Wave Dissipation by Bottom Friction on the Inner Shelf of a Rocky Shore

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Abstract

Approximately 32% of the measured wave energy flux by sea and swell waves was dissipated over distances less than 130 m, outside of wave breaking on the inner shelf, over a rocky shore in southern Monterey Bay, CA. The bottom roughness of the rocky shore is defined by the standard deviation of bottom vertical variability, $\sigma_b$, that is 0.9 m, which is of similar magnitude to previously measured $\sigma_b$ for rough coral reefs. Spectral wave energy flux balanced by bottom friction is modeled and compared with observations. Measured average wave reflection was 0.08 and is neglected in the model. The average energy dissipation owing to bottom friction over the rocky shore results in energy friction factors, $f_e$, ranging 4 to 34. The observed $f_e$ are larger than previously measured $f_e$ on coral reefs. An empirical power law relationship is developed for $f_e$ as a function of the ratio of wave orbital excursion amplitude, $A_o$, and $\sigma_b$, based on combined data from coral reefs, rocky platforms, and this rocky shore. As $\sigma_b$ increases, $f_e$ increases.

Numerical simulation by Yu et al. (2018, https://doi.org/10.9753/icce.v36.waves.57) of waves over large bottom variations, similar to observed on coral reefs, suggests that drag forces do not account for the large observed $f_e$. Therefore, it is hypothesized that bottom friction on rocky shores is a function of multiscale physical and biological roughness.

Plain Language Summary

During the spring and the fall of 2018, two experiments were conducted to examine how waves change as they progress from offshore to onshore over a rough rocky reef on the Central California coast. Wave statistics from a buoy offshore, which marked the edge of the rocky reef, were compared to wave measurements acquired farther onshore located seaward of wave breaking. Wave heights decreased between the edge of the reef and the corresponding onshore wave measurement stations. Since the observations are outside of wave breaking, the decrease in wave height is attributed to energy dissipation by bottom friction. Bottom friction is a function of bottom roughness, which was found to be as much 7 times larger than the roughest coral reef. Wave energy dissipation is hypothesized to be a result of the large and small rocks and biological growth on the rocky reef. Understanding wave transformation due to friction is important as this region is the home to a diverse ecosystem.

1. Introduction

It is estimated that 75% of the world’s coastlines can be described as rocky (Bird, 2000). Only a few wave transformation measurements have been made on rocky shorelines, and those studies have been limited to rocky shore platforms, which constitute about 20% of rocky shorelines (Emery & Kuhn, 1982; Kirk, 1977; Poate et al., 2018; Trenhaile, 2002). This leaves 60% of the world’s shorelines unstudied. Rocky shore platforms are near planar beaches composed of erodible rock (e.g., sandstone, mudstone, and limestone) (Kennedy & Beban, 2005; Marshall & Stephenson, 2011; Sunamura, 1992). Bottom roughness is a distinguishing characteristic between rocky platforms and rocky shores. The roughness of five platforms was measured by Poate et al. (2018) using a terrestrial laser with 0.1-m horizontal resolution and 3-mm vertical resolution. They found the roughness, measured as the standard deviation about the mean profile, $\sigma_b$, ranged 0.02 to 0.04 m (Table 1). This is compared with the measured roughness of $\sigma_b = 0.9$ m for the site of the experiment on a rocky shore in southern Monterey Bay, California, described in this paper. This is the roughest morphology reported to date. Rocky shores slope from offshore to the shore ending in a rocky beach, intertidal reef, or cliff. The rocky shore can be described as quasi-random undulations of rock mounds that result in quick transitions forming bathymetric highs and lows.
Coral reefs also are composed of a rough bottom that have been the subject of a number of wave transformation studies. However, the morphology of the coral reef is flat, with a width of tens of meters and is always submerged. Lowe et al. (2005) measured the bottom profile of the coral reef at Kaneohe Bay, Hawaii, with a resolution of 0.05 m in the horizontal and ±2 mm in the vertical and measured $\sigma_b = 0.04$ m (Table 1).

Jaramillo and Pawlak (2011) measured the bottom of a reef offshore of Honolulu, Hawaii, using side scan mounted on an autonomous underwater vehicle (AUV). They found that $\sigma_b$ varied from 0.03 to 0.07 m between bare reef areas and rough reef areas with spectral wavelengths between 0.2 and 6 m. Lentz et al. (2016) measured the bathymetry of a coral reef in the Red Sea using a floating downward looking Aquadopp resulting in a resolution of 0.2 m in the horizontal and 2 cm in the vertical and measured $\sigma_b = 0.13$ m.

To help put roughness in perspective, Yu et al. (2018) simulated the bottom roughness of coral reefs in a numerical study with 0.5‐m diameter hemispheres separated 0.75, 1, and 2 m apart. The corresponding $\sigma_b = 0.08, 0.06$, and 0.04 m, which are comparable to measured roughness on coral reefs (Table 1). It is noted that the standard deviation of the bottom roughness, $\sigma_b$, is a robust statistic that contains both height and length scale information (Duvall et al., 2018). In the example above, the height of the hemispheres is the same 0.25 m, but as the separation between hemispheres increases, $\sigma_b$ decreases inferring information about spacing.

