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Multi-Array Constrained 50 mHz Rayleigh-Wave Microseism Sources: Global Distribution and Ocean–Solid Earth Coupling

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Abstract

Microseisms, as the most energetic component of the Earth's background noise field, represent a forefront area of research where precise location of their sources is paramount. This study systematically investigates the spatiotemporal characteristics of 50 mHz Rayleigh wave microseisms using the dense Shandong array deployed in eastern China, through beamforming and a multi-array combined analysis. The results reveal that the incident direction of the Rayleigh waves exhibits distinct temporal and seasonal variations, primarily originating from four back-azimuth sectors. To further constrain the source regions, we integrate background noise data from the Alaska array and the Venezuela array (supplemented by the Indonesia array). The multi-array product back-projection, by cross-constraining back-azimuths from geographically separated arrays, mitigates the inherent ambiguity of single-array analyses and enables robust global source localization. This approach not only improves the reliability of source attribution but also demonstrates the potential of using microseismic noise as a passive tool for monitoring ocean wave activity and investigating solid-Earth structure. The combined analysis identifies four microseism source regions (M1–M4): the Bering Sea–Gulf of Alaska–Aleutian Islands, the central South Pacific, the southwestern Indian Ocean off southern Africa, and the northeastern North Atlantic–Northern Europe. These source regions fundamentally correspond to areas of elevated significant wave height, confirming the coupled ocean–solid Earth excitation mechanism. These findings provide a methodological basis for future applications of multi-array microseismic monitoring in ocean–climate studies and seismic imaging.

Keywords: Shandong province; microseisms; Rayleigh wave; multiple seismic arrays; beamforming; ambient seismic noise; source localization

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1. Introduction

Background Rayleigh waves in the frequency range of 2–20 mHz, referred to as the background free oscillations, have now been unequivocally confirmed [1,2]. The phenomenon of background free oscillations should be understood as occurring within a unified system comprising the atmosphere, oceans, and solid Earth. Hasselmann [3] identified the characteristics and excitation mechanisms of background noise, and in 1976, he proposed a stochastic climate model (the Hasselmann model). Ambient noise is a ubiquitous low-amplitude signal generated by atmospheric coupling processes [4]. With advancements in seismic interferometry techniques, ambient noise recorded by seismometers can be

transformed into usable signals through cross-correlation, forming the basis for ambient noise tomography used to study the structure of the crust and upper mantle [5–7].

Longuet-Higgins [8] proposed a theory for the origin of microseisms: ocean waves generated by wind induce pressure variations on the seafloor, thereby producing microseisms. The relevance of ocean wave activity also extends beyond microseism excitation, as wave induced loading and cyclic marine exposure are known to affect the durability and mechanical response of coastal and marine infrastructure, as shown by recent studies combining wave impact simulation and acoustic monitoring of concrete in marine environments [9]. The excitation of primary microseisms can be attributed to the topographic coupling of ocean waves with seismic waves over subtle bathymetric variations on continental shelves; the excitation of secondary microseisms can be attributed to nonlinear forces generated by oceanic standing waves on the sea surface in both pelagic and coastal regions [4]. Research on microseisms date back to the late 19th century, when seismologists recognized the coincidence between microseismic activity and ocean weather conditions [10,11]. Under severe weather conditions, microseismic activity increases simultaneously over spatial scales on the order of 1000 km around storms [12,13]. However, whether the source of this excitation was atmospheric or oceanic remained controversial until the 1950s. For instance, Gherzi [14] argued that microseisms were excited by the “pumping” action of storms associated with low-pressure systems. Sezawa and Kanai [15] also discussed the possibility of atmospheric excitation. Shimura et al. [16], for the first time, systematically analyzed the statistical relationship between extreme waves excited by tropical cyclones and secondary microseism energy by combining long-term microseism records from the Japanese seismic observation network with 20-year high-resolution wave hindcast data, thereby compensating for the lack of direct wave observations in the open ocean. Xiao et al. [17] employed distributed acoustic sensing (DAS) technology using ocean-bottom seismometer data collected offshore Oregon to investigate microseism sources across a range of frequency bands. The results revealed that high-frequency microseisms (0.35–1.5 Hz) primarily originate from wind waves near the coastline, with only a minor contribution from the deep ocean. Although microseisms coincide with weather conditions, atmospheric excitation alone cannot fully explain all observations of these waves.

Microseisms are categorized into two types: (A) primary or single-frequency microseisms (PM) in the 20–100 mHz range, and (B) secondary or double-frequency microseisms (SM) in the 100–500 mHz range [8,18]. The dominant frequency of PM is approximately 70 mHz, identical to the typical frequency of ocean waves [4]. PM are generated by ocean waves impinging on shallow seafloors [3]. The dominant frequency of SM is approximately 150 mHz, about twice the typical frequency of ocean waves [4]. The excitation of SM requires nonlinear interaction between ocean waves propagating in opposite directions with similar periods [8,18,19]. The energy of SM is significantly higher than that of PM [20].

In microseism research, a primary objective is to determine the sources of microseisms. Gao et al. [21] followed the classic technical route established by Cessaro [22] using power spectral density analysis and $f - k$ analysis combined with triangulation to identify secondary microseism sources excited by typhoons in the Indian Ocean and North Pacific. In contrast, the study by Liu et al. [23] represents a modern evolution in this field: by combining two large-aperture, continent-spanning dense arrays from China and the United States, they applied a back-projection method to jointly constrain P-wave microseism sources using multiple arrays, effectively reducing location biases caused by the array response function of a single array. Additionally, Guo et al. [24] employed polarization analysis to systematically identify multiple source regions of single-frequency and double-frequency microseisms along the eastern coast of the United States.

Array beamforming has been widely applied in microseism studies [24]. For example, single arrays often utilize time-domain beamforming methods, which include techniques using single three-component stations [25], cross-correlation combined with back-projection [26,27], analysis of azimuthal variations [28], and seasonal variations [26]. Since microseismic energy propagates as Rayleigh waves and is often incoherent in frequency but coherent in space, frequency-domain beamforming is frequently employed. Utilizing all stations in a region enables observations of microseismic sources with higher temporal, frequency, and azimuthal resolution than previous studies.

Previous researchers have utilized beamforming methods to analyze ambient noise across different regions over multiple time periods. Gerstoft et al. [29], in their analysis of ambient noise in the Gulf of Mexico, suggested that both surface waves and P waves are generated in shallow water, possibly due to wave shoaling or breaking. Xiao et al. [30], using continuous waveforms from three arrays along the southern coast of China, found that Rayleigh and Love waves have distinct sources, which are frequency-dependent and show no significant seasonal variation in the dominant energy arrival directions. Liu et al. [31] detected SV and SH waves generated by storms in the North Atlantic using seismic array methods, while Zhang et al. [32] studied pelagic and coastal sources of P-wave microseisms under tropical cyclones. Nishida and Takagi [33] identified teleseismic S-wave microseisms. Fang and Xue [34] systematically revealed the energy characteristics and source azimuths of double-frequency microseisms in the north-central North Atlantic through power spectral density and polarization analysis. Gualtieri et al. [35] found, through global spectral-element simulations, that Love-wave generation in the secondary microseism band (3–10 s) must rely on three-dimensional heterogeneous structures of the crust and mantle. Herrmann et al. [36] obtained short-period phase and group velocity dispersion curves for Rayleigh and Love waves across the contiguous United States and southern Canada. Lin et al. [37] revealed that Typhoon Hagupit generated double-frequency microseisms through two distinct mechanisms in open ocean and nearshore regions. Nishida and Takagi [38] constructed a global catalog of P-wave microseismic CSF sources using Hi-net data, which generally agrees with wave model predictions. Pelaez Quiñones et al. [39] performed three-component polarization beamforming on hurricanes in the Northwest Atlantic and found that microseismic signals are intermittent and concentrated in shallow-water regions. Shabtian et al. [40] found pronounced seasonality in primary and secondary microseisms in California, with winter noise levels 15.4 dB higher than summer. John and West [41] showed that the Gulf of Alaska is the primary source region for secondary microseisms (5–10 s), while short-period microseisms (1–2 s) are strongly modulated by sea ice. Xu et al. [42] found that seabed sediments can amplify secondary microseism energy by up to a factor of 100. Aster et al. [43] confirmed that microseism amplitudes exhibit seasonal variations with opposite polarities between the Northern and Southern Hemispheres. Kabanda and van der Lee [44] analyzed the direct influence of seasonal storm activity in three major ocean basins surrounding Africa on the azimuthal distribution of microseism energy. Sheen and Shin [45] found that during the boreal winter, most seismic stations in Asia observed spectral peaks near 0.2 Hz for several consecutive months, and back-projection suggested that these microseismic P waves are likely generated near the center of the northern Pacific Ocean.

