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Shallow-Water Broadband OBS Seismology

by Spahr C. Webb and Wayne C. Crawford

Abstract The recent development of broadband ocean-bottom seismometers that can be deployed for more than a year has led to the construction of large ocean-bottom seismometer (OBS) fleets and to many successful experiments studying Earth structure and tectonics beneath the oceans. However, ocean surface waves raise noise levels at deep ocean-floor sites far above those at continental sites in the microseism band between 0.2 and 10 sec period, and currents and ocean waves raise noise levels at longer periods. Broadband OBSs are rarely deployed in shallow water because of a fear of loss due to bottom trawling and an expectation of very high noise levels from strong currents and the nearby ocean surface. However, these noise sources can be overcome such that shallow OBS deployments may provide noise levels that are comparable to deep-water sites at periods >10 sec and lower than deep-water sites at shorter periods. Burial of the instrument into the sediments can shield the seismometer from current noise, while the noise from deformation under wave loading can be removed using pressure gauge data. We predict the noise levels can be reduced to allow the detection of Rayleigh waves from 20 to 200 sec period with good signal-to-noise ratio (SNR) from teleseismic earthquakes as small as M_w 5. Short-period (<2 sec) noise levels will be 20–30 dB lower in shallow water than in deep water because short-period microseisms are greatly attenuated during propagation from deep to shallow water. Short-period (0.5–2 sec) teleseismic body waves should be detected with good SNR from events as small as M_w 4.5.

Introduction

Continental margins are some of the most interesting areas for potential seismological study, but there is little published data on seismic observations from broadband ocean-bottom seismometers deployed in shallow water (depths corresponding to the continental shelf or continental slope). Can useful broadband measurements be obtained from the continental margins? The characteristics of seismic noise in shallow water are expected to be quite different from the deep ocean. Deep-ocean sites are already more noisy than their continental counterparts because of a much stronger microseism peak, the noise from currents tilting the sensor, and motion of the seafloor under loading by long-period ocean waves (see Webb, 1998 for a review of deep ocean-floor seismology). While seismic noise at any site on Earth is dominated by the microseism peak near 0.15 Hz, the microseism peak on the deep seafloor is 20 to 30 dB higher than the typical continental site and extends higher in frequency, reaching at least up to 5 Hz. On the deep seafloor in the Pacific, the typical current is 5–10 cm/sec and is mostly tidal. This small current is sufficient to cause subtle tilting of any sensor deployed above the seafloor, which elevates long-period horizontal component noise levels up to 60 dB above the levels observed at good continental sites (e.g., Crawford *et al.*, 2006).

In shallower water, seafloor currents are expected to be stronger than in deep water, and the ocean surface that generates much of the other noise is much closer. However, this does not always mean that noise from ocean waves is higher. The microseism peak at shallow-water seafloor sites may be more similar to continental sites, and smaller in amplitude than deep-water sites, with relatively little energy at short periods (<3 sec). The other noise sources, currents and longer period ocean waves, will often be stronger in shallow water, but their effect can be removed through proper experiment design. Any seismic sensor deployed in shallow water must be buried into the sediments or otherwise shielded from seabed currents to avoid long-period flow noise. Strong currents can elevate short-period noise levels as well, but this noise has been primarily associated with current interacting with objects such as radio antennas that wobble the sensor and can be avoided by isolating the sensor from the rest of the instrument package (e.g., Trehu, 1985). With decreasing water depth, noise due to deformation of the seafloor under the loading of ocean waves becomes larger and affects higher frequencies, thus long-period (>10 sec) noise levels are expected to be higher at shallow-water sites than at deep-water sites. In 4400 m water depth, wave loading is detectable only at periods longer than about 35 sec (Crawford *et al.*,

2006) and is of small enough amplitude that it is usually ignored in seismological studies using OBSs. In 2500 m water depth, this noise can be detected to 30 sec period. At 700 m depth wave loading is detectable to periods as short as 20 sec and is found to reduce the signal-to-noise ratio (SNR) for 25 sec period Rayleigh waves by nearly 25 dB (Webb and Crawford, 1999). In this article, the seismic noise from loading under wave deformation is modeled for water depths as shallow as 50 m and compared with observations of vertical component spectra from three American west coast shallow-water (283–500 m) seafloor sites.

Webb and Crawford (1999) show pressure measurements can be used to measure the ocean waves causing this long-period seafloor deformation allowing the wave deformation noise to be predicted and removed from vertical component seismic data. Application of this method to three sites between 283 and 500 m depth results in 25 to 30 dB reductions in long-period vertical component noise levels. This method should be applicable to data from any water depth except right on the coast where the nonlinear components of the ocean waves become significant. However, seismic body waves in the band from 0.05–0.1 Hz may be more difficult to detect in shallow water because of a more energetic single frequency microseism peak generated during intervals when there is an energetic, local ocean swell.

Seafloor Pressure and Vertical Acceleration Spectra in Shallow Water

Deep seafloor spectra of vertical acceleration (Fig. 1a) are dominated by the main microseism peak from 0.1 to 5 Hz that is usually 20–40 dB more energetic than the microseism peak observed at land sites, including island sites. The second, smaller single frequency microseism peak from 0.05 to 0.1 Hz is observed to be of similar amplitude at deep seafloor sites as at many land sites. The small peak from 0.01 to 0.03 Hz seen in vertical component spectra from the deep seafloor is caused by deformation of the seafloor under loading by low-frequency ocean waves (infragravity waves). The deformation noise under wave loading is observable even for sensors installed a few hundred meters below the seafloor in a deep sea borehole (Araki *et al.*, 2004). A similar wave loading peak is seen in horizontal component noise spectra from borehole sensors as wave loading causes measurable tilting of the sensors in the borehole (Araki *et al.*, 2004). For horizontal component sensors on the seafloor, the wave loading noise is always hidden by high noise levels from flow noise due to tilting of the sensor package by ocean currents (Crawford and Webb, 2000; Crawford *et al.*, 2006).

