Retrieving PmP Travel Times From a Persistent Localized Microseismic Source

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Abstract Ocean swells or storms can generate persistent localized microseismic sources in deep oceans, which emanate body wave energy traveling through the Earth's interior, carrying abundant information about subsurface structures as earthquake waves do. However, body waves from localized microseismic sources have not been fully exploited to map the internal structures of the Earth. Here, we report that the travel times of body wave reflections from the Moho can be obtained from body waves generated by localized microseismic sources. We further demonstrate that body waves from these sources can be utilized to constrain the Moho morphology through synthetic experiments and comparisons with earthquake receiver function results. Our work suggests that localized microseismic sources can be exploited to map the internal structure of the Earth without relying on earthquake data.

Plain Language Summary Movements of ocean waves can trigger the vibration of the solid earth, generating seismic waves which can propagate downward into the deep interior of the Earth and be recorded by a distant seismograph. Such seismic waves are called body waves, which carry information about the interior structures of the Earth. In the past, body waves generated by oceanic swells or storms have not been exploited to map the internal structures of the Earth. In this study, we report that body waves generated by sources related to oceanic swells or storms can be utilized to map the boundary between the crust and the upper mantle of the Earth. Our method may be also applicable to ground motion data triggered by large storms in Mars to image Mars's interior.

1. Introduction

Seismic ambient noise is ubiquitous background motion generated by various sources and has been observed since the early 20th century (Gutenberg, 1947). It is believed that high frequency ambient noise (>1 Hz) is generated by human activities, however, ambient noise at periods longer than 1 s is related to oceanic swells and atmospheric storms. Especially, ambient noise at periods of around 5–20 s, the so-called microseisms, is the most energetic noise, which is believed to be generated by nonlinear interaction of oceanic waves with solid earth in oceans.

Surface waves at microseisms are typically much stronger than body waves as most of sources are located on the surface of the Earth. Surface waves from microseisms can propagate inland thousands of kilometers from coastlines. An interferometry method has been developed to retrieve empirical Green's functions (EGFs) of surface waves from microseisms (e.g., Shapiro, et al., 2005). Such method leads to the development of ambient noise tomography capable of imaging crust and upper mantle structures without relying on earthquake data (e.g., Lin et al., 2008; Luo et al., 2013; Moschetti et al., 2010; Wang et al., 2019; Yang et al., 2007; Yao et al., 2008).

Similar efforts have been devoted to extracting body waves from ambient noise, especially the reflection body waves from various interfaces within the Earth, which can be utilized to constrain structural interfaces that cannot be constrained by surface waves (e.g., Feng et al., 2017; Zhan et al., 2010). However, due to the fact that the energy of body waves in microseisms is much weaker than that of surface waves and sources of ambient noise are mostly located on the surface of the Earth, retrievals of body waves from ambient noise have been less successful and are only limited to some specific areas (e.g., Poli, Campillo, et al., 2012; Poli,
Pedersen, & Campillo, 2012; Zhan et al., 2010; Zhang et al., 2020). Such obstacles limit the applications of body waves from ambient noise in imaging the internal structures of the Earth.

To circumvent the limitation of weak energy of body waves from ambient noise, various studies have adopted spatial stacking methods to improve the retrieval of reflected body wave signals, such as those from transition zones (e.g., Feng et al., 2017; Poli, Campillo, et al., 2012) and the retrieval of body wave propagation on the global scale (e.g., Boué et al., 2013; Lin et al., 2013; Nishida, 2013; Poli et al., 2015). Recent works however have indicated that the body wave features in cross-correlation functions (CCFs) on the global scale are mainly attributed to coda waves from large earthquakes (e.g., Lin et al., 2013; Tkalčić et al., 2020), suggesting the body wave features on the global scale are not independent from earthquakes. On the other hand, ocean swells or storms can generate strong and localized microseismic sources (e.g., Fan et al., 2018; Fan et al., 2019), exciting P wave (e.g., Pyle et al., 2015; Lin et al., 2017; Retailleau et al., 2018) and S wave energy (e.g., Nishida & Takagi, 2016; Liu et al., 2020). These localized microseismic sources can be utilized to obtain body wave features through cross-correlations (Euler et al., 2014; Gerstoft et al., 2008; Liu et al., 2016; Wang et al., 2018).

