Crust–mantle structure difference across the gravity gradient zone in North China Craton: Seismic image of the thinned continental crust

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Abstract

The Archean North China Craton (NCC) has been tectonically modified and lost its thick lithospheric keel during the Mesozoic and Cenozoic. The processes and mechanisms of the lithospheric modification and its appearance in and the relation between different subregions of the NCC are still poorly understood. Seismic data from 45 stations along a 470-km long profile cross the Bohai Bay Basin (BBB) and the Taihangshan Mountain Range (TMR) in the NCC were employed to construct a coherent structural image of the crust and uppermost mantle. An integrated receiver function imaging technique combining the common conversion point stacking approach with waveform inversion and forward modeling was proposed to extract the structural information beneath the study region. Modeling of Bouguer gravity anomalies was also applied to constrain the density distribution. The imaging result reveals distinct structural features between the mountain range and the basin area, and presents a picture of uneven crust thinning within the study region. In the east BBB the crust is significantly thinned due mainly to the reduction in the thickness of the lower crust including the crust–mantle transition zone, by up to ~12 km. The west TMR, in contrast, is characterized by a relatively thick lower crust of ~20 km. The teleseismic waveform data and the gravity observation suggest a thicker crust and a buoyant mantle lithosphere beneath the TMR compared with the BBB. The contrasting crustal structural features appear coupled with the lithospheric processes and possibly reflect that different tectonic mechanisms and deformation regimes dominated the evolution of the two regions. The North-South Gravity Lineament, lying between the TMR and BBB, might represent a deep intra-continental boundary separating the NCC into topographically and tectonically different regions.

Keywords: North China Craton; Receiver function imaging; Gravity modeling; Lower-crust thinning

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1. Introduction

The North China Craton (NCC), a cratonic nucleus composing of Archean and Proterozoic rocks in the eastern China continent, is as old as 3.8 Ga (Liu et al., 1992). The Archean nucleus of eastern China was tectonically stable until the late Mesozoic rejuvenation which is evidenced by the development of large sedimentary basins, high heat flow and extensive magmatism. Therefore, the eastern China continent has become a natural laboratory to study the modification process of the old Archean craton (Griffin et al., 1998; Menzies and Xu, 1998; Gao et al., 1998a; Fan et al., 2000; Xu, 2001; Zhang and Sun, 2002; Zhou et al., 2002). There are two important tectonic events intimately linked to the Mesozoic and Cenozoic modification of the NCC. One is the removal or transformation of about 80–120 km of the Archean lithosphere (Fan and Menzies, 1992; Menzies et al., 1993), which is evidenced by the petrologic and isotopic data from Ordovician kimberlites and Tertiary basalts, as well as xenoliths enclosed in them. The other is the separation of the NCC into two topographically and tectonically different regions by a giant linear zone of gravity gradient, known as the North-South Gravity Lineament (Ma, 1987).

Several hypotheses have been proposed to explain the lithospheric evolution process, such as rifting (Tian et al., 1992), delamination (Gao et al., 1998a), thermal and chemical erosion (Griffin et al., 1998), mantle plume (Deng et al., 1998, 2004) and basaltic underplating (Zhang and Sun, 2002). To identify and certify these hypotheses, more complete knowledge of the structure and composition of the cratonic lithosphere is required. Studies of the Cenozoic and Mesozoic magma and their mantle xenoliths provide direct information about lithospheric composition and physical properties because these rocks directly sample the mantle. The limited spatial and temporal distribution of the xenoliths, however, may lead to multiple solutions and hence give rise to large uncertainties in the interpretation. On the other hand, geophysical surveys may provide fine-scale structural information of the present crust and mantle. In the late 1980s, the Xiangshui-Mandal Geoscience Transect as a part of the Chinese Geoscience Transect Program traversed North China from the coast of the Yellow Sea to the border between China and Mongolia (Fig. 1), which gives a primary geophysical understanding of the NCC (Ma et al., 1991). More recently, a temporary deployment of broadband seismic stations under the Northern China Interior Structure Project (NCISP) (Fig. 1) (Zheng et al., 2005) offers an excellent opportunity to construct a crust image in terms of shear wave velocity ($V_s$) distribution using teleseismic receiver functions.

