9-21-2020

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Accepted to Seismological Research Letters, Sept. 21, 2020

Abstract

Local seismic events recorded by the large-N IRIS Community Wavefield Experiment in Oklahoma are used to estimate Moho reflections near the array. For events within 50 km of the center of the array, normal moveout corrections and receiver stacking are applied to identify the *PmP* and *SmS* Moho reflections on the vertical and transverse components. Corrections for the reported focal depths are applied to a uniform event depth. To stack signals from multiple events, further static corrections of the envelopes of the Moho reflected arrivals from the individual event stacks are applied. The multiple-event stacks are then used to estimate the pre-critical *PmP* and *SmS* arrivals, and an average Poisson's ratio of 1.77±.02 was found for the crust near the array. Using a modified Oklahoma Geological Survey (OGS) velocity model with this Poisson's ratio, the time-to-depth converted *PmP* and *SmS* arrivals resulted in a Moho depth of 41±.6 km. The modeling of wide-angle Moho reflections for selected events at epicenter-to-station distances of 90 to 135 km provides additional constraints, and assuming the modified OGS model, a Moho depth of 40±1 km was inferred. The difference between the pre-critical and wide-angle Moho estimates could result from some lateral variability between the array and the wide-angle events. However, both estimates are slightly shallower than the original OGS model Moho depth of 42 km, and this could also result from a somewhat faster lower crust. This study shows that local seismic events, including induced events, can be utilized to estimate properties and structure of the crust, which in turn can be used to better understand the tectonics of a given region. The recording of local seismicity on large-N arrays provides increased lateral phase coherence for the better

identification of pre-critical and wide-angle reflected arrivals.

Keywords: Large-N seismic arrays, Moho reflections, Crustal Poisson's ratio

Introduction

New advances in seismic instrumentations have made it possible to deploy arrays with a large number of seismometers (large-N arrays). The seismic sensors used in these large-N experiments are small, inexpensive systems which allow for fast deployments, and incorporate a geophone, a data logger, a GPS clock, and power (Freed, 2008; Karplus and Schmandt, 2018; Ringler et al., 2018). Some of the first academic applications of large-N arrays used single-component seismic nodes, including experiments conducted near Long Beach, California (Lin et al., 2013; Schmandt and Clayton, 2013; Nakata et al., 2015; Riahi and Gerstoft, 2015; Z. Li et al., 2018), Mount Saint Helens (Hansen and Schmandt, 2015; Hansen et al., 2016; Kiser et al., 2016, 2019; Wang et al., 2017; Glasgow et al., 2018), and along the San Jacinto fault zone (Hillers et al., 2016; Roux et al., 2016). Recent seismic studies involving three-component nodes have also demonstrated the wide application of large-N arrays for the fault zones (Qin et al., 2018; Wang et al., 2019), geothermal areas (Brenguier et al., 2016; Ward and Lin, 2017; Wu et al., 2017), for earthquake mechanism studies (Fan and McGuire, 2018), and for the analysis of cultural signals (C. Li et al., 2018).

A number of array based approaches have been used to study seismic body waves from the Moho, including controlled-source wide-angle reflection and refraction methods (e.g. Cook et al., 2010; Prodehl and Mooney, 2012; Carbonell et al., 2013). The *PmP* and *SmS* phases from earthquake sources have also been recorded on dense arrays being used for active-source studies, for example by Mechie et al. (2012) for determining crustal *S*-velocities and Poisson's ratio along the INDEPTH IV profile in northeast Tibet. Mori and Helmberger (1996) used closely spaced aftershocks from the 1992 Landers earthquake in southern California to observe *SmS* Moho reflections. Wang et al. (2018) used measurements of *P* and *PmP* arrivals from aftershocks and local earthquakes to perform crustal tomography in western Japan. Griffin et al. (2011) investigated the velocity structure of the Tibetan Lithosphere using travel-times from *P* and *PmP* arrivals from regional earthquakes recorded on the large-scale Hi-CLIMB array.

Array based methods have also been applied for seismic receiver functions from teleseismic earthquakes (e.g. Levander and Nolet, 2005; Rondenay, 2009; Nowack et al., 2010). Virtual deep seismic sounding (VDSS) using the *SsPmp* phases has been used for array based imaging of the crust and Moho from earthquake sources (Tseng et al., 2009; Yu et al., 2013; Thompson et al., 2019). Seismic interferometry has also been applied to teleseismic earthquake sources recorded by seismic arrays using global-phase seismic interferometry (Ruigrok and Wapenaar, 2012; Nishitsuji et al., 2016).

Shen and Ritzwoller (2016) utilized observations from ambient noise and earthquakes, receiver functions and Rayleigh wave ellipticity using the USArray transportable array to study the crust and upper mantle beneath the continental U.S., including the broader features beneath Oklahoma. Evanzia et al. (2014) conducted Vp and Vs tomography of the crust and upper mantle beneath Texas and Oklahoma, and Zhu (2018) performed tomography of ambient noise for crustal structure of North Texas and Oklahoma. In these studies, features in southern, eastern and southwestern Oklahoma including the Anadarko basin, Wichita uplift, the Arkoma basin and the

Ouachita uplift could be observed in the tomographic images.

A state-wide velocity model in Oklahoma was developed by Marsh (2018) using ambient noise tomography down to about 20 km in depth, where slower regions were associated with basin structures and faster regions with shallower basement. Ratre and Behm (2019) developed a 3D model of central Oklahoma from the combination of 1D travel-time curves resulting from the stacking and inversion of Pg waveforms. They found a generally homogeneous crust with some velocity variations related to regional geological structures. They also estimated a higher speed lower crust inferring a more mafic lower crust, and used this to assess the extent of the mid-continent rift in Oklahoma. Pei et al (2018) performed *Pg* tomography with anisotropy to investigate crustal seismogenic features in central Oklahoma and correlated these with recent seismic activity. Wang et al. (2019) applied P-to-S receiver functions from teleseismic earthquakes to investigate the crust of Oklahoma.

Inamori et al. (1992) utilized seismic reflection methods to image mid-crustal structure using aftershock waveforms from the 1984 Western Nagano Prefecture earthquake in Japan. Quiros et al. (2007) applied reflection imaging of aftershocks from the Mw 5.8, 2011 Virginia earthquake recorded by several dense seismic arrays to image crustal reflections. Vertical seismic profiling (VSP) was utilized for earthquake events followed by VSP-CRP transformations to account for offset and depth of the events. Eddy and Harder (2018) investigated the correlation of local earthquakes from the IRIS Community Wavefield Experiment in north-central Oklahoma to form a reflection image of reflectors in the basement and sedimentary section.

In this study, later arrivals from frequent local seismic events near the IRIS Community Wavefield Experiment in Oklahoma (Anderson, et al., 2016; Sweet et al., 2018) are used to estimate Moho reflections and crustal Poisson's ratio near the array. This is followed by a normal move-out correction and stacking to an equivalent zero offset. Laterally homogeneous velocities and Moho depth in the vicinity of the array are assumed, with static corrections applied to account for small velocity variability and uncertainties in the earthquake locations. Since the event depths are much shallower than the Moho depth, a reflection methodology is applied rather than full VSP processing.

Seismic events with locations from the Oklahoma Geological Survey Earthquake catalog (OGS Earthquake Catalog, 2016, Walter, et al., 2020) near the array are utilized. Waveform-relocation and template-matching studies have also been performed by Schoenball and Ellsworth (2017) and Skoumal et al. (2019). Many of the seismic events in Oklahoma are now identified as induced events from oil and gas activities, and in particular from waste water injection (Ellsworth, 2013; Keranen et al., 2014; Walsh and Zoback, 2015; Weingarten et al., 2015; Langenbruch and Zoback, 2016; Hincks et al., 2018).

We first analyzed the *P*- and *S*-wave later arrivals on the vertical and transverse components for selected events within 50 km from the center of the array. We identified the Moho reflections on the vertical and transverse components for individual events by applying normal moveout corrections and receiver stacking. *P*-wave and *S*-wave Moho reflection signals were enhanced by multiple-event stacking after static corrections and polarity sign corrections were applied. These stacks were then used to infer average crustal Vp/Vs ratio, Poisson's ratio and Moho depth near the array. We also modeled wide-angle *PmP* and *SmS* Moho reflections for selected events at epicenter-to-station distances of 90 to 135 km, which provides additional constraints. The analysis of local seismicity, including induced seismicity in areas with less natural seismicity, can be used to estimate the properties and structure of the crust, which in turn can be utilized to better understand the tectonics of a given region. The recording of local seismicity on large-N arrays allows for increased lateral phase coherence for the better identification of smaller seismic arrivals.

Seismic Data

The seismic data used in this study are selected from the Oklahoma Geological Survey (OGS) Earthquake Catalog from June 24 to July 20, 2016 (OGS Earthquake Catalog, 2016). The seismic events were recorded by the seismic stations of IRIS (Incorporated Research Institutions for Seismology) Community Wavefield Experiment in Oklahoma. Fig. 1a) shows the locations of seismic events with epicenters less than 50 km from the center of the array. Selected events in this distance range are then used to investigate P- and S-wave Moho reflections. The seismic array and a 50 km radius region around it, including the selected events, are also shown in Fig. 1b), along with the major geological features in Oklahoma (Johnson, 2008).

The IRIS Community Wavefield Experiment included 363 nodal sensors deployed along three lines, a 7-layer gradiometer nodal array, and 18 broadband stations collocated with 9 infrasound stations (Sweet et al., 2018). The nodes deployed in this experiment were Z-Land Fairfield nodal 5 Hz three-component sensors with ~35 days of battery life and a 250 samples/s sampling rate.

