Local modification of the lithosphere beneath the central and western North China Craton: 3-D constraints from Rayleigh wave tomography

Mingming Jiang a,⁎, Yinshuang Ai a, Ling Chen b, Yingjie Yang c

a Key Laboratory of the Earth's Deep Interior, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing 100029, China
b State Key Laboratory of Lithospheric Evolution, Institute of Geology and Geophysics, Chinese Academy of Sciences, Beijing 100029, China
c ARC Centre of Excellence for Core to Crust Fluid Systems/GEMOC, Dept. Earth and Planetary Sciences, Macquarie University, North Ryde, NSW 2109, Australia

Abstract

We have imaged the lithospheric structure beneath the central and western North China Craton (NCC) with Rayleigh wave tomography. The Rayleigh waveforms of 100 teleseismic events recorded by 208 broadband stations are used to yield high-resolution phase velocity maps at 13 periods from 20 s to 143 s. A 3-D S-wave velocity model is constructed based on the phase velocity maps. Our S-wave velocity model is broadly consistent with the results of previous tomography studies, but shows more detailed variations within the lithosphere. The Trans-North China Orogen (TNCO) is generally characterized by low-velocity anomalies but exhibits great heterogeneity. Two major low-velocity zones (LVZs) are observed in the north and south, respectively. The northern LVZ laterally coincides with sites of Cenozoic magmatism and extends to depths greater than 200 km. We propose that a small-scale mantle upwelling is present, confined to the north of the TNCO. A high-velocity patch in the uppermost mantle is also observed between the two LVZs adjacent to the narrow transtensional zone of the Cenozoic Shanxi–Shanxi Rift (SSR). We interpret this as the remnant of a cratonic mantle root. The Ordos Block in the western NCC is associated with high-velocity anomalies, similarly reflecting the existence of cratonic mantle root, but a discernible low-velocity layer is observed at depths of 100–150 km in this location. We interpret that this mid-lithospheric structure was probably formed by metasomatic processes during the early formation of the NCC. Based on the observations from our S-wave velocity model, we conclude that the current highly heterogeneous lithospheric structure beneath the TNCO is the result of multiphase reworking of pre-existing mechanically weak zones since the amalgamation of the craton. The latest Cenozoic lithospheric reworking is dominated by the far-field effects of both Pacific plate subduction and the India–Eurasia collision.

© 2012 International Association for Gondwana Research. Published by Elsevier B.V. All rights reserved.

1. Introduction

The North China Craton (NCC) is traditionally divided into two major Archean blocks, namely the Eastern Block (EB) and the Western Block (WB), separated by a Paleo-Proterozoic orogen called the Trans-North China Orogen (TNCO) or the Central Orogenic belt (Fig. 1) (Zhao et al., 2001; Kusky and Li, 2003; Zhao et al., 2005; Santosh, 2010). The whole craton was tectonically stable until the Middle Ordovician (e.g., Menzies et al., 1993; Griffin et al., 1998; Gao et al., 2002). From the Ordovician to Cenozoic, the cold, thick and refractory lithosphere of the EB was intensively reactivated and destructed, and has been replaced by a hot, thin and fertile lithosphere (e.g., Fan and Menzies, 1992; Menzies et al., 1993; Griffin et al., 1998; Fan et al., 2000; Xu, 2001; Wu et al., 2005; Chen et al., 2006; Menzies et al., 2007; Chen et al., 2008; Wu et al., 2008; Zhang et al., 2011; Zhang, 2012). In contrast, the central and western NCC remains stable and exhibits sparse magmatism, relatively low heat flow (J.Y. Wang et al., 1996; Hu et al., 2000), and generally thick crust (Ma, 1989; Li et al., 2006; Chen et al., 2010; Wei et al., 2011) and lithosphere (Chen et al., 1991; Zhu et al., 2002). These remarkable tectonic contrasts make the NCC unique amongst cratons worldwide, and ideal for studying the stabilization and destruction of old cratons.

Previous research interest in the NCC has primarily focused on the lithospheric architecture of the EB due to the abundant petrological and geochronological data derived from its widespread suites of xenoliths and xenocrysts (e.g., Menzies et al., 1993; Griffin et al., 1998; Fan et al., 2000; Xu, 2001; Wu et al., 2005; Menzies et al., 2007; Zhang et al., 2011; Zhang, 2012). Subsequent geophysical investigations (e.g., Chen et al., 2006; Zhao et al., 2007; Chen et al., 2008; Zheng et al., 2008a,b; Chen, 2009; Li et al., 2009) have provided important constrains on the spatial extent of lithospheric thinning and destruction. In combination with multidisciplinary observations, several model, including the deep subduction of the Paleo-Pacific plate (Griffin et al., 1998; Wu et al., 2003; Sun et al., 2007), collision of an amalgamated...
North China–Mongolian plate with the Siberian plate (Davis et al., 2001), enhanced mantle temperatures associated with plumes (Deng et al., 2004), the India–Eurasia collision (Menzies et al., 1993) and the North–South China collision (Yin and Nie, 1993; Zhang, 2007), have been proposed to interpret the dynamic processes and mechanisms involved in lithospheric destruction of the EB. However, these models have difficulty in interpreting the contrasting responses of the three NCC domains to the lithospheric reactivation processes from the Late Mesozoic to Cenozoic. How the lithosphere of the central and western NCC evolved in response to this reactivation or other tectonic events.
remains controversial. Detailed images of the lithosphere can provide key insights into this issue. Large-scale seismic tomography results show that high-velocity anomalies extend to deeper than 200 km beneath the central and western NCC, suggesting the presence of a thick cratonic mantle root in this region (e.g., Lebedev and Nolet, 2003; Huang and Zhao, 2006; Li and van de Hilst, 2010). As more and more geological and geophysical investigations have been conducted in the central and western NCC, great heterogeneities in the lithosphere are revealed. For instance, a thinned lithosphere is observed in the north of the TNCO by magnetotelluric soundings (Wei et al., 2008). High-resolution images derived from dense seismic-array data indicate that the dramatically thinned lithosphere is mainly present beneath the TNCO and the Inner Mongolian Suture Zone (IMSZ), and the thick cratonic mantle root is preserved beneath the Ordos Block (L. Chen et al., 2009; Chen, 2010; Bao et al., 2011). In addition, recent petrological and geochemical studies in the TNCO provide direct evidences that the underlying lithospheric mantle is Archean or Paleoproterozoic (e.g., Xu et al., 2004a; Tang et al., 2006; Wang et al., 2006; Xu, 2007; Xu et al., 2008; Ying et al., 2010). These heterogeneities are attributed to the localized modification and thinning of the lithosphere (Chen, 2010; Xu et al., 2010; Zhu et al., 2011). However, most of these existing studies are confined to sparse spots, discrete areas or linear profiles, and hence are difficult to yield integrated lithospheric images for the entire central and western NCC. A detailed 3-D structural model is needed for delineating the spatial extent of the lithospheric thinning and to understand the geodynamic processes driving localized lithospheric modification in this region.

