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Surface Wave Impacts on Air-Sea Momentum Flux and Upper Ocean Turbulence Under Tropical Cyclones

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SURFACE WAVE IMPACTS ON AIR-SEA MOMENTUM FLUX AND UPPER
OCEAN TURBULENCE UNDER TROPICAL CYCLONES

BY

BRANDON G. REICHL

A DISSERTATION SUBMITTED IN PARTIAL FULFILLMENT OF THE
REQUIREMENTS FOR THE DEGREE OF
DOCTOR OF PHILOSOPHY

IN

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UNIVERSITY OF RHODE ISLAND

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DOCTOR OF PHILOSOPHY DISSERTATION
OF
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2015

ABSTRACT

Air-sea energy and momentum fluxes constrain the available energy for tropical cyclone intensification. The surface wave field modifies these fluxes in numerous ways. In this study we focus on two sea-state dependent processes. The first is the sea-state dependence of the wind stress over the ocean under tropical cyclone force winds. The second is the sea-state dependence of the upper-ocean turbulence due to Langmuir turbulence, which modifies the mixed-layer deepening and cooling of the ocean surface during the passage of a tropical cyclone. In this study we utilize the WAVEWATCH III (WW3) surface wave model, the one-dimensional General Ocean Turbulence Model (GOTM), and the Princeton Ocean Model (POM) to investigate these processes.

The impact of the surface wave field (sea-state) on the wind stress over the ocean is investigated with fetch-dependent seas under uniform wind and with complex seas under idealized tropical cyclone winds. Two different approaches are employed to calculate the wind stress and the mean wind profile. The near-peak frequency range of the surface wave field is simulated using the WAVEWATCH III model. The high frequency part of the surface wave field is empirically determined using a range of different tail levels. The results suggest that the drag coefficient magnitude is very sensitive to the spectral tail level but is not very sensitive to the drag coefficient calculation methods. The drag coefficients at 40 m/s vary from 1×10^{-3} to 4×10^{-3} depending on the saturation level. The misalignment angle between the wind stress vector and the wind vector is sensitive to the stress calculation method used. In particular, if the cross-wind swell is allowed to contribute to the wind stress, it tends to increase the misalignment angle. Our results predict similar amounts of sea state dependence of the drag coefficient regardless of the approaches taken. More sea-state dependence of the drag coefficient is predicted for tropical cyclones than for aligned growing wind conditions, this enhancement of the drag coefficient may be attributed

to swell that is significantly misaligned with local wind. The amount of sea-state dependence of the drag coefficient is sensitive to the tail level and the translation speed of the tropical cyclone.

The upper-ocean turbulence is significantly modified by the Stokes drift of the surface waves because of the Craik-Leibovich vortex force (Langmuir turbulence). Under tropical cyclones the contribution of the surface waves varies significantly depending on complex wind and wave conditions. Therefore, turbulence closure models used in ocean models need to explicitly include the sea-state dependent impacts of the Langmuir turbulence. In this study the K Profile Parameterization (KPP) 1st moment turbulence closure model is modified to include the Langmuir turbulence effect, and its performance is tested against concurrent Large Eddy Simulation (LES) experiments under tropical cyclone conditions. First, the KPP model is tuned to reproduce LES results in conditions of shear turbulence only (KPP-ST). Next, KPP is tuned to typical ocean conditions (with typical Langmuir turbulence) but includes no explicit sea-state dependent modifications (KPP-iLT). Next, the Eulerian currents are replaced by the Lagrangian currents in the KPP equations for calculating the bulk Richardson number and the vertical turbulent momentum flux (KPP-Lag). Finally, an enhancement to the turbulent mixing is introduced as a function of the non-dimensional turbulent Langmuir number (KPP-LT). KPP-LT, with the Lagrangian currents replacing the Eulerian currents and the turbulent mixing enhanced, significantly improves the prediction of the upper-ocean temperature and currents compared to the default (unmodified) KPP model under tropical cyclones. This modified KPP model also shows improvements over the default KPP at constant moderate winds (10 m/s).

We then examine differences between simulations with KPP-ST, KPP-iLT, and KPP-LT under tropical cyclones in POM. Both KPP-iLT and KPP-LT enhance sea surface cooling due to vertical mixing compared to KPP-ST, but KPP-iLT signifi-

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ACKNOWLEDGMENTS

The success of this work was only possible thanks to the contributions of many mentors, colleagues, and collaborators. I gratefully thank my major professors, Tetsu Hara and Isaac Ginis, for being excellent advisers and inspiring me during my doctoral studies at the University of Rhode Island. I also thank Kathleen Donohue and Alexander Meyerovich for serving on my core doctoral committee. I thank David Ullman for serving as a dissertation committee member and Stephen Licht for serving as chair of my dissertation defense. I thank Biju Thomas for continuous developmental support throughout my studies. I thank all other members of the University of Rhode Island's Air-Sea Interaction and Numerical Modeling Laboratory. I thank Mark Donelan of the University of Miami for helpful discussion related to Chapter 1. I thank Dong Wang and Tobias Kukulka of the University of Delaware, who were instrumental to Chapter 2, including running many computer simulations and participating in many helpful discussions.

I thank the National Oceanographic Partnership Program through the National Oceanic and Atmospheric Administration and the Naval Research Laboratory for funding the work that led to the first chapter and the National Science Foundation for funding the work that led to the second and third chapters. I thank the University of Rhode Island's Graduate School of Oceanography and the Graduate School of Oceanography Alumni Association for funding additional support including travel to scientific meetings.

DEDICATION

This dissertation is dedicated to my parents, Jeffrey Reichl and Judith Reichl. My success has only been possible thanks to your support. From my first day of pre-school to my last day of graduate school, I always knew I could count on you for anything.

I also dedicate this work to the memory of Steven J. DeStefano. I was fortunate to take a similar journey from Pennsylvania to Rhode Island with the companionship of an incredible friend. I miss you every day, but I will always treasure the memories.

PREFACE

This dissertation is written in manuscript format and consists of three scientific journal articles. The first chapter was published in the Journal of Geophysical Research Volume 119 Issue 1 in January 2014. The second chapter is in review with the Journal of Physical Oceanography as of December 2015. The third chapter has been prepared to be submitted for publication to the Journal of Physical Oceanography.

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MANUSCRIPT 1

Sea state dependence of the wind stress over the ocean under hurricane winds

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Abstract

The impact of the surface wave field (sea-state) on the wind stress over the ocean is investigated with fetch-dependent seas under uniform wind and with complex seas under idealized tropical cyclone winds. Two different approaches are employed to calculate the wind stress and the mean wind profile. The near-peak frequency range of the surface wave field is simulated using the WAVEWATCH III model. The high frequency part of the surface wave field is empirically determined using a range of different tail levels. The results suggest that the drag coefficient magnitude is very sensitive to the spectral tail level but is not as sensitive to the drag coefficient calculation methods. The drag coefficients at 40 m/s vary from 1×10^{-3} to 4×10^{-3} depending on the saturation level. The misalignment angle between the wind stress vector and the wind vector is sensitive to the stress calculation method used. In particular, if the cross-wind swell is allowed to contribute to the wind stress, it tends to increase the misalignment angle. Our results predict enhanced sea-state dependence of the drag coefficient for a fast moving tropical cyclone than for a slow moving storm or for simple fetch-dependent seas. This may be attributed to swell that is significantly misaligned with local wind.

1.1 Introduction

The wind stress (or the drag coefficient) at the ocean surface is one of the most important parameters needed for ocean, atmosphere, and surface wave models. In particular, accurate predictions of tropical storm (hurricane) track and intensity require detailed knowledge of the spatial and temporal development of the wind stress that is strongly modified by complex surface wave fields (sea-states).

Many previous studies have investigated how the wind stress is modified by different sea-states. They all start with the momentum conservation constraint that the wind stress is equal to the sum of the momentum flux into surface waves (form drag of surface waves) and the momentum flux directly into the subsurface currents (through viscous stress). The momentum flux into the waves is then expressed as an integral of the wave variance spectrum multiplied by the wave growth rate. Beyond this common basic framework, however, the studies significantly diverge in a few key aspects. These aspects include the parameterization of the high frequency part (tail) of the wave variance spectrum, calculation of the wave growth rate due to wind, the feedback of the wave form drag on the mean wind profile, and wave breaking impacts.

In order to estimate the sea-state dependent wind stress, a wave variance spectrum must be specified first. Often the wave variance spectrum is defined empirically. For example, several studies (*Makin and Kudryavtsev*, 1999; *Kudryavtsev and Makin*, 2001; *Makin and Kudryavtsev*, 2002; *Mueller and Veron*, 2009) used a wave spectrum based on the spectrum introduced by *Elfouhaily et al.* (1997). This spectrum is a combination of a low frequency model that is dependent on the wind speed and fetch, and a high frequency model that is dependent on the wind friction velocity. The studies of *Kudryavtsev and Makin* (2001) and *Makin and Kudryavtsev* (2002) defined the high frequency spectrum using the energy balance model proposed by *Kudryavtsev et al.* (1999). The study of *Moon et al.* (2004b) (hereafter MGHBT) and *Donelan*

et al. (2012) (hereafter DCCM) considered more complex sea-states, including those under tropical cyclone winds, by explicitly simulating the wave variance spectrum. MGHBT modeled the wave variance spectrum using the WAVEWATCH III (*Tolman*, 2009) (hereafter WW3) ocean wave model, while DCCM modeled the wave spectrum using the University of Miami wave model (UMWM). It is well known that the high frequency part (tail) of the spectrum has a significant impact on the air-sea momentum flux. While the tail is included in the empirical parameterization of *Elfouhaily et al.* (1997), numerical wind-wave models typically resolve the spectrum to a certain frequency (not far from the spectral peak) and require empirical tail parameterizations to extend the spectrum to high frequencies. The study of MGHBT used the resolved spectrum up to $3 \times f_{pi}$ (f_{pi} is the peak input frequency, which is the peak frequency of the wind sea portion of the wave spectrum and is one of the standard outputs of the wave model) and then applied the equilibrium tail model developed by *Hara and Belcher* (2002). DCCM resolved the spectrum to 2 Hz and extended the spectrum using an empirical relationship between the spectral slope of the tail and the wind speed.

Once the spectrum is specified, the next step is to calculate the wave form drag using the wave growth rate. Previous studies parameterized the wave growth rate either using the wind speed (*Snyder et al.*, 1981; *Donelan et al.*, 2006; *Tsagareli et al.*, 2010, DCCM) or the wind stress (after *Plant*, 1982). The parameterizations with the wind stress are further divided into those using the total wind stress (*Janssen*, 1991), and those using the reduced stress, which accounts for the reduction of the wind forcing in the presence of larger scale waves (e.g. *Makin and Mastenbroek*, 1996; *Makin and Kudryavtsev*, 1999; *Hara and Belcher*, 2004; *Mueller and Veron*, 2009; *Banner and Morison*, 2010, MGBHT). One early study also considered a hybrid growth rate that is a function of both the full wind stress and wind speed (*Makin*

et al., 1995). Another unresolved aspect of the growth rate is the impact of swell that may or may not be aligned with the local wind. Observational evidences show that the wind stress may be modulated by the presence of swell (e.g. *Donelan et al.*, 1997; *Drennan et al.*, 1999; *Grachev et al.*, 2003; *García-Nava et al.*, 2009, 2012). DCCM explicitly included the impact of swell on the form drag calculation by considering different growth rates for waves that are faster than or opposite to the wind. MGHBT neglected the impact of swell on the growth rate.

The final step of the drag coefficient calculation is to model the feedback of the wave form drag on the mean wind profile. This step is needed to establish a relationship between the wind stress and the wind speed (normally at 10 m height). The neutral wind profile in some studies is simply approximated using log-layer vertical wind profiles (e.g. *Kudryavtsev and Makin*, 2001; *Mueller and Veron*, 2009, DCCM). In this case the wind profile is dependent only on the surface roughness parameter, z_0 , that is, the feedback appears only in the parameterization of the sea-state dependent z_0 . Other studies account for the feedback of the modified turbulent stress (due to wave form drag) on the mean wind shear in the wave boundary layer using various turbulence closure methods (e.g. *Makin and Mastenbroek*, 1996; *Makin and Kudryavtsev*, 1999; *Hara and Belcher*, 2004, MGHBT). Some studies (*Hara and Belcher*, 2004, MGHBT) explicitly ensure that energy remains conserved in the wave boundary layer.

When surface waves break, the airflow may separate at the wave crest and apply increased form drag on the waves. While earlier studies did not explicitly account for the impact of breaking waves on the form drag (e.g. *Makin and Kudryavtsev*, 1999, MGHBT), many recent studies separated the breaking wave form drag from the form drag of non-breaking waves (*Kudryavtsev and Makin*, 2001; *Kudryavtsev et al.*, 2001; *Makin and Kudryavtsev*, 2002; *Kukulka and Hara*, 2008a,b; *Mueller and*

Veron, 2009; Banner and Morison, 2010). The inclusion of an explicit breaking-wave component into a wind stress model introduces additional complexities and uncertainties into the wind stress model. First, the presence of breaking waves is often represented by a breaking wave distribution. This distribution is not well known and often parameterized from the spectrum and the growth rate (*Kudryavtsev and Makin, 2001; Makin and Kudryavtsev, 2002; Mueller and Veron, 2009*). Second, the momentum exchange between breaking waves and the wind must be modeled. This is often simplified by assuming that the shape of the wave can be approximated as a backward facing step where the airflow separates (*Kudryavtsev and Makin, 2001; Makin and Kudryavtsev, 2002; Kukulka and Hara, 2008a,b; Mueller and Veron, 2009*). Third, the feedback of the breaking and non-breaking wave form drag on the mean wind profile needs to be included.

Kukulka and Hara (2008a,b) included the breaking wave effect on the drag coefficient over a wide range of sea-states from laboratory conditions to open ocean conditions. They allowed large uncertainties in the breaking distribution, the breaking wave form drag, as well as the wave spectrum at high frequencies (tail) that is limited by wave breaking. They found that in open ocean conditions the most significant breaking wave impact on the drag coefficient appears in reducing the level of the spectral tail rather than in enhancing the wave form drag due to flow separation. This is because the occurrence of dominant scale breaking waves is relatively rare in the open ocean. (In contrast, the dominant scale breakers determine the drag coefficient in laboratories, since almost all dominant waves break.) An alternative (simpler) approach to implicitly including the breaking wave effect is to define the average growth rate of all waves (including non-breaking and breaking waves). In fact, the study of DCCM tuned their growth rate to observations that included both breaking and non-breaking conditions.

In summary, the impact of different sea-states on the wind stress is still an unresolved question, particularly with complex wave fields such as those under tropical cyclone winds. The main objective of this study is to investigate the effect of different sea-states on the wind stress with the help of numerical experiments. We will conduct two types of experiments. The first type will be fetch-dependent simulations under constant uniform winds. The second type will use idealized tropical cyclone wind fields to force simulations of the waves. We will evaluate two distinct approaches of estimating the momentum flux, one proposed by DCCM and the other based on MGHBT, using identical wave spectra. We seek to determine to what degree the estimations of the wind stress depend on different assumptions regarding the two key aspects: wave growth rate and feedback on the wind profile. Furthermore, we will investigate the impact of the different spectral tail parameterizations and examine misalignment between the wind stress vector and the wind speed vector in hurricane conditions.

1.2 Methods

1.2.1 Wave Spectrum Simulations

In order to investigate the sea-state dependent wind stress, we first specify the wave spectrum. In this study all wave spectra are simulated using the wind-wave model, WW3. WW3 is a third-generation model maintained by the National Oceanic and Atmospheric Administration (NOAA) through the National Center for Environmental Prediction's (NCEP) Environmental Modeling Center (EMC). The latest operational model version, 3.14, is used in the simulations. The model accepts the 10 meter wind speed (in time and in space) as an input, and calculates the wind stress using its own drag coefficient parameterization. It then calculates the wind forcing using its own wave growth rate parameterization. In principle, this WW3 growth rate should be consistent with the growth rate we use in our calculation of the sea-state

dependent momentum flux, and the WW3 drag coefficient should be consistent with our own estimates as well. However, the WW3 drag coefficient and the growth rate have been empirically adjusted (together with the wave dissipation parameterization) to produce wave spectra that are consistent with observations, mainly in low to moderate wind conditions. It has been known that the model tends to overestimate the significant wave height in tropical cyclone conditions. Recently, *Fan et al.* (2009) have shown that the WW3 wave prediction significantly improves if the WW3 drag coefficient is replaced by the parameterization developed in *Moon et al.* (2004b,a), which yields a lower drag coefficient at high wind speeds. In their study the WW3 results were directly compared to Scanning Radar Altimeter (SRA) observations, and it was shown that the model accurately reproduced large variability of the significant wave height under complex hurricane wind conditions. We therefore employ the same drag coefficient as in *Fan et al.* (2009) for the wave simulations in this study. However, the feedback of the sea-state dependent drag coefficient obtained in this study on surface wave simulations is not pursued here. For a complete description of WW3 including the governing equations see *Tolman* (2009).

WW3 explicitly simulates waves up to $3\times$ the peak input frequency. Internally the model attaches a spectral tail that is highly sea-state dependent. This parameterized tail is largely a residual of the tuning process and has not been thoroughly validated against observations (Dr. Hendrik Tolman, personal communication). The dependences of the spectral tail level on the wind speed and the wave age ($c_p/u_\star$, where c_p is the phase speed at the spectral peak and $u_\star$ is the wind friction velocity) are still poorly understood, because very few direct observations exist. Since the stress calculation is very sensitive to the tail level, we need to parameterize the tail level and investigate its impact on the drag coefficient calculation.

There are few observations of the spectral level of the tail in the field. Ob-

servations by *Romero and Melville* (2010) in the Gulf of Tehuantepec showed that the directionally integrated saturation spectrum ($B(k) = \Psi(k) \times k^4$, where $\Psi(k) = \int_{-\pi}^{\pi} \Psi(k, \theta) d\theta$, and the directional wavenumber spectrum $\Psi(k, \theta)$ is defined such that the mean square surface displacement is equal to $\int_{-\pi}^{\pi} \int_0^{\infty} \Psi(k, \theta) k dk d\theta$) in the tail is roughly independent of k and falls between 6×10^{-3} and 10×10^{-3} for wind speeds between 11 and 20 m/s and wave ages between 17 and 32. Their results were roughly consistent with previous findings over similar wavenumbers (*Forristall*, 1981; *Banner et al.*, 1989). They did not find a significant wave age dependence of the tail level within the range of their observations. Note that their data were collected under a gap flow and the resulting wave field may be different from typical ocean conditions. In higher wind conditions there are no reliable observations of the spectral tail levels.

Since our knowledge of the spectral tail is very limited, particularly at high wind conditions, we will test a wide range of B values and investigate its impact on the drag coefficient. Since there is no consensus regarding the dependence of the tail on the wind speed or wave age, we set the saturation spectrum B in the tail as constant in k and with no systematic variation with the wave age or wind speed in each experiment. We will simulate the peak region of the wave spectrum using WW3, and smoothly transition the result to the parameterized tail at high frequencies. Specifically, the WW3 spectrum is used up to $1.25\times$ the peak input frequency. From $1.25\times$ to $3\times$ the peak input frequency the spectrum is linearly varied from the explicit model calculation to the predetermined B value. The spreading function is also changed from the explicit calculation result to simple cosine squared dependence around the direction of the near surface wind vector.

In our preliminary analysis it became apparent that any single B value for the tail does not reproduce the value of the drag coefficient given by the COARE 3.5 air-sea flux parameterizations at its full range of wind speeds (*Edson et al.*, 2013), suggesting

that the tail level systematically increases with wind speed between 0 and 20 m/s. The COARE 3.5 is a recent and comprehensive drag coefficient parameterization and is therefore chosen as a benchmark to compare our results to. (We also include comparisons to the classic *Large and Pond* (1981) drag coefficient parameterization for reference.) We therefore vary the B values over a sufficiently large range such that our models can reproduce the observed drag coefficient in all wind speeds up to 20 m/s. Specifically, we will test three different values; the low B value ($B = 2 \times 10^{-3}$) yields better agreement with the drag coefficient of the COARE 3.5 algorithm at 5 m/s, the medium B value ($B = 6 \times 10^{-3}$) shows better agreement with the COARE 3.5 at 10 m/s, and the high B value ($B = 12 \times 10^{-3}$) shows a better agreement at 20 m/s. This range of B values encompasses the variability reported in the observations (*Forristall*, 1981; *Banner et al.*, 1989; *Romero and Melville*, 2010). It is certainly possible that B may become either greater than 12×10^{-3} at higher wind speeds and/or at very young seas or less than 2×10^{-3} under lower wind speeds. Nevertheless our choice of the three different B values is sufficiently broad to systematically study the impact of the tail on the drag coefficient.

Directionally integrated saturation spectra for two wind speeds at a fetch of 100 km (the experimental set up is presented in the next section) are shown to demonstrate the transition from the WW3 spectrum to the three parameterized tail levels (Figure 1.1). Note that at high winds the WW3 tail parameterization is higher than even the highest of our 3 tail levels. The empirical spectrum of *Elfouhaily et al.* (1997) is also shown in the figure for reference. While their spectrum is very high for the gravity-capillary waves at high wind speeds, their saturation spectrum of the short gravity waves (k up to 50 rad/m or so) is within the range of our investigation. In this study we do not account for the enhancement of B in the gravity-capillary range and simply truncate the spectrum at a fixed wavenumber $k = 400$ rad/m for simplicity.

The impact of the spectral enhancement in the gravity-capillary range and a different cutoff wavenumber on our drag coefficient calculation is relatively small compared to the large variation of the tail level examined.

To confirm that our choice of the B value range is reasonable it is useful to examine the resulting mean square slope. In Figure 1.2, the calculated mean square slope from our fetch-dependent wave spectra (WW3 spectra with the 3 levels of the tail) at different wind speeds are shown. It is clear from this exercise that the mean square slope has a strong dependence on the tail level. To reproduce the mean square slope measured by *Cox and Munk* (1954) it is clear that the B value would need to increase from about 4×10^{-3} at 5 m/s to nearly 12×10^{-3} at 15 m/s. It is interesting that these B values are quite consistent with the B values that yield drag coefficients similar to the COARE 3.5 algorithm. At higher wind speeds the mean square slope given by a linear extrapolation of the *Cox and Munk* (1954) is much higher than that calculated using our highest tail. This suggests that we have either underestimated the B value at high winds, or that the *Cox and Munk* (1954) relationship does not apply well to higher winds. Unfortunately, we cannot validate/invalidate our mean square slope estimates at higher wind speeds because no direct observations exist.