Table 1

| Morphology/author | Location            | $\sigma_b$ (m) | $f_e$ |
|-------------------|---------------------|---------------|-------|
| Rocky shore       | Present paper       | Monterey, CA  | 0.9   | 4–34  |
| Coral reefs       | Rogers et al. (2018)| Ofu, Samoa    | 0.15–0.5 | 0.5–5 |
|                   | Rogers et al. (2016)| Palmyra Atoll | 0.5  |
|                   | Monismith et al. (2015)| Palmyra Atoll | 1.8  |
|                   | Lentz et al. (2016) | Red Sea       | 0.13  | 0.5–5 |
|                   | Lowe et al. (2005)  | Kaneohe Bay, HI | 0.04 | 0.24  |
|                   | Jaramillo and Pawlak (2011) | Honolulu, HI | 0.03–0.07 |
| Rock platforms    | Poate et al. (2018) | UK, Ireland   | 0.02–0.04 | 0.04–0.14 |
| Sandy shore       | Thornton and Guza (1983) | 0.16–1.3 · 10$^{-3}$a | 0.01–0.02 |
| Laboratory        | Simons et al. (1988) | Small-scale lab | 0.01 | 15–48 |
|                   | Yu et al. (2018)    | 0.04–0.08     | 0.001–0.03 |

*a*Calculated as the standard deviation of tightly packed hemispheres with $d_{50} = 1–8$ mm.

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Rocky shores support a highly productive and exceptionally diverse ecosystem compared with the smaller biological populations and lack of diversity on sandy beaches. For the past 50 years, wave‐swept rocky shores have served as a model system for exploring the basic mechanisms of community ecology. The diversity of plants and animals found there, the rapid turnover of individuals from wave‐induced disturbance, and the relative ease with which experimental manipulations can be implemented have served ecologists well: Much of the basic theory regarding competition, predation, recruitment, and disturbance (the fundamental drivers of community structure) has been based on (and tested in) this dynamic environment (e.g., Connell, 1961; Dayton, 1971; Denny, 1995; Menge, 1976; Paine, 1974; Paine & Levin, 1981). However, despite rocky shores’ long history of study and the central role played by wave‐induced transport and hydrodynamic forces in their ecology, the dynamics of wave transformation and induced circulation on the complex topography of these shores has not been studied.

Wave transformation across the rocky bottom, seaward of the surf zone on the inner shelf and assuming steady state, is modeled as the change in spectral wave energy flux, $F$, balanced by bottom dissipation due to bed friction, $\varepsilon_f$ (e.g., Hasselmann & Collins, 1968).
\[
\frac{dF_x(f, \theta, h)}{dx} + \frac{dF_y(f, \theta, h)}{dy} = -\varepsilon_f(f, h),
\]

where the wave energy flux \( F(f, \theta, h) \) is a function of frequency, direction, and depth with \( x \) directed onshore and \( y \) alongshore. Wave attenuation by scattering is neglected. It is assumed that the bottom contours are straight and parallel and that the waves are homogeneous in the alongshore that results in the \( y \) gradients equal to zero, which is discussed in the next section. The energy flux can then be expressed (Dean & Dalrymple, 1984)

\[
F_x(f, \theta, h) = E(f, \theta, h)C_g(f, h)\cos \theta(f, h),
\]

where \( E(f, \theta, h) \) is the total wave energy and \( C_g(f, h) \) is the group velocity given by linear wave theory.

Bottom friction dissipation is considered to be the work done against turbulent shear stresses induced at the bed by the water particle motions (Putman & Johnson, 1949)

\[
\varepsilon_f = \tau_b u_b,
\]

where the overbar indicates time averaging, \( u_b \) is the wave orbital velocity at the seabed, and \( \tau_b \) is the bottom shear stress defined by (Jonsson, 1966):

\[
\tau_b = \rho \frac{f_w u_b^2}{2 |u_b|},
\]

where \( \rho \) is the density of seawater and \( f_w \) is the wave friction factor. The wave friction factor, \( f_w \), is related to the bed friction coefficient \( C_f = f_w/2 \). \( C_f \) and \( f_w \) are friction factors commonly described in the literature, though with different nomenclatures. For energy dissipation with \( \tau_b \) described by 4, the relevant coefficient is changed to \( f_e \), referred to as the energy friction factor. While mathematically \( f_w \) and \( f_e \) are not equivalent due to a phase shift between the \( \tau_b \) and \( u_b \), when both friction factors are compared to each other, large experimental scatter exists between the two friction factors, and they are often assumed equal (Nielsen, 1992). Friction factors from this point forward will be referred to as \( f_e \) since the analysis is based on energy dissipation, substituting for the bottom shear stress formulation by (Jonsson, 1966):

\[
\varepsilon_f = \rho \frac{f_e u_b^2}{2 |u_b|},
\]

A primary objective of this study is to determine \( f_e \) for the inner shelf outside the surf zone on a rocky shore.

Wave transformation studies have found that \( f_e \) has a wide range of values (Table 1). For wave transformation studies on sandy beaches, \( f_e \) was found to range from 0.01 to 0.02 (Thornton & Guza, 1983), where bottom friction is often ignored over sandy bottoms. Poate et al. (2018) found \( f_e \) to range from 0.04 to 0.14 on rocky platforms. For coral reefs, Lowe et al. (2005) found \( f_e \) to be 0.28, while on a “remarkably” rough reef Monismith et al. (2015) found \( f_e \) to be 1.86 and Lentz et al. (2016) found \( f_e \) to range from 0.9 to 4.2. In general, as bottom roughness increases, \( f_e \) is found to increase.

In the following, wave and bottom measurements acquired over a rocky shore in southern Monterey Bay, California, are described (section 2). The bottom friction dissipation, \( \varepsilon_f \), is spectrally solved applying 1 to determine \( f_e \) for the inner shelf outside the surf zone on a rocky shore (described in sections 3 and 4). An empirical relationship between \( f_e \) and the ratio of wave orbital excursion, \( A_{ob} \), measured \( \sigma_b \), is developed for the data acquired in this study along with results from three previous experiments on coral reefs and rocky platforms to generalize the results (section 5).

