Investigating the Characteristics of Microseisms using the Australian Seismic Arrays

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Abstract

Earth’s microseisms are continuously recorded by seismographs worldwide. Yet, studies using broadband digital data to analyze microseisms in Australia have been rare. Building on initial research efforts that relied on the Warramunga array in the continent’s center, we expand the investigation of microseisms by utilizing seismic arrays in various locations and with distinct apertures and geometries, particularly spiral-arm arrays. Motivated by expanding knowledge of microseismic sources, we investigate the distribution and characteristics of microseisms. We process one-year continuous waveform data using beamforming at various periods. Using the back-projection, we then investigate the plausible source areas of surface waves and teleseismic P-waves generated by ocean activity. We also examine the seasonal variability of microseismic sources and their relationship with the ocean wave hindcast model by comparing our observations of Rayleigh (Rg) waves with modelled Rg wave sources and juxtaposing the back-projected P-waves with significant wave heights. Our results suggest that over the time interval of several months and longer, Rayleigh waves are the dominant component arriving from the nearby coastlines. They show a transition to higher mode Lg waves in the higher frequency bands. In contrast, on the time scale of days and weeks, teleseismic P-waves from the coastal and pelagic sources are observed particularly from the tropical and equatorial regions. We also identify new patterns of body waves from the perspective of the Southern Hemisphere. Our study highlights the importance of utilizing multiple arrays and elucidates the critical roles of the frequency range and bathymetry.

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Key Points:

- Short-aperture arrays in Australia are used to study microseism excitations, with observability mainly affected by frequency and bathymetry.
- Microseisms induced body waves are observed across various scales, from global to regional, on seismic arrays in Australia.
- Coastal reflection and scattering effect in sedimentary basins play an essential role in shaping microseismic surface waves.
Abstract

Earth’s microseisms are continuously recorded by seismographs worldwide. Yet, studies using broadband digital data to analyze microseisms in Australia have been rare. Building on initial research efforts that relied on the Warramunga array in the continent’s center, we expand the investigation of microseisms by utilizing seismic arrays in various locations and with distinct apertures and geometries, particularly spiral-arm arrays. Motivated by expanding knowledge of microseismic sources, we investigate the distribution and characteristics of microseisms. We process one-year continuous waveform data using beamforming at various periods. Using the back-projection, we then investigate the plausible source areas of surface waves and teleseismic P-waves generated by ocean activity. We also examine the seasonal variability of microseismic sources and their relationship with the ocean wave hindcast model by comparing our observations of Rayleigh (Rg) waves with modelled Rg wave sources and juxtaposing the back-projected P-waves with significant wave heights. Our results suggest that over the time interval of several months and longer, Rayleigh waves are the dominant component arriving from the nearby coastlines. They show a transition to higher mode Lg waves in the higher frequency bands. In contrast, on the time scale of days and weeks, teleseismic P-waves from the coastal and pelagic sources are observed particularly from the tropical and equatorial regions. We also identify new patterns of body waves from the perspective of the Southern Hemisphere. Our study highlights the importance of utilizing multiple arrays and elucidates the critical roles of the frequency range and bathymetry.
Plain Language Summary

This study explores the Earth's microseisms – continuous vibrations recorded worldwide on seismographs. Since there has been limited research on microseisms using broadband digital data in Australia, we aim to investigate their distribution and composition across the continent “girt by the sea”. We analyze year-long seismic data from various Australian seismic arrays, using array techniques to locate the sources of seismic wavefield generated by ocean-wave activities. We also examine the seasonal variations of microseismic sources and analyze the link with the ocean wave model. The results reveal the dependence of microseismic sources on frequency, bathymetry and shoreline configuration. Additionally, the study identifies new microseism sources such as P-waves arriving from French Polynesia islands in the South Pacific Ocean, and the core-sensitive phases from the north-Atlantic Ocean, which have not been reported using the seismic arrays in the southern hemisphere.
1 Introduction

The seismic background vibrations of the Earth, mainly rising from oceans, are referred to as microseisms (Haubrich & McCamy, 1969). Microseisms, long considered “noise”, are the most energetic signals of the ambient seismic field (Nakata et al., 2019). They are generated due to the continuous harmonic force by the ocean waves, mainly driven by ocean storms (Nakata et al., 2019). These fluctuating ground motions of low-amplitude seismic signals are observed globally (Ebeling, 2012) and vary in space, time and frequency.

Early in the 20th century, notable efforts were made in observational seismology to understand the source region of microseisms and the cause of their generation (e.g., Gherzi, 1932; Gutenberg, 1931; Hasselmann, 1963; Longuet-Higgins, 1950; Ramirez, 1940; Scholte, 1943). Spectral analysis shows ambient seismic noise has two distinct peaks, one at approximately 14 s called primary microseisms (PM) and the other at ~7 s called secondary microseisms (SM) (e.g. Berger et al., 2004; Stutzmann, 2000). The PM are weak-amplitude long-period oscillations, or gravity waves generated in shallow water due to direct pressure fluctuations from the interaction between the ocean and the ocean sea floor, particularly in the continental shelf region at the ocean swell period (14 s) (e.g., Hasselmann, 1963; Haubrich et al., 1963). The SM are the most energetic signal dominating the seismic noise spectra, generated by the non-linear interaction between ocean gravity waves that have similar periods and travel in opposite directions (Longuet-Higgins, 1950). Ardhuin et al. (2011) generalized three broad classes of favorable sea-state conditions that are responsible for the generation of SM sources. The first class (class I) occurs when the ocean waves generated by the local winds are present oblique to the mean wave direction, and these oblique waves interact with the waves from the opposite direction to generate seismic noise. These sources are weak and commonly observed above 0.2 Hz. The second class (class II) occurs due to coastal reflections when the ocean waves reflected from the shorelines interact with the incoming waves. These sources potentially exist along the entire coastline. The third class (class III) corresponds to the less frequent sources that occur due to the interactions of two wave systems having a similar dominant frequency and travelling in opposite directions. This can happen when a progressing wave overtakes its previously generated ocean swell within the same storm or when ocean waves from a distant storm travel long distances and
interact with the waves from another storm. This class generates the strongest noise sources, which can produce energy at a frequency as low as 0.05 Hz.

Locating the microseism sources has many practical uses in the global geophysical community. They include but are not limited to: imaging the Earth’s structure by retrieving the Green’s function between two seismic stations from cross-correlating ambient noise (Shapiro & Campillo, 2004; Shapiro et al., 2005) and Rayleigh-wave dispersion curves from energetic hurricanes (e.g., Feng & Chen, 2022; Wang et al., 2019), tracking major storms throughout their life cycle (e.g., Gualtieri et al., 2018; Retailleau & Gualtieri, 2021), investigating climate change by the analysis of microseisms and structural monitoring of volcanoes using ambient noise (e.g., Berger et al., 2004). The frequency of their occurrence and spatial distribution does not limit the microseismic signals as in the case of earthquakes, which is a long-standing problem in the exploration of Earth’s interior. Therefore, investigating the Earth’s structure using ambient seismic noise would require a detailed understanding of the noise source distribution and their compositions.