In this study, we report a discovery that persistent localized microseismic sources can be utilized to obtain travel times of Moho reflected body waves. Here, rather than aiming to directly recover EGFs of body waves between station pairs as most previous works do, we use a Common Receiver Station stacking method to obtain PmP travel times at individual stations. Our method overcomes the stringent requirement of noise sources having to be located at a narrow stationary zone for EGF retrieval, allowing this method to be applicable to a seismic array recording body waves from strong localized microseismic sources. Our method opens the door for better exploiting localized microseismic sources generated by ocean swells or storms in imaging the internal structures of the Earth.

2. Data and Methods

We use ambient noise data recorded at an L-shaped array in the Shandong Peninsula which contains 47 three-component seismographs and recorded three-component continuous waveform data for about 1 year from November 2000 to October 2001 (Figure 1a).

The procedures we use to calculate CCFs between the L-shaped stations in our study region are similar to those detailed in Bensen et al. (2007). First, all continuous waveform data recorded in June 2001 are decimated to 10 Hz sampling rate and cut into 3-hour segments. Then, instrument responses, mean and trend are all removed from the waveforms of all the segments. And, the 3-hour segments are bandpass filtered between 0.05 and 2 Hz and whitened in frequency domain. Furthermore, each segment is normalized by a running-absolute-mean method to suppress effects from earthquakes or other irregularities (Bensen et al., 2007). After the above processing, we calculate 3-hour CCFs between all stations over the 1-month period of June 2001 and stack all the 3-hour CCFs over a month to obtain 1-month CCFs among all the stations of the L-shaped array.

Since most energetic P wave signals mainly appear in the period band of 3–10 s, we bandpass filter all the CCFs at a frequency band of 0.1–0.33 Hz. Waveforms of 1-month CCFs of June 2001 between station 003 and other stations are plotted in Figure 1d, which clearly show surface waves with a move-out velocity of ∼3 km/s, as indicated by the two black dashed lines in Figure 1d. In addition to the surface wave arrivals, we also observe the near zero-time arrivals. Such near zero-time arrivals have been widely observed worldwide in CCFs before (e.g., Landès et al., 2010) and are attributed to body wave sources generated by ocean swells or storms in deep oceans.

3. Noise Source Analysis

In this study, as we intend to use localized microseismic sources to obtain the Moho reflected P wave (PmP) travel times, we first analyze the temporal and spatial distribution of ambient noise data recorded at our array to identify microseismic sources by adopting a beamforming method.
Figure 1. (a) Map of topography and station distribution in the Shandong Peninsula. The inset shows the location of our study area (the red triangle) and the location of the localized microseismic source (the red star) we use in this study. (b) Projection process of station distances in the back azimuth relative to the noise source. The location of station 003 is denoted with a blue triangle. (c) Body wave signals from focused continue source in the time window outlined by the red dashed box for the data recorded in June 2001. (d) Cross-correlation functions (CCFs) for station pairs between station 003 and all other stations in June 2001 plotted as a function of interstation distance. (e) CCFs with interstation distances longer than 40 km for station pairs between station 003 and all other stations in June 2001. They are arranged by the projection of the station-pair distance to the back azimuth of the noise source.
A seismic wave propagating from a distant location can be approximated as a plane wave when it is incident at a small seismic array as the L-shaped array in this study. Different phases from the incoming plane waves can produce signals with different time delays in CCF between a pair of stations. The delay of these signals can be calculated and signals in cross-correlation records can be stacked using the azimuth and slowness of an incoming wave as follows:

\[ P(t) = \frac{1}{N} \sum_{i=1}^{N} C_i (t - t_{\text{diff},i}). \]  

with \( C_i \) representing the \( i \)th CCF, \( t_{\text{diff},i} \) representing the differential travel time of a phase between a pair of stations which can be computed using the horizontal slowness \( u \), and \( N \) representing the number of CCFs used in the stacking.

If the true time shifts \( t_{\text{diff},i} \) are known, the signals from the same phase among different station pairs are coherent, that is, after correcting the travel time differences for different interstation distances in CCFs, the coherent signals have the same arrival times, leading to a maximum beam power \( P \) (Rost & Thomas, 2002). An incorrect selection of azimuthal direction and slowness of incoming waves lead to an incoherent summation, resulting in a degradation of the beam power \( P \).