The infragravity wave spectrum varies between ocean basins (Fig. 2b) with waves in the North Atlantic basin more variable and usually much lower amplitude than in the Pacific (Webb *et al.*, 1991). The Arctic basin spectrum is very quiet in winter because the complete ice cover prevents the generation of the ocean waves that are the source of microseisms and suppresses the infragravity waves that raise the long-period levels. These variations in the infragravity wave spectral levels between basins and over time will be mirrored in the levels of wave loading noise seen at seafloor stations.

Three pressure spectra (Fig. 2a) from the shelf north of San Diego (Fig. 3) show increasing spectral levels for the infragravity waves at long periods with decreasing depth.

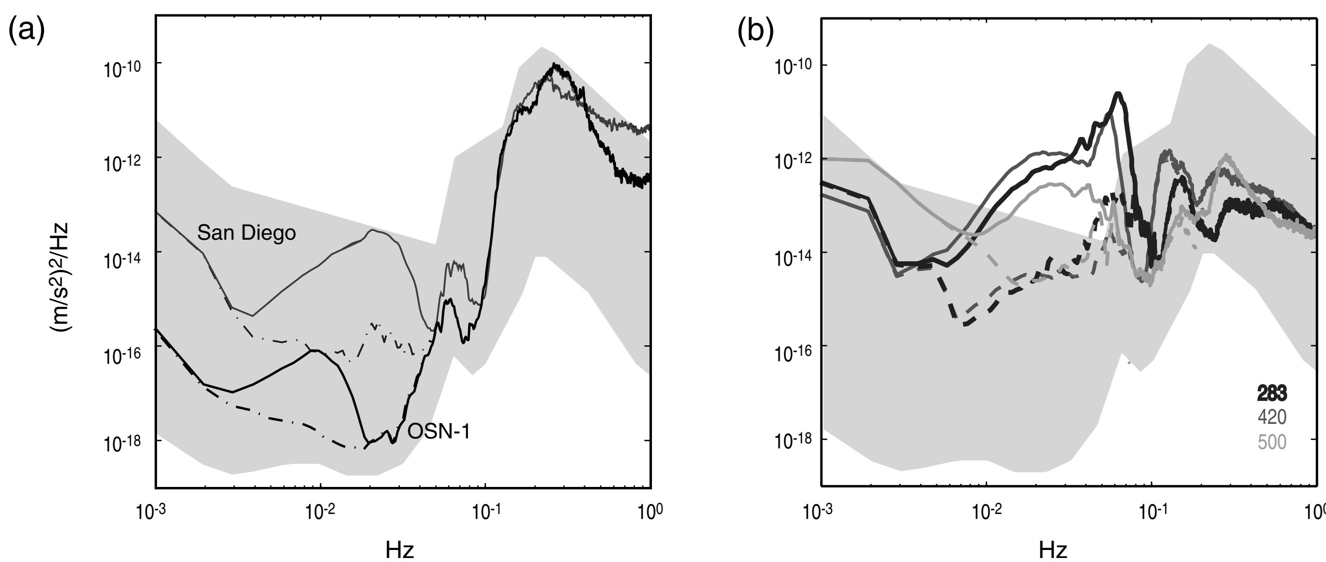

Figure 1. (a) Vertical acceleration spectra from the seafloor from 900 m depth from a site offshore of San Diego (see Crawford and Webb, 2000) and the OSN-1 site near Hawaii at 4400 m depth (Crawford *et al.*, 2006). Dashed lines show the same spectra with the wave deformation noise removed. The shaded area corresponds to the bounds of noise spectra observed at land sites (Peterson, 1993). (b) Vertical acceleration spectra from three sites offshore of San Diego in water depths of 283, 420, and 500 m (solid lines). Dashed lines show the same spectra after using pressure data to predict and remove the noise induced by wave loading of the seafloor. The 500 m site spectrum was noisy at the longest periods due to current.

