<table>
<thead>
<tr>
<th>Title</th>
<th>Observations of turbulent dynamics in the ocean surface boundary layer</th>
</tr>
</thead>
<tbody>
<tr>
<td>Author(s)</td>
<td>Sutherland, Graigory John</td>
</tr>
<tr>
<td>Publication Date</td>
<td>2014-03-26</td>
</tr>
<tr>
<td>Item record</td>
<td><a href="http://hdl.handle.net/10379/4393">http://hdl.handle.net/10379/4393</a></td>
</tr>
</tbody>
</table>

Some rights reserved. For more information, please see the item record link above.
Observations of Turbulent Dynamics in the Ocean Surface Boundary Layer

A Dissertation Submitted in Accordance with the Requirements for the Degree of Doctor of Philosophy in the College of Science

by

Graigory John Sutherland

School of Physics
National University of Ireland, Galway

Supervisor: Dr. Brian Ward

April 2014
Contents

Abstract iii

Declarations vi

List of Figures vii

List of Tables xvi

Dedication xvii

Acknowledgements xix

1 Introduction 1

2 Theory and Literature Review 5
  2.1 Turbulence Theory ............................................. 5
    2.1.1 Navier-Stokes Equation .......................... 7
    2.1.2 Turbulent Kinetic Energy ............................. 7
    2.1.3 Turbulent Kinetic Energy in the Presence of Surface Waves ............................................. 8
    2.1.4 Dissipation Rate of Turbulent Kinetic Energy .... 12
  2.2 Microstructure Measurements ................................. 13
    2.2.1 Measuring Turbulence with Airfoils ................ 15
    2.2.2 Spectral Analysis ........................................ 16
  2.3 Upper Ocean Turbulence ...................................... 20
CONTENTS

2.3.1 Mixed Layer ........................................... 23
2.3.2 Surface Stress - Law of the Wall ....................... 27
2.3.3 Wave-Turbulence Interactions .......................... 28
2.3.4 Langmuir Turbulence ................................. 33
2.3.5 Surface Buoyancy Flux ............................... 36

3 Wave-Turbulence Scaling in the Ocean Mixed Layer 39
3.1 Introduction ............................................. 40
3.2 Measurements ............................................ 44
  3.2.1 Microstructure Measurements .......................... 45
  3.2.2 Meteorological Measurements .......................... 47
  3.2.3 Mixed Layer Depth .................................... 49
  3.2.4 Wave Measurements ................................... 50
  3.2.5 Upper Ocean Parameters ............................ 53
3.3 Discussion ............................................... 57
  3.3.1 Integrated Energy Flux ............................... 58
  3.3.2 Dissipation Rate Scaling - Law of the Wall ........... 60
  3.3.3 Wave Induced Turbulence ............................. 60
3.4 Summary ................................................. 63

4 Evaluating Langmuir Turbulence Parameterizations in the
  Ocean Surface Boundary Layer 65
4.1 Introduction ............................................. 66
4.2 Observations ............................................. 69
4.3 Evolution of the OSBL ................................... 74
  4.3.1 Evaluating the Regime Diagram of Belcher et al. (2012) 75
  4.3.2 Evaluating Different Regimes ......................... 76
4.4 Depth Dependence of $\epsilon$ in the OSBL ............... 78
  4.4.1 Wind-Wave Regime ................................... 80
  4.4.2 Convective Regime ................................... 82
4.5 Summary ................................................. 84
CONTENTS

5 Mixed and Mixing Layer Depths in the Ocean Surface Boundary Layer: Buoyancy-Driven Conditions 89
  5.1 Introduction ................................................. 90
  5.2 Observations ................................................. 93
  5.3 Results and Discussion ................................. 97
  5.4 Summary .................................................... 103

6 Conclusions and Future Work 107
Abstract

Turbulence and mixing processes are investigated in the ocean surface boundary layer (OSBL) using microstructure measurements from the Air-Sea Interaction Profiler (ASIP) a vertically rising, autonomous instrument equipped with a suite of sensors designed to study the dynamics and energetics of the OSBL. These are presented in conjunction with observations of the atmospheric forcing and surface gravity wave fields to test various hypotheses for the vertical profile of the dissipation rate of turbulent kinetic energy $\varepsilon$.

The influence of surface gravity waves were not restricted to the upper few metres, but in fact extend throughout the OSBL. This large depth of influence is consistent with the presence of Langmuir circulations, which are expected to fill the OSBL. Although we didn’t have the instrument to detect whether Langmuir cells formed, observations of $\varepsilon$ were found to be consistent with results from Large Eddy Simulations (LES) which included effects of Langmuir circulations. These are some of the first observations to confirm LES studies of Langmuir turbulence in the open ocean.

The accuracy of scaling depends on the definition for the OSBL. Improved agreement with LES results were found when the depth of the active mixing layer (XLD), determined by locating the depth at which $\varepsilon$ fell to a background dissipation rate, were used for the boundary layer depth rather than the depth of the mixed layer (MLD), which is the depth at which the density exceeds a threshold value relative to the near surface density. These results
show the importance of defining the OSBL as the region of active mixing and not just the region with a quasi-homogeneous density structure.

The variations of the MLD and XLD are investigated in the subtropical Atlantic for a buoyancy-driven regime with both the MLD and XLD responding with a predictable diurnal structure. Observations show the density threshold used to estimate the MLD can be adapted to obtain the XLD, but this has a clear variation as a function of the local time of day and hence surface buoyancy flux. Mean values for the density threshold of the XLD are consistent with experiments, which suggest larger density thresholds near the equator.
Declarations

The work of this thesis is based on research carried out in the Air-Sea Physics Lab, School of Physics, NUI Galway. No part of this thesis has been submitted elsewhere for any other degree or qualification and is all my own work unless referenced to the contrary in the text.

Copyright ©2014 by Graig Sutherland
The copyright of this thesis rests with the author. No quotations from it should be published without the authors prior written consent and information derived from it should be acknowledged.
List of Figures

2.1 Stirring without mixing of a passive tracer. The initial streamlines of the 2-D turbulent flow are shown on top. The flow deforms the checkerboard pattern into elongated filaments (adapted from Welander, 1955). ................................. 6

2.2 Schematic of the upper ocean adapted from Ardhuin and Jenkins (2006). The thin arrows denote wave velocities while the thick arrows denote the turbulent momentum flux from waves and turbulence. ........................................ 10

2.3 Schematic of an airfoil shear probe. ............................... 16

2.4 The effect of spatial averaging on the shear spectrum (adapted from Macoun and Lueck (2004)). ................................. 17

2.5 Normalized universal shear spectra. ............................... 18

2.6 Schematic diagram of forcing and dynamics in the OSBL. ................................. 21

2.7 Schematic of Langmuir circulations (adapted from Smith, 2001). 34

3.1 (a) The 200 m, 1000 m, and 2000 m depth contours are shown for the North Atlantic with the deployment location shown by the large black dot. Inset map (b) shows the ASIP profile locations (black dots) with the green dot being the first profile and the red dot showing the final profile. The ship locations at three hour intervals are shown by the black triangles. All times are in local mean time. ................................. 45
3.2 Sample shear spectrum (black dots) and modelled Nasmyth spectrum (dashed line) for the depth interval $-17.6m < z < -17.1m$ of the profile corresponding to Fig. 3.3. The lower and upper integration limits are denoted by the vertical dashed lines. ................................................................. 47

3.3 Spectrogram of a profile taken at 17:15 LMT (panel (e)). The depth and corresponding rise velocity are shown in panels (a) and (b). A uniform rise speed is adopted in the upper 10 m (solid line) to filter out wave effects in the pressure signal (dotted line). Panel (c) shows the raw shear signal in volts and the calculated dissipation rate is in (d). ............. 48

3.4 Comparison of (a) significant wave height and (b) zero-upcrossing period for data collected with the ultrasonic wave altimeter (solid line) and the ERA-Interim reanalysis of ECMWF (dots connected by dashed line). Time is in local mean time. .... 51

3.5 Time evolution of wave spectra over six hour time intervals. 52

3.6 (a) Pressure (Pa), relative humidity (RH), (b) wind speed at 15.5 m ($U_{15}$), wind direction (Dir, clockwise from North), (c) air temperature ($T_a$), sea surface temperature (SST), rainfall rate (Rain), (d) significant wave height (and mean crossing period are shipboard measurements from the R/V Knorr for the deployment location in Fig. 3.1. The shaded region corresponds to the time when ASIP was profiling. ....... 54

3.7 (a) Net radiative flux (Q), short wave (SW), sensible heat (SH), latent heat (LE), and net infrared (IR) radiation during the deployment. Positive is upwards out of the ocean. The net surface buoyancy flux (B) and the relative contributions from heat and salt and shown in (b). (c) shows the mixed layer depth (D) and the Monin-Obukhov length (L). ....... 55
3.8 (a) Temperature, (b) salinity, and (c) potential density as measured by the ASIP profiler. The solid black line denotes the mixed layer depth and the time has been corrected to local mean time. ......................................................... 56

3.9 (a) Turbulent Langmuir number, (b) buoyancy and wind stress forcing, (c) Brunt-Väisälä frequency $N^2$, (d) dissipation $\epsilon$ and dissipation normalized by the law of the wall (Eq. (3.5)) (d). The solid black line in panels (c-e) denote the mixed layer depth $D$ and the dashed line in (d) and (e) is the Monin-Obukhov length $L$. All times have been converted to local mean time. ......................................................... 57

3.10 Profiles of turbulent kinetic energy dissipation (a-c) for a particular wave spectra (d-f). Five successive profiles of $\epsilon$ taken over one hour are averaged vertically into 1 metre bins with the solid black line showing the mean and the grey shaded region the 95% confidence intervals determined using a bootstrap method. The depth dependence of $\epsilon$ is compared with Eq. (3.5) (blue line), scaling of Terray et al. (1996) (green line), and the wave scaling of Huang and Qiao (2010) (red line) using portions of the wave spectra along with Eq. (3.6) (dashed lines with colours matching the corresponding spectral region marked by I, and II in (d-f)). The red line denotes the sum of the wave scaling turbulence profiles from section I and II, i.e. the dashed orange line plus the dashed green line. The values of $H_s$ and $T_0$ for each wave spectra are computed for the entire spectra. The mixed layer depth is the black horizontal line denoted by $D$. ......................................................... 58
LIST OF FIGURES

3.11 Depth integrated dissipation rate in the OSBL as a function of wind speed at 10 metres. Panel (a) shows the integrated dissipation as a function of input wind energy while panel (b) shows this quantity as a function of local time. The individual profiles are averaged into 1 hour by 1 metre bins and the error bars represent the limits of the integrated 95% confident limits as determined with a bootstrap method. 61

4.1 Location of each of the four ASIP deployments, numbered 1 to 4, during the Knorr 11 campaign. 70

4.2 Time series of the collected meteorological data including a) radiative fluxes including the total (Q), shortwave (SW), sensible heat (SH), latent heat (LE) and infrared (IR), b) wind speed and direction, c) significant wave height ($H_s$), zero up-crossing period ($T_z$) and period of the peak of the spectrum ($T_p$), d) the Stokes drift ($u_{s0}$) divided by 10, and the convective velocity ($\omega_*$) and friction velocity ($u_*$) respectively, e) turbulent Langmuir number (La) and wave age ($C_p/U_{10}$), and e) the mixed layer ($h_\rho$) and mixing ($h_\epsilon$) layer depths. The periods where there are ocean microstructure measurements are shaded grey and wave measurements are only shown for when the ship was on station. The dashed red and blue lines in e) correspond to $La = 0.35$ and $C_p/U_{10} = 1.2$ respectively. 71
4.3 Measured values of $\epsilon$ at $z = h/2$ normalized by the dissipation of B12 ($\epsilon_B$) using the a) mixed layer depth $h_\rho$ and b) mixing layer depth $h_\epsilon$ for the turbulent length scale respectively. The black lines show the contours for $\log_{10} \epsilon_B$ from Eq. (4.4) with the heavy black lines denoting where a single forcing accounts for 90% of the total dissipation. The dashed black line shows the $h/L_L = 1$ threshold chosen to separate Buoyancy regime from the Wind-Waves regime. 

4.4 Histogram of $\log_{10} \epsilon/\epsilon_B$ from Fig. 4.3. The blue and red lines denote the Gaussian curve calculated from the mean $\mu$ and standard deviation $\sigma$ for the $h_\rho$ and $h_\epsilon$ cases respectively.

4.5 Dissipation values normalized by $u^3_*/z$ at depths normalized by the a) mixed and b) mixing layer depths respectively. The red dots denote unstable buoyancy forcing, the blue dots denote neutral buoyancy forcing, and the green dots denote stable buoyancy forcing. The solid line is the geometric mean calculated at depth intervals of 0.05z/h and the dashed line corresponds to $\epsilon = u^3_*/z$.

4.6 Dissipation values normalized by $u^2_* u_{s0}/h$ at depths using the a) mixed $h_\rho$ and b) mixing $h_\epsilon$ layer depths respectively. The red dots denote unstable buoyancy forcing, the blue dots denote neutral buoyancy forcing, and the green dots denote stable buoyancy forcing. The solid black line shows the geometric mean and the dashed black line shows the curve from the LES results of Grant and Belcher (2009) (their Fig. 5).
4.7 Dissipation values normalized by $u^3_*/z$ as a function of La for various depth intervals normalized by $h_\rho$. The colour denotes the relative ratio of $h_\rho/L_L$. Only values where $h_\rho/L_L < 1$ are used. The slope $m$ and $R^2$ statistic are written on each plot respectively. 80

4.8 Dissipation values normalized by $u^3_*/z$ as a function of La for various depth intervals normalized by $h_\epsilon$. The colour denotes the relative ratio of $h_\epsilon/L_L$. Only values where $h_\epsilon/L_L < 1$ are used. The slope $m$ and $R^2$ statistic are written on each plot respectively. 82

4.9 A summary of the fit $m$ (a) and $R^2$ (b) for the mixed (black) and mixing (red) layer depths. The Langmuir case where $m=2$ is shown by the vertical dotted line in a. 83

4.10 Dissipation values normalized by $u^3_*/z$ as a function of $h_\rho/L_L$ for various depth intervals in a convective regime where $h_\rho/L_L > 1$. Values for La are shown by the colour scale. The slope $m$ and $R^2$ statistic are written on each plot respectively. 84

4.11 Dissipation values normalized by $u^3_*/z$ as a function of $h_\rho/L_{mo}$ for various depth intervals in a convective regime where $h_\rho/L_{mo} > 3$. Values for La are shown by the colour scale. The slope $m$ and $R^2$ statistic are written on each plot respectively. 85

4.12 A summary of the fit $m$ (a) and $R^2$ (b) for the ratio of the mixed layer to the Langmuir stability length (black) and Monin-Obukhov length respectively. The classic convection case where $m=1$ is shown by the vertical dotted line in a. 86

5.1 ASIP tracks for the five deployments during the STRASSE experiment. All deployments are within a radius of 1°. The grid lines on inset map denote 20' intervals. 94
5.2  (a) Wind speeds measured at 10 m (orange line) and surface buoyancy flux (blue region), (b) the Monin-Obukhov length $L$ (brown), and $h_\sigma/L$ (green-blue) and also time-depth plots of (c) potential temperature $\theta$, (d) $\log_{10}$ of the buoyancy frequency squared $N^2$, and (e) $\log_{10}$ of the turbulent dissipation rate $\epsilon$. The green-blue shaded region in (b) shows the region where $-1 \leq h_\sigma/L \leq 1$. The black solid and dashed lines in (c-e) shows the mixed layer depth as calculated using a threshold of $\sigma_\theta = 0.03$ and $0.09$ kg m$^{-3}$ respectively. The grey solid and dashed lines in (c-e) show the depth where $\epsilon$ first falls to $10^{-9}$ and $10^{-8}$ m$^2$s$^{-3}$ respectively.

5.3  Observations and diurnal phase averages of (a) $B_0$ (blue) and $U_{10}$ (orange), (b) $L$ (brown) and $h_{\sigma 1}/L$ (blue-green), (c) mixed $h_{\sigma 1}$ (green) and mixing layer depths ($h_{\epsilon 1}$ (grey) and $h_{\epsilon 2}$ (red) respectively), (d) the ratio of the mixing to the mixed layer depth, and (e) the absolute difference between mixing and mixed layer depths for the two mixing layer depth definitions. The black dashed lines in (d-e) denote $h_{\epsilon} = h_\sigma$. Panels (f-h) show the histograms for the observations in (c-e) with the box plots showing the stats for the entire record.

5.4  Observations and diurnal phase averages of (a) $B_0$ (blue) and $U_{10}$ (orange), (b) $L$ (brown) and $h_{\sigma 2}/L$ (blue-green), (c) mixed $h_{\sigma 2}$ (green) and mixing layer depths ($h_{\epsilon 1}$ (grey) and $h_{\epsilon 2}$ (red) respectively), (d) the ratio of the mixing to the mixed layer depth, and (e) the absolute difference between mixing and mixed layer depths for the two mixing layer depth definitions. The black dashed lines in (d-e) denote $h_{\epsilon} = h_\sigma$. Panels (f-h) show the histograms for the observations in (c-e) with the box plots showing the stats for the entire record.
5.5  a the density difference from the reference level to $h_{e1}$ (grey) and $h_{e2}$ (red) along with the diurnal phase averaged mean and 95% confidence intervals. b shows the histogram over the entire period using 0.01 kg m$^{-3}$ bins. The shaded region in a shows the 95% confidence interval for the mean and the box plots in b are for the entire record. 

............. 101
List of Tables

2.1 Examples of mixed and mixing layer depth criteria. . . . . . 26
Dedication

This dissertation is lovingly dedicated to my wife Jessica and son Woodrow.
It wouldn’t have been possible without them.
Acknowledgements

First I must thank my supervisor, Brian Ward, for all his support and guidance over the course of this work. After the many travels and closely working together on four field campaigns we have forged a working relationship to which I am extremely grateful and would be lucky to find again in the future.

I also extend my gratitude to Kai Christensen from the Norwegian Meteorological Institute for all his time and assistance. The scholarship in this thesis has been greatly improved by our interesting talks with regards to surface gravity waves.

A benefit of so many field campaigns is the comraderie developed with colleagues and I am grateful for the company and support of my fellow students João de Almeida, Sebastian Landwehr, Niall O’Sullivan, Brian Scanlon, Anneke ten Doeschate, Leonie Esters and postdocs Xavier Sanchez-Martin, Adrian Callaghan, Danielle Wain, Marc Defossez, Chiara Uglietti, and Kieran Walesby.

I am grateful to the exceptional work of the scientific teams and crew of the Johan Hjort, R/V Knorr, N/O Thalassa and B/O Sarmiento de Gamboa. These were all memorable and successful expeditions due to the excellent people involved and so much of the data quality is tied in with their tireless work.

I must also acknowledge the role of the Air-Sea Interaction Profiler (ASIP) in this work, which has taught me first and foremost that there are no shortcuts in observational oceanography. Throughout the constant tweaking
and retweaking I have learned many lessons in patience and perseverance and this thesis is richer for it.

My family has always been a source of support and their help and contributions can not be overestimated. My mum, Kerri, who has taught me so much and has always been a source of inspiration. My brother, Steve, who has always been an inspiration of truth and integrity. And foremost I am grateful to my wife, Jessica, who has lived and breathed this thesis with me every step of the way for which I can not thank her enough. In addition, we have welcomed a little boy, Woodrow, into the world who is an endless source of joy and wonder who never ceases to amaze us.

This work was predominantly funded through a NSERC Postgraduate scholarship and I am grateful for their support. I am also grateful to the Research Council of Norway who also contributed to funding this work.
Chapter 1

Introduction

The ocean surface boundary layer (OSBL) plays a pivotal role in the physical, biological and chemical dynamics of oceans. The OSBL is the quasi-homogeneous upper portion of the ocean, which ranges in depth from a few metres to hundreds of metres, and is characterized by high levels of turbulence. The high levels of turbulence act to enhance diffusion and keep the OSBL well mixed. Turbulence is also an important factor in controlling the transfer and storage of heat, momentum and trace gases between the atmosphere and the ocean, in addition to the distribution of nutrients and plankton and other nonmotile aspects of the biological cycle. The OSBL is bounded on the top by the ocean surface and below by a sharp increase in the water density, which acts to strongly inhibit diffusion across this boundary into the deep ocean. This sharp increase in density acts to inhibit transport of heat, momentum, biological organisms, and chemical properties. Transport of these properties from the OSBL to the deeper ocean require relatively large levels of turbulence. Therefore, enhanced turbulence levels from surface forcing and the depth to which they affect dynamics are also important aspects to variations on climate scales.

The evolution of the OSBL is directly influenced by a combination of buoyancy flux and mechanical mixing at the air-sea interface. The
OSBL is forced by the surface wind stress $\tau$ and buoyancy flux $B_s$ and surface waves through wave breaking and through interactions between the Stokes drift $u_s$ and the turbulent component of the wind creating Langmuir Circulations (Langmuir, 1938). Surface wave breaking generates enhanced levels of turbulent dissipation within a relatively short distance, on the order of a significant wave height $O(H_s)$, from the surface. However, the effects of wave breaking appear to be confined to this near surface region and do not appear to affect processes deeper in the mixed layer (Terray et al., 1996; Sullivan et al., 2007). Below the wave breaking region, turbulence dynamics are controlled by the relative strength of the surface wind, buoyancy and waves via Langmuir circulations (Grant and Belcher, 2009; Belcher et al., 2012).

Boundary layer processes occur on scales ranging from centimetres to tens of metres and it is often impossible to include the detailed dynamics of small-scale turbulent motions while resolving the large-scale regional dynamics. Thus, these small-scale turbulent processes must be parameterized in terms of the large scale forcing. Often these parameterizations are developed through an understanding of the turbulent dynamics and verified with detailed observations of the macro forcing in conjunction with turbulent observations in the OSBL. Due to the difficulty in making measurements in the open ocean, most of the boundary layer theory comes from laboratory experiments with a rigid boundary or the bottom atmospheric layer (e.g. von Kármán, 1930; Wyngaard, 1973). This layer near a rigid surface is often referred to as the law of the wall and can be accurately parameterized from the surface stress, buoyancy flux and distance from the boundary (Monin and Yaglom, 1971). This universality, more commonly known as similarity theory (Foken, 2006), has been applied very successfully in the atmospheric boundary layer when the surface buoyancy force is destabilizing (Wyngaard and Coté, 1971), but relatively poorly under stable conditions (Mahrt, 2014).

