Wave Energy Delivery and the Shape of Rocky Coasts:
Part 1 – Introduction and Background on Wave Theory

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Though responsible for many breathtaking landscapes, the coastal geomorphology of active margins has received much less scientific attention than the depositional landforms and sedimentary processes responsible for decorating passive margins. Some studies have focused on dating marine terraces (Merrits and Bull, 1989; Perg et al., 2001), which form as a result of wave action, tectonic uplift and sea level fluctuation (Anderson et al., 1999), but little work has been done toward understanding the processes responsible for shaping the planform appearance of active margin coasts like that of central California. Some of the factors responsible for sculpting the features we recognize on active margin coasts are depicted on Figure 1 of this section.

When considering geomorphic “damage” inflicted upon a rocky coast, the variable of interest is the delivery power, $P_D$, of waves. This power is strongly controlled by oceanographic conditions and bathymetry. To better understand the delivery power, one must examine the deep-water wave power, $P_0$, and how it is modified through wave transformation.

Deep-Water Wave Height

Deep-water wave height, $H_0$, should exert the greatest control on delivery power. The energy density, $E_0$, of a deep-water wave is given by

$$E_0 = \frac{1}{8} \rho g H_0^2$$

(1)

where $\rho$ is the density of seawater and $g$ is gravitational acceleration. Deep-water wave power, $P_0$, is defined as the energy flux per unit length of wave crest averaged over one wave period (Sunamura, 1992; Komar, 1998),

$$P_0 = E_0 C_0 n = \frac{1}{8} \rho g C_0 n H_0^2$$

(2)

where $C_0$ is deep-water wave celerity and $n$ describes the shape evolution of a wave as it shoals, a hyperbolic function whose value is 1/2 in deep water and 1 in shallow water. The transformation of waves as they interact with the shelf results in evolution of both their height and celerity. These transformations involve changes in wave geometry both in plan view and in cross section (Fig. 2).

Wave Shoaling

Airy wave theory assumes that in the absence of refraction and bottom friction, wave power is conserved from deep to shallow water. Changes in wave height must therefore result in changes of the opposite sign in celerity (equation 2). This wave-shape evolution can be
expressed as a shoaling coefficient, where $H$ and $C$ are the local wave height and celerity, computed at the breaking wave depth:

$$K_s = \frac{H}{H_0} = \sqrt{\frac{1}{2n}} \frac{C}{C_0}$$

**Wave Refraction**

As waves approach a coast obliquely, refraction bends the wave crests toward a more coast-parallel orientation. Wave crests can be significantly stretched (Fig. 2), allowing straight-crested offshore waves to distribute their power to a coastline whose shape is irregular and of greater length. Wave-crest stretching decreases wave height, thereby decreasing wave power. This effect is captured in a refraction coefficient, $K_R$, that further transforms offshore wave height:

$$K_R = \frac{H}{H_0} = \sqrt{\frac{S_0}{S}} = \sqrt{\frac{\cos \alpha_0}{\cos \alpha}}$$

where $S_0$ and $S$ are the wave-crest lengths between two wave rays in deep and shallow water, respectively, and $\alpha_0$ and $\alpha$ are angles between wave crests and depth contours in deep water and breaking-wave depth shallow water, respectively (Fig. 2). Incorporating both shoaling and refraction, wave height at the coast can be expressed as.

$$H = H_0 K_s K_R$$

The ratio of the delivery power of waves to their deep-water power simplifies to

$$\frac{P_D}{P_0} = K_R^2$$

**Energy Dissipation**

Energy dissipation by bottom interaction is dictated by both the depth at which waves begin to feel bottom and the path length over which dissipation occurs (Anderson et al, 1999). Two oceanographic variables dictate the location at which waves first begin interacting with the sea floor: The wavelength, $L$ (set by wave period, $T$) and the tide. The water depth, $h$, to which there is significant wave orbital motion is approximately $L/2$. Wave period influences $P_D$ by affecting the wavelength and therefore the water depth at which energy dissipation begins. From Airy wave theory, the wavelength is related to wave period through the dispersion equation (Komar, 1998):

$$L = \frac{g}{2\pi} T^2 \tanh\left(\frac{2\pi h}{L}\right)$$
where $h$ is the water depth. In deep water the hyperbolic tangent function approaches unity, and the water depth at which dissipation begins increases as the square of the wave period: $h_f = (g/4\pi)T^2$. Longer-period waves feel bottom earlier and should lose a larger fraction of their energy to bottom friction.