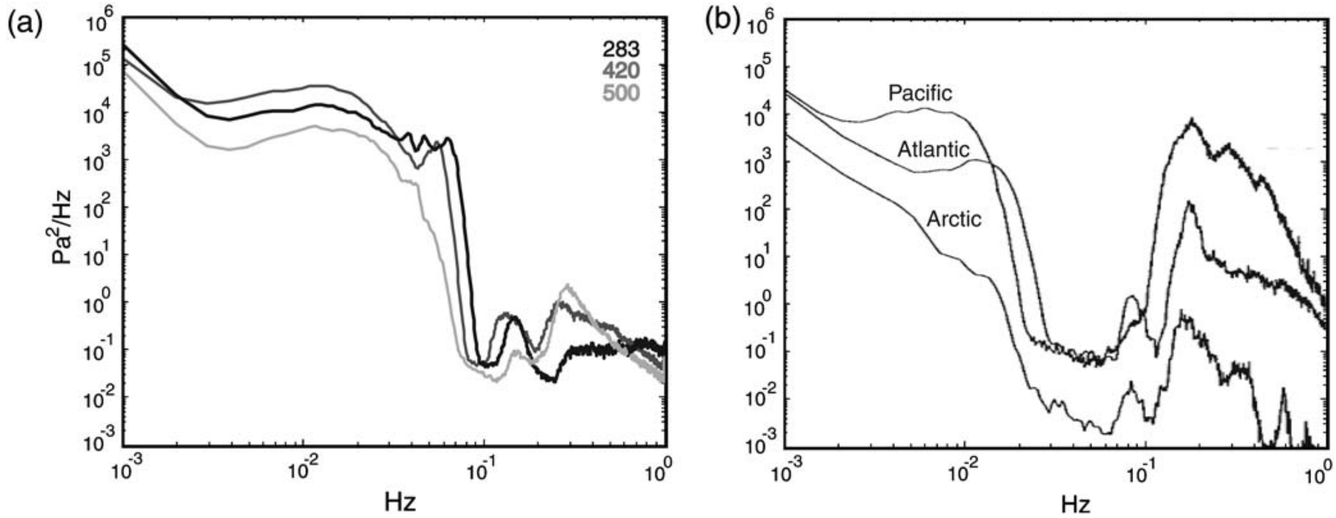

Figure 2. (a) The spectra of pressure at the three offshore San Diego sites corresponding to the acceleration spectra in Figure 1b. (b) Pressure spectra from three deep-water sites in the Pacific, North Atlantic, and Arctic oceans. The infragravity wave signal is lower in the Atlantic compared with the Pacific (Webb *et al.*, 1991). The Arctic is considerably quieter, both in the infragravity wave band (<0.03 Hz) and in the microseism peaks (0.05 to 0.1 and 0.1 to 5 Hz) because the ice cover greatly attenuates infragravity waves and prevents the propagation of the swell and wind-driven waves that are the source for the microseisms (Webb and Schultz, 1992). The multiple peaks in the microseism band at 0.15, 0.6, 0.9 Hz in the Arctic data are associated with different Rayleigh wave modes.

The wave deformation induced peak extends to higher frequency in shallower water because the pressure signals from shorter wavelength waves reach the seafloor in shallower water. These spectra show infragravity amplitudes increase toward shallower water, although some of the difference in spectral level is due to day-to-day changes in the height of swell incident on the coast, the primary source of the infragravity wave energy. Energy transport is roughly conserved during propagation into shallow water, requiring wave amplitudes to increase as the inverse of the fourth root of depth. Spectral levels at 500 m are expected to be about twice the levels seen at 2000 m depth and 75% of the levels seen at 283 m. (The infragravity wave spectrum at the 420 m site is more energetic between 0.003 and 0.07 Hz than the spectrum at the 283 m because the 420 m data are from a day with a more energetic local wave climate). An additional effect increasing long-period wave ocean energy near the coast is the trapping of locally generated infragravity wave energy within the waveguide formed by the sloping bathymetry. Outward traveling waves propagating other than normal to

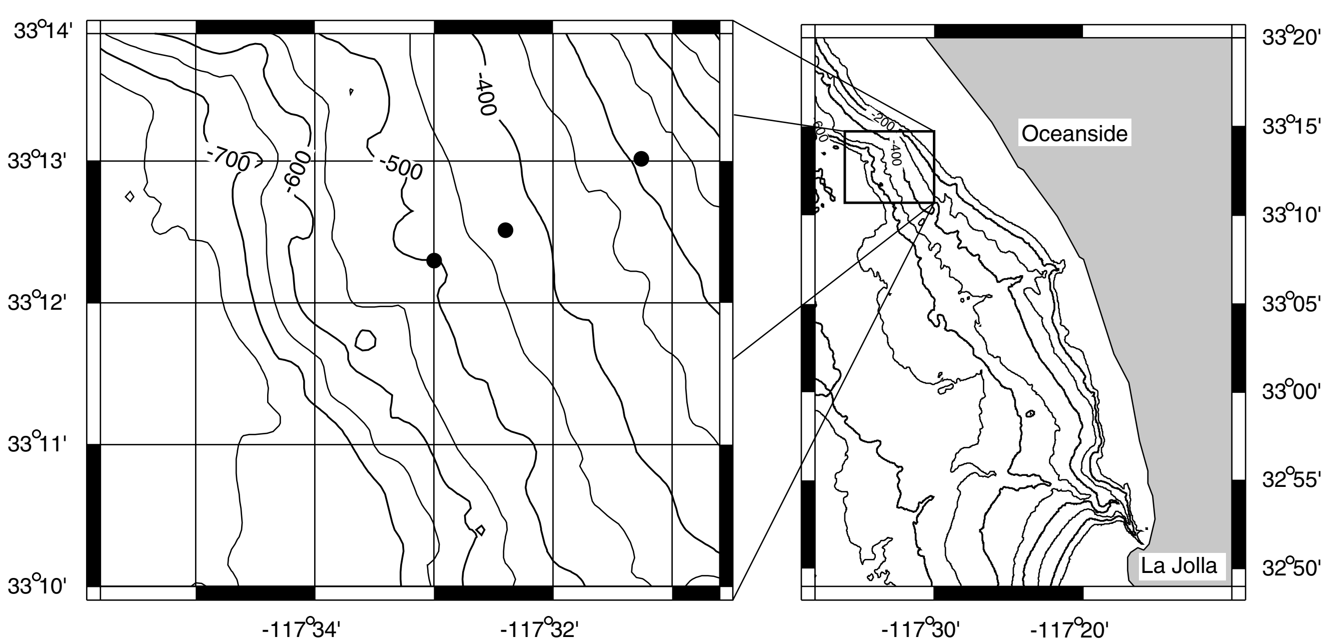

Figure 3. The three sites offshore of San Diego are shown.

the coast are refracted back toward the coast because of increasing wave speed with depth, thus elevating infragravity wave levels near the coast (Okihiro *et al.*, 1992).

