RESEARCH ARTICLE

10.1002/2017JC013525

Key Points:

- Open ocean turbulence observations presented do not support commonly applied scaling approaches
- An empirical scaling relationship based on wind and wave forcing is suggested for daytime conditions
- When convection is important additional buoyancy forcing should be considered

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Citation:

Esters, L., Breivik, Ø., Landwehr, S., ten Doeschate, A., Sutherland, G., Christensen, K. H., et al., (2018). Turbulence scaling comparisons in the ocean surface boundary layer. *Journal of Geophysical Research: Oceans*, 123, 2172–2191. <https://doi.org/10.1002/2017JC013525>

Received 2 OCT 2017

Accepted 6 FEB 2018

Accepted article online 7 MAR 2018

Published online 26 MAR 2018

Turbulence Scaling Comparisons in the Ocean Surface Boundary Layer

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Abstract Direct observations of the dissipation rate of turbulent kinetic energy, ϵ , under open ocean conditions are limited. Consequently, our understanding of what chiefly controls dissipation in the open ocean, and its functional form with depth, is poorly constrained. In this study, we report direct open ocean measurements of ϵ from the Air-Sea Interaction Profiler (ASIP) collected during five different cruises in the Atlantic Ocean. We then combine these data with ocean-atmosphere flux measurements and wave information in order to evaluate existing turbulence scaling theories under a diverse set of open ocean conditions. Our results do not support the presence of a “breaking” or a “transition layer,” which has been previously suggested. Instead, ϵ decays as $|z|^{-1.29}$ over the depth interval, which was previously defined as “transition layer,” and as $|z|^{-1.15}$ over the mixing layer. This depth dependency does not significantly vary between nonbreaking or breaking wave conditions. A scaling relationship based on the friction velocity, the wave age, and the significant wave height describes the observations best for daytime conditions. For conditions during which convection is important, it is necessary to take buoyancy forcing into account.

1. Introduction

Upper ocean turbulence plays a crucial role in ocean dynamics on various scales and is important for the horizontal and vertical dispersion of various properties in the ocean. In the highly turbulent ocean surface boundary layer (OSBL), turbulence is predominantly generated due to the ocean’s interaction with the atmosphere by exchange of heat and freshwater fluxes, and the transfer of momentum through wind stress. Waves and wave breaking are an additional source of turbulence in the top meters of the ocean (Gemmrich & Farmer, 2004). These surface waves can cause Langmuir cells, which further modifies the upper ocean turbulence characteristics (Belcher et al., 2012). In addition, breaking of internal waves at the pycnocline (Thorpe, 2007; Wain et al., 2015) and submesoscale processes, including the onset of gravitational and symmetric instabilities resulting from steady winds blowing over ocean fronts, can cause turbulence in the OSBL (D’Asaro et al., 2011; Thomas & Taylor, 2010).

These processes convert mean and wave kinetic energy into turbulent kinetic energy (TKE), which will partially be dissipated into heat by viscous forces and partially contribute to the mixing of the surface ocean. This mixing works against the stratification of the upper ocean and controls the transport of heat, gases, and particles across the air-sea boundary. For example, turbulence controls the frequency of surface renewal events, which are significant for the gas transfer across the air-sea interface (e.g., Esters et al., 2017; Tokoro et al., 2008; Zappa et al., 2007). Small-scale mixing in the upper ocean also has an impact on biology by defining the level of dispersion of pollutants and phytoplankton, which itself controls the growth of marine organisms.

As it is difficult to measure turbulent fluxes and kinetic energy directly, measurements of the dissipation rates of TKE, ϵ , are conventionally used to describe turbulence within the OSBL (e.g., Moum, 1997). But even those measurements are difficult to conduct due to the strong temporal and spatial intermittency of turbulence (Gemmrich & Farmer, 2004). Another obstacle stems from the inherent difficulty of making measurements near the wave-dominated ocean surface. Therefore, improved estimates of ϵ have been proposed as an essential step toward a more complete understanding of near-surface turbulence (e.g., Belcher et al., 2012, hereafter B12; Huang & Qiao, 2010, hereafter HQ10; Terray et al., 1996, hereafter T96). The main objective of this study is to investigate if any of the existing scalings compare favorably with measurements in the open ocean. Such an agreement between

Table 1
Presented Scaling Approaches With Their Described Sources of Turbulence

Scaling	Source of turbulence
LOW	Wind
LG89	Wind and buoyancy
T96	Waves via the energy input F
HQ10	Wave-turbulence interaction via the Stokes shear
B12	Wind, buoyancy, and Langmuir turbulence

theoretical scaling laws and observations is important as the energy profiles can be used to infer which processes are the leading source of the prevailing turbulence.

A universally applicable scaling of ϵ would allow for accurate parameterizations of air-sea gas exchange (Esters et al., 2017; Katul & Liu, 2017) and the determination of characteristic length scales of upper ocean mixing (B12), as well as for modeling of rates of many ecological processes (e.g., Umlauf & Burchard, 2003; B12).

This paper uses an extensive data set with measurements of ϵ from the upwardly rising Air-Sea Interaction Profiler (ASIP; see Ward et al., 2014) from five different field campaigns in the open Atlantic Ocean. These measurements of ϵ are combined with hourly wave information from dedicated integrations of the European Centre for Medium-Range Weather Forecasts (ECMWF) stand-alone ECWAM wave model, proving an unprecedented opportunity to evaluate and compare scaling assumptions and theories with direct open ocean measurements. This paper focuses on applying this integrated set of measurements of meteorological data, wave information, and ϵ observations in the ocean mixing layer to assess the ϵ profiles, their slopes with depth, and the coefficient α scaling the wind input to surface waves as a function of the wave age. This is compared to the ϵ scaling assumptions presented in Table 1: Law of the wall (LOW) and scaling approaches proposed by LG89, T96, HQ10, and B12. The conclusions are used to predict our own empirical scaling relationship.

Existing scalings and theories are summarized in section 2. Section 3 gives a brief overview of the field campaigns and description of the methodology. The measured dissipation is compared to existing ϵ scaling approaches in section 4. Section 4.1 focuses on shear and surface waves as the main driver for near-surface turbulence, where the LOW and the “constant-dissipation layer” scalings (suggested by T96) are compared to observations, as well as turbulence-wave interaction and ϵ induced by Stokes drift shear as proposed by HQ10. Section 4.2 analyzes scaling approaches that consider buoyancy forcing as the generating mechanism for turbulence. These results are discussed in section 5. Finally, a summary is given in section 6.

2. Turbulence Scaling and Theory

The Reynolds decomposition splits the velocity of a turbulent flow into a time-averaged value $\bar{u}_i$ and a zero-mean fluctuating part u'_i : $u_i = \bar{u}_i + u'_i$, where $u_i = (u, v, w)$ is the velocity in direction $x_i = (x, y, z)$.

The TKE is the kinetic energy per unit mass associated with the energy of the fluctuating part in a flow. Its evolution equation can be written as (a full derivation is, e.g., given in Kundu and Cohen (2001)):

$$\underbrace{\frac{D}{Dt} \left(\frac{1}{2} \overline{u_i'^2} \right)}_{\text{Total rate of change in TKE}} = \underbrace{- \frac{\partial}{\partial x_j} \left(\frac{1}{\rho_0} \overline{p' u_j'} + \overline{u_j' E} - 2\nu \overline{u_i' s_{ij}'} \right)}_{\text{Transport of TKE}} - \underbrace{\overline{u_i' u_j'} \frac{\partial \bar{u}_i}{\partial x_j}}_{\text{Shear production term}} - \underbrace{\overline{b_o' w'}}_{\text{Buoyancy production}} - \underbrace{2\nu \overline{s_{ij}' s_{ij}'}}_{\text{Viscous dissipation } \epsilon}, \quad (1)$$

where ρ_0 is the reference water density, E the TKE, b_o the buoyancy, p the pressure, ν the molecular viscosity, t the time, and $s_{ij}' = \frac{1}{2} \left(\frac{\partial u_i'}{\partial x_j} + \frac{\partial u_j'}{\partial x_i} \right)$ is the strain rate tensor.

Turbulence in the ocean is often described by the viscous dissipation rate ϵ , which under a steady state solution balances the shear production and buoyancy flux (Osborn, 1980):

$$\epsilon = -\overline{u' w'} \frac{\partial \bar{U}}{\partial z} - \overline{b_o' w'}, \quad (2)$$

where z is the depth, $\bar{U}$ the mean velocity, and $\overline{u' w'} = u_*^2$ with u_* representing the water-side friction velocity, related to the Reynolds stress $\tau = \rho \overline{u' w'} = u_*^2 \rho$, with ρ as the water density.

