Presence of an intralithospheric discontinuity in the central and western North China Craton: Implications for destruction of the craton

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ABSTRACT
Detailed knowledge of lithospheric structure is essential for understanding the long-term evolution and dynamics of continents. We present an image of lithospheric structure across the central and western North China Craton (NCC), derived using S and P receiver functions from a dense seismic array. A negative velocity discontinuity is identified at ~80–100 km depth within the thick lithosphere (~160–200 km), similar to that observed in many other cratonic regions and roughly at the same depth as the base of the lithosphere in the eastern NCC. The intralithospheric discontinuity may indicate an ancient, mechanically weak layer within the overall strong cratonic lithosphere, and probably also existed beneath the eastern NCC before the Mesozoic. The presence of such a weak layer could have facilitated simultaneous lithospheric modification at the base and the middle of the lithosphere in the eastern NCC, especially under the strong influence of the Mesozoic Pacific subduction, leading to the severe lithospheric thinning and destruction recorded in this region. The weak layer probably did not strongly affect the stability and evolution of the central and western NCC and other cratonic regions where effects from plate boundary processes were weak. Our seismic images, integrated with geological data, provide new insights into structural heterogeneities in the subcontinental lithospheric mantle and their roles in the dynamic evolution of continents.

INTRODUCTION
Structural heterogeneities are ubiquitous in the continental lithosphere. How such heterogeneities have influenced the tectonic evolution of continents is an area of active research. Lateral structural heterogeneities, such as preexisting weak zones in the lithosphere, have been extensively studied and are thought to have functioned as the primary control on the evolution of continents (e.g., Begg et al., 2009; Chen, 2010). In particular, the anisotropic behavior of the weak zones in terms of thermal diffusivity and mechanical deformation can lead to intense heating and strain localization within the lithosphere, facilitating repeated reactivations of the ancient fabric during successive tectono-thermal events (e.g., Tornamasi et al., 2001). While vertical heterogeneities in the continental crust have also been widely recognized and are closely related to regional tectonics and evolution (e.g., Christensen and Mooney, 1995), structural variations with depth in the subcontinental lithospheric mantle (SCLM) remain much less understood.

Recent seismic studies commonly find evidence for a strong velocity drop or a low-velocity layer at ~100 km depth beneath continents (e.g., Fischer et al., 2010, and references therein). Such a shallow-mantle velocity drop is consistent with a thin lithosphere and was interpreted as the lithosphere-asthenosphere boundary (LAB) in tectonically active regions (e.g., Chen et al., 2008; Lekic et al., 2011); however, it is well within the lithosphere of stable cratons, where thick mantle keels commonly extend to ~200 km depth (e.g., Chen et al., 2009). This structure was thus considered to be an intralithospheric discontinuity (ILD) and a sign of vertical structural heterogeneities in the SCLM of cratons. The nature and formation of the ILD have been widely investigated, and it has been interpreted as either a remnant signature of the Archean formation of cratons (e.g., Chen et al., 2009) or resulting from later metasomatic refertilization as either a remnant signature of the Archean formation of cratons (e.g., Savage and Silver, 2008). However, the potential effects of the ILD on the later geodynamic evolution of cratons have seldom been studied, and identifying the ILD and LAB simultaneously beneath cratons remains a seismological challenge.

The North China Craton (NCC) is an ideal place to study structural heterogeneities and their influences on the lithospheric evolution, because of its marked deep structural variations that are well correlated with Phanerozoic tectonics and the spatially uneven reactivation and destruction of the cratonic root (e.g., Zhu et al., 2011). In this study we have constructed a continuous two-dimensional (2-D) image of both the ILD and LAB in the central and western NCC using teleseismic receiver function (RF) data. Combining this new image with previous seismic observations and geological and geochemical data, we can link the lithospheric structure to the regional tectonics and evaluate the possible role of the ILD during the Mesozoic reactivation and destruction of the craton.

