Estimation of Kinetic Energy Dissipation from Breaking Waves in the Wave Crest Region

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ABSTRACT

Wave-induced turbulence due to breaking in the absence of surface shear stresses is investigated experimentally. A high-fidelity particle image velocimetry (PIV) technique is used to measure the turbulence near the water surface, inside the wave crests. The spatial velocity vector fields of the breaking waves acquired from PIV provide accurate vertical velocity profiles near the air–water interface, as well as wavenumber velocity spectra beneath the breaking waves at different depths. These velocity spectra exhibit a Kolmogorov interval at high wavenumbers, indicating the presence of isotropic turbulence and permitting an estimation of energy dissipation rates. The depth dependence of dissipation rates of the breaking waves generated in the laboratory shows a scaling similar to that found in wind-forced breaking waves in the field. A phase dependence in the dissipation rates of turbulence kinetic energy is also observed, which should be considered to improve the accuracy of the estimated and modeled wave energy dissipation.

1. Introduction

Turbulence due to breaking waves plays an important role in air–sea interaction processes. For example, the transfer of energy and momentum over the air–sea interface and the enhancement of mixing under wave crests are both closely related to wave-induced turbulence. Also, the breaking of waves is the main energy dissipation mechanism for wind-generated waves (e.g., Babanin 2011; Sutherland and Melville 2015). The dissipation rate close to the water surface is critical because the dissipation is a measure of mixing near the water surface and is important for understanding wave dynamics and gas transfer, and also for improving spectral wave models and ocean mixing models. Numerous laboratory and field experiments and theoretical and numerical studies have been performed and important progress has been made in understanding the physics of wave-breaking turbulence. However, the great range of parameters in the ocean environment and the complexity of turbulent fluid flows require further investigation to improve predictions of dissipation in breaking waves. Whitecapping (breaking) dissipation is one of the three major source terms for wave-forecast models, but it is very difficult to measure because of the sporadic nature of the wave-breaking events. To estimate the total dissipation, the volumetric dissipation rate through the water column is required and then integrated. However, the profile of the volumetric dissipation rate diverges near the surface, and hence artificial constraints were always introduced in close proximity to the mean water level. Therefore, measurements in the water, but above the wave troughs and certainly within the wave crests, are critical for such practical applications, but are most challenging.

A classical wall (surface) layer similarity scaling for turbulence in a boundary layer gives an interpretation of the dissipation rate of turbulence kinetic energy \( \varepsilon = \frac{U_*^3}{\kappa z_{\text{wall}}} \), where \( U_* \) is friction velocity, \( \kappa \) is the Von Kármán constant, and \( z_{\text{wall}} \) is the vertical distance from a solid surface. However, it has been shown that the presence of wave breaking with strong wind enhances the dissipation close to the water surface (Agrawal et al. 1992; Melville 1994; Terray et al. 1996; Drennan et al. 1996; Soloviev and Lukas 2003; Gemmrich 2010;...
Sutherland and Melville (2015; Thomson et al. 2016). For example, Terray et al. (1996) proposed a scaling for the rate of dissipation based on the conclusion that the dissipation rate under breaking wave conditions depends on wind input and distance from the water surface. They obtained higher dissipation values than expected from the classical wall layer scaling and attributed this to wave breaking. Drennan et al. (1996) observed similar dissipation rates to the wave-dependent scaling of Terray et al. (1996) from field data obtained in oceanic conditions. Soloviev and Lukas (2003) also observed enhanced dissipation rates under breaking waves and showed that the dissipation profiles agreed well with the scaling in Terray et al. (1996). However, because of the limitation of experimental techniques, measurements near the water surface were not available to draw strong conclusions about the dissipation rate near the wavy surface. Recently, Gemmrich (2010), Sutherland and Melville (2015), and Thomson et al. (2016) have successfully made field measurements close to the wave crests. Gemmrich (2010) showed the first direct field measurements of the turbulent kinetic energy dissipation beneath deep water wave crests, resulting in a parameterization of the depth dependence of dissipation rates. Sutherland and Melville (2015) observed an increase in the dissipation rate above the classical scaling, which is attributed to non-air-entraining microbreakers. Thomson et al. (2016) found agreement with the depth scaling reported by Gemmrich (2010) and Sutherland and Melville (2015) when the vertical reference frame is matched. Gemmrich (2010) and Thomson et al. (2016) used a fixed vertical reference frame and a wave surface following reference frame to demonstrate that the depth scaling of the turbulent kinetic energy dissipation rates depends on the vertical reference frame. Throughout this paper, the wave-surface-following reference frame will be used, with \( z_w \) denoting the distance from the wave surface and \( z_f \) denoting depth in the fixed reference frame, where \( z_f = 0 \) is the mean water level. All these studies involved observations of waves in the presence of wind forcing.