Models of Seafloor Spectra as a Function of Depth

The amplitude of the vertical deformation under wave loading depends on the elastic structure beneath the seafloor (Webb and Crawford, 1999). One can model the expected seafloor deformation (Fig. 4b) as a function of water depth starting with a model for the transfer function and a model for the ocean wave height spectrum. We start with a model for the infragravity wave spectrum that rises at very long periods (below 0.002 Hz) and that is flat at higher frequencies (Fig. 4a). A wave height spectrum of 10^{-4} m^2/Hz corresponds to a near-surface pressure spectrum of about 10^4 Pa^2/Hz. This is a good approximation to deep-water Pacific pressure observations (Filloux, 1983; Webb *et al.*, 1991). The infragravity wave component of the spectrum is expected to increase toward shallower water and to vary from day to day in shallow water (as reflected in the observations in Fig. 2b), but this model is a useful starting point. We add to the wave spectrum a long-period swell peak near 0.06 Hz corresponding to a 16 sec period swell with a significant wave height of 5 cm. Swell amplitudes and periods will also vary greatly from day to day; this peak was chosen to roughly match the seafloor spectrum observed at the site at 283 m depth (Fig. 5b). We also add to the model (Fig. 4a) a wind-driven wave field using a Joint North Sea Wave Project (JONSWAP) style wave model (Hasselmann *et al.*, 1973) with a significant wave height near 2 m. The long-period (0.06 Hz) swell is not significant at sites deeper than 500 m and the wind-driven wave component of this model cannot be detected at depths deeper than about 200 m.

We estimate seafloor motions under the ocean waves at different depths using a linear fit to the transfer function recorded at the 283 water depth site (Fig. 5c). For all water depths only the infragravity wave component below 0.04 Hz (25 sec period) is relevant to the detection of teleseismic Rayleigh waves. Because the elastic moduli change only slowly with depth below the seafloor, this transfer function resembles that of an elastic half-space, which increases linearly with frequency. A flat ocean wave pressure spectrum between 0.002 Hz and 0.06 Hz (Fig. 4a, 5b) therefore results in a steeply increasing vertical acceleration spectrum with frequency in this same band. At shorter periods, water depth significantly affects the shape of the predicted wave loading noise spectrum. Noise levels in the single frequency microseism band (0.05–0.1 Hz) are predicted to increase by more than 40 dB between shallow and deep sites from wave loading noise. In water depths less than 200 m, the noise near 0.15 Hz from wave loading by swell can be much larger than the normal microseism peak.

The wave deformation signal is not important at periods shorter than 3 sec (0.33 Hz) at any site deeper than 50 m. (It probably makes no sense to make seismic measurements at water depths less than about 50 m because of wave loading noise). Short-period noise at the seafloor will be associated with microseisms propagating primarily as Rayleigh waves. In deep water, noise levels are very high near 1 Hz (Fig. 1a), but the shallow-water measurements from near La Jolla, California (Fig. 1b), show 1 Hz noise levels that are much lower than typical deep-water noise levels. The spectral levels observed at these three offshore sites are very similar in the microseism band (0.1–5 Hz) to spectra measured onshore in La Jolla (e.g., fig. 8 from Webb *et al.*, 2001).

Removing Deformation Noise from the Vertical Component

Long-period (>20 sec) ocean-floor vertical component seismic noise levels can be reduced substantially by predicting and removing the wave loading noise using pressure measurements. The vertical acceleration is related to the