Since the pioneering work of Langston (1979), many geophysicists have used receiver function imaging to study the velocity structure of the crust and upper mantle, and great progress has been made in the application of this approach to various regions during the last decade. Some data processing techniques, such as the move-out correction and the common conversion point (CCP) stacking techniques (Zhu, 2000), have been routinely applied to improve the signal-to-noise ratio for receiver function imaging (Dueker and Sheehan, 1997; Li et al., 1998; Owens et al., 2000). Receiver functions are very sensitive to velocity contrasts beneath a seismic recording site. The radial receiver function isolates the P to S mode–converted phases at the Moho or interfaces within the crust and allows the travel times from the interfaces to the surface to be determined. The amplitudes of the converted phases are sensitive to the impedance contrasts at the interfaces. However, the strong model dependence of the time-to-depth conversion in CCP stacking often gives rise to large uncertainties in the imaging result, and hence needs to be considered in interpretation. Particularly, when the lateral variations of the overburdens are substantial, incorrect velocity models may cause incoherent stacking and incorrect translation from the time domain to the depth domain of the receiver functions, hence reducing the stacking-based image quality and even apparently distorting the structural feature of the target discontinuity (Chen et al., 2005; Rost and Weber, 2001; Schlindwein, 2006). Therefore, a correct velocity model is essential in constructing high-quality stacking-based depth images for subsurface discontinuities that are overlain by strong lateral structural anomalies.

Unfortunately, thinned continental crust associated with lithospheric extension is generally overlain by thick sediments, such as that of the Bohai Bay Basin (BBB) in the eastern NCC. The highly heterogeneous crust of the NCC means that crustal structure imaging is more difficult using receiver functions. To reduce the uncertainty and increase the reliability of receiver function imaging, we combined the advantages of CCP stacking and waveform inversion by seeking a consistent result. Waveform inversion and extensive forward modeling were used to construct a velocity model for reliable CCP stacking and to judge the signals appearing in the CCP image; on the other hand, CCP imaging provided useful information to constrain the parameter space for waveform inversion. Furthermore, the density distribution was also jointly constrained through modeling the Bouguer gravity anomaly and based upon the structure model from receiver function inversion and CCP stacking. By apply-
Fig. 1. Topographic map of the eastern NCC and adjacent regions, showing major tectonic units including the BBB and the TMR. Triangles represent broadband seismic stations used in this study, with some station numbers marked alongside. The right inset shows the distribution of teleseismic events used. The bottom inset shows the subdivision of the craton into the Eastern Block (EB), the Western Block (WB) and the Trans North-China Orogen (TNCO). Quadrangles mark the locations of the Xiangshui-Mandal Geoscience Transect (XMGT).

2. Geologic background

The NCC, the Chinese part of the Sino-Korea Platform, covers most of the North China Block. The Qilianshan Orogen and the Tianshan-Inner Mongolia-Daxinganling Orogen bound the craton to the west and the north, respectively, and to the south the Qinling–Dabie–Su–Lu ultrahigh-pressure metamorphic belt separates the craton from the Yangtze Block (Fig. 1). The Trans North-China Orogenic Belt, a Paleoproterozoic collision boundary trending roughly north–south (Zhao et al., 2001), separates the NCC into two major tectonic block, the western and the eastern blocks (Fig. 1). The NCC consists of Early Archean to Paleoproterozoic basement overlain by Mesoproterozoic to Cenozoic cover. The relatively flat-lying Paleozoic sedimentary cover of the NCC shows that it was stable until the Jurassic, before the collision between the NCC and the...
Yangtze block to the south. The western block of the NCC remains as a stable craton with thick lithosphere and low heat flow, and has experienced little internal deformation since the Precambrian. The eastern block, on the other hand, underwent extensive magmatism and large-scale basin formation during the Late Mesozoic and Cenozoic. The reactivity of the eastern NCC could be related to the closure of a Mongolo-Okhotsk sea to the north, or the collision of the Yangtze block to the south (Zhang et al., 2003). The convergent motion of the subducted Paleopacific oceanic lithosphere is also a possible tectonic cause (Griffin et al., 1998).

One of the most intriguing structural features of the NCC is the NNE–SSW trending North-South Gravity Lineament separating the mountain range (Taihangshan Mountain Range, or TMR) in the Trans North-China Orogenic belt and the basin (BBB) in the eastern NCC. The two geological units on the opposite sides of the North-South Gravity Lineament differ significantly not only in surface topography but also in deep tectonics, as manifested by striking contrasts in altitude, gravity as well as lithologic stratum (Ma, 1987). Such a specific structural feature is obviously a key exploration target for the tectonic evolution of the rejuvenated craton. In this paper the study profile (Fig. 1) traverses both the BBB and the TMR, which allows us to systematically compare the structural difference between the two geological units.

The previous geological investigations have shown that the BBB was developed mainly during the Cenozoic; however, it also received various depositions and underwent strong destruction before its major development. The recent seismic imaging result (Zheng et al., 2005), which reveals different sedimentary features between the Mesozoic and Cenozoic, suggests that the sediments were not successively developed and the compression associated subsidence of the Mesozoic basin was interrupted after the Middle Mesozoic. Considering the evolution of the basin during the Early Cenozoic, a two-stage evolution model is popularly accepted, which includes Paleogene rifting and differential subsidence and Neogene post-rift thermal subsidence (Ye et al., 1985).

