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# Spatiotemporal correlation analysis of noise-derived seismic body waves with ocean wave climate and microseism sources

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## Key Points:

- Time variations of a noise-derived  $P$ -type phase are compared with those of the ocean wave heights and microseism sources.
- Do not equate a positive correlation with a causal relation when studying the links between noise sources and noise-derived signals.
- The derivation of seismic signals from ambient noise relies on the competition between the effective and ineffective sources.

## Abstract

Seismic signals can be extracted from ambient noise wavefields by the correlation technique. Recently, a prominent  $P$ -type phase was observed from teleseismic noise correlations in the secondary microseism period band. The phase is named  $P_{\text{dmc}}$  in this paper, corresponding to its origin from the interference between the Direct  $P$  waves transmitting through the deep Mantle and the Core ( $P$  and  $PKPab$  waves). We extract the phase by correlating noise records from two seismic networks in the northern hemisphere, and locate the microseism sources that are efficient for the  $P_{\text{dmc}}$  construction in the south Pacific. We investigate the spatiotemporal links of the  $P_{\text{dmc}}$  signal with global oceanic waves and microseism sources. Interestingly, the correlation with wave height is higher in several regions surrounding the effective source region, rather than in the effective source region. The  $P_{\text{dmc}}$  amplitude is highly correlated with the power of the effective microseism sources. Also, it is apparently correlated with ineffective sources in the southern hemisphere, and anti-correlated with sources in the northern hemisphere. We ascribe the correlation with the ineffective southern sources to the spatiotemporal interconnections of the southern sources. The anti-correlation with northern sources can be explained by the reverse seasonal patterns between the southern and northern sources, and by that the northern sources impede the signal construction. The signal construction from noise correlations relies on the

39 competition between the effective and ineffective sources, not just on the power of the effective  
40 sources. This principle should be valid in a general sense for noise-derived signals.

41

## 42 **Plain Language Summary**

43 Earth is experiencing tiny but incessant movement induced by natural forces, particularly, storm-  
44 driven ocean waves. While this ambient seismic noise (microseism) was deemed a nuisance in  
45 the past, it can be turned into signals via the seismic correlation technique.

46 Recently, a new *P*-type phase was derived from the noise correlations between two regional  
47 seismic networks. The noise-derived phase originates from the correlation between *P* waves that  
48 propagate through the deep mantle and outer core of the Earth.

49 The temporal amplitude variations of the noise-derived signals are compared with the variations  
50 of microseism sources in the oceans. We show that the signal emergence depends on the  
51 competition between the sources in a specific region that contribute to the signals and sources in  
52 other regions. The conclusion can be generalized to other noise-derived seismic phases.

53 We also analyze the links of the noise-derived signals to ocean waves. In our case, the ocean  
54 waves in the contributing source region are dominated by wind seas forced by local winds,  
55 whereas the excitation of microseisms is primarily owing to the freely traveling swells generated  
56 by oceanic storms in surrounding regions.

57

## 58 **1 Introduction**

59 The incessant background vibrations of Earth had been observed as early as the birth of  
60 seismometers in the later 19th century (Bernard, 1990; Dewey & Byerly, 1969; Ebeling, 2012).  
61 They were termed “microseisms” due to their tiny amplitudes. With more apparatus deployed  
62 worldwide, it was soon recognized that microseisms are ubiquitous and irrelevant to seismicity.  
63 The observation of microseisms aroused interests from various disciplines. Researchers linked  
64 the generation of microseisms to atmosphere processes and ocean wave activity. Meteorologists  
65 tried to employ land observations of microseisms to track remote oceanic storms (e.g., Harrison,  
66 1924). Since the mid-twentieth century, it has been well known that microseisms are excited by  
67 storm-driven ocean waves. The most energetic microseisms that dominate the seismic noise  
68 spectra, namely, the so-called secondary microseisms at seismic periods around 7 s (Peterson,  
69 1993), are excited by the nonlinear interactions between nearly equal-frequency ocean waves  
70 propagating in nearly opposite directions (Longuet-Higgins, 1950; Hasselmann, 1963). The  
71 periods of the excited secondary microseisms are half those of the colliding ocean waves. The  
72 excitation source is equivalent to a vertical pressure applied to the water surface, which is  
73 proportional to the product of the heights of the opposing equal-frequency waves. Due to this  
74 second order relation, moderate sea states can sometimes generate loud microseism noise  
75 (Obrebski et al., 2012). Thus, the presence of a strong microseism event does not necessarily  
76 imply a locally intense sea state.

77 By coupling the excitation theory of secondary microseisms proposed by Longuet-  
78 Higgins (1950) with the ocean wave action model, Kedar et al. (2008) modeled the secondary  
79 microseism excitations in the north Atlantic, and validated the numerical modeling by comparing  
80 with inland seismological observations. Afterwards, more authors simulated the oceanic

81 microseism sources and some reported the consistency between predictions and observations  
82 (e.g., Arduin et al., 2011, 2015; Hillers et al., 2012; Stutzmann et al., 2012; Nishida & Takagi,  
83 2016). Stopa et al. (2019) compared the microseism simulations with real observations to  
84 validate their corrections to the global reanalysis wind fields, which systematically reduced the  
85 residuals in the wave hindcast over the past decades.

86 The seismic excitation by an oceanic microseism source is essentially akin to that by an  
87 earthquake, in that the seismic wavefield recorded at any point is a convolution of the source  
88 time function with the Green function of the propagating medium between source and receiver.  
89 Their main difference lies in the source process. For earthquakes, the sudden rupture of faults  
90 leads to short-duration, impulsive source time functions. Isolated seismic phases are generally  
91 distinguishable from the seismograms. In contrast, the excitation of microseisms, approximated  
92 as Gaussian random process by some authors (Peterson, 1993; Steim, 2015), is incessant, leading  
93 to long, random-like source time functions. The convolution mixture signals are not directly  
94 discernible from the seismograms. With array beamforming (Rost & Thomas, 2002) or  
95 correlation technique (Campillo & Paul, 2003; Shapiro & Campillo, 2004), specific phases from  
96 distant microseism sources have been identified from microseism noise records (e.g., Gerstoft et  
97 al., 2008; Landès et al., 2010; Zhang et al., 2010; Euler et al., 2014; Reading et al., 2014; Gal et  
98 al., 2015; Liu et al., 2016; Nishida & Takagi, 2016; Meschede et al., 2017, 2018; Retailleau &  
99 Gualtieri, 2019). The correlation technique is advantageous in that, by correlating the noise  
100 records at two receivers, explicit seismic signals can be derived. Noise-derived surface waves  
101 have been used to infer the azimuthal and seasonal changes of noise sources (e.g., Stehly et al.,  
102 2006). Noise-derived body waves can provide better constrains in imaging the noise sources  
103 (Landès et al., 2010). Recently, deep body waves that propagate through the mantle and core  
104 have been extracted from ambient noise (e.g., Boué et al., 2013; Lin et al., 2013; Nishida, 2013;  
105 Poli et al., 2015; Xia et al., 2016; Spica et al., 2017; Retailleau et al., 2020). The noise-derived  
106 body waves are valuable for surveying the deep structure and for understanding the links  
107 between seismological observations and atmospheric/oceanographic phenomena.

108 Hillers et al. (2012) made the first global-scale comparison between the oceanic  
109 microseism sources derived from seismological observations and oceanographic modeling. The  
110 seismologically derived data (time resolution: 13 days; spatial resolution:  $2.5^\circ$  latitude  $\times$   $5^\circ$   
111 longitude) are the global back-projections of near-zero-lag  $P$  signals generated from the cross  
112 correlations of microseism  $P$  waves at seismic array (Landès et al., 2010). The modeled data  
113 (time resolution: 3 hours; spatial resolution:  $1^\circ$  latitude  $\times$   $1.25^\circ$  longitude) are a global extension  
114 of the numerical simulation by Kedar et al. (2008). The two datasets are resampled to common  
115 resolutions for comparison. For the seismologically derived data, the back-projection is based on  
116 the relationship between the source-receiver distance and the horizontal slowness of teleseismic  
117  $P$  wave. However, seismic phases that have common slownesses (e.g.,  $P$  and  $PP$  waves) cannot  
118 be discriminated in this method (Gerstoft et al., 2008; Landès et al., 2010). Thus, the imaged  
119 sources are somewhat ambiguous. For the modeled data, coastal reflections of ocean waves, that  
120 can play a role in the ocean wave-wave interactions at near-coast regions (Longuet-Higgins,  
121 1950; Arduin et al., 2011), are neglected. Due to the resonance of seismic waves in the water  
122 columns, bathymetry can have significant effect on the excitation of microseisms (Longuet-  
123 Higgins, 1950; Kedar et al., 2008; Hillers et al., 2012). The importance to account for the  
124 bathymetric effect on the microseism  $P$ -wave excitations has been addressed in several studies  
125 (e.g., Euler et al., 2014; Gal et al., 2015; Meschede et al., 2017). Hillers et al. (2012) considered

126 the bathymetric effect, but using the amplification factors derived by Longuet-Higgins (1950) for  
127 surface waves.

128 Rasche and Ardhuin (2013) established an oceanographic hindcast database that includes  
129 global oceanic secondary microseism sources of a 3-hour time resolution and a  $0.5^\circ$  spatial  
130 resolution. Coastal reflections were accounted for in the modeling (Ardhuin et al., 2011).  
131 Regarding the bathymetric effect on microseism excitations, Gualtieri et al. (2014) proposed the  
132 formulae for body waves based on ray theory. Concerning the localization of noise sources, Li et  
133 al. (2020) developed a double-array method that can estimate the respective slownesses of the  
134 interfering waves, and thereby, provide better constrains for the determination of the correlated  
135 seismic phases. The microseism sources that are effective for the derivation of seismic signals  
136 from noise records, can be mapped by back-projecting the noise-derived signals along the ray  
137 paths of the correlated phases. The double-array configuration eliminates the ambiguity in  
138 determining the effective source region (Fresnel zone). In this study, we integrate these new  
139 progresses to survey the associations of noise-derived body waves to ocean wave activity and  
140 microseism excitations.

