3-D velocity model beneath the Middle–Lower Yangtze River and its implication to the deep geodynamics

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A R T I C L E   I N F O

Article history:
Received 30 August 2012
Received in revised form 28 February 2013
Accepted 18 March 2013
Available online 28 March 2013

Keywords:
Middle–Lower Yangtze River
Teleseismic tomography
Mineralization mechanism
Upwelling asthenosphere
Detached lithosphere

A B S T R A C T

A mineralization zone exists in the Middle and Lower Yangtze River (MLYR) region. Previous studies have shown that this zone might be caused by the rich ejection of magma in the Mesozoic. We have applied the teleseismic tomography method to determine a 3-D P-wave velocity structure of the mantle down to 500 km depth beneath this region by using 14,740 P-wave arrival times collected from 519 teleseismic events recorded by 46 portable and 47 static seismic stations located in this region. The relative residual times used for the tomography are calculated by the modified multi-channel cross-correlation method which increases not only the efficiency but also the data precisions up to 0.01 s. The grid space is set as 1° × 1° horizontally and 50–100 km vertically. Our tomographic results show that the lithosphere with high velocity anomalies is separated into two parts: one locates above the depth of 100 km and the other at the depths between 250 km and 400 km, and the asthenosphere with low velocity anomalies locates between these two parts of lithosphere, which indicates that the lithosphere delaminated together with the asthenosphere upwelling. The detachment of lithosphere might relate to the subduction of the western Pacific slab. Therefore, our study provides clear evidence to geoscientists for understanding the deep dynamic process beneath this region and a three-stage geodynamic process is developed based on the new evidence.

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1. Introduction

The Middle–Lower Yangtze River (MLYR) region locates in the east of China where more than 200 kinds of polymetallic (Cu, Au, Mo, Fe, Zn, Pb, Ag and so on) deposits exist (Mao et al., 2006; Pan and Dong, 1999). Most deposits are clustered in such mineralization zones as Ningzhen, Ningwu, Luzong, Tongling, Anqing, Guichi, Jiurui, Edongnan and so on (Fig. 1). In these zones, the period of mineralization is different, for example, the orebodies within Jurassic–Cretaceous granitoid intrusions, the skarn orebodies related to the late Mesozoic intrusions and the late Paleozoic–early Mesozoic carbonates, and the stratiform massive sulfide orebodies in the late Paleozoic–early Mesozoic sedimentary strata (e.g. Chang et al., 1991; Chen et al., 2006; Mao et al., 2006; Pan and Dong, 1999; Yang et al., 2011). In the view of geology, the mineralization zone of MLYR is restricted by several deep faults, mainly including the Xiangfan–Guangji, Yangxin–Changzhou, and Tancheng–Lüjiang (Tanlu) faults (Fig. 1), in which the Tanlu fault may run through over the crust (Chen et al., 2006), and looks like a reversed “L” shape.

Why does such narrow zone contain so rich mineral resources? What is the deep dynamics? How does the deep process control the derivation, the evolution and the intrusion of magma for mineralization? It is necessary to explore the deep structure to find out the law of intracontinental mineralization.

In the map of Bouguer gravity anomalies (Lü et al., 2005), the positive anomaly within the mineral zone may be caused by the rise of mantle, which is consistent with the results of deep seismic reflection (Lü et al., 2005) and receiver function (Chen et al., 2006). The isotope geochemical and petrological studies suggest that many intermediate-acid intrusive rocks in the MLYR region have similar geochemical characteristics to adakitic rocks and these adakite-like rocks are most likely derived from the partial melting of the thickened lower crust rather than the partial melting of subducted oceanic crust (Lü et al., 2005; Xu et al., 2002). Some deep seismic reflection profiles show that a layered lower crust with strong reflective exists beneath the MLYR region. Lü et al. (2005) explained this strong reflection as the underplating basic or ultrabasic rocks under the compress condition, which coincides with the opinion of Yang and Wang (2002) as studying the Sulu area. These evidences indicate the underplating of magma is very normal in the mineralization zone of MLYR.

The result of receiver function migration reveals a 60–80 km thick lithosphere thinned from about 180 km thick in the Paleozoic right below the Tanlu fault zone, which provides very significant evidence for lithosphere thinning (Chen et al., 2006). Some geochemical
studies also make us believe that the Tanlu fault zone plays an important role in the lithosphere thinning during the Mesozoic–Cenozoic (e.g. Xu, 2001; Xu et al., 2004).