2. Field Experiments

Waves were measured on the rocky shore off of Stanford’s Hopkins Marine Station (HMS), Monterey Bay, California. HMS is located at the southern end of the Monterey Bay and is characterized by an irregular rocky coastline (Figure 1a). Two wave transformation experiments were conducted: (1) Experiment A was from 7 March until 4 April 2018 (yeardays 68–94), and (2) Experiment B was from 12 October until 1 November 2018 (yeardays 285–305).
2.1. Bathymetry

A 2012 bathymetric survey was provided by the Sea Floor Mapping Laboratory (SFML) at California State University at Monterey Bay. The SFML survey used a Reson Seabat 7125 multibeam echosounder, which has resolutions of 6 mm in the vertical and 1 m in the horizontal. The accuracy associated with the inertial motion unit and GPS is estimated as ±0.15 m in the horizontal and ±0.05 m in the vertical (Barnard et al., 2011). The bathymetry from 18 to 5 m water depths is shown in Figure 1b. The experiment site is outlined by gray lines. The bathymetry offshore of the experiment site is constant in the alongshore direction (y) with no abrupt changes in bottom elevation (z), where z = 0 m at MSL. In the cross-shore direction (x) approaching the rocky shore, the smooth bottom is composed of sand (Eittreim et al., 2002). Conversely, within the experiment site (denoted by insert, Figure 1b) there are significant elevation changes in x as well as in the y directions. From x = 200–175 m, variability about the mean profile starts to increase. The Edge Of the Rocky shore (referred to as EOR) is identified at x = 175 m in a depth 13.4 m (dotted lines, Figures 1b and 1c). Approaching the rocky coastline (x = 175–50 m), there is large variability in bottom elevation owing to large rock features.

Cross-shore bathymetric profiles were measured every meter in the alongshore in the experiment site (Figure 1d) and averaged to create a mean profile (black line, Figure 1c). The standard deviation of the mean profile and the maximum and minimum bottom elevations are provided in Figure 1c with corresponding instrument locations. Based on the mean profile of the experiment site (Figure 1c), the slope prior to the EOR is mild at 1/65 and shows little variation. From the EOR (black vertical line) to the end of the experiment site, the mean slope increases and is steeper at 1/27. The variability in the mean profile (black line) is indicative of the undulations due to the Miocene age, igneous rocks that are common on the Central California coast (Eittreim et al., 2002). The standard deviation of the mean profile of approximately 1 m is...
relatively consistent along the profile. The bathymetry of the rocky shore is complex with large rock features throughout the field site.

A histogram of the vertical variations (\(z'\)) about the mean bottom profile is calculated to better portray the bottom roughness (Figure 2). \(z'\) have a standard deviation of 0.9 m with 87% of the undulations ±1 m from the mean profile. The remaining 13% of the vertical variations are distributed in the larger more extreme undulations of up to +4 and −3 m highlighting the roughness and spatial variability of the bathymetry. The measured roughness values are the largest seen to date in the field.

A two-dimensional autocorrelation was performed on \(z'\) in Figure 1d to examine the ensemble-averaged horizontal spatial scales (not shown). Characteristic horizontal \(x\) and \(y\) scales are specified by the \(e\)-folding decorrelation scales, which are approximately 14 and 8 m. On average, the bottom is composed of large rock features that have approximate horizontal scales of 14 m long and 8 m wide with rms height of 2.5 m.

### 2.2. Waves

Waves were measured offshore in 17.8 m of water by the National Data Buoy Center Waverider Buoy, Station 46240 at Cabrillo Point (referred to as Cabrillo Point Buoy, CPB), which is hosted by the by the Coastal Data Information Program at the University of California at San Diego. The buoy is located 255 m seaward of EOF (Figure 1b). The NDBC provides 30-min estimates of significant wave height, \(H_{\text{m0}}\), spectrally weighted wave period, \(T_{\text{m0}}\), peak wave period, and wave direction associated with the peak wave period. Spectral energy estimates are provided by CDIP every 30 min with a variable frequency resolution for nine frequency bins spanning 0.0455 to 0.5 Hz and an associated mean wave direction for each frequency band. Hence, in the application of measurements, the frequency-direction energy spectrum is approximated by

\[
S_b(f, \theta, h) \equiv S_b(f, \bar{\theta}, h),
\]

where \(\bar{\theta}\) is the mean direction at frequency \(f\).

Experiments A and B, while 6 months apart, experienced similar offshore wave conditions with \(H_{\text{m0}}\) ranging from 0.3 to 2 m (Figures 3a and 3b). The waves at this location are predominantly swell from west to southwest (Figure 3c) owing to the narrow aperture between the headlands at Point Santa Cruz to the north and Point Pinos to the south (see Figure 1a). About 88%, 97%, and 95% of the mean incident wave directions measured at the offshore buoy during Experiments A, B, and C were within ±15° (Figures 3c and 3e). About 76%, 88%, and 90% of the mean incident wave directions measured at the offshore buoy during Experiments A, B, and C were within ±10° (Figures 3c and 3e).

The narrow aperture also has the effect of narrowing the frequency spectrum. The spectrally weighted wave period, \(T_{\text{m0}}\), ranged from 5–12 s. Tides in the Monterey Bay are mixed, mainly semi-diurnal, where the low-low tide always follows the high-high tide with a tidal range of approximately 2 m (Broenkow & Breaker, 2005).

Inshore wave estimates were obtained using bottom-mounted RBR Solo-D pressure sensors located well outside of wave breaking in 7.1-m (A) and 8.8-m (B) water depths (Figures 1b–1d). The pressure sensors measured 1-hr records sampled at 2 Hz. Spectra were calculated applying a Hanning window to 10-min sample records overlapped by 50% that were ensemble averaged to obtain spectral estimates with 32° of freedom at a resolution of 0.0017 Hz. The pressure spectra were converted to surface elevation spectra applying a linear wave theory transfer function (Guza & Thornton, 1980).

