1 Predicting short-period, wind-wave-generated seismic 2 noise in coastal regions

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8 **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

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can be studied in detail from seismic observations, including under specific conditions such as in the presence of sea ice.

9 *Keywords:* ocean waves; seismic noise; directional spectra; wind

10 **1. Introduction**

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

19 Microseisms are particularly useful for studying Earth structure using
20 noise cross correlation techniques (Campillo and Paul, 2003; Bowden et al.,
21 2015), and better knowledge of their characteristics is needed to avoid spuri-
22 ous artifacts caused by spatial and temporal variations in noise sources (Tsai,
23 2009). Numerous investigations have been conducted to better understand
24 the ocean processes that create microseismic noise and their spatiotemporal
25 characteristics. Maximum amplitudes of the secondary microseism peak (4-8
26 s) have been successfully predicted (Kedar et al., 2008; Ardhuin et al., 2015)
27 by combining numerical ocean wave models with the Longuet-Higgins (LH)
28 theory (Longuet-Higgins and Ursell, 1948; Longuet-Higgins, 1950), as later
29 revisited by Hasselmann (1963). The maximum energy in that period range
30 is mainly caused by strong ocean swell populations with periods of typically

31 8-16 s, i.e. wavelengths of 100-400 m, that travel in nearly opposite direc-
32 tions either in coastal or deep-ocean regions as a result of their generation by
33 distant storms, by single but fast moving storms or by coastline reflections.

34 In contrast to these previous findings, little attention has been devoted to
35 understanding how ocean processes cause the relatively shorter period (<4
36 s) seismic noise discussed in various recent studies (Zhang et al., 2009; Tsai
37 and McNamara, 2011; Beucler et al., 2014). In contrast to the longer (4-8
38 s) periods at which secondary microseisms are observed almost everywhere
39 in continental areas (Berger et al., 2004), shorter period ocean-induced noise
40 is expected to be more restricted to coastal regions since seismic waves are
41 more attenuated at these shorter periods. However, short-period ocean noise
42 is increasingly used for high-resolution imaging of shallow Earth structure
43 in coastal regions (Lin et al., 2013; Bowden et al., 2015). Moreover, Tsai
44 and McNamara (2011) suggested that sea-ice mechanical properties could be
45 continuously monitored from the analysis of coastal ground motion in the
46 0.6-2 s period band. Before such a goal can be achieved, though, an accurate
47 understanding of coastal seismic noise amplitude and frequency scaling under
48 sea-ice free conditions is needed: this is the main purpose of this study.

49 As in Webb (1992), Farrell and Munk (2008) and Duennebier et al. (2012),
50 we adopt a simple analytical approach that predicts secondary microseisms
51 from the interaction of short-wavelength (1-25 m) wind-waves. In contrast to
52 these previous studies who limited their analysis to the modeling of acoustic
53 pressure and ground floor displacement at the ocean bottom, we conjointly
54 model pressure spectra recorded at the ocean bottom together with accelera-
55 tion spectra recorded by seismic stations in coastal areas. This joint analysis

Figure 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.

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 figure 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 on figure 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-hr 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 frequency f_s and over a given time window of duration T as

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

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

96 frequencies f and f' , generate a resultant wave of horizontal wave number
 97 $\mathbf{K} = \mathbf{k} + \mathbf{k}'$. Of all wave types resulting from all possibly interacting OSGW
 98 pairs, only those that satisfy $|\mathbf{K}| \approx 0$, i.e. $\mathbf{k} \approx -\mathbf{k}'$, and consequently
 99 $f \approx f'$ contribute to seismic wave generation in deep water (Longuet-Higgins,
 100 1950; Hasselmann, 1963; Ardhuin and Herbers, 2013). The frequency of the
 101 surface forcing is $f_s = f + f' \approx 2f$ and its amplitude is proportional to the
 102 amplitude and fraction of nearly-oppositely traveling wave pairs within the
 103 broad OSGW spectrum $E(f, \theta) = E(f)M(f, \theta)$ (with dimension $\text{m}^2 \cdot \text{Hz}^{-1}$),
 104 where $E(f)$ is the ocean surface wave elevation PSD and $M(f, \theta)$ is the
 105 directional distribution of OSGWs that depends on azimuth θ , and satisfies
 106 $\int_{-\pi}^{\pi} M(f, \theta) d\theta = 1$ (Mitsuyasu et al., 1975; Ewans, 1998). The fraction of
 107 interacting wave pairs can be represented by the overlap function $I(f)$ defined
 108 as

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

109 so that the PSD $P(\mathbf{K} \approx 0, f_s)$ of pressure fluctuations in frequency-2 di-
 110 mensional (2D) wavenumber space (with dimension $\text{N}^2 \cdot \text{m}^{-2} \cdot \text{Hz}^{-1}$, see Has-
 111 selmann, 1963) can be approximated around $\mathbf{K} \approx 0$ as (Ardhuin et al., 2013)

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

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

113 We assume that seismic energy is dominated by seismic surface waves,
 114 and we thus neglect the contribution of direct P and S waves. This as-
 115 sumption is appropriate for coastal stations that are mainly sensitive to local
 116 oceanic sources (Ardhuin and Herbers, 2013), but would be less appropriate
 117 for farther-inland stations, where P and S waves can significantly contribute

118 to the observed noise (Zhang et al., 2009). As in Gualtieri et al. (2013), we
 119 integrate the contribution of pressure fluctuations resulting from all interact-
 120 ing OSGW pairs within the area Γ_i of each element number i of the wind
 121 grid by considering an equivalent point force acting in its center $\mathbf{x}_i$. The
 122 equivalent point force PSD $F_i(f_s)$ (with dimension $\text{N}^2 \cdot \text{Hz}^{-1}$) resulting from
 123 the pressure PSD $P_i(\mathbf{K} \approx 0, f_s)$ can be written as

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

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

$$\begin{aligned} A^{of}(f_s, \mathbf{x}, H) &= \sum_{j=1}^{\infty} A_j^{of}(f_s, \mathbf{x}, H) \\ &= (2\pi f_s)^4 \sum_{i=1}^N F_i(f_s) \sum_{j=1}^{\infty} G_j\left(f_s, (\mathbf{x}, H); (\mathbf{x}_i, 0)\right)^2 \end{aligned} \quad (5)$$

129 where $A_j^{of}(f_s, \mathbf{x}, H)$ is the PSD of ocean floor acceleration in the vertical
 130 direction for mode j surface waves, $G_j\left(f_s, (\mathbf{x}, H); (\mathbf{x}_i, 0)\right)$ is the vertical dis-
 131 placement Green's function at $(\mathbf{x}, H)$ for a vertical point force at the ocean
 132 surface $(\mathbf{x}_i, 0)$ and the prefactor $(2\pi f_s)^4$ results from converting displacement
 133 to acceleration power.

