Air-Sea Interaction Dynamics Under Hurricane Wind Conditions

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AIR-SEA INTERACTION DYNAMICS UNDER HURRICANE WIND CONDITIONS

By

Sanchit Mehta

A DISSERTATION

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AIR-SEA INTERACTION DYNAMICS UNDER HURRICANE WIND CONDITIONS

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Understanding turbulent fluxes of momentum, mass, and energy across the air-sea boundary are fundamental to our ability to model and parameterize a number of multidimensional geophysical processes, such as wind-wave generation, oceanic circulation, and air-sea gas transfer. The physical nature of the near surface boundary layer remains less known, especially under high winds due to the development of an intermediate substrate layer of large spray droplets known as spume, between the atmosphere and ocean surface. Presence of these spume droplet effects the aerodynamic resistance of the prevailing winds over the surface and thus the behavior of surface drag coefficient. The size-dependent vertical distribution of spume particles in high wind conditions is necessary to understand their effect on air-sea fluxes of heat and momentum.

Given spume’s role in mediating air-sea exchange at the base of tropical cyclones or other storm events, the predominant focus of present literature studies on spray dynamics has been within the marine environment. In contrast, spume production in non-seawater bodies have not been extensively studied and potential differences between sea and freshwater are neglected. Thus any significant differences between sea and freshwater
remain unquantified. Direct measurements of the physical processes happening at this interface remains scarce till date due to difficulty in making robust measurements in the field. Laboratories on the other hand remains the primary means for directly observing spray processes near the surface, and offers promising aspects for improving our understanding by learning these processes in a controlled environment.

There is no standardization on the water type used for these experiments and any potential effects water masses have on the spume generation process is unknown. This adds uncertainty in our ability to make physically realistic spume generation functions that are ultimately applied to the geophysical domain. To address this gap, we have conducted a series of laboratory experiment at the Air-Sea Interaction tank facility (ASIST) of the University of Miami, directly comparing spume concentrations, and surface drag coefficient behavior above fresh and real seawater for 10-m equivalent wind speeds up to 54 m/s. Direct measurements of the near-surface processes were made and directly related to local sources of variance. Droplets in the air above the intensely breaking wind-waves were optically observed and their distribution as functions of wind speed, height, and droplet radius was compared between the two water types. Drag coefficient was calculated using the eddy covariance method on the three-dimensional wind data observed using a sonic anemometer.

Our results show significant differences in the spume generation as well as in the surface drag coefficient behavior for the two water types. Substantially higher concentrations of seawater spume were observed as compared to freshwater across all particle sizes and wind speeds. The seawater particles’ vertical distribution was concentrated near the surface, whereas the freshwater droplets were more uniformly
distributed. Statistical analysis of these findings suggest significant differences in the size- and height-dependent distribution response to increased wind forcing between fresh and seawater. Drag coefficient values for seawater were found less than that of freshwater at all wind speeds suggesting modulation of momentum fluxes in the near surface layer due to the presence of spray droplets.

These findings were generally unexpected and point to an unanticipated role of physiochemical processes in the spume generation mechanism which may impact spray-mediated flux parameterization over water bodies of different salinities. 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 the Mehtas’ (Sofia, Ashley, Savita, and Sakesh)
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Chapter 1

Overview

1.1 Opening Remarks

Under high wind conditions, small liquid water droplets ejected from the water surface following the breaking of surface waves get disseminated into the near surface boundary layer. There they interact with the ambient environment and play an important role in exchanging momentum, heat, and moisture over a wide range of spatial and temporal scales (Andreas et al. 2008; Jeong et al. 2012; Veron 2015). Therefore, it may be necessary to explicitly consider these spray mediated effects in order to better forecast the intensity of storms and air-sea flux exchange.

Spray produced by bubbles is driven by the dynamics of the sub-surface bubble population, which has been observed to differ significantly between fresh and saline water masses (Monahan and Zietlow 1969; Hanes and Johnson 1995; Slauenwhite and Johnson 1999). However, it remains unclear whether or not the physio-chemical differences between the two media are important for more mechanically driven processes such as spume (droplet radius > 25 µm) production.

The transfer of momentum, mass, and sensible heat across the air-sea interface plays an important role in numerous geophysical phenomena like tropical cyclone development, Upper Ocean mixing, and wave growth. Hurricanes are fueled by the transfer of moist enthalpy from the ocean through evaporation and convection, and they lose energy through the frictional drag of the atmosphere on the ocean surface roughness. Air-sea interfacial turbulent fluxes are especially important in the formation and
development of hurricanes. Hurricane tracks are driven primarily by meteorological phenomena that occur on scales large enough to be well resolved by current operational forecast models. Hurricane intensity changes are in addition strongly affected by the coupling between the atmosphere and ocean. There has been a significant reduction in the mean forecast track error for hurricanes over the last twenty years (NHC, https://www.nhc.noaa.gov/verification/verify5.shtml) while the corresponding intensity error has not decreased appreciably (DeMaria et al. 2014).

Therefore, one of the primary ways to accurately predict the development of TCs is to precisely estimate heat and momentum fluxes across the sea surface beneath the TCs. The transfer coefficients are usually parameterized against wind speed $U_{10}$ at a height of 10m, using the field measurement data. In spite of a long history of field measurements (Friese and Schmitt 1976; Large and Pond 1982; DeCosmo et al. 1996; Pedreros et al. 2003; Drennan et al. 2007; Zhang et al. 2008; Richter and Stern 2014) and laboratory experiments using a wind wave tank (Garratt and Hyson 1975; Hasse et al. 1978; Ocampo-Torres et al. 1994; Haus et al. 2010; Jeong et al. 2012; Komori et al. 2018), a large degree of uncertainty still remains in the transfer coefficients, particularly at extremely high wind speeds.

Here I present results from the laboratory experiments conducted in the state-of-the-art ASIST (Air Sea Interaction Saltwater tank) facility at the University of Miami addressing some of the dynamical aspects of atmosphere-ocean interaction under high wind conditions. This will be divided into various research topics. The first one involves quantifying the differences in spume generation between fresh and sea water. The second
one is the investigation of the effects of sea-spray on the surface drag coefficient behavior. Both of these fit together as a part of the initiative to more fully understand the mechanism by which momentum is transferred across the air-sea interface under high wind conditions and its consequences in the context of processes that occur at the ocean-atmosphere interface.

1.2 Specific Research Objectives

An improved body of laboratory measurements and analysis of air-sea interaction under hurricane wind speeds has strong implications for a number of applications. The aim of this section is to lay out several scientific objectives that I feel are rich enough to form the basis of dissertation caliber research.

1.2.1 To quantify the differences in spume generation between fresh and sea water

Under high-wind conditions, the presence of spume particles in the marine boundary layer may significantly mediate the exchange of heat and momentum at the air-water interface (Andreas et al. 2008; Jeong et al. 2012; Veron 2015). While spray dynamics have been the focus of many investigations in the ocean, spray production in freshwater bodies has been far less studied and remains poorly understood. The results of this unique study will demonstrate how salinity plays an important role in the spume generation at high wind, which ultimately holds implications for characterizing spray-mediated air-water flux over varying water masses. Mehta et al. (2019) represents an important step towards completing this objective.
1.2.2 Investigation of the effects of sea-spray on the drag coefficient behavior in high winds.

Tropical cyclones (henceforth TCs) develop from the balance between the supply of thermal energy from the ocean and the loss of kinetic energy caused by the drag acting on the sea surface. Their intensities are very sensitive to the small-scale physical processes such as air–sea momentum and heat transfer. Therefore, one of the primary ways to accurately predict the development of TCs is to precisely estimate the heat and momentum fluxes across the sea surface beneath the TCs (Donelan 2004; Komori et al. 2018). I will focus here on the momentum transfer across the air-sea interface by exploring the behavior of surface drag coefficient ($C_D$) at extreme high winds. To investigate the effects of sea spray on $C_D$, experiments will be conducted in both fresh and sea water. Since we already know that sea water produces considerably more spume droplets than fresh water at high winds Mehta et al. (2019), would this presence of enhanced spume droplet concentration translate to differences in the behavior of $C_D$?

1.3 Structure of this Dissertation

This dissertation is structured around two chapters, each dealing with a different aspect for investigating air-sea exchanges. Each chapter is focused on addressing one of the research objectives posed in Section 1.2. Each chapter contain the relevant introductory and methodological material for understanding how the experiment was conducted, the data acquisition and processing, how results were acquired, interpreted, and their significance.
Each chapter will also end with a final conclusion 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

Spume Generation under High Winds for Fresh and Seawater

Re-statement of Hypothesis: Large droplet spray (Spume) generation which is important for spray-mediated fluxes in high winds is a purely mechanical process and is independent of the physiochemical properties of the water.

The focus of this work will be on laboratory observations of large spray droplets (radius >25 µm), typically referred to as spume above fresh and seawater. This hypothesis aims to address the following questions:

- What is the size-dependent, and vertical distribution of spume droplets under high winds for fresh and seawater?
- Are there any significant differences between the distributions among two water types?
- If yes, what is the underlying physical mechanisms for these differences?
- What are the implications of these results?
2.1 Introduction

2.1.1 Motivation

Sea spray, or ocean spray, comprises liquid droplets are ejected from the sea surface generally due to breaking waves and related phenomena, such as bubble bursting in whitecaps. The ejected droplets contain seawater as well as biological and chemical constituents from non-marine sources. The spray composition is usually considered to be similar to that of bulk seawater, although water vapor exchange with the atmosphere and chemical reactions subsequent to spray formation will influence and change the composition. Once separated from the surface, these droplets transported and dispersed within the atmospheric surface layer (ASL) for fractions of a second to weeks. There they interact with the ambient environment and mediate the air-sea fluxes of sensible and latent heat over a wide range of spatial and temporal scales (Andreas 1992).

The smallest spray droplets (radii in nanometers) can have chemical compositions that are significantly different from that of bulk seawater because their production mechanism might accumulate hydrophobic matter or surface-active material. Small marine aerosols scatter short-wave radiation, decreasing the amount of solar radiation reaching the ocean’s surface by $O(1–5) \, \text{W/m}^2$ (Lewis & Schwartz 2004), or absorb long-wave radiation. Sea salt particles represent approximately 90% of the aerosols in the marine boundary layer, almost half of the total natural aerosol flux, and more than one-third of the global total flux (Seinfeld & Pandis 1998). Estimates of the global emission of sea salt are on the order of $10^{12}–10^{14} \, \text{kg per year}$ (Textor et al. 2006). Consequently, there has been renewed interest in the past decade or so in the smallest sea spray droplets and their contribution to global and regional climates.
Comparatively, our understanding of the role of the larger sea spray droplets is not significantly better. The larger droplets remain suspended in the atmosphere from fractions of a second to several minutes and typically settle back into the ocean under the effect of gravity. In fact, these droplets transfer heat and momentum with the atmosphere through direct exchange (Andreas 1992, 2004; Edson & Fairall 1994; Fairall et al. 1994; Andreas et al. 2008). The magnitude of these exchanges is still under debate, but it is presently thought that the sensible and latent fluxes directly attributed to the sea spray may be critical in the development of large tropical storm systems (Andreas & Emanuel 2001, Andreas 2011, Bao et al. 2011, and Bianco et al. 2011). However, large uncertainties linger because these spray-mediated heat fluxes depend strongly on the rate of sea spray droplet formation at the surface. For large spray drops, the formation process is poorly understood; therefore, the formation rate of large sea spray droplets is not adequately described. Spume droplets are larger spray droplets and have radii typically between 25 µm to several millimeters. These drops are generally assumed to be produced by the mechanical disruption of wave crests by the tangential wind stress. Large numbers of spume droplets are generated under very high wind speed conditions (e.g. storms and tropical cyclones). These spume droplets have been shown to significantly mediate the interfacial fluxes of heat, moisture (Andreas 1994), and momentum (Andreas 2004) across the air-sea interface. 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).

Several field and laboratory studies have shown that the aerodynamic drag coefficient tends to saturate as the 10-m, neutral wind speed approaches 30 to 40 m/s
(Powell et al. 2003; Donelan et al. 2004; Potter et al. 2015). This regime coincides with the initiation of intense wave breaking and significant spume generation, which leads to the development of a distinct spray layer within the ASL. The spray acts as an important thermodynamic layer that can mediate the balance between moist enthalpy input and wind energy dissipation at the air-sea interface (Haus et al. 2010). These processes have been the focus of much investigation in the marine environment (Andreas et al. 2015; Monahan et al. 2017).

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. In a recent laboratory experiment, Jeong et al. (2012) demonstrated that the unmeasured spray-mediated fluxes may account for up to a 38% increase in the overall moist enthalpy transfer coefficients at high winds. However, large uncertainties remain in our understanding of the spume generation and distribution within the ASL simply due to the challenges of making accurate measurements in such extreme conditions.

2.1.2 Mechanism of spray generation

In general terms, spray droplets are known to form via two main pathways. The first one occurs when bubbles, previously entrained in the water column by breaking waves, rise to the surface and burst. This bursting event is a very energetic phenomenon that in turn creates film and jet drops through two different processes. Film drops are generated when the bubble bursts and the surface film shatters; the receding rims become unstable and eject O (10–100) small droplets. Jet drops are generated when the bubble cavity collapses
and shoots up a central jet that also becomes unstable and breaks up into several daughter droplets. The second pathway generates so-called spume drops and occurs when the wind shear at the surface is sufficiently high for water droplets to be literally torn off the surface waves. Figure 2.1 shows the two main spray generation pathways and mechanisms originally described in the work of Andreas et al. (1995),

![Figure 2.1 Mechanisms for sea spray generation. Adapted from Veron (2015).](image)

In the first pathway, droplets are generated by bubble bursting at the surface. Following surface wave breaking, a significant amount of air gets entrained into the rising water column (Thorpe 1992; Melville 1996). This large fraction of the total amount of air entrained, the void fraction, makes back to the surface layer in the form of bubbles. A typical breaking wave crest in the ocean is thus followed by a turbulent wake capped with a foamy layer, the whitecap (Monahan 1971). Blanchard (1963) showed through his work
that two distinct types of droplets form when the bubbles in whitecaps burst. This bursting event is a very energetic phenomenon that in turn creates film and jet drops through two different processes. Field measurements of these jet and film droplets generation in the surface layer is very challenging and difficult. The present literature have necessarily been obtained in the laboratory studies (Wu 1973; Resch et al 1986; Fuentes et al. 2010).

In film drop generation, when bubble reaches the ocean surface, its upper surface protrudes from the interface. This outwardly convex bubble surface film drains, thins and eventually shatters. The receding rims become unstable and eject anywhere from a few to a few hundred film droplets. Overall, film drops have been reported with radii from 10 nm (Martensson et al. 2003, Sellegr et al. 2006) to several 100 μm (Afei & Resch 1990), but most are less than 1 μm in radius (Gong et al. 1997, de Leeuw et al. 2011). The velocity of ejection in the film drops can reach up to 30 m/s because of the large accelerations involved in the rupture of the film cap (Spiel 1998).

The second category of bubble-generated sea spray are jet droplets. Once the bubble cap has shattered, the remaining cavity collapses violently, and a vertical jet forms (MacIntyre 1972). The collapse of the bubble is a rapid and energetic process that leads to a rapid acceleration of the jet. The resulting drop ejection speeds can reach 8 m/s for the smallest top drop to 0.3 m/s for the largest of secondary drops (Spiel 1995, 1997). This results in ejection heights on the order of 10–20 cm for large jet drops (Blanchard 1989). Jet drops are generated usually one to six in number (Resch and Afeti, 1991), following the collapse of each bubble cavity collapses which then shoots up a central jet that also becomes unstable and breaks up into several daughter droplets (MacIntyre 1972;
This column, or Rayleigh jet, results from the rebound, or overshoot, of the sea surface caused by its surface tension. Though these droplets can have radii in the range 2-200 µm, however droplets typically found to be dominating the spray droplet flux spectrum lies in the 3-20 µm radius range.

In contrast to the above two kinds of indirectly produced droplets, in spume generation wave breaking can directly produce other droplets without the mediation of bubbles. Spume droplets are formed when the wind stress is sufficiently large to tear off water from the crest of the waves (Figure 2.1). After being ejected from the breaking wave crest, the droplets are then sweep downwind of the crest. Spume droplets result from the mechanical tearing of the sharpened wave crests by the wind (Monahan 1986). Less is known about the spume drop generation mechanism than the generation of film and jet drops. On formation, spume droplets typically have radii larger than 25 µm. The wind speed threshold for spume generation is 7-11 m/s (Monahan et al. 1983; Andreas et al. 1995; Andreas 2002). Along with the wind speed, the actual threshold depends on other physical parameters like water temperature, wave field, and the turbulence intensity in the near-surface air.

Spume generation was first observed by Koga (1981). He detected drops as large as \( r_0 = 400 \mu m \) and measured their velocity in the airflow. Subsequent studies later showed that spume droplets can have radii in the range \( r_0 = 10-500 \mu m \) with a peak centered at \( r_0 = 100 \mu m \) (Smith et al. 1993; Wu 1993; Fairall et al. 1994; Andreas et al. 2010). Bigger size spume drops were observed in the last decade by numerous laboratory experiments. Drops up to \( r_0 = 700 \mu m \) were measured by Fairall et al. (2009). Veron et al. (2012) measured a significant number of spume drops with radii as large as \( r_0 = 2 \) mm in
the laboratory. Ortiz-Suslow et al. (2016) reported spume droplets up to 1.4 mm in the same experimental facility as this current study. Spume droplets with radii as large as $r_0 = 6$ mm have also been observed in the laboratory (Anguelova et al. 1999). These extremely large spume drops are likely to fall back within a fraction of a second, except in the most extreme winds. Whereas it was previously thought that spume drops larger than approximately $r_0 = 500\mu m$ would have no appreciable impact on the heat, moisture, and momentum flux (and certainly would not take part in the aerosol cycle), their influence on the air-sea momentum balance could be substantial if they happen to be generated in significant numbers.