Building on previous findings, in this study, we first investigated the spatiotemporal characteristics of Rayleigh wave fundamental mode (PM) using the dense broadband seismic array deployed in Shandong, China (hereinafter referred to as the SD array). Subsequently, we performed beamforming and Rayleigh wave amplitude vs. azimuth analysis to examine the seasonal variations and azimuthal distribution of microseisms. Next, we back-projected the beam energy along great-circle paths to identify potential source regions. Following that, by combining the SD array with the YG array (hereinafter referred

to as the YG array) and the Venezuela array (hereinafter referred to as the YU array, supplemented by the Indonesian array at the same latitude, collectively referred to as the YC array), we applied a multi-array back-projection technique to constrain possible source zones. Finally, based on the multi-array combined images, we inferred the global locations of PM sources.

Although previous studies have performed beamforming using single arrays or global networks, this study uniquely combines the SD array in eastern China with three geographically distinct arrays—the YG array, the YU array, and the YC array—to achieve tight spatial constraints on Rayleigh wave microseism sources through multi-array product back-projection. This approach reduces the ambiguity inherent in source localization based solely on azimuth from a single array. In this study, we focus on the lower limit of the primary microseism frequency band (50–100 mHz), i.e., the frequency range of 50–80 mHz. We first used the SD array to investigate the spatiotemporal characteristics of Rayleigh wave PM. We then performed beamforming and Rayleigh wave amplitude-versus-azimuth analysis to examine seasonal variations and azimuthal distributions of the microseisms. Next, we back-projected the beam energy along great-circle paths to identify potential source regions. Subsequently, by combining the SD, YG, and YU arrays (with the YC array at similar latitudes as a supplement), we applied multi-array back-projection to constrain possible source zones. Finally, based on the multi-array combined images, we inferred the global locations of the PM sources.

2. Materials and Methods

In this study, we first use data from 83 broadband stations of the SD array deployed across the Liaoning segment of the Anhui in eastern China (Figure 1). Three-component continuous seismic data were recorded for 10 months from October 2017 to July 2018. In addition, we downloaded vertical continuous records of the YG array of the same period from the YU array, on this basis, we supplemented the data with records from the YC array for the period from April to July 2018. Due to differences in aperture, inter-station spacing, and geographic settings between the two arrays, inconsistencies may arise in the localization results. To address this limitation, we investigated the relevant parameters of the two arrays. The YU array data are available only from October 2017 to April 2018. For the period from May to July 2018, we used data from the YC array (Indonesia) as a substitute because of data unavailability. The YC array is deployed in the northern Borneo region, within the Sunda plate collage zone, and was originally designed for seismic surveys of the Northern Borneo Orogenic Belt. In contrast, the YU array is located in the Mérida Andes and surrounding areas, at the boundary between the Caribbean Plate and the South American Plate, with complex geological structures including major fault systems. The two arrays differ substantially in aperture, inter-station spacing, and geographic setting. As a supplementary data source for the period from May to July 2018, the YC array can only serve as a compromise and cannot fully replace the YU array. This substitution may introduce systematic biases, particularly in the source localization results for the April–June 2018 transition period, as discussed in Section 4. Based on the data from the SD array and three arrays of YG, YU and YC, we employed a multi-array back-projection method to constrain source locations of Rayleigh wave microseisms.

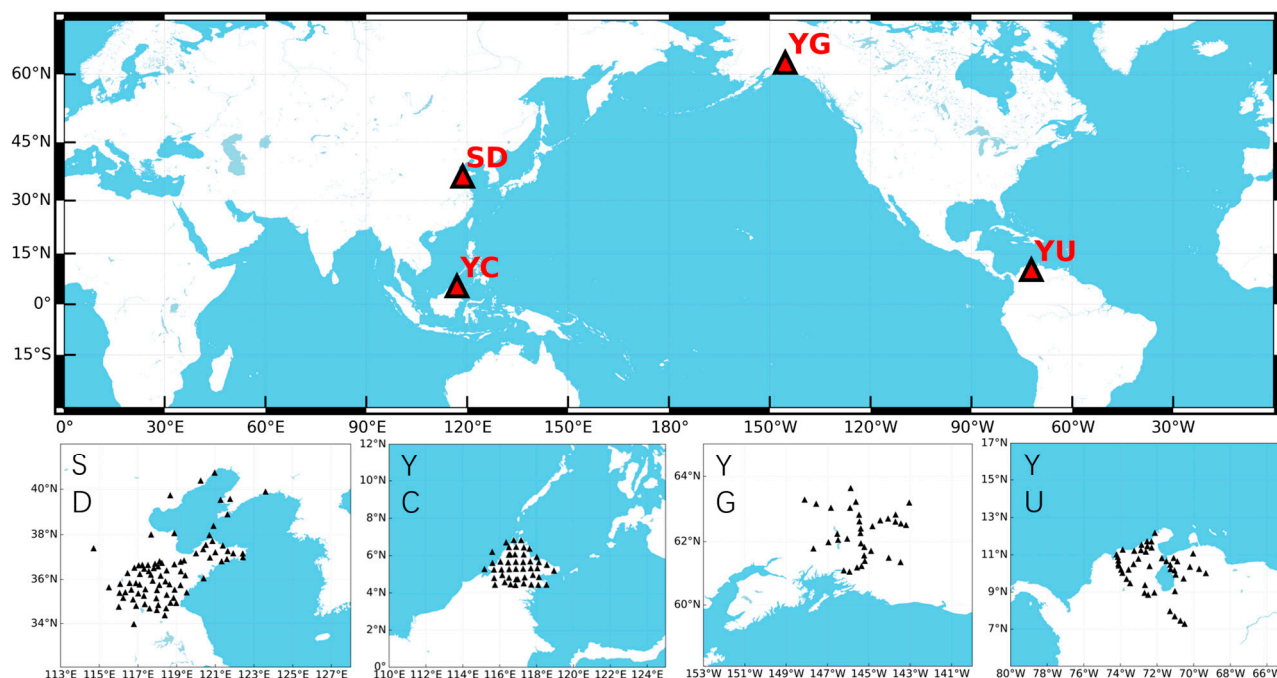


Figure 1. Locations of the arrays used in this study are shown in the upper panel. We use four arrays, SD array (83 stations) in eastern China, YG array (34 stations), YU array (45 stations) and YC array (46 stations). The small black triangles in the figure represent the station distribution.

2.1. Beamforming

We first use the beamforming method to obtain the directionality of surface wave microseisms. Continuous vertical-component data are cut into one-day time series, resampled at 1 Hz, and filtered from 10 to 400 mHz. Then the data are split into non-overlapping 512-s long segments. We compute the standard deviation σ of each segment and truncate signal amplitude with a clipping threshold of σ , that is, all data points with absolute values greater than σ are set to $\pm\sigma$. For reproducibility, we list the key preprocessing parameters here. Continuous vertical-component data were cut into one-day time series, resampled at 1 Hz, and band-pass filtered from 10 to 400 mHz. The filtered data were then divided into non-overlapping 512-s segments. For each segment, we computed the standard deviation σ and clipped the amplitude at a threshold of 3σ (i.e., all data points with absolute values $> 3\sigma$ were set to $\pm 3\sigma$). The slowness search grid spans ± 0.5 s/km in both the x- and y-directions, with an increment of 0.002 s/km, producing a 501×501 grid image for each frequency. The beam-power spectrum is computed as the time-averaged cross-spectral density matrix over 10 segments (approximately 1.5 h). For the surface-wave analysis, we use a slowness window of 0.2–0.4 s/km (apparent velocity 2.5–5.0 km/s). The chosen slowness window corresponds to the expected phase-velocity range of fundamental-mode Rayleigh waves in the primary microseism band (~ 50 mHz) for continental and oceanic paths. However, it should be recognized that the Earth's crust is distinctly layered, and surface waves in this frequency band are dispersive: their phase velocity depends on the elastic properties and thicknesses of the underlying crustal layers. In a layered medium, the long-wavelength limit of Rayleigh–Lamb waves cannot be determined straightforwardly and generally requires a full solution of the Rayleigh secular function [46]. For the regional paths relevant to this study (continental and oceanic–continental mixed paths), the phase velocity of fundamental-mode Rayleigh waves at 20-s period typically lies between 2.8 and 3.8 km/s, with values near the lower end for oceanic paths and near the upper end for continental paths [47]. Our chosen slowness window (0.2–0.4 s/km) conservatively brackets this range, allowing for path-averaged velocity