The right inset of Fig. 1a) shows the geometry of the seismometers in the array. The triangles are the north-south and east-west linear nodal array stations, the diamonds denote stations from the nodal gradiometer array. The spacing between the nodes along the north-south and east-west lines was about 100 m. The length of the east-west line is about 13 km, and the length of the north-source lines are about 5 km. The linear array nodal sensors were deployed from 20 June to July 20 for about 30 days (Sweet et al., 2018). The squares are the 18 broadband stations, which were deployed from June 20 to November 10, 2016 for about five months. For this study, the east-west linear nodal array stations were utilized to analyze the seismic events near the array.

A reference velocity model for Oklahoma from the Oklahoma Geological Survey (Darold et al., 2015) is shown in Figure 2 and is denoted by solid lines, with the S-wave velocities on the left and P-wave velocities on the right. The OGS velocity model assumes a constant 1.73 Vp/Vs ratio. A modified S-velocity model with a Vp/Vs ratio of 1.77 obtained from the analysis of *PmP* and *SmS* arrivals in this study is also shown in Figure 2 denoted by the dashed line. For comparison, the velocity model for locations near the array from CRUST1.0 (Laske et al., 2013) is also shown in Figure 2 by the dotted lines. The depth of the Moho is at 42 km for the OGS velocity model and also for CRUST1.0 in the region near the array.

Move-out Analysis

Alignment of Moho reflected seismic signals is performed in order to enhance the signal to noise ratio (SNR). Fig. 3 illustrates the move-out corrections performed for the alignment of the Moho-reflected signals.

The dashed line in Fig. 3a) is a schematic of Moho reflected arrival-times based on ray tracing for the OGS velocity model (Darold et al., 2015) for a standard focal depth of 5 km. The predicted Moho reflection arrival-times have offsets ranging from 0 to 50 km. The asterisks illustrate the Moho reflection arrival-times for events of different focal depths reported in the OGS earthquake catalog. The offsets are calculated from the latitudes and longitudes of the receivers and OGS located events. To account for event depth variability, each trace is shifted in time by a time difference between the predicted Moho reflection arrival-times ray traced for a 5 km depth source and the OGS reported event depths. Each trace is then move-out corrected by the time between the dashed line and the flat line. In this way, the delay times resulting from offset are compensated for. The total time shift for each trace is denoted by the arrows in Fig. 3a), integrating the time shifts for the variability in both event depth and offset. The Moho reflection arrival-times are then approximately flattened for stacking along the receiver array for a given event.

Fig. 3b) illustrates the ray tracing for the Moho reflections. The asterisks and triangles represent seismic events and receiver array, respectively. The schematic ray paths are denoted by black lines between seismic events and receiver array. The move-out corrections then flatten the arrival-times to the zero-offset time for a standard 5 km event depth. This then allows for stacking of the arrivals across the array for each event.

Stacking for Moho Reflections

Single-Event Stacking

We analyzed the later *P*- and *S*-wave arrivals for the receiver gathers from events within 50 km of the center of the array. Move-out corrections and receiver stacking were then applied to identify the Moho reflections on the vertical and horizontal transverse components. Corrections for the reported focal depths were applied to a standard depth of 5 km.

Fig. 4 shows receiver gathers for two seismic events recorded on the vertical component by the east-west linear array nodal stations. Each trace is low-pass filtered to 2 Hz and gained by a power of t to approximately equalize the amplitude decay with time. The solid and dashed lines are the predicted arrival-times for the *P* and *PmP* phases from ray tracing, after a move-out correction for the *PmP* arrival-time has been applied. The rightmost traces for each single event receiver gather are the move-out corrected stacked traces. The arrows on the right denote the estimated arrival-times of the *P*, *S*, and *PmP* on the stacked trace.

Fig. 5 shows receiver gathers for two events recorded on the transverse component. Each trace is low-pass filtered to 1.5 Hz and gained by a power of t to approximately equalize the amplitude decay with time. The solid and dashed lines are the predicted *S* and *SmS* arrivals after a move-out correction for the predicted *SmS* phase. The rightmost traces in each single-event receiver gather is the move-out corrected stacked traces. The arrows on the right denote the estimated arrival-times of the *S* and *SmS* on the stacked trace.

The filtering, gaining, move-out corrections, and stacking are used to bring out the Moho reflected signals. The observed Moho reflection signals could arrive earlier or later than the predicted arrival-times from the ray tracing. The deviation between observed and predicted Moho reflected arrival-times could result from variations in Moho depths, location and focal depth variability in the OGS earthquake catalog and velocity variability.

Multiple-Event Stacking

In order to further enhance the Moho reflection signals, 17 seismic events within 50 km of the center of the array were selected based on signal to noise level and event size to perform multiple-event stacks for the P- and S-wave Moho reflections. We also looked for events with offsets from the array where there was no interference between the Moho reflections and other phases. The selected pre-critical seismic events can be found in the supplementary material with distances less than 50 km from the array, and are shown by the dark black dots in Figure 1.

Fig. 6 illustrates the stacking of multiple events on the vertical component and the improvement of stacking by incorporating static and polarity corrections. Each trace on the left in Fig. 6a) is a single-event receiver gather stacking result. The rightmost trace is the multiple-event stacked trace. The line on top of each trace is the trace envelope. The Hilbert transform is used to form the analytic signal which is a complex time signal whose real part is the original signal and imaginary part is the Hilbert transform of the original signal (Shin and Hammond, 2008). The magnitude of analytical signal is the instantaneous amplitude, i.e. the envelope of the original signal.

After the envelope is calculated, the local maximum of the envelope around the predicted Moho reflection, within ± 1.5 seconds of the predicted arrival-time is determined. The average of the observed Moho reflection arrival-times is calculated, and each trace is shifted to align the peaks of the envelopes. The standard deviation about the average line is .29 s, and could result from lateral variability of the velocities, variations in the Moho depth, and uncertainties in the event locations and depths.

Sign corrections of the waveforms were also applied at the peak of the envelopes. Traces with negative amplitudes at the peak of the envelopes are multiplied by a negative one to account for the variability of polarities, resulting from the event focal mechanisms and possible effects from local structure. The corrected receiver stacked traces are then stacked in Fig. 6b). The dashed line is the predicted moveout-corrected arrival-times from the OGS model, and the solid line is the mean of the envelope peaks. The location of the direct *S*-wave is shown by an arrow. After the static and polarity corrections, the stacked trace is shown on the rightmost trace. The estimated *PmP* is shown by the arrow. Also shown by the lower arrow on the stack in Fig. 6b) is the possible location of the surface reflected *sPmP* arrival, although this is less evident on the individual event stacks.

Figure 7 shows a similar plot for the multiple-event stacking on the transverse component. The Moho reflection for the *S*-wave can also be enhanced by the multiple-event stacking after similar static corrections and sign corrections for single-event receiver stacks, where the inferred *SmS* is shown by the arrow. The lower arrow in Fig. 7b) is the possible arrival of the surface reflected *sSmS* arrival, however this is less evident on the individual event stacks.

The average absolute deviations about the average lines for the *PmP* and *SmS* arrivals are .29 s and .40 s, respectively. The average standard errors for the selected events from the OGS catalog are 0.57 km for the horizontal locations and 1.1 km for the depths. For a 25 km offset and 5 km depth earthquake, the largest arrival-time deviations obtained from ray tracing for the errors in horizontal location and depth are 0.18 s for the *PmP* and 0.32 s for the *SmS*. These are comparable with the deviations about the average lines for the *PmP* and S*mS* arrivals, with additional components of the deviations resulting from lateral velocity variations and Moho depth.

The multiple-event stacks are used to identify the Moho reflections from the vertical P-waves and transverse S-waves. The zero-offset results from multiple-event stacks can then be used to estimate the average P- and *S*-wave velocity ratio, Vp/Vs, and the corresponding Poisson's ratio for the crust in this area. The normal moveout corrected P- and S-wave arrival-times from the multiple-event stacks after static and polarity sign corrections are 12.57 s and 22.28 s for *P*- and *S*-waves, respectively. The uncertainties for the stacked zero-offset Moho reflection arrival-times are approximately 0.15 s on the stacked traces. This results in an average crustal Vp/Vs ratio of 1.77 ± .02 and a corresponding Poisson's ratio of 0.266 ± .009.

The OGS velocity model of Darold et al. (2015) used an assumed Vp/Vs ratio of 1.73. The velocity model from CRUST1.0 (Laske et al., 2013) near the array has a variable Vp/Vs ratio with depth, with an average Vp/Vs for the crust of 1.75. The average Poisson's ration for the continental crust has been estimated as 0.265 from rock sample measurements (Christensen, 1996). Also an early study by Zandt and Ammon (1995) estimated a Poisson's ratio of 0.27 for platform areas of the continental crust.

Receiver functions can be used to estimate the crustal Vp/Vs using the H-κ method (Zhu and Kanamori, 2000). From the EarthScope Automated Receiver Survey (EARS) (Crotwell and Owens, 2005; IRIS, 2010), the Vp/Vs and Poisson's ratio are quite variable in the study area. Individual station values and a map of Vp/Vs values across the U.S. are given by IRIS (2010). Using the broadband stations from the IRIS Community Wavefield Experiment in Oklahoma, the Vp/Vs values range from 1.73 to 2.10 and Poisson's ratios range from 0.25 to 0.35. The Vp/Vs values from receiver functions, however, can have significant variability resulting from difficulties in the measurements of later arrivals of the receiver functions at specific stations. Based on results from well-log information, Schoenball and Ellsworth (2017) used a Vp/Vs of 1.78 to revise the OGS velocity model for their study. This is similar to the average crustal Vp/Vs of 1.77 found in this study from the *PmP* and *SmS* arrivals. The modified Vs velocities assuming a Vp/Vs of 1.77 is shown by the dashed line in Figure 2.