Several studies in the past have suggested detailed explanations for spume droplet formation mechanism. Mueller & Veron (2009) postulated that drops were generated from the turbulent fragmentation of large globules resulting from a transverse instability of a breaking wave crest (Marmottant & Villermaux 2004). Veron et al. (2012) partially confirmed this phenomenon but also revealed that the bag breakup (Villermaux 2007) of small lenticular canopies of water inflated by the wind can be a potential spume droplet generation mechanism. Soloviev et al. (2017) showed through their numerical experiments that microscale wave breaking caused disruption of the air-sea interface due to the presence of Kelvin-Helmoltz (KH) instabilities. At high winds, these instabilities causes formation of foam and spume droplets covering most of the sea surface. Another laboratory study by Troitskaya et al. (2018) revealed that the generation of spume droplets near the wave crest is caused by several local phenomena. Statistically, bag breakup was found to be the dominant mechanism for spume droplet generation at high winds, when the friction velocity exceeds 1 m/s.
Figure 2.2 Spume generation mechanism reported by Troitskaya et al. (2018). The formation and rupture of a bag in different views. A single bag: a) Side view, b) top view. A multi-bag: c) side view, d) top view. $U_{10} = 25$ m/s. Figure adapted from Kandaurov et al. 2019.
During a bag breakup event, an increase in the small-scale elevation of the water surface results in the formation of a kind of small sail. This sail is then inflated into a canopy bordered by a thicker rim which then ruptures to produce spray (figure 2.2). Other alternative mechanisms beside bubble bursting and the tearing off of spume drops from the ocean surface also known to exist, even though they are believed to be much less efficient. When a plunging wave impinges on the surface and bounces back in the form of a splash-up (Kiger & Duncan 2012). This idea of so-called splash drops was first postulated by Andreas 2002. In the absence of wind, these splash drops are likely to fall back into the ocean very quickly unlike the drops generated by the free fall of a plunging crest.

However in the presence of wind, they could be further fragmented into daughter drops and transported downwind in the near surface layer. Although these drops might be roughly included in the spume population, it might be relevant to discriminate among the different generation mechanisms in the hopes of accurately parameterizing drop ejection sizes and velocities based on wave parameters (Mueller & Veron 2009). Also, spume droplets are created at the high wind crest of waves, while splash droplets are created in the sheltered lee of the crest (at lower height). Further studies are needed to develop physically based parameterizations of spray generation in order to differentiate between large radii drops that are ejected from the surface, be it by splash or direct tearing, and the daughter droplets resulting from the breakup of the initial drops (Pilch & Erdman 1987).

2.1.3 Sea spray fluxes

To investigate the effects of sea spray, it is crucial to consider their effects on both thermal and momentum flux. Below, a detailed description of the thermal and momentum
effects of sea spray droplets is provided. There are a number of alternative explanations for the potential role of spray in altering the air-sea flux exchange, which are given below as:

1. The return of spray droplets to the water surface suppresses the short waves that carry much of the stress (Andreas 2004). 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 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.

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). 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 40 m/s, with a maximum value at 42 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.

3. The onset of Kelvin-Helmholtz 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). The microstructure of the air-water interface under hurricane force wind resembles Kelvin-Helmholtz shear instability between fluids with a large density difference, resulting in suppression of short gravity-capillary waves in the two-phase environment thus altering the aerodynamic properties of the sea surface. The unified wave-form and two-phase parameterization model shows the well-known increase of the drag coefficient ($C_d$) with wind speed, up to 30 m/s. Around 60 m/s, the new parameterization predicts a local peak of $C_k/C_d$, under constant enthalpy exchange coefficient $C_k$. This peak may explain rapid intensification of some storms to major tropical cyclones and the previously reported local peak of lifetime maximum intensity (bimodal distribution) in the best-track records. The bimodal distribution of
maximum lifetime intensity, however, can also be explained by environmental parameters of tropical cyclones alone.

Figure 2.3 Schematic illustration of the spray evaporation layer showing the interfacial, total, and spray-mediated heat fluxes.

Thermal effects of sea spray: When sea spray droplets are lofted into air from relatively warmer ocean surface compared to air, they can exchange both heat and moisture. Also, as the sea spray are saline, when evaporated, they would either result in saline crystals or, as suggested by Andreas (1995), attain a temperature and radius at which they are in a quasi-equilibrium state with their environment.
Andreas and DeCosmo (1999) suggested that at wind speeds greater than 5 m/s, there exist a droplet evaporation layer within the vicinity of the ocean surface; extending from the ocean surface to one significant wave height. Within this layer, the thermal fluxes can be separated into interfacial and sea-spray-mediated fluxes (figure 2.3). Here the interfacial fluxes are the thermal fluxes with no sea spray influence. At the top of the layer, the total fluxes would be the combination of sea spray-mediated fluxes and the thermal fluxes from the ocean surface. Also, the majority of sea spray droplets lofted in the droplet evaporation layer would fall back to the ocean, unless they are fully absorbed or carried further aloft by the turbulent eddies, where they can act as cloud condensation nuclei. 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 (Wu 1979; Andreas and Decosmo 1999, 2002; Andreas et al. 2008; Emanuel 2003) that the spray contribution becomes important at wind speeds above 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.
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 m to 1 mm). Droplets with radii less than about 25 µm are likely produced by bubble bursting (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 (Marmottant and Villermaux 2004; Soloviev et al. 2014).

### 2.1.4 Spray generation dependence on water properties

Spray generation by bubble bursting is driven by the dynamics of the sub-surface bubble population and is highly sensitive to the differing physiochemical properties (like surface tension, viscosity, density etc.) in fresh and seawater (Monahan 1966, 1967). The size distribution and number of bubbles produced via bubble bursting have been observed to differ significantly between fresh and seawater as demonstrated by several experimental studies. Monahan (1966) found that the size spectra of fresh-water and salt-water bubbles produced by the same mechanical mixing process were markedly different. His observation led to the suggestion that there might be observable differences between the whitecaps that form on fresh-water bodies and the whitecaps that form on the sea surface. Monahan & Zietlow (1969) confirmed this through their laboratory experiment. They found relatively more numerous bubbles with radii below 500 micron for sea water.
than fresh water. Their results showed that salt water whitecap areas decays almost exponentially with a time constant of 3.85 s as compared to 2.54 s for fresh water. Hence they concluded that the whitecap lifetimes should be greater on the oceans than on the lakes.

In further experiments conducted by Haines and Johnson (1995), significant differences between bubbles of sizes up to 100 µm in sea and freshwater populations was observed. They used a photographic method to investigate the size spectrum of bubbles in the plume generated by falling water. Bubbles were more numerous, smaller, and resided longer in the seawater plume than in freshwater. In another laboratory experiment by Slauenwhite and Johnson (1999), effects of breaking waves on the bubbles populations in fresh and seawater was explored. Air bubbles were expanded through an orifice at constant pressure drop, shattering them into clouds of smaller bubbles. Bubble from breaking waves in seawater were found to break up in up to 4-5 times more bubbles as compared to freshwater. The number of bubble produced in shattering were found to a function of salt concentration along with very sensitive to the types of ions present. Effects of adding an external solvent and changing temperature were also explored in their study. Addition of medium (marine diatom Phaeodactylum tricornutum) was found to further increase the production by 200 %, and a drop in temperature from 30 to 20 degrees, increased the bubble production in sea water by nearly 50 %. The authors argued that these effects are separate from coalescence inhibition which is known to be fundamental reason for differences in bubble production in fresh and sea water. Exact reasons remains unexplored and present an active area of research.
In the past, the observed differences in bubble populations in seawater and freshwater were attributed to coalescence inhibition in seawater, but more recent studies have suggested that other factors like ionic strength, surface tension, structure making/breaking ability, viscosity, or density may be important as shown in the Table 1 below. Seawater is approximately 2.5% more dense, 7% more viscous, and possesses 2% higher surface tension than freshwater, which may explain differences in bubble populations produced by breaking waves in the two media.

Table 2.1: Differences between the bulk physical properties of freshwater and seawater.

<table>
<thead>
<tr>
<th>Property</th>
<th>Seawater (35 PSU)</th>
<th>Freshwater</th>
</tr>
</thead>
<tbody>
<tr>
<td>Density, g cm⁻³, 25°C</td>
<td>1.02412</td>
<td>1.0029</td>
</tr>
<tr>
<td>Specific conductivity, ohm-1 cm⁻¹, 25°C</td>
<td>0.0532</td>
<td>0.00005</td>
</tr>
<tr>
<td>Viscosity, millipoise, 25°C</td>
<td>9.02</td>
<td>8.90</td>
</tr>
<tr>
<td>Isothermal compressibility, vol./atm, 0°C</td>
<td>46.4 x 10⁻⁶</td>
<td>50.3 x 10⁻⁶</td>
</tr>
<tr>
<td>Temperature of max density, 0°C</td>
<td>-3.25</td>
<td>3.98</td>
</tr>
<tr>
<td>Freezing point, 0°C</td>
<td>-1.91</td>
<td>0.00</td>
</tr>
<tr>
<td>Surface tension, dyne cm⁻¹, 25°C</td>
<td>72.74</td>
<td>71.97</td>
</tr>
<tr>
<td>Velocity of sound, ms⁻¹, 0°C</td>
<td>1450</td>
<td>1407</td>
</tr>
<tr>
<td>Specific heat, J g⁻¹ oC⁻¹, 17.5°C</td>
<td>3.898</td>
<td>4.182</td>
</tr>
</tbody>
</table>

Unlike spray formation through bubble bursting, spume generation differences between water types are way less known. In the present literature only two studies exists discussing spume production in a laboratory setting for different types. Lai & Shemdin (1982) were the first one to report some results regarding the effects of wind speed and wave height on the vertical distribution of spray over freshwater and saltwater using the wind & wave facility at the University of Florida. Their aim was to improve understanding of the spray generation mechanisms and to obtain reliable information for the total production rate and size spectral distribution of droplets at different winds speeds ($U_{10}=10-18$ m/s) and elevation above the mean water level. A hot-film anemometer was used in a 46 m long wind flume to measure the size and number of
water droplet over the air-water interface under the action of wind waves and wind & swell (mechanically generated) waves with for two water types. They found that the vertical distribution and the drop size distribution of the total horizontal flux of spray droplets can be expressed as a logarithmic distribution and negative power law of drop diameter respectively. Their showed that spectral shapes for the two water types differ with the freshwater drop size distribution exhibiting a steeper trend than $d^{-2}$ for saltwater. Also the smaller freshwater drops would fall below the equilibrium line. They attributed these differences to a number of factors that differ in the differences i.e. surface tension and the differences in bubble spectra between freshwater and saltwater.

In another experiment by Fairall et al. (2009), some observations of spume ($r_0 = 15$-600 µm) volume spectra were reported in fresh and saline water (salty water with salinity 24 ppt) in their SPANDEX experiment (Spray Production and Dynamics Experiment) conducted at the University of New South Wales Water Research Laboratory in Australia. The goals of SPANDEX were to illuminate physical aspects of spume droplet production and dispersion; verify theoretical simplifications used to estimate the source function from ambient droplet concentration measurements; and examine the relationship between the implied source strength and forcing parameters such as wind speed, surface turbulent stress, and wave properties.

Their observations of droplet profiles gave reasonable confirmation of the basic power law profile relationship that is commonly used to relate droplet concentrations to the surface source strength. They also indicated considerable droplet mass may be present for sizes larger than 1.5 mm diameter. Phase Doppler Anemometry was used to reveal significant mean horizontal and vertical slip velocities closer to the surface. A cloud
imaging probe (CIP) was used to measure spume droplets at three different wind speeds and heights above the mean water level. While they showed some results regarding droplet volume spectra for fresh and salt water, a descriptive and detailed analysis comparing the two water types was lacking in their analysis. Other limiting factors were their use of artificial seawater at a salt concentration much lower than typically found in the ocean, as well as their combination of the effects of wind and mechanically generated waves. Further study is needed to expand on this work.

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 (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. For the purposes of studying
spume generation in models and laboratories, it is generally assumed that this process is
primarily mechanical and that any physiochemical effects are secondary or negligible.
However, as far as the authors are aware, this has never been experimentally verified.

To test this hypothesis, a laboratory experiment was conducted that compares
spume generation in fresh and seawater under hurricane-force wind conditions. To our
knowledge, this is the first quantitative assessment of the varying rates of spume
generation between the two water types using real seawater. The findings of this work
directly test the general assumption that spume generation is controlled primarily by the
mechanical wave-breaking process. This holds significant implications, not only for the
theoretical or laboratory-based study of spume generation, but also for the distinct role of
spume generation in fresh versus seawater.

2.2 Methods

A brief summary of the experimental design and analytical methodology with
specific details relevant to this study has been described in this section.

2.2.1 The laboratory facility

![Diagram of the ASIST facility set-up](image)

Figure 2.4 A schematic of the ASIST facility set-up for this study. Color visualizes the water level, 0.42 m.
The experiment was conducted in the University of Miami’s Air-Sea Interaction Saltwater Tank (ASIST), which is an acrylic wind-wave-current flume with a cross section of 1 x 1 m and length 15 m (Figure 2.4). This facility is capable of generating both wind waves with a single turbine recirculating fan, as well as one-dimensional mechanical waves. In this study, wind was the primary forcing on the system. This experiment did not assess the impact of mechanically-generated waves on spume generation.

2.2.2 Water characteristics

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). The seawater used here was sourced from a nearby tidal inlet (Bear Cut) with a salinity of ~34 PSU, which is fully exposed to Atlantic Ocean water with no nearby freshwater sources. The seawater was pumped into the facility and stored in a reserve tank on top of the building. This allowed suspended particulates to settle out before being pumped through a sand filter and then a 10-micron cartridge filter before pumping into ASIST. The freshwater source was the local municipal water system.

2.2.3 Instrument locations

Spray observations were made 11.05 m downwind of the air inlet. Maximum winds ($U_{10}$) measured by a three-dimensional sonic anemometer positioned 6.60 m downwind of the inlet were 54 m/s. The 10-m referencing was done following earlier ASIST work presented by Haus et al. (2010) and Donelan et al. (2004). An important physical reference in spray observations is the wave height, typically generalized via the significant wave height, $H_s$. The wave heights reported in Table 2.1 in the tank were
measured at the location of the spray observations using a downward looking ultrasonic distance meter (UDM) sampling at 10 Hz. Figure 2.5 shows the time series of the observed water surface elevation in the tank; while Figure 2.6 shows the corresponding spectra from the 5 wind speeds tested stacked together.

Figure 2.5 Time series of water surface elevation (demeaned and detrended) as observed from an ultrasonic distance meter fastened to the roof of the ASIST flume. Wind speed increases from 36-54 m/s top to bottom. Spectra were calculated from these time series (see figure 2.6)
Figure 2.6 1-D power spectral density of the water surface elevations given in figure 2.3, with corresponding significant wave height estimates provided in the legend. Only the frequencies between the vertical lines were used in the calculation of the 0th order moment.

From the spectra, we can see an increase in the 1-D spectral peak and a slight shift to lower frequencies with increasing wind speed, reflecting a more developed wave state within the tank. This trend is not as substantial as might be expected over the ocean because of the limits of the control volume and none of these wave states reach a truly mature condition because of the very strong winds. In calculating the significant wave height, a sub range of frequencies (figure 2.6) was used to avoid contamination from very low frequency variability that may not be caused by direct wind-wave interaction.
Wind-wave development over this fetch of 11.05 m was characteristic for each wind speed. The conditions in the tank are fully reproducible as shown by various studies so far (Donelan et al. 2004; Savelyev et al. 2011; Jeong et al. 2012). It’s the same basin of water filled to the same level, driven by the same fan. It reproduces the wave field to sub-mm scales in terms of the mean or significant wave height.

2.2.4 Camera setup

Spume droplets were imaged using a Dantec Dynamics particle image velocimetry (PIV) acquisition system modified to be used in a shadow imaging mode. A camera (JAI CV-MSCL, 1.9 MP, 30fps) was positioned outside of the tank and oriented to be looking into a high intensity strobe, also mounted outside of the tank, but directly opposite the camera. A wiper was placed on the inside of the tank’s acrylic wall, facing towards the camera to minimize the accumulation of droplets on the acrylic. The image acquisition timing and strobe pulse were controlled and synchronized using the Dantec system. As spray was ejected into the air volume in the tank, the droplets would appear as shadows (or silhouettes) in the camera images. The geometry of the set-up was such that the camera imaged a 55 mm x 75 mm plane in the middle of the tank aligned with the along-tank direction. One sample image from the shadow imaging acquisition system for freshwater, at $U_{10}$=49.5 m/s, centered at 145 mm from the mean water level has been shown in Figure 2.7. A total of five wind trials were conducted in the tank, with $U_{10}$ ranging from 36 to 54 m/s (Table 1.1). For each $U_{10}$ and water type, the camera-strobe system was set-up at two vertical levels relative to the water surface (an additional height was done for freshwater).
Figure 2.7: A sample camera image from the shadow imaging acquisition system for freshwater, at $U_{10}=49.5$ m/s, centered at 145 mm from the mean water level.