141 This paper is organized as follows. In section 2, we review the main results of Li et al.  
142 (2020) who reported the observation of a prominent  $P$ -type phase from the noise correlations  
143 between two regional seismic networks at teleseismic distance. The noise-derived phase has its  
144 spectral content concentrated in the period band of the secondary microseisms that are excited by  
145 the nonlinear ocean wave-wave interactions. In this paper, we denote the phase as  $P_{\text{dmc}}$ ,  
146 corresponding to the fact that the phase originates from the correlation between the Direct  $P$   
147 waves that transmit through the deep Mantle and the outer Core (microseism  $P$  and  $PKPab$   
148 waves). In section 3, we estimate the temporal variations in the  $P_{\text{dmc}}$  amplitude and refute the  
149 associations to seismicity. In section 4, correlation analysis is used to unveil the spatiotemporal  
150 links of the  $P_{\text{dmc}}$  signal with the global oceanic wave climate and microseism sources. Last, we  
151 discuss the significance of this study in seismology, oceanography and climate science.

## 152 **2 Noise-derived $P_{\text{dmc}}$ phase**

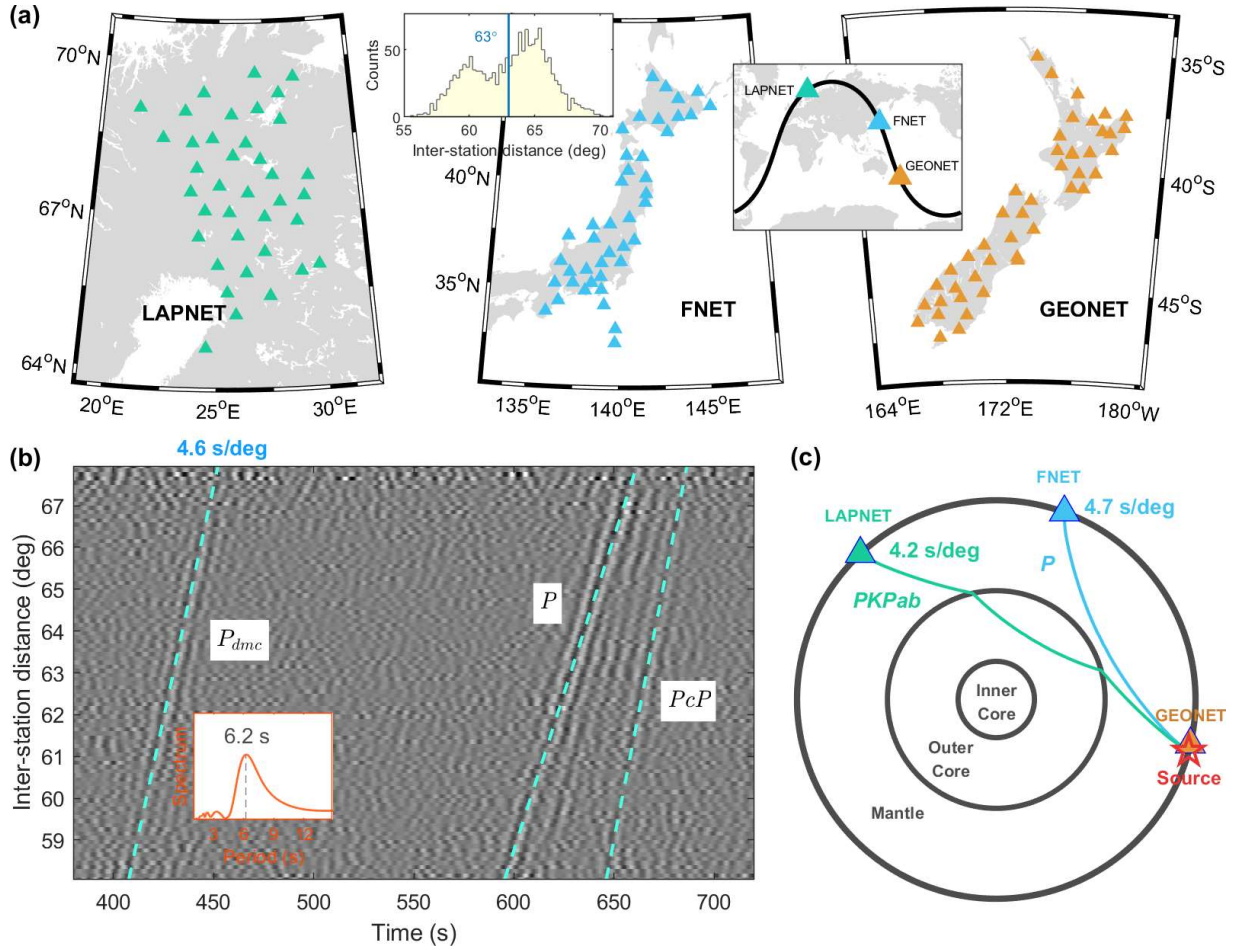
153 Li et al. (2020) correlated the seismic noise records from two regional seismic networks  
154 at teleseismic distance: the FNET array in Japan and the LAPNET array in Finland (Fig. 1a). The  
155 continuous seismograms were divided into 4 h segments and whitened in the frequency domain.  
156 Segments with large spikes (like earthquakes) were discarded. The available segments of each  
157 FNET-LAPNET station pairs were correlated. For more technical details, see section 2 of Li et  
158 al. (2020). From the vertical-vertical components of the FNET-LAPNET noise correlations, they  
159 observed coherent spurious arrivals (the  $P_{\text{dmc}}$  phase named in the previous section) that emerged  
160  $\sim 200$  s earlier than the direct  $P$  waves (Fig. 1b). By estimating the respective slownesses of the  
161 interfering waves and their time delay, it is unveiled that a quasi-stationary phase interference  
162 between the teleseismic  $P$  waves at FNET and the  $PKPab$  waves at LAPNET, emanating from  
163 noise sources in the ocean south of New Zealand (NZ), lead to the noise-derived  $P_{\text{dmc}}$  phase (Fig.  
164 1c). The quasi-stationary phase condition refers to that the interfering waves have no common  
165 path or common slowness, but the stack of correlation functions over a range of sources can still  
166 be constructive as an effect of finite frequency. This observation contrasts with the strict  
167 stationary phase condition that has been employed by Pham et al. (2018) to explain the spurious  
168 body phases in the earthquake coda correlations. The strict condition implies the existence of  
169 sources in the stationary-phase region, or say, the correlated waves have common ray paths or

170 common slownesses. Li et al. (2020) substantiates the explanation of quasi-stationary phase for  
171 the observed  $P_{\text{dmc}}$  signals with numerical experiments based on ray theory and based on spectral-  
172 element modeling, and highlighted the discrepancies between (microseism) noise correlations  
173 and coda correlations.

174 The  $P_{\text{dmc}}$  phase has an apparent slowness of 4.6 s/deg, while the slownesses of the  
175 interfering  $P$  and  $PKPab$  waves are 4.7 s/deg and 4.2 s/deg, respectively. The dominant period of  
176 the  $P_{\text{dmc}}$  phase is 6.2 s, typical for secondary microseisms. The observation of the  $P_{\text{dmc}}$  phase is  
177 time-asymmetric (Fig. S1a). Its absence from the mirror side is ascribed to the faintness of the  
178 corresponding source in the low-latitude Atlantic (Fig. S1b).

179 There are several advantages to investigating the links between noise-derived signals and  
180 microseism sources with the  $P_{\text{dmc}}$  phase. First, the correlated  $P$  and  $PKPab$  waves are both  
181 prominent phases in the ballistic microseism wavefields. The  $P_{\text{dmc}}$  phase is easily observable  
182 from noise correlations, even between some single station pairs and on some single days (Fig.  
183 S2). Second, the isolation of  $P_{\text{dmc}}$  signals avoids potential bias caused by other prominent  
184 signals. Third, the effective sources are confined in a limited, unique region (Fresnel zone). In  
185 contrast, noise-derived surface waves have a broad Fresnel zone around the line across the  
186 correlated stations, and noise-derived  $P$  waves can have multiple Fresnel zones (see fig. 5 of  
187 Boué et al., 2014 for instance). The uniqueness of the effective source region can facilitate the  
188 study on the correlation between the noise-derived signals and the effective sources. Fourth, the  
189 correlated FNET and LAPNET networks are next to the northern Pacific and Atlantic,  
190 respectively, while the effective source region locates in the southern Pacific. The northern  
191 oceans have consistent seasonal variation pattern distinct from (reverse to) that of the southern  
192 oceans (Stehly et al., 2006; Stutzmann et al., 2009; Landès et al., 2010; Hillers et al., 2012;  
193 Reading et al., 2014; Turners et al., 2020). These geographical configurations make the  
194 observations easier to interpret. Last, there happens to be a seismic array (GEONET) in NZ next  
195 to the effective source region for the  $P_{\text{dmc}}$  phase. The seismic data from GEONET provide extra  
196 support to our study.

197



198

199 **Figure 1.** (a) Three regional broadband seismic networks used in this study: left, the LAPNET  
 200 array in Finland (38 stations); center, the FNET array in Japan (41 stations); right, the GEONET  
 201 array in New Zealand (46 stations). The histogram inset shows the distribution of the separation  
 202 distances between the 1558 FNET-LAPNET station pairs. The center-to-center distance is  $63^\circ$   
 203 between LAPNET and FNET, and  $85^\circ$  between FNET and GEONET. The global inset shows the  
 204 geographical locations of the three networks that are aligned on a great circle (dark line). (b)  
 205 Annual FNET-LAPNET noise correlations that are filtered between 5 s and 10 s and stacked  
 206 over time and in  $0.1^\circ$  inter-station distance bins. The spectrum inset indicates that the  $P_{dmc}$  phase  
 207 has a 6.2 s peak period. (c) Ray paths of the interfering waves that generate the  $P_{dmc}$  phase. The  
 208 effective source region is close to GEONET.