In the absence of waves and their related processes, the ocean surface could be seen as a flat, rigid wall (i.e., LOW), beneath which a purely shear-driven boundary layer evolves where the shear is given by

$$\frac{\partial \bar{U}}{\partial z} = \frac{u_*}{\kappa |z|}, \quad (3)$$

where $\kappa \approx 0.41$ is the von Kármán's constant. Neglecting the buoyancy term in (2) and expressing the shear through (3), leads to ϵ being proportional to $|z|^{-1}$ (decaying away from the surface):

$$\epsilon_{\text{LOW}}(z) = \frac{u_*^3}{\kappa |z|}. \quad (4)$$

During the day, solar radiation heats the ocean surface. This stabilizes the upper ocean and, therefore, suppresses mixing. During the night, however, the ocean loses heat to the atmosphere (oceanic heat loss is defined as a positive heat flux). This surface cooling gives rise to convective forcing. By arguing that both wind stress and convection enhance dissipation LG89 account for this additional source of turbulent forcing by linearly adding the buoyancy flux at the surface $B_0 = b'_o w'$ to the wind forcing expressed by the LOW:

$$\epsilon_{\text{LG89}} = 0.87 \times (1.76 \epsilon_{\text{LOW}} + 0.58 B_0). \quad (5)$$

More recent theoretical extensions and refinements primarily deal with the effect of surface waves. In the presence of waves, wave breaking is suggested to increase ϵ at the air-sea interface relative to values predicted by the LOW (e.g., Agrawal et al., 1992; Anis & Moum, 1995; Drennan et al., 1996; Feddersen et al., 2007; Gemmrich & Farmer, 2004; Greenan et al., 2001; Kitaigorodskii et al., 1983; Soloviev & Lukas, 2003; T96). To account for wave breaking at the ocean surface, the energy input from the wind to the waves F can be determined. In their turbulent closure model, Craig and Banner (1994) parameterized F as $F = \alpha u_*^3$ with a constant α . Craig and Banner (1994) found that the TKE flux is relatively insensitive to sea state and $\alpha \approx 100$ for wave ages from very young wind seas to fully developed seas.

To introduce turbulence injected from the waves themselves rather than the wind alone, Gemmrich et al. (1994) introduced an effective wave speed $\bar{c}$. This $\bar{c}$ corresponds to the speed of the waves transferring the most energy to the ocean. By arguing that the wind stress is transferred to the ocean on particular spatial and temporal scales, they define $F = \bar{c} u_*^2$. Gemmrich et al. (1994) determined $\bar{c}$ to be dominated by the short wind waves and found $\bar{c} \approx 1 \text{ m s}^{-1}$. These formulations of F are evaluated by Thomson et al. (2016) using a comprehensive data set gained from several Surface Wave Instrument with Tracking (SWIFTs) in the North Pacific. Their data support a constant $\bar{c}$ of 2 m s^{-1} .

In order to account for waves in scaling ϵ with depth, T96 considered both expressions of F and applied it to ϵ data collected in Lake Ontario. They divided the vertical decay structure of ϵ under breaking waves into three depth dependent regions. Directly beneath the water surface down to a "breaking depth" $z_b = 0.6 H_s$, there is a layer, which is directly influenced by wave breaking. In this "breaking layer," ϵ is assumed to be constant and an order of magnitude larger than predicted by the LOW. The energy in this "breaking layer" is dissipated downward to a "transition depth," $z_t = 0.3 H_s \frac{\kappa \bar{c}}{u_{*a}}$. Within the "transition layer" ϵ behaves as

$$\frac{\epsilon H_s}{F} = 0.3 \left(\frac{|z|}{H_s} \right)^b, \quad (6)$$

where $b = -2$ in T96.

Substituting $F = \alpha u_*^3$ into (6), leads to

$$\frac{\epsilon H_s}{u_*^3} = 0.3 \alpha \left(\frac{|z|}{H_s} \right)^b. \quad (7)$$

In the following, (7) is investigated using open ocean observations in order to determine optimal coefficients for b and α . Using data from T96 and Drennan et al. (1996), Wang and Huang (2004) formulated α as a function of the wave age c_p/u_{*a} :

$$\alpha = \begin{cases} 0.5 \left(\frac{c_p}{u_{*a}} \right) \left(\frac{\rho_w}{\rho_a} \right)^{1/2} & \text{for } c_p/u_{*a} \leq 11 \\ 12 \left(\frac{c_p}{u_{*a}} \right)^{-1/3} \left(\frac{\rho_w}{\rho_a} \right)^{1/2} & \text{for } c_p/u_{*a} > 11, \end{cases} \quad (8)$$

where u_{*a} is the air-side friction velocity and c_p is the phase velocity of the waves.

T96 found that below the “transition layer” local shear production dominates and ϵ follows the behavior of a shear-driven flow, proportional to $|z|^{-1}$:

$$\epsilon_{T96}(z) = \begin{cases} 0.3 \alpha \frac{u_*^3}{H_s} \left(\frac{z_b}{H_s} \right)^{-2} & \text{above } z_b \\ 0.3 \alpha \frac{u_*^3}{H_s} \left(\frac{|z|}{H_s} \right)^{-2} & \text{between } z_b \text{ and } z_t \\ \frac{u_*^3}{\kappa |z|} & \text{below } z_t. \end{cases} \quad (9)$$

This scaling suggested by T96 has been supported by several studies under different environmental conditions (Drennan et al., 1996; Feddersen et al., 2007; Gerbi et al., 2009; Greenan et al., 2001; Jones & Monismith, 2008; Shuiqing & Dongliang, 2016). Drennan et al. (1996) argued that H_s should be calculated from the wind sea part of the wave spectrum only, under the assumption that swell does not break.

Using a vertically rising profiler in the Pacific Ocean, Anis and Moum (1995) observed ϵ profiles, which decayed exponentially with depth. They attributed the enhanced values of ϵ to breaking waves and wave-turbulence interactions. HQ10 neglected the buoyancy term in (2) and expressed the mean flow by the Stokes drift u_s , which is a net drift in the down-wave direction caused by the open (noncircular) orbital particle motion in the wave field (Stokes, 1847).

HQ10 used the Stokes drift expression for a monochromatic wave:

$$u_s = u_{s0} e^{2kz}, \quad (10)$$

where u_{s0} is the magnitude at the ocean surface, such is $u_{s0} = c_p(ak)^2$, where a is the wave amplitude, and k is the wave number. They use the vertical shear of this Stokes drift to express a corresponding dissipation rate as

$$\epsilon_{HQ10}(z) = a_l u_*^2 \frac{\partial u_s}{\partial z}, \quad (11)$$

where a_l is a dimensionless constant associated with the surface waves. This constant was determined through data regression by HQ10 using the data of Anis and Moum (1995) as

$$a_l = 3.75 \beta \pi \sqrt{\frac{H_s}{\lambda}}, \quad (12)$$

where λ is the dominant wavelength and β is an empirical constant between 0 and 1. Using this expression for a_l , (11) is used in the following to evaluate the performance of the scaling suggested by HQ10 with our data.

The Stokes drift in equation (10) can be calculated separately for the wind sea and swell part of the wave spectrum via the wave number k . Sutherland et al. (2013) showed that the performance of the scaling suggested by HQ10 could be improved by scaling profiles of ϵ as the sum of both separately determined expressions. In our analysis, we compare the scaling suggested by HQ10 based on the wind sea and swell part of the wave spectrum only, as well as the sum of both.

It should be noted that for $k|z| < 1$, expanding (10) to second order results in

$$\frac{\partial u_s}{\partial z} \propto \frac{1}{1 + 2k|z| + 2(kz)^2}. \quad (13)$$

Hence, the HQ10 parametrization should fall somewhere between LOW and T96, depending on the average wavelength.

Compared to the Stokes drift profile for a monochromatic wave, Breivik et al. (2016a) proposed a new approximation for the Stokes drift velocity based on the exact solution for the Phillips (1958) spectrum, which has a much stronger gradient near the ocean surface. The shear of the Stokes drift velocity determined for the Phillips spectrum is

$$\frac{\partial u_{sp}}{\partial z} = \alpha_p \sqrt{-\frac{2\pi g}{z}} \operatorname{erfc}(\sqrt{-2kz}), \quad (14)$$

where $\alpha_p = 0.0083$ is the Phillips' parameter and erfc is the complementary error function, which is defined for an arbitrary x as $\operatorname{erfc}(x) = \frac{2}{\sqrt{\pi}} \int_x^\infty e^{-t^2} dt$.

The scaling proposed by HQ10 was criticized by Kantha et al. (2014) mainly because HQ10 determined their parameters from observations assuming that only waves are important in the upper ocean layer. Also, HQ10 do not take wave breaking into account.

