^{th} \) profile bin, \( \Delta V \) is the air volume of each vertical bin, \( N_I \) is the total number of images in an observation period (e.g., from Figure 4.3 would give \( N_I = 7 \times 250 = 1750 \)), and \( dr \) is the width of each radius class. Equation 4.29 gives the number of particles per unit volume of air per radius increment for each vertical bin along the reconstructed profile. These concentrations are given in Figure 4.5 in two-dimensional grids—the vertical profiles are all scaled by their respective \( H_s \) for the given wind speed (see Table 4.1). As opposed to the mass concentration, the number concentrations clearly show that the largest number of particles were observed in radius classes < 500 µm. The two-dimensional distributions \( n(r_i, z_j) \), may have gaps where for a given wind speed, vertical location, and radius increment no particles were observed (see Figure 4.5). However, with increased forcing, these gaps tend to be filled as the imaged portion of the boundary layer becomes laden with spray across the entire size spectrum. In fact, from 36 m/s to 54 m/s the amount of "empty" space in the distribution decreases from nearly 85% to 32%. Physically, this signifies the filling of the boundary layer with spume drops. At the strongest wind forcing, particles > 500 µm are consistently observed at more than four times the local significant wave height and particles around
Figure 4.5: The color bar is common across all panels and shows log scaled (base 10) number concentration, which has units number particles per unit air volume per radius class. The profiles shown are scaled by the wind regime’s corresponding $H_s$ and the empty areas signify regions where no particles were counted.
1 mm become frequently observed at nearly three times $H_s$.

In order to isolate the effect increased wind forcing has on the vertical distribution of spray, it is useful to look at radius-integrated profiles. This is done by transforming the grids given in Figure 4.5 into a fractional spray volume profiles or in other words the volume concentration at each profile bin normalized by the total spray volume across the entire profile. The radius-integrated, volume concentration at each vertical bin is calculated using,

$$V(z_j) = \frac{4}{3}\pi \int r^3 n(r, z_j) dr, \quad (4.17)$$

which is then normalized by the total water volume observed for all $z_j$. This was done to yield a volume fraction profile for each wind forcing condition (see Figure 4.6a). The concentration information has been removed from these normalized profiles, but this does isolate the overall vertical dependence of the droplet distributions. The normalized profile at 36 m/s held more of the total observed spray volume lower in the profile when compared to the 54 m/s profile where a downshift was observed, signifying movement towards a more uniform vertical distribution. This transition is evidence in the intermediate wind speeds between the 36 m/s and 54 m/s experiments, however the absolute differences observed across the five trials are relatively small. The overall profile shape was observed to be fairly consistent for all the test conditions. The apparent compression of the profile at higher wind speeds (Figure 4.6a) is primarily due to the relative distance between the profile bins (which are fixed in the laboratory frame of reference) and significant wave height, which is increasing with the wind speed. This also explains the vertical offsets between the cumulative distributions in Figure 4.6b.
Figure 4.6: Integrated spray volume fraction profiles for all of the wind speed trials using the number concentrations from Figure 4.5. These profiles are radius-integrated they have units of volume of spray at a given height per total volume of spray produced for each wind speed (i.e., cm$^3$/cm$^3$). The profile height is scaled by the appropriate $H_s$.

The validity of the assumptions used to build the spray generation parameterization presented in Section 4.2.1 can be tested by transforming the profiles $n(r_i, z_j)$ down to the source height using equation (4.12). Theoretically, for a given wind speed and radius class, this transformation should capture all of the vertical variability in the profile and measurements made at discrete heights above $H_s$ should collapse onto a single effective source value. A sub-set of the results are given in Figure 4.7 alongside similar results from the Fairall et al. [2009] study and unpublished salt water observations made in ASIST during the Jeong et al. [2012] study. The Fairall et al. [2009] study made observations in both fresh and salt water (only the latter is used for comparison here) and both wind and mechanical waves were used to generate spray droplets. The work reported here in ASIST was conducted using salt water and only testing wind-waves. For
a comparison of fresh and salt water droplet production see Ortiz-Suslow et al. [2016].

Figure 4.7: Volume concentration spectra transformed down to the theoretical source height in volume of water per volume of air per radius increment. A sub-set of the results from this study (black, ASIST-SIS) are compared to salt water observations from Fairall et al. [2009] (blue, red, and magenta) and ASIST-CIP (green) data. The curves are referenced to their height above H_s and the corresponding U_{10}. For Fairall et al. [2009], this was estimated from the friction velocity and roughness length reported in Table 1 of that paper using the Law of the Wall and assuming neutral conditions in the surface layer.

All of the concentration spectra given in Figure 4.7 have been transformed down to the source level using equation (4.12). Like these previous studies, the \( n(r_i, z_j) \) presented here have been converted to volume concentration (the non-integral form of equation 4.17) and are denoted as dvdr. Both previous studies used the Cloud Imaging Probe (CIP, see Baumgardner et al. [2001]) to observe droplet concentrations. The data collected in ASIST using the CIP will be referred to as ASIST-CIP in order to differentiate it from the shadow images observed in ASIST as part of this work (denoted ASIST-SIS). The ASIST-CIP data was collected under comparable salinity conditions.
(filtered sea water) and approximately 2.5 m upwind of the ASIST-SIS observations. Both ASIST data sets show a size-dependent fall off in the normalized volume spectra, unlike the Fairall et al. [2009] observations which level off for radii greater than 100 \( \mu m \). In that study, the authors partially attributed this phenomena to a sampling bias in the functionality of the CIP, in that it may count two medium-sized particles as one large particle if their profiles overlap in the sampling volume. However, the ASIST-CIP observations do not appear to confirm this. It should be noted that the Fairall et al. [2009] and ASIST-CIP spectra have been smoothed to fill gaps in the curves, this was not done for the ASIST-SIS spectra.

The Fairall et al. [2009] observations in Figure 4.7 come from the same wind forcing, but different measurement heights and show within one order of magnitude collapse onto an effective \( dv/dr \) value at the source level. The convergence tends to get better with increasing particle radius, however it is unclear if this is a physical phenomenon or a result of the instrument bias noted in that previous study. In comparison, the ASIST-SIS data exhibits between half to one order of magnitude convergence, which remains fairly consistent across the entire size spectrum. In their analysis, Fairall et al. [2009] considered this level of collapse as satisfactory, but this approach may be questionable in the present study given the ASIST-SIS spectra. The differences in \( dv/dr \) within the Fairall et al. [2009] and ASIST-SIS data sets seems to be most sensitive to the position of the original measurement relative to the source height. This may also explain the relative differences between all of the volume concentration spectra for the three datasets from Fairall et al. [2009], ASIST-SIS, and the ASIST-CIP. This is somewhat counter-intuitive since wind forcing (or some other physical variable) would be expected to be a stronger predictor of the volume concentration at the source height. This suggests that there is vertical variability in the profiles that cannot be removed by this transformation. This
Table 4.1: Multiple regression results for the generation function $S_0(r)$.

<table>
<thead>
<tr>
<th>$U_{10}$</th>
<th>$u_*$</th>
<th>$H_s$</th>
<th>$\Gamma$</th>
<th>$B \pm \delta$</th>
<th>$m \pm \delta$</th>
<th>p’s</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>1.75</td>
<td>29.2</td>
<td>1.788</td>
<td>6.809 ± 0.725</td>
<td>-4.350 ± 0.406</td>
<td>[-2.19 7.11 -4.08]</td>
</tr>
<tr>
<td>40.5</td>
<td>1.97</td>
<td>31.9</td>
<td>4.949</td>
<td>6.476 ± 0.648</td>
<td>-4.062 ± 0.375</td>
<td>[-2.16 7.18 -3.98]</td>
</tr>
<tr>
<td>45</td>
<td>2.19</td>
<td>35.2</td>
<td>10.68</td>
<td>6.026 ± 0.518</td>
<td>-3.715 ± 0.308</td>
<td>[-1.96 6.49 -3.10]</td>
</tr>
<tr>
<td>49.5</td>
<td>2.43</td>
<td>36.5</td>
<td>22.25</td>
<td>6.022 ± 0.523</td>
<td>-3.672 ± 0.319</td>
<td>[-2.12 7.39 -3.97]</td>
</tr>
<tr>
<td>54</td>
<td>2.66</td>
<td>37.7</td>
<td>36.37</td>
<td>5.880 ± 0.502</td>
<td>-3.569 ± 0.313</td>
<td>[-2.10 7.39 -3.83]</td>
</tr>
</tbody>
</table>

The units for $U_{10}$ and $u_*$ are m/s, while $H_s$ has units mm. $\Gamma$ is the estimated mean, integrated spray flux in thousands of particles per unit water surface area per second (m$^{-2}$s$^{-1}$). The coefficients, $p_{123}$, from equation (4.22) are also given. The error estimates for $B$ and $m$ span the 95% confidence interval.

may explain why the collapse of dvdr onto a single value was limited for the ASIST-SIS observations.