Time-to-depth conversion

The multiple-event stacking results after corrections are used to perform the time-to-depth conversions. The Moho depth is shown for both *PmP* and *SmS* after time to depth conversions in Figure 8. We use the OGS velocity model as a starting model and adjust the Vp/Vs ratio in order to make Moho depth from the *PmP* and *SmS* arrivals consistent. Deconvolutions are also performed on the multiple-event stacks in order to mitigate the effects of the later arrivals on the *PmP* and *SmS* traces.

Figure 8 shows the Moho reflection traces and their deconvolutions with both time and depth. Fig. 8a) shows the *PmP* and decon-*PmP* traces with time. The left trace is the multiple-event stack of the vertical component after static and polarity corrections. A tapering is applied between 9 to 11 seconds to mitigate the effect of direct *S* arrivals. Slight filtering is also applied to make the frequency content on the vertical component comparable to that of the transverse component. The right trace in 8a) is the deconvolution to remove the second pulse between 15 to 16 seconds. Assuming that the stacked trace $y(t)$ is comprised of the Moho reflection pulse $r(t)$ and a delayed pulse $a \cdot r(t-\tau)$, where a is the amplitude of the second pulse and τ is the time delay of the second pulse, the stacked trace $y(t)$ can be approximated as

$$y(t) = r(t) + a \cdot r(t-\tau) \qquad 1.1$$

The corresponding Fourier transform can be written as

$$Y(\omega) = R(\omega)(1 + ae^{-i\omega\tau}) \qquad 1.2$$

where ω is the radial frequency. The Fourier transform of $r(t)$, i.e. $R(\omega)$, can then be deconvolved as

$$R(\omega) = \frac{(1+ae^{i\omega\tau})Y(\omega)}{(1+ae^{i\omega\tau})(1+ae^{-i\omega\tau})+\varepsilon^2} \qquad 1.3$$

where ε is a small damping parameter used to stabilize the deconvolution.

Fig. 8b) shows the *SmS* and decon-*SmS* traces with time. The left trace is the multiple-event stack of the transverse component after static and polarity corrections. The right trace in 8b) is deconvolution to remove the second pulse between 25 to 26 seconds, in a similar fashion as for the *PmP*.

Fig. 8c) shows the *PmP* and decon-*PmP* traces with depth. The OGS P-wave velocity model is used to perform a time-to-depth conversion, where for each point of the time axis, an equivalent depth is found by ray tracing for a source with a 5 km focal depth assuming single scattering from reflectors at depth. The Moho depth calculated from the *P*-wave Moho reflection is 41 ± .6 km, where the uncertainty is based on the width of the central pulse of the *PmP* arrival converted to depth. This is slightly shallower than the 42 km Moho depth from the original OGS velocity model.

Fig. 8d) shows the *SmS* and decon-*SmS* traces with depth. The Moho depth is converted from time in a similar fashion as for the P-wave traces, and is dependent on the Vs velocity model used. The original OGS velocity model assumed a Vp/Vs ratio of 1.73, but this doesn't result in a consistent Moho depth estimation between the *PmP* and *SmS* traces converted to depth. However, using a Vp/Vs ratio of 1.77 to convert the *SmS* trace to depth results in a consistent Moho depth estimation with the *PmP* trace. Note that converting from time to depth results in a better resolution in depth for the S-wave traces compared to the P-wave traces.

There are two possibilities for the second arrivals on the *PmP* and *SmS* traces which arrive later than the Moho reflections, between 9 to 11 seconds on the *PmP* trace in time on the left plot of Fig. 8a) and between 25 to 26 seconds on the *SmS* trace in time on the left plot of Fig. 8b), respectively. The first possibility is that these are primary reflections from a deeper interface. As shown in the OGS velocity model in Figure 2, there could be a potential deeper interface at around 50 km in this model.

The left traces in Figs. 8c) and 8d) are the time-to-depth conversion of the multiple-event stacks on the vertical and transverse components. The depth range on these plots is shown from 35 to 50 km. If the later second pulses are primary reflections from a deeper interface, this interface will be deeper than 50 km shown on Fig. 8c) for the *P*-wave converted to depth and at 48 km on Figure 8d) for the *S*-wave converted to depth. The depth of a possible deeper interface for these later phases is thus inconsistent between the *P*-wave and *S*-wave traces converted to depth. We therefore exclude the possibility that these later arrivals result from a deeper interface.

The second possibility is that the later phases are surface reflected phases. However, to be consistent between the *PmP* and *SmS* traces, the later pulse on the *PmP* trace would need to be an *sPmP* arrival and the later phase on the *SmS* trace an *sSmS* arrival. For zero-offset propagation, the amplitude for the *S*-to-*P* conversion at the free surface would be zero. However, the stacked *P*-wave trace is derived from the stacking of traces with non-zero offsets, and this would allow for a non-zero amplitude on the stacked equivalent zero-offset *P*-wave trace. Also, for earthquake sources the S-wave amplitudes would be approximately (Vp/Vs)^3, or 5.5 times larger in amplitude then the *P*-waves from the source, assuming a Vp/Vs = 1.77 (Aki and Richards, 2002). The later arrival on the P-wave trace is however still about .2 s later than would be fully consistent with an *sSmS* on the *S*-wave trace. However, this could result from some phase distortion from the processing for the second arrival on the *P*-wave trace. If the later arrivals are surface reflected arrivals, the traces on the right for each of the subplots on Figure 8 would show the results for just the primary reflection arrivals, with the later arrivals removed by deconvolution.

Further sources of errors in the results would be from the assumed velocity model. For the analysis performed here, the velocity model in the vicinity of the array is assumed to be laterally homogeneous, and static corrections are implemented to account for small lateral inhomogeneities. For a variation of the velocity model in depth, this would affect the P-wave and S-wave results in a similar fashion in terms of the inferred average crustal Vp/Vs ratio. However, a slightly faster lower crust than that of the assumed OGS velocity model, as inferred for example by Ratre and Behm (2019), would make the estimates of Moho depths correspondingly deeper.

Wide-angle Moho Reflections

As an illustration of recording wide-angle reflections from the Moho for a large-N array, several seismic events were selected to form composite record sections. The selection of these seismic events was based on several criteria, including background cultural noise level and the magnitude of seismic events. Selected events for the analysis of wide-angle arrivals, with distances between 90 to 135 km from the center of the array, are listed in the supplementary material.

The transverse *S*-wave components from five seismic events recorded on the left north-south linear nodal array stations are plotted on the composite record section in Fig 9a). The events are 21, 22, 24, 25 and 26 in the Supplemental material. Each trace is low-pass filtered to 1.5 Hz. Fig 9b) shows the annotations of several observed seismic phases. The thick solid line and arrows show the predicted *SmS* arrival, where the arrival-times are ray traced from the modified OGS velocity model and the depths in the OGS seismic catalog. A Moho depth of 40 ± 1 km was inferred for the wide-angle modeling of these events with epicenter-to-station distances of 100-135 km from the array for the S phases. However due to the depth variability in the catalog, the predicted arrival-times may deviate from the observed arrival-times. The thin solid line and arrows show the direct *S* arrival, which is the first strong seismic phase on the transverse component. The light gray dashed line and arrows show the location of an arrival after the direct *S* but followed by the *SmS* arrival from the Moho at this distance range, which could be a surface reflected direct arrival *sS*, or a surface wave with an apparent speed of 3.45 km/s in a narrow band frequency 0-1.5 Hz (Zhan et al., 2010). The surface reflected Moho reflection phase *sSmS* is also denoted by the solid dashed line and arrows in Fig 9b). The amplitude of this phase varies between events, partly due to the shallow subsurface structure on the source side.

The vertical *P*-wave components from six selected seismic events recorded on the left north-south linear nodal array stations are plotted on the composite record section in Fig 10a). The events are 18, 19, 20, 23, 25 and 26 in the Supplemental material. Each trace is low-pass filtered to 2 Hz. Fig 10b) shows the annotations of several observed seismic phases. The thick solid line and arrows show the predicted *PmP* arrival, where the arrival-times are ray traced from the modified OGS velocity model and the depth in the OGS seismic catalog. Again, a Moho depth of 40 ± 1 km was inferred for the wide-angle modeling of these events with epicenter-to-station distances of 90 to 135 km from the array for the P phases The thin solid line and arrows show the direct *P* arrivals. The slight gray dashed line and arrows show the location of an arrival after the direct *P* but followed by the *PmP* arrival from the Moho, which could be the surface reflected directed arrival pP. The surface reflected Moho reflection phase *pPmP* is also denoted by the dotted line and arrows in Fig 10b). Due to the smaller energy of *P* wave from earthquakes and the affects from *SV* waves on the vertical component, the *PmP* Moho reflections are less evident than the *SmS* Moho reflections. The difference between the pre-critical and wide-angle Moho depth estimates could result from some lateral variability between the seismic array and the more distant wide-angle events. However, both estimates are shallower than the original OGS model, and this could also result from a somewhat faster lower crust.