In order to determine the depth of the OSBL in global climate models and to include surface-wave processes that force Langmuir turbulence, B12 consider wind, buoyancy, and waves as sources for ocean turbulence. Each of the three mechanisms produces a distinct type of turbulence with its own scaling laws. B12 formulate ϵ as a linear combination of these generation mechanisms at a depth of $h/2$, where h is the mixed layer depth:

$$\epsilon_{B12}(|z|/h=0.5) = 2 \left(1 - e^{-0.5\sqrt{\text{La}_t^2}} \right) \frac{u_*^3}{h} + 0.22 \frac{u_{*L}^3}{h} + 0.3 \frac{w_*^3}{h}, \quad (15)$$

where $w_* = (B_0 h_e)^{1/3}$ is the scaling velocity for buoyancy-forced turbulence and only defined for positive B_0 , e.g., destabilizing buoyancy flux, $\text{La}_t^2 = u_*^2 / u_s$ is the turbulent Langmuir number, which expresses the ratio of turbulence production due to wind forcing and wave forcing (McWilliams et al., 1997), and $u_{*L} = (u_*^2 u_{s0})^{1/3}$. This scaling approach applies under uniform and steady conditions.

Sutherland et al. (2014a) showed that the most appropriate length scale for h was the mixing layer depth which provided better results than the widely used mixed layer depth. The mixing layer is the ocean surface layer of active mixing. In this study the mixing layer depth h_e is defined coincidentally for all cruises as the depth at which ϵ falls to a level of $\epsilon = 10^{-9} \text{ m}^2 \text{ s}^{-3}$ (Brainerd & Gregg, 1995; Sutherland et al., 2014b).

3. Methods

Measurements of ϵ were conducted on five field campaigns in different areas of the Atlantic Ocean between 2010 and 2014. These cruises are summarized in Table 2 and their location is shown in Figure 1. These measurements were taken with the Air-Sea Interaction Profiler (ASIP; see Ward et al., 2014), resulting in a collated data set of 1867 ϵ profiles. Hourly wave information is provided by dedicated integrations from the ECMWF stand-alone ECWAM wave model, which allow the capture of rapid changes in the boundary conditions (Figure 2). The observed wind speeds during each field campaign are used to calculate u_* using the Coupled Ocean Atmosphere Response Experiment (COARE 3.0) algorithm (Edson et al., 2013; Fairall et al., 1996, 2003) (except for the cruise in the Labrador Sea for which we rely on the modeled wind speeds).

Table 2

Presented Field Campaigns With the Name of the Campaign, Date, Ship, the Number of ASIP Deployments During the Campaign, and the Overall Number of ASIP Profiles per Cruise

	Date	Ship	ASIP deployments	Number of profiles
Labrador	May 2010	CCGS Hudson	5	192
Knorr11	Jul 2011	R/V Knorr	4	283
STRASSE	Aug 2012	R/V Thalassa	7	528
MIDAS	Apr 2013	R/V Sarmiento de Gamboa	5	676
NICE	Sep 2014	R/V Hakon Mosby	4	188

Figure 1. Map showing the positions of all the here presented cruises. The cruise tracks are shown as red lines in the sub-figures with the orange circles indicating the ASIP deployments.

3.1. Cruises

The first of the cruises took place for 2 weeks in May 2010 onboard the CCGS Hudson as part of the annual hydrographic cruises to the **Labrador** Sea carried out by the Bedford Institute of Oceanography, Dartmouth, Canada (see Wain et al., 2015; Yashayaev & Loder, 2016, for further details).

The **Knorr11** cruise was carried out in the North Atlantic aboard the R/V Knorr from late June to mid-July 2011 (see Bell et al., 2013; Christensen et al., 2013; Esters et al., 2017; O'Sullivan et al., 2015; Scanlon et al., 2016; Scanlon & Ward, 2013, 2016; Sutherland et al., 2013, 2014a, for further details). It is shown by Bell et al. (2013) (in their supporting information) for this cruise, that the eddy covariance u_{*g} and wind speed measurements agree with the drag determined using the COARE algorithm.

The Sub-Tropical Atlantic Surface Salinity Experiment (**STRASSE**) took place from 21 August to 9 September 2012 as part of the Salinity Processes in the Upper ocean Regional Study (SPURS-I) campaign (see Reverdin et al., 2015; Sutherland et al., 2014b, 2016, for further details). The campaign investigated the high-salinity region of the North Atlantic subtropical gyre.

As part of the second leg of SPURS-I, the **MIDAS** cruise was carried out for 4 weeks in March 2013 in the region of the North Atlantic Salinity Maximum aboard the Spanish ship R/V Sarmiento de Gamboa to map the salinity structure of the upper ocean (see Landwehr et al., 2015, for further details). High-quality meteorological and wave information was available for the whole duration of the cruise, from a mooring at 24.5°N, 38°W deployed by the Woods Hole Oceanographic Institute (WHOI; Figure 2). Still, the ship's wind measurements are considered more representative of the conditions that ASIP experienced. These wind speeds are corrected for flow distortion using the mooring as reference. The corrected wind speeds were in agreement with onboard eddy covariance momentum fluxes (u_{*g}). During this campaign ASIP was programmed to descend down to a maximum depth of 30–40 m. The observations from the WHOI mooring show the seasonal mixed layer depth to be below 100 m; therefore, out of ASIP's operational range. However, the diurnal mixed layer and the mixing layer depth were within ASIP's operational range during the

Figure 2. Time series of the boundary conditions with cumulative hours of ASIP deployments on the x axis. The different cruises are separated by their background colors and the different deployments by vertical lines. (a) Modeled (red) and measured mean period Tm_{01} (green) and inverse wave age $A = u_{*a}/c_p$ (black). (b) Surface buoyancy flux B_0 (red) and surface heat flux Q_0 (black). (c) Significant wave height H_s for the full wave spectrum from wave observations (green) and from the ECWAM wave model (red) and the wind sea part of the wave spectrum H_{sw} (black) derived from runs ECWAM wave model. (d) The measured (black) and modeled (red) wind speed time series u_{10} (yellow and gray indicate MIDAS data conducted from the buoy and ship, respectively). (e) Scaling velocities, i.e., water-side friction velocity u_* calculated from the measured u_{10} using the COARE algorithm (red), as well as determined from the eddy covariance u_{*e} (green), the surface Stokes drift velocity u_{s0} (gray), and the buoyancy velocity w_B (black). (f) Levels of TKE dissipation ϵ with overlaid mixing layer depth h_e (red).

first half of the cruise. For these deployments, it could be shown that the depth of the daily mixed layer h_{σ} , defined as a density difference of $\Delta\sigma = 0.03 \text{ kg m}^{-3}$ from 2.5 m depth (de Boyer Montégut et al., 2004), agreed best with the above defined h_e . This definition for h_{σ} was applied on the observations from the mooring. The obtained h_{σ} from the mooring was then used as a reference for h_e in the cases of missing information of the actual h_e sampled by ASIP.

The 10 day long campaign **NICE** (oN thIn ICE) studied the role of mixing for the heat and biogeochemical budgets of the Arctic Ocean. The cruise was conducted on board the R/V Håkon Mosby in September 2014.

3.2. Measuring ϵ

ASIP is fully autonomous and allows for measurements up to the air-sea interface. This is a great advantage over most previous studies of turbulence measurements, which typically discard the upper 5–10 m due to contamination from the ship's wake (e.g., Brainerd & Gregg, 1993; Oakey & Elliott, 1982; LG89). Autonomous profilers such as the ARGO floats and sea gliders circumvent the problem of the ship wake (Rudnick, 2016; Wolk et al., 2009). However, to prevent their pumped conductivity sensor sucking in air, most ARGO floats stop sampling at a few meters below the ocean surface (Anderson & Riser, 2014).

ASIP is equipped with two orthogonal mounted airfoil shear probes (Lueck et al., 2002; Oakey & Elliott, 1982; Osborn, 1980). With a constant rise velocity of 0.5 m s^{-1} , ASIP fulfills Taylor's frozen field hypothesis which allows us to transform the measured velocity "time series" into a "space series" to obtain estimations of the shear. In order to estimate ϵ , the shear spectra were integrated in wave number space for 1,024-point segments, corresponding to 1 s or about 0.5 m bins of data. These segments were moved along the profile with an overlap of 512 points, which corresponds to around 0.25 m (Sutherland et al., 2013). Following the usual technique, Kolmogorov's theory for isotropic stationary turbulence is applied. This technique assumes the shear spectrum follows a universal shape of the Nasmyth spectrum that only depends on ϵ and ν . This

Nasmyth spectrum is used to extrapolate the observed shear spectrum of each segment to high and low wave numbers. Then ϵ is obtained by integrating the spectrum over the full extrapolated wave number range (Moum et al., 1995). For the analysis presented here, the ϵ profiles are temporally averaged over one hour intervals, which guarantees at least three sequential profiles per bin.

3.3. Wave Data

Wave parameters are taken from a dedicated model integration of the ECMWF stand-alone ECWAM wave model, cycle 43R1 (ECMWF, 2016), with hourly output from a global 11 km grid with a spectral resolution of 36 directions (10° resolution) and 36 frequencies (lowest frequency 0.035 Hz, and with standard logarithmic spacing toward the higher frequencies, see ECMWF, 2016). The model was run for the different cruise periods, with the wind forcing and the sea ice cover taken from ECMWF operational analyses, available every 6 h, with a resolution of 16 km. No wave data were assimilated and the runs were conducted without any surface ocean current information. For each run, a spin-up period of 10 days was used to ensure that the swell field is properly represented.