In order to analyze the source of the body wave signals, we cut out the body wave features in the time window of \(-18–18 \) s and employ the body wave features to perform the beamforming analysis. Only CCFs with distances larger than 40 km are used in the beamforming analysis to avoid the inferences of surface wave signals on the body wave arrivals. Since some stations are located in mountainous regions, the topographic effects on the travel times of direct P waves need to be corrected. In this procedure, the travel time differences of P wave due to the variations of elevations relative to a common datum level are first calculated for all stations using a P wave velocity of 6.1 km/s for the upper crust referring to CRUST1.0 model (Laske et al., 2013). This calculation is performed for each azimuthal angle and slowness in the beamforming. Then, each CCF between a pair of stations is time shifted according to the travel time differences of topographic effects between the two stations when performing the beamforming. The results of the beamforming analysis of each month from January 2001 to August 2001 are shown in Figure S1 in Supporting Information S1. Based on the analysis, it is found that there was a single energetic ambient noise source in June and July 2001; while, in other months from January 2001 to August 2001, there were more than one focused source of ambient noise. Although there was a relatively concentrated source distribution in July, since many stations have stopped operating during that month, we do not use the data during that period. This concentrated microseismic source in June 2001 is believed to be located in the South Atlantic as shown in Figure 1a where strong site effect has been identified in the work of Gualtieri et al. (2014) and persistent localized microseismic sources of body waves were identified by Euler et al. (2014).

With the known azimuth of the body wave source, we rearrange all the CCFs according to the projection distance of the station pair in the horizontal direction from the source location to the seismic array (Figure 1b). Figure 1e shows some examples of the rearranged CCFs within time windows of \(-30–10 \) s between station 003 and other stations. Clear signals with an apparent velocity of \( \sim 30 \) km/s, which are the near zero-time arrivals in Figure 1d, appear in the CCFs.

To our knowledge, such body wave energy has not been directly utilized to image the Earth structures. In addition to these strong body wave signals, a weak signal with opposite polarity is faintly visible on the negative time lag (Figure 1e) with the almost same apparent velocity as the near zero-time arrivals. Referring to the CRUST1.0 model (Laske et al., 2013), we calculate the travel time of PmP phases at station 003, which is about 10 s, coincident with the lag time of the weak signals. Therefore, we speculate that these weak signals may be related to PmP phases.

### 4. Origins of the Body Waves in CCFs

To support our interpretation that the origins of the body wave features in the CCFs are related to reflected P waves (PmP) from the Moho, we synthesize cross-correlations from a distant body wave source with the same slowness as we observe from the CCFs. As body wave signals come from distant sources, teleseismic
P wave recorded by a small-scale seismic array can be regarded as a plane wave. As shown in Figure 2a, besides a direct Pp signal of teleseismic body wave, station A and station B also record PpPmP signals that are P waves reflected first on the free surface and then re-reflected by the Moho. After cross-correlating these two records from station A and B, it is expected there are four correlated features in CCF (Figure 2c), namely $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$. The strongest feature is the near zero-time P arrivals with a lag time of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$. This CCF also contains a weak energy of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ with a similar arrival time as $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$. When the media under station A and station B are not uniform, the lag time of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ is slightly different from the lag time of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$. Considering that the energy of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ is much weaker compared to that of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$, the contribution of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ energy is negligible. The other two features besides the near-zero arrivals have a time delay of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ and $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$, respectively.

Here, we intend to obtain travel times of PmP phase, that is equivalent to the arrival time delay of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ between PpPmP phase and Pp phase received by station A. The time delay of $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ can be obtained after correcting the arrival time of the feature $B \Delta t_{PpPmP} t_{PpPmP} E - B \Delta t_{PpPmP} t_{PpPmP} E$ in a CCF.