GEOLOGICAL SETTING
The NCC is composed of two major Archean blocks, the eastern and western NCC, that were sutured along the Trans-North China orogen (TNCO, the central NCC) in the Paleoproterozoic (Zhao et al., 2005; Fig. 1). The craton was tectonically stable as a whole for more than 1 by., and had features of typical Archean cratons in the early Paleozoic (e.g., Griffin et al., 1998). However, the craton underwent an episode of severe thermotectonic reactivation and destruction and a dramatic change in lithospheric architecture in the Mesozoic. Available data suggest that...
lithospheric destruction mainly occurred in the eastern NCC, a large part of which is now underlain by a thin (<100 km) and relatively fertile lithosphere (e.g., Griffin et al., 1998; Chen, 2010). However, the present-day lithosphere beneath the central and western NCC is generally thick and stable, with only localized lithospheric thinning and modifications, particularly along two elongated Cenozoic rift systems surrounding the cratonic core represented by the Ordos block (the major part of the western NCC; Fig. 1) (e.g., Zhu et al., 2011). This unique feature leads to strong lateral heterogeneities in the lithosphere beneath the NCC (e.g., Chen, 2010). However, vertical structural variations in the lithosphere of the region and how these were related to the Mesozoic destruction of the craton did not receive much attention previously.

**IMAGES OF LITHOSPHERIC STRUCTURE**

We performed 2-D wave equation-based migrations on both the S and P RF data collected from a broadband seismic array of 64 stations (Fig. 1) for lithospheric structure imaging (for more details about the stations, data, and method, see the GSA Data Repository1). S and P RFs contain information on S to P (Sp) and P to S (Ps) conversions, respectively, from deep velocity discontinuities. A combination of S and P RF imaging can provide more compelling constraints on the discontinuity structures than individual imaging results. According to the data coverage, especially of the S RFs, we chose to perform migration along a profile rotated slightly (~3°) clockwise from the seismic array (Fig. 1). Unlike many RF studies that constrain the lithospheric structure beneath individual stations, the 2-D wave equation-based migration adopted in this study allows us to create a lithospheric image along the entire profile and thus to detect crustal and shallow upper mantle discontinuities in a continuous and coherent manner. Here we mainly focus on the SCLM structure (though the Moho is continuously detected at 40–50 km depth); more detailed analyses and discussion of the imaged crustal structure will be presented elsewhere.

Both the S and P RF-migrated images are dominated by strong negative signals below the Moho (Fig. 2), indicating a complicated SCLM structure beneath the study region. Among these signals, we identified the LAB and an ILD signals based on the consistencies between the images of S and P RFs, of different frequencies and from data and synthetics, and between the results of RF imaging and previous surface wave studies (Fig. 2; Figs. DR2–DR7). We do not intend to interpret other signals that either have no consistent appearance in the S and P RF images or are present near the edges of the study profile where the data coverage is relatively poor (Fig. DR1).

We first identified the LAB along the profile. In the S RF images, a negative signal appears at ~200 km beneath the Ordos region (Fig. 2B; Fig. DR2). While the signal gradually rises eastward to ~165 km depth near the boundary between the eastern and central NCC, it abruptly shallows to <150 km depth under the western border of the Ordos block, and becomes shallower to the west, reaching ~100 km depth within the Qilian orogenic belt. We identify this negative signal as the Sp phase of the LAB based on the following analyses. (1) The signal cannot be high-frequency scattered energies because it is clearly observed in all the S RF images of different frequency contents (Fig. DR2). (2) Synthetic tests based on the actual data coverage suggest that the observed signal resembles the Sp phase of a modeled discontinuity at similar depths (Fig. DR3). (3) A strong negative signal also appears at ~160–200 km depth beneath the Ordos block and TNCO in the P RF images. Comparisons with the synthetic P RF images show that the signal displays a feature similar to a Ps conversion from a discontinuity in that depth range, but not to the artificial image of Moho multiples (Fig. DR4). (4) The depth variations of the signal agree with previous seismic observations using different data and methods (e.g., Fig. DR7; Jiang et al., 2013, and references therein), and mantle xenolith data (e.g., Su et al., 2010) that reveal a thick cratonic keel of ~200 km beneath the Ordos block, and a thin lithosphere of 100–140 km under the orogens immediately to the west and southwest.