Wave energy reflection was measured using a Nortek Signature 1000 Acoustic Doppler Current Profiler (ADCP) during a separate Experiment C for 13 days from 11–24 June 2019. The ADCP was located in 9-m water depth (Figures 1b–1d). Surface elevation was measured using Acoustic Surface Tracking (AST). The surface elevation and wave velocities were burst sampled at 2 Hz for 17 min every 2 hr. Velocity data were rotated 13.5° from magnetic north to geographic north. Directional wave spectra, \(E(f, \theta)\), were computed using the Maximum Likelihood Method (Capon, 1969; Oltman-Shay & Guza, 1984). Spectra were...
calculated applying a Hamming window to 10 records of 102 s with 50% overlap, which resulted in 53 degrees of freedom, a resolution of 0.01 Hz with direction resolved at 4° and smoothed with a 16° directional window. The bathymetry and waves are both rotated together 15° so that the $x$ component is approximate shore normal. The rotated mean wave directions measured by the waverider buoy in 17.8-m depth, CPR, and the ADCP in 9-m depth are compared in Figures 3c, 3d, and 3e.

The reflection coefficient, $R^2(f)$, was estimated from the directional spectra as (Elgar et al., 2003):

$$R^2(f) = \frac{E_{off}(f)}{E_{on}(f)} = \frac{\int_{-90}^{90} E(f, \theta) d\theta}{\int_{-90}^{90} E(f, \theta) d\theta}$$  \hspace{1cm} (7)

where subscript on and off refer to waves moving onshore and reflecting offshore at the measurement location. The mean reflected wave energy coefficient integrated over frequency in 7 averaged 0.08 for sea and swell waves (0.04–0.2 Hz) on this rocky shore. Wave energy reflection is consistent with other wave energy reflection estimates on rough bottoms [0.1 at Palmyra atoll (Monismith et al., 2015) and 0.16 at Red Sea coral reef (Lentz et al., 2016)]. It is also consistent with field measurements of wave energy reflection (0.09 for 0.1-Hz wave frequency) from the nearby breakwater at Monterey Harbor (Dickson et al., 1995). Monismith et al. (2015) and Lentz et al. (2016) considered wave energy reflection negligible for computing wave energy transformation, as is assumed here.

### 3. Spectral Wave Dissipation

The total wave energy spectrum in 2, to order of linear wave theory, can be expressed in terms of the surface elevation spectrum $S_\eta(f, \theta)$:

$$E(f, \theta, h) = \rho g S_\eta(f, \theta, h).$$  \hspace{1cm} (8)
Substituting into 1 along with the bottom friction dissipation function 5 gives

\[
\frac{\rho gdS_b(f, \vartheta, h)C_g(f, h)\cos \vartheta(f, h)}{dx} = -\frac{\rho f u_b^2(f, h)|u_b(f, h)|}{2}. \tag{9}
\]

The energy dissipation term in 9 results in spectral-triad interactions across frequencies that make the problem nonlinear and difficult to solve. To linearize the energy dissipation, Collins (1972) substituted a characteristic velocity, \(u_b\), for one of the velocities terms to uncouple the spectral components leading to simple spectral interactions

\[
\varepsilon_f(f, h) = \rho \frac{f}{2} u_b S_{ub}(f, h), \tag{10}\]

where \(S_{ub}(f)\) is the velocity spectrum at the bed. The characteristic velocity is defined as follows:

\[
u_b = \left(\int_s S_{ub}(f, h)df\right)^{\frac{1}{2}} = u_{rms}. \tag{11}\]

The velocity spectrum at the bed is related to the surface elevation spectrum by

\[
S_{ub}(f, h) = |K_{ub}(f, h)|^2 S_h(f, h), \tag{12}\]

where \(K_{ub}(f,h)\) is the transfer function that converts surface elevation to velocity at the bed given by

\[
K_{ub}(f, h) = \frac{2\pi f}{\sinh kh}. \tag{13}\]

where \(k\) is the wave number.

A question arises as to the error incurred in using a characteristic velocity approximation in 10 to solve for the bottom friction dissipation. An alternative to the frequency spectrum description is the joint wave height/period distribution, \(p(H, T)\), which can be used to solve for the bottom friction dissipation without assuming an approximation (Madsen et al., 1988). However, the period distribution is dependent on the spectral bandwidth, \(v\), (i.e., the spectral shape), whereas the wave height distribution is only weakly dependent on \(v\) (Longuet-Higgins, 1983). Thus, the joint distribution does not provide a general description of wave processes without specifying \(v\).

However, as waves move into shallow water, waves become nondispersive and the transfer function relating the wave velocity at the bed to the surface elevation 13 in the bottom dissipation function 10 is independent of frequency (period). Therefore, waves can be completely described by just the wave height distribution, \(p(H)\), in shallow water. As shown in Appendix A, an exact solution for the bottom dissipation function to the order of linear wave theory can be obtained using the wave height distribution in shallow water. It is found that a correction factor of \(\sqrt{2/\pi} = 0.8\) is required in the spectral formulation 9 to match the more exact wave height distribution formulation in shallow water, which is applied in the analysis below. It is satisfying that the correction is relatively small.