4.3.2 Estimates of the Generation Function and Comparison to Previous Work

The following analysis is done with regards to the ASIST-SIS data set and reflects the observations of the present study. The observed $n(r_i, z_j)$ are useful for characterizing the vertical distribution of spume above the actively breaking water surface, but the ultimate goal of these types of observations is to quantify the vertical flux of droplets across a spectrum of particle sizes. $S_0(r)$ was inferred by scaling the measured number concentration at some radius class and profile bin by the appropriate deposition velocity and transforming down to the theoretical source level of $1= z/H_s$ (equation 4.12). This is done acknowledging that this is first order estimate to the spume flux, given the limiting success of this transformation in Figure 4.7. This was done in order to compare these observations to previously modeled generation functions. That the transformation used to infer $S_0(r)$ does not account for all of the observed vertical variability is expected to add some uncertainty to the estimate of the amount of spray generated rather than affecting the size-dependence, which is more indicative of the observed dynamics.
The ASIST-SIS $S_0(r)$ estimates are provided alongside estimates from various other works (Figure 4.8). A compilation of curves from Veron [2015] is provided as a relatively low wind reference (15 m/s $U_{10}$) and come from several model functions available in the literature. It should be noted that the Fairall et al. [2009] curve referenced in Figure 4.8a comes from an unpublished physically-based model developed during the original study (see the appendix of that article). The generation function from Mueller and Veron [2009] is also given for a high wind (50 m/s $U_{10}$) model comparison. Two source functions from the observational work of Veron et al. [2012] are also provided in Figure 4.8a and are inferred from the number concentrations presented in Figure 4 of that paper. This was done using equation (4.12) and since wave information was not explicitly provided in that study the appropriate $H_s$ was estimated using fetch-limited wave growth [Stiassnie, 2012]. This leads to an effective $z/H_s$ of 1.396 and 1.186 for the 41.2 m/s and 47.1 m/s spectra respectively. Following the authors’ suggestion in that article, the deposition velocity was estimated using Fairall et al. [1994].

For the ASIST-SIS curves in Figure 4.8b, $S_0(r)$ was estimated at each profile level and for each radius class and then vertically averaged to get a mean size-dependent estimate of the production,

$$S_0(r_i) = \frac{\sum_{j=1}^{N_j} S_0(r_i, z_j)}{N_j},$$

(4.18)

where $N_j$ is the total number of estimates along a profile for that particular radius class, $r_i$. From 36 m/s to 54 m/s the effective $z/H_s$ of these mean spectra ranges from 4.25 to 3.38, respectively—the decrease in effective height is due to increasing $H_s$ with increasing wind speed. More than an order of magnitude increase in the mean bulk production was observed from 36 m/s to 54 m/s $U_{10}$. The uncertainty of this estimate was quantified as the standard error of the mean at each radius class (the shaded region in Figure
Figure 4.8: A comparison of the spray generation function estimates. The color scaling is common to both panels and represents the 10-m wind speed (m/s). (a) The blue curves are adapted from Figure 6 in Veron [2015] and are model functions for $U_{10} = 15$ m/s. The model of Mueller and Veron [2009] at 50 m/s and laboratory scale fetch is also shown. The Veron et al. [2012] source function is inferred from particle concentrations observed in a laboratory. (b) The results from this study. Each curve represents a vertically averaged estimate of $S_0(r)$, with the shaded region spanning one standard error of the means at each radius class. Slopes with $r^{-3}$ and $r^{-8}$ are provided as a reference.
4.8b). This was done so that deviations from the mean $S_0(r)$ were directly comparable across radius classes and wind speeds, regardless of the number of sub-samples used to generate the vertical average. The relatively small uncertainty demonstrates that across the five different wind trials these mean spectra are fairly robust and the size- and wind speed-dependence of the spume droplet flux is well-represented. In general, the uncertainty tended to decrease with increased wind forcing because the statistics better converge as the boundary layer becomes filled with spume droplets. The largest uncertainties were observed for particles with radii exceeding 500 microns (especially at the low winds), which is most likely due to gaps in the profiles increasing the sample variability (see Figures 4.5 and 4.7).

For comparable wind speeds the estimates of $S_0(r)$ are about one half order of magnitude less than the estimates of Veron et al. [2012]. However, for both of these studies the radius-dependence of the $S_0(r)$ spectrum are remarkably similar ($R^2 > 0.98$ for all wind speeds) and exhibit a radius-dependence between $r^{-3}$ and $r^{-5}$. Some of this similarity may be due to these spectra being converted from concentration to source functions using the same parameterization. Even with less overall production observed, the ASIST-SIS results show several orders of magnitude more large particles (radii approaching 1 mm) than a production model with a size fall-off closer to $r^{-8}$ [Mueller and Veron, 2009].

The directly observed concentration spectra from both Veron et al. [2012] and ASIST provides a more explicit comparison between these two observational studies (Figure 4.9). The spectra from ASIST given in Figure 4.9 are vertically averaged over the lowest six bins of the profile and are effectively from a $z/H_s$ of $\sim 2.5$; the Veron et al. [2012] data come from imagery taken at very close to the spray generation layer, i.e. $z/H_s$ close to 1. In general, the ASIST-SIS measurements show fairly good agreement with
Veron et al. [2012] and this demonstrates that the differences in the estimated spray flux (Figure 4.8b) may be largely attributed to the transformation used to infer $S_0(r)$. This may further exemplify the bias observed when comparing the ASIST-SIS dataset to the Fairall et al. [2009] observations of dvdr at the source height.

Figure 4.9: (Left axis) Number concentration spectra are shown for Veron et al. [2012] (ASIST-SIS) at 10-m equivalent wind speeds 41.2 (47.1) m/s and 40.5 (49.5) m/s, respectively. The spectra from this study (blue and red) are vertical averages of the lowest six bins of the profile and the shaded region spans one standard error of the means. (Right axis) The corresponding $S_0(r)$ functions from this study.

The ASIST observations show a slightly shallower radius fall-off and thus significantly more circa 1 mm radius particles, which is somewhat unexpected given the relative heights of the observations. The ASIST measurements were made at more than twice the fetch of those from Veron et al. [2012] and so differences in the local wave development may reasonably explain this discrepancy. Regardless of these differences, in the context of the various model spectra given in Figure 4.8a, there is relatively strong agreement across these two studies that the large particle production is significantly under-predicted by existing model functions [e.g. Mueller and Veron, 2009].
Statistical methods can be used to quantify the wind speed- and radius-dependence of the source functions given in Figure 4.8b. This is primarily done here as an empirical exercise to quantify the dependencies within the ASIST data set and the physical implications of this analysis are largely secondary. Following Fairall et al. [1994], the spectra given in Figure 4.8 can be represented as the interaction of two independent functions of wind speed and particle radius,

$$ S_n(r, U) = W(U)f_n(r), \quad (4.19) $$

where $U$ is typically taken as a 10-m referenced wind speed and $S_n$ is the source function, $W$ is a function of the surface fraction covered by whitecaps, and $f_n$ is some mean size-dependent distribution [Fairall et al., 1994]. Equation (4.19) suggests that the shape of the spectrum is determined by $f_n$ while the amount of spray generated is determined by the size of $W$. To validate this assumption, the $S_0(r)$ spectra from ASIST were tested using a least-squares multiple regression. As a first estimate, the spectra given in Figure 4.8 can be described using a power law relationship and equation (4.19) takes the form,

$$ S_0(r, U) = W(U)r^m, \quad (4.20) $$

where $S_0$ is the source function observed in this laboratory study. For a given wind speed and radius class, this model suggests that $m$ is a constant and the amount of spume generated is determined solely by $W$ and when transformed into log-space this provides a linear relationship,

$$ S(R) = B + mR, \quad (4.21) $$

$S$ is the log base 10 scaled source function (the LHS of equation 4.20), $R$ is the log-scaled particle radius, $m$ is the power from equation (4.25), and $B$ is the log-scaled
$W$ term. In this form, $B$ and $m$ are the linear coefficients and were determined using multiple regression (Table 4.1). The $B$ parameter is well-described as a second order function of wind speed ($R^2 > 0.98$) and was observed to decrease with increasing wind speed (Figure 4.10a).

![Figure 4.10](image)

**Figure 4.10**: Results of the multiple linear regression analysis of the estimated source functions from ASIST-SIS. (a) The $B$ parameter shown as a function of wind speed with linear and quadratic fits provided, the dashed lines mark the 5% and 95% confidence intervals. (b) The $m$ parameter shown in a similar manner. $U$ refers to the 10-m equivalent wind speed.

$B$ exists over a relatively limited parameter space and tends toward a minima as the wind speed increases from 36 m/s to 54 m/s $U_{10}$. This suggests that less additional spume production is observed for each incremental increase in wind forcing. The power, $m$, was observed to vary slightly with wind speed and increased from $-4.4$ to nearly $-3.5$ (Figure 4.10b). Similar to the $B$ results, $m$ non-linearly increases with wind speed and could be well-described with a second-order polynomial ($R^2 > 0.98$). These findings demonstrate that the size-dependent distribution of particles observed in ASIST is
a function of wind forcing, which does not confirm the assumed separation of variables presented in equation (4.19).