In addition to the deep LAB signal, we focus on the most coherent signal at ~80–100 km depth beneath the study region (marked by white dashed line in Fig. 2). This signal exhibits consistent features and depth variations in the S and P RF images, no matter which frequency contents of the data or which bin sizes are adopted in the imaging (Figs. 2C and 2D; Fig. DR5). Moreover, the P RF image amplitude varies significantly with the incident directions of P waves, similar to a Ps-converted phase but differing distinctly from a crustal multiple (Fig. DR6). These features in combination with the imaged deeper LAB (~165–200 km) (Fig. 2A; Figs. DR2–DR4) suggest that the signal at ~80–100 km depth represents a real discontinuity within the SCLM of the Ordos block and TNCO. Obvious velocity reductions were recently imaged by surface-wave tomography at depths of 80–120 km within the high-velocity mantle root in both a 2-D study along the same seismic array (Fig. 2E) and a 3-D study of the region (Fig. DR7). The depths of maximum velocity

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1GSA Data Repository item 2014078, data and methods, and image reliability evaluation, is available online at www.geosociety.org/pubs/ft2014.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.
drop agree with those of the discontinuity found in our RF images, given that the 10–20 km differences between the two are well within the depth uncertainties of tens to one hundred kilometers in surface-wave tomography (Fischer et al., 2010, and references therein) and are comparable to the uncertainties of 3–15 km in RF imaging (see the Data Repository). Indeed, the agreement among all these studies attests to the presence of the ILD beneath the study region.

INTERPRETATION AND DISCUSSION

In order to understand the systematic variations of lithospheric structure across the NCC, the published lithospheric structure image for the Tanlu fault zone area along a similar latitude in the eastern NCC (Fig. 1; Chen et al., 2008) was combined with the new image for the central and western NCC (Fig. 2B). Significant structural heterogeneities are present along the extended image profile. The depth of the LAB varies substantially, from <100 km in the east to ~160–200 km in the central and western NCC, and drops abruptly to <150 km near the boundary between the Ordos block and the Qilian orogenic belt (Fig. 2B). The variation in LAB depth between the two profiles is much less constrained because of the paucity of data (Fig. 1). However, the mantle xenoliths trapped by the Cenozoic basalts in the Hebi (Fig. 1) have high temperatures and contain no garnet, indicating a thin and hot lithosphere (<100 km) beneath the area during the Cenozoic (e.g., Sun et al., 2012). Thus, the geophysical and geochemical data imply a distinct change in the LAB depth near the boundary between the eastern and central NCC. This, together with previous similar observations for areas further north (Fig. 1; Chen, 2010), suggests consistent east-west differences in the lithospheric structure of the craton.

The most distinct feature revealed in this study is the presence of an ILD at ~80–100 km depth beneath the Ordos block and TNCO, well above the imaged LAB of the region but at the same depth level as the LAB in the eastern NCC (Fig. 2B). The negative polarity of this discontinuity suggests a reduction of velocity at ~100 km depth, similar to that revealed both seismically (e.g., Fischer et al., 2010, and references therein) and petrologically (e.g., O’Reilly and Griffin, 2010) beneath many other cratons such as the Dharwar, Kalahari, Slave, western Australia, and eastern North America cratons. This indicates that vertical structural heterogeneities, as well as the known lateral heterogeneities, are common in the cratonic lithosphere.

