TELESEISMIC TOMOGRAPHY BENEATH HI-CLIMB STATION ARRAY IN WESTERN TIBETAN PLATEAU

BY

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THESIS

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Urbana, Illinois

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This study aims to explore the velocity structure of the lithosphere and upper mantle beneath the Hi-climb station array in mid-western Tibetan Plateau (TP), thus to understand the deformation scheme of the Indian-Eurasian collision. The Hi-Climb seismic array, which includes 75 seismic stations, was collecting data from late 2002 to 2005. It forms a nearly linear shape extending from the Himalayan foreland into the western-central TP. During the three-year deployment period of the array, 24 teleseisms with good signal-to-noise ratio were chosen out of a large amount of events, allowing 1320 relative travel time residuals to be picked with high accuracy in this study. P phases were recorded and contributed to the calculation of relative travel time residuals, which were finally mapped as P wave velocity perturbations with respect to the ak135 global reference model using teleseismic tomography. Checkerboard synthetic test was performed, showing good resolution at depths between 100 km to 300 km in horizontal slices. Good recovery was also shown in North-South slices at longitudes of 85.2° E and 84.2° E, respectively. The resulting horizontal tomographic images exhibit a high-slow-high velocity structure from north end of the array to the south. The two boundaries separating the velocity structures are Bangong-Nujiang suture (BNS) and the Indus-Yarlung suture (IYS). Positive velocity perturbation was found north to the Bangong-Nujiang suture (BNS), suggesting a higher velocity zone in Qiangtang Terrane compared to Lhasa Terrane. The other high velocity zone is present at the south end of
the station array, which is south to the Indus-Yarlung suture. Cross-sections on north-south slices also present a fast-slow-fast velocity, which signals a thicker and less dense lithosphere beneath the middle part of the station array, and thinner but denser lithospheres at north and south edges. The result is in good agreement with previous work and it may be a consequence of the subducting Indian plate toward Eurasian plate.
I sincerely thank my adviser, Prof. Xiaodong Song for his continuous and tireless support and guidance during my graduate study at the University of Illinois. I wish to express my appreciation for his supporting on each of my decisions and his valuable suggestions. I also wish to thank Prof. Lijun Liu, for his endless help during my defense period, as well as his careful work on my thesis. I also thank my group members, namely Jiangtao Li, Zheng Tang and Jing Jin, for their assistance and discussions on my research. Thanks to Prof. Nick Rawlinson for providing the inversion program. Finally, thanks to my family and boyfriend who have encouraged me and helped me go through each difficulty throughout my experience.
# TABLE OF CONTENTS

Chapter 1: Introduction ........................................................................................................... 1

Chapter 2: Data and data processing methods ..................................................................... 10

Chapter 3: Tomography methodology .................................................................................... 22

Chapter 4: Model resolution assessment .............................................................................. 31

Chapter 5: Results and discussion ....................................................................................... 41

Chapter 6: Conclusions and future work ............................................................................. 49

References ........................................................................................................................... 51
CHAPTER 1

INTRODUCTION

1.1 Tectonic settings

The “roof of the world” -- Tibetan Plateau, is known as the world’s highest and largest plateau, with an average elevation exceeding 4,500 meters. It stretches about 1,000 kilometers from north to south and 2,500 kilometers east to west. Among the numerous mountain ranges, it harbors the world’s highest summit, Everest. It has been suggested in modern tectonic theory that Himalays formed due to the continental collision of Indian plate and Eurasian plate since the upper Cretaceous period about 70 million years ago (Dewey et al., 1988; Yin and Harrison, 2000). What mechanism has caused such large amount of shortening that results in an uplifted Tibetan Plateau? Since the earliest work by Argand, different assumptions and models have been proposed to understand the evolution and formation of this uplifted plateau (He et al., 2010; Dewey and Burke, 1973; Platt and England, 1994; Meyer et al., 1998; Zhao and Morgan, 1987). Although many studies have been done on this project during the past decades, the sub-surface seismic structure remains controversial and requires further imaging studies.