The TMR is a part of the Trans-North China Orogenic Belt, which is a Precambrian terrane composed mainly of Late Archean to Paleoproterozoic basement rocks, covered by the Early Paleozoic carbonates and shales, and the Late Paleozoic coal layer (Luo et al., 1997; Niu and Zhang, 2005). This Precambrian terrane was uplifted in the Mesozoic and Cenozoic as indicated by the dome-shaped tectonics, the emplacement of voluminous Late Mesozoic granites scattered in the north of Taihang Mountain, and the Tertiary and Quaternary alkali basalt and tholeiitic basalt volcanism situated in the Datong region. The previous geological investigations suggested that Taihang Mountain is an extensional orogen containing tectonic–magmatic belts and complex rock belts (Liu et al., 2000).

3. Velocity structure

3.1. Seismic data

Seismic data used in this study came from linearly aligned 45 digital broad-band seismic stations from the Northern China Interior Structure Project (Fig. 1) numbered from 151 to 196 (except 158) with an average spacing of ~10 km. The operation period was from September 2001 to February 2003, with most of the stations operating for longer than 12 months. Stations 181–196 were located on the bed-rock of the TMR, and hence called bed-rock stations. Other stations were within the BBB where relatively thick sediments are present, therefore called basin stations. The seismic observation equipment was supplied by the Seismic Array Laboratory of the Institute of Geology and Geophysics, Chinese Academy of Sciences. The majority of the stations (35) were equipped with CMG-3ESP sensors (50 Hz to 30 s) and RefTek 72A-08 data acquisition system. The remaining 10 stations were equipped with Chinese domestic three-component broadband sensors (30 Hz to 20 s) and 24-bit acquisition systems. During the experiment, over 200 teleseismic earthquakes with body wave magnitude ≥5.5 and epicenter distance between 30° and 90° were recorded. Back-azimuths of the events distributed asymmetrically within the range of 30°–330°, and mainly concentrate within the range of 110°–170° (Fig. 1).

The dense seismic station array and the property of near vertical incidence of the seismic body waves from teleseismic events provided high spatial resolution data for probing the crust and upper mantle structure. We calculated the receiver functions from these data using a time-domain Wiener deconvolution method in the same manner as described by Wu and Zeng (1998) and Ai et al. (2003). A time window of 20 s before and 80 s after the P wave arrival was selected to isolate the waveforms, and a Gaussian parameter of 5 and a water level of 0.001 were adopted in the deconvolution. For most of the stations more than 80 receiver functions with high signal-to-noise ratios were selected after careful visual inspection and sorted in the order of back azimuth.

According to the variation patterns with back-azimuth, for some stations nearly all the receiver functions were stacked together to highlight the key structural
Table 1
Seismic stations for which receiver function analysis and inversion were carried out

<table>
<thead>
<tr>
<th>Station</th>
<th>Longitude (E)</th>
<th>Latitude (N)</th>
<th>Number of receiver functions</th>
<th>Back azimuth range (°)</th>
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<td>37°36'01&quot;</td>
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<tr>
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<td>66</td>
<td>112–180</td>
</tr>
</tbody>
</table>

The number and back-azimuth range of the stacked receiver function for each station are also included.

characteristics beneath individual stations. For other stations possibly underlain by laterally varying shallow structures, the receiver functions were stacked azimuth-dependently and a stacked receiver function from the dominating back azimuth range was chosen as the observation data in the waveform inversion to avoid contamination of different structures. The station locations, the numbers and back-azimuth ranges of receiver functions involved in each stacking are listed in Table 1. The resultant 45 stacked receiver functions (Fig. 2) exhibit apparent waveform discrepancies between the bed-rock stations and the basin stations. While a distinct P-to-S converted phase from the Moho is observed explicitly in the mountain range, it could barely be detected in the basin due to heavy interference of strong reverberations produced by sedimentary layers with the Moho-induced
Fig. 2. Stacked receiver function for each station from west to east (196–151). Zero time corresponds to the P wave arrival. Amplitudes of the traces are normalized. Some station numbers are labeled to the right of the plots.

phases. The complexity and variations with stations of the observed receiver functions reveal the conspicuous lateral heterogeneity of the structure beneath the study profile.

3.2. CCP imaging

CCP imaging is a robust forward imaging method based on stacking Ps phases transformed from a velocity discontinuity. The validity of the time domain receiver function stacking profile is mainly dependent on the quality of the seismic data. Because the stacking profile in depth domain is constructed by time-to-depth conversion based on a defined velocity model, it is necessary to carefully assess the ability of stacking-based CCP imaging to constrain the depths of discontinuities in a region with strong heterogeneity.

We obtained CCP depth images along the studied profile with a stacking bin of 2 km wide (parallel to the profile), 14 km long, and 0.5 km high. The receiver functions obtained for individual earthquakes are binned according to their sampling points. Fig. 3a shows the CCP depth image constructed using an average crust-upper mantle model for the NCC (Ma et al., 1991; Gao et al., 1998b), in which no sedimentary structure is included.

