Origin of deep ocean microseims by using teleseismic body waves

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Abstract.

Recent studies of oceanic microseisms have concentrate on fundamental-mode surface waves. Extraction of fundamental-mode Rayleigh and Love wave Green functions from station-station correlations of ambient seismic noise has recently been demonstrated to be a very powerful tool for imaging of the Earth's crust and uppermost mantle.

In this study we concentrate on energetic arrivals in two frequency bands around the primary (14s) and the secondary (7s) microseismic peaks that appear at near-zero times in noise cross-correlations. Thanks to a polarisation analysis of data from the the ETSE network (Turkey), we identify this "near-zero time" signal as an upcoming P wave in the secondary microseismic frequency band (5-10s). In a second step, analysing noise cross-correlations from three different arrays (in Yellowstone, in Turkey and in Kyrgyzstan), we determine the origin of these signals by means of beamforming analysis and its projection on the Earth.

Our results show that, in the 0.1-0.3 Hz frequency band, the energetic "near-zero" time arrivals in seismic noise cross-correlations are mainly formed by teleseismic P, PP, and PKP waves. Generation of this ambient body waves in the secondary microseismic band presents a marked seasonal behaviour with sources located in southern and northern oceans during summer and winter, respectively. Moreover, body wave array analysis is accurate enough to confirm that significant amount of the microseism energy is generated far from the coast in deep oceans.

1. Introduction

Recent years witnessed a strong interest in studying background seismic noise. One of the reasons for this interest is the possibility to extract deterministic Green functions from correlations of a random wavefield that can be proved mathematically with different approaches (e.g., *Lobkis and Weaver* [2001]; *Snieder* [2004]; *Gouédard et al.* [2008]) and that has been demonstrated in acoustic laboratory experiments (e.g., *Lobkis and Weaver* [2001]; *Derode et al.* [2003]). Application of this principle to large amount of continuous digital seismic records provided by modern networks provides us with new approaches for seismic tomography (e.g., *Shapiro and Campillo* [2004]; *Shapiro et al.* [2005]; *Sabra et al.* [2005]; *Yang et al.* [2007]; *Stehly et al.* [2009]) and monitoring (e.g., *Sens-Schönfelder and Wegler* [2006]; *Brenguier et al.* [2008b]; *Brenguier et al.* [2008a]). Detailed analysis of high-quality continuous records also allows us to better understand the origin of the ambient seismic noise and its relation to oceanic and atmospheric processes. New methods of noise-based seismic imaging and monitoring are based on a principle that the Green function between two points can be extracted by correlating a random wavefield recorded by receivers located in these points. In other words, one of the two receivers can be considered as a virtual source recorded by the second receiver. This principle is especially attractive when applied in context of random wavefield recorded by a network of numerous recorders. In this case, by computing all possible inter-station cross-correlations it becomes possible to place virtual sources at every receiver location and to have their records by the whole network resulting in a very dense path coverage. The deterministic waveforms (Green functions) extracted from the cross-correlations can be used then to measure and to invert travel times with different methods developed in context of earthquake or explosion based seismology.

The noise based Green function reconstruction implies, however, some strong hypothesis about the noise modal composition. A perfect reconstruction can be achieved for an ideally random and equipartitioned wavefield (*Lobkis and Weaver* [2001]; *Sánchez-Sesma and Campillo* [2006]). In a case of seismic noise it would imply that its sources should be distributed randomly and homogeneously in volume. This is obviously not the case for the real ambient seismic noise within the Earth. First, most of its sources are located on the surface resulting in stronger presence of fundamental mode surface waves and their relatively easy reconstruction from inter-station cross-correlations. For the same reasons, extracting body wave Green functions from noise cross-correlations remains challenging. Second, the distribution of noise sources is not perfectly random and homogeneous. Background seismic oscillations are mostly generated by the forcing from oceanic gravity and infragravity waves. The interaction between these oceanic waves and the solid Earth is governed by a complex non-linear mechanism (e.g., *Longuet-Higgins* [1950]) and, as a result, the noise excitation depends on many factors such as the intensity of the oceanic waves but also the intensity of their interferences as well as the seafloor topography (e.g., *Kedar et al.* [2008]). Overall, the generation of seismic noise is strongly modulated by strong oceanic storms and therefore, has a clear seasonal and non-random pattern.

Distribution of noise sources homogenizes when considered over long times (more than one year). The homogenization and randomization of the noise wavefield is also enhanced by the scattering of the seismic waves on the small-scale heterogeneity within the Earth. Also, because of the stationary phase principle, a cross-correlation of the noise recorded by two receivers is dominated by contribution from sources located in vicinity of the line connecting these receivers. Therefore, even without a perfectly homogeneous distribution, a presence of sufficient amount of favorably located noise sources results in relatively high quality reconstruction of fundamental mode surface waves. As a consequence, reconstructing surface waves from correlations of seismic noise and measuring their dispersion curves works rather well. However, further improving the accuracy of the noise based measurements needed to develop new high-resolution imaging and monitoring methods requires better understanding of the noise modal content and its evolution in space and time. Taking into account realistic distribution of noise sources is also necessary to be able to apply waveform inversion approaches to the noise correlations.

Seismic noise spectra contains two prominent peaks at 0.05 - 0.1 and 0.1 - 0.3 Hz called primary and secondary microseisms, respectively. The primary microseism originates from direct forcing of strong oceanic waves while the secondary microseism which is characterized by stronger amplitudes is produced at double frequency by a non-linear interaction of these waves as suggested by *Longuet-Higgins* [1950]. Both microseismic peaks are dominated by fundamental mode surface waves. It is currently debated whether the surface wave component of microseisms is generated primarily along coastlines (e.g., *Friedrich et al.* [1998]; *Bromirski and Duennebier* [2002]; *Essen et al.* [2003]; *Schulte-Pelkum et al.* [2004]; *Rhie and Romanowicz* [2006]; *Yang and Ritzwoller* [2008]) or if it is also generated in deep-sea areas (*Cessaro* [1994]; *Stehly et al.* [2006]; *Chevrot et al.* [2007]; *Kedar et al.* [2008]). At the same time, body waves were detected in the secondary microseismic band using dense seismic arrays (e.g., *Backus et al.* [1964]; *Toksöz and Lacoss* [1968]; *Seriff et al.* [1965]; *Iyer and Healy* [1972]; *Koper and de Foy* [2008]; *Gerstoft et al.* [2008]) and can be often associated with specific storms (e.g., *Gerstoft et al.* [2006]). Inhomogeneous distribution of noise sources is clearly revealed by the asymmetry of noise cross-correlations observed in both primary and secondary microseismic bands (e.g., *Stehly et al.* [2006]; *Yang and Ritzwoller* [2008]). According to Longuet-Higgins' theory, the generation of secondary microseisms is associated with the non-linear interaction of swells propagating in opposite directions. Such a configuration can be encountered in the coastal region where incident and reflected waves are likely present. This is the case that is considered as prominent by the seismologists based on the observations of the radiation by individual storms (e.g., *Bromirski and Duennebier* [2002]; *Gerstoft and Tanimoto* [2007]; *Bromirski* [2009]). Although there is little doubt that a part of the ambient noise is related with the interaction of oceanic waves with the coast, it is not the only situation where waves propagating in opposite directions are encountered. *Kedar et al.* [2008] used a wave action model to implement Longuet Higgins theory and found that particular regions in the deep oceans are potential sources of secondary microseism excitation. This is related to specific conditions of meteorological forcing associated with resonances of the water column. Their results indicate that secondary microseisms can be generated in specific deep-water areas with one example in the Atlantic ocean south of Greenland.

To investigate the location of the sources of the background noise, we use seismological data averaged over long time series. The Rayleigh wave part of the noise in the secondary microseism period band rapidly attenuates with distance. It is therefore difficult to assess the locations of sources when the signal is dominated by the closest source, often the closest coast, that hides the remote sources. To overcome this difficulty, we use P wave at teleseismic distances recorded in continental environments.

We compute noise cross-correlations for three seismic arrays located within continents in the Northern hemisphere. During summer months, when most of strong storms are located in the Southern hemisphere, the observed noise cross-correlations are dominated by arrivals at near-zero times. Polarization analysis clearly indicates that these arrivals are composed of teleseismic P-waves. We then use a beamforming analysis to determine precisely back-azimuths and slowness corresponding to these arrivals and to backproject them to the regions where the energy was generated based on ray-tracing in a global spherically symmetric Earth model.