At present, many dynamic models for interpreting the tectonic evolution in the MLYR have been proposed, such as the exhumation of ultra-high-pressure (UHP) metamorphic zone (Okay and Sengör, 1992), the continent indentation model (Yin and Nie, 1993), the crustal-detachment model (Li, 1994) and the underplating model (Lü et al., 2005) and so on. Because the Tanlu fault spatially adjoins the MLYR on the east and the UHP on the west, therefore these three geological structures should have common spatial and temporal origination in the same geodynamic system. Most researchers coincide that the lithosphere beneath the MLYR was thicker in the Paleozoic (about 180 km thick) and then thinned because of its detachment and/or the basalt underplating, but argue about the deep mechanism of thickening, the manner of detachment, and the relation between the magmanism and underplating.

In the present work, we have used the teleseismic tomography to determine the 3-D velocity structure down to 500 km deep beneath the mineralization zone. Our results provide new evidence for the interaction between the asthenosphere and the lithosphere. A possible geodynamic model in the upper mantle is discussed finally.

2. Data

We have used teleseismic data recorded by 46 portable and 47 static stations located in the Middle–Lower Yangtze River to study the deep velocity structure (Fig. 2). The portable stations are installed by the Institute of Mineral Resources, Chinese Academy of Geological Sciences, for the SinoProbe Project from November 2009 to August 2011. The spatial interval between stations is about 5 km. All portable stations are aligned along a line which is almost perpendicular to the Tanlu fault (Fig. 2). The data recorded by the static stations from September 2007 to April 2011 are provided by the Data Management Centre of China National Seismic Network at Institute of Geophysics, China Earthquake Administration (Zheng et al., 2010). All static stations are mainly located in Anhui and Jiangsu provinces (Fig. 2). We selected earthquakes with epicentral distances between 30° and 90°, and magnitudes larger than M 5.0. As a result, 519 teleseismic earthquakes are chosen (Fig. 3). The sample frequencies of both data are same as 100 Hz, in other words, the sample interval is 0.01 s.

For teleseismic events, the P-wave arrival times are usually picked manually and then are used to calculate the relative residuals (Zhao et al., 1994), which is defined as

\[ r_{ij} = t_{ij} - \bar{t}_i \]

where \( r_{ij} \) is the travel time residual, \( t_{ij} \) is the observed travel time at the \( i \)th station, and \( \bar{t}_i \) is the mean residual. Generally, the quantity of processed waveform data is too large to use hand-picking way to identify the arrival time. To improve the efficiency and the data precision, a new method is proposed to automatically calculate the relative residuals from the original teleseismic waveforms. The principle of our new method is based on the multiple-channel cross correlation method ("MCCC") proposed by VanDecar and Crosson (1990).

Suppose we have \( M \) waveforms in which the \( j \)th and \( k \)th waveforms, \( S_j(t) \) and \( S_k(t) \), are shown in Fig. 4. Then the offset time between these two waveforms could be calculated according to the cross-correlation formula

\[ C_{jk}(\tau) = \frac{1}{N} \sum_{n=1}^{N} S_j(t + n \cdot \delta t - \tau) \cdot S_k(t + n \cdot \delta t) \]  

where \( \delta t \) is the sample interval of waveforms, here is 0.01 s; \( \tau \) is the offset time, which is changed from −2.0 to 2.0 s in our work; \( t \) is the beginning...
time of time window. Fig. 4 shows the curve of cross-correlation values to the offset time. According to the peak value in Fig. 4, it is very easy to determine the optimal offset time as 1.28 s. For all M data, we can obtain \((M - 1)!\) offset times of any two waveforms and generate \((M - 1)!\) equations in which the arrival times are unknown. Therefore, all arrival times could be inverted from these equations (VanDecar and Crosson, 1990) and are converted into the travel times after subtracting the original times. And then, for teleseismic events, the relative residuals could be calculated since combining with the synthetic travel times (Zhao et al., 1994). During this procedure, the precision of calculation for relative residuals is largely influenced by the inverse precision of arrival times. As we know, the inverse problem mentioned above is generally over-determined and is difficult to find a unique solution. In this work, we proposed a new method to obtain the relative residual directly from the offset times without any inverse calculation.

