Seismic Attenuation of Regional Phases
in the Northern Middle East
and Eastern Tibetan Plateau

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The undersigned, appointed by the dean of the Graduate School, have examined the
dissertation entitled

**SEISMIC ATTENUATION OF REGIONAL PHASES
IN THE NORTHERN MIDDLE EAST AND EASTERN TIBETAN PLATEAU**

presented by Wenfei Ku,

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Seismic Attenuation of Regional Phases in the Northern Middle East and Eastern Tibetan Plateau

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Abstract

Temperature and composition are two major causes for subsurface seismic anomalies. Positive temperature anomalies will lead to a reduction in both attenuation and velocity; however, compositional anomalies should not necessarily produce a strong correlation in attenuation and velocity. As a result, by combining velocity and attenuation structure we can distinguish between compositional and temperature anomalies. Using efficiency tomography and Q tomography, I have constructed S_n attenuation models for two continental-continental collision zones, the northern Middle East and the eastern Tibetan Plateau.

The Tibetan Plateau was formed by the continental collision between Indian and Eurasian plates that has been going on at least since ~50Ma. Two tomographic techniques have been used to determine the attenuation structure of the uppermost mantle beneath the eastern Tibetan Plateau. I observe lateral heterogeneity of S_n attenuation beneath the southern Tibetan Plateau that indicates a complex geometry of the underthrusting Indian

continental lithosphere (UICL). Sn is blocked with relative low Q values across the Qiangtang block and Songpan-Ganzi block indicating a hot and weak lithosphere. This observation can be caused by mantle upwellings induced by the sinking slab detached from the UICL.

The Turkish-Iranian plateau and Zagros, the main tectonic feature of the northern Middle East, was formed as a result of the continental collision between the Arabian and Eurasian plates since Early Cenozoic (23-35Ma). I have collected a large Sn waveform data set in the northern Middle East that I have quality controlled using both automated and manual approaches. Two tomographic techniques have been used to determine the attenuation structure of the uppermost mantle. I observe inefficient/blocked Sn and low Q values in the Turkish-Iranian plateau indicating a hot and thin mantle lithosphere.

Intrinsic attenuation is the dominant uppermost mantle shear wave attenuation mechanism beneath the eastern Anatolian plateau and Lesser Caucasus. Partial melting appears to be the main cause of high attenuation in two of the regions. Scattering attenuation appears to be the dominant mechanism in the Zagros. The high attenuation in the Iranian plateau is likely not caused by partial melting thus the seismic anomalies in the uppermost mantle are likely compositional.

Data censorship is a common problem in seismic attenuation studies. Discarding blocked Sn paths will cause left censored data problem and the resulting model will be biased to high Q values. Using Level of Detection Divided by Two (LOD/2) technique, I am able to obtain lower Q values and smoother variations in the resulting models comparing with censored models.

Chapter 1: Introduction

The Himalaya and Zagros are two prominent mountain belts built by continental collision. The Tibetan and Iranian plateaus illustrate a similar set of processes but at differing stages of development (Hatzfeld and Molnar, 2010). Thus, comparisons of the two orogenic zones can improve our knowledge of the continental collision in general.

A number of hypotheses have been proposed to explain the uplift and deformation of the Tibetan and Iranian plateaus. Investigations of lithospheric structure using seismic waves can help us better understand the processes of continental collision at depth. On regional seismograms, Lg and Sn waves propagate within the crust and uppermost mantle and therefore provide a good measure of attenuation and velocity of seismic shear waves (Kaviani et al., 2015). In the past decades, there are a lot of Lg studies across the Tibetan and Iranian plateaus and very few Sn studies were reported until now.

Temperature and composition are two major causes for subsurface seismic anomalies. Positive temperature anomalies will lead to a reduction in both attenuation and velocity; however, compositional anomalies should not necessarily produce a strong correlation in attenuation and velocity (Kaviani et al., 2015). As a result, by combining velocity and attenuation structure we can distinguish between compositional and temperature anomalies.

In this study I present, for the first time, high resolution Sn Q models over The Tibetan and Iranian plateaus. The results are obtained from the analysis of two large data sets in the eastern Tibetan plateau and northern Middle East.

1.1 Seismic Attenuation

Seismic attenuation occurs when the energy of seismic wave decreases in its propagation due to “internal friction” such as movements along mineral dislocations or shear heating at grain boundaries (Lay and Wallace, 1995). The loss of elastic energy that a seismic wave experiences when it traverses an imperfectly elastic material is commonly described by the inverse of quality factor (Q) which can be defined as

$$\frac{1}{Q} = \frac{\Delta E}{2\pi E_{max}} \quad (1.1)$$

where ΔE is the amount of energy lost per cycle and E_{max} is the total amount of elastic energy stored per unit volume per cycle (Knopoff, 1964; Jackson et al., 1970; Mitchell, 1995). Figure 1.1 shows two seismograms in which the distances between the event and two stations are almost the same. However, the seismic phase Sn attenuates severely along the MLAZ raypath compared with the MAZI raypath indicating a relative low Q path for MLAZ and a high Q path for MAZI.

Figure 1.1 (a) Map shows the locations of an earthquake (red circle) and stations (blue triangles). (b) Seismograms recorded at Kandilli stations MAZI and MLAZ from a same event. The epicentral distance to MAZI (along a high Q path) is 1353 km and the epicentral distance to MLAZ (along a low Q path) is 1354 km.

Seismic attenuation is studied both in laboratory and seismic wave measurements. The laboratory studies are important because they can give us important constraints on what factors affect attenuation. In the case of recorded seismic waves, we don't know the causes, but in laboratory, we can demonstrate them empirically. Most laboratory measurements of seismic properties of rocks are performed as a function of temperature and pressure (Sato et al., 1989). Previous laboratory studies indicate temperature, pressure and composition are three major parameters that constrain mechanical properties of the Earth. However, grain boundary damping is the origin of attenuation in the upper mantle (Sato et al., 1989). Grain boundary relaxation and high temperature background are possible candidates for the mechanism of grain boundary damping in the peridotite sample (Jackson et al., 1970). On the other hand, seismic attenuation is also studied by seismic wave measurements. A number a seismic wave measurements (seismic velocity and amplitude) are performed on body waves, surface waves and guided waves. In this study, I will focus on seismic wave measurements using seismic amplitude.

The amplitude of seismic waves can be written as

$$A(f,d)=S(f)G(d)I(f)R(f)Q(f,d) \quad (1.2)$$

where $S(f)$ is the source term including source excitation function and radiation pattern, $G(d)$ is the geometrical spreading function, $I(f)$ is the instrument response, $R(f)$ is the site response and $Q(f,d)$ is the attenuation term.

The apparent attenuation of elastic waves may result from geometrical effects (refraction, reflection and scattering) or from actual loss mechanisms (damped resonance, static hysteresis, relaxation and viscosity) (Jackson et al., 1970).

Geometrical spreading occurs when seismic energy spreads out from a source as a spherical wavefront. The geometrical spreading function in equation 1.2 can be written as

$$G(d) = d^{-m} \quad (1.3)$$

A geometrical spreading factor (m) of 1.0 is used for Sn waves.

Scattering attenuation results from the heterogeneity of the Earth (grains, mineral boundaries, pore edges and cracks); it is usually observed at tectonic active regions like fault zones and regions with Moho steps and it is frequency dependent.

Intrinsic attenuation is due to the anelasticity of the Earth and is frequency independent. It can be used to infer the rheology and temperature of the lithosphere and it is defined as

$$Q_{intrinsic} = e^{\frac{-\pi f d}{Qv}} \quad (1.4)$$

where f denotes frequency, d denotes distance, v denotes velocity and Q denotes quality factor.

The apparent Q calculated in my study is the combination of intrinsic and scattering attenuation. It is important to note that the frequency dependence is substantially different between these two attenuation mechanisms.

1.2 S_n

In regional seismograms ($3^\circ < \text{epicentral distance} < 15^\circ$), there are four major seismic phases: P_n , P_g , S_n and L_g (Figure 1.2). They are guided waves that travel in particular channels within the lithosphere; P_g and L_g travel within the crust and P_n and S_n travel in the uppermost mantle.

Figure 1.2 (a) A typical regional seismogram recording P_n , P_g , S_n and L_g . (b) A diagram showing ray paths of regional seismic phases. (Bao, 2011)

Sn is a regional seismic shear wave, with a predominant period between 0.5 to 2 seconds and an average group velocity of 4.5 km/s, trapped between the Moho and low velocity zone in the uppermost mantle (Molnar and Oliver, 1969).

Sn is usually modeled in terms of normal mode propagation of Love waves (Stephens and Isacks, 1977) or treated as multiply reflected shear waves trapped within the uppermost mantle. There are a number of factors that can influence the propagation of Sn such as lateral velocity variation due to composition heterogeneity, temperature and melt, and thin or missing mantle lid in tectonically active regions.

The attenuation of Sn is also assumed as a proxy to understand the rheology of the lithosphere mantle (Mitchell, 1995; Kaviani et al., 2015). Its strong sensitivity to temperature variations can help us to image the thermal tectonic features within the uppermost mantle such as hot spots, mantle plumes and subduction zones. It thus can be used to help constrain regional tectonic models in combining with seismic velocity results.

Sn has been used for velocity and attenuation studies in the uppermost mantle worldwide. Pei et al. (2007) obtained a Sn velocity image of the uppermost mantle beneath China. Barron and Priestley (2009) imaged frequency dependent propagation efficiency of Sn over the Tibetan Plateau. In North America, Beghoul et al. (1993) found that Sn efficiently propagates beneath Great Plains while no Sn is observed beneath the Basin and Range and Colorado Plateau. Buehler and Shearer (2013) constructed a high

resolution image of Sn propagation beneath the western United States. In Europe, Diaz et al. (2013) imaged Sn velocity variation beneath the Euro-Mediterranean region. Sandvol et al. (2001) obtained the propagation efficiency of Sn over most of the Middle-East. Gok et al. (2003) constructed a high resolution image of Sn propagation beneath the Anatolian and Iranian plateau. Al-Damegh et al. (2004) also improved the Sn propagation image beneath the Arabian Plate.