These levels, labeled Lower, Middle, and Upper panels in Table 2.2, were centered on different heights above the still water level and arranged such that they created a continuous virtual stack. The imagery from each independent level were stitched together to create the spume droplet concentration profiles extending over the scaled height, from $z/H_s \sim 2$ to 6. Here, $z$ is the height above the MWL in the tank, and $H_s$
is the significant wave height for a particular wind speed (determined using $0^{th}$ moment of the elevation variance spectrum). Using the geometry of the set-up, any overlap between panels was removed prior to analyzing the spume droplet distributions.

### 2.2.5 Data collection protocol

For each level, the wind was allowed to ramp up and time (120 s) was allowed for the tank conditions to become stationary. Then at least 7 consecutive sets of 250 images were acquired; 250 being the computational memory limit. Unfortunately, the ambient water temperature in the tank was not recorded. However, the laboratory was climate controlled and, before using a freshly pumped volume of water, the tank was allowed to acclimatize to room temperature (~23° C). A typical set of 7x250 image acquisitions took approximately 10 minutes to complete, and after each trial the water level was returned to 0.42 m (water loss was due to spray advecting into air outlet).

### 2.2.6 Image calibration and processing

Image processing was done in two steps using the Dantec Dynamics shadow imaging software package, independently carried out on each set of 250 images. 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 image (>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.

In total, over 52,000 images were acquired and processed for this experiment. A pre-experiment calibration was done to determine that the image resolution was 42 µm per pixel. The droplet contouring found the two-dimensional projected surface area of each droplet. Assuming spherical drops, the equivalent radius, \( r_0 \), of observed drops ranged from 80 to just over 1400 µm. The smallest drop that could be measured using this method encompassed an area of ~16 pixels (which equals 28224 µm\(^2\))

2.2.7 Spray droplet identification and sizing

The efficacy of the automatic droplet contouring was tested against visual inspection of every 25\(^{th}\) image of each 250 image set. The success of the processor was defined as the number of properly identified, in-focus droplets over the number of droplets identified via inspection (see Table 2.2). In general, the processor tended to under-count the total number of droplets per image, and there was not a significant difference in the success rate between fresh and seawater. The images from the 45 m/s trial in freshwater conditions were found to be substantially more difficult for the processor to automatically count. In the post-experiment data quality control and assessment, this was visually determined to be due to poor image contrast for this particular trial (wind speed + water type). After taking additional steps to increase the contrast, the results of the visually-checked processing results (Table 2.2) remained anomalously low as compared to all the other data sets. As a result, this wind speed (for
both water types) was removed from the comparative analysis. Simply repeating these experiments was not an option because in the interim between conducting the experiments (ca. 2013) and conducting the analysis that discovered this problem (ca. 2015), the ASIST was moved to a new facility and re-configured in a way that precludes repeating these experiments.

The radius and vertical height dependent droplet concentration profile was determined by discretizing the identified droplet distributions into 50 μm wide radius classes and 3 mm vertical cells. This was done by combining the raw images taken at the lower and upper vertical levels. Therefore, for each radius class and vertical cell, the average spume droplet number concentration for a given wind speed becomes,

$$n(r_i, z_j) = \frac{\text{Count}(r_i, z_j)}{\Delta \text{Vol} \times N_{\text{total}}}$$ (1)

where \(\text{Count}(r_i, z_j)\) is the total number of observed spume droplets in the \(i^{th}\) radius class and \(j^{th}\) vertical bin; \(\Delta \text{Vol}\) is the air volume of each vertical bin; \(N_{\text{total}}\) is the total number of images in an observation period (e.g. 7 x 250 =1750 images); and \(dr\) is the width of each radius class (50 μm). Equation 1 provides the number of particles per unit volume of air per radius increment for each vertical bin along the reconstructed profile. These concentration distributions are given as two-dimensional grids in the figure 2.8.

From these individual number concentration values at each grid point as shown in the figure 2.8, their corresponding mass concentration was then calculated using equation 2 given below. Vertically-integrated mass concentration was obtained by multiplying the corresponding water density, \(\rho\), for the two media, to the volume concentration at each radius class as,

$$m(r_i) = \frac{4}{3} \pi \rho \int r_i^3 n(r_i, z) \, dz$$ (2)
Table 2.2 Summary of the in-tank, mean conditions and image processing evaluation. The 10-m equivalent wind speed ($U_{10}$) and friction velocity ($u_*$), in m/s, are given, along with the significant wave height ($H_s$) in mm. The two right-most columns give the visually-verified percent success rate of the automatic droplet identification and the number of runs. The lower, middle, and upper panels refer to the three image frame heights used to reconstruct the profiles. This table is adapted from Ortiz-Suslow et al. (2016).

<table>
<thead>
<tr>
<th>$U_{10}$</th>
<th>$u_*$</th>
<th>$H_s$</th>
<th>Seawater</th>
<th>Freshwater</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>1.7</td>
<td>29</td>
<td>60 % (7)</td>
<td>61 % (7)</td>
</tr>
<tr>
<td>40.5</td>
<td>2.0</td>
<td>32</td>
<td>75 % (7)</td>
<td>78 % (7)</td>
</tr>
<tr>
<td><strong>Lower</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>45</td>
<td>2.2</td>
<td>35</td>
<td>80 % (7)</td>
<td>67 % (7)</td>
</tr>
<tr>
<td><strong>Panel</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>49.5</td>
<td>2.4</td>
<td>36</td>
<td>82 % (7)</td>
<td>84 % (7)</td>
</tr>
<tr>
<td>54</td>
<td>2.7</td>
<td>38</td>
<td>92 % (7)</td>
<td>91 % (7)</td>
</tr>
<tr>
<td>36</td>
<td>1.7</td>
<td>29</td>
<td>62 % (7)</td>
<td>63 % (7)</td>
</tr>
<tr>
<td>40.5</td>
<td>2.0</td>
<td>32</td>
<td>80 % (9)</td>
<td>80 % (7)</td>
</tr>
<tr>
<td><strong>Middle</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>45</td>
<td>2.2</td>
<td>35</td>
<td>85 % (7)</td>
<td>76 % (7)</td>
</tr>
<tr>
<td><strong>Panel</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>49.5</td>
<td>2.4</td>
<td>36</td>
<td>85 % (9)</td>
<td>84 % (7)</td>
</tr>
<tr>
<td>54</td>
<td>2.7</td>
<td>38</td>
<td>78 % (8)</td>
<td>80 % (7)</td>
</tr>
<tr>
<td>36</td>
<td>1.7</td>
<td>29</td>
<td>-</td>
<td>61 % (7)</td>
</tr>
<tr>
<td>40.5</td>
<td>2.0</td>
<td>32</td>
<td>-</td>
<td>79 % (7)</td>
</tr>
<tr>
<td><strong>Upper</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>45</td>
<td>2.2</td>
<td>35</td>
<td>-</td>
<td>67 % (7)</td>
</tr>
<tr>
<td><strong>Panel</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>49.5</td>
<td>2.4</td>
<td>36</td>
<td>-</td>
<td>83 % (7)</td>
</tr>
<tr>
<td>54</td>
<td>2.7</td>
<td>38</td>
<td>-</td>
<td>87 % (7)</td>
</tr>
</tbody>
</table>
2.3 Results

The results presented in this section are based on spume droplet observations collected using a non-obtrusive, optical method quantifying the droplet distribution in the air mass directly above the water surface in 10-meter equivalent winds from 36 to 54 m/s in both fresh and seawater (herein, FW and SW, respectively). These will be discussed here as overall spume droplet number concentrations \( n(r, z) \), vertically-integrated number concentration, \( n(r) \), Radius-integrated number concentration, \( n(z) \), and statistical analysis of \( n(r) \) and \( n(z) \) with the wind speed.

2.3.1 Overall number concentration, \( n(r, z) \)

A two-dimensional representation of the variability in the overall spume droplet number concentration, \( n(r, z) \) for FW and SW is shown in figure 2.8 For both FW and SW across all droplet sizes, \( n(r, z) \) decreased with height above the interface and increased with \( U_{10} \). However, there were differences between the two. With the \( U_{10} \) increasing from 36 to 54 m/s, \( n(r, z) \) increases more for the SW then FW. This can clearly be seen in the corresponding SW and FW panels in figure 2.8. For example, in FW, the percentage of filled grid space that was observed increased from 16% to 62%; whereas for SW, the grid space increased from 17% to 76% between the lowest and highest measured \( U_{10} \), respectively.

While not a generalizable metric, this filling of the grid space represents the fraction of the imaged air volume where additional spray was observed for each increase in \( U_{10} \). The distribution of droplet number concentration was found to be skewed towards the smaller radii for both water types, indicating the dominance of smaller radii droplets in number spectrum.
This is intuitive, since the largest droplets would be expected to be more rarely imaged due to their low residence time in the air flow, and resistance to vertical diffusion.

Figure 2.8: Two-dimensional distribution of overall observed spume droplet number concentration as a function of observed scaled height and binned droplet radius for FW (left column) and SW (right column) are shown at different wind speeds from top to bottom. The color bar is common across all panels and shows log-scaled (base 10) droplet number concentration (number/cm$^3$/µm). Gray cells represent no droplets counted.

The total spume droplet number concentration $n_{\text{total}}$ (integrated over $r$ and $z$) increased as $U_{10}$ increased. From the lowest to highest $U_{10}$, $n_{\text{total}}$ increased by 30 and 25 times for SW and FW, respectively. Relative to $n_{\text{total}(FW)}$, $n_{\text{total}(SW)}$ was greater by a factor varying from 1.5 to 2.5, depending on the $U_{10}$ (Table 2.3).
Table 2.3 Summary of $n_{\text{total(SW)}}$ and $n_{\text{total(FW)}}$ for all wind speeds.

<table>
<thead>
<tr>
<th>$U_{10}$</th>
<th>$n_{\text{total(SW)}}$</th>
<th>$n_{\text{total(FW)}}$</th>
<th>$\Delta = n_{\text{total(SW-FW)}}$</th>
<th>$R_{sf} = n_{\text{total(SW)}}/n_{\text{total(FW)}}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>6100</td>
<td>4100</td>
<td>2000</td>
<td>1.5</td>
</tr>
<tr>
<td>40.5</td>
<td>25400</td>
<td>16200</td>
<td>9200</td>
<td>1.6</td>
</tr>
<tr>
<td>49.5</td>
<td>126500</td>
<td>51400</td>
<td>75100</td>
<td>2.5</td>
</tr>
<tr>
<td>54</td>
<td>183900</td>
<td>102700</td>
<td>81200</td>
<td>1.8</td>
</tr>
</tbody>
</table>

2.3.2 Vertically-integrated number concentration, $n(r)$

The variability in the vertically-integrated number concentration, $n(r)$, (Figure 2.9a) and mass concentration, $m(r)$, (Figure 2.9b) are shown as a function of the spume droplet radius for different $U_{10}$. Overall, $n(r)$ values for both water types steadily increased as $U_{10}$ increased and decreased as droplet radius increased. The size-dependence of $m(r)$ was more complex and nonlinear than the simple number concentration (Figure 2.9b). The ratio $n_{\text{SW}}/n_{\text{FW}}$ was generally $>1$ for all radius classes except five instances ($r_0$ = 686 μm at $U_{10}$ = 54 m/s; $r_0$ = 936-986 μm, 1236-1286 μm at $U_{10}$ = 36 m/s), which comprise ~5% of the observations (Figure 2.9c). $n_{\text{SW}}/n_{\text{FW}}$ was only ever $<1$ for 36 m/s, except for one radius class at 54 m/s. Apart from this, $n_{\text{SW}}/n_{\text{FW}}$ ranged from 1 to 4.6. While $n_{\text{SW}}/n_{\text{FW}}$ varied with radius, for all but 2 classes, this ratio increased with wind speed up to 49.5 m/s, and only a few classes exhibited this trend monotonically. Interestingly, there was a substantial drop in $n_{\text{SW}}/n_{\text{FW}}$ between 49.5 and 54 m/s for the lower half of the radius spectrum. This became less consistent when $r_0 > 900$ μm. In general, the differences between SW and FW increased with wind speed.
Figure 2.9: Vertically-integrated number concentration (number/m$^3$/µm) spectra (a), vertically-integrated mass concentration (kg/m$^3$/µm) spectra (b), the distributions of $n_{\text{SW}}/n_{\text{FW}}$ plotted against droplet radius, $r_0$, for each wind speed.
Figure 2.10: Total vertically integrated spume droplet mass concentration (kg/m^3/µm) for FW and SW plotted against the radius, r_0. The solid colored lines and dashed lines represent the best fitted slope lines for the data points. FW data has been shown in solid circles, and SW in diamonds with the corresponding U_{10} in m/s (same as figure 2.9).

For a given wind speed, m(r) increased rapidly with increasing particle radius, but at about 400 µm, all m(r) exhibited a leveling off, with the higher winds showing an apparent decrease in mass concentration with radius for r > 900 µm. In order to explore this behavior, m(r) was segregated into three radius subranges (Figure 2.10). For all wind speeds and radius classes, we observed higher values for m(r)_{sw} than m(r)_{fw}. In the first droplet size subrange (r_0 = 86-400 µm), both m(r)_{sw} and m(r)_{fw} increased with radius and U_{10}. The mean increase in m(r) was a factor of 22 and 27 for FW and SW respectively, as U_{10} increased to 54 m/s from 36 m/s. For all U_{10}, more m(r) for SW was observed than m(r) for FW. The differences between m(r)_{sw} and m(r)_{fw} increased with increasing U_{10} for this size range. For the lowest two wind speeds, the m(r)_{sw} increased more rapidly than m(r)_{fw} per incremental radius change, but this was not found for the highest two U_{10}.
(Figure 2.10). This suggests that SW droplets were more readily generated at lower wind speeds when compared to FW over this subrange.

In the second subrange \((r_0 = 400-900 \, \mu m)\), \(m(r)\) tended to be independent of droplet radius (i.e. best fine line exhibited negligible slope) for both water type. This subrange appears to be a transition region for the \(m(r)\) behavior with the increase in droplet radius size. In the third subrange \((r_0 = 900-1400 \, \mu m)\), the \(m(r)\) for both water types decreased with the increasing droplet radius for all wind speeds. This decrease was more pronounced in FW as compared to SW at the lowest two wind speeds.

2.3.3 Radius-integrated number concentration, \(n(z)\)

Profiles of radius-integrated number concentration, \(n(z)\) were analyzed to characterize the vertical distribution of FW and SW spume droplets (Figure 2.11). These profiles come directly from integrating the spume droplet number concentrations given in Figure 2.8 at each vertical bin. Overall, across all \(U_{10}\), more spume droplets were observed along each scaled height for SW then FW.

The majority of the differences in number concentrations between the water types occurred in the lower portion of the profile for all wind speeds. For the lowest two wind speeds, \(n(z)\) tends to converge at the top of the profile, and the majority of the difference across SW and FW was near the surface. Between 49.5 m/s and 54 m/s, this convergence was not observed, but this may have been an effect of a fixed observation height \(z\) with a changing wave height \(H_s\) as \(U_{10}\) increases. So, the scaled region where convergence was observed \((z/H_s >5)\) was not reached at higher winds. Furthermore, it may be that the spray layer widened (in the vertical) with increased forcing. Close inspection of the profiles of
$n(z)$ under these higher winds indicates the profiles starting to come together (Figure 2.11), but this was not fully resolved in the present experimental set-up.

![Figure 2.11: Vertical profiles of (top) radius-integrated number concentration (number/m$^3$) (bottom) radius-integrated mass concentration (kg/m$^3$) plotted against scaled height. These profiles represent the total spume droplets observed across all radius classes. The legend represents different wind speeds with colored solid circles for FW, and diamonds for SW respectively. The profile height is scaled by the appropriate $H_s$.](image)

### 2.3.4 Statistical analysis of $n(r)$ and $n(z)$ with wind speed

Regression analysis was performed to derive empirical relationships for $n(r)$ and $n(z)$, which are presented in Figures 2.9a and 2.11, respectively. For both of these analyses, the $U_{10}$-dependence was also considered. This is an important exercise to
formalizing the qualitative findings of these experiments as described above. These analyses build on a previous study conducted using just the SW data from these experiments (Ortiz-Suslow et al. 2016); furthermore, this statistical analysis provides a means of more directly and quantitatively comparing the findings from FW and SW.

The first empirical relation was functionally represented as,

\[ N_s = f(r_0, U_{10}), \]  

(3a)

where \( N_s \) is the modeled vertically-integrated droplet number concentration (i.e. \( n(r) \)) for both FW and SW (Figure 2.9a). \( n(r) \) were tested against three empirical relationships: a power law, an exponential, and a linear fit. The model that minimized the root mean squared error (RMSE) between the observed \( n(r) \) and modeled \( N_s \) was chosen as best representing the observed values, for a given \( U_{10} \). For both FW and SW, a power law performed best for all winds. The model correlation for the power law was very strong \( (R^2 \sim 0.99, p\text{-values} < 0.001 \text{ for all fits}) \) and substantially better than the two other relationships used in this test \((R^2 \sim 0.85 \text{ and } \sim 0.93 \text{ for the linear and exponential, respectively})\); furthermore, the RMSE for the power law was substantially higher than for the other model fits. Based on these results, we can define \( N_s \) as:

\[ N_s = a r_0^b \]  

(3b)

Coefficients \( a \) and \( b \) are the empirically-derived coefficients. Initially, this analysis was performed over the entire radius spectrum, but it was found that \( n(r) \) exhibits a nonlinear (in log-log space) transition in spectral slope at radii around 500 \( \mu m \). To better resolve the variability in \( n(r) \), we performed the regression over two subranges, for \( r_0 < 500 \mu m \) and for \( r_0 > 500 \mu m \), deriving coefficients two sets of \( a \) and \( b \) for these subranges independently. The RMSE-based selection criteria described above was
applied to the subrange regressions as well with the same results. The results of this analysis are provided in Table 2.4 and the regression results as compared to the observed \( n(r) \) for SW are provided in Figure 6.