Assuming, the jth waveform recorded at the ith station is chosen as the reference and all offset times relative to this reference are named as \(Y_{ij}; j_{1}, \ldots, Y_{ij}; j_{M} \) where the zero value represents the offset time of the jth waveform relative to itself, and \( Y_{ij}; j = T^{\text{sym}}_{ij} - T^{\text{obs}}_{ij} \).

We can also get the offset times between the synthetic arrival or travel times, named as, \( Y_{ij}; j_{1} \ldots Y_{ij}; j_{M} \) where the zero value represents the offset time of the jth waveform relative to itself, and \( Y_{ij}; j = T^{\text{sym}}_{ij} - T^{\text{obs}}_{ij} \).

Now, according to the above formulas, we can derive a new formula to calculate the relative residuals \( y_{ij} \) for teleseismic data as following

\[
y_{ij} = 1 \sum_{k=1}^{M} \left( Y_{ijk}^{\text{obs}} - Y_{ijk}^{\text{syn}} \right)
\]

\[
y_{ij} = 1 \sum_{k=1}^{M} \left( T_{ij}^{\text{obs}} - T_{ij}^{\text{syn}} \right)
\]

\[
y_{ij} = 1 \sum_{k=1}^{M} \left( T_{ij}^{\text{obs}} - T_{ij}^{\text{syn}} \right)
\]

The result induced from Function (3) is entirely same with that from Function (1), that’s meaning, we find another new way to calculate the relative residual if we know the offset times \( Y_{ijk}^{\text{obs}} \) and \( Y_{ijk}^{\text{syn}} \) rather than the arrival times. As mentioned above, \( Y_{ijk}^{\text{obs}} \) and \( Y_{ijk}^{\text{syn}} \) could be easily obtained directly from the waveforms.

To process the lower signal-to-noise waveforms, we introduced the phase information into the calculation of cross-correlation according to both the principle of the phase-weighted stacking (Schimmel and Paulssen, 1997) and the idea of geometric normalized

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**Fig. 2.** Map of station distributions and the study area with surface topography. Red squares and blue triangles represent the static and the portable stations, respectively. The yellow dashed line indicates the boundary of mineralization zone. The black curve lines denote the provincial boundaries. The inset map shows the location of the present study area.
cross-correlation (Schimmel, 1999). Therefore, function (2) is modified as

$$ C_{mmcc}(\tau) = \frac{\sum_{k=1}^{N} S_j(t + k \cdot \delta t - \tau) \cdot S_k(t + k \cdot \delta t) \cdot \cos \left( \phi_j(t + k \cdot \delta t - \tau) - \phi_k(t + k \cdot \delta t) \right)}{\left[ \sum_{k=1}^{N} S_j(t + k \cdot \delta t - \tau) \right] \cdot \left[ \sum_{k=1}^{N} S_k(t + k \cdot \delta t) \right]^2} $$

where $\phi_j$ and $\phi_k$ represent the instantaneous phases of the jth and the kth waveforms, respectively; mmcc in the subscript means that the cross-correlation is calculated by the modified multiple-channel cross correlation method ("MMCC"); $\nu$ in the index denotes the phase weight.

To verify the feasibility and the availability of our new method, two kinds of synthetic tests are carried out. The first test is about the “pure” waveform without any noises (Fig. 5a). There are a total of 16 teleseismic traces in which the synthetic arrival times of P phases and the residual times are random values which are shown in the 2nd and the 3rd column of Table 1, respectively. Then the “observed” arrival time could be estimated according to these two columns. We have used functions (1) and (3) to calculate the relative residuals which are shown in the 4th and 5th columns of Table 1, respectively. During the calculation, the phase weight ($\nu$) is assigned as 2.0. The root mean square (RMS) value of differences between these two columns is 0.004 s, which indicates that these two results are almost identical, meaning that our new method works very well.

The second test is about the anti-noise ability of our new method. We generated the “noisy” waveforms by adding some random values, satisfying the normal distribution with the mean value of zero and the variance of 1.0, into the “pure” waveforms (Fig. 5c). From the “noisy” waveforms, it is very difficult to identify the arrival time by eyes. The MMCC method is used to estimate the relative residuals shown in the 6th column of Table 1. Here the phase weight is also equal to 2.0. The result shown in the 7th column is obtained when the phase weight is zero (not considering the phase information). Comparing the RMS value of differences between the 6th and 7th columns with the 4th columns, respectively, we can conclude that the
phase information plays an important role in improving the precise of relative residuals, especially in the case with lower or weak signal-to-noise. In addition, to directly illustrate the validity of results in the above two tests, all waveforms are migrated according to their corresponding offset times. Fig. 5b and d shows that the P phases after migration could be aligned consistently.