With the dramatic increase of seismic data from both dense portable arrays and permanent networks, several regional high-resolution body wave tomography studies have been conducted within the tectonic blocks of the NCC and adjacent regions (e.g., Tian et al., 2009; Zhao et al., 2009, 2012). Zhao et al. (2009) observed a plume-like low-velocity anomaly in the upper mantle beneath the TNCO from images of finite frequency body wave tomography. However, their results cannot constrain the structure of this hot upwelling in the shallow part of the upper mantle because of the limited vertical resolution of body wave tomography above 200 km depth. Surface wave tomography, whose dominant resolution is confined to the depth interval of ca. 20–200 km, is suitable to complement the body wave tomography. However, previous results of surface wave tomography in the NCC (Huang et al., 2009) failed to generate high-resolution models required to identify small-scale features such as the plume-like anomaly, largely due to the limitations of the data and the tomography method used.

In this study, we construct a high-resolution structural model for the lithosphere of the central and western NCC using a novel surface wave tomography technique. Using the fundamental mode Rayleigh wave data from both a permanent network and a dense portable array, we obtain 2-D high-resolution phase velocity maps. Combined with the Moho map derived from existing studies of receiver functions (T. Zheng et al., 2009; Wei et al., 2011), we invert these data to produce a new high-resolution 3-D S-wave velocity model of the lithosphere. On the basis of the S-wave velocity model, we discuss the geodynamic processes and mechanisms controlling lithospheric modification in the central and western NCC.

2. Geological setting

The Precambrian history of the formation and stabilization of the NCC involves three major tectonic events: (1) a major phase of continental collision at ca. 2.7 Ga; (2) the amalgamation of micro-blocks and cratonization at ca. 2.5 Ga; and (3) Paleoproterozoic rifting–subduction–accretion–collision tectonics and subsequent high-grade granulite facies metamorphism and granitoid magmatism between ca. 2.0–1.82 Ga (Zhai and Santosh, 2011). Based on integrated studies of lithology, structure, geochronology, and metamorphic P–T–t paths, the Paleoproterozoic amalgamation of the NCC includes two episodes of continental collision: the collision of the EB and WB along the TNCO, and the collision of the Ordos Block with the Yinshan Block from the north along the E–W striking Inner Mongolia Suture zone (IMSZ) (Zhao et al., 2001; Kusky and Li, 2003; Zhao et al., 2005; Kusky et al., 2007a; Kusky, 2011; Santosh et al., 2011a,b; Tsunogae et al., 2011). These Archean–Paleoproterozoic blocks make up the present tectonic framework of the NCC (Fig. 1). The timing of the collision between the EB and WB and the polarity of subduction are strongly debated (Zhao et al., 2001; Kusky and Li, 2003; Wilde and Zhao, 2005; Zhao et al., 2005; Kusky et al., 2007a,b; Zhang et al., 2007; J. Zhang et al., 2009; Kusky, 2011). To answer this debate, Santosh (2010) attempted to integrate the geologic, geochronologic and geophysical data across the major zones of amalgamation of the continental blocks in the NCC, and proposed a unified tectonic model to account for the Paleoproterozoic evolution of this region.

The present boundaries of the NCC (Fig. 1) reflect several major tectonic events in Phanerozoic. The Qilianshan Orogenic Belt bounds the craton to the west, which was folded by multiphase subduction and accretion in the Early Paleozoic (e.g., Xiao et al., 2002; Pan et al., 2006; Xiao et al., 2009, 2010; Glorie et al., 2011; Rojas-Agramonte et al., 2011). In the north, the NCC lies adjacent to the Central Asian Orogenic Belt (CAOB) along the Solonker suture, which is associated with the closure of the Paleosian Ocean at the end of the Permian (Sengör et al., 1993; Davis et al., 2001; Xiao et al., 2003). The Triassic north–south collision between the NCC and the Yangtze Craton formed its southern and eastern boundary as the Qining–Dabie–Sulu Orogenic Belt (Li et al., 1993).

Two tectonic events are interpreted as major causes of the lithospheric reactivation and modification of the NCC from the Late Mesozoic to Cenozoic. In the Middle Mesozoic, the westward oblique subduction of a series of oceanic plates, including the Izangi, Kula and Pacific, initiated along the eastern margin of Eurasia (Maruyama et al., 1997; Bartolini and Larson, 2001; Smith, 2007; Itozaki et al., 2010). The deep subduction and stagnancy in the mantle transition zone of these oceanic plates are interpreted as the primary mechanisms for lithospheric reactivation and modification (Ye et al., 1987; Wu et al., 2003; Sun et al., 2007). The eastward forcing and extrusion of the Cenozoic India–Eurasia convergence are another possible cause (Molnar and Tapponnier, 1975; Menzies et al., 1993; M. Liu et al., 2004). At present, the Tibetan Plateau has directly contacted with the southwestern margin of the NCC, producing compressional intraplate tectonics in the adjacent region (Zhang, 1989).

