P- and S-wave Tomography of the Crust and Uppermost Mantle in China and Surrounding Areas

by

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Abstract

This thesis involves inverting the seismic structure of the crust and uppermost mantle in China from the P- and S-wave travel-time tomography. The main contributions of this research are: 1) introducing the adaptive moving window method to obtain 2338 1D P and S models in China; 2) introducing a tomographic method to perform the 3D body wave travel-time tomography with the Moho discontinuity included. Both horizontal and vertical resolutions are highly controlled and smooth transitions among adjacent locations are guaranteed in the final models. To achieve these objectives, the Monte-Carlo (random search) method and the Gauss-Newton method are applied iteratively to find the nonlinear least square solutions and to optimize the models in the crust and uppermost mantle. The models we obtained provide accurate travel-time calculation, ground-truth event relocation and seismogram fittings. These models can therefore be applied to reliable earthquake location.

Geological, geodynamic, and volcanic implications of our models are discussed in this thesis. Our tomographic models provide new insights into the geological structure and tectonics of the region, such as lithological variations and large fault zones across the major geological terranes. Compared with previous tomographic studies, we have used a larger, higher quality data set and applied an updated tomographic method to take into account the effects of the complex Moho geometry in this region. Our results cast a new light over the complex structure and seismotectonics of China and surrounding areas.

Thesis supervisor: M. Nafi Toksöz
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Dedications

This dissertation is dedicated to my late father, Mr. Hanbo Sun, and to my motherland, the People’s Republic of China.

谨以此论文献给先父孙汉波先生！献给我的祖国，中华人民共和国！
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Chapter 1

Introduction

Abstract

In this chapter, the available global models as well as the local and regional models in the crust and uppermost mantle in China and the surrounding area are reviewed. None of the existing models provide accurate velocities for reliable travel-time calculation and event location. Given the large set of high-quality data, a new inversion method, the adaptive moving window, is introduced to construct the P- and S-wave velocity structure of the crust and upper mantle in China and the surrounding area. The resulting 3D models are the starting models for the tomographic study. The updated tomographic method, with Conrad and Moho discontinuities taken into account, is also applied to generate the final structure of the crust and uppermost mantle. The outline of the thesis is included in this chapter.

1.1 Introduction

It is a perpetual challenge to image the earth. The imaging process essentially consists of inverting observed data to obtain new models. The quality of data used, the methodology employed, and the interpretation of the obtained results play the key roles of the imaging process. In this thesis, I will make use of the most recent good-quality data and highly
improved inversion method to generate P- and S-wave models that best represent the structure of the crust and uppermost mantle in China and the surrounding area.

China and its surrounding area are located in a geologically highly heterogeneous and seismically very active region (Figure 1-1). More than 500 earthquakes with magnitude (M) greater than 6.0 occurred in this region between January 1978 and May 2004 (Figure 1-2). The studies of crustal and upper mantle structure in this area have been undertaken over the past decades to locate earthquakes and to investigate and understand the complicated geology and tectonics. Most studies in this region are focused on certain areas. Only a few studies have investigated the whole region using limited data and methods. Among those studies, assumptions such as single-layer crust or/and flat Moho are often made to approximate the structure. The details of these studies will be discussed in later sections of this chapter.

In this study, the most recent available travel-time data are used and a highly improved and efficient inversion method is developed to provide accurate structures in China and its adjacent regions. The detailed descriptions of the data, method and the final models will be introduced in later chapters. Our results in this study are indeed revolutionary.

1.2 Literature Review

The available global and regional models developed most recently are reviewed here. The data and methods used for the different model constructions will be discussed. For the purpose of constructing an accurate 3D model in the crust and uppermost mantle for the
China area, a large set of high-quality data and a new tomographic method will be introduced.

### 1.2.1 Global Models

The available global models are: (1) CRUST 5.1 (Mooney, 1998)/CRUST 2.0 (Laske et al., 2001); (2) CUB1.0 (Shapiro and Ritzwoller, 2001); (3) 1°×1° sediment maps (Laske and Master, 1997); (4) SAIC's 5° (Stevens and McLaughlin, 2001) and 1° dispersion model (Stevens et al., 2001); (5) 3SMAC (Nataf and Ricard, 1996); (6) RUM (Gudmundsson and Sambridge, 1998); (7) S20RTS (Ritsema and Heijst, 2000); (8) SB4L18 (Master et al., 1999); (9) saw24b16 (Megnin and Romanowicz, 2000). The last four models primarily focus on the mantle structure and will not be described in this chapter. The details about those models are discussed in the corresponding reference and can be found at the following website: [http://mahi.ucsd.edu/Gabi/rem.dir/rem.home.html](http://mahi.ucsd.edu/Gabi/rem.dir/rem.home.html).

#### (1) CRUST 5.1/CRUST 2.0 (Mooney, 1998; Laske et al., 2001)

CRUST 5.1 is a global model for the earth’s crust constructed from seismic refraction data published from 1948 to 1995 with a detailed compilation of ice and sediment thickness. A total of 560 seismic refraction measurements have been used to determine the crustal structure on continents and their margins. The model consists of 2592 5°×5° tiles in which the crust and upper mantle are represented by eight layers: ice, water, soft sediment, hard sediment, crystalline upper, middle, lower crust, and uppermost mantle. Crustal layer thickness, both P- and S-wave velocities and density in the crust and uppermost mantle are included. The database for this model is available at the website:
http://quake.usgs.gov/research/structure/CrustalStructure/crust. A program is also
provided to generate the structure model at given location (longitude, latitude).

CRUST 2.0 is a significantly updated version of CRUST 5.1 by incorporating the
1°×1° digital sediment map (Laske and Master, 1997) and using all the newly available
data. More than 360 1D profiles are obtained by tomographic inversion within a 2°×2°
cell. The ice thickness is adjusted to better match the true ice on the earth’s surface
compared to CRUST 5.1. The model and related programs can be downloaded through
FTP at the following site: http://mahi.ucsd.edu/Gabi/rem.dir/crust/crust2.html.

(2) CUB 1.0 (Shapiro and Ritzwoller, 2002)
The global model CUB 1.0 was derived on a 2°×2° grid by simultaneous inversion of
group and phase velocities from a large dataset of fundamental mode surface waves. In
the dataset, there are 100,000 group velocity paths and 50,000 phase velocity paths.
Linearized inversion, simulated annealing and Monte-Carlo inversion are sequentially
performed to achieve the final model. The initial model at each grid was combined from
the sediment model (Laske and Master, 1997), CRUST 5.1 (Mooney, 1998) and shear
velocity model of the upper mantle S20A (Ekstrom and Dziewonski, 1998). At each grid
cell, the model is represented by fourteen unknown parameters in the three-layer crust
and upper mantle. The seven unknown parameters in the crust are the P and S velocities
in three crustal layers and the Moho depth. The layer thicknesses in the first two crustal
layers are unchanged during the inversion. The other seven unknown parameters are the
SV and SH velocities in the anisotropic uppermost mantle, the bottom (thickness) of the
anisotropic mantle and four coefficients for the cubic B-spline perturbations to the
average mantle S-wave velocity. In total there are 226,800 parameters to be determined during inversion. The whole computation takes about 1 month on a single current generation workstation.

(3) 1°×1° Sediment Map (Laske and Master, 1997)

This global sediment map is digitized on a 1°×1° scale. The sediment thicknesses in most of the continental areas including shelves were obtained by digitizing the Tectonic Map of the World provided by the EXXON production research group (1985). In the oceans, published digital high-resolution maps were averaged (e.g. Pacific, Indian and South Atlantic oceans). In areas where such files are not available (e.g. Arctic and North Atlantic oceans), the sediment thickness was hand-digitized using atlases and maps. The sediment at each grid cell is subdivided into one, two or three layers. A single-layer of sediment is given for the locations with sediment thickness smaller than 2 km. Two-layers of sediment are specified with the first-layer thickness of 2 km at the locations with less than 7 km of sediment. The first-layer thickness is 2 km and the second-layer thickness is 5 km for the grid cells with sediment thickness larger than 7 km.

(4) SAIC's 5° (Stevens and McLaughlin, 2001)/1° dispersion model (Stevens et al., 2001)

The research group of Science Applications International Corporation (SAIC) has developed the crust and upper mantle structure models with resolution from 10° (Stevens and McLaughlin, 1996), 5° (Stevens and McLaughlin, 1997; Stevens and McLaughlin, 2001), and to now 1° (Stevens et al., 2001) by simultaneous inversion of group and phase velocity dispersion measurements of surface waves obtained from a variety of sources.
The 5°×5° model is currently used for routine surface wave identification at the International Data Center (IDC). In this model, 149 distinct model types were derived from approximately 90,000 dispersion measurements. Starting with the 5°×5° model, the 1°×1° model is being constructed by using the larger data set with approximately 548,000 dispersion measurements. The velocity profiles for 537 distinct crust and upper mantle model types were derived.

(5) 3SMAC (Nataf and Ricard, 1996)

This is an a priori 2°×2° seismological model of the crust and upper mantle obtained by putting together the topography, sediment thickness, crustal thickness, and lithospheric temperatures. The model is compiled based on the available geophysical data with no inversion involved.

1.2.2 Regional Models

The available regional models for China and its surrounding area are: (1) Mangino's model (1999), (2) Li and Mooney's model (1998), (3) Wu's model (1997), (4) Lamont lab's model (2000), and (5) models constructed by Chinese research groups (Chen et al., 1991; Wang et al., 1991; Zhao and Zeng, 1992).

(1) Mangino's Model (Mangino et al., 1999)

The receiver structure beneath 6 CDSN (China Digital Seismograph Network) stations was constructed by using the seismograms for 89 earthquakes (M_b ≥ 5.7) between 1987
and 1994. The epicentral distances range from 30° to 90°. The 6 CDSN stations are HIA, MDJ, NJI, KMI, LZH and WMQ (Figure 1-3).

(2) Li and Mooney's Model (Li and Mooney, 1998)

The crustal structure of China was constructed based on more than 36,000 km of Deep Seismic Sounding (DSS) profiles of China since 1958. The model includes the contour map of crustal thickness, nine representative crustal columns, and maps with average crustal velocity and Pn velocity.

(3) Wu's Model (Wu et al., 1997)

The group velocity of East Asia was inverted by seismic tomography using Rayleigh waves recorded at three stations (CDSN, SRO, GSN) and one temporary seismic array in Tibet. Totally more than 1200 event-station paths were chosen for inversion. The 1D models within 12 regions in East Asia are obtained. Four regions cover the China area including Tibet, South China, North China and South Mongolia.

(4) Lamont Lab's Model (West, 2000)

The 1D models are obtained by fitting travel-times in each roughly homogeneous region using the data provided by CMR (Center for Monitoring Research). Central Asia is divided into 25 regions corresponding to different geologic provinces. The China area is covered in regions 7, 8, 10, 11, and 13. The final 3D model can be generated at a desired grid spacing. The velocities at the nodes located within the boundaries of two adjacent regions are linearly interpolated to ensure smooth transitions. The data of regionalization,

(5) **Hearn's Pn Velocity Model (2001)**

The detailed Pn velocities with anisotropy beneath the China region are determined by tomographic inversion. The data used for tomography are the first arrival travel-time data given in the Annual Bulletin of Chinese Earthquakes (ABCE) from 1983 to 1995. Over 25,000 arrivals were used in the Pn tomography.

(6) **Huang’s Model (2003)**

The model is inverted from a tomographic study of the S wave velocity structure of China and its adjacent regions. Group velocity dispersions of fundamental Rayleigh waves along more than 4000 paths were determined with frequency-time analysis. The study region was discretized with a $1^\circ \times 1^\circ$ grid, and velocities in between grid nodes were calculated by bilinear interpolation. The Occam's inversion scheme (Constable et al. 1987; deGroot-Hedlin and Constable, 1990) was adopted to invert for group velocity distributions. The detailed Pn velocities with anisotropy beneath the China region are determined by tomography.

(7) **Models Constructed by Chinese Research Groups**

1) **Tile Model (Chen et al., 1991; Fu et al., 1993; Song et al., 1991)**

This model was constructed by Song and Zhuang's research group. The surface waves, recorded at 27 standard stations in China and 4 WWSSN (World Wide Standardized
Seismographic Network) stations from 1983 to 1987, were inverted to provide a 3D structure in the crust and upper mantle. China and the surrounding area is divided into 147 tiles with a grid size of 4° × 4°. The data and models for Huanan (South China) and Huabei (North China) areas are available for FTP download, but not for West China. Only S-wave velocities are given in the models.


In this model, the structure of the upper mantle in the Huanan (South China) area is constructed by fitting the travel times and wave forms of long period P-waves in the distance range between 15° and 30°.

3) Model for Tibet Area (Zhao and Zeng, 1992)

The P and S tomography of the crust and upper mantle in Tibet is obtained by using the travel times from 150 earthquakes recorded by Tibet Seismic Network, Sichuan Seismic Network, the WWSSN and a temporary network installed in Tibet.

4) Models for Beijing Area (Jin et al., 1980; Zhu et al., 1990; Sun and Liu, 1995; Yu et al., 2003 and Huang and Zhao, 2004)

The different P-wave velocity models in Beijing-Tianjin-Tangshan area were obtained by using data from different years. Jin et al. (1980) used the data earlier than 1980. Zhu et al. (1990) used the data from the 1980s and Sun and Liu (1995) used the data from the early 1990s. Yu et al. (2003) and Huang and Zhao (2004) used the most recent data recorded by the new digital seismic network. Only Hung et al. takes the Moho discontinuity into account for the final tomography.
5) Models for Sichuan-Yunnan and Western Areas (Huang et al., 2002; Xu et al., 2002; Wang et al., 2003a)

The recent models based on the P-wave travel-time tomography by Xu et al. (2002) for western China, and by Huang et al. (2002) and Wang et al. (2003) for Sichuan-Yunnan (Southwest China), are the newest models for the crust and upper mantle in western and southwestern China. More high-quality data from many seismic networks including temporary networks are used for their tomographic studies. Both Corad and Moho discontinuities are taken into account when they performed the tomography. The final structures are presented at a fine resolution of about 30 km in the horizontal direction and 5~10 km in the vertical direction.

6) Model in Northwest China (Wang et al., 2003b)

A new crustal section across northwest China was obtained based on a seismic refraction profile and geologic mapping. The 110-km-long section crosses the southern margin of the Chinese Altai Mountains, Junggar Accretional Belt and eastern Junggar basin, easternmost Tianshan Mountains, and easternmost Tarim basin.

1.2.3 Summary

Global models such as CUB 1.0 (Shapiro and Ritzwoller, 2002) and the SAIC 1° x 1° model (Stevens et al., 2001) were constructed from group and phase velocity dispersion measurements of surface waves. The models based on surface waves are generally large-scale models that contain information about the deep structure of the Earth. The model CRUST 2.0 (Laske et al., 2001) was constructed from seismic refraction data and
developed from the CRUST 5.1 model (Mooney, 1998) and a 1° x 1° sediment map (Laske and Masters, 1997). Although CRUST 2.0 was compiled by tomographic inversion, there are too few deep seismic soundings (DSSs) from which refraction data are obtained to provide detailed models for the areas with DSS data. Regional models, including Wu et al. (1997), Lebedev and Nolet (2003), and those by Chinese research groups such as Song et al. (1991), are constructed from surface waves. Regional models by Chinese researchers such as Liu et al. (1990) and Xu et al. (2002) were constructed by regional and teleseismic tomography. The maximum and minimum grid spacings are generally large in both the horizontal direction (2° ~ 5° for Liu et al. and 1.5° for Xu et al.) and in the vertical direction (45 ~ 300 km for Liu et al. and 10 ~ 50 km for Xu et al.). These models show the crustal and upper mantle structure beneath China at a large scale. Detailed crustal structures are not shown, and local and regional travel-time calculations based on these models are not sufficient for improving earthquake locations. In general, models obtained by the teleseismic tomography technique cannot resolve the vertical variations in the shallow structure. Some models were obtained by combining available regional models. Smooth and consistent transitions between different regions are not guaranteed in these models.

For a reliable determination of earthquake locations, accurate travel-times at local and regional distances are needed; 3D velocity models based on body-wave data can provide accurate travel-times. Therefore, we will introduce a method that describes how to construct 3D P-velocity models based on observed travel-time data.
1.3 Data and Method

The data we used for P- and S-wave tomography are the earthquake phase data from January 1990 to December 2002, given in the *Annual Bulletin of Chinese Earthquakes* (ABCE) (IG-CSB, 1990-2002). In this database there are 25,000 earthquakes, 220 stations, and 500,000 ray paths in China and the surrounding area. Figure 1-4 shows earthquake epicenters, stations, and ray paths in China. We also use the seismograms recorded by the China Digital Broadband Network (Figure 1-5).

Sun (2001) and Sun et al. (2004a) relocated the events in the Sichuan area from ABCE and obtained very small hypocentral improvements. Therefore, we used the source locations given in the ABCE for our P- and S-velocity model inversion.

In this study, we introduce a new inversion method — the adaptive moving window (AMW) method (Sun et al., 2004b) to combine 3D P- and S-wave velocity models for the crust and upper mantle by quilting 2338 layered 1D models inverted from fitting first arrival travel times. The resulting 3D models then serve as the starting model for the 3D tomography. We used the tomographic method of Zhao et al. (1992) to invert for the crust and uppermost mantle velocity structure in China and the surrounding area. Zhao's method allows 3D velocity variations everywhere in the model and can accommodate velocity discontinuities. The velocity structure is discretized using a 3D grid. The velocity perturbation at each point is calculated by linearly interpolating the velocity perturbations at the eight surrounding (adjacent) grid nodes. Velocity perturbations at grid nodes are the unknown parameters for the inversion procedure. To calculate travel times and ray paths accurately and rapidly, an efficient 3D ray-tracing technique is employed to iteratively use the pseudo-bending technique (Um and Thurber,
1987) and Snell's law. Station elevations and the sediment layers are taken into account in the ray tracing. The LSQR algorithm (Paige and Saunders, 1982) with a damping regularization is used to solve the large and sparse system of equations, allowing a great number of data to be used to solve a large tomographic problem. The nonlinear tomographic problem is solved by iteratively conducting linear inversions. At each iteration, perturbations to the velocity structure are determined simultaneously. In this thesis, all the hypocentral parameters are taken from the ABCE and they are not variables of the inversion process. A detailed description of the method is given in Appendix B.

1.4 Outline of the Thesis

After the introduction in this chapter, the adaptive moving window (AMW) method (Sun et al., 2004b) is described in detail in Chapter 2. As an example of the method, China and the surrounding area is chosen to construct a 3D P-wave velocity model for the crust and upper mantle by using an extensive catalog (ABCE) that contains twelve years of earthquake travel-time data. We determine layered velocity models at 2338 points. At each point, a window centered at the point is assigned with a size that is chosen based on the ray path density surrounding the point. A 1D model is determined at each point by fitting the first-arrival travel times. We construct models consisting of a four-layer crust and one-layer uppermost mantle. The depth and body-wave velocity of each layer can be found based on the observed travel-times. Both horizontal and vertical resolutions of the obtained models can be controlled, and smooth transitions among adjacent locations are guaranteed.
In Chapter 3, we introduce the updated tomographic method of Zhao et al. (1992) to perform the P-wave tomography in the crust and uppermost mantle in the China area. Crustal heterogeneity is also discussed. The detailed comparison of Pn structure between our model and other existing models is given. The ground-truth (GT) events (nuclear test events) are relocated by a few available models to compare the hypocentral misfit. The comparison of travel-time residuals of the GT events by different models is also given. The travel-time misfit along selected earthquake profiles is examined.

In Chapter 4, we apply the updated tomographic method of Zhao et al. (1992) to perform the S-wave tomography in the crust and uppermost mantle in China. Crustal heterogeneity and intraplate volcanism is discussed.

In Chapter 5, comparison of the obtained 3D P and S models, and interpretation of the Vp/Vs ratio (and Poisson’s ratio) is discussed. Seismograms recorded by the China digital broad-band seismic network are compared with the synthetic ones generated by our models.

The conclusion of this thesis is given in Chapter 6. The key results of this research work are summarized. Future work related to this research is discussed.

The geologic and tectonic settings in China area and the detailed tomographic method are summarized in the Appendix A.
References


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Figure Captions

Figure 1-1: 25,000 earthquakes, 220 stations, active faults and major tectonic boundaries in China and the surrounding area. Earthquake epicenters are shown in red dots and stations are shown in red triangles. The yellow line shows the boundary of China. Active faults in the China area are shown in purple lines and tectonic sutures are shown in blue lines, where SoB — Songliao Basin, OB — Ordos Basin, SB — Sichuan Basin, KB — Khorat Basin, STB — Shan Thai Block, IB — Indochina Block.

Figure 1-2: Major earthquakes with M 6.0 or greater between Jan. 1978 and May 2004 in China and surrounding areas. There are in total 542 big events shown in red dots. The 1679 Sanhe M 8.0 event, the 1976 Tangshan M 7.8 event and the 2001 Ruoqiang M 8.1 event are shown as white stars.

Figure 1-3: The distribution of six CDSN (China Digital Seismic Network) stations.

Figure 1-4: 25,000 earthquakes, 220 stations, and 500,000 ray paths in China and the surrounding area. Earthquake epicenters are shown in black circles and stations are shown in red triangles. The green line shows the boundary of China.