**Figure 2.12:** (a) Results of performing least-squares, log-scaled linear regression on the SW \( n(r) \) over the two radius subranges defined about 500 \( \mu \text{m} \), the coefficients of the regressions are provided in Table 3. The scaled residuals for the \(< 500 \mu \text{m} \) and \( > 500 \mu \text{m} \) subranges are given in (b) and (c), respectively. \( N_s \) denotes the predicted size-dependent number concentration spectrum (Eqn. 3b).

For the smaller radii subrange, \( b \) (i.e. the spectral slope) for FW and SW was between approximately -2 and -1.5, but exhibited contrasting dependence on \( U_{10} \). For
FW, $b$ monotonically increased from -1.95 to -1.79, whereas for SW, $b$ decreased from -1.60 to -1.89 and the slopes for the two highest wind speeds (SW only) were effectively the same. For both FW and SW, $a$ increased with $U_{10}$, and while SW > FW, both coefficients for each water type were of a similar order of magnitude. For the larger radii ($r_0 > 500 \mu m$), unlike in the smaller droplet subrange, the spectral slopes ($b$) for both FW and SW were negatively dependent on $U_{10}$ and the transition for the former was much more substantial (-3 to -4.5). In general, we found that the there was a significant change in both $a$ and $b$ from the lowest to highest wind speeds, across both water types, and that within each water type, the differences in $a$ and $b$ become relatively small to negligible for 49.5 and 54 m/s.

Table 2.4: Regression coefficients for SW and FW to the empirical number concentration profile given in Eq. 3b. The number of samples included in the regressions for $r_0 < 500$ (> 500) $\mu m$ was 9 (17). The values indicated in the bracket below corresponds to the ± 95% confidence interval bounds.

$$
\begin{array}{cccccccc}
U_{10} & r_0 < 500 \mu m & a & -b & r_0 > 500 \mu m & a & -b \\
\hline
FW & SW & FW & SW & FW & SW & FW & SW \\
36 & 8.9e+06 & 3.7e+06 & 1.94 & 1.60 & 1.3e+10 & 5.6e+10 & 3.05 & 3.25 \\
 & (0.9e+06) & (0.7e+06) & (0.17) & (0.21) & (0.5e+10) & (0.4e+10) & (0.44) & (0.62) \\
40.5 & 4.9e+07 & 3.5e+07 & 1.92 & 1.76 & 1.9e+10 & 3.6e+12 & 2.93 & 3.64 \\
 & (0.2e+07) & (0.3e+07) & (0.25) & (0.08) & (0.4e+10) & (0.3e+12) & (0.30) & (0.41) \\
49.5 & 1.4e+08 & 3.3e+08 & 1.90 & 1.89 & 2.3e+15 & 4.0e+13 & 4.52 & 3.78 \\
 & (0.7e+08) & (0.1e+08) & (0.10) & (0.07) & (0.5e+15) & (0.3e+13) & (0.32) & (0.32) \\
54 & 1.5e+08 & 4.8e+08 & 1.79 & 1.88 & 2.3e+15 & 3.7e+13 & 4.42 & 3.72 \\
 & (0.5e+08) & (0.6e+08) & (0.08) & (0.07) & (0.2e+15) & (0.4e+13) & (0.50) & (0.34) \\
\end{array}
$$
The second empirical relationship for \( n(z) \) takes the functional form:

\[
N_v = f\left(\frac{z}{H_s}, U_{10}\right),
\]

where \( N_v \) is the modeled radius-integrated number concentration. Using a similar technique as above, a logarithmic fit for SW and a 2\textsuperscript{nd} order polynomial fit for FW was found to best represent \( n(z) \) in the two respective water types. The figure showing fitness of the fits has been shown below (figure 2.13).

Figure 2.13: Results of performing least-squares, linear regression on the vertical number concentration profiles for SW \( n(z) \) top (eqn. 4b), and FW \( n(z) \) (eqn. 4c) bottom for four wind speeds.
These relationships can be expressed as:

\[ N_{v,sw} = \frac{\log(z/H_s) - \log(c)}{d} \]  

(4b)

\[ N_{v,fw} = p_1(z/H_s)^2 + p_2(z/H_s) + p_3, \]  

(4c)

Where \( N_v \) (sw) and \( N_v \) (fw) are the predicted radius-integrated droplet number concentration profiles for SW and FW, respectively. In order to confirm that two different functions are appropriate for FW and SW, respectively, the RMSE derived from applying both the logarithmic and polynomial expressions (4b and 4c) to both FW and SW are given in Table 2.5. For all tests, the coefficients of determination were strong (\( R^2 \sim 0.96-0.99 \)), but there is a clear difference in the resulting RMSE. While this is purely an empirical exercise, this analysis highlights that there are significant differences in the vertical distribution of spume above FW and SW.

Table 2.5: Comparison of RMSE for logarithmic and power functions for \( N_v \) for FW and SW. For each regression, the total sample size was 111 vertical cells.

<table>
<thead>
<tr>
<th>( U_{10} ) (m/s)</th>
<th>SW (Logarithmic)</th>
<th>SW(Polynomial)</th>
<th>FW (Polynomial)</th>
<th>FW (Logarithmic)</th>
</tr>
</thead>
<tbody>
<tr>
<td>36</td>
<td>0.08</td>
<td>0.20</td>
<td>3.33</td>
<td>6.24</td>
</tr>
<tr>
<td>40.5</td>
<td>0.05</td>
<td>0.21</td>
<td>1.10</td>
<td>4.23</td>
</tr>
<tr>
<td>49.5</td>
<td>0.06</td>
<td>0.19</td>
<td>1.07</td>
<td>3.17</td>
</tr>
<tr>
<td>54</td>
<td>0.08</td>
<td>0.15</td>
<td>0.96</td>
<td>2.14</td>
</tr>
</tbody>
</table>

Empirically-derived coefficients for FW were in the range as bounded by the 95% confidence interval as \( p_1 (5.13e-05 \pm 5.76e-05) \), \( p_2 (-1.59e-04 \pm 1.81e-04) \), and \( p_3 (3.57e-04 \pm 3.71e-04) \) and for seawater were as \( c (3.67 \pm 0.44) \) and \( d (-0.21 \pm 0.003) \). With the
increasing $U_{10}$, coefficients $p_1$, $p_3$, and $d$ increases while $p_2$ and $c$ decreases. Noting the disparate amount of uncertainty in the $p_1$, $p_2$, $p_3$ versus $c$ and $d$, would suggest that the complexity in $n(z)$ (Figure 5) for FW makes regression over the full profile challenging.

2.4 Discussion

The aim of this study was to investigate whether water type impacts the generation of spume droplets, which has been generally assumed to be the result of the purely mechanical wave breaking process (as opposed to bubble bursting). The results of this novel experiment demonstrated that there are substantial differences in the amount and vertical distribution of spume produced in FW and SW. This was done by keeping all testing conditions and parameters constant except for the water type and thus concluding that the water type has an impact on the radius and height dependence of the developing spray layer with increasing wind speed.

2.4.1 Empirical relationships

The statistical analysis of $n(r)$ revealed that there is a strong difference in the size-dependent concentration spectrum (for all wind speeds and both water types) for the smaller droplet ranges, relative to the larger drops. Hence, there clear difference in spectral slope for radii < 500 $\mu m$ versus >500 $\mu m$. It was interesting to note that the spectral dependence with wind speed varied across these subranges between FW and SW. While there are only four wind trials to compare, the robustness of the empirical analysis (i.e. fairly low uncertainty in $a$ and $b$) is encouraging to this reflecting a physical change in the rate of spume production and the characteristics of the size-dependent distribution of entrained spray above the breaking waves.
The analysis into $n(z)$ provided mixed results, in empirical form, between FW and SW. While this does highlight that the vertical distribution of spume differs significantly across water types there are some caveats to note regarding these findings. Firstly, the concentration profiles for FW (Figure 2.11) are very complex, as compared to SW, which makes direct regression across the entire profile challenging. This was most particularly the case for the lower wind speeds. In our analysis we used RMSE, as quantifiable means of comparing the success of various empirical relations to reproducing these profiles.

The results for SW were clear, that a logarithmic profile best reproduces the profile, but for FW the results are less certain. While the 2nd order polynomial performed best, the results for were only slightly better than an exponential profile and in fact for one trial, the latter model had a lower RMSE as compared to the polynomial. It’s important to note, however, that the logarithmic profile was a very poor fit to the concentration profile for FW (see Table 2.5 results). Additional observations made with a larger range of $z/H_s$ would help to better establish the vertical-dependence of the spume distribution above FW.

Up to this point, $n(z)$ and the statistical analysis has been concerned with the absolute concentration. However, the results of the $n(r)$ analysis revealed that there is simply more spray produced in SW. To highlight the differences in the FW and SW vertical distribution while taking into account the absolute amount of spume produced, representative number concentration profiles from the subranges used in Section 2.3.2 were normalized by their total, vertically-integrated concentration (Figure 2.14). This essentially gives the fractional change in number concentration with scaled height $z/H_s$. 
Figure 2.14: Selected profiles of spume droplet number concentration from Figure 2.8 at different wind speeds for FW and SW. Each profile has been normalized by the total spume droplet number concentration value for that particular profile. The values at the top of each column mark the droplet radius class (µm). X and Y axes are common to each plot and show the scaled height and fractional number concentration respectively. FW and SW data has been shown by red and blue dots respectively. The profile height, Z has been scaled by the appropriate significant wave height, $H_s$. $U_{10}$ is the 10-m neutral wind speed.

From this, we found that when compared to their corresponding FW profile, there were proportionally fewer SW droplets entrained in the air at the top of the profile. This
was found for all but one of the profiles analyzed, $r_0 = 536 \, \mu m$ at 40.5 m/s. This suggests that, while SW droplets are more readily produced for all radii and wind speeds than in FW, they are not as uniformly distributed in the vertical. Essentially, droplet production is more facilitated in SW, but the vertical transport of these droplets is somehow inhibited relative to FW.

### 2.4.2 Comparison with previous experiments

![Figure 2.15](image)

Figure 2.15: A comparison of vertical profiles for FW and SW spume droplet number concentration (in number/cm$^3$/$\mu m$) at three radius classes from a subset in this study (at $U_{10} = 36$ m/s) and the observations reported in Fairall et al. (2009) at $U_{10} = 32$ m/s. The legend bar is common to all three panels. FW and SW data from our study is in red and blue dots respectively. Solid lines in brown and cyan show the fresh and saline water data from Fairall et al. (2009). The profile height, Z has been scaled by the appropriate significant wave height, $H_s$. $U_{10}$ is the 10-m neutral wind speed. The values at the top of each column mark the droplet radius ($\mu m$).
A subset of the results from this study at $U_{10} = 35$ m/s were compared with measurements presented by Fairall et al. (2009) for $U_{10} \sim 32$ m/s (Figure 2.15); and at $U_{10} = 40.5$ m/s and 49.5 m/s with the measurements reported by Veron et al. (2012) (Figure 2.16). These comparisons are discussed here because the works of Fairall et al. (2009) and Veron et al. (2012) represent the only other available experiments in the literature to compare the results of this study to.

Fairall et al. (2009) used freshwater and salt water by adding salt to freshwater. Here, the FW and SW results were compared to Fairall et al. (2009) FW and the salty water (SW*) used in their experiments (salinity of 24 ppm). Vertical profiles of particle number concentration at three different radius classes ($r_0 = 136, 336, \text{ and } 536 \mu m$) were constructed from the droplet volume size spectra defined as the volume of the droplets identified per unit sample air volume per unit radius increment ($cm^3/cm^3/\mu m$) in the original Fairall et al. (2009) article. The vertical gap observed in all the profiles in Figure 2.15 is a consequence of the differences in $z/H_s$ observed in Fairall et al. (2009), which has a maximum $z/H_s$ of 2.1, as compared to this study where the minimum $z/H_s$ observed was 2.43.

Though both works represent laboratory experiments, the conditions and methods used for sampling differed significantly. In general, Fairall et al. (2009) measured lower number concentrations as well as less difference between fresh and saline water than was found in the present work. In addition, while the results of this study consistently demonstrate that less freshwater spume was produced near the water surface as compared to seawater, the Fairall et al. (2009) results are mixed. The strongest agreement between
these two studies occurred for the small particles with the profiles appearing to be consistent across the \( z/H_s \) gap.

Drawing specific, quantitative conclusions from the comparison with the Fairall et al. (2009) study is challenging, given the very different conditions in both experiments. While this previous study used salted freshwater as a proxy for seawater, the significant differences in the amount of spray observed would appear to be due to the former work’s use of mechanical paddle waves in addition to wind-generated waves. This appears to be the case because the differences between water types across the various studies are much smaller than simply the aggregate difference in observed spume concentration between the two studies. In the former work, mechanical waves were used to simulate long period swell underneath a wind-generated sea. These larger waves may impact the rate of spume generation, the size of droplets produced, and their vertical transport and distribution above breaking waves.

The primary conclusion that can be drawn from this comparison is that, while Fairall et al. (2009) reported effectively little impact of salinity, the findings of this work point to substantial differences in spume generation across water types. Given the limitations of both experiments, a more robust cross-comparison was not possible, but the results do strongly indicate that more work is needed to fill in these gaps and better understand the effect of varying salinity and different sea states on spume generation.
Figure 2.16: A comparison of the number concentration spectra for FW and SW at two wind speeds ($U_{10} = 40.5$ m/s and 49.5 m/s) from this study and the observations from Veron et al. 2012 at $U_{10} = 41.2$ m/s and 47.1 m/s for FW. The spectra from this study are vertical averages of the lowest six bins of the profile.

In another comparison with the study conducted by Veron et al. (2012), directly observed spume droplet concentration spectra from our measurements provided a comparison to a similarly designed and executed experiment, but that experiment only used freshwater (Figure 2.16). For comparison, the lowest six vertical cells of the concentration profiles from our study were averaged and compared to the spectra from Veron et al. 2012 (the authors of that work provide vertically-integrated concentration
spectra for only FW). In this case, the effective scaled height of the present and previous work were \( z/H_s \sim 2.5 \) and 1, respectively. For smaller drops (up to 250-300 µm), the magnitudes of \( n(r) \) observed in Veron et al. (2012) were very similar to \( n(r)_{fw} \) from our study.

For both of these studies, the radius dependences of the \( n(r) \) were remarkably similar and exhibited dependences around \( r^{-1.96} \) and \( r^{-1.8} \) however, from medium to larger radii drops (>300 µm), the spectra from this study diverge from Veron et al (2012). Our observations showed a slightly shallower radius fall-off and thus a higher concentration of large particles approaching the 1 mm radius.

Some of these differences may be attributed to variable experimental set-up, but, relative to the general uncertainty with spume generation model in this radius range, these two experiments agree very well. Furthermore, the agreement improves when comparing just FW spectra across the two studies. It shows a closer fit with the freshwater data included here. Indirectly, this provides some evidence in favor of our general conclusion that water type is a contributing factor for spume generation. However, other factors (like the fetch upwind of the sampled air volume) were confounders that prevented us from drawing a specific conclusion in comparing these two works.

2.4.3 Differences between freshwater and seawater

The results of this experiment demonstrated that there are significant and persistent differences in the radius and wind speed dependent spume generation process between FW and SW. A limitation of this experiment was that the physiochemical mechanism(s) controlling these observations could not be directly quantified; however, it is possible to speculate on the potential factors controlling the observed spray
distributions. Further study will be needed to corroborate these hypotheses and fully establish the mechanism(s) responsible for the differences captured by these experiments. This will ultimately facilitate the more physically realistic modeling of spume generation and spray-mediated processes at the air-sea interface. There are two major components to discuss:

1. The increased rate of droplet generation in SW versus FW and
2. The relatively constrained vertical transport of entrained droplets in the air above SW relative to above FW.

Given the controlled environment of the laboratory, it is plausible that physiochemical difference(s) between FW and SW play a significant role in the spray generation process. From a purely chemical perspective, the distinct impacts of solutes ions (e.g., Na+, Cl-, SO4-2, Mg+2, and others in seawater) in water is to disrupt the inter-molecular bonding between water molecules (i.e. cohesion). These non-chemical bonds are known as van der Waals forces and they are strongest in pure water and progressively weaken with the addition of dissolved material—at the same fluid temperature. This reduction in cohesion from FW to SW could facilitate the tearing of spume droplets from the water surface by the wind or the ejection of spume from the breaking wave crest.

The more commonly discussed water property (in air-sea interaction research), surface tension, is directly related to water cohesion (Auluck & Rai 1944), but is limited to the molecular layer at the surface of the water. Surface tension is critical to capillary wave development and damping in the presence of surfactants (Jarvis et al. 1967). Though surface tension, which might inhibit spray generation, is larger in SW than FW; however, given the strongly forced and highly turbulent regime under investigation, it is
unlikely that the surface tension, and/or the presence of surfactants (more readily generated in SW) are the primary driver of the observations presented here. While the cohesive properties of the water mass are present even if the surface is disrupted by the fluid’s kinematic motion. Although the physiochemical differences of FW and SW are small in absolute terms, these are distinct liquids. While the mechanical processes of wave breaking and wind forcing are most likely the primary drivers for spume generation, these factors suggest that the liquid properties play a role, which is highlighted when contrasting the spume generation from chemically different water masses.

