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Key Points:

- Shoaling waves observed on a healthy coral reef
- Damping of waves is an order of magnitude greater than observed previously
- High friction factor observed attributed to canopy flow dynamics

Supporting Information:

Supporting Information S1

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Frictional wave dissipation on a remarkably rough reef

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Abstract We present a week of observations of wave dissipation on the south forereef of Palmyra Atoll. Using wave measurements made in 6.2 m and 11.2 m of water offshore of the surf zone, we computed energy fluxes and near-bottom velocity. Equating the divergence of the shoreward energy flux to its dissipation by bottom friction and parameterizating dissipation in terms of the root-mean-square velocity cubed, we find that the wave friction factor, f_{w} for this reef is 1.80 ± 0.07 , nearly an order of magnitude larger than values previously found for reefs. We attribute this remarkably high value of f_w to the complex canopy structure of the reef, which we believe may be characteristic of healthy reefs. This suggests that healthy reefs with high coral cover may provide greater coastal protection than do degraded reefs with low coral cover.

1. Introduction

Surface waves play an important role in the evolution and destruction of coral reefs [*Williams et al.*, 2013]. Waves drive mass transfer between the benthos and the water column above [*Falter et al.*, 2004]. They also force flows over and through reef systems [*Hench et al.*, 2008]. Finally, breakage of corals [*Madin and Connolly*, 2006; *Storlazzi et al.*, 2005] and the transport of rubble [*Scoffin*, 1993] are important to the long-term evolution of reefs. Likewise, coral reefs can play an important role in coastal regions serving as natural breakwaters dissipating as much as 97% of the incident wave energy on the reef offshore of the common shallow reef crest [*Ferrario et al.*, 2014].

When waves shoal on reefs, they are amplified by shoaling and reduced by frictional dissipation associated with wavy flow over the rough reef topography [*Symonds et al.*, 1995]. If the depth is sufficiently shallow, they will break [*Vetter et al.*, 2010]; it is this breaking that is often the dominant source of energy dissipation. In most wave models, frictional dissipation of wave energy is assumed to take the form

$$\langle \varepsilon \rangle = 0.6 f_w \rho U_{\rm rms}^3 \tag{1}$$

where $\langle \varepsilon \rangle$ is the average rate of wave energy dissipation, ρ is the density of seawater, f_w is the wave friction factor, and $U_{\rm rms}$ is the root-mean-square near-bottom wave velocity [*Dean and Dalrymple*, 1991]. From lab experiments [e.g., *Kamphuis*, 1975], f_w is found to depend on the relative roughness of the reef, defined in terms of the ratio of the typical wave orbital excursion, l, to a roughness length, k_r that characterizes the bottom roughness. Field measurements of waves on the reef flat in Kaneohe Bay [*Lowe et al.*, 2005] and on the forereef of Moorea [*Monismith et al.*, 2013] have found $f_w \simeq 0.2 \rightarrow 0.3$, values that suggest that k_r scales with the height of local corals and lie within the range seen in the lab by *Kamphuis* [1975].

In the observations given below, we document a reef for which $f_w \simeq 1.80 \pm 0.07$. We argue that this is an effect of the structural complexity of the reef and the fact that the effective friction factor for canopies can be larger than 1.

2. Methods

Our measurements were made on the south forereef of Palmyra Atoll (5°53'N, 162°5'W—Figures 1a and 1b), an atoll reef system that is thought to be representative of what reefs would be like under natural conditions, i.e., one having a complete compliment of herbivores [*Stevenson et al.*, 2006; *Sandin et al.*, 2008], little anthropogenic influence such as nutrient discharges or excess sedimentation, and no history of fisheries exploitation.

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Figure 1. Picture of Palmyra Atoll showing (a) location; (b) general features of the atoll and the location of FR3 (image courtesy of NOAA); (c) instrument layout, isobaths, and coordinate system definition; and (d) cross section of bathymetry at FR3.

