Understanding the Influence of Nonlinear Seas on Wind Generated Waves

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UNDERSTANDING THE INFLUENCE OF NONLINEAR SEAS
ON WIND GENERATED WAVES

By

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B.S. University of Maine, 2018

A THESIS
Submitted in Partial Fulfillment of the
Requirements for the Degree of
Master of Science
(in Civil and Environmental Engineering)
The Graduate School
The University of Maine
August 2020

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An understanding of the influence of wind on the surface on irregular waves is important for improving ocean forecasting models. While many studies have investigated the phenomenon of wind wave suppression on the surface of mechanically generated waves in the laboratory, few studies have investigated the occurrence of this phenomenon for irregular waves. Chen and Belcher (2000) developed the first model to predict the suppression of wind waves as a function of the steepness of the long wave on which they travel. The Chen and Belcher (2000) model however, was only validated using monochromatic waves, not irregular waves, which are more representative of real ocean sea states. Additionally, few studies have investigated turbulence under irregular waves in the presence of wind in a controlled environment.

This thesis aims to satisfy two research objectives. The first is to determine the applicability of the Chen and Belcher (2000) wind wave suppression model to irregular sea states. The second objective is to provide a procedure for selecting the appropriate method for indirect measurement of turbulence beneath waves in a laboratory. To meet these objectives, a comprehensive data set consisting of wind velocity, surface elevation, and water velocity data were collected in the Alfond W^2 Ocean Engineering Lab at the
University of Maine. The data set consisted of a variety of irregular and monochromatic wave environments and wind speeds.

Using multiple data analysis techniques, this study reveals that for the Chen and Belcher (2000) model to be directly applicable to irregular seas, a modification must be made to the long wave-induced stress term. This modification accounts for the wave energy associated with each frequency in wave spectrum for irregular waves, whereas the original model only accounts for a single wave frequency. The modified model is able to accurately predict the trend in the suppression of wind waves on the surface of irregular, long waves as a function of the long wave steepness. Additionally, in this work a case study is presented that reveals several limitations associated with the existing methods for indirect measurement of turbulence in a laboratory.

The results of this work expand the implications of the Chen and Belcher (2000) model to be more applicable to ocean waves. This can aid in better prediction of model parameters, such as the drag coefficient and the sea surface roughness length, which are controlled by the high frequency waves on the ocean surface. This work also provides a guide for planning an experiment to measure TKE dissipation, $\varepsilon$, under waves in the presence of wind in a controlled, laboratory setting, which will aid in the planning of future experiments.
ACKNOWLEDGEMENTS

This thesis was supported by the U.S. Army Engineer Research and Development Center (ERDC) under CEED-17-0018 “Engineered Energy Efficient and Low Logistic Burden Materials and Processes” executed under Contract Number W15QKN-17-9-8888.

I would like to thank my research advisor, Dr. Lauren Ross, for being such an incredible mentor to me. I am truly grateful to have found such an incredible advisor and friend to guide me through this journey. I would like to extend a huge thank you to the rest of my research committee, Drs. Kimberly Huguenard and Anthony Viselli, and Mary Bryant for all your help and support during this work. A special thank you to two of my fellow graduate students, Rick Perry and Preston Spicer, for all your help and for your friendship during these past two years. I could not have done it without you guys. Thank you to the rest of the Coastal Engineering students, including Sohaib, Rachel, Gwyneth, Zhilong, and Longhaun for all your help and friendship. Thank you to my parents, William and Roberta, and my sister, Blake, for your endless love and support during my academic career. Finally, thank you to my best friend, Tasha, for being such a supportive friend and for always being there for me even when you are 3,000 miles away.
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CHAPTER 1

INTRODUCTION

Of all the dynamic processes and phenomena that encompass the world’s oceans, one of the most fascinating is the existence of ocean waves. Most ocean waves are generated by the wind blowing over the water surface, but the mechanisms by which they grow and evolve are intricate, and still not entirely understood. The processes involved in the growth of wind waves, and the interactions between wind waves and the atmospheric boundary layer, have been topics of research for many decades. An understanding of these processes, and the development of parameterizations to estimate them, aids in the improvement of ocean wave forecasting models. The history of wave forecasting models dates to World War II, when the first models were used for the preparation of D-day military tactics (Tolman, 2008). Since then, wave models have developed significantly and have become widely applicable. Today, these models are important for predicting the wave climate based on wind conditions for many purposes such as military operations, coastal design, weather forecasting, and shipping. Examples of these wave forecasting models include WAM, WAVEWATCH, and SWAN, all of which are now third generation wave-models. The WAM model was created in an attempt to refine wave forecasting techniques and required the efforts of researchers from all over the world during the decade succeeding 1984. Since then, models have been further refined, like WAVEWATCH III for operational forecasting, and SWAN for near-shore wave modeling (Tolman, 2008).

1.1 Wind Waves

1.1.1 Wind Wave Terminology

Wind waves are sub-categorized based on the mechanism by which they form under the action of wind blowing over the water surface. Before introducing the three types of wind waves, it will be helpful to first introduce the terminology used to describe an ocean wave. Unlike the symmetric, sinusoidal waves that are often used to represent an idealized water wave, waves in the ocean are typically irregular,
meaning their wave profiles are not symmetric, but are random, meaning the waves in a wave train (a group of waves traveling in the same direction at similar speed) are not identical. Regardless, many of the same properties are used to describe idealized water waves and wind waves. Figure 1.1 shows an example of an ocean wave generated by wind. The wave crest and trough refer to the highest and lowest point on the wave respectively. The wavelength, $L$, is the distance between two consecutive wave crests. Wave height, $H$, refers to the distance between the wave crest and its adjacent wave trough, and the wave amplitude, $a$, is the distance between the wave crest and the mean water level, or $\frac{1}{2}H$ for an idealized water wave. Not shown in Figure 1 is the wave period, $T$, the time taken for a second, consecutive wave crest, to propagate past a fixed point, since the time of the passing of the first wave crest. The phase speed, $C$, of a wave is the speed of propagation and is quantified as, $C = \frac{L}{T}$.

![Figure 1.1. Properties of an ocean wave.](image)

Due to the forward propagation of a wave (from right to left in Figure 1) fluid particles underneath a water wave follow circular velocity trajectories in the direction of the wave propagation. The trajectory of the orbital velocities under the crest of a water wave are shown in Figure 1. The rotation of the orbital
velocities under the wave is in the same direction of the wave propagation. The shape of the trajectories will depend on the classification of the wave based on its relative depth, \( d/L \), which is the ratio of the water depth, \( d \), to the wavelength, \( L \). For deep water waves, \( d/L > 0.5 \), the shape of the orbital velocities will be symmetric circles, and their size will decrease exponentially with depth. As waves transition into intermediate water waves, \( d/L < 0.5 \), and shallow water waves, \( d/L < 1/20 \), and the shape of the trajectories of the fluid particles become elliptical and decrease in size in the longitudinal direction (z-direction in Figure 1.2) due to interactions with the bottom of the ocean (Dean and Dalrymple, 2002). The wave orbital velocities under a deep water and shallow water wave are shown in Figure 1.2.

![Diagram of wave orbital velocities](image)

**Figure 1.2.** Wave orbital velocities beneath the crest of a deep and shallow water wave (not to scale).

### 1.1.2 Classification of Wind Waves

There are several different types of waves in the ocean, which are classified by their periodicity and by the mechanism by which they are generated. Figure 1.3 shows the wave period and frequency associated with each category of ocean waves. The longest waves are trans-tidal waves, which are caused by variations in the atmosphere due to storms and small changes in the Earth’s crust (Holthuijsen, 2007). Tides, which are slightly shorter with periods between 3-24 h, are caused by the gravitational forces of the moon and the sun. Surges and seiches caused by atmospheric variations, can fall within this range of
periodicity as well. Tsunamis, which are a result of the disturbance caused by an earthquake, have slightly shorter periods ranging between approximately 10 min to 2 h (Toffoli and Bitner-Gregersen, 2017). Infra-gravity waves are generated by nonlinear interactions amongst waves of different frequencies, which include both wind sea and swell, and are therefore not generated directly by wind (Toffoli and Bitner-Gregersen, 2017).

Figure 1.3. Period and frequencies of waves in the ocean (adapted from Waves in Oceanic and Coastal Waters Holthuijsen, p. 4, by L. H. Holthuijsen, 2007, New York, NY: Cambridge University Press. Copyright 2007 by Cambridge University Press)

All the aforementioned wave types are generated by mechanisms other than wind forcing. Wind-generated waves are further separated into four categories based on the mechanisms by which they form and by their wave period. Capillary waves are small ocean waves that form as wind blows over the water surface and propagate due to surface tension effects. These waves typically have a maximum wavelength of approximately 1.6 - 2 cm, and typically have frequencies greater than 10 Hz (periods less than 1 s) (Figure 1.3; Holthuijsen, 2007). Capillary waves mark the initial growth of waves by the wind, but
surface tension acts as the waves restoring force, rather than the wind itself (Sorenson, 2006). For wind waves that are generated by, and continue to propagate because of the wind, two categories are commonly used by researchers interested in the study of the transfer of momentum between the wind and the waves in the atmospheric boundary layer. In order to define these categories, it is necessary to first introduce the concept of wave age.

Wave age is a term that is used to identify wind waves based on the direction of momentum transfer between the wind and the waves. The term was first developed in order to describe the speed of propagation of the waves compared to the wind speed. It originated around the time of the introduction of Jeffrey’s sheltering hypothesis, which suggests that air flow separation over steep water waves occurs at the wave crest and produces low pressure and velocity regions downwind of the crest. The pressure differential resulting due to flow separation allows for the transfer of momentum from the wind to the waves (Jones and Toba, 2001). Based on this hypothesis, waves travelling faster than the wind, $C > U$, where $U$ is the wind velocity, would not be forced by the wind, and therefore would no longer grow under the action of wind. These waves therefore are referred to as swell, or mature waves. Waves that continue to grow due to the input of momentum from wind are referred to as wind waves (Jones and Toba, 2001). From here on, waves with $C < U$ or $C/U < 1$ will be termed wind waves and swell will refer to waves with $C > U$ or $C/U > 1$. The periodicity of wind waves and swell typically falls between approximately 1-30s (Figure 1.3; Holthuijsen, 2007). The mechanism by which wind generates water waves will be discussed next.

1.1.3 Wind Influence over the Ocean

Waves form on a still water surface due to the forces induced, and energy input, from a wind source blowing over the water. In order to understand the mechanisms by which wind generates waves, it is necessary to first discuss these forces by considering a wind blowing over a still water surface (Figure 1.4). Wind speed is measured at a reference location, $U_r$, some height, $z$, above the water surface. This
reference is typically set to, or extrapolated to \( z = 10 \text{ m} \). The wind induces a shear force, \( \tau \), on the water surface in the direction of the wind,

\[
\tau = C_d \rho |U_r| U_r
\]

where \( \rho \) is the density of the air above the water, and \( C_d \) is the drag coefficient. The drag coefficient can be estimated as

\[
C_d = 1.2 \times 10^{-3} \quad 4 \leq U_{10} < 11 \frac{m}{s}
\]

\[
C_d = (0.49 + 0.065) U_{10} \times 10^{-3} \quad 11 \leq U_{10} \leq 25 \frac{m}{s}
\]

depending on the wind speed at a reference height of \( z = 10 \text{ m} \) (Large and Pond, 1981). Alternatively, the surface shear stress can also be estimated using the air friction velocity, \( u_* \), as \( \tau = \rho u_*^2 \) (Jones and Toba, 2001), as depicted in Figure 1.4. Air friction velocity at the water surface can be measured directly, or estimated by extrapolating the wind profile to the water surface using the universal log law expression,

\[
\frac{U_r}{u_*} = \frac{1}{\kappa} \ln \left( \frac{z}{z_0} \right)
\]

where \( \kappa \) is the von Karman constant, \( \kappa \sim 0.4 \) and \( z_0 \) is the roughness length, which is an estimate of the vertical elevation at which the wind speed is approximately zero.

Figure 1.4. Schematic representation of the wind profile above the water surface and the shear induced below the water.
The parameter wave age, $c_p/U$, can also be formulated in terms of the air friction velocity, $u_*$, instead of the wind speed, $U$, as, $c_p/u_*$. 

1.1.4 Wave Statistics and Wave Spectra

Many individual waves, which vary in height, period and wavelength, make up real ocean sea states. For the analysis of ocean sea states, two common techniques are used by researchers in order to represent the height and period of the sea. The first technique is a zero-crossing analysis. A zero-crossing analysis uses the surface elevation record of the waves and considers each wave individually in order to calculate a mean wave height, $H$, or mean wave period, $T$ (Figure 4a). Additionally, the significant wave height, $H_s$, can be determined from the surface elevation record. The significant wave height is the mean height of the tallest one-third waves in a record. It is also considered to be the wave height that would be estimated by an observer overlooking the sea state in the ocean (Sorenson, 2006). An alternative, and more commonly used, technique is a wave spectrum. A wave spectrum plots the Fast Fourier Transform (FFT) of a surface elevation record varying in time in order to show the wave energy density of each individual wave frequency in a sea state versus the range of wave frequencies (Figure 1.5b).

Figure 1.5. a). Example of surface elevation record for zero-crossing analyses. Dots mark the zero-crossings of the mean water level to separate individual waves. b). Example wave energy density spectrum.
The frequency of the peak energy density in a wave spectrum is the peak frequency of the sea state, \( f_p \) (Figure 1.5b). The inverse of \( f_p \) gives the peak wave period, \( T_p \), of the sea state. Additionally, the significant wave height can be determined from the wave spectrum, in which case it is denoted by the variable \( H_{m0} \), by the relation

\[
H_{m0} = 4\sqrt{m_0^2} = 4\sqrt{\sigma^2}
\]  

(1.5)

where \( m_0^2 \) is the zeroth-order moment of the wave spectrum which is equal to the variance of the surface elevation, \( \sigma^2 \). The values of \( H_s \) and \( H_{m0} \) typically only vary by approximately 5-10\% (Holthuijsen, 2007).

Theoretical, one-dimensional spectra models are often used to predict wave spectra based on sea state conditions. These models are useful for both design purposes and wave forecasting. One of the most commonly used model spectra for simulating irregular waves in the laboratory, while representing deep water, fetch-limited (i.e. the growth of the waves is limited by the distance over which the wind blows) conditions, is based on the Joint North Sea Wave Project (JONSWAP) (Hasselmann, et al. 1973). The model has the form

\[
S(f) = \frac{ag^2}{(2\pi)^4f^3}e^{-1.25(f_p/f)^4}g^a
\]  

(1.6)

where \( g \) is gravitational acceleration, \( f \) is frequency, and \( \gamma \) is the peak enhancement factor, which is typically 3.3 for design purposes. The following expressions are used for \( a, \sigma, \alpha \), and \( f_p \),

\[
a = e^{-[(f-f_p)^2/2\sigma^2f_p^2]}
\]  

(1.7)

\[
\sigma = 0.07 \text{ for } f < f_p
\]  

(1.8)

\[
\sigma = 0.09 \text{ for } f \geq f_p
\]  

(1.9)

\[
\alpha = 0.076\left(\frac{gF}{U^2}\right)^{-0.22}
\]  

(1.10)

\[
f_p = \frac{3.5g}{U}\left(\frac{gF}{U^2}\right)^{-0.33}
\]  

(1.11)

where \( \alpha \) and \( f_p \) are based on the fetch, \( F \), and the wind speed, \( U \) (Sorenson, 2006). Models such as the JONSWAP spectra are useful for predicting a sea state based on conditions such as peak wave frequency
and fetch. In order to develop and apply such models, it is important to first understand the mechanisms that govern wave growth in the ocean.

1.1.5 Fetch-Limited Wave Growth

Wind waves grow on the ocean surface due to energy input from the wind. The actual physics behind the formation of waves on a still water surface however is more complex. Theories by Phillips (1957, 1960) and Miles (1957) suggest that resonance and shear govern the initial growth of waves by the wind, but both mechanisms induce wave growth through pressure forces. According to Phillips (1957 and 1960), the initial formation of wind waves is a result of pressure fluctuations on the water surface induced by turbulent eddies in the atmospheric boundary layer. These pressure fluctuations result in the modulation of surface waves due to resonant interactions with the free waves on the water surface which are moving forward at the same speed as the pressure fluctuations (Sorenson, 2006). Based on Miles theory (1957) these waves continue to grow due to additional pressure forces on the water surface, which are caused by wind shear induced air circulation around the wave crest. These forces result in a transfer of momentum from the wind to the waves that amplifies the steeper waves in a sea state (Sorenson, 2006).

Physically, once waves have formed on the water surface, they will continue to grow in both wave height and wave period due to the input of energy from the wind. The growth of wind waves in the ocean is typically limited by one of three parameters: wind speed, duration of the wind source, or the distance over which the wind blows, termed fetch (Sorenson, 2006). Parameterizations used to predict wave growth are typically fetch or duration limited. These parameterizations are then included in the formulation of source terms used in wave forecasting models. Wave forecasting models, such as WAVEWATCH III (Tolman, 2008), utilize an energy balance equation of the form

\[ S = S_{\text{in}} + S_{\text{nonlin}} + S_{\text{dis}} \]  

(1.12)

where \( S_{\text{in}} \) is the input of wind energy, \( S_{\text{nonlin}} \) is the wave energy associated with nonlinear interactions, and \( S_{\text{dis}} \) is the loss of wave energy due to dissipation by mechanisms such as wave breaking and turbulence (Janssen, 2008).
1.1.6 Objective of the Wind Wave Study

The wind input term, $S_{in}$, has been the focus of this review so far. There have been many works over the past decades dedicated to the improvement of this source term, many of which have focused on refining models for fetch-limited wind wave growth. Even before the development of operational wave forecasting models, researchers were focused on understanding the growth and evolution of wind sea, using Phillips (1957 and 1960) and Miles (1957) explanations for wave generation as benchmarks. The goal of this work is to help further refine parameterizations involved in the formulation of $S_{in}$. To reach this goal, the objective is to investigate the dependence of wind wave growth on the properties of long waves. In doing so, the applicability of existing small fetch-limited growth models to wind sea generated over irregular waves will be evaluated.