The nature and origin of the ILD are still uncertain; it may have formed through melting or fluid-induced metasomatism of the lithospheric mantle at an early stage of evolution during Archean time (e.g., Chen et al., 2009), or as a result of long-term multiple stages of refertilization of the SCLM by magmatic events (Savage and Silver, 2008). Another possibility is that the ILD is simply a manifestation of structural (anisotropic) layering in thick cratonic lithosphere, with the layers above and below formed separately and probably under distinct tectonic regimes (Yuan and Romanowicz, 2010). All these interpretations attribute the ILD to vertical variations in composition or fabric, rather than in temperature, within the SCLM of cratons. The formation of such an ILD, whether at an early stage or during the long-term evolution of cratons, would have been related to major tectonic and/or magmatic events. In the case of the NCC, the ILD observed beneath the Ordos block and TNCO may have come into being during the Paleoproterozoic assembly of the craton, the most significant tectonic event in the region that led to the final cratonization of the NCC (Zhai and Liu, 2003).

Considering the cratonic nature of the NCC as a whole from the Paleoproterozoic to Paleozoic, and the prevalence of the ILD in cratonic regions, we suggest that a similar ILD may have existed in the SCLM of the eastern NCC before its Mesozoic destruction. This is supported by the evidence of pre-Mesozoic to early Mesozoic alkaline magmatism (e.g., syenites) sporadically distributed in the eastern NCC. These alkaline rocks are enriched in large ion lithophile elements (i.e., Ba, Th, U, and Sr) and light rare earth elements and depleted in high field strength elements (Nb, Ta, Ti, and P) with enriched Sr-Nd-Hf isotopic compositions, suggesting that they were derived from a reenriched, refractory lithospheric mantle (e.g., Yang et al., 2012). Metasomatically enriched components commonly appear at ~100 km depth in the SCLM (e.g., O’Reilly and Griffin, 2010), and pronounced metasomatism or an ILD probably was present at ~100 km depth within the SCLM of the eastern NCC before its destruction in the Mesozoic.

The presence of the ILD would indicate the existence of a mechanically weak layer in the SCLM of stable cratons. Analogous to the case of vertically extended weak zones, the intrinsic weakness could have made such a layer susceptible to intense heating, strain localization, and chemical modifications under certain circumstances during the successive evolution of cratons.

In the Mesozoic destruction of the eastern NCC, lithospheric reactivation and modification may have preferentially started at both horizontal and vertical weak zones in the region. The reactivation of an intralithospheric horizontal zone probably was facilitated by geological processes at plate margins, such as the subduction of the Pacific plate beneath eastern Asia in the Mesozoic, rather than by processes below the lithosphere. In this scenario, the ILD and the associated horizontal weak layer played a dynamic role, essentially similar to the weak tectonic belts that exemplify the lateral heterogeneities of the lithosphere (Chen, 2010). The physical and chemical modifications at the depths of the ILD, such as rehydration, viscosity reduction, and partial melting associated with the input of water sourced from subduction (Fig. 3), would have accelerated gravitational instability and triggered delamination of the lower part of the lithospheric mantle. This, together with thermo-mechanical and chemical erosion at the base of the lithosphere, may eventually have led to the destruction of the root beneath the eastern NCC.

We emphasize that, because of the presence of the weak layer, the lithospheric destruction in the eastern NCC could be hastened by modifications occurring simultaneously at the base and the middle of the lithosphere (Fig. 3). The broad change of Nd isotopic compositions of mafic magmas from enriched to depleted mantle sources at the time (ca. 120 Ma) of strong cratonic destruction (Xu, 2001) may be related to such extensive lithospheric modifications. This process might have involved partial melting both within and at the bottom of the lithosphere (Fig. 3). The present-day LAB beneath the region probably formed following the preexisting ILD, with modifications by the hot upwelling asthenospheric mantle after the loss of the lower part of the lithosphere.

In contrast, both the ILD and the LAB beneath a large part of the Ordos block and TNCO appear to have been preserved, as the influence of the Pacific subduction faded to the west (e.g., Zhu et al., 2011). The lithosphere in this region may have also been modified, as indicated by the formation of elongated rifting systems, eruptions of basaltic magmas,
and melt- and/or fluid-induced mantle refertilization during the Cenozoic (e.g., Tang et al., 2013), as well as the presence of slow and relatively thin present-day lithosphere (e.g., Jiang et al., 2013). However, all of these effects were mainly confined to preexisting weak zones surrounding the cratonic nucleus of the Ordos block (Chen, 2010). Seismic images from this and previous studies and available geological and geochemical data therefore suggest that, compared to the fundamental destruction of the eastern NCC, the lithospheric modification in the central and western NCC was much more localized and is probably an ongoing process.