In the OSBL, the surface gravity wave field contributes additional forcing
to the turbulent dynamics causing similarity theory to be a poor approximation in this region. Although there have been a few observational studies which have shown enhanced turbulence levels in the OSBL relative to the law of the wall (Kitaigorodskii et al., 1983; Anis and Moum, 1995; Terray et al., 1996; Drennan et al., 1996; D’Asaro, 2001), these studies have been unsuccessful in deriving a universal scaling function for oceanic conditions. However, none of these studies include measurements of the turbulent structure of the entire OSBL up to the ocean surface under a variety of forcing conditions and sea states, thus making the development of a universal function difficult. Indeed, most of the development of wave-induced turbulence developed through Langmuir Circulations have come from Large-Eddy Simulations (LES), which often don’t include effects such as wave breaking or convection (McWilliams et al., 1997; Li et al., 2005; Grant and Belcher, 2009).

This thesis presents observations of the dissipation rate of turbulent kinetic energy and investigates various scaling parameterizations with the bulk forcing from wind, waves and surface buoyancy flux. Such observations are imperative if improvements are to be made to the current parameterization schemes used by climate models of these small scale and highly dissipative processes. Turbulence measurements were made using the Air-Sea Interaction Profiler (ASIP), which is an unique, autonomous, upwardly rising profiler designed to measure water properties at sub-cm scales throughout the OSBL to the ocean surface. To the author’s knowledge this is the first study that presents measured profiles of $\epsilon$ to the ocean surface along with simultaneous measurements of wind, waves and surface buoyancy fluxes.

The outline for this thesis is as follows. Chapter 2 presents some of the general background and theoretical framework for this study. Next, are three chapters which represent articles written as independent studies of turbulent dynamics in the OSBL. The first article, presented in Chapter 3, describes observations during a unique case study of rapidly changing winds in the
Introduction

North Atlantic and is published in the journal *Ocean Science* (*Sutherland et al.*, 2013). Chapter 4 presents the second article, published in the *Journal of Geophysical Research* (*Sutherland et al.*, 2014), describing a systematic study of newly developed scaling laws in the OSBL with detailed observations of turbulence and surface forcing. The third article, (Chapter 5), investigates the observed mixed layer, which is easy to estimate from density profiles, with the region of active mixing, which is important for mixing parameterizations in the OSBL. The summary and conclusions for this thesis as well as suggestions for future work are presented in Chapter 6.
Chapter 2

Theory and Literature Review

The purpose of this chapter is to present a range of concepts, terms and assumptions constituting the framework for the presentation and discussion of the current work. Section 2.1 presents some of the general theory on the study of turbulence in fluids and goes through the general steps and assumptions on how to determine dissipation rates on turbulent scales in fluids. Section 2.2 outlines the general assumptions made in measuring turbulent dissipation in the ocean and the accuracy and limitations of these methods. Section 2.3 outlines the current understanding of turbulence in the OSBL.

2.1 Turbulence Theory

Turbulence is a ubiquitous phenomenon in all fluids. Turbulence is not so easy to define as it is not a material property or a state of a certain volume of fluid. Turbulence is a difficult subject to study as it is unpredictable in a deterministic way (i.e. chaotic) in that it is irreproducible in detail, but not random as statistically it does obey various laws. Although turbulence is chaotic and irreproducible there are certain aspects that are easier to observe. One defining element of turbulence, and the main reason why oceanographers are particularly interested in it, is that it enhances diffusion.
This not only applies to the diffusion of scalar properties such as temperature and salinity but also to momentum. This enhancement is due to vortex stretching, which effectively increases the surface area between two fluids. A visualization of this can be seen in the now famous checkerboard figure by Welander (1955), adapted here in Fig. 2.1.

Figure 2.1: Stirring without mixing of a passive tracer. The initial streamlines of the 2-D turbulent flow are shown on top. The flow deforms the checkerboard pattern into elongated filaments (adapted from Welander, 1955).

Although this is a simplified two dimensional case, it clearly shows the vortex stretching of fluid elements causing the elements to appear to mix
even when there is no diffusion. Depending on the level of turbulence, enhancement in effective diffusion can be several orders of magnitude greater than the molecular diffusion of the scalar quantity.

2.1.1 Navier-Stokes Equation

The exact equation of motion for a fluid is known as the Navier-Stokes equation named after the French engineer Claude-Louis Navier and the Irish mathematician George Gabriel Stokes and was first derived in the 19th century. The general form of the Navier-Stokes equation for a non-rotating frame is written as

\[ \frac{\partial \mathbf{u}}{\partial t} + \mathbf{u} \cdot \nabla \mathbf{u} = -\frac{1}{\rho} \nabla P + \nu \nabla^2 \mathbf{u}, \tag{2.1} \]

where \( \mathbf{u} \), \( \rho \), \( P \), and \( \nu \) are the velocity, density, pressure and kinematic viscosity respectively. The two terms on the left side of Eq. (2.1) are sometimes grouped together as \( \frac{D}{Dt} \), known as the material derivative, as it incorporates the local acceleration and advection. The two terms on the right are the pressure and viscous forces respectively. The advection term on the left and the viscous term on the right of Eq. (2.1), the non-linear advection and dissipation terms respectively, make the equation difficult to solve and gives turbulence many of its unique characteristics. These characteristics will be discussed in more detail later on.

2.1.2 Turbulent Kinetic Energy

As a starting point in turbulence analysis, it is assumed that the variables (e.g. velocity, pressure, etc.) can be decomposed into a mean and a fluctuating part, i.e.

\[ a = \bar{a} + a', \tag{2.2} \]

where \( a \) is the variable and \( \bar{a} \) and \( a' \) are the time-averaged component and turbulent component respectively. This technique is sometimes referred to as
a Reynolds decomposition named after the early pioneer in fluid mechanics research Osborne Reynolds. The components of Eq. (2.2) have the following mean values

\[ \bar{a} = \bar{\bar{a}} \]
\[ \overline{a'} = 0 \] (2.3)

where the over line denotes a representative time average. Assuming an incompressible Boussinesq flow, the continuity of the mean flow can be written as,

\[ \frac{\partial \bar{u}_i}{\partial x_i} = 0 \] (2.4)

where the \( i \) subscript denotes the cartesian coordinate in standard tensor notation. Multiplying Eq. (2.1) by \( u_i \), averaging, and using Eq. (2.2), Eq. (2.3) and Eq. (2.4) yields the equation for the turbulent kinetic energy (TKE), i.e.

\[ \frac{D}{Dt} \left( \frac{1}{2} \overline{u'_i u'_i} \right) = - \overline{u'_i u'_j \frac{\partial \bar{u}_i}{\partial x_j}} + \frac{g}{\rho_0} \overline{w' p'} - 2\nu \overline{e_{ij} e_{ij}} \\
- \frac{\partial}{\partial x_j} \left( \frac{1}{\rho_0} \overline{p' u'_j} + \frac{1}{2} \overline{u'_i u'_i u'_j} - 2\nu \overline{u'_i u'_j} \right) \] (2.5)

Equation (2.5) states that the rate of change of TKE is a function of (a) shear production, (b) buoyancy forces from mixing, (c) the dissipation which is often written as \( \epsilon \) and (d) the transport of turbulent quantities.

### 2.1.3 Turbulent Kinetic Energy in the Presence of Surface Waves

The shear production term, i.e. (a) from Eq. (2.5), represents the generation of TKE from the mean shear in the water column. This is the only term in Eq. (2.5) which always generates TKE as the buoyancy term (b) is a sink if there is mixing (due to the increase in potential energy when two fluids
Theory and Literature Review

of different densities mix) or stabilizing buoyancy flux at the surface and is only a source if there is a surface destabilizing buoyancy flux (e.g. cooling of the ocean surface at night), (c) is the loss of TKE to viscous forces and (d) is the transport term which only redistributes TKE and is neither a source nor a sink. This Eulerian description of the shear production is not ideal near the air-sea interface as the ocean surface is seldom flat.

The mean difference between the Eulerian and Lagrangian descriptions in the presence of surface gravity waves is referred to as the Stokes drift, first observed by Stokes (1847), and is a result of particles tending to “surf” further with the wave than against it. For a deepwater monochromatic wave, the Stokes drift velocity is given by

\[ u_s(z) = a^2 k \omega e^{-2k|z|} \]  

(2.6)

where \( a \) is the amplitude, \( k \) is the horizontal wavenumber and \( z \) is the depth. It is important to note that the Stokes drift velocity is a horizontal vector in the direction of wave propagation and that decreases with depth exponentially with an e-folding scaling of \( 2k \), which is a factor of 2 greater than the e-folding depth for the wave orbital velocities and pressure fluctuations. Therefore, the Stokes drift velocity is primarily confined to near the surface for wind-generated waves, although ocean swell with longer wavenumbers will not decrease as rapidly. Equation (2.6) is a good approximation if there is limited information with regards to the wave spectra available, but if the full directional wave spectra is available then the Stokes drift velocity may be calculated (Webb and Fox-Kemper, 2011) as

\[ u_s(z) = \frac{2}{g} \int_0^\infty \int_{-\pi}^{\pi} (\cos \theta, \sin \theta, 0) \omega^3 S(\omega, \theta) e^{-\frac{2\omega^2}{g} |z|} d\theta d\omega \]  

(2.7)

where \( S(\omega, \theta) \) is the directional wave spectra, \( \omega \) is the angular frequency, \( \theta \) is the angle of propagation, \( g \) is the acceleration due to gravity (9.81 m/s) and the dispersion relation for deepwater surface gravity waves of \( \omega^2 = gk \) is used. In the absence of a directional spectra, \( U_s \) may be calculated assuming the
wave spectra to be separable into wave direction and frequency components
and the waves are unidirectional (see Kenyon, 1969), as
\[ \mathbf{u}_s(z) = \frac{1}{\pi g} \int_0^\infty \omega^3 S(\omega) e^{-\frac{2\omega^2 |z|}{g}} d\omega, \] (2.8)
which is equivalent to Eq. (2.7) divided by 2\pi.

The shear near the surface often doesn’t align with the mean sea surface
and it is important to account for this misalignment. Figure 2.2 shows a
schematic of the upper ocean when surface gravity waves are present (figure
reproduced from Ardhuin and Jenkins, 2006). One approach is to estimate

Figure 2.2: Schematic of the upper ocean adapted from Ardhuin and Jenkins
(2006). The thin arrows denote wave velocities while the thick arrows denote
the turbulent momentum flux from waves and turbulence.

the shear stresses using the generalized Lagrangian mean (GLM) Andrews and
McIntyre (1978) (which will be denoted by an over bar and an \( L \) superscript),
which accounts for the undulation of the sea surface. From Eq. (2.5) the
mean shear production of TKE is given by
\[ P_s = u'_i u'_j \frac{\partial \bar{u}_i}{\partial x_j}. \] (2.9)
which assumes the turbulent motions and mean motions are uncorrelated.
Furthermore, if we assume that turbulent motions are also not correlated
with the phase of the wave, the shear GLM of the production term Eq. (2.9)
can be approximated as
\[ \bar{P}_s^L = u'_i u'_j L \frac{\partial \bar{u}_i}{\partial x_j}. \] (2.10)
Using the relationship of Andrews and McIntyre (1978) which relates the Eulerian mean \( \overline{\phi} \) of an arbitrary variable \( \phi \) with the GLM value of \( \overline{\phi}^L \), the GLM can be written, valid to second order in the wave slope, as

\[
\overline{\phi}^L = \overline{\phi} + \xi_j \frac{\partial \phi}{\partial x_j}
\]  

(2.11)

where \( \xi \) is the wave-induced particle displacement field. Using linear wave theory for a monochromatic wave of amplitude \( a \) and frequency \( \omega \) travelling in the x-axis and using the standard notation \( (x_1, x_2, x_3) = (x, y, z) \) and \( (u_1, u_2, u_3) = (u, v, w) \), the velocities due to the motion of the surface gravity wave are

\[
\tilde{u} = a\omega e^{-k|z|} \cos (kx - \omega t) 
\]  

(2.12)

\[
\tilde{v} = 0
\]  

(2.13)

\[
\tilde{w} = a\omega e^{-k|z|} \sin (kx - \omega t)
\]  

(2.14)

where the \( \tilde{\cdot} \) denotes the velocities from the monochromatic wave, \( t \) is time and \( k \) is the wavenumber which is related to the frequency by the dispersion relation

\[
\omega^2 = gk
\]  

(2.15)

where \( g = 9.81 \text{ m s}^{-2} \). Equation (2.15) is valid only for surface gravity waves where \( kH \gg 1 \), where \( H \) is the mean water depth, otherwise the full dispersion relation of \( \omega^2 = gk \tanh kH \) must be used which would modify the form of Eqs. (2.12–2.14). Equation (2.15) is valid for the vast majority of oceanic conditions and the form of the GLM will be limited to this general deep water case. Furthermore, the wave-induced displacements, to first order, are

\[
\xi_x = -ae^{-k|z|} \sin (kx - \omega t),
\]  

(2.16)

\[
\xi_y = 0 \text{ and}
\]  

(2.17)

\[
\xi_z = ae^{-k|z|} \cos (kx - \omega t).
\]  

(2.18)
Combining Eqs. (2.12–2.14) and Eqs. (2.16–2.18) and the definition of Eq. (2.11) to compute the GLM of term a in Eq. (2.5) yields

\[
\frac{\partial \tilde{u}}{\partial z} = \xi_x \frac{\partial^2 \tilde{u}}{\partial x \partial z} + \xi_z \frac{\partial^2 \tilde{u}}{\partial z^2} = a^2 k^2 \omega e^{-2k|z|} \quad (2.19)
\]

\[
\frac{\partial \tilde{w}}{\partial x} = \xi_z \frac{\partial^2 \tilde{w}}{\partial x \partial z} + \xi_x \frac{\partial^2 \tilde{w}}{\partial x^2} = a^2 k^2 \omega e^{-2k|z|} \quad (2.20)
\]

Equations (2.19–2.20) show that the mean wave-induced shears are each equal to half of the vertical shear in Eq. (2.6).

Substituting Eqs. (2.19–2.20) into Eq. (2.10) and using the definition of Eq. (2.11) it can be seen that there is an additional source of TKE from the mean wave induced shears

\[
P_s = -\bar{w} \bar{w}^L \frac{\partial u_s}{\partial z}. \quad (2.21)
\]

Adding Eq. (2.21) to Eq. (2.5) gives the TKE equation in the presence of surface waves as

\[
\frac{D}{Dt} \left( \frac{1}{2} u_i^u u_i^u \right) = -u_i^u u_j^u \frac{\partial \bar{u}_i}{\partial x_j} - u_i^u u_j^u \frac{\partial u_{si}}{\partial x_j} + \frac{g}{\rho_0} \frac{\bar{w} \bar{w}^L}{\rho} - \epsilon - \partial \frac{\partial}{\partial x_j} \left( \frac{1}{\rho_0} \bar{p} \bar{u}_j^u + \frac{1}{2} u_i^u u_i^u u_j^u - 2 \nu \bar{u}_i^u \bar{e}_{ij} \right). \quad (2.22)
\]

The wave-induced motions create an additional shear term due to correlations between the wave shear and particle displacements. This result assumes that the turbulent properties are not correlated with the phase of the wave and that turbulent quantities are identical in both the Lagrangian and Eulerian framework, i.e. \( u_i^u u_j^L = \bar{u}_i^u \bar{u}_j \).

### 2.1.4 Dissipation Rate of Turbulent Kinetic Energy

The dissipation rate of turbulent kinetic energy is defined from term c in Eq. (2.5), as

\[
\epsilon = 2 \nu \bar{e}_{ij} \bar{e}_{ij}, \quad (2.23)
\]
where $\epsilon$ is the common notation for the dissipation rate of turbulent kinetic energy and $e_{ij}$ is the strain rate tensor

$$e_{ij} \equiv \frac{1}{2} \left( \frac{\partial u'_i}{\partial x_j} + \frac{\partial u'_j}{\partial x_i} \right).$$

(2.24)

The dissipation can be written more explicitly using Eq. (2.23) and Eq. (2.24) giving

$$\epsilon = \frac{\nu}{2} \sum_{i=1}^{3} \sum_{j=1}^{3} \left( \frac{\partial u'_i}{\partial x_j} + \frac{\partial u'_j}{\partial x_i} \right) \left( \frac{\partial u'_i}{\partial x_j} + \frac{\partial u'_j}{\partial x_i} \right),$$

(2.25)

which would yield 36 unique terms in order to obtain the dissipation rate.

Direct measurements of $\epsilon$ would require measuring all of the terms that compose Eq. (2.25) such as the normal strain ($\left( \frac{\partial u'}{\partial z} \right)^2$, $\left( \frac{\partial v'}{\partial y} \right)^2$, etc.), the shear strain ($\left( \frac{\partial u'}{\partial y} \right)^2$, $\left( \frac{\partial v'}{\partial z} \right)^2$, etc.) and the cross-related terms ($\frac{\partial u'}{\partial y} \frac{\partial v'}{\partial x}$, $\frac{\partial w'}{\partial y} \frac{\partial v'}{\partial z}$, etc.), and this is not feasible in an oceanic setting. However, a common assumption made in turbulence observations is that the turbulent field is isotropic, i.e. the turbulent velocity field is the same in all directions. This assumption allows for a significant simplification such that Eq. (2.25) can be written as

$$\epsilon = \frac{15}{2} \nu \left( \frac{\partial u'}{\partial z} \right)^2.$$  

(2.26)

A formal derivation leading to Eq. (2.26) is found in Yamazaki and Osborn (1990). Equation (2.26) allows for one measurement, in this case the vertical shear of the turbulent velocity, to be measured in order to obtain $\epsilon$. Although there exist many instances and length scales at which the flow is anisotropic, for typical Reynolds numbers found in the ocean the assumption of isotropy to determine $\epsilon$ from Eq. (2.26) is particularly good for vertical profiles (Smyth and Moum, 2000).

## 2.2 Microstructure Measurements

It is not trivial to measure turbulent quantities in the ocean. There exists a balance between the desired sensitivity of the instrument to resolve the
spatial and temporal scales of oceanic turbulence and the robustness to withstand the forces and volatility of the ocean. This last point is especially true near the ocean surface where strong turbulence and large waves create a broad scope of dynamics and scales that the sensor must contend with. Measurements on such small spatial scales (< cm) are often referred to as microstructure measurements.

Microstructure measurements in the ocean are relatively recent with the first observations taking place in the 1950’s (Stewart and Grant, 1962). An excellent review of the history and development of microstructure measurements can be found in Lueck et al. (2002). The most common method for making velocity microstructure measurements are with airfoil shear probes, which measure the turbulent shear tangential to the plane of the airfoil (the details are expanded on in Sec. 2.2.1). The response sensitivity of airfoil shear probes is a function of the angle of attack and inversely proportional to the mean velocity pass the probe. This inverse velocity sensitivity has lead to mounting shear probes on profilers which predominantly free fall downwards from the ocean surface (Osborn, 1974; Oakey, 1982; Shay and Gregg, 1986; Mourm et al., 1995), but have also been used on rising profilers (Soloviev et al., 1988; Anis and Mourm, 1995; Ward et al., 2004; Sutherland et al., 2013, 2014) to focus on the upper portion of the water column.

Profiles of turbulence measurements are predominantly obtained with free falling profilers deployed from the side of a large ship. Although the use of free falling profilers are excellent for observing diapycnal fluxes across the pycnocline (Schafstall et al., 2010; Kock et al., 2012; Fischer et al., 2013), they are insufficient for observing the turbulent characteristics of the upper 10 metres as this region is heavily contaminated due to the presence of the ship. Even upwardly rising profilers that are deployed from the ship (Soloviev et al., 1988; Anis and Mourm, 1995) could have contaminated surface measurements as the wake of the ship may extend hundreds of metres horizontally (Hodges and Fratantoni, 2014). Autonomous platforms
are required to accurately resolve turbulent properties in the upper few meters. This includes turbulence packages mounted on ocean gliders (Wolk et al., 2009; Fer et al., 2014), but these will deal with large errors in the upper few metres as they turn to descend during while the Air-Sea Interaction Profiler (Sutherland et al., 2013, 2014), which is both autonomous and vertically rising, does not try and change its course near the surface to give accurate results within a centimetre of the ocean surface.

2.2.1 Measuring Turbulence with Airfoils

The most common and robust sensor for measuring microstructure velocity variations in the ocean has to be the airfoil shear probe. Originally developed for atmospheric and wind tunnel measurements, it uses aerodynamical lift to measure small scale velocity variations. The sensor consists of a bullet-shaped rubber tip which houses a piezo-ceramic beam, as can be seen in Fig. 2.3. A voltage is created as the beam bends from the transverse current as the airfoil is moving through the water. The voltage, $E_p$, is proportional to the velocity of the sensor, $W$, and the transverse current, $u$, and is defined as

$$E_p = \hat{s}Wu$$  \hspace{1cm} (2.27)

where $\hat{s}$ is the sensitivity of the airfoil and is determined by calibration. In order to determine shear for estimates of $\epsilon$ from Eq. (2.26) the probe voltage in Eq. (2.27) is differentiated and transformed into shear via

$$\frac{\partial u}{\partial z} = \frac{1}{W} \frac{\partial u}{\partial t} = \frac{1}{\hat{s}W^2} \frac{dE_p}{dt}$$.  \hspace{1cm} (2.28)

When measuring shear with an airfoil there will be some spatial averaging involved due to the size and shape of the airfoil (Gregg, 1999; Macoun and Lueck, 2004). The spatial averaging will underestimate the spectrum at high wavenumbers, and hence underestimate the turbulent dissipation as shown in Fig. 2.4. To correct for this, Oakey (1982) assumed that the airfoil responds as a single-pole low-pass filter and estimated the effective
wavelength of his probe to be $\lambda_c = 2 \pm 1$ cm. In turn, this gives an estimated amplitude squared response to be

$$H^2(k) = \frac{1}{1 + \left(\frac{k}{k_c}\right)^2}$$  \hspace{1cm} (2.29)

where $k$ is the wavenumber and $k_c$ is the cut-off wavenumber of the shear probe ($k_c = \lambda_c^{-1} = 50$ cpm). Equation (2.29) allows airfoils to calculate strongly turbulent regions where $\epsilon > 10^{-4}$ m$^2$s$^{-3}$.

### 2.2.2 Spectral Analysis

To quickly preface this section on spectral fitting I must mention the contributions of Kolmogorov (1941) who introduced the notion of a spectral cascade, i.e. that energy flows from small wavenumbers to larger ones. The main hypothesis of Kolmogorov was that small scale components of a turbulent field are in or near a statistical equilibrium and are independent of the mean flow and the large scale motions by which energy is fed into the field. In this equilibrium range, turbulence could be uniquely defined by the viscosity $\nu$ and the rate at which energy is being supplied $\epsilon$ which in a steady-state is equivalent to the rate at which energy is being dissipated. Therefore, from dimensional reasoning, the state of equilibrium at high wavenumbers where variations in $\nu$ and $\epsilon$ can only have the effect of changing the length and
velocity scales

\[ \lambda_\nu = \left( \frac{\nu^3}{\epsilon} \right)^{1/4} \quad \text{and} \quad \quad u_\nu = (\nu \epsilon)^{1/4} \]  

(2.30)
respectively. This also implies that the associated Reynolds number at this Kolmogorov scale is

\[ Re = \frac{u_\nu \lambda_\nu}{\nu} = 1 \] (2.32)

which is expected as this is the scale at which viscous forces begin to dominate.