1.3 Objective

Investigation of attenuation can be used in earthquake magnitude calibration, seismic hazard assessment and seismic monitoring (Pasyanos et al., 2009; Kaviani et al., 2015). In this study, I use Sn attenuation to infer properties of the uppermost mantle in active continental orogens.

Temperature and composition are two major causes for subsurface seismic anomalies. Positive temperature anomalies will lead to a reduction in both attenuation and velocity; however, compositional anomalies should not necessarily produce a strong correlation in attenuation and velocity. As a result, by combining velocity and attenuation structure we can distinguish between compositional and temperature anomalies.

Laboratory experiments have shown that seismic attenuation depends very strongly on temperature (Sato et al., 1989). Regional seismic phases travel in the crust (Pg and Lg) and uppermost mantle (Pn and Sn). This suggests that measuring regional phase Q is an approach to investigate the temperature anomalies within the lithosphere.

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Chapter 2: Methodology

2.1 Efficiency Tomography

Efficiency tomography was developed by Sandvol et al. (2001) by constructing regional seismic phase (Lg and Sn) attenuation models beneath the Middle East. This method uses Sn propagation efficiencies as input data and inverts for the reciprocal of the extinction path length which corresponds to the distance that Sn will extinct in its propagation.

Because long period Sn can propagate through the continental lithosphere, I applied a band pass filter (0.1 Hz - 0.5 Hz) to my data to characterize Sn efficiencies. I classified the efficiency of the seismograms into three groups (efficient, inefficient and blocked) visually. If the seismogram shows no evidence of an Sn wavetrain, I characterize it as “blocked Sn” (Figure. 2.1). If an Sn wavetrain is obvious, I designate it as “efficient Sn”. If there is ambiguous signal that potentially could be an Sn signal, it is classified as “inefficient Sn”.

In order to map Sn efficiency, I used the method of Sandvol et al. (2001) and Al-Damegh et al. (2004). The Sn amplitude in frequency domain can be rewritten from equation 1.2 as

$$A_{ij} = S_i G_{ij} I_j R_j Q_{ij} \quad (2.1)$$

where A_{ij} is the spectral amplitude, S_i is the source term for the i^{th} event, G_{ij} is the geometrical spreading term, I_j is the instrument response for the j^{th} station, R_j is the site term for the j^{th} station and Q_{ij} is the attenuation term. Substituting equation 1.4 for the attenuation term, I obtain

$$A_{ij} = S_i G_{ij} I_j R_j e^{\frac{-\pi f d_{ij}}{Qv}} \quad (2.2)$$

where f is the frequency, d_{ij} is the raypath length, Q is the quality factor and v is the Sn velocity. Taking the natural logarithm of equation 2.2, neglecting the geometrical spreading term and instrument response, discretizing the attenuation coefficient, I obtain

$$\ln \left(\frac{A_{ij}^0}{A_{ij}} \right) = \frac{\pi f}{v} \sum_k m_k l_{ijk} \quad (2.3)$$

where

$$A_{ij}^0 = S_i R_j \quad (2.4)$$

Then I use the wave propagation efficiencies instead of amplitude ratios

$$E_{ij} = \frac{\pi f}{v} \sum_k m_k l_{ijk} \quad (2.5)$$

where E_{ij} is the wave propagation efficiency. I then apply the LSQR algorithm (Paige and Saunders, 1982) to solve for the model parameter m_k which is the reciprocal of the extinction path length.

Figure 2.1 (a) Map shows the locations of earthquake (red circle) and stations (blue triangles). (b) Seismograms recorded at Kandilli stations MAZI, MLAZ and SENK showing Sn efficiency.

2.2 Q Tomography

2.2.1 Two Station Method (TSM)

The Two Station Method (TSM) was presented by Tsai and Aki (1969) and improved by Xie and Mitchell (1990b) (Bao, 2011). Figure 2.2 shows the geometry of Two Station paths and figure 2.3 shows an example of Q measurement using the TSM.

Figure 2.2 (a) An ideal geometry for TSM. (b) A more practical geometry for TSM. I denote the two stations as station i and j with spectra A_i and A_j , respectively. The epicentral distances of stations i and j are d_i and d_j , the inter-station distance is d_{ij} , and the azimuthal difference between stations i and j is $\delta\theta$.

The spectral amplitude ratio is

$$\frac{A_i}{A_j} = \frac{S G_i I_i R_i Q_i}{S G_j I_j R_j Q_j} \quad (2.6)$$

Substituting equation 1.4 for the attenuation term I obtain

$$\frac{A_i}{A_j} = \frac{S G_i I_i R_i}{S G_j I_j R_j} e^{\frac{\pi f d_j}{Q_j v_j} - \frac{\pi f d_i}{Q_i v_i}} \quad (2.7)$$

Assuming the velocity structure is one-dimensional and apparent Q values are identical at station i and j and assuming that the variation in R is negligible, this becomes

$$\frac{A_i}{A_j} = \frac{G_i I_i}{G_j I_j} e^{\frac{\pi f}{Q v} (d_j - d_i)} \quad (2.8)$$

Substituting the attenuation term with equation 1.3 I obtain

$$\frac{A_i}{A_j} = \frac{d_i^{-m} I_i}{d_j^{-m} I_j} e^{\frac{\pi f}{Q v} (d_j - d_i)} \quad (2.9)$$

where m is the spreading exponent.

The inter-station Q values can be derived as

$$\frac{1}{Q} = \frac{v}{\pi f (d_j - d_i)} \ln \left(\frac{A_i d_i^m I_j}{A_j d_j^m I_i} \right) \quad (2.10)$$

The frequency dependent Q values are

$$\frac{1}{Q(f)} = \frac{v}{\pi f (d_j - d_i)} \ln \left(\frac{A_i(f) d_i^m I_j(f)}{A_j(f) d_j^m I_i(f)} \right) \quad (2.11)$$

The frequency dependence of Q is also assumed in form of a power-law equation

$$Q(f) = Q_0 f^\eta \quad (2.12)$$

where η is the frequency dependence factor.

Substituting equation 2.12 with equation 2.11 and taking the natural logarithm of equation 2.11 I obtain

$$(1 - \eta) \ln f - \ln Q_0 = \ln \left[\frac{v}{\pi(d_j - d_i)} \ln \left(\frac{A_i(f)}{A_j(f)} \frac{d_i^m}{d_j^m} \frac{I_j(f)}{I_i(f)} \right) \right] \quad (2.13)$$

By applying a linear regression curve fitting of my measurements, the frequency dependence factor η can be obtained.

Figure 2.3 Example of Q measurement using the TSM method. On the geographic map (bottom left), the red circle denotes earthquake and green triangles denote stations. The bottom right graph shows the Sn amplitude spectral ratio versus frequency that is used to estimate the Sn Q_0 and η values. The three top graphs depict the Sn spectra at the near and far stations (blue and green), the Sn and noise spectra at near station (blue and red), the Sn and noise spectra at far station (green and red).

2.2.2 Reverse Two Station Method (RTM)

The Reverse Two Station Method (RTM) was initially suggested by Chun et al. (1987) as a way to efficiently avoid the effects of neglected R terms and inaccurate I terms in the TSM by involving one more event. Figure 2.4 shows the geometry of Reverse Two Station paths and figure 2.5 shows an example of Q measurement using the RTM.

Figure 2.4 (a) An ideal geometry for RTM. (b) A more practical geometry for RTM. I use A_{ai} , A_{aj} , A_{bi} and A_{bj} , to denote spectral amplitudes of S_n recorded at station i and j for events a and b , and d_{ai} , d_{aj} , d_{bi} and d_{bj} the corresponding distances, $\delta\theta_a$ and $\delta\theta_b$ denote the azimuthal difference between station i and j to events a and b .

The spectral amplitude ratios are

$$\frac{A_{ai}}{A_{aj}} = \frac{S_a}{S_a} \frac{G_{ai}}{G_{aj}} \frac{I_i}{I_j} \frac{R_i}{R_j} e^{\frac{\pi f d_{aj}}{Q_j v_j} - \frac{\pi f d_{ai}}{Q_i v_i}} \quad (2.14)$$

$$\frac{A_{bi}}{A_{bj}} = \frac{S_b}{S_b} \frac{G_{bi}}{G_{bj}} \frac{I_i}{I_j} \frac{R_i}{R_j} e^{\frac{\pi f d_{bj}}{Q_j v_j} - \frac{\pi f d_{bi}}{Q_i v_i}} \quad (2.15)$$

Like the TSM, this method assumes that the velocity structure is one-dimensional and apparent Q values are identical at station i and j. I divide the two ratios and substitute G to obtain

$$\frac{A_{ai} A_{bj}}{A_{aj} A_{bi}} = \left(\frac{d_{ai} d_{bj}}{d_{aj} d_{bi}} \right)^{-m} e^{\frac{\pi f}{Q v} (d_{aj} - d_{ai} - d_{bj} + d_{bi})} \quad (2.16)$$

Therefore, the inter-station apparent Q values can be derived as

$$\frac{1}{Q} = \frac{v}{\pi f (d_{aj} - d_{ai} - d_{bj} + d_{bi})} \ln \left[\frac{A_{ai} A_{bj}}{A_{aj} A_{bi}} \left(\frac{d_{ai} d_{bj}}{d_{aj} d_{bi}} \right)^m \right] \quad (2.17)$$

The frequency dependent Q values are

$$\frac{1}{Q(f)} = \frac{v}{\pi f (d_{aj} - d_{ai} - d_{bj} + d_{bi})} \ln \left[\frac{A_{ai}(f) A_{bj}(f)}{A_{aj}(f) A_{bi}(f)} \left(\frac{d_{ai} d_{bj}}{d_{aj} d_{bi}} \right)^m \right] \quad (2.18)$$

Figure 2.5 Example of Q measurement using the RTM method. The red circle denotes the earthquake and green triangles denote stations. The bottom right graph shows the Sn amplitude spectral ratio versus frequency that is used to estimate the Sn Q_0 and η values. The two top graphs depict the Sn spectra (dark blue and dark green) and noise spectra (light blue and light green) at near and far stations from earthquakes.