Recent advances in observational seismology studies analyzing the distribution and composition of the microseisms wavefield utilized seismic arrays across the globe using beamforming methods (e.g. Gal et al., 2015; Gerstoft et al., 2008; Landès et al., 2010; Lepore & Grad, 2020; Wang et al., 2021). It is widely acknowledged that coastal regions are the originating areas for PM (Ardhuin et al., 2015); however, the dominant source regions of SM vary in different parts of the world. Notable efforts have been made focusing on the source locations and distribution of SM, generated due to different classes of sea-state conditions discussed above, which are permanent and moving sources, attributed to both deep water and near coastal source regions (e.g., Gualtieri et al., 2018; Guo et al., 2020; Haubrich & McCamy, 1969; Zhang et al., 2010).

Body and surface waves originating from the SM sources have been identified at the global scale, such as in the northern Pacific, Atlantic Oceans, the southern Pacific and Indian Oceans (e.g., Davy et al., 2015; Gerstoft et al., 2006; Gerstoft et al., 2008; Landès et al., 21010; Pyle et al., 2015; Reading et al., 2014; Toksöz & Lacoss, 1968). Furthermore, studies have identified the localized regions around the continental slope rather than the global source areas which dominate the SM wavefield radiating mainly as surface waves (e.g., Behr et al., 2013; Bromirski, 2001;
Gal et al., 2015; Xiao et al., 2018b). However, uncertainty exists regarding the contribution of shallow and deep-water sources.

Secondary microseisms, also referred to as double-frequency microseisms (DFMs), sometimes split into two groups, with one called ‘long-period double-frequency microseisms’ (LPDFMs) and the other called ‘short-period double-frequency microseisms’ (SPDFMs) (e.g., Bromirski et al., 2005; Koper & Burlacu, 2015; Sun et al., 2013; Xiao et al., 2018a). The split of microseisms in multiple frequency groups varies from region to region and may explain the discrepancy in source locations for the SM.

Most of the microseismical research is based on the arrays in the northern hemisphere (NH) and is limited in the southern hemisphere (SH) with some exceptions (e.g., Behr et al., 2013; Davy et al., 2015; Gal et al., 2015; Reading et al., 2014). The SH regions are known for their turbulent seas and strong winds due to the limited presence of land masses, which otherwise act as barriers. These latitudes have been termed the “Roaring Forties”, “Furious Fifties”, and “Screaming Sixties” due to the strong currents and high winds (Webb, 2019). This results in the formation of large ocean waves and produces stormy conditions, which have the potential to generate strong microseisms. The central limitation and possible bias in interpretation of surface and body wave source regions in the studies of microseisms is the lack of seismic arrays in the SH. This makes identifying the distribution of SM source regions a challenge, and most of the existing studies can only explain the directionality of the microseism arrivals (Behr et al., 2013; Davy et al., 2015).

Australia has pioneered world efforts in the continent-scale imaging by deploying many transportable and temporary arrays in the last several decades (e.g., Eakin, 2018; Fontaine & Kennet, 2007; Jiang & Miller, 2020; Rawlinson et al., 2006; Tkalčić et al., 2020; Van der Hilst and Kennett, 1993). Thus, we can capitalize on the extent and different sizes of the arrays in Australia using the nationwide database AusPass (http://auspass.edu.au/). This study aims to characterize the annual variability and sources of the microseisms observed at different locations in Australia, expanding on previous studies (Gal et al., 2015; Gal et al., 2017; Reading et al., 2014). In this study, we analyze 1-year continuous data recorded by the multiple medium- and
small-aperture seismic arrays of spiral and linear geometry covering various geological settings and locations of the Australian continent to study the distribution and propagation of SM (Figure 1). In particular, the spiral-geometry arrays are optimal for the analysis of seismic wavefield, with a better control over back azimuths and slowness due to a high versatility of inter-station vectors (e.g., Kennett et al., 2015; Stipčević et al., 2017; Tkalčić, 2017).

In order to enhance the knowledge of microseismic sources in the SH, we analyze the source locations of microseisms, corresponding to the SPDFMs (0.20 – 0.35 Hz), as well as short-period microseisms (SPMs) from 0.35 – 0.5 Hz and 0.5 – 0.7 Hz. We attempt to understand the variations in microseismic source generation for different periods using multiple arrays in different parts of Australia, which has been limited to the observations of SPMs (Gal et al., 2015; Gal et al., 2017; Reading et al., 2014). Due to the seasonal variations in seismic noise wavefield, we further investigate the composition and distribution in noise sources over an annual cycle and relate our results to different independent data sets.

2 Data

2.1 Data summary

We use data from the multiple seismic arrays in Australia with different apertures and station configurations: Southern Queensland spiral array (SQspa) located in south-east Australia (Tkalčić et al., 2013), Warramunga array (WRA) located in central Australia, Pilbara seismic array (PSAR) in north-west Australia (De Kool, 2015) and Western Australia spiral array (WAspa) in south-west Australia (Stipčević, 2015) (Figure 1). SQspa and WAspa are temporary arrays deployed by the Research School of Earth Sciences, Australian National University and operated for over a year from Nov 2013 – Nov 2014 and Feb – Dec 2015 respectively. WRA is a long-established array operated by the Australian National University as a part of International Monitoring System (IMS) network. PSAR is a permanent array operated by the Geoscience Australia. We utilize vertical component continuous waveforms from SQspa, PSA and WRA from Nov 2013 to Nov 2014 which will provide a more comprehensive understanding of microseisms and elucidate their distribution patterns, and WAspa from Feb 2015 to Dec 2015.
Figure 1. The map shows the distribution of the Australian short-aperture seismic arrays used in this study. The red triangles indicate their geographic locations. The bottom panel shows the individual array’s configuration.

2.2 Properties of the arrays

The four arrays have small to medium apertures (Figure 1). SQspa is a spiral arm array with an aperture of ~50 kms and has 16 stations along 3 spiral arms. WAspa has a similar distribution with an aperture of ~44 kms. WRA is a L-shaped array with ~25 kms aperture and a total of 24 stations with an interstation spacing of 2-3 kms. PSAR has an aperture of 13 kms and 13 stations along 3 spiral arms with logarithmic spacing.

The aperture plays a crucial role in establishing the frequencies at which an array can have a reasonable slowness resolution. The Array Response Function (ARF) is useful to test the sensitivity and resolution of the arrays (e.g., Rost & Thomas, 2002; Tkalčić, 2017). We examine the ARF on the four seismic arrays of different apertures and geometries with a slowness of 0.0 s/deg at frequencies 0.20 Hz and 0.5Hz for medium-aperture arrays SQspa and WAspa covering the secondary and short-period microseisms band, and 0.35 Hz and 0.7 Hz for small-aperture arrays PSAR and WRA, for studying the short-period microseismic sources (Figure 2). We found that, for all the 4 arrays at the respective frequencies, the central peak is very strong in comparison to the side lobes, indicating a reasonable resolution of these four arrays for discriminating the signals in this frequency range.
Figure 2. Array response functions are calculated to show the slowness resolution for the arrays: Southern Queensland spiral array (SQspa) and Western Australia spiral array (WAspa) at frequencies of 0.20 Hz and 0.5 Hz, respectively; Warramunga array (WRA) and Pilbara seismic array (PSAR) at frequencies of 0.35 Hz and 0.7 Hz, respectively.