Figure 1-5: The China Digital Broadband Network (CDBN). 47 stations are plotted in red triangles.
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Chapter 2

Adaptive Moving Window Method for 3-D P-velocity Tomography and Its Application in China

Abstract

A new tomography method, the adaptive moving window, is introduced and applied to construct the velocity structure of the crust and upper mantle of China and surrounding area. More than 500,000 high-quality compressional body-wave phase data extracted from the Annual Bulletin of Chinese Earthquakes spanning from 1990 to 2002 are used. The area of interest is represented horizontally by 2338 points with 1° intervals. Each point is assigned a window (a cell or a region centered at each point) whose size is varied depending upon the ray path density. A five-layer 1-D model from the surface down to the uppermost mantle is then determined at each point by performing a Monte Carlo random search where earthquake locations are held constant. By combining and smoothing the 1-D models obtained, an equivalent 3-D model is achieved. The predicted travel-times through the 3-D model match very well with the observed travel-times from local to regional distances. The model correlates well with tectonic features and is generally consistent with the existing models constructed by other researchers. Our model
gives detailed information about structure and is feasible for application to high-quality earthquake location problems.

2.1 Introduction

The available P-wave velocity models of the crust and upper mantle in China and surrounding area have been obtained using different approaches. Global models such as CUB 1.0 (Shapiro and Ritzwoller, 2002) and the SAIC 1° x 1° model (Stevens et al., 2001) were constructed from group and phase velocity dispersion measurements of surface waves. Huang et al. (2003) conducted the Rayleigh wave tomography of China and adjacent regions and obtained results similar to CUB 1.0. The models based on surface waves are generally large-scale models that contain information about the deep structure of the Earth. The model CRUST 2.0 (Laske et al., 2001) was constructed from seismic refraction data and was developed from the CRUST 5.1 model (Mooney, 1998) and a 1° x 1° sediment map (Laske and Masters, 1997). Although CRUST 2.0 was compiled by tomographic inversion, there are too few deep seismic soundings (DSSs) from which refraction data are obtained to provide detailed models for the areas with DSS data. A Pn and Sn velocity model was presented in Ritzwoller et al. (2002) as part of a tomography study for Eurasia. Hearn and Ni’s (2001) Pn model was constructed directly from the Pn travel-times recorded in the China area. Explosions and DSSs have been taking place in different parts of China to achieve accurate crustal and Pn velocities in local areas. Regional models, including Wu et al. (1997), Lebedev and Nolet (2003), and those by Chinese research groups such as Song et al. (1991), are constructed from surface waves. Regional P-wave travel-time tomography has been done by Chinese
researchers such as Liu et al. (1990) and Xu et al. (2002). Liu et al.'s model was constructed by regional and teleseismic tomography. The maximum and minimum grid spacings are 5° x 5° and 2° x 2°, respectively, in the horizontal direction and 300 and 45 km in the vertical direction (Liu et al., 1990). The model showed the crustal and upper mantle structure beneath China in a large scale. Detailed crustal structures are not shown in this model, and local and regional travel-time calculations based on this model are not accurate enough for locating earthquake events. Xu et al. constructed the crust and upper mantle structure beneath western China through seismic tomography by using P-wave arrival times of local/regional and teleseismic events recorded on Chinese and Kyrgyzstan seismic networks. The grid spacing is 1.5° x 1.5° in the horizontal direction and about 10 km (in the crust) in the vertical direction. In general, models obtained by the teleseismic tomography technique cannot resolve the vertical variations in shallow structure. Some models were obtained by combining available regional models. Smooth and consistent transitions between different regions are not guaranteed in those models.

For reliable determination of earthquake location, accurate travel-times at local and regional distances are needed, and 3-D velocity models based on body-wave data can provide these accurate travel-times. Therefore, we will introduce a method that describes how to construct 3-D P-velocity models based on observed travel-time data.

In this chapter, we introduce a new inversion method: the adaptive moving window (AMW) method. As an example of the method, China and the surrounding area is chosen to construct a 3-D P-wave velocity model for the crust and upper mantle by using an extensive catalog that contains nine years of earthquake travel-time data. We determine layered velocity models at 2338 points. At each point, a window centered at
the point is assigned with a size that is chosen based on the ray path density surrounding the point. A 1-D model is determined at each point by fitting the first-arrival travel-times. We construct models consisting of a four-layer crust and one-layer uppermost mantle. The depth and body-wave velocity of each layer can be found based on the observed travel-times. Both the horizontal and vertical resolution of the obtained models can be controlled, and smooth transitions among adjacent locations are guaranteed.

### 2.2 Data and Method

The data we use are the earthquake phase data from January 1990 to December 1998, given in the *Annual Bulletin of Chinese Earthquakes* (ABCE) (IG-CSB, 1990-1998). In this database there are 16,642 earthquakes, 220 stations, and 345,520 ray paths in China and surrounding area. Figure 2-1 shows the earthquake epicenters, stations, and ray paths in China.

Sun (2001) and Sun et al. (2004) relocated the events in the Sichuan area in ABCE and obtained very small epicentral improvement. Therefore we used the source locations given in the ABCE for our P-velocity model inversion. Our goal is to obtain 1-D velocity models of crust and uppermost mantle in the China area based on these travel-time data.

We selected 2338 points distributed on a 1° grid in China (shown in Figure 2-1). As shown in Figure 2-2, at each point, a 1-D velocity model is obtained by fitting first arrivals (Pg or Pn) within a window (region) centered at the point. The minimum size of each window was chosen to be 4° x 4° (latitude, longitude) to guarantee sufficient Pn ray paths. The window size is increased until the required minimum numbers of Pg and Pn
ray paths are included. Every 1-D velocity model consists of four layers of crust and one layer of upper mantle (shown in Figure 2-3). The top layer is the sediment layer, and its thickness is taken from Laske et al. (2001). The other three crustal layer thicknesses, four P-wave velocities in the crust, and the Pn velocities are the eight parameters found by random search iteration (Monte Carlo inversion).

The first step in the inversion is to choose travel-times. The travel-times recorded at each station from an earthquake contain many phase arrivals. We use only first arrivals in our inversion. Figure 2-4 shows the first arrival travel-times in China within epicentral distances of 20°. Some first arrivals at large distances are Pg arrivals instead of Pn. We believe those Pn arrivals were missing in the report or too weak to be picked. We choose all the first arrivals for the events and stations located within the window centered at each point. We select a minimum of 300 travel-times, including at least 100 Pn arrivals at each point. The minimum window size is 4° x 4° and larger windows are used around some points so that there are at least 200 Pg and 100 Pn arrivals inside the window. In a few cases where data are sparse, the window size is as large as 15° x 15° and at least 50 Pg and 20 Pn arrivals are selected in each window. The maximum epicentral distance between a source and a receiver for the entire study area is 8°.

The second step is to find a 1-D velocity model for each window using a Monte Carlo algorithm. As we mentioned earlier, every 1-D model contains eight parameters to be obtained from the travel-times. For each iteration, the Monte Carlo algorithm randomly selects each variable from within some preset bound to compose a 1-D model. Pn and Pg travel-times are calculated based on the selected 1-D model and the existing event locations in the ABCE. Event locations are not changed from those in the catalog.
Travel-time residuals for the first-arrival Pg and Pn are then obtained, and Monte Carlo iteration continues until the maximum number of iterations is reached. The optimal model is the one with minimum root mean square (rms) error. Each 1-D model is determined by 100,000 iterations.

The search range for each variable is set a priori based on our knowledge of the model. We use the explosion results by DSPEM-CSB (1988) and Hearn and Ni’s (2001) Pn results as a guide to set bounds for the variables. The Monte Carlo search range for Pn is limited to ± 0.2 km/sec of the Hearn and Ni Pn velocity. For the points not covered by their model, the Pn bound is set to [7.6, 8.3] km/sec. Bounds for the thickness of each layer in the crust are generally between 0 and 20 km. The general bounds for the four crust velocities are [4, 5], [5, 6.2], [6.3, 6.6], and [6.6, 7.4] km/sec. The bounds are ± 5 km of DSPEM-CSB’s layer thickness or ± 0.5 km/sec of the DSPEM-CSB’s layer velocities.

The first arrivals at distances greater than each critical distance are Pn phases. It is possible that some observed first arrivals at large distances are Pg arrivals instead of Pn. We separate Pn and Pg phases at large distances based on the slope of travel-time versus distance curve to avoid calculating residuals between observed Pg arrivals and calculated Pn arrivals.

The third step is to apply an adaptive moving window to all points to obtain the 1-D velocity profiles at each of the 2338 points. The 1-D velocity profile at each point inverted from the phase data inside the moving window, is not exactly the velocity profile at that point, but the averaged velocity profile in the window surrounding the point. Profiles obtained from windows with large size are averaged over large distances. The
points are located at 1° intervals. We upsample all the points by a factor of 5 to a 0.2°
interval using linear interpolation. Both velocity and thickness of each layer are
interpolated. The upsampling is accomplished using a Gaussian function with a half-
length of 8 points to smooth all the models horizontally at each layer.

After all the upsampled 1-D models were smoothed, we select and combine the
original 2338 1-D models to obtain an equivalent 3-D model.

2.3 Results

Performing the AMW method in the China area, we obtain a 3-D P-velocity model of the
crust and uppermost mantle. Figure 2-5 shows the vertical velocity profiles obtained from
travel-time inversion. Figure 2-6 shows the crustal thickness and Pn velocity in China.
Figures 2-7 and 2-8 show horizontal velocity profiles at different depths in China and the
surrounding area.

The lateral heterogeneity of the crust and upper mantle beneath China is shown
clearly in Figures 2-5 through 2-8. The 3-D velocity model obtained by quilting the 1-D
velocity profiles correlates well with the tectonic regions. The velocity images at a depth
of 50 km show that the crust of the Chinese continent is divided into two parts
approximately by the 102.5° longitude (Liu et al., 1990). In the western part, the crust is
thicker and crustal velocities are lower than those in the eastern part. The velocity models
show the high velocities in the crust beneath the Precambrian regions, including the
Tarim Basin, Sichuan Basin, Ordos Basin, and Songliao Basin. The Bohai Gulf shows
both slow and fast velocity anomalies due to a Cenozoic rift system through the gulf. The
northern part of the South China Block is slower than the southern part in the lower crust
and the difference is small in the uppermost mantle. The Indochina Block shows a low-velocity anomaly in the crust and the uppermost mantle that is consistent with volcanism.

In Figure 2-9, a profile centered at (85°, 40°) that is 4° wide in longitude and 30° long in latitude was selected to compare the observed travel-times with the travel-times predicted by the 3-D model. For the epicentral distances smaller than 12°, the calculated travel-times based on the 3-D model fit the observed travel-times very well. For distances larger than 12°, a more accurate deeper structure of upper mantle is needed. We adopt the IASPEI model for the upper mantle for travel-time calculation. The upper mantle velocity in the selected profile is faster than the real one because the predicted travel-times for the farther distances are in general smaller than the observed ones. Within the selected profile, the 3-D travel-time rms error is 0.65 sec for the first arrivals at distances of 10° or less. If we average all the 1-D velocity models along the profiles and calculate 1-D travel-times based on the resulting 1-D velocity model, the rms error of travel-times is 1.1 sec. The 3-D model provides much more accurate travel-times than any available 1-D model.

2.4 Analysis

As shown in Figure 2-3, at each point, nine parameters represent a 1-D model with a four-layer crust and one-layer uppermost mantle. Eight of them are estimated by fitting first arrivals within a window centered at the point. In this section, we discuss the robustness of the random search, the uncertainty analysis, and the resolution and accuracy. We also compare our models obtained by random search with those from other researchers.
2.4.1 Robustness of Monte Carlo Search Fit

After the layered model is set up, the distances and travel-times can be calculated according to the formulas listed in Sun (2001). Based on the observed travel-times in the window centered at a point, the best eight parameters that fit the data with the minimum RMS error can be found by the Monte Carlo method (random search). The steps of a random search are as follows:

Step 1: Choose parameter ranges.
Step 2: Choose a random number, scale the [0-1] random number to the parameter interval $[\beta_{\text{MIN}} - \beta_{\text{MAX}}]$, and repeat for all parameters.
Step 3: Calculate theoretical arrival time.
Step 4: Compare theoretical ($T_{\text{model}}$) and observed ($T_{\text{observed}}$) data. The data residual $\Delta T$ is defined to be $(T_{\text{model}} - T_{\text{observed}})$.
Step 5: Stop if residual misfit is small or after $n_{\text{max}}$ trials. The residual misfit is measured by the RMS error and calculated as follows:

$$RMS = \left( \frac{\Sigma \Delta T^2}{N} \right)^{\frac{1}{2}},$$

where $N$ is the number of observations.
Step 6: Keep parameter sets associated with small residuals.
Step 7: Repeat from Step 2 if residuals large or if $n_{\text{max}}$ not reached.

A random search will continue to $n_{\text{max}}$ or will stop if RMS is smaller than the tolerance $\varepsilon$. With the random search method one can be reasonably certain of uniqueness.
if $n_{\text{max}}$ goes to infinity and/or $\varepsilon$ goes to zero. The RMS error will go to the true minimum and the parameters go to the global solution. The question is: if the optimal models are obtained by limited iterations ($n_{\text{max}}$) instead of infinity, how can we guarantee that these models are the best?

We selected a few locations in China and searched for 1-D models with both 100,000 and 50,000,000 runs. The best models found with 100,000 runs are exactly the same as the ones with 50,000,000 runs. Our random searches converge at 100,000 runs or less because there are limited travel-times for each location. We also ran 50,000 runs for all the selected 2338 points to guarantee the robustness of the random search fit.

The velocity models obtained at each location are the averaged ones in the area surrounding each point. One can imagine that the velocity models obtained at neighboring locations are similar due to a large amount of shared ray paths. To test this, we select a central location at (102, 35) in the middle of China and compare the ray paths, inverted models, and travel-time fitting at the center with those that are one degree, three degrees and 10 degrees distance from the center in both longitude and latitude. The locations are shown in Figure 2-10.

Figures 2-11 through 2-13 show the ray paths for the locations surrounding the center point (102, 35) with one degree apart, three degrees apart and ten degrees apart, respectively. Figures 2-14 through 2-16 show the results of travel-time fitting. We can see that the travel-time data are well fit for all locations. The inverted velocity models are shown in Figures 2-17 through 2-19. Figure 2-17 shows that the four velocity models that are one degree away from the center are very close to the velocity model at the center.
When the distance between the locations and the center increase, the velocity models become more uncorrelated.

We showed in Sun (2001) that Monte Carlo inversion could exactly recover the velocity models from synthetic seismic data with no noise added. For the models with noise added, a random search also reduces the uncertainty of coupled parameters. Because we know that the origin time and depth of an event are trade-off parameters, we found the true depths of three explosions by using the Monte Carlo method to fit the travel-time data. We also ran the least squares location program (Hypoinverse) and saw that, in general, the depths located were a few kilometers off.

### 2.4.2 Uncertainty Analysis

The best model at each location is obtained by minimizing the RMS error of travel-time fitting. From the previous section, we know that the Monte Carlo search fit is very robust when the number of iterations is 100,000 or above. Another important question to ask is: what is the uncertainty in each “best” model?

Unlike other inversion methods such as Least Square (LS), in which error estimates can be evaluated based on the travel-time misfit and derivative matrix, there are no simple ways to evaluate the model errors by the Monte Carlo search fit. We first estimate the uncertainty of the models obtained from the synthetic travel-times.

Figure 2-20 shows a two-layer model with a synthetic event on the surface of the earth. The crust and the uppermost mantle are separated by a dipping Moho. There are only three parameters to represent the 1-D model at the source location. Those parameters are the Moho depth, the averaged crust velocity and the averaged velocity in the
uppermost mantle. We set the Moho depth (D) beneath the event to 40 km. The averaged crust velocity (\(V_1\)) is 6.5 km/s and the averaged Pn velocity (\(V_2\)) is 8.0 km/s.

Table 2-1 shows the standard deviations of the three parameters in different ranges of the dipping angle (\(\alpha\)) of the Moho interface. The standard deviations of all the three parameters increase when the dipping angle increases. The estimated crustal velocity carries larger uncertainties than the Pn velocity. As discussed in Sun et al. (2004), the dipping angles of the Moho interface in China are smaller than 3° in a region of 4° × 4° or larger. The uncertainties for all the three parameters are smaller than one percent.

<table>
<thead>
<tr>
<th>(\alpha)</th>
<th>(\sigma_D) (km)</th>
<th>(\sigma_{V_1}) (km/s)</th>
<th>(\sigma_{V_2}) (km/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>0°</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>[0°, 5°]</td>
<td>1</td>
<td>0.1</td>
<td>0</td>
</tr>
<tr>
<td>[5°, 10°]</td>
<td>2</td>
<td>0.2</td>
<td>0.05</td>
</tr>
<tr>
<td>[10°, 15°]</td>
<td>3</td>
<td>0.3</td>
<td>0.08</td>
</tr>
<tr>
<td>[15°, 30°]</td>
<td>4</td>
<td>0.3</td>
<td>0.16</td>
</tr>
</tbody>
</table>

We can also have an idea about the uncertainty in each final model by comparing each best model with other sub-optimal models (i.e. those with bigger RMS error) in the same location. We choose the same 12 points plotted in Figure 2-10 as the locations for model comparisons. At each location, the 10 best models based on the RMS error are selected and plotted in Figure 2-21. The best models with the minimum RMS error are shown in magenta. We can see that at most locations the top ten models are close to each other. For the locations at (102, 36), (102, 38), (105, 35) and (102, 45), there are slightly larger differences for the lower crust models. The standard deviation (STD) of the
velocity in the lower crust is from 0.05 to 0.19 km/s, while the STD range is [0.01, 0.14] in the upper crust and [0.02 0.09] in the middle crust. The STD range for the uppermost mantle is [0.01, 0.03] due to the constraints we applied.

Based on the best models at the selected 12 locations, an averaged model for the entire China area is shown in Figure 2-22. From this we can see that there is considerable lateral velocity variation in China.

2.4.3 Resolution and Accuracy

As we mentioned earlier, the 1-D velocity model at each point is inverted by the travel-time data inside a window centered at that point. The size of the window depends on the number of arrivals of the ray paths in the window. The 1-D velocity model obtained at each point is an averaged layered model for the window. Therefore, even though the grid spacing between the selected points is 1° x 1°, the size of the window indicates the resolution of the model image. Smaller window sizes indicate higher resolution of the model image, and larger window sizes mean lower resolution of the model image.

The window-size distribution of all 2338 points is shown in Figure 2-23 (upper plot). About 80% of the region sizes are 8° x 8° or smaller. Most areas with coarse ray coverage and large window size are in Mongolia and along the boundaries of the selected 2338 points. Some parts of Tibet also require larger window sizes due to sparse station coverage.

Figure 2-23 (lower plot) shows the spatial resolution in terms of the normalized accuracy. We defined the accuracy in each window to be the number of selected ray paths divided by the window size. The normalized accuracy is obtained by dividing the
accuracy of each window by the maximum accuracy in the China area. The normalized accuracy represents the inversion resolution in each window. We see a similar pattern between the window-size distribution and the resolution map.

2.4.4 Model Comparison at the Moho Interface

The models for comparison are CUB 1.0 (Shapiro and Ritzwoller, 2001), the SAIC 1° x 1° model (Stevens et al., 2001), and CRUST 2.0 (Laske et al., 2001). The first two models were constructed from the group and phase velocity dispersion measurements of surface waves. The last one was constructed from seismic refraction data, and was developed from the CRUST 5.1 model (Mooney, 1998) and 1° x 1° sediment map (Laske and Master, 1997).

All the models show a good correlation with surface topography, with high elevation corresponding to a deep Moho (Figure 2-24). The outline of the Tibet Plateau is clearly depicted by all models. Though there are small differences, the large-scale features are similar in all the models, and Moho depth decreases from west to east in China. All of the models give the deepest Moho (70+ km) at the center of the Tibet Plateau, and the shallowest (about 30 km) in the coastal areas around China’s continental shelf.

The Moho depth difference for all the models in Figure 2-24 is shown in Figure 2-25. The difference is taken by subtracting the mean Moho depth for all models from each model. The Moho depth difference for most areas is close to zero. In Mongolia and the area south of the Himalaya, the Moho depth differences are large. In Mongolia, only CUB1.0 shows negative anomaly of the range from -10 km to -2 km. MIT, SAIC and
CRUST 2.0 show a positive anomaly in the range from 2 km to 4 km. In the area south of the Himalaya, both SAIC and CUB 1.0 show a strong negative anomaly from -10 km to -2 km while MIT and CRUST 2.0 show a positive anomaly of 2 km to 6 km.

Hearn and Ni’s (2001) Pn model was constructed directly from the Pn travel-times recorded in the China area (Figure 2-26). Due to Hearn’s simple method, we believe that his Pn model is the most accurate, and we took their model as a reference in our inversion. Thus, it is natural that our Pn model is similar to Hearn’s Pn model. There are large differences between the Hearn model and the two models based on surface waves. It is not clear whether these differences emerge because of limited sensitivity of the surface wave to the thin layer of the upper mantle that defines Pn, or because the Pn velocity is obtained from the shear velocities of surface wave models. In any case, we believe that Pn velocities obtained directly from P-wave travel-times are more accurate than those of surface wave models.