![Normalized universal shear spectra](image)

**Figure 2.5: Normalized universal shear spectra.**

In this equilibrium range and for a one-dimensional slice through three-dimensional isotropic turbulence, the energy spectrum for the turbulent shear velocity is written as \cite{Wolk2002}

\[ \Phi(k) = \left( \frac{\epsilon^3}{\nu} \right)^{1/4} F \left( \frac{k}{k_\nu} \right) \] (2.33)

where \( F \) is a universal function of the wavenumber \( k \) normalized by the Kolmogorov wavenumber

\[ k_\nu = \frac{2\pi}{\lambda_K}. \] (2.34)
There are two accepted equations for $F$ and these can be seen in Fig. 2.5. One, is the Nasmyth (1970) spectrum (Fig. 2.5, dashed line), which is an empirically calculated spectrum defined as

$$F_N = \frac{8.05 \left( \frac{k}{k_v} \right)^{1/3}}{1 + \left( 20 \frac{k}{k_v} \right)^{3.7}}, \quad (2.35)$$

and the other is the Panchev and Kesich (1969) spectrum, which is deduced theoretically for a one-dimensional profile through three-dimensional steady-state turbulence and is given by the equation

$$F_{PK} = 0.937 \left( \frac{k}{k_v} \right)^{0.3748} \cdot e^{-6.011 \left( \frac{k}{k_v} \right)^{1.548}}. \quad (2.36)$$

Figure 2.5 shows the $1/3$ slope (or 0.3748 for the PK function) in the inertial range with both dropping rapidly at $k \approx 0.3 - 0.4k_v$. In general the difference between the two functions is small, especially considering the accepted accuracy of a factor of 2 in measuring $\epsilon$ (Oakey, 1982; Yamazaki et al., 1990; Moum et al., 1995).

To calculate $\epsilon$ one can either integrate the shear spectrum over the resolvable wavenumbers and correct for the variance not accounted for from either one of the universal spectra (Moum et al., 1995; Wolk et al., 2002) or one can do a best fit of the measured shear spectra to one of the universal spectra and obtain $\epsilon$ from Eq. (2.33) (Ruddick et al., 2000). Benefits of the integral method is that it is a robust estimator of the shear spectra even when the entire wavenumber range isn’t resolved, while drawbacks are mostly due to contaminations in the signal either due to vibration of the profiler or detritus striking the probe. Methodologies vary of how to account for this noise, but the common aspects are: a despiking algorithm, a band-pass filter to remove low and high frequency vibrations outside the desired frequency range, and a coherent noise-removal algorithm to remove coherent energy between the shear signal and the accelerometers. The maximum likelihood estimation method is excellent in that it is unbiased estimator and only relies
on a small number of fitted parameters. However, this method requires an accurate model for the noise spectrum of the shear probe as this determines the cut-off wavenumber for the fit (Ruddick et al., 2000). For shear probes, the wavenumber dependent aspect of the noise model is expected to follow something similar to Eq. (2.29), but it may also be a function of local conditions and is difficult to accurately obtain in non-controlled conditions.

There is one further approximation that should be addressed and that is how the vertical profile time series of shear is converted to a spatial wavenumber. To do this, the assumption of Taylor (1938), commonly known as Taylor’s frozen field theory, is made. The assumption is that the turbulent velocity fluctuations are much less than the speed at which the profiler is passing through the turbulent patch and is frozen so the measured frequency $f$ can be related to the spatial wavenumber $k$ by the simple expression

$$f = kW$$

(2.37)

where $W$ is the mean speed of the profiler. This is an important assumption in converting time series measurements into a spatial domain where these turbulence models are applicable and leads to a give and take between wanting a long enough time series to accurately resolve the shear spectra while keeping it short enough not to violate the frozen field hypothesis.

### 2.3 Upper Ocean Turbulence

Turbulence near the ocean surface controls the evolution of the OSBL though a combination of buoyancy and mechanical mixing across the air-sea interface. A schematic of the processes affecting the evolution of the OSBL can be seen in Fig. 2.6. The OSBL is forced by the surface wind stress $\tau$ and buoyancy flux $B_s$ and surface waves through breaking and the Stokes drift velocity $u_s$. Surface wave breaking generates turbulence, and leads to high levels of dissipation, within a distance of the order of a significant wave height,
\( O(H_s) \), from the surface, but does not seem to be a controlling process at deeper levels (Terray et al., 1996; Sullivan et al., 2007). Stokes drift velocity associated with the non-breaking waves penetrates a deeper distance of order \( d = 1/2k \) (\( k \) is the wave number of waves at the peak in the wave spectrum). Below is a mixed layer where the character of the turbulence is controlled by the strength of the surface wind, buoyancy and waves via the Stokes drift velocity (Grant and Belcher, 2009). At the base of the mixed layer there can also be turbulence and a buoyancy flux generated through entrainment of denser fluid from below.

![Figure 2.6: Schematic diagram of forcing and dynamics in the OSBL.](image)

The depth dependence of \( \epsilon \) in the OSBL is still an ongoing point of research. Profile measurements in a lake (Dillon et al., 1981) and in the ocean (Oakey and Elliott, 1982; Soloviev et al., 1988) both found \( \epsilon \propto z^{-1} \). This is also known as “law of the wall” as this is the expected profile for shear driven turbulence along a solid boundary (Van Driest, 2012). This lead Soloviev et al. (1988) to conclude that shear was the dominant source of turbulent energy in the upper ocean and that waves were not a significant factor in the problem.

Although the law of the wall appeared to be valid in lakes and oceans,
there was an increasing amount of evidence suggesting that the air-sea interface may be more complicated than a flat wall. In a series of tower data recorded in strongly forced, fetch limited conditions in Lake Ontario, Ki-taigorodskii et al. (1983) found enhanced dissipation rates by 1-2 orders of magnitude over that predicted by wall layer theory. They proposed that enhanced dissipation was due to energy input from wave breaking in the upper ocean. This finding was later supported by other experiments (Agrawal et al., 1992; Anis and Moum, 1992, 1995; Terray et al., 1996; Drennan et al., 1996), who also were making measurements in more strongly forced conditions. Terray et al. (1996) found $\epsilon \propto z^{-2}$ at depths less than $H_s$ and that this transition to $\epsilon \propto z^{-1}$ at depth and concluded that the enhanced dissipation in the upper portion of the OSBL was due to wave breaking.

This section will review some of the more important aspects of near surface turbulence in the ocean. Section 2.3.1 begins with the definition of this near surface region, which is referred to as the mixed layer (ML). Section 2.3.2 will review the theory of turbulence generated by flow alongside a rigid boundary, also known as “law of the wall”. This is a well established theory and has some applications to the OSBL. However, the boundary of the ocean is not rigid and there is growing evidence that waves may play an important role in OSBL turbulence. Section 2.3.3 reviews some of the theory with regards to wave-turbulence interactions. There is also an indirect process in which surface gravity waves may impact OSBL dynamics and that is through Langmuir circulations (Langmuir, 1938). The mechanism for the generation of Langmuir circulations, which are generated by the interaction of the wave-induced Stokes drift velocity with the variable wind forcing, and the role they are expected to play in turbulence dynamics in the OSBL can be found in Sec. 2.3.4. This is followed by reviewing turbulence generated by gravitational instabilities at the surface through surface cooling in Sec. 2.3.5.
2.3.1 Mixed Layer

An important aspect of the turbulence dynamics of the OSBL is that the density in the near surface region is homogeneous, thereby creating a well mixed layer. This homogeneous density structure is predominantly a result of the vigorous mixing due to atmospheric forcing through wind, waves and unstable buoyancy fluxes and can range from a few metres to hundreds of metres in depth. The mixed layer is an important aspect of many processes in the upper region as it controls the amount of heat and trace gases which transfer between the atmosphere and the ocean. Also, the ML is the limiting depth to which phytoplankton can be mixed and if this depth is greater than the photic zone this lack of productivity would cascade throughout the pelagic fauna.

Many attempts have been made to define a strict criterion to determine the mixed layer depth (MLD). The vast majority of these use a density threshold relative to a reference depth, which defines an acceptable density variability within the OSBL usually in the range $\Delta \rho = 0.01 - 0.03$ kg m$^{-3}$ (Thomson and Fine, 2003; de Boyer Montégut et al., 2004). There are many other definitions such as defining a temperature threshold (Kara et al., 2000), a density gradient threshold (Lukas and Lindstrom, 1991), a linear optimal fitting approach (Chu and Fan, 2011), a split-merge method based on the profile shape (Thomson and Fine, 2003) or dissolved oxygen profiles (Castro-Morales and Kaiser, 2012). These values are all for individual profiles of density as averaging several profiles have a tendency to smooth out the pycnocline due to vertical motions associated with internal waves (de Boyer Montégut et al., 2004) and would have less strict definitions for the MLD, such as the $\delta \rho = 0.12$ kg m$^{-3}$ threshold of Levitus (1982).

One hypothesis for the broad range in mixed layer definitions is that the mixed layer is the integrated result of all previous mixing events and it may not equal to the region of the upper ocean which is currently being
mixed (Brainerd and Gregg, 1995). This mixing layer depth (we'll refer to as XLD) will be approximately equal to the MLD in steady-state conditions, but it could vary relative to the MLD by orders of magnitude when conditions are varying (Cisewski et al., 2008). Determining the XLD requires specialized microstructure measurements so there have been relatively few observations. Table 2.1 gives an overview of some of the definition criteria from experiments that have explicitly looked at both the mixed and active mixing layers. Definitions for the XLD all revolve around the idea that $\epsilon$ will fall to a background level $\epsilon_c$ sufficiently far away from the surface. Values for $\epsilon_c$ range between $10^{-9} \leq \epsilon_c \leq 10^{-8}$ m$^2$s$^{-1}$ and are estimated from observations where $\epsilon$ quickly drops by an order of magnitude.

The results of the studies in Table 2.1 are varied with no clear relation between the MLD and the XLD. One of the first studies on the relation between the MLD and XLD was by Brainerd and Gregg (1995) using microstructure profiles from the subtropical Pacific. Using direct measurements of $\epsilon$ and Thorpe scales (Thorpe, 1977), they calculated the mixing layer using the Advanced Microstructure Profiler (Moum et al., 1995). However, Thorpe scales, defined as the root-mean-square distance a parcel of water in a gravitationally unstable patch of water would have to travel to be stably sorted, should be used with care in the OSBL for a couple of reasons: first is the issue of resolution as the OSBL is quasi-homogeneous; Cisewski et al. (2008) demonstrated that the microstructure profiles accurately resolve overturns in the OSBL, although this may not be the case under more homogeneous conditions. Second, and more crucially, is that the Thorpe scale assumes the reordered profile is “the level of origin of the unstable fluid” (Thorpe, 1977), which in the case of the OSBL under convection, is clearly not the case. In short, it is not turbulence which creates a density instability, it is the buoyancy loss at the ocean surface. It is unclear if this distinction is important for relating $L_T$ with $\epsilon$ and hence mixing, but it would suggest that $L_T$ may not be an ideal proxy for turbulence in this case. Brainerd and
Gregg (1995) also use $\epsilon$ as a proxy for the mixing layer and this seems like a better choice since it is directly related to the turbulent kinetic energy in the OSBL even when the OSBL is completely homogeneous and hence no mixing involved. The mixing layer depth coincided with density differences from the surface value ranging from 0.005-0.5 kg m$^{-3}$ with no consensus for an ideal choice for the threshold.

There have been a few more recent studies that have investigated mixed and mixing layer depths at higher latitudes. During a transect across the north Atlantic at approximately 52°N, Lozovatsky et al. (2006) defined the mixed layer using a density threshold of $\Delta \sigma = 0.02\sigma_0$ where $\sigma_0$ is the surface density and the mixing layer based on $\epsilon$ falling to a background dissipation rate of $\epsilon_b = 10^{-8}$ m$^2$s$^{-3}$. They found little difference between the mixed and mixing layer depths, but their data spanned a transect of 42 stations with generally only 2-3 profiles per station (although sometimes 7-8). In the sub-arctic around Svalbard, Fer and Sundfjord (2007), using a mixed layer defined by the split-merge method of Thomson and Fine (2003) and a mixing layer using $\epsilon_b = 3 \times 10^{-8}$ m$^2$s$^{-3}$, found a mixing layer to be greater than the mixed layer for the majority of their profiles. However, this study consisted of a total of 81 profiles with 23 of these in ice-covered conditions. In the proximity of an Antarctic Polar Front, Cisewski et al. (2008), using a mixed layer threshold of 0.02 kg m$^{-3}$ and the same definition for the mixing layer as Lozovatsky et al. (2006), found mixing layers substantially less than the mixed layer.

Using an ocean general circulation model (OGCM), Noh and Lee (2008) directly compared values of mixed and mixing layer depths on a global scale. Two different density thresholds were tested, 0.1 and 0.02 kg m$^{-3}$, to determine the mixed layer depth from the OGCM. The mixing layer depth, determined directly from the vertical eddy diffusivity $K_\rho$, drops to a background value of $10^{-5}$ m$^2$s$^{-1}$ for the first time. It was found that $h_\epsilon > h_\sigma$ in regions where strong subsurface shear is present, such as the equatorial
<table>
<thead>
<tr>
<th>Author</th>
<th>Region</th>
<th>Mixed, $\Delta \rho$</th>
<th>Mixing</th>
<th>$z_r/m$</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Dewey and Moun</em> (1990)</td>
<td>mid-latitude N. Pacific</td>
<td>0.1 kg m$^{-3}$</td>
<td>$\epsilon_c = 5 \times 10^{-9}$ m$^2$s$^{-3}$</td>
<td>5</td>
</tr>
<tr>
<td><em>Brainerd and Gregg</em> (1995)</td>
<td>sub-tropical N. Pacific</td>
<td>0.05-0.5 kg m$^{-3}$</td>
<td>$\epsilon_c \approx 10^{-9}$ m$^2$s$^{-3}$</td>
<td>5</td>
</tr>
<tr>
<td><em>Lozovatsky et al.</em> (2006)</td>
<td>mid-latitude N. Atlantic</td>
<td>0.02 $\rho(0)$</td>
<td>$\epsilon_c = 10^{-8}$ m$^2$s$^{-3}$</td>
<td>10</td>
</tr>
<tr>
<td><em>Fer and Sundfjord</em> (2007)</td>
<td>sub-arctic N. Atlantic</td>
<td>split-merge</td>
<td>$\epsilon_c = 3 \times 10^{-8}$ m$^2$s$^{-3}$</td>
<td>10</td>
</tr>
<tr>
<td><em>Cisewski et al.</em> (2008)</td>
<td>mid-latitude S. Atlantic</td>
<td>0.02 kg m$^{-3}$</td>
<td>$\epsilon_c = 10^{-8}$ m$^2$s$^{-3}$</td>
<td>10</td>
</tr>
<tr>
<td><em>Noh and Lee</em> (2008)</td>
<td>global OGCM</td>
<td>0.02 kg m$^{-3}$</td>
<td>$K_\rho = 10^{-5}$ m$^2$s$^{-1}$</td>
<td>0</td>
</tr>
</tbody>
</table>

Table 2.1: Examples of mixed and mixing layer depth criteria.
ocean and western boundary current regions, and $h_e < h_o$ during early restratification and at high latitudes during convective cooling. In addition a zonal dependence for $\Delta \sigma$ was observed by minimizing the difference between the OGCM results and climatological data (Locarnini et al., 2006; Antonov et al., 2006).

2.3.2 Surface Stress - Law of the Wall

The “law of the wall” is the solution originally found by von Kármán for the velocity profile near a rigid boundary in the presence of a constant shear stress. Assuming a constant stress layer in the upper ocean, which is valid for relatively small distances from the ocean surface, yet large enough for viscous effects to be negligible, the mean velocity profile is logarithmic, i.e. the mean velocity shear is written as

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_*}{\kappa z}$$ (2.38)

$\kappa$ is the von Kármán constant ($\approx 0.4$), $z$ is the distance from the surface and $u_*$ is the friction velocity defined as

$$u_* \equiv \sqrt{\tau / \rho},$$ (2.39)

where $\tau$ is the wind stress and $\rho$ is the density of sea water. The wind stress, $\tau$, is a function of the wind speed at a reference height and a drag coefficient, i.e.

$$\tau = \rho_a C_D U_{10}^2.$$ (2.40)

Here $\rho_a$ is the reference air density, $C_D$ is the drag coefficient, and $U_{10}$ is the wind speed referenced to 10 meters height. Values of $C_D$ are on the order of $10^{-3}$, but vary as a function of wind speed (Large and Pond, 1981; Fairall et al., 1996, 2003) and wave conditions (which adds to the surface roughness) (Smith, 1988; Taylor and Yelland, 2001; Oost et al., 2002).
Using Eqs. (2.38–2.40), assuming steady-state TKE, and assuming the Stokes, buoyancy and transport terms of Eq. (2.22) are negligible, gives the definition for the “law of the wall” definition for turbulent dissipation, that is
\[ \epsilon = \frac{u^3_*}{\kappa z}. \] (2.41)

Early measurements of turbulence supported the idea of the OSBL as a purely shear driven wall layer (Csanady, 1984). Jones and Kenney (1977) found that turbulence velocity fluctuations scaled with \( u_* \) with a length scale comparable to the depth and inferred that surface wave orbital velocities act merely as “inactive” motions. Churchill and Csanady (1983), using quasi-Lagrangian drifters and drogues in a lake, reported logarithmic mean current profiles in the OSBL, consistent with a constant shear layer. However, they couldn’t definitely say whether this was due to turbulent motions or the Stokes drift velocity profile. Microstructure profile measurements in a lake (Dillon et al., 1981) and in the ocean (Oakey and Elliott, 1982; Soloviev et al., 1988) both found \( \epsilon \) to scale with Eq. (2.41). This lead Soloviev et al. (1988) to conclude that shear was the dominant source of turbulent energy in the upper ocean and that waves were not a significant factor in the problem. However, both of these experiments were in relatively calm winds where the observations of Dillon et al. (1981) were obtained in 4.8 m s\(^{-1}\) winds and were \( 1.8 < U_{10} < 6.5 \) m s\(^{-1}\) in Soloviev et al. (1988).

### 2.3.3 Wave-Turbulence Interactions

Although there is evidence that the “law of the wall” appears to hold up over some of the ocean, most of these experiments were performed in relatively calm conditions. There is an increasing amount of evidence suggesting that in the presence of breaking waves that the air-sea interface may be more complicated than a flat wall. Surface gravity waves act to increase the local TKE in the form of wave breaking (Craig and Banner, 1994; Sullivan et al.,
2007) and/or wave-turbulence interactions (Anis and Moum, 1995; Terray et al., 1996). Wave breaking effects are believed to be limited to a depth less than the height of the surface waves (Terray et al., 1996; Sullivan et al., 2007) and are stochastic in nature. This wave breaking region of the OSBL is a hostile environment and it does not lend itself easily to direct measurements of turbulent quantities, although wave effects are recognized as a required parameter in accurate bulk models of the OSBL (Craig and Banner, 1994; Kantha and Clayson, 2004; Janssen, 2012).

In a series of data recorded in Lake Ontario, Kitaigorodskii et al. (1983) found enhanced dissipation rates 1-2 orders of magnitude greater than predicted by wall layer theory. These were for velocity measurements at a fixed point it was unclear whether this enhancement was due to wave breaking or wave-turbulence interactions. The enhancement was observed at depths $5H_S$, where $H_S$ is the significant wave height of the wave field. The significant wave height is defined as the distance between the trough and the crest of the largest third of the waves. This slightly esoteric definition was introduced by Munk (1944) in an attempt to mathematically express the height estimated by a “trained observer”, but is now calculated directly from the wave energy spectrum. According to the World Meteorological Organization (1998) the significant wave height calculated using the wave energy spectrum (we’ll call $H_{m0}$) can be related the the classic definition as $H_{m0} = 1.05H_{1/3}$ where $H_{1/3}$ is the definition using the higher 1/3 of the waves as laid out by Munk (1944).

Enhanced dissipation relative to a constant shear layer was later supported by other experiments (Agrawal et al., 1992; Terray et al., 1996; Drennan et al., 1996), in strongly forced conditions with enhanced dissipation being restricted to within a few significant wave heights from the mean surface. However, these experiments were unable to devise a universal relation due to different limitations in each. The experiments of Agrawal et al. (1992) and Terray et al. (1996) use fixed point measurements of the
turbulent velocity in a lake with the velocity spectra calculated over 90 minute records. Using fixed point measurements at depths comparable to $H_S$ will give apparent enhanced velocity signals due to the logarithmic profile shape relative to the free surface (as demonstrated by Gemmrich, 2010). The experiment by Drennan et al. (1996) was similar in that it used a point measurement of velocity from the box of a ship in open ocean conditions consisting of locally generated wind waves and non-local swell. These measurements are at one fixed point so the depth scaling is only due to the variation in $H_S$. Drennan et al. (1996) found the identical enhancement as Terray et al. (1996) if only the wind generated portion of the wave spectra is used to scale the dissipation.