1.2.2 Calculation of the Air-Sea Momentum Flux, Mean Wind Profile, and Drag Coefficient

As discussed earlier, in this study we apply two different approaches to estimate the momentum flux and drag coefficient using identical wave spectra. The first approach is identical to that proposed by DCCM. The second approach (denoted by RHG hereafter) is based on MGHBT but has been modified to account for the effect of swell and to allow different parameterized tail levels. In both approaches the momentum flux from the atmosphere to the ocean is calculated from a momentum conservation constraint. To the leading order, the momentum flux at the surface is a

sum of the viscous stress, τ_ν , and the wave form stress, τ_f .

$$\boldsymbol{\tau} = \boldsymbol{\tau}_\nu + \boldsymbol{\tau}_f \quad (1.1)$$

The wave form stress at the surface includes the impact of all waves and can be written:

$$\boldsymbol{\tau}_f = \rho_w \int_{k_{min}}^{k_{max}} \int_{-\pi}^{\pi} \beta_g(k, \theta) \sigma \Psi(k, \theta) d\theta \mathbf{k} dk \quad (1.2)$$

where ρ_w is the water density, k is the wavenumber, θ is the wave direction, σ is the angular frequency, $\beta_g(k, \theta)$ is the growth rate, $\Psi(k, \theta)$ is the wave variance spectrum, and k_{min} and k_{max} are the minimum and maximum wavenumbers of contributing waves.

In DCCM the growth rate is expressed as a function of the wind speed.

$$\beta_g(k, \theta) = A_1 \sigma \frac{[u_{\lambda/2} \cos(\theta - \theta_w) - c] |u_{\lambda/2} \cos(\theta - \theta_w) - c|}{c^2} \frac{\rho_a}{\rho_w} \quad (1.3)$$

$$A_1 = \begin{cases} 0.11, & : u_{\lambda/2} \cos \theta > c, & \text{for wind forced sea} \\ 0.01 & : 0 < u_{\lambda/2} \cos \theta < c, & \text{for swell faster than the wind} \\ 0.1 & : \cos \theta < 0, & \text{for swell opposing the wind} \end{cases} \quad (1.4)$$

where A_1 is the proportionality coefficient determined empirically (so that modeled wave spectra agree with field observations), $u_{\lambda/2}$ is the wind speed at the height of half the wavelength (up to 20 m), θ_w is the wind direction, and c is the wave phase speed. Note that the wind velocity should be taken relative to the current velocity.

The wind speed is calculated using the law of the wall for rough surfaces

$$u(z) = \frac{u_\star}{\kappa} \ln \left(\frac{z}{z_0} \right) \quad (1.5)$$

where κ is the von Kármán coefficient.

The viscous stress is calculated from the law of the wall for smooth surfaces. The viscous drag coefficient, Cd_ν is adjusted to account for sheltering:

$$Cd'_\nu = \frac{Cd_\nu}{3} \left(1 + \frac{2Cd_\nu}{Cd_\nu + Cd_f} \right) \quad (1.6)$$

where Cd_f is the wave form drag coefficient.

The viscous stress can then be solved for as:

$$\tau_\nu = \rho_a Cd'_\nu |\mathbf{u}_z| \mathbf{u}_z \quad (1.7)$$

In RHG the growth rate is calculated from the wind stress as in MGHBT. In this theory the total stress is given as a function of height as:

$$\tau = \tau_t(z) + \tau_f(z) \quad (1.8)$$

where τ_t is the turbulent stress and is equal to the viscous stress very near the surface.

The wave form stress can be expressed as

$$\tau_f(z) = \rho_w \int_{k_{min}}^{k=\delta/z} \int_{-\pi}^{\pi} \beta_g(k, \theta) \sigma \Psi(k, \theta) d\theta \mathbf{k} dk \quad (1.9)$$

that is, the wave form stress at height z is equal to the integration of the wave form stress at the surface for wavenumbers below $k = \delta/z$, where δ/k is the inner layer height (*Hara and Belcher, 2004*) for waves at a wavenumber k . This expression is derived by assuming that the wave-induced stress is significant from the surface up to the inner layer height, but is negligible further above. Since at the surface

$$\tau = \tau_\nu + \tau_f(z=0) = \tau_\nu + \rho_w \int_{k_{min}}^{k_{max}} \int_{-\pi}^{\pi} \beta_g(k, \theta) \sigma \Psi(k, \theta) d\theta \mathbf{k} dk \quad (1.10)$$

the turbulent stress at a height z can be expressed as:

$$\tau_t(z) = \tau_\nu + \rho_w \int_{k=\delta/z}^{k_{max}} \int_{-\pi}^{\pi} \beta_g(k, \theta) \sigma \Psi(k, \theta) d\theta \mathbf{k} dk \quad (1.11)$$

It is assumed that the turbulent stress at the inner layer height $z = \delta/k$ determines the growth rate of waves at wavenumber k :

$$\beta_g(k, \theta) = c_\beta \sigma \frac{|\tau_t(z = \delta/k)|}{\rho_w c^2} \cos^2(\theta - \theta_\tau) \quad (1.12)$$

where θ_τ is the direction of the turbulent stress at the inner layer height. The turbulent stress at the inner layer height is used in place of the total wind stress because longer waves reduce the effective wind forcing on shorter waves.

In MGHBT, the effect of swell (slower waves and waves opposing wind) was simply ignored and the growth rate coefficient c_β was set as:

$$c_\beta = \begin{cases} 32 & : \cos(\theta - \theta_w) > 0 \text{ and } c/u_\star^l < 1/0.07 \\ 0 & : \text{otherwise} \end{cases} \quad (1.13)$$

In this study (RHG) c_β is modified to explicitly account for the swell:

$$c_\beta = \begin{cases} 25 & : \cos(\theta - \theta_w) > 0 & : c/u_\star^l < 10 \\ 10 + 15 \cos[\pi(c/u_\star - 10)/15] & : & : 10 \leq c/u_\star^l < 25 \\ -5 & : & : 25 \leq c/u_\star^l \\ -25 & : \cos(\theta - \theta_w) < 0 \end{cases} \quad (1.14)$$

The growth rate coefficient c_β varies depending on the ratio of the wave phase speed to the local turbulent friction velocity (friction velocity at the inner layer height), $u_\star^l = \sqrt{\tau_t(z = \delta/k)/\rho_a}$. As in DCCM we define three regimes. When the wind stress direction and wave direction are within 90° of each other, we distinguish the wind forced waves ($c/u_\star^l < 10$) and the swell forcing wind ($c/u_\star^l > 25$). When wind opposes swell (wind stress direction and wave direction is misaligned by more than 90°), we assume strong wave dissipation and a large form drag as in DCCM. However, since the impact of opposing wind is not well understood, we will also test weaker forcing of opposing wind in appendix B. Finally, between the wind forced wave regime and the swell forcing wind regime, we introduce a transition regime ($10 < c/u_\star^l < 25$) with a smoothly varying c_β that compares well to the data presented in *Belcher* (1999).

In DCCM the mean wind profile is assumed to be logarithmic, that is, the feedback of the waves appears only in the modified effective roughness length. In RHG the wind profile is explicitly calculated using the energy conservation constraint in the wave boundary layer following MGHBT. From the top of the viscous sublayer to

the inner layer height of the shortest waves the wind shear is expressed as:

$$\frac{d\mathbf{u}}{\partial z} = \frac{\rho_a}{\kappa z} \left| \frac{\boldsymbol{\tau}_\nu}{\rho_a} \right|^{3/2} \frac{\boldsymbol{\tau}_\nu}{\boldsymbol{\tau}_\nu \cdot \boldsymbol{\tau}_{tot}} \quad \text{for } z_\nu < z < \delta/k_l \quad (1.15)$$

Between the inner layer height of the shortest waves and that of the longest waves the wind shear is expressed as:

$$\frac{d\mathbf{u}}{\partial z} = \left[\frac{\delta}{z^2} \tilde{F}_w \left(k = \frac{\delta}{z} \right) + \frac{\rho_a}{\kappa z} \left| \frac{\boldsymbol{\tau}_t(z)}{\rho_a} \right|^{3/2} \right] \times \frac{\boldsymbol{\tau}_t(z)}{\boldsymbol{\tau}_t(z) \cdot \boldsymbol{\tau}_{tot}} \quad \text{for } \delta/k_l \leq z \quad (1.16)$$

where $\tilde{F}_w(k = \delta/z)$ is the energy uptake by surface waves:

$$\tilde{F}_w(k = \delta/z) = \rho_w \int_{-\pi}^{\pi} \beta_g(k, \theta) g \Psi(k, \theta) k d\theta \quad (1.17)$$

Finally, above the inner layer height of the longest waves the wave effect is negligible and the wind shear is aligned in the direction of the wind stress:

$$\frac{d\mathbf{u}}{dz} = \frac{u_\star}{\kappa z} \frac{\boldsymbol{\tau}_{tot}}{|\boldsymbol{\tau}_{tot}|} \quad (1.18)$$

Although the inner layer height parameter δ is estimated to be around 0.05-0.1 (*Hara and Belcher, 2004*), its exact value is not known. In MGHBT this parameter δ was effectively treated as a tuning parameter and its value was determined to match the resulting drag coefficient at low to medium wind speeds with existing empirical parameterizations. In this study we also determine δ in the same empirical manner and set $\delta = 0.03$. Note that there are some uncertainties in the value of the growth parameter c_β as well. If the value of c_β is changed from those in (1.14), the value of δ needs to be modified to obtain similar drag coefficient values.

Let us summarize the major differences between DCCM and RHG. One major difference that has already been mentioned is the calculation of the growth rate. DCCM parameterizes the growth rate from the wind speed, and RHG parameterizes the growth rate from the wind stress. Another more subtle difference is the directional dependence of the growth rate. The method of DCCM takes the projection of the

wind in the direction of the waves and uses the difference between the two values to calculate the growth rate. This means waves propagating perpendicular ($\pm 90^\circ$) to the wind will have an impact on the wind stress calculation because the phase speed of the waves is not zero. This is quite different from the method of RHG where the cosine squared of the angle between waves and wind stress is used to determine the growth rate. This value is 0 for waves propagating perpendicular to the wind. The impact of fast propagating swell misaligned at 90° to the wind is very different between the two models. Note that the value of A_1 in the DCCM model varies from 0.01 to 0.1 as the angle difference between wind and waves exceed 90° . Therefore, swell misaligned with wind by slightly more than 90° has a large impact on the wind stress calculation; such waves are particularly effective in increasing the misalignment between the wind stress direction and the wind speed direction as demonstrated later.

The breaking wave impacts on the wave form drag are not explicitly calculated in either of the theories. In RHG it is assumed that the breaking form drag of peak waves is not of the leading order (after *Kukulka and Hara, 2008a*). The breaking effect is implicitly included in the high frequency tail because the saturation spectrum value (B) is likely limited by wave breaking process. Furthermore, if the form drag of the high frequency tail is enhanced due to breaking, their effect can be accounted for by slightly raising the B value without modifying the approach. In DCCM the growth rate coefficients have been determined to match observations in the North Sea and under hurricane conditions. Therefore, their coefficients should represent the mean impact of both breaking and non-breaking waves.

The calculation of the wind profile between the two methods is another major difference. The method of DCCM does not explicitly consider energy conservation in the wave boundary layer, and assumes the wind profile to follow a logarithmic law of the wall profile. There is a feedback on the surface roughness due to the

wave form drag, but the directions of the wind is fixed (in z), that is, the direction of the wind shear is fixed and it can be misaligned with the wind stress direction (which is also constant in z) at all heights. This has a significant effect on the stress calculation, particularly when wind and waves are misaligned. The method of RHG considers energy conservation inside the wave boundary layer. It also assumes that the wind shear and the turbulent stress are aligned at all heights. Therefore, the wind speed vector can turn (change directions) with height inside the wave boundary layer. Since the wind shear and the wind stress are aligned above the wave boundary layer, misalignment between the wind direction and the wind stress direction (if it exists at the top of the wave boundary layer) decreases with height. As shown later, this explicit calculation of the wind profile leads to misalignment angles at 10 meters that are about half as large as those at the top of the wave boundary layer.

1.2.3 Experimental Design

We conduct two numerical experiments. The first experiment is a fetch-dependent simulation under a stationary and uniform wind over a large computational domain. The second experiment is an idealized tropical cyclone that is translated across a computational domain at a fixed speed.

Experiment A: Fetch-dependent Simulation

The domain is 3,000 km in the direction of the wind and 1,800 km in the direction normal to the wind. The depth is uniformly 4 km at all locations to maintain deep water conditions. The wave simulation is run for 72 hours so that the wave field reaches a steady state. The wind stress is calculated along the central transect in the wind direction where the fetch increases with distance. The experiment is conducted for wind speeds ranging from 5 m/s to 50 m/s in increments of 5 m/s. We present data up to 400 km fetch, since the effective fetch under strong wind usually does not

exceed a few hundred kilometers in typical tropical cyclone conditions.

Experiment B: Idealized Tropical Cyclone Simulation

The second experiment is an idealized tropical cyclone simulation where an axisymmetric tropical cyclone with the *Holland* (1980) wind profile is translated across the same deep-water domain described in Experiment A. The storm is prescribed with a radius of maximum wind of 70 km and a maximum wind speed of 45 m/s. It translates through the domain for 72 hours so that the wave field becomes steady state in the reference frame of the translating storm. The wind and resulting wave field for a 5 m/s and a 10 m/s translating storm are shown in Figure 1.3. The front right of the storm is exposed to prolonged forcing from wind that produces higher, longer, and older waves because the swell field propagates with the storm (resonance effect). The rear left side of the storm generally produces lower, shorter, and younger waves where the swell, wind, and translating storm vectors can be significantly misaligned. Swell generated some time earlier in the front right quadrant of the storm propagates into the left half of the storm at later times and creates conditions of large misalignment between the swell field and the wind direction. The point where the swell intersects the moving storm varies depending on the translation speed of the storm. All these features have been documented in the previous observational studies (see *Young*, 2003).

1.3 Results

1.3.1 Experiment A (Fetch-dependent)

The drag coefficients calculated in the fetch dependent experiment are presented in Figure 1.4. The results using RHG are shown in the left column and those using DCCM in the right column. The overall results show that the drag coefficient is very sensitive to the choice of the saturation level in the tail and that it is not as sensitive to the different approaches of the drag coefficient calculation (i.e., between

RHG and DCCM). As discussed earlier, if the saturation level B is fixed, neither approach reproduces the COARE 3.5 trend from 5 to 25 m/s. At low (5 m/s) winds the lowest tail level ($B = 2 \times 10^{-3}$) yields the most consistent drag coefficient, but at high (25 m/s) winds the highest tail level ($B = 12 \times 10^{-3}$) yields the values closest to the COARE 3.5 drag coefficient. It is interesting to note that the middle level ($B = 6 \times 10^{-3}$) seems to yield the drag coefficient trend that is consistent with *Large and Pond* (1981) parameterization from 5 to 18 m/s. In all cases the drag coefficient continues to increase with the wind speed if the tail level is kept unchanged. This suggests that the drag coefficient can saturate (cease to increase) or decrease with increasing wind if the tail level decreases with increasing wind. One noticeable difference between RHG and DCCM is that the DCCM drag coefficient is smaller for the highest wind speeds with the highest tail level. This is an indication that RHG drag coefficient is more sensitive to the tail level.

Let us next focus on the sea-state dependence of the drag coefficient. The sea-state dependence of the drag coefficient is displayed by color coding the data in terms of the input wave age (c_{pi}/u_* , where c_{pi} is the phase speed at the wind-sea peak frequency). At a fixed wind speed the drag coefficient varies as the fetch increases from 50km to 400km. The most significant dependence is observed with DCCM at the highest wind speed and the lowest saturation value ($B = 2 \times 10^{-3}$), where the drag coefficient decreases by about 50% as the sea develops. As the tail level increases, the DCCM results show less wave age dependence. With RHG the sea-state dependence is not as large with the lowest saturation level ($B = 2 \times 10^{-3}$). Interestingly, as the tail level increases, the sea-state dependence (input wave age) reverses; the older seas yield larger drag coefficients with the highest saturation level ($B = 12 \times 10^{-3}$). This reversal happens because the older waves have lower peak input frequencies. As explained earlier, the wave spectrum transitions from the explicit model result to

the parameterized tail between $1.25\times$ and $3\times$ the peak input frequency. Therefore, older waves adjust to the tail level at a lower frequency. This means that the drag coefficient of older waves is more dependent on the tail level and less dependent on the spectral peak. Thus, with a high tail level (at the same wind speed) the older waves yield higher drag coefficient values. Conversely, when attaching a low tail level, the older waves yield lower drag coefficient values.

1.3.2 Experiment B (Idealized Tropical Cyclone)

The tropical cyclone experiments contain wave fields where the swell and wind vector are no longer aligned leading to more complex solutions. The calculated angle between the wind vector and the wind stress vector is an important result from these simulations. In general, the largest waves are seen on the right front side of the tropical cyclone where the storm translation speed, wind vector, and wave direction are all in the same direction. The youngest seas are found in the rear and left of the storm where the wind and translation direction are against each other, and the dominant wave direction can be highly misaligned. Note that even if the 10 meter wind vector and the wind stress vector are misaligned, we have calculated the drag coefficient as a ratio of the friction velocity squared and the 10 meter wind speed squared.

As in Experiment A, the drag coefficient value is overall very sensitive to the tail level attached and is not as sensitive to the approaches of the stress calculation (Figure 1.5-1.8). The sea-state dependence of the drag coefficient for a 5 m/s translating tropical cyclone is comparable to that for growing seas in Experiment A, but it is significantly enhanced when the translation speed increases from 5 m/s (Figures 1.5, 1.7) to 10 m/s (Figures 1.6, 1.8). The largest increase in sea-state dependence occurs with the lowest tail level. Specifically, as the translation speed increases from 5 m/s to 10 m/s, the range of drag coefficient at wind speed 40 m/s increases from

roughly $1.3 - 1.6 \times 10^{-3}$ to $1.1 - 2 \times 10^{-3}$ in RHG, and from $1.4 - 1.9 \times 10^{-3}$ to $1.2 - 2.3 \times 10^{-3}$ in DCCM. This is mainly because waves that propagate against the wind (counter swell) on the rear left of the storm center have a notable impact inside the radius of maximum wind. The presence of such a region is strongly dependent on the translation of the storm and the propagation of the swell. The counter swell effect is not as strong with the 5 m/s translating storm because the swell field does not intersect the storm track at the same location. This sensitivity of the swell field to the translation speed is consistent with the modeling results of *Moon et al.* (2003). The effect of the counter swell is more pronounced in the results of DCCM than those of RHG. Consequently, the location of the maximum drag coefficient moves further to the rear right in the DCCM results (Figures 1.5, 1.6).

We next examine the misalignment angle between the 10 meter wind speed vector and the wind stress vector (Figures 1.9-1.12). There are significant differences between the RHG results and the DCCM results. The misalignment angle is generally small (up to a few degrees) in the RHG results, but it is significantly larger (exceeding 5° very close to the storm center as well as very far from the storm center) in the DCCM results. The misalignment is more enhanced when the wind speed is lower (Right panels of Figures 1.11, 1.12). The misalignment is also enhanced to the left of the storm (Right panels of Figure 1.9, 1.10) likely because misaligned swell is present there.

The significant difference of the wind stress misalignment angle predictions between RHG and DCCM is caused by two major differences in the two methods. The first difference is in the estimations of the mean wind profile. As discussed earlier, DCCM assumes that the wind direction does not change with height. Therefore, the misalignment angle between the wind and wind stress is also independent of height (at least up to 10 meter height). However, RHG imposes that the wind shear is in

the same direction as the turbulent stress at all heights. Consequently, above the top of the wave boundary layer (outside the direct wave effects) the wind shear is aligned with the wind stress, that is, the misalignment angle decreases with height. In fact, we have found that the misalignment angle is typically about half at 10 meter height compared to that at the top of the wave boundary layer (which is typically 1.5 m).

The second and more significant difference between RHG and DCCM is the directionality of the growth rate and the impacts of cross-wind swell. Because the growth rate of RHG has a cosine squared dependence on the angle between the wind stress and the waves, cross-wind swell has essentially no impact. DCCM calculates the growth rate based on the difference between the wind speed projected in the wave direction and the wave phase speed (from equation 1.3). If the wind and waves are misaligned by around 90° the growth rate approaches

$$\beta_g(k, \theta) \rightarrow A_1 \sigma \frac{-(c^2)}{c^2} \frac{\rho_a}{\rho_w} \quad (1.19)$$

instead of zero. In particular, if the misalignment is slightly larger than 90° , the coefficient A_1 is as large as that for the strongly forced wind seas (see equation 1.4). Since the form drag of the cross-wind swell applies in the direction of the swell (perpendicular to the wind direction), these waves are very effective in turning the wind stress direction. (Note that swell that propagates against the wind may increase the wind stress and the drag coefficient but it does not turn the wind stress direction.)

There are a few previous studies addressing the misalignment between the 10 meter wind vector and the wind stress vector (*Geernaert, 1988; Drennan et al., 1999; Grachev et al., 2003; Zhang et al., 2009*). While most of these studies only demonstrated statistically significant misalignment at low wind speeds (< 5 m/s) (*Geernaert, 1988; Drennan et al., 1999; Grachev et al., 2003*), recent observations by *Zhang et al. (2009)* suggest that certain conditions (such as the presence of strong horizontal current shears that can turn waves near the peak) can result in stress angles

misaligned by more than 20° from the wind direction for wind speeds up to 15 m/s. These results have not been observed over a wide range of wave conditions, so their implications for hurricane conditions are not clear. Therefore, it is difficult to validate our model results against existing observations. Nevertheless, this study clarifies how different assumptions in the drag coefficient calculations yield very different results of the wind stress misalignment.

Several studies have attempted to observe the drag coefficient under tropical cyclones (*Powell et al.*, 2003; *Black et al.*, 2007; *French et al.*, 2007; *Jarosz et al.*, 2007; *Holthuijsen et al.*, 2012). *Holthuijsen et al.* (2012) reports drag coefficients for 20 m/s winds that are about 1×10^{-3} , which is much lower than the COARE 3.5 algorithm (*Edson et al.*, 2013)(2.4×10^{-3}) and *Large and Pond* (1981)(1.79×10^{-3}). Results from the other studies (*Powell et al.*, 2003; *Black et al.*, 2007; *French et al.*, 2007; *Jarosz et al.*, 2007) show mean values at 20 m/s more comparable to the values of *Large and Pond* (1981) but are still lower than the COARE 3.5 algorithm. At higher wind speeds the observational studies generally agree that the drag coefficient saturates or is reduced, but the wind speed at which this saturation/reduction occurs is variable (30 – 50 m/s). As mentioned previously our model results do not show saturation/reduction of the drag coefficient in either method unless the tail level is reduced with increasing wind speed.