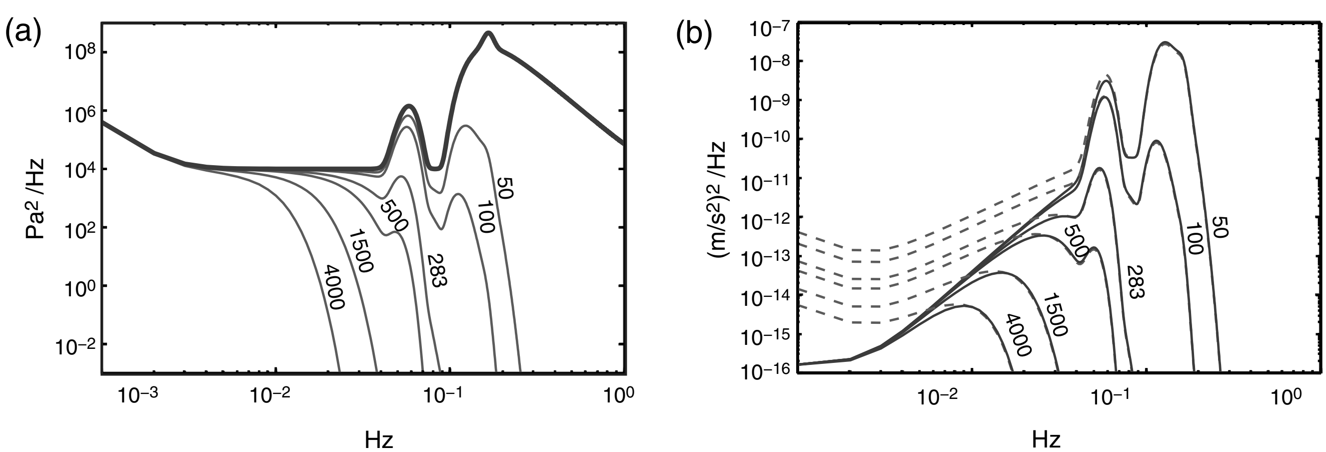

Figure 4. (a) A model for the pressure spectrum at the sea surface including the long-period (infragravity) waves, a small peak corresponding to a 16 sec swell, and an energetic wind-driven wave spectrum at higher frequencies (heavy line). Gray curves show the predicted pressure spectrum at the seafloor in different water depths. (b) Models for the vertical (solid lines) and horizontal (dashed) acceleration spectra expected at the seafloor due to deformation under wave loading in this wave model for different water depths.

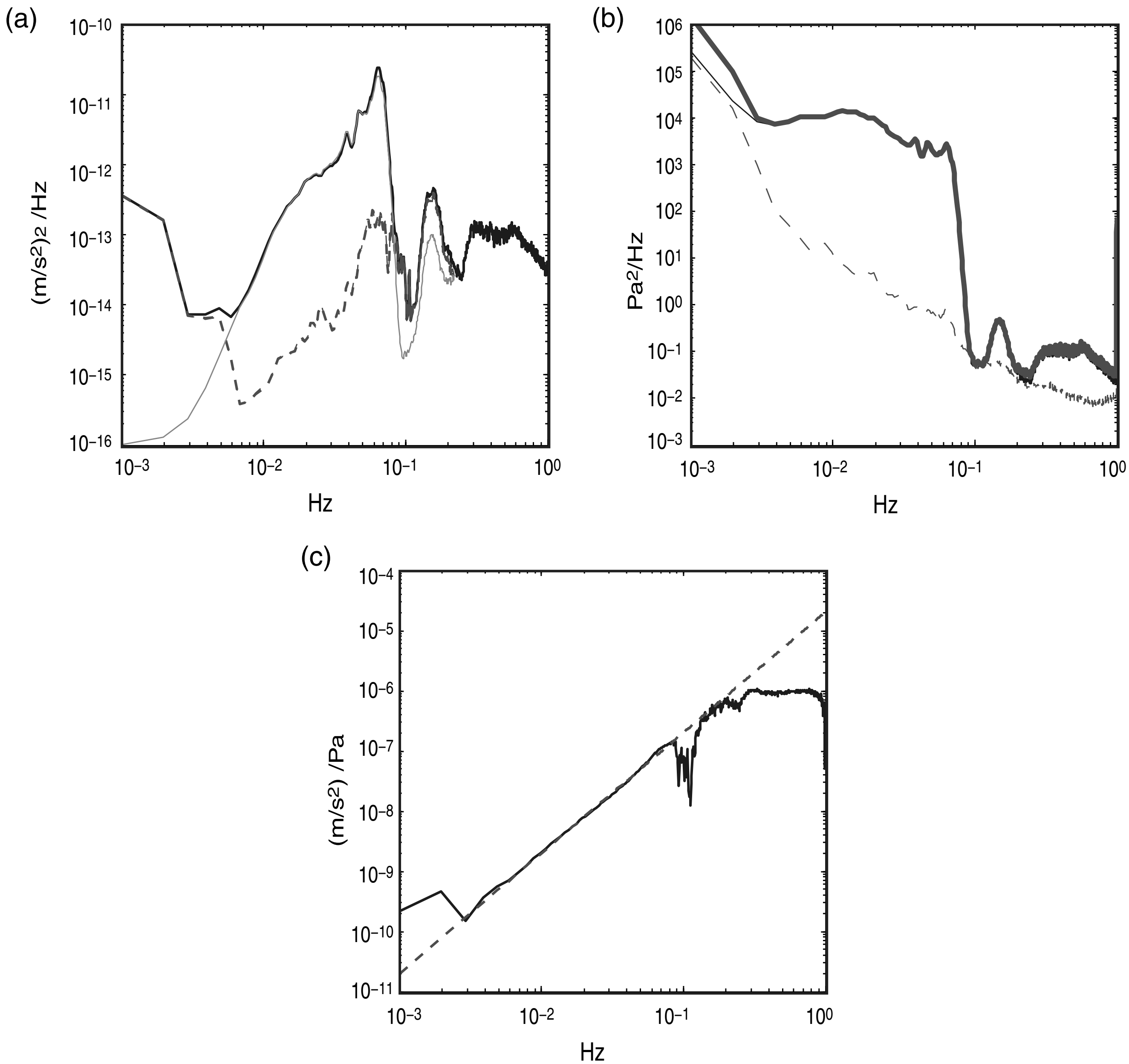

Figure 5. (a) The vertical acceleration spectrum at the 283 m site (solid), and the spectrum after removal of the wave induced deformation using the measured transfer function between pressure and vertical acceleration at this site (dashed). The predicted deformation acceleration spectrum using the simple linear model for the transfer function (gray). (b) The pressure spectrum observed at the same site by two different pressure sensors (solid and light gray), and noise levels for these pressure sensors (dashed line). (c) Measured transfer function between pressure and vertical acceleration from the same data (solid line). A simple linear fit to the same data (dashed line).