2.3 Spectral characteristics of microseisms in Australia
To investigate the characteristics of microseisms, we plot the power spectral density (PSD) probability density functions (PDFs) with mean PSD at all the four arrays located in different geographic locations to understand the noise levels and peaks during typhoon and non-typhoon periods (Figure 3). For comparison, we also plot the noise curves of new low-noise model (NLNM) and new high-noise model (NHNM) (Peterson, 1993).

To first order, background noise levels and mean PSD of SQspa (Figure 3a and 3b) and WAspa (Figure 3c and 3d) lies within the new NLNM/NHNM during the typhoon and non-typhoon periods. At the short periods of the spectrum below 1 second, the background noise at WRA and PSAR shows the decrease in power levels below the NLNM (Figure 3), which is most likely linked to the least anthropogenic noise and reflects the better network performance at short periods. The background noises of WRA (Figure 3e and 3f) and PSAR (Figure 3g and 3h) demonstrate a variability in power levels and show elevated power above the NHNM during typhoon time in the higher periods. The intensification of noise levels by typhoons is clearly observed at WAspa, WRA and PSAR at periods above 10 seconds (Figure 3), which indicates that these arrays are close to the microseism source regions.

The split of DFMs into LPDFMs and SPDFMs is clearly observed during a non-typhoon period. The LPDFMs and SPDFMs peak generally varies from region to region. Here, we observed the LPDFMs peak at ~7 seconds (0.14 Hz) and SPDFMs peak at ~4 seconds (0.25 Hz). The SPDFMs peak is not obvious during the typhoon period at WAspa, WRA and PSAR (Figure 3), which is most likely obscured by the higher noise levels, or the sources are quite far away. SQspa shows clear and higher energy levels of SPDFMs and also shows multiple noise peaks below the DFMs period band (Figure 3a and 3b).
March
(a) SQspa
(b) SQspa
(c) WAspa
(d) WAspa
(e) WRA
(f) WRA
(g) PSAR
(h) PSAR
October

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Figure 3. Power spectral density (PSD) probability density functions (PDFs) of every other PSD for the vertical component during the typhoon (March) and non-typhoon (October) period at SQspa (a) and (b), WAspa (c) and (d), WRA (e) and (f), and PSAR (g) and (h). Color bars show probability density in per cent. The Peterson new low-/high-noise (NLNM/NHNM) (Peterson, 1993) models are shown in each panel as black solid lines. Mean PSDs for each array are shown as gray line. Mean PSDs are plotted in decibels relative to ground acceleration with units of $10 \log_{10}[m^2/s^4/Hz]$.

Considering the relatively shorter aperture of the arrays, we further investigate the characteristics of SPDFMs (0.20 – 0.35 Hz), and SPMs (0.35 – 0.5 Hz and 0.5 – 0.7 Hz) by identifying their distribution and propagation using Capon method (Capon, 1969) and investigate their source regions using the back-projection method.

3 Methods

3.1 Capon method

The absence of a distinct impulsive onset in microseismic events poses a challenge for single-element station networks in accurately determining the source back azimuth and slowness. Consequently, array analysis offers significant potential for advancing the understanding of microseisms, as it facilitates the estimation of source back azimuth and slowness (e.g. Schweitzer et al., 2002; Rost and Thomas, 2009). Microseisms are superposition of travelling waves with finite frequency bands generated in the far field. Therefore, they can be characterized by frequency-wavenumber density function, which provides information on the power as a function of frequency and vector velocities of the propagating waves. One commonly used method for analyzing seismic ambient noise is the modified frequency-wavenumber (f-k) algorithm introduced by Capon (1969) known as Capon method.

In our work, we utilize the method developed by Gal et al. (2014) known as IAS (Incoherently Averaged Signal) Capon method, to determine the slowness spectrum of body and surface waves microseisms. This method is an improved implementation of the Capon method for array
analysis of seismic noise. The method prevents the occurrence of frequency mixing within the cross-power spectral density matrix (CSDM), resulting in a precise estimation of the slowness vector for the incoming seismic waves. This implementation computes individual slowness estimations for each frequency bin and then average the resulting slowness spectra within a defined frequency range.

Following the notations of Gal and Reading (2019), the resulting beam power is defined as:

\[ P(f,s) = \frac{1}{M^2} w^H(f,s)R(f)w(f,s) \]  

where, \( P \) is the resultant beam power, \( w \) is the weight of the beamformer, \( R \) is the CSDM which holds information on the auto and cross-spectra of the input data, and \( H \) denotes the conjugate transpose.

In the conventional f-k method, the weights are only depending on the array geometry and defined as:

\[ w_{fk} = a(f,s) \]  

where \( a \) is the array steering vector.

However, the data structure is an important parameter in the Capon method to estimate the power spectrum, as the weights used in the Capon method are derived from the input data (Gal and Reading, 2019). The weights in Capon method are defined as:

\[ w_{Capon} = \frac{R^{-1}a(f,s)}{a^H(f,s)R^{-1}a(f,s)} \]  

where \( C^{-1} \) is the inverse CSDM and \( a \) is the array steering vector.
The algorithm developed by Gal et al. (2014) computes slowness spectrum for different frequency bins and average the resulting spectra over a specific frequency range. Further in the IAS method, beamforming is optimized for microseisms by temporal and frequency averaging, and diagonal loading, which is especially useful in removing the influence of the transient signals such as earthquakes and obtain a more robust estimates (Gal and Reading, 2019).

We divide the time series in 1-hour time segment with 50 percent overlap between the adjacent time segments and estimate the spectra for each hour. The analysis is performed for a year in 3 frequency bands: 0.20 – 0.35 Hz, 0.35 – 0.5 Hz and 0.5 – 0.7 Hz. The 1-hour time segment is further divided into multiple non-overlapping sub-windows depending upon the number of stations and the frequency range. Finally, we compute the hourly spectra by averaging the slowness spectra of each sub window.

We incorporate days during which 75 % of the daily data is recorded on the seismic array to mitigate any anomalous contributions to the power spectrum. We test the influence of earthquakes on the power spectrum in the next section. The method resolves multiple arrivals in each time window and recover the slownesses with a better resolution. We apply the method to the SQspa, WAspa, PSAR and WRA array.

**3.1.1 Effect of earthquakes**

Earthquakes are unwanted signals in the analysis of microseisms. Earthquakes have imprint on the power spectrum and degrade the beamforming analysis. Discarding earthquake-perturbed window are the common approaches to reduce the effect of earthquake signals. We run an experiment in order to determine if the IAS Capon method can accurately resolve the microseisms arrival in the presence of earthquakes and the effect of earthquakes on the power spectrum in different time periods. At first, we include the snapshots containing earthquakes to study the influence on the microseisms, and in the second step we remove the windows that contain earthquake signals and evaluate the microseisms wavefield.
We analyze the effect of earthquakes at SQspa in 0.35 – 0.5 Hz frequency band for body waves in different time durations consisting of a day, a week, 15 days, and a month (Figure S1 and S2). To identify snapshots that are perturbed by earthquake signals, we use an event catalogue (ISC Bulletin) to identify time frames where strong earthquakes with magnitude > 5.5 occurred, and further compute the power spectrum by including and excluding them. The list of earthquakes is given in supplementary Table S1.