A couple of studies (*Black et al.*, 2007; *Holthuijsen et al.*, 2012) have explored the dependence of the drag coefficient based on the location relative to the storm translation. *Black et al.* (2007) presented observations from 3 of the 4 tropical cyclone quadrants and saw no systematic dependence of the drag coefficient. *Holthuijsen et al.* (2012), however, found a strong dependence of the drag coefficient on the location relative to the storm track. They classified their observations into three regions (left-front third, right-front third, and rear third, relative to the storm center). The

results suggest that at wind speed about 20 m/s the drag coefficient varies from 0.7×10^{-3} in the left-front third to 1.3×10^{-3} in the rear third. Their results at wind speed about 35 m/s show the drag coefficient varies from about 1.8×10^{-3} in the right-front third to 5.2×10^{-3} in the left-front third, a difference of almost 300%. In our model calculations, if the tail level is fixed or is a function of wind speed only, the variability of the drag coefficient at a fixed wind speed is at most 50 – 100%. Even if B is allowed to be sea-state dependent, it is not likely that the model results yield such large systematic variations of the drag coefficient as seen by *Holthuijsen et al.* (2012). They have attributed the large variability to the presence of cross-wind swells and counter swells. Although our model results show some sensitivity of the drag coefficient to cross-wind and counter swells, the impact of swell is never as large as the study of *Holthuijsen et al.* (2012) suggests. Furthermore, for wind speeds up to 25 m/s the model results predict that cross-wind swell and counter swell tend to increase the drag coefficient, which is opposite to the trend reported by *Holthuijsen et al.* (2012). The increase in the drag coefficient due to cross-swell for wind speeds 25-45 m/s in our model results is much less than the increase reported by *Holthuijsen et al.* (2012).

1.4 Conclusion

The sea-state dependence of the air-sea momentum flux and the drag coefficient was studied with simple fetch-dependent wave fields under uniform wind and with complex wave fields under hurricane wind conditions. Two very different approaches (RHG and DCCM) were used to calculate the wind stress and the mean wind profile. Using existing observations as guidance, a range of spectral saturation levels at high frequencies (spectral tail) were tested in the wind stress calculations.

The most important finding of this study is that the drag coefficient is very sensitive to the spectral saturation level but is not as sensitive to the wind stress

calculation approaches. If the saturation level is taken to be a function of wind speed, wave age, or both, then the drag coefficient can be constructed to match any observations or empirical parameterizations. This result supports the conclusion of *Makin et al.* (1995) and others that the waves between centimeters to meters scale have significant contributions to the air-sea momentum flux. If the spectral saturation level is assumed to remain constant at higher wind speeds, the drag coefficient continues to increase with increasing wind. Saturation or reduction of the drag coefficient at very high wind speeds occurs only if the saturation level decreases with increasing wind speed within the framework of our model study. It is possible that presence of sea sprays and sea foam, which is not considered in this study, may contribute to reducing the drag coefficient at very high wind speeds. Airflow separation, as discussed in *Donelan et al.* (2006) may also play a role in the reduction/modification of the drag.

Although both RHG and DCCM methods yield similar drag coefficient values, the results of the misalignment angle between the 10 meter wind speed vector and the wind stress vector in the tropical cyclone experiments are very different between the two methods. While the wind stress based growth rate parameterization of RHG prevents cross-wind swell (waves that are propagating perpendicular to the wind) from having a large impact on the wind stress, the wind speed based growth rate parameterization of DCCM introduces a significant contribution from cross-wind swell to the wind stress and increases the misalignment of the wind stress vector and wind speed vector. (The stress supported by cross-swell does not significantly alter the wind stress magnitude, but modifies the wind stress direction.) The 10 m/s translation speed tropical cyclone consistently gives more misalignment than the 5 m/s translating tropical cyclone.

The sea-state dependence of the drag coefficient is sensitive to the tail level.

In particular, with the RHG method the dependence of the drag coefficient on the wave age reverses as the tail level increase from the lowest level to the highest level studied. The results also show that the fetch-dependent seas and the 5 m/s translating tropical cyclone give comparable sea-state dependence, while the 10 m/s translating tropical cyclone yields a much larger sea-state dependence. The sea-state dependence is enhanced to the left of the storm track, particularly because of the presence of swell that is uncorrelated with the local wind. With the lowest tail level tested ($B = 2 \times 10^{-3}$) both RHG and DCCM show that the drag coefficient can vary by as much as 100% at wind speed 40 m/s for a 10 m/s translation speed tropical cyclone. More typically, our results show variability of the drag coefficient less than 50% at a given wind speed for a fixed tail level.

Our modeling results (both RHG and DCCM) are generally not consistent with the observational study of *Holthuijsen et al.* (2012). The magnitude of the sea-state dependence of our models is significantly smaller than that reported by *Holthuijsen et al.* (2012). Furthermore, for wind speeds up to 25 m/s the model results predict that cross-wind swell and counter swell tend to increase the drag coefficient, which is opposite to the trend reported by *Holthuijsen et al.* (2012). The increase in the drag coefficient due to cross-swell for wind speeds 25-45 m/s in the model results is much less than the increase reported by *Holthuijsen et al.* (2012).

The impact of sea-state dependent air-sea momentum flux and the drag coefficient (including the misalignment between wind and wind stress) will likely have an impact on the upper ocean mixing and resulting sea surface cooling, which will in turn have an impact on the strength of the storm. These impacts can be evaluated more thoroughly with the help of fully coupled atmosphere-wave-ocean models. Because the wind stress also serves as a bottom boundary condition in the atmospheric model, it will likely impact the tropical cyclone dynamics as well.

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Figure 1.1. Directionally integrated saturation spectrum with the peak of the wave field simulated in WW3. The first dashed vertical line (left to right) represents the peak input frequency, the second is $1.25 \times$ the peak input frequency, and the third line is $3 \times$ the peak input frequency. The original WW3 spectrum, *Elfouhaily et al. (1997)*'s empirical spectrum, and the three tail level options tested during this experiment are plotted. The fetch is 100 km and the wind speed is 10 m/s (a) and 40 m/s (b).

Figure 1.2. Mean square slope with the 3 different tail level options. The empirical linear relationships of *Cox and Munk* (1954) are plotted for reference up to wind speed 15 m/s (black), and are extrapolated to 55 m/s (gray). At each wind speed, the upper two data points are from the B=0.012 tail, the middle two data points are from the B=0.006 tail, and the lower two data points are from the B=0.002 tail.

Figure 1.3. Wind used in the experiment and resulting wave field. The upper plots (a,b) are for a 5 m/s translating tropical cyclone while the lower plots (c,d) are for a 10 m/s translating tropical cyclone. Wind vectors are superimposed in the wind contour plots. The thick black line represents the track of the tropical cyclone through the domain.

Figure 1.4. Drag coefficients ($\times 1000$) for fetch dependent simulations using 3 tail levels. Eleven simulations were conducted from wind speed 5 to 50 m/s. The left column (a,c,e) is calculated using the RHG method, while the right column (b,d,f) is calculated using the DCCM method. The saturation level is $B=0.002$ (a,b), $B=0.006$ (c,d), and $B=0.012$ (e,f).

Figure 1.5. Drag coefficients ($\times 1000$) for a 5 m/s translating tropical cyclone. The left column (a,c,e) is calculated using the RHG method, while the right column (b,d,f) is calculated using the DCCM method. The saturation level is $B=0.002$ (a,b), $B=0.006$ (c,d), and $B=0.012$ (e,f). The thick black line represents the track of the tropical cyclone through the domain.

Figure 1.6. The same as Figure 1.5, but for a 10 m/s translating tropical cyclone.

Figure 1.7. Drag coefficient ($\times 1000$) vs wind speed for a 5 m/s translating tropical cyclone. The left column (a,c,e) is calculated using the RHG method, while the right column (b,d,f) is calculated using the DCCM method. The saturation level is $B=0.002$ (a,b), $B=0.006$ (c,d), and $B=0.012$ (e,f).

Figure 1.8. The same as Figure 1.7, but for a 10 m/s translating tropical cyclone.

Figure 1.9. Misalignment angles (10-meter wind direction minus surface stress direction) for a 5 m/s translating tropical cyclone. The left column (a,c,e) is calculated using the RHG method, while the right column (b,d,f) is calculated using the DCCM method. The saturation level is $B=0.002$ (a,b), $B=0.006$ (c,d), and $B=0.012$ (e,f). The thick black line represents the track of the tropical cyclone through the domain.

Figure 1.10. The same as Figure 1.9, but for a 10 m/s translating tropical cyclone.

Figure 1.11. Misalignment angle (10-meter wind direction minus surface stress direction) vs wind speed for a 5 m/s translating tropical cyclone. The left column (a,c,e) is calculated using the RHG method, while the right column (b,d,f) is calculated using the DCCM method. The saturation level is $B=0.002$ (a,b), $B=0.006$ (c,d), and $B=0.012$ (e,f).

Figure 1.12. The same as Figure 1.11, but for a 10 m/s translating tropical cyclone.

Langmuir Turbulence Parameterization in Tropical Cyclone Conditions

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Abstract

The Stokes drift of surface waves significantly modifies the upper-ocean turbulence because of the Craik-Leibovich vortex force (Langmuir turbulence). Under tropical cyclones the contribution of the surface waves varies significantly depending on complex wind and wave conditions. Therefore, turbulence closure models used in ocean models need to explicitly include the sea-state dependent impacts of the Langmuir turbulence. In this study the K Profile Parameterization (KPP) first moment turbulence closure model is modified to include the explicit Langmuir turbulence effect, and its performance is tested against equivalent Large Eddy Simulation (LES) experiments under tropical cyclone conditions. First, the KPP model is retuned to reproduce LES results without Langmuir turbulence to eliminate implicit Langmuir turbulence effects included in the standard KPP model. Next, the Lagrangian currents are used in place of the Eulerian currents in the KPP equations that calculate the bulk Richardson number and the vertical turbulent momentum flux. Finally, an enhancement to the turbulent mixing is introduced as a function of the non-dimensional turbulent Langmuir number. The retuned KPP, with the Lagrangian currents replacing the Eulerian currents and the turbulent mixing enhanced, significantly improves prediction of upper-ocean temperature and currents compared to the standard (unmodified) KPP model under tropical cyclones, and shows improvements over the standard KPP at constant moderate winds (10 m s^{-1}).

2.1 Introduction

Tropical cyclone prediction models require sea surface temperature and currents to accurately compute air-sea heat and momentum fluxes. These air-sea fluxes are the primary contributors to the energy budget of a tropical cyclone and therefore greatly affect the storm intensity (see Emanuel, 1991, 1999). Tropical cyclones are known to generate vigorous responses in the upper-ocean, which include surface waves with amplitudes larger than 15 m, upper-ocean currents in excess of 1 m s^{-1} , and sea surface temperature cooling of several degrees (Ginis, 2002). The surface temperature and current responses to wind forcing are determined by turbulent mixing throughout the upper-ocean boundary layer, which must be carefully accounted for in ocean models that are coupled to tropical cyclone forecast models.

As a hurricane passes over a particular location, the upper-ocean mixing develops in the following manner. First, near-surface ocean currents and surface waves increase as momentum from wind is transferred into the ocean. The large vertical shear below the developing surface current generates turbulence and drives deepening of the active surface mixing-layer. Surface gravity waves also contribute to upper-ocean mixing and the turbulent kinetic energy (TKE) budget through breaking and the Stokes drift; the latter significantly modifies upper-ocean turbulence (Langmuir turbulence). The near-surface temperature cools due to entrainment of cold water from the thermocline below as the mixing-layer deepens (Price, 1981). Although these are primarily one-dimensional (vertical) mixing processes, the cool-water entrainment under a tropical cyclone can be further enhanced by three-dimensional processes, notably by upwelling due to Ekman pumping (Yablonsky and Ginis, 2009). Evaporation is another source of surface cooling, although this is generally a second-order process during strong winds and active entrainment (Ginis, 2002).

In three-dimensional ocean circulation models that utilize Reynolds-averaged

equations of motion, the upper-ocean turbulent fluxes are typically parameterized by closure models. The traditional turbulence closure models (e.g. Mellor and Yamada, 1982; Large et al., 1994) determine the evolution of the mixing-layer turbulence through inputs including the wind stress τ (or the friction velocity, $u_* = \sqrt{\tau/\rho_a}$, where ρ_a is air density) and the surface buoyancy flux B_* . These closure models do not explicitly capture the contribution of surface gravity waves to the upper-ocean turbulence. One source of upper-ocean turbulence due to surface gravity waves is the injection of TKE from wave breaking to the ocean (Melville, 1996). The elevated turbulence due to breaking waves decays within depths scaling with the significant wave height (i.e. Craig and Banner, 1994; Terray et al., 1996). Another significant source of turbulence derived from surface waves is the Langmuir turbulence. The depth scale of the Langmuir turbulence is typically larger and is determined by the mixed-layer depth and/or the Stokes drift e-folding depth (Harcourt and D’Asaro, 2008; Grant and Belcher, 2009; Sullivan et al., 2012).

First observed by Langmuir (Langmuir, 1938), it took several decades for the mechanism that drives Langmuir circulations to be identified. This is the Craik-Leibovich (CL) vortex-force, which results from interaction between the Stokes drift of surface waves and the upper-ocean Eulerian-current vorticity (Craik and Leibovich, 1976). In the high Reynolds number planetary boundary layer turbulence, the Langmuir turbulence (turbulence that is modified by the CL vortex-force) exists over scales ranging from $O(1)$ m to the mixed-layer depth, even when not organized as coherent Langmuir circulations (McWilliams et al., 1997). Langmuir turbulence has been extensively studied for the last two decades using Large Eddy Simulation (LES) models that can resolve explicitly the dominant scales of Langmuir turbulence (Skylkingstad and Denbo, 1995; McWilliams et al., 1997; Skylkingstad et al., 2000; Noh et al., 2004; Min and Noh, 2004; Polton and Belcher, 2007; Sullivan et al., 2007; Harcourt and

D’Asaro, 2008; Grant and Belcher, 2009; Kukulka et al., 2009; Teixeira and Belcher, 2010; Kukulka et al., 2010; Noh et al., 2011; Van Roekel et al., 2012). All support the conclusion that the CL vortex-force strengthens upper-ocean mixing in many different regimes. Studies that qualitatively and quantitatively compare LES results to in-situ data confirm that including the CL vortex force improves the model-observation agreement (Kukulka et al., 2009; D’Asaro et al., 2014).

LES has also been used to study stochastic momentum injection via wave breaking in the presence of Langmuir turbulence in the planetary boundary layer (Noh et al., 2004; Sullivan et al., 2007; McWilliams et al., 2012). These studies show that the impact of intermittent breaking momentum injection is mostly limited to the upper several meters when compared to constant surface momentum fluxes. Therefore, the CL vortex-force (which impacts scales of 10-200 m) is a much more effective mechanism to enhance mixed-layer deepening. Various scalings for the enhancement to the turbulence closure models based on the Langmuir number have been proposed, which will be discussed in detail later (Sections 2.2 and 2.8).

Sullivan et al. (2012) have employed LES to study Langmuir turbulence under strong, transient wind stress and Stokes drift conditions characteristic of tropical cyclones. This idealized study shows large variation in Langmuir turbulence from the right to the left side of the storm track. The right side of the storm (N. hemisphere convention) has inertially resonant wind and current rotation, which significantly deepens the mixing-layer compared to the non-resonant left side (Price, 1981; Skillingstad et al., 2000; Sanford et al., 2011). The right side also has larger waves due to resonance between the wind and the storm translation (Young, 2003; Moon et al., 2003), which results in a stronger, deeper Stokes drift profile (Sullivan et al., 2012).

Recently, Rabe et al. (2015) investigated Langmuir turbulence under Hurricane

Gustav (2008) by combining in-situ measurements of turbulence obtained by Lagrangian floats deployed ahead of the storm, and LES hindcasts of the upper-ocean turbulence at the float locations. The observations show a regime behind the eye of a tropical cyclone where the turbulent vertical velocity variance ($\overline{w'^2}$) is reduced compared to traditional wind-driven (u_*) scaling. The LES results suggest that such a regime may be due to large variability of the Langmuir turbulence, since the local wave field varies significantly (both in time and in space) under a tropical cyclone. Comparisons between the turbulent quantities from the LES and the observations are imperfect for various reasons, particularly because of uncertainty of the drag coefficient in tropical cyclone conditions (see Rabe et al., 2015). Nevertheless, the results suggest that the Langmuir turbulence is an important, spatially variable source of mixing under a tropical cyclone, and that the additional mixing due to the CL vortex-force is necessary to best match the simulations and the field observations.

To improve the comparison between models containing the Langmuir turbulence and in-situ data, large-scale three-dimensional processes such as upwelling and advection must be included. However, such models are computationally expensive, and cannot be run at resolutions that are needed to resolve the dominant Langmuir turbulence scale. Therefore, turbulence closure models that parameterize Langmuir turbulence must be developed and be included in coarse resolution ($O(10^3\text{-}10^4)$ m) basin-scale numerical ocean models. Such an approach would contribute to improved performance of coupled hurricane-wave-ocean simulation/prediction models.

The main objective of this study is to develop a robust turbulence closure model that accurately accounts for the Langmuir turbulence effects under tropical cyclone conditions. Such a model is constructed and tested using an extensive set of LES Langmuir turbulence experiments under a wide range of tropical cyclone wind and wave conditions.

2.2 Geophysical Turbulence Closure Models

In typical ocean models based on Reynolds Averaged Navier-Stokes (RANS) equations, the horizontal momentum equations, with the hydrostatic and Boussinesq approximations, can be expressed as:

$$\frac{\partial \mathbf{U}_h}{\partial t} + \left(\mathbf{U}_h \cdot \frac{\partial}{\partial \mathbf{x}_h} + W \frac{\partial}{\partial z} \right) \mathbf{U}_h + \mathbf{f} \times \mathbf{U}_h = -\frac{1}{\rho_0} \frac{\partial P}{\partial \mathbf{x}_h} - \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{U}_h}{\partial z} + \overline{\mathbf{u}_h' w'} \right) \quad (2.1)$$

where $\mathbf{x}_h$ are the horizontal coordinates, z is the vertical coordinate (positive upward), $\mathbf{f}$ is the Coriolis parameter (vector), ν is the molecular viscosity, and ρ_0 is the reference density. The instantaneous horizontal velocity vector $\mathbf{u}_h$ and vertical velocity w are decomposed into the Reynolds mean $(\mathbf{U}_h, W)$ and fluctuation $(\mathbf{u}_h', w')$, and P is the Reynolds averaged dynamic pressure. The overbar represents the Reynolds mean and $\overline{\mathbf{u}_h' w'}$ is the vertical turbulent momentum flux. The horizontal turbulent fluxes have been neglected. The temperature advective-diffusive equation can be expressed as:

$$\frac{\partial \Theta}{\partial t} + \mathbf{U}_h \cdot \frac{\partial \Theta}{\partial \mathbf{x}_h} + W \frac{\partial \Theta}{\partial z} = -\frac{\partial}{\partial z} \left(\nu_\theta \frac{\partial \Theta}{\partial z} + \overline{\theta' w'} \right) \quad (2.2)$$

where the potential temperature θ is similarly decomposed into the Reynolds mean Θ and fluctuation θ' , and ν_θ is the molecular diffusion of heat. Again, horizontal turbulent fluxes have been dropped.

In the presence of the wave-induced Stokes drift profile $\mathbf{u}_S$, the advective and planetary terms in the momentum equation contain the Stokes drift (following McWilliams and Restrepo, 1999):

$$\begin{aligned} \frac{\partial \mathbf{U}_h}{\partial t} + \left(\mathbf{U}_h \cdot \frac{\partial}{\partial \mathbf{x}_h} + W \frac{\partial}{\partial z} \right) \mathbf{U}_h + \mathbf{f} \times (\mathbf{U}_h + \mathbf{u}_S) + \boldsymbol{\omega} \times \mathbf{u}_S = \\ -\frac{1}{\rho_0} \frac{\partial \tilde{P}}{\partial \mathbf{x}} - \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{U}_h}{\partial z} + \overline{\mathbf{u}_h' w'} \right) \end{aligned} \quad (2.3)$$

where $\boldsymbol{\omega} = \nabla \times \mathbf{U}_h$ is the vertical (relative) vorticity vector, the advective Stokes drift component is represented through the vortex force $\boldsymbol{\omega} \times \mathbf{u}_S$, the planetary rotation term is modified to include the Stokes drift (the Coriolis-Stokes term), and the dynamic

pressure $\tilde{P}$ has been modified to include the Stokes drift correction. Similarly, the Stokes drift is introduced to the advective terms of the temperature equation:

$$\frac{\partial \Theta}{\partial t} + (\mathbf{U}_h + \mathbf{u}_S) \cdot \frac{\partial \Theta}{\partial \mathbf{x}_h} + W \frac{\partial \Theta}{\partial z} = -\frac{\partial}{\partial z} \left(\nu_\theta \frac{\partial \Theta}{\partial z} + \overline{\theta' w'} \right). \quad (2.4)$$

The Stokes drift should be calculated from the full wave spectra, which are prescribed based on observations or surface wave models. Therefore, the only components of the system that cannot be solved for by the horizontal momentum equation (2.3), the temperature-advection equation (2.4), the equation of state, and the continuity equation are the vertical turbulent flux terms. A turbulence closure model is used to calculate these terms, which can be expressed as:

$$\overline{\mathbf{u}_h' w'} = -K_M \frac{\partial \mathbf{U}_h}{\partial z} + \Gamma_U \quad (2.5)$$

$$\overline{\theta' w'} = -K_\theta \frac{\partial \Theta}{\partial z} + \Gamma_\theta \quad (2.6)$$

where the vertical turbulent flux of a property is equal to the vertical gradient of the mean property multiplied by an eddy viscosity K_M (or an eddy diffusivity K_θ for temperature), plus a non-local (or counter-gradient) term Γ . The K Profile Parameterization (KPP, Large et al., 1994) is one example of a turbulent closure model for geophysical applications that solves for the turbulent fluxes following this form. Recently, several studies (McWilliams and Sullivan, 2000; Smyth et al., 2002; McWilliams et al., 2012, 2014; Sinha et al., 2015) proposed modifications to the KPP model that include the Langmuir turbulence.

In McWilliams and Sullivan (2000) the Langmuir turbulence is included via an enhancement factor to the eddy viscosity (and diffusivity) through an enhanced turbulent velocity scale. The enhancement factor is parameterized from the turbulent Langmuir number:

$$La_t = \sqrt{\frac{u_*}{|\mathbf{u}_S(z=0)|}} \quad (2.7)$$

where $|\mathbf{u}_S(z = 0)|$ is the magnitude of the surface Stokes drift. This approach was later modified by Smyth et al. (2002) to include dependency on the surface buoyancy flux. Fan and Griffies (2014) showed that using this enhanced turbulent velocity scale in the MOM global ocean circulation model had significant impacts on the evolution of the temperature and mixed-layer depth in certain regions. However, the enhanced velocity scale did not properly resolve model differences against in-situ observations, suggesting the need for additional model improvements.