variations without imposing an overly restrictive assumption. The array response function is deconvolved from the observed beam-power spectrum using an accelerated Richardson–Lucy algorithm with 10 iterations, after which the beamforming output is normalized by its maximum value. Then we normalize the seismic data in the frequency domain to reduce narrow band signals that could be a non-microseismic transient [24,48]. The center of the array is calculated by averaging the latitude and longitude of all stations. Next, we apply beamforming analysis in the frequency domain under the assumption that waves recorded by the array can be represented by a superposition of plane waves [24]. The vertical-component array data are Fourier transformed to a complex-valued vector $\mathbf{v}$ containing the response of all stations in the array. In the frequency domain, the beam power value $b(\omega, t, s_x, s_y)$ for a given frequency ω and slowness (s_x, s_y) is expressed as

$$b(\omega, t, s_x, s_y) = P(\omega, s_x, s_y)^\dagger C(\omega, t) P(\omega, s_x, s_y), \quad (1)$$

As shown in Equation (1), where † denotes the conjugate transpose and $C(\omega, t)$ represents the cross-spectral density matrix, as presented in Equation (2):

$$C(\omega, t) = \frac{1}{K} \sum_{k=1}^K V V^\dagger, \quad (2)$$

As described in Equation (2) above, where k denotes the segment index, and K represents the total number of segments. Here, we set the value of K to 10, indicating time averaging over 10 segments, approximately equivalent to 1.5 h. In Equation (1), $P(\omega, s_x, s_y)$ denotes the plane-wave response of the seismic array in the frequency domain, as given by Equation (3):

$$P(\omega, s_x, s_y) = e^{i\omega \mathbf{s}^T \mathbf{r}}, \quad (3)$$

As described in Equation (3) above, where $\mathbf{s}$ is the assumed plane-wave slowness vector (s_x, s_y) , and $\mathbf{r}$ represents the station position vector measured from the array center.

In this study, to identify the incident direction of Rayleigh waves and investigate seasonal and quarterly variations in the background noise, we summarize all beamforming results for a specific frequency band over the 10-month period from October 2017 to July 2018. Respectively, we vary the values of s_x and s_y within the range of -0.5 – 0.5 s/km, with an increment of 0.002 s/km. Meanwhile, the beam-power spectrum B_0 is a 501×501 obtained as a grid image. As shown in Figure 2a–c display the beam-power images at 50 mHz for the SD-array data recorded in October 2017. In this case, owing to the blurring effect of the array response function (ARF), the resolution of the beamforming image is relatively low; removing the ARF can enhance the resolution of the observed two-dimensional spectrum [49]. The ARF for a plane wave with frequency ω and slowness (s_x, s_y) can be written as [50]:

$$a(\omega, s_x, s_y) = \left| \frac{1}{N} \sum_{n=1}^N e^{i\omega \mathbf{s}^T \mathbf{r}_n} \right|^2, \quad (4)$$

Here, N denotes the number of stations within the array. Being similar to the beam-power spectrum, the ARF spectrum is also represented as a 501×501 grid image, denoted as A . Subsequently, we deconvolve the ARF from the observed beam-power

spectrum B_0 an accelerated Richardson–Lucy deconvolution algorithm [51,52], which can be expressed as shown in Figure 2b.

$$B'_{m+1} = B_m \cdot \left(A \otimes \frac{B_0}{A * B_m} \right), \quad (5)$$

$$B_{m+1} = \begin{cases} B'_{m+1}, m = 0, \\ B'_{m+1} + \lambda_m [B'_{m+1} - B_m], m \geq 1, \end{cases} \quad (6)$$

where B_0 is the observed beam-power spectrum and also serves as the initial input for the iterative deconvolution operation. Equation (5) shows the iterative process of the conventional Richardson–Lucy algorithm, where B'_{m+1} represents the deconvolved spectrum at the current iteration. Here, we employ an accelerated Richardson–Lucy deconvolution algorithm (Equation (6)). Starting from the second iteration, a correction term is added to B'_{m+1} , which is calculated from the difference between the current and the previous iteration. We then obtain the deconvolved spectrum B_m based on the accelerated algorithm. The acceleration coefficient λ_m is defined as $\lambda_m = \sum g_m \bullet g_{m+1} / \sum g_{m+1} \bullet g_m$. The gradient λ_m at the m -th iteration is defined as $g_m = B'_{m+1} - B_m$. The value of λ_m ranges from 0 to 1. Typically, the deconvolved beam-forming spectrum stabilizes after a few iterations; therefore, we set the number of iterations to 10. The deconvolved beamforming output is further normalized by its maximum value, as shown in Figure 2c.

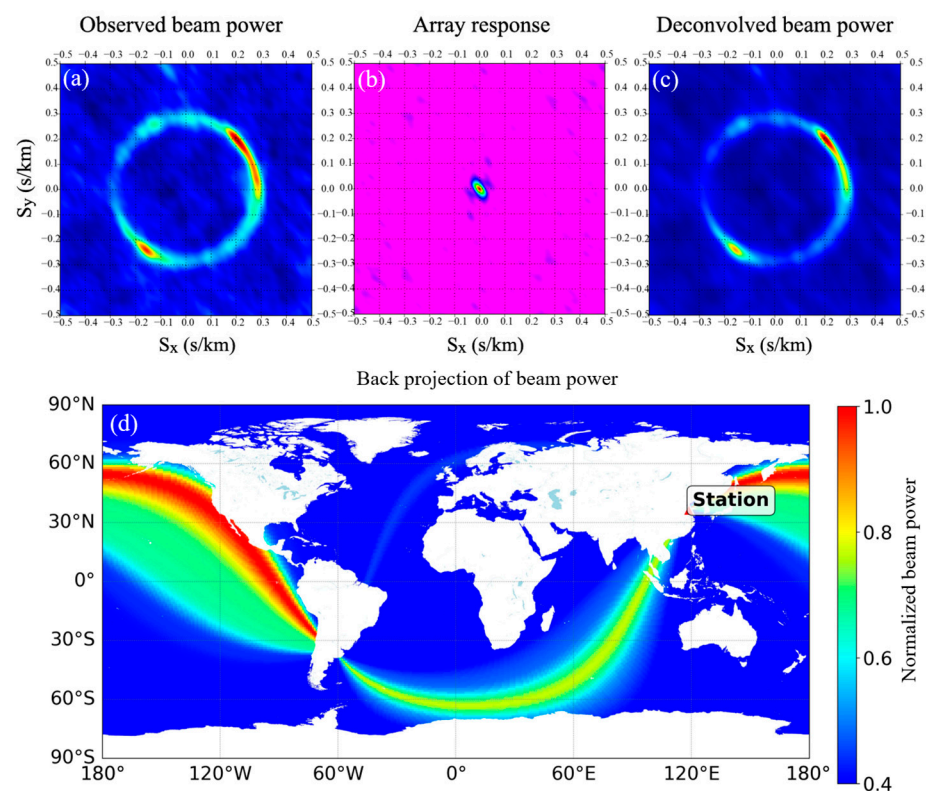


Figure 2. An example of the Richardson–Lucy deconvolution and back-projection applied to the observed beam power at the frequency band of 50 mHz using the TL array data recorded in October of 2017. (a) Observed beam power spectrum. (b) Array response function of the SD array. (c)

Deconvolved beamformer output after 10 iterations. Surface wave slowness window is indicated by white dashed circles (0.2–0.4 s/km). (d) The back-projection of deconvolved beamforming output to the global $1^\circ \times 1^\circ$ grid set based on their corresponding back azimuth. All images are normalized by the maximum energy. In panels (a–c), brighter colors indicate more concentrated energy.