1.1. Polarization analysis to detect teleseismic P-waves in noise cross-correlations

In this section, we use the data of the Eastern Turkey Seismic Experiment (ETSE) that operated a temporary network of 20 broadband stations between October 1999 and August 2001 (Figure 1) (*Sandvol et al.* [2003]).

1.1.1. Data pre-processing

Polarization analysis of noise cross-correlations requires preserving the amplitude ratio between components. Therefore, standard processing for the computation of noise cross-correlations such as one-bit normalization and spectral whitening, can not be applied. Instead, data are corrected for instrumental responses, resampled to 1 Hz and filtered between 0.01 and 0.3 Hz. A water level of 4 times the standard deviation of each record is used to decrease the amplitude effect of earthquakes on cross-correlations. Furthermore, for each station and each component, only daily records with mean energy smaller than the whole experiment mean energy are used for the polarization analysis.

1.1.2. Noise cross-correlations

Figures 2 and 3 show cross-correlations between vertical noise records (ZZ) plotted with respect to distance between stations for two different seasons and two frequency bands which correspond to primary and secondary microseismic peaks (0.05-0.1 Hz and 0.1-0.3 Hz). A propagating wave with apparent velocity close to 3 km/s (red dashed lines) is observed at negative and positive times. This time-symmetrical signal which is stronger in the 0.05-0.1 Hz bandpass (Figures 2a and 3a) is the Rayleigh wave part of the Green function reconstructed from random noise correlations. Another signal with apparent velocity larger than 10 km/s (blue dashed lines) is dominant in the 0.1-0.3 Hz bandpass (Figures 2a and 3a). This signal with very high apparent velocity is stronger during northern summer than winter. We hypothesize that those fast arrivals are P waves with steep incidence angle that are generated by very distant sources. During northern summer strong secondary microseisms are mostly expected to be generated within oceans in the southern hemisphere. For such sources, relatively short period surface-waves are attenuated because of large propagating distances. This may explain why the noise correlations are dominated by near-zero-time body wave arrivals. We test this hypothesis with a polarization analysis on cross-correlation signals. We demonstrate that the polarization of a plane wave recorded at two stations can be reliably estimated from the multicomponent cross-correlations.

1.1.3. Polarization analysis of a plane wave from its cross-correlation records : Method

Jurkevics [1988] studied the polarization of different waves emitted by an earthquake using the covariance matrix of 3-component record. He recovered the polarization angle and the azimuth from the eigenvectors of the covariance matrix. We demonstrate in the Appendix that the covariance matrix of the components at a single station (S_{CovSig}) is proportional to the covariance matrix of the cross-correlation signals ($S_{CovCorr}$) (Equation A10) for a plane P-wave propagating across a network of stations. Therefore, the eigenvectors are identical and the polarization analysis can be performed either on cross-correlations records at 2 stations or on 3-component record at a single station.

As suggested by *Jurkevics* [1988], the eigenvalues of the covariance matrix ($\lambda_1 > \lambda_2 > \lambda_3$) are used to compute the coefficient of rectilinearity R:

$$R = 1 - \frac{\lambda_2 + \lambda_3}{2\lambda_1}. \tag{1}$$

which is equal to 1 for a rectilinear polarization. The eigenvector corresponding to the largest eigenvalue λ_1 gives the polarization angle and azimuth of the plane wave. The conversion from the polarization angle (φ) to the incidence angle (I) is obtained from the displacement equations for a reflected P wave at the free surface given by *Aki and Richards* [1980].

We use a ratio between P and S waves velocities of : $V_P/V_S = \sqrt{3}$ for this conversion.

1.1.4. Polarization analysis of cross-correlations computed with ETSE data

We use only station pairs with distance larger than 50 km to compute the covariance matrix of the 3 cross-correlations ZE, ZN and ZZ. To prevent any influence of the Rayleigh wave (group velocity 3 km/s), we select time windows between -10s and 10s. Among the 127,490 cross-correlation signals available for 671 days and 190 station pairs, 23,449 signals are selected based on signal-to-noise (see section 1.1.1) and minimum distance criterion. This rather small percentage (18%) is due to time-variable data availability and quality.

Daily records including earthquakes or glitches are removed from the database due to our water-level amplitude filter.

Using eigenvalues and eigenvectors of the covariance matrix, we compute the rectilinearity coefficient (equation 1), the azimuth and the incidence angle for every inter-station cross-correlation. The rectilinearity coefficient over the whole experiment is 0.84 ± 0.12, which shows that the polarization of the studied wave is almost linear.

Figure 4 shows particle motion for 2 daily cross-correlations and 2 station pairs. Particle motion is shown in the horizontal plane (ZN as a function of ZE) and in the vertical propagation plane defined by the measured azimuth angle (ZZ as a function of ZH) where ZH is obtained by the rotation of the ZE and ZN components of the correlations using the azimuth angle measured from the polarization analysis. The rotation can be computed after the correlation because no non-linear processing such as one-bit transform or spectral whitening has been applied to the data. The red dashed lines in Figure 4 display the azimuth and incidence angle obtained from the covariance method. We observe that the displacement is stronger on the vertical component than on the horizontal ones suggesting that the signal observed on the cross-correlations at near-zero times is composed of nearly vertically incident P waves.

We then estimate the incidence angle and the azimuth of the body wave detected from correlations of ambient noise records and investigate their possible seasonal variations. Figure 5 shows the probability of occurrence of a given value of incidence angle (Figure 5a) and azimuth (Figure 5b) in a time period of 20 days evaluated from all daily records and station pairs. Figure 5a shows that we detect P waves with steep incidence angles during the whole experiment. It also documents a seasonal change of the incidence angle from an average of 15° in summer to 25° in winter, with a more accurate measurement of the incidence angle in the summer than in the winter (larger probability of occurrence). We observe the exact opposite in Figure 5b with better determined azimuths in winter than in summer, simply because the azimuth can not be evaluated for an almost vertically incident P wave.

The seasonal variation observed for the incidence angle is even clearer for the azimuth (Figure 5b). In summer, the average azimuth close to 0° shows that sources of the P waves are located south of the ETSE network. Winter noise sources are located north-west of the network as documented by azimuths close to 150°. Those observations are consistent with seasonal changes in the behavior of seismic noise sources (e.g. *Stehly et al.* [2006]; *Tanimoto et al.* [2006]). The precise location of the sources of the P wave component of the noise will be investigated in the following section.

2. Locating seismic noise sources with a beamforming analysis

To determine regions that generate these body waves, we perform a beamforming analysis of the noise cross-correlations using the whole network as an array. We use only vertical components where the body waves are mostly detected. When studying a single component, we do not need to preserve the amplitude and, for efficiency, pre-process the continuous data with spectral whitening and one-bit normalization to improve the signal-to-noise ratio (*Larosse et al.* [2004]). We analyze two frequency bands [0.05-0.1Hz] and [0.1-0.3Hz] corresponding to primary and secondary microseismic peaks, respectively.

2.1. Beamforming analysis

Our time-shift beamforming analysis consists of decomposing the body-wave part of a wavefield recorded by a network into plane waves. If a plane wave defined by its slowness vector $\vec{S}$ reaches two stations A and B, the cross-correlation of signals recorded at these stations will be shifted by:

$$\Delta T_{AB}(\vec{S}) = \vec{S} \cdot \vec{AB} \tag{2}$$

where $\vec{AB}$ is the vector connecting A and B. We approximate the network to be flat by neglecting different station elevations and project the slowness vectors into the horizontal plane considering its South-North and West-East components S_N and S_E. For a given horizontal slowness vector $\vec{S} = (S_E, S_N)$, we time-shift all inter-station cross correlations following (2) and stack them to define the function C_{stack} :

$$C_{\text{stack}}(\vec{S}, t) \quad = \quad iFFT \left(\sum_{P \in PS} e^{2i\pi\omega \ \Delta T_P(\vec{S})} \, C_P(\omega) \right) \tag{3}$$

where PS represents the ensemble of pairs of stations, C_P is the Fourier Transform of the noise correlation for pair P and $iFFT$ is the inverse Fourier transform. The characterization of the signal amplitude in the horizontal slowness domain is finally defined as :

$$A(\vec{S}) = \int_{-T}^{T} \Gamma \left(C_{\text{stack}}(\vec{S}, t) \right) \, dt \tag{4}$$

where $\Gamma[f(t)]$ returns the envelope of the function $f(t)$ using Hilbert transform. Integration limits $[-T, T]$ are used to select the part of cross-correlations centered at targeted slowness and we set T to be equal 15s and 10s for the primary and the secondary microseismic bands, respectively.