McWilliams et al. (2012) proposed further modifications to the KPP model that include an additional component of the turbulent momentum flux down the gradient of the Stokes drift

$$\overline{\mathbf{u}_h'w'} = -K_L \frac{\partial \mathbf{U}_{hL}}{\partial z} \quad (2.8)$$

where K_L is the Lagrangian eddy-viscosity, and $\mathbf{U}_{hL}$ is the Lagrangian current ($\mathbf{U}_{hL} = \mathbf{u}_S + \mathbf{U}_h$). This approach was demonstrated to improve simulations of the mean current profiles in idealized, unstratified conditions when compared to LES. McWilliams et al. (2014) suggest that for more accurate prediction of the mixing depth the KPP model may also require enhancement of the turbulent velocity scale to compute the unresolved contributions to the bulk mixed-layer shear.

Other studies investigated second moment turbulence closure models by including the CL vortex-force in the TKE equation (D'Alessio et al., 1998), the dissipation length scale (Kantha and Clayson, 2004), and the stability functions (Harcourt, 2013, 2015). The additional turbulent flux down the gradient of the Stokes drift is a key component to the modifications presented by Harcourt (2013, 2015). These studies demonstrate that prediction of the upper-ocean current is improved if the Langmuir turbulence parameterization via Stokes drift is included in the closure model.

In this study we focus on the KPP turbulence closure model under tropical cyclone conditions. First, the performance of the standard KPP model without explicit

Langmuir turbulence effects is investigated under tropical cyclone conditions. Next, we investigate how different modifications to the KPP model that include Langmuir turbulence improve the comparison with equivalent LES simulations.

2.3 Experimental Design

Although any model performance should be validated against observations in principle, it is difficult to test the KPP model using in-situ data in tropical cyclone conditions. Reliable observational data are extremely rare in high-wind environments. Furthermore, to test the KPP model it is preferable to utilize one-dimensional experiments (where horizontal variations are neglected) in order to isolate the vertical turbulent mixing from large-scale three-dimensional processes that introduce additional complications. In-situ observations are inherently three-dimensional and lack temporal and spatial distributions needed to properly test the KPP model over a robust range of conditions. Instead, we will validate the KPP model against idealized computational experiments using a LES model that explicitly resolves the Langmuir turbulence. The performance of the KPP model embedded in a one-dimensional RANS model is tested against the one-dimensional solutions derived by horizontally averaging the LES results.

The ocean surface waves are simulated using the WAVEWATCH III (WW3, Tolman, 2009) surface-wave model. The WW3 (v3.14) model is used with a modification to the wind-input source function that has been demonstrated to well predict the peak wave spectrum under tropical cyclone conditions (Fan et al., 2009). The computational domain has a uniform deep-water (4000 m) bathymetry. The horizontal dimensions are 1800 km in the direction perpendicular to the storm translation and 3000 km in the direction of storm translation, which are large enough that the boundaries are not dynamically important to the study location. The horizontal resolution of the wave model is 8.33 km and the wave spectrum is defined over 40

logarithmically spaced frequencies (with a minimum frequency of 0.0285 Hz) and 48 evenly spaced directions. Surface forcing is applied using a defined, idealized tropical cyclone wind stress. The wind field is constructed following the model of Holland (1980) with a realistic wind inflow angle model following Zhang and Uhlhorn (2012). The radius of maximum wind (RMW) is set as 50 km, and the maximum wind speed is set at 65 m s^{-1} . The wave field is initially at rest.

The identical wind stress is applied in the LES and one-dimensional RANS models, which is calculated using the bulk formula with the drag coefficient formulation as Sullivan et al. (2012):

$$Cd = \begin{cases} 0.0012, & : |u_{10}| < 11 \text{ m s}^{-1} \\ (0.49 + 0.065|u_{10}|) \times 10^{-3}, & : 11 \text{ m s}^{-1} \leq |u_{10}| \leq 20 \text{ m s}^{-1} \\ 0.0018, & : 20 \text{ m s}^{-1} < |u_{10}| \end{cases} \quad (2.9)$$

where $|u_{10}|$ is the neutral 10-meter wind magnitude. A very weak (5 W m^{-2}) destabilizing surface buoyancy flux is also applied, which is insignificant after the initial onset of turbulence. Even a realistic surface heat flux does not significantly contribute to the turbulent mixing within the peak region of a tropical cyclone (wind speeds greater than $\sim 15 \text{ m s}^{-1}$). We therefore neglect this contribution.

The idealized tropical cyclone is translated at a constant velocity through the domain at a moderate (5 m s^{-1}) and fast (10 m s^{-1}) speed. Half the translation speed vector is added to the wind speed vector to create a more realistic, asymmetric wind distribution. The resulting wind fields are shown in Figure 2.1 (left panels).

The waves are spun-up to a quasi-steady state (relative to the moving storm) through an initial 24 hour simulation with the storm translating through the domain. The resulting significant wave height (the integrated variance spectrum) is presented in Figure 2.1 (center panels) for both translation speeds. The Stokes drift is computed from the spectrum as follows:

$$\mathbf{u}_S = \int_0^{k_{UL}} \int_0^{2\pi} \Psi(k, \theta) 2\omega \exp(2kz) \mathbf{k} d\theta dk, \quad (2.10)$$

where $\Psi(k, \theta)$ is the wavenumber-direction sea-surface displacement variance spectrum, k is the wavenumber, θ is the wave direction, $\mathbf{k}$ is the wavenumber vector, $\omega = \sqrt{gk}$ is the (deep water) wave angular frequency, and k_{UL} is the upper bound of the wavenumber integration corresponding to a wavelength of 1 m. The magnitude of the Stokes drift (averaged over 20% of the mixing-layer depth given by the KPP model) is shown in Figure 2.1 (right panels).

Twenty test locations perpendicular to the translation direction are selected. They are given in Table 2.1 and plotted for reference in Figure 2.1 as white circles in right panels. At each location the time series of the wind stress and waves are used to force the LES and one-dimensional RANS models. An additional test is conducted to assess the performance of the model in lower wind conditions. A constant 10 m s^{-1} wind is applied over the entire domain. The fully developed (steady-state) wave spectrum at a fetch of 2000 km is used to calculate a Stokes drift profile. The LES and one-dimensional models then calculate the turbulence and mean ocean properties for 24 hours with this Stokes drift input, which is held constant. This ensures that waves are neither time or fetch limited for the constant wind experiments.

For the tropical cyclone experiments, one-dimensional and LES simulations are performed for 48 hours, with the wind maximum occurring at 36 hours. Comparison between the two models is performed for a duration of 24 hours at each location, from 12 hours before the wind maximum to 12 hours after the wind maximum. This restricts the comparison between the one-dimensional simulation and the LES model to periods with higher wind. The initial potential temperature profile has a homogenous mixed-layer of $\theta = 29.25 \text{ }^{\circ}\text{C}$ and a constant stratification in the interior of $-0.04 \text{ }^{\circ}\text{C m}^{-1}$, which are identical to Sullivan et al. (2012). Two initial mixed-layer depths are investigated, a shallower depth of 10 m, and a deeper depth of 32 m. The simulations are performed across all 20 test locations for the 10 m depth, and at 11

stations for the 32 m depth. The 32 m mixed-layer depth is chosen to correspond to the Sullivan et al. (2012) experiment, while the 10 m mixed-layer depth is chosen as a reasonable upper limit for a shallow, summer mixed-layer. For the constant wind simulation the initial mixed-layer depth is 10 m, but the stratification is reduced to $-0.01 \text{ }^\circ\text{C m}^{-1}$.

The LES and one-dimensional RANS model are run with identical forcing, physical parameters, and the linear equations of state, which is defined as:

$$\rho = \rho_0 + \alpha(\theta - \theta_0) \quad (2.11)$$

where the reference density ρ_0 is $1026.95 \text{ kg m}^{-3}$ for $\theta_0 = 10 \text{ }^\circ\text{C}$ and the salinity $S = 35 \text{ PSU}$. The salinity is kept constant and the thermal expansion coefficient is $\alpha = -0.2 \text{ kg m}^{-3}(\text{ }^\circ\text{C})^{-1}$. For the one-dimensional model, the General Ocean Turbulence Model (GOTM, Umlauf et al., 2005) is used, which includes the Coriolis-Stokes force and a modified KPP routine to account for the Langmuir turbulence.

The LES domain for the tropical cyclone experiments is $(x, y, z)=(750 \text{ m}, 750 \text{ m}, 240\text{m})$ with a resolution of $(dx, dy, dz)=(2.92 \text{ m}, 2.92 \text{ m}, 1 \text{ m})$. For the constant wind experiment, a smaller domain is used with $(x, y, z)=(300 \text{ m}, 300 \text{ m}, 180\text{m})$ and a resolution of $(dx, dy, dz)=(1.56 \text{ m}, 1.56 \text{ m}, 0.7 \text{ m})$. Our previous sensitivity tests indicated that a higher resolution or a larger domain size does not substantially change our results, indicating that our resolution is adequate. We also assessed the TKE partitioning between resolved and subgrid scale (SGS). The resolved TKE is usually greater than 75% of the total (resolved + SGS) TKE, which also supports that our resolution is sufficient. The LES model solves the SGS averaged equations of momentum, density, and continuity following previous LES Langmuir turbulence studies (McWilliams et al., 1997). The phase-averaged momentum equations are solved in the three-dimensional coordinate system:

$$\frac{D\mathbf{u}}{Dt} + f\hat{\mathbf{z}} \times (\mathbf{u} + \mathbf{u}_S) = -\nabla\pi - g\hat{\mathbf{z}} \left(\frac{\rho}{\rho_0} \right) + \mathbf{u}_S \times (\nabla \times \mathbf{u}) + SGS \quad (2.12)$$

where the model variables are SGS averaged, $D/Dt = \partial/\partial t + \mathbf{u} \cdot \partial/\partial \mathbf{x}$, $\mathbf{x}$ is the coordinate vector, $\mathbf{u}$ is the velocity vector, $\mathbf{u}_S$ is the Stokes drift vector (the vertical component is set to 0), f is the Coriolis parameter, π is the generalized pressure ($p/\rho + 0.5[|\mathbf{u} + \mathbf{u}_S|^2 - |\mathbf{u}|^2]$), p is the pressure, and SGS are the sub-grid scale terms described in detail by Rabe et al. (2015). Scalar (temperature) distribution is solved by the advective-diffusive equation:

$$\frac{D\rho}{Dt} + \mathbf{u}_S \cdot \nabla \rho = SGS \quad (2.13)$$

where ρ is the density (determined entirely by the temperature via equation 2.11). Volume conservation is ensured via the incompressible continuity equation:

$$\nabla \cdot \mathbf{u} = 0. \quad (2.14)$$

The Stokes drift is included in the momentum equation via the Stokes-Coriolis force (term 2 on the LHS of eq. 2.12), the CL vortex-force (term 3 on the RHS of eq. 2.12), and the generalized pressure. The Stokes drift also contributes to the temperature equation via the additional Stokes drift advection term. In general, the Langmuir turbulence has a much greater impact on the mean current profile than the Stokes-Coriolis force.

The LES is initialized in a similar manner to as described by Rabe et al. (2015). In each case stationary simulation is run with the initial wind, wave, and density profile until statistically stationary turbulence is achieved (e.g. for multiple eddy turnover times, h/u_*). The resulting field is then used as the initial condition for the transient LES simulation. Unlike Rabe et al. (2015), however, the mean current in the stationary solution is removed from the initial condition, such that the initial mean current is zero in both the LES and the one-dimensional model.

The LES model is first run at each test location with the Stokes drift set to zero. This means that both the Langmuir turbulence and the Coriolis-Stokes force

are not simulated and the turbulence is primarily driven by vertical current shear. The results from these simulations are referred to as LES-ST (Shear Turbulence).

Next, the LES model is run at each location with the Stokes drift computed from the WW3 wave variance spectrum. The Stokes drift profile is calculated directly from the spectrum on the LES model vertical grid levels and linearly interpolated from the WW3 to the LES model time step. The results of these simulations will be referred to as LES-LT (Langmuir Turbulence). The LES results are averaged horizontally into vertical profiles for comparing with results in the one-dimensional model.

2.4 LES Results

In Figure 2.2 we present the difference between the mean near-surface fields (temperature and currents at 5 m depth) in LES-LT and LES-ST, for the tropical cyclone with 5 m s^{-1} translation speed and the 10 m initial mixed-layer depth. The results with the larger translation speed and/or the deeper initial mixed-layer depth are qualitatively similar. Although the actual simulations are conducted over time at fixed locations, we use the time and translation speed of the storm to transform from the temporal coordinate to the spatial coordinate (in the direction of the storm translation) to simplify the presentation. This transformation is possible because all results are presented after the wave field has reached a quasi-steady state with respect to the frame of reference moving with the storm. Since our simulations are performed with a reasonably high spatial resolution in the direction perpendicular to the storm translation (Figure 2.1), we can present the spatial snapshots of the current and temperature fields in Figure 2.2. These spatial patterns remain independent of time after the wave field has reached quasi-steady state.

The contribution of Langmuir turbulence to cooling is greatest near the location of maximum Stokes drift shown in Figure 2.1. The additional cooling due to Langmuir turbulence reaches a maximum of nearly $0.4 \text{ }^{\circ}\text{C}$ on the right hand side in a region

where the total cooling is between 2.5 and 3 °C. This is the region where the peak waves are longest, and the Stokes drift penetrates deepest into the water column. This is also the location of rapid mixed-layer deepening, suggesting that Langmuir turbulence is a significant contributor to deepening of the mixing-layer. The region of significant enhanced cooling (0.1 °C or more) due to Langmuir turbulence is quite large, extending to about 3 (2) times the RMW to the right (left).

The near-surface current magnitude in LES-LT is smaller by as much as 0.7 m s⁻¹ than that in LES-ST near the location of the maximum current. The Langmuir turbulence significantly increases the turbulent momentum transport within the upper portion of the water column where the Stokes drift and its vertical gradient are greatest. There is a small location near the center and in the left rear of the hurricane where the currents are slightly stronger in the presence of Langmuir turbulence. Overall, Langmuir turbulence has a more noticeable impact on the mixing of currents than temperature in the upper water column, since the current gradients are strong, but temperature is well mixed and nearly uniform. This does not imply that the contribution to surface cooling is trivial, since even a temperature change of O(0.1) °C may have significant implications for the tropical cyclone development (Emanuel, 1999).

To demonstrate the vertical dependence of the Langmuir turbulence impact in this study, we present vertical transects of the upper-ocean current magnitude and temperature from 50 km to the right of the storm center (Figure 2.3). The current is significantly decreased near the surface in LES-LT compared to LES-ST in the area where the Stokes drift is large and the Langmuir turbulence is strong. The difference between the LES-LT and LES-ST currents decreases gradually from the surface down to about 50 m, suggesting that the impact of the Langmuir turbulence penetrates quite deep, compared to the Stokes drift itself that mostly decays within 10 m or so

from the surface. Near the base of the mixing-layer the LES-LT current is greater than the LES-ST currents, which is due to the increased mixing depth in LES-LT. The LES-LT temperature is cooler than the LES-ST temperature down to about 100 m, but there is less vertical dependence of the temperature difference compared to the current difference. At the base of the mixing-layer the LES-LT is warmer than LES-ST due to the increased mixing depth.

The LES-LT data are used to compare the direction of the turbulent Reynolds stress $(\overline{u'w'}, \overline{v'w'})$ with the Lagrangian and the Eulerian current shear directions (Figure 2.4). It is clear that during the tropical cyclone passage the Reynolds stress aligns better with the Lagrangian shear throughout the entire mixing-layer. There is a region of significant misalignment between the turbulent stress and the Eulerian shear in the upper 30 m. In this region, the Lagrangian shear is dominated by the Stokes drift shear and the Eulerian shear is quite small. This suggests that the stress parameterization (equation 2.8) based on the Lagrangian shear proposed by McWilliams et al. (2012) predicts the direction of the turbulent stress more accurately than the traditional parameterization based on the Eulerian shear under hurricane conditions.

During the passage of the hurricane eye there is strong misalignment of the turbulent stress from both the Eulerian shear and the Lagrangian shear. This may be caused by strong disequilibrium of the turbulence in this region as the wind magnitude and direction rapidly change. Any turbulence closure models that assume alignment of stress and current shear may not perform well in these rapidly evolving conditions. To incorporate misaligned shear/stress in the KPP model requires the rotation of the stress from the shear direction to be included in the Reynolds stress calculation following McWilliams et al. (2012). That study shows that including the misalignment information obtained from a LES model improves the KPP performance. However,

this is not a practical approach for tropical cyclone-ocean coupled forecast models.

2.5 Implicit Langmuir turbulence KPP model

Now that we have described the impact of Langmuir turbulence under tropical cyclones based on the LES results, we propose a modification to the KPP scheme that can reproduce the LES-LT results in the one-dimensional model. This modification will proceed in the following manner. First, we demonstrate that the standard KPP scheme, which implicitly includes the typical (sea-state independent) Langmuir turbulence impact, is not adequate to reproduce the simulated temperature and currents from the LES-LT (this section). Next, we remove the implicit Langmuir turbulence impact from the KPP model so that the one-dimensional simulation accurately reproduces the LES-ST results (Section 2.6). We then introduce the explicit Langmuir turbulence impact in two steps. In step one, we replace the Eulerian current by the Lagrangian current in KPP, because the LES results have clearly demonstrated that the turbulent fluxes are more closely aligned to the Lagrangian shear than the Eulerian shear. We find that this step does not produce enough mixing in the KPP model compared to the LES-LT (Section 2.7). In step two, we introduce enhancement factors of the turbulent mixing coefficient and the unresolved shear in KPP, so that the one-dimensional simulations best agree with the LES-LT results (Section 2.8).

2.5.1 Method

We first investigate the performance of the standard KPP model (without explicit Langmuir turbulence effects) in tropical cyclone conditions. We define the standard KPP model following Large et al. (1994) in which the turbulent fluxes are calculated as the product of the eddy viscosity K_M (or diffusivity K_θ) and the mean vertical gradient of the momentum (or scalar) plus an additional non-local (counter gradient) component (equations 2.5 and 2.6). Because our experiments include only an

insignificant surface buoyancy flux, the non-local components are neglected.

The vertical profile of K_x (where x can be θ or M) is a function of normalized depth, $\sigma = -z/h$, where h is the mixing-layer depth (which is defined for KPP later in this section). With the counter-gradient term dropped, $K_x(\sigma)$ is equal to the product of the mixing-layer depth, h , the turbulent velocity scale, W_x , and a non-dimensional shape function, $G_x(\sigma)$:

$$K_x(\sigma) = hW_xG_x(\sigma). \quad (2.15)$$

The non-dimensional turbulent mixing shape function is approximated by a cubic polynomial, $G_x(\sigma) = a_0 + a_1\sigma + a_2\sigma^2 + a_3\sigma^3$. The coefficients are determined following arguments of Large et al. (1994), where a_0 must be equal to zero (turbulent eddies do not cross the surface), a_1 equals one (to match the near-surface value of $K_x \sim -\kappa u_* z$), and a_2 and a_3 can be determined by matching the value and curvature of K_x to the interior. Since the interior turbulent mixing parameter is negligible in this study, $a_2 \sim -2$ and $a_3 \sim 1$, thus $G_x(\sigma) \sim \sigma(1 - \sigma)^2$ for both momentum and temperature. The turbulent velocity scale is $W_x = u_*\kappa/\phi_x(\sigma h/L)$ where κ is the von-Kármán constant, ϕ_x is the universal stability function and L is the Monin-Obhukhov length ($u_*^3/[\kappa B_*]$). In this study $L \gg h$ and $W_x \sim u_*\kappa$ for both momentum and temperature. Therefore, K_x , W_x , G_x are all identical for temperature and momentum.

The mixing-layer depth, h , is calculated by the KPP model using a bulk Richardson number (Ri_b) threshold, which compares the relative contribution of stabilizing buoyancy and destabilizing shear:

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U^r - U(z))^2 + (V^r - V(z))^2 + V_t^2(z)} < Ri_c \quad (2.16)$$

where B is the buoyancy, V_t is the unresolved turbulent velocity shear, and the superscript r denotes a reference value. At the bottom of the mixing-layer $Ri_b(z = -h) = Ri_c$. The Ekman depth ($h_E \sim 0.7u_*/f$), which is important when stratification is weak, and the Monin-Obhukov depth ($h_{MO} = L = u_*^3/B_*\kappa$), which is important dur-

ing surface heating (re-stratification) events, further restrict this mixing-layer. The value of Ri_c varies based on the model resolution, but a typical value is 0.3. The unresolved turbulence term contains the contribution of convection to mixing:

$$V_t^2(z) = \frac{C_v(-\beta_T)^{1/2}}{Ri_c\kappa^2}(c_s\epsilon)^{-1/2}|z|NW_\theta \quad (2.17)$$

where N is the stability frequency and the constants are $C_v = 1.6$, $\beta_T = -0.2$, $c_s = 98.96$, and $\epsilon = 0.1$ following Large et al. (1994). This V_t^2 term should also include the unresolved Langmuir turbulence contribution to deepening the mixing-layer (McWilliams et al., 2014). The reference variables in the bulk Richardson number (U^r , V^r , B^r) are defined as the average over the upper 10% of the mixing-layer following Large et al. (1994) and Griffies et al. (2013). This averaging is a standard procedure because it reduces the resolution dependence of the bulk Richardson number. We also find that this averaging prevents over predicting the mixing depth when the current is surface intensified, particularly under extreme wind speeds.

This version of KPP with $Ri_c = 0.3$ is defined as the implicit Langmuir turbulence version for this study, hereafter KPP-iLT. It should be noted that this version of the model implicitly includes mean wave impacts (including Langmuir turbulence) because the coefficients (particularly Ri_c) have been determined empirically from in-situ data and the real ocean always contains waves when there is significant wind.

2.5.2 Results

We first compare the mean near-surface ($z = -5$ m) values of both temperature anomaly (where the initial temperature is subtracted from the simulated temperature) and currents. We evaluate the results at the near-surface since this is most relevant for the hurricane modeling application, but the conclusions are essentially the same if we choose a deeper ($z = -30$ m) transect for comparison. The root-mean square difference between the KPP (one-dimensional GOTM) and LES results is used as a metric to evaluate the KPP performance. RMS difference is computed over 24 hours

(± 12 hours relative to the passage of the peak winds) in all test experiments (both translation speeds, both initial mixed-layer depths) and at all locations. For the 5 m s^{-1} translating storm this corresponds to the spatial study transect of 432 km, and for the 10 m s^{-1} translating storm this corresponds to the spatial study transect of 864 km. As long as the time and location are consistent, this is a useful metric to compare different KPP methods within this study. Note that because both the LES and the one-dimensional model are initialized with zero mean current, the phases of the inertial oscillations remain almost identical between the two models.

