

Dissipation of wave energy and turbulence in a shallow coral reef lagoon

Zhi-Cheng Huang,^{1,2} Luc Lenain,¹ W. Kendall Melville,¹ Jason H. Middleton,³ Benjamin Reineman,¹ Nicholas Statom,¹ and Ryan M. McCabe⁴

Received 6 April 2011; revised 14 November 2011; accepted 3 January 2012; published 10 March 2012.

[1] Simultaneous in situ measurements of waves, currents and turbulence are presented to describe dissipation rates of wave energy and turbulent kinetic energy in the windward coral reef-lagoon system at Lady Elliot Island (LEI), Australia. The dissipation of wave energy in the lagoon is tidally modulated and strongly correlates with frictional dissipation due to the presence of the extremely rough bottom boundary. The observed turbulent kinetic energy (TKE) dissipation rate, ϵ , in this wave-dominated lagoon is much larger than recently reported values for unidirectional flows over natural fringing coral reefs. The correlation between the wave dissipation and ϵ is examined. The average rate of dissipation induced by the rough turbulent flow was estimated directly from the observed ϵ coupled with both a depth-integrated approach and with a bottom boundary layer scaling. Rates of TKE dissipation estimated using the two approaches approximate well, within a factor of 1.5 to 2.4, to the surface-wave energy dissipation rate. The wave dissipation and friction factor in the lagoon can be described by a spectral wave-frictional model with a bottom roughness length scale that is approximately constant across the lagoon. We also present estimates of dissipation induced by the canopy drag force of the coral heads. The dissipation in this case is enhanced and becomes more significant for the total energy dissipation when the water depth in the lagoon is comparable to the height of the coral heads.

Citation: Huang, Z.-C., L. Lenain, W. K. Melville, J. H. Middleton, B. Reineman, N. Statom, and R. M. McCabe (2012), Dissipation of wave energy and turbulence in a shallow coral reef lagoon, *J. Geophys. Res.*, 117, C03015, doi:10.1029/2011JC007202.

1. Introduction

[2] Coral reefs are prominent features of shallow water in tropical and subtropical nearshore regions. Energy dissipation over coral reefs plays an important role in reef morphology, marine organism distribution, island shoreline stability and, in particular, nutrient uptake, which is important for sustaining coral reef communities. Nutrient uptake is positively correlated with bottom shear stress and water velocity under unidirectional [Baird and Atkinson, 1997; Baird *et al.*, 2004; Bilger and Atkinson, 1992] and oscillatory wave-induced flows [Falter *et al.*, 2004]. However, wave-induced flows often generate higher bed shear stresses than comparable unidirectional flows [Nielsen, 1992] and therefore can increase

mass transfer and the dispersion of nutrients in coral reef communities [Reidenbach *et al.*, 2009, 2006b]. The coupling of flow velocity with bottom roughness and bed shear stress leads to thinning of the diffusive boundary layers surrounding biota, resulting in the maximum nutrient uptake being proportional to $\epsilon^{1/4}$, where ϵ is the turbulent dissipation rate [Falter *et al.*, 2004; Hearn *et al.*, 2001]. Consequently, quantification of energy dissipation rates is important for monitoring and modeling coral reef communities.

[3] The geometrical structure of coral reefs produces hydrodynamic environments distinct from those of beach systems. A steep transition from relatively deep to shallow water between the fore reef and outer reef flat leads to wave transformation involving shoaling, refraction, diffraction, and wave-breaking dissipation accompanied by an enhancement of frictional dissipation [Lugo-Fernández *et al.*, 1998b; Monismith, 2007; Young, 1989]. Wave transformation and attenuation have been studied primarily in the steep transition zone to the outer reef flat in laboratory models of fringing reefs [e.g., Gourlay, 1994, 1996a; Gourlay and Colleter, 2005; Massel and Gourlay, 2000] or over natural fringing and barrier reefs [e.g., Hardy and Young, 1996; Lowe *et al.*, 2005b; Lugo-Fernández *et al.*, 1998a, 1998b; Young, 1989]. Few studies of wave processes across the reef rim into a

¹Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California, USA.

²Now at Graduate Institute of Hydrological and Oceanic Sciences, National Central University, Taoyuan, Taiwan.

³School of Aviation, University of New South Wales, Sydney, New South Wales, Australia.

⁴School of Oceanography, University of Washington, Seattle, Washington, USA.

Table 1. Summary of Studies on Wave Dissipation and Turbulence Dissipation Over Coral Reefs^a

Study Focus	Reference	Site	Study Region	Hs Ocean (m)	Hs Local (m)	Depth h (m)	Bottom Roughness (cm)
Wave and turbulence dissipation	Present study	Lady Elliot Island, GBR, Australia	Lagoon/ Fringing reef flat	0.7–3	<0.48	<1.4	$H_{bed} = 27.2$, $\sigma_b = 14$, $k_w = 109$
Turbulence dissipation under unidirectional flow	<i>Reidenbach et al.</i> [2006a]	Gulf of Aqaba, Red Sea, Israel	Fringing reef	<0.2	<0.2	9–13	$H_{bed} = 20$
Wave dissipation	<i>Lowe et al.</i> [2005b] <i>Falter et al.</i> [2004] <i>Nelson</i> [1996]	Kaneohe Bay, Hawaii Kaneohe Bay, Hawaii John Brewer Reef, GBR, Australia	Barrier reef flat Barrier reef flat Barrier reef flat	0.5–1.6 0.9–1.3 –	<0.35 0.6–0.28 0.19–0.62	1.6 1–3 0.83–2.76	$H_{bed} = 7.2$, $\sigma_b = 3.6$, $k_w = 14$ $H_{bed} = 5–30$, $k_w = 20$ $k_w = 6–7$

^aHs, significant wave height; h, local water depth; H_{bed} , mean roughness height; σ_b , standard deviation of bottom elevation; k_w , equivalent Nikuradse roughness.

lagoon currently exist. The wave energy decay results in spatial gradients in radiation stress [Longuet-Higgins and Stewart, 1964], which is balanced by an across-shore pressure gradient and consequently induces wave setup [Gourlay, 1996a, 1996b; Jago *et al.*, 2007; Massel and Gourlay, 2000] which leads to wave-driven currents and circulation on the reef flat and further into the lagoon behind [Hearn, 1999; Lowe *et al.*, 2009a, 2009b; Symonds *et al.*, 1995].

[4] Depth-limited wave breaking has been considered a substantial source of energy dissipation at the seaward reef edge [Gourlay, 1994; Lowe *et al.*, 2005b; Massel and Gourlay, 2000; Young, 1989]. By analogy to the critical wave breaking parameter on beaches [Thornton and Guza, 1983], the ratio of wave height to water depth on the reef flat is a primary determinant of the occurrence of depth-limited wave breaking [Gourlay, 1994; Hardy and Young, 1996]. The energy dissipation caused by wave breaking on mild beaches was parameterized empirically by Thornton and Guza [1983] and Battjes and Janssen [1978]. These formulae have been successfully included in numerical models to simulate the variation of wave height, wave setup [Massel and Gourlay, 2000] and wave-induced circulation [Lowe *et al.*, 2009b] over coral reefs.

[5] On reef flats with rough bottoms in shallow water, frictional dissipation has been considered a primary component of the total dissipation [Falter *et al.*, 2004; Lowe *et al.*, 2007, 2005b; Nelson, 1996]. Rough reef bottom surfaces comprised of limestone and benthic organisms of coral colonies perturb the flow, generate turbulent shear stress and dissipate wave energy in the rough turbulent boundary layer [Monismith, 2007; Nielsen, 1992; Tennekes and Lumley, 1972]. Recently, Reidenbach *et al.* [2006a] measured boundary layer turbulence for unidirectional flows over a fringing coral reef in the Red Sea. They showed that turbulent boundary layer theory could be applied to flows over the rough bottom of those coral reefs, in which the corals occupied a small fraction of the water column. Few studies have estimated the turbulence dissipation over wave-dominated coral reefs; however, Reidenbach *et al.* [2009] showed that a wavy turbulent flow may increase the dispersion of nutrients, gametes and larvae in coral communities.

[6] It is difficult to measure the near-bed turbulence to account for frictional dissipation, but estimates have been achieved through measurements of wave energy-flux gradients with the hypothesis that the loss of total wave energy equals the frictional dissipation on the reef flat [e.g., Falter *et al.*, 2004; Lowe *et al.*, 2005b; Nelson, 1996]. Previous studies on wave dissipation and turbulence dissipation over coral reefs are summarized in Table 1. These studies estimated the frictional dissipation over coral reef flats, but the subsequent propagation and attenuation of waves across the reef rim into a rougher fringing reef lagoon and the relationship to coincident turbulent kinetic energy dissipation is largely unexplored. In view of the large wave losses and wave setup over the reef flat, the lagoon may be a very different wave and current regime. The measured bottom roughness height H_{bed} was approximately 7–20 cm on the reef flat of Kaneohe Bay Barrier Reef, Oahu, Hawaii [Falter *et al.*, 2004; Lowe *et al.*, 2005b] and approximately 6–7 cm on the reef platform of John Brewer Reef, Australia [Nelson, 1996]. Energy dissipation rates in a nearshore environment with

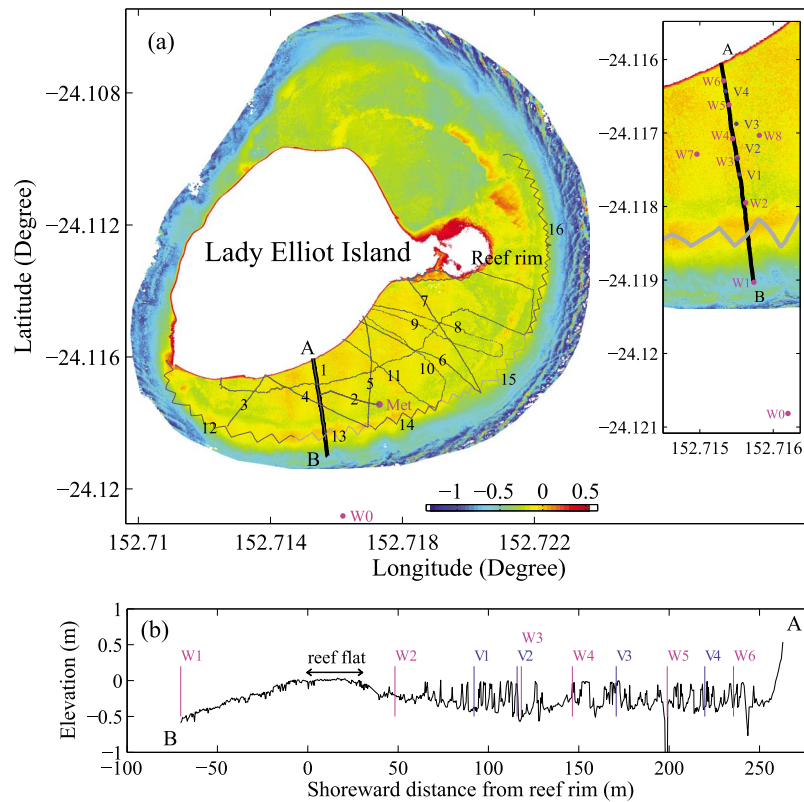


Figure 1. (a) Lady Elliot Island (located off the southeast coast of Queensland, Australia, in the Great Barrier Reef), surrounding wavefields, reef rim and transect of instrument deployment (A-B). The topographic map was constructed from multiple passes at low tide using an airborne portable scanning lidar system, described by *Reineman et al.* [2009]. Gray lines with marked numbers are manual bathymetric surveys. (b) Bathymetric transect of line A-B and the locations of the instruments. Wave gauges and ADV locations are shown as W0–W8 and V1–V4, respectively. The elevation is referenced to the mean rim elevation. Wave measurements at W1 and W2 failed.

shallower water depth and greater bottom roughness are largely unexplored.

[7] This paper presents field measurements of dissipation rates in the windward reef-lagoon system at Lady Elliot Island (LEI), Great Barrier Reef (GBR), off the coast of Queensland, Australia. Wave energy dissipation rates were measured and parameterized, and turbulence dissipation rates were measured using turbulence inertial subrange techniques with acoustic Doppler velocimetry (ADV) data. The possible equivalence of energy dissipation rates measured by these independent techniques was tested. Results are compared with linear wave and boundary layer theories and a bottom friction model. For very shallow depths, where the coral heights are comparable to the water depth, the canopy drag dissipation is also evaluated and discussed.

2. Field Experiments

2.1. Study Site

[8] Three weeks of field experiments (3–24 April 2008) were conducted in the windward coral reef lagoon at LEI (24.11°S, 152.72°E) located off the southeast coast of Queensland, Australia, in the Great Barrier Reef. The overall reef platform is kidney-shaped and about 1400 m wide and long, with an area of 1.9 km² [*Chivas et al.*, 1986]. Figure 1a

shows a topographic map of the island, the reef rim and its surrounding wavefields at spring low tide in 2008, as measured by an airborne scanning lidar system [*Reineman et al.*, 2009]. The line A-B oriented in a cross-reef direction was selected to measure the energy dissipation rate over the reef and lagoon. A manual GPS survey of the bathymetry is presented in Figure 1b. This manual survey used a Leica (Heerbrugg, Switzerland) GPS 1200, postprocessed, kinematic differential GPS with the base station located at the southern end of the island runway. The measured accuracy of the GPS unit was a few centimeters in each of the three coordinate directions. From the line A-B, it is observed that the barrier-reef lagoon system is comprised of the fore reef (seaward from the reef rim at $x = 0$), the reef rim (reef crest) ($0 < x < 30$ m) and the lagoon between the beach and the reef rim. Figure 2 shows images of the windward lagoon at LEI at spring low and high tides during the experiment. Depth-limited wave breaking is observed on the fore reef to the reef rim, with less breaking inside the lagoon.