The MMCC method is used to process all teleseismic waveforms to obtain the relative residuals. Before conducting the tomography, we simply analyze the velocity structure beneath the study region according to the average relative residuals at each station, which could be gained as following

\[ r_i = \frac{1}{n_i} \sum_{j=1}^{n_i} r_{ij} \]

where \( n_i \) is the number of data observed at the \( i \)th station. Fig. 6 shows the distribution of average result. There exist early arrivals at stations to the northwest of mineralization zone, and late arrivals to the northeast. The average relative residuals are very weak at stations within this zone. We magnify the average relative residuals 2 times at portable stations, and then see that the distribution of early and late arrivals is alternate. All patterns represent that the heterogeneity beneath the study region might be very strong.

### 3. Method and results

After obtaining the relative residuals, we have used the teleseismic tomography method (Zhao et al., 1994, 2006) to determine the 3-D velocity structure extending to 500 km deep beneath the MLYR region. During the inversion, the iasp91 model (Kennett

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**Table 1**

An example for testing the MMCC method.

<table>
<thead>
<tr>
<th>No.</th>
<th>Synthetic arrival time (s)</th>
<th>Synthetic residual (s)</th>
<th>Relative residual1 (s)</th>
<th>Relative residual2 (s)</th>
<th>Relative residual3 (s)</th>
<th>Relative residual4 (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>444.838</td>
<td>0.485</td>
<td>-0.454</td>
<td>-0.46</td>
<td>-0.42</td>
<td>-0.46</td>
</tr>
<tr>
<td>2</td>
<td>443.328</td>
<td>0.724</td>
<td>-0.216</td>
<td>-0.22</td>
<td>-0.26</td>
<td>-0.08</td>
</tr>
<tr>
<td>3</td>
<td>458.794</td>
<td>0.981</td>
<td>0.041</td>
<td>0.04</td>
<td>0.02</td>
<td>0.01</td>
</tr>
<tr>
<td>4</td>
<td>462.642</td>
<td>1.576</td>
<td>0.636</td>
<td>0.64</td>
<td>0.59</td>
<td>0.53</td>
</tr>
<tr>
<td>5</td>
<td>436.463</td>
<td>1.113</td>
<td>0.173</td>
<td>0.18</td>
<td>0.27</td>
<td>0.35</td>
</tr>
<tr>
<td>6</td>
<td>446.174</td>
<td>1.613</td>
<td>0.674</td>
<td>0.68</td>
<td>0.68</td>
<td>0.78</td>
</tr>
<tr>
<td>7</td>
<td>441.086</td>
<td>0.176</td>
<td>-0.764</td>
<td>-0.77</td>
<td>-0.73</td>
<td>-0.88</td>
</tr>
<tr>
<td>8</td>
<td>443.722</td>
<td>1.302</td>
<td>0.362</td>
<td>0.36</td>
<td>0.38</td>
<td>0.45</td>
</tr>
<tr>
<td>9</td>
<td>438.948</td>
<td>0.493</td>
<td>-0.447</td>
<td>-0.45</td>
<td>-0.53</td>
<td>-0.57</td>
</tr>
<tr>
<td>10</td>
<td>447.489</td>
<td>0.262</td>
<td>-0.678</td>
<td>-0.68</td>
<td>-0.71</td>
<td>-0.65</td>
</tr>
<tr>
<td>11</td>
<td>451.538</td>
<td>0.17</td>
<td>-0.77</td>
<td>-0.77</td>
<td>-0.67</td>
<td>-0.62</td>
</tr>
<tr>
<td>12</td>
<td>449.816</td>
<td>1.022</td>
<td>0.082</td>
<td>0.08</td>
<td>0.10</td>
<td>0.05</td>
</tr>
<tr>
<td>13</td>
<td>442.002</td>
<td>1.843</td>
<td>0.904</td>
<td>0.91</td>
<td>0.90</td>
<td>0.97</td>
</tr>
<tr>
<td>14</td>
<td>446.852</td>
<td>0.764</td>
<td>-0.175</td>
<td>-0.17</td>
<td>-0.24</td>
<td>-0.31</td>
</tr>
<tr>
<td>15</td>
<td>452.003</td>
<td>1.431</td>
<td>0.492</td>
<td>0.49</td>
<td>0.45</td>
<td>0.38</td>
</tr>
<tr>
<td>16</td>
<td>437.211</td>
<td>1.08</td>
<td>0.14</td>
<td>0.14</td>
<td>0.14</td>
<td>0.05</td>
</tr>
<tr>
<td>RMS</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
<td>/</td>
</tr>
</tbody>
</table>