In the section beneath the TMR we can clearly identify a strong positive phase at 40–50 km depth, with an apparent increase of depth from east to west. We interpret this as the conversion from the Moho of the TMR. The depth distribution, however, is apparently deeper than the previous estimation of 32–42 km from seismic refraction observations of the Xiangshui-Mandal Geoscience Transect (Ma et al., 1991), largely due to the incorrect crustal velocities used in CCP imaging. In the section beneath the BBB the conversion phases and multiple reverberations of the basin sediments are so strong that they dominate the top 20-km image. Again the improper velocity model fails to give a correct image of the sedimentary cover, and cannot restore the structural feature of the Moho and intra-crustal discontinuities, if any, beneath the BBB.

By superposing the reconstructed sedimentary structure of the BBB by Zheng et al. (2005) onto the average NCC model used in Fig. 3a, we reconstructed the crustal image of the study profile (Fig. 3b). All the signals are obviously shallowed, with the depth of the Moho beneath the TMR roughly coincident with the seismic refraction data and the sedimentary basement of the BBB identified at similar depths to that inferred previously (Zheng et al., 2005). Comparison of Fig. 3b with Fig. 3a indicates that considerable improvement in image quality has been achieved with the sedimentary structure correction. In addition, an intra-crustal interface can be continuously traced around the depths from 10 to 20 km, although the Moho beneath the BBB area is still invisible in Fig. 3b.

To further reduce the thick sediment influence and to extract the structural information for the Moho of the BBB, we performed CCP stacking on the PpPs multiples instead of the Ps phases. Positive signals are more coherently observed at about 30 km depth beneath the BBB in the PpPs image (Fig. 3c) compared with their Ps counterparts (Fig. 3a and b), possibly marking the Moho of the basin area.

It is not possible to unambiguously determine the nature of these strong phases and observed intra-crustal signals solely by CCP imaging. In the next section, we explore the detailed crustal structural features of the study region and evaluate the reliability of the CCP images from various velocity models through forward synthetic modeling and receiver function waveform inversion.

3.3. Receiver function waveform inversion

In both forward modeling and waveform inversion, synthetic receiver functions are constructed from synthetic seismograms in the same way as from real data.
The synthetics were calculated based on a reflection matrix method (Kennett, 1983). An adapted hybrid global waveform inversion method (Liu et al., 1995a,b; Ai et al., 1998) was used in the inversion process, which has proven to be a flexible and reliable way to extract structural information of the crust (Zheng et al., 2005). Similarly as described by Zheng et al. (2005), the objective function in the inversion is defined by the degree of fitting between the synthetics and the stacked receiver functions from the data for both waveforms and amplitudes. The best fitting model was searched from one-dimensional models parameterized as a stack of layers beneath each station. The model parameters include shear wave velocity $V_s$, thickness, and $V_p/V_s$ in each layer. Forward modeling shows that the waveforms of Ps phases are highly sensitive to these parameters. At the beginning of the inversion, we designed a stratified structure model based on the crust model from the Xiangshui-Mandal Geoscience Transect (Ma et al., 1991) and the sedimentary structure by Zheng et al. (2005). The subsurface structures are divided into 9–11 layers, including 3–5 layers of sedimentary cover, the upper, middle and lower crust, and two uppermost mantle layers overlying a half space.

For bed-rock stations, the Moho is initially taken as a first-order velocity discontinuity. In this case, the waveforms of the synthetic receiver functions show significant discrepancies with the observed ones (curve 3 versus curve 1 in Fig. 4a). The Ps phases converted at the Moho appear weaker and narrower, while the PpPs multiples are much stronger but with different waveforms compared with the real data. These differences indicate that gradual transitions rather than a sharp change may exist between the crust and the upper mantle beneath the TMR region. Based on the characteristics of the observed waveforms, crust–mantle transition zone models with gradually increased velocity (see Table 2) were tested by trial-and-error forward modeling. Consequently, the layer thicknesses were re-determined through the similar global inversion of the receiver functions for each station. Through these procedures, waveform matching between the synthetics and the data is remarkably improved, not only for the Ps phases but also for the PpPs multiples from the Moho (Fig. 4a).

At the stations filled by thick sediments, the waveforms of receiver functions within the first 5 s after the direct P arrival are dominantly controlled by the sedi-
Fig. 4. Examples of waveform comparisons between the observed receiver functions and synthetics for: (a) bed-rock stations and (b) basin stations. The station numbers are noted above the waveforms. Three grouped curves for each station are data (curve 1), synthetics from models with smooth crust–mantle transitions (curve 2), and synthetics from models with sharp crust–mantle transitions (curve 3), respectively.

imentary deposit (Fig. 2). It is difficult to extract the structural information of any intra-crustal discontinuity from the Ps phases except for the sediments. Employing the reconstructed sediment structure (Zheng et al., 2005) that well constrains the shallow section of the crust beneath the stations, we retrieved the structural features of the underlying crust and upper mantle through modeling the PpPs phases. The large difference between curves 3 and 1 in Fig. 4b again indicates that sharp velocity contrasts at the base of the crust would not be applicable, and