The central and western NCC, which includes the TNCO and WB, is generally termed the Loess Plateau. It reaches altitudes of 500–3500 m, in contrast to the lowland topography of the EB. Combined with sharp variations in the gravity field, thickness of crust and lithosphere, and heat flow between these two regions, a remarkable tectonic boundary can be defined coinciding with the North–South Gravity Lineament (NSGL) (Fig. 1). The WB is further divided into the Ordos Block in the south and the Yinshan Block along its northern margin. The IMSZ marks the collisional suture between these two blocks as the NCC was incorporated within the Columbia supercontinent during Paleoproterozoic (Kusky and Li, 2003; Zhao et al., 2005; Santosh et al., 2006, 2007; Santosh, 2010; Santosh et al., 2011a,b). The Ordos Block is the stable cratonic nucleus of the NCC with very low heat flow, little seismicity, and no internal deformation (Zhai and Liu, 2003; Kusky et al., 2007b). Surrounding the Ordos Block, the S-shaped Shanxi–Shanxi Rift (SSR) to the east and southeast, and the arc-shaped Yinchuan–Hetao Rift (YHR) to the northwest (Fig. 1) represent the products of Cenozoic rifting (Zhang et al., 1998). A remarkable feature of the SSR is its S-shaped geometry, with two broad extensional domains in the north and south and a narrow transtensional zone in the middle. Extension in the TNCO evolved episodically from the Late Mesozoic to Cenozoic (Zhang et al., 2003), together with uplift of the Taishan Mountains (Liu et al., 2000) and widespread magmatism and intrusions. Cenozoic xenolith-bearing basalts are found at Hannueba, Wutai, Jining, Datong and Fansi in the
north, and Fushan and Hebi in the south (Fig. 1) [e.g., Zheng et al., 2001; Tang et al., 2008; Xu et al., 2008; Liu et al., 2010; Xu et al., 2010; Zhang et al., 2012].

3. Fundamental mode Rayleigh wave data

Rayleigh wave data used in this study were collected from both the China National Seismic Network (CNSN) [Zheng et al., 2010] and the North China Interior Structure Project (NCISP). A total of 144 broadband stations from the CNSN in the central and western NCC are selected as the 2-D backbone array (Fig. 1). The CNSN stations are equipped with several types of broadband to ultra-broadband seismometers. The seventh linear sub-array of NCISP (NCISP-VII) is deployed from the CAOB in the north to the Yangtze Craton in the south running through the eastern edge of the WB and most of the TNCO (Fig. 1). It contains 64 broadband stations equipped with either Guralp GMG-3ESP or 3T seismometers (50 Hz to 30 s, 60 s or 120 s). The average station interval is ca. 10 km. The high-density coverage of seismic stations introduced by the NCISP-VII array greatly improves the lateral resolution of the lithospheric structure beneath the TNCO.

The fundamental mode Rayleigh waveforms were extracted from teleseismic events occurring from June 2008 to August 2009, the operation period of the NCISP-VII array. Firstly, we first selected 84 events that satisfied three criteria: (1) Ms ≥ 6.0, where Ms is the surface wave magnitude; (2) 30° ≤ δ ≤ 120°, where δ is the epicentral distance from the center of the study area; and (3) D < 100 km, where D is the focal depth. A considerable number of these events are located on the western margins of the Pacific and Philippine plates, in the southeast of the study area. Back azimuthal gaps occur for both east and northwest directions. To fill the azimuthal gaps, we added 16 more events to the dataset by slightly loosening the criteria for event selection. The final dataset provides good azimuthal coverage (Fig. 2), although it is still not entirely evenly distributed. With the large number of stations, these teleseismic events generate very dense raypaths in the study area.

We follow the data processing procedures first outlined in Forsyth and Li (2005) and later modified by Jiang et al. (2011). The instrument responses of all seismometer types are first normalized to the standard type: the Streckerisen STS-2. The vertical component seismograms are then filtered with a series of 10 mHz-wide, four-pole, double-pass Butterworth filters at 13 frequency bands with central frequencies ranging from 7 to 50 mHz (20 to 143 s for periods). The fundamental mode Rayleigh waveforms are visible on the filtered seismograms. We pick out the Rayleigh waveforms with good signal-to-noise ratios and strong coherence from station to station for each event. The selected waveforms are isolated by applying boxcar time windows with cosine tapers at both ends. The widths of the time windows are determined according to the duration of the Rayleigh waveforms. In addition, the waveform data resulting in large misfits in the tomography for the 2-D phase velocity maps are further removed. The remaining Rayleigh waveforms reach a maximum of ca. 13,000 pieces at 50 s period, and slightly diminish towards both longer and shorter period ends (Fig. 3). The filtered and windowed waveforms are transformed to the frequency domain to obtain the amplitudes and phases, which are two classes of independent input data directly used in the tomography. Finally, before tomography is performed, the amplitudes are corrected for geometrically spherical spreading of surface waves, as well as anelastic attenuation using the attenuation coefficients for the stable continent regions of Mitchell (1995).

4. Methods

We adopt the multiple-plane-wave tomography technique (MPWT) (Yang et al., 2008) for inversions of the phase velocities, which is a revised version of the two-plane-wave tomography technique (TPWT) specified for large-scale regional tomography. To account for the non-planar wave energy caused by surface wave scattering and multi-pathing along the propagation path, the TPWT simulates the incoming wavefield as the interference of two plane waves and inverts for the wavefield parameters and phase velocities simultaneously (Li et al., 2003; Forsyth and Li, 2005). The TPWT is further improved by using 2-D sensitivity kernels (Yang and Forsyth, 2006) instead of the original Gaussian sensitivity functions to represent the sensitivities of amplitudes and phases of surface waves to structural heterogeneities. An example of these sensitivity kernels with the smoothing length of 65 km is shown in Fig. 4. The single two-plane-wave assumption tends to fail when applied to the large-scale region of the NCC, with an aperture close to 1000 km. To solve this problem, we divide the study area into four sub-regions based on size constraints and the distribution of seismic stations (Fig. 5), and then fit the wavefield in each sub-region by a pair of plane waves following the MPWT (Yang et al., 2008).

We parameterize the study area using a 2-D rectangular grid with a node spacing of 0.5° in both longitude and latitude. Two additional outermost loops of grid nodes are added and less damped when solving for the phase velocities in order to absorb the effects of very complex wavefields that are poorly fitted by eight plane waves. In the inversion, regularization is applied to damp phase velocities by applying a priori short deviation of 0.25 km/s, and to laterally smooth phase velocities by applying a Gaussian smoothing length of 65 km. We first invert for the average phase velocity at each period. The resulting average phase velocities are used to calculate the 2-D frequency-sensitivity kernels, and are taken as the initial values in the inversions for 2-D phase velocity maps. We perform the inversions for 2-D phase velocity maps in two steps. In the first step, Rayleigh wave data with large misfits are eliminated from the dataset. In the second step, the retained dataset is then inverted again to produce the final phase velocity maps.