2.4.5 Comparison of Vertical Crustal Velocity Profiles

Taking ten CDSN stations (Figure 2-27) addressed in Mangino et al. (1999) as reference points, comparisons for 1-D vertical profiles are performed. Most stations shown in Figure 2-27 are in the regions with high ray density and good model accuracy (Figure 2-23).

The comparison is shown in Figure 2-28. The thick red lines indicate the results from our 1-D model. We can see that the differences from the other models are not very large at the reference points, even though the other models are obtained from diverse
datasets. This supports the fact that our 1-D model is in general agreement with previous work.

2.5 Discussion and Conclusion

Our 1-D Monte Carlo inversion is performed for regions with sizes ranging from 4° x 4° to 15° x 15°. Nearly 80% of the region sizes are 8° x 8° or smaller. The 1-D velocity models represent an average of the window centered at each selected location, and adjacent 1-D velocity models are based on data with considerable overlap. Given this, one would expect a smooth transition from one velocity profile to adjoining profiles. We observed this smooth transition.

The earthquake source locations we used to construct the P-velocity models are the ones given in the ABCE. The location uncertainty may have an impact on our final velocity model. Since the ray paths at each selected location are dense, we believe the event location uncertainty does not play a significant role in each 1-D model inversion.

For the areas with coarse ray density, the window size goes up to 15° x 15°. The velocity models obtained in these regions are strongly averaged, have poor spatial resolution, and are less accurate compared to the ones in the areas with dense ray coverage. Most areas with coarse ray coverage are in Mongolia.

Compared to the global 1-D velocity model (IASPEI), the inverted 1-D velocity models better represent the structure of the crust and uppermost mantle in China. The travel-time residuals between the observed travel-times and travel-times calculated from our models are much smaller than those between the observed and calculated travel-times.
for the IASPEI model. For the region centered at (102°, 34°) shown in Figure 2-2, the rms is 1.4 sec for the IASPEI model and only 0.69 sec for our inverted 1-D model.

The Moho depths we obtained are similar to those in the CUB 1.0 (Shapiro and Ritzwoller, 2002), the SAIC 1° x 1° (Stevens et al., 2001), and the CRUST 5.1 models (Mooney, 1998). All the models show excellent correlation between Moho depth and surface topography.

In order to validate our methodology and models, we compared the travel-times over long ray paths in China (Figure 2-9). The 3-D travel-times based on our models agree with the observed travel-times very well. We also compared the travel-times of a few ground truth events in both western and central China. The travel-time residuals are in general smaller than 0.6 sec when the epicentral distance is between 6° and 12°. Accurate travel-times for larger epicentral distances require a better knowledge of the deep structure.

The 3-D model obtained by the AMW method provides accurate body-wave travel-times and illustrates tectonic structures. The model can be used to locate earthquake events and calculate travel-times for other purposes as well. The AMW method provides a new and efficient method of performing 3-D tomography. This method works very well for areas with dense ray coverage and reliable earthquake location catalogs. The S-velocity model can be obtained by applying the AMW method as well. A 3-D Poisson's ratio model can be obtained once both P- and S-velocity models are constructed.

It is important to mention that this 3-D model is not a model obtained through proper tomographic inversion, which remains to be done. The model we generated
would serve as a good starting model for a tomography inversion. The horizontal resolution of the model will be better in 3-D than in 1-D because 1-D inversion obtains models averaged over a large area.

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Figure Captions

Figure 2-1: Upper: Locations of 25,000 earthquakes (red dots), 220 stations (red triangles), and 500,000 ray paths in China and surrounding area. The green line shows the boundary of China. Lower: 2,338 points (red stars) in China and surrounding area.

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Chapter 3

Crustal Structure of China and Its Surrounding Regions from P-wave Travel-time Tomography

Abstract

A 3-D P-wave velocity structure model is developed for the crust and uppermost mantle of China and the surrounding area using 345,000 high quality P-wave first arrivals extracted from the Annual Bulletin of Chinese Earthquakes (ABCE). A preliminary 3-D model is generated by combining 2400 1-D models (Sun et al., 2004b). We apply the tomography method of Zhao et al. (1992) and obtain a detailed 3-D P-wave velocity model of the crust and uppermost mantle in China. The spatial resolution is $1^\circ \times 1^\circ$ in the horizontal direction and 10 km in depth. This method is applicable to a 3-D velocity structure which includes several complex-shaped velocity discontinuities. These velocity discontinuities represent known geological boundaries such as the Moho discontinuity and subducting slabs. The velocity structure is discretized using a 3-D grid. The velocity perturbation at each point is calculated by linearly interpolating the velocity perturbations at the eight surrounding grid nodes. Velocity perturbations at grid nodes are the unknown parameters for the inversion procedure. To calculate travel-times and ray paths fast and accurately, an efficient 3-D ray-tracing technique is employed to iteratively use the
pseudo-bending technique (Um and Thurber, 1987) and Snell's law. Station elevations and the sediment layers are taken into account in the ray tracing. The LSQR algorithm (Paige and Saunders, 1982) with a damping regularization is used to solve the large and sparse system of equations.

Our tomographic model provides new insights into the geological structure and tectonics of the region, such as lithological variations and large fault zones across the major geological terranes. The velocity images of the upper crust correspond well with the surface geologic and topographic features such as basins and the Tibetan Plateau. High velocity anomalies are found in the lower crust beneath the Precambrian regions (Tarim Basin, Ordos Basin, Sichuan Basin and western half of Songliao Basin). The highest velocity anomaly is beneath the Sichuan Basin. High and low velocity anomalies imaged beneath the Bohai Gulf are associated with the presence of a major Cenozoic rift system. In the lower crust beneath the South China Block, P-wave velocities are lower in the north compared with the south. The Indochina Block shows low velocities both in the crust and in the uppermost mantle due to volcanism. The Pn velocities in the Tibet area are higher than those in other areas largely due to thicker crust. 3-D travel-times calculated for our model closely match the observed data, showing that the model is well constrained and can be applied to determine the source parameters of earthquakes and to generate synthetic travel-times.

Eleven high-quality ground truth (GT) events with a large number of observations have been collected from nuclear testing explosions in China. We relocated the GT events using our P-wave model, and the averaged hypocentral mislocation is less than 1 km with a standard deviation of 0.3 km. The mean hypocentral misfit is about 10 km if
the averaged 1-D velocity in China is used and it is about 20 km if the global AK135 model is used. We calculated the travel-times from each GT event to the stations located within an epicentral distance of 20° and the predicted travel-times match well with the observed data. The mean travel-time residuals at each station from all the GT events are smaller than 1 s with a standard deviation less than 0.2 s. Both the mean and the standard deviation of the travel-time residuals at each station are smaller than 0.6 s and 0.15 s, respectively, when the epicentral distance is smaller than 15°. We also compared the observed travel-times with the predicted travel-times for a well-located earthquake in northern China and the obtained travel-time residuals are reasonably small. The relocation of GT events and the comparisons of travel-times from the GT events and the earthquake suggest that our P-wave model is reliable for accurate 3-D travel-time calculation and event location.

Keywords: seismic tomography, p-wave velocity, crustal structure, continental earthquakes, seismotectonics
3.1 Introduction

China and the surrounding area is located in a seismically very active region (Figure 3-1). More than 500 earthquakes with magnitude (M) greater than 6.0 have occurred in this region between January 1978 and May 2004 (Figure 3-2). Historically, large and destructive earthquakes took place frequently both in east and west China. Four large earthquakes in east and west China are shown in Figure 3-2. In 1679, an earthquake of M 8.0, one of the largest known earthquakes in this region, occurred in the Sanhe county of Beijing. The 1976 Tangshan earthquake (M 7.8), with the epicenter approximately 160 km southeast of Beijing, totally destroyed the Tangshan city (population then 1 million) and killed about 240,000 people. It was perhaps the most destructive earthquake in human history. The largest earthquakes recorded in western China are the 1951 Dangxiong earthquake (M 8.0) in Tibet and the M 8.1 2001 Ruoqiang earthquake (also known as the Kunlun earthquake) in Xinjiang-Qinghai area. Most large earthquakes in this region are located at depths of 10 to 20 km. A detailed investigation of the crustal structure and seismotectonics of this region is very important for understanding the physics of continental earthquakes and for assessing and mitigating seismic hazard.

From a geological point of view, there are four Precambrian platforms (Figure 3-1) in China and the surrounding area (Lebedev and Nolet, 2003): the North China Block, the South China Block, the Tarim Basin and the Indochina Block. The North China Block, also known as the Sino-Korean Craton, consists of two major Archean (older than 2.5 Ga) continental nuclei surrounded by Paleoproterozoic (about 1.8 Ga) orogenic belts. One nucleus is approximately within the boundaries of the Ordos Basin (Ordos Plateau), and the other, larger one, is beneath and around the Bohai Gulf. Younger orogenic belts are
located along the margins of the block. The South China Block includes two major Precambrian elements: the Yangtze Craton and the Cathaysia Block. The Archean nucleus of the Yangtze Craton is approximately within the boundaries of the Sichuan Basin. The Paleoproterozoic basement is found in the vicinity of the basin and to the southwest; the rest of the basement is probably of the Grenville age (about 1.0 Ga). The Cathaysia Block is situated along the coast (partly in the South China Foldbelt and partly underwater to the east) and is separated from the Yangtze Craton by orogenic belts of different ages, from about 1000 to 150 Ma. The Tarim Basin consists of a nucleus of Archean through Proterozoic age covered by thick Cenozoic sediments. The Indochina Block also has a Precambrian core and probably extends to the northwest as the Shan Thai (Simao) Block.

The available P-wave velocity models of the crust and upper mantle in China and the surrounding area have been obtained using a variety of approaches. Global models such as CUB 1.0 (Shapiro and Ritzwoller, 2002) and the SAIC 1° x 1° model (Stevens et al., 2001) were constructed from group and phase velocity dispersion measurements of surface waves. The models based on surface waves are generally large-scale and intended to map the deep structure of the earth. The global model CRUST 2.0 (Laske et al., 2001) was constructed from seismic refraction data and developed from the CRUST 5.1 model (Mooney, 1998) and a 1° x 1° sediment map (Laske and Masters, 1997). Although CRUST 2.0 was created by tomographic inversion, there are too few deep seismic soundings (DSSs) to provide enough refraction data for a detailed model. Regional models were constructed by Pn and/or Sn tomography (Ritzwoller et al., 2002; Hearn and Ni, 2001; Pei et al., 2004a), from surface waves (Wu et al., 1997, Lebedev and
Nolet, 2003; Huang et al., 2003; Song et al., 1991; Zhu et al., 2002), and from P-wave travel-time tomography (Liu et al., 1990). The model of Liu et al. (1990) was constructed by regional and teleseismic tomography. The maximum and minimum grid spacings of Liu et al. (1990) are 5° x 5° and 2° x 2°, respectively, in the horizontal direction and 300 km and 45 km in the vertical direction. They parameterized the crustal and upper mantle structure beneath China at a large scale. Because of the lack of detail, local and regional travel-time calculations based on this model are not accurate enough for locating earthquakes. P-wave tomography has been performed in several local regions in China (Xu et al., 2002a; Huang et al., 2002; Yu et al., 2003; Huang and Zhao, 2004; Sun and Liu, 1995; Zhu et al., 1990; Pei et al., 2004b). Xu et al. used P-wave arrival times of local, regional and teleseismic events recorded by Chinese and Kyrgyzstan seismic networks to tomographically map the crustal and upper mantle velocity structure beneath western China. The grid spacing is 1.5° x 1.5° in the horizontal direction and about 10 km in the crust and 50 km in the uppermost mantle in the vertical direction. Crustal structures beneath north and east China including Beijing and surrounding regions were obtained by Zhu et al. (1990), Sun and Liu (1995), Yu et al. (2003) and Huang and Zhao (2004) as the result of their P-wave tomographic studies. Huang et al. (2002) inverted the lithospheric structure in southwest China from local/regional travel-time data. Pei et al. (2004b) imaged the mantle beneath East Asia by performing P-wave travel-time tomography using regional and teleseismic events. These models show detailed crustal structures only in a few regions. A detailed map for the whole China area remains to be developed.

We need 3-D velocity models that cover the whole China with good resolution everywhere to provide accurate travel-times for reliable determination of earthquake
locations. Models obtained by the teleseismic tomography technique generally cannot resolve vertical variations in the shallow structure. Regional models can be combined in order to cover a large area, but such models cannot guarantee smooth and consistent transitions between different regions. During the last a few years, there have been many digital seismic stations installed in the China area (Figure 3-1). The large database of high-quality recorded arrival times provides an unprecedented opportunity to determine a detailed 3-D crustal structure under the region. Therefore, we introduce a method that constructs 3-D P-velocity models for the whole China area based on observed travel-time data.

The starting model for the tomographic study is a 3-D model created with the adaptive moving window (AMW) method (Sun et al. 2004) using an extensive catalog including thirteen years of earthquake travel-time data. The model has a four-layer crust and a one-layer uppermost mantle at each one degree intersection of longitude and latitude. The thickness and body-wave velocity of each layer can be found from the observed travel-times. Both horizontal and vertical resolutions of the obtained models can be controlled, and smooth transitions between adjacent locations are guaranteed. Using the starting model, we then apply a tomographic method to a large dataset of local/regional earthquake arrival times to determine a high-resolution 3-D P-wave velocity structure of the crust and uppermost mantle under this region. Compared with previous tomographic studies, we have used a larger, higher quality dataset and applied an updated tomographic method to take into account the effects of the complex Moho geometry in this region. Our results cast new insights into the complex structure and seismotectonics of China and the surrounding area.
3.2 Data and Method

In this work, we use the earthquake phase data from January 1990 to December 2002, from the ABCE (IG-CSB, 1990-2002). In this database there are 25,000 earthquakes, 220 stations, and 500,000 ray paths in China and the surrounding area. Figure 3-1 shows earthquake epicenters and stations in China. We also incorporated 129 earthquakes with high-quality records and 3 quarry blast events (ground-truth events) from the Sichuan Province network. Each quarry blast located on the surface has at least 30 P arrivals and a small uncertainty (< 1 km) for the epicenter locations. Given that the ray coverage is denser in some areas and redundant calculation is involved, to assemble the best set of the earthquake data, we adopted the method described below.

The study area is divided into cubic (more accurately parallelepipedic) blocks with a spatial size of 10 km x 10 km x 2 km. Among the earthquakes within each block, we select the event with the greatest number of first P-wave arrivals and the smallest hypocentral location uncertainty. This is the final set of data for our study, which contains 16,000 events with more than 300,000 ray paths used for the 3-D tomography. The final ray coverage has a better (more uniform) distribution in the study area. Most aftershocks occur in similar locations with smaller magnitude, and therefore there are in general bigger reading errors for the phase arrivals. The final dataset contains the least number of aftershocks possible. Most of the events in the final dataset satisfy the earthquake location criteria with a local seismic network proposed by Bondar et al. (2004) for a hypocentral location accuracy better than 5 km.

The 1-D initial model significantly influences the final result of the tomographic study. An inappropriate initial reference model may not only affect the quality of the
three dimensional images by introducing artifacts, but it may also influence the confidence calculations by underestimating the uncertainties of the results (Yu et al., 2003; Kissingling et al., 1994). We used all the available high-quality arrival time data in our selected dataset and inverted for a 1-D velocity model representing the whole study area by minimizing the root mean square (RMS) error of the travel-times. Finally we obtained the 1-D model shown in Figure 3-5 which gave the best fit to the observed data. We used this averaged 1-D model as the reference velocity model for our tomographic inversions. The initial and final 3-D models are therefore represented by velocity perturbations relative to the reference 1-D model.

The discontinuities represent known geological boundaries such as the Conrad and the Moho discontinuities and/or a subducting slab boundary. Previous studies were able to map the Moho discontinuity in the study area and revealed its significant lateral depth variations (China Seismological Bureau, 1986; Zhang, 1998; Li et al., 2001; Sun et al. 2004). The Conrad discontinuity is clear in some regions of the study area but not in others. We only incorporate the Moho discontinuity in this study. Figure 3-6 shows the geometry of the Moho discontinuity we compiled from previous results (Sun et al. 2004). The Moho discontinuity is shallower in eastern China, ranging from 30 km to 42 km, compared to western China where it is between 50 km and 78 km.

We modified the tomographic method of Zhao et al. (1992) to invert for the crustal and uppermost mantle velocity structure in China and the surrounding area by adding the sediment layer and station corrections. Zhao's method allows 3-D velocity variations everywhere in the model and can accommodate several complex-shaped velocity discontinuities. The velocity structure is discretized using a 3-D grid. The
velocity perturbation at each point is calculated by linearly interpolating the velocity perturbations at the eight surrounding (adjacent) grid nodes. Velocity perturbations at grid nodes are the unknown parameters for the inversion procedure. To calculate travel-times and ray paths accurately and rapidly, an efficient 3-D ray-tracing technique is employed to iteratively use the pseudo-bending technique (Um and Thurber, 1987) and Snell's law. Station elevations and the sediment layers are taken into account in the ray tracing. The LSQR algorithm (Paige and Saunders, 1982) with a damping regularization is used to solve the large and sparse system of equations, allowing a great number of data to be used to solve a large tomographic problem. The nonlinear tomographic problem is solved by iteratively conducting linear inversions. At each iteration, perturbations to hypocentral parameters and velocity structure are determined simultaneously. In this study, hypocentral parameters are taken from the ABCE and only the velocities at the grid nodes are the unknowns. A detailed description of the method is given by Zhao et al. (1992, 1994) and Zhao (2001) and is also summarized in Appendix B.

3.3 Analysis

After resolution analyses, we adopt a grid spacing of 1° in the horizontal direction, and 10 km in depth (Figure 3-7). We add grid nodes at depths of 1 km, 2 km, 5 km, and 7 km to discretize the sediment layer. We have conducted many inversions using different values of the damping parameter (Figure 3-8). We found the best value of the damping parameter to be 25.0 considering the balance between the reduction of travel-time residuals and the smoothness of the 3-D velocity model obtained (Eberhart-Phillips, 1986). Convergent solutions were obtained for all the inversions. For the inversion with a
damping parameter of 25.0, the RMS travel-time residual is reduced from 0.91 s to 0.60 s, and the variance reduction is 57%. Figure 3-9 shows the change in the distribution of travel-time residuals before and after the inversion. Over 80% of the rays have residuals smaller than 0.45 s after the inversion (Figure 3-9). Figure 3-9 also shows the RMS travel-time residual is about 1.2 s if the 1-D reference model is used to calculate the travel-times. The variance reduction is about 70% between our final 3-D model and the averaged 1-D model.

The distribution of hit counts (number of rays passing through each grid node) for every layer is shown in Figures 3-10 and 3-11. Most parts of the study area are well sampled by the rays. The Tianshan area, the Sichuan-Yunnan area including the eastern Tibet, and Southeastern China including Taiwan are the regions with the most ray coverage at all depths. Ray coverage in the crust under North China is also very dense. The eastern and southern Tibet have better ray coverage than the western Tibet. Intuitively, given such good ray coverage, we expect a model with good spatial resolution. We will examine the model resolution in a systematic manner later.

Figures 3-12 and 3-13 show the obtained 3-D P-wave velocity structure together with tectonic structures. Figures 3-15 through 3-18 show vertical cross sections of the velocity images along the profiles denoted in Figure 3-14. The tomographic images are shown in areas with hit counts greater than 8.

We also conducted an inversion with a flat Moho geometry (Figures 3-19 through 3-22). We set Moho depth to 50 km, which is the average depth in the study area (Figure 3-6). Comparison of Figures 3-19 through 3-22 with Figures 3-15 through 3-18 shows that overall patterns are similar, but there are considerable changes in lower crust
anomaly amplitudes. The final RMS travel-time residual is 0.63 s for the inversion with the flat Moho, 5% higher than the 0.60 s RMS residual of the case where the Moho topography is considered. Statistic analyses show that this amount of residual reduction is significant (Huang and Zhao, 2004). Therefore, we prefer the results with the varying Moho depth model (Figures 3-15 through 3-18). The importance of taking into account the Moho depth variations and other discontinuities in the tomographic inversion has also been demonstrated in the earlier studies of the Japan and Tonga subduction zones (Zhao et al., 1992, 1997a), Southern Carpathians, Romania (Fan et al., 1998), and Southwest China (Huang et al., 2002). When the discontinuity topography is taken into account, ray paths and travel-times can be computed more accurately, and the degree of nonlinearity of the tomographic problem is reduced, thus a better tomographic result is expected.

Before analyzing the results of the tomographic inversion, we evaluate the resolution of the tomographic image. We perform a test with synthetic data to evaluate the resolution of the tomographic result. We calculate a set of travel-time delays by tracing the corresponding rays through a synthetic structure such as checkerboard, then we invert the synthetic data for the velocity structure, and finally we compare the inversion result with the initial synthetic model. In this study we adopt the checkerboard resolution test (Zhao et al., 1992) to assess the adequacy of the ray coverage and to evaluate the resolution. To make a checkerboard velocity model, we assign ±3% velocity perturbations to the 3-D grid nodes. Random errors from a normal distribution with a standard deviation of 0.1 s are added to the synthetic travel-times calculated from the synthetic models. Leveque et al. (1993) showed that in some cases small structures in a checkerboard test can be retrieved effectively while large structures are poorly imaged.
To test for such behavior in our tomographic study, we conducted checkerboard tests with different grid spacings.