Figure 6a shows results of the beamforming analysis applied to one-month cross-correlations of the noise recorded by the ETSE network during August 2000 and filtered around the secondary microseismic peak (0.1-0.3 Hz). Energy distribution on the horizontal slowness plane is clearly not random and homogeneous. Two clear patches indicate that during this month most of body-wave energy recorded by ETSE stations is arriving with rather fast apparent velocities (> 20 km/s) and is coming from two preferential directions south and south-east of the network. A similar analysis made during February 2001 also shows two very localized patches of body wave energy with fast apparent velocities (Figure 6b). However, during this winter month the waves are coming from the north. These observation are in good agreement with seasonal variations of the location of sources of microseisms deduced from the polarization analysis and from previous studies (e.g. *Stehly et al.* [2006]).

We then analyzed seismic noise recorded by a network operated in 2000-2001 in North America, i.e., by the Yellowstone PASSCAL experiment (*Fee and Dueker* [2004]). Results of the beamforming analysis for August 2000 and February 2001 shown in Figures 6c and 6d, respectively, are similar to observations made with the ETSE network. Localized noise sources are seen south of the network during the Northern hemisphere summer and north of the network during the winter.

2.2. Locating P-wave noise sources on the Earth's surface.

In a next step, we project the results of the beamforming analysis on the Earth's surface. Following the results of the polarization analysis of the near-zero-time arrivals at cross-correlations and their fast apparent velocities, we assume that they are mostly composed of teleseismic P-waves. For a given slowness and back-azimuth we can back-project a seismic wave using a ray tracing in a spherically symmetric Earth's model. We suppose that diffracted and reflected phases are less energetic than direct and refracted waves and, therefore, the waves that we take into account are P, PP, PKP, PKiKP and PKIKP waves and we use the IASPEI91 tables (*Kennett and Engdahl* [1991]) to relate the slowness with the source-receiver distance for all considered phases (Figure 7).

For every point on the Earth's surface, we identify all phases that may propagate from this point to the network location and determine their horizontal slowness. Therefore, for each position we determine the "energy" of all the waves that are considered from the horizontal slowness plane. The energy is evaluated from the function A determined by the beamforming analysis (equation 4). Finally, we select the phase corresponding to the maximum value of the beamforming map and attribute this amplitude to the projection at the considered geographical position. By repeating this procedure for all points on a $5°$ longitude $\times\, 2.5°$ latitude geographical grid we construct a map of what can be considered as probability density of noise sources during the considered period.

This process is illustrated for three geographical locations shown with stars in Figure 8b and ETSE Network (triangle in Figure 8b). White lines are the great circles and their projections in the horizontal slowness plane in Figure 8b and 8a. For the location 1, the possible seismic phases are PP, PKP (branches **ab** and **bc**) and PKiKP (red line in 8a) and the branch **bc** of the phase PKP corresponds to the strongest amplitude that is then selected for the projections. Similarly, for locations 2 and 3, possible phases are PKiKP and P or PP, respectively, and the latter correspond to stronger amplitudes and are used for the projection. The larger patch is projected as a P wave into the Indian ocean and as a PP wave into the southern Pacific. The smaller patch corresponds to a PKP wave originated in the vicinity of New Zealand.

Maps of P-wave noise source densities corresponding to beamfoming results from Figure 6 are shown in Figure 9. Some source areas such as the region south of Africa during August 2000 and Northern Atlantic south of Iceland during February 2001 are well illuminated by both networks. This suggests that strongest sources of P-waves microseisms are seen by multiple networks distributed around the world. Therefore, we decided to combine observations from different networks to improve the accuracy of the location of main sources of P-wave microseisms. We used stations from three networks shown in Figure 1: 46 stations from Yellowstone park, 29 stations from ETSE, and 14 stations of the Kirgyz Seismic Network (KNET). We interpret the projection map obtained from every individual networks as a probability density and just multiply them to find the combined distribution.

Location of P-wave sources of the primary and secondary microseisms during different seasons are shown in Figures 10 and 11. For every map we used correlations of one month of data. In the secondary microseismic band, regions that generate P-waves are well defined (Figure 10) and are mostly in deep oceans. Also, a clear seasonal migration can be observed with strongest sources located in the northern hemisphere during the northern hemisphere winter and in the southern hemisphere during the summer. In the primary microseismic band (Figure 11) we also observe a similar seasonal variation. However, uncertainties of the source location are much larger than in the secondary microseismic band. The reason for this is that the signal-noise ratio of the near-zero-time arrival in cross-correlations is much lower in the primary microseismic band than in the secondary microseismic band. Also, reliability of the array analysis degrades at longer wavelengths. Despite these large uncertainties, it is clearly seen in Figures 10 and 11 that, in most cases, sources of primary microseisms do not coincide geographically with sources of secondary microseisms pointing to different physical mechanism of generation of these two microseismic peaks.

The maps of Figure 11 are similar with the results of *Stehly et al.* [2006] in terms of seasonnality. The location of noise sources based on the backpropagation of Rayleigh waves can not provide a resolution comparable to the one we achieved with body waves. Nevertheless, even with P waves, the uncertainties of location of sources of the primary microseism do not allow to clearly conclude unambiguously that they are located in deep parts of the ocean, in near-coastal regions, or both. While strongest identified source areas tend to extend to deep oceanic parts, they also cover coastal areas.

3. Discussion and conclusion

Results of our polarization and beamforming analyses of the continuous noise records demonstrate that significant part of the microseismic noise is composed of P-waves generated by distant sources. These sources show a clear seasonality in correlation with the seasonal migration of the strong oceanic storms between the southern and the northern hemisphere suggesting that the observed teleseismic P-waves are generated by the interaction of waves produced by these storms with the seafloor. Location of P-wave sources in the primary and in the secondary microseismic bands do not coincide with each other indicating that these two peaks are generated in different regions and possibly by different physical process. P-waves are more easily identified in the secondary microseismic band than in the primary microseismic band. While we cannot exclude that this difference is related to the more efficient mechanism generating secondary microseismic P-waves than primary ones, a simple explanation of this observation can be related to difference in wave propagation. Strong noise sources generate both body wave and surface waves. For the latter, their attenuation is much stronger at higher frequencies. Therefore, surface waves in the secondary microseismic peak propagate much less efficiently over very long distances than in the low-frequency primary microseismic band. As a consequence, for distant noise sources, the relative part of the body waves in the recorded seismic noise is relatively high in the secondary microseismic band while the primary microseismic band remains largely dominated by surface waves, making observation of P-waves more difficult.

Using array-based processing of the teleseismic P-waves to locate regions generating strong microseisms has significant advantages relative to using surface waves (e.g., *Stehly et al.* [2006]). The latter yields only a determination of their back-azimuths at the array location while with body waves we can measure both backazimuths and slownesses that can be converted into distances. Therefore, we can locate the source regions more accurately with body waves than with surface waves.

We can compare our maps of the source density in the secondary microseismic band (Figure 10) with results by *Gerstoft et al.* [2008] who applied beamforming to the noise records of the Southern California seismic network. We find similar source locations in southern Pacific and Indian oceans during summer months and in northern Pacific and Atlantic ocean during winter months. Using three networks simultaneously allows us to image source regions with higher reliability. One of most important consequences of this improved reliability is that we clearly see that strongest sources of P-wave microseisms are located in deep oceans far from coasts for the secondary microseism. Also, the observed source regions are significantly smaller than areas affected by significant wave heights. Overall our observations are consistent with the generation of microseisms by non-linear interaction of ocean waves propagating in opposite directions that create a pressure distribution on the seafloor at twice the frequency of the interfering waves (*Longuet-Higgins* [1950]). Following *Kedar et al.* [2008], this wave-wave interaction occurs in deep oceans and the geographical intensity of this interaction may be computed from oceanic wave action models. Moreover, the efficiency of the coupling between the interfering oceanic waves and the seafloor may depend on the depth of the water column, i.e., on the bathymetry. As a result, an efficient transfer of energy from oceanic to seismic waves occurs over geographically very limited and specific areas. It is in particular interesting to note that the source area near Iceland seen in Figure 10 during October and January coincides with the strong source of Rayleigh wave microseisms computed by *Kedar et al.* [2008] based on Longuet-Higgins's theory and oceanic wave action models.