The atmospheric fluxes were obtained from ERA-Interim (Dee et al., 2011), using the short range forecasts at 3 hourly intervals (forecast steps 3, 6, 9, and 12) to calculate the different fluxes.

The surface Stokes drift is standard output from ECWAM (ECMWF, 2016). The computation of the low-frequency (resolved) part in arbitrary water depth is

$$\mathbf{u}_s(z=0) = \int_0^{2\pi} \int_{f_0}^{f_c} \frac{2gk}{2\pi f \tanh kH} \mathbf{k} F(f, \theta) df d\theta, \quad (16)$$

where θ is measured clockwise from north (going to), f_0 is the lowest resolved frequency, and f_c is a high-frequency cutoff beyond which the spectrum is treated as diagnostic. In deep water the Stokes drift velocity profile can be written as

$$\mathbf{u}_s(z) = \frac{16\pi^3}{g} \int_0^{2\pi} \int_{f_0}^{f_c} f^3 \hat{\mathbf{k}} e^{2kz} F(f, \theta) df d\theta, \quad (17)$$

and the f^3 factor makes it evident that the near-surface Stokes drift is disproportionately affected by the high frequencies. The high-frequency tail has been shown by Breivik et al. (2014) to account for about 30% of the surface Stokes drift. This means that a large part of the Stokes drift is supported by the unresolved part of the spectrum. ECWAM computes this high-frequency tail by assuming that the water depth is large compared to the high-frequency wavelength. A standard Phillips spectrum f^{-5} tail is fitted to the frequency bin which is $2.5\bar{f}_{\text{windsea}}$, where $\bar{f}_{\text{windsea}}$ is the mean frequency of the wind sea. The high-frequency contribution is treated as a diagnostic addition to the low-frequency (resolved) Stokes drift associated with the prognostic part of the spectrum as described by Breivik et al. (2014).

A vertical Stokes drift velocity profile was fitted to the Stokes transport and the surface Stokes drift, which are both taken from the wave model integration. The procedure is described by Breivik et al. (2016b) and essentially rests on the assumption that for high frequencies the Phillips spectrum (Phillips, 1958) is a reasonable match to the wave model spectrum, and that the higher frequencies are responsible for the bulk of the near-surface Stokes drift. This allows us to recreate the near-surface shear of the Stokes drift with only the integrated parameters u_s , H_s , and $\bar{f}$ (Breivik et al., 2016b).

4. Results

4.1. Wind-Wave-Induced Turbulence

Figure 3 shows the hourly averaged observations of ϵ , which were scaled using (7) and H_{sw} calculated from the wind sea part of the spectrum (Drennan et al., 1996; T96). The “breaking depth” z_b and the averaged “transition depth” $\bar{z}_t$ ($\pm$ its standard deviation) are indicated in Figure 3. Following T96 the depth is normalized with H_{sw} , whose magnitudes are shown in Figure 2.

The coefficient α based on F in water-side quantities and the exponent b in (7) were determined for the depth range above z_t and separately for the range above h_e using a linear regression. As shown in Figure 3a, the best fit through the observations above z_t is obtained with $\alpha = 8.07$ and $b = -1.29$, where $|z|^{-b}$

Figure 3. (a) Vertical distribution of the measured ϵ normalized with the ratio of the significant wave height based on the wind sea part of the wave spectrum H_{sw} to the cubic water-side friction velocity $\frac{H_{sw}}{u_*^3}$, versus the normalized depth z/H_{sw} for the full 70-min time-averaged data. The data are divided into subsets according to the inverse wave age $A = \frac{u_{*a}}{c_p}$: (b) $0.03 < A < 0.035$. (c) $0.035 < A < 0.04$. (d) $0.04 < A < 0.045$. (e) $0.045 < A < 0.05$. (f) $0.05 < A < 0.055$. (g) $0.055 < A < 0.06$. (h) $0.06 < A < 0.065$. The color scale gives the range of A . The dashed horizontal lines show the “breaking depth” z_b , the mean “transition depth” $\bar{z}_t$ (with its standard deviation as vertical thick black lines), and the mean mixing layer depth $\bar{h}_\epsilon$ as green dashed line. Linear regressions through the data are performed for all the data above h_ϵ (red) and above z_t as suggested by T96 (dashed in blue). For reference the T96 scaling is plotted in black and the LOW in gray.

describes the depth dependency of the ϵ profiles (summarized in Table 3). This depth dependency of ϵ is closer to the actual wall layer dependency of $|z|^{-1}$ than to the dependency proposed by T96 of $|z|^{-2}$. We conclude that the observed magnitude and depth dependency of ϵ profiles indicate that the LOW follows our measurements more closely than the scaling proposed by T96.

As it is suggested that ϵ is enhanced in the presence of breaking waves, relative to the LOW, potential differences between nonbreaking and breaking wave conditions should be investigated. With this intention, the data set is divided into ensembles, based on the ambient wind speed: $u_{10} < 4 \text{ m s}^{-1}$ and $u_{10} > 8.5 \text{ m s}^{-1}$ (Figure 4). Wind speed is here used as a proxy for the fraction of breaking waves. No wave breaking is expected to occur at wind speeds below 4 m s^{-1} , and 8.5 m s^{-1} was chosen as a threshold because it lies in the range of wind speeds for which whitecapping parameterizations start to diverge (Scanlon & Ward, 2016, Figure 3). Figure 4 shows that for both the low and higher wind speeds, the depth dependency of the observed ϵ follow closer to the classical wall layer scaling of $|z|^{-1}$ than a suggested $|z|^{-2}$ ($|z|^{-1.14}$ above h_ϵ and $|z|^{-1.34}$ above z_t for low-wind speeds and $|z|^{-1.12}$ above h_ϵ and $|z|^{-1.20}$ above z_t for higher wind speeds). However, our data were mostly collected in situations of lower wind situations; therefore, not many wave-breaking events are expected. Hence, there could still be regimes of higher dissipation under larger waves than observed in this study. T96 collected their data in wind conditions of $7\text{--}16 \text{ m s}^{-1}$, which includes less low-wind conditions but similar maximum wind speeds as observed in this study.

The data distribution of the scaled ϵ measurements in Figure 3a spans over 2 orders of magnitude above z_t . This range in magnitudes could be explained by the wave energy parameter α being a function of wave age. Dividing the data set into subsets based on the inverse wave age $A = \frac{u_{*a}}{c_p}$ (Figures 3b–3f), and determining α and b for each of these subsets using a linear regression as before, shows a clear dependency of α on A : $\alpha = 26\text{--}(388 A)$ above z_t and $\alpha = 24\text{--}(361 A)$ above h_ϵ (Figure 5). These functions are valid for a range of A between 0.03 and 0.065. For the presented range, these functions describe the values of α well, with determination coefficients of $R^2 = 0.96$ and $R^2 = 0.97$ for depth levels above z_t and above h_ϵ , respectively. Here the

Table 3

Values of the Exponent b and Coefficient α in (7) Determined Via Linear Regression Through the Measurements Above h_e and Above z_t for the Full Data, the Data Separated Into Subsets According to Their Inverse Wave Age A as in Figure 3, α as a Function of A as in Figure 5, and Separated Into Low and High Wind Speeds as in Figure 4

	Above h_e	Above z_t
<i>All</i>		
b	-1.15	-1.29
α	6.04	8.07
R^2	0.97	0.93
<i>Inverse wave age subsets</i>		
$0.03 < A < 0.035$		
b	-1.15	-1.33
α	11.53	13.89
R^2	0.97	0.43
$0.035 < A < 0.04$		
b	-1.29	-1.26
α	12.90	13.70
R^2	0.95	0.59
$0.04 < A < 0.045$		
b	-1.32	-1.34
α	9.52	10.20
R^2	0.98	0.91
$0.045 < A < 0.05$		
b	-1.29	-1.26
α	7.15	6.78
R^2	0.95	0.93
$0.05 < A < 0.055$		
b	-1.20	-1.29
α	5.02	6.65
R^2	0.91	0.93
$0.055 < A < 0.06$		
b	-1.23	-1.29
α	3.93	4.43
R^2	0.89	0.73
$0.06 < A < 0.065$		
b	-0.95	-1.08
α	2.12	2.87
R^2	0.99	0.94
<i>In sum</i>		
$\bar{b}$	-1.20 [± 0.127]	-1.27 [± 0.087]
α	26–388 A	24–361 A
<i>Wind speed subsets</i>		
$0 \text{ m s}^{-1} < u_{10} < 4 \text{ m s}^{-1}$		
b	-1.14	-1.34
α	17.48	31.91
R^2	0.86	0.98
$u_{10} > 8.5 \text{ m s}^{-1}$		
b	-1.12	-1.20
α	2.73	2.97
R^2	0.68	0.73

coefficient of determination, R^2 (the square of Pearson's linear correlation coefficient), is weighted by the uncertainties in α . The 95% confidence interval of α is determined from a linear regression based on (7). All the values for α and b are summarized in Table 3.