Waves approaching perpendicular to the bathymetric contours should lose the smallest fraction of their deep-water wave power. In addition, a lower shelf slope increases the ray-path length over which dissipation occurs. Ignoring refraction of the wave, the dissipative path length, $R$, varies inversely with both the slope of the shelf, $\theta$, and the angle between wave crests and bottom contours, $\alpha_0$ (Fig. 1), and goes as the square of the wave period:

$$R = \frac{h_f}{\sin(\theta)\cos(\alpha_0)} = \frac{gT^2}{4\pi\sin(\theta)\cos(\alpha_0)} \quad (8)$$

Increases in wave period and deep-water approach angle, and decreases in shelf slope, should lower the fraction of deep-water wave power reaching the coast.

Finally, tide affects water depth and therefore the offshore distance at which waves begin to dissipate energy (Trenhaile, 2000). At high tide, a deep-water wave travels farther unhindered by dissipative interaction with the bottom than at low tide and should result in greater energy imparted to the sea cliff (Fig. 1B). Finally, at low tide, waves break farther offshore, expending most of their energy in the surf zone, severely reducing the energy imparted to the cliffs. We seek quantification of these effects.
Wave Energy Delivery and the Shape of Rocky Coasts:
Part 2 – Microseismic Shaking of Sea Cliffs

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The morphology of a rocky coast along a tectonically active margin results from the interaction of uplifted resistant coastal bedrock and the destructive energy delivered to the coast by waves. Rocky coasts inspire interesting geomorphic questions about embayment shape, marine-terraces, and the relative roles of climate and lithology in coastline evolution? Lithology of coastal sea cliffs provides one control on rocky-coast evolution and offshore ocean climate provides another. Storm systems generate waves whose power is reduced by energy dissipation during shoaling, the remaining power being expended in the surf zone and at the sea cliffs (Komar, 1998). As a first step toward addressing rocky-coast evolution, we explore the utility of seismically sensed shaking of the sea cliffs as a measure of how offshore wave conditions, shelf bathymetry, and tides dictate the delivery of geomorphically useful energy to a rocky coast.

Deep-water waves begin losing energy through friction when their orbital motions extend to the sea floor (Fig. 1). Long-period waves and those occurring at low tide sense the sea floor further offshore and dissipate a greater fraction of deep-water energy than do short-period waves or those occurring at high tide. The deep-water swell direction and refraction of a wave’s ray path dictate the length of wave travel during which energy is dissipated. Wave power is also diminished by stretching of wave crests during refraction. In essence, deep-water wave power, \( P_0 \), is transformed into delivery power, \( P_D \), through a filter that depends on bathymetry and several oceanographic variables. Here we attempt to characterize this filter by using a novel method.

Seismologists have long recognized that microseisms complicate measurements of earthquakes (Longuet-Higgins, 1950; O’Hanlon, 2001). Standing waves on the shelf generate a seismic signal from the constructive interaction of waves reflected from the coast with incoming waves of the same period. Ground motions from the breaking of nearshore surf have also been noted. Distracting as microseisms may be to the seismologists, they are useful to oceanographers as a proxy for wave height, and they provide a record of ground motion in response to wave breaking. Zopf et al. (1976) demonstrated that microseisms could be used to measure wave heights when conventional pressure sensors are unavailable. Tillotson and Komar (1997) compared microseismically measured wave heights to those measured by buoys. Recently, researchers have employed historical seismographic records to hindcast (i.e., statistically predict past) changes in wave climate in the northeast Pacific (Bromirski et al., 1999) and in the North Atlantic (Grevemeyer et al., 2000). To date, however, few studies of coastal geomorphology have employed this valuable data source.

We assembled time series of wave heights and periods to characterize deep-water wave power, and swell directions and tidal elevations to calculate expected energy dissipation. These data were obtained from an NDBC buoy that records deep-water wave statistics and from a NOAA tidal gage. We compared time series of oceanographic variables with cliff shaking observations made with a portable broadband seismometer deployed at the edge of the sea cliff and coupled to the bedrock. We then attempted to define that combination of offshore wave climate and near-coast characteristics that best explains the shaking of the sea cliff.
Microseismic monitoring of wave-energy delivery to sea cliffs provides a rich data set against which to test theories of wave-energy dissipation. Given that seismologists working in coastal regions must commonly filter out the effects of waves, this is truly a case of one scientist’s noise being another’s signal. With a single stationary instrument, we were party to a natural experiment in which the effects of a wide set of oceanographic variables could be properly explored. Quantitative prediction of cliff shaking requires knowledge of these oceanographic variables and a model that accounts for (1) wave transformation due to shoaling and refraction and (2) dissipation through drag on the seafloor and through nearshore wave-breaking processes. We note that the tide strongly modulates the delivery of energy by controlling the location of wave break relative to the cliff. This experiment places on firmer footing any future modeling of long-term coastal evolution, including the generation of marine terraces and the embayment of coastlines.
Wave Energy Delivery and the Shape of Rocky Coasts:  
Part 3 – Strain of Sea Cliffs

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Steep sea cliffs of tectonically-active coastal landscapes are the product of wave-induced erosion. Rock uplift delivers fresh rock to the nearshore zone to be attacked by waves delivering their geomorphically effective energy. While one can sense qualitatively the energy of impact by waves crashing against a rocky coast, we have yet to fully understand the processes by which waves impart their energy to the shore and how this energy is transformed into erosion.