Equation (4.21) is a first order approximation to the ASIST spectra from Figure 4.8b, which is a convenient simplification to make for testing the assumption built into equation (4.19). However, $S(R)$ may be a nonlinear function of $R$ and equation (4.21) can be represented as a second-order function,

$$S(R) = p_1 R^2 + p_2 R + p_3. \quad (4.22)$$

![Figure 4.11:](image)

**Figure 4.11:** of the nonlinear regression analysis. (a) The $S_0(r)$ spectra from Figure 4.8b (dots) alongside the corresponding regression curves (lines). Color denotes 10-m wind speed. (b) The corresponding slope functions. The the black solid (41.2 m/s) and dashed (47.1 m/s) curves are slope functions determined from polynomial fits to the Veron et al. [2012] source functions given in Figure 4.8.

Each wind condition was tested independently to determine the $p_n$ regression coefficients (Table 4.1) and this quadratic relation was observed to capture most of the variability in the mean $S_0(r)$ spectra, see Figure ??a. Higher order polynomials (up to order 6) were also tested, but no statistical advantage was observed with using a polynomial above order 2. The change in the size-dependence of the inferred source function
with increased wind forcing was quantified directly as the slope of this new $S(R)$, i.e. the first derivative (Figure 4.11b). A clear positive wind speed dependence was observed and slopes ranged from $-1$ to $-6$; similar analysis was applied to the Veron et al. [2012] spectra and revealed a similar trend. This empirical relationship suggests an altered version of equation (4.25), which can be found by transforming equation (4.22),

$$S_0 = 10^{p_1 R^2} \left( P_3 r_{p_2} \right),$$

(4.23)

here $P_3$ is $10^{p_3}$. This is a highly nonlinear relationship, largely due to the leading term on the RHS. The terms in parenthesis are similar to the Fairall et al. [1994] relationship, but as a whole this empirically-derived form of $S_0$ highlights that the ASIST observations diverge significantly conventional parameterizations.

Up to this point, wave-state dependent parameters have not been explicitly considered. However, the observed $H_s$ or other similar wave phase averaged quantity would not be a meaningful addition to the multiple regression model. This is because the wind speed is assumed steady for each experimental condition and the waves in ASIST are solely forced by the wind. In the laboratory, Fairall et al. [2009] show some relationship between the spray mass flux and the surface wave energy flux, but these experiments were conducted in the presence of both wind-forced and simulated swell waves, creating a surface condition not solely dependent on the wind speed in the tank. Wave related processes may play a role is isolating some of the observed variability, but this cannot be determined from the present experimental data.

As an additional note, the recent review by Veron [2015] includes a slip factor parameterization, $f_s$, in the calculation of the surface generation function (equation (4.9) of this paper). This coefficient is more of a counter-diffusion parameter since it char-
acterizes how the inertia of the large particles reduces their diffusion due to turbulence [Rouault et al., 1991]. Fairall et al. [2009] approximates this term as of $O(1)$ for spume droplets; and for lack of a better scheme, this was also used for this study. By inserting realistic values from this study into equation (4.9), $f_s$ may actually range from near 0 to roughly 1, depending on the particle size considered. A simple test is devised to analyze the sensitivity of estimates of $S_0$ to changing values of $f_s$ (Figure 4.12). The effective droplet flux for the largest particles becomes negligible as $f_s$ approaches 0, i.e. these particles do not diffuse and thus are not transported beyond the theoretical source height. This illustrates that there is significant outcome sensitivity embedded in this model, which may be overlooked simply by assuming $f_s$ $O(1)$. However, the observations presented in this study contradict the physical implications of this test because very large particles are readily observed well above the $H_s$ and in significant concentrations (Figure 4.5). This suggests that the basis from which this parameter is defined does not realistically capture the dynamics of these large spume droplets in these very strongly forced conditions.
Figure 4.12: $S_0(r)$ spectra for wind speed trials 45 m/s (left panel) and 49.5 m/s (right panel). The source function was calculated using equation (4.12), but the different colored curves represent different slip factor applications—the curves have been normalized to a common intercept. The slip factor was calculated as $f_s(r) = 1 - a$, where $a$ was the test parameter. The $f_s$ values were set as monotonically decreasing with increasing particle size, but with the smallest radius class always having $f_s = 1$. The $\min(f_s)$ denotes the slip factor for the largest radius class. A line with a $r^{-8}$ slope is provided as reference (note: the purple-cross curve extends much further, but is cut-off by the axis limits).
Table 4.2: Regression coefficients to the empirical number concentration profile given in equation (4.28).

<table>
<thead>
<tr>
<th>$r_0$ = 86 $\mu m$</th>
<th>$r_0$ = 286 $\mu m$</th>
<th>$r_0$ = 536 $\mu m$</th>
<th>$r_0$ = 786 $\mu m$</th>
<th>$r_0$ = 1036 $\mu m$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$U_{10}$</td>
<td>$a$ $R^2$</td>
<td>$a$ $R^2$</td>
<td>$a$ $R^2$</td>
<td>$a$ $R^2$</td>
</tr>
<tr>
<td></td>
<td>$b$ MSE</td>
<td>$b$ MSE</td>
<td>$b$ MSE</td>
<td>$b$ MSE</td>
</tr>
<tr>
<td>36</td>
<td>-2862.8 0.8763</td>
<td>-5524.5 0.8977</td>
<td>-7155.7 0.9182</td>
<td>-11016.1 0.8035</td>
</tr>
<tr>
<td></td>
<td>0.8532 0.1278</td>
<td>0.8067 0.1051</td>
<td>0.7972 0.0834</td>
<td>0.7689 0.2019</td>
</tr>
<tr>
<td>45</td>
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<td>-2542.1 0.8038</td>
<td>-4195.5 0.7639</td>
</tr>
<tr>
<td></td>
<td>0.7925 0.1766</td>
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<td>0.7629 0.1431</td>
</tr>
<tr>
<td>54</td>
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</tr>
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<td></td>
<td>0.7989 0.0443</td>
<td>0.7757 0.0321</td>
<td>0.7449 0.0216</td>
<td>0.7486 0.0195</td>
</tr>
</tbody>
</table>
4.3.3 An Empirical Number Concentration Profile

The results of this study demonstrate the breakdown of equation (4.12) and the theoretical framework used to develop this parameterization (Section 4.2.1). This is clearly evident in Figure 4.7, where larger values of dvdr at the theoretical source height came from observations made closer to \( H_s \) and the wind forcing on the system was of secondary importance. This suggests a vertical dependence that is not removed by the transformation. This was also evident when comparing the estimated source functions of Veron et al. [2012] and ASIST-SIS. While both studies observe comparable number concentrations (Figure 4.9), it is clear that the estimates of \( S_0(r) \) inferred from Veron et al. [2012] concentrations were high relative to ASIST-SIS because these measurements come from much closer to \( H_s \) (Figure 4.8b). The spume generation model used to estimate \( S_0(r) \) relies on an eddy viscosity model (equation 4.8) as well as a conceptual model describing the vertical distribution of droplets about a theoretical source height [Veron, 2015]. Invoking these conditions effectively prescribes a number concentration profile which can be seen by re-arranging equation (4.12),

\[
\bar{n}(r_i, z_j) = \bar{n}(H_s) \left( \frac{z_j}{H_s} \right)^{-\frac{V_d S_0}{\kappa u^* f_s}}.
\]  

(4.24)

where the first parameter on right-hand side is the number concentration at the source height. For a particular radius class and wind speed, the power term is theoretically constant and \( \bar{n}(z) \propto z^a \), which describes a profile shape that follows a power law relationship. This profile is used to transform measurements made at some \( z \) above \( H_s \) down to the theoretical source height, effectively assuming some knowledge about the vertical concentration variability. This theoretical profile can be explicitly tested against the directly observed profiles reconstructed from the raw imagery acquired as part of this study.
Figure 4.13: Select profiles from Figure 4.5. The values at the top of each column mark the particle radius class in microns while the values on the right side of each row give the wind speed regime in m/s. Observations from this study (black dots) are given alongside exponential (red), power (blue), and linear (green) regressions to the profile.
Select number concentration profiles are given in Figure 4.13 and have been taken directly from the two dimensional arrays in Figure 4.5. These profiles represent a range of wind speeds and particle sizes and each profile was independently tested against a power, exponential, and linear relation. The regression model which performed best was determined by which minimized the mean squared error between observed and predicted values. This analysis was performed in profile-space,

$$\zeta(\bar{n}) = b \bar{n}^a$$ \hspace{1cm} \text{power law} \hspace{1cm} (4.25)

$$\zeta(\bar{n}) = b a^\bar{n}$$ \hspace{1cm} \text{exponential} \hspace{1cm} (4.26)

$$\zeta(\bar{n}) = a \bar{n} + b$$ \hspace{1cm} \text{linear} \hspace{1cm} (4.27)

where $\zeta = z/H_s$ and $a$ and $b$ are the regression coefficients. Note that the power law relationship, equation (4.25), is the expected profile shape based on equation (4.24). The exponential model generally performed the best in explaining the observed profile variability ($R^2$ values between 0.7 and 0.96). The linear model sometimes outperformed the exponential, but this may be a result of the profile length and it is unclear if this result would change with a longer profile. It is important to note that the power law relationship never performed better than either the exponential or linear models. The exponential relationship in equation (4.26), can be inverted to give $\bar{n}$ as a function of $\zeta$ as in equation (4.24),

$$\bar{n}(\zeta) = \frac{\log(\zeta) - b}{a}.$$ \hspace{1cm} (4.28)

This describes a logarithmic number concentration profile which is at odds with the expected power law relationship. The regression coefficients, $a$ and $b$, are provided in Table 4.2. Equation (4.28) cannot be analytically derived from the system of equations posed by equations (4.10) and (4.11). This suggests that for the ASIST observations
this system is ill-defined and thus equation (4.12) is limited in providing an estimate of the spume generation from the number concentration measurements. As a result, the prescribed profile shape (equation (4.24)) does not best describe the vertical variability in the observed profile data, which may explain the spectral behavior noted in both Figures 4.7 and 4.8. The empirical formulae, equation 4.25 to 4.27, were all extrapolated down to $z/H_s = 1$ and compared with the results of equation (4.24), also extrapolated down to the spray generation layer (Figure 4.14).