Another plausible factor contributing to (1) may be that the increased concentration of sub-surface bubbles in SW versus FW may disrupt the surface at the crest and wind-ward side of the wave face, thereby facilitating spume generation. It has been well-established that bubbles are more numerous (and smaller) in SW as compared to FW (Haines and Johnson 1995), due to the presence of dissolved ions reducing the surface elasticity and limiting large bubble growth through a shattering mechanism (Christenson & Yaminsky 1995). While spume is not generated by individual bubble-bursting (Veron 2015), the combine action of many bubbles breaking at the surface and/or rising to the very near-surface layer could function differently. This effect would be expected to be larger in SW due to the much larger bubble population as compared to FW.

Once entrained in the air, (2) might be explained by the turbulent air flow suppression/mediation due to the development of a significantly more densely populated spray layer in SW versus FW. The higher concentration of SW particles could disrupt the
turbulent air flow, thus inhibiting their own vertical transport as compared to FW particles forced by the same background wind speed. Mass loading percentage which can be described as the ratio of the total mass of the observed number of droplets at a given speed to the total air mass of the sampling volume, has been shown in the figure 2.17 as support of this argument. Beyond 40.5 m/s, this loading percentage is around 3 times higher for SW then FW.

A factor that may contribute to this is the larger quantities of small droplets, $r_0 < 80 \, \mu m$, generated in SW than FW due to bubble-generation mechanisms. While these droplets were not resolved in the study, they were present in the control volume and contributed to the overall density of the entrained spray layer. Theoretically, it has been hypothesized that the spray layer acts as an intermediate “third fluid” layer between the conventionally binary air-water interfaces (Lighthill 1999). Furthermore, Lykossov 2001 argues that spray would disrupt the logarithmic wind profile, which fundamentally depends on a known relationship between the free stream wind velocity and the near-surface turbulence generation. Barenblatt et al. (2005) provided a mathematical model to corroborate Lighthill’s theory and demonstrated, in an idealized scenario, that the presence of spray inhibits turbulence intensity and drastically reduces the drag coefficient. Recently, high resolution model simulations have shown that the turbulent energy within a spray-laden versus spray-free flow over waves differs significantly (Richter & Sullivan 2013; Tang et al. 2017).

Another possible contributing factor to our finding (2) is the varying density between FW and SW. Assuming that all other variables (droplet size, wind speed, etc.) are equal, the droplet deposition velocity $V_d$ is directly proportional to the difference
between the densities of the droplet and the air mass. The conventional form of $V_d$ (Fairall et al. 1994) is a simple parameterization that is fundamentally based on the terminal velocity of a quasi-spherical drop of water (Pruppacher & Klett 1997, Figure 10-123 and equations 10-198 and 10-199). This approach is flawed, but without a more suitable alternative it remains a standard approach to estimating the spray flux into the atmosphere (Ortiz-Suslow et al. 2016).

![Figure 2.17: Mass loading for both FW & SW as a function of wind speed, $U_{10}$](image-url)
There is a 2.8% difference in density between FW and SW, which translates to an equivalent difference in $V_d$, all other variables held constant. For a droplet with $r_0 = 100 \mu$m, this creates a relative deposition velocity of -0.028 m/s between FW and SW. Assuming 2 identical droplets (one FW, one SW) are ejected from the same wave crest at the same trajectory and entrained into the same air flow, with a residence time $O(1)$ second (approximation based on Andreas et al. 2010 Figure 3) they become separated by a maximum of $\sim 30$ mm, or $\sim 1 \ z/H_s$ at our lowest wind speed tested. Given our entire profile ranged from 2-6 $z/H_s$, this represents a significant spatial separation solely attributed to density differences. We would note that this disparity may be more emphasized in a laboratory-scale environment and it is unknown if this has a substantial impact, for spume drops, $80 < r_0 < 1200 \mu$m, over geophysical water bodies.

There is an important limitation in this study that must be noted. During the individual trials conducted as part of this experiment, the water temperature in the tank was not recorded. Prior to every trial and/or after changing water masses in the tank, the volume was given a period of time (at least 1 full day) to acclimate to the climate-controlled laboratory. Furthermore, because of the trials were run in a closed-circuit mode, i.e. no outside airflow, the air-sea temperature regime would be approximately thermodynamically neutral. Given these conditions, we do not feel that there would be substantial differences in temperature across the various trials. Nonetheless, this remains a caveat to the work presented here. Unfortunately, between the conclusion of these experiments and this analysis, ASIST was moved to a new facility and set-up in such a way as to preclude simply redoing the experiments.
2.5 Conclusions

Here we present the results of the first quantitative comparison between fresh and seawater spume generation in hurricane-force winds. The aim of this study was to directly test whether or not water type has an impact on spume generation via wave breaking. Using a non-intrusive optical technique in the laboratory, spume droplets were observed in the radius range of 80-1400 µm, and the dependence of droplet concentration was investigated in terms of wind speed, particle size, and height above the waves for both water types.

We have reported three primary results:

1. Seawater spume was observed in significantly higher quantities as compared to freshwater.
2. The vertical distribution of seawater spume was concentrated closer to the water surface as compared to freshwater.
3. Size-dependent distributions respond significantly differently in sea and freshwater to increasing wind speed.

Collectively, the findings of this experiment point to substantial differences in the spume concentration between these two water types, suggesting that the physiochemical properties of the medium may be of importance in this process. Accounting for spray-mediated fluxes has been shown to be an important factor in tropical cyclone modeling as it is considered a crucial in the development of hurricanes and severe extratropical storms. These spray droplets are responsible for the enhancement of energy flux from the ocean to the atmosphere as demonstrated in numerous modeling and experimental studies so far (Andreas and Emanuel 2001; Andreas 2011; Bao et al. 2011; Bianco et al. 2011;
Soloviev et al. 2014; Takagaki et al. 2012). Mass loading effects of the airborne spume droplets can be visualized in Figure 2.16 above as a function of wind speed. Mass loading which was defined as the ratio of the total mass of the spume drops observed for a given wind speed to the total mass of the volume observed as given by the camera calibration. In absolute term, the magnitudes for spume mediated inertial effects may be low (1-17\%) but this clearly shows that seawater spume drops exert comparatively higher loading onto the atmosphere as compared to freshwater. The results of this experiment hold implications for modeling spray-mediated fluxes over the real ocean, in addition to large fresh water bodies.

Recent studies (Balaguru et al. 2012; Rudzin et al. 2019) have shown that the enthalpy flux at the base of tropical cyclones is impacted when these storms travel over salinity-induced barriers like river plumes. These studies suggest that accounting for the salinity of the water types is an important factor in accurately predicting the energy flux to/from the storm. Additional work is needed focusing on better understanding the mechanism controlling these differences, as well as incorporating other processes known to impact spume generation, such as non-wind driven waves.
Chapter 3

Effects of Enhanced Spray Concentration on The Surface Drag Coefficient

Re-statement of the Hypothesis: No unique relationship exists between surface drag coefficient and wind speed.

The surface drag coefficient has been shown to reduce in the presence of sea spray in the boundary layer, but this has never been verified experimentally. The focus of this study is to measure momentum transfer across the air-sea interface by exploring the behavior of surface drag coefficient \( (C_D) \) at high winds over fresh and real sea water. Since we learned from the previous chapter that sea water produces considerably more spume droplets as compared to fresh water at high winds, would that translate to differences in the behavior of \( C_D \) as well? The relationship between the 10 m drag coefficient and the 10 m wind speed is examined by use of the data obtained by the eddy covariance method. This chapter aims to address the following questions:

- What is the wind speed dependence of the surface drag coefficient?
- Are there any differences between the distributions among two water types?
- If yes, what is the plausible physical mechanisms for these differences?
- What are the implications of these results?
3.1 Background

3.1.1 Motivation

Strong tropical cyclones (TCs), also known as super typhoons in the Pacific and hurricanes in the Atlantic Ocean, cause extensive damage to property and communities as was experienced during Hurricanes (Matthew 2016, Irma 2017, and Michael 2018) in the recent Atlantic seasons. Predicting TCs through numerical computer simulations have two main tasks—forecasting their track and intensity. Even though TC track forecasting has been noticeably improving over past decades, the progress in forecasting of cyclone intensity remains limited and hence intensity predictions have not been improved much (DeMaria et al. 2014).

This is largely due to the lack of understanding of the air–sea turbulent exchange of sensible and latent heat and momentum in a spray-laden marine atmospheric boundary layer (MABL) under high wind conditions (Rastigejev and Suslov 2019). Warmer ocean surface temperature due to global warming will provide more favorable conditions for the development of stronger TCs in future, emphasizing the urgent need for accurate and precise predictions for these storm systems (Tsuboki et al. 2015).

TCs intensities are very sensitive to the small-scale physical processes such as air–sea momentum and heat transfer across the interface (Bister and Emanuel 1998). The tracks on the other hand, are controlled by large-scale atmospheric structures surrounding TCs, such as pressure patterns or westerlies, whereas intensities are very sensitive to the small-scale physics in TCs, such as air–sea momentum and heat transfer. TCs development depends on these fluxes because a tropical storm gains its thermal energy supply primarily from extracting the ocean heat flux and loses it by virtue of momentum
transport caused by the aerodynamic drag force acting on the ocean surface. Therefore, one of the primary ways to accurately predict the development of TCs is to precisely estimate heat and momentum fluxes across the interface (Komori et al. 2018).

An important component of TC-ocean interaction is the ocean spray, which plays a significant role in mediating the interfacial flux exchanges, thus affecting a hurricane both mechanically (Rastigejev and Suslov 2011; Wu et al. 2015) and thermodynamically (Peng and Richter 2017). Many researchers make a consensus that spray droplets have a significant impact on dynamics and thermodynamics of airflow. Understanding the momentum, heat and humidity exchange process and its influence mechanism is the core problem of the study of air–sea interaction.

Anthes (1982) pointed out spray evaporation lead to the cooling of the air-sea interface atmosphere, then results in sensible heat flux increased. Bao et al. (2000) showed that the inclusion of sea spray evaporation can significantly increase hurricane intensity in a coupled air–sea model when the part of the spray that evaporates is only a small fraction of the total spray mass. Meirink and Makin (2001) investigated the effect of sea spray evaporation to the atmosphere, they found a substantial cooling and moistening due to the increase in latent and the decrease in sensible heat flux. Andreas and Decosmo (2002) demonstrated that spray contributes roughly 10% of the total turbulent flux in winds of 10 m/s and between 10 and 40% in winds of 15–18 m/s. The spray contribution to the total sensible heat flux is also at least 10% in winds above 15 m/s. Gall et al. (2008) showed that large concentrations of spray ejected into the boundary layer has the capability of significantly affecting hurricane structure. Cheng et al.(2012) numerically examined effects of sea spray evaporation on heat transfer, they
found that sea spray evaporation increases the interfacial sensible heat flux, but has little effect on the interfacial latent heat flux. Meanwhile, they indicated that the effect of spray droplets on the near-surface temperature and humidity depends on droplets amounts and their locations.

Figure 3.1 Photograph of the sea surface during a hurricane (Beaufort Force 12) taken from a NOAA “Hurricane Hunter” aircraft (Image adapted from Black et al. 1986).

The omnipresence of sea spray in the marine atmospheric boundary layer (MABL) has long been observed under hurricane-strength winds (Figure 3.1). Various studies have revealed that the air-sea interface changes fundamentally in hurricane-force
winds (> 30 m/s), where it is assumed that the presence of spray droplets dominate the wave boundary layer, affecting the aerodynamic resistance of the air flow. Thus spray acts to limit (Takagaki et al. 2012), and even reduce the aerodynamic drag coefficient (Powell et al. 2003; Jarosz et al. 2007; Sanford et al. 2011).

The evaluations of the spray droplets effect on the drag coefficient were performed in a set of studies. Andreas (2004) argued that dampening of the short waves due to the “rain” of spray droplets falling back onto the sea surface results into reduction of the atmospheric drag on the sea surface. Kudryavtsev and Makin (2011) found that, at very high wind speeds, a thin part of the surface layer adjacent to the surface turns into regime of limited saturation with the spume droplets, resulting in the levelling off of the friction velocity and decrease of the drag coefficient.

It is shown by Rastigejev and Suslov (2014) that the reduction of turbulent energy and mixing length above the wave crest level, where the spray droplets are generated, that is not accounted for by the TKE model results in a significant suppression of turbulent mixing in this near-wave layer. In turn, suppression of turbulence causes an acceleration of flow and a reduction of the drag coefficient that is qualitatively consistent with field observations if spray is fine (even if its concentration is low) or if droplets are large but their concentration is sufficiently high.

Zweers et al. (2015) examined the direct effects of spray originating from the breaking wave crests on the air-sea momentum flux in their numerical model. The found drag coefficient to decrease beyond 30 m/s. Bye & Jenkins (2016) numerically reproduce the observed saturation of the drag coefficient in high winds. They argued that due to the production of spray under extreme wind speed, flattening of sea surface occurs with the
transfer of energy to longer wavelengths, resulting in capping of drag coefficient. However, to our knowledge the influence of spray particles on surface drag coefficient has been never verified experimentally and remains poorly quantified.

To address this gap, a laboratory experiment was conducted comparing the observed surface drag coefficient behavior in fresh and seawater under low to high wind conditions since a recent study conducted in the same facility has shown comparatively greater concentrations of spray (spume) droplets over seawater than freshwater at a particular fan speed Mehta et al. (2019). To our knowledge, this is the first quantitative assessment of the effects of enhanced spray concentration present in the surface layer on the surface drag coefficient. The findings of this work directly test the general assumption that a unique relationship exists between drag coefficient and wind speed and any other effects are negligible. This holds significant implications, not only for the theoretical or laboratory-based study of air-sea momentum exchange, but also for incorporating appropriate parameterization for different water types under study.

3.1.2 Theoretical background

a) The marine atmospheric boundary layer or MABL

The marine atmospheric boundary layer (MABL) is at the heart of understanding air-sea interface. In the marine environment, reduction in wind speed near the surface is primarily a function of the wave characteristics. In the MABL, surface shear stress is controlled by surface roughness, which is influenced by wave steepness, surface current velocity, wave characteristics, the wind profile, and surface current speed and direction (Powell et al. 2003). Under neutral stability and growing seas (wind has just begun to act on the ocean surface to develop waves) the MABL is similar to that over land (i.e.,
consisting of surface (inner) and Ekman (outer) layers). The datum plane is typically placed at the SWL. Along with atmospheric stability, the MABL is also dependent on parameters of, or related to, sea state. (Kitaigorodskii and Volkov 1965; Hsu 1974; Donelan 1982; Taylor and Yelland 2001).

Figure 3.2 Schematic of Ocean and Atmospheric Boundary Layer and interrelated processes

Sea state is the condition of the ocean surface relative wave characteristics at a given time and place, such as wave period and wave height. Generally speaking, the mean wind velocity profile in the MABL is logarithmic, but the height at which the log-law becomes valid depends on the wave characteristics. Sjoblom and Smedman (2003) determined that Monin-Obhikhov (MO) theory is not applicable in regions of the marine
surface layer where air flow is influenced by ocean waves, generation of wave-induced circulations, or other wave-induced forcing (Edson et al. 1999; Atakturk and Katsaros 1999; Sjoblom and Smedman 2002). Wave influence on the turbulent structure of the MABL was first identified by Volkov (1970). The atmospheric layer where surface waves have a direct influence on the structure of the boundary layer is typically referred to as the wave boundary layer (WBL). The WBL is dependent upon the degree of wave field development. Sjoblom and Smedman (2003) divided the WBL into three layers (under near-neutral conditions) that are dependent on wave state:

1. Wave influenced layer (bottom, level 1)
2. Transition layer (middle, level 2)
3. Undisturbed (top, constant stress, level 3) layer.

MO theory is not valid in the wave influenced or transition layers but is assumed valid above these regions where a normal MABL structure exists. For growing seas, a logarithmic wind profile exists (similar to that over land), as the profile is essentially undisturbed by the waves (i.e., no WBL). However, during strong swell (long-wavelength surface waves outside the generating area) the wave influenced layer can extend to considerable heights affecting wind measurements. For mature seas a transition layer separates the wave influenced layer (shallower than during swell) from the ordinary surface layer. Typical heights for the layers are: level 1 around 10 m or below, level 2 from level 1 up to about 19 m, and level 3 from level 2 up to near 26 m (Sjoblom and Smedman 2003)
b) Classification of flow regimes

Regardless of the underlying surfaces (e.g., waves, buildings, hills) aerodynamic rough flow over an array of such objects can be classified into three categories: isolated roughness flow, wake interference flow, and skimming flow. Morris (1955) was arguably the first person to study various flow classifications (he studied flow in rough conduits). More recently, Oke (1988) examined the various flow types using urban street canyons. Over a single building immersed in boundary layer flow there are three disturbed regions (mean wind direction is normal to the long axis of the building).

1. Upstream of the building, where a vortex forms in front of the windward face. This is due to downwash and is dependent on building height.
2. Behind the building, where a vortex is formed by flow air flow separation from the sharp bluff edges of the building roof and sides.
3. Further downstream of the building is its wake, characterized by increased turbulence and slower horizontal wind speeds than the undisturbed flow (flow prior to interacting with the building).