The distinct east-west differences in the lithospheric structure and evolution of the NCC have implications for stabilization and destabilization of cratons. The destruction of cratonic lithosphere in the eastern NCC is not unique, although it is probably the most pronounced example; there are analogues elsewhere, such as in the Mojave and Wyoming cratons in North America (e.g., Carlson et al., 2005) and the western Brazilian craton in South America (Beck and Zandt, 2002), where the cratonic lithosphere has also been strongly modified or destroyed. The lithospheric processes beneath these cratons all appear to be closely related to oceanic plate subduction at continental margins that may have triggered reactivation of both horizontal and vertical weak zones in the lithosphere and escalated the craton destruction (Zhu et al., 2011). On the other hand, the major part of the central and western NCC and many other cratonic regions where the ILD and thick mantle root are preserved seem to have remained stable over long periods of geological time. This may indicate that the presence of a weak layer within the SCLM has not imposed notable effects on the tectonic stability and evolution of the cratons, probably because the dominant influences on these cratons have been from below the lithosphere rather than from plate margins, and the stronger lower part of the craton root can effectively resist such influences.

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REFERENCES CITED


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Data and imaging method

Teleseismic waveform data used in this study came from a linear seismic array deployed by the Institute of Geology and Geophysics, Chinese Academy of Sciences under the “Destruction of the North China Craton (DNCC)” project, a multidisciplinary scientific project launched by the NSF of China in 2008. The seismic array ran roughly E-W from around the boundary between the eastern NCC and the TNCO, crossed the Ordos block and ended at the Qilian orogenic belt which borders the NCC to the southwest (Fig. 1), with a total length of ~1000 km. It consists of 64 broadband stations with an average intra-station distance of ~15 km. Each station operated for a period of 10-18 months from December 2006 to May 2008.

In this study we investigated the lithospheric structure of the central and western NCC using the S and P receiver function (RF) data from this array. S and P RFs contain information of S to P (Sp) and P to S (Ps) conversions, respectively, from deep velocity discontinuities. P RF imaging has proven to be efficient in recovering
complex structural features of the Moho and deep mantle discontinuities such as those in the mantle transition zone (e.g., Zhu and Kanamori, 2000; Ai et al., 2007; Nábelek et al., 2009). However, it is difficult to produce a clear shallow upper-mantle image in many cases due to the significant interference of the crustal multiples with the Ps signals (e.g., Wittlinger and Farra, 2007; Chen, 2009). In contrast to P RFs, S RFs are free from inferences of crustal multiples and have been widely used to detect discontinuities in the shallow upper mantle, especially the LAB at local to regional scales (e.g., Li et al., 2004; Savage and Silver, 2008; Ford et al., 2010). Recent studies further suggest that a combination of S and P RF imaging can provide more compelling constraints on the lithospheric structure than individual imaging results (e.g., Rychert et al., 2007; Chen, 2009).