Of equal importance to the work on improving formulations for $S_{in}$, are those aimed to refine $S_{dis}$, the dissipation source term, which is primarily expected to contribute to Eq. 12 by extracting energy through wave breaking. When surface waves break, turbulence is generated that dissipates energy in the upper layer of the ocean and contributes to mixing (Dai, et al. 2010). The dissipation term has been regarded as the least understood source term, due to the difficulty of obtaining reliable measurements of wave breaking in the ocean, and therefore ocean wave forecasting models such as WAM and WAVEWATCH rely on theoretical parameterizations to compute $S_{dis}$ (Alves, 2002). Wave breaking induced turbulence however is not the only energy sink that is believed to contribute to $S_{dis}$. Recently, more attention has been given to wave-turbulence interactions that have been found to extract energy from the wave field (Qiao, et al. 2016; Huang and Qiao, 2010; Alves, 2002). The wind plays a key role in both aforementioned turbulence generating processes. Wind has been found to enhance surface wave breaking, thus leading to increased turbulence at the surface (Thomson, et al 2016) and to also enhance Langmuir circulation that advect turbulence downward away from the water surface (Dai, et al. 2010). Before further discussion of the existing literature and progress in the field of ocean turbulence, it is important to first provide background theory on the topic of turbulence.
1.2. Turbulence

Like wave forecasting models, researchers have been working to improve ocean circulation models for decades. Turbulence, which is characterized by random fluctuations in flow velocity or pressure, is an important process for ocean circulation. Turbulence is a driving force of mixing that influences the exchange of heat, momentum and gases between the atmospheric boundary layer and the upper ocean (Scannell, et al. 2017). Crucial to wave circulation models is an understanding of the wind energy input to the water surface, which acts to both generate waves and enhance surface wave breaking, both of which induce turbulence. The presence of waves on the water surface, which are modified by wind, have been studied for their contribution to ocean circulation by generating mixing in the upper layer of the ocean (Huang et al. 2018; Lai et al. 2018; Alberello et al. 2017). Turbulence in the ocean exists on many scales and is typically investigated using spectral methods.

1.2.1 Measuring Turbulence

The velocity of the flow in a one-dimensional turbulent regime can be broken into two components as follows,

\[ u(x,t) = \bar{u}(x,t) + u'(x,t) \]  \hspace{1cm} (1.13)

where \( \bar{u} \) is the mean flow, \( u' \) are fluctuating velocities in the flow, \( x \) is the distance in the horizontal direction, and \( t \) is time. A turbulent flow is made up of groups of vorticity, called eddies, that exist on a wide range of length scales. The largest scale eddies within a turbulent flow contain energy that is transferred to smaller, dissipative scales. Between the largest and the smallest scales exist a range over which energy is conserved, i.e. no energy is dissipated, called the inertial subrange. Within this range, inertial forces cause large eddies to separate into smaller ones. The overall transfer of energy from the largest to the smallest dissipative scales in a turbulent flow is called the energy cascade. At the smallest scales, the energy within the eddies is dissipated by viscous forces as heat (Davidson, 2015).
The degree of turbulence in the flow is represented by the amount of turbulent kinetic energy (TKE), which provides a measure of the kinetic energy per unit mass of the eddies, or the velocity fluctuations, and can be calculated as

\[ TKE = \frac{1}{2} \left( \overline{(u')^2} + \overline{(v')^2} + \overline{(w')^2} \right) \]  

(1.14)

where \( u, v, \) and \( w \) are the along-channel, across-channel, and vertical components of the velocity respectively. The overbar represents a time average, and the prime denotes the fluctuations from the mean.

Turbulence is investigated in the ocean by measuring the turbulent kinetic energy (TKE) within a flow. There are many mechanisms in the ocean that can generate and dissipate TKE. Following conservation of energy principles however, the following budget is often used to describe the sources and sinks of turbulence in the ocean:

\[ \frac{D}{Dt} TKE = P + T + B - \varepsilon \]  

(1.15)

where \( \frac{D TKE}{Dt} \) is the total rate of change of turbulent kinetic energy with time, \( P \) is the production of turbulence, \( T \) is the vertical transport of turbulence within the flow, \( B \) is the contribution due to buoyancy fluxes, and \( \varepsilon \) is the TKE dissipation rate. Turbulence is produced when large eddies form and extract energy from the mean flow and can be dissipated by mechanisms such as wave breaking and nonlinear interactions. Under the assumption that a turbulent flow is statistically homogenous, meaning the statistical properties within the flow do not vary in space, and assuming buoyancy fluxes are negligible, as is the case in an unstratified environment such as the ocean or a laboratory wave basin, the energy balance equation simplifies to

\[ P = \varepsilon \]  

(1.16)

which states that the production of TKE in the flow is equal to the dissipation of TKE by viscous forces (Davidson, 2015).

Measuring the production of turbulence in a flow is difficult because it requires the collection of velocity measurements that vary spatially (Parra et al., 2014), so turbulence is typically measured in terms
of the TKE dissipation rate, $\varepsilon$. The dissipation rate can be measured directly using instruments such as Particle Image Velocimetry (PIV) (Sheng et al., 2000) or microstructure shear probes (Lueck, 2013), or indirectly using acoustic instruments such as an Acoustic Doppler Velocimeter (ADV) (Parra et al., 2014) and an Acoustic Doppler Current Profiler (ADCP) (Scannell et al., 2017). Most indirect methods require the identification of the inertial subrange within the flow, which can be performed by utilizing an energy density spectrum and Kolmogorov’s 5/3 law, which will be discussed next.

1.2.2 Turbulent Energy Spectrum

The inertial subrange is a range of scales over which energy cascades from larger to smaller eddies without loss of energy by dissipation. According to Kolmogorov (1941), for that range, $L^{-1} \ll k \ll \eta^{-1}$, where $L$ and $\eta$ are the length scales of the largest and smallest eddies respectively, the statics of the scales are of a universal form, which is unique, and determined only by the wavenumber, $k$, and the turbulent kinetic energy (TKE) dissipation rate, $\varepsilon$. The inertial subrange only exists for turbulent flows. In order to determine if a flow is turbulent, the ratio of inertial to viscous forces in a fluid, called the Reynolds number, $Re$, is used. For example, when $Re > 2000$ the flow is considered turbulent in an open channel, whereas $Re < 2000$ is considered transitional or laminar (Davidson, 2015). The Reynolds number can be calculated as, $Re = \bar{u}L/\nu$ where $\bar{u}$ is the mean flow, $L$ is the length scale of the largest eddy, and $\nu$ is the kinematic viscosity of the fluid.

Within the range of scales that constitute the inertial subrange, the energy spectrum takes the form

$$E(k) = C\varepsilon^{2/3}k^{-5/3}$$

(1.17)

where $C$ is a universal constant, typically assumed to have a value of $\sim 1.5$ (Davidson, 2016). Figure 1.6 shows a graphical representation of the turbulent wavenumber spectrum. Shown in Figure 1.6 are the length scales of the largest and smallest eddies relative to the energy spectrum and the expected slope of the inertial subrange corresponding to Eq. 1.17.
1.2.3 Modeling Wind and Wave Induced Turbulence

When wind blows over the water surface, a shear force is induced on the water that acts in the direction of the wind. This shear force is quantified as $\tau = \rho u^2$, at the surface, and is expected to decrease exponentially following the “law of the wall” (Figure 3). In the boundary layer below the water surface over which this shear force is acting, $\varepsilon$ is estimated as

$$\varepsilon = \frac{u^3}{\kappa z} \quad (1.18)$$

where $\kappa$ is the von Karman constant and $z$ is the depth below the water surface. In the last few decades, parameterizations have begun to acknowledge the influence of surface waves on turbulence near the surface and identified a sublayer over which the estimates of $\varepsilon$ do not conform well to the law of the wall (Craig and Banner, 1994). Since then, advancements have been made to this theoretical estimate of $\varepsilon$ by accounting for the influence of surface waves in turbulence models. Surface waves can induce turbulence...
by dissipating energy through wave breaking, and by enhancing mixing in the turbulent boundary layer (Lai et al., 2018). Only recently have the properties of the surface waves been investigated to determine their influence on $\varepsilon$. Savelyev et al. (2012) used monochromatic waves generated in a laboratory to investigate the dependence of turbulence generated by wave-turbulence interaction on the wave properties. They noted an observed decrease in the length scale along the direction of the flow of near surface eddies with increasing wave steepness. More recently, Lai et al. (2018) conducted a laboratory study in which mechanically generated waves, which could represent both wind waves and swell, were used to explore the influence of wave properties on turbulence. They found larger estimates of turbulence dissipation for wind waves compared to swell waves of the same wave height, and larger dissipation rates for waves of larger wavelength and height. Lai et al. (2018) only investigated two different wave frequencies however and used monochromatic waves which are not representative of a real ocean sea state.

1.2.4 Objective of the Turbulence Study

Although many studies have focused on refining parameterizations of $\varepsilon$ in order to improve formulations of $S_{\text{diss}}$ in wave forecasting models and to improve ocean circulation models (Lai et al., 2018; Savelyev et al., 2012; Alves and Banner, 2002), several research gaps still exist as mentioned previously. To fill these gaps, the collection of more observational data in different wind and wave fields to investigate $\varepsilon$ is required. Although there are many methods by which $\varepsilon$ can be quantified in both the laboratory and the field, there are many theoretical limitations involved in these methods which should be considered prior to planning an experiment with the objective to quantify $\varepsilon$. A second goal work is to provide a guide for the selection of an appropriate indirect method for the quantification of $\varepsilon$ based on the type of measurement device and the experimental conditions. Additionally, this work aims to quantify turbulence below irregular waves in the presence of wind in a controlled laboratory setting in order to provide a better understanding of the influence of wind on wave-induced turbulence in real ocean sea states.
1.3 Overview of Coming Sections

The remainder of this thesis is divided into three additional chapters. The second chapter of this thesis investigates the ability of an existing wind wave suppression model to predict wind wave suppression on the surface of irregular waves. The third chapter describes several methods for quantifying wave-induced turbulence in the presence of wind and demonstrates one of these methods in a laboratory experiment. The final chapter of this thesis summarizes the conclusions from the two previous chapters and discusses the implications of this work.
CHAPTER 2

PREDICTING WIND WAVE SUPPRESSION ON IRREGULAR LONG WAVES

2.1 Chapter Abstract

The applicability of the wind wave suppression model developed by Chen and Belcher (2000) to irregular wave environments is investigated in this study. Monochromatic and irregular wave environments were simulated in the W² (Wind/Wave) laboratory at the University of Maine under varying wind speeds. The Chen and Belcher (2000) model accurately predicts the reduction of the energy density of the wind waves in the presence of the monochromatic waves as a function of wave steepness but under predicts this energy dissipation for the irregular waves. This is due to the consideration of a single wave frequency in the estimation of the growth rate and wave-induced stress of the monochromatic waves but cannot be applied to irregular waves because their spectra contain energy over a wide range of frequencies. A revised version of the model is proposed to account for the energy contained within multiple wave frequencies from the power spectra for the mechanically generated irregular waves. The revised model shows improved results when applied to the irregular wave environments.

2.2 Introduction

For decades researchers in the ocean engineering and modeling communities have focused on developing accurate representations of fetch-limited wave growth under wind forcing. This objective is based on a need for refinement of the source functions of the spectral energy balance used in wave forecasting models,

\[ S = S_{in} + S_{non} + S_{dis} \]  

(2.1)

where \( S_{in} \) is the input of wind energy, \( S_{non} \) accounts for the wave energy associated with nonlinear interactions, and \( S_{dis} \) is the dissipation by wave breaking. The development of parameterizations of the wind input source term, \( S_{in} \), have been complicated by an interesting phenomenon observed in laboratory
studies (Mitsuyasu, 1966; Phillips and Banner, 1974; Donelan, 1987) in which short wind waves (~2 – 6 Hz) are dampened by the superposition of steep, mechanically generated “swell”. Since these pioneering studies, it has pointed out that it is not possible to represent real ocean swell using mechanically generated waves in a laboratory due to limited fetch, which does not allow enough horizontal distance for the waves to grow to lengths that would allow them to travel faster than the wind (Makin, 2007). Here we will not make the mistake of referring to mechanically generated waves as swell, and instead will refer to these waves as long waves, i.e. waves much longer than the high frequency wind waves, but waves that have not yet developed into swell. Additionally, wind wave modification by swell is unlikely to present itself in the ocean, because swell steepness is typically much lower than the steepness of long waves in the laboratory (Chen and Belcher, 2000). Regardless, the suppression of wind waves on long waves is a real phenomenon that occurs in laboratory studies, the data from some of which is still used to validate theoretical wave growth models that have been incorporated into source term formulations of wave forecasting models today. For example, the model developed by Chen and Belcher (2000), which parameterizes a reduced turbulent stress available to grow the wind waves, due to the presence of a long wave, was validated using laboratory data in which wind wave suppression was observed (Mitsuyasu, 1966; Phillips and Banner, 1974; Donelan, 1987). The model presented in Chen and Belcher (2000) has been used to develop a new parameterization for the friction velocity, \( u_* \), which accounts for reduced growth in the high frequency tail of the wave spectrum (i.e. the portion of the wave spectrum associated with short wind waves) which is now incorporated into parameterizations of the wind input source term, \( S_{in} \), in WAVEWATCH III®, a third-generation wave forecasting model (Ardhuin et al., 2010; The WAVEWATCH III® Development Group, 2016). However, the laboratory data used to validate the model presented in Chen and Belcher (2000) included only monochromatic long waves, which do not represent real sea states as closely as irregular, random waves. This study aims to determine the applicability of the Chen and Belcher (2000) wind wave suppression model for irregular, random waves.
2.2.1 Incorporation of Wind Wave Suppression into $S_m$

Two separate routes of investigation have emerged with the overarching aim to improve parameterizations of $S_m$. The first is the exploration of the growth of both the wind waves and the long waves on which they travel with a focus on the determination of accurate measures of growth rates and speculation as to the mechanism responsible for the suppression of wind waves on long waves (Phillips and Banner, 1974; Mitsuyasu, 1982; Donelan, 1987; Peirson and Garcia, 2008). The second is the investigation of wind-generated waves with the goal of developing accurate fetch- and duration-limited models to predict their growth (Hasselmann et al., 1973; Dobson et al., 1989; Donelan et al., 1992; Young and Verhagen, 1996; Hwang, 2006; Lamont-Smith and Waseda, 2008). It was not until the Chen and Belcher (2000) model (from here on referred to as CBM) accounted for wind wave suppression on long waves in existing fetch-limited wind wave growth models that these two routes of investigation converged.

The first route of investigation mentioned above started when Mitsuyasu (1966) first observed that when high frequency wind waves, $f \sim 2.5$ Hz, travel along the surface of a long, monochromatic wave train, $f \sim 0.5 - 0.75$ Hz, the wind waves are sometimes suppressed, depending on the steepness of the long wave. No speculation as to the cause of this suppression was presented however, and only a schematic representation of a possible equilibrium model between the wind waves, long waves, and the wind was presented. Phillips and Banner (1974) observed this wind wave suppression on long waves of steepness, $ak \sim 0.02 - 0.20$, where $a$ is the wave amplitude and $k$ is the wavenumber, and speculated that the mechanism responsible for suppression is instabilities that generate breaking of the small wind waves, resulting in a transfer of energy from the wind waves to the long wave. Donelan (1987) proposed a different hypothesis for the cause of wind wave suppression, suggesting that the long waves cause a ‘detuning of the resonance’ of the nonlinear interactions amongst the wind waves (i.e. the long waves disrupt the nonlinear interactions amongst the wind waves preventing resonance from occurring), which reduces the energy in the high frequency part of the spectrum. The $ak$ values of the monochromatic waves considered by Donelan (1987) were 0.053 and 0.105 Hz, with the frequency peak of the wind waves of approximately 1.5 - 2 Hz. Neither Phillips and Banner (1974) nor Donelan (1987) make any attempt to directly provide a method by which
this phenomenon could be incorporated into wave forecasting models, though Donelan (1987) does suggest its importance.

The CBM combined the aforementioned studies to account for wind wave suppression in existing wind wave growth models in order to accurately represent the growth of wind waves in the presence of long waves, converging the two routes of study described above. They present a formulation for the growth rate of the long waves,

\[ \gamma_L = \frac{\rho_a}{\rho_w} \beta (\frac{u_\ast}{c_L})^2 \sigma_L \]  

(2.2)

where \( \rho_a \) is the density of air, \( \rho_w \) is the density of water, \( \beta \) is the growth rate coefficient, \( u_\ast \) is the air friction velocity, \( c_L \) represents the phase speed of the long waves, and \( \sigma_L \) is the angular frequency of the long wave based on formulations by Belcher (1999) and van Duin (1996). The novel contribution of the CBM is the assumption that the total surface stress, \( \tau_{tot} = \rho_a u_\ast^2 \), is composed of two individual stresses: the wave-induced stress of the long wave, \( \tau_L \), and the portion of the stress that remains to grow the wind waves, i.e. the turbulent stress, \( \tau_t \). This assumption allows for an estimate of \( \tau_t \) dependent on the properties of the long wave by calculation of \( \tau_L \) and \( \tau_{tot} \) (see section 2; Chen and Belcher, 2000). Lastly, to account for the influence of the long wave on the wind wave growth, \( \tau_t \) is used to replace \( \tau_{tot} \) in fetch-limited wave growth models, such as the two models created by Mitsuyasu and Rikiishi (1978), by reformulating the friction velocity, \( u_\ast = (\tau_{tot}/\rho_a)^{1/2} \), as \( u_\ast = (\tau_t/\rho_a)^{1/2} \) (Chen and Belcher, 2000). This newly formulated friction velocity can be incorporated into the aforementioned fetch-limited growth model from Mitsuyasu and Rikiishi (1978) which is,

\[ \frac{u_\ast \sigma}{g} = 7.48 (\frac{\sigma x}{u_\ast^2})^{-0.357} \]  

(2.3)

where \( g \) is gravitational acceleration and \( x \) is the horizontal fetch distance. Additionally, the formulation for the growth of wind waves in terms of energy density relies on the new formulation of the friction velocity as,

\[ \frac{gE}{\rho_w u_\ast^4} = 4.49 \times 10^{-5} (\frac{\sigma x}{u_\ast^2})^{1.282} \]  

(2.4)
where the energy density, $E = \rho_w g \overline{\eta^2}$, with $\overline{\eta^2}$ defined as the variance of the wind waves, instead of their peak angular frequency, $\sigma$, as in Eq. 2.3 (Chen and Belcher, 2000). Chen and Belcher (2000) speculate that the Phillips-Banner mechanism (i.e. the breaking of the small wind waves that result in a transfer of energy from the wind waves to the long wave) is the most likely cause of the suppression of the wind waves, and that the breaking of the small wind waves may enhance air flow separation (AFS) at the long wave crest. However, as mentioned above, laboratory studies on this model have so far only included monochromatic waves, which are not as representative of actual sea states as irregular waves.