For an array of buildings (or other objects), flow types are dependent on spacing and height (Stuckley 2003). If the buildings are sufficiently far apart that the wake and vortex at each building are completely developed and dissipated before the following building is reached, the flow is classified as isolated. Essentially the flow acts as if there was only a single object immersed in the flow. As the buildings get closer together, the leading and trailing vortices at each element are disrupted, resulting in complex vortex interaction and turbulent mixing.
Wake interference flow is characterized by secondary flow in the canyon space, where the downward flow of the eddy in lee of the building is reinforced by the eddy associated with downwash on the windward face of the following building. When the canyon gets very narrow, only a single vortex exists in the cavities (skimming flow). The bulk of the flow does not enter the canyon and basically moves from rooftop to rooftop, generating quasi-smooth flow.

![Schematic diagram of the flow regimes over waves: (a) isolated roughness flow, (b) wake interference flow, (c) and skimming flow.](image)

Figure 3.3 Schematic diagram of the flow regimes over waves: (a) isolated roughness flow, (b) wake interference flow, (c) and skimming flow.

Flow classification over waves (or hills) is somewhat different (Fig. 3.3). This is because waves are continuous, in that there is no ‘flat’ space between them. Isolated roughness only occurs for very shallow waves. For this case, the flow remains attached along the wave profile. As wave steepness increases (H/L), the flow will separate at some location downwind of the wave crest (wake interference flow). As shown in Fig. 3.3b, a vortex is generated within the separation flow region. The flow reattaches at some point on the following wave. For skimming flow the waves are very close together (i.e., steep
waves), and the flow basically skips from crest to crest, bypassing the troughs. In general, the skimming flow regime is relatively smooth compared to other flow regimes, as the wave surface is sheltered.

c) What is aerodynamic drag coefficient?

The drag coefficient can be thought of as a measure of sea surface roughness in the marine environment (Bye and Jenkins 2006) and is a function of sea state, atmospheric stability, and wind speed (Donelan et al. 2004). Onshore and offshore wind directions have also been shown to affect $C_D$ (Geernaert and Smith 1997; Vickers and Mahrt 1997; Feddersen and Veron 2005). The horizontal component of the wind acting on the sea surface leads to wind stress or, in other words, the vertical transfer of horizontal momentum. The surface momentum flux (wind stress) at the air-sea interface is critical for modeling oceanic and atmospheric circulation. Wind stress over the ocean $\tau_w$ is formally defined as the retarding force per unit area exerted by the sea surface on the flow, and is typically described in terms of a 10-m surface drag coefficient (Simiu and Scanlan 1986).

Traditionally, the vertical flux of horizontal momentum (wind stress) has been evaluated as a function of wind speed $U_{10}$ and drag coefficient $C_D$ as,

$$\tau = -\rho_a \overline{w'u'} \approx \rho_a C_D |U_{10}| U_{10}$$

(1)

Where $\rho_a$ is air density, and $\overline{w'u'}$ is the downward turbulent flux of horizontal velocity. Drag coefficient is the squared ratio of friction velocity and wind speed, i.e. $C_D = (u_*/U_{10})^2$, where $u_* = \sqrt{-\overline{w'u'}}$. For a flow that obeys a logarithmic law of the wall, the drag coefficient is uniquely related to aerodynamic roughness length $z_0$:
\[ C_D = \left[ \frac{k}{\ln(z/z_0)} \right]^2 \]  \hspace{1cm} (2)

Where \( k \) is the Von-Karman constant, empirically determined to be about 0.4, and \( z \) is the reference height, typically 10 m. Roughness length \( z_o \) is commonly used to parameterize wind stress in numerical weather prediction models (Davis et al. 2008). Pioneering work by Charnock (1955) established a relationship between roughness length and friction velocity:

\[ z_o = \frac{a_{us}^2}{g} \]  \hspace{1cm} (3)

Where \( g \) is the gravitational acceleration and \( a \) is a constant, whose value was determined from observations by many investigators (Garratt, 1977; Wu, 1980; Smith, 1980, 1988).

Kudryavtsev and Makin (2007) separated form drag into two components \( \tau_f = \tau_w + \tau_s \).

The first term \( \tau_w \) describes the transfer of momentum from wind to waves on the upwind portion of the wave. Form (wake) drag produced by air flow separation is described by the second term \( \tau_s \) (Kudryavtsev and Makin 2001). It has been suggested that in hurricane conditions (\( U10 > 33 \) m/s), wind stress is primarily supported by form drag, thereby affecting the momentum transfer (Donelan et al. 2004; Kudryavtsev and Makin 2007). In these conditions, waves travel at significantly different velocities than the wind, as the wave field rarely has sufficient time to adjust to the strong wind forcing.

**d) Estimating the drag coefficient**

Direct measurements of the drag coefficient are problematical and typically it must be inferred from different experimental techniques. The most common techniques available in the present literature are as follow:
1. **The profile method:** This method utilizes the logarithmic law, which is based on the flat-plate boundary layer theory of Prandtl and von Karman (Schlichting and Gersten 2000). This method assumes that a nearly logarithmic mean wind velocity profile exists in the constant flux layer (\( u^* \) is approximately constant with height throughout the surface layer) under neutral stability, horizontal homogeneity, and stationarity. The profile method can only be employed if the mean wind speed at two or more levels is known. In order to obtain reliable wind profile measurements, Wieringa (1993) suggested that all of the following criteria must be met for considering observations to be good.

- Anemometers must be spaced a sufficient vertical distance apart to avoid interference and far enough away from the mast or boom outside the range of flow distortion of the structure,
- Anemometers must be precisely calibrated and wind speed data should be averaged over at least 10 min,
- Observations should be restricted to near neutral conditions or temperature data should be collected for diabatic profile correction,
- Non-stationary situations should be avoided, and
- Several independent runs of the experiment must be conducted.

2. **Eddy correlation method:** The eddy-correlation technique calculates friction velocity using direct measurements of eddy fluxes in atmospheric surface layer. For this method to be valid, momentum flux must be essentially constant with height from near the surface to the measurement height. In addition, the flow conditions must be statistically stationary and homogeneous. Over the sea surface, these conditions are usually met for a fetch length > 1 km at an observation height \( \leq 10 \) m (Kraus and Businger 1994). This
method requires that the covariance be performed between the vertical $w'$ and horizontal $u'$ and $v'$ fluctuating components of velocity. The further details for this method have been described in section 3.2.3 ahead.

3. **Inertial dissipation:** The inertial dissipation technique is an alternative approach of measuring the friction velocity from high-frequency wind velocity measurements (Hicks and Dryer 1972; Edson et al. 1991). In this method, high-frequency wind measurements are used to determine the energy dissipation rate in the inertial subrange and thus estimate the turbulent flux of momentum. Kolmogorov hypothesized that the power spectral density $S_{uu}(k)$ of the along wind component $u$ beyond the peak of the spectrum in the inertial subrange depends only on the dissipation rate of turbulent kinetic energy (TKE) $\varepsilon$ and on the wave number $k$ as:

$$S_{uu}(k) = Kk^{-5/3} \varepsilon^{2/3}$$  \hspace{1cm} (4)

where $K$ (0.05-0.6) is the Kolmogorov constant. The inertial subrange is an intermediate range of turbulent scales that separates the energy-containing eddies from the viscous eddies in the dissipating range (Kaimal and Finnigan 1994). Using Taylor’s hypothesis of ‘frozen’ turbulence and that wave number $k = 2\pi f/U$ where $f$ is the measurement sampling frequency and $U$ is the along wind component mean wind speed the rate of energy dissipation can then be determined by calculating a mean value of $f^{-5/3}S_{uu}(f)$ over an appropriate frequency range. It follows that the wind stress can be found by using the TKE budget. However, this methodology of measuring $u_*$ has a distinct advantage over the other methods in that it is relatively insensitive to platform motion, such as the low frequency oscillations on board a ship.
4. **Bulk aerodynamic method**: This method parameterizes wind stress in terms of the mean wind speed (bulk gradient) and a bulk transfer coefficient for momentum ($C_D$). Over the ocean the parameterization of $\tau_w$ usually takes the form:

$$\tau_w = \rho_a C_D(z) \left[U(z) - U_c\right]^2$$

(5)

where $U_c$ is the mean ocean surface current.

5. **Turbulence intensity and gust factor methods**: Turbulence intensity (TI) is a measure of the fluctuating component of the wind and it characterizes the intensity of gusts in the flow. It is defined as the ratio of the standard deviation $\sigma_u$ of a given wind speed record to the mean wind speed $U$. Gust Factor (GF) is defined as the ratio between the peak and mean wind speeds Wieringa (1993). The TI method is based on two assumptions: (1) that a logarithmic wind profile exists and (2) that the ratio of the standard deviation of the wind record to the friction velocity is $\sigma/u_* = 2.5$ (Beljaars 1987; Barthelmie et al. 1993). Counihan (1974) found this method is valid in smooth terrain for $z_o < 0.1$ m. Under these assumptions, the roughness length can be estimated from the total turbulence intensity at height $z_a$ as follows:

$$z_o = z_a \exp \left[1/TI\right]$$

(6)

The TI method requires the use of anemometers that are capable of sampling at a frequency of about 1 Hz or faster. However, the GF method can be applied from maximum gust data collected at typical weather observing stations. The TI method is generally preferred over the GF method, due to the latter method having numerous parameters based on anemometer and equipment need compute $z_o$. Both methods have gained popularity over the inertial dissipation method, which suffers from much error.
e) Why determining drag coefficient is important?

According to the theory of tropical hurricanes proposed by Emanuel (1995), the surface drag coefficient of the sea surface is a critical parameter. According to this theory, a mature tropical cyclone may be idealized as a steady, axisymmetric flow whose energy cycle is very similar to that of an ideal Carnot engine, where the ocean is the hot reservoir with temperature $T_S$ and the troposphere is the cold reservoir with temperature $T_0$. The maximum surface wind velocity in a cyclone which determines its category, can be estimated from the Carnot theorem. According to this theorem, the maximum efficiency of the ideal heat engine is determined by the absolute temperatures of the hot and cold reservoirs as,

$$n = \frac{w}{Q_S} = \frac{T_S - T_0}{T_0}$$

where $Q_S$ is the heat energy entering the system from the hot reservoir and $W$ is the mechanical work done by the system. Heat energy support of the tropical cyclone comes from the ocean (heat flux from the sea surface) and mechanical energy dissipated in the marine turbulent boundary layer (Emanuel, 2003), the heat energy entering the system is the surface integral of the heat flux from the sea $F_q$ and mechanical energy dissipation rate $F_p$:

$$Q_S = \int (F_q + F_p) \, dS$$

(8)

Mechanical work done by the system compensates mechanical energy dissipation, then

$$W = \int F_p \, dS$$

(9)

The heat flux from the sea and the mechanical energy dissipation rate are determined by the bulk formula:

$$F_q = C_k \rho |\bar{V}|(k_0 - k)$$

(10)
here $k_0$, $k$ is enthalpy at the sea level and in marine atmospheric boundary layer.

$$F_P = C_D \rho |\vec{v}|^3$$  \hspace{1cm} (11)

In (7)-(8) $C_k$ is heat exchange coefficient (or the Stanton number), $C_D$ is surface drag coefficient, defined by equation (3). Considering the Carnot theorem (4) and estimating integrals (5) and (6) yields estimate for the maximum surface wind velocity in a tropical cyclone as a function of ratio $C_D/C_k$:

$$|\vec{V}|_{Max} = \sqrt{\frac{C_k \tau - T_0}{C_D k}} (k - k_0)$$  \hspace{1cm} (12)

In numerical simulations of TCs, momentum flux and latent and sensible heat fluxes at the sea surface are calculated using the following bulk equations, where the drag coefficient is $C_D$, the latent heat transfer coefficient is $C_E$, and the sensible heat transfer coefficient is $C_H$:

$$\tau = p C_D U_{10}^2$$  \hspace{1cm} (13)

$$Q_E = p L V C_E U_{10} (q_i - q_{10})$$  \hspace{1cm} (14)

$$Q_H = p C_p C_H U_{10} (T_i - T_{10})$$  \hspace{1cm} (15)

Here, $r$ is the air density, $L$ is the latent heat of vaporization, $C_P$ is the specific heat of humid air at a constant pressure, $U$ is the wind speed, $q$ is the specific humidity, and $T$ is the temperature. The subscripts $i$ and $10$ indicate the sea surface and a height of 10 m above the sea level, respectively. The transfer coefficients are usually parameterized against wind speed $U_{10}$ at a height of 10m, using the field measurement data.

3.1.3 Sea surface drag behavior in high winds

A number of fields, experimental and numerical studies have been conducted till date to explore the surface drag behavior in high winds. Wind stress over the ocean has
been a subject of study (and arguably debate) for over 50 years. Present parameterizations still have significant limitations, especially in high wind regimes. Over the last six years air-sea interaction research has focused primarily on measurements of $C_D$ rather than parameterizations.

Conducting in situ measurements of the mean wind velocity profile in extreme conditions is exceptionally difficult in the open ocean, but pose an even greater challenge in the near-shore region. Due to this vicissitude, past research efforts have estimated wind stress in only weak to moderate winds ($3 < U_{10} < 25$ m/s). Data from these measurements have been extrapolated to strong wind speeds, where it has been generally accepted that $C_D$ increases linearly with increasing wind speed (e.g. Garratt 1977; Amorocho and De-Vries 1980; Smith 1980; Large and Pond 1981; Yelland and Taylor 1996; Drennan et al. 1999). Extrapolations of this wind tendency are currently being employed in applications such as forecasting the intensity and track of tropical cyclones, determining the wind forcing for storm surge and wave forecasts, input for numerical weather prediction models, and determining extreme wind loads (Powell et al. 2003).

Prior to 1997 and the development of the Global Positioning System dropwindsonde (Hock and Franklin 1999), which measures kinematic and thermodynamic properties during its descent, and the novel airborne instruments developed during the Coupled Boundary Layer Air-Sea Transfer Experiment (CBLAST) beginning at the turn of the century (Black et al. 2007), estimates of $C_D$ in extreme wind regimes (for research purposes) were simply not possible.

Arguably the most important study of the drag coefficient in extreme winds to date is that of Powell et al. (2003), hereafter PV03. PV03 used high resolution wind
profile measurements from GPS sondes dropped in the vicinity of hurricane eyewalls to observe the behavior of $C_D$ in hurricane winds in deep water. A summary of the important work of PV03 is highlighted in the following. PV03’s study utilized detailed wind profile measurements acquired by GPS sondes from 1997–1999. During this time span, 331 eyewall (or near-eyewall) wind profiles were sampled in hurricanes located in the Atlantic and Eastern/Central Pacific basins. The GPS sonde samples pressure, temperature, humidity, and position at a frequency of 2 Hz. Sondes are generally dropped at a height of 1.5–3 km above the sea surface and fall at a rate of 10–15 m/s. Failure of these instruments is possible in regions of turbulence or heavy rainfall. Derived wind speed and height measurements have an accuracy of 0.5–2.0 m/s and 2 m, respectively (Powell et al. 2003).

PV03 estimated surface layer quantities using composite wind profiles (individual 0.5-s samples from 331 sonde profiles) based on mean boundary layer (MBL) wind speed. MBL wind speed was arbitrarily defined as the mean wind speed below 500 m. The maximum wind speed for each profile was usually contained within this layer. The 331 wind profiles were averaged for all measurements in the lowest 3 km as a function of the MBL wind speed and each profile was organized into one of five MBL groups (in m/s): 30–39 (72 profiles), 40–49 (105 profiles), 50–59 (55 profiles), 60–69 (61 profiles), and 70–79 (38 profiles). A relatively wide range of wind speeds (21–67 m/s) were measured near the surface (8–14 m). PV03 found a logarithmic increase in wind speed with height in the lower 200 m (Fig. 3.4).
Above 200 m, the mean profile continued to increase, reached a peak near 500 m, and decreased thereafter. The decrease was due to a weakening of the horizontal pressure associated with the warm core of a tropical system. PV03 attributed the apparent increased variability above 600 m to convective-scale features in the eyewall (identified by Jorgensen (1984). The aforementioned logarithmic profile below 200 m and maximum wind speed at 500 m is consistent with a neutrally stable surface layer. PV03 used the profile method in the lowest 100–150 m of each MBL group to estimate $z_o$ and $u_*$. Then, $C_D$ was estimated using Eq. 5. Results indicated that surface layer parameter dependence on wind speed is vastly different than previously believed in high wind speeds, as shown in Fig. 3.5.
Figure 3.5 Momentum exchange quantities as a function of $U_{10}$. Vertical bars represent the range of estimates based on 95% confidence limits and the symbols represent the different layer depths used in the profile method: ◦ 10–100, _ 10–150, Δ 20–100, ⋄ 20–150 m. (a) $z_o$ from GPS sonde profiles (symbols) and Large and Pond (1981) relationship (solid line) and Charnock (1955) relationships for comparison (dotted line, $\alpha = 0.035$; dashed-dot line, $\alpha = 0.018$) and (b) $C_D$ with relationships and values from tropical cyclone budget studies. Figure adapted from Powell et al. (2003).