Anis and Moum (1995) took a different approach as their measurements were using a vertically rising microstructure profiler. Although they observed many periods where $\epsilon$ was less than the law of the wall with no clear reason as to why, their analysis focused on one deployment where there was a clear enhancement in $\epsilon$ and investigated possible mechanisms which would explain the enhanced dissipation. Expanding on the Reynolds decomposition of Eq. (2.2) to include a wave component to the velocity separation, i.e.

$$ a = \bar{a} + \hat{a} + a', \quad (2.42) $$

where $\hat{a}$ is the phase-averaged wave component of the total velocity. Anis and Moum (1995) calculated the TKE equation, assuming the mean velocity is in the x-direction, i.e. $\bar{u} = (\bar{u}(z), 0, 0)$, to be

$$ \frac{D}{Dt}\left(\frac{1}{2}u'_i u'_i\right) = -\frac{\partial}{\partial z}\left(w' \left(\frac{p'}{\rho} + \frac{1}{2}u'_i u'_i\right)\right) - \frac{g w'}{\rho} - \frac{w'}{\rho u' \frac{\partial \bar{u}}{\partial z}} $$

$$ -\frac{\partial}{\partial z}\left(\bar{w} \frac{1}{2}u'_i u'_i\right) - \frac{\partial}{\partial z}\left(w' \frac{1}{2}u'_i u'_i\right) - u'_i u'_j \frac{\partial \bar{u}_i}{\partial x_j} - \epsilon. \quad (2.43) $$

Equation (2.43) is similar to Eq. (2.5), but with two additional terms from the wave field: a wave-induced transport term $\frac{\partial}{\partial z}\bar{w} \frac{1}{2}u'_i u'_i$ and an interaction of the Reynolds stress with the orbital shear $u'_i u'_j \frac{\partial \bar{u}_i}{\partial x_j}$. These two terms will be expanded upon in the following paragraphs.
The analysis of *Anis and Moum* (1995) focuses on two types of wave fields: irrotational and rotational. Irrotational waves are waves derived assuming zero curl in the velocity field ($\Omega = \nabla \times u = 0$, where $\Omega \equiv$ vorticity), resulting in the horizontal and vertical components of the wave orbital motion to be in quadrature, i.e. exactly 90° out of phase with one another while rotational waves have a non-zero vorticity and, therefore, the vertical and horizontal components of the wave orbit are no longer in quadrature. For irrotational waves, one suspected mechanism for enhanced $\epsilon$ is the wave-induced transport of TKE (i.e. the $\frac{a}{2} w^1 u' u'_i$ term) since near the surface there must be a net transport downwards due to TKE transport upwards being restricted by the air-sea interface. The main idea is that TKE generated near the surface is transported away by the swell. Determining the magnitude of this term is difficult as it relies on the correlation of the wave velocity with TKE, it should decrease at least as fast as $\tilde{w} \propto e^{-k|z|}$ and they use this to obtain an upper bound.

On the other hand, if the wave field is rotational and the vertical and horizontal wave velocities are out of quadrature by a phase shift $\phi$ there will be a Reynolds stress from the wave orbital velocities

$$\overline{w'u'} = \frac{1}{2} a^2 g k e^{-2k|z|} \sin \phi$$

where $\phi$ is assumed to be constant with depth and $a$ is the wave amplitude, $k$ is the wavenumber and the dispersion relation for deepwater waves of $\omega^2 = gk$ is used. Although this Reynolds stress from irrotational waves does not factor in Eq. (2.43), it does appear as a production term of kinetic energy of the wave component of the flow it has been shown that near the surface the Reynolds stress due to wave motions can be approximated as being equivalent to the turbulent Reynolds stress, i.e. $\overline{u'u'} = \overline{U'U'}$. Thus, the production of turbulence from the mean shear may be written as

$$-\overline{u'w'} \frac{\partial \bar{u}}{\partial z} \approx -\overline{uw} \frac{\partial \bar{u}}{\partial z} = \frac{1}{2} a^2 g k e^{-2k|z|} \sin \phi \frac{\partial \bar{u}}{\partial z}$$

(2.45)
where $\phi$ is the angle out of quadrature between the horizontal and vertical components of the wave current. Anis and Moum (1995) state that only small departures from quadrature ($\phi \sim O(1^\circ)$ on average) are necessary to account for the enhanced dissipation they observed. However, Eq. (2.44) assumes the wave field is monochromatic which is definitely not the case in reality and deviations from quadrature will be most likely a function of frequency as well as depth with the overall magnitude suspected to be less than Eq. (2.45). There is another wave-turbulence interaction term in Eq. (2.43), $u_i' u_j' \frac{\partial \tilde{u}_i}{\partial x_j}$ which is deemed unimportant on dimensional grounds as it is assumed that this term is only a factor on timescales $T_t \ll T$ where $T$ is the wave period. Assuming a time scale $T_t = l_t / u_t$ where $l_t$ and $u_t$ are representative length and velocity scales velocity scales, and that $\epsilon_t = u_t^3 / l_t$, it follows that $u_t \ll \sqrt{T\epsilon}$. Using values of $\epsilon \sim 10^{-5} \text{m}^2\text{s}^{-3}$ and $T \sim 10 \text{ s}$ gives $u_t \ll 0.01 \text{ms}^{-1}$ which is much too small to be consistent with enhanced turbulence levels. However, there are wave-induced motions which last much longer than the wave period as shown earlier in Eqs. (2.19–2.20). The results of which will be explored in the next section.

In addition to the above mechanisms for wave turbulence interactions, there may also be a dependence of the enhanced turbulence dissipation rates with the phase of the wave. Well before these measurements, Stewart and Grant (1962) argued that more than one-half of the energy is dissipated above the mean waterline. That is, that dissipation profiles below a crest and a trough of a wave should differ and hence contradicts any assumption of a constant dissipative layer. This was shown by Gemmrich and Farmer (2004) that energy dissipation is greatly enhanced in the crest region. Dissipation estimates between the mean waterline and the crest of the wave are, on average, 2.5 times greater than that between the trough and the mean waterline (Gemmrich and Farmer, 2004). Gemmrich (2010) calculated $\epsilon$ within two significant wave heights of the free surface and found dissipation rates under the wave crest region to be an order of magnitude greater.
under the crest of the wave than under the trough. The depth dependence of $\epsilon$, i.e. $\epsilon(z) \propto z^m$, was calculated and it was found that beneath wave troughs $m = -1$, independent of whether the wave is breaking or not. This is in contrast to $\epsilon$ under wave crests where $m = -1.1$ when there is no wave breaking and decreasing to $m = -1.6$ as wave breaking becomes frequent (Gemmrich, 2010). These results suggest that waves greatly enhance surface dissipation with respect to the law of the wall scaling only half the time, under wave crests, while profiles of dissipation beneath the trough of waves still follow the law of the wall.

### 2.3.4 Langmuir Turbulence

It has been speculated that enhanced turbulence in the OSBL could also be generated by the presence of Langmuir Circulations in the OSBL (Gargett, 1989; McWilliams et al., 1997; Li et al., 2005). Langmuir Circulations (LC), as first observed by Langmuir (1938), are recognized by their long streaks, or windrows, which are formed by bands of horizontal convergence due to an interaction with the wind and the Stokes drift. A schematic of a LC pattern is shown in Fig. 2.7. The basic physics is an instability mechanism in which an infinitesimal downwind jet has its vertical vorticity, with opposite sign on the two sides of the jet, tilted by the Stokes drift of the surface waves to produce longitudinal rolls (Craik and Leibovich, 1976; Craik, 1977; Leibovich, 1980). These rolls produce the surface convergence at the location of the jet, and this is in turn reinforced because of the acceleration, due to the wind stress, of the water moving towards the surface convergence.

Studies that have investigated enhanced turbulence from LC have focused on the dimensionless parameter,

$$La = \sqrt{\frac{u_s}{u_s(0)}}$$

which is known as the turbulent Langmuir number as defined by McWilliams et al. (1997), as being the dominating term whether LC’s arise or not.
Equation (2.46) gives the relative importance of wave generated turbulence to wind generated turbulence where values less than 0.3 indicate LC’s can form and that wave induced turbulence should be the dominant source (Polton and Belcher, 2007). This work has predominantly involved the use of LES models (Large Eddy Simulations) (Skyllingstad and Denbo, 1995; McWilliams et al., 1997; Polton and Belcher, 2007; Grant and Belcher, 2009) which rely on parameterizations to resolve sub-grid size dynamics such as turbulent dissipation. LC’s are expected to extend to the base of the OSBL with numerical results suggesting a La dependence on enhanced dissipation levels throughout the OSBL. However, observations in the open ocean are minimal with results suggesting slightly different dependencies on $\epsilon$ (Smith, 1992, 1998; Plueddemann et al., 1996).

A recent attempt has been made to model the enhanced dissipation rates near the surface by modelling the interaction of turbulence with the shear induced from the wind and the wave field using Rapid Distortion Theory (Teixeira, 2011a,b, 2012). What Rapid Distortion Theory (RDT) does is it assumes that the non-linear equations associated with turbulence are linear on small time scales and solves these linearized equations for small time perturbations. Teixeira (2011a,b, 2012) used this method in the current
context to model the time evolution of the Reynolds stress as

$$\bar{u}^\prime w^\prime = -u_2^2 \exp \left[ 2 \left( \frac{\partial \bar{u}}{\partial z} \frac{\partial u_s}{\partial z} \right)^{1/2} T_L \right]$$  \hspace{1cm} (2.47)

where $T_L$ is eddy turnover time associated with the the time perturbation of the RDT assumption. The Reynolds stress of the form Eq. (2.47) is true when $\partial \bar{u}/\partial z$ and $\partial u_s/\partial z$ are the same sign (Teixeira, 2011a). If the shear from the wind and the waves are opposite sign the solution is periodic. This Reynolds stress combined with the shear induced by the wind and the Stokes drift velocity

$$\epsilon = -\bar{u}^\prime w^\prime \left( \frac{\partial \bar{u}}{\partial z} + \frac{\partial u_s}{\partial z} \right),$$  \hspace{1cm} (2.48)

which is identical to the sum of the mean shear production and the stokes shear production of Eq. (2.21). Combining Eq. (2.38), Eq. (2.6) and Eq. (2.47) into Eq. (2.48) gives a relationship for the parameterization of $\epsilon$. This model has been tested with data from Terray et al. (1996) and Drennan et al. (1996) with encouraging albeit scattered results. However, it is not clear what the choice of $T_L$ in Eq. (2.47) should be as it will depend on calculating both the representative length and velocity scales of the eddies.

Huang and Qiao (2010) argue that the the most dominant term in Eq. (2.22) is Eq. (2.21) and came up with a model for the turbulent dissipation rate in the OSBL. Huang and Qiao (2010) parameterized the Reynolds stress as $-\bar{u}^\prime w^\prime = a_1 u_*^2$ where $a_1$ is a dimensionless constant associated with the surface waves

$$a_1 = \frac{15}{8} \beta \sqrt{\pi ak},$$  \hspace{1cm} (2.49)

where $0 < \beta \leq 1$ and is determined by empirical regression. This model was applied to a few microstructure data sets (Anis and Moun, 1995; Osborn et al., 1992; Wüst et al., 2000) with modest success. However, their fits required $\beta$ to vary by an order of magnitude between 0.15 and 1.0. Although the magnitude of the Stokes induced turbulence is suspect, i.e. with $\beta$
varying by such a large degree, the fact that the dissipation profiles follow $e^{-2k|z|}$ is encouraging that the Stokes shear production may be an dominant factor in the OSBL dynamics.

### 2.3.5 Surface Buoyancy Flux

Another important term in the dynamics of the near surface is the buoyancy flux (either heat or salt), which in turn affects the density driven flows in this region. A useful length scale in the study of atmospheric boundary layers (ABL) is the Monin-Obukhov length scale, which is a length scale which equates the wind induced shear production with the surface buoyancy flux $B_s$, i.e.

$$L = -\frac{u^3_*}{\kappa B_s}. \quad (2.50)$$

The Monin-Obukhov length scale arrives simply from dimensional analysis and has been a driving force in understanding convection in the ABL (see review Foken, 2006). This is also known as similarity theory where the dissipation profile is a universal function of $z/L$ and where $L < 0$ when $J_0 > 0$ (i.e. net buoyancy flux out of the ocean) for unstable conditions favourable to convection and $L > 0$ for stable conditions. Thus convection should be dominating turbulence for depths greater than $L$ and less than the mixed layer of depth (MLD) (Phillips, 1977). For depths less than $L$, a combination of wind forcing and wave-breaking will dominate the turbulent energy spectrum.

Shay and Gregg (1984, 1986) found strong evidence for large convective cells in the Bahamas and in a warm-core Gulf Stream ring. Mean values for $\tau/B_s$ were found to be 0.61 and 0.72 for the Bahamas and warm-core ring data respectively, which compared very well to the 0.64 value found in convecting atmospheric layers (Caughey and Palmer, 1979). The mixed layer depths in the Bahamas and in the ring went as deep as 100 m and 150 m respectively during night-time cooling and re-stratified during the day. In
the tropical Pacific, Lombardo and Gregg (1989) found $\epsilon$ to scale well with predictions for a convective cell and calculated a mean value for $\tau/B_s = 0.58$ and a mixed layer depth down to 50 m.

One obvious deficiency in applying similarity theory to the OSBL is that it assumes that the dissipation is a balance between a wall-layer shear dissipation and convective forcing, and omits the effects of surface gravity waves which are an important part of OSBL dynamics. The effects of waves on using Eq. (2.50) will be most pronounced when the wave dynamics are comparable to the MLD, which is the case during strong stable buoyancy forcing such as heating and/or rainfall, or effects due to Langmuir circulations. Studies by Li et al. (2005), and subsequently Belcher et al. (2012), have used numerical results to construct regime diagrams which weigh in the contributions from wind, waves and surface buoyancy flux. However, simultaneous observations of the turbulence dynamics in the OSBL along with the wind, waves and surface buoyancy flux are very rare so these proposed scaling mechanisms have not been able to be verified.

Although it is encouraging to be able to parameterize turbulent dissipation over such large depths, these unstable convection areas are relatively scarce (Gargett, 1989). It is also not trivial to be able to accurately choose the exact MLD and that this often is not the same as the depth to which active mixing is occurring (Brainerd and Gregg, 1995). In these convective regions the stratification is very small and the resolution of the hydrographic parameters used to determine the MLD (e.g. temperature change from the surface) could cause significant errors in the mixed layer depth estimate. Also, the turbulence generated from strong wind and wave forcing may be a considerable fraction of the mixed layer depth (Kitaigorodskii et al., 1983).
Chapter 3

Wave-Turbulence Scaling in the Ocean Mixed Layer

Preface

This chapter is an adapted reproduction of a paper published in Ocean Science with co-authors Brian Ward and Kai Christensen:


The right to share and adapt this work is freely available under the Creative Commons Attribution 3.0 License. The data analysis, the interpretation and synthesis of results, the production of figures and the writing were done exclusively by the author of this thesis. Dr. Ward contributed by supervising, assisting and reviewing the work and by providing the infrastructure (computer time, ship-time, instrumentation, etc.) required to carry out this research project. Data collection was jointly done by the author and Dr. Ward. Dr. Christensen contributed measurements of the wave field and assisting and reviewing the work.
3.1 Introduction

The level of turbulent kinetic energy in the ocean depends on the balance between energy production, suppression by buoyancy and dissipation. The latter term is representative of the availability of mixing, which is parameterised in large-scale numerical models that lack the resolution to directly compute dissipation. There has been considerable effort to understand the scaling of the turbulent dissipation rate of kinetic energy ($\epsilon$) in the ocean surface boundary layer (OSBL). Here, $\epsilon$ directly influences many air-sea processes such as the mixing of near-surface water properties (Garrett, 1996; Stevens et al., 2011), gas transfer across the ocean interface (Lorke and Peeters, 2006; Zappa et al., 2007) and the dynamics and evolution of plankton blooms (Denman and Gargett, 1989; Yamazaki et al., 1991). Parameterizing $\epsilon$ in the OSBL has proven to be more difficult than in the atmospheric boundary layer (ABL) due to the presence of surface gravity waves (Agrawal et al., 1992; Terray et al., 1996) and Langmuir circulations (McWilliams et al., 1997; Grant and Belcher, 2009) creating enhanced dissipation relative to what is expected from a shear driven boundary layer.

There has been a desire for more observations of near-surface values of $\epsilon$ due to the growing prevalence of using Large Eddy Simulations (LES) to model the OSBL (McWilliams et al., 1997; Noh et al., 2004; Grant and Belcher, 2009). The key parameter in balancing the sub-grid dynamics is $\epsilon$ and due to it not being resolved directly it is required to be parameterized. However, the lack of observations limits the ability to validate proposed parameterizations (see Noh et al. (2004) or Grant and Belcher (2009)).

Attempts at parameterizing $\epsilon$ in the OSBL have traditionally begun with similarity scaling, which treats $\epsilon$ in the OSBL as a shear driven wall layer. This method assumes a constant stress with the mean velocity having a logarithmic profile, so that the shear is

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_s}{\kappa z}$$

(3.1)
where \( \bar{u}, z, \kappa \) and \( u_* \) are respectively the mean velocity, depth, von Kármán constant (\( \kappa = 0.40 \)) and friction velocity, defined as:

\[
 u_* = \sqrt{\tau / \rho} 
\] (3.2)

where \( \tau \) is the wind stress and \( \rho \) is the density of seawater. The wind stress is assumed to be constant across the air-sea interface so that \( \tau = u_*^2 \rho = u_{*a}^2 \rho_a \) where the \( a \) subscript denotes the air friction velocity and density respectively.

For a steady-state solution the turbulent kinetic energy equation can be written as a balance of shear production, buoyancy flux, and turbulent dissipation rate, \( \epsilon \), \((\text{Osborn}, \ 1980)\)

\[
 0 = -u'w' \frac{\partial \bar{u}}{\partial z} + g \frac{\rho}{\rho'} w' \rho' - \epsilon 
\] (3.3)

where \( u' \) and \( w' \) are the turbulent horizontal and vertical velocities respectively and \( g \) is the acceleration due to gravity. Since the wind stress, \( \tau \), can be directly calculated from the Reynolds stress as \( \tau = \rho u'w' \) it follows from Eq. (3.2) that

\[
 -u'w' = u_*^2. 
\] (3.4)

Therefore, ignoring buoyancy, which is often an order of magnitude smaller than the other terms, and substituting Eq. (3.1) and Eq. (3.4) into Eq. (3.3) leads to the familiar ”law of the wall” scaling, i.e.

\[
 \epsilon \tau = \frac{u_*^3}{\kappa z}. 
\] (3.5)

Early measurements supported the idea of the ocean as a purely shear driven wall layer. \emph{Jones and Kenney} (1977) found that turbulence velocity fluctuations scaled with \( u_* \) with a length scale comparable to the depth. \emph{Churchill and Csanady} (1983), using quasi-Lagrangian drifters and drogues, reported logarithmic mean current profiles in the OSBL, consistent with a constant shear layer. Profile measurements in a lake (\textit{Dillon et al.}, 1981) and in the ocean (\textit{Oakey and Elliott}, 1982; \textit{Soloviev et al.}, 1988) both found \( \epsilon \) to scale with Eq. (3.5).
Observations by Kitaigorodskii et al. (1983) and subsequently by Agrawal et al. (1992), using velocity microstructure measurements at fixed depths, found enhanced dissipation by an 1 to 2 orders of magnitude relative to the law of the wall in the upper ocean. This enhancement was attributed to the presence of breaking surface gravity waves directly injecting turbulent kinetic energy into the near surface region.

Terray et al. (1996) suggested enhanced $\epsilon$ values could be scaled using parameters of the wind-wave field having a vertical structure with three distinct regions. The uppermost region from the surface down to a depth $z_b = 0.6H_s$, where $H_s$ is the significant wave height, experiences a large uniform turbulent dissipation rate due to the presence of breaking waves. The dissipation rate in this region is an order of magnitude greater than predicted by Eq. (3.5). Below this there is an intermediate region of enhanced $\epsilon$ which has decays as $z^{-2}$. This enhanced region extends to $z_t = 0.3H_s\kappa\tilde{c}/u_{*a}$ where $\tilde{c}$ is the effective wave speed equating the energy input from the wind to the waves, $F = \tilde{c}u_\kappa^2$ and $u_{*a}$ is the air side friction velocity. The ratio of $\tilde{c}/u_{*a}$ is related to the wave age $c_w/u_{*a}$ where $c_w$ is the phase velocity of the wind generated waves. Hence, it was the conclusion of Terray et al. (1996) that the depth of the enhanced region is dependent on the wave age as well as the significant wave height. Below this depth $\epsilon$ scales as Eq. (3.5). The same depth dependence of $\epsilon$ was found by Drennan et al. (1996) using fixed depth velocity measurements from the bow of a ship in a mixed swell/wind sea in the Atlantic. Further support for enhanced dissipation has been established by several ocean studies, but with various depth profiles for $\epsilon$ (Gargett, 1989; Anis and Moum, 1995; Greenan et al., 2001). In a study in the North Pacific using a microstructure profiler Gargett (1989) found $\epsilon \propto z^{-4}$ in the mixed layer during a week of intermittent stormy weather, which relaxed to $\epsilon \propto z^{-1}$ as the wind and sea states calmed.

Anis and Moum (1995), using a rising vertical profiler in the Pacific Ocean, reported enhanced dissipation with respect to Eq. (3.5) with an exponential
depth decay. A profile of $e^{2kz}$, where $k$ is the dominant wave number of the surface wave field, matched closely with the observed wave number associated with the wind generated waves during enhanced dissipation. This $e^{2kz}$ depth dependence was explained theoretically by a rotational wave field with only a small deviation from quadrature necessary, on the order of a couple of degrees, to re-create their observed dissipation levels. This $e^{2kz}$ depth dependence of turbulent kinetic energy dissipation was also found by Huang and Qiao (2010) by ignoring the buoyancy term in Eq. (3.3), i.e.

$$\epsilon_w = a_1 u_s^2 \frac{\partial u_s}{\partial z}$$  \hspace{1cm} (3.6)$$

where $u_s$ is the Stokes drift induced by surface waves. For a monochromatic wave the Stokes drift is

$$u_s = u_{s0} e^{2kz}$$  \hspace{1cm} (3.7)$$

where the magnitude at the surface is equal to $u_{s0} = c(ak)^2$ and $c$, $a$ and $k$ are the wave speed, amplitude, and wave number respectively. In Eq. (3.6), $a_1$ is a dimensionless constant associated with the surface waves and predicted by regression. Using the observations of Anis and Moum (1995), Huang and Qiao (2010) estimated $a_1$ to be

$$a_1 = 3.75 \beta \pi \sqrt{\frac{H_s}{\lambda}}$$  \hspace{1cm} (3.8)$$

where $\lambda$ is the dominant wavelength and $\beta$ is a dimensionless constant between 0 and 1. Huang and Qiao (2010) compared Eq. (3.6) with observations by Anis and Moum (1995), Osborn et al. (1992) and Wüest et al. (2000) with an order of magnitude agreement for values of $\beta$ between 0.15 and 1.

Studies are ongoing in this field (Stevens and Smith, 2004; Gerbi et al., 2009) and it is still unclear on the role surface waves play in dissipation (Babanin and Haus, 2009; Huang and Qiao, 2010; Teixeira, 2012). Observations of $\epsilon$ in the deep ocean (Greenan et al., 2001) suggest that any scaling may not be straightforward especially in the presence of complex wave fields.
Over recent years there has been a large increase in modelling the OSBL using Large Eddy Simulations (LES). LES models have shown to be very effective in the ABL, while attempts in the OSBL have been hindered by the difficulty of handling the boundary condition at the surface. The difficulty lies in the lack of observations of $\epsilon$ in the near surface and that without these measurements it is difficult to properly model these processes accurately (Noh et al., 2004; Grant and Belcher, 2009).