The objective of the current study is to improve the overall understanding of wave-induced turbulence and its influence on the turbulent kinetic energy dissipation near the water surface. In particular, we seek to address and observe depth and phase dependencies on the turbulent kinetic energy dissipation rate close to the free surface due to wave-induced turbulence in the absence of wind shear stresses along the air–water interface. To this end, mechanically generated monochromatic deep water waves without wind forcing are observed in a laboratory wave tank described below.

2. Experimental setup

Experiments are performed in the Extreme Air–Sea Interaction (EASI) facility located in the Michell Hydrodynamics Laboratory at the University of Melbourne. EASI consists of a wave tank having dimensions of \( 60\,\text{m} \times 2\,\text{m} \times 2\,\text{m} \) (length × width × height), with a wind tunnel constructed above it to allow studies on wind-forced waves (Alberello et al. 2016). The current investigation only considers mechanically forced waves, generated by a computer-controlled wave maker that can generate waves with predefined wave forms. At the downstream end of the tank, there is a minimum-reflection beach to dissipate incoming wave energy. The tank is filled to a depth of 0.9 m with freshwater. Underwater optical access is provided through a viewing window located 33.5 m from the wave maker. The stationary image acquisition system, consisting of a laser, camera, and optics, is set up in front of the viewing window to obtain images as the waves propagate downstream. Two-dimensional particle image velocimetry (PIV) is employed to determine the instantaneous velocity fields, providing streamwise/wall-normal planes of data. The PIV setup consists of a Lintron dual head Nd:YAG laser that delivers 150 mJ per pulse at 15 Hz and an Andor complementary metal–oxide–semiconductor (CMOS) camera. The camera resolution is 2560 × 2120 pixels and is equipped with a Nikor f/3.5 60 mm macro lens. The field of view obtained from this camera configuration is approximately 170 mm × 200 mm (x × z). A schematic of the wave tank facility and the PIV setup are shown in Fig. 1. Throughout this paper, x, y, and z represent the streamwise (wave-propagation direction),
spanwise, and vertical directions, respectively. The ob-
jective distance between the camera sensor plane and the
laser light sheet is approximately 1000 mm. The laser
beam is fanned out into a sheet using a series of optical
lenses. The laser sheet thickness is approximately 2 mm,
illuminating an $x$–$z$ plane along the center line of the wave
tank. Hollow glass spheres with a mean particle diameter
of $10^{-5}$ m are used to seed the tank. All images are ac-
cquired at a sampling frequency of 15 Hz. The acquired
images are processed using an in-house PIV package. A
final interrogation window size of $32 \times 32$ pixels and
with a 50% overlap are used. To minimize errors in pro-
cessed velocity vector fields, the small regions of bubble
clouds near the surface of breaking waves are masked
during the PIV processing. An in situ calibration target is
imaged to obtain an accurate pixel to physical space
conversion factor. Further descriptions of the PIV pro-
cessing algorithms and calibration technique are available
in de Silva et al. (2014).

The experiment is performed with an initially steep
monochromatic wave with peak period $T = 0.8$ s and
steepness of 0.4, which is similar to the nonseeded ex-
periments described in Babanin et al. (2010), where the
nonlinear instability emerges from the background
noise as the waves propagate downstream. The surface
elevations of the generated waves at the PIV measure-
ment location are recorded by a capacitance-type wave
gauge. The surface elevation profiles $\eta$ and corre-
sponding surface elevation spectra $P_\eta$ of the steep
monochromatic waves near to the wave maker and at
the PIV station are shown in Fig. 2, and the significant
wave height $H_s$ is 120 mm. The emergence of the in-
stability due to the growth of side bands in the spectra is
indicated in Fig. 2.