As part of a study of flows on Palmyra Atoll involving numerical modeling and long-term instrument deployments around the atoll, between 16 and 22 July 2014, we deployed a set of instruments near FR3 on the south forereef (5°52.00'N, 162°6.81'W—Figures 1c and 1d). This site is characterized by nearly 100% coral cover and a slope of approximately 9% (Figure 1d). At 11.2 m below mean sea level, we installed a 1 MHz Nortek Acoustic Doppler Profiler (ADP) sampling at 0.5 Hz and recording in 1 m bins starting 1.5 m above the substrate, along with a Brancker Research DR1050 pressure logger recording at 0.5 Hz. At 6.2 m below mean sea level, we installed a pair of DR1050 loggers facing in opposite directions, one upslope and the other downslope, both recording at 0.5 Hz.

The ADP and pressure data were first low-pass filtered (fourth order Butterworth, cutoff frequency = 0.005 Hz) to remove tidal and other longer-period variations. The high-pass filtered data were then processed as follows: Wave heights from all sensors were computed using linear wave theory [cf. *Dean and Dalrymple*, 1991] to compute water surface elevations, η , from measured pressures. These were broken into 4096 sample records from which free surface variance spectra, $S_{\eta\eta}$ as functions of frequency, *f*, were computed for 256 point segments using a Hamming window, giving 40 equivalent degrees of freedom for each spectral estimate [*Emery and Thompson*, 2004]. The significant wave height for each time interval is

$$H_{\rm s} = 4 \left(\int_0^{f_{\rm N}} S_{\eta\eta}(f) \mathrm{d}f \right)^{1/2} \tag{2}$$

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Figure 2. Mean depth and velocity at 11.2 m station: (a) low-pass filtered depth variation and (b) low-pass cross-reef (gray) and along-reef (black) velocities.

where $f_N = 0.25$ Hz is the Nyquist frequency of the pressure sampling. Wave height spectra at the shallow station were averaged together to produce single spectra at that depth, whereas only the RBR data was used at the deeper station. The ADP velocity data at z = 4.4 mab were also broken into 4096 point (8192 s) segments and spectrally analyzed as above to determine principal wave direction as a function of f. Expressed relative to the cross-reef direction, the direction θ was computed as [Bowden and White, 1966]

$$\theta(f) = \tan^{-1} \left(\frac{\operatorname{Re} \{ C_{\eta V}(f) \}}{\operatorname{Re} \{ C_{\eta U}(f) \}} \right)$$
(3)

where $C_{\eta U}$ and $C_{\eta V}$ are the cross spectra of the cross-reef (*U*) and along-reef (*V*) velocities with the free surface elevation. Frequency-dependent wave directions at the shallow station were then computed using Snell's law [*Dean and Dalrymple*, 1991] and computed values of θ for the deep station.

Using computed directions and spectra, energy fluxes, É, normal to the reef were computed as

$$\dot{E} = \rho g \int_{0}^{t_{N}} C_{gx}(f) S_{\eta\eta}(f) \mathrm{d}f.$$
(4)

where C_{gx} is the cross-shore component of the group velocity. To compute the average rate of dissipation between the 6.2 and 11.2 m stations, the 1-D wave energy flux equation [*Dean and Dalrymple*, 1991] was discretized as

$$\frac{d}{dx}(\dot{E}) \simeq \frac{\Delta E}{\Delta x} = -\langle \varepsilon \rangle \tag{5}$$

where $\Delta \vec{E}$ is the change in energy flux and $\Delta x \simeq 56$ m.