These observations confirm that the source of secondary microseisms are not confined in the coastal areas as it is often accepted by seismologists. On average, the excitation of P waves by oceanic waves is stronger in the deep oceans. It does not mean, however, that there is no excitation along the coast, particularly when storms hit the shoreline.

In the future, locating sources of P-wave microseisms can be improved with using more networks better distributed over the globe. Observing and understanding these sources is important to validate predictive models of the seismic noise generation and distribution. These models, in turn, may help us to improve the accuracy of noise based seismic imaging and monitoring.

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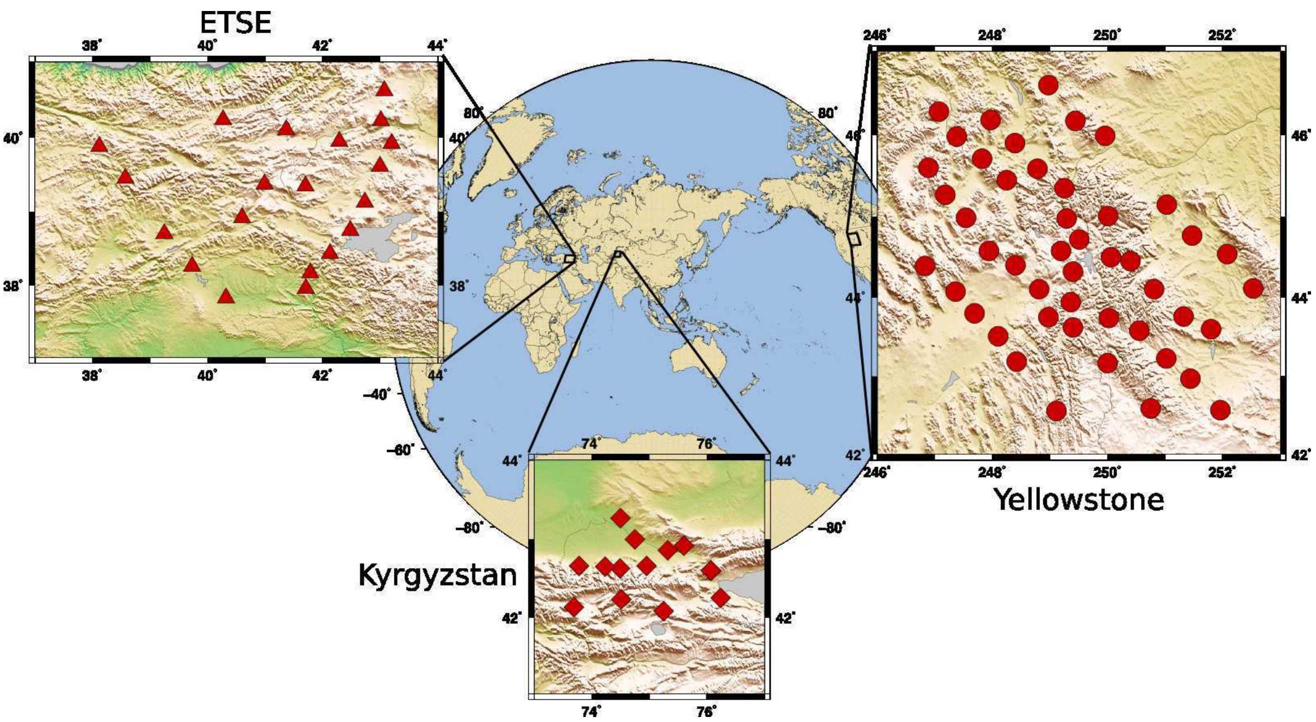

Figure 1. Maps of ETSE, Yellowstone and Kyrgyzstan networks

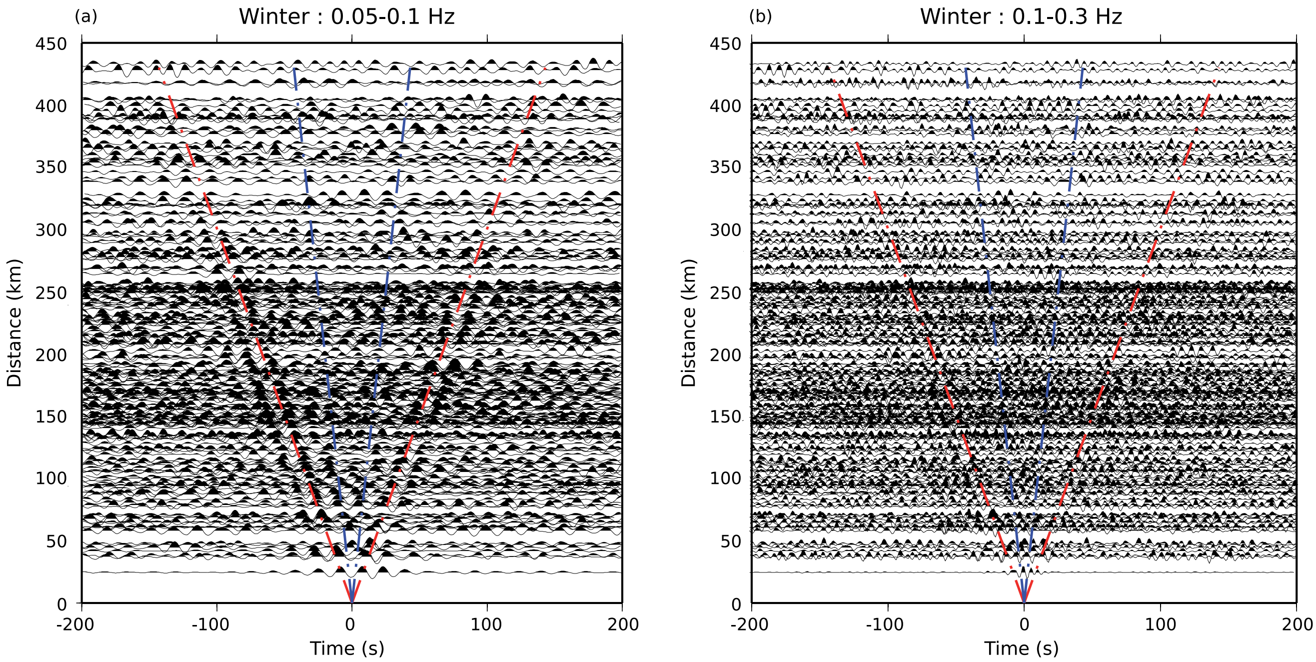

Figure 2. Cross-correlations of vertical component records between stations of the ETSE network stacked for months between September and March (winter) for years 1999, 2000 and 2001 plotted as function of the distance between the pair of stations in the 0.05-0.1 Hz (a) and 0.1-0.3 Hz (b) frequency band.