The results of the KPP-iLT and LES-LT experiments are compared in Figure 2.5. All results (both initial mixed-layer depths and both translation speeds) are included in the left panels. The small difference in the temperature anomaly between KPP-iLT and LES-LT (less than 0.1°C at most locations) suggests KPP-iLT does a reasonable job predicting the mean mixing-layer depth and surface cooling over the range of Langmuir turbulence conditions under the tropical cyclones. KPP-iLT gives a slight warm bias (Figure 2.5, bottom-center and bottom-right panels), but this bias could be corrected by adjusting the critical Richardson number. The cooling on the left hand side of the 5 m s^{-1} moving tropical cyclone is noticeably under predicted, which is discussed further in the following section.

The currents predicted by KPP-iLT are much different from those predicted by LES-LT. The error is almost as large as the contribution of Langmuir turbulence itself shown in Figure 2.2. This suggests that increased near-surface mixing from Langmuir turbulence is not captured well by KPP-iLT, and that accurate current predictions require introducing explicit Langmuir turbulence effects in the KPP model. In the upper-left panel, the time history of the current magnitude at each LES location can be seen. Initially, both LES-LT and KPP-iLT currents are zero. They increase and diverge as the wind and Stokes drift increase and the Langmuir turbulence becomes

more important. They eventually converge as the wind decreases and the turbulent mixing becomes less important. Part of the difference between the KPP-iLT current and the LES-LT current is due to the presence of the Stokes drift itself. Since the turbulent mixing occurs in response to the Lagrangian shear in LES-LT, the behavior of the Lagrangian current in LES-LT is more similar to the behavior of the Eulerian current in KPP-iLT. Nevertheless, the difference between the KPP-iLT current and the LES-LT current remains significant even below the depth of significant Stokes drift because of the enhanced Langmuir turbulence mixing.

2.6 Shear-turbulence KPP without wave effects

2.6.1 Method

Before introducing explicit Langmuir turbulence effects to rectify the under-prediction of mixing in the KPP model, it is necessary to remove the implicit wave impacts that are already included in KPP-iLT. To remove the implicit Langmuir turbulence, we retune the critical Richardson number by optimizing (reducing) the near-surface RMS difference of temperature and current between the one-dimensional simulations and the LES-ST results. An alternative way to retune the KPP model would be to keep the critical Richardson number unchanged and to reduce the turbulent velocity scale (W_x) at large Langmuir numbers (to reduce the shear driven turbulence). This method, however, is not consistent with the traditional near-surface wall-layer turbulence ($W_x \sim \kappa u_*$), and does not yield good performance in the KPP model. We find that the retuned critical Richardson number of 0.235 gives optimal agreement between the KPP and the LES-ST results for surface temperature anomaly and currents. Interestingly, this is also similar to the value of 0.25 found for atmospheric boundary layer closure schemes similar to the KPP (see e.g. Hong and Pan, 1996). This version of the KPP model is called KPP-ST (Shear Turbulence) hereafter.

2.6.2 Results

The results of the KPP-ST and LES-ST experiments are compared (Figure 2.6) in the same manner as the previous section. The RMS difference for both near-surface current and temperature anomaly are considerably reduced across the range of forcing parameters in this experiment as a result of modifying the critical Richardson number.

As in the case of KPP-iLT (Figure 2.5), the cooling on the left hand side of the 5 m s^{-1} moving tropical cyclone is noticeably under predicted (Figure 2.6, lower-middle panel). This problem is clearly unrelated to the Langmuir turbulence. It is rather related to KPP's shortcoming in handling the rapidly changing wind forcing. When the wind speed rapidly decreases after the storm passage, the near-surface current rapidly decreases as well. The KPP model predicts a shallower mixing-layer depth based on the critical Richardson number criterion, and the deepening of the mixing-layer ceases. However, the LES model shows that the turbulence near the bottom of the mixing-layer does not weaken as rapidly and the mixing-layer continues to deepen during this phase of the tropical cyclone. The KPP fails to accurately predict the vertical variation of the turbulence in such a transient and non-equilibrium condition. The problem is more evident on the left hand side because the mixing-layer is much shallower and the sea surface temperature is more sensitive to the under-prediction of the mixing-layer deepening. Because this is not related to the Langmuir turbulence effect in the KPP model we do not address it any further in this study.

2.7 KPP with Lagrangian currents

2.7.1 Method

Before adding the explicit Langmuir turbulence parameterization into the KPP-ST model we consider the effect of the Lagrangian shear (with the Stokes drift) in place of the Eulerian shear. Figure 2.4 indicates a better alignment between the Reynolds stress and the Lagrangian shear. In all diffusion equations the current shear

is replaced with the Lagrangian shear as shown in equation 2.8. In addition, the bulk Richardson number calculation is modified to account for the Lagrangian shear:

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U_L^r - U_L(z))^2 + (V_L^r - V_L(z))^2 + V_t^2(z)} < Ri_c, \quad (2.18)$$

with the critical Richardson number fixed at 0.235 as in the KPP-ST. This version of the KPP model is called KPP-Lag hereafter.

2.7.2 Results

We now compare the results of the KPP-Lag and LES-LT experiments to test the performance of the KPP model using the Lagrangian currents (Figure 2.7). Clearly, KPP-Lag improves the current predictions relative to KPP-iLT under a tropical cyclone. However, there is now significantly more undercooling relative to KPP-iLT. This suggests that the mixed-layer deepening and entrainment are under predicted, since the surface heat flux is omitted in this study,

In order to investigate the cause of the under-mixing, we compare the mixing coefficient K_M in KPP-Lag and the mixing coefficient K_{LES} that is directly estimated from the LES-LT results. Since the Lagrangian shear and the stress are mostly aligned, the K_{LES} is derived as follows:

$$K_{LES} = \frac{|\overline{u'w'} + i\overline{v'w'}|}{|\frac{\partial U_L}{\partial z} + i\frac{\partial V_L}{\partial z}|}. \quad (2.19)$$

The mixing-layer depth h is also required to obtain a non-dimensional mixing coefficient profile $G(\sigma)_{LES} = K_{LES}/(\kappa u_* h)$ from the LES results. The mixing-layer depth h is determined by applying the same bulk Richardson number threshold (equation 2.18), with the critical value of 0.235, to the LES mean field.

Examples comparing the non-dimensional mixing coefficient profiles $G_M(\sigma)$ and $G(\sigma)_{LES}$ at locations to the left and right of the storm track are presented in Figure 2.8 (from a tropical cyclone experiment with a 5 m s^{-1} translation speed, and a 10 m initial mixed-layer depth). These examples are from (upper panels) four hours prior

to maximum wind, (middle panels) the time of maximum wind, and (lower panels) four hours after the maximum wind. The LES profiles (black dots) show a clear enhancement of the mixing coefficient in the presence of Langmuir turbulence, which is missing in KPP-Lag (red line).

This analysis supports previous efforts (McWilliams and Sullivan, 2000) of enhancing the mixing coefficient in the KPP model that varies with the Langmuir number. In the next section we will explore different ways of enhancing the mixing coefficient in the KPP-Lag model.

2.8 KPP with Lagrangian currents and enhanced mixing

2.8.1 Method

In the KPP model, the enhancement to the turbulent mixing coefficient can be introduced through a factor F_{LT} :

$$K_x^{LT} = K_x \times F_{LT}, \quad (2.20)$$

where K_x is defined by equation 2.15. The unresolved turbulent shear contribution V_t^2 (eq. 2.17) must also be modified to account for the contribution of Langmuir turbulence (following McWilliams et al., 2014),

$$(V_t^{LT})^2 = V_t^2 \times F_{LT}^{Vt}. \quad (2.21)$$

We differentiate between the enhancement to the unresolved shear contribution F_{LT}^{Vt} and F_{LT} used in the mixing coefficient profile for reasons that will be discussed later in this section.

In previous studies the enhancement factor F_{LT} has been determined as a function of the turbulent Langmuir number (McWilliams and Sullivan, 2000; Van Roekel et al., 2012). However, the definition of the turbulent Langmuir number itself varies from study to study and requires some consideration. McWilliams and Sullivan (2000)

defines the turbulent Langmuir number based on the surface Stokes drift:

$$La_t = \sqrt{\frac{u_*}{|\mathbf{u}_S(z=0)|}}, \quad (2.22)$$

which is valid for aligned wind and waves. Another proposed form, which also applies to the aligned case only, is

$$La_{SL} = \sqrt{\frac{u_*}{\langle |\mathbf{u}_S| \rangle_{SL}}}, \quad (2.23)$$

where $\langle |\mathbf{u}_S| \rangle_{SL}$ is the Stokes drift magnitude averaged over the surface layer, which is defined as the top 20% of the mixed-layer (Harcourt and D’Asaro, 2008). In Harcourt and D’Asaro (2008) the surface layer averaged Stokes drift is referenced to the Stokes drift at the base of the mixed-layer, but we find this does not significantly impact the results due to the large mixed-layer depths observed in this study. We have therefore approximated this reference Stokes drift as 0, though this simplification may not be appropriate for non-hurricane conditions. The averaging over the surface layer is performed because it is unlikely that the waves that decay within the first few meters contribute significantly to the Langmuir turbulence on scales relevant to the mixed-layer. Van Roekel et al. (2012) have further modified the Langmuir number definition to accommodate misaligned wind and waves:

$$La_{SL\theta} = \sqrt{\frac{u_* \times \cos(\theta_{Wind} - \theta_{Lag})}{\langle |\mathbf{u}_S| \rangle_{SL} \times \cos(\theta_{Waves} - \theta_{Lag})}}, \quad (2.24)$$

where θ_{Wind} is the wind direction, θ_{Waves} is the direction of the average Stokes drift and (θ_{Lag}) is the direction of the Lagrangian shear over the surface layer. They also suggest that the surface layer depth should be defined as the top 20% of the mixing-layer rather than the mixed-layer. The directional components define the projection of the wind and the Stokes drift into the mean Lagrangian shear direction (used as a proxy for dominant Langmuir turbulence eddy orientation). Note that this form may yield a very small Langmuir number (strongly enhanced Langmuir turbulence) if the Lagrangian shear is significantly misaligned with wind, which is not observed in our

LES experiments. Furthermore, this Langmuir number is undefined if $|\theta_{Waves} - \theta_{Lag}|$ exceeds $\pi/2$. Although such conditions rarely occur, it is important to remove non-physical Langmuir number in complex wind and wave conditions under a tropical cyclone.

In this study we find that the misalignment between the wind and the Lagrangian shear is negligible ($\cos(\theta_{Wind} - \theta_{Lag}) \approx 1$). Therefore, for simplicity we neglect this effect and modify the definition of the Langmuir number to:

$$La_{SL\theta'} = \sqrt{\frac{u_*}{\langle |\mathbf{u}_S| \rangle_{SL}} \frac{1}{\max(\cos(\theta_{Waves} - \theta_{Lag}), 10^{-8})}}, \quad (2.25)$$

that is, the misalignment between the waves and the Lagrangian shear is limited to $\pi/2$ in the Langmuir number calculation. We also define the surface layer as the top 20% of the mixing-layer instead of the mixed-layer. Note that when the misalignment between the waves (Stokes drift) and the Lagrangian shear exceeds $\pi/2$, there is a possibility that the upper-ocean turbulence is suppressed instead of enhanced due to surface waves (Rabe et al., 2015), that is, the enhancement factor may become less than 1. However, we find that such occurrences are rare (short-lived at a particular location) and does not significantly affect the mixed-layer deepening process. Therefore, we do not find it necessary to accommodate such cases in the modified KPP model.

The three forms of the Langmuir number are compared in Figure 2.9. At the surface, all resolved spectral wavenumbers ($k \leq k_{UL}$) contribute to the Stokes drift. The decay rate is an exponential function of the wavenumber, so that waves shorter than 10 m decay to less than 10% of their surface value by $z = -2$ m. This means a larger magnitude for the surface-based definition of the Stokes drift (La_t , left panels) and thus smaller Langmuir number. The center panel shows the surface-layer averaged definition, La_{SL} , where the surface layer is defined as the top 20% of the mixing-layer. This Langmuir number is weighted toward the magnitude of longer

waves that contribute to the Stokes drift over a larger fraction of the mixing-layer. The projected surface-layer averaged Langmuir number ($La_{SL\theta'}$, right panels) gives significantly higher values than La_{SL} when the Lagrangian shear becomes misaligned with the waves, mainly on the left of the storm and near the eye. For the transient, turning wind in a moving tropical cyclone, the projected definition of the Langmuir number is likely more desirable. We have also tested the projected Langmuir number $La_{SL\theta}$ and have found that the results are almost identical to the $La_{SL\theta'}$ except for very small areas (mainly inside the RMW) where $La_{SL\theta'}$ becomes extremely small or undefined.

Next, we consider the enhancement factor F_{LT} to the turbulent eddy viscosity (or diffusivity) K_x as a function of the turbulent Langmuir number. There have been many previous attempts to scale the enhancement to the vertical velocity variance $\langle w'^2 \rangle$ (averaged over the entire mixing-layer or mixed-layer) normalized by the friction velocity squared, and the enhancement factor F_{LT} in the KPP model. Some LES studies support the scaling argument based on the ratio of the Stokes production to the dissipation and suggest $\langle w'^2 \rangle \propto u_*^2 La_t^{-4/3}$ (Min and Noh, 2004; Harcourt and D'Asaro, 2008; Grant and Belcher, 2009; McWilliams et al., 2014), when the Stokes production is significant (small Langmuir numbers). McWilliams and Sullivan (2000) suggest an enhancement factor of $F_{LT} = (1 + 0.08 La_t^{-4})^{1/2}$. Recently, Van Roekel et al. (2012) have introduced a new scaling based on the projected Langmuir number:

$$\frac{\langle w'^2 \rangle}{u_*^2 \cos^2(\theta_{Wind} - \theta_{Lag})} = 0.6 [1 + (1.5 La_{SL\theta})^{-2} + (5.4 La_{SL\theta'})^{-4}]. \quad (2.26)$$

While the enhancement of $\langle w'^2 \rangle / u_*^2$ and the enhancement factor F_{LT} should be related, the exact relationship between the two is not trivial. This is because the Langmuir turbulence likely modifies not only the velocity scale, but also the length scale in the KPP model. In this study, we do not attempt to separate these two contributions, but rather focus on the total combined impact on the mixing coefficient.

We also separate the enhancement to the mixing coefficient F_{LT} and the enhancement to the unresolved turbulent shear contribution, F_{LT}^{Vt} , because these two may be related to different length and velocity scales.

Since there is no general consensus regarding the form of F_{LT} , we determine the enhancement factor empirically by using the LES experimental results. As described in the previous section, a non-dimensional mixing coefficient profile $G(\sigma)_{LES} = K_{LES}/(\kappa u_* h)$ can be obtained from the LES results. The maximum value of $G(\sigma)_{LES}$ within the mixing-layer is then compared to the maximum value of $G_M(\sigma)$ that is used by the KPP model (i.e., $\max(G_M(\sigma)) \sim \max(\sigma(1 - \sigma)^2) \sim 0.1481$ for $0 \leq \sigma \leq 1$). The ratio of these quantities is the LES enhancement factor that should be consistent with F_{LT} in the modified KPP model. In addition, the LES vertical velocity variance $\langle w'^2 \rangle$ is readily available from the LES results and the enhancement to $\langle w'^2 \rangle / u_*^2$ can be compared to the LES enhancement factor.

First, it is beneficial to perform this exercise using the LES-ST experiments to confirm that the KPP-ST K_M profile agrees with the K_{LES} profile derived from the LES-ST. In this case, the Lagrangian current is equal to the Eulerian current in equations 2.18 and 2.19. The calculated LES enhancement factor ($F_{LT}^{LES} = \max[G(\sigma)_{LES}]/0.1481$, the ratio of the maximum value of $G(\sigma)_{LES}$ to 0.1481) is 0.81 ± 0.20 (the latter number is the standard deviation over $n = 6337$) for all times and locations except with wind speed less than 5 m s^{-1} and the locations inside the radius of maximum wind (RMW). These results suggest the KPP mixing coefficient is indeed consistent with the effective mixing coefficient in the LES without wave effects (no Langmuir turbulence) as expected from the good agreement between the KPP-ST and LES-ST results, except for very low wind conditions and very complex wind conditions inside the RMW.

Figure 2.10 shows the LES enhancement factor (ratio of maximum normalized

LES mixing coefficient to maximum normalized KPP mixing coefficient) as a function of the surface-layer averaged Langmuir number La_{SL} (middle left panel) and the projected surface-layer Langmuir number $La_{SL\theta'}$ (lower-left panel) for all LES experiments with waves. The results are distinguished by different colors depending on the locations relative to the storm. The projected Langmuir number reduces scatter relative to the surface Langmuir number. The improvement is most prevalent on the left hand side of the storm where wind-wave misalignment more often occurs. The LES enhancement factor clearly exceeds 1 at most locations, suggesting that Langmuir turbulence does indeed enhance the mixing coefficient. By bin-averaging our data, we find that the LES enhancement factor scales well with a simple empirical form of

$$F_{LT} = 1 + La_{SL\theta'}^{-1}, \quad (2.27)$$

shown by a solid black line, for Langmuir numbers above 0.8. The mixing coefficient enhancement appears to level off around 2.25 for Langmuir numbers below 0.8, possibly even decaying for lower Langmuir numbers. This is qualitatively consistent with McWilliams et al. (2014), who also find that the eddy viscosity magnitude decreases at very low Langmuir numbers. However, after close examination we have found that the leveling off or reduction of the enhancement factor only occurs in the rear right of the storm, where the mixing-layer depth h can be significantly overestimated based on the Richardson number criterion applied to the LES results. It is likely that the overestimation of h contributes to the apparent reduction of the LES enhancement factor. Because of this uncertainty we cap the empirical enhancement factor at 2.25 for Langmuir numbers below 0.8 as shown by the dashed black line.

Next, we examine the enhancement of $\langle w'^2 \rangle / u_*^2$ directly obtained from the LES results (right panels). This exercise further demonstrates the importance of the misalignment correction of the Langmuir number, supporting the results of Van Roekel

et al. (2012) and Rabe et al. (2015). It is interesting that the scaling of Van Roekel et al. (2012) (equation 2.26 with $\cos(\theta_{Wind} - \theta_{Lag}) = 1$, green dash line in Figure 2.10) is a very good fit to our LES results as well. We have also found that our LES-ST results yield $\langle w'^2 \rangle / u_*^2 = 0.57 \pm 0.24$, which is consistent with the large Langmuir number limit of 0.6 by Van Roekel et al. (2012). At very low Langmuir numbers the LES $\langle w'^2 \rangle / u_*^2$ starts to deviate from the scaling of Van Roekel et al. (2012), possibly because of the overestimation of h as discussed earlier. If h is overestimated, the mixing-layer average of $\langle w'^2 \rangle$ is performed including a region of weak turbulence below the real mixing-layer and the result may be underestimated. We do not compare our results with the enhancement factor given by McWilliams and Sullivan (2000) since their form was obtained using the very different Langmuir number La_t based on the surface Stokes drift. The scaling of $\langle w'^2 \rangle / u_*^2$ presented in Harcourt and D'Asaro (2008) avoids the asymptotic breakdown as $La \rightarrow 0$ (see McWilliams and Sullivan, 2000; Harcourt and D'Asaro, 2008) unlike the scaling of Van Roekel et al. (2012). However, the difference between these two is appreciable only at very low Langmuir numbers where our results are not reliable. We also note that the values of $\langle w'^2 \rangle$ can be quite different if the averaging is done over the entire mixed-layer (Van Roekel et al., 2012; Rabe et al., 2015) instead of the entire mixing-layer (this study).

We now investigate the relationship between the LES enhancement factor and the enhancement of $\langle w'^2 \rangle / u_*^2$. If we assume that the length scale of the KPP mixing coefficient K_M is not affected by the Langmuir turbulence and the velocity scale of K_M is enhanced in the same manner as the square root of the vertical velocity variance $\langle w'^2 \rangle$, the enhancement factor F_{LT} should be identical to the square root of $\langle w'^2 \rangle / u_*^2$ divided by its limiting value at large Langmuir numbers (no Langmuir turbulence). Then, the scaling by Van Roekel et al. (2012) (equation 2.26) suggests that

$$F_{LT} = [1 + (1.5La_{SL\theta})^{-2} + (5.4La_{SL\theta})^{-4}]^{1/2}. \quad (2.28)$$

However, this scaling significantly underestimates the LES enhancement factor (blue dash line in the left panels). If we instead assume that F_{LT} is identical to the enhancement to $\langle w'^2 \rangle / u_*^2$, the scaling of Van Roekel et al. (2012) suggests

$$F_{LT} = 1 + (1.5La_{SL\theta})^{-2} + (5.4La_{SL\theta})^{-4}. \quad (2.29)$$

This scaling (red dash line the left panels) is more consistent in terms of the order of magnitude but still underestimates the LES enhancement factors except for very low Langmuir numbers. These exercises suggest that the enhancement to the velocity and length scales of K_M is not simply related to the enhancement of the vertical velocity variance $\langle w'^2 \rangle$, and support our approach of determining F_{LT} empirically.

The ratio of K_M and K_{LES} is not constant with depth (see Figure 2.8). Comparing many similar profiles, we find that this ratio roughly peaks at the maximum of the K_M profile, and approaches 1 at the top and bottom boundaries. For that reason we apply the enhancement factor at its maximum where the non-dimensional profile also reaches its maximum. The enhancement factor is then reduced to 1 approaching both the top and bottom boundaries.

In summary, based on this analysis we set our KPP Langmuir turbulence enhancement factor (F_{LT}^K) as:

$$F_{LT}^K(\sigma) = 1 + (F'_{LT} - 1) * G_x(\sigma) / \max(G_x(\sigma)), \quad (2.30)$$

$$F'_{LT} = 1 + La_{SL\theta'}^{-1}, \quad La_{SL\theta'} \geq 0.8, \quad (2.31)$$

(black solid line in Figure 2.10) and

$$F'_{LT} = 2.25, \quad La_{SL\theta'} \leq 0.8, \quad (2.32)$$

(black dashed line in Figure 2.10). Referring back to Figure 2.8, the KPP profile with the enhancement (blue) clearly does a better job reproducing the LES turbulent mixing coefficient profile compared to the KPP profile without the enhancement

(red). The impact of reducing the enhancement toward the bottom and the top (blue-dashed) greatly improves agreement near the surface, and helps avoid over-entrainment of cool water at the base of the mixing-layer.