We constructed the S and P RFs for the 64 stations in the same way as we did previously (Chen et al., 2006; 2008; 2009; Chen, 2009). Teleseismic data from 193 events with epicentral distances 55°-85° (Fig. DR1a) and from 536 events with epicentral distances 28°-92° (Fig. DR1d) were selected respectively, resulting in a total of 1250 S RFs and 10700 P RFs with high signal-to-noise ratios for lithospheric structural imaging. The polarity and the time axis of the S RFs were reversed to make them directly comparable with the P RFs. Arrival times of both Sp conversion phases relative to the direct S waves and Ps phases relative to the direct P waves, and the corresponding piercing points of the conversion at depth, were calculated using the average 1-D velocity model for eastern China as we did previously. All the S and P RFs were also moveout-corrected to the case of vertical incidence (horizontal
slowness $p = 0$) as required for wave equation-based depth migration (Chen et al., 2005; 2008). We then applied the 2-D wave equation-based poststack migration method for Sp (Chen et al., 2008) and Ps phases (Chen et al., 2005) to the resultant S and P RFs, respectively, to image the lithospheric structure along profile A-A’. The migration for both phases consists of two basic procedures: common conversion point (CCP) stacking and backward wavefield extrapolation (Chen et al., 2005; 2008). The minimum S RF number in each bin in CCP stacking was set to be 50 for effective noise suppression. To make the migrated P and S RF images have compatible lateral resolutions, much larger numbers (> 300) but similar bin sizes were adopted for CCP stacking of P RFs. Figs. DR1b and DR1e give the numbers of S and P RFs in the bins (blue traces) and bin widths (red traces) used in the CCP stacking at 100 and 200 km depths, and Figs. DR1c and DR1f show the corresponding stacked sections of S and P RFs with frequencies 0.03 – 0.5 Hz. Besides the coherent Moho phases at ~5 s, the most pronounced common feature in the two stacked sections is the dominance of strong negative signals in the time range of ~8-12 s. To convert the time-domain stacked sections to depth images, backward wavefield extrapolation was then conducted using a frequency-space domain phase screen propagator (Chen et al., 2005). During this second step of migration, the same 1-D velocity model used in calculation of delay times and piercing point locations was adopted and different ranges of frequency contents of data were superposed to construct the images (Figs. 2, DR2-DR6).
Evaluation on the reliability of the lithospheric image

Both S and P RF images would be affected by many factors, such as, the spatial data coverage, frequency contents of the data, noise, and lateral seismic heterogeneities. We evaluated the robustness of the lithospheric structural image, especially the intra-lithospheric discontinuity (ILD) and lithosphere-asthenosphere boundary (LAB) identified beneath the Ordos block and TNCO, by frequency analysis and detailed comparisons between the S and P RF migrated images, data images and synthetic modeling results based on the actual data coverage, as well as between results of RF imaging and previous surface wave studies (Figs. DR2-DR7).

We calculated the synthetic seismograms by a 2-D hybrid method (Wen and Helmberger, 1998; Chen et al., 2005) that naturally takes into account the effect of lateral heterogeneities on the waveforms. Given that the dominant periods of our S and P RFs peak at ~4 s and 1 s, respectively, we adopted 4 s and 1 s individually in the calculation of these two types of data. We constructed the synthetic S and P RFs, picked them according to the coverage and density of the real data, and performed wave equation poststack migration to construct the corresponding migrated images in the same way as for the real data. The depths of the Moho and the LAB in the synthetic models were designed to resemble those of the imaged real structure. The sharpness of the two discontinuities was set to be uniform over the whole transverse range. We considered a first order Moho discontinuity with the same impedance contrasts as the IASPI91 global model and a 20-km gradient zone with a velocity drop of 4% for the LAB in the synthetic modeling. In some cases, an ILD with a 4%
velocity drop over a 10-km depth range at depths similar to the ILD identified in the data images was also incorporated into the synthetic models. The corresponding synthetic images were compared with those without the ILD and the data images to investigate the effects of the ILD to the migrated images, especially of the LAB (e.g., Fig. DR3).