1.2 Teleseismic tomography
A key to understanding the ongoing tectonic processes and surface deformation is to refine the underground structure. Seismic tomography provides a most effective way to image the Earth’s deep structure. By gathering the travel time measurements of compressional wave (P-wave) and shear wave (S-wave), and planting the data into an inverse problem, we can compile a 2D or even 3D subsurface velocity structure.

There are three approaches of performing seismic tomography according to the distance of earthquake events: local earthquake tomography, regional earthquake tomography, and teleseismic tomography. Local and regional earthquake tomography collect the rays with large incident angles from nearby sources, and could determine shallow structures (e.g., crust) well. However, there are also restraints involved. Since the location of earthquake sources are perfectly unknown, certain relocation processes are required during the iterative inversion to improve the location as well as velocity model. The stations must cover the study area and, together with the earthquakes, the volume of interest should be well sampled. Due to the small distance between stations and earthquakes, low signal-noise ratio always happens, which will decrease the quality of seismic data. Teleseismic tomography, however, provides us with high quality seismic data due to the large ray travelling distance. The accurate locations of earthquake sources are hence less crucial because differential travel times can effectively eliminate errors from the source side. Moreover, different from local and regional earthquake tomography, teleseismic tomography presents a better resolution at a deeper depth due
to large epicentral distance involved. However, one limitation of teleseismic tomography is the low resolution and accuracy at shallow depths right beneath the stations due to the less dense criss-cross.

The basic idea of teleseismic event is to use the relative travel time between the stations from the same event to cancel out the effect from the source and the ray paths outside the study region below the stations. This is based on the approximation that teleseismic rays close to the source are very close to each other, such that we could assume all the rays travel along the same path and go through the same structure before entering the volume on the receiver side, leading to the same travel time in the area between the source and our study area. Hence the observed travel time difference only reflects the effect from the study region other than the total area from the source to the stations.

1.3 Hi-CLIMB station array

Over the past decades, many large-scale seismic station array such as INDEPTH-I, II, III during 1991-2000 (Zhao et al., 1993; Brown et al., 1996; Kind et al., 1996; Nelson et al., 1996), HIMNT during 2001-2002 (Schulte-Pelkum et al., 2005), have been built in order to explore the mechanism of Indian-Eurasian collision. Many studies have been done on crustal structure or even deeper structures, which largely revealed the existence of subduction of the Indian Plate beneath the Tibetan Plateau. However, most of the
research areas are in central and eastern Tibetan Plateau, leaving the western side less well understood.

Hi-CLIMB (Himalaya-Tibetan Continental Lithosphere During Mountain Building) experiment is a newly deployed project from 2002-2005. It is located in the western Tibetan Plateau, extending from the Himalayan foreland into the western-central Tibetean Plateau. The seismic stations are closely-spaced (~5 km) with a wide aperture (~800 km). It forms an approximately linear shape extending northwards (Figure 1.2), cutting across Indus-Yarlung Suture (IYS) and Bangong-Nujiang Suture (BNS). There are limited work done on deeper P wave velocity structure beneath Hi-CLIMB station array (Zhang et al., 2012; Hung et al., 2010; Hung et al., 2011), and there still exist some intriguing deeper structures need to be decoded. Therefore, we aim to further explore the underground structure by using P wave teleseismic tomography, and to provide a supplement to current study on Indian-Eurasion collison.
1.4 Figures

Figure 1.1 Map of Hi-CLIMB station array (in black triangles) and topography of Tibetan Plateau in the color background. Major sutures are in gray lines: ATF (Altyn Tagh Fault), KF (Kunlun Fault), JRS (Jinsha River Suture), BNS (Bangong-Nujiang Suture), IYS (Indus-Yarlung Suture).
Figure 1.2 Strike of Hi-CLIMB station array calculated by least square method. The strike is about 340, labeled in the blue line. Green triangles indicate the distribution of Hi-CLIMB seismic stations.
Figure 1.3 Basic assumption and principles of teleseismic tomography. Red star represents for the teleseismic source, red lines are the ray paths, and the box is the study area.
### 1.5 Tables