2.2.3 Seismic Attenuation Tomography

Seismic tomography is a data inference technique that exploits information contained in seismic records to constrain 2D and 3D models of the Earth's interior (Rawlinson et al., 2010). It was first proposed by Keiiti Aki (1976) by constructing 3D P wave velocity variation beneath California from local earthquakes. In the past four decades, this technique has been developed and include local, regional and global studies. The basic idea of seismic tomography can be written as

$$\mathbf{d} = \mathbf{G}\mathbf{m} \quad (2.19)$$

where $\mathbf{d}$ denotes data, $\mathbf{m}$ denotes the model parameters, and $\mathbf{G}$ is a large sparse matrix recording the geometry of raypaths. We can predict $\mathbf{d}$ for a given model $\mathbf{m}$, the seismic tomography problems solve $\mathbf{m}$ such that $\mathbf{d}$ explains the data observations $\mathbf{d}_{\text{obs}}$.

Seismic attenuation tomography is mainly used to map the lateral variation of Q using the amplitude of seismic waves. The advantage of attenuation tomography over other tomographic methods is its strong sensitivity to temperature variations and therefore its potential to image hot spots, mantle plumes and subduction zone (Rawlinson et al., 2010). The technique used in my study was originally developed by Xie and Mitchell (1990a) in order to construct a 2D Lg coda Q model beneath Africa. If there are a total of N paths in my study area then the distance and Q along the n^{th} path ($n=1,2 \dots N$) are denoted by d_n and Q_n . I will divide the study area into identical cells and assume that the Q in each cell is constant. Denoting the length over which the n^{th} path crosses the m^{th}

cell as d_{mn} and the total number of cells crossed by the n^{th} path as M , equation 1.4 can be rewritten as

$$Q_{intrinsic} = e^{\frac{-\pi f}{v} \frac{d_n}{Q_n}} = e^{\frac{-\pi f}{v} \sum_{m=1}^M \frac{d_{mn}}{Q_m}} \quad (2.20)$$

Comparing with equation 2.19, I obtain

$$\frac{d_n}{Q_n} = \sum_{m=1}^M \frac{d_{mn}}{Q_m} \quad (2.21)$$

where data $\mathbf{d}$ is the vector of d_n/Q_n , the model parameter $\mathbf{m}$ is the vector of $1/Q_m$, and the $\mathbf{G}$ is a matrix of ray lengths storing all non-zero d_{mn} values.

Equation 2.21 can be solved using the LSQR algorithm presented by Paige and Saunders (1982).

A checkerboard test is commonly used to estimate the resolution of an attenuation model. This method uses an input model consisting of an alternating pattern of high and low Q values to generate data with the same source-receiver geometry as the observational experiment. The ability to recover the input model can be used as a measure of solution robustness (Rawlinson et al., 2010). In figure 2.6, the checkerboard pattern can be well recovered with good ray coverage. In cases where ray coverage is not dense or with similar azimuth, the recovered model can leak or show severe smearing.

Figure 2.6 An example of checkerboard test. (a) Synthetic model with ray coverage, black triangles denote stations. (b) Retrieved model from the inversion of the synthetic data of (a).

A bootstrap test is used to further estimate the stability of the resulting attenuation model. In this statistical analysis technique, the data are randomly resampled until there are the same number of observations as in the original data, and then the resampled data are inverted. I repeated this procedure 100 times and the resulting set of bootstrap solutions was used to estimate the variance of the solutions. Typically, small uncertainty indicate a stable solution and high uncertainty usually coincide with the regions of poor ray coverage (Kaviani et al., 2015).

In order to estimate the frequency dependence of my solution, I also tomographically mapped the frequency dependence factor η . However, because of the non-linear relationship between η and the distance along the ray path in equation 2.13, I cannot use the same method as for Q to map the lateral variation in η . Instead, I measured η in each cell based on the power law assumption. The equation 2.12 can be rewritten as

$$\ln Q(f) = \ln Q_0 + \eta \ln f \quad (2.22)$$

Using Q values at three frequency band (0.5 Hz, 1 Hz and 2 Hz), I can linearly fit a straight line, while the slope is η and the intercept is $\ln Q_0$. Then I can use η values to obtain the frequency dependence factor model. Typically, low η indicate intrinsic attenuation is the dominant mechanism and high η indicate scattering attenuation is the dominant mechanism. As discussed in chapter 1, scattering attenuation is resulted from the heterogeneity of the Earth and intrinsic attenuation is due to the anelasticity of the Earth.

2.3 References

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Chapter 3: Sn Attenuation in the Eastern Tibetan Plateau

3.1 Background

The Tibetan Plateau (Figure 3.1) is the largest and highest plateau on Earth, with an average elevation of 5000m (Tapponnier et al., 2001; Kind et al., 2002; Yue et al., 2012). It was formed by the continental collision between Indian and Eurasian plates that has been going on at least since ~50Ma (Molnar and Tapponnier, 1975; Yin and Harrison, 2000). Previous studies have proposed four geodynamic models trying to explain the uplift and deformation of the plateau: (1) the wholesale underthrusting model (Ni and Barazangi, 1984; Tilmann et al., 2003), (2) uniform thickening and shorting model (Dewey and Burke, 1973; Houseman and England, 1986), (3) block extrusion model (Tapponnier et al., 1982; Tapponnier et al., 2001), (4) channel flow model (Royden et al., 1997; Clark and Royden, 2000; Clark et al., 2005). It is important to note that these models are not mutually exclusive. The underthrusting model proposes the Indian lithosphere is underlying much of the southern Tibetan Plateau. The uniform thickening and shorting model suggests that the convergence of continental collision is accommodated by lithospheric and crustal thickening and horizontal shortening. The block extrusion model involves the northward convergence of Indian plate being accommodated by the localization of deformation along major strike slip faults. The channel flow model suggests the surface and upper crustal deformation on the Tibetan Plateau is produced by the motion of weak lower crust; however, the mechanism is still under debate. After comparing these models, I have found that understanding the seismic structure of the lithosphere of the Tibetan Plateau can help to reconcile which of these models, or possibly a combination, might best explain what I observe today.

Figure 3.1 Simplified tectonic map in the Tibetan Plateau. The red triangles represent Indepth IV network; the black triangles represent Namche Barwa network; the blue triangles represent MIT-China network; the green triangles represent NETS network; the yellow triangles represent ASCENT network. TB: Tarim basin; QB: Qaidam basin; SB: Sichuan basin; QL: Qilian mountain; ATF: Altyn Tagh Fault; KF: Kunlun fault; JS: Jinsha suture; BNS: Bangong-Nujiang suture; IYS: Indus-Yarlung suture; MFT: Main frontal thrust; IP: Indian plate; SGFB: Songpan-Ganzi block; QTB: Qiangtang block; LB: Lhasa block. Pink shades represent Cenozoic volcanism.

The major tectonic features in the eastern Tibetan Plateau from north to south are Qaidam Basin, Songpan-Ganzi block, Qiangtang block, Lhasa block and Himalaya that are separated by Kunlun fault, Jinsha suture, Bangong-Nujiang suture and Indus-Yarlung suture respectively (Yue et al., 2012). Furthermore, the Songpan-Ganzi block and Qiangtang block are characterized by Cenozoic volcanism in the eastern Tibetan Plateau (Ding et al., 2003).

During the past decades, there have been numerous geophysical studies on the Tibetan Plateau. As we know, the northward subduction of Indian plate beneath Eurasian plate may play a key role in the evolution of Tibetan Plateau. A number of seismic images have strongly suggested the existence of the underthrusting Indian continental lithosphere (UICL) beneath southern Tibet. High P and S wave velocities are observed beneath portions of southern Tibet (Wittlinger et al., 1996; Tilmann et al., 2003; Huang and Zhao, 2006; Li et al., 2008; Liang et al., 2012; Liang et al., 2016), high Pn velocity (Liang and Song, 2006) and efficient Sn phases (Barazangi and Ni, 1982; McNamara and Owens, 1995; Rapine and Ni, 1997) are also observed nearly exclusively for paths in southern Tibet. Fast surface wave velocity anomalies (Ceylan et al., 2012) are also observed.

Seismic tomographic models show east-west lateral variations in the geometry and thickness of the UICL (Li et al., 2008; Ceylan et al., 2012; Liang et al., 2012; Liang et al., 2016). In western Tibet, the UICL reaches as far as Tarim Basin. In the central Tibet, the UICL stops at the Bangong-Nujiang suture; in the eastern Tibet, the UICL extends no further than the Indus-Yarlung suture. Low velocity zones are also found within the UICL that are explained by the fragmentation of the UICL (Li et al., 2008; Ceylan et al., 2012; Liang et al., 2012; Liang et al., 2016), however, it is possible these anomalies are due to compositional variations within the UICL.

In the Songpan-Ganzi block and Qiangtang block, Low P and S wave velocities are observed beneath northern Tibet (Wittlinger et al., 1996; Tilmann et al., 2003; Huang and Zhao, 2006; Li et al., 2008; Liang et al., 2012; Liang et al., 2016). Low Pn velocity (Liang and Song, 2006) and blocked Sn phases (Barazangi and Ni, 1982; McNamara and Owens, 1995; Rapine and Ni, 1997) are observed in this region. Slow surface wave velocity anomalies (Ceylan et al., 2012) are also observed. Three main mechanisms are proposed to explain the origin of the low velocity zone: strain heating (e.g. Nabelek et al., 2010; Ceylan et al., 2012), small scale convection induced by the subducted Indian lithosphere (Tilmann et al., 2003) and delamination (Molnar et al., 1993).

The Moho depth beneath the Tibetan Plateau shallows from south (~80km) to north (~60km) with Moho offsets along Kunlun fault and Jinsha sutures. The lithosphere-asthenosphere boundary (LAB) in southern Tibet is ~200km and ~160km in the northern Tibet, with a ~50km gap between the Indian and Eurasian lithosphere (Zhu and Helmberger, 1998; Kind et al., 2002; Kumar et al., 2006; Zhao et al., 2010; Yue et al., 2012). In addition, Kind et al. (2002) proposes the Eurasian lithosphere is subducting southward beneath the northern Tibet; however other studies fail to image this structure (Li et al., 2008; Ceylan et al., 2012; Yue et al., 2012; Liang et al., 2012).