134 The PSD $A^c(f_s, \mathbf{x}, 0)$ of ground acceleration in continental regions likely
 135 differs from that defined above for ocean floor acceleration because (i) dis-
 136 placement is no longer evaluated underwater (at depth H) but rather at the
 137 surface (Tanimoto, 2013), (ii) the seismic structure of continental crust is ex-
 138 pected to differ from that of oceanic crust, (iii) wave scattering and seismic

139 energy loss is expected at the ocean-continent boundary due to a relatively
 140 sharp change of topography and structure there (McGarr, 1969), and (iv)
 141 seismic wave attenuation is expected to be larger in continental areas than in
 142 ocean-covered areas (regardless of changes in crustal structure), since seismic
 143 surface waves in ocean-covered areas propagate partly in water and inelastic
 144 attenuation is much smaller for water than for rocks (Gualtieri et al., 2013).
 145 For simplicity, hereafter we only account for (iv) in our model, and we later
 146 discuss in section 6.1 the potential prediction bias resulting from neglect-
 147 ing the other points. With this simplification, analytical expressions for the
 148 PSDs $A^{of}(f_s)$ and $A^c(f_s)$ are provided below by first defining the ocean sur-
 149 face wave elevation PSD $E(f)$ (which is needed to calculate the PSD $F(f_s)$,
 150 see equations 3 and 4) and the appropriate Green’s functions for oceanic and
 151 continental situations.

152 *3.1. The surface wave elevation spectrum*

153 OSGW amplitudes depend on wind speed, wind duration, fetch, wave-
 154 wave interactions, wave breaking and the presence of swell. As discussed
 155 in the introduction, ocean swell causes seismic noise at longer periods than
 156 those of interest here (Kedar et al., 2008; Ardhuin et al., 2015). Thus, we
 157 restrict our analysis to wind-waves, i.e. waves that are set by the local wind.
 158 Fetch distance and wind duration are major controls on the development
 159 and amplitude of those waves (Pierson and Moskowitz, 1964). For large
 160 enough fetch distances and long enough wind durations, though, wind waves
 161 eventually reach their fully developed stage, i.e. their amplitudes saturate
 162 at all frequencies down to a peak frequency f_p at which the wave velocity
 163 approaches the wind velocity. Following Alves et al. (2003), we estimate

$f_p \approx 0.159gU_{10}^{-1}$, where U_{10} is the wind velocity 10 meters above sea level.
 Fully developed waves are expected for fetch distances larger than about 8
 km and 600 km, respectively, for the wind-speed extrema of 2 m/s and 18
 m/s considered later (Holthuijsen, 2010). These minimum fetch distances
 for fully developed waves correspond to minimum wind durations (for the f_p
 frequency wave to travel the minimum fetch distance) of about 1 hour for 2
 m/s wind speeds and about 10 hours for 18 m/s wind speeds. We assume
 that fully developed waves always occur and we discuss the potential model
 prediction bias due to this assumption in sections 5 and 6.1.

An expression for the fully developed OSGW spectrum was first proposed
 by Phillips (1958), who suggested that the maximum amplitude of the OSGW
 spectrum is controlled by wave breaking, and is thus unlikely to depend
 on near-surface wind speeds. From dimensional arguments, Phillips (1958)
 proposed that the surface elevation spectrum $E(f)$ scales as $g^2 f^{-5}$, which has
 since been referred to as saturation-range scaling (see figure 2). While the
 seminal paper of Pierson and Moskowitz (1964) first showed good agreement
 between Phillip’s saturation-range scaling and observations, other studies
 later suggested that there exists a frequency range, often referred to as the
 equilibrium range, for which the amplitude spectrum deviates from a f^{-5} -
 dependence and is a function of near-surface wind speed such that $E(f)$
 rather scales as $U_{10}g f^{-4}$ (Mitsuyasu et al., 1975; Forristall, 1981; Kahma,
 1981; Kitaigorodskii, 1983; Donelan et al., 1985; Long and Resio, 2007). More
 recently, Battjes et al. (1987) and Alves et al. (2003) revisited the database
 of Pierson and Moskowitz (1964) and concluded that the f^{-4} -scaling of the
 equilibrium range captures the observations better than a f^{-5} -scaling does

189 in the frequency range that they investigated, i.e. between $1.5f_p$ and $4f_p$.

190 Such a f^{-4} -scaling for the wind-sea spectrum has also been theoreti-
191 cally supported by Kitaigorodskii (1983). Based on Hasselmann’s previous
192 work on non-linear wave-wave interactions (Hasselmann, 1962) and assum-
193 ing a spectral energy influx from wind that is concentrated relatively close
194 to the peak spectrum, i.e. concentrated at low frequencies, Kitaigorodskii
195 (1983) showed that a Kolmogorov-like transfer of energy from low to high
196 frequencies by wave interactions in the deep ocean causes the surface eleva-
197 tion spectrum to scale with frequency as f^{-4} within that range. This equi-
198 librium range extends from the peak frequency up to the cut-off frequency
199 f_x at which the sea surface acceleration reaches the limit for wave breaking,
200 which Phillips (1958) hypothesized to correspond to the downward gravita-
201 tional acceleration g (see figure 2). Above the cut-off frequency f_x , Phillip’s
202 saturation range is expected. Kitaigorodskii (1983) roughly estimated the
203 non-dimensional equilibrium-range frequency cut-off $\tilde{f}_x = f_x U_{10}/g$ to be at
204 most $\approx 0.25 - 0.5$.

205 We interpret the abrupt change in the frequency scaling of the acoustic
206 power spectra reported by Duennebie et al. (2012) at ALOHA (see figure
207 S1(a)) to be due to the transition from Kitaigorodskii’s equilibrium range
208 to Phillip’s saturation range. We infer $\tilde{f}_x$ by reporting the cut-off frequency
209 values observed at ALOHA as a function of wind speed. We find $\tilde{f}_x = 0.75$
210 (see figure S1(b)) and use this value throughout. This value is of the same
211 order as those observed by Forristall (1981) ($\tilde{f}_x \approx 0.65$ for wind speeds
212 varying from 0 to 28 m/s) and Long and Resio (2007) ($f_x \approx 2.1 - 3.5f_p$, i.e.
213 $\tilde{f}_x = 0.33 - 0.56$, for wind speeds varying from 2 to 24.5 m/s) and discussed

Figure 2: Unidirectional OSGW elevation spectrum $E(f)$ used in this study (thick red line) as a function of non-dimensional frequency f/f_p . The blue crosses correspond to the data analyzed by Forristall (1981). The dotted line shows the classical f^{-5} Pierson-Moscowitz (PM) spectrum built on Phillip’s saturation-range scaling. The gray region shows the potential deviations from the PM spectrum discussed by Hasselmann et al. (1976) and Kahma (1981). The second x-axis provides the seismic frequencies to which this spectrum corresponds for a wind speed of 10 m/s.

by Kahma (1981) ($\tilde{f}_x \approx 0.8$) and Hwang and Wang (2001) ($f_x \approx 2.1 - 2.9f_p$,
i.e. $\tilde{f}_x \approx 0.33 - 0.46$).