**Table 1.1**

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**Table 1.1** List of station names, latitudes, longitudes, and elevation. Data acquired from IRIS Data Center.
2.1 Data descriptions

Our dataset were obtained from IRIS Data Center, containing the earthquake catalog from MHDF over a time span of 2004 to 2005. Of the numerous events recorded by Hi-climb, 24 teleseimic events, and a total of 1320 seismic rays with good signal-to noise ratios were selected from hand picking. Figure 2.1 shows the distribution of these events on an equidistant projection centered on Hi-climb station array in Tibetan Plateau. Given the array approximately forming a linear shape, we hence picked the events with backazimuth within +/ - 15 degrees with respect to the array strike. A good ray coverage in both azimuth and incident angles were fetched from southeast, with many events occurring along the coastline of southern Indonesia and its offshore area; fewer earthquakes were detected from northwest, mainly in the North Atlantic Ocean, however it still remains a good azimuth coverage. Table 2.1 shows a detailed earthquake event catalog used in this study.

The distances from the earthquake sources to the array are within 30° to 90°. It assures highly recognizable P phases, which originate from the sources, turn back to the Earth’s surface from the upper mantle and represent for the initial arriving phase of teleseismic
rays. The depth of the earthquake sources varies from 10 km to 150 km. We have tried to acquire more events with deeper sources, because the seismic rays from deeper earthquakes are less affected by the noise in the shallower crust and surface waves travelling along the Earth’s surface. But it is also crucial to have the events with different depths, which provides different incident angles beneath the seismic station. Normally, deeper earthquakes produce body waves with steeper angles of incidence compared to those from shallower earthquakes. This wide range of earthquake depths, together with the variable distances, have guaranteed a good ray coverage inside our study area.

2.2 Data processing and measurements

In teleseismic tomography, relative arrival times of teleseismic rays provide the primary source of data for the modeling of velocity structure (Aki et al., 1977; Humphreys et al., 1984; Nolet, 1987). In this study, we first picked all the P phases of each single ray. Due to the high signal-to-noise ratio, after applied a WWSSN short period instrument, the P phases were then clearly shown. Figure 2.2 shows all the ray traces from the earthquake event occurred in North Atlantic Ocean on March 6th, 2005. They have been only rearranged in the order of distance. The clearly-shown first arrival signals are the P phases. Afterwards, a cross-correlation method was applied to each pair of P phases from the same event, aiming to calculate the relative travel time of each pair of rays.
Cross-correlation (Bungum and Husebye, 1971) is used to evaluate the similarity between two waveforms as a function of time lag applied to one of them. In discrete form, the truncated estimate of the cross-correlation function between the ith and jth traces is (Vandecar and Crosson, 1990)

$$\phi_{ij}(\tau) = \frac{\delta t}{T} \sum_{k=1}^{\tau/\delta t} x_i(t_i^P + t_o + k\delta t + \tau)x_j(t_j^P + t_o + k\delta t)$$

Where $x_i$ = digital data from ith trace; $t_i^P$ = ith trace’s preliminary arrival time estimate; $\tau$ = time lag relative to preliminary arrival time estimates; $T$ = length of correlation window (sec); $t_o$ = time between preliminary arrival time estimate and when correlation window begins; and $\delta t$ = sample interval. As a result, the time lag $\tau$ gives the amount that we need to shift the ith trace to make it best fit the jth trace.

Figure 2.3 gives an example of two traces from the same earthquake event occurred on March 6th, 2005, in North Atlantic Ocean. They were recorded by stations H1270 and H1280, respectively. T1 marker indicates the P phase for each trace, which was picked visually. The digital P phase arrival times for these two traces are 82.943 sec and 82.606 sec, which give a visual estimate relative arrival time of $82.943 - 82.606 = 0.337$ sec. Figure 2.4 shows the cross-correlation coefficient as a function of $\tau$. The $\tau$ value ($\tau^{max} = -0.34$ sec) at the maximum peak (labeled as PTPMAX) represents the cross-correlation derived relative P phase arrival time with the largest coefficient between
these two traces, which is almost the same with the result coming from the visual estimate.