pressure signal through the transfer function. A few days of data are sufficient to measure this transfer function to an accuracy of about 1%. A digital filter constructed from this transfer function when applied to the pressure data accurately predicts the deformation noise so that this noise can be subtracted from the vertical acceleration record and the noise removed from the record (Fig. 1b, 5a). Figure 5a shows that between 25 and 30 dB improvement in noise levels is obtained for the site in 283 m depth for frequencies between 0.01 and 0.08 Hz. Webb and Crawford (1999) similarly demonstrated a 25 dB improvement in SNR at the 25 sec period at a site at 700 m depth using this method. Crawford *et al.* (2006) obtained 35–40 dB reduction in long-period noise at a site on the seafloor at 4400 m depth, and a 15–25 dB reduction in 25 sec period noise levels for a borehole sensor data at the same site.

Application of this technique to remove the deformation induced noise to data from the three sites offshore of San Diego results in noise level reductions from between 20 and 30 dB in the band from 0.006 to 0.07 Hz (Fig. 1b).

(The 500 m depth site shows high noise levels below 0.01 Hz that are thought to be due to ocean current). Achieving the low vertical component noise levels at long periods shown in these figures from instruments deployed directly onto the seafloor (without shielding or burial) requires first removing tilt noise from currents from the vertical component (Crawford and Webb, 2000). The vertical component includes a tilt noise component, because vertical component sensors are never aligned perfectly with true vertical. A 0.5° error in leveling results in roughly 1% of the tilt noise variance measured on the horizontal components being rotated into the vertical component. The process used to remove the tilt noise from the vertical component is equivalent to determining the rotation matrix necessary to rotate the components such that the vertical component is perfectly vertical. This results in typically a fraction of 1% of the horizontal component signals being added or subtracted from the vertical component, but because the horizontal component noise levels are often 50 dB or more noisier than the vertical components because of tilt noise, this subtraction produces a significant difference in perceived vertical component noise levels.

It is not yet known what the limit is for removing the deformation (and tilt) induced signals from vertical component data from the seafloor. It is likely that the noise reductions shown here (Fig. 1a) are limited by statistical uncertainty because of the relatively short (a few days long) data sets used to calculate the transfer functions. As noted previously, Crawford *et al.* (2006) demonstrated up to 40 dB reduction in long-period noise level for data from the OSN-1 (ocean seismic network) site using a longer data set than those used here. Oceanographic signals other than infragravity waves produce pressure signals that may ultimately limit this noise removal process.

Horizontal Deformation under Wave Loading

The loading of the seafloor by ocean waves also causes noise on the horizontal components of a seismometer at the seafloor, although this signal is obscured by flow noise for sensors not buried beneath the seafloor or placed in boreholes. The horizontal displacements are smaller than the vertical deformation under wave loading by roughly the square of the ratio of the shear velocity to the compressional velocity near the seafloor (Crawford, 2004); however, the noise from tilting of the seafloor under wave loading is usually much larger because the tilting rotates the acceleration of gravity into the horizontal components (Araki *et al.*, 2004).

The horizontal acceleration a_h signal beneath a propagating plane wave is approximately equal to:

$$|a_h| \approx |a_v|[1/3 + gk/\omega^2], \tag{1}$$

where a_v is the vertical acceleration due to the deformation if we assume a shear velocity to compressional velocity ratio of $(1/3)^{1/2}$. The first term corresponds to the horizontal acceleration directly related to the wave loading, the second term to the apparent acceleration associated with tilting of the sensors. This second term asymptotes to one at high frequencies corresponding to the dispersion relation for ocean waves in deep water ($\omega^2 = gk$), and at low frequencies to $g/(\omega\sqrt{gh})$, corresponding to the shallow-water limit of the dispersion relation $\omega/k = \sqrt{gh}$. The horizontal component deformation signal is $+90°$ or $-90°$ out of phase with the pressure signal, depending on the direction of propagation.

The infragravity wave field consists of waves traveling at many different azimuths and is best described by its directional spectrum as a function of frequency. The vertical acceleration spectrum A_V is related to the seafloor pressure spectrum $P(\omega)$ by the vertical compliance η (Crawford *et al.*, 1998). The E, N, and vertical acceleration spectral components of the deformation noise are related to the near-surface ocean wave directional pressure spectrum G:

$$\begin{aligned}
A_E(\omega) &= \eta^2(\omega)[1/3 + gk/\omega^2]^2 \int_0^{2\pi} G(\omega,\theta)\cos^2(\theta)d\theta \\
A_N(\omega) &= \eta^2(\omega)[1/3 + gk/\omega^2]^2 \int_0^{2\pi} G(\omega,\theta)\sin^2(\theta)d\theta \\
A_V(\omega) &= \eta^2(\omega)P(\omega) \\
P(\omega) &= \int_0^{2\pi} G(\omega,\theta)d\theta.
\end{aligned} \tag{2}$$

In Figure 4b the directional spectrum of the infragravity waves is assumed to be isotropic in direction, so that the integrals over the directional spectrum generate a factor of ½. Isotropy may be a good model for infragravity waves in deep water (Webb *et al.*, 1991), but is likely to be a poor model in shallow water where refraction will tend to turn wave propagation perpendicular to the slope of the seafloor. One should expect in shallow water that the cross shore noise levels will be higher than the along shore noise levels, but cross shore noise levels should not exceed the spectra shown in Figure 4b by more than a factor of 2. The wave loading noise is larger on the horizontal components (dashed lines, Fig. 4b) than on the vertical component at all depths at long periods because of the large effect of tilting of the seafloor under loading.