\(^a\) In the 6th column, considering the phase information, \( \nu = 2 \).  
\(^b\) In the 7th column, not considering the phases, \( \nu = 0 \).
and Engdahl, 1991) is chosen as the 1-D velocity model and the grid spacing is determined according to the checkerboard test. Fig. 7 shows the optimal grid set in which the spacing is 1° × 1° in the horizontal and 50–100 km in the vertical. Velocity perturbations based on the 1-D model at each grid node are considered as unknown parameters. A 3-D ray tracing technique is used to compute ray paths and synthetic travel times (Zhao et al., 1992). As a result, a large and sparse system of observation equations can be constructed which relates the observed relative residuals to the unknown velocity parameters and are resolved by using a conjugate-gradient algorithm LSQR (Paige and Saunders, 1982) with damping and smoothing regulations (Zhao, 2001, 2004). The station elevations are taken into account in the 3-D ray tracing and inversion.

Fig. 8 shows the plan views of our optimal tomographic results. The velocity perturbations are changed from −2% to +2%. At depth of 50 km (Fig. 8a), an obvious high-velocity (high-V) belt exists beneath the locations of portable stations and two low-velocity (low-V) zones locate to the north and the south of high-V belt. According to the resolution test at this depth (Fig. 7a), we can infer that the high-V anomaly is just right beneath the portable stations where the teleseismic rays cross better than other places at shallow depth. At depths from 100 to 200 km, the low-V anomalies denote the primary feature (Fig. 8b and c). The surface-wave tomography also shows a low-velocity zone at the depth from 85 to 250 km in the eastern part of East China (Zhu et al., 2002) as well as our result. As the depth increasing, the high-V anomalies replace the low-V ones beneath the center of our study region at 300–400 km deep (Fig. 8d and e). When the depth is up to 500 km, the low-V anomalies appear again (Fig. 8f). Comparing with the resolution results, the distribution of velocity anomalies is reliable.

To understand the velocity distribution pattern better, we design two cross-section lines: one is along the locations of portable stations (profile AA′) and the other is parallel to AA′ but locates to the northeast of AA′ (profile BB′). Both profiles are perpendicular to the Tanlu Fault (Fig. 9a). Fig. 9b and c shows the cross-sections along AA′ and BB′, respectively, in which the feature of velocity anomalies is presented more clearly. There are two high-V layers and one low-V layer. The upper high-V layer with thickness of about 80 km may represent the present lithosphere which might be thinned from about 180 km thick in the Paleozoic (Chen et al., 2006). The lower high-V layer is dipping from 200 km to 400 km deep, which could be interpreted as the detached lithosphere or the metasomatism layer. The low-V layer locating between these two high-V layers looks like a wedge plugging into the separated lithospheres, which may be explained as the hot asthenosphere.

To further verify the reliability of velocity anomalies, we have conducted the restored test. In this test, the velocity model inverted from the real data is considered as the initial model to generate the synthetic travel times, and then the synthetic travel times are inverted to output a new velocity model. If the output model is similar with the initial model, then this test could prove that the inverted model by the real data is convincing. Fig. 10 shows the restored results. Comparing Fig. 9 with Fig. 10, the primary feature is restored very well except the amplitude of anomalies, which denotes that the spatial pattern of velocity anomalies is reliable. Based on this test and other geoscientific results, we have enough confidence to explain the results geologically: the low-V zone represents the upwelling asthenosphere and the high-V zone at the lower (deep) part denotes the detached lithosphere or the metasomatic material.
4. Discussion

Our tomography results suggest three features which are (1) the lithosphere is thinned and delaminated, (2) the detached part has sunk down to the upper boundary of the mantle transition zone, and (3) the upwelling asthenosphere filled in the gap left by the detached lithosphere. Wu and Sun (1999) attributed the thinning of lithosphere to the consumption of huge ejection of magma in the