Figures 3-23 through 3-30 show the input checkerboard and the corresponding resolution test results with grid spacings of 1°, 2°, 4°, and 8°. The 1° velocity model is best recovered in the Tianshan, the Sichuan-Yunnan and the Southeastern China areas to depths smaller than 40 km. This correlates well with hit count distribution patterns (Figure 3-10). When depths are greater than 50 km, the resolution is high in most of mainland China. While the resolution is good for most Tibetan regions at large depths, part of the western Tibet has relatively low resolution. The synthetic model for the Mongolia region and India area can be recovered at all depths with the grid separation larger than 8° (Figure 3-30). From these resolution tests, we infer that for mainland China and the surrounding area, the tomographic images obtained have a spatial resolution of 1° in the horizontal direction and 10 km in depth, and large-scale structures are well resolved. For the Mongolia and India areas, large grid spacing such as 8° is needed to resolve the structure. We used the 1° grid spacing in this study.

### 3.4 Results

Strong P-wave velocity variations of more than 6% found in the study area indicate the existence of significant structural heterogeneities in the crust and uppermost mantle in this region. At shallow depth (the 10 km slice), velocity images in the North China Block, the South China Block, the Sichuan Basin, the Ordos Basin, the Tarim Basin, and the Tianshan regions exhibit different patterns. In North China Block, a low velocity (low-V) anomaly is found beneath the Songliao Basin and a high-velocity (high-V) anomaly is
imaged beneath the Bohai Gulf region. Most of the Tianshan region and Tarim Basin are underlied by low-V zones. A high-V zone exists in the center of the Sichuan Basin. In the South China Block we find a high-V anomaly in the south and low-V zones in the north. Beneath the mountains surrounding the Ordos Basin in the north, west and east, a high-V anomaly is found at the depth of 10 km. The center of the Ordos basin shows a low-V anomaly. As a whole, the velocity images of shallow layers correlate well with the surface geology, topography, and lithology.

At the depth of 40 km, low-V zones are visible in the western part of China and high-V zones exist in the eastern part. A clear dividing line appears around 105°E separating the east and west seismic zones in mainland China. At depths of 50 km, 60 km and 70 km, the Tibetan plateau is clearly outlined as a low-V zone. At the depth of 80 km, the root of the Tibetan plateau disappears and high-V zones appear in the west and south part of the Tibetan Plateau. There are a few low-V zones sandwiched between the high-V zones. The deeper velocity slices (60 km, 70 km and 80 km) show scattered low-V zones in the eastern China indicating the extension of the area. These results are consistent with those obtained by a joint inversion of the DSS data from multiple profiles (Li et al., 2001).

The Pn velocity image has a high resolution in the whole study area. Pn velocities are low at the center of the Songliao Basin, but high under most of the Tarim Basin, the Sichuan Basin and the Ordos Basin. This result is generally consistent with the recent Pn tomography by Pei et al. (2004a). The detailed discussion of the Pn structure is given in 3.5.3.

Velocity changes are visible across some of the fault zones such as the Sanjiang Folding Belt. Such a feature is visible from the depth of 10 km to the depth of 30 km,
suggesting that some of the faults may have cut through the crust and reached to the middle or lower crust. No velocity contrast is visible across most of the faults, particularly in the middle to lower crust, suggesting that most of the faults may be shallow features in the upper crust. It is also possible that our tomographic model has insufficient spatial resolution to image the fault zones.

3.5 Discussion

We use the most recent earthquake data and construct a model of the velocity structure in the crust and uppermost mantle in China and the surrounding area. The Moho discontinuity is incorporated in our model. The spatial resolution is 1° in the horizontal direction and 10 km in depth. Our tomographic results reveal a large lateral heterogeneity in the crust beneath the China region. Previous tomographic studies for this region such as Liu et al. (1990) used earlier data and/or grid sizes larger than 1° in the horizontal direction. Their tomographic results have a lower resolution than the present model. Even though Zhu et al. (2002) recently obtained surface wave tomography in this region, the resolution is high only at depth larger than 70 km. Most of the previous studies in this region did not incorporate the Moho discontinuity in the tomography inversion and their final images in the lower crust and around the Moho depth therefore carry large uncertainties. In this section, we test our model by relocating the GT events and fitting the travel-times of the GT events and earthquake profiles. We use our P- and S-wave models to calculate synthetic seismograms and fit the waveforms of big events in Chapter 5.
3.5.1 Relocation of GT Events

Eleven GT events are collected from the nuclear explosions in northwestern China from 1990 to 1996 (Figure 3-31), recorded by nearly one thousand stations in China and the surrounding area. Only P-wave travel-times are available in the dataset.

The eleven GT events are closely clustered. Table 3-1 shows the parameters for these nuclear explosions. With our new P-wave velocity model for the crust and upper mantle in China and the surrounding area, we relocate the GT events using the travel-time data recorded at stations within 20° epicentral distance. The relocation errors are also listed in Table 3-1.

Table 3-1: The eleven GT events in northwest China. The relocation errors (hypocentral) are listed in km. Events are relocated using our new P-wave model, the averaged 1-D model in China and the surrounding area and the standard AK135 (global).

<table>
<thead>
<tr>
<th>Event No.</th>
<th>Date (Y/M/D)</th>
<th>Time (H/M/S)</th>
<th>Latitude (degree)</th>
<th>Longitude (degree)</th>
<th>Error (km)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>3D</td>
</tr>
<tr>
<td>1</td>
<td>1990/05/06</td>
<td>07:59:59.25</td>
<td>41.5618</td>
<td>88.7183</td>
<td>0.4</td>
</tr>
<tr>
<td>2</td>
<td>1990/08/16</td>
<td>04:59:59.26</td>
<td>41.5392</td>
<td>88.7445</td>
<td>1.2</td>
</tr>
<tr>
<td>3</td>
<td>1992/05/21</td>
<td>04:59:59.06</td>
<td>41.5419</td>
<td>88.7678</td>
<td>0.8</td>
</tr>
<tr>
<td>4</td>
<td>1992/09/25</td>
<td>08:00:00.02</td>
<td>41.7167</td>
<td>88.3767</td>
<td>1.3</td>
</tr>
<tr>
<td>5</td>
<td>1993/10/05</td>
<td>01:59:57.92</td>
<td>41.5922</td>
<td>88.7035</td>
<td>1.1</td>
</tr>
<tr>
<td>6</td>
<td>1994/06/10</td>
<td>06:25:59.46</td>
<td>41.5271</td>
<td>88.7074</td>
<td>0.9</td>
</tr>
<tr>
<td>7</td>
<td>1994/10/07</td>
<td>03:25:59.44</td>
<td>41.5741</td>
<td>88.7256</td>
<td>1.1</td>
</tr>
<tr>
<td>8</td>
<td>1995/05/15</td>
<td>04:05:59.38</td>
<td>41.5545</td>
<td>88.7516</td>
<td>0.8</td>
</tr>
<tr>
<td>9</td>
<td>1995/08/17</td>
<td>00:59:59.35</td>
<td>41.5402</td>
<td>88.7533</td>
<td>0.7</td>
</tr>
<tr>
<td>10</td>
<td>1996/06/08</td>
<td>02:55:59.37</td>
<td>41.5780</td>
<td>88.6875</td>
<td>1.0</td>
</tr>
<tr>
<td>11</td>
<td>1996/07/29</td>
<td>01:48:59.62</td>
<td>41.7162</td>
<td>88.3757</td>
<td>0.6</td>
</tr>
</tbody>
</table>

The averaged hypocentral error of relocation is only 0.9 km with a standard deviation of 0.3 km for our new 3-D P-wave model. The mean hypocentral misfit is about 10 km if the averaged 1-D velocity in China is used and it is about 20 km if the global
AK135 model is used. Even though the GT events are located in northwestern China with most stations distributed in southern and eastern China, the small relocation errors clearly suggest that our P-wave model provides accurate travel-times for event location.

### 3.5.2 Travel-time Validation

Another way of validating velocity models is to compare the observed travel-times from the GT events and earthquakes with the predicted travel-times calculated with the new model. The eleven explosions essentially occurred in two testing areas (41.55°, 88.73°) and (41.72°, 88.38°). The distance between these two areas is less than 40 km, which is relatively small compared to the average source-receiver distance of 12° from all the selected stations. Instead of comparing the travel-time residuals at each station from each individual event, we calculate the averaged travel-time residual at each station from all the GT events. The mean travel-time residuals at all the selected stations are shown in Figures 3-32, 3-34, 3-36 and 3-38 when the AK135, averaged 1-D, the input 3-D and final 3-D models are used, respectively. The standard deviations of travel-time residuals for each event are plotted in Figures 3-33, 3-35, 3-37 and 3-39. We see that both the travel-time residuals and the standard deviations decrease when the models are used in order of AK135, averaged 1-D, the input 3-D and the final 3-D.

Figure 3-38 shows that the mean travel-time residual (our final 3-D model) is smaller than 0.6 s when the stations are within 15° epicentral distance. The travel-time residuals increase to about 0.9 s when the stations are about 20° from the events. The larger travel-time residuals observed at larger distances are due to the fact that our final 3-D model is obtained based on the travel-times from the source-receiver distance smaller
than 15° and the deeper structures (deeper than 200 km) of the model are less constrained in our tomographic study from regional travel-time data. The standard deviations are smaller than 0.2 s. Figures 3-38 and 3-39 clearly confirm that the P-wave models we obtained from travel-time tomography in the crust and uppermost mantle are accurate.

It is also important to compare the travel-times from the earthquakes with well constrained hypocenters. Since all the GT events we collected are located in the northwest part of China, with a 20° limit in epicentral distance from station to event, the model structure in the eastern part of China is not validated by relocating the GT events or comparing the travel-times from the GT events. To test the eastern China velocity model, we select one large, well-located earthquake (M 6.4) in the northern part of China. The event is recorded by more than 100 stations (Figures 3-40 and 3-41). The event occurred on May 03, 1996 with the epicenter at (40.72N, 109.57E), depth of 28 km and origin time 03-32-46.3 (hour-minute-second), given in the ABCE. Our final location of this event is (40.72N, 109.58E) at a depth of 25 km and origin time 03-32-45.9. The hypocentral difference for this event between ABCE and our model is about 3.5 km, larger than the averaged hypocentral misfit of 0.9 km for the northwestern China GT events.

The observed and predicted P-wave travel-times within 20° source-station separation are plotted in Figure 3-40 (AK135) and in Figure 3-41 (final 3-D model). The travel-time residuals are much smaller in Figure 3-41 than in Figure 3-40. The mean travel-time residual is 1.89 s for AK135 and it is only 0.09 s for our final model. The standard deviation of travel-time residuals is 1.2 s for AK135 and it is 0.49 s for our final model.
3.5.3 Comparison of Pn Structure

Our initial Pn velocities were constructed based on the Pn structure of the 1-D Monte-Carlo search (Chapter 2). Figure 3-42 shows the Pn velocities obtained by Tom Hearn (2001), the 1-D Monte-Carlo search, the initial 3-D and final 3-D structure. The prominent low Pn zones in eastern China are consistent throughout all the Pn models. Southern and eastern Tibet show a high Pn velocity. We also observe a high Pn velocity beneath the Sichuan Basin, Ordos Basin and Tianshan area. The discrepancies are mostly in the Tibetan region. Tom Hearn (2001) observed a high Pn velocity beneath the Qilian and the Kunlun area. We only observe a high Pn beneath the eastern Qilian in our final model. Some part of the Kunlun area shows high Pn in our 1-D Pn result. This discrepancy might be caused by the over-averaging effect of the 1-D inversion. The high Pn in the eastern Tibet is carried over to the western part due to the sparse ray coverage in the western part and very large windows that are selected to provide averaged 1-D profiles in the Monte-Carlo inversion.

3.5.4 Effects of Moho Discontinuity on Tomography

The Moho discontinuity plays an important role in tomography. The Moho depths are fixed in our inversion. In this section we allow the Moho depths to change, and both Moho depths and velocities are unknowns in the inversion. We modify Zhao’s code by adding a Monte-Carlo search for the initial Moho depths. We first generate a Moho depth by randomly selecting a value between -5 km and 5 km of the initial model. We do this at every 4° x 4° grid. We inverted a model that gives the minimum RMS travel-time
residual when the Moho depths are fixed during the inversion. The model chosen represents one realization. We repeat this 10,000 times and the final (best) model is selected with the minimum RMS residual from the 10,000 realizations. Figure 3-43 shows the uniform distribution of Moho depth perturbations at 10,000 Monte-Carlo realizations. Figure 3-43 also shows the best 20 RMS travel-time residuals sorted from the 10,000 Monte-Carlo iterations. The Moho depth difference and Pn velocity difference are shown in Figure 3-44. We see the Moho depth difference is within ±1 km and the Pn velocity difference is within ±1 km/s. We also show both the velocity and the Moho depth differences along a vertical cross section (Profile E in Figure 3-16) in Figure 3-45. The final velocity difference is within ±1 km/s in the crust and uppermost mantle and it is smaller than ±0.5 km/s for more than 90 percent of the cross section. The Moho depth difference is within ±1 km throughout the cross section. The final Moho depth and velocity tomography obtained by 3-D random search are consistent with the ones obtained with the Moho depths fixed. It is not only due to the fact that both models are obtained using the Monte-Carlo search, but also the fact that the two models are close to the true structure of the study area.

3.6 Conclusions

We use a large number of high-quality arrival times to determine a detailed 3-D P-wave velocity model of the crust and uppermost mantle beneath China and the surrounding area. This model has a higher resolution than previous results and provides new information about the complex structure and seismotectonics of the region. The main findings of the present work are summarized as follows.
The seismic velocity images are characterized by block structures corresponding to geological features bounded by large fault zones. This region consists of a few geological structures: the North China Block including Songliao Basin, the South China Block, the Sichuan Basin, the Tarim Basin, and the Tianshan area. Those areas exhibit different patterns of velocity distribution in the tomographic images. The trend of velocity anomalies is consistent with the trend of regional tectonics.

The velocity images of the shallow crust (at a depth of 10 km) accurately reflect the surface geology and topographic features. Basins and depressions are underlied by low-velocity anomalies, while high-velocity zones are imaged beneath the uplift and mountainous areas.

A clear dividing line along the 105° parallel separates China into a low-V zone in the west and a high-V zone in the east at a depth of 40 km.

Our tomographic imaging has revealed significant velocity heterogeneities in the middle and lower crust, some of which are consistent with those detected by deep seismic soundings and other geophysical investigations.

Acknowledgments

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and M. Willis from MIT-ERL provide great help on improving the manuscript. All the figures in this work are made by using GMT (Wessel and Smith, 1995).
References


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Figure Captions

Figure 3-1: 25,000 earthquakes, 220 stations, active faults and major tectonic boundaries in China and the surrounding area. Earthquake epicenters are shown in red dots and stations are shown in red triangles. The yellow line shows the boundary of China. Active faults in the China area are shown in purple lines and tectonic sutures are shown in blue lines, where SoB — Songliao Basin, OB — Ordos Basin, SB — Sichuan Basin, KB — Khorat Basin, STB — Shan Thai Block, IB — Indochina Block.

Figure 3-2: Major earthquakes with M 6.0 or greater between Jan. 1978 and May 2004 in China and the surrounding area. There are in total 542 big events shown in red dots. The 1679 Sanhe M 8.0 event, the 1951 Dangxiong M 8.0 event, the 1976 Tangshan M 7.8 event and the 2001 Ruoqiang M 8.1 event are shown as white stars.

Figure 3-3: 25,000 earthquakes, 220 stations, and 500,000 ray paths in China and the surrounding area. Earthquake epicenters are shown in black circles and stations are shown in red triangles. The green line shows the boundary of China.

Figure 3-4: Distribution of seismic stations and epicenters of selected earthquakes used in this study. Totally 16,572 events from M 3.0 above are selected and plotted in black circles. All the 220 stations are used and plotted in red triangles.

Figure 3-5: The averaged 1-D velocity model used for the tomographic inversion.

Figure 3-6: Depth distribution of the Moho discontinuities in the present study area which were constructed in Chapter 2 by inverting the 1-D layered models from first arrivals of P-wave travel-time. The Moho depths are shown in contours.

Figure 3-7: Three-dimensional configuration of the grid adopted in the present study.
Figure 3-8: Trade-off curve for the variance of the velocity perturbations and root-mean-square travel-time residuals. Numbers beside the black dots denote the damping parameters adopted for the inversions. The largest black dot denotes the optimal damping parameter for the final tomographic model.

Figure 3-9: Number of rays versus travel-time residuals. The red line denotes the result for the averaged 1-D model. The green line denotes the result for the starting model (before the inversion). The blue line denotes the result after the inversion.

Figure 3-10: Distribution of the number of the rays passing through each grid node (hit counts). The depth of the layer is shown at the lower-left corner of each map. The color bars on the right side of each map show the hit count scale.

Figure 3-11: Similar to Figure 3-10. A different color bar at the bottom shows the scale of hit count less than 100. Purple areas show where there are more than 100 hits.

Figure 3-12: P-wave velocity anomaly at each depth slice (in percent from the average velocity). The depth of each layer is shown at the lower-left corner of each map. Red and blue colors denote low and high velocities, respectively. The velocity perturbation scale is shown at the bottom.

Figure 3-13: P-wave velocity image at each depth slice. The depth of each layer is shown at the lower-left corner of each map. The color scales are shown in the middle (for the depths of 10 km, 20 km, 30 km, and 40 km) and at the bottom (for the depths of 50 km, 60 km, 70 km and 80 km).

Figure 3-14: Locations of the vertical cross sections shown in Figures 3-14 through 3-17.
Figure 3-15: Vertical cross sections of P-wave velocity when the Moho depth variations (Figure 3-6) are taken into account. Cross sections A, B and C at the longitudes of 80°, 90° and 100° are plotted. The surface topography along each profile is shown on the top of each cross section. The black curved lines show the Conrad (dashed) and Moho (solid) discontinuities. Each grid in the region between the two white lines has a raypath hit count of 200 or above.

Figure 3-16: Similar to Figure 3-15. Cross sections D, E, and F at the longitudes of 110°, 120° and 130°.

Figure 3-17: Similar to Figure 3-15. Cross sections G, H, I and J at the latitudes of 15°, 20°, 25° and 30°.

Figure 3-18: Similar to Figure 3-15. Cross sections K, L, M and N at the latitudes of 35°, 40°, 45° and 50°.

Figure 3-19: The same as Figure 3-15, but when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 3-20: The same as Figure 3-16, but when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 3-21: The same as Figure 3-17, but when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 3-22: The same as Figure 3-18, but when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 3-23: The input checkerboard when the grid shown in Figure 3-7 is adopted. The depth of the layer is shown at the lower-left corner of each map. Blue and
red squares denote high and low velocities, respectively. The velocity perturbation scale is shown at the bottom.

Figure 3-24: Results of checkerboard resolution test of the input checkerboard shown in Figure 3-23.

Figure 3-25: The input checkerboard with the grid spacing of 2°.

Figure 3-26: Results of checkerboard resolution test of the input checkerboard shown in Figure 3-25.

Figure 3-27: The input checkerboard with the grid spacing of 4°.

Figure 3-28: Results of checkerboard resolution test of the input checkerboard shown in Figure 3-27.

Figure 3-29: The input checkerboard with the grid spacing of 8°.

Figure 3-30: Results of checkerboard resolution test of the input checkerboard shown in Figure 3-29.

Figure 3-31: The collected ground-truth (GT) events in northwest China from 1990 to 1996 and the stations in China and surrounding areas. A total of eleven GT events took place in the same area at a maximum distance of 40 km from each other. The red star denotes the central location of the events. The stations are plotted as red triangles.

Figure 3-32: The averaged travel-time residuals at the stations within 20° epicentral distance to the center of the eleven GT events. The mean of the travel-time residuals from each station to all the GT events is plotted in red (positive) and blue (negative) circles. The star denotes the center of the GT events. The reference model is the AK135.
Figure 3-33: The standard deviations (STD) of the travel-time residuals at the stations within 20° epicentral distance to the center of the eleven GT events. The STD of the travel-time residuals from each station to all the GT events is plotted in blue circles. The star denotes the center of the GT events. The reference model is the AK135.

Figure 3-34: Similar to Figure 3-32. The averaged 1-D model is used to calculate travel-times.

Figure 3-35: Similar to Figure 3-33. The averaged 1-D model is used to calculate travel-times.

Figure 3-36: Similar to Figure 3-32. The input 3-D model is used to calculate travel-times.

Figure 3-37: Similar to Figure 3-33. The input 3-D model is used to calculate travel-times.

Figure 3-38: Similar to Figure 3-32. The final 3-D model is used to calculate travel-times.

Figure 3-39: Similar to Figure 3-33. The final 3-D model is used to calculate travel-times.

Figure 3-40: The travel-time residuals at the stations within 20° epicentral distance to the earthquake (M 6.4). The travel-time residuals from each station to the event are plotted in red (positive) and blue (negative) circles. The star denotes the epicenter of the event. The reference model is the AK135. The selected earthquake occurred on May 03, 1996 in Northern China with the epicenter of (40.72°N, 109.57°E).

Figure 3-41: Similar to Figure 3-40. The final 3-D model is used to calculate travel-times.
Figure 3-42: Comparison of Pn structure.