Our ILD and LAB images appear robust based on the results of frequency analysis and images comparisons: 1) The numbers of the S RFs (mostly ≥ 50) and P RFs (mostly ≥ 300) were sufficiently large in each CCP stacking bin, which allow for good noise suppression and signal enhancement; 2) The Sp signals of both discontinuities were consistently imaged using different frequencies of the data (Fig. DR2, except for the image with frequencies lower than 0.2 Hz in which the side lobe of the Moho Sp conversion phase significantly interferes with the Sp phase of the ILD, making it an ambiguous feature); 3) The features of the S and P RF images were consistent with each other and agreed with synthetic modeling results (Figs. DR3–DR5); 4) The nature of the ILD Ps signal was further verified by the close resemblance of its image feature to a true Ps phase but large differences from a crustal multiples when different incident directions of data were considered in imaging (Fig. DR6); and 5) The depth distributions of the ILD and/or the LAB agree broadly with previous seismic observations on the lithospheric structure of the study region from both 2-D surface wave tomography along the same seismic array (Fig. 2E; Wei, Z.G. et al., unpublished result) and other surface wave studies of various scales (Fig. DR7, Jiang et al., 2013; Huang et al., 2009; Bao et al., 2011) as well as body wave
We estimated the depths of the ILD and the LAB mainly from the highest-frequency S RF image (Fig. 2B, Fig. DR2d) because of it relatively high resolution and little influence from crustal multiples compared to the P RF images. For the ILD, the P RF image of same frequency contents (Fig. 2D) was also considered, since the Ps phase of the ILD is identified without obvious interference from crustal multiples. We mainly considered the portions of the profiles where the numbers of the S RFs are sufficient (50 and above) in the individual CCP stacking bins. To investigate the uncertainties in the depths of the two discontinuities, we adopted different velocity models in imaging, including that incorporating the lateral variations in the crustal and uppermost mantle structure along the profile (e.g., Zhu and Zheng, 2009; Wei et al., 2011; Jiang et al., 2013). The images thus obtained are very similar to those presented here (Figs. 2B, Fig. DR2). We found that the depth uncertainties of the imaged ILD and LAB due to different velocity models adopted in migration are mostly 3~8 km and 5~15 km, respectively, which are comparable to what can be resolved with the real S RF data (~10 km). These values are analogues to the results of our previous studies (Chen et al., 2008; 2009; Chen, 2009) and also to those estimated by other authors for shallow upper mantle discontinuities (e.g., Li et al., 2007; Rychert and Shearer, 2009; Ford et al., 2010; Lekic et al., 2011).

References

Ai Y., Chen Q., Zeng F., Hong X., Ye W., 2007. The crust and upper mantle structure