<table>
<thead>
<tr>
<th></th>
<th>$V_p$ (km/s)</th>
<th>$V_s$ (km/s)</th>
<th>Density (g/cm)</th>
<th>Thickness (km), BBB</th>
<th>Thickness (km), TMR</th>
<th>$V_p$ (km/s), searching range for uncertainty estimation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sediment</td>
<td>1.70–5.98</td>
<td>0.68–3.23</td>
<td>2.05–2.65</td>
<td>5.8</td>
<td>2.1</td>
<td></td>
</tr>
<tr>
<td>Upper crust</td>
<td>6.02</td>
<td>3.46</td>
<td>2.70</td>
<td>6.4</td>
<td>2.9</td>
<td>5.87–6.19</td>
</tr>
<tr>
<td>Middle crust</td>
<td>6.35</td>
<td>3.65</td>
<td>2.80</td>
<td>6.6</td>
<td>6.5</td>
<td>6.24–6.46</td>
</tr>
<tr>
<td>Lower crust</td>
<td>6.61</td>
<td>3.80</td>
<td>2.86</td>
<td>3.8</td>
<td>5.1</td>
<td>6.47–6.68</td>
</tr>
<tr>
<td>Transition zone</td>
<td>6.79</td>
<td>3.90</td>
<td>2.94</td>
<td>2.2</td>
<td>10.3</td>
<td>6.69–6.89</td>
</tr>
<tr>
<td>Uppermost mantle</td>
<td>7.57–8.01</td>
<td>4.30–4.50</td>
<td>3.19–3.36</td>
<td>–</td>
<td>–</td>
<td>6.90–7.10&lt;sup&gt;a&lt;/sup&gt;</td>
</tr>
</tbody>
</table>

<sup>a</sup> Searching for the layer with the P velocity of 6.96 km/s.
synthetics (curve 2) with gradual crust–mantle transition models fit the data better.

Due to the inherent trade-off between absolute seismic velocities and depths of discontinuities, the non-uniqueness of seismic inversion should be considered. The imaging results shown in Fig. 3 indicate that the CCP depth image is helpful to assess the validity of the velocity structure model adopted in CCP stacking, and thus can be utilized to constrain the model space in waveform inversion of receiver functions. Based on the independent constraints from CCP stacking and waveform inversion, the non-uniqueness and uncertainty of the results can be reduced. We implemented this idea in an iterating process seeking for the optimal consistency of CCP image and waveform inversion result. In CCP stacking, the inverted velocity model from the preceding waveform inversion was applied to construct the depth image. The signals that could be coherently detected beneath most of the stations in the image and appeared at depths (determined at the maximum amplitudes of the image) comparable to the layer interfaces defined in the inverted velocity model used were considered to represent real structures. The depth distributions of the identified discontinuities in the CCP depth images were then utilized to constrain the searching ranges of the corresponding layer thicknesses in the next inversion step; the inversion result was in turn applied as a new model to the CCP stacking of receiver functions. Such a two-step procedure was carried out iteratively until the inverted velocity model and the CCP depth image gave consistent depth distributions for all the identified discontinuities.

To illustrate this iterating process, we present examples of CCP images in Fig. 5 and plot the corresponding waveforms for some bed-rock stations in Fig. 6. For clarity, only depths of the interface between middle and lower crust and the Moho are marked with dots in the
Fig. 6. Waveform comparisons between the observed receiver functions and synthetics for some bed-rock stations. The station numbers are labeled above the waveforms. Grouped five curves for each station are data (curve 1) and synthetics from the inverted models same as those used in CCP stacking in Fig. 5: curves 2–5 from the models used in Fig. 5d, c, b, and a, respectively.

CCP images (Fig. 5). For all the inversion runs, the sedimentary structure from Zheng et al. (2005) was fixed to limit possible variations in other model parameters. In the first step, the global inversion was carried out in a wide searching range for each model parameter. The minimum thickness of a layer is set to be 0, while the maximum can reach up to 30 km. Shear wave velocity was searched from 3.0 to 5.0 km/s, and $V_p/V_s$ ratio was permitted to vary from 1.6 to 1.9. The 45 velocity models derived for individual stations were employed for time-to-depth conversion in CCP stacking (Fig. 5a and curve 5 in Fig. 6). In the second step, $V_s$ and $V_p/V_s$ in each layer were fixed to be the average values of all the stations obtained in the first step, and only the layer thicknesses are re-determined through similar global inversions across the array (Fig. 5b and curve 4 in Fig. 6). In the following step, model structures were modified for some stations when necessary based on forward modeling and CCP image analysis. For example, as mentioned before, a crust–mantle transition zone instead of a sharp Moho discontinuity was adopted in the inversion and the searching ranges of layer thickness were adjusted according to the CCP imaging results. The processes of waveform inversion and CCP stacking were carried out iteratively, and finally the optimal models were determined for all the stations, which provide the best fitting between synthetics and observations (cf., curve 2 with curve 1 in Fig. 6) and consistency between the CCP image and waveform inversion result (Fig. 5d).