Figure 3-43: Top: Distribution of input Moho depth difference. The input Moho depth variations are taken to be ±5 km of the initial 3-D model. There are in total 10,000 Monte-Carlo realizations. Bottom: The best 20 minimum RMS travel-time residuals over 10,000 Monte-Carlo realizations. The RMS residual is 0.60 s when Moho depths are fixed in the inversion.

Figure 3-44: Effects of Moho discontinuity on velocity tomography. Top: The Moho depth difference between the fixed Moho and the varying Moho. The initial (reference) Moho depths are taken from the inversion results in Chapter 2. All the tomographic results shown in this chapter are obtained with Moho depths fixed in the inversion. If the Moho depths are allowed to change in the inversion with the input depths to be within ±5 km of the reference model, the output Moho depths are still within ±1 km of the reference model. Bottom: The Pn velocity difference.

Figure 3-45: Effects of Moho discontinuity on velocity tomography. (A): Elevation plot of Profile E shown in Figure 3-16. (B): Vertical cross section of P-wave velocity at Profile E in Figure 3-16 when the Moho depths are fixed in the inversion. (C): The same vertical cross section when the input Moho depth variations are taken to be ±5 km of the initial 3-D model. (D): The velocity difference between the plots B and C. (E): The Moho depth difference.
Figure 3-1: 25,000 earthquakes, 220 stations, active faults and major tectonic boundaries in China and the surrounding area. Earthquake epicenters are shown in red dots and stations are shown in red triangles. The yellow line shows the boundary of China. Active faults in the China area are shown in purple lines and tectonic sutures are shown in blue lines, where SoB – Songliao Basin, OB – Ordos Basin, SB – Sichuan Basin, KB – Khorat Basin, STB – Shan Thai Block, IB – Indochina Block.
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Figure 3-29: The input checkerboard with the grid spacing of $8^\circ$. 
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Figure 3-31: The collected ground-truth (GT) events in northwest China from 1990 to 1996 and the stations in China and the surrounding area. A total of eleven GT events took place in the same area at a maximum distance of 40 km from each other. The red star denotes the central location of the events. The stations are plotted as red triangles.
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Figure 3-35: Similar to Figure 3-33. The averaged 1-D model is used to calculate travel-times.
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Chapter 4

S-wave Tomography of the Crust and Uppermost Mantle in China

Abstract

China and the surrounding area is characterized by complex Cenozoic structures and active seismotectonics. In this study, we have used over 260,000 arrival times from 12,215 local earthquakes recorded by 220 seismic stations to determine a detailed three-dimensional (3-D) S-wave velocity structure of the crust and uppermost mantle in this region. We have taken into account the complex morphology of the Moho discontinuity when discretizing the model for the tomographic inversions. This approach leads to a better result than that with a flat Moho, as in previous studies. Our results show that large velocity variations of more than 6% exist in the crust and upper mantle in the China region. The velocity image of the upper crust correlates with surface geological features. The crustal heterogeneity is clearly observed. Velocity changes are visible across some of the large fault zones, and the faults and some large crustal earthquakes seem to occur at the boundary areas between slow and fast velocity anomalies. Some of the faults, such as the Red River fault, may have cut through the crust and reached up to the upper mantle. Low velocity zones beneath volcanic sites and the rifts are clearly observed in our tomographic results. Under the Tengchong volcanic area, strong low-velocity zones are
visible down to 100 km depth, with a lateral extent of about 100 km, suggesting the existence of magma chambers under the volcano. Low velocity zones beneath other volcanic sites and the rifts are clearly observed in our tomographic results.

KEYWORDS: lithosphere, seismic structure, seismic tomography, active volcanoes, active faults, China
4.1 Introduction

The present study region includes China and its surrounding area (Figure 4-1). Due to the complicated tectonic movements and continental deformation (Yin and Chen, 2004), abundant seismicity has been recorded in Chinese history. Historically, large and destructive earthquakes took place frequently, both in east and west China (Figure 4-2). In 1833 the Songming earthquake (M > 8.0) killed tens of thousands of people (Gu, 1983). The 1976 Tangshan earthquake (M 7.8), with the epicenter approximately 160 km southeast of Beijing, totally destroyed the Tangshan city (then population 1 million) and killed about 240,000 people. It was perhaps the most destructive earthquake in human history. The Lijiang earthquake (M 7.0) on 3 February 1996 caused over 1000 deaths and extensive property losses. The 2001 Ruoqiang earthquake (also known as the Kunlun earthquake) (M 8.1) in Xinjiang-Qinghai area caused huge damage to the highways and railroad under construction (Qiao et al. 2002; Xu et al., 2002b). Most large earthquakes in this region are located at the depths of 10 to 20 km. A detailed investigation of the crustal structure and seismotectonics of this region is very important for understanding the physics of continental earthquakes and for assessing and mitigating seismic hazard.

The available S-wave velocity models of the crust and upper mantle in China and the surrounding area have been obtained using different approaches. Global models such as CUB 1.0 (Shapiro and Ritzwoller, 2002) and the SAIC 1° x 1° model (Stevens et al., 2001) were constructed from group and phase velocity dispersion measurements of surface waves. The models based on surface waves are generally large-scale and intended to map the deep structure of the Earth. The global model CRUST 2.0 (Laske et al., 2001) was constructed from seismic refraction data and developed from the CRUST
5.1 model (Mooney, 1998) and a 1° x 1° sediment map (Laske and Masters, 1997). Only P-wave velocities are inverted by travel-time tomography. The S-wave velocities in the model are obtained by empirical Vp/Vs ratios or compiled from other sources. There are too few deep seismic soundings (DSSs) to provide enough refraction data for a detailed model. Regional models were constructed by Sn and/or Pn tomography (Ritzwoller et al., 2002; Hearn and Ni, 2001; Pei et al., 2004), and from surface waves (Wu et al., 1997, Lebedev and Nolet, 2003; Huang et al., 2003; Song et al., 1991; Zhu et al., 2002). Joined with P-wave tomographic inversion, S-wave tomography has been performed in several local regions in China (Xu et al., 2002a; Huang et al., 2002; Yu et al., 2003; Huang and Zhao, 2004). Xu et al. used P- and S-wave arrival times of local, regional and teleseismic events recorded by Chinese and Kyrgyzstan seismic networks to tomographically map the crustal and upper mantle velocity structure beneath western China. The grid spacing is 1.5° x 1.5° in the horizontal direction and about 10 km (in the crust) and 50 km (in uppermost mantle) in the vertical direction. Crustal structure models beneath north and east China, including Beijing and surrounding regions, were obtained by Zhu et al. (1990), Yu et al. (2003) and Huang and Zhao (2004) as the result of a P- and S-wave tomographic study. Huang et al. (2002) inverted the lithospheric structure in southwest China from local/regional travel-time data. These models provide a detailed crustal structure only in a few regions. A detailed map for the whole China area remains to be developed.

We need 3-D velocity models that cover all of China with good resolution everywhere to provide accurate travel-times for reliable determination of earthquake location. Models obtained by the teleseismic tomography technique generally cannot
resolve the vertical variations in the shallow structure. Regional models can be combined to cover a large area, but such models cannot guarantee smooth and consistent transitions between different regions.

During the last few years, there have been many digital seismic stations installed in China (Figure 4-1). The large database of high-quality recorded arrival times provides an unprecedented opportunity to determine a detailed 3-D crustal structure under the region. Therefore, we will introduce a method that describes how to construct 3-D S-velocity models based on observed travel-time data.

In the present work, we have applied a new tomographic method to a large data set of local earthquake arrival times to determine a high-resolution 3-D S-wave velocity structure of the lithosphere under this region. Compared with the previous tomographic studies (Liu et al., 1989; Sun et al., 1991), we have used a much larger data set and an updated tomographic method (Zhao et al., 1992) to take into account the effects of the complex Moho geometry in this region. Our results cast a new light on the complex seimotectonics and volcanic activity in China.

### 4.2 Data and Method

For this work, we selected arrival time data from the ABCE (IG-CSB, 1990-2002). In this database there are 25,000 earthquakes, 220 stations, and 450,000 S-wave ray paths in China and the surrounding area, from January 1990 to December 2002. Figure 4-3 shows earthquake epicenters and stations in China. The selection of earthquakes is based on the following criterion: (1) Events that occurred in the study area with magnitudes greater than M 3.0. For the northwest area with less seismicity, we selected some earthquakes
with M 2.5-3.0. (2) All the selected events were recorded by at least 15 stations. (3) Earthquakes were selected to provide a uniform distribution of hypocenters in the study area. As a result, 262,000 S-wave arrival times from 12,215 earthquakes were selected to determine the S-wave velocity structure. The picking accuracy of arrival times is in general 0.1-0.4 s for the S-wave. Figure 4-4 shows the epicentral distribution of the 12,215 earthquakes selected.

To analyze the arrival time data, we have used the tomographic method of Zhao et al. (1992). Although the conceptual approach of this method parallels that of Aki and Lee (1976), it has the following additional features:

1. It is adaptable to a velocity structure that includes several complex-shap

2. To calculate travel-times and ray paths accurately and rapidly, an efficient 3-D ray-tracing technique (Zhao et al., 1992) is employed that iteratively uses the pseudobending technique (Um and Thurber, 1987) and Snell’s law. Station elevations are taken into account in the ray tracing.

3. The LSQR algorithm (Paige and Saunders, 1982) with a damping regulariza

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LSQR algorithm has been used by a number of researchers (Nolet, 1985; Spakman and Nolet, 1988; Papazachos and Nolet, 1997) and turns out to be an efficient algorithm to solve a large and sparse system of equations.

4. The nonlinear tomographic problem is solved by iteratively conducting linear inversions. In each iteration, perturbations to hypocentral parameters and velocity structure are determined simultaneously. In this study, only the velocities are the unknowns and the hypocentral parameters are taken from the ABCE. A detailed description of the method is given in Appendix B.

4.3 Analysis and Results

Figure 4-5 shows the 3-D configuration of the grid. After resolution analyses, we adopted a grid spacing of 1° in the horizontal direction, and 10 km in depth. We also set grids at the depths of 1 km, 2 km, 5 km and 7 km to better sample the sediment layer. A previous study by Sun et al. (2004b) provides the starting 1-D model (Figure 4-6) for our tomographic inversions. According to the studies by Sun et al. (2004b) and Li and Mooney (1998), the Moho discontinuity in this region is not a simple flat plane but has a complex geometry. Based on the study by Sun et al. (2004b), we constructed a map of Moho depths, as shown in Figure 4-7. The Moho depths in this region determined by different researchers are quite consistent with each other in both pattern and depth values. The difference between different models is less than 4 km. In the study area, the Moho depth ranges from 30 km in the south and east to 78 km in the west. During the tomographic inversion, the Moho geometry is fixed and only the velocities at grid nodes are determined. We have conducted over 10 inversions using different values of damping
parameters. We found the best value of the damping parameters to be 25.0, considering
the balance between the reduction of travel-time residuals and the smoothness of the 3-D
velocity model obtained (Figure 4-8). A convergent solution was obtained through the
inversion. The root mean square (RMS) travel-time residual was reduced by 35% from
0.98 s to 0.72 s (Figure 4-9). An average of 70% RMS reduction has been observed,
compared to the average 1-D velocity model in the study area. Figures 4-10 and 4-11
show the resulting 3-D S-wave velocity structure as a perturbation plot and as an absolute
velocity plot. The ray coverage distribution in the study region is shown in Figures 4-12
and 4-13. Figures 4-15 through 4-18 show vertical cross sections of the velocity images
along the profiles denoted in Figure 4-14. We also conducted an inversion with a flat
Moho geometry (Figures 4-19 through 4-22). The detailed discussions about the results
with irregular Moho and flat Moho will be given in Section 4.4.1.

Before describing the obtained tomographic results, we first evaluate the
resolution of the tomographic image. A direct way to evaluate the resolution of a
tomographic result is to calculate a set of travel-time delays that result from tracing the
corresponding rays through a synthetic structure as though they were data, and then to
compare the inversion result with the initial synthetic structure. In this study we adopt the
checkerboard resolution test (Spakman and Nolet, 1988; Zhao et al., 1992) to assess the
adequacy of the ray coverage and to evaluate the resolution. To make a checkerboard,
positive and negative 3% velocity perturbations are assigned to 3-D grid nodes that are
arranged in an alternating pattern in the model space. Therefore, by just seeing the image
of the synthetic inversion of the checkerboard, one can easily understand where the
resolution is good and where it is poor. Random errors in a normal distribution with a
standard deviation of 0.3 s are added to the theoretical travel-times calculated for the synthetic models. Leveque et al. (1993) showed that in some cases small-sized structures in a checkerboard test can be retrieved effectively while larger structures are poorly retrieved. To know whether such a problem occurs in our tomographic results, we conducted six checkerboard tests with different grid spacings.

Figures 4-23 through 4-30 show the results of checkerboard resolution tests with grid separations of 1°, 2°, 4°, and 8°. The resolution is generally high in the Tianshan area, Sichuan-Yunnan area and Southeast China. This result is consistent with the ray coverage plot in Figure 4-12. In the deeper regions the checkerboard pattern is, in general, correctly reconstructed, although the amplitudes of the anomalies are not completely restored. At the edge of the study area the resolution is poor, and the pattern is not correctly reconstructed, as expected. We can see that the resolution is quite high and the checkerboard pattern and anomaly amplitude are correctly reconstructed for most of the study area, with a grid separation of 1°. The Mongolia region and the Indian plate require grid separations of 8° to fully recover the synthetic patterns in the areas. From these resolution tests we can infer that the tomographic images obtained in the China region have a spatial resolution of 1° in the horizontal direction and 10 km in depth, and that large-scale structures can also be well resolved.

In the following section, we describe major features of the tomographic images we obtained. Strong S-wave velocity variations of more than 6% are found in the study area, indicating the existence of significant structural heterogeneities in the crust and upper mantle in this region. At shallow depths (the 10 km slice), velocity images in the North China Block, the South China Block, the Sichuan Basin, the Ordos Basin, the
Tarim Basin and the Tianshan region exhibit different patterns. In the North China Block, a low velocity (low-V) anomaly is found beneath the Songliao Basin and a high-velocity (high-V) anomaly is imaged beneath the Bohai Gulf. Most of the Tianshan region and Tarim Basin are underlain by low-V zones. A high-V zone exists in the center of the Sichuan Basin. In the South China Block we find a high-V anomaly in the south and low-V zones in the north. Beneath the mountains surrounding the Ordos Basin in the north, west and east, a high-V anomaly is found at a depth of 10 km. The center of the Ordos basin shows a low-V anomaly. As a whole, the velocity images of shallow layers correlates well the surface geology, topography, and lithology.

At the depth of 40 km, low-V zones are visible in the western part of China, and high-V zones exist in the eastern part. A clear dividing line appears around 105°E separating the east and west seismic zones in mainland China. At depths of 50 km, 60 km and 70 km, the Tibetan plateau is clearly outlined as low-V zones. At a depth of 80 km, the root of the Tibetan plateau disappears and high-V zones appear in the western and southern part of the Tibetan Plateau. There are a few low-V zones sandwiched between the high-V zones. The deeper velocity slices (60 km, 70 km and 80 km) show scattered low-V zones in eastern China, indicating the extension of the area.

The Sn velocity image has a high resolution in the whole study area. Sn velocities are low at the center of the Songliao Basin, but high under most of the Tarim Basin, the Sichuan Basin and the Ordos Basin. This result is generally consistent with the recent Sn tomography by Pei et al. (2004). The detailed discussion of the Sn structure is given in section 4.4.3.
Velocity changes are visible across some of the fault zones such as the Sanjiang Folding Belt. This feature is visible from the depth of 10 km to the depth of 30 km, suggesting that some of the faults may have cut through the crust and reached to the middle or lower crust. No velocity contrast is visible across most of the faults, particularly in the middle to the lower crust, suggesting that most of the faults may be just a shallow feature in the upper crust. It is also possible that our tomographic model has insufficient spatial resolution to image fault zones.

Liu et al. (1989) and Sun et al. (1991) found a similar feature in their earlier tomographic studies. Beneath the Tengchong volcanic area, a strong low-velocity anomaly of up to 7% is visible from the crust down to 100 km depth (Profile C, Figure 4-32). The slow anomalies have a lateral extent of over 100 km. This result suggests that high-temperature magma chambers exist under the Tengchong volcano, which have deep origins, probably from a deep portion of the lithosphere or the asthenosphere. The Tengchong volcanism is discussed extensively in section 4.4.2.

### 4.4 Discussion

#### 4.4.1 Effects of Moho Topography on Velocity Tomography

As mentioned above, there is a large lateral variation in the thickness of the crust in the China region, ranging from 30 km in the east to 78 km in the west (Figure 4-7). The previous tomographic studies did not take into account the depth variations of the Moho discontinuity. In the tomographic inversions of the present study, we have taken into account the Moho topography. To understand the effect of using a variable crustal
thickness, we compared these results (Figures 4-19 through 4-22) with the results (Figures 4-15 through 4-18) of a tomographic inversion performed with a flat Moho at a depth of 50 km, the average for this region. The 10, 20 and 80 km depth slices show little change in the velocity images (Figures 4-16 and 4-20). However, considerable changes in velocity images are found in the 40, 50 and 60 km depth slices because the Moho depth varies in this range. The low-V anomaly under the Tengchong volcano is clearly visible in Figure 4-15, but it is not obvious when the Moho becomes flat (Figure 4-19).

Note that the RMS travel-time residual is 0.72 s when the Moho topography is considered and 0.81 s when the Moho is flat. Statistical analyses show that the change of travel-time residuals is significant (Huang and Zhao, 2004). These results suggest that it is important to consider the changes of the crustal thickness in tomographic inversions to determine a detailed seismic velocity structure. The importance of taking into account depth variations of the Moho and other discontinuities in the tomographic inversion has also been demonstrated in the earlier studies of the Japan and Tonga subduction zones (Zhao et al., 1992, 1997) and Southern Carpathians, Romania (Fan et al., 1998). When the topography discontinuity is taken into account, ray paths and travel-times can be computed more accurately, and the degree of nonlinearity of the tomographic problem is reduced. Thus a better tomographic result is expected.

4.4.2 Intraplate Volcanism

Tengchong is one of a few active volcanic areas in the study region (Profile C in Figure 4-31). Its most recent eruption occurred in 1609 (Qin et al., 1996). There are also many hot springs in this region. This volcanic area is situated between 24.67-25.50N latitude
and 98.25-98.67E longitude and covers a rectangular area of 90 km x 40 km (Huang et al. 2002).

The seismic activity and heat flow in the area south of the Tengchong volcanic center are higher than north of the volcanic center (Qin et al., 1996). The 1976 Longling earthquakes took place south of Tengchong. Our tomographic images show that a larger portion of the low-V zone exists under the southern part of the volcanic center than under its northern part (Figure 4-32), being quite consistent with the surface geothermal activity and crustal seismicity. In addition to the low seismic velocity and high heat flow, the Tengchong area exhibits a negative gravity anomaly and low electric resistivity in the crust and upper mantle (Sun et al., 1989; Kan et al., 1996). Analyzing seismic surface wave and gravity data Kan et al. (1996) estimated that the lithosphere has a thickness of 60-64 km under Tengchong, much thinner than that under other regions of mainland China (about 100 km) (Liu et al., 1989).

Similar low-V zones exist beneath the volcanic sites (Profiles A, B, F, I and J) and beneath the rifts (Profile H). Evidence of the presence of the Hainan plume has been shown in Lebedev and Nolet (2003). The Vp/Vs ratios and Poisson’s ratios clearly show a positive anomaly in the above profiles (in Chapter 5).

### 4.4.3 Comparison of Sn Structure

The Sn velocities obtained by 1-D Monte-Carlo inversion, the initial 3-D model and final 3-D tomography are shown in Figure 4-35. Similar features are observed in three plots. The prominent low Sn zones in eastern China consistently exist in all the Sn models. Southern and eastern Tibet clearly show high Sn anomaly. We also observe high Sn
anomaly beneath Sichuan Basin, Ordos Basin and Tianshan area. The discrepancies between our initial and final 3-D Sn are minor. Some part of the Kunlun area shows high Sn in our 1-D result but not in the final 3-D model. This discrepancy might be caused by the over-averaging effect of the 1-D inversion. The high Sn in eastern Tibet was carried over to the western part due to the sparse ray coverage in the western part and very large windows were selected to provide averaged 1-D profiles in Monte-Carlo inversion. The comparison between the final Sn and Pn structure will be discussed in Chapter 5.

4.5 Conclusion

In the recent study we have determined detailed S-wave tomographic images for China and its surrounding area. Our results shed new light on the complex structure and tectonics of this region. The most significant results of this work can be summarized as follows.

1. The velocity image of the upper crust is correlated with surface geological features. Basins and depressions are in low-velocity anomalies, while high-velocity zones are imaged beneath the uplift and mountainous areas.

2. A clear dividing line along the 105° parallel separates China into a low-V zone in the west and a high-V zone in the east, at depths of 40 km.

3. Large crustal earthquakes generally occurred at the boundary between slow and fast velocity anomalies or above low-velocity zones in the lower crust and uppermost mantle. The low-V zones may represent high temperature anomalies or magma chambers or contain fluids, which may cause the weakening of the seismogenic layer in the upper
crust. The weak sections of the seismogenic crust are subject to tectonic stress and hence prone to large crustal earthquakes.