Fig. DR1 Distribution of teleseismic events used (a, d), numbers of RFs in the bins (blue traces) and bin widths (red traces) used in the CCP stacking at 100 and 200 km depths (b, e) and stacked RF sections (c, f) with frequencies 0.03 – 0.5 Hz after moveout corrected to p = 0. (a, b, c) for S RFs and (d, e, f) for P RFs. The Moho conversion phases (Sp in S RFs and Ps in P RFs) and strong multiple (PpPs in P RFs) are marked by arrows in (c, f).
Fig. DR2 Migrated S RF images for profile A-A’ obtained by using different frequency contributions of the data: (a) 0.03 – 0.2 Hz; (b) 0.03 – 0.28 Hz; (c) 0.03–0.35 Hz; (d) 0.03 – 0.5 Hz. The ILD and the LAB are marked as white and black dashed lines, respectively.
Fig. DR3 Migrated S RF images of different frequency contents for two synthetic models (a-d) and from the real data for profile A-A’ (e, f). (a, c, e) 0.03 – 0.2 Hz; (b, d, f) 0.03 – 0.35 Hz. The synthetic models contain a Moho and a LAB, with (c, d) or without (a, b) an ILD built in. All the discontinuities in the synthetic models were designed to resemble the imaged real structure, and the synthetic images were constructed based on the actual data coverage. Both the modeled and observed LABs are marked as black dashed lines. The modeled Moho and ILD and the observed ILD
are marked as white dashed lines. Although the amplitudes show some changes, the overall feature of the synthetic images, especially that of the LAB is essentially the same no matter an ILD was built into the model or not. The deformed and/or intermittent LAB image in the left part of both the synthetic and data images might be a result of limited data coverage (Fig. DR1b).
Fig. DR4 Migrated P RF images with frequencies 0.03 – 0.35 Hz for three synthetic models (a-c) and from the real data for profile A-A’ (d). The synthetic models contain no ILD and no LAB (a), no ILD but a LAB (b), or both an ILD and a LAB (c), respectively. The synthetic models in (b) and (c) are the same as those used in Figs. DR3a-DR3b and DR3c-DR3d, respectively. The Moho and the ILD in the synthetic models are marked as white dashed lines and the LAB as black dashed lines. Artifacts induced by the positive PpPs and negative PsPs + PpSs multiples of the Moho are marked in (a). Without a LAB built in the synthetic model (a), the negative multiples of the Moho are much weaker and parallel to slightly steeper than the positive multiple of the Moho. In the data image (d), the negative signal at 150-200 km depths appears equally or even stronger and has a smaller dipping angle, especially in the middle part of the profile, than the above positive Moho multiple artifact. These
image features bear considerable resemblance to the synthetic images with a LAB at the similar depths (b, c). The presence of an ILD has no significant effects on the image feature of the modeled LAB, although it affects the amplitudes of the LAB Ps phase and other signals, especially in the left part of the image (compare b and c).
Fig. DR5 Migrated S RF images (a, c, e) and P RF images (b, d, f) for profile A-A’. Different frequency contributions of the data and stacking bin widths (lateral smoothing) were considered in RF imaging: (a, b) 0.03 – 0.35 Hz with 100-km bin width at 100 km depth; (c, d) 0.03 – 0.4 Hz and (e, f) 0.03 – 0.5 Hz with 60-km bin width at 100 km depth. White dashed Line in each panel marks the ILD identified in the RF images. (e, f) are same as Fig. 2C and 2D.
Fig. DR6 Migrated P RF images from two subsets of the real data in non-overlapping back azimuth ranges: 180° – 360° (a, c) and 0 – 180° (b, d). All the 3061 P RFs in the first subset and randomly selected 3061 out of 7639 P RFs in the second subset were used to construct the images. Different frequency contributions of the data were considered in imaging: (a, b) 0.03 – 0.5 Hz; (c, d) 0.03 – 0.7 Hz. Gray dashed lines are same as the white dashed lines in Fig. DR5. Comparisons of the images provide further constraints on the nature of the two negative signals observed at ~80 -120 km depths in Fig. 2D, Figs. DR5f and DR5d. For seismic rays coming from east (back azimuths within 0° - 180°), the eastward dipping part of the shallower signal is clearly identified, but the similarly eastward dipping signal below it appear much weaker, even invisible in the image (b, d). Using the subset of data from west (180° - 360°), on
the other hand, the former weakens, while the latter becomes more predominant (a, c). The image strength of the westward dipping part of the shallow signal changes with incident direction of P waves in a similar way to its eastward dipping counterpart. Moreover, this westward dipping part was imaged at different locations using the two subsets of data, being more to the east with the data coming from east (compare b, d with a, c). All these image features of the shallow signal are consistent with the theoretical predictions for Ps phases and the previous synthetic and data images for dipping discontinuities (Chen et al., 2005; 2006). The image of the deeper signal appears to be affected by the coming direction of data in an opposite way, resembling the artifacts induced by crustal multiples but bearing large differences from the modeled Ps images (Chen et al., 2006). Using all the P RFs or other groups of randomly selected 3061 P RFs in the second subset results in very similar images to those shown in (b, d). These image comparisons combined with the consistent depths in the S and P RF images (e.g., Figs. 2C and 2D) suggest that the shallower negative signal likely represents a real structure in the upper mantle, whereas the deeper signal may be an artifact induced by crustal multiples.
Fig. DR7 Shear-wave velocities along four E-W profiles crossing the TNCO and the Ordos block (latitudes 36°N, 37°N, 38°N, 39°N) and vertical velocity gradients along the 36°N profile obtained by 3-D regional surface wave tomography (Jiang et al., 2013). A relatively low-velocity layer was observed at 80-150 km depth within the high-velocity mantle root along all the profiles, with the depths of the maximum velocity reductions marked by the dashed curves.