Therefore, the goal of this study is to determine the applicability of the CBM to irregular waves. To reach this goal, the following specific research questions will be answered: 1) Does the CBM accurately predict the decrease in the energy of the wind waves in the presence of irregular waves of increasing steepness? 2) Can the CBM model be improved by accounting for the long wave energy at each wave frequency of the wave spectrum for an irregular wave environment? These research questions will be answered through analysis of data collected in a laboratory experiment conducted in the Harold Alfond Wind/Wave (W$^2$) Ocean Engineering Laboratory at the University of Maine Advanced Structures and Composites Center (ASCC). The data set consists of both irregular and monochromatic waves with a variety of wave heights and periods in addition to a range of wind speeds.

The remainder of this paper will include a description of the W$^2$ test facility at the University of Maine, the experimental set-up and a description of the test campaign description. This is followed by a characterization of the wind field in the W$^2$. Section 2.4 describes the CBM model, which is followed by methodology in Section 2.5. Section 2.6 will present the results of the CBM for the monochromatic and irregular long wave environments, as well as the results of the computation of the growth rate coefficient for these waves. The results will be interpreted in a discussion presented in Section 2.7, and a modified model will be introduced to expand the applicability of the CBM to irregular waves. Finally, Section 2.8 will provide conclusions about the applicability of CBM to irregular waves generated in the laboratory and implications for the modified model.
2.3. Experimental Set-up and Data Collection

2.3.1 Test Facility

The laboratory experiments were performed at the Harold Alfond W² (Wind/Wave) Ocean Engineering Laboratory at the University of Maine Advanced Structures and Composites Center (ASCC). The 30 m long by 9 m wide basin has a 16-flap paddle wave generator below a 5 m by 3.5 m by 6 m open-jet wind tunnel (Figure 2.1). At the opposite end of the basin is an energy absorbing elliptical beach designed to minimize wave reflection. The basin possesses a moveable concrete floor that was set to allow a water depth of 4.5 m for the duration of the experiment.

![Figure 2.1. W² Basin in the ASCC at the University of Maine. The wind machine is located above the 16-paddle flap wave generator over the 30 m long by 9 m wide basin.](image)

2.3.2 Instrumentation & Data Collection

Two separate data collection campaigns were conducted in the W² in order to obtain a comprehensive data set with a wide range of wave steepness values. A combined total of 15 long wave environments were simulated under five different wind speeds. The first campaign included four test wind
speeds: \( U = 0, 7, 8.5 \) and 10 m/s and only irregular wave environments. The second campaign expanded the original data set to include more waves of varying steepness and an additional wind speed, \( U = 5.5 \) m/s, as well as monochromatic waves. Data were also collected for wind-only seas created in the absence of long waves under each test wind speed, in order to observe the production of wind waves on the water surface. All wave environments considered in this study are outlined in Table 1.1, including the irregular wave environments, the monochromatic waves (denoted with the symbol M), and wind only seas. Wave environments are defined by the significant wave height, \( H_s \), and the peak wave period, \( T_p \), the period that corresponds to the frequency at the peak of the wave spectrum. Also presented in Table 1.1 is the steepness parameter \( a_k \), where \( a = \frac{1}{2} H_s \) is the approximate wave amplitude, and \( k = \frac{2\pi}{L} \) is the wavenumber where \( L \) the wavelength at the peak frequency. A JONSWAP gamma, \( \gamma \), of 3.3 was used to characterize the frequency spectrum for the irregular waves. Each of these wave environments were coupled with all of the wind speeds used during their respective test campaign. In addition to the parameters presented in Table 1.1 that defined the wave environments, for all irregular environments a cosine 2S shape function was defined, using an S value of 4.0, as well as a directional (angular) spread of +/- 10 degrees. All test conditions in this experiment (i.e. wind velocities and wave characteristics) are full scale.

<table>
<thead>
<tr>
<th>ID</th>
<th>( H_s ) (m)</th>
<th>( T_p ) (s)</th>
<th>( a_k L )</th>
<th>( U ) (m/s)</th>
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Table 1.1. Wave parameters used to characterize the JONSWAP for the wave environments.
The wind and wave environments were examined using a three-dimensional wind/wave array assembled with thirteen wave probes, four hotwire anemometers and one ultrasonic anemometer (Figure 2.2). This array was used for both test campaigns with dimensions varying only for the anemometer heights above the water surface as shown in Figure 2.2b and 2.2c. Wind velocities were collected using four TSI 8455 hotwire anemometers at a sampling rate of 20 Hz with a response frequency of 5 Hz, and one R.M. Young 81000 ultrasonic anemometer sampling at 32 Hz. Surface elevation was measured with 13 Akamina AWP-24-3 capacitive style wave height gauges with accuracy up to 1.35 mm. The wave probes were calibrated and zeroed to a mean water level before the data collection began and were set to sample at a rate of 32 Hz. The setup of the wave gauges and anemometers mounted on the instrumentation array are shown in Figure 2.2b and 2.2c for the first and second test campaigns respectively. The instrumentation array collected surface elevation data and wind speed data at three fetch distances: 4.96 m, 10.32 m, and 14.31 m with respect to the anemometer staff location, though only data collected at a fetch of 14.31 m will be considered for this study, because the longest fetch distance allows for the most development of the wind waves. This array position is shown in Figure 2.2a. The wind data collected at the other two fetch distances, 4.96 m and 10.32 m will be used to aid in the characterization of the wind field, which will be presented in Section 2.3.3.

Data were collected using a directional array assembly similar to the design proposed by the Coastal Engineering Research Center (CERC) (Panicker and Borgman, 1970), but only Probe C, along the center line of the basin was used for post-processing for this study. This is due to its close proximity to the anemometer staff, providing the surface elevation record that most closely corresponds to the reported values of wind speed, air friction velocity and roughness length. The wave absorbing beach shown in Figure 2.2a is expected to dissipate approximately 90% of the wave energy (Ouslett, 1986) and the flap wave
generators are also designed to actively absorb reflected wave energy. Additionally, a settling time of three minutes was also allotted between wave runs in order to reduce the presence of wave components of the previous test in the next wave record. Each irregular wave environment outlined in Table 2.1 was run for a 20-minute time period. Monochromatic wave environments were run for 15 minutes.

Figure 2.2. Test set-up and Instrumentation. (a) Array position in the basin with respect to the wave maker and wind jet. The anemometer staff on the array is used as the horizontal reference for the array position to the wave maker. Dimensions shown are in meters. Side view of the instrumentation array with dimensions of (b) the first test campaign and (c) the second test campaign. Only center wave probes are displayed. Measurements shown are vertical distances from the mean water level to the anemometer locations. All dimensions are in meters.
2.3.3 Characterizing the Wind Field

Using wind data from both test campaigns, the mean wind speed in the horizontal direction, $\bar{U}$, collected by each anemometer on the staff was used to obtain a profile for $U = 7$ m/s, $U = 8.5$ m/s and $U = 10$ m/s for both campaigns, and additionally $U = 5.5$ m/s for the second campaign. These profiles of $\bar{U}$ over the still water surface were used to obtain estimates of $u_*$ via a least-squares fit following the log law,

$$\frac{\bar{U}}{u_*} = \frac{1}{\kappa} \left( \frac{z}{z_0} \right)$$

(2.5)
in a manner similar to Mitsuyasu and Yoshida (2005). In Eq. 5, $\kappa$ is the von Karman constant ($\kappa \sim 0.4$), $z$ is the vertical elevation above the water surface, and $z_0$ is the aerodynamic roughness length (Jones and Toba, 2001). Figure 2.3 shows an example of the wind profiles measured for $U = 10$ m/s at the three fetch distances ($x = 4.96$ m, 10.32 m, and 14.31 m) at which data was collected and the log law extrapolation to the water surface to obtain estimates of $u_*$ and $z_0$. Additional data was collected with the instrumentation array moved approximately 46 cm closer to the water surface in order to extend the wind profile to obtain more accurate estimates of $u_*$ and $z_0$. In Figure 2.3 this data is referred to as ‘shear data’. This data was combined with the test data, which refers to data collected for the wind in the absence of waves during the testing of this study. The $u_*$ values obtained for the wind speeds, $U = 5.5 – 10$ m/s, are on the order of 0.51 – 0.99 m/s. It is evident from the profiles in Figure 2.3 that the wind profile generated by the wind machine is more like that of a jet than a logarithmic profile, typical of real atmospheric conditions, especially at $x = 4.96$ m. The wind profile is most like a logarithmic profile for the largest fetch, $x = 14.31$ m.
Figure 2.3. Example of wind profiles for $U = 10$ m/s measured at three fetch distances: $x = 4.96$ m, 10.32 m, and 14.31 m. Green symbols represent the additional measurements collected in a wind shear study to extend the wind profile to the water surface. Yellow symbols represent wind data collected during the testing in the absence of waves.

Due to the jet-like geometry of the wind machine in the W$^2$, the values of roughness length, $z_0$, are on the order of $10^{-2}$ m, which are higher than what is observed in many other laboratories for similar wind speeds. Mitsuyasu and Yoshida (2005) obtained values of $z_0$ on the order of $10^{-4}$ m using a similar method of extrapolation of the wind profile via fit to the log law and wind speeds on the order of $U = 10$ m/s. The values of $z_0$ obtained in the W$^2$ are high and are therefore more representative of storm sea states observed in the field in which wind waves are more likely to be well developed (Toba and Ebuchi, 1991).

### 2.4 The Chen and Belcher (2000) Model

The model presented in Chen and Belcher (2000) to estimate the effect of long waves on wind waves was applied in this study for irregular waves. The CBM model assumes that the momentum associated with
the total stress at the water surface, $\tau_{tot}$, is composed of the long wave induced stress, $\tau_L$, and the turbulent wind stress that induces wind wave growth, $\tau_t$ so that $\tau_{tot} = \tau_t + \tau_L = \rho_0 u^2_*$. The influence of the long wave on wind wave growth is then determined via the reduction of the turbulent wind stress, $\tau_t$. This is done considering only four independent parameters: the long wave steepness in the absence of wind, $a_L k_L$, the atmospheric pressure coefficient, $\alpha_p$, dimensionless frequency, $\sigma_* = \sigma_L u_*/g$, and dimensionless fetch, $X_* = gx/u_*^2$, where the subscript, $L$, denotes the properties of the long waves in the absence of wind. This is expanded upon in the following sections.

2.4.1 Long Wave-Induced Stress

The long wave-induced stress, $\tau_L$, can be quantified by considering the long wave wind induced growth rate, $\gamma_L$, which is defined by

$$\frac{\partial \Phi_L(\sigma)}{\partial t} = \gamma_L \Phi_L(\sigma),$$  \hfill (2.6)

where $\Phi_L$ is the long wave spectral density, and the formulation for $\gamma_L$ for monochromatic waves is as expressed in Eq. 2.2. Based on the rate of change of the momentum density of the long wave over time, $\tau_L$ is expressed as,

$$\tau_L = \rho_w \int_0^\infty \gamma_L \sigma_L \Phi_L(\sigma) d\sigma.$$  \hfill (2.7)

Therefore, if $\tau_L$ is quantified for long waves in the absence of wind and $u_*$ is known, $\tau_t$ can be calculated as $\tau_t = \tau_{tot} - \tau_L$. In order to accurately quantify the long wave induced stress, the growth of the long wave, $\gamma_L$, as expressed in Eq. 2.2 must be quantified.

2.4.2 Growth Rate of the Long Wave, $\gamma_L$

Belcher (1999) investigated non-separated sheltering induced wave growth and van Duin (1996) investigated wave growth induced by turbulent air flow over waves. Both studies arrived at similar formulations of the wave growth rate due to these mechanisms Eq. 2.2 after making several important assumptions. The first major assumption is that the wave steepness is on the order of $ak << 1$. The second major assumption is that the long waves fall within the regime of slow waves or fast waves which are
defined by the wave age thresholds of \(c/u^* < 15\) and \(c/u^* > 25\) respectively for \(kz_0 \sim 10^{-4}\). Finally, the model of Belcher (1999) was formulated based on the energy equation for deep water waves, therefore Eq. 2.2 can only be applied to waves that can be considered short, based on their relative depth, \(d/L\), where \(d\) is the water depth and \(L\) is the wavelength. The origin of the growth rate coefficient, \(\beta\), is an integral parameter in the CBM, which describes the growth of waves by wind. This parameter, presented in Eq. 2.2, has evolved over decades through consideration of different mechanisms by which it is possible for the wind to transfer energy and momentum to the waves (Plant, 1982; Belcher and Hunt, 1993; Belcher, 1999). The development of \(\beta\) used in the CBM is based on the physical model by Belcher (1999) of non-separated airflow sheltering over slow waves, which are defined by Cohen and Belcher (1999) as waves for which,

\[
kz c = kz_0 e^{kc/u^*} \ll 1,
\]

where \(z_c\) is the critical layer height. For slow waves, for which the wind speed is greater than the wave speed, the critical layer is small, and therefore \(kz_c\) is small, indicating that the waves are forced by wind and not vice versa (Cohen and Belcher 1999).

The model presented in Belcher (1999) to quantify \(\beta\) is based on the criterion that the waves fall within this slow wave regime, indicating that there is non-separated sheltering immediately above the water surface. Although the values of \(z_0\), on the order of 0.01 m, obtained in the \(W^2\) are high compared to other laboratory wind fields (Mitsuyasu and Yoshida, 2005) for similar wind speeds, the values of \(kz_c\) for the highest wind speed, \(U = 10\) m/s, are on the order of \(10^{-1}\) m. This indicates that the wave environments of this study fall within the slow wave regime.

The non-separated sheltering assumption guarantees the existence of an inner layer (just above the water surface, where the wave speed is greater than the wind speed, \(c > U\)) and a middle layer (within the outer layer, where \(c < U\)) (Figure 2.4) that are characterized by different flow behaviors, yet do not experience boundary layer separation at the wave crests. In order to calculate an appropriate value of \(\beta\) to use for the model for the irregular waves, the inner layer, \(l_i\), and middle layer, \(h_m\), heights were calculated following the methodology of Belcher and Hunt (1993) using the long wave data under the action of
wind. To iteratively quantify these length scales, the assumption was made that both \( l_i \) and \( h_m \) were greater than \( z_c \) (or the matched height, \( z_m \), according to Belcher and Hunt, 1993) which is defined by Belcher and Hunt (1993) to be the vertical location at which the wind speed and the wave speed are equivalent.

![Diagram of the inner, middle, and outer layer](image)

**Figure 2.4.** Diagram of the inner, middle, and outer layer as described in Belcher and Hunt (1993) (adapted from Belcher 1999).

To calculate the \( \beta \), five components must be considered (Belcher 1999),

\[
\beta = \beta_{sz_i} + \beta_{su_i} + \beta_{p_o} + \beta_{\eta s} + \beta_{u_s}. \tag{2.9}
\]

The first two of the five components account for non-separated sheltering effects, including the influence of shear stress on the inner region due to undulations on the water surface, \( \beta_{sz_i} \), and due to changes of velocity at the water surface, \( \beta_{su_i} \). The third is due to the variations in pressure in the outer region, \( \beta_{p_o} \).