A cross-correlation coefficient is used to evaluate the quality of cross-correlation, which is given by (Vandecar and Crosson, 1990)

\[ r_{ij} = \frac{\Phi_{ij}(\tau^{\text{max}})}{\sigma_i \sigma_j} \]

Where \( \Phi_{ij}(\tau^{\text{max}}) \) is the maximum magnitude of cross-correlation function, \( \sigma_i^2 \) is the sample variance of the ith trace data computed over the appropriate correlation window. The cross-correlation coefficient for the two traces above is 98.73%, which is very high, indicating a precise cross-correlation result.

### 2.3 Least square method

Due to the existence of noise, and the fact that waveforms can never be coherent from station to station, the cross-correlation derived relative travel time measurement can not be perfectly consistent (Vandecar and Crosson, 1990). For example, \( t_{AB} + t_{BC} \neq t_{AC} \), where A, B, and C are stations; \( t_{AB}, t_{BC}, \) and \( t_{AC} \) are the cross-correlation derived relative arrival times between ray A and B, B and C, A and C. For n stations, an overdetermined system will be generated by \( c_n^2 = n(n - 1)/2 \) equations.
We added one more constraint equation to this system,

\[ \sum_{i=1}^{n} t_i = 0 \]

This equation indicates a reference marker that forces the mean value of arrival times to be zero. In this case, the new system contains \( n(n - 1)/2 + 1 \) equations. Taking \( n=4 \) as an example, the new system can be written as the following matrix form, \( A t = \Delta t \), making it nonsingular. We can therefore calculate \( t_i \) by least-square method.

\[
\begin{bmatrix}
1 & -1 & 0 & 0 \\
1 & 0 & -1 & 0 \\
1 & 0 & 0 & -1 \\
0 & 1 & -1 & 0 \\
0 & 1 & 0 & -1 \\
0 & 0 & 1 & -1 \\
1 & 1 & 1 & 1
\end{bmatrix}
\begin{bmatrix}
t_1 \\
t_2 \\
t_3 \\
t_4 \\
\end{bmatrix}
= 
\begin{bmatrix}
\Delta t_{12} \\
\Delta t_{13} \\
\Delta t_{14} \\
\Delta t_{23} \\
\Delta t_{24} \\
\Delta t_{34} \\
0
\end{bmatrix}
\]

After adding the new constraint condition, the new system is nonsingular. A least square method was applied to calculate the arrival times \( t \), which is given by

\[
t = (A^T A)^{-1} A^T \Delta t
\]

Since \( A^T A = nI \), the equation above can be rewritten as

\[
t = (1/n) A^T \Delta t
\]

Or simply,
We will be able to get optimized relative arrival times $\Delta t_{ij}$ from this new set of $t_i$, and they are consistent. Figure 2.5 shows the ray traces after shifting each ray to the reference marker $\sum_{i=1}^{n} t_i = 0$ with an amount of $\frac{\sum_{i=1}^{n} t_i}{n} - t_i$. The well-aligned P phases indicate a precise relative travel time measurement.
2.4 Figures

Figure 2.1 Event distribution on an equidistant projection centered on Hi-CLIMB station array in western Tibetan Plateau. The blue triangle represents for the array, and red dots represent for all the events used in this study.
Figure 2.2 Raw seismograms from an event occurred in North Atlantic Ocean on March 6th, 2005. Data are arranged in distance.
Figure 2.3 Relative arrival time obtained by cross-correlation. (a) Seismograms showing the input two waveforms with P phase marked in T1. (b) The output of cross-correlation result. Peak with maximum cross-correlation coefficient is labeled by PTPMAX, and the corresponding time lag on X-axis indicates the relative arrival time between two waveforms.
Figure 2.4 Realigned seismograms after shifting relative arrival time calculated from cross-correlation of each single ray. Orange line marks the well-aligned P phases.
### Table 2.1