In this chapter we describe a set of observations taken during a research cruise in the North Atlantic with the upwardly-rising Air-Sea Interaction Profiler (ASIP) microstructure instrument, an ultrasonic wave altimeter, and a high quality suite of meteorological sensors. Details of all the available measurements and processing algorithms are presented in Sec. 3.2. Dissipation scaling in the upper ocean, including wave-induced scaling and comparisons of the integrated dissipation with the wind input, is discussed in Sec. 3.3. A summary of the results are presented in Sec. 3.4.

3.2 Measurements

Measurements were conducted during a field campaign in the North Atlantic (Fig. 3.1a) aboard the R/V Knorr from late June to mid-July 2011. Presented here is one deployment of the Air-Sea Interaction Profiler (ASIP), an autonomous microstructure profiler designed to study the OSBL (manuscript in preparation). A total of 54 profiles were made spanning 16 hours from 2 July 2011 14:38 to 3 July 2011 6:26. The drift track of ASIP and the ship position over the course of the deployment can be seen in Fig. 3.1b. The location is determined via the GPS receiver on ASIP which obtains a position at the surface after each profile. There are two gaps in the profiling, from 16:43 to 19:01 LMT and 20:09 to 21:47 LMT, where ASIP was unable to obtain a valid GPS location and profiling was temporarily suspended. All times are in local mean time (LMT), which aligns noon with the maximum
solar angle. For the measurement location (Fig. 3.1b) this corresponds to a time difference of 3 hours and 8 minutes behind UTC.

Figure 3.1: (a) The 200 m, 1000 m, and 2000 m depth contours are shown for the North Atlantic with the deployment location shown by the large black dot. Inset map (b) shows the ASIP profile locations (black dots) with the green dot being the first profile and the red dot showing the final profile. The ship locations at three hour intervals are shown by the black triangles. All times are in local mean time.

### 3.2.1 Microstructure Measurements

ASIP is equipped with two FP07 micro-scale temperature sensors, one SBE 7 micro-conductivity sensor, and two SPM-38 vertical shear microstructure sensors from which $\epsilon$ was computed (Macoun and Lueck, 2004). In addition to these, there is an accurate (CTD-standard) temperature and conductivity
sensor manufactured by Neil Brown Ocean Sensors Inc. (NBOSI), a Licor LI-92 Photosynthetic Active Radiation (PAR) sensor which measures incoming shortwave radiation (between 400 and 700 nm), a Keller pressure sensor, and accelerometers and orientation sensors. ASIP is positively buoyant rising upwards with a nominal speed of 0.5 $\text{ms}^{-1}$. In the wave affected region in the upper ocean, the rise velocity is calculated from a linear fit of the pressure record for the upper 10 metres. A linear fit of the pressure gradient ensures that the effects of waves are filtered out of the pressure signal. This is identical to the method adopted by Stips et al. (2005) to calculate rise velocities of a rising vertical profiler in the presence of surface waves.

The location of the ocean surface is determined for each profile from the micro-conductivity record, using a surface detection method similar to Stips et al. (2005). The uncertainty in the surface location is estimated to be $\pm 2$ cm, determined from examining hundreds of profiles in various sea states, and this result is consistent with Stips et al. (2005). The other sensors are aligned to the micro-conductivity sensor so each measurement is referenced to the same depth. Details of the surface detection and calibration algorithms can be found in Ward et al. (2014).

Turbulence parameters are calculated from the measured vertical shear (Osborn, 1974; Moum et al., 1995) over segments of 1024 points with a 512 point overlap. The vertical resolution depends on the sampling rate and rise velocity and in our case the resolution was approximately 0.5 m. Various segment lengths were tested and the 1024 segment length was found to provide a good balance between statistical significance and homogeneity. A sample spectrum taken from a profile at 2 July 2011 17:15 LMT during moderate wind speeds can be seen in Fig. 3.2. To ascertain the presence of any persistent noise artifacts, a spectrogram is calculated (Fig. 3.3e) from the raw shear signal (Fig. 3.3c) for each profile. There is a faint signal at 30 Hz, which corresponds to a wave number of $\approx 60$ cpm, but this is only prominent where the signal is low and the calculated $\epsilon$ is below the adopted
Figure 3.2: Sample shear spectrum (black dots) and modelled Nasmyth spectrum (dashed line) for the depth interval $-17.6\,\text{m} < z < -17.1\,\text{m}$ of the profile corresponding to Fig. 3.3. The lower and upper integration limits are denoted by the vertical dashed lines.

noise floor of $5\times10^{-10}\text{m}^2\text{s}^{-3}$ (Fig. 3.3d).

### 3.2.2 Meteorological Measurements

Meteorological data represent the direct forcing to the OSBL. This includes wind forcing and buoyancy forcing. The buoyancy forcing is represented as the change in density at the surface from radiative forces changing the temperature and hydrological forces, such as rain and evaporation, altering the salinity. These data were recorded continuously throughout the campaign with the on-board data acquisition system on the R/V Knorr. Measurements were recorded at one minute intervals and these were averaged into 30 minute bins.

Wind measurements were recorded using two Vaisala WXT520 weather sensors mounted at 15.5 metres above the waterline on the forward mast.
Figure 3.3: Spectrogram of a profile taken at 17:15 LMT (panel (e)). The depth and corresponding rise velocity are shown in panels (a) and (b). A uniform rise speed is adopted in the upper 10 m (solid line) to filter out wave effects in the pressure signal (dotted line). Panel (c) shows the raw shear signal in volts and the calculated dissipation rate is in (d). Measurements are sampled at 2 Hz and these are averaged and recorded at one minute intervals. The wind measurements at 15.5 metres are corrected to the standard 10 metres above sea level using the TOGA COARE 3.0 algorithm which assumes a logarithmic profile with height (Fairall et al., 1996, 2003).

The density flux \(Q_p\) into the ocean from the atmosphere was computed as (e.g. Zhang and Talley, 1998)

\[
Q_p = \rho_s (\alpha F_T + \beta F_S)
\]

where \(\rho_s\) is the surface density and \(\alpha\) and \(\beta\) are the thermal expansion
Wave-Turbulence Scaling in the Ocean Mixed Layer

and saline contraction coefficients respectively. Here $F_T = -Q_{net}/\rho_s C_p$ and $F_S = (E - P) S / (1 - S/1000)$ where $C_p$ is the specific heat of sea water, $E$ and $P$ are the evaporation and precipitation rates respectively in m/s and $S$ is the sea surface salinity in psu.

The net radiative heat flux at the ocean surface is calculated from the combination of the incoming shortwave ($SW$), net incoming and emitted longwave ($IR$), sensible heat ($SH$) and latent heat ($LE$), i.e.

$$Q_{net} = SW + IR + SH + LE. \quad (3.10)$$

Shortwave and longwave components were measured from the deck of the R/V Knorr (S. Miller 2012, personal communication) while both SH and LE were computed using TOGA COARE 3.0 flux algorithm (Fairall et al., 1996, 2003).

The buoyancy flux, $B$, is a function of the density flux at the surface, i.e.

$$B = -gQ_p \quad (3.11)$$

where $g$ is the acceleration due to gravity. The minus sign indicates that the upper surface becomes less buoyant (i.e. more dense) when there is a positive buoyancy flux out (i.e. upwards) of the ocean surface. $B > 0$ implies the density flux is negative (i.e. into the ocean) and is destabilizing which may lead to convection. If $B < 0$ then the mass flux is out of the sea, leading to stabilizing conditions.

**3.2.3 Mixed Layer Depth**

The mixed layer depth represents the depth at which the surface properties such as temperature and salinity are deemed homogeneous. However, the mixed layer is a dynamic region and there are many different methods for calculating the depth of this layer (see Thomson and Fine (2003); de Boyer Montégut et al. (2004); Stevens et al. (2011) for a brief literature review of some of the methods used historically). In determining the mixed layer depth,
\( D \), we use the same threshold value as de Boyer Montégut et al. (2004) of a 0.03 kg m\(^{-3}\) increase in the potential density from a reference depth. de Boyer Montégut et al. (2004) use a reference depth of 10 metres to avoid effects of diurnal heating in their selection criterion. For our measurements we found 5 metres was an adequate reference depth to avoid diurnal influences.

An important term in determining stability in the water column is the Monin-Obukhov length scale, defined as

\[
L = -\frac{u^3}{\kappa B}. \tag{3.12}
\]

Equation (3.12) is a measure of the relative importance of wind forcing to buoyancy forcing and is negative for destabilizing conditions and positive for stabilizing conditions. The Monin-Obukhov length is often compared with the mixed layer depth \( D \) as a bulk stability parameter in similarity scaling. Small values for the ratio \( |D/L| \) indicate stability with increasing values of \(-D/L\) leading towards greater instability and eventually overturning.

### 3.2.4 Wave Measurements

Surface gravity waves directly force the OSBL and are comprised a combination of non-locally generated swell and local wind waves. The wave measurements were made using an ultrasonic altimeter mounted at the bow. The altimeter was combined with an accelerometer to correct for ship motion and a time series of the sea surface elevation was obtained. This time series was then bandpass filtered (0.05-0.5 Hz) and consecutive half hour periods were used to calculate one dimensional surface wave spectra. For each 30 minute spectra the significant wave height \( H_S \) and the zero-upcrossing period \( T_0 \) can be approximated from the \( n^{th} \) spectral moment, defined as

\[
m_n = \int_0^\infty f^n S(f) df \tag{3.13}
\]
where $S(f)$ and $f$ are the variance spectrum and frequency respectively. Using Eq. (3.13) we can write $H_S$ and $T_0$ as

$$H_S \approx 4\sqrt{m_0}$$  \hspace{1cm} (3.14)$$

and

$$T_0 \approx \sqrt{m_0/m_2}$$  \hspace{1cm} (3.15)$$

respectively (Bouws et al., 1998). Data collected while the ship was cruising were discarded. The method has been tested successfully in previous field tests, yielding good agreement with data from a waverider buoy (Christensen et al., 2012).

![Wave Altimeter ECMWF](image)

Figure 3.4: Comparison of (a) significant wave height and (b) zero-upcrossing period for data collected with the ultrasonic wave altimeter (solid line) and the ERA-Interim reanalysis of ECMWF (dots connected by dashed line). Time is in local mean time.

In the absence of independent wave measurements we have compared our data with the ERA-Interim reanalysis of ECMWF (Dee et al., 2011). Values of significant wave height and zero-upcrossing period agree well
Wave-Turbulence Scaling in the Ocean Mixed Layer

(see Fig. 3.4). The wave model data are six hour averages over a spatial range of approximately 120 km$^2$ (i.e. 0.1° resolution) and our measurements contain significantly more variability. Our measured values are typically within +/- 10% of the wave model data for the entire cruise period.

![Wave spectra](image)

Figure 3.5: Time evolution of wave spectra over six hour time intervals.

We consider here that the significant wave height $H_S$ varied between 1.8 and 2.8 m (Fig. 3.4a), and the zero-upcrossing period $T_0$ varied between 4.5 and 6.5 s (Fig. 3.4b). On 02 July 2011 22:00 local time, the sea state was dominated by short swell with a peak period of 9.0 s. As the wind increased the sea state was later on dominated by wind waves, at 03 July 2011 01:00 the peak period had dropped to 5.8 s, but with a second peak at 8.6 s. Towards the end of the period we found mixed seas and a wide
Wave-Turbulence Scaling in the Ocean Mixed Layer

spectrum with a peak period of 7.7 s (see Fig. 3.5).

### 3.2.5 Upper Ocean Parameters

During this deployment an intense low pressure system passed over the R/V *Knorr* as can be seen in the drop and subsequent rise in atmospheric pressure (Fig. 3.6a). The leading edge of the low pressure system (Fig. 3.6a) was accompanied by moderate to heavy rainfall (Fig. 3.6c) and a uniform wind velocity of 11 ms$^{-1}$ (Fig. 3.6b). The wind speed rapidly decreased to $< 4$ ms$^{-1}$ at 2 July 21:00 LMT, accompanied by a 2 °C drop in air temperature (Fig. 3.6c) until at 3 July 04:00 LMT where the wind speed increased to $> 17$ ms$^{-1}$. The wave field changed little during the deployment with the zero-upcrossing time slightly increasing as the wind forcing decreased (Fig. 3.6d).

The net radiation (Fig. 3.7a) and buoyancy fluxes (Fig. 3.7b) show a net loss of heat and mass from the ocean to the atmosphere during the course of the deployment. The one exception to this was the salt component of the buoyancy flux (Fig. 3.7b) due to a large rain event on the leading edge of the storm.

Although the buoyancy forcing was unstable for the majority of the deployment (Fig. 3.7b), the ratio of the Monin-Obukhov length to the mixed layer depth was greater than unity (i.e. $| - L|/D \gg 1$), which implied that the wind forcing was significantly greater than the destabilizing buoyancy forcing. Only during the lull in the wind from 2 July 21:00 to 3 July 05:00 LMT does the ratio of $-L/D < 1$, indicative of a convective overturning regime.

The temperature (Fig. 3.8a) and salinity (Fig. 3.8b) from ASIP indicates that the deployment began with relatively warm and fresh water with the latter due to the rain at the beginning of the deployment (Fig. 3.6c). Overnight the upper 20 metres were relatively homogeneous with little
Figure 3.6: (a) Pressure (Pa), relative humidity (RH), (b) wind speed at 15.5 m ($U_{15}$), wind direction (Dir, clockwise from North), (c) air temperature (Ta), sea surface temperature (SST), rainfall rate (Rain), (d) significant wave height (and mean crossing period are shipboard measurements from the R/V *Knorr* for the deployment location in Fig. 3.1. The shaded region corresponds to the time when ASIP was profiling.

variation in the mixed layer depth. At 05:00 the wind rapidly increased (Fig. 3.6b) which resulted in an increase of the mixed layer depth from 15 metres to 40 metres in about 0.5 hours.

The response of the upper mixed layer to atmospheric buoyancy and wind stress (Fig. 3.9b) can be seen in the evolution of $\epsilon$ (Fig. 3.9d). In the mixed layer, the regions of high turbulent dissipation follow very closely with the mixed layer depth except during the evening when the wind is calm. The most likely cause of the discrepancy was due to inadequacies of using a single mixed layer depth parameterization to encompass all the various
Figure 3.7: (a) Net radiative flux (Q), short wave (SW), sensible heat (SH), latent heat (LE), and net infrared (IR) radiation during the deployment. Positive is upwards out of the ocean. The net surface buoyancy flux (B) and the relative contributions from heat and salt and shown in (b). (c) shows the mixed layer depth (D) and the Monin-Obukhov length (L).

forcing conditions encountered. The stratification in the mixed layer was complex (Fig. 3.9c) and often there were small gradients in the upper few metres which were strong enough to inhibit turbulence under mild forcing. Also, the use of a reference depth of 5 metres limits values for the mixed layer depth to be $D \geq 5 \text{ m}$.

The turbulent Langmuir number, defined as $L_{t} = \sqrt{u_{s}/u_{s0}}$, is shown in Fig. 3.9a. $L_{t}$ is used as an indicator for when Langmuir circulation begins to be the dominant mechanism for shear driven turbulence in the OSBL. Values of $L_{t} < 0.3$ are expected to correspond with Langmuir dominated turbulence Grant and Belcher (2009). The wave field does not vary much
Figure 3.8: (a) Temperature, (b) salinity, and (c) potential density as measured by the ASIP profiler. The solid black line denotes the mixed layer depth and the time has been corrected to local mean time.

over the deployment (see Fig. 3.6d) thus \( L_a \) closely follows the wind stress in Fig. 3.9b. During the evening \( L_a \) drops from about 0.4 to 0.2 indicating that Langmuir generated turbulence to be dominating. However, during this period (from around 2 July 22:00 to 3 July 2011 04:30 LMT) most of the turbulence is restricted to the upper 10 metres (see Fig. 3.9d) and it is not clear if there is any enhancement in this region. This shallow region of large turbulence levels is contrary to what would be expected from Langmuir circulations which are expected to extend to the base of the mixed layer. The wind direction is rotating throughout this lull in the wind (Fig. 3.6b) which may inhibit the formation of Langmuir cells (Van Roekel et al., 2012).
Figure 3.9: (a) Turbulent Langmuir number, (b) buoyancy and wind stress forcing, (c) Brunt-Väisälä frequency $N^2$, (d) dissipation $\epsilon$ and dissipation normalized by the law of the wall (Eq. (3.5)) (d). The solid black line in panels (c-e) denote the mixed layer depth $D$ and the dashed line in (d) and (e) is the Monin-Obukhov length $L$. All times have been converted to local mean time.

3.3 Discussion

We begin our analysis in Sec. 3.3.1 with a look at the overall energy budget in Sec. 3.3.1 by comparing the vertically integrated dissipation rate with the energy input from the wind. This is followed by our discussion of the profile shapes of $\epsilon$ by comparing these profiles with that expected from the law of the wall (Sec. 3.3.2) and from a wave-induced shear (Sec. 3.3.3).
Figure 3.10: Profiles of turbulent kinetic energy dissipation (a-c) for a particular wave spectra (d-f). Five successive profiles of $\epsilon$ taken over one hour are averaged vertically into 1 metre bins with the solid black line showing the mean and the grey shaded region the 95% confidence intervals determined using a bootstrap method. The depth dependence of $\epsilon$ is compared with Eq. (3.5) (blue line), scaling of Terray et al. (1996) (green line), and the wave scaling of Huang and Qiao (2010) (red line) using portions of the wave spectra along with Eq. (3.6) (dashed lines with colours matching the corresponding spectral region marked by I, and II in (d-f)). The red line denotes the sum of the wave scaling turbulence profiles from section I and II, i.e. the dashed orange line plus the dashed green line. The values of $H_s$ and $T_0$ for each wave spectra are computed for the entire spectra. The mixed layer depth is the black horizontal line denoted by $D$.

### 3.3.1 Integrated Energy Flux

It is often convenient to discuss dissipation in terms of the total dissipation rate in the mixed layer. This is defined as the vertically integrated dissipation
Wave-Turbulence Scaling in the Ocean Mixed Layer

rate, i.e.

$$\epsilon_I = \int_{-D}^{0} \rho \epsilon dz,$$  \hspace{1cm} (3.16)

where the units of $\epsilon_I$ in Wm$^{-2}$. Equation (3.16) can be conveniently used to compare the total energy input from the surface wind field. The input wind power may be estimated from the wind speed reference to 10 m height, i.e.

$$E_{10} = \tau U_{10},$$  \hspace{1cm} (3.17)

where $\tau$ is the wind stress and $U_{10}$ is the wind speed referenced to 10 metres. Comparing Eq. (3.17) with Eq. (3.16) can demonstrate what percentage of the wind power was going into mixing the OSBL. Most of the energy flux in the lower atmosphere is dissipated in the air before it ever reaches the surface so this ratio is expected to be small. Using a combination of field and laboratory results, Richman and Garrett (1977) estimated that 4-9% of $E_{10}$ would be dissipated in the mixed layer. Later direct measurements of $\epsilon$ by Oakey and Elliott (1982) found $\epsilon_I$ to be 1% of $E_{10}$. Computing Eq. (3.16) for this deployment found excellent agreement with the 1 % value of Oakey and Elliott (1982) with the results shown in Fig. 3.10. The one discrepancy is during the night where a combination of an overestimation of the mixed layer depth and buoyancy-induced turbulent dissipation create a slightly greater integrated dissipation level relative to 1 % of the wind power.

This result that 1% of the wind power is dissipated in the mixed layer is similar to previous observations Stewart and Grant (1962); Dillon et al. (1981) and a shear-driven wall layer where $\epsilon$ follows Eq. (3.5). However, there are many cases (Kitagorodskii et al., 1983; Greenan et al., 2001) where this ratio is closer to the 4-9 % predicted by Richman and Garrett (1977) suggesting a greater input of energy into mixing the upper ocean. Anis and Moum (1995) generally found the total dissipation of turbulent kinetic energy to be consistent with Oakey and Elliott (1982), but found a few occasions where integrated dissipation was much closer to 10 % suggesting that this may not be a constant value.
3.3.2 Dissipation Rate Scaling - Law of the Wall

The measured turbulent dissipation rate from ASIP was compared to the estimates from Eq. (3.5), with this ratio shown in Fig. 3.9e. For the majority of the deployment $\epsilon$ scales within an order of magnitude with Eq. (3.5) in the mixed layer. The exception to this was during the night where $\epsilon$ was larger than expected from Eq. (3.5) between 2 and 15 metres. As the night progressed, starting at 3 July 2012 03:00 LMT, there was a region of enhanced $\epsilon$ in the near surface water that was slowly extending deeper into the mixed layer depth. This time corresponded with a region of small -$L$ (Fig. 3.6c) as well as $L_{\epsilon} < 0.3$, which indicates that conditions were favourable for turbulence enhancement from convective overturning as well as Langmuir circulation.

3.3.3 Wave Induced Turbulence

Profiles of $\epsilon$ were also modelled using Eq. (3.6) as proposed by Huang and Qiao (2010), who investigated profiles of turbulent dissipation induced by shear created by Stokes drift. This leads to a depth dependence of $\epsilon \propto e^{2kz}$ where $k$ is the wave number of the wave field. Figure 3.11a-c show the wave scaling for dissipation profiles averaged over one hour in time (i.e. 5 successive profiles) and one metre in depth. The grey shaded region represents the 95 % confidence intervals for each bin using the bootstrap method (Efron and Gong, 1983).

In an attempt to differentiate between locally generated wind waves and swell waves the wave energy spectra (Fig. 3.11d-f) was divided into two sections (labelled I and II) and $k$ was calculated for each section from the mean spectral period for that section and the dispersion relation for deep water gravity waves, i.e. $\omega^2 = gk$ where $\omega$ is the angular frequency. The significant wave height was also calculated separately for each spectral bin. A value for $\beta$ of 0.15 was adopted for all subsequent estimates using Eq. (3.6),
Figure 3.11: Depth integrated dissipation rate in the OSBL as a function of wind speed at 10 metres. Panel (a) shows the integrated dissipation as a function of input wind energy while panel (b) shows this quantity as a function of local time. The individual profiles are averaged into 1 hour by 1 metre bins and the error bars represent the limits of the integrated 95% confident limits as determined with a bootstrap method.

consistent with the findings of Huang and Qiao (2010).