Thus, we use a f^{-4} -scaling for $E(f)$ to describe the equilibrium range from the peak frequency f_p up to f_x , and include a saturation range with a f^{-5} -scaling at higher frequencies. By imposing continuity of the spectrum at

219 f_x , we write $E(f)$ as

$$\begin{aligned} E(f) &= \gamma U_{10} g f^{-4} \exp \left[- \left(\frac{f}{f_p} \right)^{-4} \right] & \text{for } f < \frac{g}{U_{10}} \tilde{f}_x \\ E(f) &= \tilde{f}_x \gamma g^2 f^{-5} & \text{for } f \geq \frac{g}{U_{10}} \tilde{f}_x. \end{aligned} \quad (6)$$

220 where γ is a constant and the exponential term accounts for a frequency
 221 cut-off at f_p . By adjusting the f^{-5} -scaling of the form $\gamma' g^2 (2\pi f)^{-5}$ to fit
 222 observations, Phillips (1958) originally suggested that $\gamma' = 0.0074$ was a uni-
 223 versal constant. Pierson and Moskowitz (1964) later obtained $\gamma' = 0.0081$
 224 from fitting the observations of the Joint North Sea Wave Project, and pro-
 225 posed the classical Pierson-Moskowitz (PM) spectrum shown by the dotted
 226 line in figure 2. Since then, other investigators have shown that γ' is in fact
 227 not a universal constant and may vary by about a factor of 2 depending on
 228 situations (e.g., Hasselmann et al., 1976; Kahma, 1981). The range of ocean
 229 wave spectra for these expected γ' -variations is shown by the gray regions in
 230 figure 2. It is important to note that the γ' -values and thus the gray region in
 231 figure 2 were constrained by adjusting the f^{-5} ocean wave spectrum scaling
 232 to match observations only in the relatively low frequency range of $1.5f_p$ to
 233 $2.5f_p$ (Phillips, 1958; Pierson and Moskowitz, 1964; Hasselmann et al., 1976).

234 We determine the constant γ of equation 6 so that it agrees with the
 235 amplitude of previously proposed formulations in the $1.5f_p$ - $2.5f_p$ frequency
 236 range (see figure 2). We use $\gamma = 1.8 \cdot 10^{-5}$ and obtain the OSGW spectrum
 237 shown by the red line in figure 2. Our result as well as the observations
 238 of Forristall (1981) (see blue crosses in figure 2) roughly agree with previous
 239 parametrizations in the $1.5f_p$ to $2.5f_p$ frequency range, but deviate from the
 240 later at larger frequencies as a result of a f^{-4} rather than f^{-5} -scaling for
 241 frequencies lower than f_x . Accounting for these differences in scaling and

242 amplitude of the omnidirectional spectrum with respect to the classical PM
 243 spectrum has important implications at the higher frequencies of $2f_p$ to $5f_p$
 244 (0.5 to 1.5 Hz for 10 m/s wind speeds, see figure 2) that we mostly investigate
 245 here. In particular, we show that the use of this modified f^{-4} -scaling strongly
 246 affects the overlap function I inferred from acoustic and seismic observations.

247 3.2. Acoustic and seismic noise

248 3.2.1. The acoustic spectrum

249 Ocean-floor acoustic measurements are affected by direct acoustic waves
 250 propagating down the water column and by pressure fluctuations resulting
 251 from seismic waves excited by more distant seas (Ardhuin and Herbers, 2013).
 252 The PSD $P_D^a(f_s)$ of directly created acoustic pressure can be obtained fol-
 253 lowing Ardhuin and Herbers (2013) by integrating the contribution of all
 254 waves with a non-zero down-propagating component in the water column,
 255 i.e. all waves that satisfy $|\mathbf{K}| < 2\pi f_s/\alpha_w$. To do this, the pressure PSD
 256 $P(\mathbf{K} \approx 0, f_s)$ of equation 3 is integrated within the circle of radius $2\pi f_s/\alpha_w$
 257 in 2D-wavenumber space to obtain

$$P_D^a(f_s) \approx \pi \frac{(2\pi f_s)^2}{\alpha_w^2} \rho_w^2 g^2 f_s E^2(f) I(f), \quad (7)$$

258 where $\alpha_w = 1500$ m/s is the sound speed in water.

259 The PSD $P_I^a(f_s, \mathbf{x}, z)$ of acoustic pressure at depth z below the water sur-
 260 face resulting indirectly from more distant seas through ocean floor shaking
 261 at location $\mathbf{x}$ can be evaluated by solving for mode resonance in the water
 262 column. The pressure $p(z)$ equals the water stiffness K_w times the strain
 263 $l_j \delta$ due to the acoustic wave, where δ is the ocean floor displacement and
 264 $l_j = 2\pi f_s \sqrt{\frac{1}{\alpha_w^2} - \frac{1}{v_j^2}}$ is the vertical wave number of the mode j surface wave

265 propagating at horizontal phase velocity v_j . By using the sound wave re-
 266 lation $\alpha_w = \sqrt{K_w/\rho_w}$ and applying boundary conditions of zero pressure
 267 at the ocean surface (free surface) and predicted displacement at the ocean
 268 bottom, one can express $P_i^a(f_s)$ as a function of the ocean-floor acceleration
 269 PSD $A_j^{of}(f_s, \mathbf{x}, H)$ as (Ardhuin et al., 2013)

$$P_I^a(f_s, \mathbf{x}, z) = \sum_{j=1}^{\infty} \left[\rho_w \frac{\sin(l_j z)}{l_j \cos(l_j H)} \right]^2 A_j^{of}(f_s, \mathbf{x}, H). \quad (8)$$

270 The final PSD $P^a(f_s)$ of acoustic pressure at the hydrophone may be cal-
 271 culated as $P^a(f_s) = P_D^a(f_s) + P_I^a(f_s)$, and thus involves the independent
 272 calculation of ocean-floor acceleration power $A_j^{of}(f_s)$, which is performed in
 273 the following section.

274 3.2.2. The ocean floor acceleration spectrum

275 In order to calculate the PSD A_j^{of} of vertical ocean floor acceleration due
 276 to mode j surface waves (see equation 5), we express the vertical-vertical
 277 Rayleigh-wave Green's function G_j as (Aki and Richards, 2002)

$$G_j(f_s, (\mathbf{x}, H); (\mathbf{x}_i, 0)) = \frac{r_j^v(0)r_j^v(H)}{8u_jv_j\mathbb{I}_{1j}} \sqrt{\frac{2}{\pi k_j d_i}} e^{-\pi f_s d_i/(u_j Q)} \quad (9)$$

278 where $r_j^v(H)$ and $r_j^v(0)$ are the vertical eigenfunctions of the mode j surface
 279 wave at the ocean bottom and at the ocean surface, respectively, u_j and v_j are
 280 the group and phase velocities, $k_j = 2\pi f_s/v_j$ is the horizontal wavenumber, Q
 281 is the quality factor that we assume to be mode independent, $d_i = |\mathbf{x} - \mathbf{x}_i|$ is
 282 the lateral source-to-station distance and $\mathbb{I}_{1j} = 1/2 \int_0^\infty \rho(z)[r_j(z)^2 + r_j'(z)^2]dz$
 283 is the integral of horizontal r_j^h and vertical r_j^v displacement eigenfunctions
 284 with depth z , $\rho(z)$ being either equal to water density ρ_w in the ocean, ρ_s in
 285 the sediments or ρ_c in the crust. The general Green's function expression as

defined in equation 9 can be mapped to that originally proposed by Longuet-Higgins (1950, see equation 183 therein) by taking the ocean crust of uniform density ρ_c and shear-wave speed β_c as a reference, and rewriting equation 9 as

$$G_j(f_s, (\mathbf{x}, H); (\mathbf{x}_i, 0)) = f_s \frac{c_j}{\rho_c \beta_c^{5/2}} \sqrt{\frac{1}{d_i}} e^{-\pi f_s d_i / (u_j Q)} \quad (10)$$

where

$$c_j = \frac{N_j(H, 0) \beta_c^{5/2}}{4u_j v_j^{3/2}} \quad (11)$$

and $N_j(H, 0)$ is defined as in Tsai and Atiganyanun (2014) as

$$N_j(H, 0) = \frac{\rho_c r_j^v(0) r_j^v(H)}{k_j \mathbb{I}_{1j}}. \quad (12)$$

By substituting the Green's function of equation 10 into the definition of the PSD $A^{of}(f_s, \mathbf{x})$ provided in equation 5, one can rewrite the total PSD $A^{of}(f_s, \mathbf{x}, H)$ as in Ardhuin and Herbers (2013) by assuming that all modes propagate at the same group velocity u so that

$$A^{of}(f_s, \mathbf{x}, H) = (2\pi f_s)^4 \sum_{i=1}^N T_i(f_s) \frac{1}{d_i} e^{-2\pi f_s d_i / (uQ)} \quad (13)$$

where

$$T_i(f_s) = f_s \frac{C^2}{\rho_c^2 \beta_c^5} F_i(f_s) \quad (14)$$

and $C = \sqrt{\sum_{j=1}^{\infty} c_j^2}$ are coefficients.