[9] The seabed flow features are composed of the coral colonies, which include a wide variety of coral types, the boundary layer, and the larger reef geometry ranging from $O(10^{-2})$ to $O(10^3)$ m with irregular coral structures. *Grant and Madsen* [1986] have pointed out that for boundary layer models at different scales, a spatial average over many

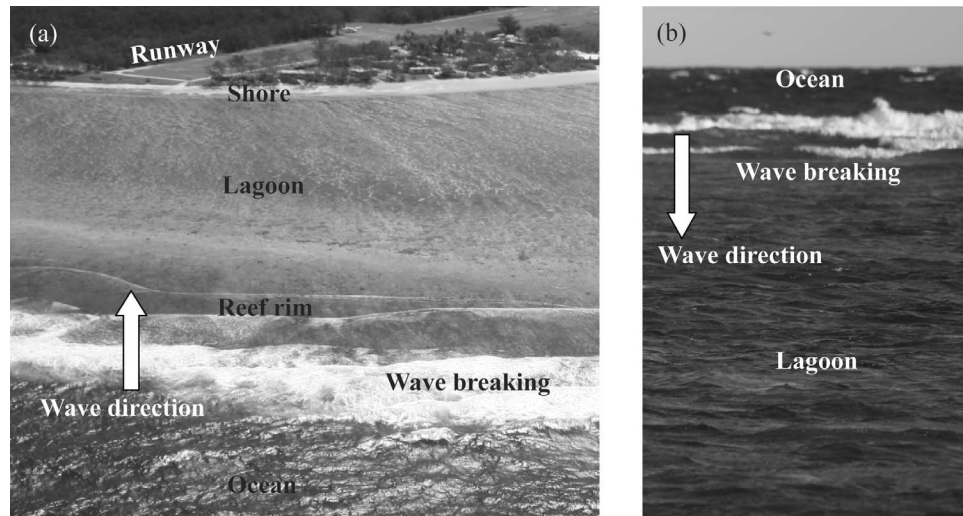


Figure 2. Pictures of the windward lagoon surrounding Lady Elliot Island at the spring (a) low and (b) high tides in April 2008. Note that most of the wave breaking is depth limited and occurs between the fore reef and the reef flat with minimal breaking inside the lagoon.

roughness elements is needed because a boundary layer flow is related to the total force acting over all roughness elements rather than to the details of the flow around an individual roughness element. A detailed bathymetric survey and statistical analysis is therefore needed to estimate the bottom roughness over the coral reef system. Detailed manual GPS bathymetric surveys were conducted over multiple lines in the windward lagoon. Standard deviations of the measured bottom elevation, σ_b , for sixteen surveyed lines are shown in Figure 3, with marked numbers corresponding to the gray lines in Figure 1a. Because of the geometrical complexity of the coral reefs, the surveys were performed with irregular horizontal spatial steps ranging from several meters down to decimeters. The highest average horizontal resolution is approximately 20 cm while the lowest is approximately 2.5 m. For the eleven cross-lagoon manual bathymetric surveys σ_b ranges from 11.4 to 15.6 cm, with an average value of 14 cm. In addition, the mean height of the bottom roughness element from the sand to the top of the coral, H_{bed} , along the surveyed line A-B is 0.27 m. The use of a single length scale of bottom roughness will be tested below to estimate the frictional dissipation rate for the windward lagoon. By contrast, the measured σ_b on the windward reef flat surrounding LEI is about 7–8 cm with individual roughness elements up to about 20 cm in different areas on the reef flat. This σ_b is much larger than the measured σ_b on the reef flat in Kaneohe Bay, Oahu, Hawaii (approximately 3.6 cm) [Lowe *et al.*, 2005b].

2.2. Observations

[10] A suite of instruments was deployed along the line A-B shown in Figure 1a to characterize waves, currents, and turbulence in the lagoon and the atmospheric conditions. An RBR Ltd. TWR2050 pressure and temperature (PT) sensor was deployed at site W0 in a water depth of approximately 18.5 m on the fore reef to measure the offshore tide, wave elevation and the incident wave energy flux. This sensor was set to record a 17 min data burst at 4 Hz every

hour. Six Banner Engineering QT50U ultrasonic wave gauges mounted on anchored tripods (W4, W4-1, W4-2, and W5, W7, W8) were deployed inside the lagoon to measure the wave energy flux (see Figure 1a). All ultrasonic wave gauge tripods were equipped with a wireless RF module (Digi Xtend), continuously sending the wave measurement to a recording ground station on shore. A triangular array of three ultrasonic wave gauges was mounted on the same tripod at W4, to measure the wave directional spectrum inside the lagoon. The additional PT sensors, built at the Hydraulics Laboratory, Scripps Institution of Oceanography,

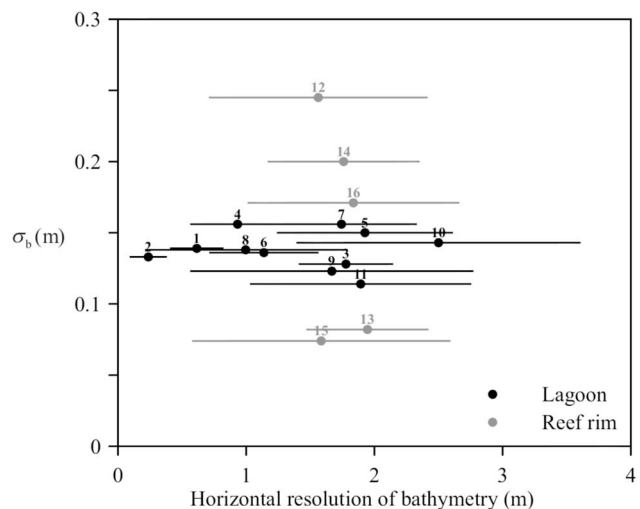


Figure 3. Standard deviation of the measured bottom elevation in the lagoon (transect numbers 1–11) and on the reef rim and flat (numbers 12–16), σ_b , versus average horizontal resolution of each GPS survey transect as shown in Figure 1a.

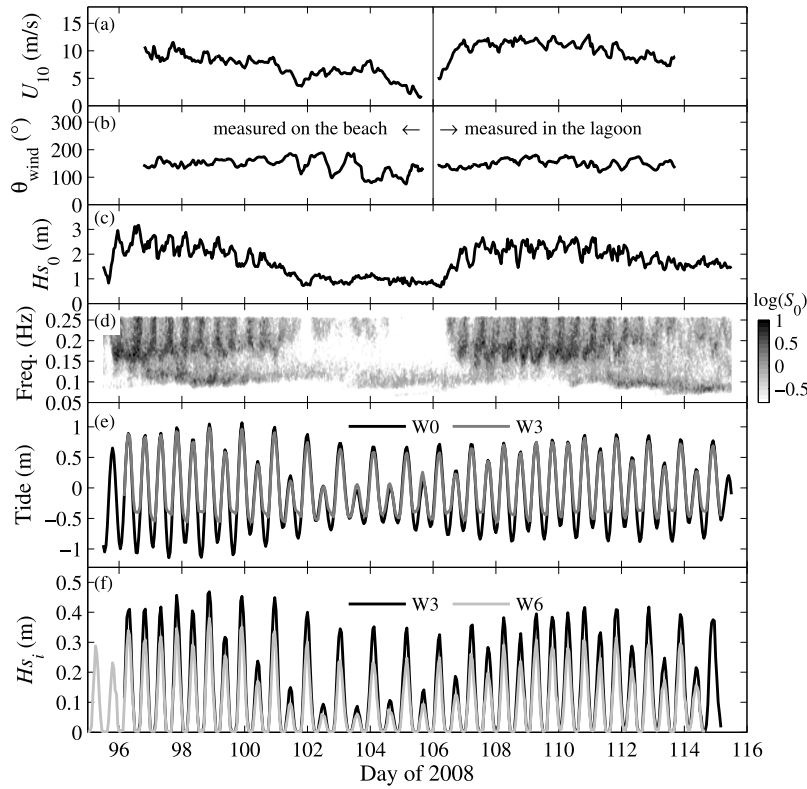


Figure 4. Time series of wind, wave and tide conditions. (a) Wind speed U_{10} and (b) wind direction θ_{wind} (from). (c) Significant wave height Hs_0 and (d) spectrogram of surface displacement in the ocean (site W0). (e) Tidal elevation (represented as a deviation from the mean) at W0 (black line) and W3 (dark gray line). (f) Significant wave height inside the lagoon Hs_i at W3 and W6 (bright gray line).

continuously sampled at 4 Hz while the wave gauges were set to record at 5 Hz for a 20–55 min burst every hour, at W3 and W6. Three 6 MHz Nortek Vector ADVs (V1, V2 and V4, oriented downward) and one 16 MHz Sontek Micro ADV (V3, oriented upward) were deployed in the lagoon to continuously sample the near-bed flow velocities at 32 Hz (Nortek Vectors) and 25 Hz and at 0.13 (V1), 0.13 (V2), 0.32 (V3) and 0.12 m (V4) above the sandy bottom, respectively.

[11] An eddy covariance system (denoted as Met in Figure 1a) was deployed to acquire meteorological and air-water flux measurements. The system includes an ultrasonic anemometer (Campbell CSAT3) measuring wind speed and direction, an open-path infrared gas analyzer to measure water vapor and CO_2 (Licor LI7500), a relative humidity and temperature sensor (HMP45C), shortwave and longwave net radiometers (Kipp & Zonen CNR1), and pH and oxidation-reduction-potential probes (Campbell CSIM11 and CS511, respectively), both compensated for temperature variations in the lagoon. All meteorological instruments were sampled continuously at 20 Hz, and then averaged to produce 30-min samples (Figure 4). A discussion of some of these data and the LEI lagoon heat budget is contained by McCabe *et al.* [2010]. The eddy covariance system was initially located on the beach at the end of the runway on the southern part of the island, without radiation measurements. A lagoon-based tower was constructed later just inside the southern lagoon reef flat where we relocated the meteorological package for

the remainder of the experiment. A summary of the locations and settings of the deployed instruments is listed in Table 2.

3. Wave Transformation

[12] The measured sea surface elevations were analyzed to evaluate the significant wave height, Hs , which is defined as [Young, 1999]

$$Hs = 4 \left[\int S(f) df \right]^{1/2}, \quad (1)$$

where $S(f)$ is the spectral density of the surface displacement and f is the frequency in Hz. Approximately 15-min records containing 4352 samples (wave gauges) and 3584 samples (PT sensors) in each hourly burst were split into 33 and 27 blocks, respectively, using 50% overlap to calculate the individual spectra with Hanning windows after mean removal and linear detrending. Ensemble-averaged spectra for each hourly burst were computed from averaging the spectra with 66 and 54 degrees of freedom. Figure 5 shows the variation of the spectra at site W3 for one tidal cycle during low wind conditions. Surface wave energy in the 0–0.8 Hz frequency band increases as the tide rises. At low tides, the PT sensors showed significant noise at higher frequencies, so only spectral energies up to 1 Hz were considered for the wave analysis. Wave gauge spectra also show a background noise level of $S(f)$ of approximately $10^{-6} (m^2 s)$

Table 2. Location and Instrument Settings^a

Location	Instrument	Settings	Measurements
W0	RBR TWR2050 PT sensor	SR = 4 Hz, recording for 17 min per burst, B/H	Wave elevation, tide, temperature
W1, W2, W3, W6	Hydraulics Laboratory (SIO) PT sensors	SR = 4 Hz, Cont.	Wave elevation, tide, temperature
W4, W4-1, W4-2, W5, W7 W8	Ultrasonic wave gauges	SR = 5 Hz, recording for 20–55 min per burst, B/H	Wave elevation, tide
V1, V2, V4	Nortek Vector ADV	SR = 32 Hz, Cont.	3D velocity, turbulence, backscatter
V3	Sontek micro ADV	SR = 25 Hz, Cont.	3D velocity, turbulence, backscatter
Met	Eddy flux system (Campbell Scientific), includes CSAT3 anemometer, Licor LI7500, Vaisala HMP45C, Kipp and Zonen CNR1, Hemisphere GPS heading sensor, pH, dissolved O ₂ sensors (Campbell CSIM11-L and CS511-L)	SR = 20 Hz, Cont.	u, v, w, H ₂ O, CO ₂ , gps heading, temperature, relative humidity, pressure, downward/upward long and short wave radiation, and corresponding fluxes, pH, dissolved O ₂ , water temperature

^aSR, sampling rate; B/H, one burst every hour; Cont., continuous recording. Wave measurements at W1 and W2 failed.

at low tides. Measurement uncertainty of the instruments affects the determination of the wave height, wave frequency, and wave dissipation rate at low-tide conditions. Except for the low tides, the frequency range of 0–1 Hz contained over 95% of the total wave energy density at all sites, indicating that the 1 Hz cutoff is a reasonable upper limit for the wave analysis. The lowest frequency band spectra in Figure 5, <0.04–0.05 Hz, were approximately constant except for the lowest tide in which H_s was already down in the noise and O(0.01) m. In a very recent study of wave dissipation over a fore reef and flat in Guam [Péquignot *et al.*, 2011], the authors found data at similar low frequencies (0.02–0.06 Hz) which they acknowledged could be wind sea or swell, or their subharmonics, or perhaps infragravity waves, deferring the analysis for future studies.