### 209 3 Temporal variations

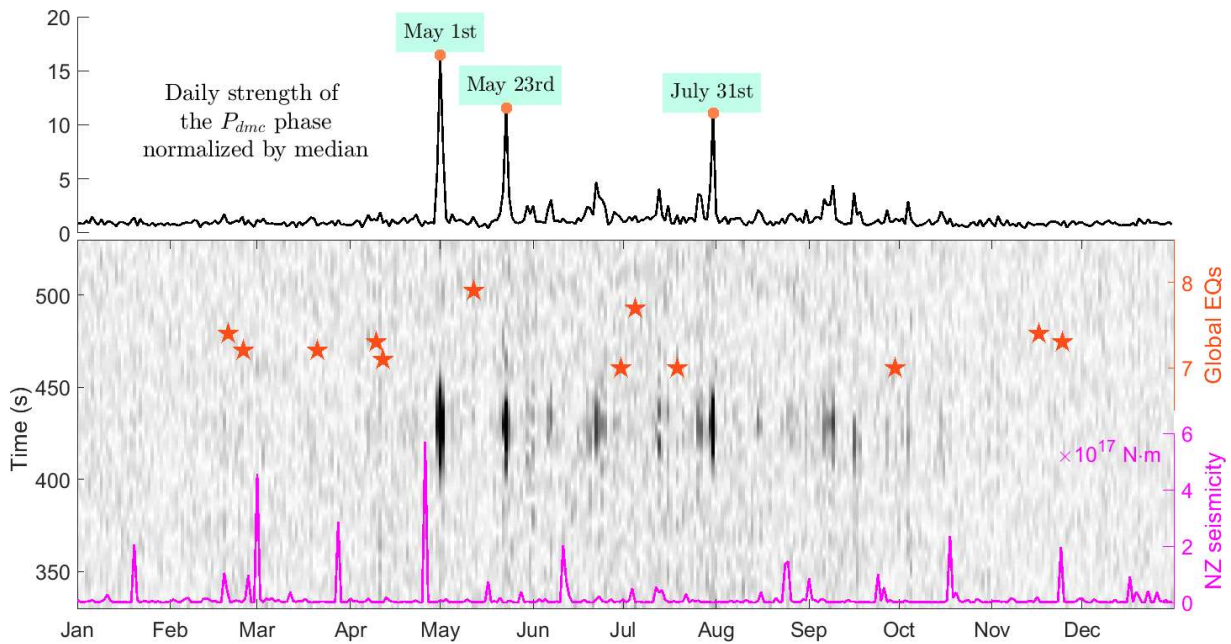
210 We extract the temporal variations of the  $P_{dmc}$  signals by beamforming the FNET-  
 211 LAPNET noise correlations on a daily basis. The daily noise correlations are shifted and stacked  
 212 by

$$213 B(t) = \langle C_{ij}(t + (d_{ij} - d_0) \cdot p) \rangle, \quad (1)$$

214 with  $\langle \cdot \rangle$  the mean operator,  $C_{ij}$  and  $d_{ij}$  the correlation function and the distance between the  $i$ th  
 215 FNET station and the  $j$ th LAPNET station,  $d_0$  the reference distance ( $63^\circ$ ),  $p$  the apparent

216 slowness of the  $P_{\text{dmc}}$  phase (4.6 s/deg), and  $t$  the time. The image in Fig. 2 shows the envelopes  
 217 of the daily beams computed from the Hilbert transform of Eq. (1), with the daily  $P_{\text{dmc}}$  strength  
 218 by averaging the envelope amplitudes plotted in the top panel. The strength of daily  $P_{\text{dmc}}$  signals  
 219 varies strikingly, extremely strong on some single days (see the labeled dates in the  $P_{\text{dmc}}$  strength  
 220 curve for examples), but indiscernible on most other days.

221 Considering that the region of effective source is tectonically active, one should  
 222 investigate the plausible connection between the  $P_{\text{dmc}}$  signals and seismicity. From Fig. 2, it is  
 223 obvious that  $P_{\text{dmc}}$  is decorrelated with the NZ seismicity. Also, it shows no connection with  
 224 global large earthquakes as has been observed for coda-derived core phases at periods of 20 to 50  
 225 s (Lin & Tsai, 2013; Boué et al., 2014). That again demonstrates the substantial difference  
 226 between ambient noise correlations and earthquake coda correlations, as emphasized by Li et al.  
 227 (2020). The  $P_{\text{dmc}}$  strength exhibits an obvious pattern of seasonal variation. The seasonal pattern  
 228 does not favor a tectonic origin because of the lack of a seasonal pattern in seismicity. Instead, an  
 229 oceanic origin is more favored because of the well-documented fact that oceanic wave activity  
 230 and microseism excitations show similar seasonal pattern: more powerful during the local winter  
 231 (e.g., Stehly et al., 2006; Stutzmann et al., 2009; Landès et al., 2010; Hillers et al., 2012; Reading  
 232 et al., 2014). Next, we analyze the correlations between  $P_{\text{dmc}}$  signals and oceanographic data at a  
 233 global scale.



234

235 **Figure 2.** Temporal variations in the strength of daily  $P_{\text{dmc}}$  signals, in comparisons with the daily  
 236 cumulative seismic moments in NZ (pink line at bottom; for earthquake magnitudes above 2.0 in  
 237 GEONET catalogue) and global large earthquakes (stars; magnitudes above 7.0 in USGS  
 238 catalogue; see Table S1 for a full list of earthquakes in 2008 above magnitude 5.5). The  
 239 background image is composed of columns of daily envelopes of beamed FNET-LAPNET noise



240 correlations. Darker color represents larger amplitude. The curve on the top shows the daily  $P_{\text{dmc}}$   
 241 strength derived from the daily envelopes. Dates of the three largest peaks are labeled.

#### 242 **4 Correlation analysis**

243 The sea state is composed of ocean waves at various frequencies and propagation  
 244 directions. The nonlinear interaction between nearly equal-frequency ocean waves traveling in  
 245 nearly opposite directions is equivalent to a vertical random pressure applied to the ocean surface  
 246 (Longuet-Higgins, 1950; Hasselmann, 1963), so that microseisms are generated. Figure 3(a)  
 247 shows a global map of average Power Spectral Density (PSD) of the equivalent surface pressure  
 248 for a seismic period of 6.2 s, during the northern winter months of 2008. The hindcast PSD data  
 249 are simulated by Arduin et al. (2011) and Rasclé & Arduin (2013), based on the microseism  
 250 excitation theory of Longuet-Higgins (1950) and Hasselmann (1963). The most energetic  
 251 microseism excitations occur in the northern Atlantic south of Greenland and Iceland (near  
 252 LAPNET), and in the northern Pacific between Japan and Alaska (near FNET). Figure 3(b)  
 253 shows the map for the austral winter months, with the strongest excitations occurring between  
 254 NZ and Antarctic (near GEONET). The seasonal pattern of oceanic microseism excitations  
 255 results from the same pattern of global wave climate (Figs 3e-f). The seasonal pattern of the  $P_{\text{dmc}}$   
 256 strength agrees with that of the microseism excitation and wave climate in the effective source  
 257 region south of NZ.

258 We compute the correlation coefficient (denoted as  $r$ ) between the  $P_{\text{dmc}}$  strength and the  
 259 source PSDs at each grid point, and thereby obtain a global correlation map (Fig. 3c). The largest  
 260  $r$  value for  $P_{\text{dmc}}$  and source PSD arises at [47°S, 177°E] in the effective source region ( $E$  in Fig.  
 261 3c). The corresponding time series of daily source PSDs is plotted in Fig. 4, in parallel with the  
 262  $P_{\text{dmc}}$  strength. Large peaks in the  $P_{\text{dmc}}$  series have good correspondence with large peaks in the  
 263 source PSD series. From Fig. 3(c), one can observe a broad region of positive  $r$  values (red  
 264 colors; roughly, south Atlantic, south Pacific, and Indian ocean). However, the positive  
 265 correlation does not imply a causality between the  $P_{\text{dmc}}$  phase and the sources outside the  
 266 effective region  $E$ . We ascribe the apparent positive correlation to the spatial correlation of the  
 267 time-varying microseism excitation. As shown in Fig. 3(d), the source at [47°S, 177°E] in region  
 268  $E$  exhibits a similar pattern of apparent correlations with global sources as in Fig. 3(c). Despite  
 269 the microseism excitations at varying locations are independent (Hasselmann, 1963), we note  
 270 that the independence refers only to the phase information. The time variations of microseism  
 271 source power are spatially associated. That is not surprising since the interacting ocean waves  
 272 that excite microseisms could be driven by the same storms and swells can propagate freely over  
 273 thousands of kilometers away (Arduin et al., 2009). We also notice there are high- $r$  regions that  
 274 may not be fully explained by the spatial association. These regions are characterized by low  
 275 intensity of microseism excitations in Figs 3(a-b). A striking example is around [12°N, 88°E] in  
 276 the Bay of Bengal ( $F$  in Fig. 3c). From Fig. 4, it can be seen that the source PSD series for  
 277 [12°N, 88°E] is dominated by a single peak around May 1<sup>st</sup>, coincident with the largest  $P_{\text{dmc}}$   
 278 peak. This coincidence leads to a high value of correlation coefficient. However, the Bay of  
 279 Bengal is far away from the FNET-LAPNET great circle, which is inconsistent with the source  
 280 imaging shown later in Fig. 5. Thus, the high correlation is spurious and does not imply a  
 281 causality relationship between the microseism sources in the Bay of Bengal and the  $P_{\text{dmc}}$  signals.  
 282 Figures 3(g-h) show the correlation maps for  $h_s$ , which will be discussed later.