![Fig. 7. Results of a checkerboard resolution test. Open and solid circles denote fast and slow velocity perturbations, respectively. The scale is shown at the bottom. The layer depth is shown at the top of each subfigure.](image)
Mesozoic. We think that it is possible to consume the lithosphere in some degree as the ejection of magma, but we argue against that the lithosphere with 100 km thick (>180 km thick in the Paleozoic, and now just ~80 km thick) were completely transformed into the magma and then be ejected because the crust is not obviously thickened. According to our tomographic results, the average thickness of deep high-V layer is about 100 km, which rightly equals the thickness of disappear lithosphere. Therefore, it is reasonable to interpret the deep high-V zone as the detached lithosphere. Combining our results with other previous geoscientific studies, we can further discuss what caused the detachment of lithosphere and how the deep tectonic process controlled the rich mineralization in the MLYR.

There are some possible mechanisms to cause the thinning of lithosphere, such as lithosphere root and de-rooting (Deng et al., 1994), thermal-tectonic destruction/chemical erosion (Xu, 1999), upwelling-erosion-replacement of deep and hot material (Lu et al.,

Fig. 8. Plan views of P-wave velocity images obtained in this study. Red and blue colors denote the slow and fast velocities, respectively. The depth of each layer is shown at the upper right corner of each subfigure. The perturbation scale is shown at the bottom.
the abrupt delamination (Xue et al., 2010), and so on. Based on our results, we favor the delamination as the thinning mechanism of lithosphere. The asthenospheric material would upwell as the lithosphere delaminating and filled the gap left by the detached lithosphere. When the upwelling material reached the bottom of lower crust of MLYR along the existing lithospheric weakness, huge heat energy provided by the asthenospheric flow caused a large-scale melting of the lower crust and then formed the mixed magma. The ore-bearing magmatic rocks are generally adakitic-like with high K₂O and MgO contents which indicates that these rocks might be derived from the partial melting mafic materials at the base of the continental crust contaminated by the mantle component (Xu et al., 2002). In addition, we can estimate the time of lithosphere detachment according to the age of ejected magma.

There occurred two intense activities of post-Paleozoic magmatism in Eastern China. The first one was from the Middle Jurassic to the Late Cretaceous (Zhang et al., 2002, 2004) and the second one occurred in the Cenozoic (e.g., Li, 2000; Zhi et al., 1990). The geochronological data from the Late Mesozoic Cu–Au–Mo-related magmatic rocks suggest that the peak ages of magmatic event in Jiurui, Edongnan–Tongling, and Luzong–Ningwu districts are successively younger between 148 and 125 Ma range (Yang et al., 2011). Therefore, the time when the lithosphere began to be detached should be earlier than the first episode of magmatism. Some researchers consider that these magma activities

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**Fig. 9.** Vertical cross-sections of P-wave velocity images. (a) The locations of the profile lines AA′ and BB′. The white triangles denote the portable stations. The blue thin lines illustrate the faults. The yellow thick lines represent the Tanlu fault and the Yangxin–Changzhou Fault (YCF). (b) Along the profile line AA′. The inferred geological structures are surrounded by dashed lines. Red and blue colors denote slow and fast velocities, respectively. The horizontal dashed line represents the 410-km discontinuity (the upper boundary of mantle transition zone). (c) Similar to (b), but along the profile line BB′.
might be caused by the subduction of paleo-Pacific plate (Li and Li, 2007; Yan et al., 2008; Yang et al., 2011). When the paleo-Pacific plate subducted into the Yangtze craton, it was firstly blocked off by the margin of the Dabie orogenic belt near the Jiurui district which triggered the first intense magmatic event. However, some researchers argue against the triggering of paleo-Pacific plate (Ren et al., 1998; Wang and Mo, 1995) because they consider that the paleo-Pacific plate was located 800 km far away from the MLYR region during the Late Mesozoic, therefore it is impossible for mantle wedge metasomatism to extend over such a large distance. Even though, in our opinion, the paleo-Pacific plate played an important role in the tectonic framework, even controlled the magmatic genesis and the tectonic evolution of MLYR region in the Late Mesozoic because it is the only known subduction event in the Asian region at that time.

In summary, we consider the delamination as the thinning mechanism of lithosphere. So then, what caused the delamination? To answer this question, the geodynamic process should be described. Simply, the whole geodynamic evolution of the MLYR region could be separated into three stages from the Mesozoic to the Quaternary, as well as proposed by Lü et al. (2005).