Conclusions

The deployments of large-N seismic arrays have become increasingly popular, enabling advances in subsurface imaging. In this study, the *P*- and *S*-wave later arrivals from local seismic events near the large-N IRIS Community Wavefield Experiment in Oklahoma are used to illustrate the analysis of Moho reflections and average crustal thickness and Vp/Vs ratio near the array. The pre-critical *PmP* and *SmS* Moho reflections for events within 50 km of the center of the array were first found on single-event stacks by applying normal moveout corrections, low pass filtering and gaining methods. Multiple-event stacks were then applied to further enhance the Moho reflection signals from the stacking of the single-event receiver gathers. Static corrections and polarity corrections were applied to account for lateral variability of Moho depth and velocity structure, and uncertainties in the event locations and focal depths. The arrival-times of the *PmP* and *SmS* on multiple-event stacked traces after NMO, static, and polarity corrections were used to estimate an average crustal Vp/Vs ratio of 1.77 ± .02, with a Poisson's ratio of 0.266 ± .009 near the array. The inferred crustal Vp/Vs and Poisson's ratio are in good agreement with the previous studies for the average continental crust, although the results in the study area from receiver functions are more variable. Using a modified Oklahoma Geological Survey (OGS) velocity model with this Poisson's ratio, the time-to-depth converted *PmP* and *SmS* arrivals resulted in a Moho depth of 41 ± .6 km, which is slightly shallower than the 42 km for the original OGS model.

For selected events at epicenter-to-station distances of 90 to 135 km, the modeling of wide-angle arrivals provided additional constraints, and a Moho depth of 40 ± 1 km was inferred. The difference between the pre-critical and wide-angle Moho depth estimates could result from some lateral variability between the seismic array and the more distant wide-angle events, as well as event depth uncertainties. However, both estimates are slightly shallower than the original OGS model, and this could also result from a somewhat faster lower crust.

This study shows that local seismic events, including induced events in areas with less natural background seismicity, can be used estimate properties and structure of the crust, which can then be used to better understand the tectonics of a given region. The recording of local seismicity on large-N arrays provides increased lateral phase coherence and this allows for better identification of both pre-critical and wide-angle reflected arrivals.

Data and Resources

The seismic waveform data used in this study were collected as part of the IRIS Community Wavefield Experiment in Oklahoma. Data can be obtained from the IRIS Data Management Center at www.iris.edu (last accessed February 2020). Supplemental Material for this article includes a list of the seismic events used for this study. The hypocenter data for the events are from the Oklahoma Geological Survey at http://wichita.ogs.ou.edu/eq/catalog/2016/2016.csv (last accessed February 2020).

MATLAB software was used for the visualization, and several codes were modified from the CREWES MATLAB toolbox of the Consortium for Research in Elastic Wave Exploration Seismology, University of Calgary, Calgary, Alberta, CANADA.

Acknowledgments

The authors thank the two anonymous Reviewers, The Associate Editor, Dan McNamara, and the Editor for their constructive reviews, which substantially improved the manuscript. The study was partially supported by US National Science Foundation Grant EAR-1839322, and a Purdue EAPS travel grant for the first author.

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Figure 1. a) Map of near offset seismic events from the OGS catalog over a period from June 24 to July 20, 2016. The events are denoted by dots, and the dark black dots were selected for the analysis. The dashed circles have radii of 10 km, 30 km and 50 km from a central station of the array of the IRIS Community Wavefield Experiment. The right inset map shows the geometry of the receivers in the array. The triangles are the north-south and east-west linear nodal array stations, the small square denoted stations from the nodal gradiometer array, and the squares are the broadband stations (Sweet et al., 2018). b) shows the major geologic structures in Oklahoma (adapted from Johnson 2008). The triangle in central north Oklahoma denotes the location of the array. The dashed circles around the triangle have radii of 10 km, 30 km and 50 km from a central station of the array.

Figure 2. Reference velocity models of the study area. The velocity model from the Oklahoma Geological Survey (Darold et al., 2015) is denoted by solid lines, with S wave velocity on the left and P wave velocity on the right. The OGS velocity model assume a constant 1.73 Vp/Vs ratio. Using a Vp/Vs of 1.77 instead, we derive a new shear wave velocity model, denoted the dashed line. The velocity model from CRUST1.0 (Laske et al., 2013) is shown by the dotted lines for comparison.

Figure 3. a) Move-out corrections for the Moho reflections. The dashed line is the predicted Moho reflected arrival-times for a reference 5 km depth source, from the OGS velocity model (Darold et al., 2015). b) shows a schematic diagram of the Moho reflected rays. The asterisks represent

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Figure 4. a) A single-event receiver gather of the vertical components recorded by the east-west linear array nodal stations for event 5 in the supplementary material. Each trace is low-pass filtered to 2 Hz and gained by a power of t to approximately equalize the amplitude decay with time. The solid and dashed lines are the predicted arrival times for the *P* and *PmP* from ray tracing after a move-out correction for the *PmP* arrival. The rightmost trace is the event stack. The arrows on the right denotes the observed arrival-times of the *P*, *S*, and *PmP* on the stacked trace. b) is a similar plot for event 9.

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Figure 6. a) shows the multiple-event stacking of the vertical component stacks. Each trace on the left is a single-event receiver stack result. The rightmost trace is the multiple-event stacked trace. For each trace the envelope is also shown. The dashed line is the predicted time of the normal moveout corrected predicted *PmP* arrivals for the OGS model. The direct *S* is also shown for one

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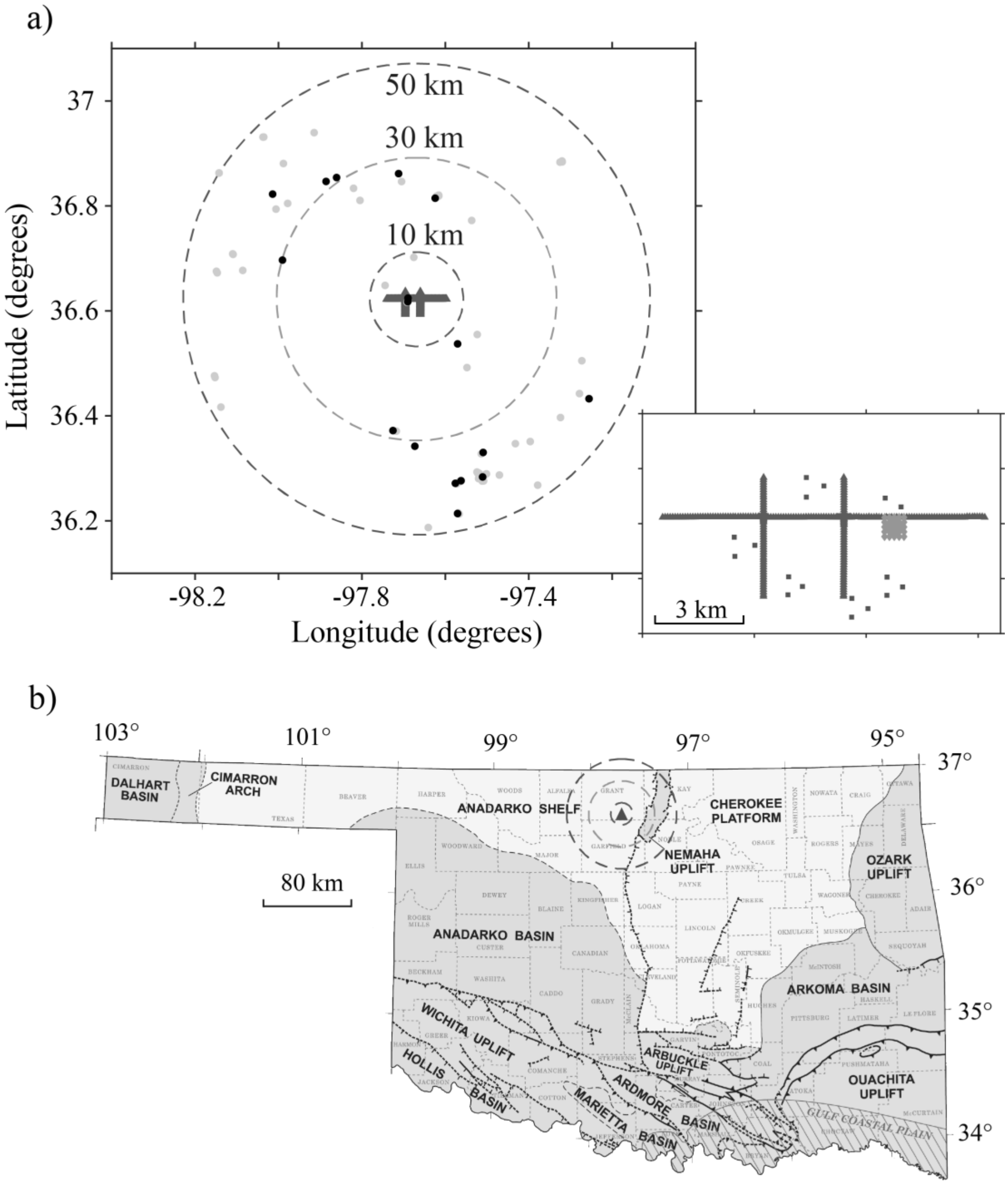
Figure 8. a) shows the *PmP* and decon-*PmP* traces with time. The left trace is the multiple event stack of the vertical component after static and polarity corrections. A tapering is applied between 9 to 11 seconds to mitigate the effect of direct *S* arrivals. Slight filtering is also applied to make the frequency content on the vertical component comparable to that of the transverse component.