There have been a large number of studies on the propagation of the regional seismic phase Lg across the Tibetan Plateau; however, there have been far fewer studies of Sn because of the extensive blockage of this phase in tectonically active regions. There have been a few studies of Sn blockage across the Tibetan Plateau.

(1) Molnar and Oliver (1969) studied the propagation of Sn on a worldwide scale using WWSSN seismic ray paths. They found the propagation of Sn is very efficient in stable regions like continental shields and deep-ocean basins indicating that a high-Q uppermost mantle is continuous, but the propagation of Sn is very inefficient across the concave side of most arcs and across the crests of the mid-ocean ridge system implying that the high-Q upper layer of the mantle is discontinuous. They also observed efficient propagation of Sn across the Indian shield and inefficient propagation of Sn across the Tibetan Plateau.

(2) Barazangi and Ni (1982, 1983) qualitatively classified the Sn amplitudes relative to those of Pn and found that high frequency Sn waves propagate efficiently across most of the Tibetan Plateau, except the north-central part of the Tibetan Plateau (the Qiangtang terrane). They also concluded their observations support the shallow-angle underthrusting model instead of crustal shortening and thickening models

(3) McNamara and Owens (1995) studied the propagation of Sn across the central Tibetan Plateau by determining Sn propagation efficiency by comparing the maximum tangential amplitude of Sn to the average vertical amplitude of the P wave coda. They found that the Sn propagation is frequency dependent. At high frequency, Sn is severely attenuated across the northern plateau, but Sn at longer periods is able to propagate throughout the entire plateau.

(4) Rapine and Ni (1997) studied high frequency Sn propagation in China using the same method of McNamara and Owens (1995). They found efficient Sn propagation in Tarim platform and southern Tibet and inefficient Sn propagation in the north central Tibet which can be explained by partial melting.

(5) Barron and Priestley (2009) measured the ratio of the Sn amplitude to the Pg coda amplitude as the efficiency of Sn propagation and analyzed data at different frequency band. They found the similar observation with MacNara and Owens (1995) that at low frequency the propagation of Sn is efficient over the entire plateau while at higher frequencies the propagation becomes significantly inefficient in the northern plateau and the margin of the area of inefficient propagation migrates further southwards. They also indicated that their observation supported models involving large-scale underthrusting while models involving delamination is not reliable.

In this study, I have collected a large data set in the eastern Tibetan Plateau and developed both Sn efficiency tomography at low frequency band (0.1Hz - 0.5Hz) and Sn Q tomography at different frequencies (0.5Hz, 1Hz, 2Hz). My results show high attenuation across the Songpan-Ganzi block and Qiangtang block and lateral heterogeneity within the southern Tibet. These observations can greatly improve the knowledge of Sn attenuation across the eastern Tibetan Plateau.

3.2 Data Collection and Processing

The data analyzed in my study are collected from 168 permanent and temporary broadband stations from 5 seismic networks: Indepth IV, Namche Barwa network, MIT-China network, NETS network and ASCENT network (Figure 3.1). 350 regional events are used in my study with magnitudes greater than 4.5 and hypocentral depths less than the Moho. The epicentral distances are limited between 3 ° to 15 °. 5737 seismograms (Figure 3.2) are picked with good signal noise ratio and clear Pn arrivals.

I have applied a band pass filter (0.1Hz - 0.5Hz) to maximize signal to noise in order to characterize the efficiencies of Sn phases (Figure 3.2). Then I used the method of Sandvol et al. (2001) and Al-Damegh et al. (2004) to tomographically map regions of Sn blockage (Figure 3.3a). I also perform synthetic tests (Figure 3.3b) to assess the resolution in the Sn efficiency tomography.

In addition to looking at Sn blockage, I used TSM and RTM to measure Sn Q at 0.5Hz (0.1Hz-1.0Hz), 1Hz (0.5Hz-1.5Hz) and 2Hz (1.5Hz-3.0Hz). Then I applied the LSQR algorithm to tomographically map the lateral variations of Q in the eastern Tibetan Plateau (Figure 3.4 and Figure 3.8). I also performed checkerboard tests (Figure 3.6) and bootstrap error estimations (Figure 3.7) to examine the resolution and stability of the Sn attenuation tomography.

3.3 Results

3.3.1 Sn Efficiency Tomography

The Sn efficiency tomography is shown in figure 3.3a. The most prominent feature is that Sn propagates efficiently in the southern Tibet and inefficient or blocked in the northern Tibet. This feature is consistent with previous attenuation studies (Molnar and Oliver, 1969; Barazangi and Ni, 1982; McNamara and Owens, 1995; Rapine and Ni, 1997; Barron and Priestley, 2009). Efficient Sn propagation in the southern Tibet also correlates well with fast Pn waves (Liang and Song, 2006) and a fast body wave velocity anomaly (Liang et al., 2012) and it can be interpreted as the underthrusting Indian

continental lithosphere (UICL). Further north, a blocked Sn zone is observed (BSZ1) within the UICL. Some seismic velocity studies (Figure 3.11) also confirm the UICL is not uniform and low velocity zones are observed within the UICL (Ceylan et al., 2012; Liang et al., 2012; Liang et al., 2016). It is notable that an efficient Sn zone (ESZ) is observed at $\sim 91^{\circ}\text{E}$ – 31°N and bounded north by BNS. This feature is consistent with high P wave velocity anomalies (Li et al., 2008) and fast Rayleigh waves in this region (Ceylan et al., 2012). In the QTB and SGFB, Sn is blocked or inefficient and shows east-west lateral variations. The BSZ2 extends southward to BNS and the ISZ extends further south to the IYS. This observation correlates with Pn wave studies (Figure 3.10) very well because they also observe a low velocity zone (LVZ) beneath QTB and SGFT in northern Tibet. This positive correlation extends further south into Indo-China. The Sn blockage in the BSZ2 also coincides with Cenozoic volcanism. In the stable QB, efficient Sn is observed.

I performed a checkerboard test (Figure 3.3b) to assess the resolution of my model by using the same ray geometries that are used for the inversion of actual data. A 300km blockage path length and 10% noise was assigned to my Sn synthetic test. The checkerboard test shows my model can be well resolved by $4^{\circ} \times 4^{\circ}$ cells covering the eastern Tibetan Plateau and it is good enough to show first order features. The Sn smearing is observed at the edges of my model where ray paths are confined in the same direction. The Sn leakage is also observed (for example at Sichuan Basin) because of poor ray coverage.

Figure 3.2 A map of ray paths with Sn propagation efficiencies

Figure 3.3 (a) Map showing Sn efficiency tomography results. Circles represent seismic attenuation anomalies. (b) Resolution test using a synthetic checkerboard for extinction

path length of 300km. Black lines are major tectonic boundaries. Red triangle represents volcano.

3.3.2 Sn Q_0 Tomography

The TSM Sn Q_0 map is shown in figure 3.4a. From north to south, I observe high Q (~ 500) values in the QB. In the SGFB and QTB, I observe relative low Q (~ 300) values. Another high Q value zone (HQZ) is observed between 91°E - 95°E and bounded north by BNS. Further southward in the LB, a low Q value zone (LQZ) is observed.

The RTM Sn Q_0 map is shown in figure 3.4b. As discussed in chapter 2, the RTM Q measurements are more accurate because it is not affected by inaccurate instrument responses or relative differences in site terms that may bias the TSM Q measurements; however it dramatically reduces the ray coverage due to more strict recording geometry. As a result, I use RTM measurements to verify the accuracy of the TSM measurements. The RTM Sn Q_0 map (Figure 3.4b) shows very similar features with those observed on TSM Sn Q_0 map (Figure 3.4a). In the QB, I observe high Q (~ 500) values. In the SGFB, I observe low Q (~ 150) values. A different pattern between these two models is that in RTM Sn Q_0 map, I observe a relative high Q value zone in the QTB while relative low Q values in this region are observed in the TSM Sn Q_0 map; however, this is very likely caused by artificial effect due to the poor ray coverage in this area (Figure 3.6b). In order to obtain more reliable RTM Sn Q_0 model in the Tibet, I have added results from Pan J. J. (2016) to my model (Figure 3.5). I observe high Q (~ 500) values in the QB, QL and SB. I also observe low Q (~ 150) values in the SGFB. Further south, a high Q value zone is

observed along the BNS. A relative low Q (300) value zone is also observed in LB which corresponds with the BSZ1 in figure 3.3a.

I also performed checkerboard tests to examine the resolution of my models. The checkerboard test illustrates that Sn Q_0 anomalies with a size of $2^\circ \times 2^\circ$ can be well resolved by the TSM ray coverage. Figure 3.6a shows TSM Sn Q_0 model is well resolved in most of my study area. The synthetic anomalies are smeared and distorted in regions lacking dense ray coverage such as the eastern part of the model. Figure 3.6b shows the RTM Sn Q_0 map is not well resolved because of poor ray coverage.

I performed a bootstrap resampling technique to test the stability of my tomographic models (Figure 3.7). In this test, I randomly resample my data until I reach the same number of observations as in my original data and then I invert the resampled data for a bootstrap model. I repeated this procedure for 100 times and the resulting set of bootstrap solutions is used to estimate the variance of the solutions. The result of the bootstrap test for the TSM Sn Q_0 model is shown in figure 3.7a. It shows an uncertainty less than 50 within the Tibetan plateau and high uncertainty at the edge of the model. The result of the bootstrap test for the RTM Sn Q_0 model (Figure 3.7b) shows low uncertainty within most of the Tibetan plateau.

I also construct Sn Q maps at different frequencies (0.5HZ, 1HZ, 2HZ) both using TSM and RTM (Figure 3.8). The main features are consistent with the Sn Q_0 maps. At low frequency (0.5HZ), the HQZ extends further northward beneath QTB. At high

frequency (2HZ), the HQZ vanishes. This feature is consistent with Barron and Priestley (2009) and can be explained by low frequency Sn can dive deeper in the uppermost mantle while high frequency Sn travels more like head wave beneath the Moho. There are discrepancies at the edges of these models that may be caused by poor ray coverages.

The frequency dependent factor η map of TSM is shown in figure 3.9a. Across most of my model, the frequency dependent factor is low to normal (0-0.5) indicating the intrinsic attenuation is the dominant mechanism. In figure 3.9b, I observe relative high η along the Kunlun fault indicating that scattering attenuation is the dominant mechanism.