We find that using the same form of the enhancement factor in both the turbulent mixing profile (K_x) and the unresolved shear (V_t) does not work. This supports the conclusion of McWilliams et al. (2014) that the scale of Langmuir turbulence that contributes to the near-surface mixing is different from the scale of Langmuir turbulence that drives mixed-layer deepening. The deepening of the mixing-layer (the contribution of Langmuir turbulence to the V_t term) is under-predicted if the same enhancement factor (equations 2.31 and 2.32) is used. To address this, we empirically modify the enhancement factor for V_t to

$$F_{LT}^{V_t} = 1 + 2.3La_{SL\theta'}^{-1/2}, \quad (2.33)$$

so that the bulk Richardson number calculation is now:

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U_L^r - U_L(z))^2 + (V_L^r - V_L(z))^2 + (V_t(z) \times F_{LT}^{V_t})^2}. \quad (2.34)$$

This form (2.33) has been found by optimizing the agreement between the one-dimensional model and the LES results of mixing-layer depth evolution and surface cooling, by varying both the slope and magnitude of the enhancement factor while maintaining that the enhancement factor approaches one in the large Langmuir number limit.

The KPP model with the Lagrangian shear (instead of the Eulerian shear) and with the enhancement factors, in the form of equations (2.30)-(2.32) for K_x and equation 2.33 for V_t , is called KPP-LT hereafter.

2.8.2 Results

The results of the comparison of the KPP-LT experiment against the LES-LT experiment (Figure 2.11) show much improved agreement compared to the KPP-iLT

experiment (Figure 2.5). Most noticeably, the prediction of the near-surface current has improved from an RMS difference of 0.15 m s^{-1} (with local differences greater than 0.5 m s^{-1}) to a much reduced RMS difference of 0.03 m s^{-1} (with local differences smaller than 0.15 m s^{-1}). Even as the eye of the storm passes, the prediction of the cooling remains within $0.1 \text{ }^{\circ}\text{C}$ of the LES-LT results in the center. After the eye has passed some under-cooling in the KPP-LT experiment is noticeable particularly on the left-hand side. However, this corresponds to the region where the KPP performs worse even in shear-only turbulence experiments, as discussed in Section 2.6.

We have also conducted identical experiments with slightly different enhancement factors to the turbulent velocity scale, and have found that the results are not sensitive to their detailed form, provided (1) the enhancement factor for K_x is capped at low Langmuir numbers, and (2) the enhancement factors for K_x and V_t are set differently, with higher values for V_t . The latter condition (2) is consistent with the previous scaling suggestions regarding the differences between the near-surface Langmuir mixing and the thermocline entrainment.

While the surface current and temperature are the most important quantities for air-sea interaction affecting tropical cyclone dynamics, the mixing scheme should also provide accurate prediction of the subsurface current and temperature. We find that the mixing-layer depth can exceed 200 m on the right side of a slow-moving storm. To examine performance at depth we present a typical example of the vertical profiles of current magnitude and temperature from the one-dimensional model with the various KPP methods and from LES-LT (Figure 2.12). The current direction is not greatly impacted by the different parameterizations and is not shown. Here, the results from the tropical cyclone with the moderate (5 m s^{-1}) translation speed and the 10 m initial mixed-layer depth are presented, but they are qualitatively similar for all experiments. The KPP-LT results show best agreement with the LES-

LT current profiles (top panels), with some minor discrepancy in the current shear profile near the base of the mixing-layer. The KPP-Lag does not correctly predict the mixing in the upper 30 m, although it performs better than the KPP-iLT. For the temperature anomaly profile (where the initial stratification has been subtracted from the simulated temperature profile), the KPP-LT performs slightly better than the other KPP methods, although the differences are small.

We examine the bulk impact of the various KPP mixing schemes by comparing the mixed-layer depth, h_{ML} , from the one-dimensional model and from the LES-LT (Figure 2.13). Here, the definition of h_{ML} is the depth where the maximum vertical temperature gradient ($\partial\Theta/\partial z$) occurs. These results show that h_{ML} from the KPP-LT agrees best with h_{ML} from the LES-LT compared to KPP-iLT and KPP-Lag (giving the smallest RMS difference). The reason why the LES mixed layer is deeper than the KPP prediction at later times is likely because in the LES there is still some residual turbulence that contributes to mixing near the base of the mixed layer even after the mixing layer depth has shoaled. We have also examined the mixing-layer depth, h , with the different KPP mixing schemes and have found that the results are similar to those of the mixed-layer depth, h_{ML} .

2.9 Constant moderate wind experiment

An additional test is performed to assess the performance of the KPP models under a constant moderate wind (10 m s^{-1}) condition as described in Section 2.3. The results at the end of the experiment are shown in Figure 2.14. The right panel shows the temperature anomaly, where the initial temperature profile for the constant wind experiment (defined in Section 2.3) is subtracted from the temperature profile. While the KPP-Lag and KPP-iLT models perform similarly in this test, they both under predict the cooling and deepening of the mixed-layer. The KPP-LT results yield the best agreement with the LES model, even if the model tuning has been done in

high-wind, tropical cyclone conditions. This experiment suggests that the improved performance of the KPP-LT model is largely due to the enhancement factors (both in the mixing coefficient and V_t) rather than using the Lagrangian shear in place of the Eulerian shear.

Global climate models that use the standard KPP (no Langmuir turbulence) tend to produce positive temperature biases and shallow mixed-layer biases in the Southern Ocean compared to observations (see Fan and Griffies, 2014). These biases are consistent with the warm temperature bias and under predicted mixed-layer depth produced by KPP-iLT in this constant wind experiment. Fan and Griffies (2014) suggest that Langmuir turbulence may be a missing component of the KPP model in current global climate models that would lead to under-predicted mixing. The Southern Ocean is where this impact is primarily observed due to the consistent wind forcing and resulting large swells. The best performance of the KPP-LT model in the constant wind experiment suggests that it may be a suitable replacement of KPP-iLT in climate models to reduce biases in these regions.

2.10 Conclusions

We have developed a series of KPP models and have tested their performance against equivalent LES model simulations in idealized tropical cyclone conditions. The standard KPP model, without explicit effects of the Langmuir turbulence, does a reasonable job in predicting the mixing-layer depth and the surface cooling under an idealized tropical cyclone, but it yields inaccurate prediction of the near-surface turbulent mixing and the mean current profile. When the KPP is retuned to remove the implicit Langmuir turbulence impact, it adequately reproduces the mean current and temperature simulated by the LES model without the wave effects (shear only mixing).

Replacing the Eulerian currents by the Lagrangian currents in this retuned KPP

model improves the currents, but degrades the surface cooling compared to the standard KPP model. The turbulent mixing coefficient estimated from the LES results is significantly larger than in the KPP model, that is, the KPP with the Lagrangian currents under-predicts turbulent mixing. The KPP parameterization of the mixing coefficient is improved by introducing an enhancement factor of the turbulent velocity scale, which is dependent on the Langmuir number. Separate enhancement factors are needed for the turbulent mixing coefficient and for the unresolved turbulent shear contribution.

This new modified version of the KPP model, with the Lagrangian currents replacing the Eulerian currents and the enhanced turbulent velocity scales, performs much better than the standard version of the KPP for predicting both temperature and current profiles under a tropical cyclone. This modified KPP model also performs best under constant moderate wind conditions (10 m s^{-1}). The improvement compared to the standard KPP model suggests that the new model may reduce under-mixing biases that have been previously documented in global climate model simulations in the regions with consistent wind forcing and large swells.

Here we present the summarized KPP-LT scheme (as implemented in this study). This KPP scheme is used to solve for the vertical turbulent momentum and heat (scalar) fluxes:

$$\overline{\mathbf{u}_{\mathbf{h}}'w'} = -K(\sigma)\frac{\partial\mathbf{U}_{\mathbf{h}L}}{\partial z} \quad (2.35)$$

$$\overline{\theta'w'} = -K(\sigma)\frac{\partial\Theta}{\partial z}. \quad (2.36)$$

The mixing coefficient is set:

$$K(\sigma) = hWG(\sigma)F_{LT}^K(\sigma). \quad (2.37)$$

where h is determined as the shallowest depth at which the bulk Richardson number

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U_L^r - U_L(z))^2 + (V_L^r - V_L(z))^2 + (V_t(z) \times F_{LT}^{Vt})^2} \quad (2.38)$$

exceeds the critical value (0.235 for this study). The enhancement to the unresolved turbulence is set:

$$F_{LT}^{Vt} = 1 + 2.3La_{SL\theta'}^{-1/2}, \quad (2.39)$$

and the enhancement to the turbulent mixing coefficient is set:

$$F_{LT}^K(\sigma) = 1 + (F_{LT}' - 1) * G(\sigma) / \max(G(\sigma)), \quad (2.40)$$

$$F_{LT}' = 1 + La_{SL\theta'}^{-1}, \quad La_{SL\theta'} \geq 0.8, \quad (2.41)$$

$$F_{LT}' = 2.25, \quad La_{SL\theta'} \leq 0.8. \quad (2.42)$$

The next step of this research will be to introduce this modified KPP model into three-dimensional ocean models. This will allow the impact of the Langmuir turbulence to be more thoroughly evaluated against available in-situ observations.

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Table 2.1. Location of 20 test sites for one-dimensional and LES models.

Site number	1 ^a	2 ^a	3 ^a	4 ^a	5 ^a	6 ^a	7 ^a	8 ^{ab}	9 ^{ab}	10 ^{ab}
Location (km)	300	200	150	130	110	90	70	50	30	10
Site number	11 ^{ab}	12 ^{ab}	13 ^{ab}	14 ^{ab}	15 ^{ab}	16 ^{ab}	17 ^{ab}	18 ^a	19 ^{ab}	20 ^a
Location (km)	0	-20	-40	-60	-80	-100	-120	-140	-200	-300

^a10 m mixed layer.

^b32 m mixed layer

Table 2.2. Description of various KPP model experiments.

KPP-iLT	KPP with implicit Langmuir turbulence
KPP-ST	retuned KPP to shear turbulence
KPP-Lag	retuned KPP with Lagrangian currents
KPP-LT	retuned KPP with Lagrangian currents and turbulent velocity scale enhancement

Figure 2.1. Wind speed (left), significant wave height (center), and magnitude of surface layer averaged Stokes drift (right) for a 5 m s^{-1} (upper) and 10 m s^{-1} (lower) translating idealized tropical cyclone. The horizontal and vertical axes indicate the distance from the storm center position normalized by the radius of maximum wind (RMW, 50 km). The white open circles in the right panels show the test site locations. The dashed circle represents RMW and the solid black line represents the storm track (moving from right to left).

Figure 2.2. LES results of near-surface (at 5 m depth) current magnitude (upper) and temperature anomaly (temperature minus initial temperature, lower) for a 5 m s⁻¹ translating tropical cyclone. The results are shown for: LES-ST (left), LES-LT (center), and the difference between the two (right). Note the white regions in the upper-right panel have small negative values. The horizontal and vertical axes indicate the distance from the storm center position normalized by the RMW (50 km).

Figure 2.3. LES transect (at 50 km to right of storm center) of current magnitude (upper) and temperature (lower) for a 5 m s⁻¹ translating tropical cyclone. The results are shown for: LES-ST (left), LES-LT (center), and the difference between the two (right). The horizontal axes indicates the distance from the storm center position normalized by the RMW (50 km).

Figure 2.4. Cosine of the misalignment angle between the directions of the Reynolds stress (θ_τ) and vertical gradient ($\theta_{\partial \mathbf{U}_h / \partial z}$) of the Lagrangian current (upper) and the vertical gradient of the Eulerian current (lower) for 5 m s^{-1} translating tropical cyclone from the LES-LT simulations. The test sites are from 60 km to the left (left panels), the storm center (middle panels), and 50 km to the right (right panels). The horizontal axis indicates the distance from the storm center position normalized by the RMW (50 km). Note that the color scale saturates at 0, but negative values can exist within the RMW. The solid black line represents the mixing-layer depth using the bulk Richardson number criterion. Values beneath the mixing-layer have been filtered out.

Figure 2.5. Results of current magnitude (upper panels) and temperature anomaly (temperature minus initial temperature, lower panels) at 5 m depth. Left panels: results from the one-dimensional model with the KPP-iLT on the vertical axis vs results from the LES-LT model on the horizontal axis. The ‘E’ value is the RMS difference. Center panels: difference between results from KPP-iLT and LES-LT for the 5 m s⁻¹ translating tropical cyclone. The horizontal and vertical axes indicate the distance from the storm center position normalized by the RMW (50 km). Right panels: same as center panels but for the 10 m s⁻¹ translating tropical cyclone.

Figure 2.6. Same as Figure 2.5, but KPP-ST is compared to LES-ST.

Figure 2.7. Same as Figure 2.5, but KPP-Lag is compared to LES-LT.

Figure 2.8. Vertical profile of non-dimensional mixing coefficient at 4 hours prior to peak wind (upper), at peak wind (middle) and 4 hours after peak wind (lower) at locations -40 km (left) and +50 km (right) as derived from the LES turbulent stress (black dots) compared to the parameterized profile in the KPP model without any enhancement (solid red) and with the enhancement proposed in equation 2.27 (solid blue). The dashed blue shows the KPP model result where the enhancement is reduced near the surface and near the base of the mixed layer (equation 2.30).

Figure 2.9. Spatial distribution of Langmuir number for a 5 m s^{-1} (top) and 10 m s^{-1} (bottom) translating tropical cyclone. The horizontal and vertical axes indicate the distance from the storm center position normalized by the RMW (50 km). Left: La_t (eq 2.22) based on the surface Stokes drift (lower wavelength limit of 1 m); Center: La_{SL} (eq 2.23) based on the surface layer (top 20% of the mixing-layer) averaged Stokes drift; Right: $La_{SL\theta'}$ (eq 2.25) based on the surface layer averaged Stokes drift and corrected for misalignment between current shear and waves. The white open circles in the right panels show the test site locations. The color scale saturates above 0.8, but the Langmuir number can be much larger when the Stokes drift is small and/or the wind stress is large.

Figure 2.10. LES enhancement factor (ratio of maximum normalized LES mixing coefficient to maximum normalized KPP mixing coefficient) (left panels), and mixing-layer averaged vertical velocity variance scaled by the friction velocity squared (right panels). Both quantities are plotted against the surface layer Langmuir number La_{SL} (upper panels) and the surface layer projected Langmuir number $La_{SL\theta'}$ (lower panels). The marker color locations are given in the map above. White circles with vertical bars are bin averages with standard deviations. (The lines labeled in the legends are described in the text.)

Figure 2.11. Same as Figure 2.5, but KPP-LT is compared to LES-LT.

Figure 2.12. Vertical profiles (from experiments listed in legend) of current magnitude (upper) and temperature anomaly (temperature minus initial temperature, lower) at time of peak wind. The test sites are 60 km to the left (left panels), the storm center (middle panels), and 50 km to the right (right panels).

Figure 2.13. Mixed layer depth (defined as depth where maximum $\partial\Theta/\partial z$ occurs) from all simulations comparing KPP-iLT vs LES-LT (left), KPP-Lag vs LES-LT (center), and KPP-LT vs LES-LT (right) for both translation speeds and both initial mixed layer depths. The value E is the RMS difference computed over all points.

Figure 2.14. Vertical profiles at 24 hours for current magnitude (left) and for temperature anomaly (temperature minus initial temperature, right) for a constant 10 m s^{-1} experiment. The colors represent the same experiments as in Figure 2.12.

MANUSCRIPT 3

Impact of sea-state dependent Langmuir turbulence on the ocean
response to a tropical cyclone

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Abstract

Tropical cyclones are fueled by the air-sea heat flux, which is reduced when the ocean surface cools due to mixed-layer deepening and upwelling. Wave-driven Langmuir turbulence can significantly modify these processes. This study investigates the response of the ocean to tropical cyclones using three versions of the KPP vertical mixing scheme. The first, KPP-ST, is tuned to conditions of shear turbulence only. The second, KPP-iLT, is tuned to typical ocean conditions (with implicit typical Langmuir turbulence) with no explicit sea-state dependent modifications. The third, KPP-LT, includes explicit sea-state dependent Langmuir turbulence effects by parameterizing the vertical turbulent momentum flux using the Lagrangian current (Eulerian current plus Stokes drift) with an enhanced vertical mixing coefficient that depends on the turbulent Langmuir number. Both KPP-iLT and KPP-LT enhance sea surface cooling due to vertical mixing compared to KPP-ST, but KPP-iLT significantly underestimates the cooling particularly on the left hand side of propagating storms. While KPP-LT significantly reduces and homogenizes currents inside the mixing-layer by enhanced vertical momentum mixing, KPP-iLT has little impacts on currents. Therefore, KPP-iLT introduces further error in predicting horizontal advection of heat near the cold wake as well as sea surface cooling due to upwelling for stationary and slow moving storms. These results suggest that accurate predictions of the Langmuir turbulence effects on upper-ocean response to a tropical cyclone requires an explicit sea-state dependent Langmuir turbulence parameterization.

3.1 Introduction

The energy budget of a tropical cyclone is primarily dictated by air-sea fluxes of heat and momentum (Emanuel, 1991). The total heat flux between the ocean and atmosphere under tropical cyclone force winds is dominated by latent heat flux. Generally, in the warm tropical waters where a hurricane forms, the latent heat flux pumps energy into the atmosphere, occasionally exceeding $1,000 \text{ W m}^{-2}$ over warm sea surface temperature (SST) during high winds. SST can be greatly reduced under a tropical cyclone due to rapid mixed-layer deepening due to vertical mixing and upwelling of cold, sub-pycnocline waters in the presence of horizontally divergent currents. Then, the air-sea heat flux and the energy available to the storm can be significantly reduced (see Price, 1981; Emanuel, 1999; Bender and Ginis, 2000; Ginis, 2002). For typical tropical cyclone atmospheric conditions, the latent heat flux can be reduced by an order of magnitude if SST cools by only a few degrees, because the relative humidity of the fully saturated air adjacent to the air-sea interface significantly decreases as the SST cools. It is therefore vitally important to accurately model the upper-ocean response to a tropical cyclone and the resulting SST cooling for accurate forecasts of hurricane wind structure, wind intensity, and storm track.

The ocean surface temperature under a tropical cyclone is controlled by both one-dimensional (vertical mixing/diffusion) and three-dimensional (upwelling and horizontal advection) processes, with each process dominating in certain locations relative to the storm center and for certain storm characteristics (such as size, intensity, and forward translation speed). The one-dimensional process is primarily driven by the shear-driven vertical mixing of cold, sub-thermocline water into the mixed-layer, which is achieved when the mean current shear overcomes the stable density gradient associated with the thermocline. The cooling due to large-scale three-dimensional processes can become important under tropical cyclones due to the strong spatial vari-

ability of the surface forcing (i.e. Yablonsky and Ginis, 2009; Vincent et al., 2012). Vertical advection, or upwelling, is a prominent feature of a slowly moving tropical cyclone where horizontal current divergence drives a vertical current that advects sub-thermocline water into the near-surface mixed-layer. Horizontal fluxes of heat is also an important mechanism, particularly near the cold wake associated with a tropical cyclone, because cool water can be advected outward by the cyclonic, wind-driven currents and redistribute heat (Vincent et al., 2012). The effects of upwelling and horizontal advection are not as important when the translation speed becomes too fast ($>\sim 5 \text{ m s}^{-1}$) for three-dimensional processes to establish (Yablonsky and Ginis, 2009).

The vertical turbulent mixing of momentum and scalar quantities such as temperature and salt can be significantly enhanced due to the Langmuir turbulence that is driven by the interaction between the Stokes drift (due to surface waves) and the Eulerian current vorticity (this is often referred to as the CL2 vortex force, originally proposed by Craik and Leibovich (1976)). This mechanism enhances the vertical mixing over the entire mixed-layer even if the Stokes drift is confined in a very thin surface-layer, as demonstrated by McWilliams et al. (1997) and many subsequent Large Eddy Simulation (LES) studies (e.g. Noh et al., 2004; Polton and Belcher, 2007; Kukulka et al., 2009). Because the intensity of the Langmuir turbulence depends on the relative importance of the wind forcing and the wave forcing, it strongly depends on the surface wave field (or sea states). Therefore, existing upper-ocean mixing parameterizations (without explicit sea-state dependence) may introduce significant errors in conditions where the surface wave field is not in equilibrium with local wind forcing (Fan and Griffies, 2014; Li et al., 2015). The spatial and temporal variability of the Langmuir turbulence intensity is particularly significant in tropical cyclone conditions because the ocean surface wave field is complex, often dominated by large

waves misaligned with local wind (Sullivan et al., 2012; Rabe et al., 2015). This has led to the development of numerous modifications to existing mixing parameterizations to account for variability of the Langmuir turbulence due to sea states. These include modifications to the Mellor-Yamada scheme (Mellor and Yamada, 1982), such as Kantha and Clayson (2004) and Harcourt (2013, 2015), and modifications to the K Profile Parameterization (KPP, Large et al., 1994), such as McWilliams and Sullivan (2000); Smyth et al. (2002); McWilliams et al. (2012); Reichl et al. (In Review).

A recent modification to the KPP model by Reichl et al. (In Review, hereafter RWHGK) has been developed by matching the performance of the KPP, imbedded in a one-dimensional ocean model, with concurrent LES results with identical initial conditions and wind and wave forcing, in a wide range of transient wind and wave conditions under tropical cyclones. The study has confirmed that the intensity of the Langmuir turbulence is correlated with the turbulent Langmuir number that characterizes the significance of the wave forcing relative to the wind forcing, as suggested by previous studies. The study has also demonstrated that the Langmuir turbulence significantly reduces current magnitude inside the mixed-layer due to vigorous momentum mixing. The existing (standard) KPP parameterization is tuned to include typical (average) Langmuir turbulence effects. Therefore, it is able to predict the (typical) mixed-layer deepening events and SST cooling reasonably well; it does not contain the sea state dependence of such events. For example, it incorrectly predicts the spatial pattern of SST cooling. In addition, the existing KPP appears to entirely miss the enhanced momentum mixing and current magnitude reduction due to the Langmuir turbulence. As stated previously, these mixed-layer currents control the three-dimensional response of the ocean to the hurricane, including the upwelling and the horizontal advection of heat. Therefore, the reduced currents due to the Langmuir turbulence may play a significant role in modifying the SST cooling.

In this study, we investigate the impact of the sea-state dependent Langmuir turbulence on the three-dimensional ocean response to a hurricane. We achieve this by introducing the newly modified KPP by RWHGK in a three-dimensional ocean model coupled with a surface-wave prediction model. Numerical experiments are conducted using idealized tropical cyclone wind forcing with varying storm translation speeds. The results are then compared with the experiments using the KPP tuned for shear-only turbulence (no Langmuir turbulence) to clarify the overall impact of the Langmuir turbulence. The results are also compared with the experiments using the standard KPP that includes the typical (average, independent of sea states) Langmuir turbulence effects. This addresses an important practical question of whether the explicit (sea state dependent) Langmuir turbulence parameterization is necessary to accurately predict the SST cooling under tropical cyclones. In our investigation, we also explore the impact of the Coriolis-Stokes acceleration (CS) on the ocean response to the tropical cyclone. The CS is due to the interaction between the Stokes drift and the planetary vorticity in the water column (Hasselmann, 1970; Polton et al., 2005) and is another mechanism by which the Stokes drift can modify the mean current and possibly the SST cooling.