The right trace is the deconvolution of the left trace to remove the later pulse between 15 to 16 seconds. b) shows the *SmS* and decon-*SmS* traces with time. The left trace is the multiple-event stack of the transverse component after static and polarity corrections. The right trace is the deconvolution of the left trace to remove the later pulse between 25 to 26 seconds. c) shows the *PmP* and decon-*PmP* traces with depth. The OGS *P* wave velocity model is used to perform the time to depth conversion. The Moho depth obtained for the *P*-wave Moho reflection is at 41 km. d) shows the *SmS* and decon-*SmS* traces with depth. The S-wave velocity calculated from OGS *P* wave velocity using a 1.77 Vp/Vs ratio is used to perform the time to depth conversion. The Moho depth obtained from the S-wave Moho reflection is also at 41 km consistent with the estimate from the *PmP* traces with depth.

Figure 9. a) Composite record section of the wide-angle *SmS* Moho reflections on the transverse component, recorded on the left north-south linear nodal array stations. Each trace is low-pass filtered to 1.5 Hz. The inset map shows the location of earthquakes, denoted by gray dots. The black dots are selected for the plot. b) shows the composite record section with identified phase shown. The thick solid line and arrows show the predicted *SmS* arrival. The arrival-times are ray traced from the modified OGS model and the depths in the OGS seismic catalog. The thin solid line and arrows show the direct *S* arrival. The surface reflected Moho reflection phase *sSmS* is also denoted by the dotted line and arrows. The light gray dashed line and arrow show either a surface reflected direct wave *sS* or a surface wave with a 3.45 km/s apparent velocity in a narrow frequency band.

Figure 10. a) Composite record section of the wide-angle *PmP* Moho reflections on the vertical component, recorded on the left north-south linear nodal array stations. Each trace is low-pass filtered to 2 Hz. The inset map shows the location of earthquakes, denoted by gray dots. The black dots are selected for the analysis. b) shows the composite record section with identified phase shown. The thick solid line and arrows show the predicted *PmP* arrival. The arrival-times are ray traced from the modified OGS velocity model and the depth in the OGS seismic catalog. The thin solid line and arrows show the direct *P* arrival. The surface reflected Moho reflection phase *pPmP* is also denoted by the dotted line and arrows. The light gray dashed line and arrows show a possible surface reflected direct wave *pP*.



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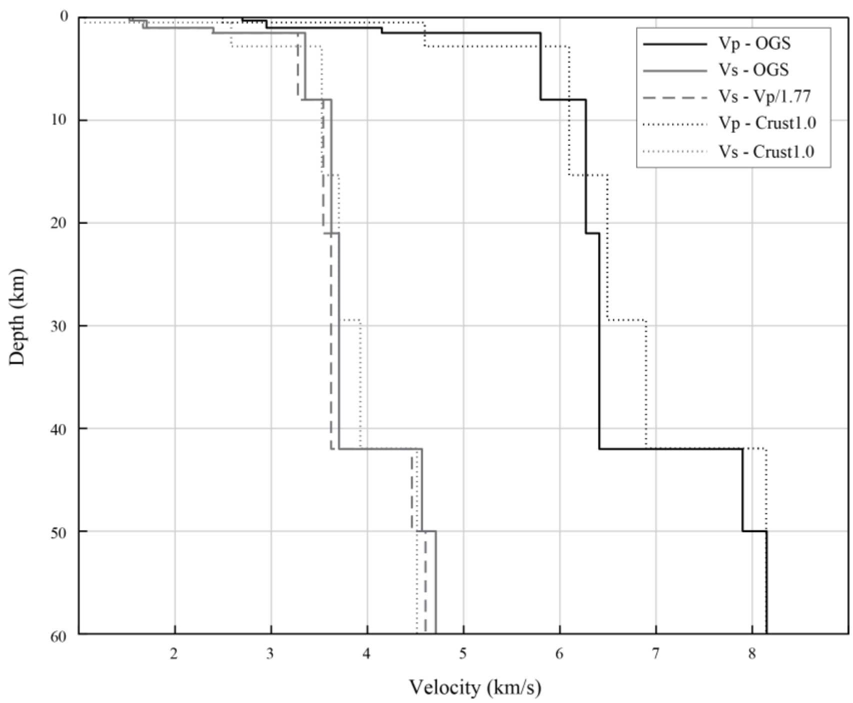

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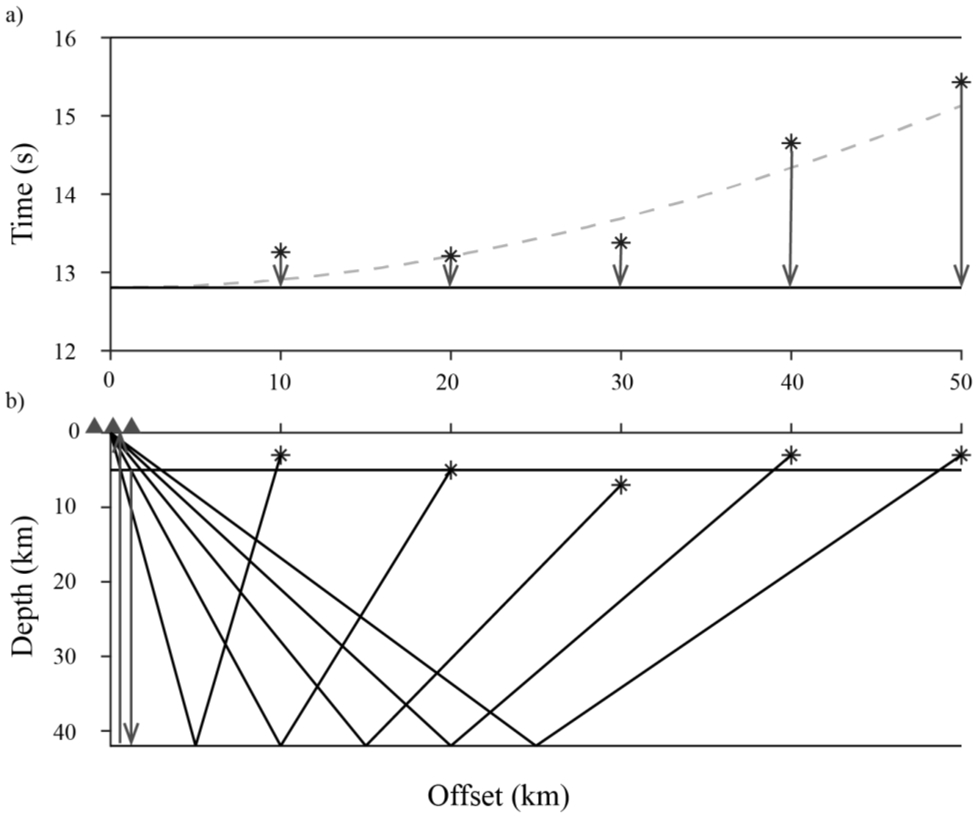

Figure 3. a) Move-out corrections for the Moho reflections. The dashed line is the predicted Moho reflected arrival-times for a reference 5 km depth source, from the OGS velocity model (Darold et al., 2015). b) shows a schematic diagram of the Moho reflected rays. The asterisks represent seismic events with different offsets and focal depths from the receiver array depicted by triangles. Move-out corrections are performed for each seismic event and receiver pair. The move-out time shifts flatten the Moho reflected arrivals and also correct for the event depths to a standard depth.