3.3.3 A Comparison of Efficiency and Q Tomographys

Comparing the results of Sn efficient tomography and Sn Q tomography, the Sn efficient tomography can resolve more features than the Sn Q tomography because the Sn Q tomography requires more strict geometry of recording geometry and incorporates far more information than the efficiency tomography. Recall that the efficiency tomography essentially reduces Sn amplitudes to a binary data set; either blocked or efficient. Still the main features between these models are very similar. I observe both efficient Sn and high Q values in the QB indicating a cold and strong lithosphere. I also observe blocked Sn and low Q values in the SGFB and QTB that suggests a hot and weak lithosphere. Another important feature is that at $\sim 91^{\circ}\text{E}$ $\sim 31^{\circ}\text{N}$ I observe efficient Sn and high Q values. At LB, I both observe blocked Sn and a low Q value zone. There are, however, clear discrepancies between these models. The Sn efficient tomography shows strong lateral variations in the SGFB and QTB while I do not see this feature in the Sn Q tomography.

This may be caused by the poor ray coverage in the Sn Q tomography and thus I should be skeptical about this part of the Sn Q model.

Figure 3.4 (a) Two Station (TSM) Sn Q₀ tomographic map. Circles represent seismic attenuation anomalies. (b) Reverse Two Station (RTM) Sn Q₀ tomographic map. Black lines are major tectonic boundaries.

Figure 3.5 Reverse Two Station (RTM) S_n Q_0 tomographic map based on Pan J. J. 2016. Black lines are major tectonic boundaries.

Figure 3.6 Checkerboard tests for the Two Station (TSM) $\text{Sn } Q_0$ tomographic map (a) and Reverse Two Station (RTM) $\text{Sn } Q_0$ tomographic map (b). Black lines are major tectonic boundaries.

Figure 3.7 Map showing the standard deviations computed by bootstrap analysis for the TSM (a) and RTM (b) $\text{Sn } Q_0$ tomographic models. Black lines are major tectonic boundaries.

Figure 3.8 Two Station (TSM) Sn tomographic map at 0.5Hz (a), 1Hz (c), 2Hz (e). Reverse Two Station (RTM) Sn tomographic map at 0.5Hz (b), 1Hz (d), 2Hz (f). Black lines are major tectonic boundaries.

Figure 3.9 Tomographic map of frequency-dependent (η) of (a) Two Station Method Sn Q measurements and (b) Two Station Method Sn Q measurements. Black lines are major tectonic boundaries.

Figure 3.10 Pn wave velocities. The gray lines show major tectonic features of Tibet. (Liang and Song, 2006)

Figure 3.11 Shear wave velocities for depth of 100km. The gray lines show major tectonic features of Tibet. (Ceylan et al., 2012)

3.4 Discussion

3.4.1 The Geometry of UICL

Previous studies have confirmed the existence of the UICL. However, the geometry of UICL is still under debate. Ni and Barazangi (1984) proposed a wholesale underthrusting Indian lithosphere beneath the south Tibet while Tilmann et al. (2003) shows in the central Tibet, the UICL is subducting vertically at BNS. Yue et al. (2008) fails to see the subvertical downwelling feature, instead, they found north dipping discontinuity beneath BNS and interpreted it as underthrusting Lhasa terrace or remnant oceanic slab. Recently, a number of seismic studies (Li et al., 2008; Ceylan et al., 2012; Liang et al., 2012; Liang et al., 2016) have proposed a fragmented underthrusting Indian lithosphere beneath the south Tibet. It suggests the UICL has a sub-horizontal geometry, and tears laterally into two fragments. The west branch appears to be detached and vertically sinking into the asthenosphere.

I observe efficient S_n in the most southeastern Tibet indicating the UICL. A prominent feature is at $\sim 91^\circ\text{E} \sim 31^\circ\text{N}$ I observe efficient S_n and high Q values. This feature coincides with high P_n velocity anomalies (Liang and Song, 2012) and fast Rayleigh wave (Ceylan et al., 2012). Typically, efficient S_n and high seismic velocities indicate cold and strong lithosphere (Barron et al. 2009). As a result, I interpret this feature as a sinking slab detached from the UICL.

In addition, I also observe high Q zone extends as far as QTB at low frequency; at high frequency, I do not see this feature. According to previous studies (Molnar and Oliver, 1969; Barron et al. 2009), low frequency S_n can dive deeper than high frequency

Sn. It can be interpreted that the UICL is dipping northward rather than vertically subduction.

The Sn efficient map also shows east-west lateral variation of the UICL, A BSZ extends southward to BNS and the ISZ extends further south to the IYS. This observation correlates with Pn wave studies (Liang and Song, 2006) very well. My results show the UICL extends further north in the central Tibet than it extends in east Tibet.

3.4.2 SGFB and QTB

The blocked Sn and low Q values in the SGFB and QTB is consistent with low Pn velocity anomalies (Liang and Song, 2012) and slow Rayleigh wave (Ceylan et al., 2012). It thus suggests a hot and weak lithosphere. The blocked Sn and low Q values also coincide with Cenozoic volcanism. A number of studies (Li et al., 2008; Ceylan et al., 2012; Liang et al., 2012; Liang et al., 2016) propose that the low velocity zone in the SGFB and QTB is caused by upwellings induced by the sinking slab detached from the UICL. My observations can also support this hypothesis.

3.4.3 KL and QB

I observe efficient Sn and high Q values in the QB indicating a cold and strong lithosphere. The KL outlines a boundary between efficient Sn and blocked/inefficient Sn, I also observe high frequency dependent factor η and negative site term north of KL suggesting a strong scattering attenuation. These observations can be caused by the Moho offset along the KL.

3.5 Conclusions

I have collected a large data set in the eastern Tibetan Plateau. Two tomographic techniques have been used to determine the attenuation structure of the uppermost mantle beneath the eastern Tibetan Plateau. My primary conclusions are as follows:

- (1) I observe efficient S_n in the southern Tibet and I interpret this feature as the UICL.
- (2) I also observe efficient S_n with high Q values (~ 450) at $\sim 91^\circ\text{E}$ – 31°N and it suggests a sinking slab detached from the UICL.
- (3) S_n is blocked with relative low Q values (~ 300) across the QTB and SGFB indicating a hot and weak lithosphere. This observation can be caused by upwellings induced by the sinking slab detached from the UICL.
- (4) The lateral heterogeneity of S_n attenuation indicates a complex geometry of the UICL. In the central Tibet, The UICL is dipping northward and can extend further north beneath the QTB.
- (5) I observe efficient S_n and high Q values (~ 500) in the QB indicating a cold and strong lithosphere.

3.6 References

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Chapter 4: Sn Attenuation in the Northern Middle East

4.1 Background

The Turkish-Iranian plateau and Zagros, make up the main tectonic feature of the northern Middle East, is formed as a result of the continental collision between the Arabian and Eurasian plates since Early Cenozoic (23-35Ma) (Hatzfeld and Molnar, 2010). Previous studies (Hatzfeld and Molnar, 2010; Bao et al., 2011; Kaviani et al., 2015) have revealed the Turkish-Iranian plateau and Zagros (Arabian and Eurasian collision) is still at the early stage of continental collision compared to the more mature Tibetan plateau and Himalaya (Indian and Eurasian collision). We can better understand the early stages of continental collision by creating detailed images of the uppermost mantle structure across the Zagros Mountains, Iranian plateau and Anatolian plateau. Three hypotheses (underthrusting, slab break-off and delamination) have been proposed in order to explain the uplift of the plateau. Simmons et al. (2011) suggests the Arabian plate is extensively underthrusting beneath the Iran. Agard et al. (2011) proposes a slab break-off model that the sinking Tethyan slab detached from Arabian margin. Maggi et al. (2005) proposes partial delamination of the lithosphere mantle beneath central Iran as a result of lithosphere thickening. In order to reconcile the debate, it is critical to better understand the lithospheric structure beneath the Turkish-Iranian plateau and Zagros. More specifically images of upper mantle attenuation will help us determine whether seismic anomalies are likely to be a function of temperature or compositional variations.

The major tectonic feature in my research area is Turkish-Iranian plateau (Figure 4.1). A number of studies have indicated a thin lithosphere and warm upper mantle

beneath the plateau including slow Pn velocity (Hearn and Ni, 1994; Amini et al., 2012) and high Sn attenuation (Sandvol et al., 2001; Gok et al., 2003; Gok et al., 2011). Low surface and body wave velocity anomalies (Hafkensheid et al., 2006; Kaviani et al., 2007; Simmons et al., 2011; Maggi and Priestley, 2005; Gok et al., 2007; Gok et al., 2011) in the upper mantle are also observed. The Zagros orogenic belt is comprised of the following sub-parallel domains from SW to NE: The Zagros fold and thrust belt (ZFTB), the Sanandaj-Sirjan zone (SSZ), the Urumieh-Dokhtar Magmatic Arc (UDMA), central Iran and far field deformation areas (Alborz, Kopet Dagh). The ZFTB is formed from buckling and subsequent detachment folding of a 12km thick sediment cover (Mouthereau, 2011). The SSZ is bounded by the Main Zagros Thrust (MZT) to the SW which is considered as the major boundary between Arabia and Iran (Stocklin, 1968; Agard et al., 2005; Paul et al., 2006). The UDMA is characterized by Eocene volcanism (Agard et al., 2011). Central Iran is a relatively flat, aseismic and rigid block (Amini et al., 2012). The regions surrounding the Iranian plateau (Alborz and Kopet Dagh mountains) accommodate part of the Arabian-Eurasian convergence (Agard et al., 2011; Vernant et al., 2004). Another key feature is the southern Caspian Sea basin, a deep (~20km) sedimentary basin, which is thought to be underlain by a rigid oceanic type crust (Allen et al., 2003; Gok et al., 2011; Kaviani et al., 2015). Previous studies (Jackson et al., 2002; Allen et al., 2003; Knapp et al., 2004) suggest it is moving westward relative to Iran and subducting at its northern and western margins.