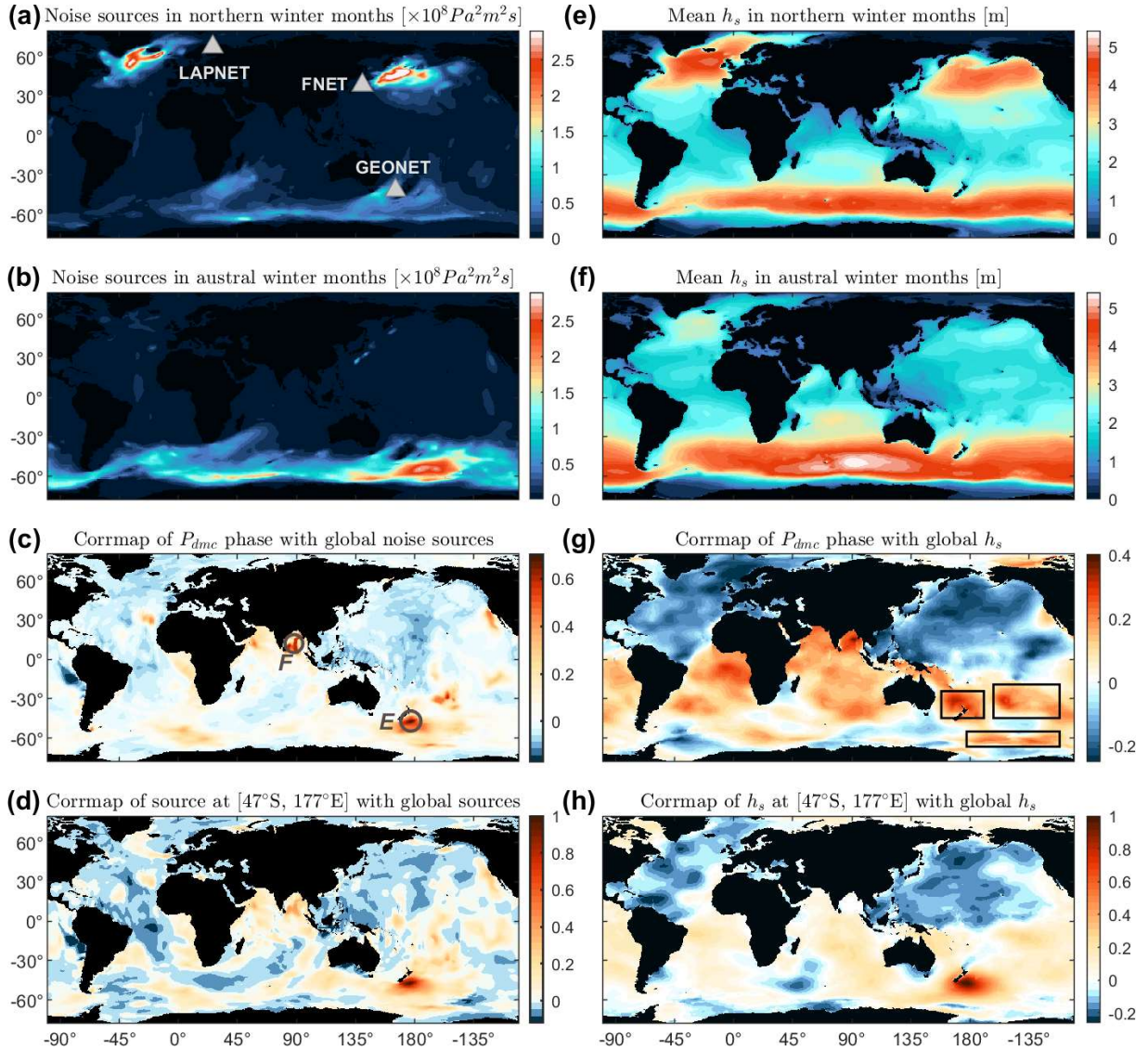
283 As shown in Fig. 4, prominent peaks in the  $P_{\text{dmc}}$  series have correspondence in the source  
284 PSD series for the effective source at [47°S, 177°E]. However, there are some peaks in the latter  
285 without correspondence in the former (see the labeled dates in Fig. 4b for examples). Note that  
286 here the  $P_{\text{dmc}}$  strength is compared to the microseism source PSD at single point in Fig. 4,  
287 whereas the effective sources spread over a region. One needs to verify if the peak disparities  
288 observed from Figs 4(a-b) can be ascribed to the neglect of the spreading of the effective source  
289 region. To evaluate an overall microseism excitation in the effective source region, the  
290 bathymetric effect on  $P$ -wave excitation should be considered (in the previous analysis for single  
291 point locations, the consideration of bathymetric effect is unnecessary because a scaling over the  
292 source PSD series does not change the value of the correlation coefficient between  $P_{\text{dmc}}$  and  
293 source PSD). Using the equations proposed by Gualtieri et al. (2014) and the bathymetry around  
294 NZ (Fig. 5a), we compute the bathymetric amplification factors for  $P$  waves at a period of 6.2 s  
295 (Fig. 5b; see Fig. S3 for comparisons between the factors calculated following Longuet-Higgins,  
296 1950 and Gualtieri et al., 2014). The factors vary largely with locations. Also, note that the  $P_{\text{dmc}}$   
297 phase has different sensitivity to the sources in the effective region, or say, the sources make  
298 varying contributions to the  $P_{\text{dmc}}$  signal. The power of sources should be weighted in the  
299 averaging. We obtain the weights by back-projecting the beam power of noise correlations onto a  
300 global grid (Fig. 5c; see Supplementary for technical details). Figure 5(d) shows the map of  
301 annually averaged source PSDs surrounding NZ and Fig. 5(e) shows the map after the  
302 modulation of the bathymetric amplification factors in Fig. 5(b). The spatial patterns are altered  
303 significantly, indicating the importance to account for the bathymetric effect. The final source  
304 imaging that has been weighted by Fig. 5(c), is plotted in Fig. 5(f). It agrees well with the  
305 effective source region  $E$  determined from the correlation map in Fig. 3(c). Replacing the annual  
306 PSD map in Fig. 5(d) with daily PSD maps, we obtain maps like Fig. 5(f) for each date.  
307 Averaging over the map leads to the time series of daily intensity in the effective source region  
308 (labeled as effective source intensity in Fig. 6). Averaging over a wide region has the advantage  
309 that the effects of potential source location errors due to the simplification of Earth model for fast  
310 travel time calculation, which have been addressed in some single array back-projection studies  
311 (e.g., Gal et al., 2015; Nishida & Takagi, 2016), can be largely reduced. From Fig. 6, one can see  
312 that the new effective source intensity series has almost the same peaks as the source PSD series  
313 for [47°S, 177°E] in Fig. 4(b), suggesting that the observed peak disparities are caused by other  
314 reasons. Next, we investigate if the disparities are caused by errors in the simulation of hindcast  
315 data or if there are other physical explanations.

316 The microseism source PSD data are simulated from the hindcast data of ocean wave  
317 directional spectra base on the excitation theory of Longuet-Higgins (1950) and Hasselmann  
318 (1963), which have no constrains from seismological observations. One should consider the  
319 accuracy of the simulation: can we ascribe the peak disparities in Fig. 4 to the simulation error or  
320 not? The seismic noise records from the GEONET array adjacent to the effective source region  
321 provide the opportunity to validate the simulation. To obtain the daily microseism noise levels at  
322 GEONET, we apply the Hampel filter, a variant of the classic median filter, to the continuous  
323 seismograms to discard earthquakes and anomalous impulses. The filter replaces outliers with  
324 the medians of the outliers' neighbors and retains the normal samples. Technical details are  
325 provided in section S4 of the Supplementary. The resultant GEONET noise level exhibits a good  
326 correlation with the effective source intensity ( $r = 0.7$ ). We thus deem that the numerical  
327 simulations are statistically reliable. When the effective source intensity is high, the GEONET  
328 noise level should also be high (see the peaks marked by dots in Fig. 6 for examples). However,

329 due to the great spatiotemporal variability of noise sources in the effective region and the  
330 complexity of seismic waves propagating from ocean to land (Ying et al., 2014; Gualtieri et al.,  
331 2015), a larger peak in the source intensity series does not necessarily imply a larger peak in the  
332 noise level time series (e.g., see diamonds in Fig. 6 for examples). We also emphasize that a high  
333 GEONET noise level does not need to always have a correspondence in the source intensity (see  
334 squares in Fig. 6 for example), because the GEONET stations record microseisms emanating  
335 from noise sources all around, not only from the effective source region.

336 The above analysis explains the observed disparities between the  $P_{\text{dmc}}$  strength and the  
337 effective source intensity. From Fig.6, one can see that the disparities primarily emerge in the  
338 shaded period when dominant microseism sources shift to the north hemisphere. The shading  
339 roughly separates the northern winter from the austral winter. The correlation between  $P_{\text{dmc}}$   
340 strength and effective source intensity is low in the shaded period ( $r = 0.16$ ), in contrast to the  
341 high correlation during the unshaded period ( $r = 0.74$ ). Large  $P_{\text{dmc}}$  peaks always emerge on dates  
342 during the austral winter when the effective source intensity is much higher than its median, and  
343 meanwhile, noise levels at FNET and LAPNET are below their respective medians (see dots in  
344 Fig. 6 for examples). The seasonal variations of oceanic sources in the southern hemisphere are  
345 less strong than in the northern hemisphere (Fig. 3). On some dates (see triangles in Fig. 6 for  
346 examples), the effective source intensity can be considerable, but relevant  $P_{\text{dmc}}$  peaks are still  
347 missing. We notice that the corresponding microseism levels at FNET and LAPNET are  
348 obviously above their medians. Intensive ocean activity and microseism excitations in the north  
349 Pacific and Atlantic, lead to increased microseism noise levels at FNET and LAPNET. The  $P_{\text{dmc}}$   
350 strength is anti-correlated with microseism noise levels at FNET ( $r = -0.12$ ) and LAPNET ( $r = -$   
351  $0.18$ ). We hereby conjecture that the microseism energy from the distant effective source region  
352 is dwarfed by the energetic microseisms excited by oceanic sources closer to the correlated  
353 FNET and LAPNET arrays, and consequently,  $P_{\text{dmc}}$  signals are overwhelmed by the background  
354 noise in the FNET-LAPNET cross-correlations. Last, we mention that the median threshold in  
355 Fig. 6 separates the major features of the time series described above, but there is no guarantee  
356 that it is a perfect threshold due to the nonlinear relationships between the  $P_{\text{dmc}}$  strength and the  
357 noise levels at the arrays.