There are 6 events in a one-day beamforming stack, 13 events in a weekly beamforming stack, 25 events in 15 day beamforming stack and 51 events on a monthly beamforming stack. In a one-day stack, earthquake energy is strong and visible in the north-west direction, however in the beamforming analysis of 7 days, 15 days and 1 month, the energy from the earthquakes diminishes and don’t have effect on the microseisms wavefield. The power spectrum is similar to the spectra in the absence of earthquakes (Figure S1 and S2).

3.2 Back-projection method

Back-projection is an array-based method for searching for source locations (Ishii et al., 2005). This technique involves mapping the source distribution by backprojecting observed body waves using travel time information. The back-projection method relies on a straightforward delay-and-sum beamforming algorithm, where time delays are parameterized using the theoretical travel time calculated for a particular seismic phase and source location (Zhang et al., 2023). It has been shown that the back-projection method is useful for improving the ability to determine source locations of P-wave microseisms (e.g., Gal et al., 2015; Gerstoft et al., 2008; Euler et al., 2014; Liu et al., 2016).

We apply a back-projection method to perform a grid search for possible source locations of P-wave microseisms in the SM at SQspa, WAspa and WRA where body waves are observed using Capon method and determine the most probable source regions from which the globe microseismic energy is coming. In contrast to the Capon method, this method does not assume plane wave propagation across the array, and it relies on a 1-D reference velocity model.
For the purpose of backprojecting body waves, we consider the assumption of P-wave propagation within the distance range of 30°– 90° for the propagation of both regional and teleseismic P-waves. Since P and PP have the same slowness at different distances, this could lead to ambiguity in source localization. Therefore, we only back-project for P-waves, which are the arrivals most apparent in SM period (Astiz et al., 1996).

We process each array data individually and only consider the time window which is recorded by at least 75 % of all the array stations. We remove the mean and trend and filter each trace in different frequency bands: 0.20 – 0.35 Hz and 0.35 – 0.5 Hz for the waveforms recorded at SQspa and WAspa arrays, and 0.35 – 0.5 Hz and 0.5 – 0.7 Hz for WRA array. We do not explicitly remove earthquake signals from the data before doing the back-projection. Instead, we apply median across multiple time windows to dampen the transient earthquake energy and amplify the stationary microseism energy. We verified our method by testing an earthquake (2014-04-13T12:36:19.000000Z) of magnitude 7.4 at 34.2-km depth, located in the Solomon Islands region (-11.47 N, 161.96 E) using the multi-array back-projection approach. The method works successfully by determining the plausible source area of earthquake location by utilizing SQspa and WRA simultaneously (Figure S3). The spatial uncertainty in earthquake location arises from the limitations imposed by the aperture of the seismic arrays. We further attempt different stacking intervals on the waveforms from 1 to 6-hour and obtain similar results in each time interval, suggesting that the P-wave microseismic sources are quite stable as a function of time. We found that 3-hour represents an optimal time interval for stacking to obtain a stable location map of microseisms, in terms of resolution and computation cost. We normalize the time windows and compute the fourth-root beams which reduce the influence of anomalous waveforms and enhance the coherency (Pyle et al., 2015). We consider a 5° × 5° geographical grid on the globe and compute the beam power by back-projecting the seismic energy from continuous waveforms to each grid point, assuming the P-wave travel time calculated using 1-D velocity model ak135 (Kennett et al., 1995).

We then study body waves microseismic sources in monthly stacks at various seismic periods to understand the spatio-temporal distribution of microseism source regions and their variation with frequency. We further compare our observations with the global significant wave height derived
from ocean wave hindcast model (Ardhuin et al., 2011) and explore how SM sources are related to the local sea conditions.

### 3.3 Multi-array Back-projection of Surface waves

The use of simultaneous multiple arrays can provide better constraints on the source regions in terms of resolution and more accurately determine the source of incoming energy (Stipčević et al., 2017). To locate the potential source regions of surface waves along the great circle path, we utilize the multi-array back-projection method for surface waves (Wang et al., 2021). Firstly, we exploit the Capon beamforming results, and compute the annual hitcount plots of surface waves (22 – 40 s/deg) at each array. The hitcount is defined as the number of times of surface wave arrivals in a grid cell of slowness and back azimuth, and it is computed for each hour by considering the sources in the top 15% of the power levels.

Next, we discretize the earth's surface as $1^\circ \times 1^\circ$ grids for the region surrounding Australia. For each grid, we compute the back azimuths from the grid to each array center. Finally, we back-project the hitcount results of each array to the regional $1^\circ \times 1^\circ$ grid based on their corresponding back azimuth. We back-project the maximum power in the annual hitcount plot to the grid points along a particular back azimuth. Then we utilize simultaneous multiple seismic arrays (SQspa, WAspa, WRA and PSAR) and employ the multi-array back-projection to compute the total beam power as the product of individual beam power of all the arrays and investigate the source regions of surface waves. We further compare our observations with the modelled Rg wave sources averaged over a year (Dec 2013 – Nov 2014) based on the IFREMER model (Ardhuin et al., 2011) and understand the excitation of surface wave source regions.

### 4 Results

#### 4.1 Beamforming results

To understand the microseismic wavefield observed in Australia, we investigate the typical pattern of incoming seismic signals on the spiral-arm and linear arrays situated in different parts
of Australia. We plot a summary diagram for SQspa (Figure 4), PSAR (Figure 6), WRA (Figure 7) and WAspa (Figure 8) in different frequency bands to observe the detailed variation of microseisms with frequency. We also explore the frequency-azimuth properties of body waves with focus on SQspa (Figure 5) and WAspa (Figure 9), in order to understand the capability of these spiral-arm arrays for detecting body waves microseisms. We identify the dominant and weaker arrivals by computing the slowness spectrum for each quarter and observe different modes of propagation of SM arrivals and their azimuthal distribution.