4. Under the Tengchong volcanic area, strong low velocity anomalies are visible down to 100 km depth and have a lateral extent of about 100 km. This feature is consistent with other geophysical observations in this area, such as high heat flow, negative gravity anomaly, and low electric resistivity, which suggest the existence of high temperature magma chambers under Tengchong.

5. Low velocity zones beneath other volcanic sites and the rifts are clearly observed in our tomographic results.

Acknowledgments

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References


Figure Captions

Figure 4-1: 25,000 earthquakes, 220 stations, active faults and major tectonic boundaries in China and the surrounding area. Earthquake epicenters are shown in red dots and stations are shown in red triangles. The yellow line shows the boundary of China. Active faults in the China area are shown in purple lines and tectonic sutures are shown in blue lines, where SoB — Songliao Basin, OB — Ordos Basin, SB — Sichuan Basin, KB — Khorat Basin, STB — Shan Thai Block, IB — Indochina Block.

Figure 4-2: Big earthquakes with M 6.0 or above between Jan. 1978 and May 2004 in China and the surrounding area. There are in total 542 big events shown in red dots. The 1679 Sanhe M 8.0 event, the 1951 Dangxiong M 8.0 event, the 1976 Tangshan M 7.8 event and the 2001 Ruoqiang M 8.1 event are shown in white star.

Figure 4-3: 25,000 earthquakes, 220 stations, and 450,000 S-wave ray paths in China and the surrounding area. Earthquake epicenters are shown in black circles and stations are shown in red triangles. The green line shows the boundary of China.

Figure 4-4: Distribution of seismic stations and epicenters of selected earthquakes used in this study. Totally 12,215 events from M 3.0 above are selected and plotted in black circles. All the 220 stations are used and plotted in red triangles.

Figure 4-5: Three-dimensional configuration of the grid adopted in the present study.

Figure 4-6: The averaged 1-D velocity model used for the tomographic inversion.

Figure 4-7: Depth distribution of the Moho discontinuities in the present study area which were constructed in Chapter 2 by inverting the 1-D layered models from first arrivals of S-wave travel-time. The Moho depths are shown in contours.
Figure 4-8: Trade-off curve for the variance of the velocity perturbations and root-mean-square travel-time residuals. Numbers beside the black dots denote the damping parameters adopted for the inversions. The largest black dot denotes the optimal damping parameter for the final tomographic model.

Figure 4-9: Number of rays versus travel-time residuals. The red line denotes the results for the averaged 1-D model in the study area. The green line denotes the result for the starting model (before the inversion). The blue line denotes the result after the inversion.

Figure 4-10: S-wave velocity anomaly at each depth slice (in percent from the average velocity). The depth of each layer is shown at the lower-left corner of each map. Red and blue colors denote low and high velocities, respectively. The velocity perturbation scale is shown at the bottom.

Figure 4-11: S-wave velocity image at each depth slice. The depth of each layer is shown at the lower-left corner of each map. Red and blue colors denote low and high velocities, respectively.

Figure 4-12: Distribution of the number of the rays passing through each grid node (hit counts). The depth of the layer is shown at the lower-left corner of each map. The color bars on the right side of each map show the hit count scale.

Figure 4-13: Similar to Figure 4-12. A different color bar at the bottom shows the scale of hit count less than 100. Purple areas show where there are more than 100 hits.

Figure 4-14: Locations of the vertical cross sections shown in Figures 4-15 through 4-18.
Figure 4-15: Vertical cross sections of S-wave velocity when the Moho depth variations (Figure 4-6) are taken into account. Cross sections A, B and C at the longitudes of 80°, 90° and 100° are plotted. The surface topography along each profile is shown on the top of each cross section. The black curved lines show the Conrad (dashed) and Moho (solid) discontinuities. Each grid in the region between the two white lines has a raypath hit count of 200 or above.

Figure 4-16: Similar to Figure 4-15. Cross sections D, E, and F at the longitudes of 110°, 120° and 130°.

Figure 4-17: Similar to Figure 4-15. Cross sections G, H, I and J at the latitudes of 15°, 20°, 25° and 30°.

Figure 4-18: Similar to Figure 4-15. Cross sections K, L, M and N at the latitudes of 35°, 40°, 45° and 50°.

Figure 4-19: The same as Figure 4-15 but for the inversion result when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 4-20: The same as Figure 4-16 but for the inversion result when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 4-21: The same as Figure 4-17 but for the inversion result when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the location of the true Moho.

Figure 4-22: The same as Figure 4-18 but for the inversion result when a flat Moho is used. The Moho depth is taken to be 50 km. Black curved lines show the
location of the true Moho.

Figure 4-23: The input checkerboard when the grid shown in Figure 4-7 is adopted. The depth of the layer is shown at the lower-left corner of each map. Blue and red squares denote high and low velocities, respectively. The velocity perturbation scale is shown at the bottom.

Figure 4-24: Results of checkerboard resolution test of the input checkerboard shown in Figure 4-23.

Figure 4-25: The input checkerboard with the grid spacing of 2°.

Figure 4-26: Results of checkerboard resolution test of the input checkerboard shown in Figure 4-25.

Figure 4-27: The input checkerboard with the grid spacing of 4°.

Figure 4-28: Results of checkerboard resolution test of the input checkerboard shown in Figure 4-27.

Figure 4-29: The input checkerboard with the grid spacing of 8°.

Figure 4-30: Results of checkerboard resolution test of the input checkerboard shown in Figure 4-29.

Figure 4-31: Locations of the vertical cross sections shown in Figures 4-32 through 4-34.

Figure 4-32: Cross sections A, B, and C.

Figure 4-33: Cross sections D, E, and F.

Figure 4-34: Cross sections G, H, I, and J.

Figure 4-35: Comparison of Sn structure.
Figure 4-1: 25,000 earthquakes, 220 stations, active faults and major tectonic boundaries in China and the surrounding area. Earthquake epicenters are shown in red dots and stations are shown in red triangles. The yellow line shows the boundary of China. Active faults in the China area are shown in purple lines and tectonic sutures are shown in blue lines, where SoB – Songliao Basin, OB – Ordos Basin, SB – Sichuan Basin, KB – Khorat Basin, STB – Shan Thai Block, IB – Indochina Block.
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Chapter 5

Discussion

Our tomographic study of the crust and uppermost mantle in China and the surrounding area generated 3-D P- and S-wave velocity models for a large continental region. Validation and interpretation of these models are part of this thesis. P-wave travel-time validation was discussed in Chapter 3 for both GT events and earthquakes. Both travel-time residuals and hypocentral misfit of GT events are significantly reduced compared to any other models applied, by using our final 3-D P-wave model. To validate P- and S-velocity models, we compare recorded seismograms with synthetics generated through the new models. Then, we use Vp/Vs ratios, along selected profiles to characterize the heterogeneities in the crust and uppermost mantle.

5.1 Waveform Validation

A reliable validation can be performed by comparing recorded seismograms with the synthetic ones calculated from our tomographic model. Both P- and S-wave velocities can be tested by fitting body waves and surface waves. We have collected seismograms for six events (M5.0 to M6.0) recorded by the China Digital Broadband Network (CDBN) (Figure 5-1) between 2001 and 2003, in different areas of China (Figure 5-2). We select three events and four profiles in eastern and southern China for seismogram
fitting (Figure 5-3). Each event was recorded by up to 47 broadband stations in the CDBN. For each event, we select waveforms recorded by stations at different azimuths and at distances ranging from 50 km to 1000 km. Since the areas where the profiles were selected are laterally uniform we used the discrete wavenumber method (DWM) (Bouchon, 2003) to generate seismograms based on the 1-D averaged velocity profile between the source and the station. The mechanism parameters are provided by the Harvard CMT (Centroid-Moment-Tensor) Project.

Figures 5-4 through 5-7 show the radial, transverse and vertical components for both synthetic and observed seismograms. In eastern China, the agreement between observed and calculated seismograms is good for all profiles (A, B, C and D) for both body waves and surface waves. The first arrivals of P and S waves match very well the observed arrival times and the directions of the particle motion are consistent. This good agreement suggests that our P- and S-wave models are accurate in eastern and southern China. The best agreement has been observed in the transverse component for all four profiles. The amplitude and polarity of both Rayleigh and Love waves are correctly matched. Discrepancies between transverse and radial and vertical components are primarily due to the inaccurate source mechanism, the presence of anisotropy, and the reflections of scattered waves, converted waves and multiples.
5.2 Pn and Sn Comparison

Our P-wave tomography and S-wave tomography are independently obtained by fitting first arrivals of P- and S-wave, respectively. Only the same set of events and stations are the common parameters in the inversion. The final Pn structure and Sn structure show great similarity (Figure 5-8). Low Pn and Sn velocities in eastern China near Bohai gulf clearly indicate the effect of Cenozoic rifts through the region. High Pn and Sn velocities are observed in the Southern Tibetan plateau and beneath the Sichuan basin as well as the Ordos plateau.

5.3 1-D Profile Comparison

As mentioned in Chapters 3 and 4, the combined 1-D models (the pseudo 3-D models) are the input 3-D models for our final P- and S-tomographic study. The input 3-D models have demonstrated good vertical resolution as discussed in Chapter 2. The final 3-D models are properly obtained by full 3-D tomography. We expect that the final models are smoother than the input models and the Moho discontinuities stay unchanged.

Figure 5-9 shows twelve locations in different colors surrounding the center location (102°, 35°). These locations are one degree, three degrees or ten degrees from the center location. The input and final profiles at these selected locations are shown in Figure 10. The final profiles are indeed smoother than the input ones and the final Moho discontinuities clearly match the input discontinuities.
5.4 Vp/Vs Ratios and Poisson’s Ratios

We have shown the P- and S-velocities along different profiles in longitudinal and latitudinal directions in Chapters 3 and 4. In this section, we will show the Vp/Vs ratios and Poisson’s ratios along these profiles (Figure 5-11). Figures 5-12 through 5-15 show the profiles of the Vp/Vs ratios and Figures 5-16 through 5-19 show the profiles of Poisson’s ratios. Along the same profile the Vp/Vs ratios are consistent with the Poisson’s ratio. High Poisson’s ratios exist beneath the East and South China Sea, Japan Sea and Pacific Ocean. Along Profiles J and K high Poisson’s ratios in the crust beneath part of the Tibetan plateau suggest the presence of high temperature and/or fluid. This observation is consistent with results of previous studies in these areas based on seismic soundings and gravity surveys.

To show the relationship of volcanism and Poisson’s ratios, we plot the Vp/Vs ratios and Poisson’s ratios along local profiles in different regions shown in Figure 5-20. Profiles A, B, C, F, H, I, and J are cross sections through active and dormant volcanic sites. Prominent high Vp/Vs and Poisson’s ratios are observed beneath the active volcanic and geothermal sites in the Tengchong region (Profile C), in the Changbai region (Profile I), and in the Hainan region (Profile F). Our Vp/Vs and Poisson’s profiles clearly indicate the presence of high temperature beneath these well-known volcanic sites.

From this study, we conclude that the structure and P- and S-wave velocities we obtained for the crust and uppermost mantle are significant for better understanding of tectonic features in China and the surrounding area.
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Chapter 6

Conclusion

6.1 Summary

In this thesis, a large set of high-quality travel-time data given in the ABCE from 1990 to 2002 was used to construct the P- and S-wave models of the crust and uppermost mantle in China and surrounding areas. About 2400 1D models were created by using the adaptive moving window method (Sun et al., 2004). We combined all the 1D models and obtained 3D P- and S-wave velocity models. Using the quilted 3D velocity structures as the starting models and applying the updated tomographic method (Zhao et al., 1992), we inverted for a full 3D structure in the crust and uppermost mantle in China and surrounding areas. The obtained models provide accurate travel-times that relocate the GT events with small hypocentral error. The comparisons between the observed travel-times and calculated travel-times for both the GT events and earthquake profiles show that our models are reliable to generate 3D travel-times for the epicentral distances smaller than 20°. The synthetic seismograms generated by our models match well the observed seismograms recorded by the broadband stations. Our tomographic results also show the relationship between large earthquakes and velocity anomalies, the effect of fluid in seismotectonics and the volcanic implications. The major findings of this research are:
(1) The seismic velocity images are characterized by block structures corresponding to geological features bounded by large fault zones. The study region consists of a few geological structures: the North China Block including Songliao Basin, the South China Block, the Sichuan Basin, the Tarim Basin, and the Tianshan area. Those areas exhibit different patterns of velocity distribution in the tomographic images. The trend of velocity anomalies is very consistent with the trend of regional tectonics.

(2) The velocity images of the shallow crust (at the depth of 10 km) reflect well the surface geology and topographic features. Basins and depressions are represented by low-velocity anomalies, while high-velocity zones are imaged beneath the uplift and mountainous areas.

(3) A clear dividing line along the 105° parallel separates China into a low-V zone in the west and a high-V zone in the east at the depth of 40 km.

(4) Our tomographic imaging has revealed significant velocity heterogeneities in the middle and lower crust, some of which are very consistent with those detected by deep seismic soundings and other geophysical investigations.

(5) Large crustal earthquakes such as the 1679 Sanhe earthquake (M 8.0), the 1976 Tangshan earthquake (M 7.8), and the 2001 Ruoqiang earthquake (M 8.1), are generally located in high-velocity or relatively high-velocity areas in the upper to middle crust. In the lower crust and the uppermost mantle under the source zones of the large earthquakes, however, there are low-velocity and high-conductivity anomalies. Those anomalies are considered to be associated with fluids, similar to the 1995 Kobe earthquake (M 7.2) in Japan and the 2001 Bhuj earthquake (M 7.8) in India. The fluids in
the lower crust may cause the weakening of the seismogenic layer in the upper and middle crust and thus contribute to the rupture nucleation of the large crustal earthquakes.

(6) Under the Tengchong volcanic area, strong low velocity anomalies are visible down to 100 km depth and have a lateral extent of about 100 km. This feature is consistent with other geophysical observations in this area, such as high heat flow, negative gravity anomaly, and low electric resistivity, which suggest the existence of high temperature magma chambers under Tengchong.

(7) It is still unclear how the Tengchong intraplate volcanism was generated. There are three possibilities. The first is that it is related to the collision processes between the Indian plate, the Burma microplate and the Eurasian plate, and the possible subduction of the Burma microplate under the Eurasian plate. The second is that it was caused by the extensional fractures of the lithosphere and the upward intrusion of the hot asthenospheric materials. The third possibility is that the Tengchong intraplate volcanism represents a hot spot with a lower mantle origin. More detailed future studies with geological, geophysical and geochemical approaches are needed to clarify the formation mechanism and magma origin of this unusual volcanic center within the continental Eurasian plate.

### 6.2 Future Work

The Tibetan area carries relatively low resolution compared to other parts of the study area. With more networks installed in Tibet, more high-quality data will be available for resolution improvement.
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Appendix A

Geology and Tectonic Background

The geology of China consists of Precambrian platforms surrounded by accreted terrenes and fold belts of various ages. These tectonic elements were first assembled in the Paleozoic Era, but have been further deformed and rearranged in multiple episodes throughout the Mesozoic and Cenozoic. The following summary of the geologic and tectonic history of China is mostly after Li (1998) and Lebedev et al. (2003).

There are four Precambrian platforms in China and surrounding areas: the North China Block, the South China Block, the Tarim Basin and the Indochina Block. The North China Block, also known as the Sino-Korean Craton, consists of two major Archean (older than 2.5 Ga) continental nuclei surrounded by Paleoproterozoic (about 1.8 Ga) orogenic belts. One nucleus is approximately in the boundaries of the Ordos Basin (Ordos Plateau), and the other, larger one, is beneath and around the Bohai Gulf. Younger orogenic belts are located along the margins of the block. The South China Block includes two major Precambrian elements: the Yangtze Craton and the Cathaysia Block. The Archean nucleus of the Yangtze Craton is approximately within the boundaries of the Sichuan Basin. The Paleoproterozoic basement is found in the vicinity of the basin and to the southwest; the rest of the basement is probably of the Grenville age (about 1.0 Ga). The Cathaysia Block is situated along the coast (partly in the South China Foldbelt and partly underwater to the east) and is separated from the Yangtze
Craton by orogenic belts of different ages, from about 1000 to 150 Ma. The Tarim Basin consists of a nucleus of Archean through the Proterozoic age covered by thick Cenozoic sediments. The Indochina Block also has a Precambrian core and probably extends to the northwest as the Shan Thai (Simao) Block (Metcalfe, 1996).

The history of the North China and South China Blocks may be traced as far as 1.0 Ga, when both were a part of the supercontinent Rodinia; Cathaysia first joined the Yangtze block during the assembly of Rodinia. Rifting with Rodinia started soon after 900 Ma; a rift formed but failed between Yangtze and Cathaysia. The rift was mostly closed during the “Caledonian orogenic episode”, by 400 Ma. The closure completed later, in the Mesozoic. Active margins existed in the Paleozoic along both the south and north boundaries of the Sino-Korean Craton.

The North China Block joined with the Mongolian terranes, presently situated to the north of the block, in the Permian (at 280-250 Ma), and around the same time continental collisions began between the core of Eurasia, North China, and South China Blocks, the latter already joined with Indochina. The initial contact between the North and South China Blocks occurred at their eastern ends, probably resulting in the subduction of some of the Yangtze lower continental crust and lithosphere; the main phase of suturing followed at 240-170 Ma. Foldbelts now found to the northwest of the Yangtze Craton were formed in Paleozoic-early Mesozoic. The suturing between North China – Mongolia and Siberia was complete around 150 Ma.

As the relative motion between the continental blocks slowed down and finally stopped, a convergent margin developed along the southeast Asia coast. It caused
widespread magmatism at 160-100 Ma and the development of a “basin and range province” along the coast (in the South China Foldbelt).

The India-Eurasia collision started at about 55 Ma and controlled the Cenozoic tectonic history of SE Asia. The collision caused the extrusion of the Indochina block southeastward between 50 and 23 Ma and the opening of the South China Sea to accommodate it (Briais et al., 1993). Around 15 Ma the compressional regime and eastward motion of Indochina changed to extension and westward motion retreat. According to the escape-tectonics scenario (Molnar and Tapponnier, 1975), the subsequent eastward displacement of southern China and widespread rifting in the Sino-Korean Craton represents the next episode of the extrusion. Major North China’s rift systems are located in the area of the Bohai’s rift systems are located in the area of the Bohai Gulf on the east and along the perimeter of the Ordos Basin on the west, with more rifts in between.

Figure 1-1 shows the active faults and tectonic sutures in China and surrounding areas.
References


Appendix B

Theory and Methodology of Seismic Tomography

All the studies of body wave seismic tomography include the following operations (Mishra, 2004): (1) modeling the earth velocity structure; (2) calculating ray paths and travel times; (3) solving the large linear system of observation equations and (4) evaluating the resolution of the obtained tomographic image. In this appendix, the approach utilized in the tomographic method of Zhao (1991) is described. The following summary of the theory and methodology of seismic tomography is mostly after Mishra (2004), Paige and Saunders (1982a, b), Zhao (1991), and Zhao et al. (1992).

B.1 Model parameterization

The problem of imaging the earth’s structure has been investigated by several researchers using different approaches to model parameterization: 1) velocity model with constant velocity layers (Crosson, 1976a, b); 2) large number of constant velocity blocks in three-dimension (Aki and Lee, 1976); 3) velocity model with a number of “ideal averaging volumes” (Chou and Booker, 1979); 4) three-dimensional analytic function defined by a small number of parameters for a velocity model (Spencer and Gubbins, 1980); 5) constant-velocity layers with layer boundaries expressed by power series of spatial
positions (Horiuchi et al., 1982a, b); and 6) large number of grid nodes in three-dimension (Thurber, 1983). Each approach has the pros and cons.

In this study, we adopted the model parameterization of Zhao (1991) and Zhao et al. (1992), which consists of a velocity structure having a number of complicated-shaped velocity discontinuities; the velocity at any location between these discontinuities changes everywhere three dimensionally (Figure B-1). This model can accurately represent the earth’s velocity structure. The depth distribution of velocity discontinuities has been expressed in two ways. One is a continuous function such as power series, which is used in the studies of Horiuchi et al. (1982a, b). The depth to the i-th discontinuity is expressed as

\[ H_i(\phi, \lambda) = \sum_k C_{ik} h_{ik}(\phi, \lambda) = C_{i1} + C_{i2}\phi + C_{i3}\lambda + C_{i4}\phi^2 + \ldots \] (B-1)

where \( \phi \) and \( \lambda \) are latitude and longitude, respectively. \( C_{ik} \) (I = 1, 2, ..., m, k = 1, 2, ..., n) are coefficients, m is the number of discontinuities, n is the number of the coefficients of the power series.