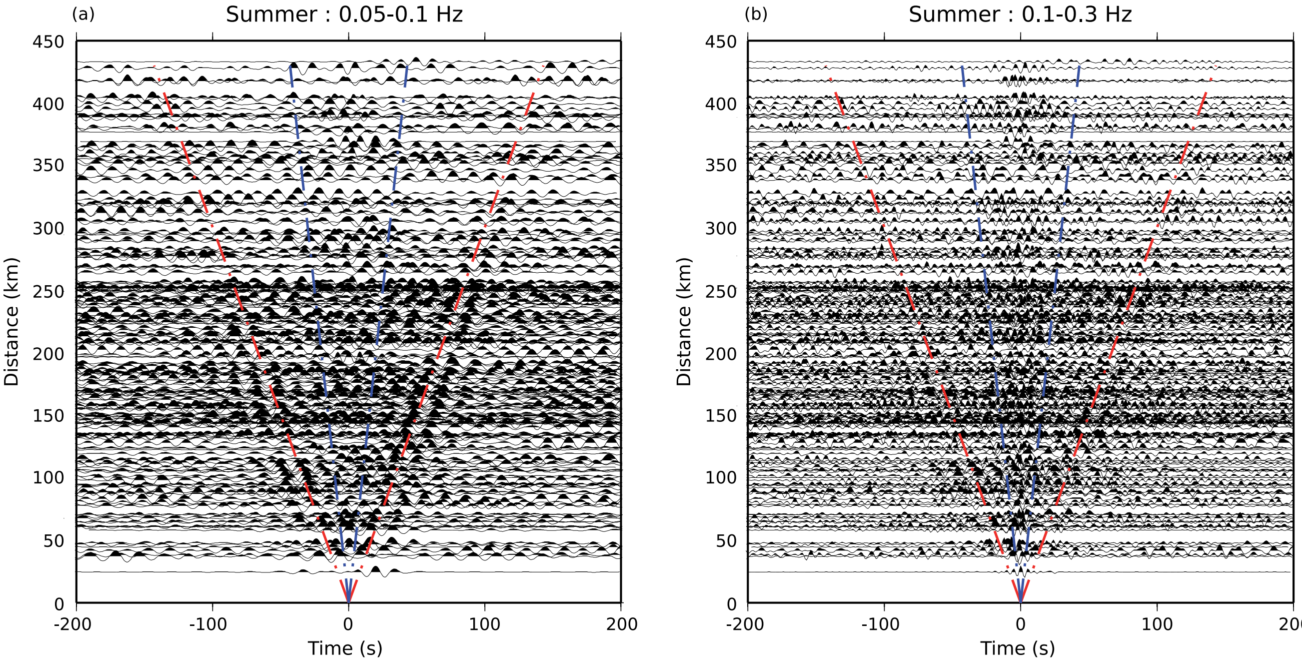

Figure 3. Cross-correlations of vertical component records between stations of the ETSE network stacked for months between April and August (summer) for years 1999, 2000 and 2001 plotted as function of the distance between the pair of stations in the 0.05-0.1 Hz (a) and 0.1-0.3 Hz (b) frequency band.

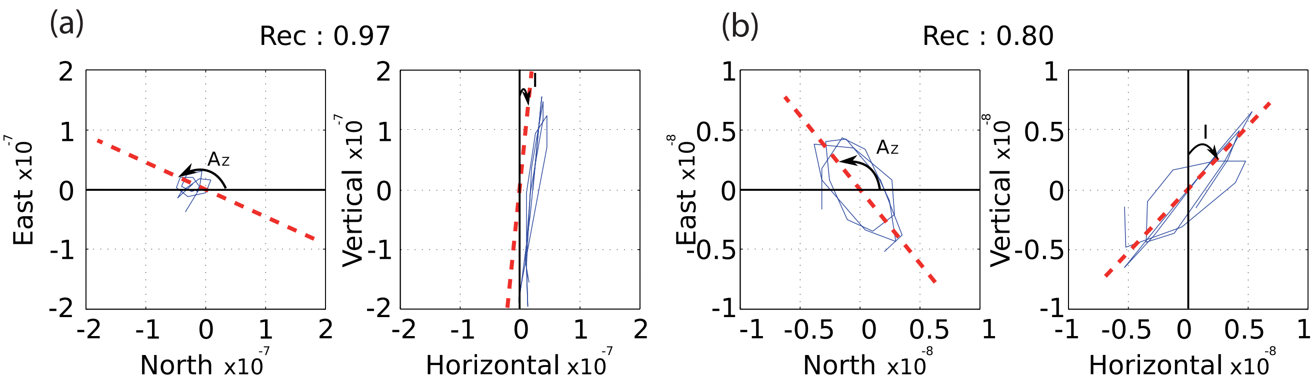

Figure 4. Particle motion for 2 station pairs from noise correlations in the $[-10s; +10s]$ time window for the band 0.1-0.2 Hz. (a) : cross-correlations between KARS and KOTK (distance 391 km) for the 10/30/1999 ; (b) : cross-correlations between KARS and ILIC (distance 404 km) for the 03/23/2000. *Rec* is the rectilinearity coefficient defined by Eq . 1. Red dashed lines show the incidence angle and the azimuth measured from the covariance method.

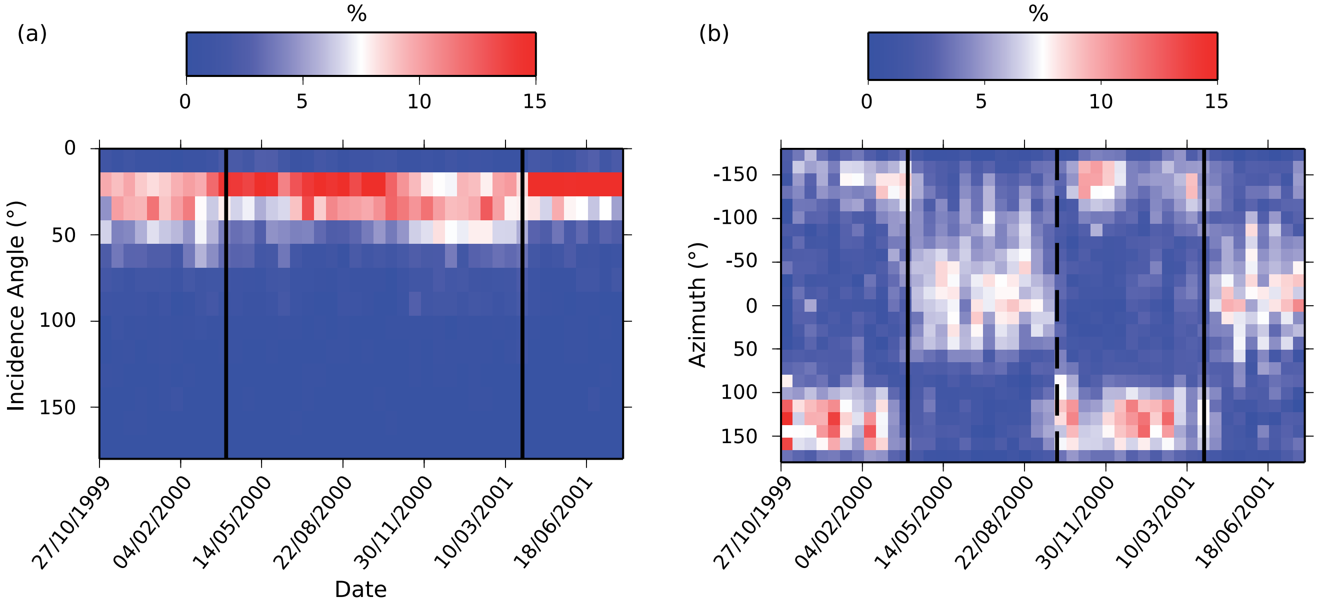

Figure 5. Probability of occurrence of a given value of the incidence angle (a) and of the azimuth (b) for 20 days time periods for all station pairs of the ETSE network. Continuous black lines correspond to 04/01/2000 and 04/01/2001 and black dashed lines to 10/01/2000