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**Table 2.1** List of teleseismic events in this study, including the description of occurring time, location (longitude, latitude and depth), distance and back azimuth from the array, and magnitude. Data acquired from IRIS Data Center.
3.1 Travel time residual

In this study, we used teleseismic relative travel time residuals to invert for the velocity variations. The relative travel time residual is not only a scale used to evaluate the difference between real model and predicted model, but also an important component of the inversion equation. During the inversion process, the residuals are aimed to be minimized. The smaller the residual is, the less difference exists between the real model and predicted model.

For the \( j \)th event at the \( i \)th station, raw travel time residual can be written as

\[
t_{ij} = T_{ij}^{obs} - T_{ij}^{cal}
\]

where \( T_{ij}^{obs} \) represents for the observed travel time from \( j \)th event to the \( i \)th station; \( T_{ij}^{cal} \) indicates the calculated travel time from the \( j \)th event to the \( i \)th station through a predicted model. Then the relative travel time residual is calculated by

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

The mean travel time residual is given by
\[ \bar{t}_j = \frac{1}{n_j} \sum_{i=1}^{n} t_{ij} \]

where \( n_j \) is the total number of rays in \( j \)th event.

Figure 3.1 shows the observed arrival time scattering (Figure 3.1.a) and relative travel time residuals (Figure 3.1.b) with respect to the ak135 global reference model (Kennett et al. 1995) from an event occurred northwest to the station array. In Figure 3.1.a, the observed arrival time scatterings are mostly within +/- 1 sec. However, after subtracting the predicted arrival time scattering, relative travel time residuals in Figure 3.1.b are almost within +/- 0.4 sec. The majority positive relative residuals suggest that our observed travel times are mostly larger than the predicted travel times, indicating a thicker crust beneath the station array, which is in good agreement with the tectonic structure of Tibetan Plateau. Since the events used in this study are either from northwest or from southeast to the array, they form an approximately 2-D plane including all the ray paths. The patterns from completely opposite directions may provide insightful information on underground velocity structures. Therefore, as another example, Figure 3.2 shows the observed arrival time scattering and the relative travel time residuals from another event occurred southwest to the station array, which has the similar pattern with Figure 3.1.
3.2 Ray tracing

The observed travel time residuals are relatively easier to acquire, since they are directly measured from our dataset. However, in order to calculate the predicted travel time residuals, we need to manage the ray paths first, which requires a precise ray tracing scheme. In this study, we set up 3-D grids throughout our study area, in which each grid has an initial velocity. Outside this area, we assume that the heterogeneities have no effect on ray paths.

We traced the rays from the sources listed in Table 2.1, to the location of each station (Table 1.1) on the Earth’s surface through a spherically symmetric Earth (ak135 model). We then used ray tracing to compute the travel time from the surface of the model back to the base of the model, and subtracts this value from the ak135 travel times (Rawlinson et al., 2006). In this way, we acquired the travel times from sources to the base of our study area. The next step is to trace the rays inside our study area, where we applied fast marching method. It is a grid-based eikonal solver used to solve the forward problem of travel time prediction (de Kool et al., 2006).

After all the ray paths were determined, we hence could calculate the travel time of the ray segments inside each grid, therefore able to calculate the relative travel time residuals.
3.3 Inversion equation

The travel time of one single ray inside our study area could be written as

\[ t = \int \frac{dl}{v} \]

where \( dl \) is the distance that the ray travels inside each grid, and \( v \) is the velocity inside that specific grid. The integration starts from the base of the study area, and ends at the surface. The relative residual time \( \delta t \) can be written as

\[ \delta t = \int \delta \left( \frac{dl}{v} \right) = \int - \frac{\delta v}{v^2} dl = - \int \frac{\delta v}{v} \frac{dl}{v} = - \int \frac{\delta v}{v} dt \]