3.2.3. Sensitivity of ground motion to ground properties

We evaluate the sensitivity of ground motion amplitude to ground properties by calculating the Green's function pre-factor $C^2/(\rho_c^2 \beta_c^5)$ for different structures. We first consider the case of a water layer on top of the crust as an

Figure 3: Green's function pre-factor $C^2/(\rho_c^2 \beta_c^5)$ (see equation 13, summed over all modes) for an ocean layer (a) over a homogeneous half-space with various shear-wave speeds β_c and (b) over a sedimentary layer of depth-average shear-wave speed β_s on top of a homogeneous half-space with various shear-wave speeds β_c . We use $\rho_c = 2500 \text{ kg/m}^3$, $\rho_s = 1700 \text{ kg/m}^3$, $\alpha_c = \sqrt{3}\beta_c$ and $\alpha_s = 1500 \text{ m/s}$. Left panels show $C^2/(\rho_c^2 \beta_c^5)$ as a function of dimensionless frequency Hf_s/α_w , with the gray zones delimiting the region of interest in this study given an ocean depth range that mainly varies between 4000 and 6000 m around KIP and AIS (see figure 1) and a frequency range of interest varying between 0.5 and 1.5 Hz. The average values of the pre-factor $C^2/(\rho_c^2 \beta_c^5)$ within the grey regions are shown in the right panels as a function of (a) β_c and (b) H_s for various values of β_c .

302 ‘equivalent’ homogenous half-space (i.e. no explicit sediment layer) of con-
 303 stant density $\rho_c = 2.5 \text{ kg/m}^3$ and different constant shear and compressional
 304 wave speeds β_c and $\alpha_c = \sqrt{3}\beta_c$ where β_c is varied from 2 to 3.5 km/s (see fig-
 305 ure 3, in which C is calculated by solving equation 179 of Longuet-Higgins,
 306 1950). At small values of Hf_s/α_w , i.e. for a relatively thin ocean layer,
 307 $C^2/(\rho_c^2\beta_c^5)$ strongly depends on β_c . In contrast, due to surface waves being
 308 more sensitive to the water layer for relatively deeper oceans, $C^2/(\rho_c^2\beta_c^5)$ is
 309 less sensitive to β_c as Hf_s/α_w increases. In the special case of high-frequency
 310 noise at island stations surrounded by a deep ocean ($f_s \approx 0.5 - 1.5 \text{ Hz}$ and
 311 $H \approx 4 - 6 \text{ km}$, see grey region in figure 3(a)), uncertainties in β_c only causes
 312 a $\pm 1.5 \text{ dB}$ bias in model predictions (see right panel of figure 3(a)).

313 Next, we investigate the role of an explicit sedimentary layer of constant
 314 and uniform density $\rho_s = 1.7 \text{ kg/m}^3$ and compressional wave speed $\alpha_s = 1.5$
 315 km/s but varying thickness H_s and depth-averaged shear-wave speed β_s . In
 316 this situation, a more realistic crustal range is used for β_c (2.9 to 3.5 km/s)
 317 and β_s is averaged for values of H_s varying from 10 m ($\beta_s = 162 \text{ m/s}$)
 318 to 750 m ($\beta_s = 757 \text{ m/s}$) using the generic shear-wave-speed depth profile
 319 of Hamilton (1976). An example of the variation of $C^2/(\rho_c^2\beta_c^5)$ with Hf_s/α_w
 320 for various H_s -values (and thus various β_s -values) is shown in figure 3(b)
 321 using $\beta_c = 2.9 \text{ km/s}$ and $H = 4.5 \text{ km}$. As already noted by Ardhuin et al.
 322 (2013), the sediment layer modifies the amplitude and shape of each mode,
 323 causing the C -coefficients to exhibit a more complex variation with frequency.
 324 However, the average values of the pre-factor $C^2/(\rho_c^2\beta_c^5)$ remain of the same
 325 order of magnitude as those obtained without an explicit sediment layer. In
 326 the following, we consider the simple situation of no sediment layer and an

‘equivalent’ homogeneous crust of shear velocity $\beta_c = 2.6$ km/s. We note that uncertainties in ground structure cause a -3 dB to a +4 dB uncertainty in model predictions (see right panels of figure 3(b)).

3.2.4. The acceleration spectrum in continental regions

To predict ground acceleration power in continental regions, we define two distinct quality factors Q_w for water-covered areas and Q_g for grounded areas and write the PSD P_a^c as

$$A^c(f_s, \mathbf{x}, 0) = (2\pi f_s)^4 \sum_{i=1}^N \frac{T_i(f_s)}{d_i^{sc} + d_i^{cs}} \exp\left(-2\pi f_s \left(\frac{d_i^{sc}}{uQ_w} + \frac{d_i^{cs}}{uQ_g}\right)\right), \quad (15)$$

where d_i^{sc} is the source-to-coast distance and d_i^{cs} is the coast-to-station distance along the considered ray path. Here, we assume that seismic surface waves propagate at the same group velocity u over water-covered and inland regions. In reality, the velocity of seismic waves is significantly affected by the presence or absence of the water layer on top of the crust. Tsai and Atiganyanun (2014) suggested that group velocities of fundamental-mode Rayleigh waves vary from 1-2 km/s for generic continental crust and in the frequency range of interest. In water-covered areas and for the crust as an ‘equivalent’ homogeneous half-space, group velocities may vary between α_w and β_c depending on wave mode and ocean depth (e.g., see Ardhuin and Herbers, 2013). For simplicity, we approximate u by α_w , and note that this assumption could lead to at most a factor of 2 overestimate of the absolute values of Q_w and Q_g (compared to using $u \approx \beta_c$) inferred in the following (see section 4), and would not change the final predictions.