4.1.1 Distribution of SM arrivals on SQspa

We analyze the microseism arrivals on SQspa in two frequency bands: 0.20 – 0.35 Hz corresponding to the SPDFMs, and 0.35 – 0.5 Hz corresponding to the SPMs (Figure 4).
Figure 4. The summary of Capon plots for the slowness (along the radial direction from the origin) and back azimuth (clockwise direction from the north) of body and surface wave microseism in the following frequency ranges: (a) 0.20 – 0.35 Hz and (b) 0.35 – 0.50 Hz recorded at the Southern Queensland spiral array (SQspa). Only the vertical component seismograms were utilized in the analysis. From top to bottom, Quarter 1 (Q1), or the NH winter
Our results suggest that the arrivals of surface waves span over a wide range in the southern directions in SPDFMs frequency band (Figure 4a). The surface wave peak is observed at the low phase velocity of around 3.3 km/s, (slowness of 35 s/deg), which corresponds to the fundamental mode Rayleigh (Rg) waves, and they are the dominant microseism arrivals. The Rg waves mainly arrive from south to southeast, while there are seasonally varying weak arrivals from southwest, for instance, they are the strongest in the winter months (Q3). There is a fundamental bimodality between the body and surface waves. Strong body waves are present with strength comparable to that of surface waves in 0.20 – 0.35 Hz frequency range (Figure 4a). Body wave sources are more significantly present between 4.5 and 12 s/deg and mainly arrive from the south direction (southeast to southwest) in all the seasons except Q1 (Figure 5a), which is consistent with teleseismic P-waves. In Q1, body wave arrivals with slowness less than 4.5 s/deg, corresponding to the core-sensitive phases are prominent and arrive from the northern direction, which is consistent with the enhanced seismic noise from the distant storms in the NH winter (Figure 5a). Based on the monthly azimuth-slowness spectrum from the beamforming results, we observe signals with slowness less than 4.5 s/deg from the north-eastern direction in Feb month of SPDFMs microseisms at SQspa (Figure S4), which likely corresponds to the PKP core phases because PcP has a much lower amplitude (Gerstoft et al., 2008). However, we do not focus on the branch of PKP phase. We do not observe any other signal from P-phase slownesses in beamforming results (Figure S4) during this time. To the best of our knowledge, PKP energy from the north Atlantic Ocean are observed for the first-time using arrays in Australia. Further classification of these arrivals is beyond the scope of the current study.
Figure 5. Same as Figure 4, zoomed-in Capon plots for the slowness and back azimuth of body wave microseism. The outer and inner circles are drawn at 12 s/deg and 4.5 s/deg corresponding to teleseismic P-wave phases.
While in the SPMs, Lg waves are the dominant arrivals, mainly arriving from the southeast direction. Lg waves are the supercritical S waves trapped in the crust, attributed to the apparent velocity of 4.1 km/s (slowness of 25 s/deg) and separated by 1-1.5 km/s from Rg waves. The dominant direction of Lg waves shifted with a minor change in back azimuth (Figure 4b) in comparison to SPDFMs (Figure 4a). The Lg waves are predominantly observed for continental source-receiver paths because Lg waves do not propagate in the oceanic crust (Zhang & Lay, 1995). Furthermore, teleseismic body waves are coming from the southwest direction in all the seasons (Figure 5b). Body wave from the south-southeast varies in strength and direction in different seasons. However, in Q1 strong arrivals are from the north direction which is consistent with the strong ocean-wave activity during the NH winter (Figure 5b). These arrivals are most likely associated with the teleseismic P-waves, diving deep into the mantle (slowness ~ 4.5 s/deg) arriving from the epicentral distance of 90°.

4.1.2 Distribution of SM arrivals on PSAR

Pilbara Seismic Array in northwestern Australia is almost half the size of SQspa and WAspa, having an aperture of ~13 kms. By analyzing the ARF, PSAR has a good resolution in the higher frequency bands 0.35 – 0.5 Hz and 0.5 – 0.7 Hz, which corresponds to the SPMs.

In both the frequency bands, Rg waves are the dominant arrivals and we do not observe body waves. Rg waves mainly arrive from the north, northwest and southwest direction (Figure 6). In 0.35 – 0.5 Hz, the dominant Rg waves arrive from the southwest in Q1, Q2 and Q4, while in Q3 we observe a seasonal change in dominant arrival, and it shifts to the north to northwest direction during the SH winter (Figure 6a). In higher frequency band of 0.5 – 0.7 Hz, Rg waves from the north are dominant throughout the year. Furthermore, strong Rg wave signals are arriving from northwest during SH winter (Q3), which becomes weak in the rest of the year (Figure 6b). The directionality of surface waves in the frequency bands (0.35 – 0.5 Hz and 0.5 – 0.7 Hz) are consistent with the previous study on the PSAR by Gal et al. (2017). Seasonality is prevalent for Rg waves during the winter months in both frequency bands, and there are no body waves and Lg waves observed at PSAR.
Figure 6. Similar to Figure 4, the data are analyzed for the PSAR during Dec 2013 – Nov 2014.

4.1.3 Distribution of SM arrivals on WRA
Warramunga Array (WRA) has a good slowness resolution in higher frequency bands, similar to PSAR. We analyze the microseism arrivals on WRA in SPMs: 0.35 – 0.5 Hz and 0.5 – 0.7 Hz during the time period same as those on SQspa and PSAR, Dec 2013 – Nov 2014. However, the data for Q3 months are not available. Therefore, we compute the summary plots of slowness-back azimuth spectrum for rest of the year: Q1, Q2 and Q4, and observe the variability of microseismic sources at WRA.

At WRA, teleseismic P-waves are the dominant arrivals with strength comparable to the surface waves in both frequency bands, and mainly come from the north and south directions (Figure 7), having the similar pattern observed at SQspa (Figure 5). The P-wave sources from the north are strong in Q1, and sources from the south are stronger in Q2 and Q4 (Figure S5), showing seasonal variations. In 0.35 – 0.5 Hz, Rg waves arrive from the south, northwest and north-northeast directions, and varies seasonally (Figure 7a). In the higher frequency range of 0.5 – 0.7 Hz, we observe a transition from Rg to Lg waves arriving from the north-northeast direction with similar back azimuth as Rg waves (Figure 7b). Both Lg and Rg waves shows seasonal variabilities in this frequency band. Our results of WRA agree well with the observations of Gal et al. (2015).
Figure 7. Similar to Figure 4, the data are analyzed for the WRA during Dec 2013 – Nov 2014.

4.1.4 Distribution of SM arrivals on WAspa

WAspa has a similar array configuration and aperture as SQspa, and therefore we analyzed the microseisms in the same frequency bands as for SQspa. This entails the SPDFMs and SPMs: 0.20 – 0.35 Hz and 0.35 – 0.5 Hz.
Figure 8. Similar to Figure 4 legend except the data are analyzed for the WAspa during Feb 2015 – Dec 2015.

In SPDFMs, the sources are not stable over the 1 year of analysis, having varying slownesses (Figure 8a). In Q1 and Q2, we observe the fundamental mode Rg waves, while in Q3 and Q4 higher modes Rg waves are observed (Figure 8a). Teleseismic P-waves from the southwest
direction in Q1 are observed in SPDFMs (Figure 9a), consistent with the Rg wave direction (Figure 8a). In other seasons, body waves observed are likely the regional P-waves arriving from the north-west and south-west directions (Figure 9a).

In SPMs, we observe dominant Rg waves from the southeast and southwest direction, which varies seasonally (Figure 8b). The dominant Rg wave signals arrive from the southeast during Q2 and Q3, while in Q4 the dominant direction changes to southwest and remains active in Q1 (Figure 8b). Moreover, the observed body waves have varying slownesses between 4.5 and 15 s/deg in Q1, corresponding to regional and teleseismic P-wave sources (Figure 9b). Except Q1, during the rest of the year, we observe body waves with slowness between 10 s/deg and 15 s/deg from the northeast direction, which corresponds to regional P-wave sources mostly generated due to local wind condition in the nearby ocean (Figure 9b). Weak body wave arrivals are observed from the northwest with a little seasonality (Figure 9b). At WAspa, the slowness of body and surface waves vary substantially which reduces the coherency in signals.
Figure 9. Similar to Figure 5, the data are analyzed for the WAspa during Feb 2015 – Dec 2015.