Several profiles of $\epsilon$ (Fig. 3.11a-c) are compared with observations including Eq. (3.5), Terray et al. (1996), and the wave scaling of Huang and Qiao (2010). For the scaling of Terray et al. (1996) II is used to determine the wave age and hence the depth of the transitional layer, $z_t$. Any errors associated with this choice for the wave age will only affect the depth of the transitional layer and will not affect the depth dependence of $\epsilon$. Early in the deployment profiles were taken in steady wind forcing and developed seas ($c_w/u_{*a} \approx 80$). In this case $\epsilon(z)$ follows an exponential depth dependence
(Fig. 3.11a, solid red line) even below the mixed layer depth suggesting that sheared generated by the Stokes drift of swell with long wavelengths may be a mechanism for mixing below the mixed layer depth, even though these dissipation levels are relatively low (see Kantha, 2006; Ardhuin and Jenkins, 2006).

The profile of $\epsilon$ during the night when the wind dropped to below 5 ms$^{-1}$ is shown in Figure 3.11b. The wave age is still $\approx 80$ but the direct wind forcing has died down and the buoyancy forcing, which is on the order of $10^{-8}$ m$^2$s$^{-3}$ is now comparable to measured dissipation rates in the mixed layer. Between 5 and 18 metres the slope of $\epsilon$ follows the wave induced dissipation profiles accurately, but above this $\epsilon$ decays more closely to that predicted by Terray et al. (1996). This is likely as a result of the low dissipation levels encountered during the night.

When the wind increased at the end of the deployment, the dissipation profile became less continuous with more discrete jumps in $\epsilon$, as shown in Fig. 3.11c. While the wave field was developing (wave age $\approx 10$) the high winds appeared to reveal a more incremental approach to a rapidly decreasing mixed layer, as is seen in the order of magnitude drops in $\epsilon$ at $\approx 15$ m, $\approx 27$ m and $\approx 32$ m (Fig. 3.11c). The dissipation rate between 2 metres and the remnant mixed layer depth of 15 metres had a near uniform dissipation rate with little variability as denoted by the confidence intervals. There were subsequent drops in $\epsilon$ at 27 metres and 32 metres with near uniform values of $\epsilon$ in between suggestive of incremental steps in eroding the mixed layer. None of the dissipation models do particularly well in this scenario with an increasing wind with Eq. (3.5) appearing to be the best in the upper 10 metres and Huang and Qiao (2010) faring better below.
3.4 Summary

Measurements of the dissipation rate of turbulent kinetic energy along with measurements of atmospheric fluxes and wave spectra are presented in detail for a field campaign in the north Atlantic during July 2012. Accurate observations of all of these parameters simultaneously in the open ocean are extremely rare, especially in the presence of mixed seas where it is difficult to distinguish between the swell and wind generated waves. Dissipation measurements were made with the Air-Sea Interaction Profiler (ASIP), which is an unique instrument designed for profiling the mixed layer of the ocean. ASIP is a vertically rising profiler which functions autonomously allowing the profiler to be sufficiently far away from any ship induced effects to allow for measurements up to the ocean surface. Direct measurements of $\epsilon$ were obtained during the transit of an intense low pressure system allowing for a wide range of sea states to test various scaling laws.

The results were used to test various scaling laws proposed for the depth dependence of $\epsilon$. Specifically, the classic wall layer where $\epsilon \propto z^{-1}$, the scaling of Terray et al. (1996) who found a transitional layer with $\epsilon \propto z^{-2}$ and the scaling proposed by Huang and Qiao (2010) who use the shear induced by the Stokes drift from surface waves. For the conditions encountered, the wall layer scaling in the mixed layer was within an order of magnitude of our $\epsilon$ estimates with a tendency to overestimate the observed dissipation. Although the scaling of Terray et al. (1996) has been confirmed previously in the ocean with the presence of swell (Drennan et al., 1996; Gerbi et al., 2009) in cases where the wave field could easily be separated into wind and swell components. Results from ASIP, the ultrasonic wave altimeter and the meteorological measurements suggest that the scaling of Terray et al. (1996) does not match the data well. This result was also reached in a similar experiment by Greenan et al. (2001) where observations deviated from the Terray et al. (1996) scaling when the sea becoming mixed and
there was no clear separation between wind and swell. Our results, early in the deployment when the wind and waves were more constant, coincided well with the exponential depth dependence proposed by Huang and Qiao (2010). However, the scaling proposed by Huang and Qiao (2010) did not agree as well with our observations during the evening with low winds and swell present nor later in the experiment during the rapid increase in wind speed.

Estimates of the integrated dissipation in the mixed layer indicated that ≈ 1 % of the wind power at 10 metres is dissipated in the mixed layer. This is identical to the results of Oakey and Elliott (1982) and consistent with results that found the theory of a wall layer to hold in the mixed layer (Soloviev et al., 1988).

Understanding how energy is dissipated in the OSBL is fundamental to accurate parameterizations of these processes. There has been some excellent advances in modelling $\epsilon$ in the presence of breaking surface gravity waves (Craig and Banner, 1994; Burchard, 2001) and Langmuir circulations (Grant and Belcher, 2009; Teixeira, 2012; Janssen, 2012), but often these models are presented with largely varying empirical coefficients to fit the limited data available. This is especially true in the open ocean where measurements of $\epsilon$ in the upper few metres and free from ship contamination are very hard to obtain.

Due to the intermittent nature of turbulence any scaling law will always have some limitations when comparing snapshots of the turbulent dissipation in the upper mixed layer. Taking this into consideration, along with a dearth of accurate dissipation estimates in the mixed layer of the open ocean, make attempts at parameterizing the profile of $\epsilon$ very challenging. More comprehensive data sets under various sea states and conditions are necessary to determine the conditions under which certain scaling can and may hold true.
Chapter 4

Evaluating Langmuir
Turbulence Parameterizations
in the Ocean Surface Boundary Layer

Preface

This chapter is an adapted reproduction of a paper published in *Journal of Geophysical Research: Oceans* with co-authors Brian Ward and Kai Christensen:


The right to reuse this work was retained by the authors and nonexclusive copyright was granted to the American Geophysical Union. The data analysis, the interpretation and synthesis of results, the production of figures and

64
the writing were done exclusively by the author of this thesis. Dr. Ward contributed by supervising, assisting and reviewing the work and by providing the infrastructure (computer time, ship-time, instrumentation, etc.) required to carry out this research project. Data collection was jointly done by the author and Dr. Ward. Dr. Christensen contributed measurements of the wave field and assisting and reviewing the work.

4.1 Introduction

The exchange of heat, trace gases, and momentum from the atmosphere to the ocean occurs over a region near the surface often referred to as the ocean surface boundary layer (OSBL). The OSBL is defined as that part of the upper ocean with a nearly uniform density, active vertical mixing, and high rate of dissipation. The ability of the ocean to buffer variability from the atmosphere, through the storage of heat and trace gases into the deep ocean, is predominantly controlled through the dynamics and the depth of the OSBL. The OSBL also imposes the boundary conditions on the dynamics and stratification of the deep ocean thus regulating more permanent storage of heat by the world’s oceans, which is a critical part of the short and long term dynamics of global climate (Belcher et al., 2012, (henceforth B12)).

The depth of the OSBL is determined from the integrated effects of destabilizing processes which work to mechanically mix the water (e.g. wind induced shear, wave breaking, converging fronts, radiative heat loss) and stabilizing processes due to surface buoyancy flux (e.g. rainfall or surface heating). The depth of the OSBL over the global oceans varies from a few meters under a large stabilizing buoyancy flux up to 500 m in the winter in sub polar latitudes (de Boyer Montégut et al., 2004). Accurate determination of the depth of the OSBL has large implications on short and long term ocean circulation, heat uptake, and oceanic flora and fauna, therefore it is important to understand the processes which regulate this region.
Most parameterizations assume that the depth of the OSBL is actively being mixed, i.e. that the mixed layer is equal to the mixing layer. Differences between the mixed and mixing layers will depend partially on the definition for each (see Stevens et al. (2011) for a review), but this does not seem to account for all the variability observed between the two (Brainerd and Gregg, 1995; Cisewski et al., 2008). Certain processes may have a mixing layer depth either greater than (corresponding to entrainment or regions with large shear across the pycnocline) or less than (usually due to a sudden decrease in surface forcing) the mixed layer depth. Simultaneous observations of the mixed and mixing layer depths require specialized instruments and there are relatively few studies which have differentiated between mixed and mixing layer depths.

Although recent parameterizations, which include the effects of submesoscale eddies (Fox-Kemper et al., 2008), have greatly reduced global biases in the OSBL depth from climate models, there still exist many regions where the depth of the OSBL is poorly resolved (Fox-Kemper et al., 2011) compared with observations (de Boyer Montégut et al., 2004). It has been argued that current climate models are lacking key surface-wave processes in their OSBL dynamics which is leading to systematic biases in certain regions of the world oceans (B12).

Wind blowing over the sea surface generates turbulence through the direct action of wind stress on the surface and indirectly through surface gravity waves via two mechanisms: wave breaking and interactions between the wave induced Stokes current and the wind-forced surface shear. The first mechanism, breaking waves, are expected to be a major source of enhanced dissipation near the ocean surface (Agrawal et al., 1992), the dissipation levels are expected to decay rapidly with depth and have a significantly reduced impact on turbulence levels in the bulk of the OSBL relative to the surface. The second mechanism, the interaction between the Stokes drift and the wind-forced surface shear, leads to the wave field causing the
vertically aligned wind-induced vortices to align into the down-wave direction leading to an instability in the wind-forced shear (Craik and Leibovich, 1976). This vortex force, known as the Craik-Leibovich vortex force (type 2), or CL2 for short, is known to generate Langmuir circulations (LC), named after Langmuir (1938) who first observed these horizontally counter-rotating vortices and can be recognized by patterns of surface convergence and downwelling. These Langmuir cells are expected to be transient and require detailed observations to resolve the vortical velocity to determine if Langmuir circulations are prevalent (Plueddemann et al., 1996; Smith, 1998; Kukulka et al., 2009).

Due to these observational difficulties, studies on LC-generated turbulence have been predominantly restricted to Large Eddy Simulations (LES). These have either assumed a monochromatic wave forcing for LC in the presence of wind and convective forces (Skyllingstad and Denbo, 1995; Li et al., 2005), or have calculated LC with a modelled wave spectrum in the absence of convection (Sullivan et al., 2007; Harcourt and D’Asaro, 2008; Grant and Belcher, 2009; McWilliams et al., 2012). Langmuir circulations are simulated through the LC vortex force, plus an additional dynamic pressure term from the Stokes drift, in an idealized OSBL of uniform density. These studies have led to parameterizations involving La as an indicator for LC and enhanced turbulence in the OSBL.

There are relatively few studies which simultaneously measure surface waves and turbulence in the OSBL (e.g. Plueddemann et al., 1996; Greenan et al., 2001; Huang et al., 2012; Sutherland et al., 2013) and it remains unclear the extent to which surface gravity waves impact the OSBL dynamics. This leaves a large gap in verifying the abundance of numerical simulations with accurate observations.

Recently, D’Asaro et al. (2014) found the LES results of Harcourt and D’Asaro (2008) to be consistent with detailed measurements of turbulence and the wave and wind field at two sites: Ocean Station Papa (ocean
weather station (OWS)-P 50°N, 145°W and Lake Washington near Seattle, Washington. Turbulence levels in the ocean were equated with the turbulent vertical velocity variance as measured with a neutrally buoyant Lagrangian profiler D’Asaro (2003). They estimated the depth of the OSBL as twice the average depth of their Lagrangian profiler as Lagrangian particles are expected to be evenly distributed in the OSBL. This depth would correspond to the active mixing layer and may be different than the mixed layer in many parts of the ocean (Brainerd and Gregg, 1995; Cisewski et al., 2008).

Presented here are detailed observations of the OSBL in the North Atlantic from the Knorr_11 experiment (Fig. 4.1) during July and August 2011 (see also Bell et al., 2013; Sutherland et al., 2013). Details of the observations include profiles from an upwardly rising microstructure profiler in the OSBL, surface waves, and atmospheric forcing, over a broad range of conditions are explained in Sec. 4.2. The evolution of turbulent kinetic energy (TKE) in the OSBL is explained in Sec. 4.3 along with comparisons between observations with the parameterizations of B12 along with depth-dependent parameterizations of the “law of the wall” and Grant and Belcher (2009). The depth dependence of turbulent dissipation as a function of La is explored in Sec. 4.4. A summary of our results can be found in Sec. 4.5.

4.2 Observations

Ocean measurements of microstructure shear, temperature, and salinity were obtained with the Air-Sea Interaction Profiler (ASIP), an autonomous, ascending, microstructure profiler specifically designed to measure turbulence properties in the OSBL (for further details see Sutherland et al., 2013). These data provide estimates of the turbulent dissipation rate $\epsilon$ and hydrographic parameters such as temperature and salinity. Four ASIP deployments, consisting of 283 profiles, were made during Knorr_11 (Fig. 4.1). Profiles of $\epsilon$ were calculated using 1 second bins, corresponding to a vertical resolution
of about 0.5 m, and averaged over three successive profiles to obtain mean values approximately every hour. Figure 4.2 shows atmospheric and wave observations obtained during Knorr_11 with the ASIP deployments denoted by the shaded regions and numbered along the top panel.

![Figure 4.1: Location of each of the four ASIP deployments, numbered 1 to 4, during the Knorr_11 campaign.](image)

The surface buoyancy flux is calculated using $B_s = -gQ_p$ where $g$ and $Q_p$ are the acceleration due to gravity and density flux respectively. The density flux is a combination of heat and salt inputs and is defined as $Q_p = \rho_s (\alpha F_T + \beta F_S)$ where $\alpha$ and $\beta$ are the thermal expansion and saline contraction coefficients respectively and $\rho_s$ is the density at the ocean surface. Here $F_T = -Q_{net}/\rho_s C_p$ and $F_S = (E - P) S/(1 - S/1000)$ where $C_p$ is the specific heat of sea water and $E$, $P$ and $S$ are the evaporation, precipitation and sea surface salinity respectively. The net radiative heat flux at the ocean surface (Fig. 4.2a) is calculated from the combination of the incoming shortwave ($SW$), net incoming and emitted longwave ($IR$), sensible heat ($SH$) and latent heat ($LE$), i.e. $Q_{net} = SW + IR + SH + LE$. Positive $Q$ denotes an increase in surface density and hence unstable conditions, while a negative $Q$ denotes a decrease in surface density and stable conditions.
Shortwave and longwave components were measured from the deck of the R/V *Knorr* (S. Miller 2012, personal communication) while both *SH* and *LE* were computed using the TOGA COARE 3.0 flux algorithm (*Fairall et al.*, 2003).

Figure 4.2: Time series of the collected meteorological data including a) radiative fluxes including the total (Q), shortwave (SW), sensible heat (SH), latent heat (LE) and infrared (IR), b) wind speed and direction, c) significant wave height (*H*ₜ), zero up-crossing period (*T*ₚ) and period of the peak of the spectrum (*T*ₚ), d) the Stokes drift (*u*ₚ₀) divided by 10, and the convective velocity (*u*ₚ) and friction velocity (*u*ₚ) respectively, e) turbulent Langmuir number (La) and wave age (*C*ₚ/₁₀), and e) the mixed layer (*h*ₚ) and mixing (*h*ₑ) layer depths. The periods where there are ocean microstructure measurements are shaded grey and wave measurements are only shown for when the ship was on station. The dashed red and blue lines in e) correspond to La = 0.35 and *C*ₚ/₁₀ = 1.2 respectively.
Wind measurements, shown in Fig. 4.2b, were recorded using two Vaisala WXT520 weather sensors mounted at 15.5 meters above the waterline on the forward mast on both port and starboard sides respectively. Measurements are sampled at 2 Hz and these are averaged and recorded at 20 minute intervals. The wind measurements at 15.5 meters are corrected to the standard 10 meters above sea level where the wind stress is calculated using the TOGA COARE 3.0 algorithm (Fairall et al., 2003).

Observations of the surface gravity wave field in Fig. 4.2c are calculated from the 1-D spectra as measured by an ultrasonic acoustic transducer mounted on the bow of the ship. The altimeter is mounted on the end of a steel pole on the bow and measures the sea surface elevation in front of the ship. The data are corrected for ship motion (heave and pitch) using data from an accelerometer mounted immediately above the altimeter. Wave measurements are only shown for the times when the R/V Knorr was stationary to avoid any affects due to a Doppler Shift of the signal when the ship is in transit (for a detailed description see Christensen et al., 2013).

Figure 4.2d shows the dominant velocity scales of the OSBL: Stokes drift ($u_s$, red line), friction velocity ($u_*$, blue line) and the convection velocity ($w_*$, green line). The friction velocity is related to the wind stress, $\tau$, such that $\tau = \rho u_*^2$ where $\rho$ is the density of sea water and $\tau$ is computed from the wind speed using the TOGA COARE 3.0 algorithm. The convection velocity is defined as $w_* = (B_s h)^{1/3}$ where $h$ is the depth of the OSBL and only exists in destabilizing conditions, i.e. when $B_s > 0$.

The surface Stokes drift is calculated from the 1-D wave spectra (Kenyon, 1969) as $u_s = \frac{16\pi^3}{g} \int_{f_{\text{min}}}^{f_c} f^3 S(f) df$ where $f_{\text{min}}$ and $f_c$ are 0.05 and 0.40 Hz respectively (Christensen et al., 2013). There are two opposing uncertainties with calculating the Stokes drift from a 1-D wave spectra using a finite frequency range; the lack of a measured high frequency spectral tail and the lack of the directional spreading of the wave energy. The former will lead to a systematic underestimate of $u_s$ up to 30% (Rascle et al., 2006),
with the exact amount dependent on the slope of the spectral tail, while the latter leads to a systematic overestimate of $u_{s0}$ up to 30%, (Webb and Fox-Kemper, 2011) dependent on the directional spread of wave energy. Lacking measurements of the high frequency component of the wave spectrum and directional information we assume that the two assumptions should closely cancel each other and note that this could lead to a maximum uncertainty in $u_{s0}$ of 30%, but that the true error is most likely much less. Error estimates for the surface Stokes drift have been estimated using data from a previous cruise when we had access to independent measurements from a waverider and were found to be 14.5%.

Figure 4.2e shows the turbulent Langmuir number, which is often used as an indicator for LC, with LC strength increasing with decreasing $L_a$ (McWilliams et al., 1997). For well developed seas $L_a \approx 0.35$ (Fig. 4.2e, red dashed line), although B12 found $L_a \approx 0.4$ when frequencies greater than $f_c$ were not included in calculating $u_{s0}$. Misalignment between $u_*$ and $u_{s0}$ is expected to broaden the observed range of $L_a$, but without directional wave measurements we adopt the assumption, as does B12, that $u_*$ and $u_{s0}$ are aligned. Figure 4.2e also shows the wave age $C_p/U_{10}$, where $C_p$ is the phase velocity of the peak of the wave spectrum, is equal to 1.2 for well developed seas (Donelan et al., 1985, Fig. 4.2e, blue dashed line), where $C_p/U_{10} < 1.2$ denotes young wind seas and $C_p/U_{10} > 1.2$ indicate older swell dominated seas.

The mixed ($h_\rho$) and mixing ($h_\epsilon$) layer depths are shown in Fig. 4.2f as the red and blue lines respectively. The mixed layer depth $h_\rho$, as calculated using a density threshold criterion of 0.03 kg m$^{-3}$ (de Boyer Montégut et al., 2004) relative to the density at 5 m depth, is shown in red. The other length scale of importance is the mixing layer depth, denoted $h_\epsilon$ (Fig. 4.2f, blue line), which is the region of the OSBL actively being mixed and is defined here as the depth at which $\epsilon$ falls to a background level of $10^{-9}$ m$^2$s$^{-3}$, which is similar to previous definitions used for the mixing layer (Dewey and Moum,
4.3 Evolution of the OSBL

The evolution of the OSBL is directly influenced by a combination of buoyancy and mechanical mixing at the air-sea interface. Although there are regions where lateral mixing may be a factor in OSBL dynamics (Fox-Kemper et al., 2008) we shall assume a horizontally homogeneous flow and thus consider only the direct effects of atmospheric forcing on the OSBL.

The time evolution of the TKE equation, where the TKE is defined as $q^2 = 0.5(u'^2 + v'^2 + w'^2)$, for horizontally homogeneous flow in the constant stress layer of the OSBL, as given by B12, is

$$\frac{Dq}{Dt} = -\left(\partial \frac{\bar{u}h}{\partial z} \right) - \left(\frac{w'}{w'} \frac{\partial u_h}{\partial z} \right) + \left(\frac{w'v'}{w'} \frac{\partial u_s}{\partial z} \right) + \frac{w'v'}{b'} \frac{\partial u_s}{\partial z} - \frac{\partial}{\partial z} \left( \frac{1}{2} gw' + \frac{1}{\rho_0} w'p' \right) + \epsilon \, .$$

(4.1)

Here we assume that each variable may be written as a mean and fluctuation part, such as

$$a = \bar{a} + a'$$

(4.2)

where $\bar{a}$ and $a'$ are the mean and fluctuating part of $a$ with the properties $\bar{a} = \bar{a}$ and $\bar{a'} = 0$ where the operator denotes the time average. Furthermore, we assume the wave field is propagating in the x-axis direction and that the mean flow is in the horizontal (i.e. x and y) axis. The time evolution of the TKE is a combination of (a) shear production (b) Stokes shear production, (c), buoyancy flux, (d) TKE transport and work by pressure and (e) the dissipation of turbulence.
4.3.1 Evaluating the Regime Diagram of Belcher et al. (2012)

Using dimensional arguments B12 derived an expression for the turbulent dissipation rate $\epsilon_B$ in terms of the velocity scale $u_*$, length scale $h$ such that

$$\epsilon_B = \frac{u_*^3}{h} f \left( \frac{z}{h}, L_a, \frac{h}{L_L} \right)$$  \hspace{1cm} (4.3)

where $f \left( \frac{z}{h}, L_a, \frac{h}{L_L} \right)$ is a universal function. Here $L_L = -u_*^2 u_s / B_s$ is the Langmuir stability length as coined by B12, which can be related to the Monin-Obukhov length (while omitting the von Kármán constant) through $L_m = L_L L_a^{-2}$. Assuming $\epsilon_B$ to be a linear combination of wind, wave and buoyancy forcing at $z = h/2$ (which is an arbitrary depth where the three forms of turbulence were considered to be well established by B12), Eq. (4.3) becomes

$$\frac{\epsilon_B (z/h = 0.5)}{u_*^3/h} = A_s + A_L L_a^{-2} + A_c L_a^{-2} \frac{h}{L_L}.$$  \hspace{1cm} (4.4)

Here B12 derived the constants $A_s = 2 \left( 1 - e^{-L_a/2} \right)$ and $A_L = 0.22$ from LES results of Grant and Belcher (2009) and $A_c = 0.3$ from simulations of the atmospheric boundary layer (Moeng and Sullivan, 1994).