While the present data do not permit us to directly compute reflection of wave energy by the sloping reef, past work on beaches, rubble-mound breakwaters, and reef-like geometries suggest that reflection is likely small: Given observed wave heights ($H \approx 1$ m),wave lengths ($L \approx 100$ m), and a bottom slope, $s_0 \approx 0.1$, the Irribarren number [*Dean and Dalrymple*, 1991] $I = s_0 / \sqrt{H/L} \approx 1$. From the work of *Seelig* [1983], $I \approx 1$ implies a reflection coefficient $R \approx 0.06$. Since the reflected energy flux is proportional to R^2 , reflection by the *whole*



Figure 3. Wave spectral properties at the 11.2 m station as a function of time: (a) power spectral density of water surface variations, (b) wave direction, and (c) shore-normal wave energy flux.

reef slope should account for approximately 4% of the total incident energy flux. In contrast, the observed energy flux change over a short section of the reef was approximately 40% of the flux observed at the deeper station. Hence, the effects of reflection amount at most to an overestimate of $\langle \varepsilon \rangle$, by 10% although probably much less if one accounts for dissipation over the whole forereef. Moreover, calculation of wave direction using velocity and pressure cospectra as described by *Herbers et al.* [1999] gave nearly identical values of mean wave direction as did equation (1), suggesting that reflections were not important.

Finally, to compute the friction factor, $U_{\rm rms}$ is needed. Because near-bed velocities were only measured at the 11.2 m station, and because the ADP velocities are of uncertain accuracy given beam spread with height, $U_{\rm rms}$ for both sites were computed from wave height spectra using linear wave theory as

$$U_{\rm rms} = \left[\int_{0}^{f_{\rm N}} 4\pi^2 f^2 S_{\eta\eta}(f) \frac{1}{\sinh^2(kh)} df\right]^{1/2}$$
(6)

and then averaged together.

3. Results

Mean flows at the site (Figure 2) reflect the superposition of the North Equatorial Counter Current with local tides [*Gove et al.*, 2015; J. Rogers, unpublished data, 2014]. Analysis of the coherence between the free surface elevation and depth-averaged velocities shows only a weak dependence ($r^2 = 0.1$) of flows at FR3 on tidal variations in the free surface.

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Figure 4. Wave properties for 17:00 17 July 2014 (Figures 4a–4d) and 13:20 20 July 2014 (Figures 4d, 4c, and 4f). (a and e) Wave height spectra at 11.2 m including 95% confidence intervals (shading). (b and f) Wave direction at 11.2 m—16 point average (solid line) of individual estimates (grey points). (c and g) Shore-normal energy fluxes at the 11.2 m (solid line) and 6.2 m (dashed line) isobaths and (d and h) percent loss in shore-normal energy flux between the 11.2 and 6.2 m isobaths.

During the experiment, the wave field was primarily the result of superposition of longer-period (12–20 s) waves with shorter period (7 to 8 s) ones (Figures 3a, 4a, and 4e). This is because the southwards facing forereef is exposed to both short, local trade wind-driven wind waves propagating to the northwest and to longer swell generated in the Southern Ocean propagating more nearly due north across the reef (Figures 3b, 4b, and 4f). Wave directionality is more clearly seen in Figures 4b and 4f, plots of wave direction for two specific periods (17:00 17 July 2014 and 13:20 20 July 2014): short-period motions propagate at approximately 40° to the west of the cross-shore direction. Using these wave height spectra and frequency-dependent wave direction, the spectral distributions of shore-normal energy fluxes for these two periods (Figures 4c and 4g) closely resemble the wave height spectra themselves.

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Figure 5. Wave properties on Palmyra forereef at FR3: (a) significant wave height at 11.2 m (solid) and 6.2 m (dashed); (b) average $U_{\rm rms}$ for the reef between the 11.2 and 6.2 m stations (solid) and the depth-averaged speed at the 11.2 m (dotted line); (c) shore-normal wave energy fluxes at 11.2 m (solid) and 6.2 m (dashed); and (d) dissipation computed from wave energy flux divergence (o) and from equation (1) using $f_w = 1.80$ (solid).

Significant wave heights throughout the experiment varied between 1 and 1.7 m (Figure 5a), well under the breaking heights at these depths. These waves produce near-bed orbital velocities that are larger than the mean current speeds at the site (Figure 5b), suggesting that mean flows played a small role in setting near-bed frictional dissipation of wave energy. Most importantly, there is a significant decrease in the wave energy flux between the 11.2 and 6.2 m stations (Figure 5c), giving values of $\langle \varepsilon \rangle$, that reach nearly 30 W m⁻² (Figure 5d).