[13] Time series of the wind and incident wave conditions during the experiment from 3 to 24 April 2008 (year days 95–116) are shown in Figures 4a–4d. During this time, the wind speed U_{10} ranged from 1.5 to 13 m s⁻¹ and the wind direction varied from 80 to 190°. Steady and strong winds (U_{10} approximately 7–11 m s⁻¹) were predominantly from the direction θ_{wind} approximately 140–170° for days 97–101 and 107–114, and the wind was weak and variable (U_{10} approximately 2–7 m s⁻¹, θ_{wind} approximately 70–190°) for days 102–106. The wave height at site W0 (on the fore reef in the ocean) varies with a similar tendency to the

wind speed and oscillates with the tide, being larger at low tide. The PT sensor at site W0 was close to the surf zone in low tides; the oscillation is likely associated with the change of the surf zone location between high and low tides, with the location moving offshore at low tides; however, the amplitude of the oscillation is not explained by simple models of wave height proportional to the water depth to the 1/4 power: $H_s \propto h^{1/4}$. The significant wave height on the fore reef H_{s0} was in the range 1.5–3 m and 0.7–1.2 m, and the peak frequency in the spectrogram was about 0.2 Hz and 0.1 Hz for days corresponding to the strong and weak wind fields, respectively. The recorded tidal elevation is presented in Figure 4e and covers an entire spring-neap cycle. These data also show that the differences between the water elevations outside and inside the lagoon can reach up to approximately 0.6 m at low tide. The lagoon is isolated from the ocean during the spring low tides because the elevation of the reef rim is higher than the sea level in the ocean [McCabe *et al.*, 2010].

[14] For $U_{10} \leq 5$ m s⁻¹ H_{s0} is mainly due to distant swell and independent of the local wind field. For higher winds there is a clear correlation with the local wind speed which is evident in Figure 4. However, the surface wave height within the lagoon H_{si} is less correlated with the wind speed or with the oceanic significant wave height (not shown here for brevity). The wave height in the lagoon is correlated with the modulated tidal elevation as shown in Figures 4e and

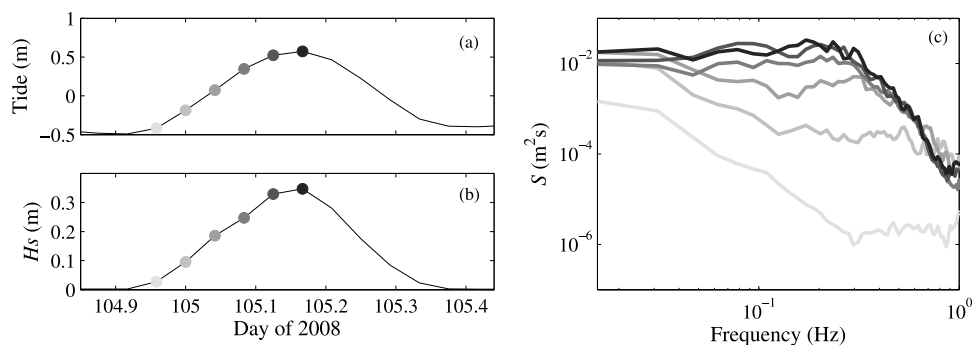


Figure 5. One-tidal-cycle variation of (a) tide (m), (b) significant wave height, H_s (m), and (c) spectra of surface elevation, S (m² s), for the half tidal cycle represented as gray lines corresponding to the marked dots in Figures 5a and 5b at site W3 for a low wind speed condition ($U_{10} = 2.2\text{--}3.7$ m s⁻¹).

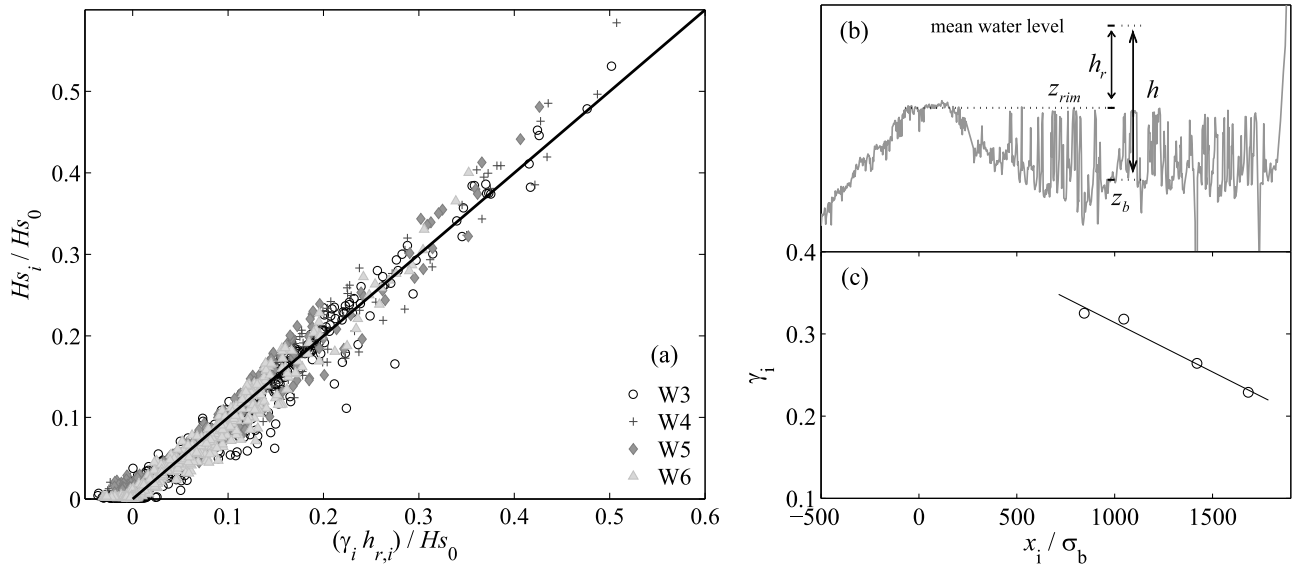


Figure 6. (a) Dimensionless wave height in the lagoon as a function of dimensionless $h_{r,i}$ (sites W3–W6), the water depth in the lagoon referenced to the rim elevation, (b) defined as $h_{r,i} = h_i - (z_{rim} - z_{b,i})$, where h_i is the local water depth in the lagoon, z_{rim} the elevation of the rim and $z_{b,i}$ the bottom elevation. The results are normalized by the attenuation parameter γ_i obtained by fitting $Hs_i/Hs_0 = \gamma_i h_{r,i}/Hs_0$ (R^2 values for W3 to W6 are 0.96, 0.96, 0.95 and 0.95, respectively). (c) Variation of the attenuation parameter γ_i as a function of the normalized distance away the reef rim. Parameter γ_i shows a decay with the distance from the reef rim as represented in the linear solid line: $\gamma_i = -0.00012x_i/\sigma_b + 0.434$.

4f. Because of the occurrence of depth-limited wave breaking on the fore reef ahead of the rim [Hardy and Young, 1996; Lowe et al., 2005b; Massel and Gourlay, 2000], the wave height on the rim and further into the lagoon, is related to the local water depth on the reef rim, h_{rim} , with a relationship $Hs < \gamma_c h_{rim}$, where γ_c is a critical breaking parameter [Thornton and Guza, 1983]. Hardy and Young [1996] suggested $\gamma_c = 0.5$. In general, the wave height in the lagoon will also be a function of the wind speed and direction in the lagoon, but the relatively short fetch in the lagoon did not lead to any significant local wind-wave generation.

[15] We computed a water depth in the lagoon referenced to the elevation of the reef rim, $h_{r,i} = h_i - (z_{rim} - z_{b,i})$ as shown in Figure 6a, where h_i is the local mean water depth in the lagoon, z_{rim} the elevation of the rim, and $z_{b,i}$ the bottom elevation (Figure 6b); subscript i denotes the instrument number in the lagoon, and $h_{r,i} = h_{rim}$ if $x = 0$ (x is the shoreward distance from the reef rim). Note that the variance of the measured mean water level was less than 2 cm, and the measured mean currents were only a few mm s^{-1} in the lagoon. As a result, the effects of the wave setup in determining $h_{r,i}$ are small in the lagoon, at least at our instrument locations. Wave setup that relates to the wave-driven currents and radiation stresses through the mean momentum equation may be greater at locations closer to the reef rim and shoreline [see Monismith, 2007]. Significant wave height as a function of $h_{r,i}$ is shown in Figure 6a. It is clear that the wave height in the lagoon increases with increasing water depth in the lagoon, similar to the results reported by Kench and Brander [2006]. Note that Hs_{rim} approximates as $\gamma_c h_{rim}$ if $Hs_0 > \gamma_c h_{rim}$ and Hs_{rim} approaches Hs_0 if $Hs_0 \leq \gamma_c h_{rim}$

[Hardy and Young, 1996]. Since the ratios of Hs_0 to h_r are larger than the critical value γ_c during the experimental period, this implies that the incident waves are filtered by the depth-limited wave breaking. When the transformed waves continually propagate into the lagoon behind the rim, Hs_{lagoon} positively correlates with h_r rather than with Hs_0 . Note that $h_r = h_{rim}$ at the reef rim ($x = 0$).

[16] An attenuation of wave height from W3 to W6 is also observed in Figure 6c. This reveals the wave dissipation in the lagoon. After normalizing with the incident wave height in the ocean Hs_0 , we find a region of linear dependence between the two parameters, Hs_i/Hs_0 and $h_{r,i}/Hs_0$. The linear dependence shows a different slope γ_i from sites W3 to W6. A linear least squares fit was used to obtain the attenuation parameter γ_i (with all R^2 larger than 0.95 for W3 to W6). Figure 6a shows the dimensionless wave height in the lagoon as a function of the dimensionless water depth. All data satisfactorily collapse into one single line when normalized by the attenuation parameter γ_i . Indeed, γ_i is related to the dissipation of wave energy in the lagoon, and it is a function of the normalized distance x_i/σ_b from the reef rim as shown in Figure 6c.

4. Wave Energy Dissipation

4.1. Observations of the Wave Energy Dissipation Rate

[17] Balancing the energy in a control volume, the total wave energy dissipation rate D can be determined from the spatial gradient of the measured wave energy flux F

$$D = -\frac{\Delta F}{\Delta x \cdot \cos \theta}, \quad (2)$$

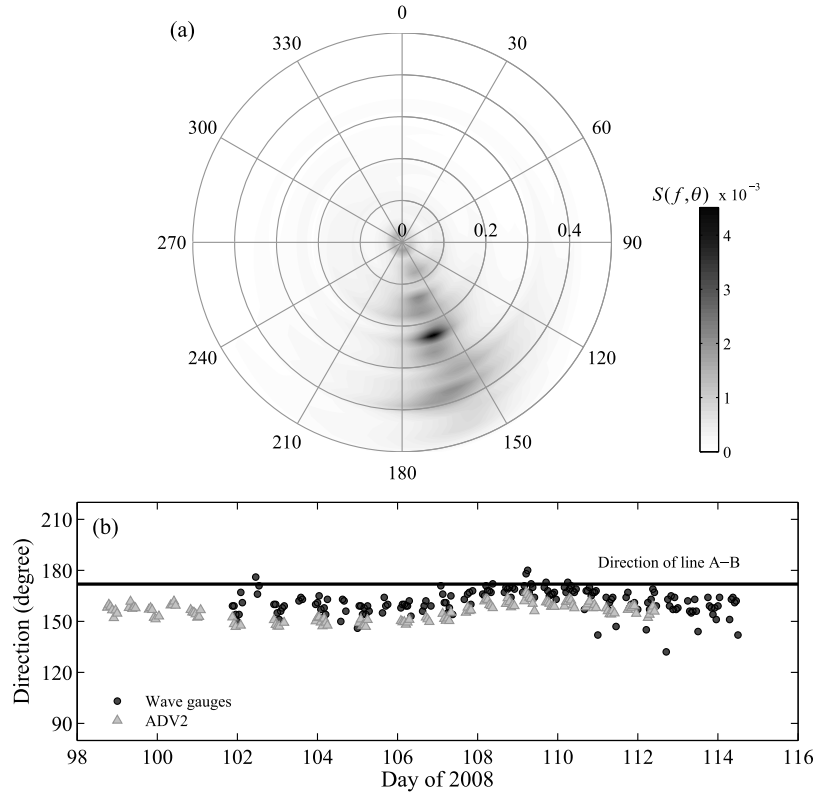


Figure 7. (a) Example of wave directional spectrum (from) obtained from the wave gauge array located at W4 (UTC time: 106.29 day of 2008). (b) Comparison between wave directions determined from the wave directional spectra (at W4) and wave-induced flow velocities from the ADV at site V2. The solid line is the direction of line A-B of in situ instrument deployment.

where Δx is the distance between two adjacent sites and θ is the angle of the wave propagation direction intersecting the line connecting the two sites. The wave energy flux is defined as

$$F = EC_g, \quad (3)$$

where E is the total wave energy density per unit area and C_g is the wave group velocity. Total wave energy density can be computed from the depth-integrated kinetic and potential energies, or can be well approximated by equipartition as twice the potential energy density. For wave spectra in the field, the total wave energy is expressed as an integral of the spectral density of the surface displacement

$$E = 2\rho g \int S(f)df, \quad (4)$$

where ρ is the water density and g is gravity. For wave spectra, the group velocity is determined using a spectral weighted group velocity [Drazen *et al.*, 2008]

$$C_g = \frac{\int C_{gn}S(f)df}{\int S(f)df}, \quad (5)$$

where C_{gn} is the characteristic group velocity of the n th component of the waves

$$C_{gn} = \frac{\partial \omega}{\partial k} \Big|_n = \frac{1}{2\omega_n} \left[\frac{\omega_n^2 + gk_n^2 h (1 - \tanh^2(k_n h))}{k_n} \right], \quad (6)$$

with ω_n the n th component of the radian frequency, k_n the n th component of the wavenumber, and h the local water depth. The total wave energy dissipation rate may be computed using equations (2)–(6).