**Figure 2.** (a) Ray paths of teleseismic P waves incident under stations. Teleseismic P waves incident nearly vertically produce corresponding reflection signals from the Moho discontinuity. (b) Waveforms recorded by stations A and B. Due to the small reflection coefficient, the signal of the reflected wave PpPmP is weak. (c) The cross-correlation function between stations A and B. After the cross-correlation, signals that represent different phases in waveforms appear in the cross-correlation function. (d) The cross-correlation function after time correction. With the time offsetting of $t_{Pp} - t_{Pp}$ corrected to zero, the arrival time of reflected signal $t_{Pp} - t_{Pp}$ in the negative causal part is consistent with PmP beneath station A, which equivalent to the autocorrelation record of station A. (e) Corrected cross-correlation functions (CCFs) for station 003 with other stations, which all contain a reflected signal with the same arrival time as $t_{Pp} - t_{Pp}$. (f) A stacked record of station 003 after stacking all time corrected CCFs in (e).
with a time offset of \( B_{AtPp} t_{Pp} E - B_{AtPp} t_{PpE} \). The time delay of \( B_{BtPp} t_{PpPmP} E - B_{BtPp} t_{PpPmPE} \) can also be obtained by this method. Figures 2c and 2d shows the procedures of correcting this time offset by shifting the waveform toward the positive time side by a time offset of \( B_{AtPp} t_{Pp} E - B_{AtPp} t_{PpE} \). After the shifting, the time of the reflected feature (the feature appearing in the negative time lag of the CCF) beneath station A becomes \( t_{PmP} = t_{PpPmP} - t_{Pp} \), equivalent to the travel time of \( PmP \) beneath station A (Figure 2d). Similarly, the feature in the positive time lag of the CCF is equivalent to the \( PmP \) beneath station B with the arrival time of \( t_{PmP} = t_{PpPmP} - t_{Pp} \). Therefore, the travel times of the Moho reflection \( PmP \) phases at the two stations can be obtained from their CCF after the time offset is corrected. Because reflections in individual CCFs are very weak, to boost the SNR of features related to the \( PmP \) phases, we stack CCFs between station A and other stations after correcting the time shifting of the near-zero arrivals to obtain the stacked CCF. Such a stacked CCF after the time shifting correction is equivalent to an auto-correlation record of station A. In this process, station A is regarded as a receiving station, and we name this stacking method as a Common Receiver Station stacking method (CRS).

In the time shift correction of our CRS, we use the lag time of signal \( B_{AtPp} t_{Pp} E - B_{AtPp} t_{PpE} \) (the strongest near-zero-time signal in Figure 2c). The lag time of this signal is caused by the differences of propagation paths of direct P waves and the differences in crustal structures at different stations. Using the lag time of \( B_{AtPp} t_{Pp} E - B_{AtPp} t_{PpE} \) for the time shift correction can remove the influences of the two aforementioned factors on the travel time of the \( PmP \) feature, allowing us to obtain the CRS record equivalent to an auto-correlation at a target station even when the Moho is dipping. The detailed result of the CRS records for the case of a dipping Moho is illustrated by a conceptual model shown in Figure S2 in Supporting Information S1.

Figure 2e shows the CCFs with the time offset corrected for station pairs between station 003 and other stations in the L-shaped array. Due to the small incident angle of teleseismic body waves, the reflection coefficient of body waves at the Moho is small. The negative time lags of the CCFs show visible features related to \( PmP \) phase at station 003 (Figure 2e), and after stacking, a clearer feature is obtained with a travel time of \( \sim 10 \) s (Figure 2f). It should be noted that, in the CRS, only those CCFs with distances longer than 40 km are utilized in the stacking to avoid interferences from surface wave signals.

5. Illuminating the Moho Morphology Using the Retrieved \( PmP \) Travel Times

With the obtained travel times of \( PmP \) phases from our CRS method, we here demonstrate how we can constrain the Moho morphology using the retrieved \( PmP \) travel times. We design an experimental model to illustrate our method. We set up a L-shaped array of 25 stations spaced at 10 km apart (Figure 3a), similar
as the practical L-shaped array in our study area. The Moho depth beneath station 01–07 is 30 km while the Moho depth below station 08–25 is 35 km. The Vp of the crust and upper mantle are set to 6.1 km/s and 8.1 km/s, respectively. Since the teleseismic P wave energy is mainly at the band of 3–10 s, we use a Ricker wavelet with a dominant period of 6 s as the incident P plane wave. The incident angle of this plane wave is set to 10°, and the azimuth is set to 50°. We then add Gaussian random noise with the 25% amplitude relative to the amplitude of P wave to the waveforms recorded at each station. Cross-correlations between any two stations are then calculated. Figure S3 in Supporting Information S1 shows the stacking processes for station 01. After applying the aforementioned CRS method to all stations, we arrange these stacked records for individual stations along the seismic line from station 1 to station 13 (Figure 3b). Clear features related to PmP phases beneath each station are observed, and a small step in these features is visible across stations 07 and 08, which is consistent with the Moho step in the designed model. All the arrival times of these features are consistent with the predicted PmP travel times (blue lines in Figure 3b) calculated from the experimental model. The results of synthetic experiments indicate that this method is reliable for imaging the morphology of Moho.

In addition to the demonstration based on the synthetic data, we also apply the time shift correction and stacking method to the CCFs from the L-shaped array in the Shandong Peninsula. The stacked records at each station are plotted in Figure 4 along the southern part of L-shaped array. Pronounced features related to Moho reflections (PmP) are very clear beneath each station along the profile. The arrival times of PmP features (the green line in Figure 4) display a concave shape at the middle section of this profile, which appears to be related to the location of the Tanlu fault zone.

