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Predicting short-period, wind-wave-generated seismic noise in coastal regions

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ABSTRACT

Substantial effort has recently been made to predict seismic energy caused by ocean waves in the 4–10 s period range. However, little work has been devoted to predict shorter period seismic waves recorded in coastal regions. Here we present an analytical framework that relates the signature of seismic noise recorded at 0.6–2 s periods (0.5–1.5 Hz frequencies) in coastal regions with deep-ocean wave properties. Constraints on key model parameters such as seismic attenuation and ocean wave directionality are provided by jointly analyzing ocean-floor acoustic noise and seismic noise measurements. We show that 0.6–2 s seismic noise can be consistently predicted over the entire year. The seismic noise recorded in this period range is mostly caused by local wind-waves, i.e. by wind-waves occurring within about 2000 km of the seismic station. Our analysis also shows that the fraction of ocean waves traveling in nearly opposite directions is orders of magnitude smaller than previously suggested for wind-waves, does not depend strongly on wind speed as previously proposed, and instead may depend weakly on the heterogeneity of the wind field. This study suggests that wind-wave conditions can be studied in detail from seismic observations, including under specific conditions such as in the presence of sea ice.

1. Introduction

Ocean waves generate seismic waves either directly by interacting with the shoreline (primary microseisms, Hasselmann, 1963; Ardhuin et al., 2015) or indirectly through wave–wave interactions (secondary microseisms, Longuet-Higgins, 1950; Hasselmann, 1963; Ardhuin et al., 2013). It is well known that these two different processes cause two distinct peaks in seismic noise spectra with maximum amplitudes at periods of 8–16 s for primary microseisms and of 4–8 s for secondary microseisms (McNamara and Buland, 2004; Berger et al., 2004).

Microseisms are particularly useful for studying Earth structure using noise cross correlation techniques (Campillo and Paul, 2003; Bowden et al., 2015), and better knowledge of their characteristics is needed to avoid spurious artifacts caused by spatial and temporal variations in noise sources (Tsai, 2009; Fichtner, 2014). Numerous investigations have been conducted to better understand the ocean processes that create microseismic noise and their spatiotemporal characteristics. Maximum amplitudes of the secondary microseism peak (4–8 s) have been successfully predicted (Kedar et al., 2008; Ardhuin et al., 2015) by combining numerical ocean wave models with the Longuet-Higgins (LH) theory (Longuet-Higgins and Ursell, 1948; Longuet-Higgins, 1950), as later revisited by Hasselmann (1963). The maximum energy in that period range is mainly caused by strong ocean swell populations with periods of typically 8–16 s, i.e. wavelengths of 100–400 m, that travel in nearly opposite directions either in coastal or deep-ocean regions as a result of their generation by distant storms, by single but fast moving storms or by coastline reflections.

In contrast to these previous findings, little attention has been devoted to understanding how ocean processes cause the relatively shorter period (<4 s) seismic noise discussed in various recent studies (Zhang et al., 2009; Tsai and McNamara, 2011; Beucler et al., 2014). In contrast to the longer (4–8 s) periods at which secondary microseisms are observed almost everywhere in continental areas (Berger et al., 2004), shorter period ocean-induced noise is expected to be more restricted to coastal regions since seismic waves are more attenuated at these shorter periods. However, short-period ocean noise is increasingly used for high-resolution imaging of shallow Earth structure in coastal regions (Lin et al., 2013; Bowden et al., 2015). Moreover, Tsai and McNamara (2011) suggested that sea-ice mechanical properties could be continuously monitored from the analysis of coastal ground mo-

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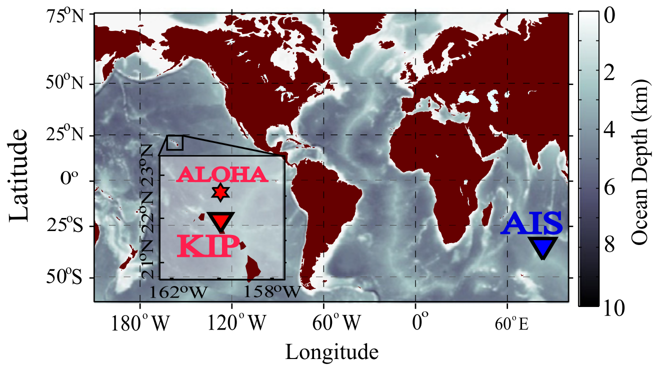

Fig. 1. Locations of the seismic stations KIP and AIS (triangles) and the marine hydrophone station ALOHA (star) used in this study. The color scale indicates ocean depth. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

tion in the 0.6–2 s period band. Before such a goal can be achieved, though, an accurate understanding of coastal seismic noise amplitude and frequency scaling under sea-ice free conditions is needed: this is the main purpose of this study.

As in Webb (1992), Farrell and Munk (2008) and Duennebier et al. (2012), we adopt a simple analytical approach that predicts secondary microseisms from the interaction of short-wavelength (1–25 m) wind-waves. In contrast to these previous studies who limited their analysis to the modeling of acoustic pressure and ground floor displacement at the ocean bottom, we conjointly model pressure spectra recorded at the ocean bottom together with acceleration spectra recorded by seismic stations in coastal areas. This joint analysis allows us to independently constrain the key ocean-wave and seismic model parameters through their control on the amplitude and variability of acoustic and seismic noise records.