3. Results and discussions

The results obtained from the PIV measurements
provide two-dimensional instantaneous velocity fields of
streamwise ($U$) and vertical ($W$) components. The mean
water column velocity $U$ of the breaking wave below the
wave crests as a function of the distance from the surface
$z_w$ normalized by $H_s$ (=120 mm) is obtained by aver-
aging over 20 tests and is shown in Fig. 3. With this
spatial dataset, the velocity spectra of the breaking
waves are directly computed and the corresponding
wavenumber $k$ is obtained without having to convert
from the temporal domain. The range of water depth
from the wave crest for which spectra could be computed
is $5 \leq z_w \leq 110$ mm. The spatial velocity fields permit an
insight into the distribution of wave-generated turbulence
inside the breaking crest region. The wavenumber spectra
of the breaking waves as a function of the depth below the
wave crest are shown in Fig. 4. As expected, the magni-
tude of the spectra decreases with the distance away from
the water surface. It should be noted that since the
streamwise field of view obtained from the PIV mea-
surements is less than the wavelength, a spectral peak
due to large-scale orbital motion is not captured. The
velocity spectra at high wavenumbers ($k > 500 \text{ rad m}^{-1}$)
exhibit a trend close to $k^{-5/3}$ (shown as a solid line in the
figure) throughout the depth considered, indicating a
Kolmogorov interval and the presence of isotropic tur-
bulence. The energy level of the Kolmogorov interval is
used to estimate the volumetric turbulence kinetic en-
ergy dissipation rate $\varepsilon$ using the relation suggested by
Veron and Melville (1999).

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P(k) = \frac{18}{55} \left( \frac{8\varepsilon}{9\alpha} \right)^{2/3} k^{-5/3},
$$

where $\alpha \approx 0.4$ is the Heisenberg constant. Using Eq. (1),
the profile of volumetric kinetic energy dissipation rate

![Fig. 2](image-url)  
Fig. 2. Time history of (left) the surface elevation $\eta$ and (right) the corresponding spectra $P_\eta$ of the monochromatic waves at (top) the wave maker and (bottom) the PIV station.

![Fig. 3](image-url)  
Fig. 3. Mean water column velocity $U$ profile under the crest of a breaking wave, obtained from PIV. Water depth is scaled with the significant wave height $H_s$. 

![Fig. 4](image-url)
\( \varepsilon(\eta) \) as a function of the distance from the water surface can be estimated, as shown in Fig. 5. The depth dependence of \( \varepsilon \) has previously been investigated in field and laboratory experiments (Soloviev et al. 1988; Terray et al. 1996; Drennan et al. 1996; Babanin et al. 2005; Gemmrich 2010; Sutherland and Melville 2015; Thomson et al. 2016). The parameterizations of the vertical dissipation rate \( \varepsilon \) have been determined to predict the total dissipation rate based on water depths in the literature.

Soloviev et al. (1988) obtained \( \eta^{-1} \) dependence, consistent with boundary layers over a solid wall, whereas Terray et al. (1996) showed that in the presence of strong winds, the depth dependence of dissipation follows \( \varepsilon(\eta) \propto \eta^m \), with \( m = -2 \), and attributed this stronger dependence to wave breaking. To make the total energy dissipation converge, they imposed the limit of \( \eta > 0.6H_w \), where \( H_w \) is a significant wave height, and assumed that \( \varepsilon(\eta) \) remains constant when the measured height is below the limit \( \eta \leq 0.6H_w \). Babanin et al. (2005) observed both \( \eta^{-1} \) and \( \eta^{-2} \) dependencies; the former one corresponds to light winds (no breaking) and the latter to stronger winds and therefore persistent wave breaking. They also had to assume that the constant dissipation layer exists, in their case for \( \eta < 0.4H_w \), and the volumetric rate of total turbulent kinetic energy dissipation for \( \eta > 0.4H_w \) was determined by independently measured wind energy inputs. More recently, Gemmrich (2010) performed field observations to investigate velocity profiles beneath deep water wave crests of wind-driven wave fields for breaking and nonbreaking cases. They determined that the depth dependence of the kinetic energy dissipation rate was related to wave saturation and phase. Also, on average, the vertical profiles of the dissipation rate approximately show \( \varepsilon(\eta) \propto \eta^{-1.1} \) for the nonbreaking case and \( \varepsilon(\eta) \propto \eta^{-1.6} \) for the breaking case. Sutherland and Melville (2015) made measurements within \( O(10) \) cm of the air-sea interface and calculated the turbulent kinetic energy dissipation throughout the wave-affected surface layer. They also observed that the dissipation rate approached \( \varepsilon(\eta) \propto \eta^{-2} \) below one significant wave height from the sea surface and \( \varepsilon(\eta) \propto \eta^{-1} \) closer to the surface. They concluded that this enhanced energy dissipation rate was caused by non-air-entraining microbreakers. It is reminded that all previous parameterizations of the volumetric dissipation rate are based on data from forced waves.