It will not be possible to use a single pressure gauge to remove the deformation noise from the horizontal components unless a single direction of propagation is dominant such that the wavefield at any frequency is essentially a single-plane wave. A possible solution to this problem would be to measure two orthogonal horizontal components of pressure gradient along the seafloor by differencing measurements between pairs of pressure gauges separated by short distances along the seafloor. The transfer functions between pressure gradient and horizontal acceleration can then be measured and the horizontal component deformation noise removed using the same method as for the vertical component. The SNR for these pressure gradient measurements will depend on the frequency, water depth, and distance of separation between gauges. A few meters of separation should be adequate for shallow water (100 m) and for frequencies above 0.03 Hz, based on the noise levels of seafloor pressure gauges.

The Microseism Peak in Shallow Water

The pressure perturbation from Rayleigh waves become negligible at the free surface because of the small acoustic impedance of the atmosphere, whereas the free surface is a local maximum for vertical acceleration. Thus, the microseism peak (0.1 and 1 Hz) becomes smaller in shallower water when resolved in pressure spectra in water depths that are a small fraction of an acoustic wavelength (Fig. 2a), while vertical acceleration spectra are not expected to show a similar effect. However, the main microseism peak in vertical acceleration spectra (Fig. 1b) from the three shallow-water sites offshore of San Diego (Fig. 4) is 20 to 30 dB lower than the microseism peak in spectra from deep-water Pacific sites (Fig. 1a). In the deep ocean, microseisms longer than about 4 sec period propagate primarily as fundamental mode Rayleigh waves, whereas higher frequency microseisms are mostly higher mode waves (Webb and Cox, 1986; Webb, 1992). These waves are attenuated during propagation onto shore such that noise levels near 1 Hz at coastal continental sites are much lower than observations from the deep seafloor. Wilcock *et al.* (1999) found noise levels onshore in Iceland near 1 Hz in the windy North Atlantic were between 10 and 20 dB lower than typical spectral levels observed during the Melt experiment in the tropical Pacific. Spectra from shallow water offshore of San Diego show similar noise levels as Iceland near 1 Hz. The low noise levels could be a consequence of unusually quiet wind conditions; however, this seems an unlikely explanation despite the mild wind climate of near shore San Diego. Microseisms near 1 Hz are forced by 2 sec period ocean waves. The ocean wave spectrum at this period saturates at low wind velocities leading to a saturated microseism spectrum near 1 Hz virtually all the time in deep water (Webb 1992; Webb, 1998).

The diminution of the microseism peak during propagation from deep water to shallow water is likely a consequence of faster phase and group velocities and a deeper extent to the Rayleigh mode eigenfunctions in shallower water for frequencies >0.1 Hz. The energy in microseisms in the deep ocean is carried primarily by Rayleigh waves with most of the energy carried within the ocean layer. Hasselmann (1963) shows the energy in the microseism peak builds up during propagation of these Rayleigh waves across broad regions of the ocean. Webb (1992) predicts the microseism spectrum eventually reaches a saturation spectrum (at least at short periods), for which the excitation from the ocean surface is balanced by dissipation within the Rayleigh waves. (A correction due to an omitted term discovered by Tanimoto, 2007 may explain the discrepancy seen between the predicted and observed microseism spectrum at short periods in the modeling of Webb, 1992).

Latham and Sutton (1966) found microseism amplitudes observed on the deep (4280 m) seafloor near Bermuda were much larger than the amplitudes measured at a station onshore in Bermuda. Using calculations of the energy density and group velocity for models of Earth structure at the two sites, they showed this difference in vertical acceleration of the ground was consistent with a similar flux of microseism energy at the two stations, assuming the energy was propagating as fundamental mode Rayleigh waves. The vertical accelerations at the surface required to enable the same energy flux in the Rayleigh was lower at the onshore site, both because the Rayleigh wave group velocity was faster onshore and because the energy density per unit area in the Rayleigh wave was higher for a given vertical acceleration at the surface.

Deep ocean microseisms near 0.2 Hz propagate at phase and group speeds near 1.5 km/sec or less (e.g., Chiaruttini *et al.*, 1985). Thinning of the water layer results in faster group and phase velocities for Rayleigh waves (unless there is very large thickening of the sediment layer on the shelf). Rayleigh wave surface displacements must diminish to conserve energy flux (e.g., Tromp and Dahlen, 1992) because the phase and group velocities, wavelength, and depth extent of the eigenfunction all increase as waves propagate into shallower water.

In addition, some microseism energy should be refracted back out to deep water if the phase velocity increases for Rayleigh waves propagating obliquely into the coast. The literature of surface wave propagation across continental shelf boundaries has focussed to date only on wave propagation at periods longer than 10 sec, but large refraction effects are expected near 10 sec period (Meier and Malinschewsky, 2000). Some conversion between Love and Rayleigh wave modes may also occur for obliquely incident waves (Gregersen and Alsop, 1976).

The wave–wave interaction mechanism that forces the microseisms produces a more energetic pressure spectrum near the sea surface (Cox and Jacobs, 1989) and in very shallow water (Herbers and Guza, 1992). This strong coastal source, however, acts over only a small distance (and area) compared with the sources over the deep ocean; thus, deep ocean sources should control the peak energy in the microseism peak. More observations will be required with sensors both in shallow and deep water to confirm whether this statement applies to all frequencies within the microseism peak at shallow-water sites.