Observations of $C_D$ are consistent with other investigation in weak to moderate wind speeds. However, above 40 m/s $C_D$ behaved markedly different than what was anticipated. PV03 observed a leveling off at 40 m/s and a large decrease when $U_{10}$ increased to 51 m/s. Similarly, $z_o$ and $u_*$ increased up to 40 m/s (this behavior was roughly predicted by previous studies), leveled off around 40 m/s, and decreased as $U_{10}$ increased to 51 m/s. In summary, $C_D$ plateaus near wind speeds of 35 m/s, corresponding to a saturation value near 0.0026 Powell et al. (2003). Their surface drag depended significantly on the sector of the tropical cyclone, where it was measured. Their results showed clear leveling-off of the drag coefficient, and an indication of its decrease in
hurricane-force winds despite the large confidence intervals and uncertainties associated with the lagrangian nature of derived wind profiles.

Shortly after the work of PV03, Donelan et al. (2004) examined the effect of high wind speeds on $C_D$ in a laboratory setting. Donelan et al. (2004), hereafter DO4, utilized the unique Air-Sea Interaction Facility at the University of Miami. They offered supporting evidence to the observational results found by Powell et al. (2003). The wave tank at the University of Miami is 15 m long, 1 m wide, and 1 m high. The tank was equally divided between air and water with the water side being 0.5 m deep. Wind speed in the tank was set by a programmable fan. Vertical profiles of wind speed and turbulence were not provided.

Reynolds stresses were measured directly using an x-film anemometry. Direct measurements of Reynolds stress were made at measured elevations and corrected to surface values using the measured horizontal pressure gradient in the tank. For high wind speeds, DHR04 computed surface stress from a momentum budget using sections of the tank as a control volume (Fig. 3.6). The momentum budget method was accomplished assuming steady state conditions. The control volume, defined by the shaded area $A_1B_1B_2A_2$ in Fig. 3.6, extended a unit width into the paper. Under this assumption, wind stress $\tau_a$ produced a downwind current near the surface which resulted in an increase in mean surface elevation downwind $h_2$. 
A return flow in the bottom of the water column in the upwind direction and stress $\tau_b$ applied to the bottom of the tank was caused by an induced pressure gradient due to the elevation difference. Momentum fluxes through the tank are respectively denoted as $I_1$ and $I_2$ for sections $A_1B_1$ and $A_2B_2$. They verified the momentum budget method by comparing results for this method to the other methods utilized, which show good agreement. Verification of this method was important since it allows for the estimate of surface stress in high winds in a way that was insensitive to airborne water droplets (problematic for x-film measurements).

Using this method, DHR04 showed that wind stress, $\tau$ is given by:

$$
\tau = h(\rho_w g \gamma + \frac{\Delta P}{L}) + \frac{\Delta S_{ss}}{L} - \tau_b
$$

(16)
where $h$ is the average water depth, $\Delta P$ is the air pressure difference, $\Delta S_{xx}$ is the radiation stress for deep water, and $s$ is the slope of the water surface. Sidewall stresses were neglected since they tend to cancel out assuming steady state conditions (Donelan et al. 2004).

DO4 conducted several experiments using the programmable fan in the wave tank. The fan was set to fixed speeds for 300 second runs to obtain a stable estimate of wind stress. DO4 noted a loss of water droplet spray to the return duct which contributes to wind stress on the surface. They accounted for water loss due to spray by calculating an upper limit of 1.5% lost momentum due to spray, which they noted as “generous”.

Results from DO4 compared to those of Large and Pond (1981) and Ocampo-Torres et al. (1994) are shown in Fig. 3.7. To bring the results to scale, the height of the wind speed measured in the tank (10 cm) was extrapolated to the standard meteorological height of 10 m using a logarithmic dependence on height. Again, it was not mentioned what the actual wind profile was like above the wave/water surface. The initial decrease in $C_D$ to a minimum at a wind speed of 3 m/s was a result of an aerodynamically smooth surface (Fig. 3.7). As the waves increased and the surface became aerodynamically rough, $C_D$ increased with increasing wind speed between 3–33 m/s. At 33 m/s, $C_D$ leveled off, corresponding to a limiting value of 0.0025. Results were only slightly lower than the open ocean measurements of Large and Pond (1981) for weak to moderate wind speeds.
Figure 3.7 Laboratory measurements of the neutral stability drag coefficient by profile, Reynolds Stress and momentum budget methods. The drag coefficient refers to the wind speed measured at the standard anemometer height of 10 m. The frequently cited drag coefficient formula derived from field measurements by Large and Pond (1981) is also shown. From Donelan et al. (2004).

Later Jarosz et al. (2007) strengthened this saturation argument by showing similar dependencies of the surface drag coefficient on the wind speed. They estimated momentum transfer using ocean currents measurement data in their analysis and showed a decreasing trend of $C_D$ for wind speeds greater than 32 m/s. So, field measurements of the wind stress both from the atmospheric and ocean sides of the air-sea interface showed that the sea surface drag coefficient is significantly reduced at hurricane wind speeds in
comparison with the extrapolation of the experimental data obtained at “normal” wind speeds and even decreases for $U_{10}$ exceeding 35 m/s.

In another summary of the field observations for drag coefficient reported by Holthuijsen et al. (2012), hereafter HN12, $C_D$ was observed to reach its maximum at a value ranging from $\sim 1.5 \times 10^{-3}$ to $3 \times 10^{-3}$, consistent with the Donelan et al. (2004) laboratory results. HN12 performed image processing analysis of 86 nadir-looking photographs from low-level flights in nine hurricanes, and quantified the whitecap and streak coverage as function of wind speed. They also estimated drag coefficient using the same methodology as Powell et al. (2003). They found that while the whitecap coverage remains under 15% with increasing winds, the streaks of foam and spray gradually increase until approximately 40 m/s winds beyond which they dominate the surface with the so-called white out. HN12 suggested that the near-saturation of streak coverage results in aerodynamically smooth surface and reduction of $C_D$ to below $1.0 \times 10^{-3}$ in $U_{10} > 60$ m/s (Fig. 3.8).

They also found higher drag in the front-left quadrant of the hurricane where counter- and cross-wind swell were prevalent. However, to date there are no in situ measurements in such conditions, and there may be uncertainty or bias associated with these results owing to the Lagrangian nature of dropsondes wind measurements and the distance that they travel during the fall. The ability to reproduce wind-wave misalignment in the controlled laboratory setting may thus be critical to assess the role of this process for air-sea momentum exchange.
Figure 3.8 Figure adapted from Holthuijsen et al. (2012). (a, b) Observed drag coefficient $C_D$ and (c) white cap and streak coverage as a function of surface wind speed $U_{10}$. The solid green line in panel (b) represents the approximation for cross swell and the solid black line for following and opposing swell. $C_D$ values from the previous studies (Garratt, 1977; Wu, 1982; Black et al., 2007; Powell et al., 2003; Jarosz et al., 2007; Smith and Banke, 1975; Large and Pond, 1981; Petersen and Renfrew, 2009) are also shown in panel (a). Blue and red dots in panel (c) represent bin mean of the observations of this study for each 2 m/s wind speed bin. Shaded area represents white cap coverage $W$ from 19 previous studies compiled by Anguelova and Webster (2006), curved blue line represents analytical approximation, and horizontal red line represents mean value for $U_{10} > 24$ m/s. Vertical bars represent 90% confidence interval of mean value. Numbers at the bottom indicate sample sizes used in computing the data points with the same color directly above as determined by the number of wind speed measurements in the 25 m height bin.

Recent laboratory experiments conducted by Troitskaya et al. (2012) and Takagaki et al. 2012 showed similar results too. Troitskaya et al. (2012) showed surface drag coefficient saturation above 25 m/s accompanied by saturation of wind wave slopes as well. Takagaki et al. 2012 used an eddy correlation method and showed the saturation level above 35 m/s with $C_D$ approaching to a constant value of 0.00255. They attributed their finding due to interface flattening and slipping associated with the generation of huge quantities of bubbles and spray as a result of wave breaking.

The results described above (Powell et al. 2003; Donelan et al. 2004; Black et al. 2007; Jarosz et al. 2007) are significantly different from those previously accepted, hence seeking explanations of this phenomenon are critical. Donelan et al. (2004), using a study by Reul (1998) who observed vorticity contours for gentle and steep-sloped waves, suggested that this qualitative change in the surface drag coefficient is a consequence of air flow separation from breaking waves which acts to decouple the MABL from the chaotic (rough) sea surface. Air flow experiences separation from the surface and
reattachment near the crest of the following wave, indicative of skimming flow where the flow ultimately bypasses the troughs. Powell et al. (2003) offered an explanation via the role of sea foam and spray that has been observed on the sea surface in wind speeds above hurricane force. He suggested that the thick layer of foam that typically develops in winds in excess of approximately 40 m/s forms a ‘slip’ surface which impedes the transfer of momentum from the wind to the sea. These explanations were supported by the findings of Kudryavtsev and Makin (2007).

Using modeling techniques, Kudryavtsev (2006) investigated the role of sea spray on sea drag at high wind conditions. They showed that the production of spray results in the acceleration of airflow, and the suppression of turbulence and surface drag. At wind speeds of 50–80 m/s, the model predicted the effects of a ‘slip’ surface, with $C_D$ on the order of 0.0001 (Kudryavtsev 2006). Although both studies offer valuable insight, it is still to be determined if these explanations hold in the near-shore region, where conditions are markedly different.

So, it can be concluded both from field and laboratory data that the growth of the aerodynamic roughness of the water surface with wind speed is significantly reduced at sufficiently high winds in spite of increasing of surface wave heights. Several mechanisms are present in the literature for explanation of this finding and are given in the next section.

3.1.4 Possible mechanisms of drag saturation

A number of physical mechanisms have been suggested so far by various studies attempting to hypothesize the existence and level of drag saturation which can be described as below:
1. **Enhanced dissipation at the surface due to wave breaking:** In sufficiently high wind speeds, the tops of steep wave crests can be blown off due to direct impact of wind onto a sloped water surface of the wave. This process acts to effectively transfer part of wave energy into kinetic and gravitational potential energy of spray and spume, which eventually re-enter the ocean. This is manifested as a decrease of wave spectrum peakedness in high winds and has been observed in the laboratory by Takagaki et al. (2012, 2016). However, applicability of this argument remains uncertain in hurricane conditions in the open ocean where length scales of surface wave characteristics can reach to the order of 10-100 m.

2. **Sheltering effects leeward of breaking waves:** Aerodynamic properties of the airflow has been shown to modify in high wind speeds, due to the geometry of the steep breaking waves. Reul et al. (1999) observed this in a laboratory setting using a PIV (particle image velocimetry) technique and then was postulated by Donelan et al. (2004) as the likely cause for drag saturation in high winds. Sudden changes in the water surface slope of a breaking wave cause the flow separation from the surface of every consecutive crest, trapping stagnant high vorticity flow structures in the wave troughs. Buckley and Veron (2016) used a PIV method along with a LIF (laser induced fluorescence) in their laboratory experiment to confirm this existence of sheltering downwind from the crest for young wind sea, and upwind from the crest for swell outrunning wind (figure 3.9).
Figure 3.9 Examples of instantaneous velocity fields $u/U_{10}$ plotted over LFV images: (a) $U_{10}=0.86$ m/s, (b) $U_{10}=1.40$ m/s with MGR waves, (c) $U_{10}=5.00$ m/s, (d) $U_{10}=9.41$ m/s and (e) $U_{10}=14.34$ m/s. Figure adapted from Buckley & Veron (2016)

3. **Presence of spray/foam in the boundary layer causing gradients in vertical density above**: Wave breaking and wind blowing spume off the wave crests cause ejection of water droplets (spray) into the air and injection of air bubbles into the water. These two processes effectively act to reduce the density gradient across
the air-sea interface. This was first coming into light following Barenblatt et al. (2005) numerical study of Lighthill’s sandwich model.

![Figure 3.10 Schematic of ocean spray formation. Figure adapted from Barenblatt et al. (2005)](image)

Soloviev et al. (2014) postulated that at wind speeds higher than 30 m/s, spume and bubbles dominate the wavy interface and form the so-called white out (Holthuijsen et al., 2012), under which the individual waves become suppressed as roughness elements and the form drag decreases with wind speed. They argue that the two-phase regime, in which the there is a continuous density gradient from air to water, leads to a local minimum of drag coefficient at winds between 60-70 m/s.

4. **Physical limitations of experimental laboratories:** Scale limitations of the wave tanks can contribute to the behavior of wave growth, dissipation, and ability to fully simulate spray-mediated fluxes due to loss of droplet particles from the tank in very high winds. In a tank flume, the wave spectrum development is highly
limited by its fetch (length of the tank, O (10 m) for Donelan et al. (2004) and Takagaki et al. (2016) and by water depth. In comparison, the open ocean surface waves forced by hurricane winds formed under an effective fetch of O (100 km), have peak wavelengths of O (100 m), and significant heights of O (10 m) (Wright et al. (2001); Walsh et al. (2002)).

3.1.5 Effects of sea spray on sea surface drag behavior

A number of studies (predominantly numerical modeling) have been conducted so far focusing on the effects of sea spray on the momentum transfer by exploring the surface drag coefficient dependencies on the wind speed. The generation of sea spray and the effect of sea spray on the wind turbulence in extreme winds have drawn strong interests in the past few decades. Andreas (2004) suggested that the spray droplets re-entering the water surface suppress the shortest gravity waves that support a large fraction of stress. High concentrations of spray have also been postulated to create a stably stratified multi-phase interface that suppresses small-scale turbulence and effectively limits drag (Lighthill, 1999; Barenblatt et al., 2005; Bye and Jenkins 2006; Bye and Wolff 2008).

In a study by Kudryavtsev and Makin (2011), the impact of ocean spray on the dynamics of the marine near-surface air boundary layer in conditions of very high (hurricanes) wind speeds was investigated. They parameterized the effects of spume droplets on the turbulence mixing through stratification. They showed through their marine near-surface air boundary layer model that the presence of sea spray droplets in the near surface boundary layer leads to the wind acceleration and suppression of turbulence. Approaching very high wind speeds, a thin part of the surface layer adjacent
to the surface turned into a regime of limited saturation with the spume droplets, resulted in the levelling off of the friction velocity and decreased the drag coefficient as $U_{10}^{-2}$, $U_{10}$ being the wind speed at 10-m height.

In a similar fashion, dynamics in the marine boundary layer were investigated using the Monin-Obukhov framework by Bao et al. (2011). They explored the mechanical effects of sea spray, and showed that the mechanical impact counterbalances the thermal effect by reducing the friction velocity, and stabilizing the marine boundary layer. They also showed how the presence of sea spray intensifies the storms by decreasing the drag through numerical simulations.

Rastigejev et al. (2011) pointed out that the concentration of spray increases rapidly with the wind speed at extreme winds, which results in a significant flow acceleration, compared to the wind without sea spray droplets. Vertically momentum exchange was predominantly affected by the large spume droplets. The contribution of jet and film droplets in the momentum exchange rate was found to be negligible at wind speeds $>20$ m/s due to the very low concentration of these droplets in the boundary layer.

Rastigejev & Suslov (2011) did a number of numerical experiments modeling the effects of ocean spray on the momentum fluxes across the air-sea interface at high winds. Primarily the spume droplets resulting from the mechanical action of wind on wave crests were found to affect the vertical momentum exchange. The contribution of jet and film droplets in the momentum exchange rate was found to be negligible since their total volume production at wind speeds is much lower than that of spume droplets. They also found that the airflow in the lower part of the boundary layer can noticeably accelerate
because of the turbulent transport suppression by the ocean spray if the spray concentration increases rapidly enough with the wind speed.

Richter and Sullivan (2013) performed a series of direct numerical simulations to highlight the role of sea spray in the momentum transfer between the atmosphere and the ocean. The sea spray droplets were modeled as heavy inertial particles, suspended in a turbulent airflow. They used a common approach from CFD (Computational Fluid Dynamics) and Engineering, the Eulerian-Lagrangian approach, where the droplets were represented by solid pointwise particles whose individual trajectory was tracked over time in a Lagrangian way. At the same time, the fluid governing equations (the traditional Navier-Stokes equations plus a feedback term from the sea spray) were solved in an Eulerian framework. They discussed the fact that for typical diameters of sea spray droplets (typically from 10m to 1mm), the mechanical effect would dominate over the thermal effect. By studying the momentum budget, Richter and Sullivan introduced a “spray” stress, which compensates the decrease in Reynolds stress by providing a feedback effect to the turbulence. They explained that the drag coefficient based on the total stress remains almost unchanged in the presence of sea spray, while the drag coefficient based on the turbulent stress is reduced.

In a later study by (Rastigejev & Suslov, 2016) the significance of mechanical and thermodynamic effects spray has on the temperature, humidity, and turbulent kinetic energy of the air in the lower atmosphere above the ocean was emphasized. Results were found sensitive to the spray concentration. The mechanical effect of spray primarily leads to the decrease in average turbulent kinetic energy (hence forth TKE) level, while thermodynamically spray had almost no effect on the TKE distribution. The value of the
drag coefficient at the reference level of 10 m above the sea level was monotonically decreased as the spray concentration increased, re-confirming the so-called lubrication effect of a spray reported in Barenblatt et al. (2005) and then in Rastigejev & Suslov (2014).

Most recently, Zhang et al (2016) recently showed that the sea spray droplets play a dominant role in the momentum transfer between air and water in their model of the marine atmospheric boundary at extreme winds. Their numerical model demonstrated that the presence of sea spray droplets can strongly affect near-surface wind profiles and sea-surface drag coefficients under a range of low to high wind speeds. Low drag coefficients at high wind caused by ocean-spray production were shown to increase turbulent stress in the sea-spray generation layer, accelerating the near-sea-surface wind.