The average crustal thickness is ~45km beneath ZFTB, very close to the crustal thickness of Arabian platform (45km), suggesting ZTFB has not been thickened much by

collision yet and thus the Arabian-Eurasian collision is now in a very early stage of continental collision (Hatzfeld et al., 2003; Paul et al., 2006; Paul et al., 2010). Further northeast, the crustal thickness thickens to the maximum (~70km) beneath SSZ and then decreases to ~42km beneath the UDMA and Central Iran. A striking feature is the region (SSZ) with maximum crustal thickness is not located at the region (MZT) with highest elevation and maximum negative Bouguer anomaly. In order to interpret this discrepancy, Paul et al. (2006, 2010) proposed that the crust of Zagros is underthrusting the crust of Central Iran along MZT. Alternatively a more buoyant uppermost mantle may help support the higher topography, something we can investigate using measurements of uppermost mantle shear wave attenuation.

In the past decades, there are a few attenuation studies in the Middle East; Sn attenuation studies are even more scarce in my study area. Some significant previous seismic attenuation studies are listed below:

- (1) Sandvol et al. (2001) and Al-Damegh et al. (2004) studied the propagation efficiency of Lg and Sn in the Middle East. They observed inefficient or blocked Lg over the Turkish-Iranian plateau and blocked Lg within the south Caspian Sea basin and along the ZFTB. They also observed blocked Sn in Turkey and most of Iran. In the southern Caspian basin and Arabian plate, they observed efficient Sn.
- (2) Gok et al. (2003, 2011) improved the efficiency tomographic map of Sn in the Turkish-Iranian plateau and observed efficient Sn within southern Caspian basin and along the ZFTB. Blocked Sn is observed throughout the Anatolian plateau.

(3) Zor et al. (2007) developed the first tomographic model for Lg Q in the Turkish plateau and they observed very low Q (<100) values in the East Anatolian Plateau and it is probably due to the widespread Quaternary volcanism.

(4) Bao et al. (2011) measured the Pg Q over the Turkish plateau and adjacent regions. They observed low Q values within Turkish plateau and high Q values within Arabian plate. The low Q values in the Turkish plateau are consistent with crustal partial melting and suggesting the intrinsic attenuation is probably the primary reason.

(5) Pasyanos et al. (2009) presented a broad-band tomographic model of Lg attenuation in the Middle East using single-station absolute amplitude measurements. They observed high attenuation over most of the Turkish-Iranian plateau ($Q < 200$) and along the Zagros Mts. The lowest Q is found in eastern turkey ($Q < 150$). They also indicated the frequency dependent of attenuation is not compatible with power-law assumption.

(6) Kaviani et al. (2015) measured the Lg Q across the Turkish-Iranian Plateau using exactly the same data set and methods I have used for Sn Q measurements. They observed very low Q (<150) values beneath Turkish plateau and relative higher Q values (150-400) over Iranian plateau. They also obtained Lg group velocity model and the Lg Q measurements strongly correlated with the measurements of Lg group velocity.

In this study, I have analyzed a large seismic waveform database in the northern Middle East and obtained both Sn efficiency tomography and Q tomography models. My observations can greatly improve the knowledge of Sn attenuation across the northern Middle East.

Figure 4.1 Simplified tectonic map in the Middle East. AP: Anatolian Plateau; BS: Bitlis suture; NAF: Northern Anatolian Fault; EAF: Eastern Anatolian Fault; ZFTB: Zagros fold and thrust belt; MZT: Main Zagros thrust; SSZ: Sanandaj-Sirjan Zone; UDMA: Urumieh-Dokhtar Magmatic Arc. The red triangles represent quaternary volcanoes. The black lines are the major active fault zones.

4.2 Data Collection and Processing

The data analyzed in this study are collected from 568 permanent and temporary broad-band and short-period stations over the Turkish-Iranian plateau (Figure 4.2). 1267 regional events are used in my study, all with magnitudes greater than 4.5 and hypocentral depths less than the Moho. The epicentral distances are limited between 3° to 15° . 15256 seismograms (Figure 4.3) are picked with good signal noise ratio and clear Pn arrivals. It is important to note that I used pre-event Pn signal to noise ratios rather than Sn signal to noise because of the issue of phase blockage which is discussed in chapter 5.

I have applied a band pass filter (0.1Hz-0.5HZ) to maximize signal to noise and characterize the efficiencies of Sn phases (Figure 4.4). Then I used the method of Sandvol et al. (2001) and Al-Damegh et al. (2004) to tomographic map the Sn efficiencies (Figure 4.4a). In order to include more data, I have added the efficiency database from Al-Damegh et al., 2004 to my model. I also did checkerboard test (Figure 4.4b) to assess the resolution in the Sn efficiency tomography.

I used TSM and RTM to measure Sn Q at 0.5Hz (0.1Hz-1.0Hz), 1Hz (0.5Hz-1.5Hz) and 2Hz (1.5Hz-3.0Hz) (Figure 4.8). Then I applied the LSQR algorithm to tomographically map the lateral variations of Q in the northern Middle East. I also performed checkerboard test (Figure 4.6) and bootstrap test (Figure 4.7) to examine the resolution and stability of the Sn Q tomographic models.

Figure 4.2 Map showing the seismic stations used in this study. The black triangles represent seismic stations.

Figure 4.3 A map of ray paths with Sn propagation efficiencies

4.3 Results

4.3.1 Sn Efficiency Tomography

My Sn efficiency tomography model is shown in figure 4.4a. The most prominent feature is that Sn is blocked or inefficient within nearly all of the Turkish-Iranian plateau. This observation is consistent with previous Sn attenuation studies (Sandvol et al., 2001; Gok et al., 2003; Al-Damegh et al., 2004; Gok et al., 2011). It also correlates well with low body wave (Figure 4.12) and surface wave velocity anomalies (Hafkensheid et al., 2006; Kaviani et al., 2007; Simmons et al., 2011; Maggi and Priestley, 2005; Gok et al., 2007; Gok et al., 2011). In particular, the eastern Anatolian plateau and Lesser Caucasus are highest attenuation regions in my model with very low Lg Q values (<100) (Zor et al., 2007) and slow Pn velocities ($\sim 7.7\text{km/s}$) (Figure 4.11) (Hearn and Ni, 1994; Amini et al.,

2012) are observed. In addition, this area coincides with extensive quaternary basaltic volcanism. Another significant feature is that I observe blocked or inefficient Sn within the Iranian plateau but the Pn velocity in the Iranian plateau is not particularly low (7.9-8.1km/s) (Hearn and Ni, 1994; Amini et al., 2012). I have extended my efficiency tomography model well to the south of the Al-Damegh et al., 2004 and Gok et al., 2011 models. I have found strong evidence of Sn blockage across the Makran subduction zone which would be consistent with retreating slab and associated upper mantle upwelling. I also observe Sn blockage all along the Dead Sea Fault system, again consistent with the presence of Neogene volcanism.

Surprisingly I do observe inefficiently Sn propagation (white colors) in central Anatolia and the central Iranian plateau. This suggests there might be some stable or thin lithosphere over relatively short distances in these regions. In addition these regions are surrounded by Sn blockage zones make the observations of efficient Sn as very unlikely or difficult. This results is also somewhat different from prior results but could be due to that the much improved data coverage in my model. My estimates of Sn Q could also help us determine whether this is the case or not.

I observe efficient Sn across Black Sea, Mediterranean Sea and Persian Gulf. These observations are consistent with the well established observation that Sn propagates very efficiently in the oceanic lithosphere. In the Arabian plate, efficient Sn is observed which is consistent with the fact that Sn can propagate efficiently across intact and stable continental lithosphere (Oliver and Molnar, 1969). In the southern Caspian Sea

basin, I also observe efficient Sn. This feature is consistent with fast Pn velocity (Hearn and Ni, 1994; Amini et al., 2012) and high surface wave velocity anomalies (Maggi and Priestley, 2005). Previous studies have proposed the southern Caspian basin is underlain by a rigid oceanic type lithosphere (Allen et al., 2003; Gok et al., 2011; Kaviani et al., 2015).

In the ZFTB, I observe efficient Sn. Further NE, inefficient or blocked Sn is observed beneath the SSZ and UDMA where widespread Eocene volcanism was found (Agard et al., 2011). It is important to note, however, the ray coverage map (Figure 4.3) shows most of rays are confined in the same direction in the Zagros.

I performed checkerboard test (Figure 4.4b) to assess the size of anomalies I can resolve in my model by using the same ray geometries that are used for the inversion of actual data. The checkerboard test shows my model can well resolve anomalies with sizes of $4^{\circ} \times 4^{\circ}$ cells covering the Turkish-Iranian Plateau and it is thus good enough to image the large blockage regions. Anomaly smearing is observed at the edges of my model where ray path coverage is relatively poor and/or the ray azimuthal coverage is relatively poor (for example at Zagros). Model smearing is also observed (for example at Arabian platform) because of poor ray coverage.

Figure 4.4 (a) Map showing Sn efficiency tomography results. (b) Resolution test using a synthetic checkerboard for extinction path length of 300km.

4.3.2 Sn Q Tomography

The TSM Sn Q_0 map is shown in figure 4.5a. Across most of the Anatolian Plateau, I observe low Sn Q values (100-300). In the East Anatolian Plateau, the Sn Q values can be as low as 100. I also observe very low Sn Q values (~100) in the Lesser Caucasus. These two areas correlate well with Neogene volcanism. The Sn Q values in the Iranian plateau are ~250 while normal Pn velocity (Hearn and Ni, 1994; Amini et al., 2012) is observed here. In the Zagros, I observe high Sn Q values (~450). I also observe high Sn Q values in the southern Caspian basin.

The RTM Sn Q_0 map is shown in figure 4.5b. As discussed in chapter 2, the RTM Q measurements are more accurate because it is not affected by inaccurate instrument responses or relative differences in site terms that may bias the TSM Q measurements; however it dramatically reduce the ray coverage due to more strict recording geometry. As a result, I use RTM measurements to verify the accuracy of the TSM measurements. I observe very similar attenuation structure in the RTM Sn Q_0 map as compared with TSM Sn Q model. In the western Anatolian plateau, I observe relative low Q values (300). In the central Turkey, I also observe low Q values (200). In the East Anatolian Plateau and Lesser Caucasus, I observe very low Q values (~100). I also observe low Q values (~250) in the central Iran. In the Zagros Mountains, I observe high Q values (>400). In other regions at the edges of the RTM Sn Q_0 model (such as the south Caspian Sea basin), my results are not necessarily very reliable because of poor ray coverage.