2.2. Locating Noise Sources Using Multi-Array Back-Projection

To locate potential sources along great-circle paths, we introduce a multi-array back-projection method. For this purpose, we first perform a beamforming analysis for each array, as described in Section 2.1. Next, we discretize the Earth's surface using a global grid of $1^\circ \times 1^\circ$. For each grid point, we compute the back-azimuth from that grid point to the center of each array. Finally, based on the corresponding back-azimuth for each array, we back-project their respective beamforming results onto the global set of $1^\circ \times 1^\circ$ grid points. Assuming a surface-wave slowness window of 0.2–0.4 s/km, the maximum beam power within this window is back-projected along the specific back-azimuth to all suitable grid points that share the correct back-azimuth, as given by Equation (7):

$$BP_i(lon, lat)(\varphi) = \max[B_i(\omega, s, \varphi)], 0.2 \text{ s/km} < s < 0.4 \text{ s/km}, \quad (7)$$

According to Equation (7) above, where $BP_i(lon, lat)(\varphi)$ is the back-projection of the beamforming results from the i -th array at the grid point (lon, lat) with back-azimuth φ . $B_i(\omega, s, \varphi)$ is the beam power for the i -th array at slowness $s = \sqrt{s_x^2 + s_y^2}$ and back-azimuth $\varphi = 90 - \arctan(s_y/s_x)$. An example $BP(lon, lat)(\varphi)$ of the resulting back-projection is shown in Figure 2d, which corresponds to the beam power displayed in Figure 2c. Since the PM originate from ocean waves, we mask the values on the continents by coloring the land areas gray. When multiple arrays are available, the total beam power is defined as the product of the individual beam powers from all the N arrays, as given by Equation (8):

$$BP_{total}(lon, lat)(\varphi) = \prod_{i=1}^N BP_i(lon, lat)(\varphi), \quad (8)$$

We selected the YG array and the YU array to form two additional arrays, whose aperture and station spacing are comparable to those of the SD array. Figure 1 shows the station locations of these two selected arrays. Given their relative proximity, we first performed a back-projection of the beamforming results from both arrays to obtain the combined beam power for all grid points. This combination is highly effective in suppressing contributions from local sources. On the other hand, microseisms generated in more distant regions are expected to retain high beam power in the combined image, as their great-circle paths to the closely spaced arrays are nearly parallel to each other. We further normalized the back-projected beam power of the SD, YG, and YU arrays by their respective maximum amplitudes to ensure that beam power from different locations contributed equally to the final result. Finally, we computed the element-wise product of the three back-projected images from the SD, YG, and YU arrays to generate the final back-projection image. Since the seismic data recorded by the YU array only covers the period from October 2017 to April 2018, we supplemented it with three months of data (May to July 2018) from the YC array at a similar latitude. All subsequent processing methods and steps remained consistent.

3. Results

We apply beamforming analysis to the SD array data at the frequency bands of PM (45–95 mHz), a detailed analysis was conducted on the frequency bands of 50 mHz, 60 mHz, 70 mHz, and 80 mHz. To investigate the seasonal variations of microseisms, we first stack beamforming results from 5120-s long non-overlapping segments to generate quarterer beamforming spectra: Quarter Fourth (October 2017 to December 2017), Quarter First (January 2018 to March 2018), Quarter Second (April 2018 to June 2018), and Quarter Third (July 2018). Details are shown in Figure 3. A comparison of Figure 3d–l shows that among the tested frequencies, 50 mHz yields the clearest focusing and the highest signal-to-noise ratio. Therefore, we selected 50 mHz for further analysis and source localization.

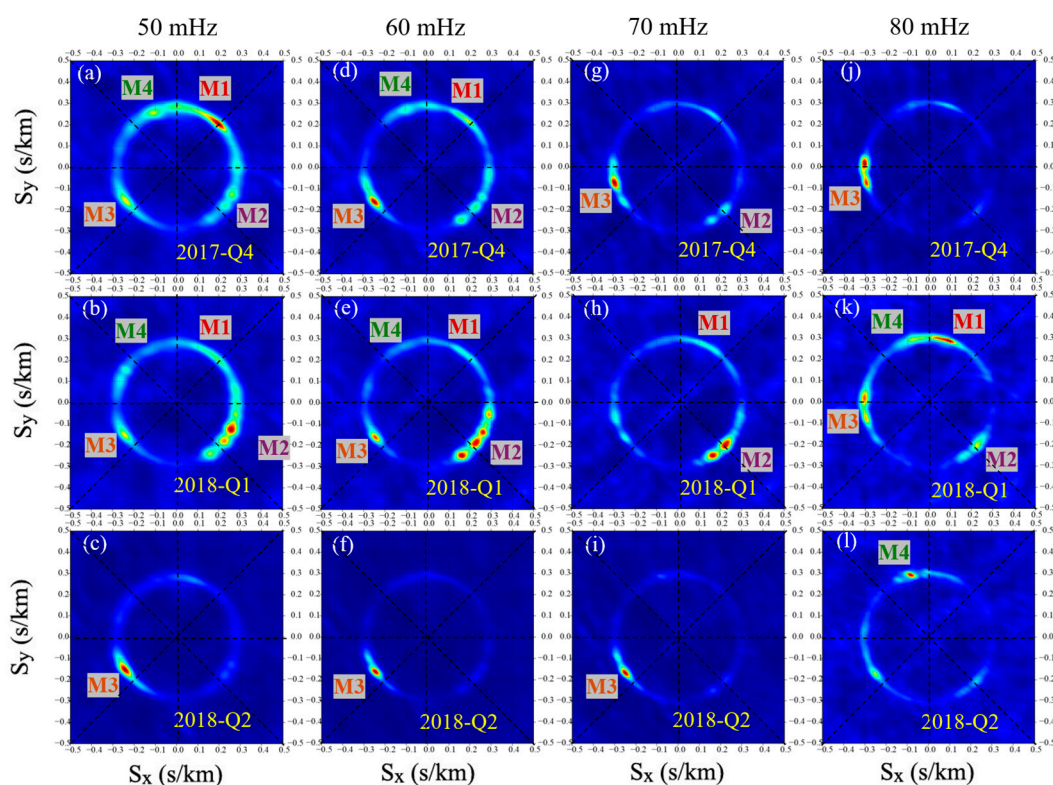


Figure 3. The seasonal and time variations of 50 mHz–80 mHz recorded at the SD array. Panels (a–c) show seasonal variations at 50 mHz, with dominant azimuths labeled M1–M4. Panels (d–l) compare frequency dependence (50–80 mHz) within the same season, demonstrating optimal focusing at 50 mHz. In the figure, brighter colors indicate more concentrated energy, and the black dashed lines are for azimuthal reference.

In the PM frequency band, the SD array beamforming images (Figure 3a–c) are dominated by arrivals with a slowness between 0.25 s/km and 0.4 s/km, that is, 2.5–4.0 km/s in apparent velocity, suggesting that they are most likely surface waves. Roughly, these surface wave arrivals come from four azimuthal bands: $\sim 0^\circ$ – 50° , $\sim 60^\circ$ – 155° , $\sim 200^\circ$ – 290° and $\sim 330^\circ$ – 360° . These four azimuthal bands are labelled M1–M4, and they provide the primary directional constraints for subsequent source localization. The strength and central azimuth vary with season. In Quarter Fourth (Figure 3a) and Quarter First (Figure 3b), the strongest arrival comes from $\sim 40^\circ$ (ENE), which we label as M1; the second strongest from $\sim 120^\circ$ (EES) is labelled M2. A weaker signal from $\sim 335^\circ$ (M4) appears mainly in Quarter Fourth. The arrival from $\sim 240^\circ$ (WSW), labelled M3, strengthens progressively and becomes the only dominant signal in Quarter Second (Figure 3c).

Based on the results from beamforming, we back-project the beam power along great-circle paths from the SD array centre (Figure 4). Figure 4 shows the great-circle paths for each azimuthal band. While these paths indicate potential source regions, they suffer from the 180° azimuthal ambiguity inherent in single-array analysis. High-beam paths in Quarter Fourth cover M1 and M3 (Figure 4a); in Quarter First, M1 and M2 (Figure 4b); in Quarter Second, only M4 remains visible (Figure 4c); and in Quarter Third, the pattern resembles Quarter Fourth (Figure 4d).