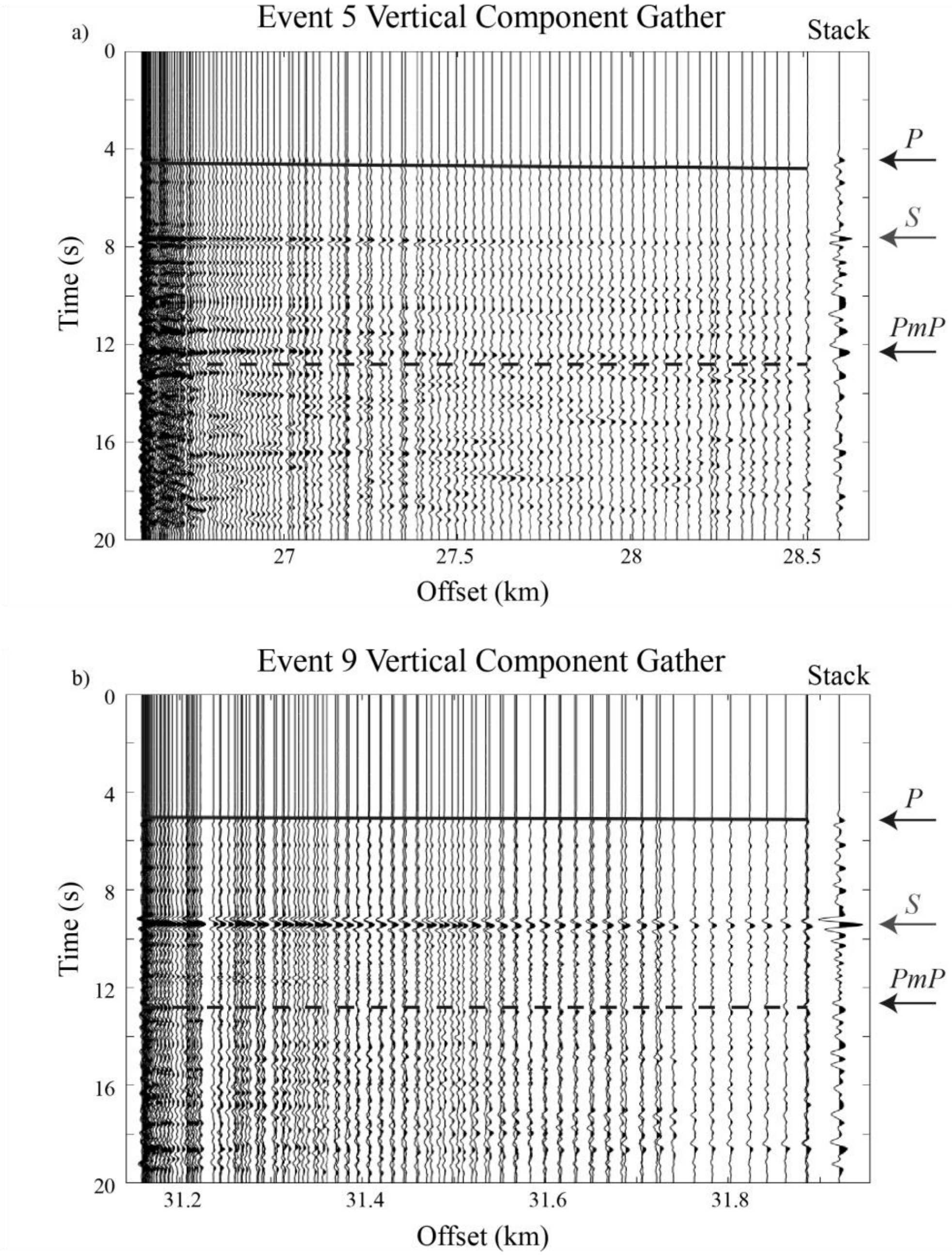

Figure 4. a) A single-event receiver gather of the vertical components recorded by the east-west linear array nodal stations for event 5 in the supplementary material. Each trace is low-pass filtered to 2 Hz and gained by a power of t to approximately equalize the amplitude decay with time. The solid and dashed lines are the predicted arrival times for the *P* and *PmP* from ray tracing after a

move-out correction for the *PmP* arrival. The rightmost trace is the event stack. The arrows on the right denotes the observed arrival-times of the *P*, *S*, and *PmP* on the stacked trace. b) is a similar plot for event 9.

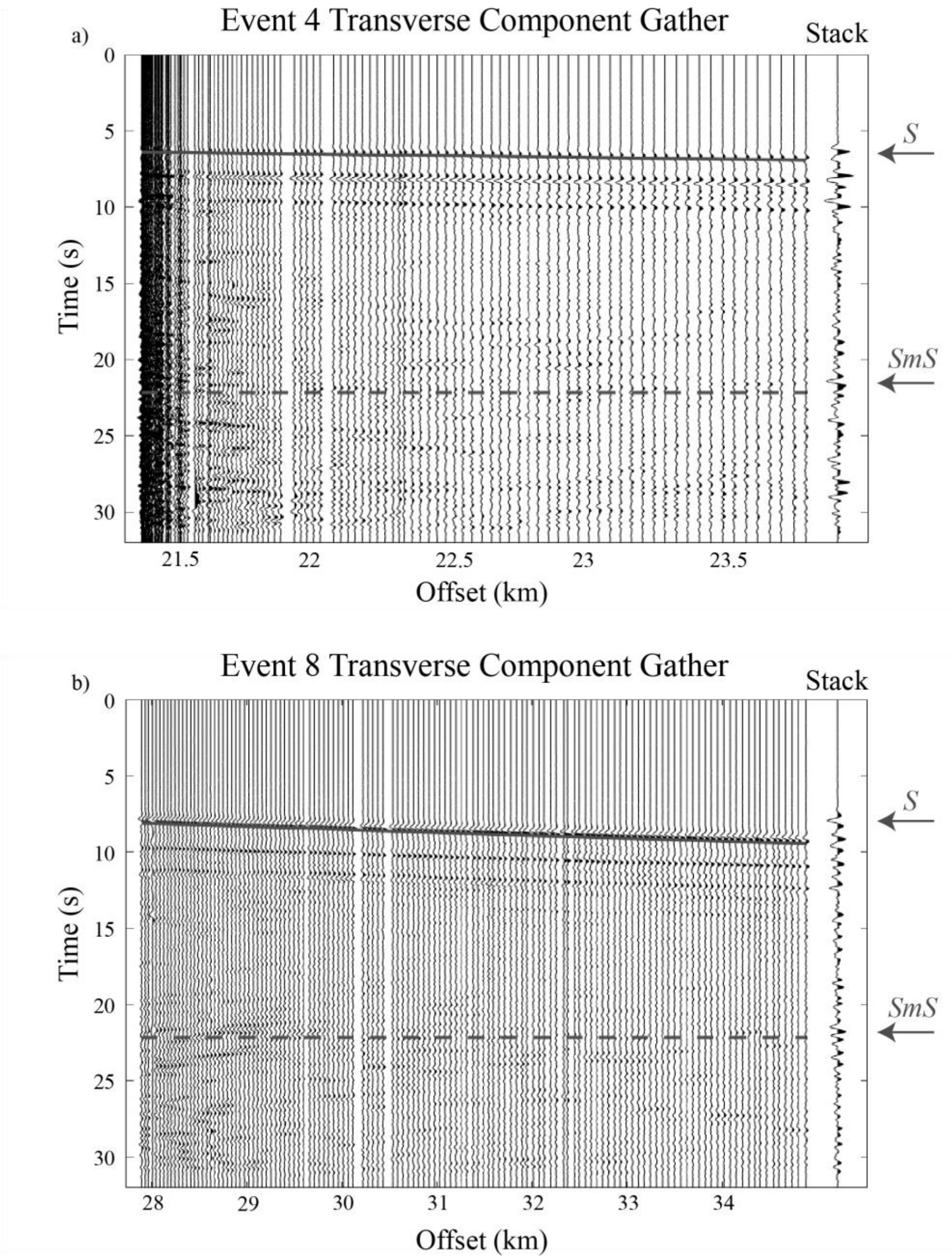

Figure 5. a) A single-event receiver gather of the transverse component recorded by east-west linear array stations for event 4 in the supplementary material. Each trace is low-pass filtered to 1.5 Hz and gained by a power of t to approximately equalize the amplitudes in time. The solid and dashed lines are the predicted *S* and *SmS* arrivals after a move-out correction for the *SmS* arrival-time. The rightmost trace in each single-event receiver gather is the stack. b) is a similar plot for

event 8.

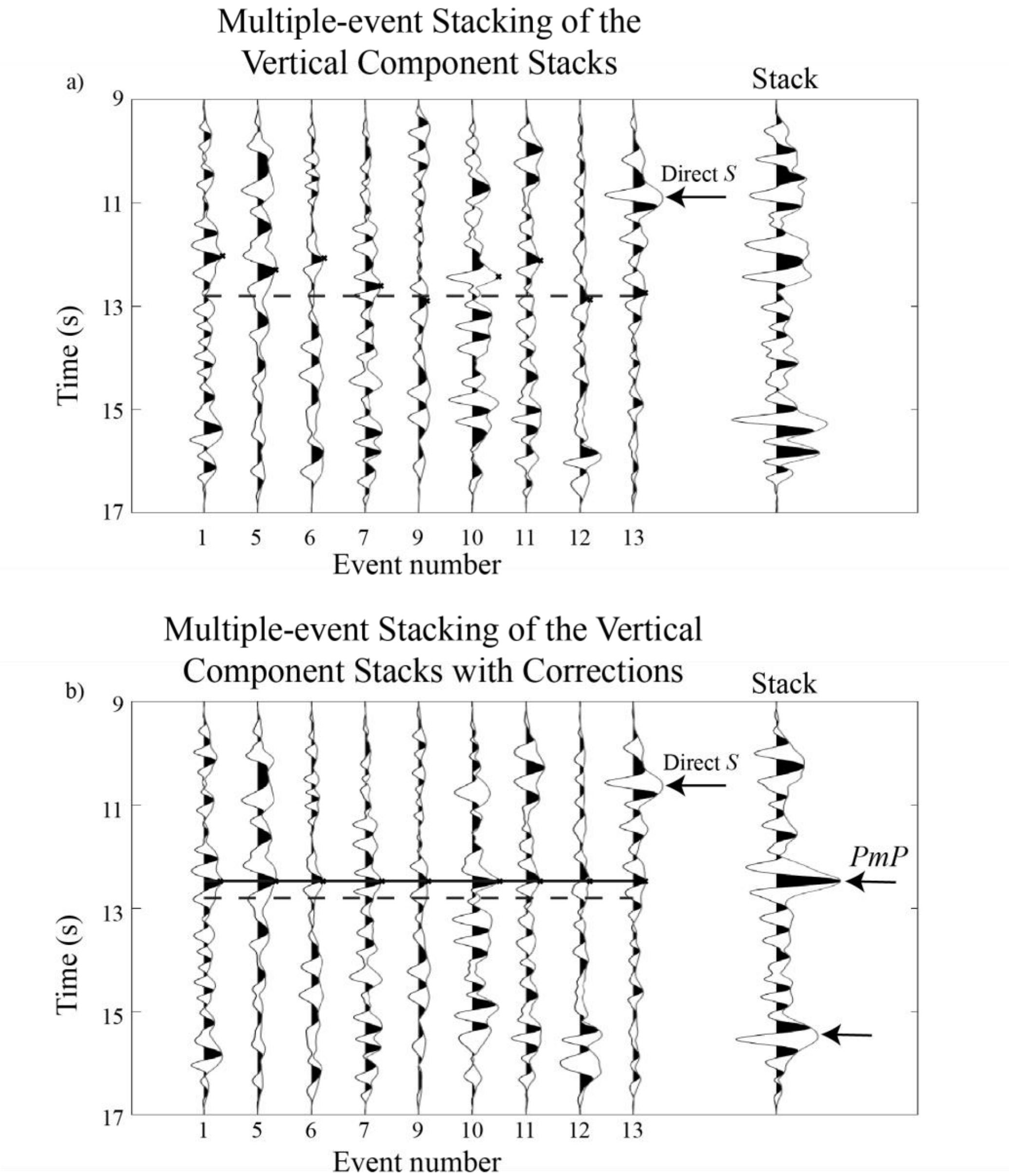

Figure 6. a) shows the multiple-event stacking of the vertical component stacks. Each trace on the left is a single-event receiver stack result. The rightmost trace is the multiple-event stacked trace. For each trace the envelope is also shown. The dashed line is the predicted time of the normal moveout corrected predicted *PmP* arrivals for the OGS model. The direct *S* is also shown for one

event. b) shows the multiple-event stack of the vertical component after static and polarity corrections. The dashed line is the predicted moveout corrected time of the *PmP* from the OGS model. The solid line is the average of the envelope peaks denoting the predicted Moho reflection time. The direct *S* wave is shown by arrows in both a) and b). Static corrections and polarity corrections are applied to the individual traces prior to stacking, with the stack shown on the right. The inferred *PmP* arrival is shown by an arrow. The lower arrow on the stack could be the location of the surface reflected *sPmP* arrival.