The final expression for the ground acceleration power A^c is obtained by substituting the expression for T_i in equation 14 into equation 15. Using

equations 3 and 4 to express the point force PSD $F_i(f_s)$ and equation 6 to express the OSGW spectrum $E(f)$, we obtain

$$\begin{aligned}
A^c(f_s, \mathbf{x}, 0) = & \sum_{i=1}^N \frac{\rho_w^2}{\rho_s^2} \frac{C^2}{\beta^5} \frac{\gamma_4^a g^4 U_{10,i}^2 I f_s^{-2}}{d_i^{sc} + d_i^{cs}} \Gamma_i \exp \left[- \left(\frac{f}{f_p} \right)^{-4} \right] \\
& \exp \left[- 2\pi f_s \left(\frac{d_i^{sc}}{u Q_w} + \frac{d_i^{cs}}{u Q_g} \right) \right] \quad \text{for } f_s < \frac{2g}{U_{10,i}} \tilde{f}_x \\
A^c(f_s, \mathbf{x}, 0) = & \sum_{i=1}^N \frac{\rho_w^2}{\rho_s^2} \frac{C^2}{\beta^5} \frac{\gamma_5^a g^6 I f_s^{-4}}{d_i^{sc} + d_i^{cs}} \Gamma_i \\
& \exp \left[- 2\pi f_s \left(\frac{d_i^{sc}}{u Q_w} + \frac{d_i^{cs}}{u Q_g} \right) \right] \quad \text{for } f_s \geq \frac{2g}{U_{10,i}} \tilde{f}_x.
\end{aligned} \tag{16}$$

with $\gamma_4^a = 4\pi^2(2\pi)^4 2^8 \gamma^2 = 5.1 \cdot 10^{-3}$ and $\gamma_5^a = 4\pi^2(2\pi)^4 2^{10} \tilde{f}_x^2 \gamma^2 = 1.1 \cdot 10^{-2}$.

Our model predictions use a downscaled version of the bathymetry grid (8-minute resolution) on which the coarser wind field has been interpolated. This relatively high-resolution grid is particularly suitable for representing coast-to-station distances over which significant attenuation occurs at the frequencies of interest (see section 4). The acoustic pressure power P^a is calculated by summing the direct acoustic wave contributions (see equation 7) and the contribution from ocean floor shaking (see equation 8). The ocean floor acceleration PSD A^{of} is obtained by propagating the modes separately from each grid source to the hydrophone and then stacking all mode contributions. The total ground acceleration PSD A^c at a seismic station is obtained by evaluating equation 16. Prior to showing model predictions against observations in section 5, we first provide constraints on model parameters in the next section.

4. Constraints on model parameters

Here we provide constraints on the overlap function I and the two quality factors Q_w and Q_g (see equations 8 and 16) from the joint analysis of seismic

369 and acoustic observations at various local wind speeds. Since ocean-floor
 370 acoustic measurements are not sensitive to the inland quality factor Q_g while
 371 seismic measurements are, the joint analysis of seismic and acoustic data
 372 provide useful constraints on Q_g . In contrast, the 2 other parameters I and
 373 Q_w affect the seismic and acoustic noise records similarly, since ocean-floor
 374 acoustic measurements are also mainly sensitive to distant sources through
 375 surface wave propagation (Ardhuin et al., 2013). However, the respective role
 376 of the 2 parameters I and Q_w in setting the resulting noise power depends on
 377 the different wind-field situations. By controlling the contribution of nearby
 378 vs. faraway sources, Q_w strongly influences the predicted noise power when
 379 local wind speeds are low but distant wind speeds are higher. On the other
 380 hand, the noise power predicted when local wind speeds are strong is only
 381 weakly affected by Q_w , since the main noise contribution in that case comes
 382 from surface waves that propagate over short distances and thus undergo
 383 little inelastic attenuation. In this case, the acoustic power is almost entirely
 384 set by I .

385 We therefore model acoustic and seismic PSDs that we bin averaged by
 386 local wind speed. We do this jointly at KIP and ALOHA, and separately
 387 at AIS. We use the wind measurements of the WHOTS buoy as the local
 388 wind-speed reference for KIP and ALOHA, while for AIS we use wind speed
 389 at the closest grid-point of the ERA-Interim dataset (located 35 km from the
 390 station). Wind bins of 0.5 m/s are used within the 2 to 14 m/s range for KIP
 391 and ALOHA, while wind bins of 1 m/s are used within the 2 to 18 m/s range
 392 for AIS. Average PSDs for each wind bin are directly taken from Duennebieer
 393 et al. (2012) for acoustic records at ALOHA, and are processed over the same

394 time period (Feb. 2007 to Oct. 2008) for ground acceleration at KIP. The
 395 entire year of 2012 is considered for AIS. To avoid potential contamination
 396 of the observed PSDs by anthropogenic noise sources at KIP (see figure S2),
 397 PSDs and wind fields are only taken from night time (8 PM to 6 AM) periods
 398 there. For consistency between observed average PSDs and PSDs modeled
 399 from average wind fields, average $U_{10,i}$ -values are calculated at each grid point
 400 i of the wind dataset as $\langle U_{10,i} \rangle = \sqrt{\left(\sum_{p=1}^n U_{10,i,p}^2\right)/n}$, where n corresponds
 401 to the number of wind-field configurations that have local-wind-speed values
 402 lying within the same bin.

403 Figure 4 shows the average wind fields computed surrounding ALOHA
 404 and KIP and at local wind speeds of 2 and 14 m/s, and figure 5 shows the
 405 average PSDs observed at all considered stations and for the various local
 406 wind speeds (see continuous lines). The direct relationship between observed
 407 noise power and local wind speed is obvious for all stations in the 0.5-1.5 Hz
 408 frequency range (i.e. periods of 0.6-2 s), while this is no longer the case at
 409 lower frequencies (around 0.2 Hz) where similar noise levels are observed for
 410 different local wind speeds (with the exception of the very high wind-speed
 411 cases).

412 In order to find the parameter values that best fit the data over all local
 413 wind speeds, we perform a grid search over all model parameter space. For
 414 each model parameter combination and each local wind speed value, we cal-
 415 culate the misfit (in dB) between the observed and predicted power averaged
 416 in the 0.5 to 1.5 Hz frequency band. The power averaging is done in dB
 417 space and using log-binned frequencies so that the power values are equally
 418 weighted for all frequencies within the 0.5 to 1.5 Hz range. We finally obtain

Figure 4: Average wind fields computed in the Pacific Ocean from Feb. 2007 to Oct. 2008 for local-wind speeds of (a) 2 m/s and (b) 14 m/s. The white dot on each plot indicates the location of the WHOTS buoy (which we use as a wind reference in the Pacific Ocean) and the red and blue circles denote distances of 1000 and 2000 km from the WHOTS buoy.

419 the total misfit by summing misfits over all wind speeds and all seismic and
 420 acoustic observations. The total misfit values from jointly analyzing ALOHA
 421 and KIP are shown in figure 6(a). 5-unit bin sizes have been used for our
 422 grid search on Q_g , but slices are only shown every 25 units in figure 6(a) for
 423 clarity. A unique minimum misfit is obtained at $I = -28$ dB, $Q_w = 800$ and
 424 $Q_g = 75$. We discuss the consistency of our inverted I -values with previ-
 425 ous work in section 6.2. The quality factors Q_w and Q_g are of similar order
 426 of magnitude compared to those proposed in previous work. Anderson and
 427 Hough (1984) suggested $Q_g = 20$ at similar frequencies, and Gualtieri et al.
 428 (2013) proposed $Q_g \approx 35$ based on noise amplitude analysis at lower fre-
 429 quencies. Gualtieri et al. (2013) also obtained Q_w -values ranging from 500
 430 to 2000 from modeling surface waves that propagate in ocean covered areas
 431 with similar ocean depths.