In contrast to α , the exponent b does not depend on A and stays constant as $b = -1.27$ [± 0.087] (above the z_t and $b = -1.20$ [± 0.127] above the h_e) for the whole range of sea states. The determined coefficient for b and the relation between α and A , which are summarized in Table 3, allow us to formulate a refined scaling approach based on (7):

$$\epsilon_{\text{wind},s}(z) = (7.2 - 108.3A) \left(\frac{u_*^3}{H_{sw}} \right) \left(\frac{z}{H_{sw}} \right)^{-1.15} \quad (18)$$

in this case determined for the the range above h_e .

HQ10 modeled the effect of wave-induced turbulence on ϵ with (11). We normalized our observations with the scaling approach proposed by HQ10 based on separated wind and swell seas (Figures 6a and 6b), respectively, as well as the sum of both (Figure 6c). The HQ10 wind sea scaling (Figure 6a) underestimates our observations close to the sea surface above a depth corresponding to H_s but describes the observations better below (albeit with a scatter of 2 order of magnitude). The most appropriate value for β in (12) is determined to be 1.53 (Table 4). The normalized data based on the swell (Figure 6b) deviates considerably from the observed ϵ , with highest underestimation below H_s (several orders of magnitudes). The sum of both the wind and swell part of the spectrum (Figure 6c) describes the observed ϵ better than the separated sea states. The determined $\beta = 0.97$ (Table 4) agrees with HQ10 who suggested that $\beta \leq 1.0$. However, there remains significant scatter which is a function of the surface Stokes drift velocity u_{s0} (see the color coding in Figure 6).

The surface Stokes drift is sensitive to errors in the wind field, but not more so than other wave parameters. By comparison, the significant wave height is proportional to the square root of the zeroth moment m_0 of the wave spectrum and is more sensitive to errors in the swell representation. The surface Stokes drift is more directly related to the wind sea spectrum since it is heavily weighted toward the higher frequencies of the spectrum, which respond quickly to changes in the wind field. We find that the analyzed wind field is generally in good agreement with the observations from the cruises for which we avail wind measurements (see Figure 2d).

The values of higher Stokes drift shear at the surface from the Phillips approximation relative to the ones for a monochromatic wave, give higher estimates for ϵ in (11). This is reflected in a lower value of $\beta = 0.21$ (Figure 6d) relative to the HQ10 scaling for monochromatic waves (Figure 6c), where $\beta = 0.97$ (Table 4). For the approach based

on the monochromatic wave, most of the variability among the different parameterizations can be explained by u_{s0} . For lower u_{s0} ($u_{s0} < 0.06 \text{ m s}^{-1}$) the scaled profiles underestimate the measured ones and vice versa for higher u_{s0} as these surface values directly influence the values of the modeled ϵ in (11). Using the Phillips approximation results in less variability (root-mean-square error [RMSE] of 3.26 compared to 3.79 for the monochromatic wave), especially closer to the surface, and the parameterizations for periods of lower and higher u_{s0} fall closer together than based on the conventionally proposed approach (Figure 6d). Applying $\beta = 0.21$ in (12) and using the Stokes drift shear from the Phillips approximation (14) allows us to reformulate HQ10's scaling (11) as

Figure 4. Same as in Figure 3 but with the color scale giving the range of u_{10} : (a) for the full 70 min time-averaged data, (b) for wind speeds less than 4 m s^{-1} , and (c) for wind speeds higher than 8.5 m s^{-1} .

$$\epsilon_{\text{wave}}(z) = 0.79 \pi \sqrt{\frac{H_s}{\lambda}} u_*^2 \alpha_p \sqrt{-\frac{2\pi g}{z} \text{erfc}(\sqrt{-2kz})}. \quad (19)$$

4.2. Buoyancy-Induced Turbulence

To identify periods during which convective forcing dominates over wind-wave forcing, the ratio of h_ϵ to the Langmuir stability length $L_L = u_*^2 u_{s0} / B_0$, where a positive B_0 describes destabilizing periods, is used (Sutherland et al., 2014a; B12). The threshold for a dominant convective forcing is considered to be unity ($h_\epsilon / L_L = 1$ for h_ϵ defined as being positive), so that in cases of $h_\epsilon / L_L > 1$, buoyancy forcing is considered as important.

Figure 5. The value for α in (7) determined through the linear regressions in Figure 3 as a function of the inverse wave age $A = u_{*a} / c_p$. The regression for the ϵ data above the mixing layer depth h_ϵ is given by $\alpha = 24 - 361A$ (red) and above the “transition depth” z_t by $\alpha = 26 - 388A$ (blue). The error bars represent the 95% confidence level of α and the horizontal dashed lines show the value of α determined from the full data set.

Figure 7a shows the measured ϵ normalized with the LOW ($\tilde{\epsilon}_{\text{LOW}} = \epsilon / \epsilon_{\text{LOW}}$) for both the wind-wave-dominated regime ($h_\epsilon / L_L < 1$) and the convective regime ($h_\epsilon / L_L > 1$). The near uniformity of $\tilde{\epsilon}_{\text{LOW}}$ ($\tilde{\epsilon}_{\text{LOW}} = 0.90$) of the wind-wave-dominated regime indicates that when convection is minor, ϵ approximately follows the LOW scaling (Figure 7a). The observed ϵ during periods of strong convection is better scaled by including B_0 (Figure 7b). At depths below H_s , the ϵ measured for the convective regime normalized by B_0 ($\tilde{\epsilon}_{B_0} = \epsilon / \epsilon_{B_0}$) stays relatively uniform over depth. This is typical as convectively forced turbulence has not been found to display a significant depth dependency (Callaghan et al., 2014; LG89). The higher values of $\tilde{\epsilon}_{B_0}$ near the surface could be explained by the presence of shear.

In order to yield a refined formulation, $\epsilon_{\text{buoyancy}}$, to describe the observed ϵ in a similar form to LG89, we apply the averaged values of $\tilde{\epsilon}_{\text{LOW}}$ and $\tilde{\epsilon}_{B_0}$ as scaling coefficients (summarized in Table 5). In periods where the wind-wave forcing dominates, $\epsilon_{\text{buoyancy}}$ is given by the averaged $\tilde{\epsilon}_{\text{LOW}}$. In this case a value of 0.90 of $\tilde{\epsilon}_{\text{LOW}}$ scales the observed ϵ best. To account for the additional buoyancy effect when convection is dominant, the similarity scaling for those periods is a linear combination of the LOW and B_0 . Thus, $\epsilon_{\text{buoyancy}}$ is extended by the averaged $\tilde{\epsilon}_{B_0}$ of 0.91 times the prevailing B_0 :

Figure 6. Vertical distribution of the measured ϵ normalized with the scaling proposed by HQ10 (ϵ_{HQ10}) versus the normalized depth z/h_ϵ for the full 200 min time-averaged data. (a) For the wind sea part of the wave spectrum. (b) Swell part of the wave spectrum. (c) Modeled ϵ from both parts added together with Stokes drift shear for a monochromatic wave as suggested by HQ10. (d) The same is calculated for the Stokes shear profile based on the Phillips approximation. The colorbar indicates the surface Stokes velocity u_{s0} . The horizontal green dashed line represents h_ϵ and the horizontal black dashed line represents the mean of the significant wave height $\bar{H}_s$ (with its standard deviation as thick vertical black line). The vertical red line gives the average through the data and is expressed by β in (12).

$$\epsilon_{buoyancy_*}(z) = \begin{cases} 0.90 \times \epsilon_{LOW} & \text{for } h_\epsilon/L_L < 1 \\ 0.63 \times (0.90 \epsilon_{LOW} + 0.91 B_0) & \text{for } h_\epsilon/L_L > 1. \end{cases} \quad (20)$$

Figure 7c shows that the extended similarity scaling reduces the variability relative to the ones based on wind or buoyancy only.

In conditions when convection dominates over wind-wave forcing ($h_\epsilon/L_L > 1$ red in Figure 7c), ϵ is well scaled with scale factors of u_* and B_0 . Even most of the observations of ϵ during situations in which wave breaking is expected to occur, which are indicated by black circles in Figure 7c, fall in the midst of the normalized data. Highest bias with up to more than an order of magnitude is found for periods during which convection is less important ($h_\epsilon/L_L < 1$ shown in green in Figure 7c). This suggests that wind and buoyancy alone do not explain all the turbulence for this period and wave-induced turbulence might contribute. The outlying profile in Figure 7c left of the mean normalized ϵ , which was taken during wave-breaking conditions, falls into the period during which wind-wave forcing dominates over convection. These data were taken during the NICE cruise when u_{10} , u_* , and H_s increased, and B_0 reached minimal negative values, and thus wind-wave-induced turbulence is suggested to be important. In comparison to the scaling in Figure 7c, the ϵ profile is well described by the regression line in Figure 4c, which additionally considers wave information in form of the significant wave height H_s . In this figure, all the high wind speed data fall within the cluster.