After obtaining the 2-D phase velocity maps at periods from 20 to 143 s, we extract a local phase velocity dispersion curve at each node from these maps. We then invert each local phase velocity dispersion curve for a 1-D S-wave velocity model. Finally, all of the 1-D S-wave velocity models are integrated to construct a 3-D S-wave velocity model. The 1-D inversion is carried out with the generalized non-linear least-squares algorithm (Tarantola and Valette, 1982; Li and Burke, 2006). The forward calculation of Rayleigh wave dispersion curve is based on the method of Saito (1988). In the inversion, the 1-D S-wave velocity structure is parameterized into a layered model. There are 17 layers with thicknesses of 20–25 km from the Earth’s surface to the bottom of upper mantle at 410 km depth. The structure deeper than 410 km is fixed to the AK135 global model (Kennett et al., 1995). Each layer contains one model parameter (i.e., S-wave velocity). Rayleigh wave phase velocity is primarily sensitive to S-wave velocity in the upper mantle, and the sensitivity to density and P-wave velocity is mostly confined to shallow depths (<40–50 km, i.e., the crust in this region). In this study, our primary interest lies in the structure of upper mantle. Therefore, we keep density invariant and let P-wave velocity vary with S-wave velocity at a constant Vp/Vs ratio of 1.73. The Moho depth is also treated as a model parameter, considering the tradeoff between the variations of S-wave velocity and undulations of the Moho.

The AK135 global model is a conventional choice for the initial model in the inversions of S-wave velocities, but it is not necessarily the most appropriate. We first invert the average Rayleigh wave phase velocities for an average 1-D S-wave velocity model using the AK135 model as the initial model. The optimized average S-wave velocity model is then used as the initial model in the inversions of phase velocity dispersion curves at individual grid nodes. The initial Moho map is derived from the studies of receiver functions (T. Zheng et al., 2009; Wei et al., 2011). We apply a correlation coefficient of 0.4 between adjacent layers to smooth the model, along with an a priori standard deviation of 0.1 km/s to damp S-wave velocities. Heavy damping is applied to the Moho depth to allow for small
variations from the initial model, because Rayleigh wave phase velocities are much less sensitive to the topography of interfaces in the Earth's interior than receiver functions.

5. Results

5.1. Rayleigh wave phase velocities

The average Rayleigh wave phase velocities for the central and western NCC are shown in Fig. 6a. The phase velocities at periods of 20–143 s are consistently lower than the global averages, which are calculated from the AK135 global model. The output standard deviations solely reflect the convergence properties of the inversions and are somewhat lower than they should be. Phase velocities at different periods roughly reflect the structure at different depths. The sensitivity kernels to S-wave velocities (Fig. 6b) indicate that the Rayleigh wave data used in this study can clearly resolve the S-wave velocity structure at depths of ca. 20–200 km. The phase velocities at periods longer than 40 s are primarily sensitive to the S-wave velocities in the mantle lithosphere and the phase velocities at periods of 30–40 s are sensitive to both the S-wave velocities and Moho undulations.

The resulting 2-D phase velocity maps are shown in Fig. 7. The velocity anomalies are calculated relative to the average phase velocities. We focus on the structure of the mantle lithosphere, and pay close attention to the phase velocity maps at the periods of 33–125 s. The Ordos Block in the WB is characterized by high-velocity anomalies at periods of 33–125 s, consistent with previous studies of body wave tomography (Huang and Zhao, 2006; Tian et al., 2009; Zhao et al., 2009; Li and van de Hiest, 2010) and surface wave tomography in this region (Huang et al., 2009). To the south, this high-velocity zone is connected to the high-velocity anomalies in the Yangtze Craton. In the north, it stops near the IMSZ. The IMSZ and Yinshan Block are dominated by low-velocity anomalies. The prominent features in the TNCO are two low-velocity zones (LVZs) in the north and south, respectively. The northern LVZ is persistently observed at the periods of 33–143 s, and the southern LVZ is only visible at the periods of 33–80 s. Between the two LVZs, a high-velocity patch is observed adjacent to the transtensional zone of the SSR, which is much narrower than the extensional domains in the north and south (Xu et al., 1993; Zhang et al., 1998).

The resolution of the 2-D phase velocity maps can be approximately assessed by the standard deviations calculated from the covariance matrices in the inversions. Assuming that the model parameters in the inversions are statistically independent, phase velocity anomalies exceeding two standard deviations are thought to be well resolved at approximately the 95% confidence level (Li et al., 2003). Two standard deviations of the phase velocities at 50 s period are shown with contours in Fig. 7b. Based on this contour map, the resolution for most of the study area is quite good, but the resolution of the northwest corner is poor due to the sparse stations and crossing raypaths. The shapes of two-standard-deviation contours for the other periods are similar except for slightly larger standard deviation values.

Routine checkerboard tests were also performed to test the resolution of the 2-D phase velocity maps. The input checkerboard is designed with a patch size of 2°×2° (ca. 200 km) and velocity anomalies of ±5%. The synthetic data with random noises are calculated based on the real raypaths using the method developed by Li et al. (2003). The checkerboard inversions share the same regularization of the inversions of the real data. The recovered models at periods of 33, 50, 80, 100 and 125 s are shown in Fig. 8. Overall, the checkerboard is well recovered at periods of 33–100 s. At 125 s period, obvious smearing at the NW corner of the study area is observed in the recovered model. In addition, we expect that small-scale (<200 km)
velocity anomalies may be resolved in the TNCO because the complementary NCISP-VII array provides very dense stations and crossing raypaths.

5.2. S-wave velocities

By inverting the average phase velocities, we obtain a 1-D average S-wave velocity model for the central and western NCC (Fig. 9a). The AK135 global model is used as the initial model, and the fit of the average phase velocities is good (Fig. 6a). Vertical resolution is evaluated by inspecting the resolution matrix (Fig. 9b). For all layers above a depth of ca. 200 km, the peak resolution values occur on the diagonal of the matrix and decrease with depth. This indicates that the S-wave velocities of these layers are resolved at the correct depths. Vertical smearing is also observed as the stretching of resolution curves beyond the peak values (Fig. 9c). Considering both the vertical resolution and sensitivity kernels of phase velocities to S-wave velocities (Fig. 6b), we therefore primarily describe and interpret the structure of the mantle lithosphere (Moho – 200 km).