The large number of rays we have in this study lead to large amount of equations, each of which describing a single ray. Therefore, we constructed a matrix equation based on the equation above:

\[ Gm = d + e \]

\( G \) is a large sparse kernel matrix containing the travel time of rays inside each grid (because for a single ray, it only travels through a little portion out of the whole study area due to the large number of the total grids), which links up the vector \( m \) and \( d \). \( d \) is a column vector whose entries are the relative travel time residuals \( (r_{ij}) \). \( m \) is also a column vector composed of velocity perturbations for each grid that we would like to obtain. \( e \) is an error vector.
We applied LSQR method (Paige and Saunders, 1982a, 1982b) to solve the equation

\[ \mathbf{Gm} = \mathbf{d} + \mathbf{e} \]

by minimizing the least square system:

\[ \| \mathbf{Gm} - \mathbf{d} \|^2 + \lambda^2 \| \mathbf{m} \|^2 + \varphi^2 \| L \mathbf{m} \|^2 \]

Because tomographic inversions are generally ill-conditioned, which leads to unstability of the inversion process. We therefore imposed regularization on this system with additional constraint, such as damping constraint \( \lambda \) in the second term and smoothing constraint \( \varphi \) in the third term. Damping parameters are often determined on the basis of a trade-off relation between data variance and model variance (Eberhart-Phillips, 1986).

The second term \( \lambda^2 \| \mathbf{m} \|^2 \), which is the model vector with a damping constraint, is used to suppress the overreaction of the model norm. In the third term, smoothing constraint allows us to get a smooth tomography result without sharp velocity changes at the edge of the grids. \( L \) is the finite difference Laplace operator (Lees and Crosson, 1989), which controls the roughness over the model space.

Seismic inversion needs to solve the equation above repetitively. A new model vector \( \mathbf{m} \) is generated after each inversion, and this new model will be served as an optimized predicted model for the next inversion. The iteration stops when the result reaches convergence.

3.4 Model parameterization
Our model area extends from 26° N ~ 37° N in north-south direction and 82° E ~ 88° E along east-west direction on the Earth’s surface. It reaches 300 km underground, and 6.5 km above sea level. We divided this 3D area into grids, which are set to be 0.25° × 0.125° horizontally and separated every 25 km on depth. Compared to the station array covering area, the model area has large margins to avoid missing points and misallocation at the bottom. We used ak135 global model as our initial 1D velocity model (Figure 3.2). After trying out different damping and smoothing parameters, we set damping parameter to be 5.0 and smoothing parameter to be 10.0.
3.5 Figures

Figure 3.1.a

Figure 2.1.b

Figure 3.1 (a) Observed arrival time scattering from an event occurred northwest to the station array; (b) Relative travel time residuals with respect to ak135 model from the same event.
Figure 3.2 (a) Observed arrival time scattering from an event occurred southeast to the station array; (b) Relative travel time residuals with respect to ak135 model from the same event.
**Figure 3.3** P and S wave velocity from ak135 global model (Kennett *et al.*, 1995). Figure acquired from www.iris.edu.
CHAPTER 4
MODEL RESOLUTION ASSESSMENT

4.1 Introduction of checkerboard test

We applied a checkerboard test (Leveque et al., 1993) to evaluate the resolution of our tomographic results. Because we can never see the real structure of Earth’s interior, the reliability of the tomographic images in inverse problems is always a concern. Checkerboard test, however, provides an effective way to know how close the seismic image is to the real structure. The basic idea of checkerboard test is that we first build up a synthetic checkerboard model served as a “real” model by varying velocity as a sinusoidal function based on a 1D velocity model. The next, perform seismic inversion iteratively with the 1D velocity model to be the initial model. Finally, compare the recovered checkerboard pattern with the input synthetic checkerboard model. The areas with high similarity indicate good recovery and high resolution, the tomographic images inside these areas are therefore more reliable. Normally, good recovery happens at the area with denser ray coverage compared to the margin area with poor ray coverage.