[18] To compute the wave energy dissipation rate, the wave direction was determined from the triangular wave gauge array located at site W4. Directional spectra were computed using the WAFO toolbox [Brodtkorb *et al.*, 2000] with the maximum likelihood method [Young, 1994]. A typical example of the wave directional spectrum, with the spectral density mainly at frequencies between 0.1 and 0.4 Hz from 160 degrees is shown in Figure 7a, while the time series of the dominant (from) wave direction, defined as the peak of the energy spectrum as a function of time is drawn in Figure 7b. The wave directional spectra were calculated for periods with significant wave heights greater than 5 cm. Results show that waves inside the lagoon are nearly aligned in a rim-to-shore (approximately south-to-north) direction. The average difference between the dominant

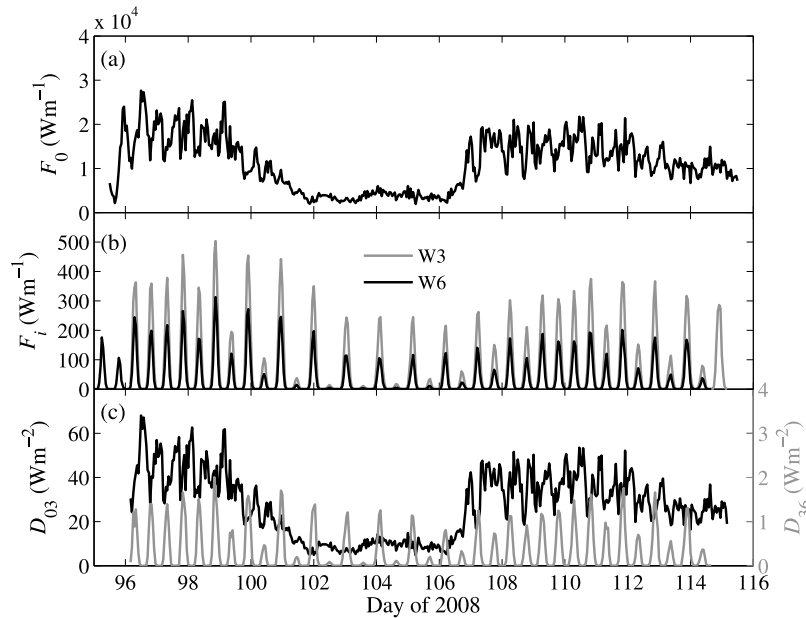


Figure 8. Time series of the measured wave energy flux F (a) in the ocean and (b) inside the lagoon. (c) Time series of wave energy dissipation rates across the fore reef (D_{03}) and inside the lagoon (D_{36}). Note the difference between right- and left-hand axes in Figure 8c.

wave direction and the line A-B is 10.8° with a standard deviation of 7° . The values verify that approximately 98% of the wave energy flux is along the direction of the array. It should be noted that the array was designed to be aligned with the dominant wavefield, but the methods employed here can be generalized for more complex wavefields and bathymetry.

[19] In addition to the wave directional spectrum observed by the wave gauge array, ADV measured velocities can also provide wave direction by calculating the principal axis of the wave-induced eastward and northward flow [Emery and Thomson, 2001]. For determining the wave directions, velocities were bandpass filtered with cutoff frequencies of 0.05 and 0.8 Hz using an FFT algorithm after mean removal and linear detrending. The principal axis of the wave-induced velocities at the site V2 is included in Figure 7b, and similar results are obtained for the other ADVs deployed in the lagoon. This confirms that wave direction in the windward lagoon is from 150 to 160° and verifies the alignment between the array of in situ instruments and the predominant wave propagation direction. Due to the limited fetch of the lagoon and the wind conditions, wind-generated waves in the lagoon make a negligible contribution to the wave energy budget.

[20] The measured energy fluxes at sites in the ocean and inside the lagoon are given in Figure 8a and 8b. Energy flux in the ocean is two orders of magnitude larger than that in the lagoon. Note the strong modulation by the tides of the wave energy flux in the lagoon. This is associated with the attenuation of the incident waves from the ocean across the fore reef correlating with the height of the tide on the fore reef. In addition, Figure 8c shows time series of the wave energy dissipation rates between two pairs of sites: between the fore reef and the rim, D_{03} ; and inside the lagoon, between

W3 and W6, D_{36} . Wave dissipation across the fore reef is an order of magnitude larger than that in the lagoon. D_{03} ranges from 10 to 60 Wm^{-2} and D_{36} varies from 0 to 3 Wm^{-2} depending on tide and wave conditions. These values are of the same order of magnitude as the dissipation across the fore reef ($8\text{--}40 \text{ Wm}^{-2}$) and on the reef flat ($0\text{--}6 \text{ Wm}^{-2}$) at Kaneohe Bay, Oahu, Hawaii [Falter *et al.*, 2004; Lowe *et al.*, 2005b].

4.2. Parameterization of Wave Energy Dissipation Rate

[21] A scatterplot of the wave energy dissipation rate across the fore reef, reef flat and the seaward edge of the lagoon, D_{03} , versus the wave height in the ocean, H_{s0} , is presented in Figure 9. Wave dissipation across this region is found to be highly correlated with the square of the wave height in the ocean. Similar quadratic dependence was observed by Lowe *et al.* [2005b] for the barrier reef in Kaneohe Bay, Hawaii. This is attributed to the fact that most of the incident wave energy in the ocean is dissipated by depth-limited wave breaking on the fore reef and reef flat. Energy flux in the ocean is proportional to the square of the wave height which is up to two orders of magnitude larger than that in the lagoon. Thus wave dissipation across the fore reef shows a quadratic dependence on the oceanic wave height.

[22] No obvious correlation of wave dissipation in the lagoon with the incident wave height is observed (not shown here for brevity). This is likely because the barrier reef rim filters out waves with wave heights higher than the maximum ratio of stable wave height to water depth on the rim through depth-limited wave breaking. For a lagoon with a rough bottom surface and less wave breaking behind the reef rim, dissipation in the lagoon may be dominated by bottom frictional effects, suggesting that the dissipation should be

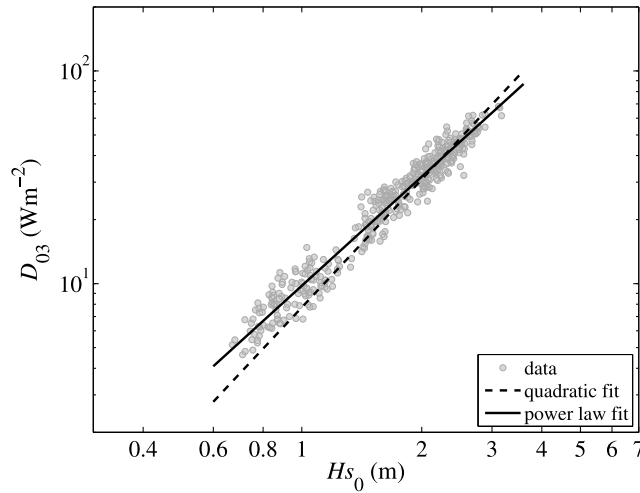


Figure 9. Wave dissipation rate across the fore reef, reef flat and the seaward edge of the lagoon (D_{03}) versus the wave height in the ocean (Hs_0). (Note that wave pressure gauge measurements at W1 and W2 failed.) The dashed line is a quadratic fit to the data, $D_{03} = \alpha Hs_0^2$, where $\alpha = 7.73$ is a dimensional fitting parameter with $R^2 = 0.92$; the solid line is the best power law fit of the data, $D_{03} = 9.77 Hs_0^{1.7055}$ with $R^2 = 0.94$.

related to the near-bed velocity. To estimate the spectral wave dissipation, we computed the RMS near-bottom velocity [Madsen, 1994; Madsen et al., 1988]

$$u_{b,r} = \sqrt{\sum_{n=1}^N u_{b,n}^2}, \quad (7)$$

with $u_{b,n}$ the near-bed horizontal orbital velocity of the n th component that is calculated using linear wave theory

$$u_{b,n} = \frac{a_n \omega_n}{\sinh k_n h}, \quad (8)$$

where ω_n is the radian frequency and a_n is the n th component of the wave amplitude as determined by $a_n = \sqrt{2S_n df}$. Average dissipation in the lagoon as a function of the average RMS near-bottom velocity predicted by linear wave theory is presented in Figure 10. The data are least squares fit to

$$D_{lagoon} = C_f \rho u_{b,r}^3, \quad (9)$$

giving $C_f = 0.0643$ and $R^2 = 0.95$. Clearly, the dissipation in the lagoon is proportional to the cube of the RMS near-bottom velocity. Rather than using a spectrally based estimate of bottom velocities, in shallow water the orbital motion is uniform with depth and scales as $a(g/h)^{1/2}$, where a is the wave amplitude and h is the local water depth. The average dissipation in the lagoon versus the scaling parameter $Hs(g/h)^{1/2}$ is least squares fit to

$$D_{lagoon} = C_w \rho Hs^3 (g/h)^{3/2}, \quad (10)$$

giving $C_w = 0.0021$, and $R^2 = 0.95$, the same correlation as obtained with the spectrally resolved bottom velocity.

4.3. Frictional Dissipation

[23] For a lagoon with a rough bottom surface and less wave breaking behind the reef rim, the wave friction factor over coral reefs can be determined by assuming that the average total wave energy loss in the lagoon D_{lagoon} is caused by the dissipation due to bottom friction $D_{f,lagoon}$ with a form [Madsen, 1994; Mathisen and Madsen, 1996]

$$D_{lagoon} \approx D_{f,lagoon} = \frac{1}{4} \rho f_{w,r} \cos \phi_r u_{b,r}^3, \quad (11)$$

where $f_{w,r}$ is a representative friction factor, $u_{b,r}$ is the average RMS near-bottom velocity in the lagoon, and $\phi_r \approx 33^\circ$ is the representative phase angle between the bottom shear stress and near-bottom horizontal orbital velocity [Mathisen and Madsen, 1999]. The wave frictional factor has been recognized as a function of the near-bed flow velocity and seabed roughness [Nielsen, 1992]. To compare the friction factor with previous empirical formulae, we need to specify the equivalent Nikuradse roughness, k_w . Manual GPS bathymetric surveys (Figures 1 and 3) showed that the bottom roughness in the lagoon can be described using a single average length scale $\sigma_b = 0.14$ m, where σ_b is the standard deviation of the measured bed elevation. Mathisen and Madsen [1999] verified that a single roughness can characterize boundary layers for both monochromatic and spectrally distributed waves. This indicates that the single length scale $\sigma_b = 0.14$ m may account for the Nikuradse roughness under the wave conditions in the lagoon. Lowe et al. [2005b] suggested that the equivalent Nikuradse roughness over a reef flat can be determined by $k_w \approx 4\sigma_b$ and

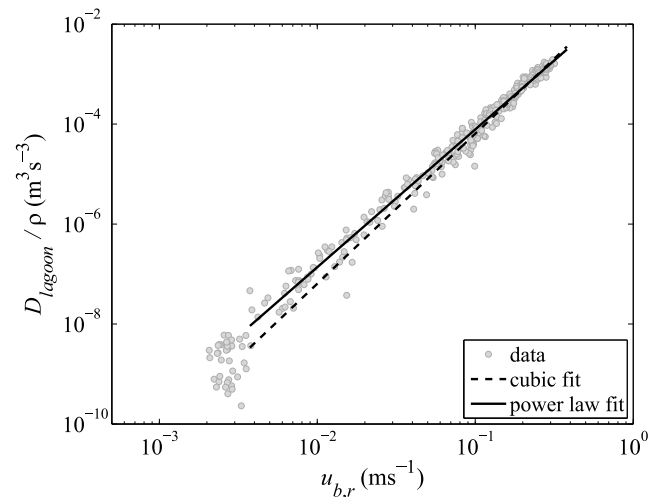


Figure 10. Average dissipation rate across the lagoon D_{lagoon} (from W3 to W6) as a function of average RMS near-bottom velocity predicted by linear wave theory $u_{b,r}$. The dashed and solid lines are a cubic fit and a best fit of the data $D_{lagoon} = 0.0643 \rho u_{b,r}^3$ and $D_{lagoon} = 0.0445 \rho u_{b,r}^{2.756}$, with $R^2 = 0.95$ and 0.96 , respectively. Only the data $D_{lagoon}/\rho > 10^{-8} \text{ m}^3 \text{ s}^{-3}$ were used for the curve fitting due to the measurement uncertainty at low tides.

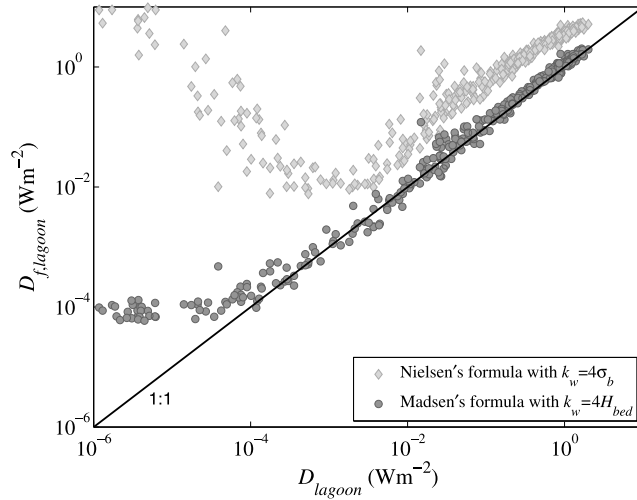


Figure 11. Scatterplot of average measured wave energy dissipation rate D_{lagoon} versus average frictional dissipation rate $D_{f, lagoon}$ estimated using Madsen's model and Nielsen's formula. The frictional dissipation rates are respectively calculated using $k_w = 4\sigma_b$ with Nielsen's formula as suggested by *Lowe et al.* [2005b] (light gray), but using $k_w = 4H_{bed}$ with Madsen's formula (dark gray). The solid line is a 1:1 ratio. Note that the scatter of the data for small values of dissipation rate is due to the measurement uncertainty at low tide.