Figure 5: Observed vs. modeled seismic and acoustic PSDs at (a) KIP, (b) ALOHA and (c) AIS. Continuous lines show observations and dashed lines show model predictions. The vertical dashed lines delimit the frequency bands over which seismic power is averaged for the misfit calculations shown in figure 6 and for the timeseries shown in figures 7 and 8. The red and purple lines in figure 5(a) for KIP and figure 5(b) for ALOHA are for wind speeds that correspond to the average wind fields shown in figure 4.

Figure 6: Model misfit for various values of the model parameters I , Q_w and Q_g in (a) the Pacific Ocean (stations KIP and ALOHA) and (b) the Indian Ocean (station AIS). The minimum misfit values are highlighted by the black dots on each plot, and the corresponding modeled PSDs are shown against observations in figure 5(a) for KIP and (b) for ALOHA (using $I = -28$ dB, $Q_w = 800$ and $Q_g = 75$), and in figure 5(c) for AIS (using $I = -26$ dB, $Q_w = 1500$ and $Q_g = 75$).

432 We perform a similar grid search for AIS. However, since we lack acoustic
 433 measurements there, we evaluate model misfit only with respect to seismic
 434 observations. Furthermore, since AIS is located very close (about 1 km) to
 435 the ocean, Q_g only weakly affects the seismic signal and is thus not well
 436 constrained in our grid search. Thus, we impose Q_g to be similar to that
 437 at KIP, and set $Q_g = 75$. By performing our grid search over the 2 model
 438 parameters I and Q_w , we obtain $I = -26$ dB and $Q_w = 1500$ (see figure 6(b)).
 439 The value of I inverted at AIS is slightly larger than that at KIP, and we
 440 suggest that this difference could be due to more heterogeneous winds in
 441 the Indian Ocean (see section 5). The value of Q_w is about twice as large
 442 as that at KIP, which could be due to differences in ocean floor properties
 443 between the two regions. It is important to note, though, that a significant
 444 tradeoff is obtained in figure 6 between Q_w and I at both KIP and AIS. This
 445 implies that I -values within ± 1.5 dB of those inverted here accompanied
 446 with Q_w -values within a factor of about 1/3 of those inverted here would
 447 still predict the average spectra shown in figure 5 reasonably well. The next
 448 section shows model predictions of average PSDs and timeseries using the
 449 model-parameter values constrained here.

450 5. Results

451 The predicted average PSDs are shown against observations for the vari-
 452 ous local wind speeds in figure 5. The frequency scaling and variability with
 453 local wind speed for both the acoustic and seismic power are well represented
 454 by our wind-wave-induced noise model. Good agreement between observed
 455 and predicted PSDs is obtained from 0.5-4 Hz at ALOHA, from 0.5-1.5 Hz at

456 KIP and from 0.5-1.2 Hz at AIS. The smaller frequency range of agreement
 457 between our model and the seismic observations compared to the acoustic
 458 observations at ALOHA is likely due to other noise sources overwhelming
 459 that of wind-waves at higher frequencies. This extra source is likely anthro-
 460 pogenic at KIP but is unknown at AIS (see figure S2). Interestingly, one can
 461 clearly see at station KIP and ALOHA that the noise content caused by the
 462 large swell in the Pacific Ocean is generally distinct from that of wind-waves.
 463 In contrast, wind-waves formed at high local wind speeds entirely explain the
 464 seismic energy observed in the swell frequency range at AIS. We interpret
 465 this as a result of ocean swell being smaller in the Indian Ocean due to more
 466 unsteady winds there, which allows wind-wave-induced noise to overwhelm
 467 that of swell under high local winds.

468 Figure 7(a) shows the observed (blue) versus predicted (green) seismic
 469 noise timeseries at station KIP during the 2007-2012 time period. Here ac-
 470 celeration power has been averaged over a timescale of one week and within
 471 the 0.5-1.5 Hz frequency range (see figure 5(a)). Consistently good agreement
 472 is obtained between observations and predictions. However, a weak but sig-
 473 nificant mismatch can be seen between predictions and observations during
 474 the winter, i.e. during time periods for which the wind field is relatively more
 475 heterogeneous (see supplementary video). We crudely estimate the degree of
 476 heterogeneity of the wind field by evaluating the standard deviation $\sigma(U_{10})$ of
 477 the wind values that occur within 2000 km of KIP, i.e. within the red circle
 478 shown in figure 4. This measure is obviously not a physical nor a generic
 479 measure (for example it depends on the wind-mesh resolution), but it pro-
 480 vides a qualitative estimate of the relatively more heterogeneous wind fields

Figure 7: (a) Observed (blue line) versus predicted (green and red lines) timeseries of ground acceleration power over the 2007 to 2012 time period at KIP. The green line is obtained using $I = -28$ dB as inferred in figure 6, while the red line is obtained using I_{mod} (see equation 17) as empirically defined in figure 7(c). (b) Timeseries of the standard deviation σ of the wind values U_{10} lying within a disk of 2000-km radius around the KIP station (see figure 4). (c) Overlap function I (blue lines) and amplitude difference (green dots) between modeled and observed timeseries at KIP. The dashed and continuous blue lines show average I -values as inferred in figure 6 for KIP and AIS. The green crosses show the average of the green dots over binned values of $\sigma(U_{10})$, and the errorbars correspond to the standard deviation. We interpret the observed amplitude difference in terms of variations of the overlap function I with $\sigma(U_{10})$, which we account for in the final model predictions by empirically defining the modified overlap function I_{mod} as shown here by the dotted blue line.

481 in winter (see figure 7(b)). The proportionality between model mismatch and
 482 $\sigma(U_{10})$ is shown in figure 7(c), where a linear scaling is obtained between the
 483 observed-to-predicted amplitude difference and the wind standard deviation
 484 $\sigma(U_{10})$. However, it should be noted that because $\sigma(U_{10})$ is correlated to U ,
 485 the deviations between observed and predicted average spectra in figure 6, al-
 486 though weak, could explain part of the observed amplitude difference versus
 487 $\sigma(U_{10})$ in figure 7(c).

488 We interpret the dependence of the amplitude difference between pre-
 489 dictions and observations on the wind standard deviation to result from a
 490 dependence of the overlap function I on the degree of wind heterogeneity.
 491 The more heterogeneous the wind field is, the more isotropic the generated
 492 seas are expected to be, and thus the larger the average overlap function I is
 493 expected to be over the region that the seismic station is sensitive to. With
 494 this interpretation, we replace the constant I -value of -28 dB obtained in
 495 the previous section at KIP by a modified function I_{mod} that has a slight de-
 496 pendence on the wind standard deviation $\sigma(U_{10})$. From fitting the amplitude
 497 difference observed on figure 7(c), we obtain

$$I_{mod} = -29.87 + 0.80\sigma(U_{10}). \quad (17)$$

498 Such a minor modification of the overlap function allows us to correct for
 499 the amplitude differences observed at extreme values of $\sigma(U_{10})$ at KIP (see
 500 red curve in figure 7(a)). We also suggest that the relatively larger average
 501 value of $I = -26$ dB inferred at AIS is caused by the more heterogeneous
 502 wind there (see supplementary video), where $\sigma(U_{10})$ -values are significantly
 503 higher than for KIP but no dependence of I is observed with $\sigma(U_{10})$.