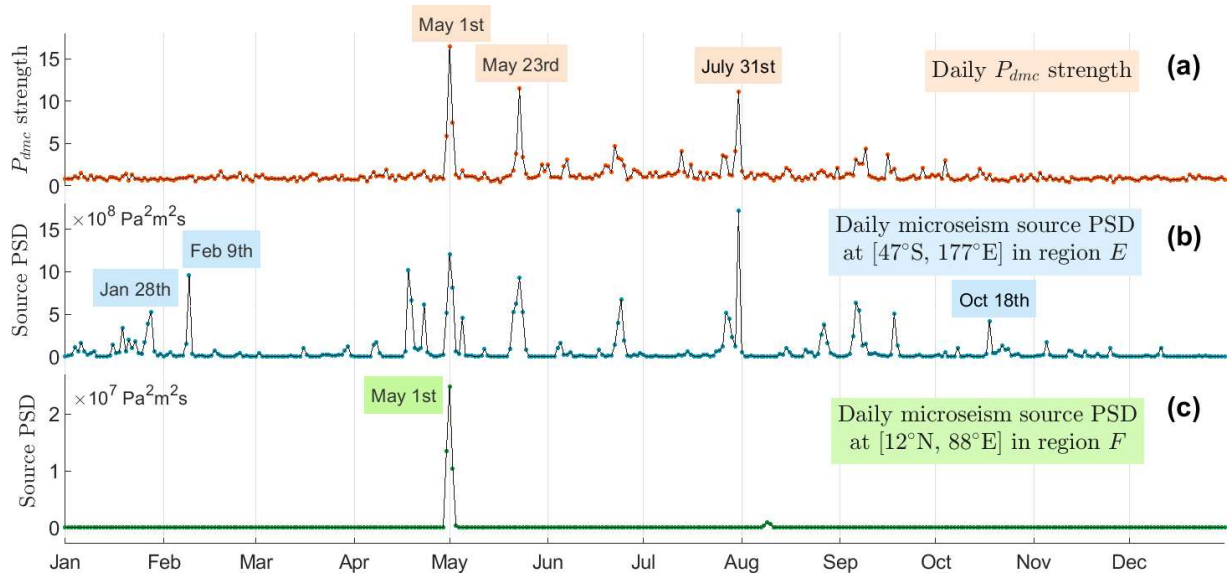
358



359

360 **Figure 3.** (a) Global map of average PSD of oceanic microseism sources in 2008 northern winter  
 361 months (Jan. to Mar. and Oct. to Dec.), for a seismic period of 6.2 s. (b) Similar to (a) but for  
 362 2008 austral winter months (Apr. to Sep.). (c) Correlation map (corrmap) for the  $P_{dmc}$  strength  
 363 and global microseism noise sources. Circles mark two regions with highest correlation  
 364 coefficients:  $E$ , effective source region surrounding  $[47^\circ\text{S}, 177^\circ\text{E}]$  south of NZ;  $F$ , fake highly-  
 365 correlated region surrounding  $[12^\circ\text{N}, 88^\circ\text{E}]$  in the Bay of Bengal. (d) Correlation map for the  
 366 source at  $[47^\circ\text{S}, 177^\circ\text{E}]$  and global sources. (e) Mean significant wave height ( $h_s$ ; four times the  
 367 square root of the zeroth-order moment of ocean-wave frequency spectrum) in northern winter  
 368 months. (f) Similar to (e) but for austral winter months. (g) Correlation map for the  $P_{dmc}$  strength  
 369 and global wave heights. (h) Correlation map for wave heights at  $[47^\circ\text{S}, 177^\circ\text{E}]$  and global wave  
 370 heights. The oceanographical hindcast data are provided by the IOWAGA products (Rascle &  
 371 Ardhuin, 2013).

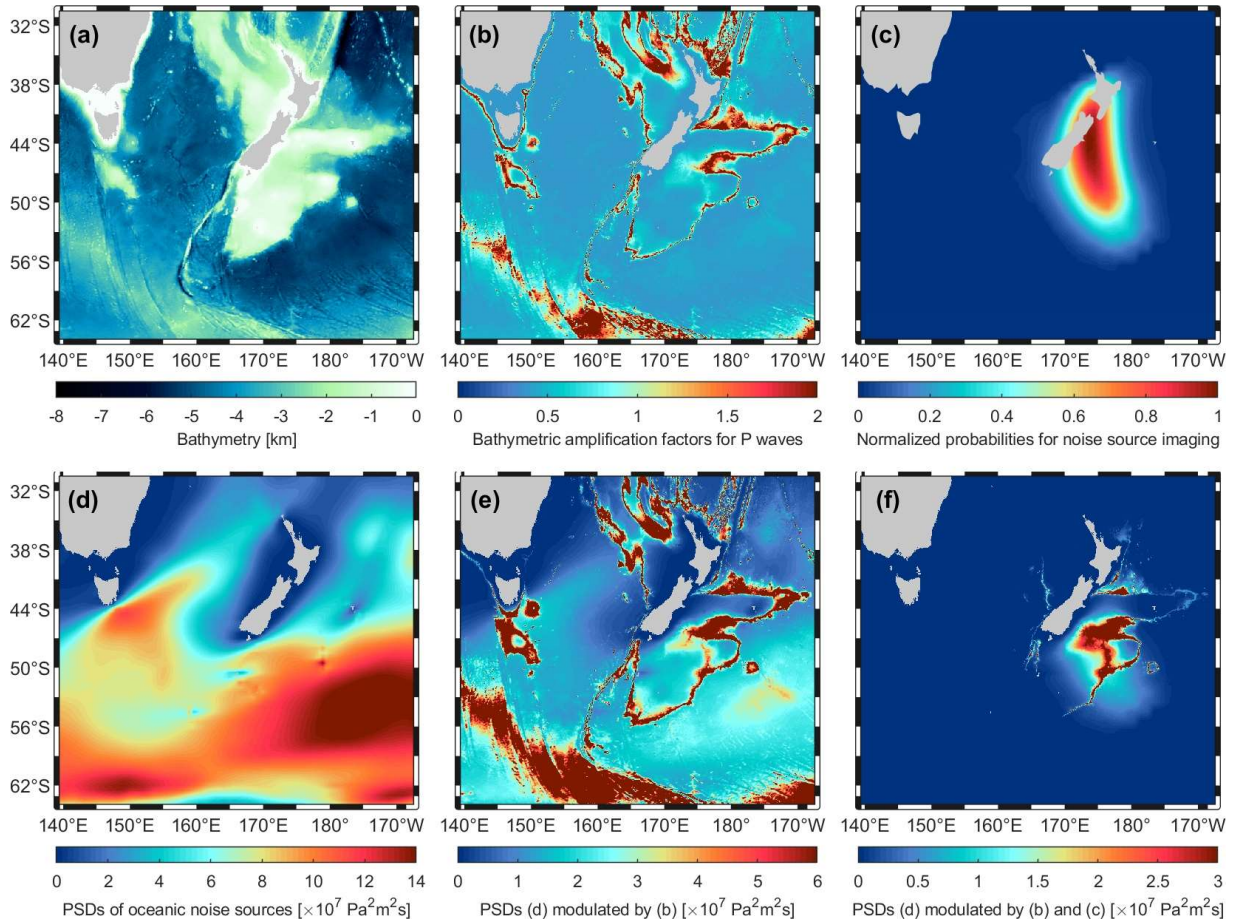
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373

374 **Figure 4.** True correlation ( $r = 0.73$ ) between (a) the  $P_{dmc}$  strength from Fig. 2 and (b) the power  
 375 of source at  $[47^\circ\text{S}, 177^\circ\text{E}]$  in the effective source region ( $E$  in Fig. 3c), and spurious correlation  
 376 ( $r = 0.71$ ) between  $P_{dmc}$  and (c) the power of source at  $[12^\circ\text{N}, 88^\circ\text{E}]$  in the Bay of Bengal ( $F$  in  
 377 Fig. 3c).

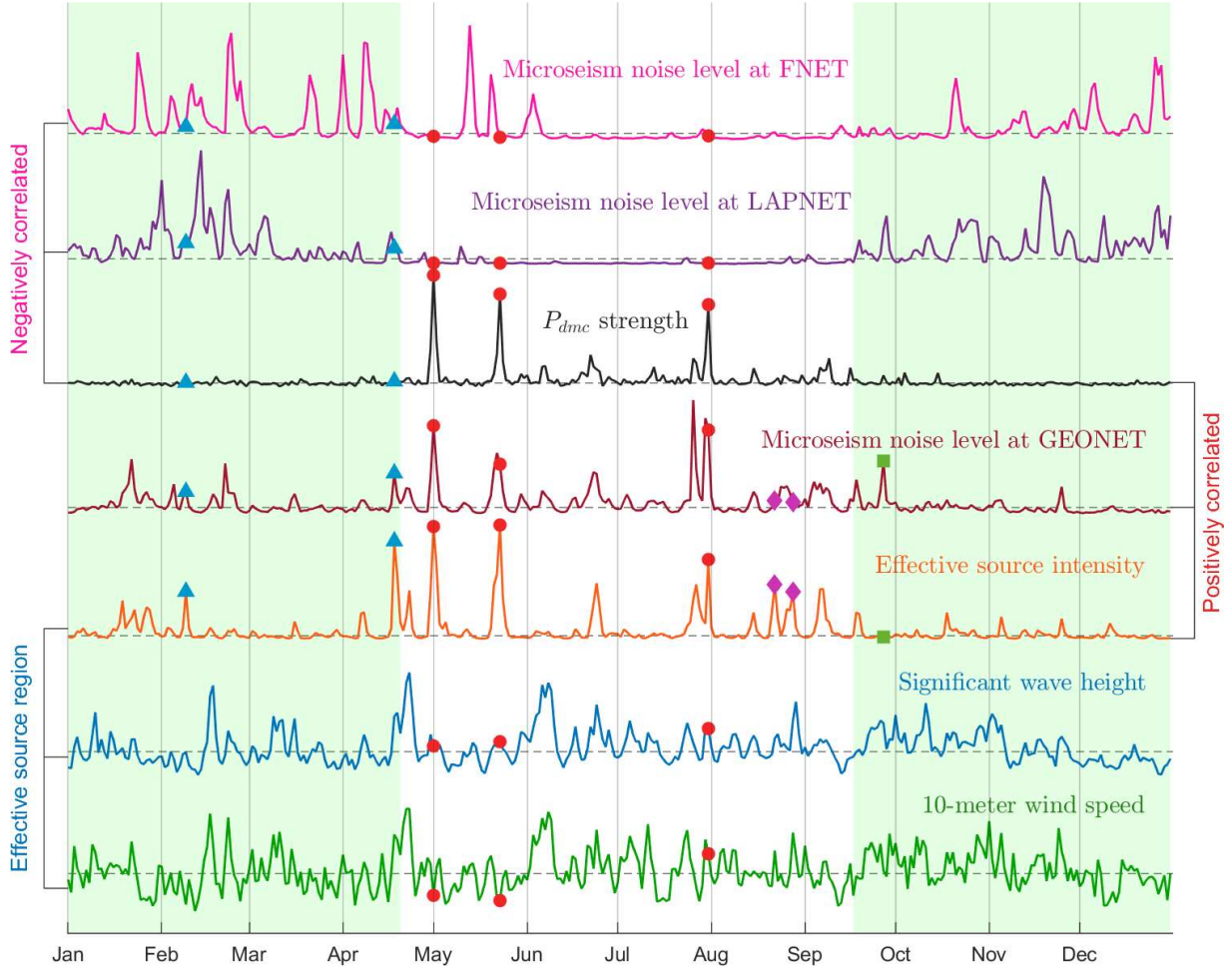
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379

380 **Figure 5.** (a) Bathymetry around NZ. (b) Bathymetric amplification factors for *P*-type waves.  
 381 (c) Imaging of effective sources obtained from the back-projection of the FNET-LAPNET noise  
 382 correlations. (d) Annual average of source PSDs in 2008. (e) Source PSDs in (d) modulated by  
 383 the factors in (b). (f) Source PSDs in (e) further modulated by the weights in (c).