To image the Moho discontinuity along the same profile, Chen et al. (2006) performed receiver functions using teleseismic earthquake events recorded during an eight-month period. Based on the Moho depth results and crust velocity model of Chen et al. (2006), we can calculate the arrival times of the PmP phases obtained in our method. Equation 2 describes this conversion process

\[ t_{\text{pmp}} = 2H\eta_p \]

where \( H \) is the crust thickness, \( \eta_p \) is the P wave vertical slowness (e.g., Zhu & Kanamori, 2000). P wave vertical slowness is calculated with the horizontal slowness from our beamforming analysis and P velocity model used in Chen et al. (2006). The dashed blue line in Figure 4 shows these calculated arrival times of PmP from receiver functions. We notice that the arrival times of PmP from our method are very similar to the calculated ones from receiver function results, even though we only use one month of continuous waveform data. There are some slight differences in the arrival times obtained by the two methods, which may be due to the imperfections of the P velocity model of Chen et al. (2006).

6. Discussions and Conclusions

The first step in obtaining travel times of reflected body waves from ambient noise is to analyze the locations of concentrated oceanic sources. In our application to field data, we use an L-shaped array to analyze the source locations of oceanic sources. If the source locations are known previously, we can use our CRS method to obtain signals related to the Moho reflected body waves without the need to perform source analysis. Many strong deep ocean sources have been discovered in many areas before, such as the persistent localized 26 s noise source near the Gulf of Guinea (Shapiro et al., 2006) and the 10 s oceanic noise source in the Kyushu Island, Japan (Zeng & Ni, 2010). These sources can be directly utilized to image the internal

![Figure 4. Results of features of PmP signals from our CRS stacking method along profile 1. The location of profile 1 is shown in Figure 1. Each trace is normalized with the root-mean-square (RMS) calculated within a time window of 0–30 s. The shaded features associate with PmP signals show a concave shape under the Tanlu Fault Zone. The dashed blue line presents the arrival time of PmP calculated based on the model of Chen et al. (2006). Note that we change the lag time of our Common Receiver Station (CRS) stacks from the negative times shown in Figures 1 and 2 to the positive leg times here.](image-url)
interfaces of the Earth if these localized sources excite body wave signals. In this paper, we only report the method to image the Moho morphology using our method. In principle, our method can also be extended to obtain travel times of other interface signals, such as the lithosphere-asthenosphere boundary, the 410 and 660 discontinuities of the mantle.

The microseismic source we use in this study is a persistent localized source generated by ocean swells during a one-month period of time (Euler et al., 2014). In principle, our method can be readily applied to energetic sources generated by strong oceanic storms with much shorter durations such as those excited by hurricanes and typhoons (Zhang et al., 2010). Such applications using short-duration sources generated by individual oceanic storms will be pursued in future studies.

As we mention in the preceding section, the stacked record at a station we obtain from our method is equivalent to an auto-correlation record. The reason why we do not directly obtain the record by auto-correlating noise records at a station is that there is strong surface wave energy in the noise recorded at the microseismic band. Surface waves together with body wave signals result in dominant signals at near-zero arrival times in an auto-correlation function. The energetic near-zero arrival time signals obscure the arrivals of PmP, making it hard to retrieve the accurate arrival time of PmP from an auto-correlation record.

In conventional methods, travel time information of internal interfaces of the Earth is obtained based on earthquake events through methods such as receiver function (Langston, 1977; Zhu, 2000), virtual deep seismic sounding (Tseng et al., 2009; Yu et al., 2013), and auto-correlation (Phạm & Tkalčić, 2017). Since most large earthquakes occur at plate boundaries, the lack of earthquake events in many regions, such as in deep oceans and inland areas, leads to a large gap of data coverage. On the other hand, when using earthquake data to obtain interface information, one needs to deploy instruments over a prolonged period of time in order to record a sufficient number of large earthquakes. Storms in the oceans are alternative data to earthquakes and help fill gaps in seismic data coverage. The method illustrated in this study can utilize the localized sources generated by ocean swells or storms to obtain the reflection information of interfaces. For regions lack of large quakes but experiencing intensive atmospheric activities, such as Mars, our approach may also offer a potential way to study the subsurface structures there.

**Data Availability Statement**

Open Research: The raw seismic data used in this study are archived at the IRIS Data Management Center (https://doi.org/10.7914/SN/YY_2000). All the cross-correlation data are available at an online repository (https://doi.org/10.6084/m9.figshare.16306176.v1).

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