3.2 Methods

3.2.1 Description of models

MPIPOM-TC with KPP

A two-way coupled ocean and wave model has been implemented for this study. The ocean component is the recently updated MPIPOM-TC (Yablonsky et al., 2015a), which is a branch in the hierarchy of the Princeton Ocean Model (POM, Blumberg and Mellor, 1987; Mellor, 2004). This version of POM utilizes MPI (message passing interface) capabilities in addition to having both one-dimensional (vertical only) and three-dimensional (vertical+horizontal) run options (Yablonsky and Ginis, 2009). In this study, the vertical resolution of the model is increased from the operational

version of POM in the Hurricane Weather and Research Forecasting model (HWRf, Yablonsky et al., 2015b), which has 23 sigma levels, to 60 levels (with a constant ocean depth of 2500 m). A near-surface resolution of 4.5 m is kept constant over the upper 200 meters of the water column to better resolve the near-surface boundary layer mixing and the evolution of the mixing-layer depth in the KPP model. The traditional Mellor-Yamada (Mellor and Yamada, 1974, 1982) vertical mixing routine of POM has been replaced by a turbulent mixing subroutine based on the KPP model (Large et al., 1994), which is described below.

The standard KPP model solves for the vertical turbulent flux terms as:

$$\overline{\mathbf{u}_h'w'} = -K_M(z) \frac{\partial \mathbf{U}_h}{\partial z} + \Gamma_U \quad (3.1)$$

$$\overline{\theta'w'} = -K_\theta(z) \frac{\partial \Theta}{\partial z} + \Gamma_\theta \quad (3.2)$$

where $\mathbf{U}_h$ is the mean horizontal current, Θ is the mean potential temperature, K_X is the eddy diffusivity of heat ($X = \theta$) and momentum ($X = M$), z is the vertical coordinate (positive upward), and $\overline{\mathbf{x}'w'}$ is the covariance of the perturbation component of either the horizontal velocity ($\mathbf{x} = \mathbf{u}_h$) or the temperature ($\mathbf{x} = \theta$) with the perturbation of the vertical (w) velocity. In this study, we neglect the surface buoyancy flux and the resulting non-local Γ terms because the contribution of the surface buoyancy flux to the turbulence is relatively small in a high-wind region of a hurricane. We also assume that K_θ and K_M are equal and simply express them as K hereafter, which is consistent with the original KPP formulation when the surface buoyancy flux is negligible. In the KPP model, the vertical profile of K within the surface mixing-layer is parameterized as:

$$K(\sigma) = hWG(\sigma), \quad (3.3)$$

where h is the mixing-layer depth, W is the turbulent velocity scale, $G(\sigma)$ is the non-dimensional turbulent mixing shape function, and $\sigma = -z/h$. In this study

W is approximated as $W = \kappa u_*$ (where κ is the von Kármán constant and u_* is the magnitude of the surface friction velocity) and $G(\sigma)$ is approximated as $G(\sigma) = \sigma(1 - \sigma)^2$.

In the KPP model, h is defined as the shallowest depth where the bulk Richardson number exceeds the critical value:

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U^r - U(z))^2 + (V^r - V(z))^2 + V_t^2(z)} < Ri_c, \quad (3.4)$$

where B is the buoyancy, U and V are the horizontal components of the mean current, and V_t is the unresolved turbulent shear contribution. In this study, the reference values (superscript r) are defined as averages over the upper 10% of the mixing-layer. The unresolved turbulent shear contribution is solved for as:

$$V_t^2(z) = \frac{C_v(-\beta_T)^{1/2}}{Ri_c \kappa^2} (c_s \epsilon)^{-1/2} |z| N W, \quad (3.5)$$

where N is the stability frequency and the constants are $C_v = 1.6$, $\beta_T = -0.2$, $c_s = 98.96$, and $\epsilon = 0.1$ following Large et al. (1994).

We employ three versions of the KPP model in this study. The first, KPP-ST, is tuned to conditions of shear turbulence only (no Langmuir turbulence). The second, KPP-iLT, is tuned to typical ocean conditions (with typical Langmuir turbulence) but includes no explicit sea-state dependent modifications. The third, KPP-LT, includes explicit sea-state dependent Langmuir turbulence effects. These three versions have been proposed in RWHGK. In the study the performance of each KPP model was evaluated under a wide range of tropical cyclone wind and wave conditions in the one-dimensional General Ocean Turbulence Model (GOTM, Umlauf et al. 2005) by comparing with concurrent LES simulations with identical mean initial conditions and forcing parameters. The first and second versions are the standard KPP model as described above, but with different critical Richardson numbers. It is known that the critical Richardson number used in KPP is dependent on the vertical resolution

of the ocean model. Since the vertical resolution of the ocean model in this study is coarser than that used in RWHGK, we have retuned the critical Richardson numbers to 0.27 for KPP-ST and 0.35 for KPP-iLT.

The development of the third version, KPP-LT, is discussed in detail by RWHGK and is briefly summarized below. This model differs from the KPP-ST model in two ways. First, the KPP-LT model utilizes the Lagrangian current in place of the Eulerian current (where the Lagrangian current is the Eulerian current plus the Stokes drift) in the calculation of the turbulent momentum flux:

$$\overline{\mathbf{u}_h' w'} = -K(z) \frac{\partial \mathbf{U}_L}{\partial z}, \quad (3.6)$$

and in the calculation of the bulk Richardson number:

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U_L^r - U_L(z))^2 + (V_L^r - V_L(z))^2 + V_t^2(z)} < Ri_c, \quad (3.7)$$

where the subscript $_L$ is used to differentiate between the Eulerian and Lagrangian currents.

Second, the KPP-LT model introduces an enhancement factor to the eddy viscosity profile and the unresolved turbulent shear contribution to the bulk Richardson number. The enhancement to the eddy viscosity profile, F_{LT} , is given by:

$$K(\sigma) = hWG(\sigma)F_{LT}(\sigma), \quad (3.8)$$

with

$$F_{LT}(\sigma) = 1 + (F'_{LT} - 1) * G(\sigma) / \max(G(\sigma)), \quad (3.9)$$

$$F'_{LT} = 1 + La_{SL\theta'}^{-1}, \quad La_{SL\theta'} \geq 0.8, \quad (3.10)$$

$$F'_{LT} = 2.25, \quad La_{SL\theta'} \leq 0.8. \quad (3.11)$$

The enhancement to the unresolved turbulent shear, F_{LT}^{Vt} , is different from the enhancement to the eddy viscosity, and takes the form:

$$F_{LT}^{Vt} = 1 + 2.3La_{SL\theta'}^{-1/2}. \quad (3.12)$$

So that the bulk Richardson number calculation is now:

$$Ri_b(z) = \frac{(B^r - B(z))|z|}{(U_L^r - U_L(z))^2 + (V_L^r - V_L(z))^2 + (V_t(z) \times F_{LT}^{Vt})^2}. \quad (3.13)$$

The surface-layer averaged turbulent Langmuir number, $La_{SL\theta'}$, is defined as the square root of the ratio of the friction velocity to the Stokes drift averaged over the surface-layer (upper 20% of the mixing-layer), corrected for the misalignment between the Stokes drift and the Langmuir turbulence:

$$La_{SL\theta'} = \sqrt{\frac{u_*}{\langle |\mathbf{u}_S| \rangle_{SL}} \frac{1}{\max(\cos(\theta_{Waves} - \theta_{Lag}), 10^{-8})}}, \quad (3.14)$$

where θ_{Waves} is the direction of the Stokes drift averaged over the surface-layer, and the direction of the Langmuir turbulence is approximated by the direction of the Lagrangian shear, θ_{Lag} , averaged over the surface-layer following Van Roekel et al. (2012). The definition of the surface-layer averaged turbulent Langmuir number in Harcourt and D'Asaro (2008) takes the surface-layer averaged Stokes drift relative to the Stokes drift at the base of the mixing-layer, which we neglect as in RWHGK.

In the ocean model of this study, the Buossinesq horizontal momentum equation is solved including the surface wave impact,

$$\frac{\partial \mathbf{U}_h}{\partial t} + \left(\mathbf{U}_h \cdot \frac{\partial}{\partial \mathbf{x}_h} + W \frac{\partial}{\partial z} \right) \mathbf{U}_h + \mathbf{f} \times (\mathbf{U}_h + \mathbf{U}_S) = -\frac{1}{\rho_0} \frac{\partial P}{\partial \mathbf{x}_h} - \frac{\partial}{\partial z} \left(\nu \frac{\partial \mathbf{U}_h}{\partial z} + \overline{\mathbf{u}_h' w'} \right), \quad (3.15)$$

where $\mathbf{f}$ is the Coriolis vector (0,0,f) using the f-plane definition centered at 22.4° N, ρ_0 is the mean density, P is the Reynolds averaged dynamic pressure, ν is the molecular diffusivity, and the instantaneous horizontal ($\mathbf{u}_h$) and vertical (w) current components are decomposed into mean ($\mathbf{U}_h$, W) and perturbation ($\mathbf{u}_h'$ and w') components. (In the actual MPIPOM-TC implementation, these calculations are performed in a σ -level coordinate system, which is identical to the Cartesian expression given here with the constant bathymetry of this study.) In this equation (equation

3.15) the Stokes drift ($\mathbf{U}_S$) appears in the Coriolis term (Coriolis-Stokes acceleration). In addition, the KPP vertical momentum flux parameterization is modified by the sub-resolved-scale Stokes-Vortex force (Langmuir turbulence) as described earlier. Although more complete equations include the resolved scale Stokes-vortex force in the mean equations (interaction of the Stokes drift and relative vorticity, see e.g., McWilliams and Restrepo (1999)), we follow the scaling arguments presented in Fan et al. (2010) and include the Stokes drift in the Coriolis term only. As discussed by McWilliams and Fox-Kemper (2013), including just the Coriolis-Stokes acceleration can introduce a pseudo-counter-Stokes current that is on the same order as the mean current. Therefore, we also include the impact of the Stokes drift on the mass conservation by calculating the horizontal divergence of the vertically-integrated Stokes drift, which is used to modify the vertical velocity surface boundary condition. In a steady state the vertically-integrated Stokes drift exactly cancels the Ekman transport that drives this pseudo-counter-Stokes current resulting from the Coriolis-Stokes acceleration. In a transient state this cancellation does not occur as described below. Without including the Stokes drift effect on the mass conservation, we may introduce a non-physical upwelling or downwelling in a steady state if the vertically integrated Stokes drift has horizontal divergence or convergence.

The drag coefficient used in this study is identical to that proposed by Sullivan et al. (2012):

$$Cd = \begin{cases} 0.0012, & : |u_{10}| < 11 \text{ m s}^{-1} \\ (0.49 + 0.065|u_{10}|) \times 10^{-3}, & : 11 \text{ m s}^{-1} \leq |u_{10}| \leq 20 \text{ m s}^{-1} \\ 0.0018, & : 20 \text{ m s}^{-1} < |u_{10}|. \end{cases} \quad (3.16)$$

We have tested the sensitivity of these experiments to different drag coefficient models. While the overall cooling is obviously affected by the drag coefficient choice, the impact of the Langmuir turbulence (the difference among the three versions of the KPP) remains qualitatively similar.

WAVEWATCH III component

The wave model component of the coupled system is version 3.14 of the WAVEWATCH III 3rd generation wind-wave model (Tolman, 2009). (Although version 3.14 is not the most recent public release, the physics packages used in version 3.14 is included in the later release as well.) The wind-input source term of the version 3.14 has been modified, using a reduced, wave-age dependent drag coefficient, to optimize performance under tropical cyclone conditions as described by Moon et al. (2008) and demonstrated by Fan et al. (2009). The wave spectrum is defined over 40 logarithmically spaced frequencies (with a minimum frequency of 0.0285 Hz) and 48 evenly spaced directions. In the model the Stokes drift is calculated as:

$$\mathbf{u}_S(z) = \int_0^{k_{UL}} \int_0^{2\pi} \Psi(k, \theta) 2\omega \exp(2kz) \mathbf{k} d\theta dk, \quad (3.17)$$

where $\Psi(k, \theta)$ is the wavenumber-direction variance spectrum, k is the wavenumber, θ is the wave direction, $\mathbf{k}$ is the wavenumber vector, $\omega = \sqrt{gk}$ is the wave angular frequency, and k_{UL} is the upper bound of the wavenumber integration corresponding to a wavelength of 1 m. The depth-dependent Stokes drift values are then passed to the ocean model for calculation of the turbulent Langmuir number, the Lagrangian shear, and the bulk Richardson number used in the KPP-LT routine.

In principle many parameters should be passed back and forth in the two-way fully coupled ocean-wave model system. For this study we have disabled the impact of the ocean model on the wave model, using the model as a one-way coupled system. The communication from the wave model to the ocean model is restricted to passing the Stokes drift only. In this way, we have simplified the problem to focus on the effect of the Langmuir turbulence on upper-ocean response under prescribed wind and wave conditions eliminating feedback effects between the two models. We have ignored the effect of the ocean current on the waves, the effect of growing/decaying waves on the momentum flux into the ocean (Fan et al., 2010), as well as all the effects of waves

and currents on the atmospheric model (i.e., wind forcing). We save investigation of the fully coupled models for future studies.

3.2.2 Experiment design

A large, deep-water computational domain is identically defined for both the ocean and the wave model. The ocean and wave model domains are projected into a curvilinear coordinate system based on the Earth’s latitude and longitude, and model communication and interpolation is then based on the latitude/longitude grid. Both models are simulated on 1/12 degree resolution domains. The horizontal domain of both models is much larger than the storm size so that boundary conditions are not relevant at the time scales considered. The model is initialized with a homogenous salinity (35 PSU), such that the density structure is determined entirely by the temperature. The initial temperature profile consists of a 20 m mixed-layer of 29.25°C, a layer of constant temperature gradient of 0.1°C m⁻¹ from the base of the mixed-layer until 120 m depth, a layer of smooth transition between 120m and 1300 m depth, and a realistic lower ocean temperature (4° C) below 1300 m. Due to the short total simulation time the deep ocean is effectively isolated from the surface. The temperature profile is similar to the Gulf of Mexico Common Water temperature profile of Yablonsky and Ginis (2009).

Idealized tropical cyclone winds based on the Holland wind profile (Holland, 1980) are inserted into the domain, and translated at varying speeds from east to west. The maximum wind speed is set to 50 m s⁻¹ and the radius of maximum wind is set to 50 km. We examine three different translation speeds including the stationary case, the 0.6 deg hr⁻¹ (~ 2.85 m s⁻¹) case, and the 1.2 deg hr⁻¹ (~ 5.7 m s⁻¹) case. For each case, the storm maximum wind is spun up from 0 m s⁻¹ to the prescribed maximum wind speed over the initial 24 hours to minimize the inertial response due to the sudden onset of winds (particularly for the stationary case).

The ocean current is initially at rest, and in the KPP model the initial turbulence structure is determined entirely by the mixing-layer depth. We have also explored the sensitivity of our results by varying the initial temperature profile, the radius of maximum wind, and the maximum wind speed, which are briefly discussed later.

3.3 Results

3.3.1 Wind and wave fields

In Figure 3.1, the results of the wind stress (top panels), the surface 4.5 m layer averaged Stokes drift (middle panels), and the turbulent Langmuir number defined as equation 3.14 (bottom panels) are presented for the 0 m s^{-1} (left panels), 2.85 m s^{-1} (center panels), and 5.7 m s^{-1} (right panels) translating storms. The middle row panels also show contours of the mixing-layer depth defined in KPP, since the mixing-layer depth is utilized to compute the surface-layer averaged Stokes drift and the turbulent Langmuir number. These results are obtained using the KPP-LT including the CS acceleration. Although the mixing-layer depth and the turbulent Langmuir number vary depending on the version of the KPP used and the presence/absence of the CS, their spatial patterns remain quite similar (not shown). The stationary storm results are presented 24 hours after the initial 24 hour spin-up (48 hours total). The 2.85 m s^{-1} and 5.7 m s^{-1} results are simulated for an additional 24 hours (72 hours total) so that the wave field and current field become quasi-steady with respect to the reference frame moving with the storm. The low value of the turbulent Langmuir number on the right of the moving storms can be explained by the larger developed wave field. In the rear of the storm, the secondary minimum in the turbulent Langmuir number corresponds to the shallowing of the mixing-layer, which increases the surface-layer averaged Stokes drift.

3.3.2 Temperature and current fields

We present the results of the overall temperature and current fields in the absence of surface wave effects (KPP-ST-noCS) in Figures 3.2-3.4. Figure 3.2 shows the spatial fields of the surface temperature (top panels) and the Eulerian current at 11.25 m depth. (The choice of this particular depth is explained later in this subsection.) The vertical transects in the north-south direction at selected locations are shown for the temperature (Figure 3.3) and for the current magnitude (Figure 3.4). Figure 3.3 shows that temperature is well mixed and vertically almost uniform inside the mixing-layer. In contrast, the current magnitude near the storm center is typically stronger at the surface and decreases with depth. It is more uniform in the rear of the storm (Figure 3.4).

The top left panel of Figure 3.2 shows that the temperature is significantly cooled (by over 10° C) near the storm center in the stationary case. This is mainly caused by upwelling due to horizontal current divergence (Figure 3.2 bottom left panel). In nature, such strong cooling is unrealistic because of the feedback effect; the reduced heat flux would rapidly weaken the storm. However, in our idealized experiment the prescribed wind stress continues to force the upwelling. As the storm translation speed increases, the cooling due to the upwelling effect is reduced, but the cooling due to the mixed-layer deepening becomes more important and is stronger on the right of the storm (Figure 3.2 top middle and top right panels). The rightward bias of cooling appears because the current and the resulting shear-driven turbulence are stronger on the right of the storm due to the resonance effect (Figure 3.2 middle row center and middle row right panels). The near-surface current, excited by the wind stress, gradually turns to the right (inertial response in the Northern Hemisphere). On the right of the track the wind vector itself also turns to the right (at a fixed location) and continually forces the surface current, causing the rightward bias of the current.

For the 2.85 m s^{-1} translating storm, the current magnitude peak on the rear left is caused not only by the resonance process but also by the geostrophic component of the current induced by the pressure gradient associated with the storm upwelling (Ginis, 2002). The geostrophic current does not cause enhanced cooling. Indeed, this current peak disappears in the one-dimensional simulation discussed in Section 3.4.

Next we present the results of the overall temperature and current fields with the surface wave effects, that is, with the Langmuir turbulence (KPP-LT) and the CS (this case is denoted KPP-LT-CS) in Figures 3.5-3.7. Although the spatial patterns of the surface temperature and the current (11.25 m depth) in Figure 3.5 are qualitatively similar to those in Figure 3.2 without the wave effects, the quantitative difference is significant. The wave effects tend to enhance the surface temperature cooling and weaken the currents. Another striking difference appears in the vertical structure of the current (Figure 3.7 compared to Figure 3.4). With the surface wave effects the current is more vertically uniform (suggesting enhanced vertical mixing due to the Langmuir turbulence) and does not intensify near the surface. Figure 3.8 shows the magnitude of the Stokes drift in the same vertical transects as in Figure 3.7. Unlike the Eulerian current, the Stokes drift is mostly confined near the surface and becomes insignificant below 10 m depth. Because the turbulent mixing occurs due to the Lagrangian shear rather than the Eulerian shear (RWHGK), the Stokes drift tends to introduce an Eulerian current that cancels the Stokes drift (counter-Stokes current) in order to reduce the Lagrangian shear when the vertical mixing is strong. This tends to decrease the magnitude of the Eulerian current near the surface as seen in Figure 3.7. (The Lagrangian current does not decrease near the surface.) Since the Eulerian current is influenced by the Stokes drift near the surface, we present the horizontal current fields at 11.25 m depth below the influence of the Stokes drift throughout this study.

In the next two subsections we investigate the impact of the waves in more detail.

3.3.3 Impact of Coriolis-Stokes acceleration

Before investigating the effect of Langmuir turbulence, we first address the impact of the CS acceleration. Figure 3.9 shows the difference of the surface temperature and the current (11.25 m depth) between KPP-ST-CS (simulation with KPP-ST and with the CS acceleration) and KPP-ST-noCS (simulation with KPP-ST and without the CS acceleration). The colormap shows the difference of the temperature and the vectors show the difference of the current. For reference, the isotherms from the KPP-ST-CS results are superimposed on this plot.

For the stationary storm, the temperature at the center is lower in the simulation with the CS acceleration (left panel). This is a consequence of a slight increase of the outward component of the storm-induced current velocity, which increases the current divergence and the associated vertical upwelling of cold water. Since the Stokes drift is nearly cyclonic (in the direction of the wind), the CS acceleration enhances the near surface outward currents in the early stage. Ultimately in the steady state, this outward force would produce a steady Ekman response and the vertically integrated Ekman transport would be in an anti-cyclonic direction, exactly canceling the vertically integrated Stokes drift. However, this steady state solution is not achieved in our 48 hour simulation (including 24 hour spin up). In fact, such a steady state is unlikely to occur because a tropical cyclone rarely remains stationary for longer than a day in nature.

For the 2.85 m s^{-1} translating tropical cyclone, there is a warm anomaly to the right-rear of the storm's cold-wake of up to 0.3° C and a cold anomaly along the left side of the cold wake of up to 0.4° C (center panel). On the right side of the cold wake, the CS acceleration introduces an excess current across the background isotherms from the warm water to the north toward the cold water in the wake,

which causes the warm anomaly. There is a similar excess current (also due to the CS acceleration) on the left side of the cold wake directed from the cold wake toward the warmer water to the south, which causes the cold anomaly. The cold anomaly on the left is stronger than the warm anomaly on the right because the CS acceleration also increases the cooling due to upwelling (similar to the stationary case, although the enhanced upwelling is much weaker compared to the stationary case). For the faster 5.85 m s^{-1} translating tropical cyclone (right panel), the upwelling does not have enough time to set-up before the storm moves away. There is still a slight warm anomaly on the right and a slight cool anomaly in the rear due to the horizontal advection, though the anomalies are much weaker than those for the slower moving storm.

In summary, the impacts of the CS acceleration are two-fold. First, the CS acceleration increases upwelling due to the divergent current acceleration. This effect is the largest for the stationary storm and rapidly decreases as the storm translation speed increases. Second, for a moving storm the CS acceleration warms the right hand side of the cold wake and cools the left side of the cold wake due to modified horizontal advection. This effect also weakens as the storm translation speed increases.

3.3.4 Impact of Langmuir turbulence

Next, the impacts of the Langmuir turbulence on the near-surface temperature and the currents are investigated. This problem has been previously studied using LES and one-dimensional column models (see Sullivan et al., 2012; Rabe et al., 2015, RWHGK). The impact of Langmuir turbulence on the three-dimensional ocean response, including the impact on horizontal advection and upwelling, has not been investigated.