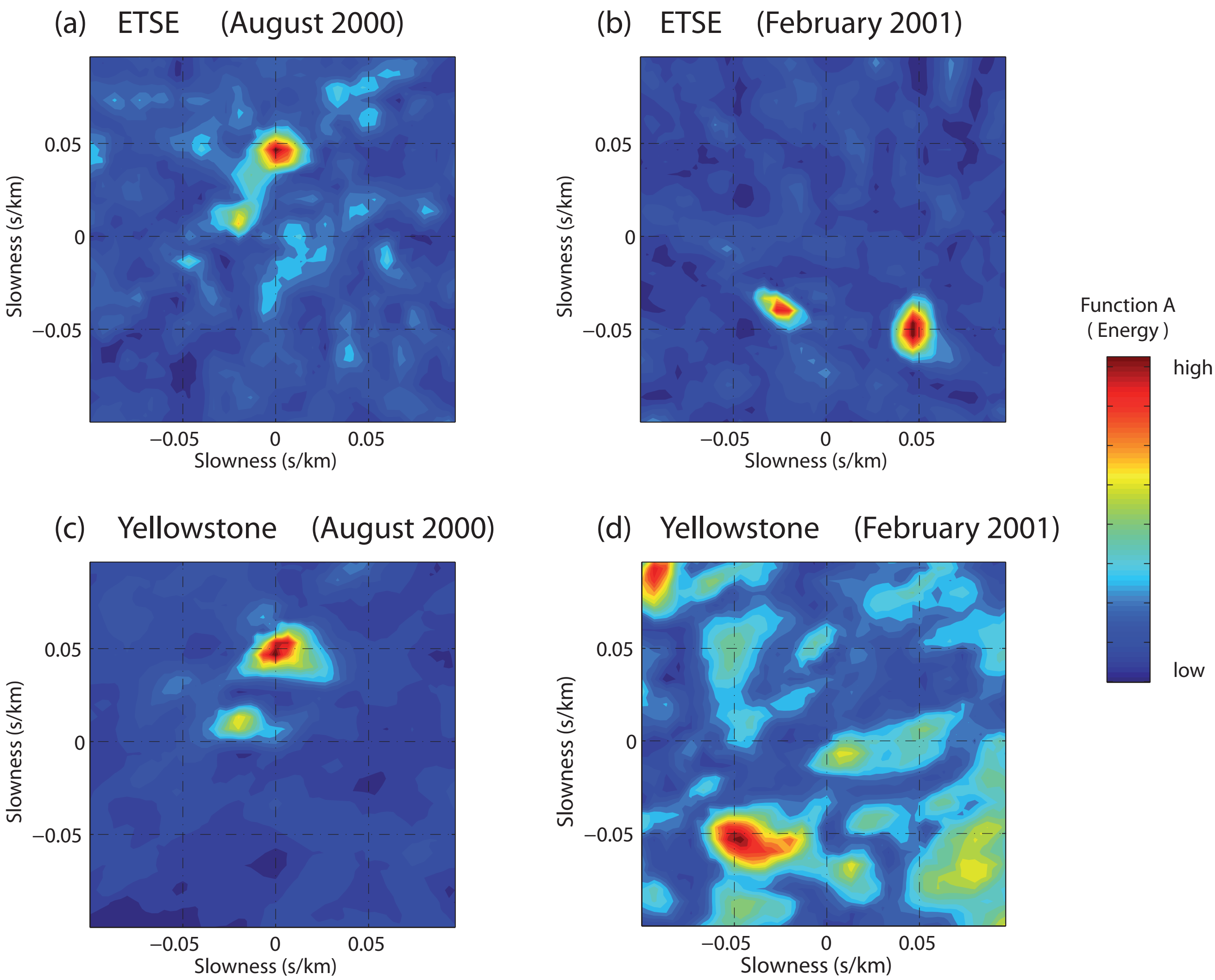

Figure 6. Beamforming result of the Yellowstone and ETSE networks for August 2000 and February 2001 around the secondary microseismic peak (0.1-0.3 Hz) . Axcis are in s/km.

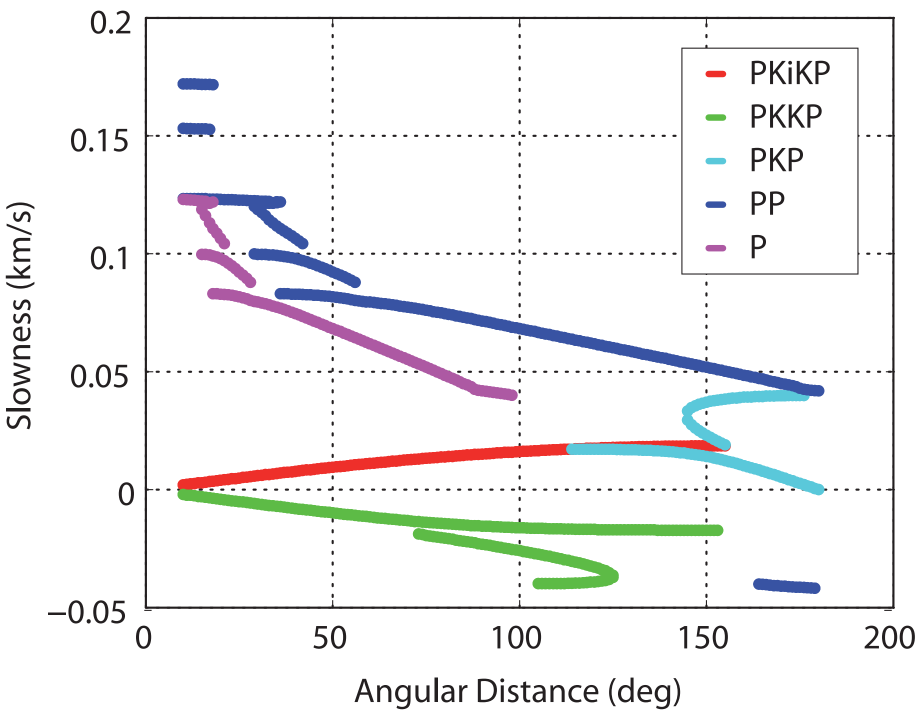

Figure 7. Variations of slowness (s/km) with respect to the angular distance (deg) for P, PP, PKP, PKiKP and PKIKP phases.

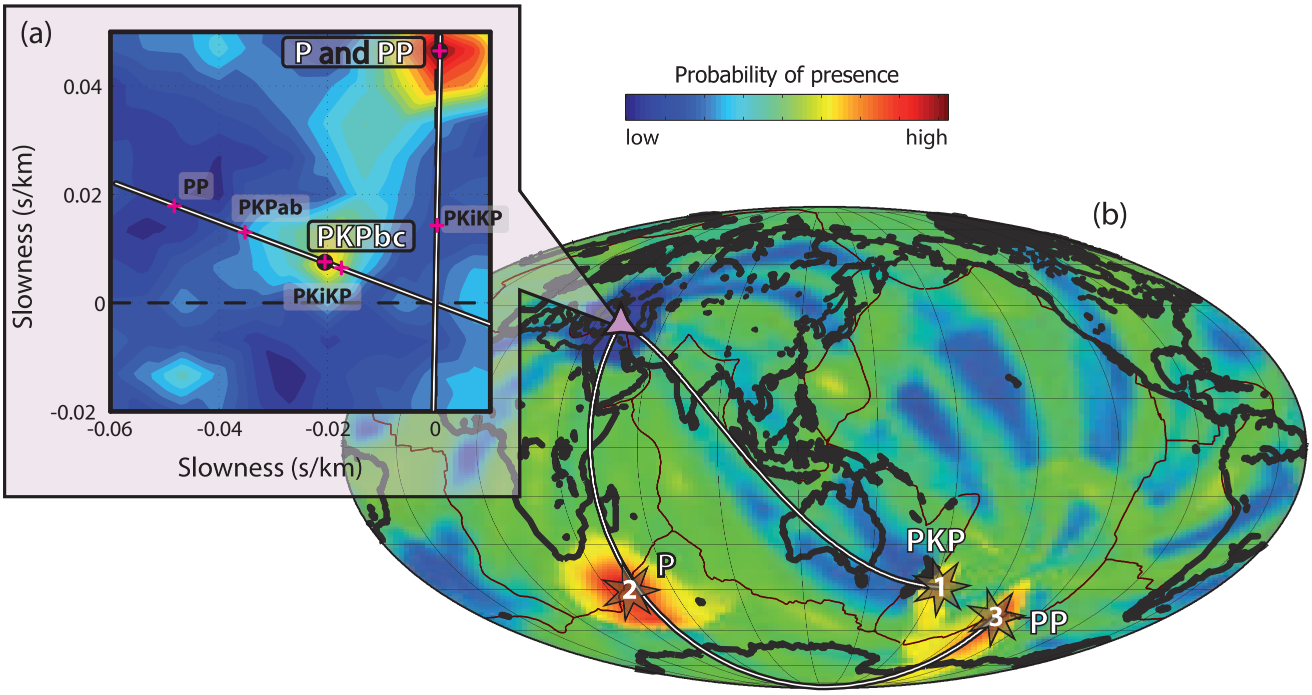

Figure 8. Illustration of projection of the beamforming results on the Earth's surface (see text for explanations). (a) Results of the beamforiming analysis of noise cross-correlations computed during August 2000 for the 0.1-0.3 Hz frequency band between stations of the Turkey network plotted as function of horizontal slowness. (b) Projection of these results on the Earth surface.