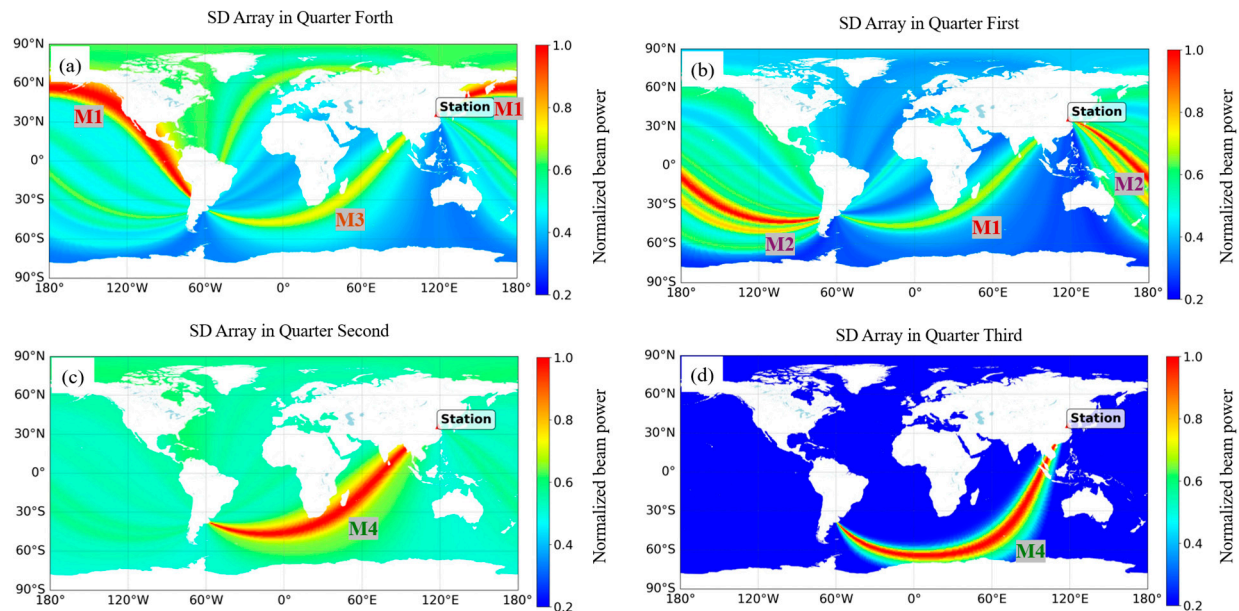


Figure 4. The back-projection of beam power along great circle paths extending from the SD array center for 50 mHz computed from Quarter Fourth (October 2017–December 2017) to Quarter Third (July 2018). Great circle paths are marked with their corresponding arrival number, M1–M4. (a) The potential source regions of M1 ($\sim 40^\circ$) locates in the region of the Bering Sea, the Gulf of Alaska, and the Aleutian Islands. The path of M3 ($\sim 240^\circ$) transects the southern part of the African continent, southwestern Madagascar, and the adjacent waters of the southwestern Indian Ocean. (b) M2 ($\sim 120^\circ$) areas cover the central South Pacific Ocean, situated at the convergent boundary between the Pacific Plate and the Indo-Australian Plate. (c,d) M4 ($\sim 340^\circ$) path spans Northern Europe, the northeastern North Atlantic, and eastern Greenland.

To resolve the ambiguity, we combine the SD array with the YG, YU, and (when YU data were unavailable) YC arrays using multi-array product back-projection. Figures 5a and 5b show the combined back-projections for October–December 2017 and January–March 2018, respectively, using SD, YG, and YU. For April–June 2018, the YU array is replaced by the YC array (Figure 5c). The multi-array product back-projection resolves the 180° ambiguity and yields well-constrained source locations for M1–M4, as shown in Figure 5. The back-projection results from the combined arrays indicate that the potential source regions corresponding to these four arrivals include the North Pacific, the north-eastern sector of the Arctic Ocean, and the southwestern sector of the Indian Ocean.

In addition, we quantify the beam powers of M1 and M2 at 50 mHz for the different quarters (Table 1). During October–December 2017, the beam power of M1 is ~ 0.0058 and that of M2 ~ 0.0038 , giving a ratio of 1.5263. In January–March 2018, M1 decreases to ~ 0.0045 while M2 increases to ~ 0.0065 , yielding a ratio of 0.6923. During April–June 2018 and July 2018, both M1 and M2 become extremely weak (close to zero). Table 1 quantifies the beam power variations of M1 and M2 across the quarters, providing additional evidence for the seasonal changes in source activity.

Table 1. Beam powers of M1 and M2 (2017.10–2018.07).

Beam Power	October 2017– December 2017	January 2018– March 2018	April 2018–June 2018	July 2018
M1	0.0058	0.0045	0	0
M2	0.0038	0.0065	0	0
M1/M2	1.5263	0.6923	0	0

As shown in Figure 3, from the perspective of the SD array, the azimuthal width of each source region also varies with season: in Quarter Fourth, M1 has a width of $\sim 25^\circ$, M2 $\sim 10^\circ$, M3 $\sim 20^\circ$, and M4 $\sim 15^\circ$; in Quarter First, M1 $\sim 5^\circ$, M2 $\sim 45^\circ$, M3 $\sim 10^\circ$, and M4 is not clearly observed; in Quarter Second, M1, M2, and M4 are barely observable, while M3 widens to $\sim 25^\circ$; in Quarter Third, M3 further widens to $\sim 30^\circ$, whereas the other sources remain undetected. These quantitative measurements, together with the back-projection images, confirm that the four source regions are geographically distinct and that their seasonal behaviours are consistent with the ocean wave activity discussed in Section 4.

These back-projection results from the combined arrays indicate that the potential source regions corresponding to these three high-amplitude arrivals include the North Pacific, the northeastern sector of the Arctic Ocean, and the southwestern sector of the Indian Ocean.

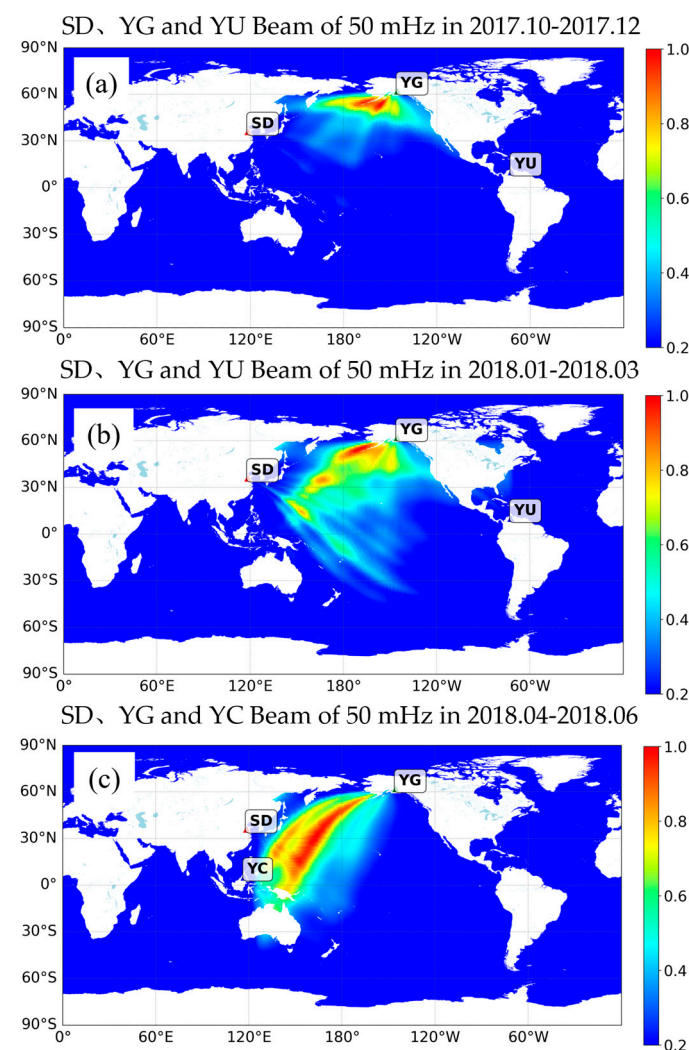


Figure 5. Back-projection analysis of microseismic data recorded by three arrays: (a) YG array, (b) YU array, and (c) YC array. For each array, four source regions (labelled M1–M4) are identified,

which correspond to the four Rayleigh-wave arrivals (M1–M4) observed at the SD array. In panels (a–c), brighter colours indicate more concentrated energy, and the black dashed lines serve as azimuthal reference lines. Overlaps in the figures do not affect the scientific interpretation; they are merely intended to save space in the figure layout. The small black triangles represent the station distribution.