2. Data

We use seismic data from the 2 stations KIP (Oahu, Hawaii; Pacific Ocean) and AIS (Amsterdam Island; Indian Ocean) shown in Fig. 1. For simplicity, we only consider these two island stations surrounded by 4–6 km deep ocean, where model predictions are relatively insensitive to uncertainties in ground properties (see Section 3.2.2). However, we also expect our modeling framework to apply to continental stations with shallow ocean nearby.

Acoustic noise records are taken from Duennebier et al. (2012), who reported broadband hydrophone measurements at the ALOHA Cabled Observatory, 100 km north of Oahu, Hawaii (see red star in Fig. 1). This station is located near the KIP seismic station, so that our acoustic and seismic noise predictions can be done jointly at this location. Details on the deployment and signal acquisition at station ALOHA are provided by Duennebier et al. (2008, 2012). We use frequency spectra that were processed by Duennebier et al. (2012) over the 20 months of continuous acoustic noise records acquired from February 2007 to October 2008. As in Duennebier et al. (2012), we consider average spectra that have been sorted by local-wind speed, which was independently measured above the ALOHA station by the WHOTS meteorological buoy (see Plueddemann et al., 2006).

We use estimates of near-surface wind speeds provided by the ERA-Interim dataset of the ECMWF (Dee et al., 2011). This model simulation includes 12-h assimilations of observations with 3-hourly model outputs on a regular grid with a 0.7° horizontal resolution. Finally, for ocean depths, we use the bathymetry map ETOPO2 provided by the NOAA data center (http://www.ngdc.noaa.gov) with a 2-minute latitude and longitude resolution.

3. Model

In this section, we calculate the ground acceleration power spectral density (PSD) $A(f_s)$ defined at seismic frequency f_s and over a given time window of duration T as

$$A(f_s) = \frac{1}{T}\left(\int_0^T a(t)e^{-2\pi i f_s t}dt\right)^2 \tag{1}$$

where $a(t)$ is the ground acceleration timeseries. For the <4 s periods of interest, the ocean surface gravity waves (OSGW) that cause the observed ground motion have wavelengths (<25 m) that are much shorter than ocean depths such that the deep water approximation is appropriate. For such 'deep water' conditions, ocean-surface pressure fluctuations are thought to generate seismic surface waves only from the interaction of OSGW pairs. Any interacting OSGW pairs with wavenumber vectors $\mathbf{k}$ and $\mathbf{k}'$, and associated frequencies f and f', generate a resultant wave of horizontal wave number $\mathbf{K} = \mathbf{k} + \mathbf{k}'$. Of all wave types resulting from all possibly interacting OSGW pairs, only those that satisfy $|\mathbf{K}| \approx 0$, i.e. $\mathbf{k} \approx -\mathbf{k}'$, and consequently $f \approx f'$ contribute to seismic wave generation in deep water (Longuet-Higgins, 1950; Hasselmann, 1963; Ardhuin and Herbers, 2013). The frequency of the surface forcing is $f_s = f + f' \approx 2f$ and its amplitude is proportional to the amplitude and fraction of nearly-oppositely traveling wave pairs within the broad OSGW spectrum $E(f,\theta) = E(f)M(f,\theta)$ (with dimension $\mathrm{m^2\,Hz^{-1}}$), where $E(f)$ is the ocean surface wave elevation PSD and $M(f,\theta)$ is the directional distribution of OSGWs that depends on azimuth θ, and satisfies $\int_{-\pi}^{\pi} M(f,\theta)d\theta = 1$ (Mitsuyasu et al., 1975; Ewans, 1998). The fraction of interacting wave pairs can be represented by the overlap function $I(f)$ defined as

$$I(f) = \int_0^{\pi} M(f,\theta)M(f,\theta+\pi)d\theta, \tag{2}$$

so that the PSD $P(\mathbf{K} \approx 0, f_s)$ of pressure fluctuations in frequency-2 dimensional (2D) wavenumber space (with dimension $\mathrm{N^2\,m^{-2}\,Hz^{-1}}$, see Hasselmann, 1963) can be approximated around $\mathbf{K} \approx 0$ as (Ardhuin et al., 2013)

$$P(\mathbf{K} \approx 0, f_s) \approx \rho_w^2 g^2 f_s E^2(f_s/2) I(f_s/2) \tag{3}$$

where g is the gravitational acceleration and ρ_w is water density.

We assume that seismic energy is dominated by seismic surface waves, and we thus neglect the contribution of direct P and S waves. This assumption is appropriate for coastal stations that are mainly sensitive to local oceanic sources (Ardhuin and Herbers, 2013), but would be less appropriate for farther-inland stations, where P and S waves can significantly contribute to the observed noise (Zhang et al., 2009). As in Gualtieri et al. (2013), we integrate the contribution of pressure fluctuations resulting from all interacting OSGW pairs within the area Γ_i of each element number i of the wind grid by considering an equivalent point force acting in its center $\mathbf{x}_i$. The equivalent point force PSD $F_i(f_s)$ (with dimension $\mathrm{N^2\,Hz^{-1}}$) resulting from the pressure PSD $P_i(\mathbf{K} \approx 0, f_s)$ can be written as

$$F_i(f_s) = 4\pi^2 P_i(\mathbf{K} \approx 0, f_s)\Gamma_i \tag{4}$$

where the $4\pi^2$ pre-factor results from the conversion from the 2D wavenumber to the 2D spatial domain. The total PSD of the vertical acceleration of the ocean floor $A^{of}(f_s, \mathbf{x}, H)$ at horizontal coordinate $\mathbf{x}$ and at depth H where H is the ocean layer thickness can be calculated by summing all surface-wave modes and cell contributions as (Aki and Richards, 2002)