The 1-D average S-wave velocity model shows a thicker crust (42 km) and a slightly slower mantle lithosphere compared with the AK135 global model, with two prominent low-velocity layers at ca. 100 and 200 km depths, respectively (Fig. 9a). The relatively thick crust is consistent with the results of receiver functions (T. Zheng et al., 2009; Tian et al., 2011; Wei et al., 2011). The intermediate S-wave velocities of the mantle lithosphere accommodate the coexistence of a cratonic mantle root beneath the Ordos Block and thinned lithosphere beneath the TNCO and IMSZ. The deep low-velocity layer is most likely associated with the lithosphere–asthenosphere boundary (LAB), while the shallow low-velocity layer is probably associated with a mid-lithospheric interface also revealed by the results of S receiver functions (Chen et al., unpublished data).

The 3-D S-wave velocity model of the mantle lithosphere is illustrated by horizontal slices in Fig. 10 and vertical slices in Figs. 11–13. The Ordos Block is characterized by velocities higher than ca. 4.5 km/s, reflecting the presence of a cold and refractory mantle root. Surrounding
the Ordos Block, the IMSZ to the north, the TNCO to the east and the Qilianshan Orogenic Belt to the west are characterized by low velocities. The variations from high to low velocities approximately delineate the boundaries of the cratonic nucleus. To the south of the Ordos Block, the Qinling Mountains (western part of the Qinling–Dabie–Sulu Orogenic Belt) are characterized by high velocities (Figs. 10 and 12). The high

![Fig. 6](image)

**Fig. 6.** (a) Average phase velocities of the fundamental mode Rayleigh waves for the central and western North China Craton. The circles with error bars denote the observed phase velocities at 13 frequency bands. The solid line denotes the predicted dispersion curve calculated from the inverted 1-D S-wave velocity model (Fig. 9a) and the dashed curve denotes the synthetic dispersion curve of the AK135 global model. (b) Phase velocity sensitivity kernels for shear velocity. Each line represents a kernel function at one period. The kernel functions at the periods of 33, 50, 67, 80, 100, 125 s are highlighted with black lines.

![Tomographic maps of Rayleigh wave phase velocity anomalies](image)

**Fig. 7.** Tomographic maps of Rayleigh wave phase velocity anomalies. The anomalies are relative to the average phase velocities (Fig. 6a). The dashed lines mark the primary tectonic boundaries and the dotted lines mark the two Cenozoic rift zones. The thick gray line denotes the North–South Gravity Lineament (NSGL). The contours in (b) are two standard deviations of phase velocities at 50 s period. The abbreviations of the tectonic provinces are explained in Fig. 1.
velocities blur the boundaries between the Ordos Block and the Yangtze Craton.

Due to limited vertical resolution, our S-wave velocity model is insufficient to accurately detect the LAB beneath the Ordos Block, which has recently been reported at a depth of ca. 200–300 km (L. Chen et al., 2009; Zhao et al., 2009; Bao et al., 2011). Nevertheless, the high velocities beneath the Ordos Block are traced continuously from the Moho to a depth of 200 km in our model (Figs. 10–12). Within the high-velocity mantle root, a relatively low-velocity layer is observed at a depth of 100–150 km, with the upper boundary at a depth of ca. 100 km (Figs. 11 and 12). A recent study of S receiver functions also finds the interface of a low-velocity layer at similar depths (Chen et al., unpublished data). This kind of interface has been observed in other stable cratons worldwide (Rychert et al., 2010, and references therein), but its nature is still unknown (e.g., Yuan and Romanowicz, 2010).

Fig. 8. Checkerboard test for the 2-D phase velocity maps. The black triangles denote the seismic stations. (a) Input 2° × 2° checkerboard of ±5% anomalies. (b–f) Recovered models at five different periods. The dashed lines mark the primary tectonic boundaries and the dotted lines mark the two Cenozoic rift zones. Refer to Figs. 1 and 7 for tectonic province names.

Fig. 9. (a) 1-D average S-wave velocities (solid line with error bars). The initial AK135 global model (red dashed line) is plotted for comparison. (b) Resolution matrix of the inversion for average S-wave velocities. (c) Rows of the resolution matrix for three layers: Moho–60 km, 120–140 km, and 185–210 km.
Surrounding the high-velocity zone beneath the Ordos Block, the Qilianshan Orogenic Belt, IMSZ and TNCO primarily exhibit low velocities in the mantle lithosphere. In particular, our S-wave velocity model is capable of resolving small-scale heterogeneities in the mantle lithosphere of the TNCO due to the dense stations and crossing raypaths. Two major LVZs are observed beneath the TNCO (Figs. 10 and 13). The northern LVZ stretches westward and southward along the Paleoproterozoic orogenic belts, and extends downward to depths greater than 200 km. Its surface position correlates well with the distribution of the Late Mesozoic to Cenozoic magmatism and intrusions (Fig. 1b). In contrast, the southern LVZ tapers off at a depth of ca. 180 km, which is consistent with new results of body wave tomography (Zhao et al., 2012). Between the two LVZs, a prominent high-velocity patch is present from the Moho to a depth of ca. 100 km, and is located adjacent to the narrow transtensional zone of the SSR. It may reflect a craton remnant preserved beneath the TNCO due to lithospheric reactivation during the Phanerozoic tectonic events. Another similar craton remnant is inferred beneath the Haiyuan region of the YHR by Bao et al. (2011).

6. Discussion

Great heterogeneities and systematic variations are present in the mantle lithosphere of the central and western NCC based on our 3-D S-wave velocity model. As only Rayleigh wave data are used in this study, the S-wave model presented here is a model of the SV components of shear waves. Nevertheless, our results display similar structural features to those imaged by previous studies of body wave tomography (Tian et al., 2009; Zhao et al., 2009) and surface tomography (Huang et al., 2009), but reveal more small-scale velocity anomalies. These small-scale anomalies are in good agreement with geological and tectonic features (Zhang et al., 2003) and the high-resolution structural images derived from receiver function analysis (L. Chen et al., 2009; Chen et al., unpublished data). The overall consistencies indicate that our S-wave velocity model is reliable and can provide important deep structural constraints on the spatial extent and geodynamics of lithospheric modification and thinning in the central and western NCC.