For the basin stations, due to the strong sediment reverberations that significantly interfere with the Moho Ps phases, the PpPs multiples were used instead of Ps phases as the main criteria to determine the Moho depth. With a similar procedure to that described above, a continuous interface appears at ∼30 km depth beneath the BBB in the CCP image constructed from the best fitting model of waveform inversion (see Fig. 7b). We interpret it as the Moho of the BBB considering that it is consistent with the Moho observed at 30–34 km depths from the Xiangshui-Mandal Geoscience Transect. There is no significant change in PpPs amplitude of the Moho when we change the bin size from 2 to 8 km, supporting a flat Moho beneath the BBB.

3.4. Results and reliability analysis

We have combined CCP imaging and waveform inversion of receiver functions to extract the velocity structure information beneath the study region. The optimal CCP images using Ps phases and PpPs multiples are plotted in Fig. 7a (same as Fig. 5d) and Fig. 7b. Taking into account the Fresnel zones of ray paths, the best-fitting shear wave velocities derived for individual stations are compiled to form the final velocity image for the study profile (Fig. 7c). The synthetic receiver functions calculated from the best-fitting models are plotted in Fig. 8, superposed upon the observed data. The model parameters of the velocity structures above 50-km for
Fig. 7. CCP image using Ps phases (a) and PpPs phases (b), and the compiled shear-wave velocity structure (c) obtained from the optimal inverted velocity models. Some station numbers are labeled on the top of the image. The dots in the CCP images mark the depths of the interfaces between upper and middle crust (yellow dots), middle and lower crust (red dots), lower crust and crust–mantle transition zone (grass green dots), and the Moho (green dots) estimated from waveform inversion. The red arrow marks the boundary between the BBB and the TMR.

The BBB and the TMR are listed in Table 2 for comparison.

Three prominent features are particularly highlighted in the shear-wave velocity image. First, excluding the sedimentary cover which is up to about 10 km thick in the basin, the crust shows roughly a four-layered structure with Vs of $\sim 3.46$ km/s and Vp of $\sim 6.02$ km/s for the upper crust, Vs of 3.65–3.80 km/s and Vp of 6.35–6.61 km/s for the middle crust, Vs of 3.90 km/s and Vp of 6.79 km/s for the upper lower crust, and progressively increasing Vs from 4.0 to 4.2 km/s and Vp from 6.96 to 7.39 km/s forming a crust–mantle transition zone (see Table 2). Second, the whole crust thins eastward from about 40 km in the west to about 30 km in the east. The base of the upper crust is relatively flat in the BBB, but fluctuates in the TMR. The thickness of the middle crust ($\sim 10$–12 km) is nearly uniform in the basin and the mountain; however the lower crust and the crust–mantle transition zone exhibit large variations in thickness along the profile. Thirdly, pronounced changes in structural property of the crust occur rapidly across the North-South Gravity Lineament, including: (1) thin sediments ($\sim 2$ km) under the west TMR vs. thick sediments ($\sim 6$ km) beneath the east BBB; (2) a striking decrease in thickness of the lower crust from $\sim 10$ to $\sim 2$ km; (3) a decrease in thickness of the crust–mantle transition zone from $\sim 10$ to $\sim 5$ km.

The quality of our seismic imaging can be assessed by the following features. (1) The strongest and most continually traced phases all appear in the best-fitting model-based CCP image, compared with those from the middle-step velocity models (Fig. 5). (2) The depth distributions of the intra-crustal interfaces and the Moho derived from waveform inversion are generally in agreement with the strong phases imaged by CCP stacking (Fig. 7). (3) Both the waveforms and the amplitudes of the synthetics from the best-fitting models match the real data quite well (Fig. 8). The general consistency between the CCP images and the waveform inversion results demonstrates that most of the retrieved structural features of the study region are robust and reliable. To further strengthen our results and to estimate the uncertainties in model parameters, in particular considering the tradeoff between absolute velocities and depths, we
performed more extensive forward modeling based on various available velocity models. First, we estimated the uncertainty of $V_p$ in the crust (not including the sedimentary cover) by testing plenty of possible values in a broad range given by the P-wave velocity models for the nine broad tectonic units in China and the global continental crust (Gao et al., 1998b). For each layer we adopted 5 $V_p$ values within the given range (see