Madsen [1994] reported $k_w \approx 4H_{bed}$ for rough rippled bed forms, where H_{bed} is the height of the bottom roughness. In our case, H_{bed} is defined as the mean height of the roughness elements. The measured value of H_{bed} is 0.272 m, which is close to $2\sigma_b$.

[24] A commonly used empirical formula to estimate the wave friction factor was proposed by *Swart* [1974] with the form

$$f_{w,r} = \exp \left[c_1 \left(\frac{u_{b,r}}{k_w \omega_r} \right)^{c_2} + c_3 \right], \quad (12)$$

where c_{1-3} are constants and ω_r is the representative (kinetic energy weighted) radian frequency

$$\omega_r = \frac{\sum_{n=1}^N \omega_n u_{b,n}^2}{\sum_{n=1}^N u_{b,n}^2}. \quad (13)$$

Nielsen [1992] suggested $c_1 = 5.5$, $c_2 = -0.2$ and $c_3 = -6.3$ for monochromatic waves; *Madsen* [1994] and *Mathisen and Madsen* [1999] extended the monochromatic friction formulas to spectral wave conditions with $c_1 = 7.02$, $c_2 = -0.078$ and $c_3 = -8.82$. The analyzed friction factor, $f_{w,r}$, not shown here, agrees much better using equation (12) with Madsen's coefficients rather than Nielsen's. In fact, Nielsen's coefficients were obtained by fitting laboratory data measured in monochromatic turbulent flows for $0.5 - 1 < u_b/(k_w \omega_r)$. However, Madsen's coefficients can be applied to a range of $0.2 < u_{b,r}/(k_w \omega_r) < 10^2$ for wave

spectra, which is much closer to the rougher bottom and spectral wave conditions in the LEI lagoon.

[25] The energy dissipation rate due to bottom friction for waves in the presence of currents was generalized by *Madsen et al.* [1988] and *Madsen* [1994] by defining representative parameters using weighted averages of discrete components. From our ADV current measurements, the effects induced by mean currents are negligible compared to wave orbital velocities inside the lagoon. Accordingly, the model for spectral waves in the absence of a current was used to estimate the frictional dissipation rate [*Mathisen and Madsen*, 1999]

$$D_{f,n} = \frac{1}{4} \rho f_{e,n} u_{b,r} u_{b,n}^2, \quad (14)$$

where $f_{e,n}$ is an energy dissipation factor defined as $f_{e,n} = \sqrt{f_{w,r}} \sqrt{f_{w,n}} \cos \varphi_n$, where $f_{w,n}$ is the friction factor of the n th wave component and φ_n is the phase angle between the bottom shear stress and near-bottom horizontal orbital velocity for the n th component. The friction factor $f_{w,n}$ is calculated using equation (12) with a replacement of ω_n by ω_r . The total frictional dissipation is determined by summing the dissipation calculated for each frequency component. *Lowe et al.* [2005b] applied *Madsen's* [1994] spectral wave model with $k_w = 4\sigma_b$ to estimate the frictional dissipation. They used Nielsen's formula to determine the friction factor instead of Madsen's. Although their results showed a consistency between the estimated and measured values, it may be questionable to apply Nielsen's formula to estimate the frictional dissipation over a rougher lagoon system. Figure 11 shows a scatterplot between the averaged measured wave energy dissipation rate in the lagoon and the frictional dissipation rates estimated by equation (14). The frictional dissipation accounts for the energy dissipation induced by flow passing over the rough bottom surface, and thus will test whether the bottom-shear-generated turbulent dissipation is a dominant mechanism for wave dissipation in the lagoon. The frictional dissipation in the lagoon determined by *Madsen's* [1994] spectral wave model was estimated using $k_w = 4\sigma_b$ and Nielsen's formula as suggested by *Lowe et al.* [2005b] using $k_w = 4H_{bed}$, and Madsen's formula as originally given by *Madsen* [1994] and *Mathisen and Madsen* [1999]. The frictional dissipation rate estimated using $k_w = 4H_{bed}$ and Madsen's formula is typically found to approximate the total energy dissipation in the lagoon. The results do not agree with *Lowe et al.* [2005b] but show that Madsen's frictional factor formula and $k_w = 4H_{bed}$ can be applied to a rougher lagoon system. Because Madsen's model was developed based on modeling the turbulent boundary layer flow with an eddy viscosity assumption over the rough bed, the agreement between the measurements also indicates that the dissipation in the lagoon is dominated by the rough turbulent flow generated by the bottom friction.

4.4. Dissipation due to Drag Within the Coral Canopy

[26] When the bottom roughness elements, the coral heads, are comparable in height to the depth of the water, it is appropriate to consider the dissipation caused by the form drag of these "canopy" elements. The roughness elements are simulated as cylindrical columns, and the resistance

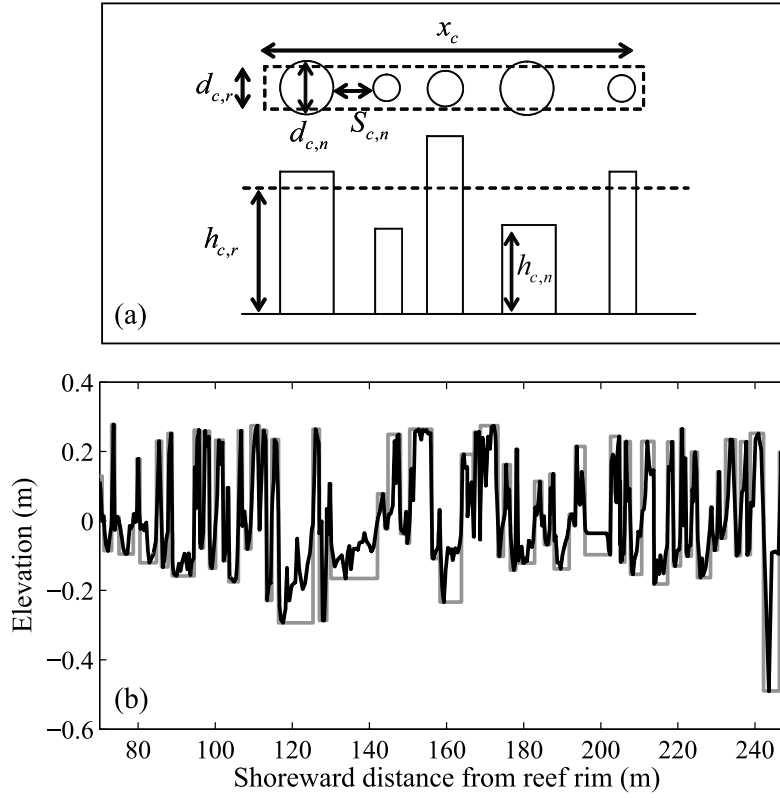


Figure 12. (a) Conceptual schematic of the irregular cylinder geometries. (b) Manual bathymetric survey of line A-B (gray line) and simulation of canopy elements using 1D distribution of cylindrical columns (black line).

force f_d for flow-passing cylindrical columns [Lowe *et al.*, 2005a] is written as

$$f_d = \frac{C_d \lambda_f}{2h_c(1 - \lambda_p)} \tilde{U}^2, \quad (15) \quad \text{with}$$

where C_d is an empirical drag coefficient $O(1)$ due to the spatially averaged in-canopy flow $\tilde{U}$ [Coccal and Belcher, 2004], where h_c is the canopy element height, and λ_f and λ_p are parameters defined as [Britter and Hanna, 2003]

$$\begin{aligned} \lambda_f &= A_f / A_T, \\ \lambda_p &= A_p / A_T. \end{aligned} \quad (16)$$

Here A_f is the frontal area of canopy elements, A_p is the plan area of canopy elements and A_T is the underlying surface area of canopies, i.e., A_T is the total area that the elements occupy divided by the number of canopy elements. Consider N irregular individual canopy elements aligned in one direction in the lagoon, representative parameters of A_f , A_p and A_T are written as

$$A_f = \frac{1}{N} \sum_{n=1}^N A_{f,n}, \quad (17)$$

$$A_p = \frac{1}{N} \sum_{n=1}^N A_{p,n}, \quad (18)$$

$$A_T = \frac{(d_{c,r} + S_{c,r})x_c}{N}, \quad (19)$$

$$d_{c,r} = \frac{1}{N} \sum_{n=1}^N d_{c,n}, \quad (20)$$

$$S_{c,r} = \frac{1}{N} \sum_{n=1}^N S_{c,n}, \quad (21)$$

where $A_{f,n} = h_{c,n}d_{c,n}$, $A_{p,n} = (\pi d_{c,n}^2/4)$, $d_{c,n}$ is the diameter of the n th element, and $S_{c,n}$ the spacing between the two adjacent elements. A sketch of the definition of the irregularly distributed cylinder geometries is given in Figure 12a. The manual bathymetric survey of the line A-B was simulated as a 1D distribution of cylindrical columns with a calculated $\lambda_f = 0.032$ as shown in Figure 12b. Based on the dissipation rate for monochromatic waves $D_d = \rho h_c(1 - \lambda_p)f_d\tilde{U}$, the dissipation rate induced by the drag force for a wave spectrum can be estimated by [Lowe *et al.*, 2007]

$$D_d \approx \frac{1}{4} \rho C_d \lambda_f U_{rms}^3. \quad (22)$$

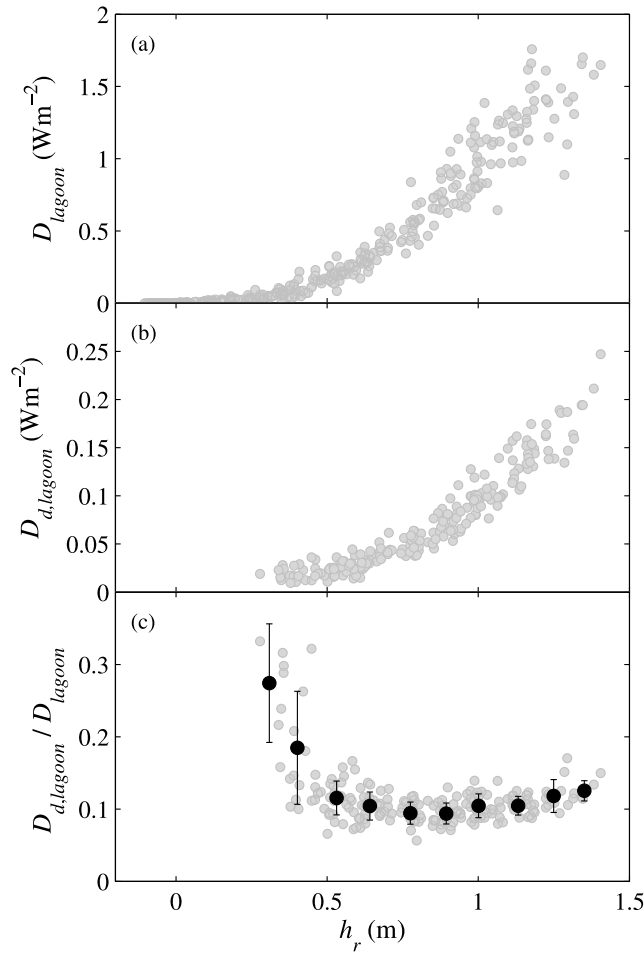


Figure 13. (a) Average wave energy dissipation D_{lagoon} , (b) average dissipation rate induced by drag force $D_{d,lagoon}$, and (c) ratio of $D_{d,lagoon}$ to D_{lagoon} as a function of water depth over the rim h_r . The large black dots represent the bin-averaged data. Note that lack of data for $D_{d,lagoon}$ at low tide is due to high noise level for ADV measurements at low tide.

The magnitude of the dissipation induced by the drag force depends on the coefficient $C_d/4$. Unfortunately, no empirical formula is available to determine the drag coefficient for oscillatory flow over canopy structures. The drag coefficient C_d is found to be $O(1)$ and was set at 2.5 in equation (22) [Coccal and Belcher, 2004; Lowe *et al.*, 2005a].

[27] Figure 13 shows the ratio of the average dissipation induced by the canopy drag force $D_{d,lagoon}$ to the wave dissipation rate in the lagoon D_{lagoon} as a function of the water depth over the reef rim, h_r , displayed with $D_{d,lagoon}$ and D_{lagoon} . The canopy drag dissipation is typically much smaller than the observed wave energy dissipation for mid to high tides. This indicates that most of the dissipation occurred in the region from the top of the canopy to the water surface for mid to high tides. The result agrees with previous field observations reported by Falter *et al.* [2004] and Lowe *et al.* [2005b], but differs from those reported by Lowe *et al.* [2007, 2008]. The discrepancy may be explained by the fact that the parameter λ_r , which depends on the distribution of coral structures, is quite different between

the two sets of observations. Here λ_r is 0.032, which is one order of magnitude smaller than the value of 0.32 used by Lowe *et al.* [2007, 2008]. This implies that the distribution and density of coral elements is crucial for determining the contribution of canopy drag dissipation to total energy dissipation, and therefore should play an important role in turbulent mixing and transport of nutrient and larvae in coral reefs. In the analysis, we found that $D_{d,lagoon}/D_{lagoon}$ is dependent on the water depth at the reef rim, h_r . As might be expected, dissipation induced by the canopy drag force becomes more significant when the water depth decreases.