**Figure 4.14:** Number concentrations extrapolated to the source height for the three empirical models tested for this study: power law (equation 4.25), logarithmic (equation 4.26), and linear (equation 4.27). Also shown are the results equation (4.24), which appeared as a volume concentration in Figure 4.7. The number concentrations from Veron et al. [2012] and Mueller and Veron [2009] are also shown. It should be noted that these have not been transformed down to the source height.

All of the curves in Figure 4.14 are derived from the observed number concentrations at some height $z$ above $H_s$, but their prediction of $n(r, z/H_s = 1)$ differ drastically.
Just comparing the two power law models, equation (4.24) and equation (4.25), reveals an apparent breakdown in the physical basis of the governing equations used to predict spray production. The latter model, derives a power law relationship from a conceptualization of the vertical distribution of the spray generation and the coefficients of the model are determined based on some physical arguments. The empirical model, used here to test the observations, does not prescribe the functional dependence of the power law coefficients, but it assumed that the shape of the profile determined by the theory was correct. Clearly, from Figure 4.14, the coefficients derived from fitting a power law profile to the observed number concentrations, do not coincide with the theoretically prescribed coefficients determined from the physical arguments used to develop equation (4.24). Interestingly, when extrapolated to the spray generation layer, there is relatively little difference between a logarithmic and linear concentration profile results. Strong similarity between the observations presented here and those from Veron et al. [2012] was observed and suggests that both laboratory studies measured similar dynamics and thus equation (4.28) may be applicable in some form across these independently conducted experiments. Unfortunately, there is no profile data available from Veron et al. [2012] so a direct comparison cannot be made to confirm this hypothesis.

As an empirical solution to the number concentration profile, equation (4.28) could suggest that the $V_d \pi$ term in the spray balance equation is negligible. Removing this term from equation (4.7) and solving for $\pi$ yields a logarithmic solution of the form of equation (4.28). From a physical perspective, it does not seem reasonable that vertical deposition of particles is negligible for spume, even though this would be an empirically justified conclusion given the statistical analysis conducted on the data. It is more likely that the $V_d \pi$ term has some vertical dependence that is being neglected. This phenomenon may be partially a result of using a solely size-dependent $V_d$, which may
not be appropriate for spume (E. Lewis 2015, personal communication). Recent modeling work by Mueller and Veron [2014] supports this claim, as they examined the size and height dependence of the spume transport and also found significant vertical dependence. Theoretically the deposition velocity used for this study is height dependent (equation 4.13), but the dependence was found to be negligible across our size regime. The significance of equation (4.28) is that it suggests some gaps in the theoretical framework used to model spume generation, but as a purely empirical finding, generalizing this result may be challenging and drawing direct conclusions about what this implies for spume transport equations goes beyond the scope of this work.

Figure 4.15: Two dimensional distributions of number concentration as a function of height and observed particle radius for fresh (left column) and salt (right column) water. The lowest wind speed (upper row) and highest wind speed (lower row) trials are provided for comparison. Color refers to the log-scaled number of particles per unit air volume per radius class, the color scales are equivalent across the four panels. Grayed cells represent no particles counted, white regions represent unsampled physical space.
4.3.4 Fresh versus Salt Water

Spray concentration was observed to increase with wind speed and decrease with height above the surface. This was quantified following Fairall et al. [2009] as the number of particles per unit air volume per discrete radius interval,

\[ n(r_i, z_j) = \frac{\text{Count}(r_i, z_j)}{a(z_j) Udtdr}, \]

where \( i \) and \( j \) are simply indices. The \( \text{Count}(r_i, z_j) \) is the total number of particles for a particular vertical bin and radius class, \( a(z_j) \) is the cross-sectional area of the vertical bin, \( U \) is the wind speed, \( dt \) and \( dr \) are the time interval and radius increment (here 50 \( \mu m \)), respectively. The along-tank and vertical extent of the total sampling volume was 55 \( mm \) and 75 \( mm \), respectively, with a corresponding across-tank extent of 70 \( mm \). This across-tank extent was determined in post-calibration as the operational depth of field. This was quantified using a standardized target with circles of known diameters between 1 and 2.5 mm. Investigations into the particle detection algorithm results revealed that the edge-detection method could not discriminate in and out-of-focus particles to the precision of the depth of field of the equipped lens (i.e., \( \pm 1.5 \) mm). Thus the across-tank dimension used here can be considered the effective region sampled within each image by the particle detection method. By examining the grayscale gradient on the edge of out-of-focus standardized circles, this across-tank dimension (i.e., operational depth of field) was found to be largely size-independent. The vertical dimension used to estimate the air volume was 3 \( mm \) or the resolution of the profile.

The \( n(r, z) \) for the highest (54 m/s) and lowest (36 m/s) wind conditions in both fresh and salt water (Figure 4.15) provide a two-dimensional view of the droplet concentrations and show the filling of the laboratory air boundary layer with spume particles. For all particle sizes, the number concentration tends to decrease with height
above the surface and increase with with wind speed. The vertical gradients in number concentration for the smallest observed particles (<200 µm) tend to reduce with increasing wind speed, as in the $z$ dependence becomes negligible. However this never occurs for the particle radii approaching 1 mm as their strong vertical gradients persist even during the strongest wind forcing. The height above the MWL here is scaled by the respective $H_s$,

$$H_s = 4\sqrt{m_0},$$

(4.30)

where $m_0$ is the $0^{th}$ moment of the surface elevation variance spectrum derived from time series collected via an Ultrasonic Distance Meter (UDM) which was sampled at 10 Hz and located 5 m upwind of the imaging system used here.

In comparing fresh water to salt water, the distributions in Figure 4.15 appear qualitatively similar, but some subtle differences exist which warrant highlighting. The number concentrations are presented here as a two dimensional particle density distribution that is a function of $z/H_s$ and $r$ (the latter is the observed radius). For fresh water, this distribution is concentrated in the small particle regime of the observed spectrum, while for saline conditions this distribution is far broader. This is evidenced by the relatively higher number concentrations observed in salt water opposed to fresh water out to an observed radius near 600 µm. This is seen at 36 m/s, but becomes more obvious in the 54 m/s wind trials (Figure

<table>
<thead>
<tr>
<th>$U_{10}$ [m/s]</th>
<th>FW</th>
<th>SW</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>85.3</td>
<td>84.5</td>
</tr>
<tr>
<td>40.5</td>
<td>70.6</td>
<td>65.5</td>
</tr>
<tr>
<td>49.5</td>
<td>53.3</td>
<td>37.0</td>
</tr>
<tr>
<td>54</td>
<td>44.9</td>
<td>32.4</td>
</tr>
</tbody>
</table>

Table 4.3: Percent of the $n(r, z)$ that was observed to be "empty" for fresh and salt water in all wind speed conditions considered in this study. Examples of the distributions are given in Figure 4.15.
This may be quantified by the "emptiness" of the distribution. Given the discretization done in the averaging, there are portions of these arrays where no particles are actually counted for a given \((z/H_s, r)\) coordinate, this information in terms of percentage for both water states are given in Table 4.3. These percentages were only calculated over of the shared vertical extent between the two data sets. There is a clear trend of decreasing emptiness with increasing wind speed, however this transition occurs faster in the saline water. From the distributions it is evident that this filling occurs most significantly in the larger particle regime. This may provide some evidence of the limiting effect of evaporation in the salt water, which can only lose a certain amount of their volume during their aerial transport Fairall et al. [2009].

Figure 4.16: (Upper row) The total spray volume fraction profile for fresh and salt water across all wind speed conditions. (Lower row) Corresponding cumulative summation. Note, the fresh water was normalized up to the top of the salt water profile in order to make a fair comparison across data sets—this explains the \(>1\) cumulation.
To highlight the vertical dependence, the radius-integrated $n(r, z)$ can be scaled into spray volume fraction profiles, as in the volume of spray observed in a vertical bin per total spray volume observed across the entire profile (Figure 4.16). In order to reduce noise due to the discretization, these profiles have been smoothed with a running median filter two vertical bins wide. The profiles for fresh and salt water are generally similar in shape, but consistently in fresh water more spray volume is lost vertically than is observed to occur in salt water. This can also be seen cumulative summations where the salt water profiles take longer than the fresh water to reach an equivalent fraction. These differences are fairly small and the profiles converge with increased forcing.