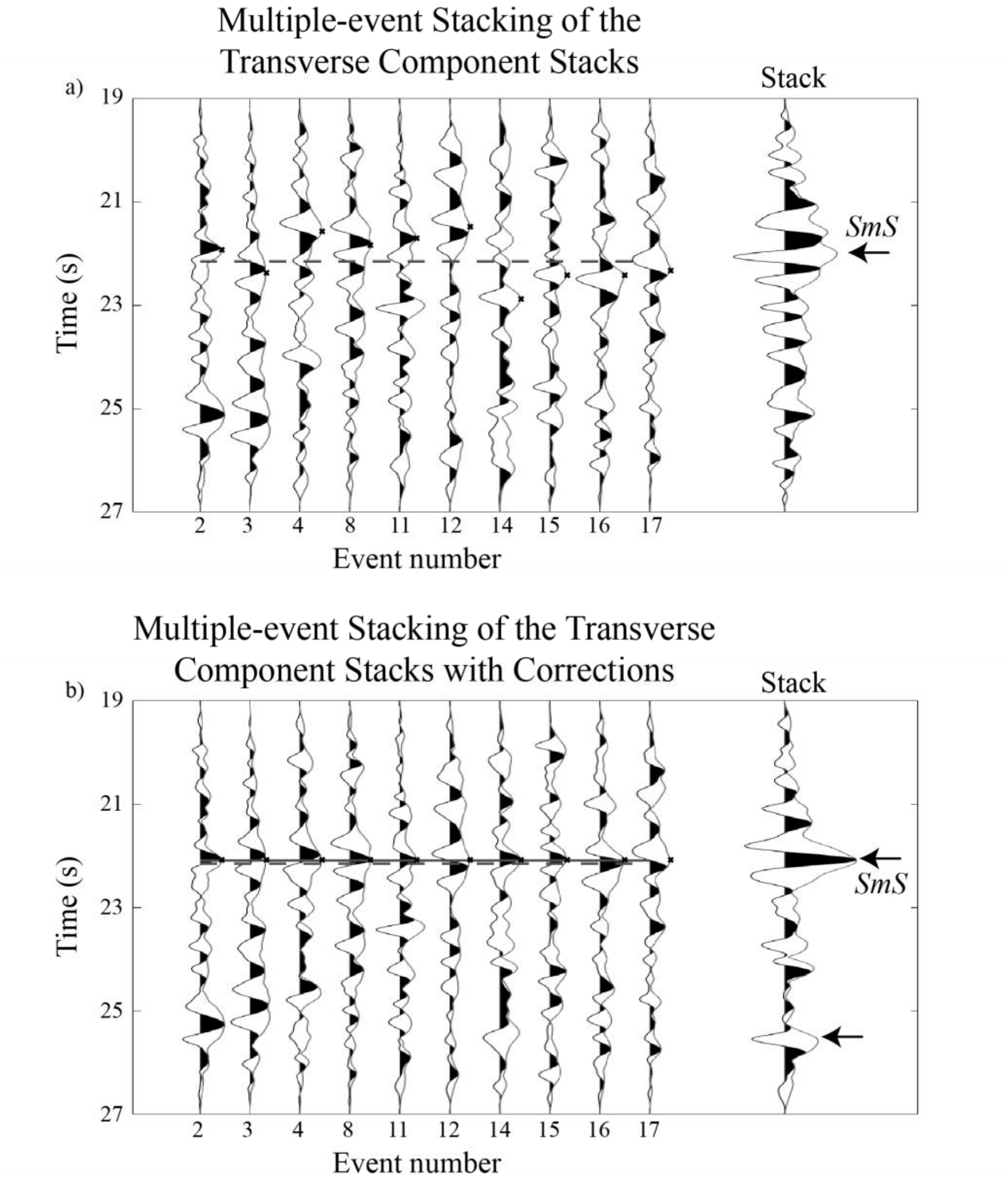

Figure 7. a) shows the multiple-event stacking of the transverse component stacks. Each trace on the left is a single-event receiver stack result. The dashed line is the predicted time for the moveout corrected *SmS* from the OGS model. The rightmost trace is the multiple-event stacked trace. b) shows the multiple-event stack of the transverse component after static and polarity corrections. The dashed line is the predicted time for the moveout corrected *SmS* from the OGS model. The solid line is the average of the envelope peaks denoting the inferred Moho reflection time. Static corrections and polarity corrections are applied to the individual traces prior to stacking, with the

stack shown on the right. The inferred *SmS* arrival is shown by an arrow. The lower arrow on the stack could be the location of the surface reflected *sSmS* arrival.

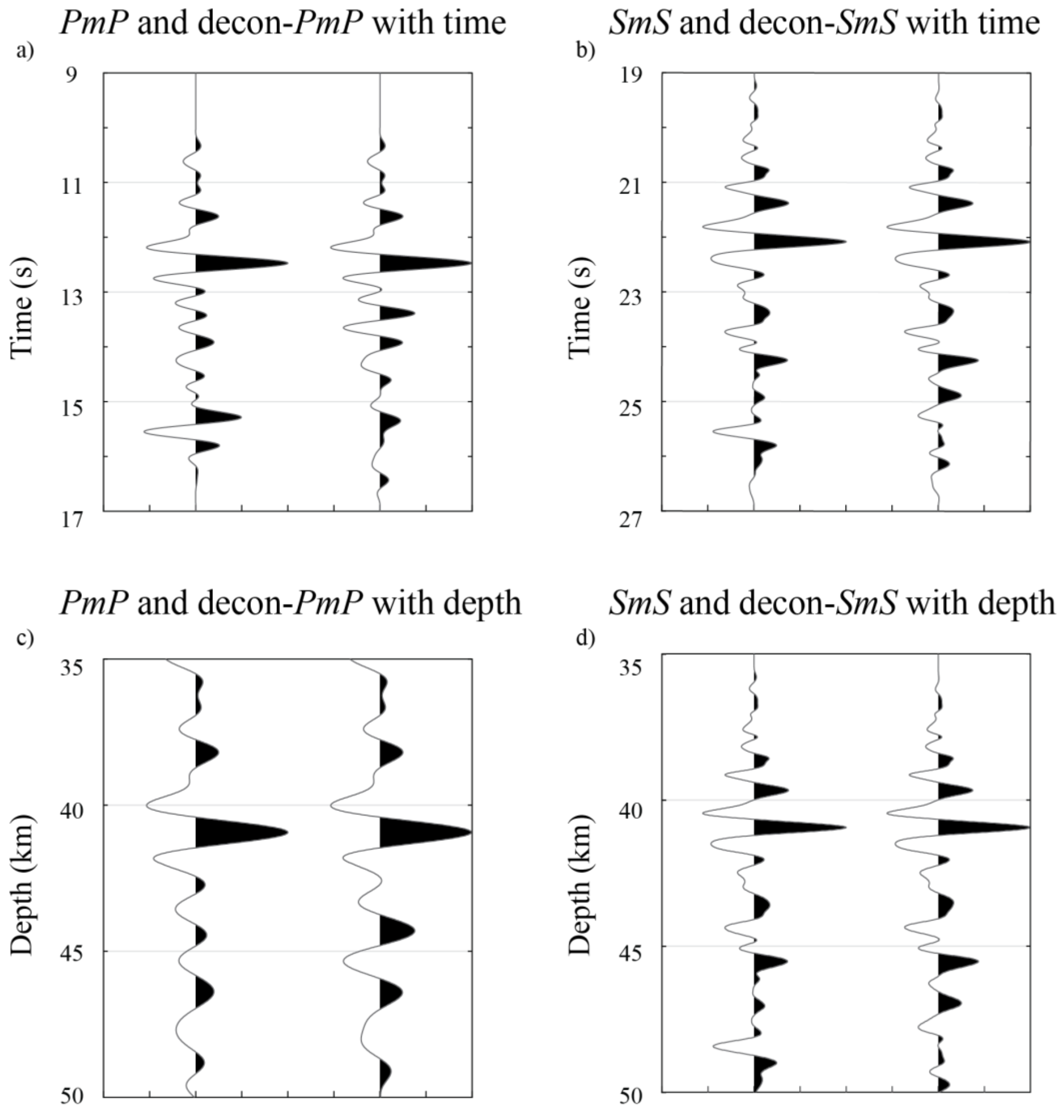

Figure 8. a) shows the *PmP* and decon-*PmP* traces with time. The left trace is the multiple-event stack of the vertical component after static and polarity corrections. A tapering is applied between 9 to 11 seconds to mitigate the effect of direct S arrivals. Slight filtering is also applied to make the frequency content on the vertical component comparable to that of the transverse component. The right trace is the deconvolution of the left trace to remove the later pulse between 15 to 16 seconds. b) shows the *SmS* and decon-*SmS* traces with time. The left trace is the multiple-event stack of the transverse component after static and polarity corrections. The right trace is the deconvolution of the left trace to remove the later pulse between 25 to 26 seconds. c) shows the *PmP* and decon-*PmP* traces with depth. The OGS *P*-wave velocity model is used to perform the

time to depth conversion. The Moho depth obtained for the *P*-wave Moho reflection is at 41 km. d) shows the *SmS* and decon-SmS traces with depth. The *S*-wave velocity calculated from OGS *P*-wave velocity using a 1.77 Vp/Vs ratio is used to perform the time to depth conversion. The Moho depth obtained from the *S*-wave Moho reflection is also at 41 km consistent with the estimate from the *PmP* traces with depth.