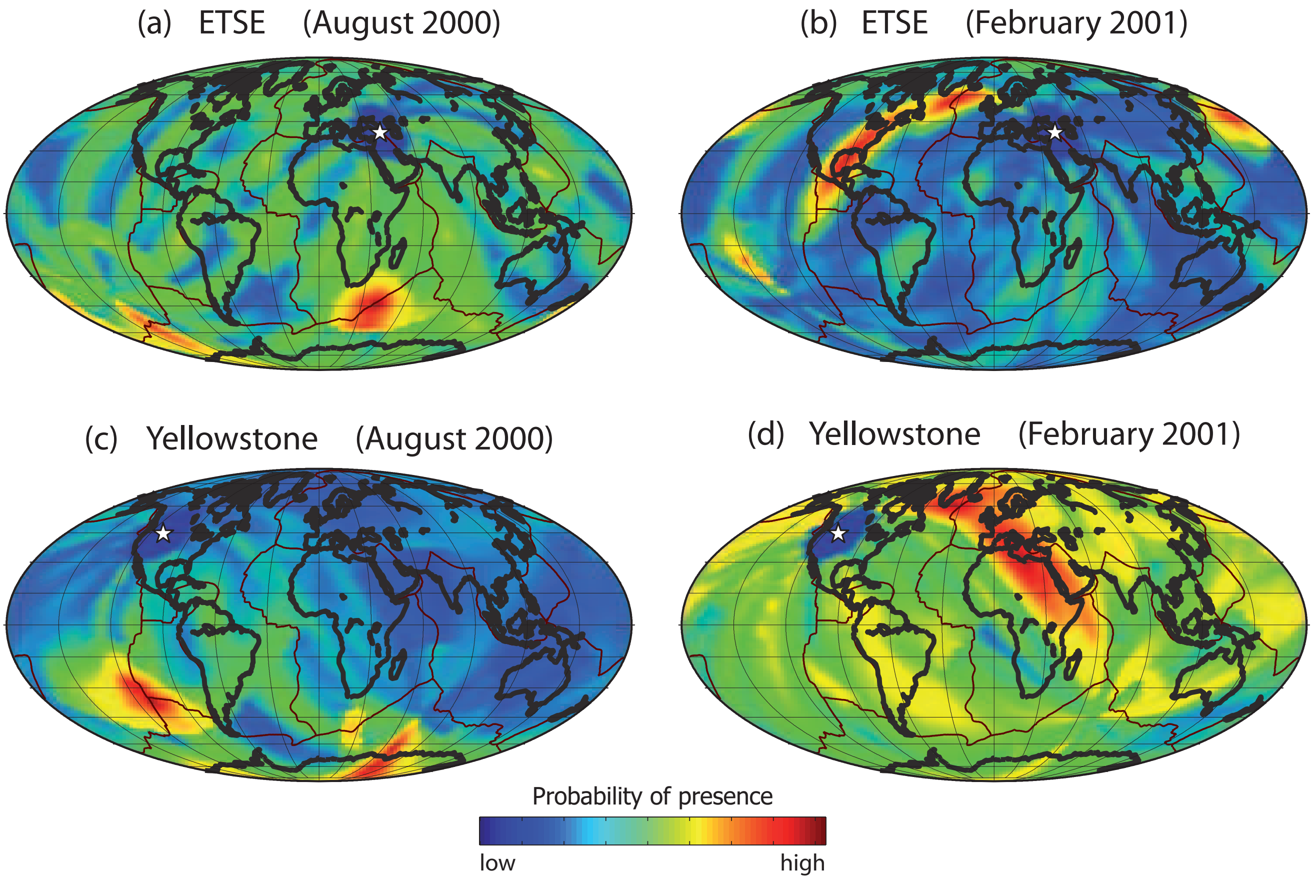

Figure 9. Projection results of the Yellowstone and ETSE networks for August 2000 and February 2001 around the secondary microseismic peak (0.1-0.3 Hz).

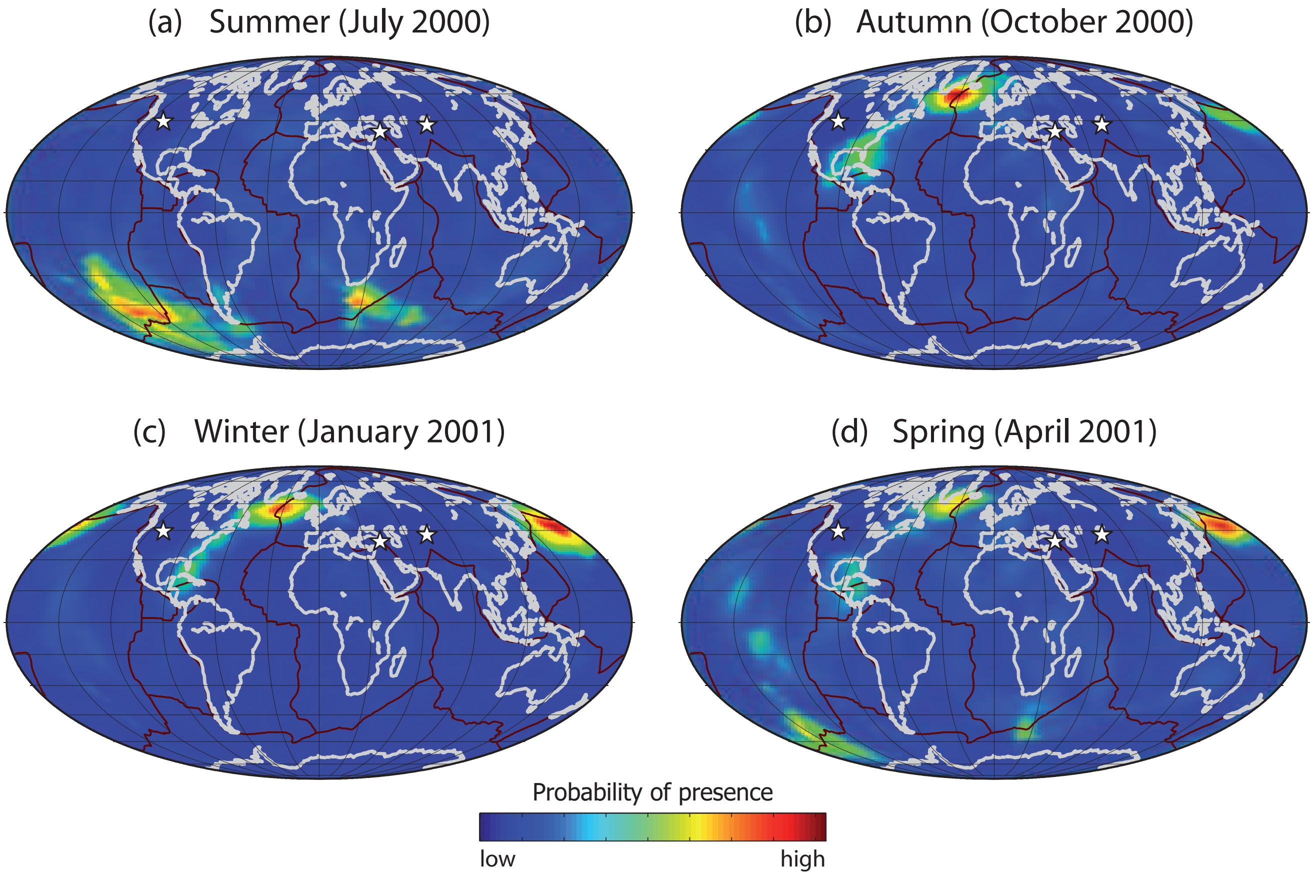

Figure 10. Seasonal variation of the location of P-wave seismic noise sources in the secondary microseismic band (0.1-0.3 Hz).