Another way to express the depth distribution of a discontinuity is to use a two dimensional grid. When the depth of a discontinuity at grid points is determined, the depth of that discontinuity at any position within the study area can be determined by using a simple interpolation function

\[ H_i(\phi, \lambda) = \sum_{j=1}^2 \sum_{k=1}^2 h(\phi_j, \lambda_k) \left[ \left( 1 - \frac{\phi - \phi_1}{\phi_2 - \phi_1} \right) \left( 1 - \frac{\lambda - \lambda_2}{\lambda_1 - \lambda_2} \right) \right] \] (B-2)

where \( \phi_j \) and \( \lambda_k \) represent the coordinates of the four grid points surrounding the point \((\phi, \lambda)\). This is a continuous function, a product of linear functions in \( \phi \) and \( \lambda \).
Three-dimensional grids have been set up independently for every layer to express the three-dimensional velocity structure for layers, which are bounded by two adjacent discontinuities (Figure B-2). For each grid within one layer, sheets of grid points are densely set within the layer, except the outermost sheets of grid points which are set up far away from the internal sheets of grid so that all the seismic rays in that layer are completely included in the grid net. Velocities at interior grid points are unknown parameters. Velocities at the outermost sheets of grids are used only to interpolate velocities outside of the modeling space. The velocity at any position in m-th layer is calculated by using an interpolation function

\[ V_{m}(\phi, \lambda, h) = \sum_{i=1}^{2} \sum_{j=1}^{2} \sum_{k=1}^{2} V_{m}(\phi_i, \lambda_j, h_k) \left[ \left( 1 - \frac{\phi - \phi_i}{\phi_2 - \phi_1} \right) \left( 1 - \frac{\lambda - \lambda_j}{\lambda_2 - \lambda_1} \right) \left( 1 - \frac{h - h_k}{h_2 - h_1} \right) \right] \] (B-3)

where \( h \) is the depth from the earth’s surface; \( \phi_i, \lambda_j, \) and \( h_k \) represent the coordinates of the eight grid points (\( \phi, \lambda, h \)). \( V_{m}(\phi, \lambda, h) \) is the velocity at the grid point of the grid net set up for the m-th layer.

The model parameterization of Zhao (1991) considers velocities at grid points in every layer as unknowns. The depth distribution of the velocity discontinuities is determined from priori information, and the discontinuities are fixed in the inversion process.

### B.2 Fast Ray Tracing

Computation of travel times and ray paths in a heterogeneous earth model requires very fast and accurate ray tracing for tomographic inversion. Several algorithms for three-dimensional ray tracing have been developed by seismologists and mathematicians...
(Jacob, 1970; Wesson, 1971; Julian and Gubbins, 1977; and Pereyra et al., 1980). However, these algorithms are found to be very time consuming for use in tomographic problems. Thurber and Ellsworth (1980) and Horie (1980) separately introduced an inexpensive scheme for computing approximate travel times by determining a precise ray path in a one-dimensional, local approximation to the three-dimensional velocity structure. Thurber (1983) developed another approximate ray tracing (ART) technique, which adopts circular arcs connecting the source and receiver to represent the ray path. The two ART techniques work well when the epicenter distance is shorter than about 50 km. When the epicenter distance becomes larger, travel times and ray paths computed by the two ART techniques considerably deviate from the exact ones. This limitation makes the two ART techniques inapplicable for tomographic studies of large aperture arrays.

Um and Thurber (1987) proposed an approximate algorithm for fast 3-D ray tracing in a velocity model where the velocity is continuous everywhere. Zhao (1991) and Zhao et al. (1992) extended this technique to models containing velocity discontinuities. This technique is described briefly as follows:

The ray equation can be written as

$$ -\frac{d^2\vec{r}}{ds^2} = \left( (\text{grad}V) - \frac{dV}{ds} \frac{d\vec{r}}{ds} \right) \frac{1}{V} $$

where $\vec{r}$ is the position vector along the ray, $s$ is the length of the ray path. The first term of the right side of equation (B-4) is the velocity gradient, and the second term is a component of the velocity gradient parallel to the ray vector (Figure B-3). Equation (B-4) states that the component of the velocity gradient normal to the ray vector is always anti-parallel to the ray path curvature. Such geometric interpretation of the ray equation was first given by Cerveny et al. (1977).
Consider three adjacent points along the ray path as shown in Figure B-4. The two points $X_{k-1}$ and $X_{k+1}$ are temporarily fixed. A new point $X'_k$ instead of the previous point $X_k$ is sought to minimize the travel time along the segment of the ray path from $X_{k-1}$ to $X_{k+1}$. In order to define the new point $X'_k$, two variables, the direction $\vec{n}$ and $R$, the offset from the mid-point $X_{\text{mid}}$, are calculated. The local ray direction at $X'_k$ is approximately given by the direction of the line connecting the two end points of the ray segment $X_{k-1}$ and $X_{k+1}$. According to equation (B-4), the component of the velocity gradient normal to that direction defines the curvature direction. Hence, this direction gives the correct offset direction for the point $X'_k$ (Zhao, 1991). The equation is given by

$$\vec{n} = (\text{grad}V) - \left[ (\text{grad}V) \cdot (\overrightarrow{X_{k+1}} - \overrightarrow{X_{k-1}}) \right] \cdot \frac{\overrightarrow{X_{k+1}} - \overrightarrow{X_{k-1}}}{||\overrightarrow{X_{k+1}} - \overrightarrow{X_{k-1}}||^2} \quad (B-5)$$

Using the Taylor expansion of the velocity at the mid-point, $V_{\text{mid}}$, the velocity $V'_k$ at the new point $X'_k$ is approximated by

$$V'_k = V_{\text{mid}} + [\vec{n} \cdot (\text{grad}V)_{\text{mid}}]R \quad (B-6)$$

where $(\text{grad}V)_{\text{mid}}$ is the velocity gradient at the mid-point $X_{\text{mid}}$. Then the amount of perturbation $R$ along the direction $\vec{n}$ is obtained by minimizing the travel time along the ray segment connecting the three points, $X_{k-1}$, $X'_k$ and $X_{k+1}$, yielding

$$R = \frac{C V_{\text{mid}} + 1}{4 \vec{n} \cdot (\text{grad}V)_{\text{mid}}} + \left[ \frac{(C V_{\text{mid}} + 1)^2}{4 \vec{n} \cdot (\text{grad}V)_{\text{mid}}^2} \right] + \frac{L^2}{2 C V_{\text{mid}}} \right] \quad (B-7)$$

where $L = [X_{k+1} - X_{k-1}]$ and $C = \left( \frac{1}{V_{k+1}} + \frac{1}{V_{k-1}} \right) \times \frac{1}{2}$. The derivation of equation (B-7) has been described in detail by Um and Thurber (1987).
This three-point perturbation scheme has successively been extended to all the points along the ray path. The perturbation is iteratively performed until the travel time converges to a specified limit. Um and Thurber (1987) and Zhao (1991) carried out many tests and showed that this algorithm successfully found an accurate travel time much faster than the exact 3-D ray tracing programs. This algorithm also finds an accurate ray path in a fully 3-D form even when velocity variations are severe and hypocenter distance is large (Zhao, 1991).

The algorithm of Um and Thurber (1987) is called “pseudo-bending technique”, because it takes the form of two-point bending approach but does not solve exactly the ray equation (Zhao, 1991). The ordinary exact ray tracing methods (Wesson, 1987; Pereyra et al., 1980) solve the equation exactly, which requires huge computation time. A disadvantage of the pseudo-bending method of Um and Thurber (1987) is that it fails to work when a velocity discontinuity exists, and therefore it can be applicable only to a continuous velocity model. This largely limits the application of the pseudo-bending method. Zhao (1991) solved this problem by adapting the method for a velocity model that contains complex velocity discontinuities and the velocity on both sides of a discontinuity changes three-dimensionally (Figure B-5).

MM’ in Figure B-5 represents a velocity discontinuity. Velocities, \( V_1(\phi, \lambda, h) \) and \( V_2(\phi, \lambda, h) \) on both sides of MM’ are taken as continuous. At two points A and B on both sides of MM’ the velocities are \( V_a \) and \( V_b \), respectively. C is the intersection of the seismic ray AB and the discontinuity MM’.
The velocities at C on both sides of MM' are $V_{c1}$ and $V_{c2}$, given by the input 1-D or 3-D reference model, respectively. The arithmetic average $V_1$ of $V_a$ and $V_{c1}$, and arithmetic average $V_2$ of $V_b$ and $V_{c2}$ can be expressed as,

$$V_1 = \frac{V_a + V_{c1}}{2}, \quad V_2 = \frac{V_b + V_{c2}}{2}$$

(B-8)

where $V_1$ and $V_2$ represent the mean velocity around AC and BC respectively. This approximation may be poor when the points A and B are far away from the discontinuity MM'. However, the approximation is considered to be appropriate if the points A and B are close to MM'. Then this approach iteratively uses the dichotomy to find the accurate location of point C that satisfies Snell’s law (Zhao, 1991).

$$\frac{\sin \theta_1}{V_1} = \frac{\sin \theta_2}{V_2}$$

(B-9)

The new ray tracing algorithm proposed by Zhao (1991) is schematically shown in Figure B-6. A simple model with two discontinuities has been considered. The triangle denotes a station and the star represents a hypocenter. The straight line $A_1-A_5$ connecting the station and the hypocenter is assumed to be the initial ray path. The intersection points such as $A_2$ and $A_3$ of the ray path with the model discontinuities are labeled “discontinuous points (DPs)”, and the points along the ray path between the two adjacent discontinuities are “continuous points (CPs)”. Snell’s law is used to perturb the DPs, while the pseudo-bending technique is adopted to perturb the CPs in the new algorithm of Zhao (1991). Snell’s law is used to find new DPs $A_2'$ (from $A_1$ and $A_3$) and $A_3'$ (from $A_2$ and $A_4$), and the pseudo-bending is used to find the new CPs $B_1$ (from $A_1$ and $A_2'$), $B_2$ (from $A_2'$ and $A_3'$), and $B_3$ (from $A_3'$ and $A_4'$) (Figure B-6). Using the Snell’s law again, we find $A_2''$ from $B_1$ and $B_2$, and $A_3''$ from $B_2$ and $B_3$. After a number of iterations the ray path is considered to converge gradually to its true location.
B.3 Inversion

B.3.1 Preamble

Starting with the initial hypocenter locations and origin times for a set of earthquakes, we estimate correction terms for the source and the medium parameters, in such a way that all the arrival time data can be best explained in the least squares sense. As mentioned in Chapters 3 and 4, Sun (2001) and Sun et al. (2004) have shown that the source locations from China are reliable with small uncertainties. In this study, the source parameters are taken from the ABCE (1990-2002) and only the medium parameters are variables to be obtained from inversion. In general, the equation for an observed arrival time can be written as:

\[
T_{ij}^{OBS} = T_{ij}^{CAL} + \left( \frac{\partial T}{\partial \phi} \right)_{ij} \Delta \phi_i + \left( \frac{\partial T}{\partial \lambda} \right)_{ij} \Delta \lambda_i + \left( \frac{\partial T}{\partial h} \right)_{ij} \Delta h_i + \Delta T_{ai} + \sum_{n=1}^{N} \frac{\partial T}{\partial V_n} \Delta V_n + e_{ij} \quad (B-10)
\]

where

\( T_{ij}^{OBS} \): observed arrival time for i-th earthquake at j-th station;

\( T_{ij}^{CAL} \): calculated arrival time based on the initial model;

\( \phi_i, \lambda_i, h_i, T_{ai} \): latitude, longitude, focal depth and origin time of i-th earthquake;

\( \Delta \phi_i, \Delta \lambda_i, \Delta h_i, \Delta T_{ai} \): perturbation of the above parameters;

\( \left( \frac{\partial T}{\partial \phi} \right)_{ij}, \left( \frac{\partial T}{\partial \lambda} \right)_{ij}, \left( \frac{\partial T}{\partial h} \right)_{ij} \): partial derivatives of the arrival time with respect to hypocenter location;

\( V_n, \Delta V_n \): velocity at the n-th grid point and its perturbation;
\[
\left( \frac{\partial T}{\partial V_n} \right): \text{partial derivative of the arrival time with respect to velocity parameter;}
\]

\[e_e: \text{higher order terms of perturbations and errors in the observation.}\]

The perturbations of the source parameters are zero in this study.

Travel time data required for equation \((B-10)\) are calculated using the fast ray tracing method described in the previous section. Partial derivatives in equation \((B-10)\) can be calculated if we know the velocity model and the ray path from the earthquake to the station. In the case when the source parameters are unknown, partial derivatives with respect to hypocenter location can be written as

\[
\begin{align*}
\frac{\partial T}{\partial \phi} &= -(R_0 - h) \sin i \cos \alpha / V_e \\
\frac{\partial T}{\partial \lambda} &= -(R_0 - h) \sin i \cos \phi \sin \alpha / V_e \\
\frac{\partial T}{\partial h} &= \cos i / V_e 
\end{align*}
\]

where \(R_0\) is the radius of the earth, \(i\) is the takes-off angel at the hypocenter, \(\alpha\) is the azimuth from the epicenter to the station, and \(V_e\) is the velocity at the hypocenter (see Engdahl and Lee, 1976).

Partial derivatives with respect to velocity parameters can not be calculated analytically because they involve integrals along the ray path as shown below (Zhao, 1991):

\[
\frac{\partial T}{\partial V_n} = \int_{\text{source}}^{\text{station}} \left[ \frac{1}{V(\phi, \lambda, h)} \right]^2 \frac{\partial V(\phi, \lambda, h)}{\partial V_n} ds \quad (B-12)
\]

In fact partial derivatives with respect to slowness are utilized for computational simplicity (Thurber, 1983). Instead of equation \((B-12)\), we have
\[
\frac{\partial T}{\partial V_n} = \int_{\text{source}}^{\text{station}} \frac{\partial U(\phi, \lambda, h)}{\partial U_n} ds
\]  
(B-13)

where \( U_n = 1/V_n \) and \( U(\phi, \lambda, h) = 1/V(\phi, \lambda, h) \). This ray path integral is approximated by dividing the ray path into \( M \) segments

\[
\frac{\partial T}{\partial U_n} = \sum_{m=1}^{M} \frac{\partial U(\phi_m, \lambda_m, h_m)}{\partial U_n} \Delta s_m
\]  
(B-14)

where \( \Delta s_m \) is the step length and \( (\phi_m, \lambda_m, h_m) \) is the midpoint of the \( m \)-th path segment.

The partial derivative \( \frac{\partial U}{\partial U_n} \) is calculated by using the interpolation function in equation (B-3).

Travel time residual can be written as:

\[
t_{ij} = T_{ij}^{\text{OBS}} - T_{ij}^{\text{CAL}}
\]  
(B-15)

All the residuals form a whole set of data with a column vector \( d \) of dimension \( N \). In the case when both source locations and velocities are unknown, the source and medium parameter corrections, a column vector \( \Delta m \), can be arranged as

\[
\Delta m^T = (\Delta \phi_1, \Delta \lambda_1, \Delta h_1, \Delta T_{01}, \ldots, \Delta \phi_M, \Delta \lambda_M, \Delta h_M, \Delta T_{0M}, \Delta V_1, \Delta V_2, \ldots, \Delta V_K)
\]

where \( M, N, K \) are the numbers of earthquakes, data and grid points, respectively. The data vector and the unknown parameter vector are related to each other and can be expressed by the following observation equation:

\[
d = G\Delta m + e
\]  
(B-16)

where \( e \) is an error vector, and \( G \) is a matrix with dimension \( N \) by \( (4M + K) \), whose elements consist of the derivatives in equations (B-11) and (B-14).
There are several approaches to solve equation (B-16). Zhao (1991) uses the damped least squares method and a conjugate gradient type algorithm LSQR as the main solver.

**B.3.2 Damped Least Squares Method**

When many earthquakes are used, the matrix $G$ in equation (B-16) becomes very large and it becomes difficult to handle. Zhao (1991) adopts the same approach as those of Pavlis and Booker (1980) and Thurber (1983). Each earthquake yields a set of $L$ observation equation like equation (B-10), which can be written in matrix notation as

$$d_i = H_i \Delta h_i + M_i \Delta m$$

(B-17)

$Kx1  \quad Lx4  \quad 4x1  \quad LxK  \quad Kx1$

where $d_i$ is a vector containing $L$ residuals, $h_i$ is the four hypocenter parameter corrections for the $i$-th earthquake. $H_i$ and $M_i$ are the matrices of hypocenter and velocity partial derivatives for the $i$-th earthquake. Adopting the parameter separation technique described by Pavlis and Booker (1980), a matrix $Q$ is constructed which has the property

$$Q \quad H_i = 0$$

(B-18)

$(L-4)xL  \quad Lx4$

Operating on equation (B-17) by $Q$ results in

$$d'_i = M'_i \Delta m$$

(B-19)

$(L-4)x1  \quad (L-4)xK  \quad Kx1$
Consider a matrix $M'$ composed of a set of submatrices $M'_i$ and the vector $d'$ composed of a set of subvectors $d'_i$. As the number of earthquakes increases, $M'$ and $d'$ abruptly grow in size, leading to a severe difficulty of computer storage. To overcome this problem, Zhao (1991) accumulates the matrices, $M'^T M'$ and $M'^T d'$ sequentially as each earthquake gets processed. This approach produces a symmetric matrix and a vector of fixed size:

$$M'^T M' = \sum_i M'_i M'_i$$
$$M'^T d' = \sum_i M'_i d'_i$$  \hspace{1cm} (B-20)

Then the damped least squares solution (Crosson, 1976a and Aki and Lee, 1976) satisfies the normal equation

$$(M'^T M' + \rho I)\Delta m = M'^T d'$$ \hspace{1cm} (B-21)

where $I$ denotes the unit (Identity) matrix, and $\rho$ is a constant called damping parameter.

The resolution matrix defined by Backus and Gilbert (1968) and Wiggins (1972) is given by

$$R = (M'^T M' + \rho I)^{-1} M'^T M'$$ \hspace{1cm} (B-22)

The covariance matrix of $\Delta m$ can be written as

$$D = \sigma^2 \left(M'^T M' + \rho I \right)^{-1} R$$ \hspace{1cm} (B-23)

where $\sigma^2 = \frac{|d' - M' \Delta m|}{N - K - 4 M}$.

The damped least squares method is discussed in detail by Aki and Lee (1976) and Aki et al. (1977). An advantage of this matrix type approach is that resolution and
covariance matrices can be computed simultaneously with the solution. The drawback is that when the number of unknown parameters increases, the matrix $M' M'$ will need much more computer memory storage. Moreover, the construction of $M' M'$ needs so many operations that even the super-computers cannot handle it. In that situation iterative inversion methods (e.g., Algebraic Reconstruction Technique (ART) (Herman, 1980), Simultaneous Iterative Reconstruction Technique (SIRT) (Humphreys and Clayton, 1988) or projection type methods (e.g., Conjugate Gradient [CG] techniques) should be used to solve the large and sparse linear system of equation (B-16). Zhao (1991) uses the CG type solver LSQR algorithm (Paige and Saunders, 1982a).

**B.3.3 LSQR Algorithm**

This algorithm was introduced by Nolet (1985) for the first time to seismology to solve the tomographic problem. Latter, it was used by Spakman and Nolet (1988), and Lees and Crosson (1989). Comparing the operational behaviors of SIRT and LSQR, Nolet (1985), van der Sluis and van der Vorst (1987) and Spakman and Nolet (1988) concluded that LSQR is superior to SIRT. Lees and Crosson (1989) showed that LSQR and SIRT gave almost the same result. Hirahara (1990) pointed that in realistic analyses these algorithms may give substantially the same results, and the CG-type solver may be slightly superior in the aspect of preferable operation behavior and fast convergence.

The LSQR algorithm is similar to the well-known method of conjugate gradients as applied to the least-squares problem (Paige and Saunders, 1982a). CG-like methods are iterative in nature. To solve $A x = b$, where $A$ is a real matrix with $m$ rows and $n$ columns ($m \geq n$) and $b$ is a real vector, the CG-like methods are characterized by their
need for only a few vectors of working storage and by their theoretical convergence with
at most n iterations. In practice such methods may require far fewer than n iterations to
reach an acceptable approximation to x. It is often possible to divide the solution
procedure into a direct and an iterative part, such that the iterative part has a better
conditioned matrix for which CG-like methods will converge more quickly.

Algorithm LSQR is based on the bidiagonalization procedure of Golub and Kahan
(1965). It generates a sequence of approximations \( \{x_k\} \) such that the residual norm \( \|r_k\|_2 \)
decreases monotonically, where \( r_k = b - Ax_k \). Analytically, the sequence \( \{x_k\} \) is identical
to the sequence generated by the standard CG algorithm and by several other published
algorithms. However, LSQR is shown by example to be numerically more reliable in
various circumstances than the other methods considered (Paige and Saunders, 1982a).

To derive the procedures of LSQR, we need to introduce the Lanczos process.

B.3.3.1 The Lanczos Process

In this section we will review the symmetric Lanczos process (Lanczos, 1950) and its use
in solving symmetric linear equations \( Bx = b \). Algorithm is then derived by applying the
Lanczos process to a particular symmetric system.

Given a symmetric matrix \( B \) and a starting vector \( b \), the Lanczos process is a
method for generating a sequence of vector \( \{v_i\} \) and scalars \( \{\alpha_i\} \) and \( \{\beta_i\} \) such that \( B \)
is reduced to tridiagonal form. A reliable computational form of the method is the
following.