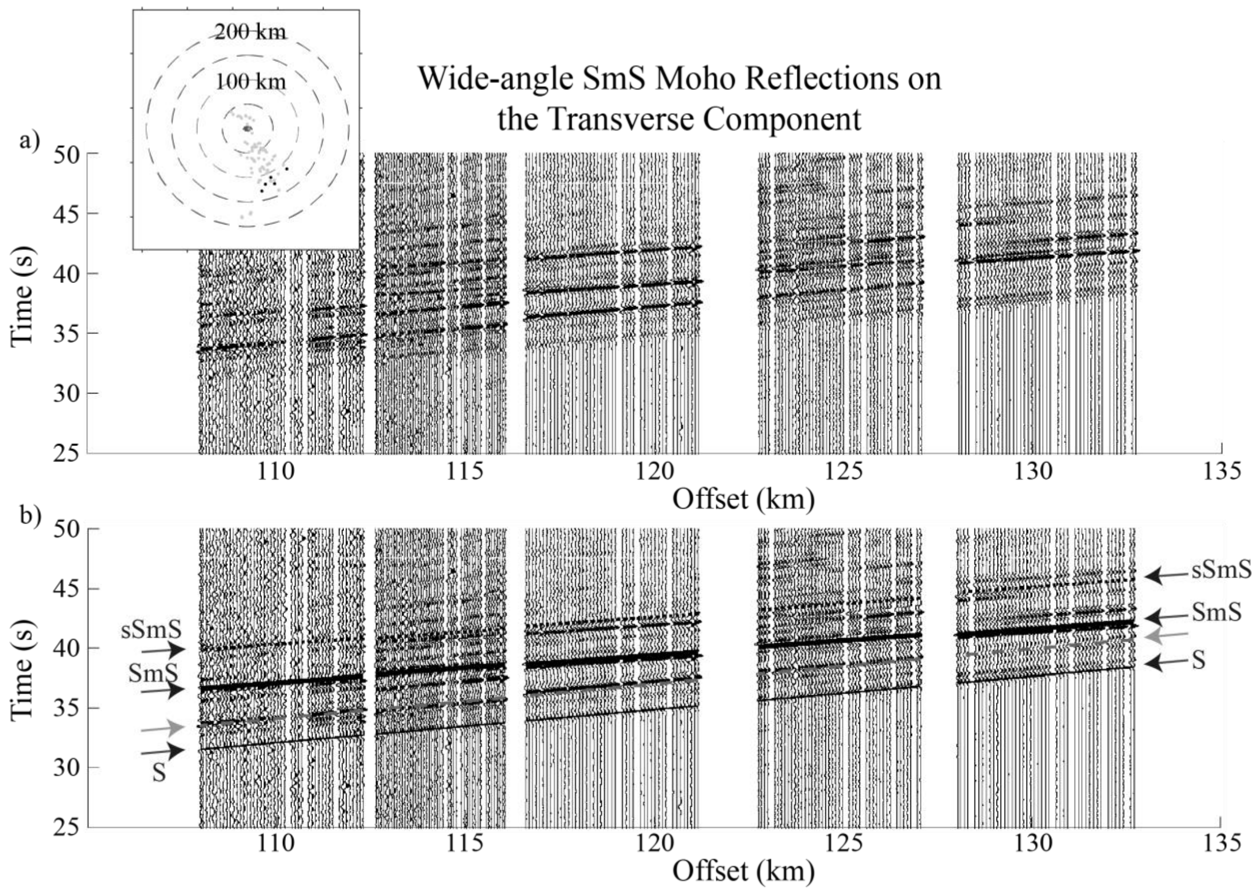

Figure 9. a) Composite record section of the wide-angle *SmS* Moho reflections on the transverse component, recorded on the left north-south linear nodal array stations. Each trace is low-pass filtered to 1.5 Hz. The inset map shows the location of earthquakes, denoted by gray dots. The black dots are selected for the plot. b) shows the composite record section with identified phase shown. The thick solid line and arrows show the predicted *SmS* arrival. The arrival-times are ray traced from the modified OGS model with a Vp/Vs of 1.77 and the depths in the OGS seismic catalog. The thin solid line and arrows show the direct *S* arrival. The surface reflected Moho reflection phase *sSmS* is also denoted by the dotted line and arrows. The light gray dashed line and arrow show either a surface reflected direct wave *sS* or a surface wave with a 3.45 km/s apparent velocity in a narrow frequency band.

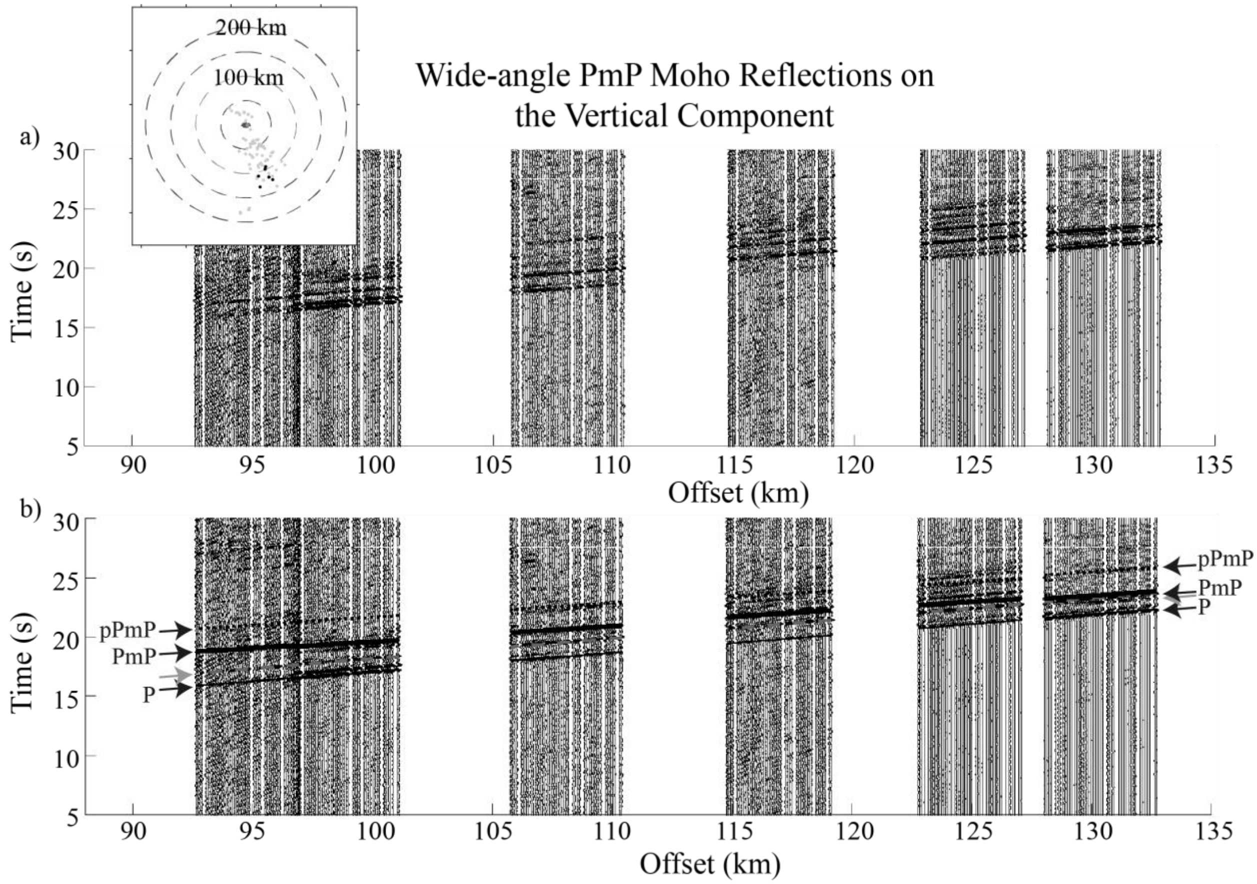

Figure 10. a) Composite record section of the wide-angle *PmP* Moho reflections on the vertical component, recorded on the left north-south linear nodal array stations. Each trace is low-pass filtered to 2 Hz. The inset map shows the location of earthquakes, denoted by gray dots. The black dots are selected for the analysis. b) shows the composite record section with identified phase shown. The thick solid line and arrows show the predicted *PmP* arrival. The arrival-times are ray traced from the modified OGS velocity model and the depth in the OGS seismic catalog. The thin solid line and arrows show the direct *P* arrival. The surface reflected Moho reflection phase *pPmP* is also denoted by the dotted line and arrows. The light gray dashed line and arrows show a possible surface reflected direct wave *pP*.