4. Results

The energy flux Equation 1 with substitution of 10 is solved. Since \(\varepsilon_f\) in 10 is a function of \(S_{ub}(f)\) and \(k\), which are functions of the local \(h\), the energy flux Equation 1 cannot be solved directly. Therefore, 1 is solved numerically one frequency at a time with an iterative forward differencing scheme over the measured bathymetric profile:

\[
F(f, \vartheta, h(x + \Delta x)) = F(f, \vartheta, h(x)) - \frac{\rho f \Delta x}{2} \sqrt{\frac{2}{\pi}} u_b(\vartheta, h(x)) S_{ub}(f, \vartheta, h(x)), \tag{14}\]

where the energy flux is given by

\[
F(f, \vartheta, h) = \rho g S_h(f, \vartheta, h)C_g(f, h)\cos \vartheta(f, h), \tag{15}\]

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The mean wave angle at each frequency is solved by Snell’s law over straight and parallel contours

\[
\theta_f = \sin^{-1}\left(\frac{C[f, h(x + \Delta x)]}{C[f, h(x)]} \cdot \sin \theta_f \right) \quad \text{(16)}
\]

It is noted that the correction of factor \( \sqrt{2/\pi} \) (see Appendix A) has been included in \( \text{14} \).

Direct spectral wave measurements at the EOR were not available for either experiment as sensors were not deployed there. The sea floor between CPB buoy and the EOR is a smooth, sandy bottom (Figure 1b). The \( \text{fe} \) are obtained by solving \( \text{14} \) between EOR and A and B. The solution starts with the calculated initial offshore condition \( F(f, h(\text{EOR})) \) and marches forward over the bottom profile at \( \Delta x = 1 \text{ m} \) intervals with a specified \( \text{fe} \) to solve for \( F(f, h(A,B)) \). \( \text{fe} \) is then iterated until the inshore calculated integral

\[
F(A, B) = \int F(f, h(A, B))df
\]

(term on left side of \( \text{14} \)) matches with the measured integral of the energy flux spectrum at A and B. The numerically determined \( \text{fe} \) represents a spatial average over the entire varying profile. In actual application of \( \text{14} \) and \( \text{15} \), it is assumed for simplicity that the waves are normally incident. Owing to the near-normal incidence of the measured waves, the error is small as discussed below.

The average reduction in \( F \) between the EOR and A was 28% over 131 m apart, and between EOR and B was 36% over 116 m apart. These reductions are substantial considering the relatively short distances between EOR and A and B.

\( \text{fe} \) is compared with the rms water-particle excursion amplitude at the bed, \( A_b \), averaged over the entire profile. For linear wave theory,

\[
A_b(h(x)) = \left[2\int (2\pi f)^{-2} S_{w1}(f, h(x))df\right]^{1/2}
\]

and averaging over the profile

\[
A_b = \frac{1}{L} \int_{\text{EOR}}^B A_b(h(x))dx
\]

where \( L \) is the length of the profile. The \( \text{fe} \) values are bin averaged over 0.5 m \( A_b \) bins (squares in Figure 4). The \( \text{fe} \) values are color coded by wave period. Bin-averaged \( \text{fe} \) values decrease from 33 to 4 as a function of increasing \( A_b \).

5. Discussion

5.1. Error in Assuming Waves Approach Normal to the Shore

It was assumed the incident waves arrive normal to the shoreline in the \( x \) direction over straight and parallel contours. The difference between that and accounting for the direction of the waves is small owing to the relatively small range of incident wave angles (Figure 3c). The sensitivity of nonnormal directionality is measured in the refraction coefficient, given by (Dean & Dalrymple, 1984)

\[
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Figure 4. \( \text{fe} \) as a function of \( A_b \) averaged over the profile for Experiments A and B. Circles represent individual hourly estimates of \( \text{fe} \) that have been color coded based on wave period, \( T_{\text{mo}} \). Large gray squares are 0.5-m bin-averaged values of \( \text{fe} \).
\[ K_{fl}(f, h) = \frac{\cos \alpha_0}{\cos \left( \sin^{-1} \left( \frac{C(f, h) \sin \alpha_0}{C_0(f)} \right) \right)} \]

(20)

where \( C \) is the phase speed, \( \alpha \) is wave angle relative to normal incidence, and the subscript 0 refers to the initial location. A difference of \( \pm 15^\circ \) results in less than a \( \pm 2\% \) difference in \( K_{fl}(f) \) for all values of \( f \), which is within the margin of error of the measurements. Therefore, since the waves angles are constrained by the relative narrow aperture, wave angle do not significantly impact the measured \( f_e \) values.

5.2. Energy Friction Factor \( f_e \)

The calculated \( f_e \) values for the rocky shore are the largest found to date. The largest \( f_e \) values are associated with the smallest \( A_b \), which are associated with the shortest period waves and/or the lowest waves (Figure 2). Qualitatively, it appears that \( f_e \) decreases with increasing wave period. Putman and Johnson (1949) compared the transformation of 6-12 s period waves with same wave height over bottom slopes of 1/300 and 1/10 owing to bottom friction dissipation. Greater dissipation was found for the short period waves and for the low sloped bottom. They argued that it is the cumulative effects of the number of wave cycles doing work in relative shallower water that results in more dissipation; that is, there are more cycles for short period waves that travel slower over the same distance and more cycles required to transit low slope beaches resulting in larger \( f_e \) values.

The largest \( f_e \) values may also be associated with the smallest waves. Simons et al. (1988) measured wave attenuation in a small-scale laboratory experiment \( (H = 0.02 \text{ m}, T = 1 \text{ s}, h = 0.3 \text{ m}) \) using a bed composed of 10-mm angular limestone. They found \( f_e \) values ranging 15–45 (Table 1) associated with small \( A_b \) values ranging 0.06–0.7. The small-scale laboratory experiment scaled geometrically and with Froude number \((H/L, h/L; u/\sqrt{gh})\) but not with Reynolds’s number. However, the angular limestone roughness may differ substantially from the rocky bottom. It is not understood why small wave heights would result in large \( f_e \) values.