The scaling approach suggested by B12 in (15) formulates a linear combination of ϵ from the three processes (i.e., shear, Stokes drift, and buoyancy) producing TKE at the ocean surface. The time series of u_* , estimated from the measured u_{10} by the COARE algorithm, and u_{s0} , determined from the ECWAM model, is

suggestive of a linear relationship between both velocity scales (Figure 2a). It is well known that the relationship between the wind sea part of the spectrum and its Stokes drift is close to linear (see Ardhuin et al., 2009; Kenyon, 1969; Li et al., 2017). This relationship is applied for regimes of constant La_t as shown in the regime diagram in Figure 8. A constant La_t predicts no variation within the composition of the dominant wind-wave-turbulence generation processes over the periods of ASIP's deployments.

A linear regression $u_{s0} = -5.5 \times 10^{-4} + 12.75 u_*$ explains 67% of the variability between u_* and u_{s0} (Figure 9). The slope is very close to the value 12 found by Gargett and Grosch (2014) for

Table 4

Values of the Factor β in (12) Indicated as Red Vertical Lines in Figure 6 and the Respective Root-Mean-Square Error (RMSE) Determined for the HQ10 Scaling in the Same Figure

	β	RMSE
<i>Monochromatic wave</i>		
Wind sea part	1.53	4.09
Swell part	158.17	47.15
Both parts added	0.97	3.79
Phillips spectrum	0.21	3.26

Figure 7. Vertical distribution of measured ϵ for destabilizing periods of positive surface buoyancy flux $B_0 > 0$ normalized with: (a) LOW; (b) normalized with B_0 ; (c) normalized with a combination of both. The data are divided into periods during which convection is important ($h_e/L_L > 1$) in red and during which convection is less important ($h_e/L_L < 1$) in green. The mean values are indicated by the vertical lines for periods during which convection is important (red) and convection is less important (green). The blue line shows the value suggested by LG89. These mean values are used to determine the normalization in Figure 7c. The horizontal dashed green line indicated the mixing layer depth h_e and the dashed black line shows the mean significant wave height $\bar{H}_s$ with the thick vertical black lines being its standard deviations. The circled data indicate situations during which u_{10} is higher than 8.5 m s^{-1} .

cases in which Langmuir circulation dominates the turbulence forcing. Also, this slope agrees with a slope of ≈ 11 – 12 by Kitaigorodskii et al. (1983) as well as a derived slope between 9.2 and 13.8 from the Pierson-Moskowitz (PM) spectrum by Li and Garrett (1993). Forcing the described linear regression between the water-side friction velocity u_* and u_{50} to pass through the origin, changes the relation to

$$u_{50} = 12.04 u_*, \quad (21)$$

which does not significantly reduce the statistics as the coefficient of determination is still $R^2 = 0.67$.

Considering this linear relation between u_{50} and u_* allows us to substitute u_{50} in (15) with (21) as

$$\epsilon_{B12} = \left(2 - 2e^{-0.5\sqrt{1/12.04}} + 2.65 \right) \frac{u_*^3}{h} + 0.3B_0. \quad (22)$$

By setting La_t to a constant in this equation, the scaled turbulence is forced to have its origin in a specific wind-wave regime as discussed above. This means that no distinction between a Langmuir or wind-dominated regime exists, but rather u_{50} is greater than the wind stress by a constant factor.

The formulation in (22) of the scaling suggested by B12 has the same form as that suggested by LG89 in (5). Both scalings overestimate the observed ϵ profiles (Figure 10). The $\epsilon_{\text{buoyancy}}$ describes these observations with a RMSE of 2.50 compared to 3.11 for the B12 scaling from (22) and compared to 3.25 for the LG89 scaling.

4.3. Evaluation of Coefficients and Scaling Relationships

The coefficients for α , b , β , $\bar{\epsilon}_{\text{LOW}}$, and $\bar{\epsilon}_{B_0}$, as summarized in Table 6, were extracted for the whole data set (i.e., the data from all five cruises). To determine the optimal scaling, these coefficients are compared

to the ones gained from the same analysis based on training data sets, which consist of four of the five cruises with respectively one (the testing data set) being withheld from each run. This procedure results in five sets of coefficients summarized in Table 6.

The RMSE of these tests is highest for the ϵ_{wave} scaling (with the respective coefficients) for the tests on the Labrador and the Knorr11 cruise for negative B_0 conditions. Also, the ϵ_{wind} scaling for negative B_0 and the ϵ_{wave} for positive B_0 (with the respective coefficients) show

Table 5

Values of the Factors $\bar{\epsilon}_{\text{LOW}}$, $\bar{\epsilon}_{B_0}$, and Their Combination $\bar{\epsilon}_{B_0} \times B_0 + \bar{\epsilon}_{\text{LOW}} \times \epsilon_{\text{LOW}}$ in Figure 7 and Shown by the Vertical Lines in the Same Figure

	$\bar{\epsilon}_{\text{LOW}}$	$\bar{\epsilon}_{B_0}$	$\bar{\epsilon}_{B_0} \times B_0 + \bar{\epsilon}_{\text{LOW}} \times \epsilon_{\text{LOW}}$
$h_e/L_L < 1$	0.9		1.00
$h_e/L_L > 1$		0.91	0.63
LG89	1.76	0.58	0.87

Figure 8. (top) The regime diagram for mixing in the OSBL as introduced by B12 for (a) the Labrador cruise in orange, (b) the Knorr11 cruise in blue, (c) the STRASSE cruise in green, (d) the MIDAS cruise in red, and (e) the NICE cruise in black. The thick solid black lines divide the diagram into regions where the single forcing produce more than 90% of the total ϵ . As the regime diagram is only defined for periods of $B_0 > 0$. (bottom) The remaining data as a function of the turbulent Langmuir number La_t to fall between wave- and wind dominated turbulence. During the NICE campaign, all data fall in the $B_0 > 0$.

clearly higher RMSE in relation to the other testing data sets. Both of these cruises were located along the Canadian continental shelf.

For both the Knorr11 and Labrador cruise the ϵ_{wave} , consistently overpredicts the measured ϵ by more than 3 orders of magnitude. This suggests that the waves induced by the Stokes shear affect the growth of turbulence less in these shelf areas than in the other sites of the Atlantic. This was already indicated in the regime diagram in Figure 8, where the data of those both cruises reach least far into the wave regime during negative B_0 conditions. Therefore, other processes are expected to cause the prevailing turbulence. In the case of the Knorr11 cruise this might be wind. The ϵ_{wind} , (with the α and b from Table 6) under-predicts the measured ϵ for periods of negative B_0 , but describes them within 1 order of magnitude for positive B_0 conditions.

The overall RMSE (averaged) is lowest for the ϵ_{wind} , scaling for positive B_0 conditions with $RMSE = 7.07$, for which we only compare the scalings based on wind-wave-induced turbulence. This ϵ_{wind} , and the $\epsilon_{buoyancy}$, scaling collapse to the measured ϵ of the respective training data sets within an order of magnitude as shown in Figure 11. Thereby, the $\epsilon_{buoyancy}$, describes the magnitudes of the measured ϵ better than the other scalings for $B_0 > 0$ conditions with $\overline{\epsilon_{buoyancy}} = 1.18 [\pm 2.339]$ being closest to unity (Figure 11b and Table 7). For $B_0 < 0$ conditions, $\overline{\epsilon_{wind}} = 0.85 [\pm 3.089]$ is closest to unity (Figure 11a and Table 7). The $\epsilon_{buoyancy}$, scaling describes our data best for conditions when convection dominates over wind-wave-induced turbulence ($h_e/L_L < 1$). This is represented by the relatively higher RMSEs, for example, for the Knorr11 cruise ($RMSE = 14.68$ in Table 6), where during positive B_0 conditions most of the data are spread in the wind-waves domain, and thus convection is considered to be less important (Figure 8b).

The performance of these scaling approaches seems suitable even though not all possible sources of turbulence in the OSBL are included. For example, internal waves (breaking or nonbreaking) enhance the observed turbulence by redistributing energy vertically or by releasing energy when they actually break. Additionally, submesoscale processes and inertial shear are further sources of turbulence that are expected to enhance ϵ relative to the introduced scaling approaches.

The averaged coefficients from the analysis on all training data sets are similar (within their range of error) to those found for the whole data set, which gives confidence in these coefficients determined through the analysis of this study. Excluding the coefficients gained from the two training data sets with the highest RMSE yields similar values for the coefficients of α , b , β , $\overline{\epsilon_{LOW}}$, and $\overline{\epsilon_{B_0}}$.