![Figure 4.17: Number concentration spectra for fresh and salt water at two different wind speeds. The $r^{-2}$ is also given for reference.](image)

The radius-integrated profiles are useful for garnering a broad sense of the vertical dependence of the number concentration profile, however, the primary interest in spray
observations comes in the radius spectra. Figure 4.17 provides a direct comparison between height-averaged number concentration spectra for both fresh and salt water. The fresh water spectra were observed to be consistently lower than the corresponding salt water size-dependent concentrations, for both the lowest and highest wind speed trials. This difference tends to increase with wind speed and decrease with particle size. The latter dependence may be partially caused by the diminishing statistical likelihood of observing supra 1 mm radius particles. However, some sensitivity with wind speed was observed with this large particle behavior, suggesting a physical phenomena (see Figure 4.17). This comparison between fresh and salt water was further characterized simply by looking at the size-dependent percent difference relative to fresh water number concentrations (Figure 4.18).

Figure 4.18: Percent difference between fresh and salt water at two difference wind speeds given as a function of particle radius.
With regards to the ASIST data, fresh and salt water trials exhibit general similarity, but some marked differences are noted especially in terms of the radius-dependence. Specifically, the concentration spectra for fresh water were consistently observed to be less than corresponding salt water spectra. Suggesting enhanced spume production in salt water versus fresh water. However, it should be mentioned that the spectral slope, i.e. radius-dependence, were markedly similar between the two water types. This may suggest that the fundamental mechanism of spume generation is similar in fresh versus salt water, except that the rate of production is increased in the latter. This somewhat agrees with the Fairall et al. [2009] observations where fresh water and salt show markedly different behavior for particles >200 \( \mu m \) (see Figure 7 of that paper). However, the nature of the difference observed is opposite to the behavior observed in that previous study. In Fairall et al. [2009], fresh water spectra in the larger particle portion of the spectrum exhibited higher number concentrations than was observed in their laboratory for salt water. The differences between this study and the previous may be the fact that the salt water used here was filtered physical sea water, whereas Fairall et al. [2009] conducted their studies with salty fresh water. Given that the various fresh and salt water observations made at different wind speeds and heights above of the surface were all done independently, it seems that the observations of the present work is purely a function of systematic bias in the observations. In summation, some of the results of the fresh and salt water comparisons may provide indirect evidence that bubble production of spray is not a significant spume generation mechanisms. However, the observed differences in number concentration suggest that their may be some physical mechanism related to spume production that hinges on the water mass chemistry.
4.4 Conclusions

The hypothesis posed at the beginning of this chapter was: **Large droplet spray production is important for spray-mediated fluxes in high winds.** Based on the laboratory studies conducted and the analysis done, a direct affirmation or rejection of this hypothesis is indeterminate. However, the results of this work strongly suggest that large droplet production, as observed in the laboratory, could have a significant effect on the air-sea fluxes. The hypothesis, which was proposed to be evaluated here, represents one of the most significant gaps in the understanding of spray-mediated fluxes. And in preparing for this study, we soon realized that one of the reasons for this lingering gap was a lack of information regarding the production of spray spume droplets and their distribution above breaking waves. Therefore, in order to provide a rigorous evaluation of this hypothesis, the first order questions regarding spray production needed to be addressed\(^1\). It was determined that spume particle production is generally underestimated by production models and this was especially true for very large particles approaching 1 mm in radius. Furthermore, it was determined that the vertical distribution of particles produced in the laboratory was not predicted by the theoretical vertical dependence of the spray concentration. Both of these findings suggest that our understanding of the rate of spume production and the transport of spume particles may be flawed, or at least misrepresenting the entire spectrum of spray production. Fundamentally, this motivates a reconsidering the *a priori* assumptions made about spume production and potentially revisiting the theoretical equations used to describe this complex and difficult to observe process. While the focus of this study has been on particle concentrations and spatial distribution, the implications of those findings do suggest that any models at-

\(^1\)And this, in and of itself, turned out to be a dissertation-amount of work.
tempting to account for spray effects on the fluxes may be misrepresenting the actual effects—especially if they rely on functions that produce too little spray.
Chapter 5

Final Remarks

5.1 General Conclusions

This dissertation is wide in breadth of topic and method. Using field observations, numerical simulation, and laboratory measurements this work sought to expand the understanding of air-sea coupling to coastal environments and strongly wind-forced conditions. On the surface, these two regimes seem diametrically opposed, though during recent and major natural disasters, not mutually exclusive. However, I would argue that they are fundamentally related in that they both reveal the terminus of the air-sea interface, as it is typically conceived.

Along the depth-limited coastal margins, the air-sea interface is literally coming to an end. This persistent spatial heterogeneity creates variance in the turbulent atmospheric boundary layer that cannot be seen with such strength over the open oceans and thus presents a domain where many of the assumptions built into momentum flux parameterizations breakdown. In a region where nonstationarity is the norm, novel methods of analysis and new perspectives on interpretation must be developed. For my part, the field work I have presented here represents one of the few air-sea interaction studies specifically focused on the coastal domain. Of those studies that have specif-
ically investigated air-sea interaction in coastal and nearshore environments, our work was the only to combine wind, wave, and current measurements. Furthermore, we were able to map out the spatial variability in the air-sea momentum flux and suggest local, coastal processes which could explain these signals. This air-sea coupling work was continued, and in a sense more generalized, via the modeling of the hydrodynamics of the New River Inlet. Essentially, the Delft 3D work was designed to address the skepticism we faced from some nearshore investigators: "OK, the air drag coefficient is higher at the inlet, so what?". This query summarizes the conventional wisdom of some in coastal studies: that any potential air-side forcing is second order or trivial in comparison to wave breaking or strong nearshore currents. Intuitively and, depending on the nearshore region of the observer, visually, this refrain would seem valid. However, armed with our observations (and a ready-made model domain), the New River Inlet presented an excellent opportunity to explicitly test these ideas. The numerical study results demonstrated that this conventional wisdom is not always, or necessarily true, and in fact the local wind forcing on the surface can play a significant role in the water-side dynamics. This has significant implications for accurately simulating coastal flows. However, equally important is that this work demonstrates that even hydrodynamics-focused observational campaigns should accurately represent the local wind forcing because, even with surf and currents, the wind can be an important source of variability.

While coastlines represent the physical end of the air-sea interface, as wind forcing increases to cyclone levels the binary nature of the air-sea interface physically breaks down. In these conditions, the amount of spray entrained into the lower atmosphere begins to form an intermediate layer that is some fraction air and some fraction water. This spray layer disrupts the vertical exchange from atmosphere to ocean and
thus plays an important role in air-sea coupling during storm events. Critical to our understanding of how spray affects the air-sea fluxes, is quantifying the rate of spray production and the vertical distribution (i.e., shape of the spray layer) of the entrained spray. To-date, the laboratory measurements presented in this dissertation represent the most comprehensive data set of directly observed spume droplet concentrations above the wavy surface, in tropical cyclone-like conditions, ever collected. There is only one other data set (laboratory as well) available in the literature that is comparable and it was conducted in lower winds and could not resolve the vertical distribution. From this data set, we were able to demonstrate that spray production models significantly underestimate spume droplet production rates and their vertical distribution is misrepresented by droplet transport theory. The implications of these findings suggest that even more work needs to be done to better understand how spray is generated and how these droplets are transported in the turbulent boundary layer. Only once these questions are answered, can significant headway be made in quantifying the effect of spray on the air-sea fluxes.

Coastal environments and very high winds both present regimes where our general understanding of the air-sea interface breaks down. While this dissertation has primarily taken an observational perspective, the ultimate aim of all of this work is to improve the predictive ability of numerical, air-sea coupled models. Fundamentally, these simulations must represent, either parametrically or directly, the vertical exchange across the air-sea interface. And if to be generally useful, these models must be applicable over a wide range of geophysical conditions. Continuing to feed these models empirical parameterizations that are themselves limited in scope, works against these ultimate goals and constrains the ability to represent and predict more general near-surface atmospheric and oceanic flows. Expanding our empirical knowledge base to new conditions
or domains will help improve model capabilities, which is significant for growing environmental and human concerns.