5. Turbulence Dissipation

5.1. Observation of Turbulence Dissipation Rate

[28] Three components of near-bed flow velocities inside the lagoon were measured by three Nortek Vectors and one Sontek ADV. High-quality data were collected in the mid- to high-tidal levels while the data at low tide were inadequate for further analysis because the noise levels were comparable to observed flow velocities. This is because the lagoon is isolated or nearly isolated from the ocean at low tide, so that wave-driven flows and tidal currents are too weak to suspend particles adequate for reflecting acoustic backscatter signals. Velocity data were quality controlled based on correlation coefficients and signal-to-noise ratios, but even so, some despiking of the data was required with outliers detected and interpolated using the method proposed by Goring and Nikora [2002]. Despiked velocities were then transformed into a cross-shore, alongshore and vertical coordinate system.

[29] Acoustic Doppler velocity records of approximately 15-min duration each hour containing 28672 (Nortek Vector ADVs) and 22528 (Sontek ADV) samples were split into 27 and 21 segments, respectively, using 50% overlap. Individual spectra of each segment were calculated using an FFT of 2048 samples with a Hanning window after mean removal and linear detrending. Spectra were then averaged to give 54 and 42 degrees of freedom, respectively. Typical spectra of the three velocity components ($S_{uu}(f)$, $S_{vv}(f)$, $S_{ww}(f)$ for the cross-shore, alongshore and vertical components, respectively, measured by the Sontek ADV (V3) at high tide are shown in Figure 14. Note that the noise floor increases for S_{uu} and S_{vv} at frequencies in the range 6–12.5 Hz when the tidal elevation decreases. The three components of the spectra are essentially equal in the range 0.7–6 Hz, above the surface-wave band, with a slope of $-5/3$ indicating the presence of an inertial subrange. In the surface-wave band, the energy of the vertical velocity is smaller than that of the horizontal cross-shore and alongshore velocities with $S_{uu}(f)$ and $S_{vv}(f) \propto f^{-3}$, which is similar to shallow water wave data in the surf zone [Thornton, 1979]. Note that the cross-shore and alongshore velocity components of the Nortek Vector ADVs were affected by measurement noise as discussed in section 4.3.

[30] The total flow kinetic energy density in the lagoon is computed from the measured ADV velocities

$$K = \frac{\rho}{2} \overline{(u^2 + v^2 + w^2)}, \quad (23)$$

where the overbar denotes a time average over a 15-min burst of data every hour. The temporal and spatial variation

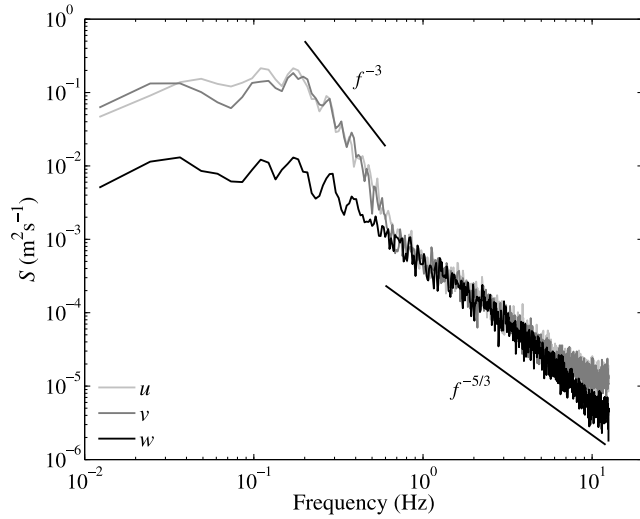


Figure 14. Typical velocity spectra of cross-shore (u), alongshore (v) and vertical (w) velocities measured by the Sontek ADV (V3) at a high tidal condition (UTC time: 98.875 day of 2008, $U_{10} = 8.5 \text{ m s}^{-1}$, H_s approximately 0.38 m, $h_r = 1.4 \text{ m}$, $\sigma_u = 0.204 \text{ m s}^{-1}$, $\sigma_v = 0.195 \text{ m s}^{-1}$, $\sigma_w = 0.065 \text{ m s}^{-1}$).

of the total flow kinetic energy, with the variation of the total wave energy density (assuming equipartition) is illustrated in Figure 15c. Both the flow kinetic energy and the wave energy density are strongly modulated by the tide, increasing and decreasing with the rise and fall of lagoon water level,

and decreasing shoreward due to energy dissipation caused by the rough bottom surface. Can the wave dissipation rate be related to the viscous dissipation of energy inferred from the inertial subrange of the turbulence?

[31] Estimates of the rate of turbulence dissipation, ϵ , were obtained by fitting the vertical velocity spectra S_{ww} to inertial-subrange turbulence spectra. The method proposed by Gerbi *et al.* [2009] was used to estimate turbulence dissipation rates (Appendix A). Calculated mean velocities ($\bar{u}$, $\bar{v}$, $\bar{w}$) and standard deviations of the wave-induced velocities (σ_u , σ_v , σ_w) show that the wave orbital motions are much greater than mean currents in the lagoon, which are very small, almost zero. The observed wave parameters and flow velocities were used to evaluate the turbulence dissipation rates in the inertial subrange.

[32] The observed tidally phase-averaged time series record of the TKE dissipation rates $\tilde{\epsilon}_i$, for the Sontek ADV and the three Nortek Vector ADVs, are presented in Figure 16 along with the phase averaged tides for the entire experiment. The dissipation rates vary with a trend similar to the time series of wave orbital velocity and are modulated by the tide. Comparing dissipation rates of the four ADVs at the same phases reveals that ϵ tends to decrease shoreward from the reef rim to the beach at high tide, consistent with the distribution of the wave dissipation rate. Under a comparable flow velocity magnitude ($<0.2\text{--}0.3 \text{ m s}^{-1}$) and significant wave height ($<0.6\text{--}1 \text{ m}$), the observed ϵ in the lagoon is $O(10^{-4} - 10^{-3}) \text{ m}^2 \text{ s}^{-3}$, which is much larger than reported values over a sandy nearshore bottom: $O(10^{-6} - 10^{-4}) \text{ m}^2 \text{ s}^{-3}$ [e.g., Feddersen *et al.*, 2007; Jones and Monismith, 2008]. These values are also larger than recently observed values of ϵ of

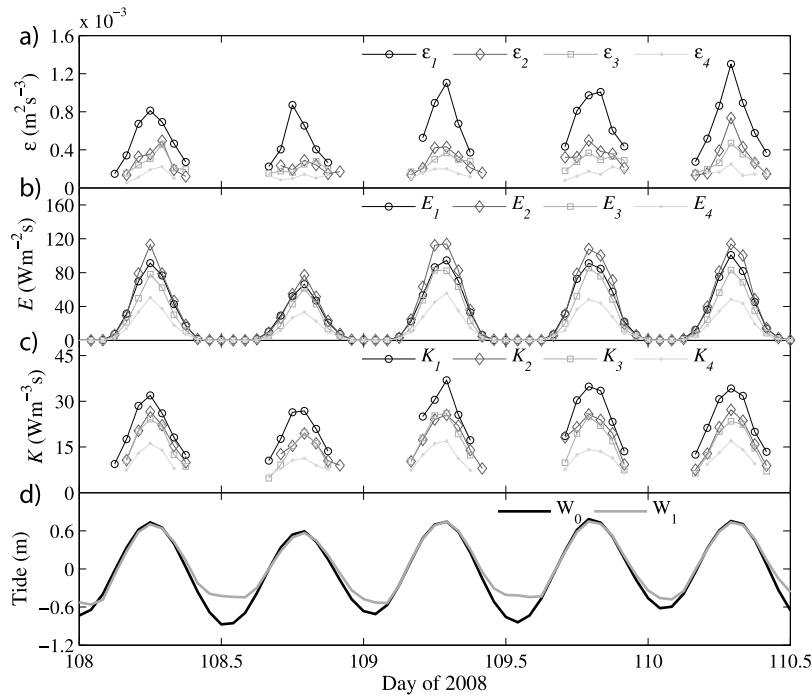


Figure 15. (a) Observed turbulent dissipation rate ϵ , (b) total wave energy E and (c) flow kinetic energy K at four locations V1 (Nortek Vector ADV), V2 (Nortek Vector ADV), V3 (Sontek Micro ADV), and V4 (Nortek Vector ADV). (d) Tidal elevation in the ocean (black, W_0) and in the lagoon (gray, W_1). Note that the data of the ADVs at low tide were inadequate for analysis because the noise level was comparable to the observed flow velocity.

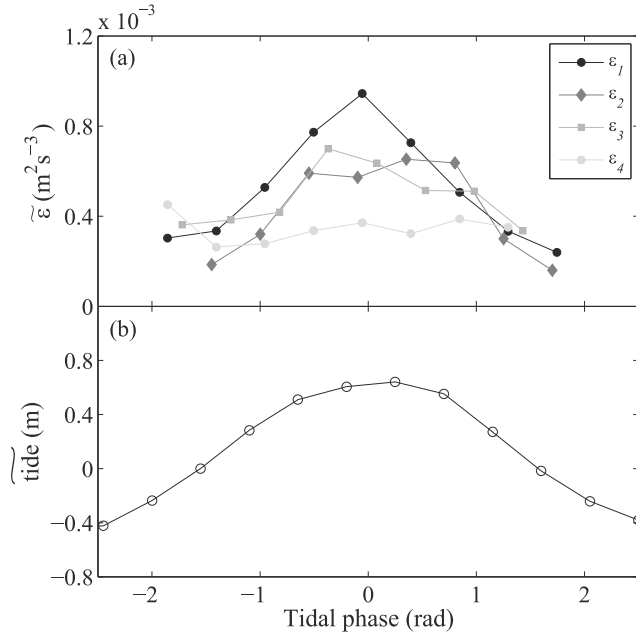


Figure 16. (a) Tidally phase averaged dissipation $\tilde{\epsilon}$ at four locations V1 (Nortek Vector ADV), V2 (Nortek Vector ADV), V3 (Sontek Micro ADV), and V4 (Nortek Vector ADV) computed using the entire experiment record. (b) The corresponding tidally phase-averaged water level is also shown.

$O(10^{-7} - 10^{-4}) \text{ m}^2 \text{s}^{-3}$ in boundary layer turbulence for unidirectional flow over a fringing coral reef in the Red Sea [Reidenbach et al., 2006a], but comparable to those of approximately $2 \times 10^{-4} \text{ m}^2 \text{s}^{-3}$ found by Baird et al. [2004] on the Warraber Island reef flat. Under comparable flow-velocity magnitudes (approximately 0.2 m s^{-1}) and bottom roughnesses (approximately 20 cm) for observations in the Red Sea [Reidenbach et al., 2006a] and in the LEI lagoon, the much higher values of ϵ in the lagoon suggests that turbulent mixing and dissipation is enhanced for oscillatory flows over a shallow wave-dominated lagoon environment.

5.2. Contribution of Turbulence Dissipation Rate to Total Energy Loss

[33] The measured ϵ provides us an alternative way to directly estimate the rate of dissipation induced by the rough bottom surface of coral reefs. The direct evaluation of the total dissipation rate by molecular viscosity would require time-resolved measurements of $\epsilon(z, t)$ profiles throughout the water column, including the rough turbulent boundary layer. Here we assume that the depth-averaged dissipation may be approximated by the measured ϵ in the flow. This is supported by nearshore observations of vertical profiles of ϵ over sandy bottoms being nearly uniform or the same order of magnitude throughout most of the water column [e.g., Feddersen et al., 2007; Jones and Monismith, 2008]. Thus we define

$$\epsilon^* = \rho \epsilon h. \quad (24)$$

Because of the local inhomogeneity of the rough coral heads, we average over the larger horizontal scale of the

wave-gauge and ADV measurements in the lagoon: Sites 3–6 along the line AB. Figure 17 shows a scatterplot of the spatially averaged wave energy dissipation measured by the wave gauges versus the rate of spatially averaged dissipation estimated using equation (24). The results show that D_{lagoon} and $\epsilon_{\text{lagoon}}^*$ are approximately proportional to one another with $D_{\text{lagoon}} \approx (1.46 \pm 0.74) \epsilon_{\text{lagoon}}^*$.

[34] An alternative to assuming the TKE dissipation rate is approximately constant with depth, an alternative is to assume a constant stress layer and the turbulence production being balanced by local dissipation. This classical bottom boundary layer scaling [Feddersen et al., 2007; Fredsøe and Deigaard, 1992; Grant and Madsen, 1986; Reidenbach et al., 2006a] for the turbulence dissipation rate is

$$\epsilon = \frac{u_*^3}{\kappa z}, \quad (25)$$

where u_* is the friction velocity, κ is von Kármán's constant, and z is the height above the (sandy) bottom. Coupling this friction velocity, described by the measured turbulence dissipation with boundary layer scaling, the average rate of energy dissipation may be expressed as

$$\epsilon_{BL} = \tau_b u_b \approx \rho u_*^2 \left(\frac{3\pi}{8} u_{b,r} \right) = \frac{3\pi}{8} \rho (\epsilon \kappa z)^{2/3} u_{b,r}, \quad (26)$$

where $\tau_b = \rho u_*^2$ is the bed shear stress [Grant and Madsen, 1986; Huang et al., 2010] and $u_b = (3\pi/8) u_{b,r}$ is defined for wave spectra [Madsen et al., 1988]. The computation of $u_{b,r}$ at the ADV sites in the lagoon uses equations (7) and (8) with spatially interpolated tidal elevation and spectral densities of the surface displacements from the wave measurements.