5. Discussion

Our observations support a depth dependency of ϵ described by $b = -1.29$ above z_r , which is less than that proposed by T96 ($b = -2$), which was supported by observations in Lake Michigan (Wang & Liao, 2016), in the Pacific Ocean at depths below approximately one H_s (Sutherland & Melville, 2015), and in the laboratory (Siddiqui & Loewen, 2007) at depths below $0.4 H_s$. For measurements in coastal waters of the South China Sea, Shuiqing and Dongliang (2016) find an even more gentle slope with $b = -2.11$. Feddersen et al. (2007) show a slightly smaller decay rate of $b = -1.9$ in nearshore regions, which still is higher than the b found in the presented study. The studies that we refer to are summarized with their measurement technique and depth in Table 8.

Greenan et al. (2001) found that the T96 scaling holds for the case in which the swell can be easily separated. However, for open ocean conditions in the

Figure 9. The surface Stokes velocity u_{s0} determined from the ECWAM wave model versus the water-side friction velocity u_* determined from the measured u_{10} using the COARE algorithm. The red line indicates the linear fit through the data, which is given by $u_{s0} = 12.04 u_*$.

$z_b = 0.01 H_s$ (Thomson et al., 2016, for a wave-following reference frame). However, these studies observe ϵ closer to the surface than ϵ could be determined from ASIP measurements.

In the wave age range covered by T96, α -values between 90 and 250 were found, where α is determined from the water-side normalized TKE flux from the wind to the waves F . The highest values of α were found at levels of A around 0.0667. These values were confirmed by Feddersen et al. (2007) ($\alpha = 250$) and Gerbi et al. (2009) ($\alpha = 168$), which are larger than used in the modeling study of Craig and Banner (1994). Other observational studies within the same wave age regime find slightly smaller values of α , e.g., Jones and Monismith (2008), $\alpha = 54 \pm 6$. This lower α is explained by a more limited conversion of the wind energy to wave breaking.

Figure 10. Bin-averaged vertical distribution of the measured ϵ normalized with the scaling suggested by LG89 (green circles), B12 (red stars), and $\epsilon_{buoyancy*}$, determined for this study in Figure 7c (blue diamonds). Colored background areas give the 95% confidence interval and the respective root-mean-square error (RMSE) is shown. The horizontal dashed green line is h_ϵ . The horizontal dashed black line the significant wave height $\bar{H}_s$ with the thick black vertical line being the standard deviations of the H_s .

presence of more complex sea states, where wind sea and swell interact, the T96 scaling does not hold and ϵ seems to decay rather with $|z|^{-1}$ than with $|z|^{-2}$. T96 measured in Lake Ontario with young seas in the absence of swell. The T96 scaling was confirmed in the open ocean by Drennan et al. (1996) but most of the cases were under wind sea conditions. Similarly, Gemmrich (2010) observed ϵ profiles that decayed with $|z|^{-1}$ in a depth range between $\approx 2 H_s$ and $\approx 0.3 H_s$. Only above $\approx 0.3 H_s$ they found a layer where ϵ was enhanced with respect to the profile decaying with $|z|^{-1}$. Thomson et al. (2016) found values for b that fall between the depth dependency of the LOW and T96 as -1.56 ± 0.03 for a fixed reference frame at depth shallower than $z/H_s = 10^{-1}$.

Near the surface, above $z_b = 0.6 H_s$, the T96 scaling predicts a layer of constant dissipation. However, this is not supported by our 1867 open ocean profiles of ϵ . Soloviev and Lukas (2003) and Gemmrich and Farmer (1999, 2004) question the feasibility of scaling a “breaking layer” with H_s using wave-following observations. T96 never actually observed this layer of constant ϵ as their sensors were located below wave troughs from a fixed tower-based system. In fact, they determined this depth by assuming that the integrated water column dissipation would equal the energy input of the wind. The “breaking layer” would then be the depth at which half of the wind energy input has dissipated. Some studies find shallower depth levels for this “breaking depth,” e.g., $z_b = 0.4 H_s$ (Jones & Monismith, 2008; Siddiqui & Loewen, 2007; Wang & Liao, 2016), $z_b = 0.25 H_s$ (Young & Babanin, 2006) and

$z_b = 0.01 H_s$ (Thomson et al., 2016, for a wave-following reference frame). However, these studies observe ϵ closer to the surface than ϵ could be determined from ASIP measurements.

In the wave age range covered by T96, α -values between 90 and 250 were found, where α is determined from the water-side normalized TKE flux from the wind to the waves F . The highest values of α were found at levels of A around 0.0667. These values were confirmed by Feddersen et al. (2007) ($\alpha = 250$) and Gerbi et al. (2009) ($\alpha = 168$), which are larger than used in the modeling study of Craig and Banner (1994). Other observational studies within the same wave age regime find slightly smaller values of α , e.g., Jones and Monismith (2008), $\alpha = 54 \pm 6$. This lower α is explained by a more limited conversion of the wind energy to wave breaking.

We find smaller values of α than Jones and Monismith (2008), and significantly smaller (1 order of magnitude) ones than those by T96. These low values of α are related to our lower levels of ϵ , see (7). The $\bar{\alpha}_m$ in the ECWAM (see Breivik et al., 2015; ECMWF, 2016; for details on the calculation of α in ECWAM) lies in the range of 143.74 ± 112.70 with low-wind outliers being removed. Compared to T96’s parameterization of α on A from Wang and Huang (2004), we find a different behavior, as we see a consistent decrease for α with A even for A that are below 0.0833. In a similar way, Shuiqing and Dongliang (2016) observed a decrease of α with A and found higher values of α when using their complete data set; when focusing only on young seas with A above 0.0286, α was lower. However, their values of $\alpha = 1,700$ (full data) and $\alpha = 230$ (young seas) are still higher than the ones here. Similar to the results presented here, Thomson et al. (2016, Figure 6) suggests a decrease in $\bar{c}$ with inverse wave age, defined as u_{10}/c_p , for young seas (c_p/u_{10} smaller than 2, which corresponds to $A \sim 0.0148$). They show that the assumption of a constant $\bar{c}$ is supported by histograms showing $\bar{c}$ centering around 2 m s^{-1} . This $\bar{c}$ is related to α as

Table 6

Values of the Coefficients α , b , β , $\overline{\epsilon_{\text{LOW}}}$, $\overline{\epsilon_{B_0}}$, and $\overline{\epsilon_{B_0} \times B_0 + \overline{\epsilon_{\text{LOW}}} \times \epsilon_{\text{LOW}}}$ From the Five Training Data Sets With the Name Referring to the Respective Testing Data Set of Each Run Analysis, the Averaged Coefficients of These Runs, and the Coefficients Gained From the Analysis Using the Whole Data Set (Including Data From the Five Here Presented Cruises)

	Labrador	Knorr11	STRASSE	MIDAS	NICE	Averaged	Whole data
b (above z_t)	−1.30	−1.29	−1.19	−1.38	−1.30	−1.29 [± 0.068]	−1.29
b (above h_e)	−1.14	−1.13	−0.88	−1.42	−1.20	−1.154 [± 0.193]	−1.15
α (above z_t)	26–395 A	26–388 A	24–362 A	28–425 A	25–363 A	25.8 [± 1.48]–392 [± 36.6] A	26–388 A
α (above h_e)	24–371 A	24–365 A	18–288 A	36–549 A	22–316 A	24.8 [± 6.72]–377.8 [± 101.7] A	24–361 A
RMSE ϵ_{wind_+} ($B_0 < 0$)	208.92	66.74	2.23	9.87		38.31	
RMSE ϵ_{wind_+} ($B_0 > 0$)		10.20	2.90	6.79	7.35	7.07	
RMSE ϵ_{wind_+} (all)	208.92	7.49	1.93	4.33	7.35	7.35	
β	0.22	0.24	0.18	0.22	0.2	0.212 [± 0.022]	0.21
RMSE ϵ_{wave_+} ($B_0 < 0$)	543.71	318.62	5.08	6.49		162.56	
RMSE ϵ_{wave_+} ($B_0 > 0$)		46.04	6.37	4.98	16.90	11.64	
RMSE ϵ_{wave_+} (all)	543.71	21.59	3.43	3.37	16.90	16.90	
$\overline{\epsilon_{\text{LOW}}}$	0.90	0.94	1.12	0.81	0.82	0.917 [± 0.126]	0.90
$\overline{\epsilon_{B_0}}$	0.91	0.96	1.14	0.60	0.89	0.90 [± 0.195]	0.90
$\overline{\epsilon_{B_0} \times B_0 + \overline{\epsilon_{\text{LOW}}} \times \epsilon_{\text{LOW}}}$	0.59	0.61	0.61	0.65	0.59	0.610 [± 0.025]	0.63
RMSE $\epsilon_{\text{buoyancy}_+}$ ($B_0 > 0$)		14.68	7.03	3.08	9.54	8.29	

Note. The root-mean-square error (RMSE) refers to the tests of the gained coefficients on the testing data sets and the RMSE for the averaged data refers to the median of the single RMSEs.