The respective equations are,

\[
\beta_{sz_i} = 2 \left( \frac{U_{ml} - c}{u^*} \right)^4 \left\{ 2 - \frac{c}{U_{il}} \right\} \tag{2.10}
\]

\[
\beta_{su_i} = -2 \left( \frac{U_{ml} - c}{u^*} \right)^2 \frac{c}{U_{il}} \tag{2.11}
\]

\[
\beta_{p_o} = 2 \kappa \delta^2 n \left( \frac{U_{ml} - c}{u^*} \right) \tag{2.12}
\]

where \( \overline{U_{ml}} \) and \( \overline{U_{il}} \) are the mean wind speeds at \( h_m \) and \( l_i \) respectively, calculated as,

\[
\overline{U_i}(z) = \left( \frac{U_*}{k} \right) ln \left( \frac{z}{z_o} \right) \tag{2.13}
\]
where \( z \) is set equal to \( h_m \) and \( l_i \), and the subscript \( l \) is set to \( ml \) or \( il \), respectively. In Eq. 2.12, \( \delta = \kappa / |\ln (kz_e)| \) for slow waves (Cohen and Belcher, 1999) and \( n \) is a model coefficient between 0 and 1 (Belcher, 1999), set here to 0.5. The fourth and fifth terms in Eq. 2.9 arise from the wave-induced surface shear stress, where the first contribution is from the waves, \( \beta_{\eta_s} \), and the second from variations in surface velocity, \( \beta_{u_s} \), where

\[
\beta_{\eta_s} = \frac{2(\bar{u}_m - c)^2}{(\bar{u}_m - c)\bar{u}_d}
\]

(2.14)

\[
\beta_{u_s} = -\frac{2c}{\bar{u}_d}.
\]

(2.15)

According to Belcher (1999), the term \( \beta_{sz_l} \) should dominate the growth rate coefficient for slow waves, which is what is expected for this study. Once \( \beta \) has been quantified, \( \gamma_L \) can be determined and used for calculation of the long wave induced stress, \( \tau_L \) (Eq. 2.7). Combining Eq. 2.2 and Eq. 2.7, an expression for the dependence of the long wave amplitude development with fetch (under a steady wind), denoted, \( \epsilon_L \), is formulated as

\[
\epsilon_L = \exp \left[ 2 \frac{\rho_a}{\rho_w} \frac{\alpha_p}{1 + (a_p k_L)^2} \sigma_p^4 (X_* - X_{0*}) \right],
\]

(2.16)

where the subscript, \( o \), denotes a value measured at the location of the initial fetch, meaning the horizontal distance at which the wind first impacts the long wave, \( \sigma_* = \sigma_L u_* / g \) is the dimensionless frequency of the long wave, \( X_* = g x / u_*^2 \) is the dimensionless fetch, and \( x \) is the fetch. The result from combining Eq. 2.7 and Eq. 2.16 is a ratio which describes the distribution of the turbulent stress to the total stress,

\[
\frac{\tau_t}{\tau_{tot}} = \frac{1}{1 + (a_L k_L)^2 \sigma_p \epsilon_L}.
\]

(2.17)

To assess the influence of the long waves on the wind waves through the reduction of the turbulent stress, the CBM utilizes the ratio of the wind wave energy in the presence of long waves, \( E_{L+ww} \), to the wind wave energy in the absence of long waves, \( E_{ww} \), to compare the model results to data collected in the laboratory. Assuming that in the presence of the long wave \( u_* \) can be rewritten in terms of the turbulent stress as \( u_* = (\tau_t / \rho_a)^{1/2} \), the ratio \( \frac{E_{L+ww}}{E_{ww}} \) is derived from Eq. 2.4 as,
\[
\frac{E_{L+ww}}{E_{ww}} = \left( \frac{\tau_t}{\tau_{tot}} \right)^{1.36}
\]

(2.18)

which for short fetch distances such as the conditions in the laboratory, can be simplified to

\[
\frac{E_{L+ww}}{E_{ww}} = \left( \frac{1}{1 + \frac{1}{4}(a_Lk_L)^2\alpha_p} \right)^{1.36}
\]

(2.19)

due to the dependence of Eq. 2.19 on \(a_Lk_L\) and \(\alpha_p\). The energy ratio presented from the model (Eq. 2.20) will be compared to estimates made directly from laboratory data. Data analysis techniques used to determine the ratio from data will be described in the following section. For further details on the model derivation the reader is referred to Chen and Belcher (2000).

2.5 Methods: Data Analysis

2.5.1 Experimental Energy Ratio, \(\frac{E_{L+ww}}{E_{ww}}\)

This section describes the data analysis performed in order to obtain experimental values of \(\frac{E_{L+ww}}{E_{ww}}\) directly from the laboratory data to compare to the values of \(\frac{E_{L+ww}}{E_{ww}}\) obtained from the model. The procedure for obtaining the values of \(E_{ww}\) and \(E_{L+ww}\) will be discussed separately, as additional data analysis techniques were required to obtain \(E_{L+ww}\) due to the presence of the long wave in the surface elevation records.

To calculate \(E_{ww}\) directly from the data, a spectral analysis and a wavelet analysis were performed on the surface elevation records collected at Probe C for wind only runs for all wind speeds. Using the wave gauge sampling interval of 0.0313 s, a Fast Fourier transform (FFT) of 1024 points was applied to the surface elevation record with the first two minutes of the record removed as contribution to start up time for the wind waves to develop fully. Frequencies lower than 0.1 Hz were cutoff in order to accurately quantify the peak period from the spectra and a high pass filter with a cutoff at \(f = 2\) Hz was performed to remove residual low frequency energy observed in the wave spectra. The variance of the wind sea, \(\bar{\eta}^2\).
was quantified from the spectral results as $\overline{\eta^2} = S_f \Delta f$, where $S_f$ is the spectral energy at each frequency separated by $\Delta f = 0.0078$ Hz. The energy of the wind sea alone was then quantified as $E_{ww} = \rho_w g \overline{\eta^2}$.

To calculate $E_{L+ww}$ from the data, a wavelet reconstruction method, outlined in Torrence and Compo (1998) was applied. First, a wavelet transform was performed on the data in order to analyze the energy within the high frequencies and how it varies with time. The Morlet mother wavelet function was used with a spacing of 0.01 between discrete scales to provide a fine resolution. All other input parameters were set to default (Torrence and Compo, 1998). The sampling interval in this case was the same as the sampling interval of the surface elevation record, which was approximately 0.0313 s. Surface elevation records used in the analysis were unfiltered in order to capture the high frequency oscillations associated with wind waves. Based on the observed peak frequencies of the wind sea alone and the results of the wavelet analysis, the energy associated with frequencies, $f > 2$ Hz, were deemed to be associated predominantly with energy input from the wind. Therefore, the wavelet reconstruction was used to recreate the surface elevation records for each case using only the energy associated with $f > 2$ Hz.

Following the reconstruction of the surface elevation record of the high frequencies for the irregular waves, despiking of each individual record was performed, using a threshold wind wave amplitude of 1 cm to identify spikes, and replacing the spikes with the mean wave amplitude of the record, which is one option for spike replacement as suggested by Goring and Nikora (2002). Prior to obtaining the variance of each high frequency record for the irregular and monochromatic waves, a second method to ensure removal of any potentially remaining low frequency long wave energy was applied, which consisted of a high-pass filter with a cutoff frequency of $f = 2$ Hz. Following the high-pass filter, the spectral analysis was performed on the high frequency record and the energy of the wind waves in the presence of the long wave, $E_{L+ww}$, was calculated using the same methods as described to calculate $E_{ww}$. For the mixed sea states of monochromatic waves and wind, the code WaveSpectraFun from the OCEANLYZ Ocean Wave Analyzing Toolbox was used to perform the spectral analysis on the high frequency records (Karimpour, 2017). From the spectral analysis of the monochromatic waves, the same method was then used to
quantify $E_{L+ww}$ as outlined for the irregular waves, despite the smaller spread of the peak spectral energy. This was performed for consistency and was decided based on the knowledge of the distribution of spectral energy for the wind sea alone.

2.5.2 Quantifying the Growth Rate Coefficient and the Atmospheric Pressure Coefficient

Data from several laboratory studies (Mitsuyasu, 1966; Phillips and Banner, 1974; Donelan, 1987) are used to validate the CBM. A value of $\alpha_p = 80$ was selected based on a fit to the data, and values of $\beta$ were quantified from the relation,

$$\beta = \frac{\alpha_p}{1 + \frac{k_L^2 \eta_L^2}{\alpha_p}},$$

(2.20)

which can be deduced from Eq. 2.2. The values of $\beta$ obtained via this method were considered acceptable because they fell within the range of $\beta = 34 \pm 16$, which was the range obtained by Plant (1982). This practice was adopted here in the selection of an appropriate $\alpha_p$ value for the wave data of this study. A $\alpha_p$ value was selected based on fit of the model results to the experimental energy ratios, then $\beta$ quantified via Eq. 2.21 and verified to be within the range of values obtained by Plant (1982).

2.6 Results

2.6.1 CBM Results for the Monochromatic Waves

As was found in Chen and Belcher (2000) the monochromatic waves in this study indicate that as wave steepness increases, energy is transferred from the long wave to the wind waves ultimately decreasing the energy ratio. This was found for all wind speeds considered in this study, except $U = 5.5$ m/s (Figure 2.5). The experimental results deviate from the model at $U = 5.5$ m/s, due to reduced wind wave growth leading to a lower steepness threshold for wind wave suppression. The model does well to predict the energy ratio for $U = 5.5$ m/s up until a value of $a_t k_L \approx 0.12 – 0.14$. 
For the model (Eq. 2.19), $\alpha_p$ was set to 100 based on best fit to the data and were compared to values of $\alpha_p$ used to assess previous experimental data (Phillips and Banner, 1974; Mitsuyasu, 1966) in Chen and Belcher (2000), for which $\alpha_p \sim 80 – 160$.

Figure 2.5. The CBM model results for the experimental monochromatic wave data. The red, solid line represents the model results for $\alpha_p = 100$, using the simplification of $\frac{\tau_c}{\tau_{tot}}$ in Eq. 2.19 and a range of $a_L k_L \sim 0 – 0.2$. Red, open symbols represent model results using Eq. 2.17 to calculate $\frac{\tau_c}{\tau_{tot}}$ for the $a_L k_L$ values of our long waves with no wind. The shape of the open symbols correspond to the same wind speeds as the closed symbols. Black, closed symbols represent results from laboratory data.

2.6.2 CBM Results for Irregular Waves

Here the CBM is applied to the irregular waves of this study, and it is shown that the decrease in the energy ratio with increasing steepness occurs far more quickly than is predicted by the model (Figure 2.6). Using the same value of $\alpha_p = 100$ as was used in the model for the monochromatic waves, the model results were computed for the irregular waves. In order to apply Eq. 2.2 and Eq. 2.17 to the irregular wave environments, the values of $c_L$, $k_L$ and $\sigma_L$ are based on the significant properties of the waves, as opposed
to the single frequency of a monochromatic wave. Here, the dispersion relation was used to iteratively quantify $L$ based on the peak period, $T_p$, of the wave spectra, $c_L$ was calculated as $c_L = L/T_p$, $k_L$ as $k_L = 2\pi/L$, and $\sigma_L = 2\pi f_p$, where $f_p$ is the peak frequency of the wave spectrum. The variance of the long wave, $\overline{\eta^2_L}$, was quantified as the zeroth-moment of the wave spectrum.

The results in Figure 2.6 show that the slope of the model curve, calculated from Eq. 2.19, and the model points, computed from Eq. 2.17 in Eq. 2.18, is too small to fit the trend of the experimental energy ratios with increasing steepness for the irregular waves. It was estimated that an $\alpha_p$ of approximately 300 would be required in order to fit the model results to the experimental energy ratios of the irregular waves (not shown), which for the range of steepness values considered ($a_L k_L \sim 0.03 – 0.15$) produce corresponding $\beta$ values between 69 – 264. The values of $\alpha_p$ and $\beta$ are far too large to be reasonable, falling outside the range obtained by Plant (1982).

![Figure 2.6](image1.png)

Figure 2.6. The CBM model results for the experimental irregular wave data. The red, solid line represents the model results for $\alpha_p = 100$, using the simplification of $\frac{\tau_L}{\tau_{tot}}$ in Eq. 2.20 and a range of $a_L k_L \sim 0 – 0.2$. Red, open symbols represent model results using Eq. 2.17 to calculate $\frac{\tau_L}{\tau_{tot}}$ for the $a_L k_L$ values of
our long waves with no wind. The open symbols correspond to the same wind speeds as the closed symbols. Black, closed symbols represent results from laboratory data.

The waves with $a_0 k_L > 0.1$ do not follow the model curve, nor the trend of the model energy ratios for waves of $a_0 k_L \leq 0.1$, due to breaking of the tallest waves, which induces high frequency energy (Banner and Peregrine, 1993) that results in an increase in the variance of the high frequency record. Wave breaking of the significant waves of these cases was confirmed visually in test video of the experiment (not shown). Because the value of $\alpha_p$ required to fit the model to the experimental data is far too large to be reasonable, it was necessary to directly quantify values of $\beta$ and the associated $\alpha_p$ for the irregular waves following Belcher (1993) and Belcher (1999), as outlined in Section 2.4 rather than using $\alpha_p$ as a fitting parameter. These results will be presented next.

2.6.3 The Growth Rate Coefficient for the Irregular Waves

Following the methodology of Belcher (1999) summarized in Section 2.4.2, values of $\beta$ were quantified using Eqns. 2.10-2.12 and 2.14-2.15 based on the irregular wave data collected in the experiments. It was first verified that the assumptions of the model by Belcher (1999) were met for the wave environments of this study; the long wave environments fall within the slow wave regime, i.e. $c/U_* < 15$, are of steepness, $a k \ll 1$, and can be considered deep water, i.e. a relative depth, $d/L$ of approximately 0.5 or greater. The long wave data in the absence of wind was then used to obtain the properties of the waves to convert the actual values of $\beta$ to $\alpha_p$ through iterative calculation of Eq. 2.20. For the irregular waves of this study, under $U = 5.5 - 10$ m/s, $\alpha_p \sim 27-49$, and $\beta \sim 25-34$. Figure 2.7 shows the estimates of $\beta$ versus wave age, $c/U_*$, for the irregular long waves of this study. The values of $\beta$ obtained for the irregular waves fall within the range outlined by Chen and Belcher (2000) based on values determined by Plant (1982) of approximately $\beta = 34 \pm 16$. The corresponding values of $\alpha_p$ for the irregular long waves versus $c/U_*$ are shown in Figure 2.8. The values of $\alpha_p$ for the irregular waves of this study are low compared to the values used for fit to the laboratory data obtained by Mitsuyasu (1966) ($\alpha_p = 80$) and Phillips and Banner ($\alpha_p =
160), and are therefore more representative of ocean conditions as opposed to the conditions in the laboratory. Chen and Belcher (2000) suggest that in the ocean, $\beta$ is small and because of the small slopes associated with ocean swell and imply that $\alpha_p \approx \beta$. The waves of this study are not representative of ocean swell, as discussed in Section 2.2, but are more representative of long wind waves which are typically steeper than ocean swell (Bailey, et al. 1991). The wind field conditions simulated in the W$^2$ however are more similar to ocean conditions during storm seas, as discussed in Section 2.3.3, due to large roughness values of $z_0 \sim 10^{-2}$ m. These results for $\alpha_p$ and $\beta$ suggest that the conditions of this study are representative of long waves coupled with strong wind fields ($U \sim 7 - 10$ m/s) in the ocean.

![Figure 2.7. Growth rate coefficient, $\beta$, versus wave age for all irregular waves.](image)

Figure 2.7. Growth rate coefficient, $\beta$, versus wave age for all irregular waves.
Figure 2.8. Atmospheric pressure coefficient, $\alpha_p$, calculated based on $\beta$ using Eq. 2.13 for the irregular waves versus wave age.

The results for the application of the CBM model to the monochromatic and irregular wave data of this study have been presented, as well as values of $\beta$ associated with the conditions of the wind field and the irregular wave environments. It was found that the model was able to accurately predict the reduction of wave energy with increasing steepness in the monochromatic wave environments, but not the irregular wave environments. In the latter case, using the atmospheric pressure coefficient ($\alpha_p$) and growth rate coefficient ($\beta$) as fit parameters, the model underestimates the energy transfer from the wind to development of wind waves and requires unrealistic values of $\alpha_p$ and $\beta$. Implications of these results will be discussed next, and a modified version of the CBM model will be presented that adapts the original model for use with irregular wave environments using the values of $\beta$ directly quantified.

2.7 Discussion

The goal of this study is to determine if the CBM for wind wave suppression on the surface of long waves is applicable to irregular waves. The CBM model has been applied here to monochromatic waves
simulated in the W\textsuperscript{2} Wind Wave Basin and the results show that it does well to predict the variation in the experimental energy ratio with increasing long wave steepness for these waves. The CBM was then applied to the irregular wave environments simulated in the W\textsuperscript{2}. The model results for the irregular waves underpredict the slope of the experimental energy ratios with increasing long wave steepness. An unreasonably high value of $\alpha_p$ (~300) is required for use in the model in order to fit the model results to the experimental energy ratios, which for the range of steepness values considered ($a_Lk_L \sim 0.03 – 0.15$) produce corresponding $\beta$ values between 69 – 264. For slow waves, $c/u_* < 15$, which is the regime within which the waves of this study fall, these $\beta$ values are much higher than values observed in both laboratory and ocean environments (Plant, 1982; Chen and Belcher, 2000; Pierson and Garcia, 2008). Belcher (1999) concluded that for slow waves, $\beta \approx 20$.

The reason for the discrepancy in the model’s applicability to monochromatic and irregular waves, is that the model is formulated to account for only the momentum of the long wave at a singular frequency, $\sigma_L$. This is shown namely in Eq. 2.7, where $\tau_L$ is based on the rate of change of momentum of the long wave at $\sigma_L$. While this formulation is accurate for monochromatic waves, whose wave energy exists at a single frequency, it is inapplicable to irregular wave environments due to the spread of wave energy over a range of frequencies. Therefore, direct application of the CBM model to irregular long waves underestimates the rate of change of $\frac{E_L+ww}{E_{ww}}$ with $a_Lk_L$, resulting in an overestimation of $\alpha_p$. In order to apply the model to irregular waves, the long waves must be treated as a superposition of monochromatic waves that occur at a wide range of frequencies. A modified version of the CBM model that invokes this principle is introduced next.

2.7.1 Modifying the CBM for Irregular Long Waves

In order to modify the CBM for direct application to irregular long waves, Eq. 7 must be modified to account for the momentum associated with each wave frequency in the wave spectrum. The modified expression for $\tau_L$ for irregular waves becomes
\[ \tau_L = \rho_w \int_0^N \left[ \gamma_{L1} \sigma_{L1} \Phi_{L1}(\sigma) + \gamma_{L2} \sigma_{L2} \Phi_{L2}(\sigma) + \cdots \gamma_{LN} \sigma_{LN} \Phi_{LN}(\sigma) \right] d\sigma, \tag{2.21} \]

where the integration is performed over the range of frequencies of the wave spectrum. To calculate this expression, the wavelet reconstruction method described by Torrence and Compo (1998) was used. The reconstruction method allows the original surface elevation record, \( \eta_L \), to be broken down into individual records, \( \eta_{Lj} \), based on the energy contained within each discrete frequency scale, \( j \), where \( j = 1, 2, \ldots N \), and \( N \) is the number of scales of the wavelet, and \( \sigma_{Lj} \) is the frequency associated with each scale (schematic representation in Figure 2.9).