4.2 Source locations of surface waves
We further estimate the annual hitcount plots of incoming energy and infer the location of surface wave source region (in the slowness range 22 – 40 s/deg) using the multi-array back-projection method (e.g., Wang et al., 2021) in the SPMs 0.35 – 0.5 Hz. Despite WAspa was not operational simultaneously with SQspa, PSAR and WRA, we assume that the averaged source over a year remains consistent. Considering surface waves generation due to coastal reflections in the nearby coastlines as the primary cause and also observed by several studies in the past (e.g., Behr et al., 2013; Gal et al., 2015; Xiao et al., 2018b), the dominant sources are stationary over the years. Integrating WAspa in the multi-array analysis, can potentially facilitate inferring the location of microseism sources over an annual scale. We plot the annually stacked hitcount for SQspa, WAspa, PSAR and WRA to identify the dominant sources of surface waves that are strong throughout the year for SPDFMs (Figure 10a, 10c, 10e and 10g). Next, we plot the surface wave arrival distribution for SPMs by backprojecting the annual hitcount result along the great circle path for each array (Figure 10b, 10d, 10f and 10h). This gives us an overview of the distribution of surface wave arrivals.

Our result suggests that the surface waves mainly arrive from the south: covering from southeast to southwest at all the arrays (Figure 10), except at SQspa where Lg waves are observed (Figure 9g and 9h). In addition, surface waves arrive from the north and north-west direction at both PSAR (Figure 10a and 10b) and WRA (Figure 10c and 10d) as these arrays are located close to the northern coastline of Australia. For WAspa, moderate surface wave arrivals are observed from south and north-west (Figure 10e and 10f). At SQspa, Lg wave sources arrive from the entire range of east coast of Australia (Figure 10g and 10h).
Figure 10. Annual stacked hitcount results in 0.35 – 0.55 Hz at: (a) PSAR, (c) WRA, (e) WAspa and (g) SQspa. Summary plot of surface waves arrival directional distribution in SPMs frequency bands: 0.35 – 0.5 Hz along the great circle path using the back-projection technique at: (b) PSAR, (d) WRA, (f) WAspa and (h) SQspa. Red triangles denote the location of the seismic arrays.

Based on the distribution of arrivals and employing a combination of three arrays, namely WAspa, WRA, and PSAR, within the frequency band of 0.35 – 0.5 Hz, we apply the multi-array back-projection technique to delineate potential source regions of surface waves (Figure 11). Given that only Lg waves are detected in SPDFMs at SQspa, SQspa is excluded from the multi-array back-projection analysis. Our analysis reveals six distinct source regions proximal to the coastline surrounding Australia denoted as S1, S2, S3, S4, S5 and S6. Specifically, S1 is identified in the vicinity of the Great Australian Bight region, S2 near the southwest tip of Australia, S3 in proximity to Shark Bay, S4 from the Islands in Indonesia northwest of Australia, while S4 and S5 are situated to the north of PSAR, within the Indian Ocean and among the Indonesian islands.
Figure 1. Surface waves source regions inferred from the multi-array back-projection. Red triangles denote the seismic arrays. S1, S2, S3, S4, S5 and S6 represent the plausible source areas of surface waves generated from the nearby coastal region surrounding Australia.

S1, located within the Great Australian Bight region (Figure 11), is a strong microseismic source region corroborated by previous investigations (Reading et al., 2014; Gal et al., 2015). S3 was identified as a strong source near the Shark Bay region in a comprehensive study by Gal et al. (2017). Our findings from multiple arrays indicate that the microseisms observed in Australia primarily originate from nearby coastal regions rather than global sources.

4.3 Source locations of body waves

The monthly back-projections of the beam powers of the SQspa, WAspa and WRA arrays in the SM frequency band are shown in Figures 12, 13 and 14. The back-projection images of the three arrays are different in the prominent source distribution. SQspa and WRA show the source distribution is dominated by global sources, while WAspa is mostly dominated by regional sources. Further, we compared back-projected P-waves with the quarterly stacked beamforming result and found a good correlation except for a few seasons.

4.3.1 Seasonal variability of P-wave microseism source areas observed at SQspa

Monthly observations from Dec 2013 – Nov 2014, shows a seasonal variability of P-wave microseism source areas in both SPDFMs and SPMs (Figure 12). In SH summer, Bering Sea and North Pacific Ocean are active regions of P-waves in SPDFMs (Figure 12a), and Sea of Okhotsk and east coast of Japan generates strong P-waves in SPMs (Figure 12b). In SH winter, South Pacific Ocean and South Indian Ocean are strong source areas of SPDFMs (Figure 12a), while in SPMs strong P-wave sources are generated near French Polynesia Islands and Kerguelen Plateau (KP) (Figure 12b). During the rest of the year (autumn and spring), South Pacific Ocean (southeast of New Zealand) is the dominant source region for SPDFMs (Figure 12a). For SPMs, southern Indian Ocean (coast of Antarctica), KP, Southern Ocean (south of Australia) and South Pacific Ocean (southeast of New Zealand) are active P-wave source regions (Figure 12b).
In addition, we observe P-wave microseisms from the Philippines Sea in SPMs during the late SH winter until spring (Figure 12b). In Apr, strong P-wave arrivals from the north were observed near the Solomon Islands (Figure 12b), which correlates with the tropical cyclone Ita, which was an intensity-5 tropical cyclone in Apr 2014. This suggest the strong influence of cyclone activity on SPMs (Figure 12b) while it is not observed in SPDFMs (Figure 12a). The impact of cyclones on the microseism wavefield has been intensively studied and observed in different regions of the globe (Davy et al., 2015; Xiao et al., 2018a). The shifting locations of the P-wave arrivals in SPDFMs and SPMs during SH summer are most likely due to the ocean site effects, as those similarly observed by Gal et al. (2015). The back-projection result (Figure 12) is consistent with the beamforming result (Figure 5) in both frequency bands. Further, the strong seasonal variability in the P-wave source regions at SQspa is consistent with the strong ocean wave activity in the NH and SH.

To the best of our knowledge, the P-wave microseisms near the Polynesia Islands has been identified as one of the source areas for Rg waves in SM (Wang et al., 2021), however it has never been reported as a dominant body wave SPMs sources in SH winter (Figure 12b).
Figure 12. P-wave sources from monthly averaged back-projection analysis observed at SQspa during Dec 2013 – Nov 2014 in frequency bands: (a) SPDFMs (0.20 – 0.35 Hz) and (b) SPMs (0.35 – 0.5 Hz).

4.3.2 Seasonal variability of P-wave microseism source areas observed at WRA

During the period from Dec 2013 – Nov 2014 at WRA, the data is missing from Jun 2014 to Aug 2014. We identify the P-wave microseism source regions in the rest of the months and address the variability in dominant sources identified in SPMs: 0.35 – 0.5 Hz and 0.5 – 0.7 Hz.