$\bar{c} = \alpha u_*$. When applying an average $u_* = 0.0066 \text{ m s}^{-1}$ for the data presented here, $\bar{c}$ of 2 m s^{-1} corresponds to an α of 303.

Determining ϵ from ASIP's shear measurements as described does not allow for estimates of ϵ closer than 0.25 m to the air-sea interface. This is for example a limitation for studies on the air-sea gas exchange as highlighted in Esters et al. (2017). Other instruments, such as SWIFTS (Thomson, 2012) measure ϵ acoustically up to the air-sea interface. These buoys resolve the upper 0.6 m with a 5 min temporal resolution and, therefore, yield higher-resolution time series of ϵ . However, these buoys do not allow for any assertion of ϵ deeper than 1 m. Therefore, a combination of the advantages of both types of instruments would be desirable: ASIP's ability to gain turbulence information down to the mixing layer depth and SWIFTS ability to gain high spatial and temporal resolution ϵ directly at the air-sea interface.

The data analyzed in the preceding sections were conducted at different locations within the Atlantic Ocean. Different sources of turbulence could be expected in the tropics (STRASSE and MIDAS), the Canadian continental shelf region (Labrador and Knorr11), and the Arctic region (NICE). The regime diagram introduced by B12 shows regions where a single forcing (wind, waves, or buoyancy) produces the majority of the total ϵ . The top plots in Figures 8a–8e cover periods of positive B_0 . As the diagram is not defined for negative B_0 , those periods are shown in the bottom plot (Figure 8) as a single function of La_t . Most of the data are cluttered in the regime range where no single forcing is expected to explain the total of the prevailing turbulence. Some of the ϵ data from the MIDAS cruise are mainly driven by buoyancy effects, which explains the $\epsilon_{\text{buoyancy}_+}$ to scale the observed ϵ profiles well during periods of positive B_0 , in particular for this

Table 7

Mean [$\pm$ Standard Deviation] and Root-Mean-Square Error (RMSE) of the ϵ Observations From the Testing Data Sets Normalized by the Different Scaling Assumptions as Determined in Figure 11 Using the Coefficients From the Respective Training Data Sets for Daytime ($B_0 < 0$) and Nighttime Conditions ($B_0 > 0$) As Well As for All the Data

	ϵ_{wind_+}		ϵ_{wave_+}		$\epsilon_{\text{buoyancy}_+}$	
	Mean $\pm$ std	RMSE	Mean $\pm$ std	RMSE	Mean $\pm$ std	RMSE
$B_0 < 0$	0.85 [± 3.089]	38.31	0.49 [± 2.926]	162.56		
$B_0 > 0$	1.72 [± 3.431]	7.07	1.37 [± 3.556]	11.64	1.18 [± 2.339]	8.29
All data	1.24 [± 3.080]	7.35	0.87 [± 4.033]	16.90

Figure 11. Bin-averaged vertical distribution of the measured ϵ normalized with the improved wind-wave based scaling $\epsilon_{wind,}$ (green); wave-based $\epsilon_{wave,}$ (blue); buoyancy-based $\epsilon_{buoyancy,}$ (red). The profiles are averaged over the five respective profiles from the “testing” analysis and the symbols indicate the mean values for periods of (a) negative surface buoyancy flux B_0 ; (b) positive B_0 ; (c) all data.

cruise. Only data gained in the Arctic during the NICE cruise are expected to be mainly driven by wave effects. The data from the STRASSE cruise cover the smallest range of Lq_t and are most constrained to the wave region, in particular for negative B_0 , whereas the data from the other tropical cruise (MIDAS) cover a much wider range of Lq_t . Additional turbulence sources, not considered in the B12 regime diagram, include internal waves that might break along the continental shelves, as well as submesoscale processes, and inertial shear, which could occur at these sites; therefore, shape the local turbulence characteristics differently.

Table 8

Studies Cited in This Paper Listing the Location of Experiments, Number of ϵ Measurements, the Measurement Technique, and Respective Depth Range

	Location	Number of ϵ estimates	Measurement technique	Depth range
Lombardo and Gregg (1989)	North Pacific Ocean	Almost 700 profiles	Advanced Microstructure Profiler (AMP)	≈ 5 –300 m
Anis and Moum (1995)	Off the Oregon coast	Total of 160 profiles	Vertical profiler Chameleon	0.5 m to about 120 m
Terray et al. (1996)	Lake Ontario	13 runs with a total of 42 estimates	Three types of current meters	0.153–2.151 m
Drennan et al. (1996)	Off Maryland Coast	20 runs	Acoustic travel-time current meters (Minilab SD-12)	2 m
Greenan et al. (2001)	Scotian Shelf	Total of 324 profiles	Tethered free-fall vertical profiler (EPSONDE2)	5 m to sea bed
		75 profiles	Quasi-horizontal profiler (EPSONDE-Glider)	2 m to sea bed
Siddiqui and Loewen (2007)	Wind-wave flume	8 runs with 13 depth levels	Digital particle image velocimetry (PIV)	1–25 mm
Feddersen et al. (2007)	Nearshore region of Atlantic Ocean	56 h of data	Three ADVs	0.56, 1.32, 1.86 m above the bed (water depth 3.2 m)
Jones and Monismith (2008)	Grizzly Bay in San Francisco Bay	3,240 measurements	Four Acoustic Doppler velocimeter (ADV)s	0.15, 0.5, 1.5, 2 m above bed (water depth 2.5 m)
Huang and Qiao (2010)			Same as Anis and Moum (1995)	
Sutherland and Melville (2015)	Pacific Ocean	70	Array of acoustic Doppler profiler (subsurface)	0.06–7
			Stereo IR particle image velocimetry (surface)	0 m
Shuiqing and Dongliang (2016)	Coastal waters of South China Sea	≈ 150	ADV	4–6 m
Thomson et al. (2016)	North Pacific (50°N, 145°W)	2 522 (10 min burst length)	SWIFTs	0–0.5 m
Wang and Liao (2016)	Lake Michigan	6 runs	Free-floating PIV (FPIV)	0–15 cm

6. Conclusion

Parameterizing turbulence in the OSBL is of great importance as observations of turbulence in the ocean are still limited. Microstructure measurements of ϵ from five cruises in different parts of the Atlantic Ocean have been presented. ASIP took ϵ profiles from the mixing layer depth all the way up to the sea surface; a significant advantage over traditional microstructure profilers. The observations were used to evaluate and improve existing parametrizations of ϵ , summarized as follows:

1. T96 proposed wind and wave forcing to model profiles of ϵ . The data presented here do not support the idea of the near-surface region comprising three different layers, with a constant dissipation layer near the surface. In addition, our open ocean measurements do not show the existence of a “breaking layer” as proposed by T96. Instead, the depth dependency of the observed ϵ follows a decay of $|z|^{-1.15}$, which is closer to the slope of the traditional LOW ($|z|^{-1}$) than to the transition layer as suggested by T96 with $|z|^{-2}$.
2. Wave-turbulence interaction as a mechanism for ocean turbulence generation was proposed by HQ10. They used the Stokes shear profile of a monochromatic wave to model profiles of ϵ . Here we found that the scaling improved when used with the Stokes shear profile based on the Phillips approximation as suggested by Breivik et al. (2016a).
3. A linear relationship of u_* with the surface Stokes drift velocity u_{s0} allowed us to substitute u_* for u_{s0} in the scaling of B12. This permitted for a direct comparison between the scaling suggested by B12 and the similarity scaling proposed by LG89. Both approaches consider a destabilizing buoyancy forcing, and therefore could be applied under conditions of surface cooling.
4. During daytime conditions of negative buoyancy flux, our observation are best described by a scaling relationship:

$$\epsilon(z) = (7.2 - 108.3A) \left(\frac{u_*^3}{H_{sw}} \right) \left(\frac{z}{H_{sw}} \right)^{-1.15}, \quad (23)$$

which considers the inverse wave age A and the friction velocity u_* .

5. In conditions when convection dominates over wind-wave forcing with $h_e/L_L > 1$, the observations are best described by

$$\epsilon(z) = 0.63 \times \left(0.90 \times \frac{u_*^3}{\kappa|z|} + 0.91 B_0 \right). \quad (24)$$

Acknowledgments

Funding for this research was provided by a College of Science (NUIG) PhD Fellowship. B.W., K.H.C., and G.S. were supported by the Norwegian Research Council under projects 233901. Ø.B. and K.H.C. gratefully acknowledge the support of the Research Council of Norway through the CIRFA (grant 237906) and RETROSPECT (grant 244262) projects. Travel grants were provided by the Ryan Institute at NUIG and the Marine Institute. We also thank the captains and crew of the CCGS Hudson, R/V Knorr, R/V Thalassa, R/V Tangorua, and R/V Håkon Mosby. We thank the two anonymous reviewers for their thoughtful reading of this manuscript. Data are available at <https://www.dropbox.com/sh/w3vmbtywgu5t34y/AABpRYv270tfVcm1ouhD8Mr1a?dl=0>.

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