Figure 2.9. Schematic representation of the wave spectrum decomposed into individual surface elevation records via the wavelet reconstruction method.

The terms within the bracket of Eq. 2.21 were quantified for each scale of the wavelet, \( j \), which are separated by a spacing, \( d_j = 0.1 \). For each \( \eta_{Lj}, \gamma_{Lj} \) was calculated by quantifying \( c_{Lj} \) for the waves of the record of scale \( j \). The \( u_* \) associated with each wind speed was used, and the \( \sigma_{Lj} \) value is quantified as the inverse of the Fourier period, \( T_j \), times \( 2\pi \). The value of \( \beta \) used in Eq. 2.2 to calculate \( \gamma_{Lj} \) was the calculated values of \( \beta \) based on the properties of the entire long wave surface elevation record. To quantify \( \Phi_{Lj} \), the power at each scale was integrated over time, in order to provide an estimate of \( \Phi_{Lj} \) for each \( \sigma_{Lj} \).
The results of the modified CBM model described above, which accounts for the energy associated with range of frequencies of the wave spectrum, show better agreement with the trend of the experimental energy ratios with increasing wave steepness for the irregular waves for $a_i k_L = 0.03 – 0.09$ (Figure 2.10). Also shown in Figure 2.10 are the model curves generated by the original CBM using the values of $\beta$ (and thus $\alpha_p$) quantified directly from Eqs. 2.10-2.12 and 2.14-2.15. The model curves shown in the figure, correspond to the maximum and minimum $\alpha_p$ values obtained from CBM for the irregular waves. Physically, the maximum and minimum $\alpha_p$ values represent the model's prediction for the longest (in terms of wavelength) and shortest waves in the irregular data set, respectively. Despite the direct quantification of $\beta$ and $\alpha_p$, the model continues to under predict the slope of the trend in the experimental energy ratios. The original CBM results are shown in order to emphasize the improvement of the results of the modified model using the same values of $\beta$.

For the updated model, the values of $\frac{E_{L+ww}}{E_{ww}}$ are set to 0 when $\tau_L > \tau_{tot}$, because based on the momentum equation, $\tau_t = \tau_{tot} - \tau_L$, a value of $\tau_L > \tau_{tot}$ gives a negative estimate of $\tau_t$, which makes the value of $\frac{E_{L+ww}}{E_{ww}}$ based on Eq. 2.19 complex. Physically, $\tau_t$ cannot be negative since as the horizontal momentum associated with the wave-induced stress, $\tau_L$, increases, the portion of the total momentum associated with $\tau_t$ decreases if $\tau_{tot}$ is constant (Chen and Belcher, 2000), which prevails in the reduction of available momentum to grow the wind waves. If $\tau_L = \tau_{tot}$, the total momentum from the wind is absorbed by the long wave, and $\tau_t$ is completely reduced to 0, at which point the wind waves are completely suppressed by the long wave.
Figure 2.10. Results of the modified CBM model which accounts for the long wave stress associated with the range of frequencies of the irregular wave spectrum. Blue, open symbols represent the new model results, i.e. the model energy ratio as quantified from Eq. 2.18 using the new formulation for $\tau_L$ given in Eq. 2.21, and using $\tau_t = \tau_{tot} - \tau_L$ to calculate the turbulent stress. Black symbols represent the energy ratios directly from experimental data as shown in Figure 2.5. Model results for which $\frac{E_{L+ww}}{E_{ww}} = 0$ indicate a value of $\tau_L > \tau_{tot}$ or $\tau_t = 0$, representing total wind wave suppression. The red, dashed line and red, dot-dashed lines indicate the maximum and minimum limits of the model results, respectively, using Eq. 2.7 instead of Eq. 2.21 to quantify $\tau_L$ using directly quantified values of $\alpha_p$.

The modified model does well to predict the decrease in $\frac{E_{L+ww}}{E_{ww}}$ for the high wind speeds ($U = 7, 8.5, 10$ m/s) until $a_ik_L \geq 0.1$. To provide some explanation for this threshold, the modified model results, which are based on the measured $u_*$ and calculated $\beta$ values for each wind speed, $\tau_t$ and $\tau_L$ are investigated in relation to steepness, $a_ik_L$ (Figure 2.11). The $a_ik_L$ at which $\tau_t$ and $\tau_L$ are nearly equal is approximately 0.075 but this threshold increases slightly with increasing wind speed. Until the steepness of the long wave environment is greater than approximately 0.075, $\tau_t$, which is responsible for wind wave...
growth on the long waves, is the dominant fraction of the total surface stress. Once this transition \( a_l k_L \) value is reached, the long wave is steep enough to begin to receive the majority of the wind energy input, reducing the available \( \tau_t \) for wind wave growth, which results in the suppression of the wind waves. For steepness values above 0.075, the model is no longer predicting the transfer of wind energy to the long wave, but rather the suppression of the wind waves by the long waves, and therefore breaks down.

![Figure 2.11](image)

Figure 2.11. Modified model results for \( \tau_t \) and \( \tau_L \) as a function of \( a_L k_L \) for (a) \( U = 5.5 \) m/s, (b) \( U = 7 \) m/s, (c) \( U = 8.5 \) m/s, and (d) \( U = 10 \) m/s.

It has been shown that by modifying the CBM model to incorporate the growth of individual wave components and directly quantified values of \( \beta \), it can be used to predict the trend in the experimental energy ratios with increasing steepness. To physically understand the mechanisms contributing to the growth of the long wave, the individual contributions to \( \beta \) will be discussed next.

2.7.2 Analysis of the Growth Rate Coefficient

As discussed in Section 2.4.2, there are five physical mechanisms that contribute to \( \beta \) and ultimately cause the waves to grow. As noted by Belcher (1999), for the slow wave regime, i.e. \( c/u_* < 15 \), it is the
contribution from the shear stress on the inner region due to undulations on the water surface associated with non-separated sheltering, $\beta_{sz1}$, that will dominate the growth of the long waves under wind action. This claim is supported by the calculated components of $\beta$ for the irregular long waves of this study (Figure 2.12). In fact, the contribution from $\beta_{sz1}$ is nearly ten times larger than the contribution from the wave-induced surface stress component, $\beta_{\eta_s}$, which is believed to be the largest contributor to wave growth in the fast wave regime, i.e. $c/u_s > 25$ (Belcher, 1999).

Figure 2.12. Contributions to the growth rate coefficient, $\beta$, for all irregular wave cases. Green, red and blue symbols represent $U = 7$ m/s, $U = 8.5$ m/s and $U = 10$ m/s respectively.

The contribution from the varying surface velocity associated with the non-separated sheltering, $\beta_{szu}$, is negative because it accounts for a decrease in the pressure at the surface due to negative wave-induced stress over the inner region. This negative portion of the wave-induced stress is caused by the orbital motions of the waves (Belcher, 1999). The contribution from the varying surface velocity associated with the surface stress, $\beta_{us}$, is also negative, but for slow waves, this contribution to the growth coefficient is typically small, on the order of 1 (Cohen and Belcher, 1999). The terms $\beta_{szu}$ and $\beta_{us}$ therefore counteract
the contributions from the other terms, nearly cancelling the contributions from $\beta\eta_s$ and the contribution from changes in pressure in the inner region associated with non-separated sheltering, $\beta_{p_o}$.

This study has investigated the mechanisms that contribute the growth of the long wave by analyzing the contributions to $\beta$. A research question that has been posed by many studies but remains unanswered is, what physical mechanism is responsible for the suppression of wind waves on the surface of long waves? Although this study cannot draw conclusions that answer this question, it does support the conclusion drawn by Chen and Belcher (2000) that the momentum available to grow the wind waves is dependent on the amount of momentum absorbed by the long waves due to direct coupling between the long waves and the wind. This suggests that the suppression of the wind waves is a result of a reduced turbulent stress, which is proportional to the long-wave induced stress which increases with increasing $a_1k_L$. Additionally, Chen and Belcher (2000) mention that for application of the model (Eq. 2.18), the fetch law presented in Eq. 2.4, from which Eq. 2.18 is derived, must be replaced with a fetch law obtained from observational measurements in the ocean. To investigate the applicability of the modified model to open ocean conditions, the energy ratio was also derived from fetch laws presented in Komen et al. (1994) for open ocean sea states and produced results with negligible difference to Eq. 19 (not shown). The agreement between the energy ratio of Eq. 2.18 derived by Chen and Belcher (2000) and that derived from the fetch laws for open ocean conditions is contributed to the wind fields generated by the wind machine in the W$^2$ Basin, which compares well to measured conditions in the ocean (Toba and Ebuchi 1991).

2.7.3. Application to Ocean Waves

The irregular waves of this study correspond to a range of wave steepness values, $ak \sim 0.03 – 0.14$. The irregular wave environments were generated using a JONSWAP spectrum, because it has been shown to be representative of ocean wave environments (Sorenson 1993), with a JONSWAP gamma of 3.3, which is typical for design applications (Sorenson 1993). Both the wave and wind environments
simulated in this experiment are full scale. To demonstrate the applicability of the results of this study to real ocean environments, Figure 2.13 shows a map of significant wave height, adapted from Figure 3.3b in Hanley (2008), in the ocean with regional steepness values. To quantify the steepness values, the map of global wave phase speed in Figure 3.3c in Hanley (2008) was converted to wavelength using the relation between phase speed and wavelength for deep water waves. According to Sorenson (1993), for deep water waves

\[ c = \frac{gL}{\sqrt{2\pi}} \]  
(2.22)

where \( c \) is the wave peak phase speed, and therefore, by rearranging Eq. 2.22 the deep-water wavelength can be calculated as,

\[ L = \frac{c^22\pi}{g}. \]  
(2.23)

The global map of phase speed in Hanley (2008) was converted to wavelength using Eq. 2.23, which could then be used to estimate the regional wave steepness shown based on the global \( H_s \) values provided by the map in Figure 3.3a in Hanley (2008). To covert from steepness in terms of \( H_s/L \) to \( ak \), values of wave steepness were multiplied by \( \pi \),

\[ ak = \frac{H_s2\pi}{2L} = \frac{H_s\pi}{L}. \]
Figure 2.13. Regional wave steepness based on significant wave height in the ocean (adapted from Figure 3.3b in Hanley 2008). Significant wave height data calculated from the comprehensive ERA-40 data set covering 1958 – 2001 (Uppala et al. 2005). The wave steepness in terms of $ak$ is characterized regionally based on the phase speed map data in Hanley (2008). Arrows above steepness values denote the direction of increasing $ak$.

The map of regional $ak$ shown in Figure 2.13 suggests that the waves of this study ($ak \sim 0.03 – 0.14$) represent waves of typical steepness in the ocean. The modified CBM was shown to work well for waves within the range of $ak \sim 0.03 – 0.09$, which is typical of waves in the northern Atlantic, Pacific and Indian oceans according to Figure 2.10. Additionally, Figure 2.14, adapted from Hanley (2008), shows the global mean wind speed measured at an elevation of 10 m above the water surface. According to this figure, the wind speeds of this study are typically of the far northern regions of the Atlantic and Pacific oceans. The waves of this study are therefore most likely to emulate the conditions in these regions.
Figure 2.14. Global mean wind speed and direction measured at a height above the surface of $z = 10$ m (adapted from Figure 3.3a in Hanley 2008) calculated from the comprehensive ERA-40 data set covering 1958 – 2001 (Uppala et al. 2005). Wind speed is in m/s.

2.8 Conclusions

The Chen and Belcher (2000) wind wave suppression model was applied to a comprehensive data set of long monochromatic and irregular waves of varying properties ($H_s$ and $T_p$). The goal of this study is to investigate the use of the CBM for wind wave suppression on the surface of irregular waves. The model accurately predicts the variation in the energy ratio $\frac{E_{L+ww}}{E_{ww}}$ with increasing wave steepness, $\alpha_p k_L$, for monochromatic waves under wind speeds varying from $U = 7 – 10$ m/s, which contain wave energy at only a single frequency. When applied to the irregular wave environments, unreasonably large estimates of $\alpha_p$ are required to fit the CBM results to the trend of the experimental energy ratios with increasing $\alpha_p k_L$. The CBM was modified for application to irregular waves by considering a wide range of frequencies in the expression for the long wave-induced stress, as opposed to a single frequency.
consistent with monochromatic waves, in order to account for the growth of the individual wave components of an irregular sea state. When this modification is considered, the CBM model predicts the trend in $\frac{E_{L+ww}}{E_{ww}}$ with $\alpha_L k_L$ well using the directly quantified values of the atmospheric pressure coefficient ($\alpha_p \sim 25\text{-}50$).

This study further validates the CBM model for monochromatic waves in wind environments of elevated roughness $z_0$ ($z_0 \sim 10^{-2}$ m) and suggests a correction to the model to make it more applicable to realistic sea states that are better represented by irregular waves. Application of this modified model to field conditions or to irregular waves simulated in a laboratory of longer fetch could help to further validate this model and refine the conditions for which it is applicable. Given more observational data, improved thresholds, in terms of long wave steepness and wind speed, could be determined and used in wave models to identify sea states that are susceptible to the suppression of wind waves.
CHAPTER 3

TURBULENCE GENERATED BY IRREGULAR WAVES AND WIND

3.1 Chapter Abstract

Turbulence generated by wind and irregular waves was investigated through a laboratory experiment performed in the W² Laboratory at the Advanced Structures and Composites Center (ASCC) at the University of Maine. Turbulence is described in terms of turbulent kinetic energy (TKE) and friction velocity. It is shown that for the irregular waves of this study ($H_s = 0.25$ m, $T_p = 2.5$ s), the wind reduces TKE measured below the waves, by ‘flattening’ (reducing the wave steepness) the tallest waves, which reduces wave breaking. In addition to this finding, this study provides a guide for planning a laboratory investigation of turbulence by highlighting several theoretical limitations associated with common methods for indirect measurement of turbulence to aid in the planning of future laboratory experiments. In fact, several theoretical and experimental limitations were revealed that hindered the ability to quantify TKE dissipation, $\varepsilon$, from the data collected in this study. This chapter will detail how to avoid these issues in future experiments.
3.2 Introduction

An understanding of turbulent mixing, which drives vertical exchange flows, is essential for the prediction of pollutant fate and transport, and for the improvement of ocean circulation models (Wiles et al., 2016). In addition, the characterization of turbulence at site specific locations is becoming increasingly important with new developments in tidal energy converters and turbines (McMillan et al. 2016). Within the past two decades, there has been increased interest in the study of wind and wave-induced turbulence (Thais and Magnaudet, 1996; Denissenko et al., 2007; Huang and Qiao, 2010), as more attention is brought to the possibility that turbulence is generated in the upper ocean through more mechanisms than wave breaking alone (Diao and Qiao, 2010). Wind-induced shear on the water surface is one mechanism that has gained attention for its potential to induce turbulent motions in the upper layer of the ocean. Csanady (1979; 1984) proposed a shear layer model based on the idea that the application of a sudden wind stress on the water surface imparts momentum into the water column, causing the flow velocity profile below the water surface to deviate from the traditional wall layer layer structure.

While Csanady (1979; 1984) did bring to light the importance of accounting for the wind influence on the water surface in turbulent models, the role of surface waves on a wind-driven water surface was not addressed. Craig and Banner (1994) adopt the idea of a surface shear layer, suggested by Csanady (1974; 1984) and other researchers (Wu, 1975; Richman et al. 1987) and expand it to account for the influence of surface waves. They develop a model with solutions for the variation of turbulent kinetic energy (TKE), TKE dissipation, $\varepsilon$, and flow velocity, $u$, induced by surface waves and wind with depth. Several studies have since investigated wave-induced turbulence and found that the motions associated with the waves themselves, even when no breaking is observed, can induce significant turbulence. For example, Babanin and Haus (2009) investigated wave-induced turbulence in the absence of wind and other possible forcing mechanisms, under long monochromatic waves in the laboratory. They used a particle image velocimetry (PIV) system to obtain velocity measurements below the waves, then quantified $\varepsilon$ using Kolmogorov
theory (Babanin and Haus, 2009). In their study, no wave breaking was observed, but TKE dissipation, $\varepsilon$, was still fairly high, on the order of $10^{-3} \text{ m}^2 \text{s}^{-3}$.

Lai et al. (2018) more recently studied wave induced turbulence in the laboratory with the objective of investigating the influence of the wave’s stage of development, i.e. wind waves, swell, or mixed sea. This study drew several major conclusions. Amongst waves of similar characterization, they found larger TKE dissipation rates associated with waves of larger height and wavelength. In addition, they found enhanced rates of wave induced turbulence associated with wind waves as opposed to swell, and found that the characterization of the wave varied the dissipation rates even amongst waves of similar properties; wind waves induce more turbulence than swell waves of similar wave height (Lai et al, 2018). Similar to Babanin and Haus (2009), Lai et al. (2018) studied the turbulence below long monochromatic waves and used Kolmogorov theory to obtain estimates of $\varepsilon$. Instead of a PIV system, Lai et al. (2018) used an acoustic Doppler velocimeter (ADV) to obtain velocity measurements below the waves.