Traditionally, researchers have concluded that waves perform geomorphic work on rocky coasts by mechanical abrasion and by plucking or quarrying of blocks (Bagnold, 1939; Stephenson and Kirk, 2000; Sunamura, 1992; Trenhaile, 1987). Mechanical abrasion employs sediment grains, entrained by wave orbital motion, as tools to grind away the face of the sea cliff as the water of breaking waves impacts the vertical surface. Quarrying of blocks is thought to occur by hydraulic action when waves striking the sea cliff compress air in cliff face cracks. This exerts an outward stress on the surrounding rock mass which when repeated cause cracks to grow, ultimately detaching blocks (Sanders, 1968). Importantly, in this view the rate-limiting process is the growth of cracks by wave-induced hydraulic blasts. A major focus of rocky coastline geomorphic research is the documentation of the relative efficacy of each of these processes, and the detailed exploration of the physics of each process.

Ocean microseisms, first proposed by Weichert (1904), are ground motions generated by shallow water waves in coastal regions, and are subdivided into primary and secondary types (Bormann, 2002). Primary ocean microseisms involve the conversion of water wave-induced pressure variations to seismic energy, and hence have roughly the same period as the incoming water waves (Haubrich, 1963). Secondary ocean microseisms record pressure variations beneath a standing wave of half the period of the incoming ocean waves. This standing wave exists because the incoming waves are reflected back off the coast, causing superposition of waves traveling in opposite directions (Longuet-Higgens, 1950). This phenomenon was employed by Zopf et al. (1976) to measure ocean waves with a seismometer. The relationship linking microseismic energy and wave climate has been explored in detail by Bromirski and Duennebier (2002). Recently, several research groups have successfully hindcast ocean wave climates by examining records of long-term microseismic energy collected at coastal seismic stations (Tillotson and Komar, 1997; Bromirski et al., 1999; Bromirski, 2001; Grevemeyer et al., 2000).

To understand how changes in offshore wave climate modulate geomorphic energy delivery to a rocky coast, Adams et al. (2002) used a broadband seismometer to measure ground motions associated with waves impacting a sea cliff. The purpose of this previous study was to document how offshore wave conditions, shelf bathymetry and tides dictate the temporal pattern of energy delivery to the sea cliffs. Their study focused on velocity of ground motion, high pass filtering the data to explore the signal in the frequency band between 1Hz and the Nyquist frequency (50Hz) (Adams et al., 2002). Over the course of that several month-long study, a strong signal in the 0.05–0.1 Hz frequency band (10–20 second period) persisted throughout the time series, but was cautiously ignored, as it was not relevant to the outlined research objectives. In the present study reported here, we revisit this signal explicitly, in a process geomorphology context, asking the questions: (1) How is the long-period (10–20 sec, 0.05–0.1 Hz) ground motion signal related to the incident nearshore wave field? (2) What are the details of the motion exhibited by the seacliff at this frequency? (3) What, if any, strain is the seacliff experiencing at this frequency?
Rock uplift from slip on plate boundary faults (convergent or transform margin)

Lithologic contrast in seafloor rock strength

Wave energy imparted to seafloors

Water level fluctuations: short term (tidal) and long term (changes in eustatic sea level)

Role of sediment in littoral zone armoring seafloors

Wave transformation from deep water swell to nearshore surf

Deep water wave field with spectrum of wave heights, periods, and directions

Figure 1

Figure 2

Figure 3
Figure 4. An eight-day record from March 2001. A: Microseismic shaking at cliff edge (three components of ground motion). B: Tidal elevation. C: Offshore significant wave height. D: Swell direction. E: Wave period. Horizontal shaking is considerably stronger than vertical. Note strong correspondence between times of high shaking and times of high tide (shown with arrows) over an interval of large wave heights.

Figure 5. Modeled time series of power delivery for three periods of eight days each, along with microseismic shaking (dark solid line). Note different scales of shaking magnitude for the three plots. Simple model (light gray solid line) employs linear scaling of deep-water wave power, whereas advanced model (dark gray dashed line) incorporates wave shoaling and refraction, tide, dissipation from shelf drag, and temporally-dependent seismic attenuation.
Figure 6.

Figure 7.

Figure 8.

Figure 9.