I also performed checkerboard tests to examine the resolution of the Sn Q models the same as I did for the efficiency tomography models. The checkerboard test illustrates that Sn Q₀ anomalies with a size of 2°×2° can be well resolved by the TSM ray coverage. Figure 4.6 shows the TSM Sn Q₀ model which is well resolved in most of my study area. The checkerboard anomalies are distorted in regions lacking dense ray coverage such as Arabian Plate and the Black Sea. The Sn Q₀ anomalies appear to be somewhat poorly resolved along the Zagros where most ray paths are parallel to the strike of the belt. Figure 4.7 shows the RTM Sn Q₀ map is not well resolved except in the central Iranian plateau and central Turkey because of poor ray coverage.

I performed a bootstrap resampling technique to test the stability of my tomographic models (Figure 4.8). In this test, I randomly resample my data until I reach the same number of observations as in my original data and then I invert the resampled data for a bootstrap model. This will result in some observations being omitted and other duplicated in my bootstrapped data sets. I inverted each of these bootstrapped data sets thus obtaining 100 bootstrap solutions. I use these solutions to estimate the variance of the model parameters. The result of the bootstrap test for the TSM Sn Q₀ model is shown in Fig figure 4.8a. It shows an uncertainty less than 50 within the Turkish-Iranian plateau as well as the southern Caspian basin and high uncertainty in the south of Iran. The result of the bootstrap test for the RTM Sn Q₀ model (Figure 4.8b) show low uncertainty within most of the Turkish- Iranian plateau.

In order to determine the frequency dependence of S_n Q, I also construct S_n Q models at different frequencies (0.5HZ, 1HZ, 2HZ) both using TSM and RTM (Figure 4.9). The main features are consistent with the S_n Q_0 models; however, I also observe some significant differences at different frequencies. In general I see that Q generally increases with increasing frequency. This observation is consistent with the power law assumption which is described in chapter 2; however, I found Q values in the eastern Anatolian plateau is not much change with different frequency indicating that the frequency dependent factor is near zero in this region. On the other hand, I also observe some significant differences at different frequencies. At low frequency (0.5HZ), I observe relative low attenuation within central Turkey. At high frequency (2HZ), I observe high attenuation beneath central Turkey. This feature can be explained by low frequency S_n can dive deeper in the uppermost mantle while high frequency S_n travels more like head wave beneath the Moho. There are discrepancies at some regions (such as southern Caspian Sea basin) showing strong frequency dependence and this is possibly caused by scattering. Another possible explanation for the inconsistency of frequency dependence of attenuation is that the power law assumption used in the calculation of Q values can be incompatible in the Middle East (Pasyanos et al., 2009).

The frequency dependent factor (η) map is shown in figure 4.10. Across most of my model, the frequency dependent factor is low to normal (0-0.5) indicating the intrinsic attenuation is the dominant mechanism. In the eastern Turkey, I observe very low η values corresponding with very low Q values. In the Zagros, I observe high η values (~0.9) indicating that scattering attenuation is the dominant mechanism for S_n attenuation.

Furthermore a Moho step beneath the Zagros can also potentially disrupt the Sn wavefield leading the scattering attenuation of Sn (Hatzfeld et al., 2003; Paul et al., 2006; Paul et al., 2010).

Figure 4.5 (a) Two Station (TSM) Sn Q_0 tomographic map. (b) Reverse Two Station (RTM) Sn Q_0 tomographic map. Black lines are major tectonic boundaries.

Figure 4.6 Checkerboard test for the Two Station (TSM) $\text{Sn } Q_0$ tomographic map, black lines are major tectonic boundaries. (a) Original checkerboard model. (b) Inverted checkerboard model using the TSM ray paths.

Figure 4.7 Checkerboard test for the Reverse Two Station (RTM) $S_n Q_0$ tomographic map, black lines are major tectonic boundaries. (a) Original checkerboard model. (b) Inverted checkerboard model using the RTM ray paths.

Figure 4.8 Map showing the standard deviations computed by bootstrap analysis for the TSM (a) and RTM (b) Sn Qo tomographic models. Black lines are major tectonic boundaries.

Figure 4.9 Two Station (TSM) Sn tomographic maps at 0.5Hz (a), 1Hz (c), 2Hz (e). Reverse Two Station (RTM) Sn tomographic map at 0.5Hz (b), 1Hz (d), 2Hz (f). Black lines are major tectonic boundaries.

Figure 4.10 Tomographic map of frequency-dependent (η) of (a) Two Station Method Sn Q measurements and (b) Two Station Method Sn Q measurements. Black lines are major tectonic boundaries.

Figure 4.11 Pn wave velocities. The black lines show major tectonic features of Middle East. (Amini et al., 2012)

Figure 4.12 P wave velocity anomalies. (Simmons et al., 2011)

4.3.3 A Comparison of Efficiency and Q Tomographys

Comparing the results of Sn efficient tomography and Sn Q tomography, the Sn efficient tomography can resolve more features than the Sn Q tomography because the Sn

Q tomography requires a more strict ray path geometry. The main features between these models are very similar. In the Turkish-Iranian plateau, I both observe blocked or inefficient Sn and low Q values which indicate a hot and thin lithosphere. I also observe efficient Sn and high Q values in the southern Caspian basin that support the presence of a cold and strong oceanic lithosphere. In the Zagros, I observe strong lateral variations in Q_{sn} . In the ZFTB, I observe efficient Sn and high Q values, further to the NE, the Q values decrease and the propagation of Sn shifts to inefficient and blockage. Another important feature is that I observe blocked Sn and low Q values in the Lesser Caucasus. The presence of quaternary volcanism indicates partial melting may cause the high attenuation in this region. I also observe discrepancies between these models. The Sn Q tomography shows some lateral variations within the Iranian plateau while I do not see this feature in the Sn efficiency tomography. This may be caused by the poor ray coverage in the Sn Q tomography.

4.4 Discussion

Temperature and composition are two major causes that can generate seismic anomalies. Positive temperature anomalies will lead to a reduction in both attenuation and velocity; however, compositional anomalies should not necessarily produce a strong correlation in attenuation and velocity (Kaviani et al., 2015). As a result, by combining velocity and attenuation structure we can distinguish between compositional and temperature anomalies.

The Sn attenuation model presented in my study is consistent with previous Sn attenuation studies (Sandvol et al., 2001; Gok et al., 2003; Gok et al., 2011) but with higher resolution and more reliable estimates. In the Anatolian plateau, I observe blocked/inefficient Sn and relative low Q values. Previous studies have found slow Pn velocity (Hearn and Ni, 1994; Amini et al., 2012) and low shear wave velocity anomalies (Kaviani et al., 2007; Maggi and Priestley, 2005; Gok et al., 2007; Gok et al., 2011) within the Anatolian plateau. The correlation between attenuation and velocity observations indicate a hot and thin lithosphere beneath the Anatolian plateau. One striking observation is the high attenuation and very low Q values in the eastern Anatolian plateau and Lesser Caucasus. The low frequency dependent factor (η) indicates that intrinsic attenuation is likely the dominant mechanism in the two regions. The presence of widespread quaternary volcanism suggests partial melting in the uppermost mantle is the main cause of high attenuation in the eastern Anatolian plateau and Lesser Caucasus.

I also observe lateral variation of Sn Q within the Anatolian plateau (Figure 4.5). The Sn Q values decreases from the western Anatolian plateau (~300) to the eastern Anatolian plateau (<100) and the Sn Q values in the central Anatolian plateau is ~200. Typically, low Q values and low velocity anomalies can be explained by the presence of partial melting. The good correlation between low velocity anomalies and low Q values in the eastern Anatolian plateau is confirmed by the widespread quaternary volcanism. Due to the lacking of high resolution velocity anomalies in the western and central Anatolian plateau, it is difficult to infer whether partial melting exists in these regions.

Kaviani et al. (2015) points out that intrinsic attenuation in partial molten rocks seems to be weakly frequency dependent and hence in regions where I observe low Q and low frequency dependence, I may infer the presence of partial melting with strong intrinsic attenuation. In figure 4.10, I observe low frequency dependent factor in the central Anatolian plateau and relative high frequency dependent factor in the western Anatolian plateau indicating the presence of partial melting in the central Anatolian plateau and a lack of partial melting in the western Anatolian plateau.

In the Iranian plateau, I observe blocked/inefficient S_n and relative low Q values. But the P_n velocity (Figure 4.11) is normal (Hearn and Ni, 1994; Amini et al., 2012). This observation indicates a lack of partial melting beneath the Iranian plateau. The hot lithosphere can be explained by localized mantle upwelling caused by the detachment of the subducted Neo-Tethys slab during the late Miocene (Al-Damegh et al., 2004).

In the ZFTB, I observe efficient S_n and high Q values, further to the NE, the Q values decrease and the propagation of S_n shifts to inefficient and blockage in the SSZ and UDMA. The high frequency dependent factor (η) indicates that scattering attenuation is possibly the dominant mechanism. Receiver function studies (Hatzfeld et al., 2003; Paul et al., 2006; Paul et al., 2010) have found Moho step beneath the Zagros and proposed that the crust of Zagros is underthrusting the crust of Central Iran along MZT. Simmons et al. (2011) confirmed this hypothesis by using P wave tomography (Figure 4.12). This type of strong lateral heterogeneity can cause scattering of the S_n wavefield.

My observation supports this underthrusting model, however, due to the poor ray coverage in my Q model, more work is required in the future.

In the southern Caspian basin, I observe efficient Sn and high Q values. Fast Pn velocity (Hearn and Ni, 1994; Amini et al., 2012) and high surface wave velocity anomalies (Maggi and Priestley, 2005) are also observed. The correlation between attenuation and velocity structure implies a cold and thick mantle lithosphere. Other studies have proposed the southern Caspian basin is underlain by a rigid oceanic type lithosphere (Allen et al., 2003; Gok et al., 2011; Kaviani et al., 2015).