For wave fields of only a single frequency, such as those simulated in the studies by Babanin and Haus (2009) and Lai et al. (2018), it is fairly simple to differentiate between the motions associated with the waves and those associated with low frequency motions and turbulence. It becomes more difficult to discern turbulent motions from wave motions when wave motions exist over a wide range of frequencies, as is the case with real ocean sea states and irregular waves. For these wave environments, a robust method of wave-turbulence decomposition is required. In a recent study, Huang et al. (2018) introduced a new method for quantifying $\varepsilon$ from ADV data contaminated by surface waves. The method utilized an empirical mode decomposition (EMD) analysis that separates turbulence from the wave modes and white noise in the velocity record. This method is one of several that exist in literature for wave-turbulence decomposition for ADV data.
The appropriate method for surface wave removal varies depending on the type of instrument used to collect the velocity measurements. Acoustic Doppler current profilers (ADCPs) are becoming an increasingly popular choice for field measurements due to their ability to be deployed for long periods of time with little attention required throughout their sampling period (Nystrom et al. 2002). Wiles et al. (2006) introduced a novel method for quantifying $\varepsilon$ from ADCP measurements. These instruments however, like ADVs, are susceptible to velocity contamination by surface waves, which was not a consideration in the model of Wiles et al. (2006). Scannell et al. (2017) later expanded the novel structure function technique introduced by Wiles et al. (2006) in order to remove surface wave bias to ensure accurate estimates of $\varepsilon$. The method introduced by Scannell et al. (2017), like many other indirect methods for quantifying turbulence, relies on several theoretical and practical assumptions, including statistical isotropy, satisfaction of Taylor’s frozen field hypothesis, and the use of an adequate bin size to resolve turbulent motions.

Wave-turbulence decomposition is one of several difficulties that arise when attempting to study wind and wave-induced turbulence in a laboratory setting. For example, in many laboratories, even when a ‘sufficiently high Reynolds number’ is obtained, a threshold which in itself is ambiguous, turbulence may still be considered anisotropic (Pope, 2000). Statistical isotropy is one of the many underlying theoretical assumptions associated with Kolmogorov’s hypotheses, which are inherently the theory used to develop many of the methods that exist for indirect measurement of turbulence. According to Denissenko et al. (2007), a minimum basin width to depth ratio of one is required in order to achieve statistical isotropy and ensure the existence of a range of inertial scales. In addition to the required sizing of the tank, the wind-wave forcing should be contained within low frequencies.

In addition to measures of turbulence, such as $TKE$ and $\varepsilon$, it is often of interest to quantify turbulent shear stress, which is typically represented by the friction velocity, $u^*$. Turbulent shear stress is an important consideration when investigating turbulent flow because friction can induce turbulent mixing, which contributes to the overall momentum balance (Shaw and Trowbridge, 2001; Trowbridge, 1998).
There is often overlap between methods for quantifying the friction velocity and turbulence. For example, Stapleton and Huntley (1995) demonstrate the use of the inertial dissipation method (IDM) for estimation of $u^*$ at the bottom. This method has been used by several other studies (Parra et al., 2014; Walter et al., 2011) to quantify $\epsilon$, and relies on the same theoretical assumptions as many of the aforementioned methods, such as the assumption that Taylor’s hypothesis holds and that statistical isotropy can be assumed.

Several methods for indirect quantification of turbulence and friction velocity, and some of the associated theoretical assumptions, have been highlighted. The choice of an appropriate method is contingent on factors such as the type of instrumentation available for data collection, the experimental conditions (i.e. laboratory or field), and, to some degree, an understanding of the characteristics of the turbulence that will be measured. The objective of this Chapter is to provide a helpful guide for selection of a method to calculate turbulence in a laboratory when surface waves are present, demonstrate one of these methods, discuss how wind speed and the spread of the wave spectrum effect estimates of turbulence and friction velocity and explain how the experiment could be improved to better estimate turbulence properties in the future. The remainder of this Chapter will be broken up with Section 2 which summarizes two methods for quantifying turbulence from velocity measurements collected by an acoustic Doppler current profiler (ADCP) and an acoustic Doppler velocimeter (ADV), two commonly used instruments for indirect measurement of turbulence in the field. Section 3.2 will include the details of the underlying assumptions associated with each method and describe the wave-turbulence decomposition techniques that can be used when the removal of wave motions from the data is necessary. Section 3.3 will present a case study of an experiment performed in the W2 Laboratory in order to demonstrate how the experiment was limited by the assumptions of the methods for quantifying turbulence described in Section 3.2. Section 3.4 will discuss the results from the case study and Section 3.5 will draw conclusions from these results, and present a procedure for planning a laboratory investigation of turbulence below waves.
3.3 Methods for Quantifying Turbulence from ADCPs and ADVs

A variety of methods exist for the measurement and calculation of turbulence in the presence of surface waves in field and laboratory settings using ADVs and ADCPs. An assessment of the experimental conditions prior to the selection of an instrument and a method for calculation is essential to ensure the ability to quantify turbulence. Often, when buoyancy fluxes can be considered negligible based on the experimental conditions, production of TKE can be assumed to equal the dissipation of TKE, and an estimate of $\varepsilon$ can be used to describe turbulence (Monismith, 2010). If the theoretical assumptions associated with the instrument-dependent methods available to quantify $\varepsilon$ are not satisfied, then $TKE$ can be used to describe turbulence instead.

This section will provide a summary of methodologies for quantifying turbulence, describe wave-turbulence decomposition techniques and outline the limitations associated with two methods for quantifying $\varepsilon$ in the presence of surface waves. The first method is a modified structure function method applicable to data collected by an ADCP. The second method is the IDM applicable to data collected by an ADV.

3.3.1 Modified Structure Function for ADCPs

A four-beam ADCP can be a useful tool for collecting velocity measurements due to ease of deployment and the ability of ADCPs to be deployed for long time periods (Nystrom et al., 2002). When surface waves are present, it is necessary to first remove wave bias before quantifying turbulence. Scannell et al. (2017) introduced a method for calculating $\varepsilon$ from ADCP data that inherently separates the wave bias from estimates of $\varepsilon$. The derivation of the modified structure function method (MSFM) can be found in Scannell et al. (2017) but an overview of the method will be given here followed by a summary of the underlying theoretical limitations associated with the method.

3.3.1.1 Summary of the MSFM
To estimate $\varepsilon$, the MSFM uses a least squares fit of second order structure function, $D_{LL}(x, r)$, measurements quantified for each bin of data collected by the ADCP against measures of $r^{2/3}$, where $r$ is the separation distance between estimates. The structure function, $D_{LL}(x, r)$, is of the form,

$$D_{LL}(x, r) = < [u'(x + r) - u'(x)]^2 >$$  \hspace{1cm} (3.1)

where $u'$ is the along-beam fluctuating velocity and $x$ is the distance along the beam where $u'$ is calculated. In Eq. 3.1 the angle brackets are used to indicate a mean over a statistically valid sampling period. The separation distance, $r$, is set to be at least the length of two bins. Measures of $u'$ are quantified for each bin as the instantaneous velocity minus the mean over a sampling period of several minutes. The following equation is then used for a least squares fit of estimates of $D_{LL}(x, r)$ against the values of $r^{2/3}$

$$D_{LL}(x, r) = A_0 + A_1 r^{2/3} + A_2 (r^{2/3})^3$$  \hspace{1cm} (3.2)

from which the coefficients of $A_0$, $A_1$ and $A_2$ can be determined. The coefficient $A_0$ represents noise, while the coefficients $A_1$ and $A_2$ represent turbulent and wave motions respectively. Once a value of the coefficient $A_1$ is determined from the fit, the relation,

$$\varepsilon = \left( \frac{A_1}{c_2} \right)^{3/2}$$  \hspace{1cm} (3.3)

can be used to estimate $\varepsilon$.

### 3.3.1.2 Theoretical Assumptions of the Structure Function

The modified structure function method (MSFM) introduced by Scannell et al. (2017) is derived from Kolmogorov's similarity hypotheses, and therefore involves several theoretical assumptions. The first is that Taylor’s frozen field hypothesis can be assumed valid. The use of Taylor's hypothesis depends on the presence of a mean advective flow, $U$, that is much larger than the fluctuating velocities, $u'$, (i.e. $U \gg u'$) where the velocity, $u$, is composed of the mean velocity and it's fluctuating components, $u = U + u'$ (Lien and D’Asaro, 2006; Lai et al., 2018). The threshold for $U \gg u'$ is not well defined in literature. Lien and D’Asaro (2006) provide the guideline that $U$ must be 10 times greater than $u'$ for Taylor’s hypothesis to be satisfied. For grid turbulence, satisfaction of this criteria is enough to validate the use of Taylor’s hypothesis, but for free shear flows Taylor’s hypothesis may still not be applicable (Pope, 2000).
Additionally, for ADCP data collected in the presence of surface waves, removal of surface wave bias according to the MSFM takes place after satisfaction of Taylor’s hypothesis has been assumed. For this scenario, an understanding of the experimental conditions at the site location or in the laboratory is essential in order to determine if the assumption of a frozen field is appropriate.

The second assumption of the MSFM is the existence of an inertial subrange, which is defined by Kolmogorov’s five-thirds law, as a range over which the energy contained within turbulent eddies is conserved over the range between large energy containing eddies and small, dissipative scales (Pope, 2000). For an inertial subrange to exist, the Reynolds number, \(Re\), of the flow must be sufficiently high (though the limit for a ‘sufficiently high’ Reynolds number is not well defined in literature) in order to ensure statistically isotropy holds for the small-scale \(l \ll L\) motions associated with turbulence (Pope, 2000).

The MSFM also requires that the turbulence be isotropic, meaning that the properties of the flow field do not vary depending on the direction in which it is measured (Pope, 2000). Lu and Lueck (1999) developed an expression for an anisotropic ratio, \(\alpha\), which can be used to determine if the flow is anisotropic. Their equation for \(\alpha\) is,

\[
\alpha = \frac{\bar{w}^2}{\bar{u}^2 + \bar{v}^2} \tag{3.4}
\]

where a prime denotes the fluctuating component of the velocity and \(u\), \(v\), and \(w\) are the velocity components in the along-channel, across-channel, and vertical directions respectively. Since the \(w\) component of the velocity cannot be resolved for a four-beam ADCP, which does not have a beam oriented along the instrument’s axis, Eq. 3.4 can only be used to determine the degree of anisotropy when using a five-beam ADCP (Lu and Lueck, 1999). For a four-beam ADCP, determination of isotropy is more difficult. For some ADCPs, isotropy is accounted for through the error velocity recorded by the instrument, and measurements for which isotropy is not satisfied are marked and can be removed in post-processing (Bender and DiMarco, 2009). In the field, isotropy is often assumed based on knowledge of
the experimental conditions. A sufficiently high Reynolds number is typically enough evidence to suggest that isotropy exists. In the laboratory however, it is less appropriate to assume isotropic flow (Pope, 2000). As mentioned in Section 3.1, a minimum basin width to depth ratio of one is necessary for statistical isotropy to be assumed in laboratory flows (Denissenko et al., 2007). Finally, the MSFM also assumes that the wave motions and the turbulent motions do not interact.

3.3.1.2 Instrument and Data Collection Considerations for the Structure Function Method

Before conducting an experiment in the field or laboratory with the objective of quantifying $\varepsilon$ via the MSFM there are several things to consider relative to the setup of, and collection of data by, an ADCP. To be able to resolve turbulence, the bin size of the ADCP must be set to a length smaller than the length scale of the largest eddies, $L$ (Nystrom et al., 2002). There are two ways to ensure that the selected ADCP bin size is appropriate, however both rely on some prior knowledge of the experimental conditions. For these reasons, some experience with turbulent measurements can be helpful. The first method for determining the appropriate bin size is by use of the definitions of Kolmogorov’s microscales if an estimate of the Reynolds number is available,

$$\frac{\eta}{L} \sim Re^{-3/4}. \quad (3.5)$$

Eq. 3.5 can be rearranged used to estimate $L$ if an understanding of $\eta$, the length scale of the eddies at the dissipative scales, and the $Re$ for the experimental conditions are known. Alternatively, the $L$ can be estimated directly from the velocity spectrum by identifying the limits of the inertial subrange by fit to a slope of -5/3 (Pope, 2000).

The theory used to derive the structure function is also based on the assumption that the flow is statistically steady over the duration of the experiment. For long sampling intervals (on the order of hours to days) the data set is divided into intervals over which statistical properties of the flow must be invariable, typically on the order of several minutes for ADCP data (Scannell et al., 2017). It is important to consider
the experimental conditions in order to ensure that this condition can be met prior to deployment of the ADCP.

Several theoretical assumptions and practical limitations associated with indirect measurement of turbulence with an ADCP by use of the MSFM have been discussed. While an ADCP is a useful instrument to collect data over a range of depths, it is susceptible to contamination by surface waves. There are few methods detailed in literature to date for the removal of wave bias from measurements (Scannell et al., 2017; Whipple et al., 2005). Many more methods that exist for the removal of surface wave bias from measurements collected by an ADV. In the next section, a commonly used method for the indirect measurement of turbulence with an ADV will be discussed, followed by a summary of several methods that exist for the removal of surface wave bias from ADV measurements.

3.3.2 Inertial Dissipation Method for ADVs

An ADV is a commonly used device for indirect measurement of oceanic turbulence, due to the ease of assembly and deployment, and to the availability of a variety of methods by which $\varepsilon$ can be estimated. One common method for estimating $\varepsilon$ from ADV velocity measurements is the IDM. A summary of the IDM will be given next, followed by a discussion of the theoretical assumptions used in the derivation of the method.

3.3.2.1 Summary of the IDM

The IDM relies on the existence of the inertial subrange, which is defined by a slope of the velocity shear spectrum that fits Kolmogorov’s -5/3 law:

$$\Phi(k) = \alpha \varepsilon^{2/3} k^{-5/3},$$

(3.6)

where $\alpha$ is the empirical Kolmogorov constant, typically set to 0.71 (Huang et al., 2018), and $k$ is the radian wavenumber (Huang et al., 2018). To apply to ADV data, which is recorded in frequency space, Taylor’s frozen field hypothesis is applied in order to convert to frequency space,
\[ \Phi(f) = \alpha \varepsilon^{2/3} f^{-5/3} \left( \frac{U}{2\pi} \right)^{2/3}, \]  

(3.7)

where \( f \) is frequency and \( U \) is the mean horizontal advection velocity (Huang et al., 2018). To determine \( \varepsilon \), Eq. 3.7 can be rearranged as

\[ \varepsilon = \frac{2\pi}{U} \alpha^{-3/2} \langle f^{5/2} \phi^{3/2}(f) \rangle \]  

(3.8)

where angle brackets represent an average and the values of \( f \) and \( \Phi(f) \) are selected to be within the inertial subrange (Liu and Wei, 2007). The satisfaction of Taylor’s frozen field hypothesis is essential for the use of the IDM for velocity data, but it is just one of several limitations to the IDM. These limitations will be discussed next.

3.3.2.1 Theoretical Assumptions

The IDM method, like many indirect methods for measuring turbulence, is derived from theory that only holds in specific conditions. Many of the underlying assumptions of the IDM method are similar to those of the MSFM detailed in Section 2.1.1. Taylor’s hypothesis is one of the theories from which the IDM method is derived, and for it to be valid, the same criterion for a frozen field as detailed in Section 2.1.1 applies. For ADV measurements biased by surface waves however, it is necessary to first remove the wave bias before checking if the criterion is satisfied if the wave orbital velocities are larger than the mean flow velocity. In addition to the existence of a mean flow, the flow must also be statistically steady over the sampling period. Often, for long deployments (on the order of hours or days), 10-minute sampling periods are usually considered statistically steady (Parra et al., 2014) and therefore the mean must not vary across these sampling periods. In addition to Taylor’s hypothesis, Kolmogorov’s hypotheses are used to derive the IDM, and therefore a sufficiently high Reynolds number is required for the inertial subrange to exist. If no inertial subrange can be identified from the power spectrum, this suggests that the flow is laminar and the IDM method cannot be applied. Finally, \( \varepsilon \) must be proportional
to TKE production, \( P \), which based on the TKE budget, is contingent on negligible buoyancy fluxes, which is not always the case in stratified environments (Monismith, 2010).

As mentioned previously, in an environment where surface waves are present, it is important to first remove the wave orbital velocities from the velocity measurements collected by the ADV, otherwise the scales at which turbulence and wave motions exist may overlap and bias the turbulence estimates. Unlike the MSFM for ADCPs, which inherently separates wave motions from turbulent motions, the IDM does not account for wave bias. Therefore, an additional wave-turbulence decomposition technique is required, prior to quantification of \( \varepsilon \) using the IDM. There are several methods that have emerged in literature recently that can be used for this purpose. Three of these methods will be introduced next.

3.3.2.2. Removal of Surface Wave Bias

A velocity record collected by an ADV in the presence of surface waves can be considered a superposition of the contribution of the mean flow, \( U \), fluctuating velocities associated with turbulent motions, \( u' \), and wave orbital velocities, \( \tilde{u} \), as

\[
u = U + u' + \tilde{u},
\]  

(3.9)

where \( u \) is the instantaneous velocity (Bricker and Monismith, 2007). To be able to accurately quantify \( \varepsilon \) from the instantaneous velocity record, it is necessary to first separate the wave motions from the mean flow and turbulence. There are several methods that have been proposed for this purpose over the past couple of decades (Shaw and Trowbridge, 2001; Bricker and Monismith, 2007; Huang et al., 1998). The choice of an appropriate method depends on the number of instruments used and the assumptions that can be made based on the experimental conditions. For example, the wave-turbulence decomposition method introduced in Shaw and Trowbridge (2001) requires the use of two ADV sensors, while the Bricker and Monismith (2007) method requires only one sensor, but is derived from the assumption that the wave motions and turbulence do not interact in a wave dominant field.
3.3.2.2.1 Covariance Method

Shaw and Trowbridge (2001) introduce a unique wave-turbulence decomposition method that is useful for removing wave bias from turbulent shear stress estimates when two sensors are available for measurement. The method requires that the two sensors spaced in the vertical dimension, a distance equal to five times the height above the bottom of the lower sensor. The method expands the wave-turbulence decomposition method proposed by Trowbridge (1998), by reducing the wave motions to the measured covariances between the velocity components of the two sensors, which can provide an estimate of turbulent shear stress with significantly less wave bias. Additionally, a filtering technique is implemented that improves the accuracy of the covariance technique when there is a significant contribution to the flow field from wave energy. The method provides an estimate of the wave-induced velocity as

\[ \bar{U}_{(1)}(t) = \int_{-\infty}^{\infty} \hat{h}(t') U_{(2)}(t - t') \, dt' \quad (3.10) \]

where the subscripts \( (1) \) and \( (2) \) represent the location of the lower and upper sensor respectively, \( t \) is time, \( U \) is the horizontal velocity, and \( \hat{h} \) is the filter weight. The filter weight is estimated as

\[ \hat{h} = (A^T A)^{-1} A^T U_{(1)} \quad (3.11) \]

where the superscript, \( T \) represents the transpose of the matrix and \( A \) is a matrix of dimensions, \( M \times N \), where \( M \) and \( N \) are the number of data points and filter weights respectively (for the derivation of Eq. 3.11 see Shaw and Trowbridge 2001). The covariance of the velocity measurements between the two probes can then be computed from the filtered records, \( \bar{U}_{(1)} \) and \( \bar{W}_{(1)} \). The covariance from the filtered records, \( cov[\bar{U}_{(1)}, W] \) and \( cov[U, \bar{W}_{(1)}] \) can then be decomposed into the turbulent fluctuations and the wave components, based on the equation,

\[ cov(U, W) = \bar{u}' \bar{w}' + \bar{u} \bar{w} + \theta [ \bar{w}'^2 - \bar{u}'^2 ] + \theta [ \bar{w}^2 - \bar{u}^2 ] \], \quad (3.12) \]

where \( \bar{u}' \bar{w}' \) is the turbulent covariance, i.e. the parameter of interest, \( \bar{u} \bar{w} \) is the wave covariance, and \( \theta \) is the angle of instrument rotation. In Eq. 3.12, an overbar represents a mean, and the turbulent and wave
components are represented with prime and a tilde respectively. The full methodology of the covariance method can be found in Shaw and Trowbridge (2001).