4. Discussion

We use the beamforming method to investigate the distribution of Rayleigh-wave microseisms. Seasonal variations are observable for 50 mHz. The local ocean activities have little contribution to microseisms recorded by the SD array. The back-projections of beam power show potential source regions transecting large swaths of ocean. By combining the SD array with YG array and YU array (YC array serving as a supplement) in southern California, we locate four strong and persistent microseisms at a global scale. M1 locates in the region of the Bering Sea, the Gulf of Alaska, and the Aleutian Islands. M2 areas cover the central South Pacific Ocean, situated at the convergent boundary between the Pacific Plate and the Indo-Australian Plate. The path of M3 transects the southern part of the African continent, southwestern Madagascar, and the adjacent waters of the southwestern Indian Ocean. And M4 spans Northern Europe, the northeastern North Atlantic, and eastern Greenland (Figure 6).

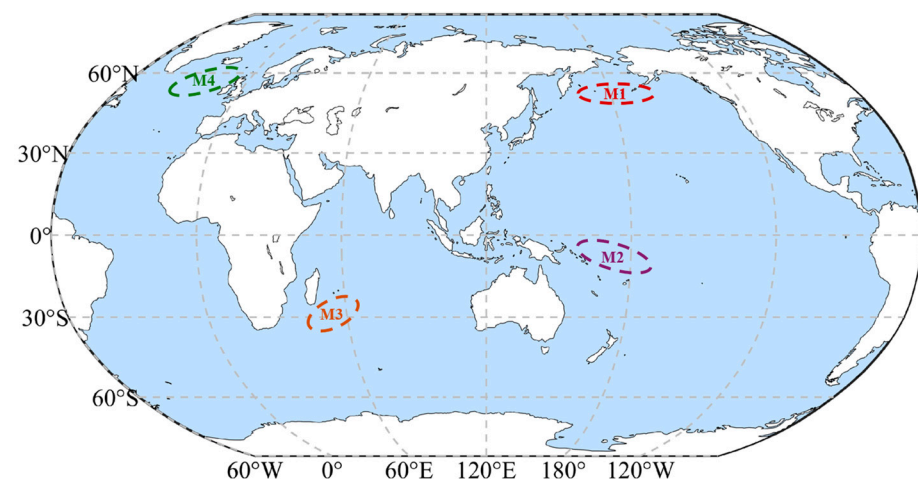


Figure 6. Source regions of Rayleigh-wave microseisms located by multi-array beam power back-projection (M1–M4). M1: the eastern sector covers the Norwegian Sea, the North Sea, and extends to the Scandinavian Peninsula, the western sector encompasses eastern Greenland, Iceland, and stretches into the open ocean surface of the northeastern North Atlantic, it is classified as the North Atlantic and Northern Europe source region (red dotted ellipse). M2: one of the most intense zones of the global prevailing westerlies belt, situated in the open ocean area of the central South Pacific (purple dotted ellipse). M3: located along the core path of the Southern Hemisphere’s “Roaring Forties,” it belongs to the South Indian Ocean source region (orange dotted ellipse). M4: situated along the southern Bering Sea, the Gulf of Alaska, and the Aleutian Islands, it is classified as the North Pacific source region (green dotted ellipse).

The reason why 50 mHz gives the clearest focusing is that it lies at the core of the global primary microseism band (approximately 50–100 mHz), where the ambient seismic noise energy is intrinsically high. Within this frequency band, seismic waves propagate as high-quality plane waves, and the wavelength is generally larger than the array aperture, leading to low sidelobe energy and accurate slowness estimation. Moreover, the plane-wave approximation typically holds below about 50–60 mHz, avoiding multipath

complications. Thus, 50 mHz is the optimal frequency for our beamforming analysis, and we have used it for all subsequent source-localization procedures.

According to the theory proposed by Longuet-Higgins [8], microseisms are generated by the coupled interaction between ocean waves and the solid Earth. As shown in Figure 7, we downloaded the global mean significant wave height data from the Global Ocean Fusion Dataset 1.0 (CGOF1.0_Significant_Wave_Height) provided by the National Marine Data and Information Service (NMDIS), covering the period from October 2017 to July 2018. The yellow areas in Figure 7 represent regions with generally higher wave heights. Based on oceanic climate characteristics, the North Atlantic exhibits relatively high mean significant wave heights of 3–5 m during winter and spring (October–December 2017, January–March 2018), while the mean wave height decreases to 1.5–2.5 m in summer (April–June 2018 and July 2018) [53]. In the central South Pacific, influenced by the westerly wind belt, mean wave heights remain high throughout the year, but become even higher during the austral winter (corresponding to April–July 2018 in the Northern Hemisphere), with averages of approximately 3–5 m [54]. The South Indian Ocean/Southern Ocean exhibits a similar pattern, with mean wave heights of 4–6 m in winter and 2–4 m in summer [55]. In the Bering Sea–Gulf of Alaska region, mean wave heights range from 3–4.5 m in winter and spring, and drop to 1.5–2.5 m in summer [56]. As shown in Table 1, the variations in beam power of microseism sources M1 and M2 are generally consistent with these wave height patterns. The preliminary conclusion is that higher mean significant wave heights correspond to higher beam power of the microseism sources. Regarding the specific causes responsible for the generation of M1–M4, we conducted a detailed discussion based on previous studies.

Before discussing the individual source regions, we first assess the robustness of the M1–M4 assignments. The multi-array product back-projection reduces the directional ambiguity of single-array beamforming by cross-constraining back-azimuths from geographically separated arrays. For the periods when the YU array was available (October 2017 to April 2018), the consistency between the SD, YG, and YU results supports the reliability of the M1–M4 source assignments. However, for the period from May to July 2018, the YU array had to be replaced by the YC array. The two arrays differ significantly in aperture, inter-station spacing, and geographic setting, which introduces a potential systematic bias in the back-projection results during this transitional period. This is particularly evident for M2 and M3 in the April–June 2018 quarter, where the beam power dropped sharply (Table 1). While we cannot fully exclude the possibility that this drop reflects a genuine decrease in source activity, the array substitution likely contributed to the observed change. We therefore advise caution when interpreting the temporal evolution of source intensities across the April–June 2018 boundary. In addition, the spatial resolution and detection capability of the beamforming are inherently constrained by the array response function of each array. Some source regions may fall into blind zones of a particular array, which could lead to an underestimation of their energy. Despite these uncertainties, the multi-array product approach still provides a more robust source attribution than any single array alone, because the product of back-projections from different arrays suppresses spurious local sources and retains only those that are consistently detected from multiple viewpoints.

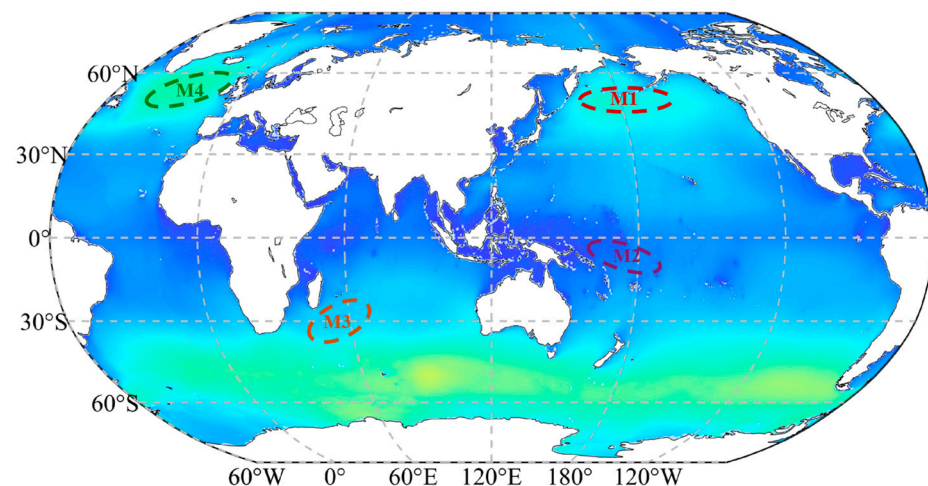


Figure 7. Monthly averaged global significant ocean wave heights in October 2017–July 2018, the yellow areas represent regions where wave heights are generally elevated, the blue areas represent the sea regions, the dashed ellipses denote the identified microseism source regions, labeled M1–M4.