4.2 Checkerboard test results
Due to the nearly linear shape of Hi-climb station array, we applied a checkerboard test not only in horizontal slices, but also in north-south cross-section slices. The test involved using the identical receivers, sources and type of phases from the raw dataset. The travel time residuals were calculated from the checkerboard model with maximum +/- 4% of velocity perturbation based on ak135 model (Figure 4.1.a, Figure 4.2.a). The travel time residuals were then inverted through the same tomographic procedure stated above. The recovered structures are shown in Figure 4.1.b-g and Figure 4.2.b-c, from which we can directly see the resolution throughout the model area.

Figure 4.1. b-g show the test results on horizontal slices, ranging from 50 km to 300 km. Overall, the recovery of the pattern is good at the area close to the station array, which is due to the dense ray coverage. The recovery is generally good between 100 km and 250 km. For shallow part (0-50 km), there are less criss-cross of the teleseismic rays, which leads to a relatively lower resolution. Since the rays are all from teleseismic events, they converge beneath the station array, which means greater portion of the recovered pattern appears as the depth increases. At the depth of 300 km, the smearing of the pattern is due to it hits the bottom edge of the model area.

Figure 4.2.b-c show the test results of north-south cross-section slices. Due to the special geometry of the station array, we picked two longitudes matching the array segments.
from high latitude area and low latitude area respectively. Compared with the
geometry of Hi-Climb station array in Figure 1.1, stations with latitude ranging from 29°
N to 32° N gather together and form a nearly north-south striking line at the longitude
of ~85.2° E, which also explains why good recovery of the pattern appears in the low
latitude area in Figure 4.2.b but no patterns in the high latitude area. Similarly, the
recovery pattern in Figure 4.2.c mainly appears at high latitude area (32° N to 34° N)
which is also due to the horizontal bounds of the station array.

The checkerboard test was terminated after the sixth inversion, giving a result of the
final RMS residual reduced significantly by 70.45%, and the data variance was reduced
by 91.28% compared to the original synthetic data. Figure 4.3 shows how RMS residual
and variance change with the iteration. After the first iteration, both have reduced
significantly and gradually converge.
4.3 Figures

Figure 4.1.a
Figure 4.1.b

Figure 4.1.c

Depth = 50km

Depth = 100km
Figure 4.1.d

Figure 4.1.e

Depth = 150km

Depth = 200km
Figure 4.1 (a) Input synthetic checkerboard model in horizontal slices with maximum velocity perturbation +/- 4% based on ak135 global model. (b)-(f) Output recovered checkerboard models at depth of 50km, 100km, 150km, 200km, 250km, 300km. Red triangles are Hi-CLIMB station array.
Figure 4.2.a
Figure 4.2 (a) Input synthetic checkerboard model in N-S cross-section, with maximum velocity perturbation of +/- 4% based on ak135 global model. (b) Recovered checkerboard model at longitude of 85.2° E. (c) Recovered checkerboard model at longitude of 84.2° E.
Figure 4.3 RMS data residuals and variance after six iterations. RMS residual is reduced by 70.4%, and variance reduced by 91.3%.
5.1 Results

After six iterations with ak135 as initial model, both RMS residuals and their variance gradually converged, indicating a stable inversion system. RMS value has reduced by 36%, with a final value of 0.33 sec. And the variance of the data decreased by 59%. Velocity images are mapped in velocity perturbation with respect to ak135 model. Horizontal and cross-section velocity structures beneath Hi-climb station array are shown in Figure 5.1 and 5.2, which exhibit some distinctive patterns.