504 Figure 8(a) shows timeseries of ground acceleration power averaged daily

Figure 8: Predicted (thick red lines) vs. observed (thin blue lines) ground acceleration power timeseries at KIP and AIS over (a) the year and over (b) the month of October (see colored rectangles in figure 8(a)). Colored stars in figure 8(a) correspond to randomly selected days over which associated PSDs are averaged and shown in figure 9. Figure 8(c) shows scatter plots of predicted vs. observed noise power for the yearly predictions, where each dot corresponds to a single day. The black line indicates the 1-to-1 correspondence.

505 and in the appropriate frequency bands, i.e. 0.5-1.5 Hz for KIP and 0.5-1.2
 506 Hz for AIS (see figure 5). An overall good agreement is obtained over the
 507 year between observations and predictions, with correlation coefficients of
 508 0.87 at KIP and 0.79 at AIS between the observed and predicted power (see
 509 scatter plots in figure 8(c)). We note, though, that the minimum noise levels
 510 at AIS are significantly overestimated by our model predictions. This could
 511 be due to an overestimate of Q_w in figure 6, or to a complex dependence of I
 512 with wind spatial heterogeneity. However, ground acceleration power remains
 513 reasonably well predicted at shorter timescales as shown in figure 8(b), which
 514 shows that the observed daily to weekly fluctuations in ground acceleration
 515 power are nicely captured by the model (see figure 8(b)). At even shorter
 516 timescales, there exists a slight time delay between predicted and observed
 517 noise timeseries. While predictions are always ahead of observations, the
 518 time lag is not constant over the year and can vary from 0 to 9 hours. As
 519 already suggested by Duennebier et al. (2012), this time delay likely results
 520 from our crude assumption that ocean waves fully develop instantaneously,
 521 while a time delay is expected in reality (see section 3.1). We attribute time
 522 periods of significantly larger time delays to time periods at which relatively
 523 higher wind speeds (either local or distant) mainly contribute to the resulting
 524 noise.

525 Finally, observed and modeled seismic PSDs are shown in figure 9 for
 526 3 randomly selected days corresponding to the colored stars shown in fig-
 527 ure 8(a). One can see that the daily noise spectral content as well as its
 528 variations due to temporal changes in the wind-field spatial structure is rel-
 529 atively well captured by the model.

Figure 9: Observed (continuous lines) versus modeled (dashed lines) daily-averaged PSDs at KIP (Top) and AIS (Bottom). Each panel corresponds to the randomly selected times shown by stars in figure 8(a) (with corresponding colors). The vertical dashed lines delimit the frequency bands over which seismic power is averaged for the timeseries shown in figures 7 and 8.

530 6. Discussion

531 6.1. Uncertainties on inverted model parameters

532 Due to our imperfect knowledge of OSGWs and ground properties, sev-
533 eral assumptions and approximations have been made. Mainly, we assumed
534 that wind waves are always fully developed, we modeled the crust as a ho-
535 mogeneous half space of constant shear-wave speed $\beta_c = 2600$ m/s with no
536 sediment layer, and we did not account for energy loss and 3D effects as
537 seismic waves propagate across the ocean-continent boundary.

538 Previous numerical modeling of Ardhuin et al. (2013) suggests that the
539 PSD $E(f)$ obtained under non-fully developed situations can be up to 2 dB
540 smaller than that predicted in the fully developed case. Thus, accounting for
541 non-fully developed waves could cause values of I to be up to 3.5 dB larger
542 than those presently obtained. Uncertainties in ground structure (crustal and
543 sediment properties) typically introduce a -3 dB to +4 dB uncertainty in the
544 predicted noise power (see figure 3). Under the reasonable assumption of
545 spatially homogeneous crustal and sediment properties nearby stations, this
546 uncertainty should have limited effect on Q_w and Q_g , and should entirely
547 translates into a -4 to +3 dB uncertainty in the overlap function I . Energy
548 loss at the ocean-continent boundary is unlikely to cause any bias in our
549 inverted I -values, since I is obtained so that the amplitude of ocean bottom
550 pressure is reproduced at ALOHA. Instead, significant energy loss at the
551 ocean-continent boundary would artificially lower our inverted Q_g -values.
552 Interestingly though, the value of $Q_g = 75$ obtained here is not unexpectedly
553 low as compared to values suggested in other studies (Anderson and Hough,
554 1984; Gualtieri et al., 2013). By considering the extreme case where no

555 attenuation occurs in continental areas (i.e. $Q_g = \infty$), the reduction of
 556 seismic power due to energy loss at the ocean-continent boundary can be
 557 estimated to be at most 3 dB. However, such an estimate could be biased by
 558 our neglect of the difference between the ocean floor and the continent in the
 559 Green’s function (Tanimoto, 2013). By assuming that uncertainties on I add
 560 randomly between those due to uncertainties in the wave spectrum amplitude
 561 (+3.5 dB) and those due to uncertainties in the ground structure (−4 dB
 562 to +3 dB), we conclude that the I -values can potentially lie between −33 dB
 563 and −21 dB in the most extreme scenario, but with one-sigma uncertainties
 564 ranging from perhaps more like −30 to −23 dB.

565 *6.2. The overlap function I and implications for the directionality of wind-* 566 *waves*

567 The I -values inverted here (approximately −28 and −26 dB for the chosen
 568 model parameters but potentially as low as −33 or as high as −21 dB including
 569 all model uncertainties) are significantly smaller than those of −23 to −8 dB
 570 inferred by Duennebier et al. (2012) and of −21 to −18 dB inferred by Ardhuin
 571 et al. (2013) in a similar frequency range and for similar wind speeds (see
 572 figure 10). This difference has two main origins. First, as in Ardhuin et al.
 573 (2013), we account for finite ocean depth. Since it is surrounded by an ocean
 574 of large extent, the acoustic noise at ALOHA from ocean-induced ground
 575 vibrations due to distant sources overwhelms by about an order of magni-
 576 tude that from direct waves, which causes a decrease in I by about 10 dB as
 577 compared to the infinitely deep ocean case considered by Duennebier et al.
 578 (2012). Second, we use the equilibrium-range scaling, which causes an extra
 579 5 to 10 dB decrease in I as compared to the previous estimates of Ardhuin

et al. (2013) due to larger OSGW amplitudes predicted at higher frequencies (see figure 2). Use of the equilibrium-range scaling should be appropriate since it is an inherent feature of OSGW spectra (Forristall, 1981; Kahma, 1981; Hwang and Wang, 2001; Long and Resio, 2007), and is theoretically supported by the occurrence of wave-wave interactions (Hasselmann, 1962; Kitaigorodskii, 1983). Interestingly, our representation of the equilibrium range allows us to capture the acoustic observations at ALOHA without involving complex variations in the overlap function I with wind speed as suggested by Duennebier et al. (2012), which as yet has no theoretical justification.