This method relies on the assumption that the degree of statistical similarity (coherency scale) of the turbulent motions is much smaller than that of the wave motions. Shaw and Trowbridge (2001) show that this method works well to remove high frequency surface wave motions from turbulent shear stress estimates but is less efficient for the removal of low frequency, internal wave motions.

3.3.2.2.2 Phase Method

Bricker and Monismith (2007) introduce a method for wave-turbulence decomposition, called the Phase Method, which requires only one sensor and assumes that there is no interaction between the turbulence and the wave motions, meaning that in a wave dominant field, the turbulent motions are not stretched as a result of the strain induced by the wave field. The Phase Method interpolates estimates of turbulence within the range of the peak wave frequencies the lie within the inertial subrange. This is performed by quantifying the wave stresses in this region based on the phase lag between the $u$ and $w$ velocity components. Bricker and Monismith (2007) derived the following expression for the wave stresses,

$$
\overline{uw} = \sum_{j=\omega_{p,\text{wave}}} |U_j||W_j| \cos (\angle W_j - \angle U_j)
$$

(3.13)

Where the subscript, $j = \omega_{p,\text{wave}}$ represents summation over the frequency scales beneath the wave peak, $W_j$ and $U_j$ are the Fourier coefficients, $\overline{W_j}$ and $\overline{U_j}$ are the wave-related Fourier coefficients, and $\angle W_j$ and $\angle U_j$ are the phase of the components of $w$ and $u$ respectively and are of the form,

$$
\angle W_j = \arctan \left[ \frac{\text{Im}(W_j)}{\text{Real}(W_j)} \right]
$$

(3.14)

$$
\angle U_j = \arctan \left[ \frac{\text{Im}(U_j)}{\text{Real}(U_j)} \right].
$$

(3.15)
The Fourier coefficients of $W_j$ and $U_j$ then take the form,

$$W_j = |W_j| e^{i\omega W_j} \quad (3.16)$$

$$U_j = |U_j| e^{i\omega U_j}. \quad (3.17)$$

The wave-related Fourier coefficients, $\widetilde{W}_j$ and $\widetilde{U}_j$, are quantified as the difference between the raw values of $W_j$ and $U_j$ and the turbulent part, $W_j'$ and $U_j'$, which are determined by linear interpolation of the power below the wave peak in the power spectral density (PSD) of the raw velocities (see Bricker and Monismith, 2007 for further details on the linear interpolation). Once the wave stresses, $\overline{\overline{u}\overline{w}}$, have been determined, they can be removed from the measured stress, $\overline{\overline{u}\overline{w}}$, leaving only the turbulent stress,

$$\overline{\overline{u}'\overline{w}'} = \overline{\overline{u}\overline{w}} - \overline{\overline{u}\overline{w}}. \quad (3.18)$$

Several studies have found the Phase Method is efficient for removing wave bias from velocity estimates prior to estimating $\varepsilon$ via the IDM (Walter et al., 2014; Stocking et al., 2016; MacVean et al., 2014). The accuracy of this method is reduced however in a wave dominant field if the turbulent motions exist at the same scales as the waves, which can lead to a straining of the turbulent motions. The full methodology of the Phase Method is described in Bricker and Monismith (2007).

3.3.2.2.3 Empirical Mode Decomposition (EMD) Method

A relatively new technique for the removal of wave bias from ADV velocity measurements is the empirical mode decomposition (EMD) method, which breaks down the velocity record into intrinsic mode functions (IMFs), which represents the different oscillatory modes within the velocity signal (Huang et al., 1998). The modes associated with the wave motions can be identified, based on the peak frequencies of the individual signals, and removed, leaving only the modes associated with high frequency noise and turbulence, low frequency motions, and white noise (Bian et al., 2020). The EMD is fairly simple to implement and alleviates the need for two velocity sensors to be used. Since this method
was introduced by Huang et al. (1998) for the purpose of non-stationary and nonlinear time series analysis, it has been used in several studies for the purpose of wave-turbulence decomposition for ADV data (Qiao et al., 2018; Bian et al., 2020; Huang et al., 2018).

The first step in the EMD method is to quantify the fluctuating velocities for each velocity component, $u$, $v$, and $w$ as

$$
\begin{align*}
  u' &= u - \bar{u} \\
  v' &= v - \bar{v} \\
  w' &= w - \bar{w}
\end{align*}
$$

where overbars represent a mean and the prime denotes the fluctuating component. The fluctuating components $u'$, $v'$ and $w'$ will, when surface waves are present, contain both wave, turbulent, and low frequency motions. To separate these motions, an EMD is performed to decompose the fluctuating velocities into IMFs that represent random stationary processes and the residual motions as

$$
\begin{align*}
  u' &= u'_{IMF_1} + u'_{IMF_2} + \cdots + u'_{IMF_n} + u'_{Residual} \\
  v' &= v'_{IMF_1} + v'_{IMF_2} + \cdots + v'_{IMF_n} + v'_{Residual} \\
  w' &= w'_{IMF_1} + w'_{IMF_2} + \cdots + w'_{IMF_n} + w'_{Residual}
\end{align*}
$$

The IMFs associated with wave motions can then be determined by comparison of the frequency peak of the IMFs with the peak wave frequency. The remaining IMFs are then summed to recreate a fluctuating velocity record for each component free of surface wave bias (Bian et al., 2020).

A method for indirect estimation of turbulence in the presence of surface waves was described for both an ADCP and an ADV. In the next section, a case study will be presented in which turbulence below waves simulated in the $W^2$ was investigated, in order to demonstrate the importance of determining the appropriate method for quantifying turbulence in a laboratory prior to performing the experiment.
3.4 Case Study

3.4.1 Experimental Set-Up and Instrumentation

The test was initially set up to collect velocity data with two Nortek ADVs and one Workhorse Sentinel 1200 kHz ADCP and to collect surface elevation data with 13 Akamina AWP-24-3 capacitive style wave height gauges with accuracy up to 1.35 mm. The ADVs were positioned on the instrumentation array adjacent to wave probes A and D (Figure 2), with the transducer heads located approximately 0.6 m below the water surface. During the testing however, unanticipated instrument noise interference occurred between the two ADVs and could not be resolved. Therefore, the ADV located adjacent to Probe A (Figure 2) was turned off, and data was only collected using the ADV adjacent to Probe D. Without two velocity sensors, the covariance method described in Section 2.2.2.1 was no longer an option for wave-turbulence decomposition in post-processing.

The wave gauges were calibrated and zeroed to a mean water level before the data collection began and were set to sample at a rate of 32 Hz. All instrumentation on the array in Figure 2, with the exception of ADV 1 located adjacent to Probe A, was used to collect data during the test cases discussed in this section. Data was collected for two array positions. The working ADV (ADV 2) was located at a horizontal distance, $x = 16.54$ m from the wave maker at array position 1 and $x = 20.54$ m at array position 2. The ADCP was bottom mounted at $x = 15.5$ m from the wave maker and was not moved during the testing. In order to improve the resolution of the ADCP, cloud seeding was used in the water. The ADCP bin size was set to 0.25 m, in order to collect multiple velocity measurements throughout the water column despite the shallow water depth of 4.5 m, however it was determined during post-processing that this bin size was larger than the length scale of the largest eddies, as described in Section 3.3.1.2, and therefore turbulent motions could not be resolved. This inhibited the ability of the modified structure function method to be used to quantify turbulence. Additionally, during post-processing of the velocity data collected by the ADV, it was determined that the mean flow of the laboratory basin was much smaller than that of the fluctuating velocities, and so the system was essentially treated as one with
no mean flow. In the absence of a mean flow, Taylor’s hypothesis is not valid, and so the conversion from frequency space to wavenumber space, which is integral for the IDM method, was not possible. Therefore, turbulence could only be quantified in terms of $TKE$ and $u^*$ using the ADV data.

In addition to the limitations that have resulted due to the theoretical assumptions of the methods for indirect measurement of turbulence outlined in Section 3.2, there were several other factors limiting this study. The ADV transducer head was located approximately 0.6 m under the water surface, while the $H_s$ of the waves in this study was only 0.25 m. Placement of the ADV transducer head closer to the water surface would have allowed for the collection of elevated wave-induced turbulence due to closer proximity to the surface layer within which wave motions had stronger influence. Additionally, noise interference between the wave gauges and the working ADV limited the number of cases for which data could be collected. These cases will be outlined in Section 3.4.2.
Figure 3.1. (a) Array positions in the basin with respect to the wave maker and wind jet. The anemometer staff on the array is used as the horizontal reference for the array position to the wave maker. Dimensions shown are in meters. Side view of the instrumentation array. Measurements shown are vertical distances from the mean water level to the anemometer locations. All dimensions are in meters.

The wave absorbing beach shown in Figure 3.1a is expected to dissipate approximately 90% of the wave energy (Ouslett, 1986). Additionally, a settling time of three minutes was also allotted between wave runs in order to reduce the presence of wave components of the previous test in the next wave record. Test video was also collected for the experiment.

3.4.2 Wave Environments

Like the wave environments of Chapter 2, the wave cases used for this analysis were characterized by a JONSWAP spectrum. The peak enhancement factor, $\gamma$, was varied for these cases, changing the shape of the wave spectrum, however the significant wave height and peak wave period is the same for these cases. Varying the $\gamma$ values allowed for investigation of the spread of the wave spectrum, i.e. a larger value of $\gamma$ corresponds to a smaller range of wave frequencies, with more waves at frequencies close to
the spectral peak. Setting $\gamma$ to 1.0 reduces the wave spectral shape to that of a Pierson-Moskowitz type spectrum, which is representative of fully developed seas. A $\gamma$ of 3.3 is typically used for laboratory investigation of irregular seas and for design (Sorenson, 1993). Setting $\gamma$ equal to 100 narrows the spectral bandwidth and increases the peak of the spectral energy, creating wave envelopes with more gradual slopes. A sea with a $\gamma$ of 100 is idealized.

**Table 3.1.** Wave parameters used to characterize the JONSWAP for the wave environments.

<table>
<thead>
<tr>
<th>ID</th>
<th>$H_s$ (m)</th>
<th>$T_p$ (s)</th>
<th>$\gamma$</th>
<th>$U$ (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2a</td>
<td>0.25</td>
<td>2.5</td>
<td>1.0</td>
<td>0, 10</td>
</tr>
<tr>
<td>2b</td>
<td>0.25</td>
<td>2.5</td>
<td>3.3</td>
<td>0, 10</td>
</tr>
<tr>
<td>2c</td>
<td>0.25</td>
<td>2.5</td>
<td>100</td>
<td>0, 10</td>
</tr>
</tbody>
</table>

The test cases in Table 3.1 were run with no wind, and with a wind speed, $U$, of 10 m/s, providing a total of 12 test cases for analysis.

This study was limited by the availability of data for which velocity measurements were collected by the ADV. Due to noise interference with the wave gauges, the ADVs were only used to collect data for the wave cases listed in Table 3.1. Although the properties of the wave cases ($H_s$ and $T_p$) did not vary, the varying $\gamma$ of the wave environments allowed for the investigation of the influence of spectral spread on measured values of $TKE$ and $u_*$. Additionally, it is of interest to investigate the trend in the wave steepness under wind with distance along the basin for each wave environment of varying $\gamma$ in order to determine the influence of the spectral spread on wave shape. To do this, a wave shape analysis will be performed on each test case.

### 3.4.3 Data Analysis

Prior to analysis, velocity data collected by the ADV was preprocessed using the despiking method outlined by Goring and Nikora (2002). Similar to the test cases in Chapter 1, 20 minutes of data was collected for each wave environment in Table 3.1.
3.4.3.1. Removal of Surface Waves

In order to investigate the wind and wave induced turbulence, it is necessary to separate the energy associated with the wave orbital velocities from the turbulent portion of the spectra. When two velocity sensors are available, the covariance method described in Section 3.3.2.2.1 can be used to separate the wave orbital velocities from the turbulence. Because only one ADV was used to collect data during this study, the Covariance Method cannot be used. Additionally, during data processing, it was determined that the peak frequencies of the wave motions overlapped with the frequencies of the turbulent motions, which reduces the accuracy of the Phase Method described in Section 3.3.2.2.2. In the absence of a second velocity sensor, and when the frequency of turbulence is near the frequency of the wave motions, the EMD method described in Section 2.2.2.3 is a useful tool. The EMD method was therefore used for wave-turbulence decomposition in this case study. An EMD analysis was performed on the fluctuating part of each velocity component, $u'$, $v'$, $w'$, for each test case. The EMD analysis decomposed each test record into 10 IMFs and a residual component (Figure 3.2). A spectral analysis was then performed on each individual mode, in order to obtain the peak frequency of the mode (Figure 3.3). A Fast Fourier transform (FFT) of 4096 points and the wave gauge sampling interval of 0.0313 s was used for the spectral analysis.
Figure 3.2. Example of the original record of horizontal velocity fluctuations and the IMFs and residual resulting from the EMD.
Based on the peak frequency of the IMFs, the modes corresponding to wave motions could be identified. The IMFs that were associated with white noise could be identified by the characteristic flat tail of the spectrum in the high frequency region ($f > 10^{0.4}$ Hz) (Bian et al., 2020). Low frequency motions could be identified by a low frequency spectral peak, in this case a peak at frequencies lower than that of the waves (Bian et al., 2020). Once the modes associated with the wave motions were identified, the
record was reconstructed by superimposing all other modes and excluding the wave modes. For all test cases, only one mode corresponded to high frequency energy which could be associated with turbulence. From the spectral analysis of this mode, it was also clear that noise was captured in this IMF as well, which could be identified by the flat tail at frequencies greater than approximately $f = 10^{0.4} \text{ Hz}$ (Bian et al., 2020).

3.4.3.2. Quantifying $TKE$ and $u^*$

Using the reconstructed records for the fluctuating velocity components, $TKE$ and $u^*$ could be quantified. According to Pope (2000), $TKE$ can be quantified from the velocity components as

$$TKE = \frac{1}{2} (\overline{(u')^2} + (v')^2 + (w')^2) \quad (3.25)$$

where the overbar indicates a time average and the prime denotes the fluctuating component of the velocity. Several methods exist for the quantification of friction velocity from ADV measurements, but many of them rely on the satisfaction of Taylor’s hypothesis (Stapleton and Huntley, 1995; Salehi and Storm, 2012). Since Taylor’s hypothesis is not satisfied for the laboratory flows of this study, the filtered fluctuating velocity components were quantized using Reynolds stresses, following the equation from Walter et al. (2011),

$$u_*^2 = -\sqrt{(-\overline{u'w'})^2 + (-\overline{v'w'})^2}. \quad (3.26)$$

3.4.3.3 Wave Shape Analysis

To investigate the variation in steepness, $ak$, of the waves as they evolve along the basin, a zero-crossing analysis was performed for all surface elevation records collected by the wave probes for each test case. Prior to the zero-crossing analysis performed in this study, the original surface elevation record collected by each wave probe was separated into a high frequency record and a low frequency record using the reconstruction method introduced by Torrence and Compo (1998) (example of reconstruction method in Figure A.2). As discussed in Chapter 2, wind waves were observed on the surface of the long
waves. To investigate the variation in the steepness of the long waves under the action of wind, it was of interest to first remove the high frequency wind waves from the surface elevation records. The high frequency record contained the energy associated with the wind waves, identified from the wavelet transform, while the low frequency record contained the energy associated with the long wave. A zero-crossing analysis was then performed on both of these records for each test case considered in this study. A modified version of the code WaveZeroCrossingFun from the OCEANLYZ Ocean Wave Analyzing Toolbox was used to perform the zero-crossing analysis on the high and low frequency records (Karimpour, 2017). The code isolates the wave height, $H$, of each individual wave in a surface elevation record, from which the wave amplitude, $a$, was estimated as $1/2H$. The code also estimates the wave period, $T$, of each individual wave, which was used to quantify individual wavelength, $L$. From these measures of $a$ and $L$, the mean wave steepness, $ak$, could be quantified for the low frequency surface elevation record.

3.5 Results

3.5.1. $TKE$ and $u_*$

The results of $TKE$ and $u_*$ for all wave cases are presented in Table 3.2. There is little trend observed in the estimates of $TKE$ and $u_*$ for the different $\gamma$ values. The only notable trend is consistently higher values of $TKE$ measured for the case of $\gamma = 100$ compared to the other cases measured at the same location and under the same wind condition. This is likely a result of the enhanced spectral peak of the case of $\gamma = 100$, meaning that for this wave environment, there are a larger number of waves with heights and energy close to that of the peak wave frequency. Because there are a larger number of waves with greater wave energy in the wave trains of this environment, it is expected that the increased $u_*$ estimates are due to enhanced wave-turbulence interactions.