5.2 Future Efforts

Observational work is incredibly useful as a physical check on theories. However, it is also incredibly sparse. Further work will be focused on providing a fuller and more comprehensive observational study into coastal air-sea interaction and the effects of nearshore processes on the air-sea fluxes. Specifically, the goal will be to map out the transition region of the atmospheric boundary layer between land and sea. Orographic effects on the local wind and wind stress field need also be assessed. Some of this work is on-going with the CLASI data set and I intend to keep working as part of the investigatory team for this project. Additionally, this coastal work has helped to further reveal the dynamics of wind-swell interactions and this could pose a fruitful avenue of research into this topic, specifically focusing on coastal environments. For high wind conditions, further laboratory work is needed (and some is planned) to tackle the problem of spray spume production. The focus must be on understanding the actual generation mechanisms, how this relates to wave phase, and how this effected in a variety of conditions, i.e. with the addition of swell and mixed-seas. The ultimate goal of all of this work must be up-scaling. As a first step, larger laboratories (e.g. the RSMAS SUSTAIN Laboratory) would help to expand the ASIST work. However, continuing work should be done on developing a reliable field measurement technique for capturing droplet concentrations in tropical cyclone conditions. From a numerical perspective, in both of these regimes, more effort must be spent on incorporating these findings into existing model frameworks and assessing the sensitivity of the model output.
Appendix A

Comments on the Origin of Parameterized Air-sea Momentum Flux

The fundamental dynamical equation in continuum mechanics is known as the Cauchy momentum equation [Acheson, 1990]:

\[ M \dot{a}_i = f_i + V \frac{\partial \sigma_{ij}}{\partial x_j}, \] (A.1)

where \( M \) is the mass the control volume \( V \), \( F \) are the external forces, and \( \sigma_{ij} \) is the total, symmetric stress tensor. This relationship is essentially Newton’s Second Law for a non-rigid body whose deformation must be explicitly considered. The total stress tensor is given as,

\[ \sigma_{ij} = -p\delta_{ij} + \mu \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right), \] (A.2)

where \( p \) is the thermodynamic (conventional) pressure and the second term on the right is the shear stress, \( \tau_{ij} \), for a Newtonian Fluid, \( u \) is the fluid velocity and \( \mu \) is the shear (dynamic) viscosity. For a Newtonian fluid, \( \mu \) is a constant and independent of the strain acting on the fluid volume [Tritton, 1988]. From a geophysical fluid mechanics perspective, the more familiar Navier-Stokes Equations comes directly from
putting equation (A.2) into (A.1). After applying some simplifications and expanding the acceleration into the total material derivative, the elementary momentum balance becomes:

\[
\frac{\partial \rho u_i}{\partial t} + u_j \frac{\partial \rho u_i}{\partial x_j} = f_i - \frac{\partial p}{\partial x_i} + \mu \frac{\partial}{\partial x_j} \left( \frac{\partial u_i}{\partial x_j} + \frac{\partial u_j}{\partial x_i} \right) \tag{A.3}
\]

Along with the continuity equation (mass conservation),

\[
\frac{\partial \rho u_i}{\partial t} + \frac{\partial \rho u_i}{\partial x_i} = 0, \tag{A.4}
\]

these equations fully describe the motion of fluids in three dimensions and through time. For equation (A.3), from the left-to-right the terms are, time variation, convection, external body forces (e.g., gravity), the pressure gradient, and viscous diffusion. For simplicity, the non-intertial terms, i.e. Coriolis, have not been included. Bundled up in this seemingly innocuous set of equations is all of the complexity of a fully turbulent geophysical flow, e.g. a 1 m/s tidal current passing underneath Bear Cut Bridge.

In order to gain some insights into these flows, the game of fluid mechanics is to craft simplifications to the governing equations that, combined with observations, provide insights into the natural dynamics. The Reynolds approximation is a famous example of using statistical arguments to simplify these equations.

Assuming that any instantaneous value can be decomposed into fluctuations around a stationary mean,

\[
u_i = U_i + u'_i, \tag{A.5}\]

where \(U\) is the mean value and \(\prime\) denote fluctuations, the incompressible Reynolds-Averaged Navier Stokes (RANS) equations the form:

\[
U_i \frac{\partial U_i}{\partial x_i} = \frac{1}{\rho} F_i - \frac{1}{\rho} \frac{\partial P}{\partial x_i} + \frac{1}{\rho} \frac{\partial}{\partial x_j} \left( \mu S_{ij} - \rho u'_i u'_j \right). \tag{A.6}\]
All of the terms have been decomposed into mean and fluctuating components, temporal stationarity is assumed by definition of there being a $U$, $S_{ij}$ is mean strain rate, and the last term on the RHS is the covariance of fluctuating velocity components known as the Reynolds stress tensor:

$$
\begin{vmatrix}
    u_i' u_i' & u_i' u_j' & u_i' u_k' \\
    u_j' u_i' & u_j' u_j' & u_j' u_k' \\
    u_k' u_i' & u_k' u_j' & u_k' u_k'
\end{vmatrix}
$$

The diagonal elements are normal stresses and the off-diagonals (which are symmetric) are the shear stresses [Tennekes and Lumley, 1972]. It is not expressly noted here, but each element in the tensor represents the mean product of $u_i' u_j'$ components. Visually, the RANS equations help to make the original equations more comprehensible, but they also highlight the complexity of fluid turbulence. The original four equations (three-dimensional velocity field and continuity) described four unknowns, $[u_i, u_j, u_k, p]$, but now the Reynolds stress tensor introduces nine more variables without anymore corresponding equations. The RANS equations have a closure problem and a major focus of computational fluid dynamics is determining the best way to close this loop in numerical models.

The off-diagonal components of the Reynolds stress tensor are critical to the mean transport of momentum in sheared, viscous flows, where as the diagonals plays a more minor role [Tennekes and Lumley, 1972]. The $u_i' u_j'$ terms are statistical quantities stating that these fluctuating velocity components are correlated in time and so, on average, transport momentum in some direction. Turbulent eddies are the source of this covariance between velocity components; these eddies are generated through strain in the fluid and so interact with the mean flow to extract energy. But they lose this energy through their own viscous dissipation, creating small eddies, which extract energy from the larger eddies and so an energy cascade develops [Tennekes and Lumley, 1972].
This points to a complex relationship between the mean flow and the turbulence and so the Mixing Length Model (MLM), originally proposed by Prandtl [1925], means to describe this relationship in a simplified, but common fluid mechanics scenario: a one-dimensional sheared flow.

\[ \Delta M = \rho u_1(x_2, T) - \rho u_1(0, 0). \]  \hspace{1cm} (A.7)

Performing a Reynolds decomposition and assuming that turbulent fluctuations of
the particle do not effect the mean change in momentum, equation (A.8) becomes,

\[ \Delta M = \rho[U_1(x_2) - U_1(0)], \tag{A.8} \]

which can be estimated to first order as: \( \rho x_2 \partial U_1 / \partial x_2 \). Because we are interested in the momentum flux, we re-write this change in momentum per unit volume as:

\[ \tau_{12} = \rho x_2 \partial U_1 / \partial x_2. \tag{A.9} \]

Expanding the new absolute derivative in time of \( x_2 \), it can be shown that this term is the covariance between a characteristic velocity scale \( u'_2 \) and a transverse scale \( L \). This length scale must be the decorrelation scale of the covariance term, otherwise \( \tau_{12} \) would continually increase. \( L \) is known as the mixing length scale, hence the name of this model. Now, equation A.9 can be re-written in more familiar terms,

\[ \tau_{12} = \rho \nu_T \frac{\partial U_1}{\partial x_2}, \tag{A.10} \]

and by introducing the eddy viscosity \( \nu_T \) the shear stress on the fluid particle is directly proportional to the transverse gradient of the mean flow. The eddy viscosity is a turbulent exchange coefficient for momentum [Tennekes and Lumley, 1972] and it is defined as, \( \nu_T \equiv C u'_2 L \). The MLM makes for a relatively straight-forward set of coefficients and an equation for the shear stress in terms of a linear gradient of the mean flow. However, \( \nu_T \) and \( L \) are not fluid properties, like \( \mu \) or \( \rho \), so they may easily vary in space and time and are essentially determined by the local flow conditions. Therefore, the MLM is only a feasible option when these coefficients can be regarded as constant.

Adding a rigid boundary at the \( x_1 \) axis in Figure A.1 creates the classic fluid mechanics scenario of wall-bounded flow. All gradients in \( x_1 \) are neglected and the \( x_2 \) velocity scale, \( u_2 \), is constant (this could be, for example, a constant mass transfer ve-
locity). Thus, the momentum equation is:

\[ u_2 \frac{\partial U_1}{\partial x_2} = \frac{1}{\rho} \frac{\partial}{\partial x_2} \Sigma_{12}, \]  
(A.11)

where \( \sigma_{12} \) is the total mean stress. This can be integrated easily,

\[ \rho u_2 U_1 = \sigma_{12} - \Sigma_{12}(x_2 = 0). \]  
(A.12)

At the wall, \( U_1 \) must vanish. Out of convenience, the mean stress at the surface can be re-written in terms a velocity scale,

\[ \Sigma_{12}(x_2 = 0) = \rho u_*^2. \]  
(A.13)

This defines the well known friction or shear velocity, \( u_* \), which is a characteristic velocity scale of the total mean stress at the wall surface. For large Reynolds numbers, the mean viscous stress component, \( \mu S_{ij} \), can be neglected and so equation A.12 becomes,

\[ u_2 U_1 = -u'_1 u'_2 - u_*^2, \]  
(A.14)

and if \( u_2 \) can be neglected this provides the definition of the constant stress layer assumption typically applied near to turbulent boundary layers (e.g., the Atmospheric Boundary Layer). In this scenario, there is a direct relationship between the eddy vorticity and the mean flow vorticity,

\[ \frac{u_*}{L} = \alpha \frac{\partial U_1}{\partial x_2}, \]  
(A.15)

which essentially states that the turbulent eddies in the flow maintain their because of their interaction with the mean flow vorticity [Tennekes and Lumley, 1972]. If no other constraints are placed on the system, the \( \alpha \) should be order one and fluctuations in the mean should translate directly to the turbulence. The mixing length for
the wall-bound flow is defined as \( L = C x_2 \), where the constant of proportionality is the commonly known von Karman constant, \( \kappa \). Thus, the stress-gradient relationship from equation (A.10) for this scenario is,

\[
\tau_{12} = -u_1' u_2' = \kappa u_* x_2 \frac{\partial U_1}{\partial x_2},
\]

which can be integrated to yield the widely recognized Logarithmic Profile for the mean flow within a turbulent boundary layer. The drag coefficient comes into play if the drag force is introduced as,

\[
F_D = \frac{1}{2} \rho U^2 C_D A.
\]

This is the force a projected surface area, \( A \), feels due to the flow \( U \) outside of the boundary layer (at the object surface) with some density \( \rho \). The drag coefficient, \( C_D \) is the transfer coefficient that is a function of the Reynolds number of the system. If \( \tau = F_D / A \), then one can see how \( C_D \) can be related to the mean flow above the wall, the characteristic eddy velocity, and a transverse length scale. In micrometeorology, this transverse length scale would be the measurement height above the mean surface.