6.1. Locally modified lithosphere beneath the TNCO

In contrast to the widespread thinning of the EB lithosphere evidenced by many previous multidisciplinary studies (Chen et al., 1991; Griffin et al., 1998; Fan et al., 2000; Xu, 2001; Zhu et al., 2002; Chen et al., 2006; Menzies et al., 2007; Chen et al., 2008; Huang et al., 2009; Santosh, 2010; Zhang et al., 2011), our tomography results suggest the coexistence of both thinned lithosphere and a preserved cratonic mantle root in the central and western NCC, in agreement with previous seismological results (L. Chen et al., 2009; Chen, 2010; Bao et al., 2011). The low-velocity anomalies in the mantle lithosphere are
mainly confined to the Paleoproterozoic orogenic belt of the IMSZ and TNCO, whereas the mantle lithosphere beneath the Ordos Block is characterized by prominent high velocities (Fig. 10). Furthermore, marked small-scale structural heterogeneities are found beneath the TNCO. For example, a high-velocity patch is embedded in the prevailing low-velocity anomalies, corresponding to the narrow transtensional

![Diagrams showing S-wave velocities and vertical gradients along profiles A-A’ and B-B’](image)

Fig. 11. (a) S-wave velocities and (b) vertical gradients along the E–W striking profile A–A’ (see Fig. 10a for profile location). This profile starts from the Qilianshan Orogenic Belt (QOB) in northeastern Tibet, runs through the Ordos Block in the western Block (WB) of the North China Craton, the Lvliang Mountains (LM), the Shanxi–Shanxi Rift (SSR), the Taihang Mountains (TM) in the Trans-North China Orogen (TNCO), and ends at the Bohai Bay Basin (BBB) in the Eastern Block (EB). This traverse roughly coincides with the NCISP-V array (a sub-array of the North China Interior Structure Project). The featured abbreviations of the tectonic provinces are explained in Fig. 1.

Fig. 12. (a) S-wave velocities and (b) vertical gradients along the N–S striking profile B–B’ (see Fig. 10a for profile location). This profile starts from the Central Asian Orogenic Belt (CAOB) in the north, runs through the Yinshan Block (YB), the Inner Mongolia Suture Zone (IMSZ), the Ordos Block in Western Block (WB), and ends at the Yangtze Craton (YTC) to the south (Fig. 10a). The abbreviations of the featured tectonic provinces are explained in Fig. 1.
zone of the SSR on the surface (Figs. 10a,b and 13). This suggests a close
correlation between deep lithospheric structure and surface geology.
Even the low-velocity anomalies show distinct north–south variations.
The southern LVZ tapers off at a depth of ca. 180 km, whereas the northern LVZ can be traced to a depth of 200 km (Figs. 10f and 13) or deeper (Zhao et al., 2012). According to the structural variations from Earth’s surface to the upper mantle derived from our surface wave tomography, geological surveys (Xu et al., 1993; Zhang et al., 1998) and high-resolution body wave tomography (Zhao et al., 2012), we divide the TNCO lithosphere into three segments along a north–south direction (Fig. 10a).

One of the most significant structures in our results is the upper mantle LVZ beneath the northern segment of the TNCO (Fig. 10). Combined with recent results of body wave tomography (Zhao et al., 2009, 2012), a plume-like upwelling is interpreted to originate from the deep upper mantle and to ascend into the lithosphere. Based on a synthesis of the earlier results of body wave tomography (Tian et al., 2009; Xu and Zhao, 2009) and their correlation with the surface geological features, Santosh et al. (2010) suggested that the hot upwelling was immediately below the Ordos Block, rising up through the TNCO to shallow levels. Although its deep origin is still debatable, the hot upwelling is believed to rise up through the TNCO as a corridor. As thermal anomalies are predominantly responsible for velocity anomalies within the upper mantle (Cammarano et al., 2003), the consecutive low velocities in our results from the Moho to 200 km depth beneath the northern segment of the TNCO indicate that this hot upwelling impacts the whole lithosphere and even the crust, consistent with a thinned lithosphere (L. Chen et al., 2009).

Major element and isotopic analysis of mantle peridotite xenoliths and xenocrysts indicate that there is compositional modification of the lithospheric mantle beneath this region by massive addition of asthenospheric melts (Tang et al., 2006, 2008; H.F. Zhang et al., 2009).

Fig. 13. (a) S-wave velocities and (b) vertical gradients along the N–S striking profile C–C′ (see Fig. 10a for profile location). This profile primarily traverses the Trans-North China Orogen (TNCO), and roughly follows the NCISP-VII array (Fig. 10a). The abbreviations of the featured tectonic provinces are explained in Fig. 1.

Fig. 14. Schematic diagram of local lithospheric modification in the Trans-North China Orogen (TNCO). The basemap is the S-wave velocity image along the profile C–C′ (Fig. 13). The question mark indicates that the geodynamic process driving lithospheric modification in the southern segment of the TNCO is still unclear.
Although obvious crustal thickening beneath the northern segment of the TNCO is not observed by receiver function imaging (Zheng et al., 2008a; T. Zheng et al., 2009; Wei et al., 2011), magmatic underplating due to mantle upwelling is inferred by the presence of a thick crust–mantle transition zone (Zheng et al., 2008a) and a distinctly anisotropic lowermost crust (Cheng et al., 2012). U–Pb age and Hf isotopic data from Phanerozoic zircon populations of the granulite and pyroxenite xenoliths demonstrate that magmatic underplating in the lower crust beneath this region is episodic through the Paleozoic to Cenozoic (Wilde et al., 2003; Y.S. Liu et al., 2004; J.P. Zheng et al., 2009; Ying et al., 2011; Zhang et al., 2012). All of these observations support the inference that thermal or chemical ‘bottom-up’ erosion and modification of the lithosphere may happen beneath the northern segment of the TNCO.

Another significant LVZ is confined to the southern segment of the TNCO. In contrast to that in the north, this LVZ appears to taper off at a depth of ca. 180 km revealed by both our surface wave tomography study (Figs. 10f and 13) and a new body wave tomography study (Zhao et al., 2012). Rare Cenozoic magmatism and a noticeably thinned crust (Chen et al., 2010; Wei et al., unpublished data) appear to reflect no magmatic underplating in the lower crust. These observations suggest that lithospheric modification in the south may not be induced by the thermal or chemical ‘bottom-up’ erosion, but requires a different geodynamic process.