In this subsection the CS acceleration is always included in order to isolate the Langmuir turbulence impact. The top panels of Figure 3.10 show the difference

of the surface temperature and the current (11.25 m depth) between KPP-LT-CS (simulation with KPP-LT and with the CS acceleration) and KPP-ST-CS (simulation with KPP-ST and with the CS acceleration). The sum of the fields shown in Figure 3.9 and the top-row panels of Figure 3.10 is the result of the total wave impact (due to Langmuir turbulence and CS), which is equal to the difference between Figure 3.5 (KPP-LT-CS) and Figure 3.2 (KPP-ST-noCS).

The impact of the Langmuir turbulence is significantly stronger than the impact of the CS acceleration. This means that the overall impact of surface waves is predominantly due to the Langmuir turbulence. The Langmuir turbulence enhances surface cooling during moving storms by up to 0.7° C, particularly on the left side. For the stationary storm the Langmuir turbulence effect is more subtle, with both warming and cooling occurring at different locations. The vector current difference between KPP-LT-CS and KPP-ST-CS is mostly in the opposite direction of the current vector itself (Figure 3.2 lower panel). This shows that the Langmuir turbulence reduces the current at 11.25 m depth.

In order to better understand the Langmuir turbulence effect, we carry out two more simulations of KPP-LT-CS-1d and KPP-ST-CS-1d. These two simulations are identical to KPP-LT-CS and KPP-ST-CS, respectively, except that the ocean model is run in one-dimensional mode (without horizontal advection of momentum and heat and gradient of pressure). The difference of the surface temperature between KPP-LT-CS-1d and KPP-ST-CS-1d is shown in the middle row panels in Figure 3.10. These panels show the impact of the Langmuir turbulence on the one-dimensional processes only (which is the cooling due to entrainment of cooler water from below the mixed-layer).

The Langmuir turbulence enhances the cooling everywhere in the one-dimensional simulation because the surface waves always decrease the turbulent Lang-

muir number and enhance the vertical mixing compared to the no wave (shear only) case, which agrees well with the previous studies. One interesting result from the one-dimensional case is that the contribution of the Langmuir turbulence to cooling is larger on the left than on the right for moving tropical cyclones, even though the turbulent Langmuir number tends to be lower (the Langmuir turbulence tends to be stronger) on the right side of the tropical cyclone due to the larger Stokes drift (Figure 3.1 lower panels). This is because the left side of the storm has less shear-driven turbulent mixing, and therefore is more sensitive to the Langmuir turbulence, as explained in the following sentences. In KPP-ST-CS-1d (without Langmuir turbulence), the mixing-layer depth peaks at 120 m on the right hand side, compared to 70 m on the left hand side (not shown). The mixing-layer depth continues to increase on the right for a few hours after the peak wind forcing occurs, because the surface current continues to increase due to resonance between wind forcing and the inertial current response. This maintains a high level of shear-driven turbulence in the water column throughout the storm passage. On the left, the mixing-layer depth peaks much earlier, and shear-driven mixing begins to decrease prior to the onset of maximum wind. This is because the maximum current occurs in front of the storm and the current quickly reduces, which leads to lower levels of shear-driven turbulence in the water column. In KPP-LT-CS-1d (in the presence of Langmuir turbulence), however, mixing is no longer only due to the shear contribution, but continues to increase beyond the time of maximum wind on the left due to the presence of Stokes drift. This allows the mixing-layer depth in KPP-LT-CS-1d to overtake the KPP-ST-CS-1d mixing-layer depth by about 14 m on the left. On the right hand side, KPP-LT-CS-1d only increases the mixing-layer depth by about 7 m from KPP-ST-CS-1d, since the shear-driven mixing alone is sufficiently strong. This means that despite the lower Langmuir number on the right, Langmuir turbulence has a larger impact on the total

cooling on the left. This result is consistent with previous LES results (see RWHGK, Figure 2, bottom-right panel).

The difference between the top row panels and the middle row panels of Figure 3.10 shows that Langmuir turbulence significantly modifies the three-dimensional processes (upwelling and horizontal advection). To understand the three-dimensional effect, in the bottom row we present the difference of the temperature between the top row and the middle row. Since the sum of the middle row (one-dimensional effect) and the bottom row is equal to the top row (total effect), the bottom row panels can be interpreted as the Langmuir turbulence effect on three-dimensional processes. This interpretation is valid if the one-dimensional effect and the three-dimensional effect are not strongly (nonlinearly) coupled. We superimpose the vector current difference and the background temperature field (isotherms) from the KPP-LT-CS simulation in the bottom row panels (both are copied from the top row panels) to show the impact of upwelling and horizontal advection.

With the stationary storm (bottom left panel), the Langmuir turbulence raises the surface temperature by over 0.7° C around the radius of maximum wind. This is due to a large reduction of the upwelling due to the decreased divergent horizontal current. This three-dimensional effect (reduction of upwelling) in the bottom row left panel nearly cancels the one-dimensional effect (increase of vertical mixing) in the middle row left panel and the resulting total impact is relatively small (top row left panel).

For the 2.85 m s^{-1} case, the impact of the Langmuir turbulence in the three-dimensional case is to raise the temperature on the right-side of the cold wake by about 0.2° C, while it cools the temperature on the left-side of the cold wake by about 0.2° C. These temperature differences in the three-dimensional case appear because the reduced current due to Langmuir turbulence reduces horizontal advection of the cold

wake. As in the case of the CS acceleration, the reduction of the current (vector current difference) is directed inward towards the cold wake on the right, warming the surface temperature, and is directed outward away from the cold wake on the left, cooling the surface temperature. For the 5.7 m s^{-1} case the advection effect is reduced.

In conclusion, the Langmuir turbulence impacts are summarized as follows. First, Langmuir turbulence always enhances the vertical mixing, the mixed-layer deepening, and the resulting SST cooling. Although the intensity of Langmuir turbulence is determined by the turbulent Langmuir number, its impact on the sea surface cooling is more complex, depending on the local mixing-layer depth. Second, Langmuir turbulence decreases the current magnitude inside the mixing-layer (because of the enhanced vertical momentum mixing). Hence, it modifies the horizontal heat advection pattern and the resulting cold-wake spatial structure. It also weakens the upwelling due to the horizontal current divergence and reduces the resulting sea surface cooling for stationary and slow moving storms.

3.3.5 Explicit vs Implicit Langmuir Turbulence Model

The previous subsection has focused on the Langmuir turbulence effects compared to the shear-only mixing results. Here, a more practical question is addressed: How well can an implicit Langmuir turbulence scheme commonly used in ocean circulation models perform relative to an explicit scheme? As discussed earlier, the mean Langmuir turbulence impacts can be included in an implicit manner by tuning the critical Richardson number used in the KPP (KPP-iLT). The computational requirements of a surface wave model are not nominal, so the explicit Langmuir turbulence scheme is more costly to employ. Therefore, we aim to determine whether the explicit scheme improves the hurricane upper-ocean simulations significantly enough to justify such an effort.

We now repeat the same numerical experiments using KPP-iLT instead of KPP-LT. (The results are denoted KPP-iLT-CS.) In Figure 3.11 the same set of plots are produced as in Figure 3.10 but using KPP-iLT. The top row panels of Figure 3.11 show the difference of the surface temperature and the current (11.25 m depth) between KPP-iLT-CS and KPP-ST-CS. It is apparent that these results are significantly different from the top row panels of Figure 3.10. For the moving storms KPP-iLT-CS significantly underestimates the sea surface cooling, particularly on the left hand side of the storm. For the stationary storm both warming and cooling are underestimated with KPP-iLT-CS.

To better understand the difference between KPP-iLT-CS and KPP-LT-CS, we now focus on the middle row panels of Figure 3.11, which show the effect of the implicit Langmuir turbulence parameterization on the one-dimensional process (difference between KPP-iLT-CS-1d and KPP-ST-CS-1d). Again the difference of these results from the middle row panels of Figure 3.10 is striking. In the moving tropical cyclone cases the Langmuir turbulence effect with KPP-iLT-CS is remarkably simple; the Langmuir turbulence simply enhances the cooling in proportion to the cooling with the shear-driven turbulence only. The spatial patterns of the enhanced cooling (middle row center and right panels of Figure 3.11) are very similar to those of the cooling due to the shear-driven turbulence only (top row center and right panels of Figure 3.2). This simple response is not surprising because the KPP-iLT scheme just modifies the critical Richardson number compared to the KPP-ST scheme. This is in stark contrast to the rather complex Langmuir turbulence effect with KPP-LT as explained earlier.

The cooling is also reduced in the stationary case with KPP-iLT-CS-1d (middle row left panel of Figure 3.11) compared to KPP-ST-CS-1d (middle row left panel of Figure 3.10). This difference is due to a similar reason as on the left hand side of

the moving storms. The Langmuir turbulence effect is weak with KPP-iLT-CS-1d because the current is weak and vertical mixing is not as strong with the KPP-ST scheme. The Langmuir turbulence effect is stronger with KPP-LT-CS-1d because the shear-driven turbulence is weak and even a small enhancement of vertical mixing due to the Stokes drift makes a large impact.

As in Figure 3.10, we can take the difference between the top row and the middle row to approximate the Langmuir turbulence modification to the horizontal advection and the upwelling (bottom rows of Figure 3.11). The huge difference between KPP-iLT-CS (bottom left panel of Figure 3.11) and KPP-LT-CS (bottom left panel of Figure 3.10) for the stationary storm case can be explained as follows. As discussed earlier, KPP-LT significantly reduces and homogenizes the current magnitude inside the mixing-layer because of the enhanced vertical momentum mixing (Figure 3.7). KPP-iLT hardly modifies the current magnitude. This is clearly seen in Figure 3.12, where the vertical transect of the current magnitude is shown. The results in Figure 3.12 are almost identical to the results of the shear only case (KPP-ST-noCS) in Figure 3.4 rather than the results of KPP-LT-CS in Figure 3.7. Because KPP-iLT-CS does not reduce the current, it does not appreciably reduce the horizontal current divergence and the resulting upwelling, as seen in the bottom row left panel of Figure 3.11. This is in contrast to the significant reduction of the upwelling effect (i.e., warming) in the bottom row left panel of Figure 3.10. Because of the reduced current modification by KPP-iLT-CS, the horizontal advection effects in the moving storm cases are also reduced/modified (bottom row middle and right panels of Figure 3.11) compared to the KPP-LT-CS results (bottom row middle and right panels of Figure 3.10).

Finally, we investigate the difference between the explicit and implicit parameterizations by directly comparing KPP-LT-CS and KPP-iLT-CS in Figure 3.13. The top

row panels indeed confirm that KPP-iLT-CS significantly underestimates the cooling (particularly on the left side of the moving storm) and underestimates the current magnitude reduction (i.e., overestimates the current magnitude). The difference between KPP-LT-CS and KPP-iLT-CS in Figure 3.13 (top panels) is not smaller than the difference between KPP-iLT-CS and KPP-ST-CS in Figure 3.11 (top panels). This suggests that the implicit parameterization (KPP-iLT) is not very skillful in predicting the overall the Langmuir turbulence effect on the upper-ocean response under a tropical storm, compared to the explicit parameterization (KPP-LT).

3.4 Discussion

As discussed earlier, the most significant feedback mechanism from the upper-ocean to the tropical cyclone is through changes in the SST and the resulting surface latent heat flux. We have demonstrated that the SST cooling is significantly modified by the surface-wave induced Langmuir turbulence. Since this study specifies wind forcing fields and does not allow the feedback from the upper-ocean to the tropical storm, it is difficult to estimate the impact of the Langmuir turbulence on the heat flux, since the near surface air temperature and humidity are modified in response to the changes in the SST in a real storm. Nevertheless, estimates can be made of the impact on the heat flux if the near surface air temperature and humidity are specified and assumed unaffected by the sea surface cooling. For example, using an air temperature of 23°C and a humidity of 95% (comparable to values used in previous literature), the reduction of the heat flux can be greater than 50 W m^{-2} (roughly 10% of the total heat flux) in some locations comparing the KPP-LT and KPP-iLT. We certainly expect that this large modification of the heat flux would have a significant impact on the tropical cyclone evolution.

We have so far investigated the Langmuir turbulence effect using a storm with one size and one intensity, and with one initial ocean temperature profile. We have

also examined the sensitivity of the results to different storm sizes, storm intensities and different initial ocean profiles. In general, the Langmuir turbulence effect is more significant if the storm is larger and more intense due to the increased waves (Stokes drift). The Langmuir turbulence effect is also stronger if the initial mixed-layer is shallower and if the temperature gradient below the mixed-layer is larger. For example, the Langmuir turbulence is more significant with the typical temperature profile in the Gulf of Mexico compared to the typical profile in the Caribbean Sea with a larger mixed-layer depth.

3.5 Conclusions

We have explored the modification of the ocean response to tropical cyclones due to Langmuir turbulence and the Coriolis-Stokes acceleration. The Coriolis-Stokes acceleration increases the upwelling in a slowly moving tropical storm, contributing to the total cooling in the storm's cold wake. It also modifies the horizontal advection of the cold-wake by the storm-induced current. However, the impact of the Coriolis-Stokes acceleration is much weaker than the impact of Langmuir turbulence.

Next, we investigated the impact of the explicit Langmuir turbulence parameterization compared to the turbulent mixing parameterization that accounts only for shear induced turbulence. Langmuir turbulence always enhances the vertical mixing, the mixed-layer deepening, and the sea surface cooling. It also reduces and homogenizes currents inside the mixing-layer because of the enhanced vertical momentum mixing. Although the intensity of Langmuir turbulence is determined by the turbulent Langmuir number, its impact on the sea surface cooling is more complex, depending on the local mixing-layer depth. The reduction of the horizontal currents significantly modifies the horizontal advection of heat and reduces the cooling due to upwelling for stationary and slowly moving storms.

We have also compared the impacts of the explicit (sea state dependent) and im-

PLICIT (independent of sea states) Langmuir turbulence parameterizations. Although the implicit parameterization also introduces some enhanced SST cooling, it significantly underestimates the cooling, particularly on the left of a propagating storm, and overestimates the currents. Overall, the implicit scheme does not adequately predict the Langmuir turbulence effects on the three-dimensional upper-ocean response to a tropical storm.

The next step of this research is to include the explicit Langmuir turbulence parameterization (KPP-LT) in a fully coupled hurricane-wave-ocean system and to investigate the feedback of the modified SST to the storm evolution. This study suggests that explicitly resolving the sea-state dependent Langmuir turbulence will lead to increased accuracy in predicting the air-sea fluxes and thus the intensity and track forecasts of tropical cyclones.

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Figure 3.1. Wind stress (top row), surface 4.5 m layer averaged Stokes drift (middle row), and turbulent Langmuir number (bottom row) for a stationary tropical cyclone at 48 hours (left), a tropical cyclone translating at 2.85 m s^{-1} at 72 hours (center), and a tropical cyclone translating at 5.7 m s^{-1} at 72 hours (right), for model runs with explicit Langmuir turbulence and CS (KPP-LT-CS). In the top and middle rows colors indicate magnitude and arrows indicate direction and relative magnitude. In the middle row panels, the mixing-layer depth contours are also shown in 20 m increments starting at 10 m (lightest gray line) and ending at 90 m (thin black line), where the line color is continuously darker for the deeper mixing layers. The white circles represent the storm center location at 6 hour increments and the thick black line is the radius of maximum wind.

Figure 3.2. Mean surface temperature (top row) and current at 11.25 m depth (bottom row) with KPP-ST-noCS. The black circles represent the storm center location at 6 hour increments and the thick black line is the radius of maximum wind. In the bottom row colors indicate magnitude and arrows indicate direction and magnitude. The same 3 storms are presented as in Figure 3.1.

Figure 3.3. Vertical transect of temperature in the north-south direction with KPP-ST-noCS. Top row panels show transects centered at the storm center and bottom row panels show transects centered at the storm center 18 hours prior. The same 3 storms are presented as in Figure 3.1. The thick black line shows the mixing-layer depth given by KPP.

Figure 3.4. Same as Figure 3.3, but for current magnitude.

Figure 3.5. Same as Figure 3.2, but for model runs with Coriolis-Stokes force and explicit Langmuir turbulence (KPP-LT-CS).

Figure 3.6. Same as Figure 3.3, but with KPP-LT-CS.

Figure 3.7. Same as Figure 3.6, but for current magnitude.

Figure 3.8. Same as Figure 3.6, but for Stokes drift magnitude.

Figure 3.9. Difference between KPP-ST-CS and KPP-ST-noCS for surface temperature (colormap) and currents at 11.25 m depth (vectors), with the gray contours representing isotherms from KPP-ST-CS in 2° C increments for the stationary case and 1° C increments for the moving cases. The black circles represent the storm center location at 6 hour increments and the thick black line represents the radius of maximum wind. The same 3 storms are presented as in Figure 3.1.

Figure 3.10. Top row panels show difference between KPP-LT-CS and KPP-ST-CS for surface temperature (colormap) and currents at 11.25 m depth (vectors), with the gray contours representing isotherms from KPP-LT-CS in 2° C increments for the stationary case and 1° C increments for the moving cases. Middle row panels show surface temperature difference between KPP-LT-CS-1d and KPP-ST-CS-1d. Bottom row panels show difference between top rows and middle rows for surface temperature, with the same arrows and gray contours as in the top row panels. The black circles represent the storm center location at 6 hour increments and the thick black line represents the radius of maximum wind. The same 3 storms are presented as in Figure 3.1.

Figure 3.11. Same as Figure 3.10, but with KPP-iLT-CS instead of KPP-LT-CS and with KPP-iLT-CS-1d instead of KPP-LT-CS-1d.

Figure 3.12. Same as Figure 3.4, but with KPP-iLT-CS instead of KPP-LT-CS.

Figure 3.13. Difference between KPP-LT-CS and KPP-iLT-CS for surface temperature (colormap) and currents at 11.25 m depth (vectors), with the gray contours representing isotherms from KPP-LT-CS in 2° C increments for the stationary case and 1° C increments for the moving cases. The black circles represent the storm center location at 6 hour increments and the thick black line represents the radius of maximum wind. The same 3 storms are presented as in Figure 3.1.

APPENDIX A

Wind Stress Calculations with Empirical Wave Spectra and Comparison with Results of *Mueller and Veron (2009)*

It is of interest to apply the well known empirical wave spectra of *Elfouhaily et al. (1997)* in the two RHG and DCCM approaches to estimate the wave age dependent drag coefficient. In particular, this allows us to compare another stress calculation approach of *Mueller and Veron (2009)*, who included an explicit breaking-wave stress impact and applied the method to the *Elfouhaily et al. (1997)* spectra. In *Mueller and Veron (2009)* the stress is given as the sum of the viscous component, the wave form component, and a breaking-wave separation stress component.

$$\tau(z = 0) = \tau_\nu + \tau_w + \tau_s \quad (\text{A.1})$$

The growth rate of non-breaking waves is calculated similarly to RHG. To calculate an explicit breaking-wave stress term, they must infer a breaking-wave distribution. They parameterize their breaking-wave distribution in terms of the saturation spectrum and the growth rate. Ultimately this leads to a calculation of the breaking-wave stress that is similar to the non-breaking stress calculation. There is also no feedback mechanism considered for the breaking or non-breaking wave-form drag on the mean wind profile.

The results of RHG, DCCM, and the *Mueller and Veron (2009)* methods are shown in Figure A.1. The three methods yield a similar trend in the drag coefficient with wind speed. One notable difference between the results with the WW3 wave spectrum and those with the empirical wave spectrum is that the latter tends to saturate (or increase more slowly) at higher wind speeds. This is mainly because the empirical saturation spectrum of *Elfouhaily et al. (1997)* decreases with wind speed

beyond the spectral peak and outside the gravity-capillary range. This is consistent with our earlier statement that saturation or reduction of the drag coefficient at high wind speeds can be realized if the tail level is systematically reduced as wind speed increases.

The method of DCCM yields the lowest drag coefficients at higher wind speeds. This is attributed to the DCCM method being less sensitive to the shortest (highest frequency) waves where the *Elfouhaily et al.* (1997) spectrum can become quite high (see Figure 1.1). The methods of RHG and *Mueller and Veron* (2009) yield quite similar results at all wind speeds. These results with the empirical tail are generally consistent with our earlier results using WW3 and the parameterized tail, at least in simple fetch-dependent conditions.

Bibliography

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Figure A.1. (A1) Drag coefficients ($\times 1000$) calculated using the 3 different methods with the empirical spectrum of *Elfouhaily et al. (1997)*. The red symbols are values calculated using the RHG method, the green symbols are values calculated using the DCCM method, and the blue symbols are values calculated using the *Mueller and Veron (2009)* method. The triangles are calculated at a fetch of 10 km, the squares at a fetch of 100 km, and the circles are from fully developed (infinite fetch) spectrum.

APPENDIX B

Effect of Counter-swell on the Drag Coefficient

Although we have assumed that the counter-swell effect is large in RHG, being consistent with DCCM, the effect of the counter-swell is not well understood. In fact, a different growth rate parameterization with a small contribution from the counter swell has been suggested (Dr. Stefan Zieger and Dr. Alexander Babanin, personal communication). Here, we examine the impact of the counter-swell growth rate on the drag coefficient using the RHG approach, by setting the counter swell forcing $c_\beta = -10$ instead of -25 . The tail level is fixed at the middle level ($B = 6 \times 10^{-3}$).

The results are shown in Figures B.1-B.4. Although the 5 m/s translation results (left panels) show no significant changes (upper and middle images), the 10 m/s translation results (right panels) show a more noticeable difference on the left of the storm track. The RHG results with a smaller counter-swell coefficient do not contain the large drag coefficient on the inner left-rear of the tropical cyclone seen in the original RHG (middle) and the DCCM (bottom). Figures B.1 and B.2 also shows that with a large (10 m/s) translation speed the counter swell may significantly increase the sea state dependence of the drag coefficient. However, Figures B.3 and B.4 show that the misalignment angle between the wind speed vector and the wind stress vector is hardly affected by the counter swell in RHG. This is expected since cross-wind swell has little impact on the drag coefficient in RHG as discussed earlier.

Figure B.1. Drag coefficients ($\times 1000$) for a 5 m/s (a,c,e) and 10 m/s (b,d,f) translating tropical cyclone with saturation level $B=0.006$. The RHG drag coefficient with the growth rate of the counter swell equal to 40 % of the growth rate for wind sea (a,b), the RHG drag coefficient with the growth rate of the counter swell equal to that of the wind sea (c,d), and DCCM drag coefficient (e,f). The thick black line represents tropical cyclone's track through the domain.

Figure B.2. The same as Figure B.1, but for drag coefficient ($\times 1000$) vs wind speed.

Figure B.3. The same as Figure B.1, but for misalignment angle (10-meter wind direction minus surface stress direction).

Figure B.4. The same as Figure B.1, but for misalignment angle (10-meter wind direction minus surface stress direction) vs. wind speed.