384



385

386 **Figure 6.** Temporal variations of daily  $P_{dmc}$  strength, microseism noise levels at three networks,  
 387 and average wind speeds, wave heights and microseism excitations in the effective source  
 388 region. The curves are normalized by their own maximums. Dashed horizontal lines denote their  
 389 respective medians. Symbols mark some dates cited in the main text. When computing the  
 390 effective source intensity, the bathymetric factors in Fig. 5(b) and weights in Fig. 5(c) are used.  
 391 When computing the average wind speeds and wave heights, weights in Fig. 5(c) are used.

## 392 5 Discussions and conclusions

393 In this study, we explore the relations between the noise-derived  $P_{dmc}$  signals and global  
 394 oceanic microseism sources using spatiotemporal correlation analysis. The effective source  
 395 region  $E$  for the  $P_{dmc}$  phase is successfully identified from the correlation map in Fig. 3(c), which  
 396 is consistent with that determined from the seismological back-projection in Fig. 5(c). The  
 397 correlation map provides a convenient way to identify the effective sources of noise-derived  
 398 seismic signals.

399 In our case, the seismic networks used for noise correlation are located in the northern  
 400 hemisphere, while the effective source region is in the southern hemisphere. Ideally, we expect a  
 401 correlation map with the following features: positive correlation with sources in the effective  
 402 region, and negative or insignificant correlations with other inefficient sources. Positive

403 correlation indicates a contribution to the construction of  $P_{\text{dmc}}$  signal from noise correlations,  
 404 negative correlation implies an adverse impact, and insignificant correlation (decorrelation)  
 405 means a negligible effect on the signal construction. However, we obtained a correlation map  
 406 roughly showing that, the  $P_{\text{dmc}}$  signal is correlated with the southern sources and anti-correlated  
 407 with the northern sources. The correlation with southern sources outside the effective region can  
 408 be interpreted with the spatiotemporal correlation of the power of the microseism sources in the  
 409 southern oceans, due to the large span of ocean storms and the long-range propagation of swells.  
 410 The anti-correlation with the northern sources, can partly be explained by the well-known  
 411 reverse seasonal patterns of oceanic microseism excitations in the south and north hemispheres  
 412 (Stutzmann et al., 2009; Landès et al., 2010; Hillers et al., 2012; Reading et al., 2014). Another  
 413 important reason is that compared to the remote effective sources in the south hemisphere, the  
 414 northern sources closer to the correlated stations have larger impacts on the microseism noise  
 415 levels at stations. Strong energy flux from the northern sources outshines the microseism energy  
 416 coming from the distant effective sources. That deteriorates the construction of the  $P_{\text{dmc}}$  phase.  
 417 The noise-derived  $P_{\text{dmc}}$  signals are primarily observable in the austral winter. That can be, on one  
 418 hand, attributed to the stronger effective source intensity during that period, and on the other  
 419 hand, to the relative tranquility in the northern oceans.

420 In Fig. 7, we summarize the classification of noise sources, the decomposition of  
 421 wavefields, and the associations to the constituents of the inter-receiver noise correlation  
 422 function. The diagram of Fig. 7(a) explains the relationships using the case study of the  $P_{\text{dmc}}$   
 423 phase discussed above. We generalize Fig. 7(a) to the derivation of an arbitrary signal (referred  
 424 to as the target signal for convenience) from ambient noise wavefields (Fig. 7b). The noise  
 425 correlation function is composed of the target signal, any other signals and background noise. A  
 426 source or a wave is called effective if it contributes to the construction of the target signal from  
 427 noise correlations. Otherwise, it is called ineffective. The construction of the target signal is  
 428 exclusively ascribed to the interference between the effective waves. Stronger effective sources  
 429 (relative to ineffective sources) imply more effective waves in the total wavefield, and thereby, a  
 430 better quality for the noise-derived target signal. Note that not all waves emanating from the  
 431 effective sources, but only those following specific ray paths, are effective. There might be  
 432 multiple pairs of seismic phases that could contribute to the construction of the target signal.  
 433 However, their relative strength matters. As for the case of the  $P_{\text{dmc}}$  phase, the effective waves  
 434 are  $P$  and  $PKPab$ , which are both prominent phases in the ballistic wavefield. Li et al. (2020)  
 435 showed that the  $PcP$ - $PKPab$  correlation and the  $PcS$ - $PcPPcP$  correlation, could also lead to a  
 436 signal at around the  $P_{\text{dmc}}$  emerging time. However, the  $PcP$ ,  $PcS$ , and  $PcPPcP$  waves are weak  
 437 phases in the ballistic wavefield, and thereby have minor contributions to the  $P_{\text{dmc}}$  signals. We  
 438 emphasize that the sketch in Fig. 7(b) is only suitable for the ambient noise wavefields that are  
 439 dominated by ballistic waves.

440 From Fig. 6, one can observe a high correlation between wind speed and wave height in  
 441 region  $E$  ( $r = 0.74$ ). It indicates that the ocean waves in region  $E$  are likely dominated by the  
 442 waves forced by local winds. The correlation between wave height and microseism excitation is  
 443 low ( $r = 0.25$ ), implying a dominant role of the freely propagating swells in exciting the  
 444 microseisms. Extreme sea state does not guarantee strong microseism excitation. That is not  
 445 surprising according to the microseism excitation theory (Hasselmann, 1963; Longuet-Higgins,  
 446 1950): the excitation is proportional to the product of the heights of the colliding equal-frequency  
 447 ocean waves. In lack of equal-frequency waves coming from opposite directions, even extreme  
 448 wave climate cannot incite strong secondary microseisms. In contrast, for large peaks in the



449 microseism excitation, the corresponding wave heights are generally moderate (e.g., on May 1st  
450 and 23rd). On these two dates, the low wind speeds but moderate wave heights in region *E*  
451 suggest that the ocean waves are dominantly the freely travelling swells from elsewhere, as also  
452 illustrated in the supplementary movie S1. Oppositely propagating equal-frequency swells  
453 collide with each other and incite strong microseisms. Our analysis and observations agree with  
454 those of Obrebski et al. (2012), who investigated a specific case that small swells from two  
455 storms meeting in the eastern Pacific generate loud microseism noise. There are also examples  
456 showing that wind waves can play a role in the excitation of microseisms, for instance, around  
457 July 31st when the local winds, wave height, and microseism excitations are all strong. Such  
458 examples are few. The good consistency between the temporal variations in the  $P_{\text{dmc}}$  strength, the  
459 effective source intensity and the NZ microseism noise level (Fig. 6), provides extra supports to  
460 the analysis of the  $P_{\text{dmc}}$  observations and the quasi-stationary phase arguments proposed by Li et  
461 al. (2020). It also gives credits to the validity of the numerical modeling of oceanic microseism  
462 sources by Arduin et al. (2011) and Rascle & Arduin (2013).

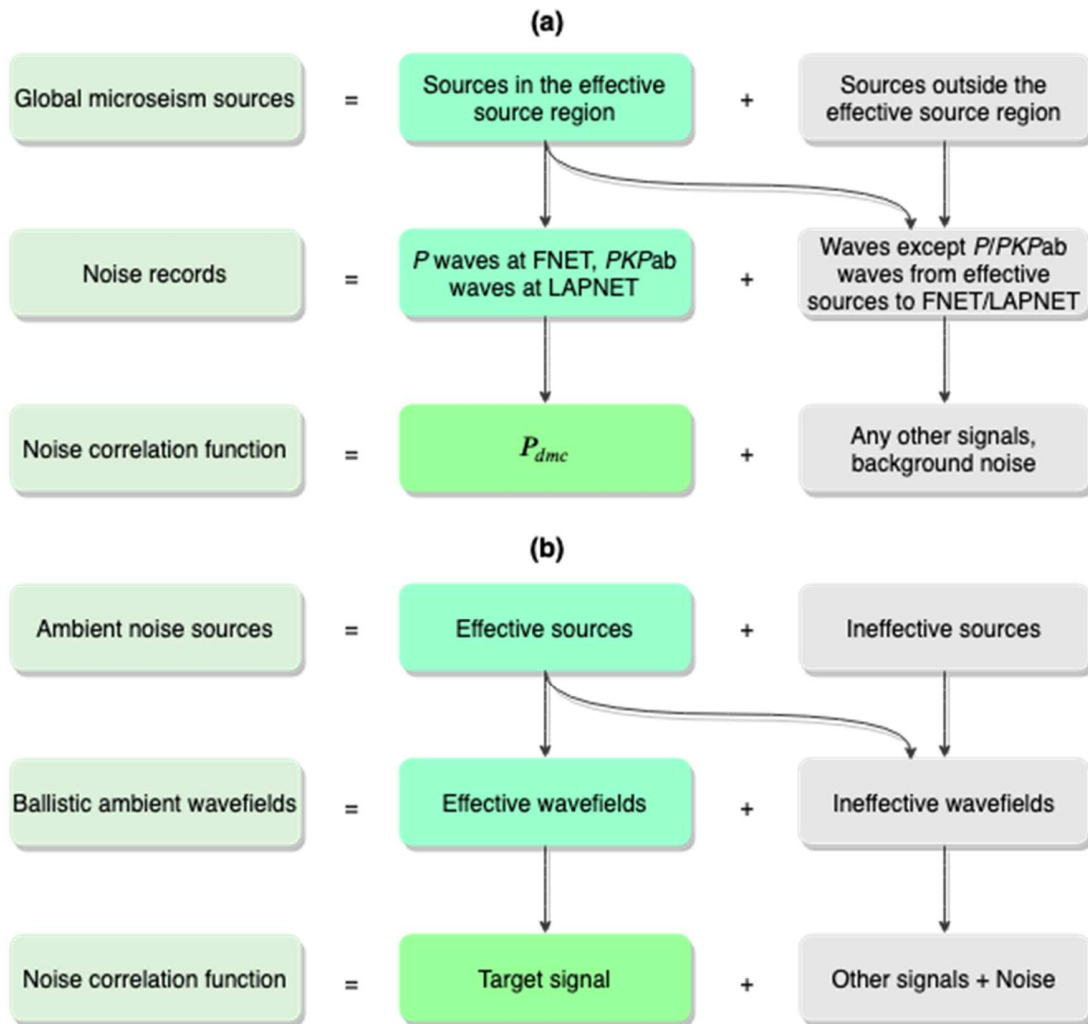
463 We have described above the implications of this study in seismology and in  
464 understanding the process of microseism excitation. Now, we discuss the significance in  
465 oceanography and climate science. Well-documented historical ocean storms and wave climate  
466 are valuable for improving our understanding of climate change and global warming (Ebeling  
467 2012). However, modern satellite observations of ocean waves and storms have a history of  
468 merely decades. Microseisms are induced by storm-driven ocean waves (Arduin et al., 2015;  
469 Hasselmann, 1963; Longuet-Higgins, 1950). The records of microseisms contain the imprint of  
470 climate (Aster et al., 2010; Stutzmann et al., 2009). Instrumental observation of microseisms has  
471 an over-century history, and started much earlier than the modern observations of ocean waves  
472 and storms. It has been a long-lasting effort for the seismological community to digitalize the  
473 historical analog seismograms (Bogiatzis & Ishii, 2016; Lecocq et al., 2020). Researchers expect  
474 that past seismic records can be used to recover undocumented historical ocean storms and wave  
475 climate (Ebeling 2012; Lecocq et al., 2020).