In SH summer, intense wave activity across the North Pacific dominates P-wave microseism generation in both frequency bands, mainly arriving from the Philippine Sea, east coast of Japan and Sea of Okhotsk (Figure 13). During the autumn months, open sea near the Hawaii Islands emerges as the predominant P-wave source region in SPMs lower frequency band, shifting to southern Pacific Ocean in May (Figure 13a), whereas in the higher frequency band, deep waters of the North Pacific ocean is a strong P-wave source region and moderate P-wave arrivals are coming from the Solomon Islands and East China Sea near Taiwan (Figure 13b). In the spring months, South Indian Ocean (south of Australia) near the coast of Antarctica is a dominant P-wave source region in the lower frequency band (Figure 13a), and in the higher frequency band deep waters in the South Indian Ocean and Southern Ocean also dominate as P-wave source regions (Figure 13b). In Apr, we observe strong P-wave sources from the Solomon Islands which dominates the microseismic wavefield in 0.5 – 0.7 Hz (Figure 13b) and in the low frequency band 0.35 – 0.5 Hz (Figure 13a). This correlates with the observation of tropical cyclone Ita similarly observed at SQspa (Figure 12b). Interestingly, at WRA the P-wave arrival from cyclone dominates the prominent microseismic wavefield, which was not the case at SQspa. This suggests that ocean storms are observed at arrays differently, even they were at the same proximity to the storm location, and the predominant direction of microseisms changed greatly. The cyclone has varying degrees of influence on the microseismic wavefield at SQspa (Figure 12) and WRA (Figure 13) in different frequency bands. This was similarly observed by Xiao et al. (2018a), where they studied the influence of different typhoons on the OBS’s and land stations in China.
Seasonality is prevalent for the P-wave sources observed at WRA (Figure 13). The back-projection result is consistent with the observations of beamforming results (Figure S5) in both bands, and the P-wave sources are consistent with the intense ocean activity between the NH and SH. The P-wave source regions such as north of Japan, KP, southern Pacific Ocean, Taiwan align with the findings of Gal et al. (2015), who identified P-wave microseismic sources within the SPMs frequency range of 0.325 – 0.725 Hz, based on WRA data.
Figure 13. Similar to Figure 12, the data are analyzed for the WRA array in frequency bands: (a) 0.35 – 0.5 Hz and (b) 0.5 – 0.7 Hz.

4.3.3 Seasonal variability of P-wave microseism source areas observed at WAspa

WAspa was operated during Feb 2015 – Dec 2015. Since it has the similar aperture like SQspa, we compute the back-projection SPDFMs and SPMs and determine their seasonal variability. Our analysis suggests that the P-wave microseisms observed are dominated by the regional
sources in both frequency bands throughout the year generated mainly around the coastlines (Figure 14). Indian Ocean near Madagascar and Antarctica, KP, coastal regions near the Papua New Guinea and Solomon Islands and south-east Asian region near Singapore and Malaysia are the strong P-wave source regions (Figure 14). In addition, Philippine Sea during Dec, French Polynesia Islands during Mar and Oct, east coast of Australia during Apr, Arabian Sea during Aug, north of Japan during Nov are active P-wave source regions in SPMs (Figure 14b). Polynesia Islands are source regions of body waves in SPMs at SQspa as well during SH winters (Figure 12b).

In SPDFMs, the dominant slowness peak observed in beamforming results from the southwest direction during Q1 (Figure 9a) corresponds to back-projected P-waves from KP in Dec (Figure 14a), indicating teleseismic P-wave sources. For the remainder of the year (Q2, Q3, and Q4 in Figure 9a), regional sources prevail, consistent with back-projected P-waves from the Indian Ocean northeast and southeast of WAspa (Figure 14a). In SPMs, beamforming results during Q1 reveal strong teleseismic P-wave arrivals from the northeast and south directions (Figure 9b), aligning with back-projected P-wave source regions in Dec originating from the Philippine Sea and South Indian Ocean near Antarctica (Figure 14b). Conversely, beamforming during the rest of the year exhibits robust regional P-wave arrivals from the northwest (Q2, Q3, and Q4 in Figure 9b), corresponding to back-projected P-waves from Papua New Guinea (Figure 14b).

WAspa is dominated by the regional P-wave microseism sources rather than the global source regions observed at SQspa and WRA. There is no seasonality, and the coastal component is stronger for P-wave microseism source regions in both the bands (Figure 14).
Figure 14. Similar to Figure 12, the data are analyzed for the WAspa array during 2015 (Feb – Dec).
5 Discussion

We firstly discuss the limitation in identifying the source regions of surface and body waves. In general, microseismic noise sources are characterized by broad source regions. However, due to the relatively short-aperture arrays used in our study, the slowness-back azimuth estimates have a greater spatial uncertainty. This introduces substantial errors in the surface waves source location S1 (Figure 11) determined using the multi-array back-projection from WAspa and WRA. Moreover, the uncertainty in back-projecting P-wave sources exist due to the same reason, which significantly reduces the accuracy in locating P-wave source regions. Apart from the aperture of the seismic arrays, the grid spacing finer than $5^\circ \times 5^\circ$ (used in this study) is required to delineate the source regions (pelagic vs. coastal) in different frequency bands. Although the location accuracy is relatively poor and the noise sources generally have a large geographic extent, our analysis reveals from which part of the globe the P-wave microseisms arrive in Australia. We note that we only focus on relative amplitudes because we are primarily interested in the location of body wave sources.

Furthermore, we observe Lg waves at SQspa and WRA in the higher frequency bands: 0.35 – 0.5 Hz and 0.5 – 0.7 Hz, respectively, while no Lg waves are observed at PSAR and WAspa. Using the IMS arrays in short-period microseisms, Koper et al. (2010) suggested that the irregular morphology of coastlines mainly contributes to the generation of Lg waves. Furthermore, using two decades of continuous waveforms from the WRA array, Gal et al. (2015) suggested irregular morphology or scattering effect could contribute to the Lg-wave generation. However, we observe the variation in the Lg wave propagation in the eastern and western parts of Australia. It has been shown that eastern Australia has a relatively high attenuation of Lg waves compared to western and central Australia, which is in good agreement with the thinner crust and sedimentary basins in the east (Wei et al., 2017). The sedimentary basins enhance the crustal heterogeneity and promote scattering effects. A numerical study by Xie & Lay (1994) that used explosive seismic sources found that scattering effects can transfer Rg into Lg waves. In our study, Rg and Lg waves arrive from the similar back azimuth at WRA while their back azimuths at SQspa. Is different. We suggest that the Rg to Lg wave conversion is most likely due to the scattering
effects in the crustal heterogeneities or a combination of both the scattering effects and irregular morphology of the coastlines.

One puzzling feature of this study is the unstable observations of surface and body waves at WAspa. One potential reason may be the closer locations of microseismic sources, which are poorly approximated by a plane wave. The surface and body waves in beamforming result correspond to the higher slowness of higher modes Rg waves (Q3 and Q4 in Figure 8a) and regional P-waves (Figure 9), respectively. This can also be observed from back-projection results. The proximity of back-projected P-wave sources in the Indian Ocean, Southeast Asian regions, east coast of Australia and KP indicate that these are regional microseismic sources at a distance of less than 60° (Figure 14) and consistent with the beamforming result (Figure 9).

Additionally, the slowness varies substantially within the span of 3 months, which reduces the stability and coherency of the averaged signals. The variation in slowness can be confirmed from the monthly back-projection stacks of body waves, where P-wave source regions are not fixed within the same seasons (Figure 14). Monthly beamforming stacks of SPDFMs further confirms the variation of slownesses in the same quarters (Figure S6). The observation of higher mode Rg waves at WAspa further confirms the influence of the nearby coastlines on microseismic sources. Source regions around the continental slope are commonly observed for surface waves, however our analysis suggests that the ocean waves reflected from the coastlines and interacting with the incoming waves can be a dominant generating factor for body waves as well, which implies that the interference of waves far from the storm activity significantly contributes to the microseismic body wave spectrum at WAspa. This suggests that at WAspa, the influence of ocean wave interactions is more noticeable or significant compared to the influence of bathymetry.