Fitting $\langle \varepsilon \rangle$ as a function of $U_{\rm rms}^3$ per equation (1), we find that $f_w = 1.80 \pm 0.07$ ($r^2 = 0.83$). As seen in Figure 5d, while much larger than is typically found on reefs, i.e., $f_w \simeq 0.25$ [Monismith et al., 2013], the inferred value of f_w provides a good estimate of the measured dissipation rate.

4. Discussion

The value of f_w determined observationally for the Palmyra forereef is nearly an order of magnitude larger than values found previously for rough boundaries [Kamphuis, 1975; Mirfendereska and Young, 2003]. Indeed, many models for calculating f_w are limited to a maximum value of 0.23 [Mirfendereska and Young, 2003]. The value of f_w is also striking in that it can be shown that with this large value of f_w the rate of energy dissipation by bottom friction can be larger than the rate at which energy is dissipated by wave breaking, behavior not previously seen for reefs (see supporting information).



Figure 6. Comparison of reef topography (a) the forereef at FR3 on Palmyra near the 10 m isobaths ($f_W \simeq 4.55$); (b) the north shore Moorea forereef (near the 6 m isobath) in December 2006 ($f_W \simeq 0.3$).

Given that the value of f_w is so different from what has observed before, it is important to consider alternative explanations for the decline in shoreward energy flux. First, modeling of wave shoaling on steep reef faces shows that wave-wave interactions can be important [*Sheremet et al.*, 2011]. However, the gain in energy flux between the two stations for the infragravity wave band (T > 20 s) is only 0.3% of the total incident energy flux, suggesting that transfer of swell band energy to subharmonics by difference wave-wave interactions is not generally important on the Palmyra forereef. On the other hand, the effect of sum interactions, which transfer energy to short-period waves (T < 6 s), is less clear because of noisiness, although this band appears in general to be dissipative as well (Figures 4d and 4g). However, even if sum interactions extract energy from the swell band, this energy is also ultimately dissipated by friction. Alternatively, it is possible that the change in energy flux we observe is an effect of scattering by the rough reef topography (T. Janssen, personal communication, 2014). How this might work is not clear in that scattering theory relies on the assumption that the wave field varies on scales long compared to typical wavelengths [e.g., *Mei*, 1985]. In the present case, complete wave shoaling takes place within one wavelength of the shore (for waves near the spectral peaks), and so scattering is probably too "slow" to account for the observed change in energy flux.

An explanation for the observed high dissipation rates can be had by considering flows through canopies such as exist on coral reefs. In particular, a formulation similar to (1) can be developed by accounting for frictional drag associated with the canopy elements, e.g., coral branches [*Huang et al.*, 2012]. In this case, f_w can be written as [*Lowe et al.*, 2007]

$$f_w = \left[f_0 + C_d \lambda_f \alpha_w^3 \right] \tag{7}$$

Here f_0 is the friction factor associated with substrate at the bottom of the canopy, which is typically in the range of 0.01 to 0.1, $C_D \approx 1$ is the drag coefficient appropriate to the canopy elements, and λ_f is the ratio of canopy element frontal area to the underlying surface area. The ratio of the velocity in the canopy to the velocity just above the canopy, a_{wr} depends on the ratio of the spacing of the canopy elements to *l*. For dense canopies and longer waves, $0.5 < a_w < 0.7$ [Lowe et al., 2005]. Depending on the coral reef structure, it may be possible that $\lambda_f > > 1$, suggesting that f_w for complex reef matrices may also be larger than 1, as we have observed here.