To compare Eq. (4.4) with observations we will use two definitions for $h$: the mixed layer depth $h_\rho$ and the mixing layer depth $h_\epsilon$. Observations of $\log_{10} \epsilon/\epsilon_B$ measured at $z = h/2$ for the mixed and mixing layer depths are shown in Fig. 4.3a and Fig. 4.3b respectively. The contour lines show the regime diagram of B12 as calculated using Eq. (4.4). Observations of $\epsilon$ are predominantly within an order of magnitude of $\epsilon_B$ with the distribution of $\log_{10} \epsilon/\epsilon_B$ shown in Fig. 4.4. Using the mixed layer depth $h_\rho$ gives a negative bias and larger standard deviation relative to $\epsilon_B$ than using the mixing layer depth $h_\epsilon$. 

74
4.3.2 Evaluating Different Regimes

To differentiate between the wind-wave and convective regimes we use the ratio $h/L_L$ to investigate separately the extreme cases when convective forcing is strong or not. The cutoff between wind-wave forcing and convective forcing is chosen to be $|h/L_L| = 1$ as this is similar to the strong convective regime ratio of $h/L_{mo} \approx 10$ of Lombardo and Gregg (1989) assuming $\text{La} \approx 0.3$.

Figure 4.5 shows profiles of the normalized dissipation measurements where the depth is normalized by the mixed (a) and mixing layer (b) respectively. The convective and wind-wave regimes are shown in red and blue respectively and the geometric mean, calculated at normalized depth...
Evaluating Langmuir Turbulence in the OSBL

Figure 4.4: Histogram of $\log_{10} \epsilon/\epsilon_B$ from Fig. 4.3. The blue and red lines denote the Gaussian curve calculated from the mean $\mu$ and standard deviation $\sigma$ for the $h_\rho$ and $h_\epsilon$ cases respectively.

intervals of 0.05, is shown by the black line. The stabilizing regime is also shown in Fig. 4.5 (green dots), but this data is not used in order to restrict our analysis to events which lead to a deepening of the OSBL. Although mean dissipation is similar to that expected from the law of the wall the variability spans several orders of magnitude with slight differences in the mean profile shape. However, there is a clear distinction between the neutral buoyancy forcing and the strong unstable forcing with the unstable forcing corresponding to consistently greater dissipation rates.

Figure 4.6 is similar to Fig. 4.5, but with $\epsilon$ normalized by $u_0^2 u_s / h$, which was the scaling presented by Grant and Belcher (2009) with the dashed line showing their LES results (from their Fig. 5). The depths at which using the mixed and mixing layer depths give differing results in both Fig. 4.5 and Fig. 4.6 are predominantly limited to $0.2 < z/h < 1$. Although the mean values appear to collapse onto the curve of Grant and Belcher (2009) there is still significant scatter which appear to be grouped into a wind-wave regime and a buoyancy regime respectively.

76
Evaluating Langmuir Turbulence in the OSBL

Figure 4.5: Dissipation values normalized by $u^3_*/z$ at depths normalized by the a) mixed and b) mixing layer depths respectively. The red dots denote unstable buoyancy forcing, the blue dots denote neutral buoyancy forcing, and the green dots denote stable buoyancy forcing. The solid line is the geometric mean calculated at depth intervals of $0.05z/h$ and the dashed line corresponds to $\epsilon = u^3_*/z$.

4.4 Depth Dependence of $\epsilon$ in the OSBL

Although it has proven effective to evaluate $\epsilon$ at one depth in the atmospheric boundary layer (Moeng and Sullivan, 1994, B12) this is most likely to be insufficient given our current understanding of turbulent dissipation OSBL (Anis and Moum, 1995; Terray et al., 1996; Sutherland et al., 2013). Specifically, the wind, wave and buoyancy terms of Eq. (4.1) have different depth dependencies (Dillon et al., 1981; Agrawal et al., 1992; Terray et al., 1996).

Away from the ocean surface, where turbulence is not directly influenced by wave breaking, it is expected that turbulence from wind generated shear
should scale as $\epsilon \propto z^{-1}$. As far as wave generated turbulence is concerned, there is first the contribution from Langmuir turbulence, which is expected to scale as a function of the mixed layer depth (Grant and Belcher, 2009; Teixeira and Belcher, 2010). Secondly, it can be shown using generalized Lagrangian-mean theory (e.g. Ardhuin and Jenkins (2006)), that the waves give rise to a shear production term in the TKE equation, which for deep water waves scales as the Stokes drift shear, i.e. $\epsilon \propto (1/k)e^{-2k|z|}$, where $k$ is the wave number. The exact profile in the latter case will then depend on the shape of the wave spectrum. It is not yet clear how these mechanisms should be distinguished in ocean models, e.g. Janssen (2012) models the combined...
effect of both mechanisms using a parameterization based on Stokes drift shear only.

### 4.4.1 Wind-Wave Regime

Figure 4.7 shows the normalized measured dissipation as a function of Langmuir number at various depths $z$ relative to $h_\rho$ when buoyancy forces are small (i.e. $|h_\rho/L_L| < 1$). The turbulent dissipation is normalized by $u^3/z$ where $z$ is selected rather than $h_\rho$ of Eq. (4.4) for the length scale, as $z$ is the expected length scale for shallow depths ($z < h_\rho/2$) (Grant and Belcher, 2009) and also for wall bounded turbulence omitting the von Kármán constant (Csanady, 1984).

![Figure 4.7: Dissipation values normalized by $u^3/z$ as a function of La for various depth intervals normalized by $h_\rho$. The colour denotes the relative ratio of $h_\rho/L_L$. Only values where $h_\rho/L_L < 1$ are used. The slope $m$ and $R^2$ statistic are written on each plot respectively.](image)

A linear fit of $\log_{10} \epsilon z/u^3 = m \log_{10} La + b$ is calculated with the corresponding slope $m$ and $R^2$ value, computed in log-log space, shown in each
Evaluating Langmuir Turbulence in the OSBL

panel of Fig. 4.7. The slope $m$ can be related to the velocity scale

$$u = u_{\infty 0}^{\frac{|m|}{6}} u_\infty^{1 - \frac{|m|}{6}},$$  (4.5)

where $-6 \leq m \leq 0$ with the limits of $m = 0$ corresponding to the law of the wall scaling $u = u_\infty$ (Dillon et al., 1981) and $m = -6$ corresponding to $u = u_{\infty 0}$ in the presence of strong Langmuir circulations (Smith, 1998). However, most models and observations of the OSBL under wave and wind forcing expect a velocity scale of either $u = (u_\infty^2 u_{\infty 0})^{1/3}$ (corresponding to $m = -2$) (Harcourt and D’Asaro, 2008; Grant and Belcher, 2009, B12) or $u = (u_\infty u_{\infty 0})^{1/2}$ (corresponding to $m = -3$) (Plueddemann et al., 1996).

At shallow depths, a $m = -2$ dependence is observed (Figs. 4.7a–c) which slowly vanishes in the mid-depths of the OSBL (Figs. 4.7d–g) only to reappear at depths near the base of the OSBL (Fig. 4.7j). The small correlation in the mid-depths may be due to uncertainties in $h_\rho$ from the definition used (de Boyer Montégut et al., 2004) or variations between the mixed layer depth and the actively mixing layer depth (Brainerd and Gregg, 1995). It is unclear why the small correlation in the middle of the mixed layer increases as $z \rightarrow h_\rho$. The coefficient of determination, $R^2$, as shown in Fig. 4.7, is low at all depths suggesting that the variance of the normalized dissipation when $h_\rho$ is used is not well explained by a simple La dependence. In the upper quarter of the mixed layer the La dependence only accounts for 15-25% of the signal variance, which quickly drops to zero in the middle of the mixed layer.

Figure 4.8 shows the normalized dissipation rate at depth intervals relative to the mixing depth $h_\epsilon$, similar to Fig. 4.7. The La dependence of the normalized dissipation is consistently between $-3 < m < -2$ for all depths greater than $0.05z/h_\epsilon$ with $R^2$ between 0.17 and 0.33. A summary of the fits of Figs. 4.7–4.8 can be found in Fig. 4.9.
4.4.2 Convective Regime

There is only one period (i.e. the first 60 profiles of deployment 2) with relatively large convective forcing, as can be seen by comparing where \( w_\ast \gg u_\ast \) in Fig. 4.2d. During this period there is little difference between the mixed and mixing layer depths so \( h_\rho \) is used to normalize the depth. The normalized turbulent dissipation rate as a function of \( h_\rho/L_L \) for convective forcing when \( h_\rho/L_L > 1 \) is shown in Fig. 4.10. A linear fit of \( \log_{10} \varepsilon z/u_\ast^3 = m \log_{10} h_\rho/L_L + b \) is calculated with the corresponding slope \( m \) and \( R^2 \) value, computed in log-log space, similar to the previous analysis with \( \text{La} \). At shallow depths there is no strong relationship between \( \varepsilon z/u_\ast^3 \) and \( h_\rho/L_L \), but when \( z > 0.25h_\rho \) it appears that \( m \approx 1 \), which is consistent with Eq. (4.4).

In Fig. 4.7 the variance of the normalized dissipation as a function of \( \text{La} \) was best explained for \( z < 0.25h_\rho \), which may suggest that wind and waves
Evaluating Langmuir Turbulence in the OSBL

Figure 4.9: A summary of the fit $m$ (a) and $R^2$ (b) for the mixed (black) and mixing (red) layer depths. The Langmuir case where $m=2$ is shown by the vertical dotted line in a.

dominate dissipation at shallow depths.

It is important to note that $L_L$ is the ratio of buoyancy forces to the composite wind-wave velocity $(u^2 u_s)^{1/3}$, so this includes the effects of waves. To determine if $L_L$ improves scaling over $L_{mo}$, the data in Fig. 4.10 are reproduced according to the criterion $h_{ρ} / L_{mo} > 3$ and shown in Fig. 4.11. For shallow depths, $εz / u^3_s$ scales poorly with $h_{ρ} / L_{mo}$. They are in fact worse than in Fig. 4.10 suggesting that the turbulent Langmuir number may be a better scaling parameter in the near surface. For $z > 0.25h_{ρ}$ there is good agreement between the normalized dissipation and $h_{ρ} / L_{mo}$ with a greater coefficient of determination than observed in Fig. 4.10.

The greater $R^2$ values may be explained if $ε \propto B_s$ as is expected for purely convective turbulence (Shay and Gregg, 1986) since both the dissipation and $B_s$ are normalized by $u^3_s$, leading to a slope $m = 1$. However, large buoyancy forcing is only observed for a small range of $La$ and more observations in

82
convective regimes are required to investigate the role that surface gravity waves play in convective regimes. A summary of the fits of Figs. 4.10–4.11 can be found in Fig. 4.9.

4.5 Summary

Presented here are observations of the turbulent dissipation rate, $\epsilon$, in the OSBL over a broad range of atmospheric and wave forcing during the Knorr_11 campaign in the North Atlantic during June and July 2011. Observations of turbulent dissipation and hydrographic parameters were made with the Air-Sea Interaction Profiler (ASIP), which is a vertically rising profiler designed to operate autonomously so to be sufficiently away from any ship-induced turbulence in the OSBL. These observations are compared with the regime diagram of B12 evaluated at the mid-depth of the
OSBL. Observations were generally lower than that predicted by B12 and spanned two orders of magnitude. However, when using the mixing layer depth (determined directly from the $\epsilon$ profiles) as the turbulence depth scale, observations were more consistent with the model.

When surface buoyancy forces are small, near surface dissipation observations are consistent with a velocity scale of $u = (u_s^2 u_0)\,^{1/3}$ as previously estimated in the OSBL from LES (Harcourt and D’Asaro, 2008; Grant and Belcher, 2009, B12). This dependence disappears deeper into the OSBL and it is unclear whether this arises from the intermittency of turbulence, uncertainties associated with calculating the mixed layer depth or the Stokes drift, or missing physics. When the depth intervals are normalized by the mixing layer depth ($h_e$), the La dependence is much more consistent with a higher coefficient of determination. The best fits of $La^m$ show $-3 < m < -2$ suggesting a velocity scale between $u = (u_s^2 u_0)\,^{1/3}$ and $u = (u_s u_0)\,^{1/2}$ with
Figure 4.12: A summary of the fit m (a) and $R^2$ (b) for the ratio of the mixed layer to the Langmuir stability length (black) and Monin-Obukhov length respectively. The classic convection case where $m=1$ is shown by the vertical dotted line in a.

the latter value being consistent with observations from (Plueddemann et al., 1996) and LES results (Grant and Belcher, 2009).

Strong destabilizing surface buoyancy flux are only present during one ASIP deployment with dissipation profiles proportional to $h_p/L_L$, which is consistent with $\epsilon \propto B_s$ as expected for purely convective driven turbulence (Shay and Gregg, 1986). This trend is consistent for depths greater than 0.25$h_p$ and explains approximately 50% of the observed variance. When the Monin-Obukhov length $L_{mo}$ is used instead of $L_L$ for the convective scaling, where the Monin-Obukhov length is essentially $L_{mo} = L_L \lambda^{-2}$ (while omitting the von Kármán constant), not only does the coefficient of determination increase from 0.50 to 0.80, but the slope also increases to a value closer to 1.5 than to 1.0. However, this is only true for $z > 0.25h_p$ as the correlation is worse at shallow depths when $L_{mo}$ is used instead of $L_L$.
Evaluating Langmuir Turbulence in the OSBL

We have shown that agreement between proposed scaling laws and observations increase when only the active mixing layer is considered instead of the mixed layer. While this is not surprising in itself, it is interesting to note how the parameterizations did not perform well when using the mixed layer and the role this difference has with regard to turbulent dissipation rate parameterizations in the OSBL. More observations in open ocean conditions are required to assess when mixing and mixed layer depths are different and the effects this has on parameterizing the turbulent dissipation rate in the OSBL.
Chapter 5

Mixed and Mixing Layer Depths in the Ocean Surface Boundary Layer: Buoyancy-Driven Conditions

Preface

The material in this chapter will be submitted to Geophysical Research Letters with co-authors Brian Ward and Gilles Reverdin. The data analysis, the interpretation and synthesis of results, the production of figures and the writing were done exclusively by the author of this thesis. Dr. Ward contributed by supervising, assisting and reviewing the work and by providing the infrastructure (computer time, ship-time, instrumentation, etc.) required to carry out this research project. Data collection was jointly done by the author and Dr. Ward. Dr. Reverdin provided contributed by assisting and reviewing the work.
5.1 Introduction

The upper ocean typically exhibits a surface mixed layer, which can range from a few metres to several hundred metres defined by a quasi-homogeneous density structure. The presence of turbulent mixing from wind, surface gravity waves, and convective cooling create a well-mixed layer with little variation in density. This surface mixed layer, also referred to as the ocean surface boundary layer (OSBL), is an important component of the global climate system as it controls the transfer of heat, momentum and trace gases between the atmosphere and the ocean (Sutherland et al., 2014; Belcher et al., 2012).

Many attempts have been made to define a strict criterion to determine the mixed layer depth (MLD). The vast majority of these use a density threshold relative to a reference depth, which defines an acceptable density variability within the OSBL usually in the range $\Delta \sigma = 0.01 - 0.03 \text{ kg m}^{-3}$ (Thomson and Fine, 2003; de Boyer Montégut et al., 2004). There are many other definitions such as defining a temperature threshold (Kara et al., 2000), a density gradient threshold (Lukas and Lindstrom, 1991), a linear optimal fitting approach (Chu and Fan, 2011), a split-merge method based on the profile shape (Thomson and Fine, 2003) or dissolved oxygen profiles (Castro-Morales and Kaiser, 2012). We hypothesis that one of the leading causes to so many definitions is that they are trying to parameterize a dynamic layer where there are remnants of mixing processes.

In the OSBL, the primary mechanism controlling the vertical transfer of heat, momentum, and material in the OSBL is turbulent mixing and not necessarily density. Therefore, a more important parameter in OSBL dynamics is the mixing layer depth (XLD) where there exists active turbulent mixing. Although this may appear to be a minor distinction, as the homogeneous density distribution within the OSBL is a direct result of high levels of turbulent mixing, it is an important distinction in relation to the
response of processes of the OSBL (Croot et al., 2007).

Observations of differences between the mixed layer and the mixing layer in the ocean have been commented on in early microstructure experiments (Shay and Gregg, 1986; Dewey and Moun, 1990). However, there have been relatively few studies that have investigated the nature of this difference, as it requires specialized instruments to quantify turbulence. Brainerd and Gregg (1995) was one of the first studies to use microstructure profiles from the subtropical Pacific to distinguish between MLDs and XLDs. Using direct measurements of the rate of dissipation of turbulent kinetic energy, $\epsilon$, and Thorpe scales, $L_T$ (Thorpe, 1977), they calculated the mixing layer using the Advanced Microstructure Profiler (Moun et al., 1995). However, Thorpe scales should be used with care in the OSBL for a couple of reasons: first is the issue of resolution as the OSBL is quasi-homogeneous; Cisewski et al. (2008) demonstrated that the microstructure profiles accurately resolve overturns in the OSBL, although this may not be the case under more homogeneous conditions. Second, and more crucially, is that the Thorpe scale assumes the reordered profile is “the level of origin of the unstable fluid” (Thorpe, 1977), which in the case of the OSBL under convection, is clearly not the case. In short, it is not turbulence which creates a density instability; it is the buoyancy loss at the ocean surface. It is unclear if this distinction is important for relating $L_T$ with $\epsilon$ and hence mixing, but it would suggest that $L_T$ may not be an ideal proxy for turbulence in this case. Brainerd and Gregg (1995) also use $\epsilon$ as a proxy for the mixing layer and this seems like a better choice since it is directly related to the turbulent kinetic energy in the OSBL even when the OSBL is completely homogeneous and hence no mixing involved. The mixing layer depth coincided with density differences from the surface value ranging from 0.005-0.5 kg m$^{-3}$ with no consensus for an ideal choice for the threshold.

There have been a few more recent studies that have investigated mixed and mixing layer depths at higher latitudes. During a transect across the
Mixed and Mixing Layer Depths in the OSBL

north Atlantic at approximately 52°N, Lozovatsky et al. (2006) defined the mixed layer using a density threshold of $\Delta \sigma = 0.02\sigma_0$ where $\sigma_0$ is the surface density and the mixing layer based on $\epsilon$ falling to a background dissipation rate of $\epsilon_b = 10^{-8}m^2s^{-3}$. The observed difference between the mixed and mixing layer depths was minimal across a transect of 42 stations. However, with generally only 2-3 profiles per station (although sometimes 7-8) it is impossible to say anything about the evolution of the mixed layer which would be more closely related to differences between the MLD and XLD. In the sub-arctic around Svalbard, Fer and Sundfjord (2007), using a mixed layer defined by the split-merge method of Thomson and Fine (2003) and a mixing layer using $\epsilon_b = 3 \times 10^{-8}m^2s^{-3}$, found a mixing layer to be greater than the mixed layer for the majority of their profiles. However, this study consisted of a total of only 81 profiles with 23 of these in ice-covered conditions. In the proximity of an Antarctic Polar Front, Cisewski et al. (2008), using a mixed layer threshold of 0.02 kg m$^{-3}$ and the same definition for the mixing layer as Lozovatsky et al. (2006), found mixing layers substantially less than the mixed layer.

Using an ocean general circulation model (OGCM), Noh and Lee (2008) directly compared values of mixed and mixing layer depths on a global scale. Two different density thresholds were tested, 0.1 and 0.02 kg m$^{-3}$, to determine the mixed layer depth from the OGCM. The mixing layer depth, determined directly from the vertical eddy diffusivity $K_{\rho}$, drops to a background value of $10^{-5}m^2s^{-1}$ for the first time. It was found that $h_{\epsilon} > h_\sigma$ in regions where strong subsurface shear is present, such as the equatorial ocean and western boundary current regions, and $h_{\epsilon} < h_\sigma$ during early restratification and at high latitudes during convective cooling. In addition a zonal dependence for $\Delta \sigma$ was observed by minimizing the difference between the OGCM results and climatological data (Locarnini et al., 2006; Antonov et al., 2006).

Many of the parameterizations on energy dissipation and turbulence are
based on the assumption that MLD=XLD, but this is the exception rather than the rule (Noh and Lee, 2008; Sutherland et al., 2014). No studies to date have derived a relationship between the MLDs and XLDs in regions where the two depths differ. This paper attempts to derive such a relation for OSBL measurements in the subtropical Atlantic. An overview of the conditions and observations can be found in section 5.2. Section 5.3 investigates the time-dependent nature of \( h_c \) and \( h_\sigma \). This is followed by a summary of the results in section 5.4.

## 5.2 Observations

Measurements are presented from the SubTRopical Atlantic Surface Salinity Experiment (STRASSE) (Fig. 5.1) aboard the N/O Thalassa, which took place during August and September 2013. Radiative fluxes and wind speed measurements were recorded aboard the N/O Thalassa with the wind stress and buoyancy flux calculated using the TOGA COARE 3.0 algorithm (Fairall et al., 2003). An example of the 10 meter wind speed \( U_{10} \) and the surface buoyancy flux \( B_0 \) for the STRASSE campaign is shown in Fig. 5.2a. The sign convention for \( B_0 \) is negative (positive) when the surface buoyancy flux is into (out of) the ocean i.e. during restratification the surface water becomes more buoyant (i.e. less dense) and therefore there is a flux of buoyancy into the ocean.

The buoyancy flux and the wind speed are used to calculate the Monin-Obukhov length

\[
L = \frac{-u^3_\ast}{\kappa B_0}
\]

where \( u_\ast \) is the friction velocity (in water), which is related to the wind stress as \( u_\ast = \sqrt{\tau/\rho_0} \), \( \kappa = 0.4 \) is von Kármán’s constant and \( \rho_0 \) is the density at the ocean surface. During convection \( L < 0 \) and during restratification \( L > 0 \) and is expected to give the relative length scale at which buoyancy and
shear forces are equal (Fig. 5.2b, brown region). Therefore, \( h_\sigma/L \ll -1 \) for buoyancy dominated convection and \( h_\sigma = L \) during restratification (Large et al., 1994). The green-blue line in Fig. 5.2b shows the ratio of \( h_\sigma/L \) where the shaded horizontal band denotes the region \(-1 \leq h_\sigma/L \leq 1\). During daytime restratification \( h_\sigma/L \) reaches 10-20 times what expected during restratification suggesting Eq. (5.1) may not be valid during this diurnal phase.