There is considerable variability of \( f_e \) values for a particular \( A_b \) (Figure 4). To calculate \( f_e \), the energy flux measured inshore at A and B with a pressure sensor is matched in the wave transformation differencing scheme and then divided by the dissipation function in 14. The inshore energy flux is obtained by multiplying the pressure spectrum by a spectral transformation function \( iK_p(f,h) \), and the dissipation function is obtained by multiplication of the surface elevation spectrum by spectral transform 13, such that

\[ f_e = \frac{|K_p(f, h)|^2}{|K_{ab}(f, h)|^2} \left( \frac{\cosh^2 kh}{(2\pi f / \sinh kh)^2} \right) \]

(21)

that has values in 8.8-m water depth of 35 for 5-s waves and 1.8 for 10-s waves. The combined transfer functions in 18 have the effect of amplifying any noise in the measurements. This may explain the larger variability at small values of \( A_b \) associated with short period waves (Figure 4) but does not explain the order two variability for longer period waves.

5.3. Relating \( f_e \) to \( \sigma_b \)

\( f_e \) has been related to hydraulic roughness (\( \sigma_b \)) and \( A_b \) through several empirical relationships based on laboratory results (e.g., Grant & Madsen, 1982; Nielsen, 1992; Soulsby, 1997). Soulsby (1997) finds the power law relationship:

\[ f_e = 1.39 \left( \frac{A_b}{\sigma_b} \right)^{-0.52} \]

(22)

Previous experiments did not measure \( \sigma_b \) requiring investigators to apply one of the empirical relationships in order to determine an optimized value of \( \sigma_b \) from known values of \( f_e \) and \( A_b \) (Lentz et al., 2016). It is assumed \( \sigma_b = k_n / \beta^3 \) where \( k_n \) is the geometric bottom roughness scale. Henceforth, \( \sigma_b \) represents \( k_n \), \( \beta \) is a proportionality coefficient that relates \( k_n \) to \( \sigma_b \) based on the environment whose roughness is being quantified (Bagnold, 1946). \( \beta \) has been reported anywhere between 2.5 and 100 for river, ocean, and atmospheric boundary layers depending on the given environmental conditions (Britter & Hanna, 2003; Jimenez, 2004;
The calculated $f_e$ values are an order of magnitude greater than previously reported for field observations, and the bottom roughness $\sigma_b$ are among the largest ever reported. This leads to a paradox. Yu et al. (2018) used Large Eddy Simulations to calculate turbulent boundary layers under waves over a bottom composed of 0.5-m diameter hemispheres for three cases of evenly spaced intervals, $S = 0.75$, 1, 2 m. They partitioned the calculated bottom stress into form drag and inertial force components and found that decreased flow separation owing to the small curvature of the hemispheres resulted in decreased form drag and enhanced inertial forces. Inertial forces are due to the mass of the fluid having to be accelerated around the rocks (hemispheres). The larger the rock, the more acceleration required and the larger the inertial forces. However, the inertial forces do not contribute to wave dissipation as they are in quadrature with the work being done on the bed. The bottom dissipation coefficients $f_e$ calculated by Yu et al. (2018) due to drag forces only are plotted in Figure 5 and account for only about one half of the expected $f_e$.

Visual inspection of the bottom by swimmers found that there are multiple, unresolved scales of roughness that are not accounted for in the roughness measured by the side scan. Based on the horizontal resolution of the bathymetry by the side scan, the rocks are perceived as smooth structures. The rocks are not smooth but are jagged and have peaks and valleys and different crevices. Additionally, the ecosystem is diverse and made up of different rocky invertebrates and algae that are on centimeter and smaller scales.

It is hypothesized that the large values of $f_e$ and resulting dissipation by friction are owing to multiscale physical and biological roughness. The roughness is composed of large-scale variations owing to rock (coral) outcrops and small-scale roughness on the order of centimeters owing to roughness of rocks (corals) and marine growth on the rocks (corals). The smaller features hosted by these large rocks (corals) cause flow separation resulting in form drag, which may partially explain the large amount of dissipation and resulting large $f_e$.

The unresolved small-scale roughness also may be the result of marine growth on the rocks composed primarily of seaweeds, which are ubiquitous (see Figure 6). The contribution by benthic seaweed growing on...
the rocky bottom can be approximated by measurements of forces on the seaweeds. As water moves past benthic algae, it exerts a drag force on the algal fronds, which can be expressed as a bottom stress described by 4. Seaweeds can densely cover the seafloor. Often, the combined area of blades exceeds the area of substratum to which they are attached. The ratio, $B_f$, of the total blade area of seaweeds to the area of substratum to which the algae are attached is typically $5 < B_f < 34$ (Leigh et al., 1987). In describing the forces on the algal, an equivalent bottom friction dissipation coefficient is used,

$$f_e = C_D B_f,$$

(24)

where $C_D$ is a dimensionless drag coefficient obtained from measurements, which is roughly 0.2 for most seaweeds (e.g., Martone et al., 2012). Applying the range of values for $B_f$, the biological, smaller-scale bottom friction dissipation factor, $f_e$, could range 1 to 7. The results suggest that benthic algae growing on the rocky bottom could add substantially to $f_e$. Accounting for the multiscale bathymetry requires considerable exploration, complicated by contributions from solid rock and flexible biological roughness.