The reef at FR3 is remarkably rugose (Figure 6a) with many features that are O(1 m) in scale and with significant geometric complexity well beyond that seen, e.g., on the Moorea forereef (Figure 6b). For example, it seems likely that multilayered structures like those of the plate-like and branching *Porites* sp. colonies seen in Figure 6a produce surface areas for drag that are much larger than their footprint, i.e., effectively making $\lambda_f > 1$. Nonetheless, it is clear that translating the geometry of real reefs like that on the Palmyra forereef into hydrodynamic parameters like f_w that might be computed from simplified

configurations such as a cylinder array or a single layer of uniformly sized sand grains remains an open problem [see *Nunes and Pawlak*, 2008].

The high dissipation rate of wave energy has two implications for how the reef functions. First, as discussed by *Falter et al.* [2004], mass transfer between the reef and the overlying water column is proportional to $\langle \varepsilon \rangle^{1/3}$. Thus, larger values of $\langle \varepsilon \rangle$ imply higher rates of mass transfer and thus the possibility of supporting a higher biomass of corals as well as other benthic organisms such as sponges. Second, dissipation of wave energy reduces the setup produced by wave breaking, setup that is proportional to the wave energy flux to the 0.8 power [*Vetter et al.*, 2010]. Given that this setup is what drives mean flow through the reef inshore of the reef crest, higher dissipation rates imply weaker mean flows and thus longer residence times on the inshore reef which presumably affects a variety of ecological and biogeochemical processes [*Falter et al.*, 2013; *Teneva et al.*, 2013]. For example, longer residence times mean that coral metabolic rates are able to generate larger deviations in back reef carbon chemistry from open ocean conditions than would be possible under higher flow conditions [*Falter et al.*, 2013; *Teneva et al.*, 2013].

Palmyra Atoll reef is known to effectively be an end-member in terms of minimal human impact and thus is thought to be representative of healthy reefs [*Stevenson et al.*, 2006; *Sandin et al.*, 2008]. Given the analysis of *Ferrario et al.* [2014] highlighting the importance of wave energy dissipation as an ecosystem service provided by reefs, our observation that the Palmyra reef is much more dissipative than reefs that have had significant human modification suggests that healthy reefs are even better for coastal protection than are highly modified or damaged ones.

Finally, there is one caveat concerning the structure of healthy reefs: The Palmyra reef is not exposed to large cyclone-generated waves. In contrast, the Moorea reef shown in Figure 6b is significantly damaged periodically by cyclone-generated waves. Indeed, following cyclone Oli in May 2010, virtually all of the corals seen in Figure 6b were gone (J. Hench, personal communication, 2010). Thus, the value of healthy reefs to coastal protection may depend, at least partially, on whether or not a given reef is exposed to the very large waves that accompany cyclones and hurricanes.

References

- Bowden, K. F., and R. A. White (1966), Measurements of the orbital velocities of sea waves and their use in determining the directional spectrum, *Geophys. J. R. Astron. Soc.*, *12*, 33–54.
- Dean, R. G., and R. A. Dalrymple (1991), Water Wave Mechanics for Engineers and Scientists, Adv. Ser. on Ocean Eng., vol. 2, World Scientific, Hackensack, N. J.
- Emery, W. J., and R. E. Thompson (2004), 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, 1820–1831.
- Falter, J. L., R. J. Lowe, Z. L. Zhang, and M. McCulloch (2013), Physical and biological controls on the carbonate chemistry of coral reef waters: Effects of metabolism, wave forcing, sea level, and geomorphology, *PLoS One*, 8(1), e53303, doi:10.1371/journal.pone.0053303.

Ferrario, F., M. W. Beck, C. D. Storlazzi, F. Micheli, C. C. Shepard, and L. Airold (2014), The effectiveness of coral reefs for coastal hazard risk reduction and adaptation, *Nat. Commun.*, doi:10.1038/ncomms4794.

Gove, J., G. J. Williams, M. A. McManus, S. J. Clark, J. S. Ehses, and L. M. Wedding (2015), Coral reef benthic regimes exhibit non-linear threshold responses to natural physical drivers, *Mar. Ecol. Prog. Ser.*, 522, 33–48.

Hench, J. L., J. L. Leichter, and S. G. Monismith (2008), Episodic circulation and exchange in a wave driven coral reef lagoon system, *Limnol. Oceanogr.*, 53, 2681–2694.