As evident from the above studies that numerous attempts have been made so far to explore the effects of spray droplets on the momentum transfer across the air-water interface. The results from these studies points to the potential role these spray droplets have on altering the near surface boundary layer dynamics and the momentum transfer. However, to our knowledge, it hasn’t been verified experimentally.

Here we report results from first laboratory experiment conducted in the same test facility as described in the previous chapter, looking into spray-spume effects on the momentum transfer indirectly by comparing results in fresh and seawater, since results from the previous chapter showed significantly higher concentrations of spume droplets over seawater as compared to freshwater. Surface drag coefficient was calculated here using a direct eddy covariance method on the three-dimensional wind data obtained using a sonic anemometer inside the wind flume in fresh and seawater.
3.2 Methods

3.2.1 The laboratory setup

The high speed, state-of-the-art Air-Sea Interaction Saltwater Tank facility (ASIST) at the University of Miami was used for the experiments in both freshwater (FW) and seawater (SW). It has a 15.0 m long, 1.0 m wide and 1.0 m high glass test section (see Figure 2.4).

Figure 3.11 Sonic Anemometer (IRGASON) placed inside the ASIST facing upwind. Two capacitance-based wave gauges measuring water elevation can also be seen placed behind the anemometers’ sampling volume to avoid the flow distortion.

The end of the tank has a sloped porous surface that acts to efficiently dissipate waves to minimize reflection. Sources for the water were the same as mentioned earlier in chapter 2. The initial water depth was kept at 0.42 m, and the vertical height of the airflow above the tank was 0.58 m. In the tank, wind waves can be generated by a
programmable fan. Wind and waves were measured using a 3-D sonic anemometer, and capacitance-based wave gauges, which were placed 6.5 m, and 6.75 m from the inlet respectively as shown in figure 3.11.

3.2.2 Data collection and processing

For both FW and SW, three dimensional perturbations of wind velocity were measured by a sonic anemometer (IRGASON) sampling at 20 Hz and located at a fetch of 6.5 m from the inlet. The fan frequency was increased from 10 Hz to 60 Hz (corresponding to the measured 10 m neutral wind velocity $U_{10}=6–38$ m/s, with increments of 5 Hz (~3.2 m/s). A total of 11 runs were recorded, each lasting 360s for both water types. For the data analysis, first 60 s data were not included as it was the time allowed for fan to ramp up to reach a steady state at that particular fan frequency. The wind speed was measured at 29 cm height in the tank and extrapolated to the standard meteorological height of 10 m using the well-established logarithmic dependence on height (Donelan et al. 2004).

Here the stress at the interface was estimated by extrapolating the measured values of the Reynolds stress with the measured horizontal pressure gradient in the tank (Donelan et al. 2004). Necessary corrections were applied to the raw sonic anemometer data to produce fluctuations and means. Eddy covariance data processing has been widely investigated in previous studies (Lee et al. 2005) and consists of several steps: spike detection, coordinate rotation, and detrending to separate the turbulent signals from the mean flow. Each step of the process can be approached in various ways, and selection of specific approaches depends upon both site and weather conditions (Vickers and Mahrt,
In this study, the raw eddy covariance data were processed by the following steps:

1). *Spike detection and removal:* A point can be considered a spike and is discarded if it falls outside the prescribed limits. The prescribed limits were 40 m/s for the horizontal wind speed, 10 m/s for the vertical wind speed. A point is also considered a spike if it is greater than 4 standard deviations of the difference between consecutive data points of a 1.5 min record. The data are not treated as spikes if four or more are detected consecutively. The discarded spikes are replaced by linearly interpolated values.

2). *Averaging period:* The choice of the averaging period is considered a crucial factor in turbulent flux calculation using the eddy covariance method. We chose 1.5 min as the averaging period in our calculation of turbulent fluxes. Hence for each fan frequency, we obtained 4 values for friction velocity and drag coefficient.

### 3.2.3 The eddy covariance method

In the eddy covariance method, friction velocity can be directly calculated from the measurements of turbulent fluctuations as,

$$u_s = \left( \bar{u'} \bar{w'}^2 + \bar{v}^1 \bar{w'}^2 \right)^{1/4}$$

(17)

Here $u'$, $v'$, and $w'$ are the turbulent fluctuations of the three-dimensional component of the wind speed. Reynolds averaging has been indicated by the over bar symbol, and $u_s$ is the friction velocity. The 10 m equivalent drag coefficient under the neutral stability conditions can be estimated as,

$$C_{D10} = \frac{u_s^2}{u_{10}^2}$$

(18)
The 10 m equivalent wind speed or $u_{10}$ was estimated from the measured wind speed ‘$u$’ at the height of 29 cm from the mean water level using a logarithmic surface layer profile as,

$$u_{10} = u - \frac{u_s}{k} \cdot \ln(z / 10)$$

(19)

Here $k=0.40$ is the von Karman constant and $z$ are the measurement height (29 cm in this case). 10 m equivalent wind speed (in m/s) derived from equation 15 has been shown as a function of ASIST’s fan speed (in Hz) for both SW and FW in figure 3.12 below.

![Figure 3.12 10-m equivalent wind speed ($U_{10}$, m/s) determined from equation 15, plotted against fan speed (in Hz)](image)

### 3.3 Results and Discussion

The 10 m equivalent drag coefficients ($C_{D10}$) calculated using the eddy covariance method on the three-dimensional wind data obtained from a sonic anemometer over FW
(orange squares) and SW (blue squares) were compared with the previous laboratory studies (Donelan et al. 2004 in maroon from the same facility, Takagaki et al. 2012 in green circles) and field observations reported by Powell et al. 2003 in pink circles, and Edson et al. 2013 in grey stars. Figure 3.13 shows the distributions of $C_{D10}$ against 10-m equivalent wind speed ($U_{10}$) in m/s.

![Figure 3.13](image)

Figure 3.13 Laboratory measurements of the drag coefficient refers to the wind speed measured at the standard anemometer height of 10 m (under neutral stability) by eddy covariance method for fresh and seawater. Other commonly cited distributions (lab and field studies) for drag coefficient are also shown for comparison.

The characteristic behavior of the surface drag coefficient ($C_{D10}$ increasing with wind speed) is evident here as the surface condition goes under transition from aerodynamically smooth to rough as the wind speed is increased. For both FW and SW,
the drag coefficient increased with the increase in wind speed and then started levelling off as the wind speed was increased further, consistent with observations reported by Donelan et al 2004 and Takagaki et al. 2012. Magnitude wise, the $C_{D10}$ values from the current experiment were closer to those reported by Takagaki et al. 2012 and Edson et al. 2013 for both FW and SW. Donelan et al. 2004 and Powell et al. 2003 showed comparatively lower $C_{D10}$ values. In the wind speed range of 7 to 22 m/s, our observations were within the data limits of open ocean measurements reported by Edson et al. 2013.

In general, $C_{D10}$ values for FW were found to be always greater than of SW with the difference ($C_{D_{(FW-SW)}}$) increasing monotonically with increase in $U_{10}$. This difference became appreciable starting tropical storm winds, where there is significant amount of spray present in the near surface boundary layer. The difference in $C_{D10}$ values between FW and SW increased monotonically with the increase in $u_{10}$. Saturation of the drag coefficient can be seen once the wind speed exceeds 33 m/s for FW and 30 m/s for SW. The saturation level for the drag coefficient was 0.0027 for the FW and 0.0024 for the SW, showing a 12.5% increase in the roughness over freshwater. These consistent lower $C_{D10}$ values in seawater than freshwater indicates a physical phenomenon limiting the aerodynamic roughness state of the sea surface is of critical importance in understanding and modelling the development of hurricanes and other intense storms.

We hypothesize our results to be likely caused by the turbulent air flow intensity suppression/mediation due to the development of a significantly more densely populated spray layer in SW versus FW. Seawater has been shown to facilitate the ejection and entrainment of the water particles from the surface, and generates larger quantities of
spray droplets in the boundary layer over a wind range of size spectrum through both bubble bursting mechanism (Monahan 1967; Monahan & Zietlow 1969; Slauenwhite and Johnson 1999), and spume production (Mehta et al. 2019) droplets compared to freshwater.

These spray droplets manifest their physical presence by affecting the wind profile and turbulent structure, thereby suppressing the sea-surface drag coefficient. Theoretically, it has been hypothesized that the spray layer acts as an intermediate “third fluid” layer between the conventionally binary air-water interfaces (Lighthill 1999). Barenblatt et al. (2005) provided a mathematical model to corroborate Lighthill’s theory and demonstrated, in an idealized scenario, that the presence of spray inhibits turbulence intensity and drastically reduces the drag coefficient.

![Figure 3.14 Friction velocity (U*, in m/s) obtained via eddy covariance method for FW (orange circle) and FW (blue circle), plotted against fan speed (in Hz)](image)
Recently, high resolution model simulations have shown that the turbulent energy within a spray-laden versus spray-free flow over waves differs significantly (Richter & Sullivan 2013; Tang et al. 2017). The turbulent suspension of sea-spray droplets reduces the buoyancy and makes the surface layer more stable. As a consequence, the downward turbulent mixing of momentum is reduced and the friction velocity is lowered by around 20% (Bao et al. 2011), as compared to 10% in our analysis at wind speed around 35 m/s (see figure 3.14).

Our results also support the field observations by Xiao et al. 2013, who showed higher $C_{D10}$ values for freshwater lakes than modeled values according to the Garratt (1992) model for similar conditions in the marine environment. At wind speed around 9 m/s, they found the difference to be about 35%. They suggested that the air-water transfer coefficients taken from studies focusing on marine environment may be subject to large uncertainties when applied to inland lakes since very few studies exist on their applicability to lake environments.

Recently, Ortiz-Suslow et al. (2016) observed seawater spray generation in high winds in the ASIST laboratory at the University of Miami and found significantly more large spume droplets were generated than previously considered (Fairall et al. 2009; Mueller and Veron 2009). The larger concentration of spume droplets would likely enhance the effects of spray on the air-sea exchanges. These effects are not considered in current coupled models that evaluate $C_D$ based on a parametric form or by integrating the wave growth source function. Even in low-to-moderate winds, the observed drag coefficient has a significantly large scatter around some average value (Edson et al. 2013). This raises the question whether a unique relationship between drag coefficient
and wind speed exists. We hypothesize that the answer is no, and that the drag is governed by several physical processes that are only in part determined by surface wind. These processes include but are not limited to the presence of spray droplets in the surface layer, sheltering effects lee wards of breaking waves, enhanced dissipation at the surface due to wave breaking.

3.4 Conclusions

In this work, we have studied the effect of sea spray on the drag coefficient by measuring it indirectly over fresh and seawater using the eddy covariance method in a laboratory setup. This is the first experimental verification of the sea sprays’ microphysical effects on the turbulent intensity of the air flow in the boundary layer as suggested by numerous modelling and theoretical studies. Consistent with previous modeling studies of spray influence on the drag coefficient, the results from our experiment shows higher surface drag coefficient values in freshwater as compared to seawater.

The only difference between salt and freshwater that could contribute to the momentum flux difference is either bubble or spray concentrations. While bubbles have not been shown to directly impact momentum fluxes, we conclude that spray is the likely contributing factor. Since seawater facilitates and produce larger quantities of spray droplets in the surface layer than freshwater for a given wind speed via both bubble bursting and spume generation. Hence this enhanced spray stratification of the boundary layer can likely contribute to the reduction in interfacial momentum transport.

Our result also supports the previous that while working in a lake environment, appropriate corrections should be applied on the transfer coefficients obtained from
marine environment. Although still limited, the results derived from this unique study shed new light on the relationship between $C_{D10}$ and $u_{10}$ suggesting that the physiochemical properties of the medium may be of importance in this process. More data and analysis are needed to re-confirm our findings. In this work, just the momentum fluxes were analyzed. The importance of the heat and moisture fluxes and the lack of observations in high wind have been discussed in some studies (Richter and Stern 2014; Soloviev et al. 2014).
Chapter 4

Final Remarks

4.1 Overall Conclusions

This dissertation has a wide scope with critical implications in the understanding of air-sea interaction dynamics under hurricane conditions. The results of this work are part of the larger effort towards the better understanding of the physical coupling between the atmosphere and the ocean. In hurricane wind conditions, the amount of spray entrained into the lower atmosphere begins to form an intermediate layer that is composed in part of air, in part of 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 the quantification of 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, and the resulting surface drag coefficient in tropical cyclone-like conditions above fresh and real seawater ever collected.

Here I presented the results from the first of its kind experimental quantitative comparison of spume generation and surface drag coefficient behavior between fresh and real seawater from low to hurricane-force winds. The aim of this study was to directly test whether the inherent physiochemical properties of the water type have any impact on the surface drag coefficient and spume generation via wave breaking. The aerodynamic resistance offered by the water surface to the air flow in the boundary layer was calculated using the eddy covariance method on the three-dimensional wind fluctuation
data from a sonic anemometer. Spume droplets, on the other hand, were observed using a non-intrusive optical technique in the laboratory, in the radius range of 80-1400 µm. The dependence of droplet concentration was investigated in terms of wind speed, particle size, and height above the waves for both water types.

From this data set, I was able to demonstrate that the physiochemical properties of the water masses significantly affect the interfacial flux exchanges from that water to the atmosphere above. Seawater spume was observed in significantly higher quantities along with their vertical distribution being concentrated closer to the water surface as compared to freshwater. Size-dependent distributions respond significantly differently in sea and freshwater to increasing wind speed. Our analysis shows higher surface drag coefficient values in freshwater as compared to seawater, even at the low wind speeds when the spray concentration is relatively low. Since seawater facilitates and produces larger quantities of spray droplets than freshwater for a given wind speed, the spray stratification of the boundary layer can explain the reduction of momentum transport between the atmosphere and the ocean.

Collectively, the findings of this experiment point to substantial differences in the spume concentration between these two water types, suggesting that the physiochemical properties of the medium may be of importance in this process. Accounting for spray-mediated fluxes has been shown to be an important factor in tropical cyclone modeling as it is considered crucial in the development of hurricanes and severe extratropical storms. These spray droplets are responsible for the enhancement of energy flux from the ocean to the atmosphere as demonstrated in numerous modeling and experimental studies so far (Andreas and Emanuel 2001; Andreas 2011; Bao et al. 2011; Bianco et al. 2011; Soloviev
et al. 2014; Takagaki et al. 2012). The results of this experiment hold implications for modeling spray-mediated fluxes over the real ocean, in addition to large fresh water bodies. We have also demonstrated that, while working in a lake environment, appropriate corrections should be applied on the transfer coefficients obtained from the marine environment.

The implications of these findings suggest that appropriate consideration should be given to the water type used in the laboratory experiments. Further work will be focused on providing a fuller and more comprehensive observational study into the air-sea interaction and the effects of physiochemical properties of water type on the interfacial fluxes with focus on wave development effects on the spray generation and their transport 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.

Recent studies have shown that the enthalpy flux at the base of tropical cyclones is impacted when these storms travel over salinity-induced barriers like river plumes. These studies suggest that accounting for the salinity of the water types is an important factor in accurately predicting the energy flux to/from the storm.

4.2 Future Endeavors

Intrigued by the outcome of this very basic, simple, yet interesting study, I have certain objectives that I want to work on in the near future. Additional work will be done focusing on better understanding the mechanism controlling these differences in FW and SW, as well as incorporating other processes known to impact spume generation, such as non-wind driven waves. Approaching further extreme wind conditions (cat-5 hurricane and above), further laboratory work is planned to tackle the problem of spray spume
production in the larger laboratories such as the SUSTAIN facility. The focus will be on understanding the actual generation mechanisms. How these relates to the wave phase, and how these are affected in a variety of changing experimental conditions, i.e. with the addition of swell and mixed-seas, incremental change in salinity etc.

Figure 4.1 Spectral density (m²/Hz) from the surface elevation time series measured for the freshwater in ASIST facility at fetch=9.0 m, plotted against frequency (Hz). Color legend shows different runs corresponding to difference $U_{10}$ (in m/s).
Another very important aspect of the air-sea interaction i.e. surface gravity waves will be explored in future studies. There is an increasing need for accurate wave spectra in remote sensing applications and for accurate estimates of momentum exchanges between waves and the atmosphere and ocean in coupled modeling. When improving wave models, it is important to understand the spectral shape also known as the high frequency tail of a wind wave spectrum.

This can be approximated by two power laws as: $B_4f^{-4}$ (equilibrium range given by Toba 1973) and $B_5f^{-5}$ (saturation range given by Phillip 1958). Resio et al. (2015) found that the break between the two ranges (from equilibrium to saturation) definitely migrated towards the peak of the spectrum as the ratio of wind speed to wave phase speed ($U/C$) increases suggesting that strongly forced spectra may have different spectral dynamics than more temperate conditions.

This is potentially relevant for tropical cyclones. Some preliminary measurements for the wave spectra from the ASIST tank are shown in figure 4.1 below. Further wave measurements at higher winds will be conducted in the SUSTAIN facility for both FW and SW to see if the differences observed for spume generation and drag coefficient translates to differences in the wave characteristics as well.

Further going, continuing work should be done on developing a reliable field measurement technique for capturing droplet concentrations in tropical cyclone conditions. More data and analysis are needed to re-confirm our findings. In this work, just the momentum fluxes were analyzed.
The importance of the heat and moisture fluxes and the lack of observations in high wind have been discussed in some studies (Richter and Stern 2014; Soloviev et al. 2014). Further research should be conducted regarding that; this will be the next topic of my future studies.
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