A high-velocity patch beneath the middle segment of the TNCO is a well-resolved feature in our results. This patch tapers off at a depth of ca. 100 km and appears to be isolated from the thick high-velocity zone beneath the Ordos Block (Fig. 11a). However, it is laterally smeared with the high-velocity zone beneath the Ordos Block in most previous tomography studies (Huang et al., 2009; Tian et al., 2009; Zhao et al., 2009). The uppermost mantle in the same region is detected to be of high electrical resistivity by magnetotelluric insmearied with the high-velocity zone beneath the Ordos Block in the Phanerozoic. Related major tectonic events include the amalgamation of the NCC (Zhao et al., 2001; S.H. Zhang et al., 2009; Jiang and Guo, 2010; Zhang et al., 2010; Ying et al., 2011; Zhang et al., 2012). Unlike lithospheric reactivation in the EB from the Late Mesozoic to Cenozoic which caused fundamental destruction of the cratonic mantle root, the effects of the Phanerozoic tectonic events on the central and western NCC may be weaker and localized in pre-existing weak zones. They are therefore unable to fully erase the intrinsic heterogeneities and the imprints of earlier tectonic events on the lithosphere of this region. Hence, the present heterogeneous lithospheric structure beneath the central and western NCC may reflect the combined effects of multiple tectonic events over a long-term history from the formation of the craton to the present.

The SSR extends across the TNCO from north to south, representing the last stage of Cenozoic lithospheric thinning and modification (Zhang et al., 2003). Subsidence in this elongate rift zone began in the Late Miocene or Earlier Pliocene (N. Wang et al., 1996), although its formation is disputed. In our tomography results, LVZs in the TNCO mantle lithosphere are not restricted to the rift zone, but are well correlated with the distribution of Cenozoic basalts in the TNCO. Cenozoic magmatism in the TNCO is dated mainly at the Early Tertiary (Song and Frey, 1989; Tatsumoto et al., 1992; Zhu, 1998; Zhou et al., 2002; Tang et al., 2006; Xu et al., 2008). Therefore, we conclude that the implied hot mantle upwelling and underplating occur much earlier than the rifting of the SSR and are unlikely to be the direct dynamic trigger of this rifting. On the other hand, the extension rate of the SSR is estimated to be 0.5–1.6 mm/yr averaged over the Late Pliocene to Quaternary (Zhang et al., 1998). GPS data show no clear indication of active extension across the SSR at present (He et al., 2003). The onset of the SSR is nearly coeval with that of east–west extension in the Tibetan Plateau (Harrison et al., 1995; Yin et al., 1999; Maheo et al., 2007), suggesting that it may be directly triggered by the eastern pushing and extrusion of Tibet. If this is true, the SSR is a type of passive rift.

6.2. A cratonic nucleus beneath the Ordos Block

The Ordos Block is characterized by high velocities down to 200 km or deeper in both our results and previous tomography studies (Huang and Zhao, 2006; Huang et al., 2009; Tian et al., 2009; Zhao et al., 2009). Recent receiver function images based on dense array data also identify the LAB at ca. 200 km depth (L. Chen et al., 2009; Chen et al., unpublished data), and GPS measurements indicate that there is no internal crustal deformation within the Ordos Block (Shen et al., 2000; Wang et al., 2003). These observations suggest that a thick cratonic mantle root may have been largely preserved under the Ordos Block. However, there are no constraints on the thickness and composition of the cratonic lithosphere from petrological and geochemical data because of the thick Phanerozoic sedimentary cover in this region.

Within the high-velocity zone of the Ordos Block, our results further reveal a ubiquitous low-velocity layer at a depth of 100–150 km, with the upper boundary at a depth of ca. 100 km (Figs. 11 and 12). New high-resolution images of receiver functions beneath the Ordos Block also indentify a strong discontinuity at a depth of 80–100 km, across which velocity decreases with depth (Chen et al., unpublished data). This is consistent with our observations (Fig. 11). Such a mid-lithospheric low-velocity layer or discontinuity is not a unique feature of the Ordos Block, but has been widely reported in other Precambrian shields worldwide. Rychert and Shearer (2009) observed similar discontinuities at depths of ca. 100 km beneath many stable Precambrian shields based on receiver function analysis of global seismic data, and interpreted them as boundaries of composition, melting or anisotropy, not temperature alone. Regional studies of receiver functions also detect
shallow interfaces at a depth of ca. 100 km beneath the Dharwar Craton (Kumar et al., 2007), northeastern Brazil (Heit et al., 2007), and the Arabian Platform (Hansen et al., 2007). The spatial overlap between a seismic discontinuity from receiver functions and a conductive anomaly from magnetotelluric data at ca. 100 km depth is observed beneath the Slave Craton in Canada (C.-W. Chen et al., 2009; Griffin et al., 2009). More recently, Yuan and Romanowicz (2010) observed two discontinuities in the upper mantle of the stable North American continent by joint inversion of long-period seismic waveforms and SKS splitting data. They argued that the deeper discontinuity at a depth of 180–240 km represents the LAB, whereas the shallow one at a depth of ca. 150 km indicates the bottom of a depleted chemical layer in the cratonic lithosphere. Petrological and geochemical studies of heavy-mineral concentrates and xenoliths from the kimblerite intrusions beneath the Slave Craton also find two compositionally distinct layers in the mantle lithosphere (Griffin et al., 1999). It therefore appears that this two-layered structure is a common feature of Archean cratons worldwide.

The nature of the mid-lithospheric discontinuity beneath stable cratons is closely related to their formation and long-term evolution. A positive thermal anomaly, a layer of partial melt, or the presence of fluid can all account for the observed low velocities, but they cannot be sustained over long time scales in the cold, refractory cratonic lithosphere. An enrichment of phlogopite or a similar mineral caused by metasomatism is the most likely candidate to produce the low-velocity anomaly (C.-W. Chen et al., 2009; Griffin et al., 2009). Phlogopite is stable at pressure and temperature conditions corresponding to depths of 100–150 km in a cold cratonic mantle root (Boerner et al., 1999), and is commonly found in mantle xenoliths entrained in Cenozoic basalts of the TNCO (Xu et al., 2010). Metasomatic processes are assumed to be ubiquitous in the mantle lithosphere of Archean cratons. If metasomatism is not present, the synthetic calculations of density and seismic velocity based on the highly-depleted mineral models of the Archean lithospheric mantle would predict a more negative geoid anomaly and higher elevation than is observed (Afonso et al., 2008). The existence of a metasomatized layer at 100–150 km depth beneath the Ordos Block fits well within the context of subduction related to assembly of the NCC during the Paleoproterozoic (L. Zheng et al., 2009; Santosh, 2010; Santosh et al., 2011b). However, as mentioned by Yuan and Romanowicz (2010), additional constraints are required from seismic anisotropy measurements of the lower mantle lithosphere layer beneath the Ordos Block. This is an interesting topic that will be further investigated in future studies.