For the M1 source region, Safarkhani [57] acquired continuous vertical-component seismic records from 19 broadband seismic stations spanning January to December 2019 and January to December 2021, of which 13 stations belong to the German Regional Seismic Network (GR). Using beamforming and multi-array back-projection methods to analyze microseisms in the North Pacific and North Atlantic, it was found that during the boreal winter months (October–March), the highest mean wave heights occur in the Northern Hemisphere, particularly in the North Atlantic and North Pacific, indicating dominant oceanic activity in these regions. This pattern shifts to the Southern Hemisphere during the boreal summer (April–September), with increased storm activity in the South Atlantic and South Pacific. Safarkhani pointing observations are consistent with those of Lu et al. [58], who noted that these seismic waves originate in the Southern Hemisphere and are more readily detected during periods of reduced Atlantic wave activity in winter when Southern Hemisphere wave activity intensifies. Subsequently, Safarkhani investigated the polarization of Rayleigh and Love waves in primary and secondary microseisms using two distinct methods: array-derived rotation and three-component beamforming. For the Norwegian array, both Love and Rayleigh waves exhibit highly stable back-azimuth distributions across all seasons, indicating the presence of a persistent, dominant microseism source in the North Atlantic. Aster et al. [43] analyzed global microseism intensity in the secondary (4–10 s) and primary (14–20 s) microseism bands, as well as in eight 2 s-bandwidth period segments between 4 and 20 s, using data from 73 globally distributed stations with continuous records exceeding 20 years (from the late 1980s to 2025). The results show that 84% of the stations exhibit significant primary microseism enhancement, and 63% exhibit significant secondary microseism enhancement. The global mean energy growth trend is $0.27 \pm 0.08\%/yr$ in the primary band and $0.15 \pm 0.09\%/yr$ in the secondary band, with the most pronounced enhancement occurring in the longer-period portion of the primary microseism band. This is consistent with the broadening trend of the global wave spectrum of approximately $0.01\%/yr$, likely reflecting an increased frequency of large storm systems. Wang et al. [59] used continuous seismic data from the YL array and four AU stations during the life cycle of Tropical Cyclone TOMAS in March 2010 to perform $f-k$ analysis and polarization analysis on secondary microseisms (0.1–0.5 Hz). The results show that when the wind speed at the cyclone center reached typhoon strength, the azimuth of the maximum secondary microseism energy aligned

with the moving direction of TOMAS; high-speed wind waves not only generate high swells but also excite secondary microseisms. Furthermore, the observation of microseisms at the AU stations was more effective than at the YL array, which may be attributed to hydrophone vibrations and tilts caused by seafloor turbulence. $f-k$ analysis outperforms polarization analysis in slowness and back-azimuth estimation, although large earthquakes can interfere with its observations. Pandey et al. [60] investigated microseisms in Australia using broadband data from December 2013 to November 2014. Their results revealed new body wave phases from a Southern Hemisphere perspective, including P waves from French Polynesia and core phases from the North Atlantic. Our findings for M1 (North Atlantic–Northern Europe source) are consistent with Safarkhani’s identification of a persistent North Atlantic source and with Aster et al.’s observation that primary microseisms show long-term intensification trends. The stability of M1 across our four quarters supports the view that the North Atlantic is a year-round microseism source.

For M2, located in the open ocean area of the central South Pacific, and M3, belonging to the South Indian Ocean source region, this vast expanse of ocean is commonly referred to as the “Southern Ocean” for research convenience. Albuquerque et al. [61] employed downscaling methods based on global climate models (GCMs) under the CMIP5 framework to simulate historical (1979–2005) and future (2026–2045 and 2081–2100) wave climate around New Zealand. After validating the model reliability against buoy data, they focused on wave changes under the high-emission RCP8.5 scenario. The results show that the wave climate around New Zealand is primarily controlled by the Southern Ocean westerly wind belt, exhibiting a pronounced spatial pattern of generally increased significant wave height and wave period along the southwest coast and decreased along the northeast coast. These trends are closely linked to the positive phase intensification of the Southern Annular Mode (SAM) and the southward shift of the westerly belt. Ardhuin et al. [18], extensively citing microseismic data from the Southern Ocean, developed the first comprehensive numerical model for microseism generation by stochastic ocean waves (including wave reflection). They found that the generation of Southern Ocean microseism sources is influenced not only by the “prevailing westerlies” but also significantly by reflections from the Antarctic Peninsula, islands, and icebergs. Based on model results, the causes of microseism generation in the Southern Ocean can be categorized into three main types: wind seas with a broad directional spectrum (Type I), sea states with significant coastal reflection contributions (Type II), and the interaction of two independent wave systems (Type III). Turner et al. [62] analyzed seismic data from 11 stations along the East Antarctic coast, as well as in the Indian Ocean, Australia, and New Zealand, spanning 12 to 26 years. Combined with sea ice extent and the SAM climate index, they calculated the power spectral density of the seismic wavefield to investigate variations in microseism intensity. The results show that microseism intensity at Antarctic stations is primarily controlled by changes in sea ice extent, exhibiting a seasonal pattern with a phase shift of 4–5 months relative to stations on other continents. Extreme intensity peaks at East Antarctic stations occur in March–April (austral autumn), and the highest peaks of secondary microseisms coincide with negative SAM events. These findings reveal the significant modulation of Southern Ocean microseism generation by sea ice and climate variability. Our identification of M2 and M3 confirms the presence of strong Southern Hemisphere sources in the prevailing westerlies belt, consistent with the climatic controls described by Albuquerque et al. and Turner et al. However, our study adds a new perspective by locating these sources from geographically distant arrays (SD, YG, and YU/YC) and by quantifying their seasonal beam power variations, which were not directly reported in those earlier studies.

For M4, situated along the southern Bering Sea, the Gulf of Alaska, and the Aleutian Islands, it belongs to the North Pacific source region. Pyle et al. [63] applied a back-

projection method to analyze data recorded in January 2010 from five International Monitoring System arrays located along the Pacific coast, investigating the source locations of P-wave microseisms within a narrow frequency band (0.67–1.33 Hz) significantly higher than the classical microseismic band (~0.05–0.3 Hz). The study revealed that high-frequency P-wave energy originates from the North Pacific basin, indicating that P-wave microseisms in the North Pacific ambient noise cover a broad frequency range. Neale et al. [64] compared simulated P-wave microseism sources with those observed at an array in California and investigated the relationship between the observed sources and significant wave height. They found that the time-varying locations of North Pacific microseism sources, mapped using beamforming and back-projection of seismic data, exhibited an accuracy of better than 10° in 90% of the cases. Our M4 source, located in the Bering Sea–Gulf of Alaska–Aleutian Islands region, is consistent with the North Pacific sources reported by Pyle et al. and Neale et al. The novelty of our contribution lies in extending this observation to the 50 mHz primary microseism band (whereas previous studies focused on higher-frequency P-wave microseisms) and in providing a multi-array cross-constrained location, which enhances the reliability of the source assignment.

Based on the causative factors of the four microseismic source regions M1–M4, their excitation mechanisms can be grouped into two major categories: the Northern Hemisphere sources (M1 and M4) are primarily driven by persistent storm activity and direct ocean-wave forcing, whereas the Southern Hemisphere sources (M2 and M3) involve more complex factors, including the westerly wind belt, coastal/iceberg reflections, and multi-system wave interactions. Specifically, M1 (North Atlantic/Northern Europe source) and M4 (Bering Sea–Gulf of Alaska–Aleutian Islands, North Pacific source) are both located at mid-to-high latitudes in the Northern Hemisphere. Their common core mechanism is perennial active marine storms and wave activity. The North Atlantic source is known for its relatively stable year-round storm systems, providing a continuous background energy for microseisms. The North Pacific source (especially the Gulf of Alaska region) also experiences frequent cyclonic activity, where persistent ocean waves couple directly into the solid Earth to generate microseisms. The difference lies in the fact that M1 exhibits stronger stability, maintaining a certain level of microseismic radiation even in summer, whereas the excitation of M4 may depend more strongly on seasonal wind-wave conditions. M2 (open ocean of the central South Pacific) and M3 (South Indian Ocean source, i.e., the Southern Ocean) are both strongly driven by the prevailing westerlies in the Southern Hemisphere. Their uniqueness lies in the fact that, besides the direct effect of wind-generated waves (Type I), they also involve coastal, island and iceberg reflections (Type II) and nonlinear interactions between different wave systems (Type III). This makes the excitation mechanisms in the Southern Ocean more diverse than those in the Northern Hemisphere. On one hand, the westerlies produce the most energetic wave climate on Earth; on the other hand, the presence of the Antarctic Peninsula and surrounding islands and ice shelves significantly enhances the distribution of wave-wave interaction sources. Consequently, the microseisms generated by M2/M3 are not only strong in energy but also exhibit a broader spatial distribution and more complex temporal variations. In comparison, the Northern Hemisphere sources (M1, M4) are dominated by a “single-source” mechanism (storm-generated gravity waves coupling with the seafloor), with relatively concentrated source locations and strong seasonality (winter-dominant). The Southern Hemisphere sources (M2, M3) are characterized by “multi-process superposition”; reflections and nonlinear energy transfers make the microseismic radiation more persistent and broader in frequency bandwidth. Even during periods without strong storms, a significant microseismic background can be maintained through swell-swell interactions. Furthermore, the year-round stability of M1 (North Atlantic) distinctly differentiates it from the other three source regions, while M4 (North Pacific) and the Southern

Ocean sources (M2/M3) are both controlled by the wind-belt systems of their respective hemispheres, but the latter additionally benefit from reflection contributions due to their high-latitude near-polar environment.