It is evident that for all cases, apart from case 2b at position 2, that $TKE$ decreases when the waves are under the action of wind. This is hypothesized to be a result of the wind ‘flattening’ the waves, i.e.
decreasing wave steepness, \( ak \), and reducing wave breaking, which injects momentum into the water column and increases \( TKE \). This hypothesis is also supported by the trend observed in \( u^* \) between the two array positions: it is observed that \( u^* \) decreases from position 1 to position 2 for the cases under wind action. This is hypothesized to be a result of increased fetch distance, which allows time for the wind field to reach full development, thus increasing the potential for the wind to flatten the waves.

**Table 3.2.** Estimates of \( TKE \) and \( u^* \).

<table>
<thead>
<tr>
<th></th>
<th>( TKE ) ( 10^{-3} ) ( (m^2/s^2) )</th>
<th>( u^* ) ( (cm/s) )</th>
<th>( TKE ) ( 10^{-3} ) ( (m^2/s^2) )</th>
<th>( u^* ) ( (cm/s) )</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Pos 2</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2a</td>
<td>1.0</td>
<td>1.46</td>
<td>0.60</td>
<td>1.25</td>
</tr>
<tr>
<td>2b</td>
<td>3.3</td>
<td>1.28</td>
<td>1.11</td>
<td>1.42</td>
</tr>
<tr>
<td>2c</td>
<td>100</td>
<td>2.59</td>
<td>0.64</td>
<td>1.90</td>
</tr>
<tr>
<td><strong>Pos 3</strong></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2a</td>
<td>1.0</td>
<td>2.81</td>
<td>1.53</td>
<td>1.61</td>
</tr>
<tr>
<td>2b</td>
<td>3.3</td>
<td>1.72</td>
<td>0.55</td>
<td>1.49</td>
</tr>
<tr>
<td>2c</td>
<td>100</td>
<td>2.31</td>
<td>0.79</td>
<td>2.01</td>
</tr>
</tbody>
</table>

To investigate why \( TKE \) decreases with increasing wind speed, the test video recorded during the experiment was analyzed. Snapshots of the test video for one instance in time, capturing a large wave at position 2 are shown for all wave cases in Figure 3.4, both with and without wind. It was observed that for the largest waves in the time series, breaking would occur for the waves when there was no wind influence, but little to no breaking was observed for the same waves under \( U = 10 \text{ m/s} \). This is believed to be more evidence of the wind flattening the waves. To further investigate this trend, the \( ak \) measured along the basin for each test case was quantified along the basin. These results are presented next.
Figure 3.4. Snapshots of test video displaying the influence of wind on the water surface at position 2. Snapshots were taken at the same instance in time for the cases of $U = 0$ m/s and their respective $U = 10$ m/s for each gamma value.

3.5.2. Variation in Long Wave Steepness along the Basin

In order to quantify how the wave steepness of the long waves is altered by the wind, the surface elevation records were separated into a low frequency record, $f \leq 2$ Hz, corresponding the wave energy of the long wave, and a high frequency record, $f > 2$ Hz corresponding to the energy associated with the wind waves using the wavelet reconstruction method by Torrence and Compo (1998). From the low frequency record, the mean wave steepness $ak$, was quantified by zero-crossing analysis as described in Section 3.3.3. The mean $ak$ quantified at each location along the basin for each test case under $U = 10$ m/s was normalized with the $U = 0$ m/s wind condition, in order to give a representation of the variation in
wave steepness due to wind alone. The results of this analysis are shown in Figure 3.5. A value of $a_{k_{10}}/a_{k_0} < 1$ indicates that the wave is less steep compared to the $U = 0$ m/s $a_k$ value. In Figure 3.5 it is evident that for all cases, 2a, 2b, and 2c, the normalized $a_k$ decreases linearly with fetch.

![Figure 3.5](image.png)

Figure 3.5. The normalized long wave steepness is shown as a function of distance along the basin non-dimensionalized by the calibrated wavelength of the long waves. On the y-axis is mean wave steepness for $U = 10$ m/s, $a_{k_{10}}$, normalized to the mean $a_k$ under $U = 0$ m/s, $a_{k_0}$, of each 20-minute surface elevation recorded by the center wave probes at both array positions. The trend line is a linear fit through the data for $\gamma = 3.3$. The coefficient of determination, $R^2$, and equation for the trend line are shown.

The results presented in this section will be discussed next, with a focus on the trend of decreasing TKE with increasing wind, and comparisons of the observed TKE to theory. A guide to planning a laboratory investigation of turbulence generated by wind and waves will then be introduced.

### 3.6 Discussion

This work aimed to describe the methodology for the indirect measurement of turbulence in the presence of waves in a laboratory and reveal the importance of a standard procedure for selecting the appropriate method to do so. To achieve this objective, a case study in which turbulent measurements were very limited by the theoretical assumptions of the available methods was presented. In the case
study, the influence of wind speed and spectral spread of the wave environment on measures of turbulence was investigated. Estimates of TKE and $u^*$ were presented for three wave cases of varying $\gamma$ measured at two locations in the wave basin, with and without wind. The case study was limited by several factors, including the availability of data, the availability of instrumentation and the experimental conditions of the W2. This section will describe the most interesting finding of the case study, which is the trend of decreasing TKE with increasing wind speed. The TKE results will then be compared with theoretical models. This discussion will conclude with a brief outline for a procedure for planning a laboratory investigation of turbulence in order to aid in the planning of future experiments.

3.6.1 Observed Trend in TKE

In the results of the case study, it was observed that TKE decreases with increasing wind speed. In order for these results to have implications for future laboratory or field studies, more data would need to be collected in order to draw a firm conclusion, that steep waves of small period, $T_p \sim 2.5$ s, flatten under the influence of high winds, $U \sim 10$ m/s, and thus wave breaking is reduced which decreases the measured TKE below the waves. The measured TKE values below the waves will be compared with a theoretical model of wave-induced turbulence in the presence of wind next.

3.6.2. Comparison of Observations with Theory

Craig and Banner (1994) developed a model to describe turbulence in the surface layer of the ocean that is strongly influenced by the presence of surface waves. This model expands on wind-induced shear layer models, such as those proposed by Csanady (1979; 1984), to account for the deviation from the typical “law of the wall” profile that results from both wind-induced shear and wave motions in the surface layer. Craig and Banner (1994) proposed the following model for the variation in TKE with depth immediately below the wave-enhanced water surface,

$$TKE = u^*\alpha^{1/3}\left(\frac{3B}{\delta_q}\right)^{1/6}\left(\frac{z_0}{\epsilon_0x}\right)^{n/3}$$

(3.27)
where \( u^* \) is the water side wind induced friction velocity, \( \alpha = 100 \) is the wave energy factor, \( B = 16.6 \) and \( S_q = 0.2 \) are model constants, \( z_0 \) is the surface roughness length, and \( z \) is the depth below the water surface. As Craig and Banner (1994) point out, it is not well defined in literature how to obtain an appropriate measure of the surface roughness length. The Charnock relation,

\[
z_0 = a u^*_w / g
\]

(3.28)

where \( a \) is a constant and \( g \) is the acceleration due to gravity, is often used to obtain an estimate of \( z_0 \) using the measured \( u \) values. Several studies have used different values of \( a \) in this expression (Bourassa, 2000). Bye (1988) proposed a value of 1400 based on fit to observed velocity profile below the water surface. This value is adopted by Craig and Banner (1994) to validate their model. For the present study, a value of \( a = 250 \) is used to determine values of \( z_0 \) for use in the Craig and Banner (1994) model. The model results are shown in Figure 3.6. The single point estimates of \( TKE \) measured by the ADV at a depth, \( z = 0.6 \) m, are plotted in this Figure. The experimental measures of \( TKE \) for all cases appear to be well predicted by the model. This result suggests that the Craig and Banner (1994) model can a useful tool for predicting the depth of the wind-induced shear layer in the presence of surface waves in laboratory studies. Knowledge of the depth of the wind-induced shear layer in the presence of waves can be useful for the setup of laboratory experiments with the objective of estimating \( TKE \) in this region.
Figure 3.6. a). Comparison of measured $TKE$ at $z = 0.6$ m to the Craig and Banner (1994) model predictions. Circles indicate experimental measured of $TKE$. The lines represent the model results calculated using the values of $u^*_w$ and $z_0$ for the respective test cases. b) Zoomed representation of a).

3.6.3 Improving the Experiment in the W$^2$

The initial objective of the experiment performed in the W$^2$ was to quantify $\varepsilon$ beneath the irregular waves. This experiment was seriously limited however by the absence of a mean flow in the basin. Without a mean flow, Taylor’s frozen field hypothesis could not be used to convert the velocity measurements from frequency space to wavenumber space, and the assumptions behind the MSFM for the ADCP data could not be met. In order to improve this experiment, a mean flow would need to be simulated in the basin. Additionally, this experiment could have been improved by the collection of data by a second sensor, which would have allowed for the covariance method to be used for the wave-turbulence decomposition of the ADV data. Pre-tests in which the ADVs were used to collect data together before beginning the experiment should be performed in order to ensure noise interference did not occur. The second sensor would need to be located some distance from the bottom that allows for a vertical spacing from the existing ADV of five times the height above the bottom of said sensor. To resolve more turbulent motions induced by the waves, the existing sensor should be moved closer to the
water surface, a distance only slightly larger than the height of the largest wave trough (for example, in this case since the amplitude of the waves was approximately 0.125 m, the ADV could have been located a distance \( z = 0.2 \) m from the mean water level). Figure 3.7 shows the improved setup of the ADV sensors.

![Diagram](image)

Figure 3.7. Improved set up of the ADVs on the instrumentation array. The variable, \( d \), denotes an arbitrary depth.

In order to be able to use the ADCP to measure turbulence indirectly in this experiment, a bin size much smaller than 0.25 m is required to be able to resolve the turbulent motions. The bin size must be smaller than the length scale of the largest eddies, as described in Section 3.3.1.2. For the case study described in this work, the length scale of the largest eddy was estimated to be on the order of 0.02 m, however this estimate was made based on the upper limit of the inertial subrange, as described in Section 3.3.1.2. To determine the limits of the inertial subrange, the velocity spectrum was converted from frequency space to wavenumber space assuming Taylor’s hypothesis held, and could be used, which is
not the case in this study. In order to determine the appropriate bin size to be used, a mean flow is required in order to make Taylor’s hypothesis valid. The velocity spectrum could then be converted from frequency space to wavenumber space and upper limit of the inertial subrange could be determined and used to estimate the length scale of the largest eddy, following the method described in Section 3.3.1.2.

In order to avoid the unforeseen complications faced in this case study in future laboratory investigations of turbulence, a procedure for selecting the appropriate method for quantifying turbulence and removing surface wave bias was developed. This procedure is presented next.

3.6.4 Procedure for Planning an Experiment

In Section 3.2 two methods for the indirect measurement of turbulence, by ADCP or by ADV, were described. The experiment performed in this study and the discussion of the limitations of that experiment detailed in the previous section place emphasis on the need for procedure for planning a laboratory experiment. Figure 3.8 provides a flow chart for the decisions that need to be made when planning a laboratory investigation of turbulence generated by wind and waves. Although there are many other methods that exist in literature for the estimation of turbulence below waves, Figure 3.8 provides a good starting point for selecting the appropriate method, depending on desired objective, available instrumentation, and the experimental conditions.
Figure 3.8. Flow chart for determination of the appropriate method for quantifying turbulence below waves in a laboratory setting.
3.7 Conclusions

The goal of this study was to provide a procedure for determining the appropriate method for indirect measurement of turbulence in a laboratory. To demonstrate the importance of such a procedure, a case study was presented in which turbulence generated by irregular waves and wind in a laboratory setting was investigated. Several limitations of the case study were revealed due to the theoretical assumptions of the available methodologies for quantifying $\varepsilon$, which inhibited the ability to obtain estimates of $\varepsilon$ for the data set. Instead, $TKE$ was measured and was found to decrease with increasing wind speed, due to the flattening (decreasing $ak$) of the waves by the wind. In order to conclude that this trend could in fact prevail in real ocean environments, more data collected under irregular waves is needed. The major outcome of this chapter is a guide for determining the appropriate method for quantifying turbulence below waves in the laboratory, which will aid in the planning of future experiments.
CHAPTER 4

CONCLUSIONS

The goal of this thesis was to investigate the influence of wind on irregular wave fields in a laboratory environment, focusing on wind-wave development and suppression and turbulence. This objective was met through the analysis of a comprehensive data set collected in the W² laboratory at the UMaine Advanced Structures and Composites Center. Wind velocity, current velocity and surface elevation data collected in various test campaigns in the W² Laboratory were used to reach the goal of this study by addressing two research objectives. The first is to validate an existing model for the prediction of wind wave suppression on the surface of monochromatic long waves, and to expand said model to apply to irregular sea states, which are more representative of real ocean environments. The second is to provide a review of applicable methodologies for quantifying turbulence in the laboratory environment, present a case study in the W² laboratory to quantify the wave and wind-driven turbulent kinetic energy (TKE), and present an improved test campaign setup that would optimize turbulence measurements in the W² Laboratory.

To meet the first objective of this thesis, the applicability of the Chen and Belcher (2000) model (CBM) for wind wave suppression was first validated with monochromatic wave data collected in the W² Laboratory. The model was then shown to fail when applied to irregular wave data. This is because the equation for the long wave growth rate and the long wave-induced shear stress used in the model only accounted for a single wave frequency, which is characteristic of a monochromatic wave, but not of representative of an irregular wave environment, which contains multiple wave frequencies. The CBM was then modified by considering a wide range of frequencies in the expression for the long wave-induced stress to account for the growth of the individual wave components of an irregular sea state. When this modification was adapted, the results of the CBM for prediction of wind wave suppression for the irregular waves was significantly improved.
There have been many studies over the past several decades that have investigated the phenomenon of wind wave suppression by long waves. The CBM was the first to provide some theory to predict the occurrence of this phenomenon, based on the steepness of the long wave. The CBM however, was limited by its validation by only monochromatic waves, which are idealized, and hence cannot accurately represent a real ocean environment. The results of Chapter 2 of this thesis show that the CBM model can be made more applicable to ocean environments by modifying the major equation for the long wave shear stress to account for the range of wave frequencies that are present in an irregular sea state. This thesis has shown that through this modification, the CBM can be used for the prediction of wind wave suppression in the field, which is an important phenomenon for wave models to capture. If wind wave suppression is not accurately modeled, this can lead to overestimations of the drag coefficient and sea surface roughness, both of which are controlled by the presence of high frequency wind waves. Overestimation of these parameters could reduce the accuracy of the model source term, $S_m$. Further validation of the modified CBM with field data would benefit the modeling community by making the model directly applicable to ocean wave data, and by improving estimations of the parameters encompassing the model source term $S_m$.

The second objective of this thesis was to investigate turbulence generated by irregular waves and wind in a controlled laboratory setting. This objective was met by the analysis of velocity data collected during the experiments in the W² Laboratory, which revealed lower values of $TKE$ under waves in the presence of wind, compared to the waves with no wind influence. The trend in $TKE$ was attributed to the ability of the wind to ‘flatten’ the waves (decrease the wave steepness), which reduces wave breaking, which reduces the input of momentum into the water column and thus the $TKE$. Several limitations of the data set were revealed, which inhibited the use of many existing methods for the estimation of $\epsilon$. A major outcome of the second research objective is a procedure for determining the appropriate method for quantifying $\epsilon$ based on the objective of the experiment, which will aid in the planning of future laboratory investigations of turbulence below waves. This work has shown that without the presence of a mean flow,
which inhibits the use of Taylor’s frozen field hypothesis, it is not possible to use the methods outlined in Chapter 3 to quantify $\varepsilon$. This work has also shown that although investigations of turbulence below water waves are becoming more popular, the procedure for determining the appropriate method for quantifying $\varepsilon$ remains ambiguous. There are several theoretical assumptions that exist for many of the methods for estimating $\varepsilon$ that must be satisfied, and it is important to understand these limitations prior to performing the experiment. The results of Chapter 3 provide a comprehensive guide for the determination of the appropriate method for quantifying $\varepsilon$ that can be used to aid in the planning of future laboratory investigations of wave-generated turbulence. The collection of more data under the irregular waves simulated in the W² in an experiment planned using the guide presented in this work would allow for the trends observed in $\text{TKE}$ to be confirmed, and could aid in the improvement of the model source term $S_{dis}$ by allowing for the measurement of $\varepsilon$ below irregular waves under wind in a controlled environment.
BIBLIOGRAPHY


Figure A.1. Wavelet of the entire wave record for wind only sea for $U = 10$ m/s collected by the first wave probe on the array at $X = 14.31$ m. A 50-second interval is shown to highlight the portion of significant energy within a confidence level of 90%.
Figure A.2. Example of the wavelet reconstruction technique. (a) Original surface elevation record, (b) low frequency record \((f \leq 2 \text{ Hz})\) corresponding to the long waves, (c) high frequency record \((f > 2 \text{ Hz})\) corresponding to the wind waves, and (d) the entire reconstructed record.
BIOGRAPHY OF THE AUTHOR

Taylor Bailey was born in Waterville, Maine on August 17th, 1996. In spring of 2014 she graduated from Erskine Academy and began attendance at the University of Maine in the fall. She graduated summa cum laude from the Civil and Environmental Engineering program at UMaine in 2018. She remained at the UMaine in pursuit of her Master’s of Science degree in the Civil and Environmental Engineering program. Her expected graduation date from the Master’s program is August 2020. In the fall of 2020, she will continue her studies as a PhD student in the Civil and Environmental department at UMaine. Taylor is a candidate for the Master of Science degree in Civil and Environmental Engineering from the University of Maine in August of 2020.