To test whether our inverted I -values are consistent with previous observations of directional OSGW spectra, we use the previous parametrization proposed by Ewans (1998), which describes the directional function $M(f, \theta)$ for wind-waves (see equation 2) using a double Gaussian function

$$M(f, \theta) = \frac{1}{\sqrt{8\pi}\sigma(f)} \left\{ \exp \left[-\frac{1}{2} \left(\frac{\theta - \theta_1}{\sigma(f)} \right)^2 \right] + \exp \left[-\frac{1}{2} \left(\frac{\theta - \theta_2}{\sigma(f)} \right)^2 \right] \right\}, \quad (18)$$

where $\sigma(f)$ corresponds to the standard deviation of each Gaussian and θ corresponds to the direction of wave propagation with respect to the local wind direction, with θ_1 and θ_2 being the average values of each Gaussian distribution defined as

$$\theta_2 = -\theta_1 = \frac{1}{2} \exp \left[5.453 - 2.750 \left(\frac{f}{f_p} \right)^{-1} \right]. \quad (19)$$

This bimodal description of OSGW directionality was first proposed on the basis of inversions of directional buoy records (Ewans, 1998), but has since been supported by various observations that use different techniques, including airborne imagery (Hwang and Wang, 2001; Hwang et al., 2000). The

602 increasing departure of wave propagation direction from wind direction as
 603 frequency increases is thought to result from wave-wave interactions (Banner
 604 and Young, 1994). Good observational constraints exist on the amplitude of
 605 the angle difference between the two Gaussian distributions, i.e. on $\theta_1 - \theta_2$,
 606 for frequencies ranging from $1f_p$ to $4f_p$ (see green line in figure 10(b)). In
 607 contrast, the standard deviation $\sigma(f)$ defined in equation 18 is poorly con-
 608 strained and inverted values for it strongly depend on the observational and
 609 inversion methods (Ewans, 1998). We thus use $\sigma(f)$ as a tuning parameter to
 610 obtain our central value of $I = -27$ dB, and then compare the obtained di-
 611 rectional function $M(f, \theta)$ with those previously reported. The values of $\sigma(f)$
 612 as inverted by substituting equations 18 and 19 into equation 2 are shown in
 613 figure 10(b). In contrast to the suggestion of Ewans (1998), we obtain a stan-
 614 dard deviation $\sigma(f)$ that decreases with frequency, i.e. the directional distri-
 615 bution of each lobe narrows with frequency. This could be explained by the
 616 cancelling of waves in certain azimuthal directions due to wave-wave interac-
 617 tions. The directional function $M(f, \theta)$ inferred from our inversions is shown
 618 against the previous observations of Hwang et al. (2000) in figure 10(c) for two
 619 selected non-dimensional frequencies $f/f_p = \sqrt{3}$ and $f/f_p = 3$. We obtain
 620 good agreement between observed and modeled $M(f, \theta)$ -functions for both
 621 frequencies. We thus conclude that previously reported $M(f, \theta)$ -functions
 622 remain consistent with our inverted I -values. It is important to note that
 623 the larger values of $M(f, \theta)$ observed by Hwang et al. (2000) at $f/f_p = 3$
 624 and in the -30 to +30 degrees angle range as compared to our inverted val-
 625 ues do not significantly affect the associated I -values. In fact, the detailed
 626 shape of the $M(f, \theta)$ -function from -90 to 90° with respect to wind direction

627 is not sufficient to provide good constraints on I , and reliable observations of
 628 $M(f, \theta)$ at θ -values from -180° to -90° and from 90° to 180° are needed. As an
 629 example, the relatively larger values observed by Hwang et al. (2000) at $\pm 90^\circ$
 630 as compared to those inverted here (see figure 10(c)) would be sufficient to
 631 cause I -values to be 10 to 20 dB larger than -27 dB if they were to remain
 632 as large for all angles in the opposite-wind direction, i.e. for θ -values smaller
 633 than -90° or larger than 90° . Unfortunately, Hwang et al. (2000) were not
 634 able to solve for $M(f, \theta)$ in these wind-opposite directions, since the direction
 635 of ocean wave propagation could not be solved for from airborne imagery.
 636 In addition, measurements of realistic and accurate values of $M(f, \theta)$ in this
 637 angle range are likely to be challenging to obtain from classical devices due
 638 to the poor signal-to-noise ratio expected within this angle range.

639 *6.3. Applications*

640 We expect our modeling framework to be particularly useful for moni-
 641 toring sea-ice and for correcting noise correlation measurements from source
 642 directivity effects. In the context of the shrinking sea ice cover observed in
 643 recent years in the Arctic basin, the marginal ice zones of the ice pack are
 644 increasingly affected by storm-generated waves that break-up the ice (Ko-
 645 hout et al., 2014). Increasing ocean wave heights and the reduction of sea
 646 ice mechanical strength (Gimbert et al., 2012a,b) are major controls on sea
 647 ice disintegration, but the scientific community currently lacks observational
 648 tools from which ocean waves and their relation with ice-cover mechanical
 649 strength can be continuously monitored. Tsai and McNamara (2011) sug-
 650 gested that sea-ice mechanical strength could be monitored from land-based
 651 seismic stations through the damping of ground motion power observed in

Figure 10: Directional properties of the OSGW associated with our average inverted I -value of -27 dB. (a) I -values inverted in our study as compared to those inverted from acoustic noise in the previous studies of Duennebier et al. (2012) and Ardhuin et al. (2013). These studies both inverted I at ALOHA, i.e. for similar 2 to 16 m/s wind speeds, and I -values have been taken in a similar frequency range (0.5 to 1.5 Hz). The grey region indicates the range of possible I -values including all model uncertainties. Black stars indicate the configurations shown in figure 10(c). (b) Standard deviation $\sigma(f)$ (blue line) and angle difference $(\theta_1 - \theta_2)/2$ (green line) associated with the double-Gaussian formulation of equation 18. The angle difference $(\theta_1 - \theta_2)/2$ is set from the formulation of Ewans (1998) (see equation 19), while $\sigma(f)$ is inverted such that our average value of $I = -27$ dB is obtained (see figure 10(a)). Vertical dashed and dotted lines indicate the configurations shown in figure 10(c). (c) Directional function $M(f, \theta)$ as a function of θ for frequencies $f = \sqrt{3}f_p$ (dashed lines) and $f = 3f_p$ (continuous lines) as inverted in this study (red) and as observed by Hwang et al. (2000) (black).

the 0.6-2 s period range. The framework presented here can now be used to assess wave-amplitude changes due to sea ice from seismic noise, which one could translate to changes in sea ice mechanical strength by developing appropriate mechanical models.

More generally, we also expect our framework to be useful in the context of the growing use of dense seismic arrays for high-resolution tomography from noise correlation techniques (e.g. Bowden et al., 2015). These noise tomography techniques typically rely on the assumption of isotropic noise sources distributed around stations, which is rarely true in reality (Tsai, 2009). In particular, it is common that dense seismic arrays are placed in coastal regions (e.g. the Long Beach array, California, see Lin et al., 2013), and are thus mainly sensitive to seismic noise incoming from a single side of the network, i.e. within $\pm 90^\circ$ of azimuth (Bowden et al., 2015). The effect of anisotropic noise sources could be determined using our model, therefore improving the ability of noise correlation studies to obtain robust results.

7. Conclusions

We have shown that ocean-floor acoustic and near-coastal seismic noise measurements can simultaneously be modeled in the 0.6-2 s period band. This shows that deep-water and near-shore secondary-microseism sources are sufficient to explain the seismic noise characteristics (amplitude, frequency scaling) observed in coastal regions. There is thus no need to include other source contributions like the breaking of ocean waves on the shore as suggested by Beucler et al. (2014). We suggest that the fraction of ocean waves traveling in nearly-opposite directions is smaller than previously proposed,

676 but still consistent with the available direct observations. We expect the
677 direct and robust relationship reported herein between seismic noise content
678 in coastal regions and local wind-wave properties to be useful for various ap-
679 plications, such as for the remote monitoring of sea ice mechanical properties
680 and for correcting noise-correlation measurements for source directivity.

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