476 This study confirms that it is possible to detect remote microseism events (burst of  
477 microseism energy) with land observation of microseisms. We demonstrate that the noise-  
478 derived  $P_{\text{dmc}}$  signals can be employed to monitor microseism events in a specific ocean region  
479 (Fig. 5). The remote monitoring of microseisms is promising as an aid to improving wave  
480 hindcast, in similar manners as demonstrated by Stopa et al. (2019). The comparative analysis in  
481 Fig. 6 indicates that the remote event detection could be effective in the absence of strong  
482 sources near the stations, otherwise the detection could fail. Stations at low latitudes where wave  
483 climate and microseism excitation are relatively mild, or inland stations far from oceans, should  
484 have better performance in remote monitoring.

485 Energetic microseism excitation does not always need extreme in situ wave heights, and  
486 extreme wave heights do not necessarily produce powerful microseisms (Obrebski et al., 2012;  
487 and this study). It imply that secondary microseism events are not a perfect proxy for the  
488 extremal in situ wave climate. However, it does not mean the long-lasting attempt to monitor  
489 remote sea state and ocean storms with land observation of secondary microseisms is futile. In  
490 the  $P_{\text{dmc}}-h_s$  correlation map (Fig. 3g), the largest  $r$  values do not fall in the effective region *E* as  
491 in the  $P_{\text{dmc}}$ -source correlation map (Fig. 3c), but in surrounding regions with moderate to high  
492 ocean wave activity (the bounded areas in Fig. 3g). We speculate that these regions could be the  
493 birthplaces of the colliding swells that generate the secondary microseisms in region *E*, or the

494 ocean waves in these regions are driven by the same storms as the colliding waves in region *E*  
 495 (see the spatial links of  $h_s$  from Fig. 3h and supplementary movie S1). The detection of a  
 496 microseism event could affirm the existence of the causative storms that generated the ocean  
 497 waves propagating to the location of the microseism event, although the storms could be distant  
 498 from the events.

499



500

501 **Figure 7.** (a) Sketch explanation for the relationships between microseism noise sources and the  
 502 noise-derived  $P_{dmc}$  signal. (b) Generalization of diagram (a) for an arbitrary signal derived from  
 503 ambient noise wavefields that are dominated by ballistic waves.

504 **Acknowledgments and data**

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 508 access: June 2018), respectively. The seismic data of GEONET and the earthquake catalogue of  
 509 New Zealand were provided by the GEONET Data Center (<https://www.geonet.org.nz/>; last

510 access: June 2018). The global earthquake catalogue was provided by the U.S. Geological  
 511 Survey (<https://earthquake.usgs.gov/>; last access: June 2018). The wind hindcast data were  
 512 provided by the European Centre for Medium-Range Weather Forecasts  
 513 (<https://www.ecmwf.int/>; last access: June 2018). The hindcast data of wave heights and  
 514 microseism source PSDs were provided by the IOWAGA products (Rascle & Ardhuin, 2013).  
 515 The bathymetry data were extracted from ETOPO1 Global Relief Model (Amante & Eakins,  
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[*Geochemistry, Geophysics, Geosystems*]

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Supporting Information for

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**Spatiotemporal correlation analysis of noise-derived seismic body waves with ocean wave**

684

**climate and microseism sources**

685

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Caption for Movie S1

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**Text S1. FNET-LAPNET noise correlations**

703

The correlation function  $C_{AB}$  between two seismograms ( $S_A$  and  $S_B$ ) is given by

704

$$C_{AB}(\tau) = \frac{\sum_i S_A(i) S_B(i-\tau)}{\sqrt{\sum_i S_A^2(i) \sum_i S_B^2(i)}} \quad (\text{S1})$$

705

The resultant  $C_{AB}$  consists of an acausal part and a causal part, that correspond to the negative

706

lags ( $\tau < 0$ ) and the positive lags ( $\tau > 0$ ), respectively. For efficiency, it is routine to compute

707

the correlation function with the Fast Fourier Transform:

708

$$C_{AB}(\tau) = \frac{\mathcal{F}^{-1}[\mathcal{F}(S_A)\mathcal{F}^*(S_B)]}{\sqrt{\sum_i S_A^2(i) \sum_i S_B^2(i)}} \quad (\text{S2})$$

709

Figure S1(a) shows the acausal and causal sections of FNET-LAPNET noise correlations in 2008

710

that are filtered between 5 s and 10 s and binned in distance intervals of 0.1°. The acausal section

711

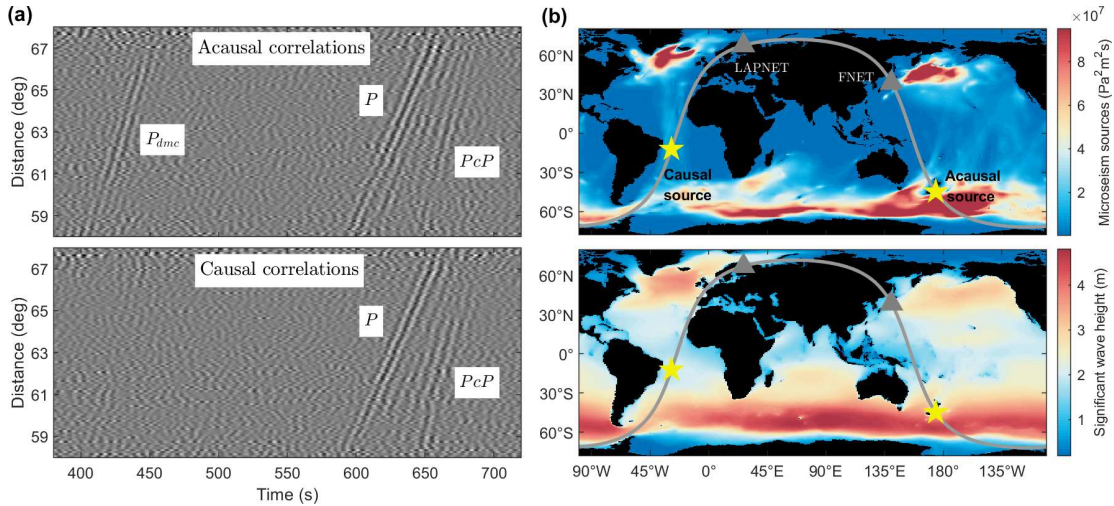
is flipped to share the time axis with the causal section. The expected locations of the acausal and

712

causal noise sources are marked by stars on the maps of global microseism source PSDs and



713 ocean wave heights in Fig. S1(b). The ocean wave activities and microseism excitations at the  
 714 acausal source region are intense, while those in the causal source region are fainter.  
 715 Consequently, the  $P_{dmc}$  phase is only observable from the acausal noise correlations.  
 716



717  
 718 **Figure S1.** (a) Acausal and causal sections of FNET-LAPNET noise correlations in 2008. (b)  
 719 Global maps of 6.2 s period secondary microseism sources and significant wave heights in 2008.  
 720

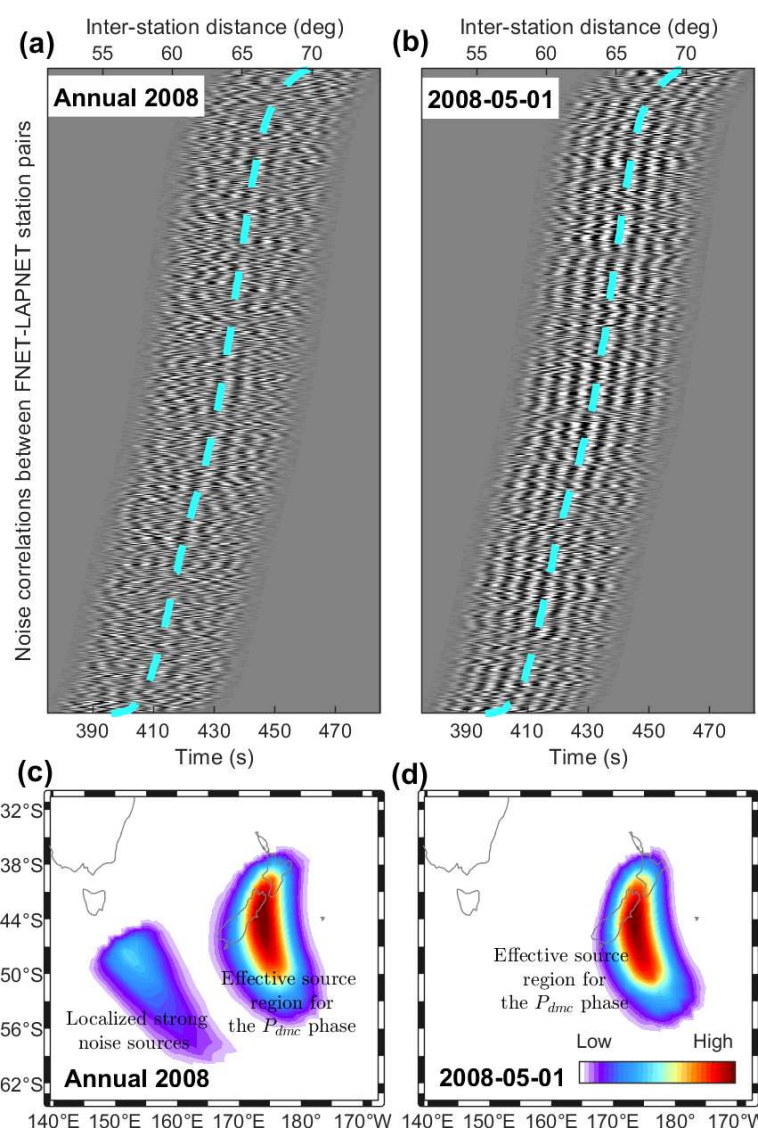
## 721 Text S2. Noise source imaging by back-projection