This brief review has gone from first principles, forces acting on mass, to a set of empirical relations meant to represent the turbulent motions in fluids. While these formulae have been presented in the general sense, equations (A.16) and () are the basis of the widely used parameterizations of the wind stress on the ocean surface\(^{13}\). These relations are the foundation of air-sea interaction observation and modeling. Decades of work has focused on developing the functional form of these empirical coefficients through observations in the field and the laboratory and then testing their generalization in numerical models. Fundamentally, these equations were derived using assumptions

\(^{13}\)All of this work is applicable to the land surface and in some cases significant steps forward in air-sea interaction study were made from over-land observations.
which highly constrain the application of the MLM and so an equal amount of investiga-
tive effort has been spent on making corrections to the empirical coefficients so that these simple, easy to digest relationships can continue to be used (e.g., Monin and Obukhov [1954]). Clearly, MLM has been successfully used in geophysical fluid dynamics and can describe some systems accurately. This review of the origin of the wind stress parameterization was given as an origin story for a widely used set of equations whose beginnings and built-in assumptions are not always appreciated or sometimes overlooked.
Appendix B

The Delft3D Governing Equations

Delft3D is multi-faceted software package developed by Deltares and the Technical University of Delft to be used for modeling coastal morphology, hydrodynamics, waves, and water quality. The Delft3D-FLOW (herein FLOW) module is the focus of this appendix because it is the engine behind the hydrodynamics in Delft3D as well as the most relevant to this dissertation. Much of this material comes directly from the Delft3D User’s Manual and from Lesser et al. [2004]. Other sources used will be cited explicitly.

The vertical coordinate system used in FLOW for this study is known as the $\sigma$-coordinate system. As opposed to a cartesian or $Z$-coordinate system, the $\sigma$-coordinate is defined along surface-following contours. The advantage is that the model grid can capture bathymetric changes better than a cartesian grid (think trapezoidal versus rectangular numerical integration), but at the expense of a variable vertical resolution in the water column. This coordinate system is related to the cartesian vertical grid by,

$$
\sigma = \frac{z - \zeta}{H},
$$

where $H$ is the total water depth and $\zeta$ is the free surface elevation above $z = 0$. This system is essentially a normalized vertical grid and $\sigma$ layers are defined as percentages of the total water depth. For numerical stability reasons, there is some guidance pro-
vided by FLOW as to how to define layer width near the bed as well as the transition in width from layer-to-layer. Below, the governing equations will be presented in cartesian coordinates, \([x, y, z]\), for compactness.

FLOW also uses the Generalized Lagrangian Mean (GLM) reference frame, which is only different from a Eulerian reference frame if waves are included in the simulation. GLM is defined as,

\[
U = u + u_s,
\]

where \(u_s\) is the Stokes velocity and \(u\) is the Eulerian water velocity. This is done for both horizontal components of the flow \(U\) and \(V\).

**B.0.1 The Shallow Water Equations**

FLOW solves the unsteady Navier-Stokes equations for shallow water in either two or three dimensions. The latter being a depth-averaged model. The system of equations is comprised of the horizontal momentum equations, continuity equation, transport equations for dissolved substances, and a turbulence closure model. The full Navier-Stokes equations, neglecting non-interial forces, are:

\[
\frac{\partial \rho U}{\partial t} + U \frac{\partial \rho U}{\partial x} + V \frac{\partial \rho U}{\partial y} + w \frac{\partial \rho U}{\partial z} = - \frac{\partial P}{\partial x} + \left( \frac{\partial \sigma_{xx}}{\partial x} + \frac{\partial \tau_{xy}}{\partial y} + \frac{\partial \tau_{xz}}{\partial z} \right)
\]

\[
\frac{\partial \rho V}{\partial t} + U \frac{\partial \rho V}{\partial x} + V \frac{\partial \rho V}{\partial y} + w \frac{\partial \rho V}{\partial z} = - \frac{\partial P}{\partial y} + \left( \frac{\partial \sigma_{yy}}{\partial y} + \frac{\partial \tau_{yx}}{\partial x} + \frac{\partial \tau_{yz}}{\partial z} \right)
\]

\[
\frac{\partial \rho w}{\partial t} + U \frac{\partial \rho w}{\partial x} + V \frac{\partial \rho w}{\partial y} + w \frac{\partial \rho w}{\partial z} = - \frac{\partial P}{\partial z} + \left( \frac{\partial \sigma_{zz}}{\partial z} + \frac{\partial \tau_{zx}}{\partial x} + \frac{\partial \tau_{zy}}{\partial z} \right) - \rho g
\]

\[
\frac{\partial \rho}{\partial t} + \frac{\partial \rho U}{\partial x} + \frac{\partial \rho V}{\partial x} + \frac{\partial \rho w}{\partial x} = 0
\]

The last equation is known as the continuity equation or mass balance. This set of equations describes any flow type, but does not contain an analytical solution [Roelvink and Reniers, 2012]. Therefore, a series of assumptions are made to derive the widely
used “shallow water equations”, which are essentially a simplified version of the Reynolds-averaged Navier-Stokes equations. Assuming incompressible flow and uniform density equation B.6 simplifies to a volume balance with $\rho \rightarrow \rho_0 = \text{constant}$. Assuming that each velocity can be decomposed into mean, fluctuating, and wave-coherent components,

$$U = \bar{u} + u' + \tilde{u}, \quad (B.8)$$

and after employing the eddy viscosity model\(^{14}\) the horizontal momentum equations can be written as,

$$\frac{\partial \bar{u}}{\partial t} + \bar{u} \frac{\partial \bar{u}}{\partial x} + \bar{v} \frac{\partial \bar{u}}{\partial y} + \bar{w} \frac{\partial \bar{u}}{\partial z} = -\frac{1}{\rho_0} \frac{\partial P}{\partial x} + \frac{\partial}{\partial x} \left( \nu_h \frac{\partial \bar{u}}{\partial x} \right) + \frac{\partial}{\partial y} \left( \nu_h \frac{\partial \bar{u}}{\partial y} \right) + \frac{\partial}{\partial z} \left( \nu_h \frac{\partial \bar{u}}{\partial z} \right) + M_x \quad (B.9)$$

$$\frac{\partial \bar{v}}{\partial t} + \bar{u} \frac{\partial \bar{v}}{\partial x} + \bar{v} \frac{\partial \bar{v}}{\partial y} + \bar{w} \frac{\partial \bar{v}}{\partial z} = -\frac{1}{\rho_0} \frac{\partial P}{\partial y} + \frac{\partial}{\partial x} \left( \nu_h \frac{\partial \bar{v}}{\partial x} \right) + \frac{\partial}{\partial y} \left( \nu_h \frac{\partial \bar{v}}{\partial y} \right) + \frac{\partial}{\partial z} \left( \nu_h \frac{\partial \bar{v}}{\partial z} \right) + M_y. \quad (B.10)$$

Now included in the governing equations are the $M_{xy}$ terms, which encompass the external sources or sinks of horizontal momentum, for example: wind stress, wave stresses, river discharge, etc. The shear and normal stresses from equation B.3, $\sigma^{15}$ and $\tau$, have been replaced by their corresponding horizontal eddy viscosity representation, an example is given here:

$$\tau_{xy} = \nu_h \frac{\partial \bar{u}}{\partial y}, \quad (B.11)$$

which relates the transverse gradient in some velocity component to the shear stress through a constant of proportionality, $\nu_h$, which must be determined using a turbulence closure model.

The vertical momentum equation is even further simplified by making these assumptions [Roelvink and Reniers, 2012]:

\(^{14}\)Delft3D does not directly resolve the turbulent motions and it uses the eddy viscosity and a turbulence closure scheme to model the turbulent dissipation in the horizontal momentum equations.

\(^{15}\)Not to be confused with the vertical coordinate system $\sigma$. Pardon the confusing notation, this only appears once here.
1. Vertical velocities are much smaller than horizontal velocities

2. The turbulent shear and vertical accelerations are small compared to the force of gravity

3. Changes in bottom topography are negligible.

Thus the vertical momentum equation reduces to the well-known, hydrostatic balance,

$$\frac{\partial P}{\partial z} = -\rho_0 g,$$  \hspace{1cm} (B.12)

which can be integrated analytically or numerically,

$$P = P_0 + \rho_0 g (\bar{\zeta} - z)$$ \hspace{1cm} (B.13)

$$P = P_0 + g \int_{z}^{\bar{\zeta}} \rho dz.$$ \hspace{1cm} (B.14)

Equation B.14 gives the hydrostatic pressure at all $z$ for a homogeneous fluid, while the latter equation must be integrated numerically for a variable density. $P_0$ is the background pressure at the surface. In most cases, this is simply the air pressure at sea level, $P_a$. Also, $\bar{\zeta}$ represents the short-wave averaged free surface elevation [Roelvink and Reniers, 2012].