From the horizontal slices (Figure 5.1.a-f), a high velocity area is present at 32° N at the depth deeper than 100 km, extending northwards until 34° N. On the opposite, the structure pattern is dominated by a slow velocity perturbation in the middle area, between ~32° N and ~30° N. The boundary between the two velocity zones is Bangong-Nujiang suture (BNS) at 32° N, which divides Qiangtang Terrane in the north from the Lhasa Terrane in the south. It is an indication that a thinner and denser lithosphere may exist beneath Qiangtang Terrane, in comparison with a thicker and less dense lithosphere beneath Lhasa Terrane. Studies from (Zhang et al., 2011) show that an average of 60~65 km thickness crust is found beneath Qiangtang Terrane, which is
slightly thinner than that of Lhasa Terran with a maximum thickness of 70 km. Another significant feature is that high velocity area also appears at southern end of the station array, stretching from 30° N to ~28° N. The transition at 30° N happens to be the location of Indus-Yarlung suture (IYS), which separates Lhasa Terrane from Himalayan block.

Same features also present in cross-section slices in Figure 5.2.a-b. The former cross-section (Figure 5.2.a) is acquired at longitude of 85.2° E, mostly representing for the structure beneath southern part of the array. It shows slow velocity perturbation ranging from 29.5° N to 32° N, a fast velocity perturbation south to 29.5° N. Because no stations were built north to 32° N at 85.2° E, thus no structure is shown in high latitude area in Figure 5.2.a. However, Figure 5.2.b gives the velocity structure at higher latitude, ranging from 32° N to 34° N with a dominant high velocity structure.

5.2 Discussion

Since Hi-CLIMB station array is a newly deployed experiment, only limited research has been done on the deep structure beneath it. Even that, we can still find some previous work with which we can compare our results. The tomographic images from He et al. 2010 show a slow velocity perturbation at ~30° N between 84° E and 86° E at
depth between 150 km and 200 km. At deeper structure between 200 km and 300 km, it also shows a fast-slow-fast velocity zone which is in good agreement with our results.
5.3 Figures

Figure 5.1.a

Depth = 50km

Figure 5.1.b

Depth = 100km
Figure 5.1.c  
Delta υ (m/s)

Depth = 150km

Figure 5.1.d  
Delta υ (m/s)

Depth = 200km
Figure 5.1 Tomographic images in horizontal slices at depth of 50km, 100km, 150km, 200km, 250km, 300km. Velocity structure is mapped in P wave velocity perturbation with respect to ak135 model. Red triangled are Hi-CLIMB station array.

Figure 5.1.e

Figure 5.1.f
Figure 5.2. Tomographic images in N-S cross-section slices. (a) Cross-section image at longitude of 85.2° E. (b) Cross-section image at longitude of 84.2° E. Velocity structure is mapped in P wave velocity perturbation with respect to ak135 model.
Figure 5.3 RMS data residuals and variance after six iterations. They gradually converge, indicating a stable system.
CHAPTER 6

CONCLUSIONS AND FUTURE WORK

6.1 Conclusions

We performed a 3D teleseismic tomography beneath Hi-CLIMB seismic station array in western Tibetan Plateau using P phases. A least square method was used to delinate the inconsistency during residual measurements. Checkerboard synthetic tests show a good recovery below 100 km in horizontal slices, and partially good recovery in N-S slices depending on the ray coverage at different longitude. The tomographic images show a fast-slow-fast velocity structure from north to south beneath the array which is in good agreement with some previous work. The slow velocity under Lhasa Terrane may suggest a thicker crust or less dense lithosphere compared to the northern Qiangtang Terrane with a thinner crust or denser lithosphere.

6.2 Future work

The initial model for this study is ak135 global model, which has a Moho depth of 35 km world wide. However, the crustal thickness of Tibetan Plateau is much lager than this average thickness. Thus based on other previous work (Nábelek et al., 2009) on crustal Vp velocity model, we could perform a crustal correction to diminish the error
coming from the shallow part. Therefore, this may further improve the tomographic results of deeper structure. Moreover, other phases such as PKP and PKiKP (Rawlinson and Kennett, 2008), could be also involved to increase the ray coverage. More teleseismic events with wider range of back azimuth could also be included to improve the ray coverage.
References