Ocean measurements of microstructure shear (from which the turbulent dissipation rate is calculated), temperature, and salinity were obtained with the Air-Sea Interaction Profiler (ASIP), an autonomous microstructure profiler which rises upwards through the OSBL to the surface and is specifically designed to measure turbulence properties (for further details see Sutherland et al., 2013; Ward et al., 2014). A summary of the ASIP deployments and locations is presented in Fig. 5.1. The time-depth evolution of potential temperature \( \theta \) and buoyancy frequency \( N^2 \) can be found in Figs. 5.2c–d.
Figure 5.2: **a** Wind speeds measured at 10 m (orange line) and surface buoyancy flux (blue region), **b** the Monin-Obukhov length $L$ (brown), and $h_\sigma/L$ (green-blue) and also time-depth plots of **c** potential temperature $\theta$, **d** $\log_{10} N^2/s^1$, and **e** $\log_{10} \epsilon/m^2s^3$, and also time-depth plots of $\theta$. The green-blue shaded region in **b** shows the region where $-1 \leq h_\sigma/L \leq 1$. The black solid and dashed lines in **c-e** shows the mixed layer depth as calculated using a threshold of $\sigma_\theta = 0.03$ and 0.09 kg m$^{-3}$ respectively. The grey solid and dashed lines in **c-e** show the depth where $\epsilon$ first falls to $10^{-9}$ and $10^{-8}$ m$^2$s$^{-3}$ respectively.

The density ratio is given by

$$R_p \equiv \frac{\alpha \theta_z}{\beta S_z},$$  \hspace{1cm} (5.2)
Mixed and Mixing Layer Depths in the OSBL

where $\alpha$ and $\beta$ are the thermal expansion coefficient and the saline contraction coefficient respectively and $\theta_z$ and $S_z$ are the vertical gradients of temperature and salinity respectively. Typically $R_\rho > 2$ throughout the OSBL and stratification is, on average, dominated by temperature.

Dissipation rates of turbulent kinetic energy are calculated using $\epsilon = 7.5\nu \langle u_z^2 \rangle$, where $\nu$ is the kinematic viscosity and $u_z$ is the turbulent vertical shear (Yamazaki and Osborn, 1990). Values for $\epsilon$ are calculated from shear spectra over one second segments, corresponding to a vertical resolution of about 0.5 m, and averaged over three successive profiles to obtain mean values approximately every hour. Figure 5.2e shows the observed turbulent dissipation rate for the five deployments.

The mixed layer depth $h_\sigma$ is calculated using two density threshold criteria relative to the density at $z_r = 2.5$ m depth from the surface. The first threshold, 0.03 kg m$^{-3}$, is a commonly used value for profiles (de Boyer Montégut et al., 2004). The MLD calculated with this threshold $h_{\sigma1}$ is shown by the solid black line in Figs. 5.2c–e. In addition, the threshold of 0.09 kg m$^{-3}$ is also tested as density thresholds are expected to be higher at lower latitudes (Noh and Lee, 2008) with this MLD estimate $h_{\sigma2}$ is shown by the dashed black line in Figs. 5.2c–e. In general, there is very little difference between the two, suggestive of a large density gradient at the base of the OSBL and in the text $h_\sigma \equiv h_{\sigma1}$ unless specifically specified.

The mixing layer depth $h_\epsilon$ is defined as the shallowest depth where the turbulent dissipation rate $\epsilon$ falls to a certain background level. Profiles of $\epsilon$ are further smoothed using a running mean filter with a half length of one metre to minimize the effect of intermittency on the mixing layer definition (Sanchez-Martin et al., 2014). In determining the mixing layer depth, two different background dissipation levels are tested: $\epsilon_{b1} = 10^{-9}$ and $\epsilon_{b2} = 10^{-8}$ m$^2$s$^{-3}$ with the mixing layer determined via each method designated as $h_{\epsilon1}$ and $h_{\epsilon2}$ respectively. The first threshold of $10^{-9}$ m$^2$s$^{-3}$ was determined after inspecting hundreds of ASIP profiles and is consistent with mixing layer definitions.
Mixed and Mixing Layer Depths in the OSBL

from previous studies (Dewey and Moum, 1990; Brainerd and Gregg, 1995). However, recent studies by Lozovatsky et al. (2006) and Cisewski et al. (2008) defined the lower bound of the mixing layer where $\epsilon$ decreased from $10^{-7}$ to $10^{-8}$ m$^2$s$^{-3}$ so this dissipation threshold is also tested. The grey lines in Figs. 5.2c–e shows the mixing layers $h_{\epsilon 1}$ (solid) and $h_{\epsilon 2}$ (dashed) along with the mixed layer $h_\sigma$ (solid black line).

5.3 Results and Discussion

To investigate the diurnal structure of the mixed and mixing layer depths, the five deployments are phase averaged as a function of the diurnal hour to create a single composite day. The surface forcing is also averaged as a function of the local diurnal time for $B_0$ (blue dots) and $U_{10}$ (purple dots) with hourly phase averages of $B_0$ and $U_{10}$ are shown by the solid blue and purple lines respectively (Fig. 5.2a). The phase averaged surface buoyancy flux was consistent throughout the campaign with only small noticeable variations, most likely due to cloud cover which were minimal. The phase averaged 10 m wind speed has a mean of $6.45 \pm 0.05$ m s$^{-1}$ and a standard deviation of $1.40$ m s$^{-1}$ during ASIP deployments. On average $-L < h_{\sigma 1}$ during the night (Fig. 5.3b) suggesting that the surface buoyancy flux is the primary forcing driving the dynamics.

Observations of the diurnal phase averaged values for $h_{\sigma 1}$, $h_{\epsilon 1}$ and to a lesser extent $h_{\epsilon 2}$ follow a typical diurnal pattern of a new mixed layer forming during the early stages of restratification followed by a gradual deepening during convection (Brainerd and Gregg, 1995). This can be seen in Fig. 5.3c as the green, grey, and red lines for $h_{\sigma 1}$, $h_{\epsilon 1}$, and $h_{\epsilon 2}$ respectively with the shaded regions denoting the 95% confidence intervals for the hourly phase-averaged estimates. Common features of the two mixing layer depths are that they both begin to reform around 9:00-10:00 local time (LT) while the mixed layer lags the mixing layers by 1-2 hours. Both $h_{\epsilon 1}$ and $h_{\epsilon 2}$ are greater
Mixed and Mixing Layer Depths in the OSBL

Figure 5.3: Observations and diurnal phase averages of a $B_0$ (blue) and $U_{10}$ (orange), b $L$ (brown) and $h_{\sigma_1}/L$ (blue-green), c mixed $h_{\sigma_1}$ (green) and mixing layer depths ($h_{\epsilon_1}$ (grey) and $h_{\epsilon_2}$ (red) respectively), d the ratio of the mixing to the mixed layer depth, and e the absolute difference between mixing and mixed layer depths for the two mixing layer depth definitions. The black dashed lines in d–e denote $h_{\epsilon} = h_{\sigma}$. Panels f–h show the histograms for the observations in c–e with the box plots showing the stats for the entire record.

than $h_{\sigma_1}$ when $h_{\sigma_1}$ is a minimum suggesting high turbulence levels during this phase. These diurnal structures are apparent in the phase averaged ratio and differences between mixing and mixed layer depths in Figs. 5.3d–e respectively.

The statistics of diurnal phase averaged values can be found in Figs. 5.3f–h. Shown are the histograms for each parameter in addition to a box plot of
the entire record. Although \( h_{\epsilon 1} \) and \( h_{\epsilon 2} \) undergo a similar diurnal progression, they differ significantly in magnitude relative to \( h_{\sigma 1} \) with \( h_{\epsilon 1} \sim h_{\sigma 1} \) and \( h_{\epsilon 2} < h_{\sigma 1} \).

Figure 5.4: Observations and diurnal phase averages of a \( B_0 \) (blue) and \( U_{10} \) (orange), b \( L \) (brown) and \( h_{\sigma 2}/L \) (blue-green), c mixed \( h_{\sigma 2} \) (green) and mixing layer depths (\( h_{\epsilon 1} \) (grey) and \( h_{\epsilon 2} \) (red) respectively), d the ratio of the mixing to the mixed layer depth, and e the absolute difference between mixing and mixed layer depths for the two mixing layer depth definitions. The black dashed lines in d–e denote \( h_{\epsilon} = h_{\sigma 2} \). Panels f-h show the histograms for the observations in c-e with the box plots showing the stats for the entire record.

Figure 5.4 is similar to Fig. 5.3, but using \( h_{\sigma 2} \) rather than \( h_{\sigma 1} \) for the MLD. The most striking difference between the two MLD definitions is that during restratification \( h_{\sigma 2} \) on average lags \( h_{\sigma 1} \) by 90 minutes with a
more rapid transition between the MLD from the previous night and the reformation during the day. This is not surprising as it takes more time to create a density difference of 0.09 kg m\(^{-3}\) than 0.03 kg m\(^{-3}\) from insolation. However, other than the timing of the restratification and the magnitude of the diurnal response of the two MLD definitions are quite similar suggesting a weak dependence on the threshold criteria for this region. These similarities can easily be seen in the statistics of Figs. 5.4f–h.

Figure 5.5 shows the observed phase averaged values for the density difference at the mixing layer depth relative to the reference depth \(z_r = 2.5\) m, i.e. \(\Delta \sigma^{1,2}_\theta = \sigma_\theta(h_{e1,2}) - \sigma_\theta(z_r)\), where the mixing layer depths \(h_{e1}\) and \(h_{e2}\) are shown in grey and red respectively. During convection there is a clear increase in \(\Delta \sigma^{e1}_\theta\), suggestive of entrainment of denser fluid from below \(h_\sigma\). This increase in \(\Delta \sigma_\theta\) is most likely due to entrainment since the ratio \(h_{e1}/h_\sigma\) varies minimally during this period (Fig. 5.3c), which suggests that the density difference is occurring at the base of the mixing layer and not at the surface. This trend reverses at 06:00 LT when \(B_0\) becomes positive and the surface layer is becoming more buoyant. This structure during convection is not observed in \(\Delta \sigma^{e2}_\theta\) as this mixing layer depth underestimates the mixed layer depth during convection (Figs. 5.3b–c). Approximately 30% of all measurements have \(0 \leq \Delta \sigma^{e2}_\theta < 0.01\) kg m\(^{-3}\), with the the vast majority of these occurring during convection, while there are zero measurements of \(\Delta \sigma^{e1}_\theta\) corresponding to this range (Fig. 5.5b).

From 12:00 to 15:00 LT \(\Delta \sigma^{e1}_\theta \sim \Delta \sigma^{e2}_\theta\) indicative of the homogeneous density between the \(h_{e1}\) and \(h_{e2}\) depths at this time. However, starting at 15:00 LT, which coincides with the peak gradient in both \(B_0\) and \(h_{e1,2}/h_\sigma\), \(\Delta \sigma^{e1}_\theta\) and \(\Delta \sigma^{e2}_\theta\) diverge as \(h_{e1,2}\) and \(h_\sigma\) begin to deepen. Since the surface buoyancy flux is still stabilizing, the likely suspect for this deepening at 15:00 LT is shear instability due to an increase in surface currents when \(h_\sigma\) is shallow (Price et al., 1986), most likely due to the wind stress only transferring momentum to the shallow mixed layer. Assuming horizontal
Figure 5.5: a the density difference from the reference level to $h_{\epsilon 1}$ (grey) and $h_{\epsilon 2}$ (red) along with the diurnal phase averaged mean and 95% confidence intervals. b shows the histogram over the entire period using 0.01 kg m$^{-3}$ bins. The shaded region in a shows the 95% confidence interval for the mean and the box plots in b are for the entire record.

Homogeneity, the momentum in the OSBL should be related to the wind stress

$$\frac{\partial}{\partial t} (hu) = u^2$$

(5.3)

where $h$ and $u$ are the depth and velocity of the OSBL respectively. Since wind speeds vary a small amount over a composite day (Fig. 5.3a) a decrease in $h$ will result in an increase in $u$, which in turn can create shear across the shallow pycnocline which could become unstable if the local shear squared, $S^2$ exceeds about 4 times $N^2$ (e.g. Miles, 1961).

Without direct measurements of the vertical shear it is impossible to say with 100% certainty, but the timing of the deepening, about 15:00 local time, does coincide with similar measurements from the equatorial Pacific (Smyth et al., 2013). A calculation of the shear using a logarithmic profile for the
Mixed and Mixing Layer Depths in the OSBL

velocity

\[
\frac{u}{u_*} = \frac{1}{\kappa} \ln \frac{z}{z_0}
\]  

(5.4)

where \(z_0\) is the roughness length, taken here to be 0.1 m \((Craig and Banner, 1994)\). Using a friction velocity of \(u_* = 0.01\) m s\(^{-1}\) \((U_{10} = 6.5\) m s\(^{-1}\)) and a depth of \(z = 5\) m, Eq. (5.4) gives the velocity difference to be \(\Delta u/u_* = 9.8\).

For shear instability to occur, assuming an instability criterion for the gradient Richardson number

\[
Ri = \frac{N^2}{\mathcal{S}^2}
\]  

(5.5)

of \(Ri < 1/4\) would require \(N^2 < 10^{-6}\) s\(^{-2}\). This is not consistent with the observed buoyancy frequency which is at least an order of magnitude greater than this during restratification. However, \(Kudryavtsev and Soloviev\) (1990) describe measurements of a slippery surface layer which can occur during daytime restratification and weak winds with values for the velocity gradient between 0.35 m and 5 m to be 20-40 times the friction velocity, which would increase our shear calculation by an order of magnitude such that \(Ri \sim 1/4\).

The surface buoyancy flux becomes destabilizing at 18:00 LT at which point there appears to be a regime which lasts until 21:00 LT where \(h_{\epsilon_1}\) and \(h_{\epsilon_2}\) are both greater than \(h_\sigma\). This appears to be a transitional period where the shear instability which eroded the restratification is now decreasing due to the deepening of the mixed layer and the surface water has not sufficiently cooled to become gravitationally unstable. After 21:00 LT \(\Delta\sigma^c_\theta \to 0\) as the convective regime takes over and \(h_{\epsilon_2} < h_\sigma\) and \(\Delta\sigma^c_\theta\) slowly increases due to entrainment as mentioned previously.

The statistics of \(\Delta\sigma^{\epsilon_1,\epsilon_2}_\theta\) over the entire record are shown in Fig. 5.5b. Both distributions are far from normal with \(\Delta\sigma^c_\theta\) appearing bimodal between convection and restratification regimes and \(\Delta\sigma^r_\theta\) only having the single peak during restratification as this dissipation level does not descend to the mixed layer during convection. The mean values for \(\Delta\sigma^c_\theta\) and \(\Delta\sigma^r_\theta\) are 0.09 and
0.03 kg m$^{-3}$ respectively with the 95th and 50th percentiles shown by the respective box plots. It is interesting to note that the mean value for $\Delta \sigma_{\epsilon}^1$ is similar to the mean value of 0.07 kg m$^{-3}$ calculated by Noh and Lee (2008) for this latitude, yet $h_{\epsilon 1} \sim h_\sigma$ during most of the deployments. This is in contrast to the mean value for $\Delta \sigma_{\epsilon}^2$, which is equal to the mixed layer depth criteria of 0.03 kg m$^{-3}$ even though $h_{\epsilon 2} < h_\sigma$ during nighttime convection.

5.4 Summary

A total of 400 profiles from the Air-Sea Interaction Profiler (ASIP) were made over five deployments with each deployment ranging from 24 to 48 hours in duration. Observations of the mixed and active mixing layer depths are analysed in a buoyancy-dominant regime in the subtropical Atlantic during summer. Two thresholds for the background dissipation rate, $\epsilon_{b1} = 10^{-9}$ and $\epsilon_{b2} = 10^{-8}$ m$^2$/s$^{-3}$ respectively, were tested to calculate the mixing layer depth (denoted by $h_{\epsilon 1}$ and $h_{\epsilon 2}$ respectively) while the mixed layer depth ($h_\sigma$) was defined using two density thresholds of $\Delta \sigma_{\theta} = 0.03$ and $\Delta \sigma_{\theta} = 0.09$ kg m$^{-3}$ relative to the density at $z_r = 2.5$ m depth respectively with little observed variation between the two. Diurnal variations in both the mixed and mixing layers were consistent over all of the deployments with $h_{\epsilon 1} > h_{\epsilon 2} > h_\sigma$ during restratification and $h_\sigma \sim h_{\epsilon 1} > h_{\epsilon 2}$ during convection.

The mixing layer depth is presented as a function of the density field for the two thresholds for the background dissipation rate. Although $h_\sigma \sim h_{\epsilon 1}$, the density difference at $h_{\epsilon 1}$, i.e. $\Delta \sigma_{\theta}^1 = \sigma_{\theta}(h_{\epsilon 1}) - \sigma_{\theta}(z_r)$ had a mean value three times the density threshold of 0.03 kg m$^{-3}$ with two peaks in the distribution at the convective maximum and restratification maximum respectively. This mean value of $\Delta \sigma_{\theta}^1 = 0.09 \pm 0.01$ kg m$^{-3}$ is similar to the 0.07 kg m$^{-3}$ value calculated for this latitude by Noh and Lee (2008) who suggested a location-based mixed layer depth threshold. This value is greater than the density threshold of 0.03 kg m$^{-3}$ for the mixed layer depth.
definition even though $h_\sigma \sim h_{c1}$ for most of the profiles. $\Delta \sigma_{\theta}^{c2}$ has only the one peak during restratification as $h_\sigma > h_{c2}$ during convection, but with a mean value over the diurnal day equivalent to the density threshold of 0.03 kg m$^{-3}$. Even though $h_{c2}$ is much less than the mixing layer depth during convection, the turbulence levels are so high during restratification that the mean value over the day is still comparable to the density threshold used to calculated the mixed layer.

The phase averaged mixed layer depth was observed to consistently deepen while the surface buoyancy flux was still stabilizing suggesting that the increased mixing came from mechanical mixing. Since it occurs at the same time every day this would suggest the mechanism to arise from a shear instability rather than from horizontal advection. For the observed stratification, larger vertical shear consistent with the observations by Kudryavtsev and Soloviev (1990) of a slippery near surface layer are required for the onset of turbulence from shear instability. Observed values of $h_\sigma/L$ reach values of 10 to 20 during restratification, while for similarity theory there should be an upper limit of $h_\sigma = L$ when buoyancy suppression of turbulence should be greater than the mechanical production (Large et al., 1994). This suggests there are other sources of mechanical mixing, with the most likely suspect being surface gravity waves, which need to be accounted for during restratification.

The mean value for $\Delta \sigma_{\theta}^{d1}$ is much greater than the accepted range of the density threshold for individual profiles of $0.01 < \Delta \sigma < 0.03$ kg m$^{-3}$ (de Boyer Montégut et al., 2004), yet is consistent with the a zonally variable $\Delta \sigma_{\theta}$ as shown by Noh and Lee (2008). However, the MLD was not found to be sensitive to the choice for the density threshold in this region. This insensitivity may be due to the magnitude of the buoyancy flux which will be a function of the season (in this case summer) and the region (sub-tropical Atlantic). It is not surprising that the MLD and XLD should be functions of the forcing, which will vary regionally and temporally (Noh and Lee, 2008).
More observations of the entire OSBL across different regions and seasons would greatly improve the current understanding of the OSBL and lead to more dynamics based parameterizations for the depth.
Chapter 6

Conclusions and Future Work

The turbulent and mixing processes associated with the Ocean Surface Boundary Layer (OSBL) have been investigated using novel microstructure observations from the Air-Sea Interaction Profiler (ASIP). These observations are the first published results using ASIP, which was developed to measure microscale features throughout the OSBL up to the water surface. ASIP is unique in that it is completely autonomous giving unprecedented observations of the turbulence dynamics throughout the OSBL at regular intervals. In addition, ASIP is an ascending profiler allowing it to make measurements up to the ocean surface. The observations presented here comprise the vast majority of direct observations of the turbulent dissipation rate in the upper ten meters for open ocean conditions.

These observations were used in conjunction with accurate measurements of the surface wave field and atmospheric forcing to examine their respective roles in the turbulent dynamics. The influence of surface gravity waves appear not to be restricted to the upper few metres, but in fact extend throughout the OSBL. This large depth of influence is consistent with the presence of Langmuir circulations, which are expected to fill the OSBL. Although it was unclear whether Langmuir cells formed, observations of $\epsilon$ were found to be consistent with LES results of the OSBL which include
Langmuir circulations. These are some of the first observations to confirm LES studies of Langmuir turbulence in the open ocean.

The accuracy of these scalings depended on accurately determining the depth of the OSBL. Improved agreement with LES results were found when the active mixing layer depth (XLD), determined by locating the depth at which $\epsilon$ fell to a background dissipation rate, was used for the boundary layer depth rather than the mixed layer depth (MLD) which is the depth at which the density exceeds a threshold value relative to the near surface density.

The variations of the MLD and XLD are investigated in a buoyancy driven regime in the subtropical Atlantic. Both the MLD and XLD have a predictable diurnal structure in a buoyancy driven regime. Observations show the density threshold used to estimate the MLD can be adapted to obtain the XLD, but this has a clear variation as a function of the local time of day and hence surface buoyancy flux. Mean values for the density threshold of the XLD are consistent with Noh and Lee (2008) who found larger thresholds near the equator and decreasing poleward.

There are few direct observations of the MLD and XLD and this would be an interesting avenue to pursue with regards to future research. The forcing in a buoyancy driven regime such as presented here is predictable, which lends itself to arranging a composite day for sufficient statistics. Wind driven forcing is stochastic in nature and would require longer time series to obtain significant statistics of the XLD and MLD. Such observations could aid greatly in obtaining accurate parameterizations for the XLD.

Another aspect of the OSBL that hasn’t been explored are the enhanced dissipation levels during restratification. Our observations show all the evidence of shear instability in the upper few metres due to the shallow OSBL depth, but there are no measurements of the velocity field and, therefore, impossible to say for certain. The elevated dissipation may serve as a type of preconditioning to nighttime convection, which has significant
Conclusions and Future Work

effects on vertical fluxes of heat, momentum and nutrients throughout the OSBL.

There is a need to quantify the turbulent dynamics of the upper few metres, which is historically very difficult to observe in the open ocean. There are aspects, such as the transport of turbulent kinetic energy, which have never been measured in the open ocean. Ongoing improvements in observational techniques and instrumentation should allow future researchers to address many of the outstanding issues in OSBL turbulence.
Bibliography


Christensen, K. H., J. Röhrs, B. Ward, I. Fer, G. Broström, O. Sætra, and O. Breivik (2013), Surface wave measurements us-


Munk, W. H. (1944), Proposed uniform procedure for observing waves and interpreting instrument records, *La Jolla, California: Wave Project at the Scripps Institute of Oceanography*.


Rasclé, N., F. Ardhuin, and E. A. Terray (2006), Drift and mixing under the ocean surface: A coherent one-dimensional description with