6.3. Geodynamic causes of Cenozoic lithospheric modification in the central and western NCC

The lithospheric modification in the central and western NCC is substantially different from the lithospheric reactivation and destruction in the EB. In the latter, widespread thinning prevails in the mantle lithosphere (Chen et al., 2006, 2008), whose composition has been altered similar to the modern convective mantle (Wu et al., 2008 and references therein). In contrast, our results and previous geophysical studies reveal that the lithosphere of the central and western NCC has been locally modified and thinned (L. Chen et al., 2009; Chen, 2010). Combined with studies of peridotite and granite xenoliths from Cenozoic basalts (Zheng et al., 2001; Gao et al., 2002; Xu et al., 2008; Ying et al., 2010; Zhang et al., 2012), our results also show that the materials of the cratonic mantle root are still preserved in the TNCO. Thus, lithospheric modification is dramatically weaker in the central and western NCC than in the EB.

The widespread lithospheric reactivation and destruction of the EB are believed to be attributed to deep subduction of the Pacific plate, based on geophysical imaging and the concordant timing of the two tectonic events (Chen, 2010 and references therein). Body wave tomography studies reveal that the stagnant slab-front of the subducted Pacific plate has reached the mantle transition zone beneath the Tanlu Fault (Huang and Zhao, 2006; Li and van de Hilst, 2010; Zhao et al., 2012), and that the voluminous LVZ in the upper mantle of the EB appears representative of large-scale upwelling. Surface geological features including major magmatism (Xu et al., 2004b; Wu et al., 2005; Zhang et al., 2011), intensive tectonic extension (Ren et al., 2002) and large-scale gold mineralization (Yang et al., 2003; Sun et al., 2007) also favor the interpretation of large-scale mantle upwelling beneath the EB caused by Pacific subduction. In contrast, the central and western NCC is further displaced from the plate margin and thus less influenced by subduction processes. The distinct lithospheric structures between the northern and southern segments of the TNCO imply very complex geodynamic processes working in this region (Fig. 14). The LVZ extending deep into the upper mantle, together with thinned lithosphere (L. Chen et al., 2009), a thick crust–mantle transition zone (Zheng et al., 2008a), and widely distributed Cenozoic magmatism (Fig. 1) and magmatic underplating in the northern segment (Zhou et al., 2002; J.P. Zheng et al., 2009; Ying et al., 2011; Zhang et al., 2012), suggests that small-scale hot upwelling has taken place and is possibly an on-going process. The far-field effect of deep subduction and stagnancy of the Pacific plate may account for this ‘bottom-up’ process (Zhao and Xue, 2010). Upwelling may be less intense or even absent in the southern segment of the TNCO, given the relatively shallow LVZ in the lithosphere and lesser Cenozoic magmatism.

The Cenozoic India–Eurasia collision has been proposed as a possible geodynamic cause of lithospheric modification in the central and western NCC (e.g., Menzies et al., 1993; Deng et al., 2004; M. Liu et al., 2004; Xu, 2007). The direct contact and forcing of the Tibetan Plateau on the WB cause the differential counterclockwise rotations of the Ordos Block and the Taihang Mountains relative to their surrounding regions (Zhang et al., 1998, 2003). This collision may also cause the eastward extrusion of the Tibetan crust and upper mantle through the Qinling Mountains (the western part of the Qinling–Dabie–Sulu Orogenic Belt). Seismic anisotropy derived from SKS splitting measurements shows generally E–W fast S-wave polarization directions in the upper mantle beneath the Qinling Mountain region, which is used as a line of evidence to support the eastward extrusion model (Huang et al., 2008; Zhao and Xue, 2010). However, no low-velocity corridor in the shallow upper mantle has been recognized by either surface wave tomography (Fig. 10) or body wave tomography (Zhao et al., 2011, 2012). This suggests that eastward extrusion may be limited to small spatial scales, if indeed it exists at all. The corresponding impacts on the lithosphere are weak in the central and western NCC.

7. Conclusions

Using high-quality broadband seismic data recorded by a dense 2-D array of permanent and portable seismic stations, we have carried out Rayleigh wave tomography in the central and western NCC. A high-resolution 3-D S-wave velocity model of the mantle lithosphere is constructed. Our model provides some new insights into the lithospheric formation and modification of the central and western NCC.

1. The mantle lithosphere of the TNCO is primarily characterized by low seismic velocities, but includes great heterogeneities. Two major LVZs are present in the north and south, respectively. The northern LVZ exhibits coherent low-velocity anomalies from the Moho to 200 km depth (the lower limit of our model) and reflects a small-scale mantle upwelling. The southern LVZ tapers off at a depth of ca. 180 km and its geodynamic explanation remains unknown. A high-velocity patch in the uppermost mantle is observed adjacent to the narrow transtensional zone of the SSR. We interpret it as a possible remnant of cratonic mantle root.
2. Prominent high seismic velocities are observed beneath the Ordos Block, reflecting the presence of cratonic mantle root. Within the high-velocity zone, a noticeable low-velocity layer is observed at depths of 100–150 km, with the upper boundary at a depth of ca. 100 km. This mid-lithospheric structure may have been produced by metasomatic processes related to assembly of the NCC during the Paleoproterozoic.

3. The present heterogeneous lithospheric structure beneath the TNCO probably reflects multiphase reworking of pre-existing weak zones by a succession of tectonic events since the amalgamation of the NCC. The latest Cenozoic lithospheric modification may be attributed to the far-field effects of both Pacific plate subduction and the India–Eurasia collision.

Acknowledgments

We express thanks for the excellent field work of the Seismic Array Laboratory in the Institute of Geology and Geophysics, Chinese Academy of Sciences. We thank the Data Management Centre of the China National Seismic Network at the Institute of Geophysics, China Earthquake Administration, for providing waveform data from the permanent seismic stations. We are grateful to Professor Zheng Tianyu and Dr. Wei Zigen for providing the crustal thickness model. We appreciate the helpful discussions of Professor Zhao Liang and Dr. Lu Gang. We express our thanks to Dr. A. Atkin, Professor M. Santosh and an anonymous reviewer for their constructive comments and suggestions. This research is supported by the National Science Foundation of China (NSFC, grant 40904022, 90914011 and 91014006). This is contribution 196 from the ARC Centre of Excellence for Core to Crust Fluid Systems (http://www.cccfs.mq.edu.au) and 838 in the GEMOC Key Centre (http://www.gemoc.mq.edu.au).

References