Our observations (Fig. 1) show that short-period noise levels at shallow-water sites can be much lower than at typical deep-water sites. It seems likely that microseism peak noise levels on the shelf will usually be comparable to noise levels at nearby coastal seismic sites, leading to much lower detection thresholds for short-period arrivals for stations on the shelf compared with adjacent deep-water sites.

The single frequency microseism peak between 0.05 and 0.1 Hz at deep-water sites is usually at most 10 dB more energetic than that observed at typical land sites (Fig. 1a). The single frequency peak is still apparent in the shallow seafloor vertical component spectra after the wave deformation noise is removed (Fig. 5a). The single frequency peaks in the shallow-water spectra are all very energetic (Fig. 1b). In one spectrum it exceeds 3×10^{-13} $(\mathrm{m/sec}^2)^2/\mathrm{Hz}$ (after removal of the wave loading noise), whereas at continental sites the peak is rarely above10^{-15} $(\mathrm{m/sec^2})^2/\mathrm{Hz}$). This energetic noise is likely propagating as seismic waves rather than being the direct result of ocean wave loading because otherwise it should have been removed with the longer period ocean wave noise. Agnew and Berger (1978) observed elevated noise levels at La Jolla in this band that they attributed to edge wave loading causing flexure of the coast. They also observed that this noise decreased rapidly with distance from the coast. It is possible that the higher noise level in these data in the single frequency peak is associated with this same process. Adams *et al.* (2005) observed flexure of coastal cliffs due to waves striking a cliff, which may be a related process. Further observations are required to determine whether these high noise levels in the 0.05 to 0.1 Hz band are a consequence of shallow water or just related to the proximity to the coastline.

Vertical Component Detection Limits for Seismic Phases at Shallow-Water Sites

Long-period noise levels are typically higher at shallow-water sites than at deep sites, while short-period noise levels should be lower. Detection thresholds for surface waves and body waves at shallow-water sites can be estimated by comparing the measured noise amplitudes in narrow frequency bands to the expected signal levels in these same bands. Figure 6 compares the amplitude of the noise in 1/3 octave bands to expected signal levels for arrivals from teleseismic (30° range) events (following Webb, 1998 and Agnew *et al.*, 1986). The model predicts a detection threshold for Rayleigh waves at the 283 m depth site of about M_{w} 6 (assuming a SNR = 4) for the vertical component before processing to

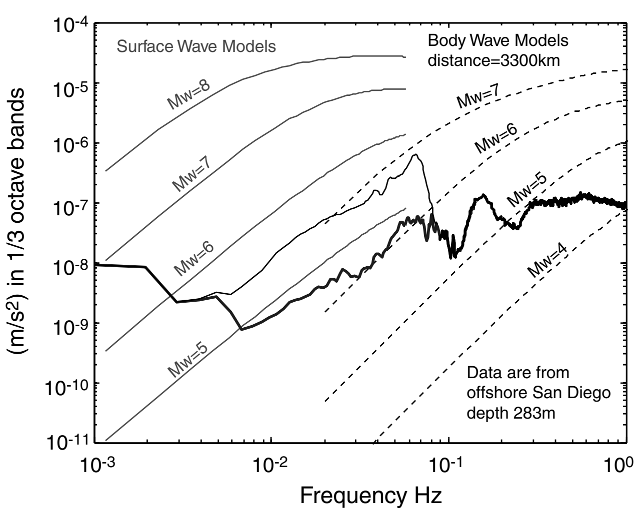

Figure 6. Modeled surface-wave and body-wave amplitudes for teleseismic earthquakes at a distance of 30° are compared with the amplitude of the noise in 1/3 octave bands for the vertical acceleration spectrum from 283 m depth before (thin line) and after processing to remove noise due to deformation under ocean wave loading (heavy line).

remove the wave deformation signal and a detection level of about M_{w} 5 for the vertical component with the deformation noise removed.

Detection thresholds for body waves near 0.08 Hz are about M_{w} 7.5 for the unprocessed data and M_{w} 6.5 for the processed data (with noise removed), suggesting one should expect high detection thresholds for long-period body waves as determined by the high noise levels observed between 0.05 and 0.1 Hz.

Short-period (1 Hz) detection thresholds are quite low ($<4.5\,M_{\mathrm{w}}$) as expected given the observation that noise levels were comparable to coastal sites on land and much quieter than deep-water sites due to lower microseism noise levels near 1 Hz.

Conclusions

Useful seismic observations of Rayleigh waves and body waves can be obtained from seismometers installed on the continental shelves at depths as shallow as 50 m, but it will be necessary to either bury seismic sensors or shield sensors action of the ocean-floor currents to obtain adequate SNR for these signals. Long-period ocean waves raise noise levels by deforming the seafloor, but this noise can be removed from vertical component data sufficiently using pressure observations to allow detection of Rayleigh waves from teleseismic events as small as about M_{w} 5. The noise levels above 0.2 Hz associated with the microseism peak in shallow water will be smaller than those observed at deep seafloor sites and more comparable to coastal onshore seismic sites, permitting the detection of short-period body waves from teleseismic events as small as M_{w} 4.5. Long-period horizontal component data will be very noisy unless it is possible to remove wave loading noise from these components using seafloor pressure gradient observations.

Data and Resources

The seismic data shown here can be obtained by contacting Wayne Crawford.

Acknowledgments

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