Fig. 8. Waveform comparisons of receiver functions from the data (solid lines) and the synthetics (dotted lines) for all the 45 stations. The station numbers are given to the left of the plots.
Table 2) and kept the $V_p/V_s$ ratio and the layer thickness the same as those in the optimal inversion model (Fig. 7c), resulting in a total of 3125 (=55) velocity models. Synthetic modeling showed that only two out of the 3125 models are acceptable in the sense that the misfit between the synthetics and the data did not exceed that of the optimal model for about 75% (~34) of the 45 stations, and no model could give a better fit for all the stations. The velocity difference between these two models and the optimal model shown in Fig. 7c defined a small uncertainty range of $V_p$ from $-0.8$ to $1.1\%$. Second, we explored the uncertainty of $V_p/V_s$ ratio in the crust by considering the average continental crustal Poisson’s ratio (0.253–0.279) (Christensen, 1996). The same number of models as above were constructed and tested except that $V_s$ were fixed and various $V_p/V_s$ values were considered. Again none of the models produced synthetics fitting the data better than the optimal model for all the 45 stations, although 32 models were accepted by the same rule as used for $V_p$ modeling. A maximum $V_p/V_s$ uncertainty of $-2.3–2.3\%$ was then obtained for different crustal layers. Accordingly, the depth uncertainty is estimated to be within 1 km for the intra-crustal and the Moho discontinuities. These synthetic tests therefore indicate the high reliability of the velocity models we have obtained.

4. Density structure

The density structure beneath the study region was constrained through modeling of the observed gravity anomalies. The gravity data come from the Xiangshui-Mandal Geoscience Transect, whose location is very close and parallel to that of the seismic imaging profile in this study (Fig. 1). As shown in Fig. 9, the observed Bouguer anomaly is characterized by a rapid gravity increase from $-125$ to $\sim 0$ mGal, mostly occurring over a width of about 100 km within the TMR. Across the North-South Gravity Lineament, only smaller gravity undulations not exceeding $\pm 30$ mGal are observed in the BBB.

The cause of the distinct gravity anomalies across the North-South Gravity Lineament is not immediately obvious. Neither the thick BBB sediments nor the topographical difference between the TMR and the BBB could be responsible for the gravity anomalies. Receiver function data are unable to provide stringent constraints on the density structure due to the insensitivity of Ps waveforms to the density variation over a discontinuity. However, with the general positive, proportional relationship between seismic velocity and density (Nafe and Drake, 1957; Berteussen, 1977; Carlson and Herrick, 1990), the capability of re-producing the observed gravity anomalies over the study region could be used as a criterion to assess the validity of the velocity model from our receiver function study.

We calculated the contributions from the crust and uppermost mantle to the Bouguer anomaly based on the optimal velocity model derived from receiver function inversion. The densities were estimated from $V_s$ and $V_p/V_s$ with an empirical velocity–density relationship: $\rho = 0.77 + 0.32V_p$ (Berteussen, 1977). The density of sediments was further modified according to the oil exploration data (Wu, 1987). The calculated Bouguer anomaly displays a rapid gravity rise of $\sim 100$ mGal from $\sim -100$ mGal in the east to $\sim 0$ mGal in the west across the North-South Gravity Lineament (thin grey curve in Fig. 9), which is the opposite to the observed gravity trend on the $\sim 100$ km-scale (black dots in Fig. 9). On the other hand, small-scale undulations especially in the BBB appear to resemble the data very well.

It is a rule of thumb that short wavelength gravity anomalies are caused by shallow density contrasts while long wavelength gravity anomalies are a function

![Fig. 9. Comparison of the predicted Bouguer gravity anomalies with the observation. Black dots are the observation records; grey curve with dot is obtained from the testing density model I; grey curve with cross is from the testing density model II; thin grey curve is obtained by a density model from receiver function inversion without any modification. Some station numbers are given on the top of the plot. The arrow marks the boundary between the BBB and the TMR.](image-url)
of broad density contrasts, either shallow or deep. The consistency of short wavelength variation between the gravity prediction and the observation therefore provide independent evidence for the correctness of the sedimentary and shallow crustal structure of the study region we have obtained. The long-wavelength misfit, however, likely results from some density difference between the west mountain range and the east basin area that may not be correctly accounted for in the above modeling due to either the inaccurate velocities, especially those of the upper mantle that could not be strictly constrained by our receiver function study, or the failure of the simple velocity–density relationship applied to the study region. We then adjusted the relative densities beneath the BBB and the TMR and performed forward modeling to find possible density models that give predictions close to the observed gravity data. By testing a large amount of models with different density modifications for the crust and upper mantle, we found only two models with which the calculated Bouguer gravity anomaly curves fit the data reasonably well (Fig. 9). Model I was obtained by reducing the density by 3% above 50 km beneath the TMR (not including the sedimentary cover) and keeping the densities of the BBB unchanged (grey dots in Fig. 9); model II (grey crosses in Fig. 9) was obtained by setting the subMoho upper mantle down to 100-km depth in the TMR 2% more buoyant than its BBB counterpart. In each case, the amplitude of the predicted Bouguer anomaly was adjusted as a whole by matching the prediction and the observation at a reference point, here the point with zero observed Bouguer anomaly, to account for possible unconstrained contributions for both the TMR and the BBB.