5.5. Further Comment About Biology

Rocky coastlines in the temperate zone, such as California, have large forests of Macrocystic kelp offshore in the approximate depth range between 15 and 7 m. This kelp extends from the bottom to the surface, has compliant foliage, and can have large canopies. The question arises as to whether the kelp dissipates waves. In a study comparing wave transformation across a kelp bed and over a nearby barren sandy bottom, Elwany et al. (1995) found that there was no measurable difference. Similar results were found by Rosman et al. (2007). It is expected that kelp has evolved to minimize the absorption of wave energy for survival. It is observed antidotally that high-frequency capillary waves are damped either mechanically or chemically. However, capillary wave-induced velocities are confined to the near surface and do not extend to the bottom. In addition, Experiments A and B were conducted in the winter and fall, after storm waves had cut back the kelp. Also, sea urchins have decimated the kelp forest in recent years. Therefore, it is concluded that Macrocystis kelp was not a factor to wave dissipation in this study.

5.6. Increased $f_e$ Owing to Increased Area of an Undulating Bottom Profile

Frictional dissipation over an undulating bottom is greater than over a planar bottom owing to the increased area the wave bottom velocities have to work. This is similar to the argument presented by Putman and Johnson (1949) that there are more wave cycles to increase dissipation for waves over less steep beaches.
The lengths of five cross-shore profiles with a 1-m horizontal resolution along the 100-m shoreline considered (Figures 1b and 1d) were measured. The undulations increase the length of travel by waves by 2–5% compared with a planar profile. The dissipation for 1 m high, 10-s period waves with $f_c = 6$ traversing the depths considered for a planar beach is compared with the undulating beach profiles. The increased lengths for the undulating profiles effectively increase $f_c$ values by 3% to 10%.

5.7. Hypothesis: Increased $f_c$ Owing to Wave Scattering

Wave scattering of the sea/swell is not well understood on a rocky shoreline. Scattering can occur from reflection off the bottom of the heterogeneous large rocks, reflection from the steepened rock shoreline, or possible Bragg scattering from the irregular rocky bottom at larger scales. Wave scattering would appear the same as bottom friction dissipation in 1. Although measures of wave reflection during Experiment C in 9-m water depth were low ($R^2 = 0.08$), the possibility that wave reflection is important nearshore is still an open question. The mean incoming wave directions were relatively constant and narrow with a mean bandwidth of ±10°. The mean reflected wave direction was near specular but diffuses with a directional bandwidth of ±45°. The reflected direction bandwidth is limited by a critical refractive turning back (caustic) of large angle directions in 9-m water depth. To resolve the importance of wave scattering, wave reflection would be needed to be measured at various cross-shore locations with colocated pressure and velocity sensors. Wave scattering in this environment remains an open question that needs to be addressed.

6. Summary and Conclusions

Results from 2-month-long wave transformation experiments conducted on the rocky shore at HMS, in the Monterey Bay, found on average 28% and 36% reduction in the wave energy flux starting at a depth of 13 to 8.8 m over a distance of 131 and to 7.1 m over a distance of 116 m, respectively, outside of wave breaking. This was an unexpected large amount of wave dissipation over a relatively short distance. The spectral wave energy flux equation balanced by dissipation by bottom friction is solved over assumed straight and parallel contours. Wave scattering (reflection) as a first approximation is assumed negligible. Bottom friction is described as the work done on the bottom by the wave velocity against a quadratic bottom stress parameterized by the bottom dissipation coefficient $f_c$. An objective of the study is to determine $f_c$ values as functions of bottom roughness and wave characteristics. The nonlinear spectral-triad bottom friction is linearized, and the approximation is shown to be reasonable. The calculated $f_c$ values ranging 4 to 34 are greater than previously reported for field observations, and the bottom roughness $\sigma_b$ of 0.9 m is among the largest ever reported. The largest $f_c$ are associated with the smallest $A_b$, which were associated with the shortest period waves. The $f_c$ values decrease with increasing $A_b/\sigma_b$.

In an attempt to generalize the results for $f_c$, field studies of wave transformation over other rough bottoms are included in the analysis. Only three data sets were found available where the bottom roughness was measured directly. Lowe et al. (2005) and Lentz et al. (2016) acoustically measured $\sigma_b$ of 0.04 and 0.13 m, respectively, on coral reefs. Poate et al. (2018) measured the morphology on five different rock platforms and found that $\sigma_b$ ranged 0.02 to 0.04 m. The largest $\sigma_b$ of 0.9 m is for the HMS rocky reef. These data sets are used to establish an empirical power law relationship that relates values of $f_c$ based on measured $A_b$ and $\sigma_b$ covering four decades over the range $0.01 < A_b/\sigma_b < 100$. The field data for large $\sigma_b$ and concomitantly large $f_c$ extend earlier power law formulations developed for laboratory experiments, where results for very low $A_b/\sigma_b$ were limited.

However, the large values of $f_c$ for large bottom roughness are contradicted by numerical wave studies leading to a paradox. Yu et al. (2018) used Large Eddy Simulations to calculate turbulent boundary layers under waves over a rough bottom composed of equally spaced, 0.5-m hemispheres. They partitioned the bottom stress into form drag and inertial force components and found that decreased flow separation owing to the small curvature of the hemispheres resulted in decreased form drag and comparable inertial forces. The inertial forces do not contribute to wave dissipation as they are in quadrature with the work being done on the bed. The bottom stress due to drag forces accounts for only about one half of the observed $f_c$. Therefore, it is hypothesized that bottom roughness is multiscale, with large-scale variations owing to rock outcrops and small-scale roughness on the order of centimeters owing to roughness of rocks and marine
growth on the rocks. The smaller features hosted by these large rocks cause increased flow separation resulting in form drag, which may partially explain the large amount of dissipation and resulting large $f_c$. Wave dissipation based on previous studies of bottom stress owing to seagrass growing on the rocks suggests that large values of $f_c$ ranging 1 to 7 may result, again partially explaining the large amount of dissipation.