Here, a relationship between the dissipation rate and depth is obtained from purely mechanically generated waves to examine the effect of the wave-induced turbulence. The water depths investigated herein are all referenced to the frame following the wave surface. In Fig. 5, the squares represent \( \varepsilon \) taken beneath the wave crest region with the depth from the water surface \( \eta \), normalized by \( H_w \). The solid line indicates the depth dependence of the dissipation rate with an average exponent \( m = -2 \). Here, it should be noted that \( \varepsilon(\eta) \propto \eta^{-1} \) is also observed (dashed line in Fig. 5) near the water surface similar to the trend reported by Sutherland and Melville (2015) from their field measurements. Again, the obtained average exponent is from mechanically generated waves only, whereas previously reported values are acquired from wind-generated waves (Terray et al. 1996; Babanin et al. 2005; Gemmrich 2010). The resulting exponents of the current study are obtained without considering the influence of bubbles on the turbulent kinetic energy dissipation rates (see Lamarre and Melville 1991; Blenkinsopp and Chaplin 2007; Lim et al. 2015). Nevertheless, the depth scaling of \( \varepsilon \) seems to agree well with the field experiment of Sutherland and Melville (2015), who measured the turbulent kinetic energy dissipation beneath breaking waves in the presence of bubbles. This could be explained by the fact that the region of bubbles covers less than 1% of the total sea surface, suggesting an estimated \( \varepsilon \) in the absence of bubble dynamics can be used to simulate most of the sea states as discussed by Schwendeman and Thomson (2015) and Thomson et al. (2016).

Measurements of the \( \varepsilon \) beneath wave troughs obtained a half period after the wave crests are shown in Fig. 5 (circles). It is observed that the rate of dissipation is lower and decreases less rapidly with depth \( \varepsilon(\eta) \propto \eta^{-1.3} \) (dotted–dashed line in Fig. 5) under the wave troughs than under the crests. This result is consistent with Gemmrich (2010), who showed that the depth-scaling exponent was lower at wave troughs compared to crests. It appears the exponent is related to the wave phase and this could be due to a stage of wave
saturation. For example, when the turbulence is induced at the breaking wave crest, the rate of volumetric kinetic energy dissipation is higher owing to the increased turbulence intensity near the wave interface. However, as the induced turbulence is distributed within waves while propagating further downstream (change in phase), the intensity of the turbulence decays as well as the dissipation rate. More recently, Thomson et al. (2016) suggested that the depth penetration of the turbulence is strongly associated with wave orbital motions. They observed that the turbulence generated by breaking waves persists longer than one wave period and thus the surface turbulence is vertically advected from crest to trough by the wave orbital motions. Finally, a constant dissipation layer is not observed at any depth considered in this study.

4. Conclusions

A high-fidelity PIV experiment investigating breaking-wave-induced turbulence was conducted. To examine the turbulence generated purely by breaking processes, the waves are mechanically generated in the absence of shear stresses from wind forcing. With the spatial velocity vector fields obtained from the PIV experiment, the velocity spectra at various depths below the water surface for a crest and a trough are computed. The spectra beneath the breaking waves show the Kolmogorov interval at high wavenumbers, indicating the existence of wave-breaking-induced turbulence. The estimated depth scaling of the volumetric kinetic energy dissipation rate in close proximity to the wave surface shows no evidence of a constant dissipation layer (at least at the depths considered in this study). In a wave-surface-following reference frame, the depth dependence of dissipation rates after the wave breaking show \( \varepsilon(z_w) \propto z_w^{-m} \) with \( m = -2 \) and \( -1.3 \) below the wave crests and troughs, respectively. In addition, the kinetic energy dissipation rate switches to \( \varepsilon(z_w) \propto z_w^{-1} \) as the surface is approached. The results are consistent with the field observations of Gemmrich (2010) and Sutherland and Melville (2015), who both considered wind-forced waves. Despite the differences in forcing (wind versus mechanical), the depth dependence of the turbulent kinetic energy dissipation rate is in good agreement with the observations in this investigation. It is therefore concluded that the depth dependence of the dissipation rate is similar regardless of the presence of wind stresses along the air–water interface, confirming the importance of wave breaking to wave-induced turbulence close to the surface. Further, a phase dependence on the energy dissipation rate is also observed, suggesting the inclusion of correlations between wave phase and dissipation rate could improve the accuracy of approximating the dissipation term in wave models.

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