We prefer model II as the optimal density model for the study region, in light of the significant lithospheric thinning (60–80 km thick) and reworking that occurred in the east, while a thicker cratonic lithospheric root (>100 km) is preserved in the west of the NCC, as suggested by recent studies (Deng et al., 2004; Zheng et al., 2001). The chemical depletion of the cratonic lithosphere could presumably cause a density deficiency of 1–2.5% (Kaban et al., 2003). The temperature effect would be insignificant here given that ~100 mGal gravity variation takes place over ~100 km beneath the TMR (Fig. 9) and the temperature variation would be small within such a limited lateral extent. We did not consider any density contrast below 100-km depth because its contribution to the observed 100-km-scale gravity anomaly is minor at best. As for model I, the density variation in the crust dominated the predicted gravity curve. Although a 3% density difference is possible between different constituent rocks in the crust, the bulk density variation is small and a linear velocity–density fit is often observed (Christensen and Mooney, 1995). In this regard, model I may not be appropriate to represent the density structure of the study region without further evidence.

5. Discussion and conclusions

The receiver function study and gravity modeling results presented in this paper show an interesting present-day crustal and uppermost mantle structural picture beneath the NCC, which might have resulted from the secular evolution of the region, in particular affected by the Late Mesozoic thinning and Cenozoic adjustment of the lithosphere in eastern China (Menzies et al., 1993; Griffin et al., 1998; Fan et al., 2000; Zhang et al., 2004). Our structural image provides useful information and has broad implications for understanding the lithospheric evolution of the region.

(1) Significant crustal thinning from more than 40 km for a typical cratonic crust to ~30 km has been observed to the east of the North-South Gravity Lineament. The thinning was mainly achieved by the decrease in the lower-crust thickness in the east basin area with thinning less evident below the west mountain range (Fig. 7c). This structural feature is consistent with previous petrologic and geochemical studies in which crustal thinning, especially lower-crust thinning, has been invoked to explain the composition variations in the NCC (Gao et al., 1998a; Zhang et al., 2004).

(2) The teleseismic waveform data and the gravity observation suggest a thicker crust and a buoyant mantle lithosphere beneath the TMR compared with the BBB, in agreement with the presence of a thicker cratonic root preserved in the west and dramatic lithospheric thinning and reworking having occurred in the east of the NCC. The distinct crustal and lithospheric mantle properties of the TMR and the BBB might have been associated with different tectonic processes and deformation regimes of the lithosphere and upper mantle on the two sides of the North-South Gravity Lineament.

The lower crust including the crust–mantle transition zone shows same variation in thickness with the whole crust beneath both the mountain and the basin areas in our seismic image (Fig. 7c), indicating that the lower crust may have played the most important role in the crustal evolution of the NCC. The presence of a thick lower crust of ~20 km in the west TMR (Table 2) corroborates petro-
logistic and geochemical observations and speculations that significant underplating of mantle-derived magma and its interaction with the continental lower crust have occurred in the late Mesozoic in this region and presumably been responsible for the vertical accretion of the lower crust and the formation of the crust–mantle transition zone (Zhou et al., 2002; Liu et al., 2001; Fan et al., 1998). In the east BBB, on the other hand, the relatively thin lower crust (<8 km, Table 2) suggests that magma underplating and crust–mantle interaction may not have taken place to the same extent as that which occurred beneath the TMR. This is also consistent with the rapid ascending of lavas, a process unlikely accompanying significant crust–mantle interaction, during the peak volcanism period of late Mesozoic observed in this area (Zhang and Sun, 2002; Zhang, 2005).

The difference in structural feature between the west TMR and the east BBB strongly suggest that the shallow crust and the deep lithospheric mantle might have tectonically coupled on both side of the North-South Gravity Lineament, but the lithosphere as a whole, probably also including the underlying upper mantle, appears to have mechanically decoupled between the west and the east sides. This may also be hinted by several other contrasting west–east differences across the North-South Gravity Lineament. For example, it was reported that the east BBB was characterized by widespread NW–SE tectonic extension from Late Mesozoic to Early Cenozoic (Ren et al., 2002), whereas the west TMR region experienced dominant NW–SE oriented compressional deformation in the lithospheric processes, as suggested by a recent regional study on seismic anisotropy of the upper mantle (Zhao and Zheng, 2005). The North-South Gravity Lineament, marked by different crustal and uppermost mantle structural features (Fig. 7), distinct surface topography (Fig. 1) and depth of the lithosphere (Deng et al., 2004), sharp gravity contrast (Fig. 9), and different mantle deformation patterns on its western and eastern sides, is therefore possibly not only a surficial geological setting but also a deep intra-continental boundary possibly developed in association with deep mantle processes of the eastern China continent.

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