Because the shallow water equations assume a scale relationship between $\bar{w}$ and the horizontal, mean velocities, the continuity equation (volume conservation) can be re-written in terms of the depth-averaged horizontal velocities,

$$\frac{\partial \bar{\zeta}}{\partial t} + \frac{\partial \bar{u} H}{\partial x} + \frac{\partial \bar{v} H}{\partial y} = 0.$$ \hspace{1cm} (B.15)

The reader is directed to [Roelvink and Reniers, 2012] (Chapter 3) for the full details of how the above relationship is derived. Equations B.10 to B.15 are known as the shallow water equations in three dimensions. FLOW also includes an advection-diffusion
equation to simulate dissolved material transport (e.g. heat or salinity),
\[
\frac{\partial HC}{\partial t} + \frac{HC}{\partial x} + \frac{HC}{\partial y} + \frac{HC}{\partial z} = H \left[ \frac{\partial}{\partial x} \left( D_h \frac{\partial C}{\partial x} \right) + \frac{\partial}{\partial y} \left( D_h \frac{\partial C}{\partial y} \right) \right] + \frac{1}{H} \frac{\partial}{\partial z} \left( D_v \frac{\partial C}{\partial z} \right) + HS,
\]
(B.16)

where \( C \) is the concentration of the dissolved substance, \( D_{h/v} \) are the horizontal and vertical diffusivities, and \( S \) encompasses all sources or sinks.

### B.0.2 Boundary Conditions

The kinematic boundary condition at the free surface and the bed stipulate that all vertical motions perpendicular to the bed go to zero,
\[
w|_{z=-d} = w|_{z=\zeta} = 0,
\]
(B.17)

where \(-d\) is the depth at the bed and the overbars have been dropped from the velocity components.

At the bed, the dynamic boundary conditions for the horizontal momentum are determined by the bed shear stresses,
\[
\frac{\nu_v}{H} \frac{\partial u}{\partial z} \bigg|_{z=-d} = \frac{\tau_{bx}}{\rho_0},
\]
(B.18)
\[
\frac{\nu_v}{H} \frac{\partial v}{\partial z} \bigg|_{z=-d} = \frac{\tau_{by}}{\rho_0},
\]
(B.19)

where \( \tau_{bxy} \) are the bed stresses in the \( x \) and \( y \) directions, respectively. These stresses are determined using some sort of bed roughness parameterization (the bed corollary to the wind stress parameterization). The dynamic surface boundary condition is defined in a similar manner:
\[
\frac{\nu_v}{H} \frac{\partial u}{\partial z} \bigg|_{z=0} = \frac{\tau_{sx}}{\rho_0},
\]
(B.20)
\[
\frac{\nu_v}{H} \frac{\partial v}{\partial z} \bigg|_{z=0} = \frac{\tau_{sy}}{\rho_0},
\]
(B.21)
where $\tau_{sxy}$ are the wind stresses in the $x$ and $y$ directions, respectively.

Open boundaries are virtually defined as water-water boundaries at the edge of the computational domain. These need to be carefully determined to maintain a physically realistic model while balancing the mathematical constraints of the numerics\textsuperscript{16}. There are several types of open boundaries that FLOW can prescribe:

- Water Level: $\zeta = F(t) + (P_{\text{average}} - P_0)/\rho g$,
- Velocity: $U = F(t)$,
- Discharge: $Q = F(t)$,
- Neumann: $\partial \zeta / \partial n = F(t)$,
- Riemann: $U \pm \zeta \sqrt{g \alpha}$.

$F(t)$ is some arbitrary function of time, $t$. $P_{\text{average}}$ is user-defined as the average pressure at the boundary (can be equivalent to the atmospheric pressure). Here, only the $U$ or normal to the boundary component is shown, but there would be an equivalent set of conditions for the tangential component. For discharge, $Q$, boundaries the shape of the profile is prescribed the user as either "uniform" or "logarithmic".

Closed boundaries are defined as the water-land boundary. For the normal component of the flow, the condition is simply "No Flow" through the boundary. For the tangential component a free slip or partial slip condition can be imposed. For simulations where the side wall effects can be neglected, the free slip is the best option. The partial slip condition should be used in small scale models where edge or side wall effects could be important to the main flow field.

\textsuperscript{16}If computation cost was not a concern, then numerical domains would not need to be constrained and every model could be run on a global scale so that water-water boundaries would not have to be imposed. Unfortunately, this is not the case.
For material transport, vertical and horizontal boundary conditions must be prescribed. For the horizontal boundary condition, only inflow conditions need to be prescribed, whereas outflow boundary conditions are dictated by the advection within the numerical domain. In other words, for material flowing into the domain from the virtual space outside of the numerical domain, the model has to know what the conditions of that material are in the virtual space. Thus, FLOW uses the Thatcher-Harlemann boundary condition,

\[ C(t) = C_{out} + \frac{1}{2} \left( C_{background} - C_{out} \right) \left( \cos \left( \pi \frac{T_{ret} - t_{out}}{T_{ret}} \right) + 1 \right), \quad 0 \leq t_{out} \leq T_{ret}. \quad (B.22) \]

\( C_{out} \) is the concentration \( C \) at the open boundary from the last outflow event, \( C_{background} \) is the background condition, \( t_{out} \) is the elapsed time since the last outflow, and \( T_{ret} \) is the return period of material with concentration \( C \). When the flow transitions from outflow to inflow, the boundary condition simply becomes \( C_{out} \). The material concentration profile can be prescribed as uniform, linear, piece-wise, or as some time series given by the user. The vertical boundary conditions, as the surface and bed, are:

\[ \frac{D_{v}}{H} \frac{\partial C}{\partial z} \bigg|_{z=0} = 0 \quad (B.23) \]

\[ \frac{D_{v}}{H} \frac{\partial C}{\partial z} \bigg|_{z=-d} = 0. \quad (B.24) \]

**B.0.3 Time Integration**

The shallow water equations above were presented analytically, but in fact Delft3D solves these equations using numerical temporal and spatial integration. Apart from the simplifications used to make the full Navier-Stokes equations more digestible, the integration methods used introduce their approximations and numerical uncertainty. The Delft3D developers use this set of criteria to judge their solver:

- Unconditionally stable.
- Second order accuracy (i.e. \( \Delta t^2 \)).
• Can solve the unsteady and steady state problems.

• Computationally efficient.

Following the first criterion, FLOW uses an implicit time integration scheme, which has the advantage of always being stable (the model will not blow-up), but can be computationally expensive. An Alternating Direction Implicit (ADI) integration scheme is used to reduce the computational cost of the implicit method. This is achieved by splitting a single integration step from time step $l$ to $l+1$, into two half-time steps. This method is second order in time and space. The ADI method applied to the horizontal momentum and surface elevation equations is given as:

Step 1:

$$
\frac{2}{\Delta t} (\Pi^{t+1/2} - \Pi^t) + \frac{1}{2} A_x \Pi^{t+1/2} + \frac{1}{2} A_y \Pi^t + \frac{1}{2} B \Pi^{t+1/2} = \Lambda, \quad \text{(B.25)}
$$

Step 2:

$$
\frac{2}{\Delta t} (\Pi^{t+1} - \Pi^{t+1/2}) + \frac{1}{2} A_x \Pi^{t+1/2} + \frac{1}{2} A_y \Pi^{t+1} + \frac{1}{2} B \Pi^{t+1} = \Lambda, \quad \text{(B.26)}
$$

with $\Pi = [U, V, \zeta]$ and $\Lambda$ as the entire RHS of the shallow water equations. The $A_x$, $A_y$, and $B$ are defined as:

$$
A_x = \begin{bmatrix}
0 & 0 & g \frac{\partial}{\partial x} \\
0 & u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} & 0 \\
H \frac{\partial}{\partial x} & 0 & u \frac{\partial}{\partial y}
\end{bmatrix}
$$

$$
A_y = \begin{bmatrix}
u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} & 0 & 0 \\
0 & 0 & g \frac{\partial}{\partial y} \\
0 & H \frac{\partial}{\partial y} & v \frac{\partial}{\partial y}
\end{bmatrix}
$$

Anecdote: I was once listening to someone discuss the pros and cons of different coastal models and this person happened to disparage Delft3D because of its reliance on an implicit time integration scheme. They said that users of Delft3D could execute a poor simulation and generate questionable data, but accept it because the model itself did not become unstable. From an observationalist perspective, this is akin to saying that nobody should use sensor X because sometimes people mount/program it incorrectly and the data could be suspect, but it does not.
\[ B = \begin{bmatrix} \lambda & 0 & 0 \\ 0 & \lambda & 0 \\ 0 & 0 & \lambda \end{bmatrix} \]

In $B$, $\lambda$ is the linearized bottom friction coefficient.
Bibliography