The continuous detection of signals from the four microseismic source regions M1–M4 by the SD, YG, and YU arrays results from the synergistic combination of source strength, propagation path, and array geometry. First, all four source regions possess stable and high-intensity microseismic radiation capabilities: the North Atlantic (M1) is driven by year-round storms, the North Pacific (M4) has persistent wave activity, and the Southern Ocean sources (M2, M3) involve the superposition of the prevailing westerlies, coastal/iceberg reflections, and multi-system wave interactions, enabling them to radiate energy continuously in the low-frequency band around 50 mHz. Second, low-frequency seismic waves attenuate slowly in the ocean-crust waveguide; this “highway-like” propagation allows signals thousands of kilometers away to still retain a sufficient signal-to-noise ratio when arriving at the receiving arrays. Finally, the YG array has a large aperture and dense station distribution, providing high angular resolution and directional sensitivity. The YU array also has a relatively large aperture, which is favorable for capturing teleseismic signals from the Atlantic direction. Although the SD array (e.g., the dense Lingshan Island array) has a limited coverage, its dense deployment facilitates extraction of coherent signals within specific frequency bands. The geometric layout of these three arrays happens to form favorable reception angles with the great-circle paths to the target source regions, enabling stable separation and localization of microseismic energy from the M1–M4 areas within the background noise. Therefore, the key to continuous detection lies in the combination of strong sources, low-attenuation paths, and the directional match of array geometry.

Taking the M2 source region as an example, as shown in Table 1 and Figure 7, during the period from April to July 2018, the source region M2 in the Southern Hemisphere entered autumn and winter; the wave height gradually increased towards its annual peak (generally greater than 4 m). However, the microseismic beam power dropped sharply from April to near zero, and completely disappeared by July. This contradicts the intuitive expectation that higher wave heights should produce stronger microseisms. Through investigation, we propose that the most likely primary reasons are as follows. M2 is located in the open ocean of the central South Pacific, where microseism generation depends critically on specific types of wave systems rather than simply on wave height. According to the theory of Ardhuin et al. [18] secondary microseisms (frequencies of approximately 0.1–0.5 Hz) are primarily generated by the nonlinear interaction of two nearly oppositely propagating wave trains. The frequency of 50 mHz lies at the lower limit of the primary microseism band, but it may also involve wave-wave interactions. If the winter waves are dominated by unidirectional wind-generated swells from the prevailing westerlies, lacking an opposing swell system (e.g., reflected waves from distant storms), the nonlinear coupling efficiency would be extremely low even under high wave heights. Moreover, primary microseisms (0.02–0.1 Hz) are mainly generated by direct coupling of ocean waves with the seabed topography, requiring wave energy to be concentrated in shallow water regions such as continental shelves or around islands. M2 is situated in a deep-ocean basin. If the winter storm paths keep the wave energy largely in deep water rather than allowing it to propagate to nearby shallow areas (e.g., the coasts of New Zealand or western South America), the coupling efficiency would be substantially reduced. In short, a mismatch between wave direction and effective coupling (winter dominated by unidirectional wind-driven swells, lacking opposing swells or shallow-water coupling) leads to poor excitation of 50 mHz microseisms even under high wave heights. We also propose a secondary reason, as noted earlier: the replacement of the YU array by the YC array after May 2018 introduced data inconsistency, which may have rendered the M2 signal

undetectable. An additional reason could be seasonal migration of the source region into an array response blind zone, or away from the optimal reception angle of the propagation path.

In summary, this study demonstrates that multi-array product back-projection is an effective tool for constraining global microseism sources at 50 mHz. The four identified source regions (M1–M4) confirm the coupling between ocean wave height and microseismic radiation, and our results are broadly consistent with previous studies. The new contributions of this work include: (i) the first systematic localization of 50 mHz Rayleigh-wave microseism sources using a multi-array product approach that combines the SD array with the YG, YU, and YC arrays; (ii) the quantification of seasonal beam power variations for each source region, which provides a baseline for future monitoring; and (iii) the explicit assessment of the uncertainties introduced by array geometry and array substitution, which offers a reference for designing multi-array experiments. These findings not only advance our understanding of ocean–solid Earth coupling but also lay a methodological foundation for using distributed seismic arrays to monitor ocean wave climate and to image Earth structure from microseismic noise.

5. Conclusions

We used data from the SD array from October 2017 to July 2018 and applied the beamforming method to investigate the distribution of Rayleigh wave microseisms. The classic microseismic frequency of 0.05 Hz, which falls within the PM band, was selected for analysis. Our results show clear temporal and seasonal variations at this frequency. Subsequently, we employed multi-array product back-projection as an effective tool for global source localization. By combining the SD, YG, and YU arrays (supplemented by the YC array), we identified four intense and persistent microseismic source regions worldwide. Their primary locations are as follows: M1 extends eastward to the Norwegian Sea, the North Sea, and the Scandinavian Peninsula, and westward to encompass eastern Greenland, Iceland, and the open ocean of the northeastern North Atlantic; M2 is located in the open central South Pacific Ocean, lying along the path of the prevailing westerlies belt; M3 belongs to the South Indian Ocean source region, also situated within the prevailing westerlies belt; M4 is situated along the southern Bering Sea, the Gulf of Alaska, and the Aleutian Islands. We complemented the analysis with global mean significant wave height data, confirming that the identified microseism sources fundamentally correspond to areas with higher wave heights.

Beyond these specific findings, the multi-array product back-projection method demonstrated in this study offers a practical framework for future microseism source localization. By cross-constraining back-azimuths from geographically separated arrays, it effectively mitigates the directional ambiguity inherent in single-array beamforming, thereby improving the reliability of source attribution. This approach can be extended to ambient seismic noise analysis in other regions and may support the development of passive monitoring tools for ocean wave activity and solid-Earth structure studies.

Nevertheless, the work has several limitations that should be acknowledged. First, the data cover only ten months (October 2017 to July 2018) and do not capture a full annual cycle; the so-called “third quarter” contains data only for July 2018, which limits the characterization of interannual variability. Second, the YU array was replaced by the YC array from May 2018 onward because of data unavailability. The two arrays differ substantially in aperture, inter-station spacing, and geographic setting, and this substitution may have introduced a systematic bias in the source localization during the transitional period (April to June 2018). Third, the spatial resolution and detection capability of the beamforming are inherently dependent on array geometry; certain source regions may fall into blind zones of the array response function and thus remain undetected. Finally, the source

attribution of the four M1–M4 regions relies on the assumption of a surface-wave slowness window (0.2–0.4 s/km), which may not fully capture the complexity of wave propagation across diverse tectonic and bathymetric settings.

Based on the above findings and limitations, we suggest that future studies should extend the temporal coverage to a full multi-year period to assess interannual variability and investigate the potential influence of climate modes such as the El Niño–Southern Oscillation (ENSO) on microseism source intensity. Moreover, future work could combine the multi-array product back-projection with more realistic 3-D Earth models and frequency-dependent slowness windows to further reduce localization uncertainties. The methodological framework developed here may also be applied to other frequency bands and to other geographic regions where dense seismic arrays are available, offering a pathway toward a more complete understanding of global microseism sources and their coupling with ocean wave climate.

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