1. Collision (from the late Permian to the early Jurassic or the Triassic). The North China Block (NCB, referring to the Sino-Korean Block) and the South China Block (SCB, containing the Yangtze Block and the Cathaysian Block) began to collide in the eastern part of the NCB during the early Triassic and the QinLing–Dabie Ocean closed initially (Fig. 11a). The large-scale collision occurred from the late Triassic to the early Jurassic (Fig. 11b–c). During this period, the paleo-Pacific plate began to subduct into the Eurasian plate (Ichikawa et al., 1990) and provided a NW compressional force which urged the SCB to rotate clockwise.

2. Extension and delamination (from the early Jurassic to the middle Cretaceous). As the relative rotation between the NCB and the SCB, there appeared the extensional zone and the compressional zone (Fig. 11c–d). The gradual extension might produce a horizontal shear stress which led the lithosphere to be delaminated. As a result, the ancient thicker lithosphere might be separated into two parts: one was reserved as the present lithosphere with thickness of 60–80 km (Chen et al., 2006) and the other was detached and has sunk down to the upper boundary of mantle transition zone until now (Fig. 9). The space left by the detached lithosphere was filled by the upwelling asthenosphere which might cause the strong underplating in the lower crust. With the rise of temperature in the lower crust, partial melting took place and the molten magma was intruded upward along the strike-slip fault such as the Tanlu fault. This magma ejection is just called the first episode of magmatism mentioned above. Although Lü et al. (2005) had already realized the reaction between the delamination of lithosphere and the upwelling asthenosphere, they do not know what happened to the detached lithosphere because they could not image the deep velocity structure. Our tomography imaging can well make up the geodynamic model proposed by Lü et al. (2005), which makes the evolution stage more completely during this period.

3. Late-stage extension and stability (from the Cretaceous to the Quaternary). The Qinling–Dabie Ocean closed completely and the suture zone between the NCB and SCB generated along the Qinling–Dabie orogenic belt. The magmatism became weak. The present tectonic framework formed in the MLYR region. In these three stages, the delamination expresses a crucial effect to the mineralization of the MLYR region. Our tomographic result provides reliable seismic evidence to support this delamination mechanism.
5. Conclusions

In the present work, we applied a 3-D teleseismic tomography method (Zhao et al., 1994, 2006) to P-wave arrival time data collected from digital seismograms of 519 teleseismic events recorded by the portable stations (supported by the “SinoProbe” Project in China) and the static stations locating in Anhui and Jiangsu Provinces. To obtain the relative residuals with higher precision, an improved method is proposed based on the multi-channel cross correlation method (VanDecar and Crosson, 1990) to directly calculate the relative residuals from the waveforms. The advantage of this new method is to avoid an overdetermination problem inverted, which increases the calculation precision. Our tomographic results show that two high-V zones associated with the separated lithospheres and a low-V zone related to the upwelling asthenosphere. The shallow part represents the thinned lithosphere and the deep part denotes the detached lithosphere.

The feature of velocity anomalies leads us to infer that the thinning mechanism of lithosphere might be the delamination which was related to the subduction of the paleo-Pacific plate. Based on the three-stage geodynamic model proposed by Lü et al. (2005), we have supplemented the detached process during the early Jurassic to the middle Cretaceous. The detached lithosphere has reached the bottom of the upper mantle which is never mentioned before. Accompanying with the descent of detached lithosphere, the asthenospheric materials ascended. The upwelling asthenosphere caused a larger-scale melting with the descent of detached lithosphere, the asthenospheric materials might provide the heat source for forming the mixed magma at the lower crust. Therefore, the rich mineralization of MLYR region is closely connected with the delamination of lithosphere in the Mesozoic.

Acknowledgments

We thank the Data Management Centre of the China National Seismic Network at the Institute of Geophysics, China Earthquake Administration, and Institute of Mineral Resources, Chinese Academy of Geological Sciences, for providing the waveform data at static and portable stations. We also thank two anonymous reviewers for their comments and suggestions. This research is co-supported by the Sino-Probe project in China (SinoProbe-03), the National Science Foundation of China (Grants 40930418, 40904021 and 40874067), and the Fundamental Research Funds for the Central Universities (2010ZD09 and 2012035). Most figures are made by using GMT (Wessel and Smith, 1998).

References