[35] The spatially averaged dissipation rate observed from the four ADVs is scaled with the boundary scaling using equation (26), and then compared to the observed wave

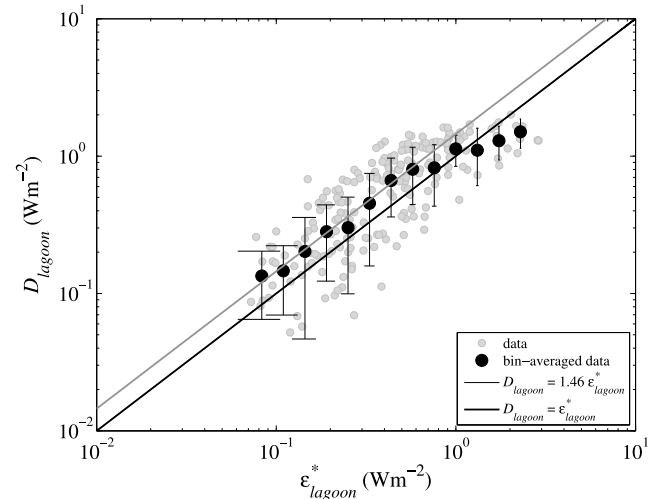


Figure 17. Scatterplot of the average wave energy dissipation rate D_{lagoon} versus the average dissipation rate estimated from the turbulent dissipation using depth-integrated approach, $\epsilon_{\text{lagoon}}^* = \langle \rho \epsilon h \rangle_{\text{lagoon}}$, where the angle brackets are a spatial averaging. The coefficient 1.46 is the mean value of $D_{\text{lagoon}}/\epsilon_{\text{lagoon}}^*$ with a standard deviation value of 0.74.

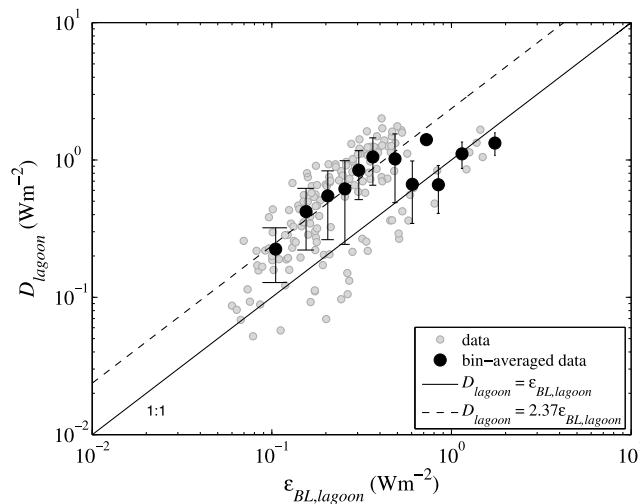


Figure 18. Scatterplot of the average wave energy dissipation rate D_{lagoon} versus the average dissipation rate estimated from the turbulent dissipation using law-of-the wall scaling $\epsilon_{BL,lagoon}$. The coefficient 2.37 is the mean value of $D_{lagoon}/\epsilon_{BL,lagoon}$ with a standard deviation of 1.2.

energy dissipation rate D_{lagoon} in Figure 18. The results indicate that in this case the boundary layer scaling is applicable to the oscillatory wave-driven lagoon system with $D_{lagoon} = (2.37 \pm 1.2)\epsilon_{BL,lagoon}$. The boundary layer scaling agrees with the eddy viscosity assumption that was used to develop a model [e.g., Madsen, 1994; Mathisen and Madsen, 1999] to account for frictional dissipation discussed in Section 4.3. It is found that the dissipation rates estimated from the two independent methods are approximately proportional to one another with $\epsilon_{lagoon}^* \approx 1.62\epsilon_{BL,lagoon}$.

6. Summary and Conclusions

[36] We have presented field measurements of wave and TKE dissipation rates over a shallow coral reef lagoon with large bottom roughness. In situ measurements of waves, currents and turbulence were achieved using wave gauge and ADV techniques, while the topography and bathymetry of the reef and lagoon was measured both manually and with airborne lidar. With these independent techniques, we quantified the dissipation of surface wave energy and the dissipation of TKE resulting from the wave orbital motion over the coral reef and lagoon.

[37] Results show that the wind-generated waves in the open ocean break on the fore reef and reef flat, with the resulting wave height in the lagoon being a function of the depth of the water over the reef rim, h_r , and a parameter γ that accounts for wave attenuation across the lagoon and decays with the distance from the reef rim.

[38] The measured wave dissipation rate is $10\text{--}60 \text{ Wm}^{-2}$ across the fore reef and reef flat, and $0\text{--}3 \text{ Wm}^{-2}$ in the lagoon. The wave energy dissipation rate in the lagoon correlates strongly with the cube of the RMS of the near-bottom velocity, $u_{b,r}$. This suggests that wave dissipation in the lagoon is strongly correlated with the frictional stress induced by the rough bottom boundary. The friction coefficient C_f of the cubic dependence between D_{lagoon} and $u_{b,r}$

is slightly larger than previous estimates made at other coral reefs sites. We showed that the frictional dissipation rate in the LEI windward lagoon can be well described with a single bottom roughness length scale σ_b , and that the spectral wave frictional model proposed by Madsen *et al.* [1988] appears to adequately estimate the frictional dissipation rate when the equivalent Nikuradse roughness k_w is described as $4H_{bed}$ (approximately $8\sigma_b$).

[39] The dissipation induced by the drag forces in the lagoon was also estimated by simulating the bottom roughness as cylindrical column elements, as in canopy-flow models. We have shown that the in-canopy drag dissipation is much smaller than the observed wave energy dissipation during mid to high tides. This in turn indicates that the dissipation in the region from the top of the canopy (i.e., the coral heads) to the water surface contributes most of the wave energy dissipation during those times. In addition, we found that the ratio of in-canopy drag dissipation to total energy dissipation rate increases when the water depth in the lagoon decreases. Dissipation induced by in-canopy drag force is enhanced and becomes more significant for the total wave energy dissipation, accounting for up to 30% when the water depth decreases to a level comparable to the bottom roughness.

[40] Estimates of the dissipation of turbulent kinetic energy ϵ were obtained using the method proposed by Gerbi *et al.* [2009] by fitting the vertical velocity spectra to the inertial-subrange turbulence spectra. Under comparable flow velocities, the observed ϵ in the lagoon is much larger than that reported in the literature for sandy beaches and, in particular, is larger than that for unidirectional flows over a fringing coral reef as recently measured by Reidenbach *et al.* [2006a]. This suggests that the turbulent mixing is enhanced for oscillatory flows over a shallow wave-dominated lagoon environment of large bottom roughness.

[41] The direct evaluation of dissipation rates in shear flows relies on the detailed measurements of profiles of turbulence dissipation rates over the entire water column including the bottom boundary layer. This was not done here, nevertheless, we used both a bottom boundary layer scaling and a depth-independent approach to estimate the average rate of energy dissipation from the observed TKE dissipation at fixed distances of $0.12\text{--}0.32 \text{ m}$ from the sandy bottom of the lagoon. It is found that both methods give average rates of turbulence dissipation that are proportional to the average surface-wave energy dissipation rate, and comparable to the wave-dissipation rate within factors of 2.37 and 1.42, respectively. This and the low current measurements in the lagoon, suggest that the wave energy dissipated in the lagoon is transformed to turbulent kinetic energy and dissipated, with little contributing to currents.

[42] Our estimates of the rate of energy dissipation due to turbulence were restricted to the point velocity measurements using ADVs. For future studies, other velocity measurement techniques such as a DopBeam [Veron and Melville, 1999] that can measure profiles of velocities and turbulence dissipation rates may provide a better understanding of the vertical and horizontal structures of turbulence properties over the coral reefs. This conclusion is compatible with the same conclusion reached by Feddersen [2011], suggesting the need for more measurements of turbulence profiles in the surf zone of sandy beaches, a region which has been much

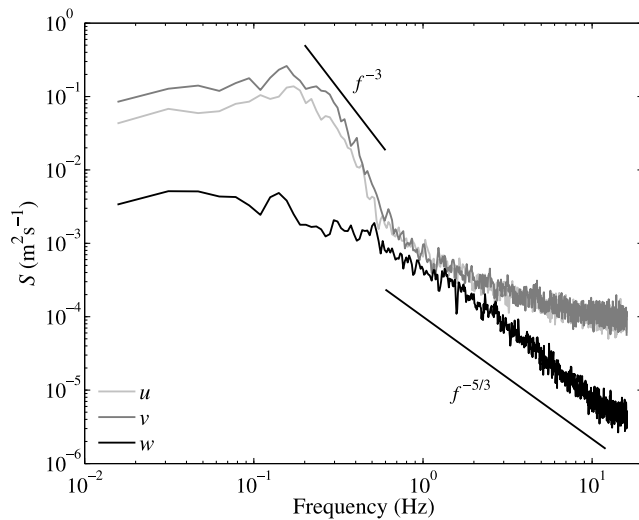


Figure A1. Typical velocity spectra of cross-shore (u), alongshore (v) and vertical (w) velocities measured by the Nortek ADV (V2) at a high tidal condition (UTC time: 98.875 day of 2008, $U_{10} = 8.5 \text{ m s}^{-1}$, $H_s = 0.47 \text{ m}$, $h_r = 1.4 \text{ m}$, $\sigma_u = 0.225 \text{ m s}^{-1}$, $\sigma_v = 0.168 \text{ m s}^{-1}$, $\sigma_w = 0.047 \text{ m s}^{-1}$).

more frequently studied than coral reef lagoons. In addition, quantifying the bottom roughness in lagoons relies on the availability of more detailed topographic and bathymetric surveys of the coral. Here visible and acoustic imaging techniques may be useful to determine the morphology of the corals and higher resolution measurements of the bottom roughness. This will in turn promote the development of three dimensional numerical models that couple the wave-field, turbulence, and mixing in reef-lagoon systems.

Appendix A: Estimation of Turbulence Dissipation Rate

[43] Figure A1 illustrates typical velocity spectra for the Nortek Vector ADV located at V1 at high tide. For frequencies above the surface-wave band, the vertical component collapses to a $-5/3$ slope; however, the cross-shore and alongshore components are affected by measurement noise, and the inertial subrange of those two components could not be identified for any of the Nortek Vector instruments. Similar measurement noise levels in the higher frequency parts of the cross-shore and alongshore velocity spectra were also observed by Gerbi *et al.* [2009]. Following Gerbi *et al.* [2009], high-frequency tails (higher than approximately 0.8 Hz) are fitted to the spectra. These tails were constructed in the inertial range of S_{ww} using the assumption of isotropic turbulence ($S_{ww} = S_{vv} = (4/3)S_{uu}$) [Tennekes and Lumley, 1972].

[44] The turbulence dissipation rate in unsteady advection for multidirectional waves was estimated using the method reported by Gerbi *et al.* [2009]

$$S_{ww}(\omega) = \frac{\alpha \varepsilon^{2/3}}{2(2\pi)^{3/2}} M_{ww}(\omega), \quad (\text{A1})$$

where $\alpha = 1.5$ is the Kolmogorov constant and M_{ww} is an integral over three-dimensional wavenumber space that depends on the mean current and waves. Indeed, M_{ww} is a function of the standard deviations of the wave velocity and the mean velocity, i.e., $M_{ww} = M_{ww}(\sigma_u, \sigma_v, \sigma_w, \bar{u}, \bar{v})$. More details of the derivations of the model are given by Feddersen *et al.* [2007] and Gerbi *et al.* [2009]. In addition, the method proposed by Bryan *et al.* [2003] was also used to estimate the turbulence dissipation rate. Good agreement between the results calculated from the two methods is obtained.

[45] **Acknowledgments.** We thank the Great Barrier Reef Marine Park Authority for granting permits for this research. The support and cooperation of Peter Gash, Wayne Fox, and the staff of both Seair and Lady Elliot Island helped make this project successful. Greg Nippard (UNSW) provided assistance in deploying and maintaining the eddy flux tower in the lagoon. We thank Ryan Bubinski for preliminary data analysis. This research was supported under the Australian Research Council's Discovery Projects funding scheme; project DP0771055 to J. H. Middleton and W. K. Melville, and from internal SIO funding to W. K. Melville. ZCH was partly supported by a fellowship from the Graduate Student Study Abroad Program (NSC97-2917-I-006-119) of the National Science Council of Taiwan. We thank two anonymous reviewers for their helpful comments.