Next, we show the wave height derived from the ocean hindcast model (Ardhuin et al., 2011) were significant during Dec 2013 to Nov 2014 (Figure 15) and compare their locations with the monthly averaged back-projected P-wave sources observed at SQspa (Figure 12). This ocean wave model considers the coastal reflections. Significant wave height is considered a good proxy for cyclone activity during different seasons and provides a relevant comparison with the source regions of teleseismic compressional waves. The monthly averaged ocean wave height (Figure
agrees well with the source regions observed at SQspa (Figure 12). The strong P and PKP-wave sources observed during the SH summers in beamforming (Figure 5 and S4) and back-projection (Figure 12) are consistent with the large wave heights in the North Pacific and North Atlantic Ocean (Figure 15). During the rest of the year, SH is a more active region and shows strong storm activity (Figure 15) and agrees well with the P-wave source regions observed in the South Indian and Pacific Oceans (Figure 12). This is most likely due to minimal continental barriers in SH and thus promulgates robust storm activity, leading to the widespread microseismic events, radiating not only P, but also surface waves. However, we observe some inconsistency between the average wave height and P-wave sources; for instance, in the SH winters, back-projection result shows a strong P-wave source near Polynesia (Figure 12b), which is not visible in the monthly averaged hindcasts (Figure 15). Furthermore, in Sep and Oct, the strong P-wave sources from the north of Australia near Papua New Guinea (Figure 12b) are not observed in the monthly averaged ocean wave height (Figure 15). This suggests that the extreme wave heights do not always excite powerful microseisms.
Figure 15. Average significant wave height in meters (m) for each month extracted from the ocean wave model (Ardhuin et al., 2011) from Dec 2013 – Nov 2014.

Finally, we compare the observed and modelled microseismic Rayleigh wave sources. The locations of microseisms obtained by back-projection of the surface wave arrivals determined using different arrays give us an estimate of source regions, but to improve the source locations, we compute the power spectrum of the vertical ground displacement for Rayleigh wave sources. The numerical modelling of microseisms is a valuable approach to understanding the generation of Rg waves (Ardhuin et al., 2011) and helps identify the coastal and pelagic dominance in different frequency bands. A brief review of Rayleigh wave source modelling is described in Text S1.
Figure 16. (a) Bathymetric model of Australia with distribution of the 4 arrays used in the study. (b) Rayleigh wave noise source regions utilizing the ocean wave model of Ardhuin et al. (2011) by considering coastal reflections of wave energy for the SM for the frequency band 0.35 – 0.5 Hz. The source regions observed through multi-array back-projection are shown by black circles.

Figure 16a shows the bathymetric model and highlights the potential of the bathymetry to generate seismic noise in SPMs (Figure 16). The bathymetry coinciding with the source regions are the continental slopes and identified as the region of high excitation. Figure 16 highlights the theoretical noise sources averaged over a year from Dec 2013 to Nov 2014, including the reflections from shorelines in the SPDFMs. The strong modelled sources of Rg waves in SPDFMs are located along the nearby coastline of Australia (Figure 16) and correlate well with the observed back-projected surface wave source regions from the nearest coastlines on different arrays, namely S1 (Great Australian Bight), S2 (Windy Harbour), S3 (Shark Bay) and S6 (Indonesia) (Figure 11).

6 Conclusions

In order to understand the microseismic wavefield observed in Australia, we have evaluated 1 year of continuous waveform data from the Southern Queensland spiral-arm array in the eastern part of Australia, Warramunga array in the central Australia, Pilbara seismic array in the northwest Australia and Western Australia spiral-arm array in the southwest Australia. This suite of relatively short aperture arrays covers the coastal areas and inland and is suitable for a comprehensive study of SM wavefield and documenting the variation in microseisms wave type. Building upon earlier research by Reading et al. (2014), Gal et al. (2015) and Gal et al. (2017), this analysis offers further understanding of the spatio-temporal characteristics of ambient seismic noise observed in Australia. We find that variations in frequency, shorelines and structure beneath the arrays are important factors influencing the sources of microseisms and signal strength.
The surface wave source region determined by the multi-array back-projection technique shows that coastal reflection plays a pivotal role in surface waves generation, which is widely observed in many studies (e.g., Gal et al., 2015; Gerstoft et al., 2008; Xiao et al., 2018b). Furthermore, modelling of Rg waves in SPMs using ocean wave model confirms that result. Our conclusion is that surface wave source regions observed in our study are primarily attributed to the class II type mechanism of the sea-state (Ardhuin et al., 2011), supported by bathymetric effects.

There is a clear distinction observed in the body wave sources at SQspa, WRA and WAspa. Both the SQspa and WRA are dominated by global P-wave source regions, while WAspa is dominated by the regional P-wave microseism sources near the equatorial regions. The P-wave microseism source areas located using SQspa and WRA show seasonal variability, in contrast to those identifying with WAspa. Further, the shared source regions identified at SQspa, WAspa, and WRA, such as those in the North Pacific Ocean, can be more precisely delineated through the application of multi-array techniques. Additionally, the identification of source regions originating from the open sea near the French Polynesia islands, observed at SQspa (Figure 12b) and WAspa (Figure 14b) during different seasons, highlight the robustness of microseismic sources in the equatorial regions.

At SQspa, SPDFMs primarily emanate from deeper waters (Figure 12a), contrasting with SPMs, which exhibit dominance in shallower regions (Figure 12b). Similarly, at WRA, SPMs predominantly occur in shallower regions (Figure 13a), while SPMs within the higher frequency band (0.5 – 0.7 Hz) prevail in deeper waters (Figure 13b). This suggest that the variation in signal strength and source regions for body waves is frequency dependent, which means different sources dominate in different frequency bands. We reveal close correlations between the observed microseismic sources and modelled Rg wave sources as well as ocean wave height and back-projected P-wave source regions. The Rg to Lg wave transition is prominently observed at SQspa and WRA, located in the eastern half of Australia, while no Lg waves are observed on the seismic arrays located on the western side. This transition is most likely due to the scattering effects caused by the crustal heterogeneities or a combination of scattering effects and irregular morphology of the coastlines. A more sophisticated numerical modelling approach is required to understand the generation mechanism of Lg waves.
Finally, we identified a new pattern of body wave arrivals using the seismic arrays in Australia, including the core-sensitive signal originating in the northern part of the Atlantic Ocean, and P-wave sources originating near the Polynesian islands in the open sea.

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Open Research

The IAS Capon package used for beamforming analysis of seismic ambient noise can be found in Gal et al. (2014), available at https://github.com/mgalcode/IAS-Capon. The documentation on the ocean wave model can be found in Ardhuin et al. (2011) and output of the model can be accessed at ftp://ftp.ifremer.fr/ifremer/ww3/HINDCAST/SISMO/. Events used in this study are from ISC Catalogue.

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