722 Assuming the interferometry between  $P$  waves at FNET and  $PKPab$  waves at LAPNET, we  
 723 image the effective noise sources through the back-projection of the FNET-LAPNET noise  
 724 correlations. We beam the FNET-LAPNET noise correlations and assign the beam power

$$725 P_s = \langle \langle C_{ij}(t + t_{si} - t_{sj}) \rangle_{ij}^2 \rangle_t, \quad (\text{S3})$$

726 onto a  $0.5^\circ \times 0.5^\circ$  grid as the probabilities of noise sources on the global surface. In the above  
 727 equation,  $\langle \cdot \rangle_x$  means the average over  $x$ ,  $C_{ij}$  is the correlation function between the  $i$ -th FNET  
 728 station and the  $j$ -th LAPNET station,  $t_{si}$  is the travelt ime of the  $P$  wave from the  $s$ -th grid point  
 729 to the  $i$ -th FNET station, and  $t_{sj}$  is the travelt ime of the  $PKPab$  waves from the  $s$ -th source to the  
 730  $j$ -th LAPNET station. The inter-station noise correlations are windowed before the beamforming  
 731 (Fig. S2a). The noise source imaging for the annually stacked noise correlations is plotted in Fig.  
 732 S2(c). Only the region surrounding NZ is shown. Outside the region, hardly can the  $P$  wave  
 733 reach FNET or the  $PKPab$  waves reach LAPNET. Besides a well-focused imaging of the  
 734 expected source region in the ocean south of NZ, we notice a secondary spot to the west. In  
 735 comparisons with the power map of oceanic microseism noise sources in Fig. 5(e), we ascribe it  
 736 to the strong microseism excitation in the ocean south of Tasmania. We also back-project the  
 737 daily noise correlations on 2008-05-01 (Fig. S2b), when the  $P_{dmc}$  phase reaches the largest  
 738 strength through the year (Fig. 2). As shown in Fig. S2(d), an exclusive source region is imaged,  
 739 which agrees with the dominant spot in Fig. S2(c).

740



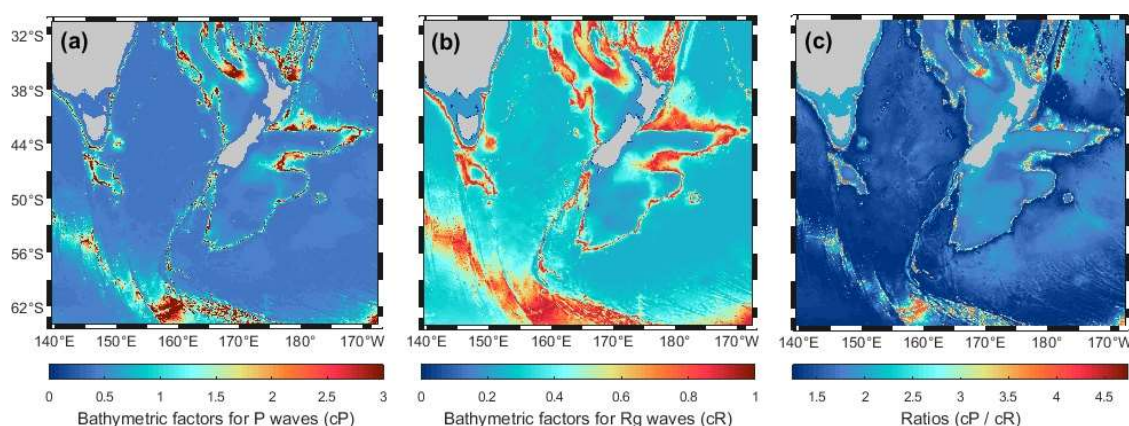
741  
 742 **Figure S2.** Inter-receiver noise correlations for all FNET-LAPNET station pairs: (a) stacked  
 743 over the year of 2008; (b) on single day of 2008-05-01. The waveforms are windowed around the  
 744  $P_{dmc}$  phase. Dashed lines indicate inter-station distances. Back-projection imaging of noise  
 745 sources: (c) using data from (a); (d) using data from (b).

746

747 **Text S3. Bathymetric amplification factors**

748 Figure S3 compares the bathymetric amplification factors surrounding New Zealand for  
 749  $P$  waves and Rayleigh waves. The factors for  $P$  waves are computed using the equations  
 750 proposed by Gualtieri et al. (2014), for a seismic period of 6.2 s and a slowness of 4.6 s/deg. The  
 751 factors for 6.2 s period Rayleigh waves are obtained by interpolating the table given by Longuet-  
 752 Higgins (1950).

753



754  
755 **Figure S3.** Bathymetric amplification factors for (a) *P* waves and (b) Rayleigh waves. (c) Ratios  
756 between the factors for *P* waves and for Rayleigh waves.

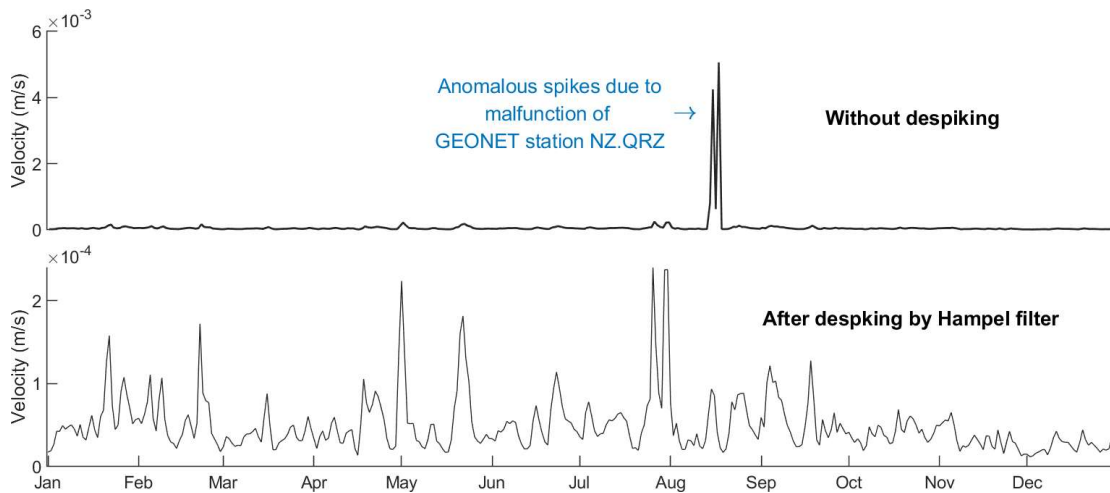
757

#### 758 **Text S4. Microseism noise levels at seismic networks**

759 The continuous seismograms record not only the background vibrations of Earth, but also  
760 ground motions induced by seismicity or other events. Instrumental malfunction also leads to  
761 anomalous (e.g., nearly vanishing or extremely large) amplitudes in the records. These extreme  
762 amplitudes (outliers) could bias the estimates of microseism noise power. It is necessary to get  
763 rid of them from the ambient noise records before the computation of noise power. Mean and  
764 median filters are the common tools for this task. However, they modify all the samples. Here,  
765 we prefer to use a variant of the median filter called Hampel filter. In contrast to the median filter  
766 that replace all samples with local medians, the Hampel filter detects outliers by compare a  
767 sample with the neighboring samples. A sample is replaced by the local median if it deviates  $k$   
768 times of the median absolute deviation (MAD) from the local median, or else, it is unchanged.

769 We filter the vertical components of the continuous seismograms around 6.2 s period. The  
770 seismograms are then divided into 15-min segments and the power of segments is computed. We  
771 apply the Hampel filter to the time series of noise power recursively. For each sample, we  
772 compute the local median and MAD of its eight neighbors (four before and four after). A sample  
773 is replaced by the median if it deviates from the median over three times of the MAD. The de-  
774 spiked time series is resampled from a 15-min interval to a 1-hour interval, by averaging over  
775 every four samples. Then, we apply the Hampel filter again and resample the time series to a 24-  
776 hour interval. The averaging of noise levels over all stations of a seismic network leads to the  
777 time series of array noise level. Before the averaging, the Hampel filter is applied again, to  
778 discard possible anomalous values at some stations (see Fig. S3 for the example of GEONET).  
779 The final time series of microseism noise levels for networks FNET, LAPNET and GEONET are  
780 shown in Fig. 6.

781



782  
 783 **Figure S4.** Comparison between the time series of daily GEONET noise levels with (lower) and  
 784 without (upper) despiking using the Hampel filter.

785

786

787 **Table S1.** List of earthquakes (magnitude above 5.5) in 2008 extracted from the USGS  
 788 catalogue, as a supplementary to the comparison between seismicity and  $P_{\text{dmc}}$  in Fig. 2 of the  
 789 main text. On some dates with earthquakes near the FNET-LAPNET great circle (e.g., events  
 790 2008-08-25T11:25:19.310 and 2008-11-21T07:05:34.940), no large  $P_{\text{dmc}}$  is present, indicating  
 791 that  $P_{\text{dmc}}$  is unrelated to earthquakes.

792

793 **Movie S1.** Daily evolutions of winds, ocean wave heights, and secondary microseism source  
 794 PSDs around New Zealand in 2008. The closed lines superposing the upper panels depict the  
 795 contour values of 0.1, 0.5, and 0.9 for the weights shown in Fig. 5(c). The source PSDs are  
 796 modulated by the bathymetric factors shown in Fig. 5(e). In the bottom panel, the time series for  
 797 the  $P_{\text{dmc}}$  strength and the weighted averages of the source PSD, wave height, and wind speed in  
 798 the effective source region, are the same as those in Fig. 6 in the main text. See captions of Figs  
 799 5 and 6 for more details.

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801