References

- Baird, M. E., and M. J. Atkinson (1997), Measurement and prediction of mass transfer to experimental coral reef communities, *Limnol. Oceanogr.*, 42(8), 1685–1693, doi:10.4319/lo.1997.42.8.1685.
- Baird, M. E., M. Roughan, R. W. Brander, J. H. Middleton, and G. J. Nippard (2004), Mass-transfer-limited nitrate uptake on a coral reef flat, Warraber island, Torres Strait, Australia, *Coral Reefs*, 23(3), 386–396, doi:10.1007/s00338-004-0404-z.
- Battjes, J. A., and J. P. F. M. Janssen (1978), Energy loss and setup due to breaking of random waves, paper presented at 16th International Conference on Coastal Engineering, Am. Soc. of Civ. Eng., Hamburg, Germany.
- Bilger, R. W., and M. J. Atkinson (1992), Anomalous mass-transfer of phosphate on coral reef flats, *Limnol. Oceanogr.*, 37(2), 261–272, doi:10.4319/lo.1992.37.2.0261.
- Britter, R. E., and S. R. Hanna (2003), Flow and dispersion in urban areas, *Annu. Rev. Fluid Mech.*, 35, 469–496, doi:10.1146/annurev.fluid.35.101101.161147.
- Brodtkorb, P. A., P. Johannesson, G. Lindgren, I. Rychlik, J. Rydén, and E. Sjö (2000), WAFO - A Matlab toolbox for analysis of random waves and loads, paper presented at 10th International Offshore and Polar Engineering Conference, Seattle, Wash.
- Bryan, K. R., K. P. Black, and R. M. Gorman (2003), Spectral estimates of dissipation rate within and near the surf zone, *J. Phys. Oceanogr.*, 33(5), 979–993, doi:10.1175/1520-0485(2003)033<0979:SEODRW>2.0.CO;2.
- Chivas, A., J. Chappell, H. Polach, B. Pillans, and P. Flood (1986), Radiocarbon evidence for the timing and rate of island development, beach-rock formation and phosphatization at Lady Elliot Island, Queensland, Australia, *Mar. Geol.*, 69(3–4), 273–287, doi:10.1016/0025-3227(86)90043-5.
- Coccal, O., and S. E. Belcher (2004), A canopy model of mean winds through urban areas, *Q. J. R. Meteorol. Soc.*, 130(599), 1349–1372, doi:10.1256/qj.03.40.
- Drazen, D. A., W. K. Melville, and L. Lenain (2008), Inertial scaling of dissipation in unsteady breaking waves, *J. Fluid Mech.*, 611, 307–332, doi:10.1017/S0022112008002826.
- Emery, W., and R. Thomson (2001), *Data Analysis Methods in Physical Oceanography*, 2nd ed., Elsevier, Amsterdam.
- Falter, J. L., M. J. Atkinson, and M. A. Merrifield (2004), Mass-transfer limitation of nutrient uptake by a wave-dominated reef flat community, *Limnol. Oceanogr.*, 49(5), 1820–1831, doi:10.4319/lo.2004.49.5.1820.
- Feddersen, F. (2011), Observations of the surfzone turbulent dissipation rate, *J. Phys. Oceanogr.*, doi:10.1175/JPO-D-11-082.1, in press.
- Feddersen, F., J. H. Trowbridge, and A. J. Williams III (2007), Vertical structure of dissipation in the nearshore, *J. Phys. Oceanogr.*, 37, 1764–1777, doi:10.1175/JPO3098.1.
- Fredsoe, J., and R. Deigaard (1992), *Mechanics of Coastal Sediment Transport*, World Sci., Singapore.
- Gerbi, G. P., J. H. Trowbridge, E. A. Terray, A. J. Plueddemann, and T. Kukulka (2009), Observations of turbulence in the ocean surface

- boundary layer: Energetics and transport, *J. Phys. Oceanogr.*, 39(5), 1077–1096, doi:10.1175/2008JPO4044.1.
- Goring, D. G., and V. I. Nikora (2002), Despiking acoustic Doppler velocimeter data, *J. Hydraul. Eng.*, 128(1), 117–126, doi:10.1061/(ASCE)0733-9429(2002)128:1(117).
- Gourlay, M. R. (1994), Wave transformation on a coral-reef, *Coastal Eng.*, 23(1–2), 17–42, doi:10.1016/0378-3839(94)90013-2.
- Gourlay, M. R. (1996a), Wave set-up on coral reefs. Part 2. Set-up on reefs with various profiles, *Coastal Eng.*, 28(1–4), 17–55, doi:10.1016/0378-3839(96)00009-9.
- Gourlay, M. R. (1996b), Wave set-up on coral reefs. Part 1. Set-up and wave-generated flow on an idealized two dimensional horizontal reef, *Coastal Eng.*, 27(3–4), 161–193, doi:10.1016/0378-3839(96)00008-7.
- Gourlay, M. R., and G. Colleter (2005), Wave-generated flow on coral reefs—An analysis for two-dimensional horizontal reef-tops with steep faces, *Coastal Eng.*, 52(4), 353–387, doi:10.1016/j.coastaleng.2004.11.007.
- Grant, W. D., and O. S. Madsen (1986), The continental-shelf bottom boundary layer, *Annu. Rev. Fluid Mech.*, 18, 265–305, doi:10.1146/annurev.fl.18.010186.001405.
- Hardy, T. A., and I. R. Young (1996), Field study of wave attenuation on an offshore coral reef, *J. Geophys. Res.*, 101(C6), 14,311–14,326, doi:10.1029/96JC00202.
- Hearn, C. J. (1999), Wave-breaking hydrodynamics within coral reef systems and the effect of changing relative sea level, *J. Geophys. Res.*, 104(C12), 30,007–30,019, doi:10.1029/1999JC900262.
- Hearn, C. J., M. J. Atkinson, and J. L. Falter (2001), A physical derivation of nutrient-uptake rates in coral reefs: Effects of roughness and waves, *Coral Reefs*, 20, 347–356, doi:10.1007/s00338-001-0185-6.
- Huang, Z. C., H. H. Hwang, S. C. Hsiao, and K. A. Chang (2010), Laboratory observation of boundary layer flow under spilling breakers in a surf zone using particle image velocimetry, *Coastal Eng.*, 57, 343–357, doi:10.1016/j.coastaleng.2009.11.004.
- Jago, O. K., P. S. Kench, and R. W. Brander (2007), Field observations of wave-driven water-level gradients across a coral reef flat, *J. Geophys. Res.*, 112, C06027, doi:10.1029/2006JC003740.
- Jones, N. L., and S. G. Monismith (2008), The influence of whitecapping waves on the vertical structure of turbulence in a shallow estuarine embayment, *J. Phys. Oceanogr.*, 38(7), 1563–1580, doi:10.1175/2007JPO3766.1.
- Kench, P. S., and R. W. Brander (2006), Wave processes on coral reef flats: Implications for reef geomorphology using Australian case studies, *J. Coastal Res.*, 221, 209–223, doi:10.2112/05A-0016.1.
- Longuet-Higgins, M. S., and R. W. Stewart (1964), Radiation stress in water waves: A physical discussion, with applications, *Deep Sea Res.*, 11, 529–562.
- Lowe, R. J., J. R. Koseff, and S. G. Monismith (2005a), Oscillatory flow through submerged canopies: 1. Velocity structure, *J. Geophys. Res.*, 110, C10016, doi:10.1029/2004JC002788.
- Lowe, R. J., J. L. Falter, M. D. Bandet, G. Pawlak, M. J. Atkinson, S. G. Monismith, and J. R. Koseff (2005b), Spectral wave dissipation over a barrier reef, *J. Geophys. Res.*, 110, C04001, doi:10.1029/2004JC002711.
- Lowe, R. J., J. L. Falter, J. R. Koseff, S. G. Monismith, and M. J. Atkinson (2007), Spectral wave flow attenuation within submerged canopies: Implications for wave energy dissipation, *J. Geophys. Res.*, 112, C05018, doi:10.1029/2006JC003605.
- Lowe, R. J., U. Shavit, J. L. Falter, J. R. Koseff, and S. G. Monismith (2008), Modeling flow in coral communities with and without waves: A synthesis of porous media and canopy flow approaches, *Limnol. Oceanogr.*, 53(6), 2668–2680, doi:10.4319/lo.2008.53.6.2668.
- Lowe, R. J., J. L. Falter, S. G. Monismith, and M. J. Atkinson (2009a), Wave-driven circulation of a coastal reef-lagoon system, *J. Phys. Oceanogr.*, 39(4), 873–893, doi:10.1175/2008JPO3958.1.
- Lowe, R. J., J. L. Falter, S. G. Monismith, and M. J. Atkinson (2009b), A numerical study of circulation in a coastal reef-lagoon system, *J. Geophys. Res.*, 114, C06022, doi:10.1029/2008JC005081.
- Lugo-Fernández, A., H. H. Roberts, and W. J. Wiseman (1998a), Tide effects on wave attenuation and wave set-up on a Caribbean coral reef, *Estuarine Coastal Shelf Sci.*, 47(4), 385–393, doi:10.1006/ecss.1998.0365.
- Lugo-Fernández, A., H. H. Roberts, and J. N. Suhayda (1998b), Wave transformations across a Caribbean fringing-barrier Coral Reef, *Cont. Shelf Res.*, 18(10), 1099–1124, doi:10.1016/S0278-4343(97)00020-4.
- Madsen, O. S. (1994), Spectral wave-current bottom boundary layer flows, paper presented at 24th International Conference on Coastal Engineering, Am. Soc. of Civ. Eng., Kobe, Japan.
- Madsen, O. S., Y. K. Poon, and H. Graber (1988), Spectral wave attenuation by bottom friction: Theory, paper presented at 21st International Conference on Coastal Engineering, Am. Soc. of Civ. Eng., Torremolinos, Spain.
- Massel, S. R., and M. R. Gourlay (2000), On the modeling of wave breaking and set-up on coral reefs, *Coastal Eng.*, 39(1), 1–27, doi:10.1016/S0378-3839(99)00052-6.
- Mathisen, P. P., and O. S. Madsen (1996), Waves and currents over a fixed rippled bed: 1. Bottom roughness experienced by waves in the presence and absence of currents, *J. Geophys. Res.*, 101(C7), 16,533–16,542, doi:10.1029/96JC00954.
- Mathisen, P. P., and O. S. Madsen (1999), Waves and currents over a fixed rippled bed: 3. Bottom and apparent roughness for spectral waves and currents, *J. Geophys. Res.*, 104(C8), 18,447–18,461, doi:10.1029/1999JC006023.
- McCabe, R. M., P. Estrade, J. H. Middleton, W. K. Melville, M. Roughan, and L. Lenain (2010), Temperature variability in a shallow, tidally isolated coral reef lagoon, *J. Geophys. Res.*, 115, C12011, doi:10.1029/2009JC006023.
- Monismith, S. G. (2007), Hydrodynamics of coral reefs, *Annu. Rev. Fluid Mech.*, 39, 37–55, doi:10.1146/annurev.fluid.38.050304.092125.
- Nelson, R. C. (1996), Hydraulic roughness of coral reef platforms, *Appl. Ocean Res.*, 18(5), 265–274, doi:10.1016/S0141-1187(97)00006-0.
- Nielsen, P. (1992), *Coastal Bottom Boundary Layers and Sediment Transport*, World Sci., Singapore.
- Péquignat, A.-C., J. M. Becker, M. A. Merrifield, and S. J. Boc (2011), The dissipation of wind wave energy across a fringing reef at Ipan, Guam, *Coral Reefs*, 30, 71–82, doi:10.1007/s00338-011-0719-5.
- Reidenbach, M. A., S. G. Monismith, J. R. Koseff, G. Yahel, and A. Genin (2006a), Boundary layer turbulence and flow structure over a fringing coral reef, *Limnol. Oceanogr.*, 51(5), 1956–1968, doi:10.4319/lo.2006.51.5.1956.
- Reidenbach, M. A., J. R. Koseff, S. G. Monismith, J. V. Steinbuck, and A. Genin (2006b), The effects of waves and morphology on mass transfer within branched reef corals, *Limnol. Oceanogr.*, 51(2), 1134–1141, doi:10.4319/lo.2006.51.2.1134.
- Reidenbach, M. A., J. R. Koseff, and M. A. R. Koehl (2009), Hydrodynamic forces on larvae affect their settlement on coral reefs in turbulent, wave-driven flow, *Limnol. Oceanogr.*, 54(1), 318–330, doi:10.4319/lo.2009.54.1.0318.
- Reineman, B. D., L. Lenain, D. Castel, and W. K. Melville (2009), A portable airborne scanning lidar system for ocean and coastal applications, *J. Atmos. Oceanic Technol.*, 26(12), 2626–2641, doi:10.1175/2009JTECHO703.1.
- Swart, D. H. (1974), *Offshore Sediment Transport and Equilibrium Beach Profiles*, Delft Hydraul. Lab., Delft, Netherlands.
- Symonds, G., K. P. Black, and I. R. Young (1995), Wave-driven flow over shallow reefs, *J. Geophys. Res.*, 100(C2), 2639–2648, doi:10.1029/94JC02736.
- Tennekes, L., and J. L. Lumley (1972), *A First Course in Turbulence*, MIT Press, Cambridge, Mass.
- Thornton, E. B. (1979), Energetics of breaking waves within the surf zone, *J. Geophys. Res.*, 84(C8), 4931–4938, doi:10.1029/JC084iC08p04931.
- Thornton, E. B., and R. T. Guza (1983), Transformation of wave height distribution, *J. Geophys. Res.*, 88, 5925–5938, doi:10.1029/JC088iC10p05925.
- Veron, F., and W. K. Melville (1999), Pulse-to-pulse coherent Doppler measurements of waves and turbulence, *J. Atmos. Oceanic Technol.*, 16(11), 1580–1597, doi:10.1175/1520-0426(1999)016<1580:PTPCDM>2.0.CO;2.
- Young, I. R. (1989), Wave transformation over coral reefs, *J. Geophys. Res.*, 94(C7), 9779–9789, doi:10.1029/JC094iC07p09779.
- Young, I. R. (1994), On the measurement of directional wave spectra, *Appl. Ocean Res.*, 16(5), 283–294, doi:10.1016/0141-1187(94)90017-5.
- Young, I. R. (1999), *Wind Generated Ocean Waves*, 288 pp., Elsevier Sci., Amsterdam.

Z.-C. Huang, Graduate Institute of Hydrological and Oceanic Sciences, National Central University, Taoyuan 32001, Taiwan.

L. Lenain, W. K. Melville, B. Reineman, and N. Statom, Scripps Institution of Oceanography, University of California, San Diego, La Jolla, CA 92093-0213, USA. (kmelville@ucsd.edu)

R. M. McCabe, School of Oceanography, University of Washington, Seattle, WA 98195, USA.

J. H. Middleton, School of Aviation, University of New South Wales, Sydney, NSW 2052, Australia.