*The Lanczos process* (reduction to tridiagonal form):

\[
\beta_1 v_1 = b,
\]
\[ w_i = Bv_i - \beta_i v_{i-1} \]
\[ \alpha_i = v_i^T w_i \]
\[ \beta_{i+1} v_{i+1} = w_i - \alpha_i v_i \]
\[ w_i = Bv_i - \beta_i v_{i-1} \]
\[ \alpha_i = v_i^T w_i \]
\[ \beta_{i+1} v_{i+1} = w_i - \alpha_i v_i \]
\[ \begin{aligned}
Bv_k &= V_k T_k + \beta_{k+1} v_{k+1} e_k^T \\
T_k &= \text{tridiag}(\beta_i, \alpha_i, \beta_{i+1})
\end{aligned} \tag{B-25}

where \( v_0 = 0 \) and each \( \beta_i \geq 0 \) is chosen so that \( \|v_i\| = 1 \) \( (i > 0) \). The situation after \( k \) steps is summarized by

\[ Bv_k = V_k T_k + \beta_{k+1} v_{k+1} e_k^T \tag{B-25} \]

where \( T_k = \text{tridiag}(\beta_i, \alpha_i, \beta_{i+1}) \) and \( V_k = [v_1, v_2, \ldots v_k] \). If there were no rounding error we would have \( V_k^T V_k = I \), and the process would therefore terminate with \( \beta_{k+1} = 0 \) for some \( k \leq n \). Some other stopping criterion is needed in practice, since \( \beta_{k+1} \) is unlikely to be negligible for any \( k \). In any event, equation (B-25) holds to within machine precision.

Now suppose we wish to solve the symmetric system \( Bx = b \). Multiplying equation (B-25) by an arbitrary \( k \)-vector \( y_k \), whose last element is \( \eta_k \), gives

\[ BV_k y_k = V_k T_k y_k + \beta_{k+1} v_{k+1} \eta_k \]

Since \( V_k (\beta_i e_i) = b \) by definition, it follows that if \( y_k \) and \( x_k \) are defined by the equations

\[ T_k y_k = \beta_i e_i \]
\[ x_k = V_k y_k \]
\[ \begin{aligned}
T_k y_k &= \beta_i e_i \\
x_k &= V_k y_k \end{aligned} \tag{B-26} \]

then we shall have \( Bx_k = b + \eta_k \beta_{k+1} v_{k+1} \) to working accuracy. Here \( x_k \) may be taken as the exact solution to a perturbed system and will solve the original system whenever \( \eta_k \beta_{k+1} \) is negligibly small. The LSQR algorithm obtains \( x_k \) after evaluating \( y_k \) by QR factorization that will be described in later section.

B.3.3.2 The LSQR Procedures
For the least squares system

\[ Ax = b \]  \hspace{1cm} (B-27)

where \( A( m \times n) \) and \( b \) are given, the solution satisfies the symmetric system

\[
\begin{bmatrix}
I & A \\
A^T & 0
\end{bmatrix}
\begin{bmatrix}
r \\
x
\end{bmatrix}
=
\begin{bmatrix}
b \\
0
\end{bmatrix},
\]  \hspace{1cm} (B-28)

where \( r \) is the residual vector \( b - Ax \). When the Lanczos process is applied to this system, we can obtain relations similar to (B-24)-(B-26).

Based on the Lanczos process stated above, Golub and Kahan (1965) proposed the following bidiagonization procedures:

\[
\begin{align*}
\beta_i u_i &= b, & \alpha_i v_i &= A^T u_i, \\
\beta_{i+1} u_{i+1} &= A v_i - \alpha_i u_i, \\
\alpha_{i+1} v_{i+1} &= A^T u_{i+1} - \beta_{i+1} v_i,
\end{align*}
\]  \hspace{1cm} (B-29)

The scalars \( \alpha_i \geq 0 \) and \( \beta_i \geq 0 \) are chosen so that \( \| u_i \| = \| v_i \| = 1 \). With the definitions

\[
U_k = [u_1, u_2, \ldots, u_k], \quad V_k = [v_1, v_2, \ldots, v_k],
\]

\[
B_k = \begin{bmatrix}
\alpha_1 & & \\
\beta_1 & \alpha_2 & \\
& \beta_2 & \ddots & \ddots \\
& & \ddots & \ddots & \alpha_k \\
& & & \beta_{k+1}
\end{bmatrix},
\]

the recurrence relation (B-29) can be rewritten as

\[
U_{k+1} (\beta_{k+1} e_1) = b, \hspace{1cm} (B-30)
\]

\[
AV_k = U_{k+1} B_k, \hspace{1cm} (B-31)
\]

\[
A^T U_{k+1} = V_k B_k^T + \alpha_{k+1} v_{k+1} e_{k+1}^T. \hspace{1cm} (B-32)
\]

If exact arithmetic were used, then we would also have \( U_{k+1}^T U_{k+1} = I \) and \( V_k^T V_k = I \), but, in any event, the previous equations hold to within machine precision.

Similar to equation (B-26), we introduce \( y_k \) and define the following quantities
\[ x_k = V_k y_k, \quad \text{(B-33)} \]
\[ r_k = b - A x_k, \quad \text{(B-34)} \]
\[ t_{k+1} = \beta_k e_1 - B_k y_k. \quad \text{(B-35)} \]

The above relations can also be written into the following expressions
\[
\begin{bmatrix}
I & B_k \\
B_k^T & 0
\end{bmatrix}
\begin{bmatrix}
t_{k+1} \\
y_k
\end{bmatrix}
= \begin{bmatrix}
\beta_k e_1 \\
0
\end{bmatrix}, \quad \text{(B-36)}
\]
\[
\begin{bmatrix}
r_k \\
x_k
\end{bmatrix}
= \begin{bmatrix}
U_{k+1} & 0 \\
0 & V_k
\end{bmatrix}
\begin{bmatrix}
t_{k+1} \\
y_k
\end{bmatrix}. \quad \text{(B-37)}
\]

It readily follows from (B-30) and (B-31) that the equation
\[ r_k = U_{k+1} t_{k+1} \quad \text{(B-38)} \]
holds to working accuracy. Since we want \( \|r_k\| \) to be small, and since \( U_{k+1} \) is bounded and theoretically orthonormal, this immediately suggests choosing \( y_k \) to minimize \( \|r_{k+1}\| \).

Hence we are led naturally to the least-squares problem
\[
\min \|\beta_k e_1 - B_k y_k\| \quad \text{(B-39)}
\]
which forms the basis for LSQR.

Computationally, it is advantageous to solve (B-39) using the standard QR factorization (Golub, 1965) of \( B_k \). This takes the form
\[
Q_k [B_k \quad \beta_k e_1] = 
\begin{bmatrix}
R_k & f_k \\
\phi_{k+1}
\end{bmatrix}
= 
\begin{bmatrix}
\rho_1 & \theta_2 & & & \\
\rho_1 & \theta_2 & & & \\
& \ddots & \ddots & & \\
& & \ddots & \ddots & \\
& & & \rho_{k-1} & \theta_k \\
& & & \rho_k & \phi_{k+1}
\end{bmatrix}
\quad \text{(B-40)}
\]
where $Q_k \equiv Q_{k,k-1} \cdots Q_{2,1} Q_{1,2}$ is a product of plane rotations designed to eliminate the subdiagonals $\beta_2, \beta_3, \ldots$ of $B_k$. The vectors $y_k$ and $t_{k+1}$ could then be found from

$$R_k y_k = f_k,$$

$$t_{k+1} = Q_k^T \begin{bmatrix} 0 \\ \phi_k \end{bmatrix}.$$

However, $y_k$ in (B-41) will normally have no elements in common with $y_{k-1}$. Instead we note that $[R_k \ f_k]$ is the same as $[R_{k-1} \ f_{k-1}]$ with a new row and column added. Hence, one way of combining (B-33) and (B-41) efficiently is according to

$$x_k = V_k R_k^{-1} f_k \equiv D_k f_k,$$

where the columns of $D_k \equiv [d_1 \ d_2 \ \cdots \ d_k]$ can be found successively from the system $R_k^T D_k^T = V_k^T$ by forward substitution. With $d_0 = x_0 = 0$, this gives

$$d_k = \frac{1}{\rho_k} (v_k - \theta_k d_{k-1}),$$

$$x_k = x_{k-1} + \phi_k d_k,$$

and only the most recent iterates need be saved. The broad outline of algorithm LSQR is now complete.

The QR factorization (B-40) is determined by constructing the $k$th plane rotation $Q_{k,k+1}$ to operate on rows $k$ and $k+1$ of the transformed $[B_k \ \beta_k e_1]$ to annihilate $\beta_{k+1}$.

This gives the following simple recurrence relation:

$$\begin{bmatrix} c_k & s_k \\ -s_k & c_k \end{bmatrix} \begin{bmatrix} \bar{\rho}_k & 0 \\ \bar{\phi}_k & \bar{\phi}_k \end{bmatrix} = \begin{bmatrix} \rho_k & \theta_{k+1} \\ 0 & \bar{\rho}_{k+1} \end{bmatrix} \begin{bmatrix} \phi_k \\ \bar{\phi}_{k+1} \end{bmatrix},$$

(B-46)
where \( \overline{\rho}_i \equiv \alpha_i \), \( \overline{\phi}_i \equiv \beta_i \), and the scalars \( c_k \) and \( s_k \) are the nontrivial elements of \( Q_{k,k+1} \) with \( c_k^2 + s_k^2 = 1 \). The quantities \( \overline{\rho}_k \) and \( \overline{\phi}_k \) are intermediate scalars that are subsequently replaced by \( \rho_k \) and \( \phi_k \). By using vectors \( w_k \equiv \rho_k d_k \) in place of \( d_k \), the procedures of LSQR are given in the following steps.

1) Initialize.

\[
\begin{align*}
\beta_i u_i &= b, \quad \alpha_i v_i = A^T u_i \\
w_i &= v_i, \quad x_0 = 0, \\
\overline{\phi}_i &= \beta_i, \quad \overline{\alpha}_i = \alpha_i.
\end{align*}
\]

2) For \( i = 1,2,3,\ldots \) repeat steps 3-6.

3) Continue the bidiagonalization.

a) \( \beta_{i+1} u_{i+1} = A v_{i} - \alpha_i u_i \)

b) \( \alpha_{i+1} v_{i+1} = A^T u_{i+1} - \beta_{i+1} v_i \).

4) Construct and apply next orthogonal transformation.

a) \( \rho_i = (\overline{\rho}_i^2 + \beta_{i+1}^2)^{1/2} \)

b) \( c_i = \overline{\rho}_i / \rho_i \)

c) \( s_i = \beta_{i+1} / \rho_i \)

d) \( \theta_{i+1} = s_i \alpha_{i+1} \)

e) \( \overline{\rho}_{i+1} = -c_i \alpha_{i+1} \)

f) \( \overline{\phi}_i = c_i \overline{\phi}_i \)

g) \( \overline{\phi}_{i+1} = s_i \overline{\phi}_i \).

5) Update \( x, w \).

a) \( x_i = x_{i-1} + (\phi_i / \rho_i) w_i \),

b) \( w_i = w_{i-1} - (\theta_{i+1} / \rho_i) w_i \).

6) Test for convergence.

Exit if some stopping criteria have been met.
For the damped least-squares problem, we need to minimize \( \| A \lambda x - b \|_2 \),

where the scalar \( \lambda \) is the damping parameter. The equation (B-36) becomes

\[
\begin{bmatrix}
  I & B_k \\
  B_k^T & -\lambda^2 I
\end{bmatrix}
\begin{bmatrix}
  t_{k+1} \\
  y_k
\end{bmatrix}
= \begin{bmatrix}
  \beta_i e_i \\
  0
\end{bmatrix}
\]

and the \( k \)th approximation to the solution \( x \) is also \( x_k = V_k y_k \) defined in equation (B-33),

where \( y_k \) solves the subproblem

\[
\min \| B_k \lambda I y_k - \begin{bmatrix}
  \beta_i e_i \\
  0
\end{bmatrix} \|. \tag{B-48}
\]

The orthogonal factorization (B-40) becomes

\[
Q_k \begin{bmatrix}
  B_k \\
  \lambda I \\
  0
\end{bmatrix} = \begin{bmatrix}
  R_k \\
  \bar{f}_{k+1} \\
  0
\end{bmatrix}.
\]

The factorization (B-49) is formed similarly to (B-40) except two rotations are required per step instead of one. For \( k = 2 \), the factorization proceeds according to

\[
\begin{bmatrix}
  \alpha_1 & \beta_1 \\
  \beta_2 & \alpha_2 \\
  \beta_3 & \beta_3 \\
  \lambda & \lambda
\end{bmatrix}
\rightarrow
\begin{bmatrix}
  \tilde{\rho}_1 & \tilde{\phi}_1 \\
  \beta_1 & \alpha_2 \\
  \beta_3 & \beta_3 \\
  \lambda & \lambda
\end{bmatrix}
\rightarrow
\begin{bmatrix}
  \rho_1 & \theta_2 & \phi_1 \\
  \beta_2 & \beta_2 & \phi_2 \\
  \beta_3 & \beta_3 & \beta_3 \\
  \lambda & \lambda & \lambda
\end{bmatrix}
\rightarrow
\begin{bmatrix}
  \rho_1 & \theta_2 & \phi_1 \\
  \rho_2 & \theta_2 & \phi_2 \\
  \rho_3 & \phi_3 \\
  \psi_1 & \psi_1 & \psi_1
\end{bmatrix}
\rightarrow
\begin{bmatrix}
  \rho_1 & \theta_2 & \phi_1 \\
  \rho_2 & \phi_2 \\
  \phi_3 \\
  \psi_1 \\
  \psi_2
\end{bmatrix}
\]

The presence of \( \lambda \) complicates the algorithm description but it adds essentially nothing to the storage and work per iteration. The damping parameter \( \lambda \) discussed above is a single value. As we can see from equations (B-48) and (B-49), LSQR also works for various \( \lambda \) when the QR factorization proceeds and eliminates both the subdiagonals of \( B_k \) and the damping parameters below the subdiagonals.
Orthogonalization used in LSQR will result in favorable convergence properties compared with iterative methods (e.g., ART and SIRT) that do not employ orthogonality properties. Moreover, van der Sluis and van der Vorst (1987) show that the LSQR algorithm starts the construction of the solution by neglecting those components belonging to the smaller eigen values of $A^T A$. The contribution of the smaller eigen values eventually enter the solution more and more as iteration proceeds. This behavior attributes to the intrinsic damping properties of the algorithm, which are similar to the SVD method with sharp eigen value cutoff.

**B.4 Error and Resolution Analyses**

Since the pioneer work of Backus and Gilbert (1968), geophysicists have commonly recognized that a solution of an inverse problem must be evaluated together with its resolving kernel as well as its a posterior variance. A result is less valuable unless its reliability is evaluated properly using these quantities. A formulation of the resolution and the covariance matrices in discrete inverse problem (equations (B-22) and (B-23)) is now widely known and is applied in tomographic studies. The tomographic method of Zhao (1991) and Zhao et al. (1992) has adopted the skillful approach of analyzing error and resolution of the data set since the calculation is not straightforward because of the large matrix size.

By the concept of the resolution we wish to know how the true structure is reconstructed in the calculated image. The most direct means of testing the resolution of the inverted solution is to first calculate the sets of travel time delays that result from tracing the actual ray set through a synthetic test structure, then invert those delays as
though they are data, and finally compare the synthetic inversion with the initial structure. In this study, we carry out two kinds of synthetic tests.

The first is the checkerboard resolution test proposed by Humphreys and Clayton (1988). This method has been applied in Grand (1987) and Inoue et al. (1990). To construct the checkerboard, positive and negative velocity perturbations are assigned to the three dimensional blocks (in the present study, grid nodes) of a homogeneous velocity model at the same interval. Such an image is straightforward and easy to remember. It is easy to visually evaluate the image of the synthetic inversion of the checkerboard in order to understand where the resolution is good and where it is poor.

The second synthetic test is named the “Restoring Resolution Test (RRT) after Zhao (1991) and Zhao et al. (1992). The tomographic images obtained by inverting the real data set are taken as the synthetic model. Then synthetic arrival time data are calculated by tracing rays three-dimensionally in the synthetic model. The same random error as that contained in the real data set are also added to the calculated synthetic data. Inverting the synthetic data by using the same inversion algorithm, the restoring image of the real result can be obtained. Comparing the two images one can know whether or not and how the main trend or some special structures in the real results are realistically restored. The RRT synthetic test is necessary in the case when the synthetic test using the simple checkerboard image may not show the resolution for the more complicated structures such as magma chambers and dipping slabs (Zhao, 1991; Zhao et al., 1992; Leveque et al., 1993). Hence this test can assess the effect of 3-D ray paths on the final tomographic images. We don’t apply the RRT in this study.
Other tests can also be done by using synthetic models with different structural geometries and grid separations, in order to evaluate whether a feature that appeared in the inversion results can reasonably be well retrieved with the data set or not (Zhao 2001). In this study we conduct the checkerboard resolution tests with grid separations of $1^\circ$, $2^\circ$, $4^\circ$, $6^\circ$, $8^\circ$ and $10^\circ$.

The resolution matrix is the best possible measure of model resolution, but the resolution matrix depends strongly on regularization, which is directly governed by data quality as opposed to limits in data coverage (Boschi, 2003). As resolution is not everywhere constant, but higher in regions of more uniform coverage, ad-hoc parameterization and regularization schemes can be designed to stabilize the solution in undersampled areas (Boschi, 2003).

**B.5 Computer Programs**

The computer program of TOMOG3D, developed by Zhao (1991) and Zhao et al. (1992), is in the Fortran language having more than 50 subroutines. It has two versions. The first version adopts the damped least squares method, while the second uses the LSQR method.

The algorithm of version 1 is as follows:

1) Program inputs:
   a) control parameters;
   b) hypocentral locations;
   c) arrival time data;
   d) initial velocity model.
2) Repeat (3) and (4) until convergence:

3) Process earthquakes 1 to M (total M earthquakes)
   a) relocate hypocenter by 3-D ray tracing;
   b) do forward problem by 3-D ray tracing to compute travel times, ray paths and their derivatives;
   c) parameter separation;
   d) construct data kernel.

4) Inversion: damped least squares method.

5) Resolution analysis.

6) Program outputs:
   a) final velocity model;
   b) hit-count;
   c) resolution matrix;
   d) standard error.

The algorithm of version 2 is as follows (Figure B-7):

(1) and (2): same as that of version 1.

(3) Process earthquakes 1 to M (total M earthquakes)
   (a) relocate hypocenter by 3-D ray tracing;
   (b) do forward problem by 3-D ray tracing to compute travel times, ray paths and their derivatives;
   (c) construct data kernel.

(4) Inversion: LSQR method.

(5) Program outputs:
(a) final velocity model

(b) hit-count.

The number of unknown medium parameters (here, the total number of grid points), NU, is an important quantity in these programs, because it marks the size of a tomographic problem to be solved. When NU is smaller than about 3000, both versions of these programs can be used (Zhao, 1991). Version 1 has an advantage that resolution matrix and the standard error of the unknown parameters can be jointly obtained with the final velocity model.

In Version 2, only non-zero elements of the very large and sparse derivative matrix are stored in which usually 99% of elements are zero. Therefore memory storage is greatly saved. Moreover, adopting the LSQR solver makes Version 2 very efficient. As a result, Version 2 needs much less memory storage and much shorter CPU time than that of Version 1 for the same problem (Zhao, 1991). The two programs give substantially the same results (Zhao, 1991). Version 2 can handle a huge tomographic problem with several millions unknown parameters and arrival time data. This version is very efficient for any other complex tectonic and geological settings as based on grid method of model parameterization. Several additional features of the 3-D tomographic method of Zhao (1991) and Zhao et al. (1992) make the methodology superior over that of others (Inoue et al., 1990; Thurber, 1992; Kissling et al., 2001).

I use version 2 of the 3-D tomographic method of Zhao (1991) and Zhao et al. (1992) to study the complicated regions of China and surrounding areas in this thesis. The resolution tests (Checkerboard and Synthetic tests) have been carried out to assess the recovery of our velocity tomograms.
References


Horie, A., 1980. Three-dimensional seismic velocity structure beneath the Kanto district by inversion of P-wave arrival times. *Ph.D. Thesis*, University of Tokyo, Tokyo.


Figure B-1. Seismic velocity model illustrating a couple of discontinuities with three-dimensional change of velocity between the discontinuities. The Conrad discontinuity and the Moho discontinuity are the two common discontinuities in the crust and uppermost mantle.
Figure B-2. In every layer, three-dimensional grid nets are set up independently (Zhao, 1991). (a) the uppermost layer bounded by the surface and the shallowest discontinuity; (b) the lowest layer bounded by the deepest discontinuity. Area down to dashed line (AB) is the modeling space. The dark dots denote grid nodes.
Figure B-3. Illustration of a geometric property of the ray equation showing the direction of the ray curvature $d\vec{r}/ds$ as antiparallel to the velocity-gradient $(\nabla v)$ and normal to the ray direction, $d\vec{r}/ds$. 
Figure B-4. Illustration of the three-point perturbation scheme in three-dimensionally continuous velocity model. An initial ray segment \((X_{k-1}, X_k, X_{k+1})\) is perturbed approximately to satisfy equation (B-4) by finding a new point \(X'_k\) (Um and Thurber, 1987).
Figure B-5. Illustration of the perturbation scheme to find the intersecting point between the ray path and the discontinuity (Zhao, 1991; Zhao et al., 1992).
Figure B-6. Illustration of the algorithm of the new ray tracing method (Zhao, 1991; Zhao et al., 1992). Pseudo bending technique and Snell’s law are jointly used iteratively to perturb an initial ray path.
Figure B-7. The flowchart of the seismic tomographic study.