4.5 Conclusions

I have collected a large Sn waveform data set in the northern Middle East that I have quality controlled using both automated and manual approaches. Two tomographic techniques have been used to determine the attenuation structure of the uppermost mantle. My primary conclusions are as follows:

- (1) I observe inefficient/blocked Sn and low Q values in the Turkish-Iranian plateau indicating a hot and thin mantle lithosphere.
- (2) Intrinsic attenuation is the dominant uppermost mantle shear wave attenuation mechanism beneath the eastern Anatolian plateau and Lesser Caucasus. Partial melting is the main cause of high attenuation in the two regions.
- (3) The high attenuation in the Iranian plateau is likely not caused by partial melting thus the seismic anomalies in the uppermost mantle are likely compositional.

(4) The scattering attenuation is the dominant mechanism in the Zagros. My observations support the crust of Zagros is underthrusting the crust of Central Iran along MZT.

(5) I observe efficient Sn and high Q values in the southern Caspian sea basin indicating the presence of oceanic lithosphere.

4.6 References

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Chapter 5: Left Censored Data Problem in Seismic Attenuation

5.1 Background

Data censoring is a common problem in statistics. Censorship occurs when data (a measurement or observation) are missing or incomplete (Oliveira, 2005). This arises either because of clipping, where all amplitudes can be determined only to exceed a given lower bound (right censored case), or because the signals are weaker than the ambient noise level and hence are not detected (left censored case) (Jih and Shumway, 1989). Left censored data problem occurs most commonly in seismology when trying to measure seismic attenuation because it is common for the observed amplitudes to be less than the background noise level. In my study, I used a signal to noise criteria to select only high quality seismic amplitude data. Amplitude ratio of Sn phase to noise below the detection threshold will be discarded which is an approach that is used in nearly every study measuring seismic attenuation. More specifically, in my Two Station Method (TSM), the Two-Station paths with blocked Sn (Sn efficiency equals to 0) for the near station will be discarded. Pasyanos et al. (2009) pointed out that by discarding non-blocked seismic data with high attenuation that have low signal to noise, the remaining data will be biased towards higher Q values or lower attenuation than what is actually occurring in the real earth. On the other hand, if I keep these noisy data, it could produce greater problems such as including amplitudes of noise and coda that could also bias my estimates of the true attenuation structure of the earth. Only tentative attempts have been made to solve this problem. Taylor et al. (2003) applied maximum likelihood data augmentation to this left censored data problem (Lg attenuation tomography) in Tibet. The maximum

likelihood data augmentation can be used to predict the missing data based on the relationship of amplitudes with other related completely observed variables (Dempster et al., 1977; Schafer, 1997; Anderson et al., 2010). They showed lower Q values and smoother variations in the resulting model using data augmentation. However, they did not consider the spatial-dependence in their method, in other words, the propagation of seismic phases can be different region by region because of the lateral heterogeneity. In fact I can be certain that this type of data censorship will be highly spatially dependent. As a result, the data censoring is more complicated in seismology. In this chapter, I present a method that substitutes half of the Level of Detection (LOD/2 technique) to deal with the left censored data problem (Sn Q tomography) and apply this technique in the Tibet and Middle East.

5.2 Simulation Test of Left Censored Seismic Amplitudes

In order to investigate the impact of seismic amplitude censorship I have chosen to create a data simulation where I know what the true attenuation structure is since I have defined it in my model. I begin by creating a synthetic seismic amplitude data set. I have chosen to focus my attenuation on the Iranian plateau because of the extensive Sn phase blockage that I observe in this region. I also have a large data set for this region that I can use for my simulation. Using this simulation and the Level of Detection Divided by Two (LOD/2) technique to augment the blocked seismic amplitudes. Finally, I apply this technique to the real data in the Tibet and Middle East in order to see how both models change.

5.2.1 Data Censoring in Seismic Q Tomography

In order to show the effect of data censorship in seismic attenuation tomography, I first construct an input Sn Q model shown in figure 5.1. The model is separated in four regions with two high Q value (500) and two low Q value (100) blocks. Then I use the ray coverage of Iranian seismic amplitude data set (Figure 5.2) to calculate the simulated amplitude reduction ratio for each path. Finally, I calculate Q values for all paths and tomographically map the results. I simulate data censorship by removing all simulated amplitudes that fall below a pre-defined value.

The amplitude reduction ratio (A^{rec}) can be shown in the equation below

$$A^{rec} = \frac{A}{A_0} \quad (5.1)$$

Where A_0 denotes the amplitude at the source and A denotes the amplitude at the station.

Substituting equation 5.1 with equation 2.2, neglecting the geometrical spreading term and instrument response I obtain

$$A^{rec} = e^{\frac{-\pi f d}{Qv}} \quad (5.2)$$

where f is the frequency (1HZ), d is the raypath length, Q is the quality factor and v is the Sn velocity (4.5km/s). Taking the natural logarithm of equation 5.2, discretizing the attenuation coefficient, I obtain

$$\ln(A^{rec}) = \frac{-\pi f}{v} \sum_k \frac{d_k}{Q_k} \quad (5.3)$$

where k denotes the index of cells that each ray travels in the model.

When I obtain the amplitude reduction ratios of all paths, I add 15% noise to these ratios and then calculate Q value for each path using equation below

$$Q = \frac{-\pi f d}{\ln(A^{rec})v} \quad (5.4) \quad Q_{blocked} = \frac{-\pi f d}{\ln(\frac{LOD}{2})v}$$

In order to show the effect of data censoring, I calculate the Q values in two groups. In the first simulated data set, I discard all blocked paths and the resulting model is a censored model. In the second simulated data set, I keep all of the paths. Then I tomographically map these Q values to obtain the resulting attenuation models.

Figure 5.1 Map showing input $S_n Q_0$ model.

Figure 5.2 Map showing ray coverage in Iran. The green lines outline the shape of the input model.

The censored model is shown in figure 5.3a. The most prominent feature is that Sn Q values (100-300) in the northeastern part of the model is higher than the Sn Q values (100) in the input model. In figure 5.3b, the other model using all paths shows similar pattern as compared with the input model. In summary, discarding blocked Sn paths will cause left censored data problem and the resulting model will be biased to high Q values.

Figure 5.3 (a) Censored model without blocked Sn paths. (b) Model with all paths.

5.2.2 LOD/2 Technique

I apply the Level of Detection Divided by Two (LOD/2) technique to deal with blocked Sn paths. In the synthetic test, I define the LOD (limit of detection) as the mean value of signal to noise ratios from all data. Then I set LOD/2 as the amplitude reduction ratio for all blocked paths and calculate Q values using equation 5.4. The tomographic map by using LOD/2 technique is shown in figure 5.4. I observe very similar pattern as compared with the input model. I also observe lower Q values (100) in the in the northeastern part of the model than the Sn Q values (100-300) in the censored model. It shows this technique recovers the model much better than the censored data set shown in figure 5.3a. It is worth noting that the LOD/2 is obviously an ad-hoc approach, however, it almost always seems to do better than the censored data sets which are almost always used in seismic attenuation studies.

In order to test the usefulness of this technique, I also calculate the residuals of amplitude reduction ratios. I repeat the process to obtain the amplitude reduction ratios in section 5.2.1, in which the input model is shown in figure 5.4. The results are shown in figure 5.5a. Most of the residuals of amplitude reduction ratio are small. I also obtain the tomographic map of residuals for Q values in figure 5.5b. I observe small Q residuals from -50 to 50 across most of the model. I also observe large Q residuals (less than -100) for paths crossing some regions (northwestern part of the model). It indicates the LOD/2 technique may introduce error in the resulting Q model by introducing noise.

For real data, I divide the amplitude of blocked seismograms by two rather than using a constant value, if the blockage is spatially dependent, which it is, the LOD/2 technique will also correct the model spatially as well.

In summary, the LOD/2 technique recovers the synthetic model very well and it can resolve left data censored problem caused by discarding blocked paths in the seismic attenuation study.

Figure 5.4 Model by using LOD/2 technique.

Figure 5.5 (a) The distribution of amplitude reduction residual by using LOD/2 technique. (b) The distribution of Q residual by using LOD/2 technique.

5.3 Application to the Eastern Tibetan Plateau and Northern Middle East

I apply the LOD/2 technique to my S_n amplitude data sets in Tibet and Middle East. Instead of removing blocked paths I use for only the far station in order to form Two Station Paths with blocked paths. As mentioned above, I divide the amplitude of blocked seismograms by two in order to avoid the effect of spatial dependence. Then I calculate Q values using Two Station Method (TSM) as described in chapter 2 even for two station paths with blocked paths. The results are shown in figure 5.6 and 5.7. The most prominent feature is that I obtain lower Q values and smoother variations in many places of the resulting models comparing with censored models. It is important to note that these models still suffer from some data censorship but I think with the LOD/2 I have reduced this effect.

In Tibet, the overall pattern is consistent with my results as described in chapter3. I observe high Q values in the Qaidam basin and low Q values in the Songpan-Ganzi block and Qiangtang block. The HQZ (high Q value zone) along Bangong-Nujiang suture shifts a little bit to the east. A LQZ (low Q value zone) is observed in the Lhasa block.

Figure 5.6 (a) Censored model without blocked Sn paths. (b) Model with all paths by using LOD/2 technique.

In the Middle East, the overall pattern is consistent with my results as described in chapter 4. I observe low Q values in the Turkish-Iranian Plateau. In particular, I also observe very low Q values in the East Anatolian Plateau and Lesser Caucasus. In the Zagros, I observe high Q values. I also observe high Q values in the southern Caspian basin.

Figure 5.7 (a) Censored model without blocked Sn paths. (b) Model with all paths by using LOD/2 technique.

5.4 Conclusions

Data censorship is a difficult problem in the seismic attenuation studies. Traditional processing is to discard blocked data and the remaining data will generate Q models that will be biased towards less attenuation. I use a LOD/2 technique to deal with the blocked data. In the synthetic test, this technique can recover the synthetic model with small residuals. In addition, I apply the LOD/2 technique to real data of Tibet and Middle East; I obtain lower Q values and smoother variations in the resulting models comparing with censored models. However, the LOD/2 technique may also introduce errors by including amplitude of noise. As a result, further study is needed in the future.

5.5 References

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VITA

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