Uncertainty in the $f_c$ values is associated with the 1-D assumption in the wave transformation calculations. Owing to the relatively narrow incident wave directional bandwidth and the wave arriving at near-normal incidence in the mean, the expected error was limited to less than ±2%. The measured wave reflection was 0.08 and was neglected. Including the wave reflection effect would reduce the wave attenuation by bottom friction by a comparable amount. This indicates that the measured $f_c$ are overpredicted by 8%.

These are the first wave transformation studies for this type of rocky shoreline and are limited in scope. It is not surprising that a number of unanswered questions arise in this initial inquiry. It is expected that $f_c$ spatially varies in the cross shore as the bottom roughness varies. An array of wave instrumentation in the cross shore would address this question. It was hypothesized that bottom is composed of multiscale physical and biological roughness. Future studies will need to measure the bottom with greater resolution to examine the importance of small-scale roughness and biologics. It is hypothesized that wave scattering, either by bottom Bragg scattering or reflection from the shoreline, contributes to wave attenuation that is not discernable from wave dissipation. The measured reflection coefficient of 0.08 in 9-m depth suggested that the scattering may be weak and was neglected in the analysis. However, Bragg scattering is expected to increase in shallower water and have a greater contribution.

The average 32% reduction in wave energy by bottom friction acts to protect the coast from wave forces, runup, and overtopping. The energy reduction could be a contributing factor in enabling the diverse intertidal ecosystem to sustain and grow as wave-generated forces, which are the leading cause of mortality, are lessened in the rocky intertidal zone (Helmuth and Denny, 2003).

**Appendix A: Correction Factor for Linearizing Bottom Friction Term**

A question arises as to the error incurred in using a characteristic velocity approximation in 10 to solve for bottom friction dissipation. Spectra can be described as an infinite sum of sinusoidal wave components with averaged variance contributions at varying frequency over the ensemble, or in practical terms, over the time of the measurement period. The sum of the spectral components is the area under the spectrum equaling the variance. An equally valid approach is the joint height/period probability distribution that describes an infinite collection of individual waves each described as a sinusoid with a height and period. Both approaches are valid to the order of linear wave theory. The total bottom friction dissipation integrated over the spectrum should equal the probability distribution summed over all wave heights and periods.

As pointed out above, the joint wave height/period distribution, $p(H,T)$, can be used to solve for the bottom friction dissipation without assuming an approximation (Madsen et al., 1988). However, the period distribution is dependent on the spectral bandwidth, $\nu$, (i.e., the spectral shape), whereas the wave height distribution is only weakly dependent on $\nu$ (Longuet-Higgins, 1983). Thus, the joint distribution does not provide a general description of wave processes without specifying $\nu$.

A simplification is to consider shallow water where the waves are nondispersive such that the velocity transfer function is independent of frequency

$$K_{ab}(f,h) \rightarrow K_{ab}(h) = \left(\frac{\nu}{H}\right)^{\frac{3}{2}},$$

(A1)

Putman and Johnson (1949) solved $\varepsilon_f$ for a single wave by describing the velocity using linear wave theory in 4 and integrating in the time domain

$$\varepsilon_f(h) = \frac{\rho f_c}{2} \left( K_{ab}(h) \frac{H}{2} \right)^{\frac{3}{2}} \frac{1}{\pi} \int H \cos^2 \omega t | \cos \omega t | dt,$$

(A2)

to give

$$\varepsilon_f(h)_{\text{dist}} = \frac{\rho f_c}{2} \frac{1}{6\pi} K_{ab}^3 H^3.$$ 

(A3)
It is noted that for shallow water waves, the dissipation due to bottom friction is independent of frequency. The joint distribution then collapses to a wave height distribution that is well described by the Rayleigh distribution, \( p_h(H) \). Following Thornton and Guza (1983),

\[
\overline{\varepsilon_r(h)} |_{dist} = \rho f_s^2 \frac{1}{3} K_{ub}(h) H^{1/2} p_h(H) dH,
\]

(A4)

Integrating over all waves in the distribution gives

\[
\overline{\varepsilon_r(h)} |_{dist} = \rho f_s^2 \frac{1}{16 \sqrt{\pi}} [K_{ub}(h)H_{rms}]^3 = \rho f_s^2 \frac{1}{2 \sqrt{\pi}} H_{rms},
\]

(A5)

where \( H_{rms} = \sqrt{8} \sigma \) is applied. Equation A5 is exact in shallow water to the order of linear theory. Now considering the total bottom dissipation in the spectral model 9 by averaging over the velocity spectrum given by 10,

\[
\overline{\varepsilon_r(h)} |_{spec} = \rho f_s^2 H_{rms} \int S_{ub}(f) df = \rho f_s^2 \frac{3}{2} H_{rms},
\]

(A6)

which if it were correct should converge in shallow water to equal the total bottom dissipation for the distribution. However, a difference is noted. Therefore, spectral representation of \( \varepsilon_r(f) \) is multiplied by a correction factor of \( \sqrt{2/\pi} \) in the numerical solution 14 that is applied to \( f_s \) to correct for the approximation of using a characteristic velocity in 10.

**Data Availability Statement**

Steve Lentz and Justin Rogers openly shared their data and provided critical insights on this work. We thank Xiou Yu for providing his data. Bathymetric data are provided by CSUMB (https://seafloor.otterlabs.org/SFMLwebDATA.htm). Offshore wave data are provided by NDBC (https://www.ndbc.noaa.gov). Data collected here in are archived online (https://doi.org/10.5281/zenodo.3572479).

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