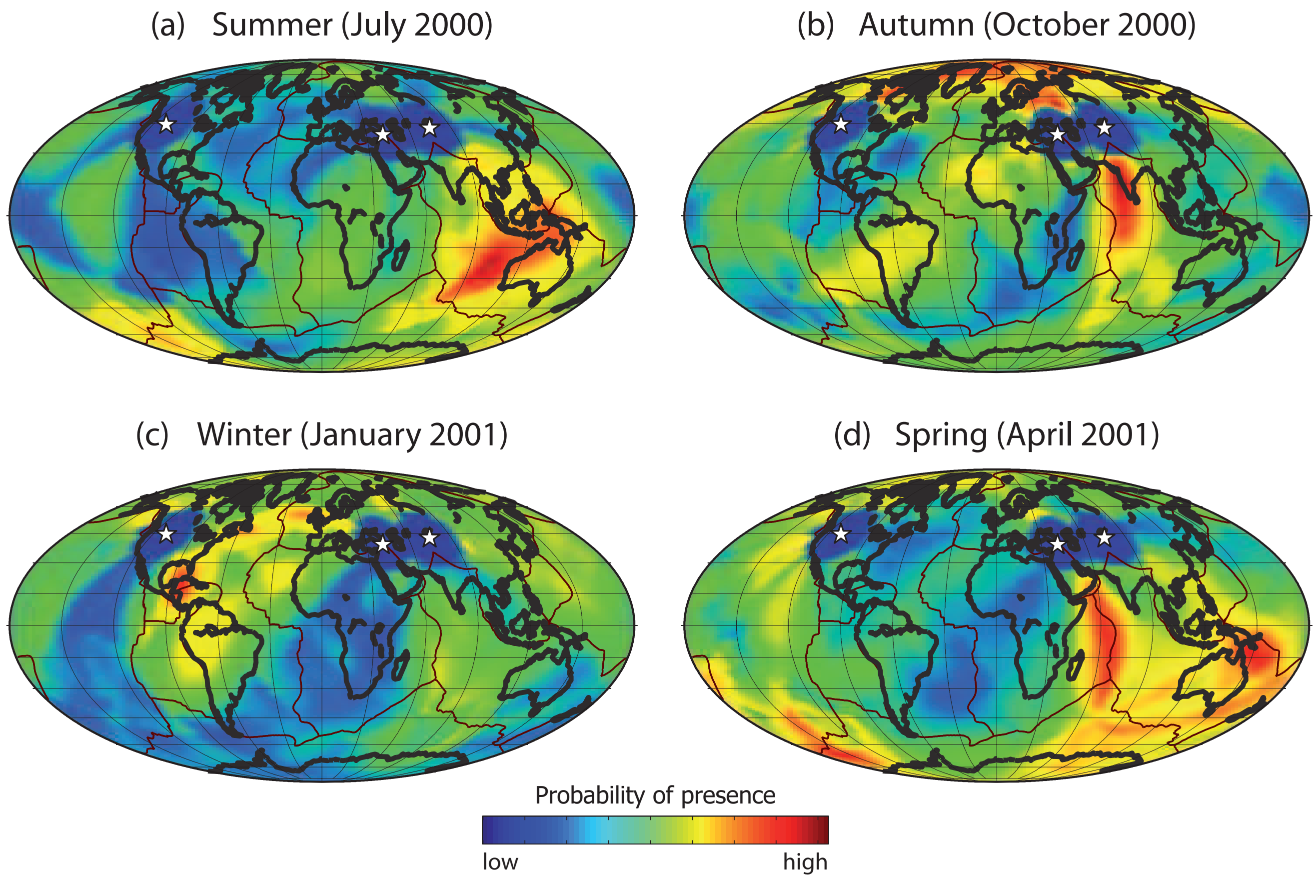

Figure 11. Seasonal variation of the localization of seismic noise in the primary microseismic bands (0.05-0.1 Hz).

Appendix A: Wave polarization analysis from cross-correlations

We consider a plane wave with amplitude A, wave vector $\vec{k}$ and pulse shape $v(t)$. The azimuth (Az) is defined as the angle between the North and the projection of the wave vector in the horizontal plane. In the vertical plane, the incidence angle (I) is the angle between the vertical and the wave vector. Equation A1 gives the general form of a 3-component record of the signal that will be used for this demonstration :

$$\begin{aligned}
S_N(t) &= A\sin(I)\cos(Az)v(t-T), \\
S_E(t) &= A\sin(I)\sin(Az)v(t-T), \\
S_Z(t) &= A\cos(I)v(t-T),
\end{aligned} \tag{A1}$$

where T is the arrival time of the signal at the station.

This plane wave is recorded at two different stations A and B at times $T = t_A$ and $T = t_B$. We compute three cross-correlations between the vertical component of station A and the 3 components of station B. They are :

$$\begin{aligned}
ZN(t) &= A^2\cos(I)\sin(I)\cos(Az)V(t), \\
ZE(t) &= A^2\cos(I)\sin(I)\sin(Az)V(t), \\
ZZ(t) &= A^2\cos^2(I)V(t),
\end{aligned} \tag{A2}$$

where :

$$V(t) = \int_{-\infty}^{+\infty} v(\tau)\overline{v}[\tau-(t_B-t_A)-t]d\tau \tag{A3}$$

is the time correlation function and $\overline{v}(t)$ the complex conjugate of $v(t)$.

The definition of the covariance between two signals (E and F) in the time window $[T_1, T_2]$ is given by :

$$Cov_{EF} = \int_{T_1}^{T_2} E(t)F(t)dt. \tag{A4}$$

For North and East component records of station A, we select a time window including the signal ($t_A \in [T_1, T_2]$) and we compute the covariance defined by equation A4 to obtain :

$$\begin{aligned} Cov_{NE} &= \int_{T_1}^{T_2} A^2 \sin^2(I)\cos(Az)\sin(Az)v(t-t_A)v(t-t_A)dt, \\ Cov_{NE} &= A^2 \sin^2(I)\cos(Az)\sin(Az)C, \end{aligned} \tag{A5}$$

where $C = \int_{T_1}^{T_2} v(t-t_A)v(t-t_A)dt$ is a non-null constant.

We apply the same computation to all pairs of records at station A and we obtain the covariance matrix of the 3-component record :

$$S_{CovS_A} = C \cdot M, \tag{A6}$$

where :

$$M = \begin{pmatrix} \cos^2(I) & \cos(I)\sin(I)\cos(Az) & \cos(I)\sin(I)\sin(Az) \\ \cos(I)\sin(I)\cos(Az) & \sin^2(I)\cos^2(Az) & \sin^2(I)\cos(Az)\sin(Az) \\ \cos(I)\sin(I)\sin(Az) & \sin^2(I)\cos(Az)\sin(Az) & \sin^2(I)\sin^2(Az) \end{pmatrix}. \tag{A7}$$

We consider the definition of the covariance (Equation A4) and the definition of the cross-correlations between 2 stations (Equation A2) to compute the covariance between ZE and ZN cross-correlations. We select a time window $[T_1; T_2]$ of the correlation which includes the correlation of the initial signal :

$$\begin{aligned} Cov_{Z_AE_B-Z_AN_B} &= \int_{T_1}^{T_2} A^4\cos^2(I)\sin^2(I)\cos(Az)\sin(Az)V(t)V(t)dt, \\ Cov_{Z_AE_B-Z_AN_B} &= A^4\cos^2(I)\sin^2(I)\cos(Az)\sin(Az)C_{Corr}, \end{aligned} \tag{A8}$$

where $C_{Corr} = \int_{T_1}^{T_2} V(t)V(t)dt$ is a non-null constant.

The covariance matrix for cross-correlations is computed from Equations A2 and A4 :

$$S_{CovCorr_{A.B}} = A^4\cos^2(I)C_{Corr}\cdot M. \tag{A9}$$

$$S_{CovCorr} = A^2\cos^2(I)\frac{C_{Corr}}{C}S_{CovS_A}. \tag{A10}$$

where $A^2\cos^2(I)\frac{C_{Corr}}{C}$ is a non-null constant. From Equation A10 we conclude that, in the case of a plane P-wave, the covariance matrix for the 3-component record at a single station and the covariance matrix for the cross-correlations between 2 stations (S_{CovS_A} and $S_{CovCorr_{A.B}}$, respectively) differ only by a scalar factor. Therefore, the eigenvectors of those matrix are the same which prove the polarization analysis can be performed either on cross-correlation or on 3 component records.