Herbers, T. H. C., S. Elgar, and R. T. Guza (1999), Directional spreading of waves in the nearshore, J. Geophys. Res., 104, 7683–7693, doi:10.1029/ 1998JC900092.

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.

Kamphuis, J. W. (1975), Friction factor under oscillatory waves, J. Waterw., Harbors Coastal Eng. Div., 101, 135–145.

Lowe, R. J., J. R. Koseff, and S. G. Monismith (2005), Oscillatory flow through submerged canopies: 1. Velocity structure, J. Geophys. Res., 110, C10016, doi:10.1029/2004JC002788.

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.

Madin, J. S., and S. R. Connolly (2006), Ecological consequences of major hydrodynamic disturbances on coral reefs, *Nature*, 444, 477–480. Mei, C. C. (1985), Resonant reflection of surface waves by periodic sandbars, *J. Fluid Mech.*, 152, 315–335.

Mirfendereska, H., and I. R. Young (2003), Direct measurements of the bottom friction factor beneath surface gravity waves, *Appl. Ocean Res.*, 25, 269–287.

Monismith, S. G., L. M. W. Herdman, S. Ahmerkamp, and J. L. Hench (2013), Waves and wave-driven flow in a coral reef, J. Phys. Oceanogr., 43, 1356–1379.

Nunes, V., and G. Pawlak (2008), Observations of bed roughness of a coral reef, J. Coastal Res., 24(2B), 39-50.

Sandin, S. A., et al. (2008), Baselines and degradation of coral reefs in the northern Line Islands, *PLoS One*, 3(2), e1548, doi:10.1371/journal. pone.0001548.

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Scoffin, T. P. (1993), The geological effect of hurricanes on coral reefs and the interpretation of storm deposits, *Coral Reefs*, *12*, 203–221.
 Seelig, W. N. (1983), Laboratory study of reef-lagoon system hydraulics, *ASCE J. Waterw. Port Coastal Ocean Eng.*, *109*, 380–391.
 Sheremet, A., J. M. Kaihatu, S. F. Su, E. R. Smith, and J. M. Smith (2011), Modeling of nonlinear wave propagation over fringing reefs, *Coastal Eng.*, *58*, 1125–1137.

Stevenson, C., L. Katz, F. Micheli, B. Block, K. Heiman, K. Weng, C. Perle, R. Dunbar, and J. Witting (2006), High apex predator biomass on remote Pacific Islands, *Coral Reefs*, 327, 1–14.

Storlazzi, C., E. Brown, M. Field, K. Rogers, and P. Jokiel (2005), A model for wave control on coral breakage and species distribution in the Hawaiian Islands, *Coral Reefs*, 24, 43–55, doi:10.1007/s00338-004-0430-x.

Symonds, G., K. P. Black, and I. R. Young (1995), Wave-driven flow over shallow reefs, *J. Geophys. Res.*, *100*, 2639–2648, doi:10.1029/94JC02736. Teneva, L., R. B. Dunbar, D. A. Mucciarone, J. F. Dunckley, and J. R. Koseff (2013), High resolution carbon budgets on a Palau back-reef

modulated by interactions between hydrodynamics and reef metabolism, *Limnol. Oceanogr., 58*(5), 1851–1870, doi:10.4319/ lo.2013.58.5.1851.

Vetter, O., J. M. Becker, M. A. Merrifield, A. C. Pequignet, J. Aucan, S. J. Boc, and C. E. Pollock (2010), Wave setup over a Pacific Island fringing reef, J. Geophys. Res., 115, C12066, doi:10.1029/2010JC006455.

Williams, G. J., J. E. Smith, E. J. Conklin, J. M. Gove, E. Sala, and S. A. Sandin (2013), Benthic communities at two remote Pacific coral reefs: Effects of reef habitat, depth, and wave energy gradients on spatial patterns, *PeerJ*, *1*, e81, doi:10.7717/peerj.81.